MAGYAR ÁLLAMI
EÖTVÖS LORÁND
GEOFIZIKAI INTÉZET

GEOFIZIKAI
ÖZLEMÉNYEK

НГЕРСКИЙ
ОФИЗИЧЕСКИЙ
НСТИТУТ
М Л. ЭТВЕША

ЕОФИЗИЧЕСКИЙ
ОЛЛЕТЕНЬ

GEOPHYSICAL
TRANSACTIONS

EÖTVÖS LORÁND GEOPHYSICAL INSTITUTE OF HUNGARY

CONTENTS

VOL. 42. NO. 1–2 NOV. 1999. (ISS N0016-7177)

BUDAPEST

TARTALOMJEGYZÉK

Editor's note

Geophysical Transactions has reached a new turning point in its history. Doubtless our readers will already have realised this from the front cover, where a subtitle now appears, and the Contents also appear there. These are the formal changes, but not only these will be new from now on.

Since it inception, Geophysical Transactions has been edited and published by the Eötvös Loránd Geophysical Institute (ELGI) of Hungary. Now, as one of the spin-off effects of the political, social and economic changes throughout Middle and Eastern Europe, the Institute's leadership thought it would be appropriate to give somewhat greater publicity to ELGI's research work, and to the results achieved both in research and in applications.

With this in mind, it was decided that a national editorial board would better serve this objective. Accordingly, the Editorial Board was reorganized in April 1998. A list of the new members can be seen on the inside front cover. The journal will continue to be published in English, the international language of Geophysics.

Taking it into account that the topics of the journal cover both pure, as well as applied geophysics, we thought that from time to time we should publish issues covering particular topics. An example of this approach is the DANREG issue, (Vol. 41. No. 3–4, 1997) which dealt with the Geophysics of the Danube Region.

The present issue, which you are now holding, is devoted to research on gravity. In the next issues, results in geomagnetics and shallow geophysical investigations will be published.

Obviously the principal concern of a journal is its readership. The changes already carried out as well as our future publishing plans reflect our attempts to present geophysics results in an easily digestible form. We do hope that our approach meets with our readers' approaval and that they will continue to gain benefit and pleasure from Geophysical Transactions.

Zsuzsanna Hegybíró
Editor in Chief

GEOPHYSICAL TRANSACTIONS 1999
Vol. 42. No. 1 – 2. pp. 5 – 27

Rock densities in the Pannonian basin — Hungary

Zoltán SZABÓ* and Zoltán PÁNCSICS*

In that density is a fundamental parameter of gravity exploration, and that Bouguer- and terrain corrections and interpretation of gravity measurements require reliable density data, the authors carried out a detailed investigation to determine the densities of all possible important geological formations in the Pannonian basin. The study was based on the laboratory measurement of 12 000 rock samples and on 145 000 linear metres of density logs. The results themselves are presented in this paper in histograms, figures and tables.

Keywords: gravity surveys, density, Pannonian basin, density logging, Hungary

1. Introduction

Up till now only sporadic density information has been published for the Hungarian part of the Pannonian basin [PINTÉR–SZABÓ 1964, SZÉNÁS 1965, KILÉNYI 1968, KOVÁCSVÖLGYI 1996]. More density information is available for the Slovakian part of the basin [ŠEFARA 1987]. Despite its incompleteness, this data set was acceptable for regional studies but nowadays — in the post-regional era — when detailed large-scale investigations are on the agenda, more accurate density data are needed for geophysical interpretations.

* Eötvös Loránd Geophysical Institute of Hungary, H-1145 Budapest, Kolumbusz u. 17–23

Manuscript received: 20 May, 1997

Rock densities can be determined directly by laboratory measurements or from gamma-gamma density logs. They can also be determined indirectly from the gravity measurements themselves (Nettleton method).

Here, we present density data obtained from laboratory measurements and gamma-gamma density logs.

2. Laboratory measurements

Density measurements on core samples can give good results especially in consolidated rocks, but they are normally available for very limited segments of the total geological column — especially in young sediments. Since the cores tend to be from the harder and more resistive part of the column, results obtained from core samples are generally higher than the actual density of the measured rock material, this is particularly true for shallow depths.

Between 1967 and 70, in ELGI's Tihany Observatory laboratory density measurements were carried out on more than 12 000 rock samples — mainly on drill cores. The samples originated from 305 localities (*Fig. 1*). The project was abruptly terminated, so the results were not evaluated and published.

The laboratory measurements were carried out by means of the buoyancy method. The samples were dried and weighed in air, then coated with paraffin and weighed again. In the next step the paraffin covered samples were immersed in water and weighed again. In water there is a loss of weight which is equal to the weight of the water displaced by the sample. The first measurements give the weight of the sample, but the second ones give a correction factor for the paraffin coating; the third ones give the volume of the sample. Knowing the density of the water and the paraffin, one can calculate the density of the sample. The average standard error of the density determination was ± 20 kg/m^3.

The laboratory measurements have a high precision but because of the lower pressure prevailing in the laboratory and the dry condition of the samples (loss of pore water) the density values obtained from the cores are lower than the original 'in situ' values. To compensate for the pressure dif-

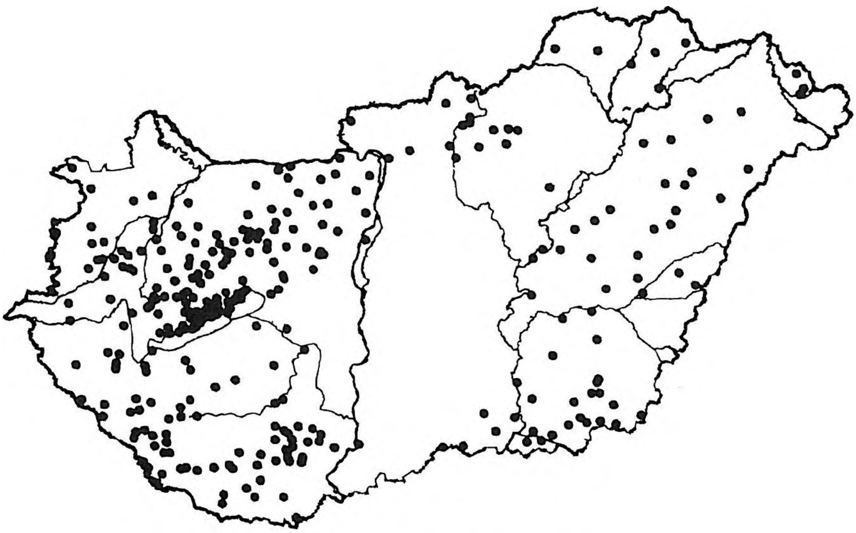


Fig. 1. Location map of rock samples
1. ábra. A mintavételi helyek eloszlása

ference, a high-pressure laboratory would be needed, but for the loss of pore water the following correction factor was applied

$$\sigma = \sigma_0 + \frac{2670 - \sigma_0}{2670}$$

where σ_0 is the air-dried density, and 2670 kg/m^3 is the specific gravity of quartz. The correction factor depends on the porosity of the samples, the higher the porosity the bigger the correction. Quartz was taken as standard because that is the main component of the Pannonian sediments.

The density values obtained were stored randomly on typed sheets in the sequence of measurements and were not analysed. The terminology (rock types, geological ages) used for the samples was far from uniform because the samples were collected by many different companies and geologists.

In order to process the data, they should first be standardized, and they then have to be converted to computer readable form. The data bank established contains all information about the samples, i.e. locality, designations of the borehole, depth, rock type, age, density.

After the completion of the preparatory phase the data were statistically analysed. The results are presented in histograms (*Figs. 2–6*). The

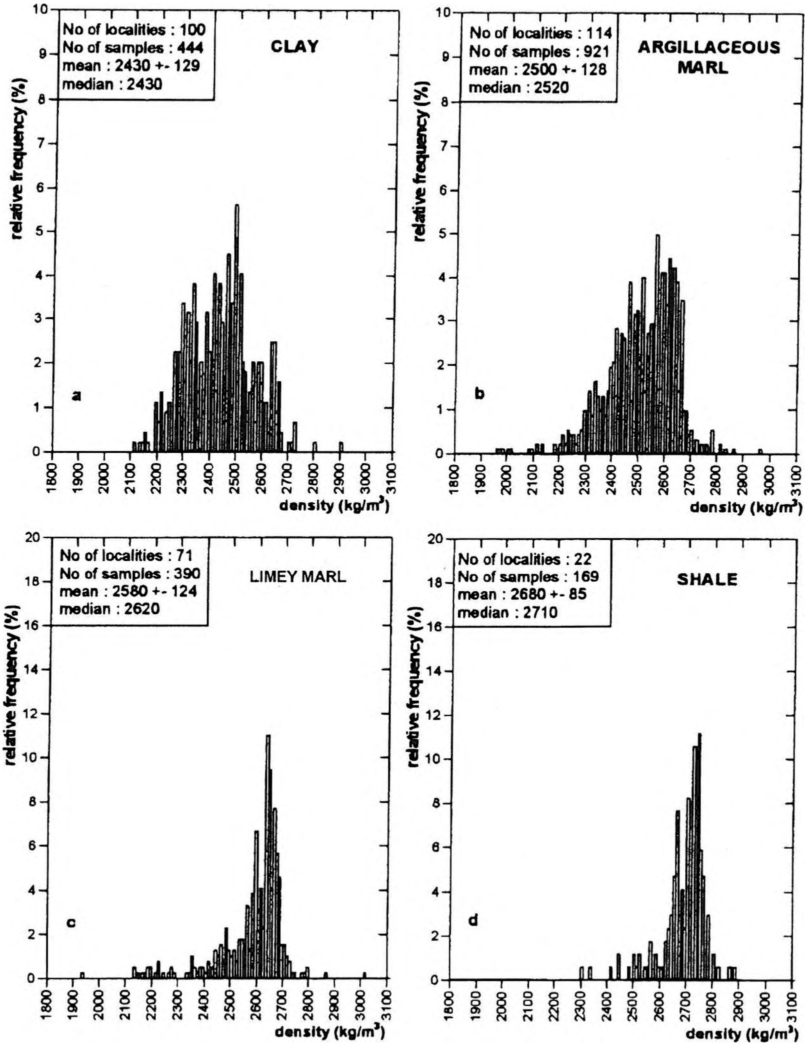


Fig. 2. Density histograms

2. ábra. Agyagok, agyagmárgák, mészmárgák és agyagpalák sűrűség hisztogramja

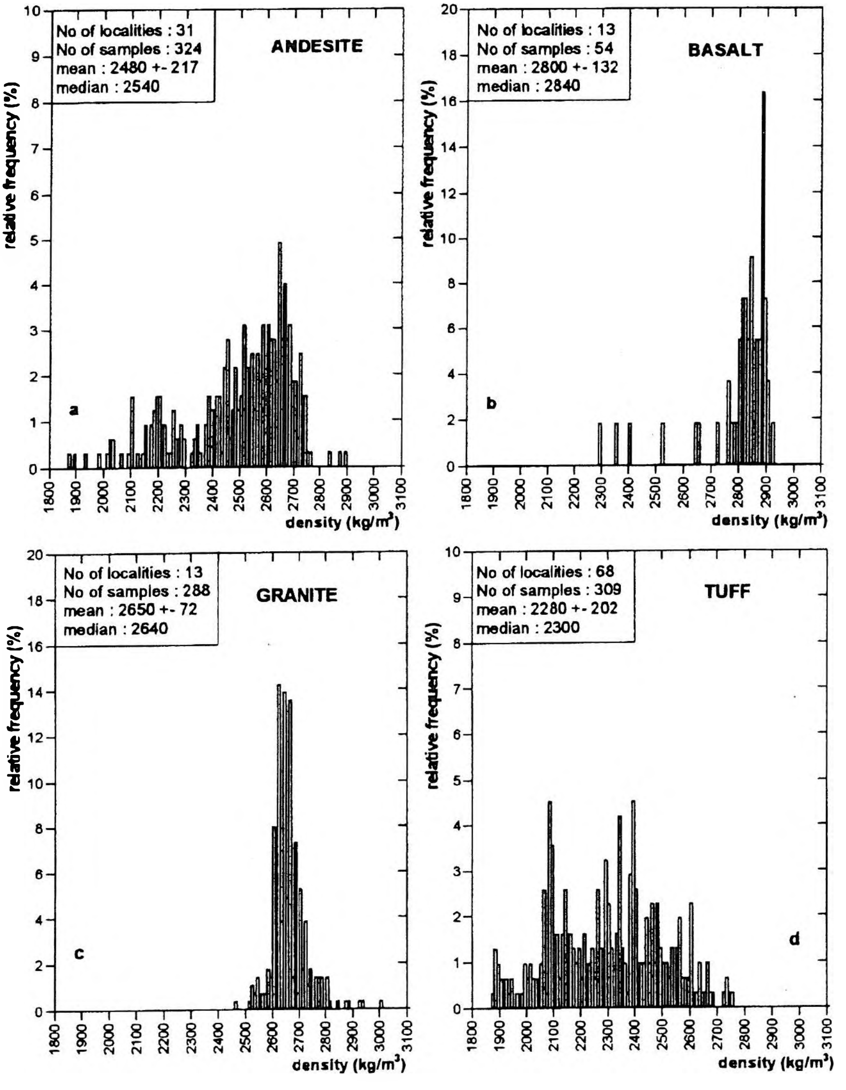


Fig. 3. Density histograms

3. ábra. Andezitek, bazaltok, gránitok és tufák sűrűség hisztogramja

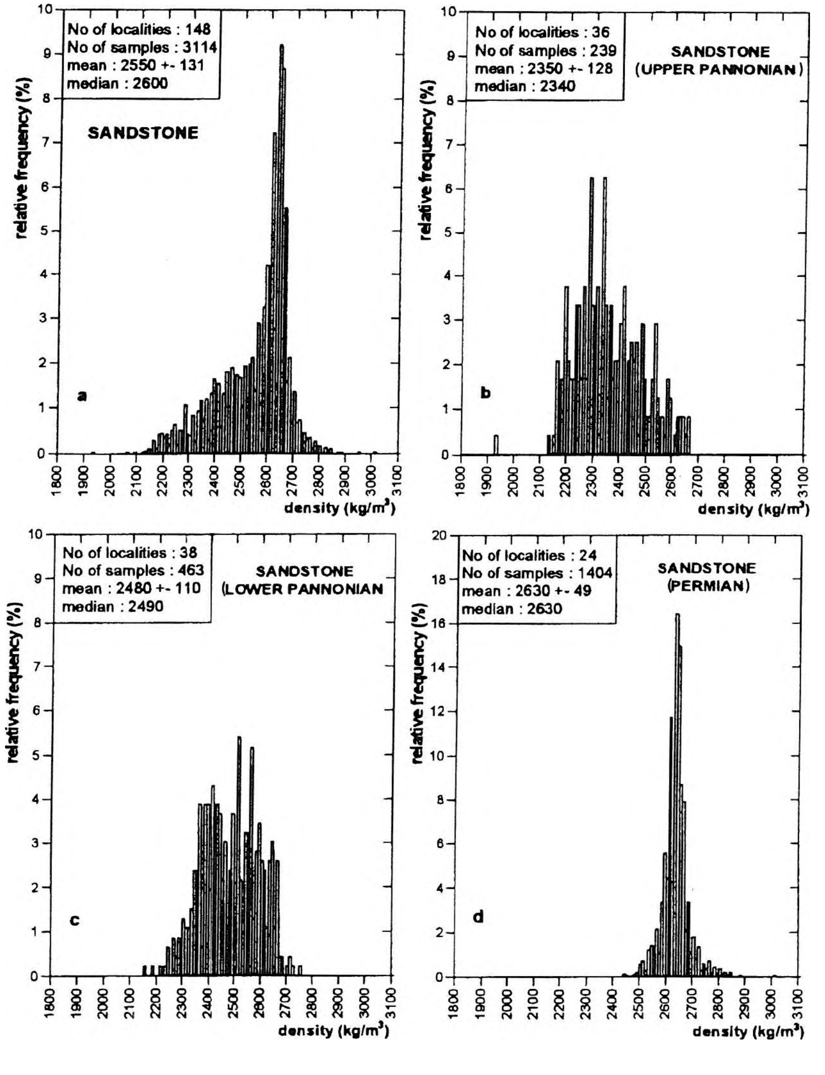


Fig. 4. Combined density histogram of sandstones and histograms of those classified according to age: Upper Pannonian, Lower Pannonian, Permian

4. ábra. Homokkövek egyesített, valamint koronként (felsőpannon, alsópannon és perm) osztályozott hisztogramjai

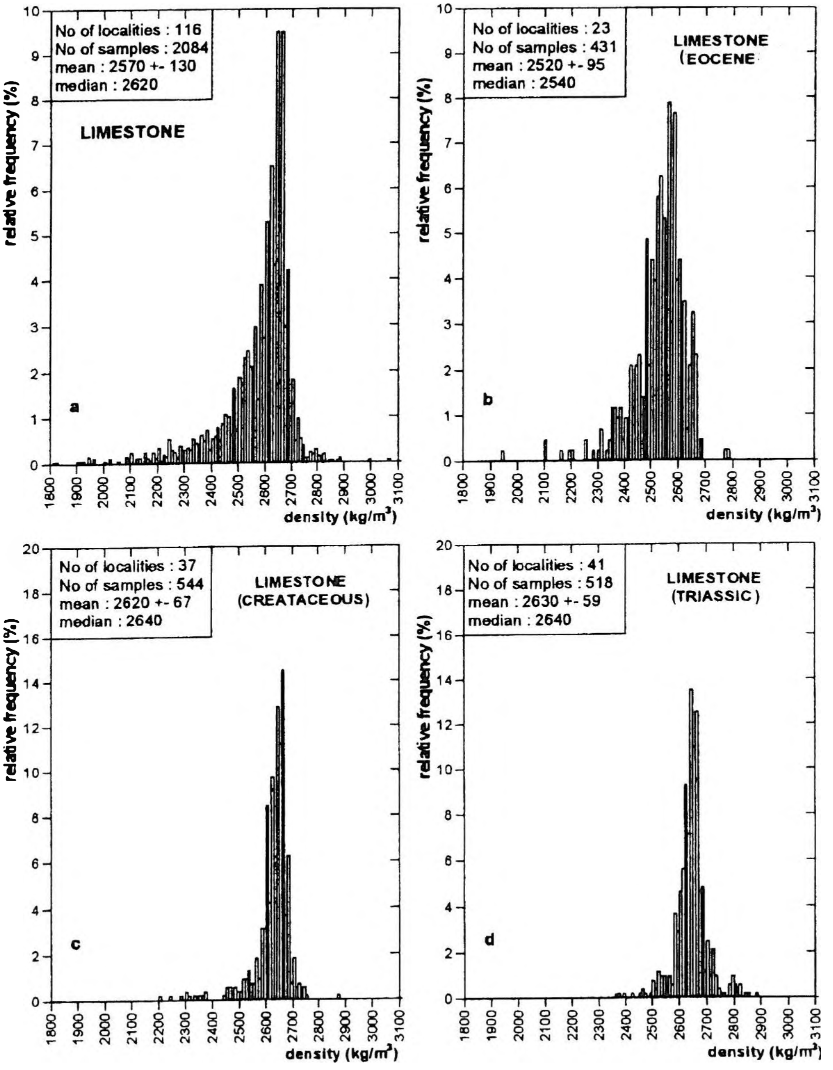


Fig. 5. Combined density histogram of limestones and histograms of those classified according to age: Eocene, Cretaceous, Triassic

5. ábra. Mészkövek egyesített, valamint koronként (eocén, kréta, triász) osztályozott hisztogramjai

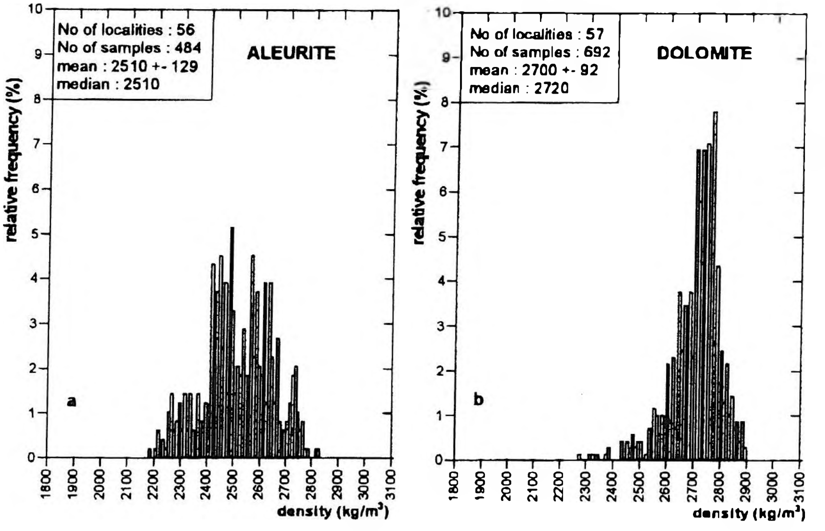


Fig. 6. Density histograms of aleurites and dolomites
6. ábra. Aleuritok és dolomitok sűrűség hisztogramja

fundamental data such as rock type, age (if it is relevant), number of samples, sample mean, sample median, and r.m.s. error are presented in the figures. The histograms are considered to be self-explanatory.

For rock types which have depth-dependent characteristics the exponential density function was determined in the same form as it was given in Athy's classical paper [ATHY 1930], viz.

$$\sigma_h = a + b(1 - e^{ch})$$

where h is the sampling depth in km. It has to be mentioned, however, that the exponential approximation seems to be somewhat artificial in certain cases, still we decided to use the well-known formula.

In some cases (e.g. sandstones and limestones) the histograms deviated from the normal distribution. After separating the samples according to their geological age and representing them in separate histograms the r.m.s. errors became smaller than in the case of the combined histograms. The sandstones could be separated into Upper Pannonian, Lower Pannonian and Permian subgroups (Fig. 4). The limestones could be separated into Eocene, Cretaceous and Triassic groups (Fig. 5). *Tables I* and *II* provide a summary of the densities based on their geological age and rock type.

Geological age	No of localities	No of samples	mean	median
UPPER PANNONIAN	59	926	2380 ± 132	2380
LOWER PANNONIAN	58	1232	2490 ± 128	2530
MIOCENE	89	1295	2430 ± 167	2470
LATE MIOCENE	18	145	2350 ± 184	2390
MIDDLE MIOCENE	40	539	2420 ± 153	2440
OLIGOCENE	23	475	2440 ± 84	2440
EOCENE	39	853	2480 ± 146	2530
LATE EOCENE	11	49	2530 ± 98	2530
MIDDLE EOCENE	17	342	2520 ± 86	2550
EARLY EOCENE	9	102	2550 ± 96	2560
CRETACEOUS	48	1038	2580 ± 101	2610
LATE CRETACEOUS	22	526	2570 ± 110	2610
MIDDLE CRETACEOUS	17	175	2570 ± 74	2600
NEOCOMIAN	13	142	2590 ± 97	2630
JURASSIC	16	285	2580 ± 189	2640
MIDDLE JURASSIC	7	86	2580 ± 137	2640
EARLY JURASSIC	7	117	2590 ± 207	2650
TRIASSIC	69	1639	2650 ± 76	2650
LATE TRIASSIC	37	345	2640 ± 67	2650
MIDDLE TRIASSIC	11	236	2660 ± 69	2655
EARLY TRIASSIC	11	638	2650 ± 59	2650
PALEOZOIC	79	2869	2640 ± 78	2640
PERMIAN	26	1780	2630 ± 69	2630
CARBONIFEROUS	21	426	2640 ± 96	2640
SILURIAN	9	136	2680 ± 72	2720
OLDER THAN LATE CRETACEOUS FORMING THE PRE-AUSTRIAN BASEMENT	154	5127	2640 ± 91	2640

Table I. Summary of laboratory density data according to their geological age
I. táblázat. A laboratóriumi sűrűség adatok földtani kor szerinti megoszlása

The unconsolidated young sediments have a wide bell curve indicating that their density is a function of depth. In these cases we present their density–depth functions (*Figs. 7–9*). Exponential functions were used to approximate their density distribution.

Rock types	No of localities	No of samples	mean	median
CLAY	100	444	2430 ± 129	2430
PANNONIAN	45	164	2400 ± 131	2365
MIOCENE	21	42	2430 ± 163	2440
OLIGOCENE	12	47	2440 ± 70	2460
CRETACEOUS	16	96	2460 ± 74	2460
ARGILLACEOUS MARL	114	921	2500 ± 128	2520
UPPER PANNONIAN	39	184	2410 ± 104	2410
LOWER PANNONIAN	46	371	2530 ± 121	2560
EOCENE	17	50	2480 ± 113	2510
CRETACEOUS	14	75	2530 ± 103	2520
TRIASSIC	8	74	2630 ± 57	2630
SHALE	22	169	2680 ± 85	2710
ALEURITE	56	484	2510 ± 129	2510
UPPER PANNONIAN	21	217	2430 ± 101	2450
LOWER PANNONIAN	19	118	2560 ± 88	2580
TRIASSIC	3	42	2680 ± 59	2705
ALEUROLITE	20	161	2620 ± 81	2640
ANDESITE	31	324	2480 ± 217	2540
BASALT	13	54	2800 ± 132	2840
BRECCIA	28	97	2580 ± 147	2610
MICA-SCHIST	13	30	2680 ± 162	2710
DIABASE	16	71	2700 ± 116	2730
DOLOMITE	57	692	2700 ± 92	2720
PHILLITE	11	88	2690 ± 78	2710
GNEISS	13	29	2730 ± 298	2690
GRANODIORITE	1	32	2640 ± 36	2645
GRANITE	13	288	2650 ± 72	2640
SANDSTONE	148	114	2550 ± 131	2600
UPPER PANNONIAN	36	239	2350 ± 128	2340
LOWER PANNONIAN	38	463	2480 ± 110	2490
MIOCENE	41	260	2450 ± 122	2470
TRIASSIC	11	186	2640 ± 68	2650
PERM	24	404	2630 ± 49	2630
CONGLOMERATE	57	256	2480 ± 138	2530
LOWER PANNONIAN	7	81	2350 ± 126	2320
PALEOZOIC	11	92	2580 ± 46	2580
QUARTZITE	16	31	2640 ± 55	2650
QUARTZPORPHYRY	9	59	2630 ± 48	2630
MARL	92	620	2540 ± 95	2550
LIMESTONE	116	084	2570 ± 130	2620
MIOCENE	29	231	2370 ± 190	2400
EOCENE	23	431	2520 ± 95	2540
CRETACEOUS	37	544	2620 ± 67	2640
TRIASSIC	41	518	2630 ± 59	2640
PALEOZOIC	6	71	2600 ± 116	2640
LIMEY MARL	71	390	2580 ± 124	2620
LOWER PANNONIAN	21	60	2550 ± 97	2575
CRETACEOUS	13	51	2550 ± 117	2590
JURASSIC	2	50	2670 ± 52	2670
TRIASSIC	15	127	2630 ± 40	2630
CALCIFEROUS SHALE	3	29	2690 ± 22	2700
TUFF	68	309	2280 ± 202	2300
MIOCENE	24	122	2310 ± 133	2305

Table II. Summary of laboratory density data according to their geological formation
II. táblázat. A laboratóriumi sűrűség adatok földtani képződmény szerinti megoszlása

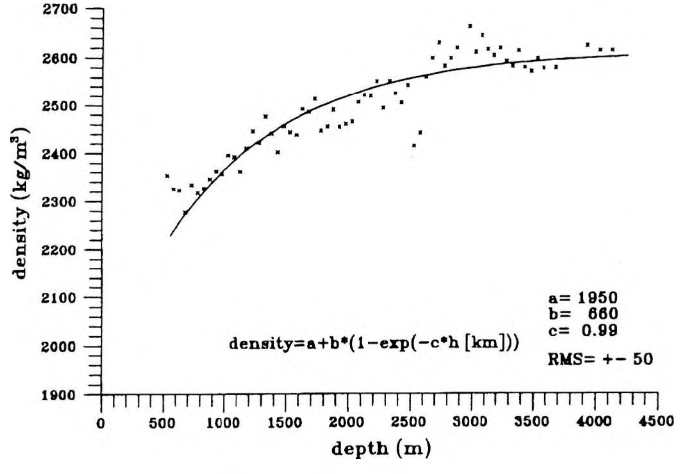


Fig. 7. Combined density function of Pannonian sediments
7. ábra. Pannóniai képződmények sűrűségfüggvénye

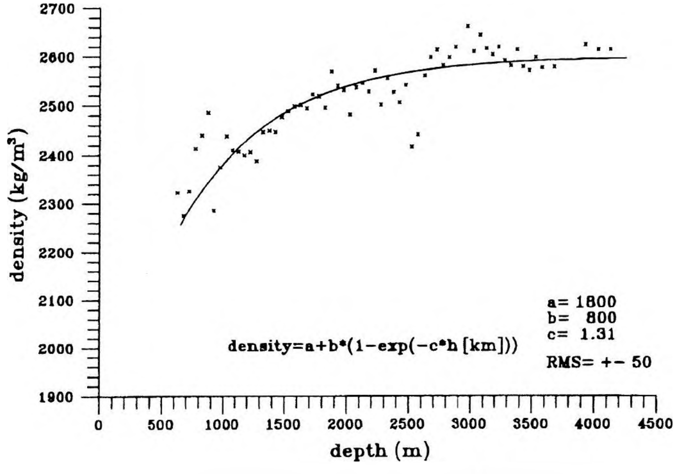


Fig. 8. Density function of Lower Pannonian sediments
8. ábra. Alsópannóniai képződmények sűrűségfüggvénye

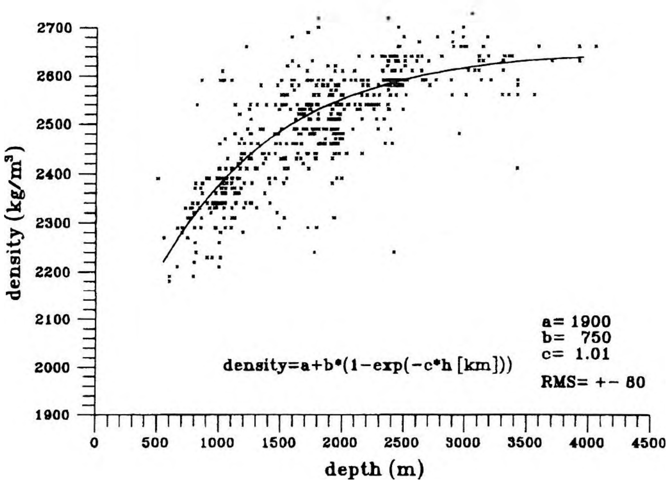


Fig. 9. Density function of Pannonian clays
9. ábra. Pannóniai agyagok sűrűségfüggvénye

For regional studies we present the densities of the whole Tertiary sequence in one diagram (*Fig. 10*) without taking into consideration the actual ages.

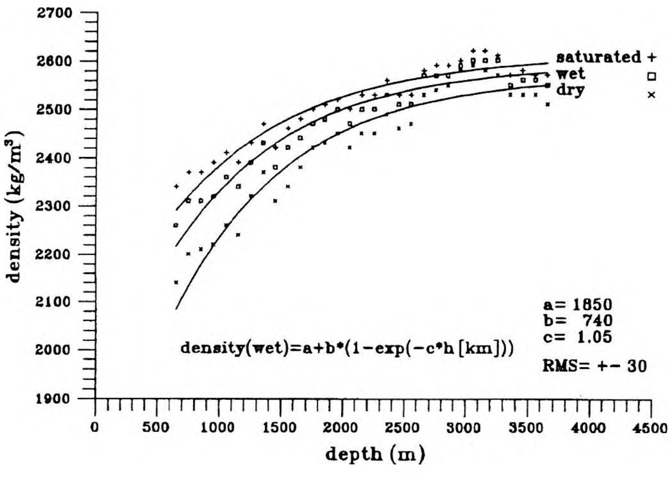


Fig. 10. Density function of Tertiary sediments
10. ábra. Harmadidőszaki képződmények sűrűségfüggvénye

3. Well-log data

The advantage of density values obtained from gamma-gamma logs over the laboratory data is that in the case of the latter, the samples are removed from their natural environment but the well-log data represent the 'in situ' parameters of the rocks. Preliminary assessment has proved that in Hungary only the well-log data taken after 1980 are of sufficiently high quality for accurate density determination. 69 boreholes were at our disposal to carry out the investigations; the total length of the logs was 145000 m; the depth of the bore-hole varied from 330–4780 m; the original sampling rate of the logs was 20 cm. For the evaluation we averaged the density data for 100 m intervals and represented them as a function of depth, separating them according to age. The different age groups could be separated further in accordance with the characteristics of their curve. The Upper Pannonian sediments (*Fig. 11*) could be separated into two groups, viz. the Hegyfalu type (*Fig. 12*) and the Pálmonostora type (*Fig. 13*). The locations of the different types are represented in *Fig. 14*.

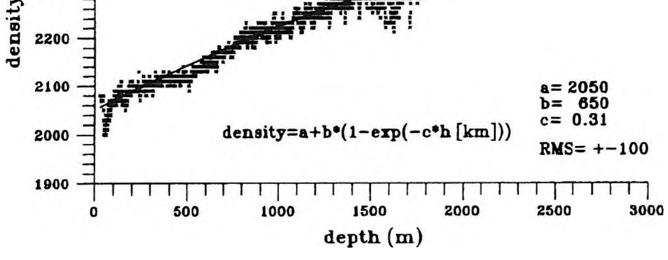


Fig. 11. Combined density function of Upper Pannonian sediments

11. ábra. Felsőpannóniai képződmények egyesített sűrűségfüggvénye

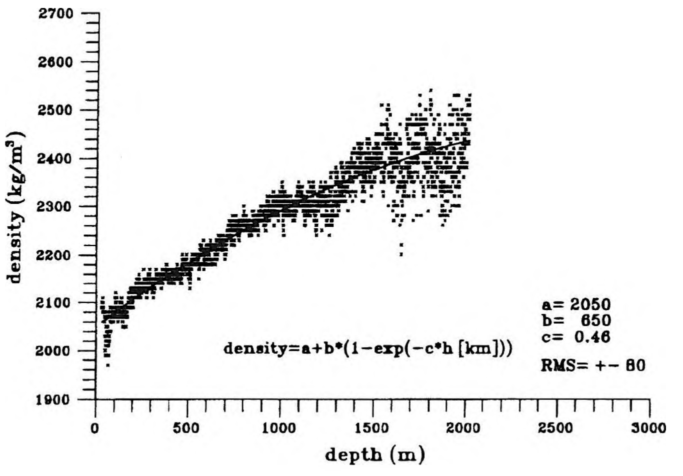


Fig. 12. Density function of Hegyfalu-type Upper Pannonian sediments
12. ábra. Hegyfalu típusú felsőpannóniai képződmények sűrűségfüggvénye

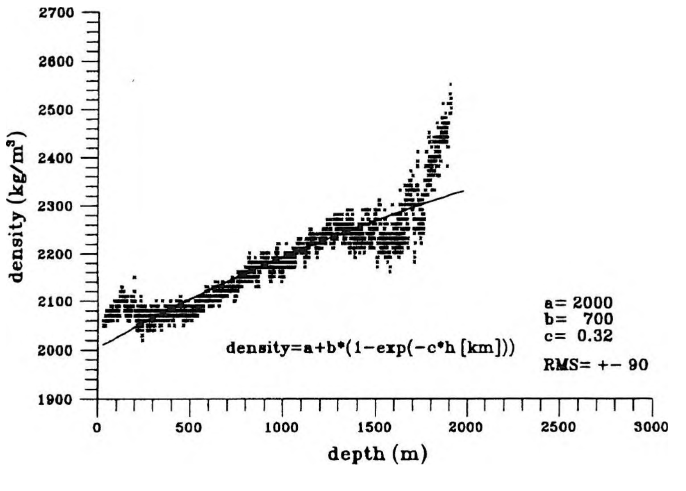


Fig. 13. Density function of Pálmonostora-type Upper Pannonian sediments
13. ábra. Pálmonostora típusú felsőpannóniai képződmények sűrűségfüggvénye

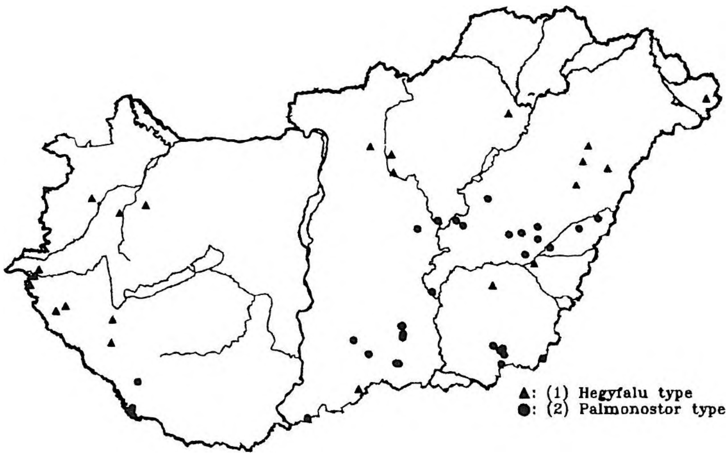


Fig. 14. Location map of Upper Pannonian sediments of different types

14. ábra. A különböző típusú felsőpannóniai képződmények eloszlási térképe

The Lower Pannonian sediments (*Fig. 15*) form three different types: Hegyfalu (*Fig. 16*), Pálmonostora (*Fig. 17*), and Kondoros (*Fig. 18*). *Figure 19* shows the location of these sediments.

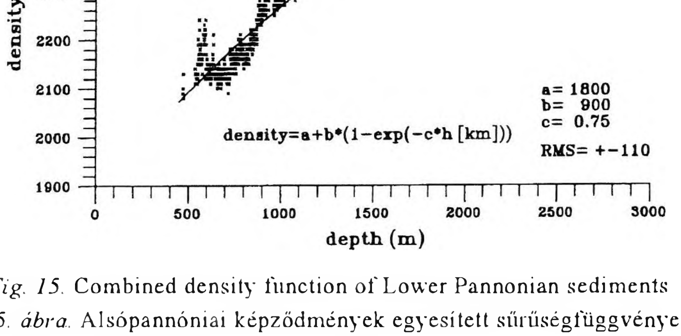


Fig. 15. Combined density function of Lower Pannonian sediments

15. ábra. Alsópannóniai képződmények egyesített sűrűségfüggvénye

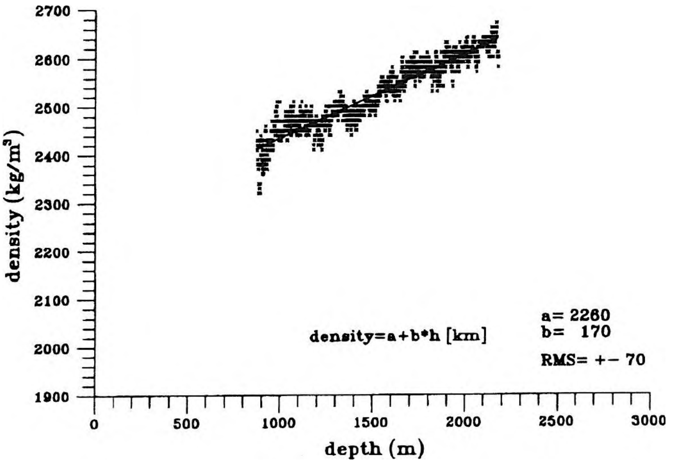


Fig. 16. Density function of Hegyfalu-type Lower Pannonian sediments
16. ábra. Hegyfalu típusú alsópannóniai képződmények sűrűségfüggvénye

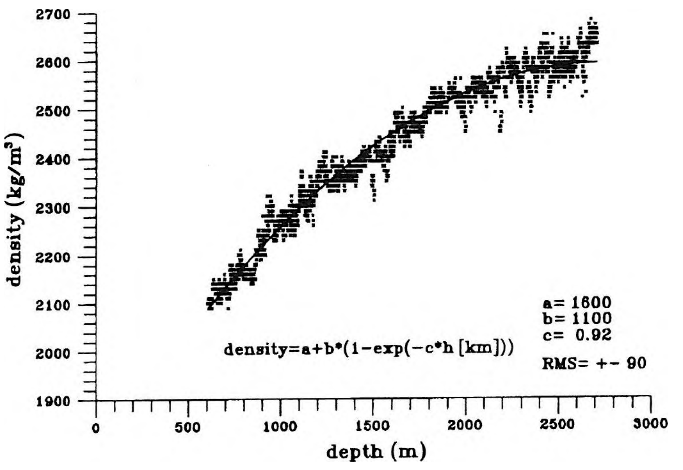


Fig. 17. Density function of Pálmonostora-type Lower Pannonian sediments
17. ábra. Pálmonostora típusú alsópannóniai képződmények sűrűségfüggvénye

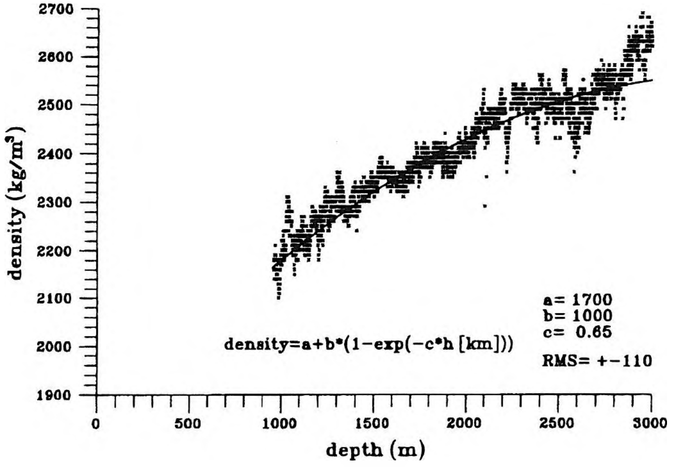


Fig. 18. Density function of Kondoros-type Lower Pannonian sediments
18. ábra. Kondoros típusú alsópannóniai képződmények sűrűségfüggvénye

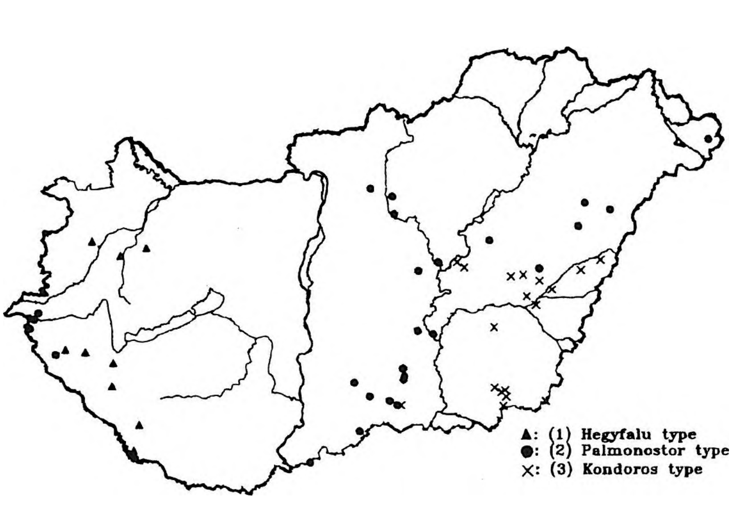


Fig. 19. Location map of Lower Pannonian sediments of different types
19. ábra. A különböző típusú alsópannóniai képződmények eloszlási térképe

Similar to the laboratory data we have prepared one diagram for the whole Tertiary sequence as well. (*Fig. 20*). Based on the characteristics of the curves we could separate the Tertiary curves into two groups: one of them is typical for the Great Hungarian Plain (*Fig. 21*), the other is typical for Transdanubia (*Fig. 22*). The second one has higher density values, indicating a possible post-Pannonian erosion which removed the least consolidated upper parts of the Pannonian layers. A summary is given in *Table III* of the laboratory and well-log data for the Tertiary sequence.

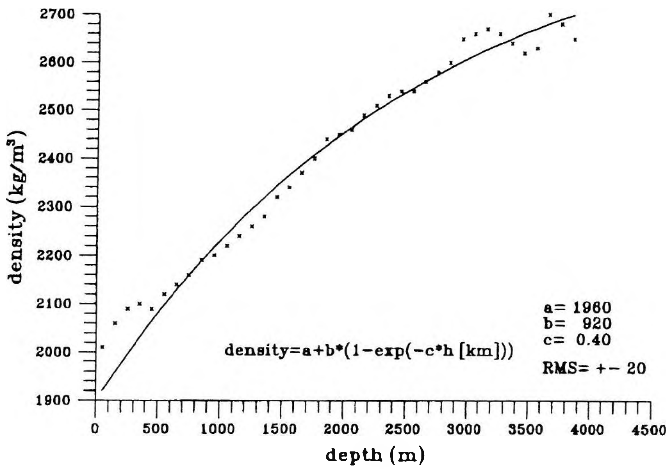


Fig. 20. Combined density function of Tertiary sediments
20. ábra. Harmadidőszaki képződmények egyesített sűrűségfüggvénye

For gravity interpretation the density of the basement is of great significance because all the model calculations based on density differences are carried out with reference to the basement. *Figure 23* presents the histograms of the density of the rocks forming the pre-Austrian basement. The laboratory measurements yielded 2640 kg/m^3 as the average density for the basement while the well-log data provided 2690 kg/m^3. The two values are very similar, their average is in good agreement with the internationally accepted value of 2670 kg/m^3. The higher value obtained from well-log data can be explained by the high hydrostatic pressure (25–100 MPa) pre-

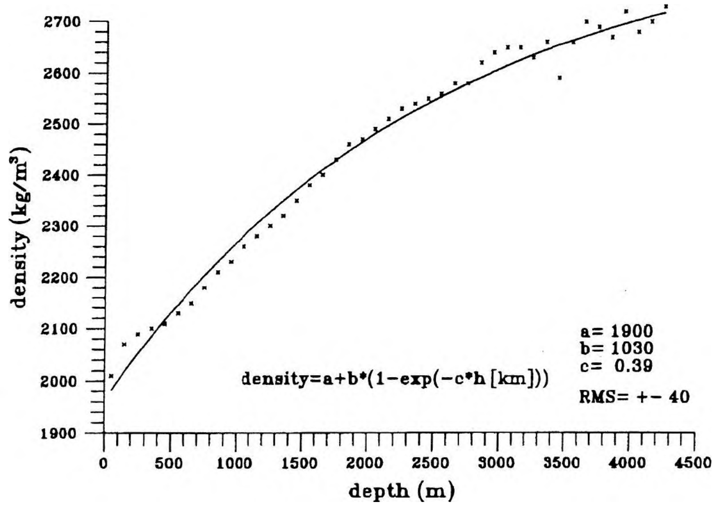


Fig. 21. Density function of Tertiary sediments of the Great Hungarian Plain
21. ábra. A Nagyalföld harmadidőszaki képződményeinek sűrűségfüggvénye

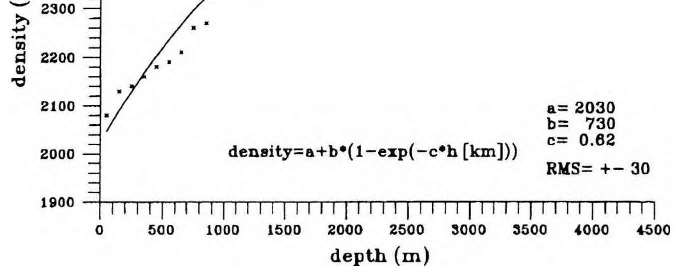


Fig. 22. Density function of Tertiary sediments of Transdanubia
22. ábra. A Dunántúl harmadidőszaki képződményeinek sűrűségfüggvénye

	Measured in lab. wet	**Measured in lab. saturated**	**Well-logging γ-γ**
depth (m)	density (kg/m^3)	density (kg/m^3)	density (kg/m^3)
0 – 100			2010 ± 120 (2223)*
100 – 200			2070 ± 110 (4726)
200 – 300			2090 ± 110 (5040)
300 – 400			2100 ± 100 (5182)
400 – 500			2110 ± 90 (5119)
500 – 600	2260 ±190 (48)*	2340 ±150	2130 ± 90 (5283)
600 – 700	2310 ±180 (66)	2370 ±160	2150 ± 90 (5533)
700 – 800	2310 ±200 (84)	2370 ±180	2180 ± 100 (5560)
800 – 900	2320 ±170 (98)	2390 ±140	2210 ± 100 (5636)
900 – 1000	2360 ±130 (110)	2410 ±110	2230 ± 120 (5709)
1000 – 1100	2340 ±170 (94)	2390 ±150	2260 ± 130 (5664)
1100 – 1200	2390 ±150 (99)	2430 ±130	2280 ± 140 (5465)
1200 – 1300	2430 ±190 (91)	2470 ±170	2300 ± 130 (5362)
1300 – 1400	2380 ±150 (83)	2420 ±130	2320 ± 130 (5214)
1400 – 1500	2420 ±150 (85)	2460 ±130	2350 ± 130 (5165)
1500 – 1600	2440 ±180 (98)	2480 ±160	2380 ± 140 (5039)
1600 – 1700	2470 ± 90 (158)	2500 ± 80	2400 ± 150 (4928)
1700 – 1800	2480 ±100 (93)	2510 ± 90	2430 ± 120 (4659)
1800 – 1900	2500 ±120 (78)	2520 ±110	2460 ± 110 (4625)
1900 – 2000	2470 ±110 (69)	2500 ± 90	2470 ± 110 (4151)
2000 – 2100	2500 ±110 (56)	2530 ± 90	2490 ± 110 (3655)
2100 – 2200	2500 ±120 (46)	2530 ±100	2510 ± 120 (3455)
2200 – 2300	2530 ±140 (38)	2560 ±120	2530 ± 100 (3340)
2300 – 2400	2510 ±140 (36)	2530 ±130	2540 ± 110 (2783)
2400 – 2500	2510 ±100 (51)	2530 ± 90	2550 ± 130 (2412)
2500 – 2600	2570 ± 80 (34)	2580 ± 70	2560 ± 100 (2235)
2600 – 2700	2570 ±100 (31)	2590 ± 90	2580 ± 90 (2239)
2700 – 2800	2570 ±140 (32)	2590 ±140	2580 ± 90 (1959)
2800 – 2900	2590 ±100 (15)	2600 ± 90	2620 ± 90 (1635)
2900 – 3000	2600 ± 90 (13)	2620 ± 70	2640 ± 90 (1149)
3000 – 3100	2600 ± 60 (26)	2620 ± 50	2650 ± 70 (902)
3100 – 3200	2600 ± 80 (33)	2610 ± 70	2650 ± 60 (776)
3200 – 3300			2630 ± 80 (607)
3300 – 3400			2660 ± 70 (509)
3400 – 3500			2590 ± 120 (504)
3500 – 3600			2660 ± 80 (490)
3600 – 3700			2700 ± 40 (400)
3700 – 3800			2690 ± 60 (352)
3800 – 3900			2670 ± 60 (197)
3900 – 4000			2720 ± 20 (100)
4000 – 4100			2680 ± 30 (100)
4100 – 4200			2700 ± 30 (100)
4200 – 4300			2730 ± 30 (100)

* No of samples

Table III. Density data of Tertiary sediments versus depth. In brackets: number of samples

III. táblázat. Harmadidőszaki képződmények sűrűség adatai a mélység függvényében

vailing in the depths in contrast to the normal pressure existing under laboratory conditions.

It is to be hoped that the density information presented in the paper will be useful to all those who are engaged in gravity interpretation.

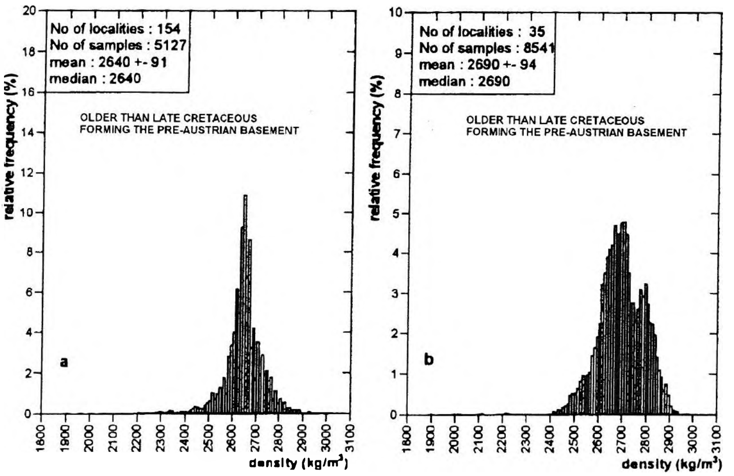


Fig. 23. Density histograms of pre-Austrian basement rocks:
a) laboratory data, b) well-log data

23. ábra. A preausztriai medencealjzatot alkotó képződmények sűrűség hisztogramjai:
a) laboratóriumi adatok, b) karotázs adatok

Acknowledgement

The authors wish to express their appreciation to G. Kövesi for standardizing the laboratory data and F. Mészáros for the preliminary processing of the well-log data. Their contributions were of great help in the final processing and analysis of the density data.

REFERENCES

ATHY L. F. 1930: Density, porosity and compaction of sedimentary rocks. AAPG Bull. **14**, 1, pp. 1–24

KILÉNYI É. 1968: Geological-geophysical considerations on the Great Hungarian Plain through a statistical analysis of density data (in Hungarian). Geophysical Transactions **17**, 4, pp. 41– 49

KOVÁCSVÖLGYI S. 1996: Density determination of near-surface formations based on gravity data – a case history (in Hungarian). Magyar Geofizika **37**, 1. pp. 38 – 45

PINTÉR A., SZABÓ G. 1964: Elevation correction of gravity measurements (in Hungarian). Geophysical Transactions **15**, 1–4 . pp. 159 – 179

ŠEFARA J. (editor) 1987: Struktúrno-Tektonicka Mapa: Vnútorných Západných Karpát pre účely prognózovania ložísk geofyzikalne interpretácie (Structural and tectonic map of the inner West Carpathians for mineral exploration). Slovensky Geologicky Urád, Bratislava, 267 p.

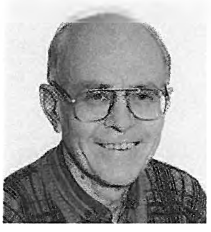
SZÉNÁS GY. 1965: The geological establishment of the geophysical mapping of Hungary (in Hungarian). Special issue of ELGI Vol. II. 167 p. Műszaki Könyvkiadó, Budapest

Kőzetsűrűségek a Pannon medence magyarországi részén

SZABÓ Zoltán és PÁNCSICS Zoltán

A gravitációs kutatások alapvető problémája a kőzetsűrűség. A Bouguer és terrén korrekcióhoz, valamint a gravitációs mérések értelmezéséhez egyaránt megbízható sűrűség adatokra van szükségünk. A szerzők részletes vizsgálatokat végeztek annak érdekében, hogy meghatározzák lehetőleg valamennyi fontosabb a Pannon medencében előforduló geológiai képződmény sűrűségét. A tanulmányt 12 000 db kőzetminta laboratóriumban meghatározott sűrűségadat és 69 mélyfúrás 145 000 folyóméter összhosszúságú gamma-gamma szelvényére alapozták. Az eredményeket hisztogramok, táblázatok és sűrűségfüggvények formájában közlik.

ABOUT THE AUTHORS



Zoltán Szabó graduated as a geophysicist at Eötvös Loránd University, Budapest, in 1955. After graduation he joined ELGI. His main fields of interest are gravity and geomagnetics and their use in structural analysis. Between 1956 and 1959 he carried out torsion balance measurements for oil exploration in China. In the period 1960 to 1963 he took part in the development of the E-60 torsion balance. From 1964 to 1967 he was project manager of mineral exploration in a neovolcanic region (Börzsöny Mts). In 1968–71 he carried out mineral exploration in Nigeria. From 1972 until 1991 he headed the Earth Physics department of ELGI, with the principal tasks of supervising the completion of regional gravity survey of Hungary, compiling a gravity data bank, and dealing with the Tihany Geophysical Observatory, the Geodynamic Station of Mátyáshegy, and the Palaeomagnetic Laboratory. He was the project manager of seismic risk assessment for the site-selection of nuclear installations and low- and intermediate-radioactive waste repositories between 1987 and 1992. Since his retirement he acts as a consultant geophysicist and the curator of the Eötvös Collection.

Páncsics Zoltán graduated from the Eötvös Loránd University, Budapest in 1991 after which he became a postgraduate student in the Department of Geophysics and Geodynamics of Friedrich Wilhelms University in Bonn, dealing with the integrated interpretation of direct current and frequency soundings. He became acquainted with modelling deformation and transportation processes and studied the compaction of young sediments in the lower-Rhine region. He joined the Earth Physics Department of ELGI in 1992, and worked on the tectonic interpretation of gravity data including the evaluation of laboratory measured and well logging density data. His main field of interest is geophysics-related regional tectonics. He is also preparing his doctoral thesis: Numerical modelling of the dynamics of the subduction shear deformations and its applications to the Pannonian basin.

Currently he is with SYSDATA — an affiliated company of SIEMENS AG Austria.

GEOPHYSICAL TRANSACTIONS 1999
Vol. 42. No. 1 – 2, pp. 29 – 40

Bouguer anomaly map of Hungary corrected using variable density

Zoltán SZABÓ* and Zoltán PÁNCSICS*

Hungary's first steps towards a regional gravity survey were taken by Eötvös in 1901. The present work is based on the data of 380 000 gravity stations recorded in the gravity data base till the time of preparing this paper.

The Bouguer anomaly map presented here is based on the complete data set. It is no easy task to construct a unified Bouguer anomaly map for a region where the density of the surface rocks varies from 2000 to 2670 kg/m^3. If incorrect density values are used for Bouguer and terrain corrections they will lead to under or over corrections in the gravity map. On the other hand the application of region dependent density may cause fictive anomalies in the transition zones. To avoid creating such anomalies the authors constructed an elevation dependent density function for the Bouguer and terrain corrections.

Some of the main features of the map and the aspects of gravity interpretation are discussed. To facilitate structural interpretation of the gravity map a gravity lineament map based on the maximum gradient method was constructed.

Keywords: gravity surveys, Bouguer anomalies, maximum gradients, lineaments, density, Hungary

* Eötvös Loránd Geophysical Institute of Hungary, H-1145 Budapest, Kolumbusz u. 17–23.

Manuscript received: 21 March, 1997

1. Introduction

Hungary's first steps towards a regional gravity survey were taken by Eötvös and his assistants on the frozen surface of Lake Balaton in 1901. The gravity measurements were carried out by torsion balances until the advent of gravimeters at the end of the 1930s. From the observed gradients the Δg values were also determined by integration, but the different survey regions could not be merged into a unified map because of the lack of a gravity base net. Since with gravimeters the relative value of gravity can only be measured between two points, the establishment of a gravity reference datum became a basic problem. The first gravity base network consisting of 168 points in the western part of the country (Transdanubia) was established by the Hungarian-American Oil Company (MAORT) between 1939 and 1940.

In the second half of the 1940s gravimeters gradually displaced torsion balances from gravity exploration so the establishment of a fundamental gravity base network for the whole territory of the country became a pressing necessity. The observations of the national gravity network had been carried out between 1950 and 1955. The network consisted of 16 first order and 493 second order points [FACSINAY and SZILÁRD 1956]. The national reference station of the survey was in the laboratory of the Department of Geodesy of the Technical University of Budapest. The reference point was tied to the absolute gravity point in Potsdam's Geodetic Institute, with relative pendulum measurements by K. Oltay in 1908 and 1915.

As a result of this connection the reference level of the first Hungarian gravity base network conformed with the Potsdam Gravity System, which is the reference point of the global gravity system as well. In later years the Hungarian Gravity Base Network was upgraded and updated so it is now converted to the absolute system [CSAPÓ 1996]. Since 1950 all gravity measurements carried out in this country have been tied to the national base network. The first regional gravity map series of Hungary was prepared in the 1960s. At that time a large part of the country was surveyed by torsion balance only so the isogams calculated from gradient values were converted into the gravity base net. The reconnaissance gravimeter survey of the country consisting of 120 000 points (1.2 point/km^2) was completed and stored in computer readable form (punched cards) at the end of the 1970s.

In later years the gravity data base was complemented by the data of detailed gravity surveys carried out partly by ELGI, partly by the oil industry. At present the gravity data base contains 380 000 points all in the same gravity system. During recent decades several national Bouguer anomaly maps based on different data bases were prepared but because of the confidential nature of gravity data they were not published.

2. The Bouguer anomaly map

'If the earth were a homogenous sphere at rest, gravity would be the same everywhere over the surface of the earth and would vary only radially, and the level surfaces would be spherical and concentric with the earth's surface. But the earth is not at rest; it is not a sphere, and its outer crust is in no way homogeneous' [BARTON 1929].

The rock bodies that make up the lithosphere are very irregular in size, density, and distribution consequently they produce varying anomalies of the gravity field. The magnitude of anomaly locally is at a maximum in the case of excess mass and at a minimum with deficiency of mass.

Those factors which affect the observed gravity are: the station elevation (because of the varying distances from the centre of the earth), irregularities of topographic form, the normal gravity field of the earth, and the density of the masses between the station and the reference elevation (generally sea level).

To make each observed gravity value comparable with every other one the effects of the above-mentioned factors have to be taken into account and the observed value should be reduced by them to a common datum, which considers not only the elevation, but also the Bouguer effect, the earth's normal gravity field, and the topography.

In order to calculate the Bouguer and terrain effects the density of the mass between the station and the elevation datum and the topography around the station have to be determined. In normal conditions the density determination is the crucial point of the reduction. The problem is even more important if the near-surface density has lateral variations between 2000 and 2670 kg/m^3 as in the case of Hungary. For an exact solution a three-dimensional density model would be required. Since density information is far from a three-dimensional model we had to restrict ourselves to estimates and approximations.

Direct determination of the density can be performed by laboratory methods measuring mass and volume of rock samples collected from outcrops or drilling cores. In gravity exploration an indirect density determination is possible by the method of Nettleton [NETTLETON 1939]. The hypothesis is based on the assumption that the Bouguer anomalies, if computed with correct density, do not correlate with the topography. The method can be applied for gravity profiles and area surveys as well. The Bouguer anomalies have to be calculated with different density values and then correlated with the topography. In other words, the correct density is where the calculated Bouguer anomalies have the least correlation with the topography.

Before constructing the Bouguer anomaly map we carried out density determinations with the Nettleton method for various hilly and mountainous regions of the country. From the results it could be concluded that the internationally accepted σ=2670 kg/m^3 density value can be regarded as a good approximation for all but one of the investigated regions. The exception is the neovolcanic area of the Zemplén Mts. where due to the predominantly tufaceous build up of the region, σ=2000 kg/m^3 was determined (*Fig. 1*).

The main problem in constructing a Bouguer anomaly map with variable density is the transition zone between geological structures with sharp density contrast. A sharp density difference in the Bouguer correction of neighbouring stations produces a high fictitious gradient in the Bouguer anomaly map which can lead to serious misinterpretations. To eliminate the distortion effect of density transitions we applied an elevation-dependent density function in the 100–435 m height interval:

$$\sigma_h = \sigma_0 + 0.01\frac{h - h_0}{5}$$

where σ_h: density in elevation h

σ_0: 2000 kg/m^3 density in h_0=100 m elevation

Fig. 1. Location map with the contours of the basement in km units
1. ábra. Helyszínrajz a medencealjzat km-es szintvonalaival

h: elevation of the point in m unit
$\sigma = 2000$ kg/m^3 if $h \leq 100$ m
$\sigma = 2670$ kg/m^3 if $h \geq 435$ m

Using these elevation-dependent density values for the Bouguer corrections one can ensure gradual transition in the Bouguer anomalies and the likelihood of fictitious anomalies arising could be prevented.

The Bouguer anomaly map presented here (*Fig. 2*) is based on 380 000 randomly distributed gravity data. The basic parameters of the map are given in the figure.

3. Some characteristics of the map

It can be seen on the Bouguer anomaly map that the deep basin areas have smooth anomalies with low gradients. In those parts of the country where the basement is on or near to the surface the anomalies become more disturbed and high gradients usually prevail.

The most characteristic features of the map are the elongated anomalies of NE–SW strike. This pattern of the anomalies served as one of the main clues to the nowadays generally accepted view of the basement topography and structure. It must be noted however that gravity anomalies reflect the effect of mass excess and mass deficiencies which may or may not be directly connected with different geological formations. It is worth mentioning that the highest anomalies are not in that region where the high-density basement (dolomite and crystalline limestone) is outcropping but in the lower middle part of the country where the basement is below sea level.

Another interesting feature of the map is the about 25–30 mGal difference between the anomalies of the deep basins of the SE part of the country (Békés basin, Makó trough) having the higher values and those of the W and NW part (Zala basin) having the lower values. This phenomenon can be regarded as originating from crustal sources, viz. from the isostatic effect of the Alps.

4. Aspects of gravity interpretation

Here, it is not our intention to interpret the Bouguer anomaly map, but to call attention to some basic characteristics that should be kept in mind when interpreting gravity maps.

The objective of the interpretation of gravity maps is to deduce the geological build-up of the subsurface from the anomalies of the gravity field. All gravity anomalies originate from horizontal density variations. If the earth were built up of layers of horizontally uniform density, there would be no gravity anomalies even if vertical variation in density were to exist.

Since the Bouguer anomaly map reflects the integrated effect of subsurface masses the anomaly map is a complex pattern of subsurface geology; however, in special cases single sources can be identified. This means that the interpretation can never provide an unambiguous answer to a given geological problem because there is no single mathematical solution to the determination of the sources of anomalies. Generally speaking, sharp anomalies are caused by near-surface sources, broader anomalies by deep ones.

To illustrate the smoothing effect of the depth we present the gravity effect and its horizontal gradients of a vertical fault with 400 m throw at four different depths (*Fig. 3*). The density contrast between the hypothetical basement and the sediment was determined from the density function of

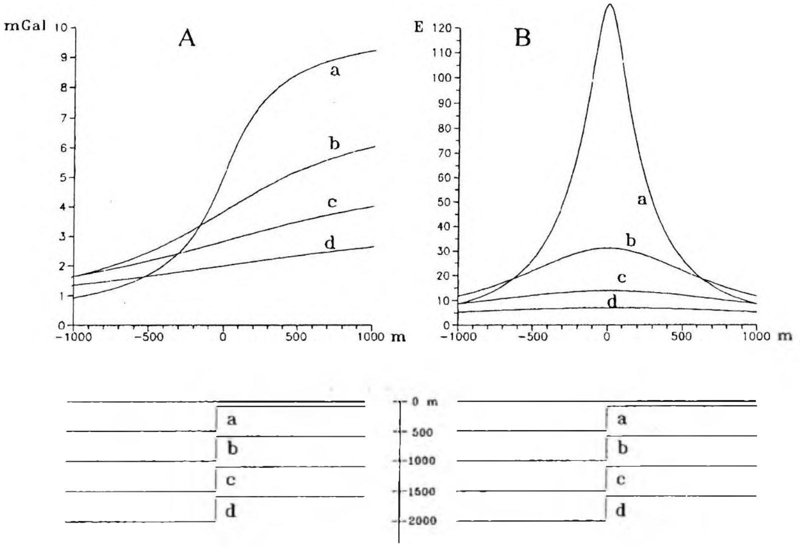


Fig. 3. Vertical fault: A – gravity effect; B – horizontal gradients

3. ábra. Függőleges vető: A – gravitációs hatása; B – horizontális gradiense

the sedimentary cover. The figure represents the masking effect of the sedimentary cover, i.e. in spite of the rugged topography of the basement the gravity contours are smooth and the corresponding gradients are low if the basement lies at a great depth.

The crucial point of gravity interpretation is to separate the effects of different sources. This is a difficult task and depends on the skill of the interpreter in choosing the most convenient procedure and parameters.

For separating the different elements many procedures have been devised: from manual 'smoothing' to the more sophisticated computer-based filtering techniques. The common aim in all procedures is to emphasize certain elements and to suppress others — which is equivalent to enhancing certain frequencies and suppressing others depending on the target of investigation.

As a result of anomaly separation we can speak about regional, residual, and derivative maps. But the designation of regional is also subjective since it refers to broad anomalies with sources normally deeper than the target of prospecting. It is difficult to differentiate between residual and derivative maps, but it is important because residual maps reflect the gravity effect of local sources relatively near to the surface and derivative maps reflect the gradient of the gravity field. These latters enhance the zones of maximum gravity variations which in most cases indicate the existence of structural lines with density contrast across them.

The resolution of a gravity survey depends on the measurement spacing but resolution decreases with increasing depth of source no matter how accurately we know the gravity field.

Filtering techniques are very sensitive to the size of the applied filter. With the combination of the matrix elements and the size of the filter many map variations can be produced. The proper designation of the resulting maps, however, depends very much on the skill of the interpreter not to mention the target of the prospecting: e.g. 'residual' has a completely different meaning in the case of ore prospecting and in the case of oil exploration.

According to PINTÉR and STOMFAI [1979] one possible way of characterizing map variations is by using the distribution of anomaly values. Maps with broad anomaly distribution curves can be regarded as residual ones, those with sharp distribution curves, as derivative ones.

To interpret a gravity map the above-mentioned characteristics and limitations have to be kept in mind otherwise one may reach wrong conclusions by interpreting a residual map as a derivative one or vice versa. Therefore it can be stated that there is no single or direct solution for eliminating regional effects and isolating local anomalies. All methods have their merits and limitations but a combination of them can provide useful information on the geological sources of the different anomalies.

The practical ambiguity of the inverse gravimetric task is always smaller than the theoretical one and it depends on a variety of other data (borehole data, seismic profiles, etc.).

As a conclusion, it has to be mentioned that Bouguer anomalies contain all information about the gravity field of a prospect area. The different transformations of the Bouguer anomaly map do not create new information but only emphasize certain elements or features of the original field to make it easier for the interpreting geologist to recognize such characteristics of the field which are not evident in the original Bouguer anomaly map.

5. Determination of lineaments based on maximum gradients of Bouguer anomalies

In basin areas, where the high-density basement is overlain by young, loose sediments, the topographical changes of the basement present themselves in a horizontal plane as density contrasts. The best way to detect a density contrast along a line is to analyse the horizontal gradient of the gravity field. The unit of the gradient is 1 E (Eötvös). Expressing this in a more practical way: 1E=0.1 mGal/km, i.e. gravity acceleration changes by 0.1 mGal within a distance of 1 km. In geological practice the gradient changes from 1–100 E. Above a fault where an abrupt density change exists, the horizontal gradient has a local maximum (Fig. 3). This phenomenon is utilized in determining structural lineaments.

As a first step, the Bouguer anomaly values were interpolated to a 1 km grid and from these interpolated values the horizontal gradients were calculated. A computer program had earlier been developed for the selection of local gradient maxima. The program's parameters can flexibly be varied depending on the radius of the neighbourhood to be studied and on the prescribed difference from the average to be selected. Moreover, the program can take into consideration the neighbouring gradients as well, when selecting the locations of gradient maxima. Since the gradients are perpendicular to the strike of the fault a new computer program has been developed to present the gravity lineaments reflecting the strikes of tectonic elements. In deep basins the Bouguer anomaly contours are smooth and have low relief because the main anomaly source, the basement, is deep and the density contrast between the sedimentary fill and the bedrock decreases with increasing depth. As a consequence the gradients of the gravity field are higher over areas where the bedrock is outcropping or is at a relatively shallow depth. In deeper basins, however, the gradients are low in spite of the possible rugged topography of the basement. This effect is illustrated in Fig. 3. It can be seen that the magnitude and the sharpness of the anomaly decrease considerably with increasing depth. To eliminate this distorting effect a depth dependent density correction was applied to the gradients. If we take the density of the near-surface sediments to be 2000 kg/m^3 and that of the basement to be 2670 kg/m^3, the near-surface density difference is 670 kg/m^3. By correction, the decreasing density difference with depth was normed to this 670 kg/m^3 value. The depths belonging to the gradients were taken from the pre-Tertiary basement topography map of Hungary [KILÉNYI et al. 1991], and the density values from the density versus depth function of Tertiary sediments [SZABÓ–PÁNCSICS 1999]. The correction enhanced the effect of deep-seated tectonic elements.

The gradient map is given in two different representations which complement each other (*Fig. 4, Fig. 5*). In Fig. 4. the lineaments are more characteristic but Fig. 5 is more expressive. Integrated study of the two maps helps one to understand the nature and importance of the different features, and to detect those structural lines taking part in the shaping of basement topography.

The gradient maps reflect the main structural lines of the country; most of them are more or less known from former studies. Here we would call attention only to the two lines marked 'a' and 'b' in Fig. 5. Line 'a' marks what is probably the eastern limit of the granitic belt of the basement extending from the western border of the country along the southern bank of Lake Balaton, its most eastern known manifestation is the granitic outcrop of the Velence Hills (No 5 in Fig. 1). Line 'b' offers an answer to the so far unresolved problem of the separation of the Mesozoic formations of the Transdanubian Central Range from the Triassic rocks of the Bükk mountains. We have stressed the importance of these two lines only, but further study of the maps — especially on a larger scale combined with geological information — will reveal more details about the geological structure of different regions. It is planned to deal with this topic in a future paper.

Acknowledgements

The authors wish to acknowledge their debt to A. Sárhidai and S. Kovácsvölgyi for developing and managing the gravity data bank, and to all the people who took part in the gravity survey of the country, without whose contributions the Bouguer anomaly map of Hungary could not been established.

REFERENCES

BARTON D. C. 1929: The Eötvös Torsion Balance Method of Mapping Geologic Structure. Geophysical Prospecting, published by A.I.M.M.E. New York pp. 416–479

CSAPÓ G. 1996: Hungary's new gravity base network (MGH-2000). Geophysical Transactions **40**, 3–4, pp. 119–143

FACSINAY L., SZILÁRD J. 1956: Hungarian gravity base network (in Hungarian). Geophysical Transactions **V**, 2, pp. 3–49

KILÉNYI É., KRÖLL A., OBERNAUER D., ŠEFARA J., STEINHAUSER P., SZABÓ Z., WESSELY G. 1991: Pre-Tertiary basement contour map of the Carpathian Basin beneath Austria, Czechoslovakia and Hungary. Geophysical Transactions **36**, 1–2, pp. 7–15

NETTLETON L. L. 1939: Determination of density for the reduction of gravimeter observations. Geophysics **4**, pp. 176–183

PINTÉR A., STOMFAI R. 1979: Gravitational model calculations. Geophysical Transactions **25**, 1, pp. 5–31

SZABÓ Z., PÁNCSICS Z. 1999: Rock densities in the Pannonian basin — Hungary. Geophysical Transactions, present issue

Magyarország változó sűrűséggel korrigált Bouguer anomália térképe

SZABÓ Zoltán és PÁNCSICS Zoltán

Magyarországon az első gravitációs felmérést Eötvös végezte 1901-ben. A cikk írásának időpontjáig a gravitációs adatbnak kb. 380 000 gravitációs állomás adatait tartalmazta, mely átlagosan 4 pont/km állomássűrűségnek felel meg.

A bemutatott Bouguer anomália térkép a teljes adatkészlet felhasználásával készült. Egységes Bouguer anomália térkép szerkesztése egy olyan területen, ahol a felszínt alkotó kőzetek sűrűsége 2000–2670 kg/m^3 között változik meglehetősen nehéz feladat. Helytelen sűrűség alkalmazása a Bouguer- és terrén-korrekció meghatározásánál a gravitációs térkép alul-, vagy túlkompenzálását okozza. Területtől függő sűrűség alkalmazása következtében viszont fiktív anomáliák jelenhetnek meg az átmeneti zónákban. A szerzők, hogy elkerüljék fiktív anomáliák kialakulását egy magasságtól függő sűrűségfüggvényt alkottak és ezt alkalmazták a Bouguer- és terrén-korrekció meghatározásánál.

Tárgyalják a térkép néhány jellegzetes vonását és a gravitációs értelmezés lehetséges szempontjait. A maximális gradiens módszerével gravitációs lineamens térképet szerkesztettek, hogy megkönnyítsék a gravitációs térkép szerkezeti értelmezését.

ABOUT THE AUTHORS

Zoltán Szabó, for a photogprah and biography, see this issue, p. 27.

Zoltán Páncsics, for a photogprah and biography, see this issue, p. 27.

GEOPHYSICAL TRANSACTIONS 1999
Vol. 42. No. 1 – 2. pp. 41 – 54

Gravity map of Hungary corrected for basin effect

Zoltán SZABÓ* and Zoltán PÁNCSICS*

More than 70 % of Hungary is basin area covered by young sediments. The thickness of the sedimentary cover reaches 7000–8000 m in the deepest parts of the basins therefore the mass deficiency caused by the sediments has a considerable influence on the Bouguer anomaly map. Based on model calculations the authors determined the optimum procedure to calculate the 'distortion effect' of the sediments. The corrected gravity map presented in this paper reflects the effect of the density variations and the thickness of the crust. To emphasize the structural lines, a gravity lineaments map was constructed as well. Based on earlier seismic information, correlation studies were carried out between the depth to the Moho derived from seismic data and the corrected gravity anomalies. In possession of the correlation parameters the contours of the Moho could be determined.

Keywords: gravity survey maps, stripping, lineaments, Mohorovičić discountinuity, basin effect

1. Introduction

Some 70 % of the territory of Hungary is covered by young sediments. The thickness of the sedimentary cover reaches 7000–8000 m in the deepest parts of the basins. Due to the unconsolidated young sediments the Bouguer anomaly map is strongly influenced by the effect of the mass deficiency of the basins. To get information about the structure of the basement

* Eötvös Loránd Gophysical Institute of Hungary, H-1145 Budapest, Kolumbusz u. 17–23.
Manuscript received: 3 April, 1997.

and of the crust the Bouguer anomaly map should be corrected for the basin effect.

A correction technique, the so-called stripping method, was proposed by HAMMER [1963] to eliminate the effect of known structures. In principle the gravity stripping technique extends the gravity interpretation to deeper horizons by calculating and subtracting the gravity effect of the overlying sedimentary cover whose structure and density relations are known in detail from former geophysical investigations and drilling activities. For areas of large extent the selection of the proper datum level of the correction is critical because of the inhomogeneity of information about the depth and density of the overlying layers. To eliminate this problem we have modified the stripping technique by selecting sea level as the datum level and calculating the gravity effect of the mass deficiency caused by the unconsolidated sediments between the sea level and the basement. This means that instead of stripping the known effects we calculated the effect of the missing masses; in other words, we substituted the young sediments by basement material. The effect obtained was added to the original Bouguer anomaly values.

The result is a version of the Bouguer anomaly map which does not reflect the effect of the young sediments but only the structure and density relations of the basement and the underlying crust. In this case the datum level is equivalent to that of the original Bouguer anomaly map.

In former years RENNER and STEGENA [1966], MESKÓ [1984], and BIELIK [1985] constructed a number of different stripped maps for Hungary but due to the lack of a proper contour map of the basement and adequate density data of the sediments at their disposal, their experiments can only be regarded as first approximations.

2. Principal factors for determining the basin effect

In order to determine the basin effect three principal factors are needed:

— Bouguer anomaly map
— contour map of the basement

— density relations of the sediments.

Thanks to the complemented gravity database of ELGI, the compilation of a basement countour map of Hungary, and the thorough investigation of the density relations of the Pannonian basin carried out in the last decade, the principal factors were at our disposal for calculating the basin effect.

Bouguer anomaly map

The Bouguer anomaly map was contoured on the basis of about 380000 gravity data. For Bouguer- and terrain corrections, elevation-dependent density values were applied, taking the highest value to be 2670 kg/m^3 (see separate paper in this volume). The accuracy of the map is more than adequate for a regional study on the structure of the crust.

Contour map of the basement

From the point of view of geophysical methods the basement of the young sediments is a first-order discordance originating from the contrast in the physical parameters of the sediments and basement rocks.

After all, it is not an easy task to contour the basement topography because integrated evaluation of many different data is needed which cannot be carried out without the close co-operation of scientists working in different branches of geological exploration. The basement contour map of Hungary [KILÉNYI, RUMPLER 1984] is based on borehole, seismic, geoelectric, and gravity data. The most reliable data come from borehole records but since the boreholes were drilled for mineral exploration purposes they are unevenly distributed. The oil-prospecting boreholes were drilled mainly on the elevated parts of the basement so in the deep basins the boreholes do not actually reach or penetrate the basement.

Since most of the geophysical measurements were carried out for mineral exploration, the geophysical coverage of different parts of the country differs. The most important regions for hydrocarbon exploration are: the Dráva basin, the Zala basin, the south-eastern part of the Danube–Tisza interfluve, and the southern part of the trans-Tisza region. In these regions the oil industry carried out detailed seismic reflection surveys and most of the boreholes were drilled in these parts of the country. The above-mentioned regions have the most detailed geophysical coverage but the different measurements are not all of the same quality. The depth range of the prospecting was determined by the specific purpose of the prospecting.

The Danube–Rába Lowland, the foreland of the Transdanubian Central Range, and the vicinity of the Mecsek–Villány mountains are in the second category.

The least explored regions at the time of the compilation of the map were: the region between Lake Balaton and the Mecsek Mts, the western part of the Danube–Tisza interfluve, and the Nyír region, where several thousand metre thick Neogene volcanic rocks and sediments cover the basement. The variable geophysical coverage means uneven reliability of the map.

Since geology does not conform with national boundaries the Pannonian Basin extends over the territory of neighbouring countries. In the second half of the 80s in co-operation with Austrian and Slovakian scientists Pre-Tertiary Basement Contour Map of the Carpathian Basin beneath Austria, Czechoslovakia and Hungary was prepared [KILÉNYI et al. 1991]. At present this is the most up-to-date basement contour map of the region. But because for the correction a larger area is needed, than is covered by the contour map, the latter was complemented by partly unpublished data (*Fig. 1*).

Density relations of the sediments

Up till now only sporadic density data have been published for the Hungarian part of the Pannonian basin [PINTÉR et al. 1964, KILÉNYI 1968, KOVÁCSVÖLGYI 1995]. More density information is available for the Slovakian part of the basin [ŠEFARA 1987]. Despite its incompleteness, for regional studies the above-mentioned data set was considered acceptable but nowadays, when detailed large-scale investigations are on the agenda, more accurate density data are needed for really meaningful geophysical interpretation.

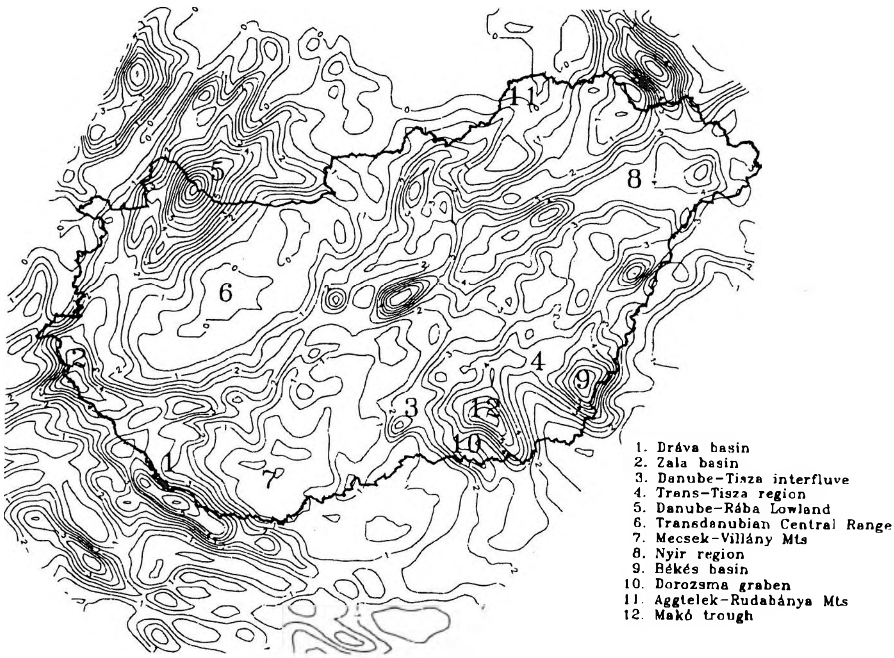


Fig. 1. Pre-Tertiary basement contours. Contour interval: 500 m

1. ábra. A neogén medence aljzatának szintvonalai. Szintvonalköz: 500 m

Rock densities can be determined indirectly from gravity data (Nettleton method) or directly either by laboratory measurements or by gamma-gamma density logs as was done by the authors (see separate paper in this volume).

To investigate the density contrast at the sediment–basement boundary we calculated the average density of the lowest 100 m density column of the sediment and that of the basement. The density values that we obtained confirmed the results given by the density functions derived from well-log data and laboratory measurements; in other words at about 3500 m depth the density of the sediments and the basement became very similar. This phenomenon calls our attention to the problems of basement contour determination from gravity data below 3500 m.

3. Preliminary model calculations

Before calculating the corrections we had carried out some preliminary investigations to sum up the limitations of the method. Because variations of the basement relief did not make it possible to approximate the sedimentary layers with infinite flat plates we divided the subsurface layers into 4000×4000×250 m blocks and calculated their gravity effect for each grid point (*Fig. 2*). The calculations were based on the method proposed by HAÁZ [1953]. In theory, the calculations should be extended over the whole surface of the Earth, but partly because of lack of data and partly to save computer time, we limited the calculations to a region of radius *R*.

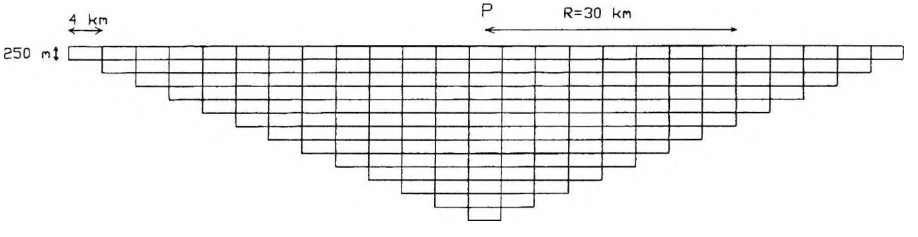


Fig. 2. Sketch of model calculations
2. ábra. A modellszámítások vázlata

To estimate *R*, i.e. the radius of the area to be taken into account in the correction, model calculations were carried out. Three basin areas were selected for the study where the basement depth was H=500, 2500 and 5000 m, respectively. For certain grid points the effect of mass deficiency of the basins was calculated by increasing *R* in 4 km steps. The effect belonging to R=42 km was taken as 100 % because the difference in the correction between the last consecutive steps was less than 0.28 % even in the case of a 5000 m deep basin. The effects belonging to different *R* values are expressed as a percentage of the total (R=42 km) effect (see *Table I*). It can be seen that the relative error is less than 1 % even in the worst case if we limit our calculations to R=30 km. That error in our case means less than

0.5 mGal regional uncertainty, which is negligible in regional interpretation.

R [km]	W_1	W_2	W_3
2	83.67	63.90	59.49
6	93.63	87.25	85.10
10	95.70	92.90	91.92
14	96.77	95.39	94.97
18	97.62	96.81	96.68
22	98.26	97.78	97.77
26	98.70	98.48	98.52
30	99.06	98.99	99.05
34	99.44	99.40	99.45
38	99.77	99.73	99.76
42	100.00	100.00	100.00

Table I. Relative gravity effect (W) as a function of R. W=percentage of total effect for a basin of 1 — H=500 m, 2 — H=2500 m, 3 — H=5000 m depth

I. táblázat. Relatív gravitációs hatás (W) R függvényében. W a teljes hatás százalékában van megadva, 1— H=500 m, 2 — H=2500 m, 3 — H=5000 m mély medencére

We were interested in what weight the 250 m thick layers of different depths have in the integrated mass deficiency. The calculations were carried out with three different density functions for the same H=500, 2500 and 5000 m deep basins as in the case of R determination. The results were very similiar: two of them are presented in *Tables II and III*.

It can be concluded that the effects of the near-surface layers dominate, and more than 90 % of the total effect originates from depths less than 2500 m. This means that the correction is sensitive to the density of the near-surface layers. Therefore the density determination of the relatively shallow geological formations requires special attention.

h [m]	$\sigma(h)$ [kg/m^3]	$\Delta\sigma(h)$ [kg/m^3]	W_1	W_2	W_3
0–250	2060	610	51.61	17.39	15.76
250–500	2110	560	46.99	15.85	14.35
500–750	2150	520	0.88	14.62	13.24
750–1000	2210	460	0.29	12.83	11.62
1000–1250	2270	400	0.13	11.07	10.02
1250–1500	2330	340	0.06	9.33	8.45
1500–1750	2390	280	0.04	7.56	6.91
1750–2000	2460	210	0.00	5.52	5.14
2000–2250	2510	160	0.00	3.51	3.86
2250–2500	2540	130	0.00	2.06	3.10
2500–2750	2570	100	0.00	0.19	2.35
2750–3000	2610	60	0.00	0.07	1.37
3000–3250	2620	50	0.00	0.00	1.09
3250–3500	2620	50	0.00	0.00	0.99
3500–3750	2630	40	0.00	0.00	0.64
3750–4000	2630	40	0.00	0.00	0.52
4000–4250	2640	30	0.00	0.00	0.24
4250–4500	2640	30	0.00	0.00	0.20
4500–4750	2650	20	0.00	0.00	0.07
4750–5000	2650	20	0.00	0.00	0.05
5000–5250	2660	10	0.00	0.00	0.02
5250–5500	2660	10	0.00	0.00	0.01
5500–5750	2670	0	0.00	0.00	0.00
5750–6000	2670	0	0.00	0.00	0.00

4. Gravity map of Hungary corrected for basin effect

Based on the above-mentioned model calculations R=30 km was accepted as the radius of the area involved in the correction. To improve the resolving power, the calculations were carried out for 1000×1000×250 m blocks instead of the 4000×4000×250 m blocks used in the model calculations. To make the procedure flexible, the effect of each block was calculated for each grid point with σ=1000 kg/m^3 density in order to reduce the further calculations to simple multiplications with the actual density and summation of the effects of different blocks. The following parameters were selected for the final calculations:

— density of the bedrock: 2670 kg/m^3

— R=30 km

— 3 different density functions, based on laboratory measurements and on gravity-derived and well log density data.

The correction factor was determined for each grid point and it was added to the mean Bouguer anomaly of the actual block. The corrected gravity values reflect that hypothetical situation as if all the young sediments were substituted by bedrock of σ=2670 kg/m^3 density. The contoured map indicates the density inhomogeneities of the basement and the undulation of the Moho. It has to be mentioned, however, that it is subject to all the uncertainties of the basement contours and the applied density relations. The corrected gravity map was prepared in three versions on the basis of the three density functions. They were compared with each other and it was concluded that they were very similar especially with regard to their anomaly pattern. The amplitudes of the respective anomalies differed by not more than 10 %, neglecting an additive constant, not influencing the anomaly pattern.

⇦ *Table II*. Relative weight of 250 m thick consecutive layers as a function of depth (h). Density function ($\sigma(h)$) from gamma–gamma logs. $\Delta\sigma(h)$=2670–$\sigma(h)$, W=percentage of total effect for a basin of 1—H=500 m, 2—H=2500 m, 3—H=5000 m depth

II. táblázat. A 250 m vastag rétegek relatív súlya (W) a mélység (h) függvényében. Sűrűségfüggvény ($\sigma(h)$) gamma–gamma karotázsból. $\Delta\sigma(h)$=2670–$\sigma(h)$, W a teljes hatás százalékában van megadva, 1—H=500 m, 2—H=2500 m, 3—H=5000 m mély medencére

h [m]	$\sigma(h)$ [kg/m^3]	$\Delta\sigma(h)$ [kg/m^3]	W_1	W_2	W_3
0–250	2050	620	54.44	21.13	19.01
250–500	2160	610	44.42	17.25	15.52
500–750	2250	520	0.74	14.10	12.68
750–1000	2320	450	0.23	11.66	10.48
1000–1250	2370	400	0.10	9.92	8.92
1250–1500	2420	350	0.05	8.19	7.37
1500–1750	2460	210	0.02	6.78	6.14
1750–2000	2500	170	0.00	5.33	4.93
2000–2250	2540	130	0.00	3.41	3.73
2250–2500	2570	100	0.00	1.87	2.83
2500–2750	2590	80	0.00	0.18	2.21
2750–3000	2610	60	0.00	0.08	1.63
3000–3250	2620	50	0.00	0.04	1.29
3250–3500	2620	50	0.00	0.03	1.18
3500–3750	2630	40	0.00	0.02	0.75
3750–4000	2630	40	0.00	0.01	0.62
4000–4250	2640	30	0.00	0.00	0.29
4250–4500	2640	30	0.00	0.00	0.25
4500–4750	2650	20	0.00	0.00	0.09
4750–5000	2650	20	0.00	0.00	0.05
5000–5250	2660	10	0.00	0.00	0.02
5250–5500	2660	10	0.00	0.00	0.01
5500–5750	2670	0	0.00	0.00	0.00
5750–6000	2670	0	0.00	0.00	0.00

The map in *Fig. 3* was based on the density function derived from well-log data (Table II). The main features of the corrected map are the conspicuous positive anomalies above the deepest sub-basins of the Pannonian basin. This phenomenon can be explained by the thinning of the crust and the upwelling of the dense mantle material. This supports the extensional origin of these basins.

A similar effect was reported by Canadian scientists [LOWE–DEHLER 1995] beneath the Queen Charlotte basin where up to 6 km marine and non-marine sedimentary rocks with interbedded volcanic material were deposited on a Mesozoic basement. Another feature of the map worth mentioning is that the level of anomalies in the south-eastern part of Hungary is higher by about 15 mGal than in the other parts.

5. Gravity lineament map

To emphasise the structural lines reflected in the corrected gravity map, the maximum gradient method was applied for the anomalies interpolated to a 1 km grid network. The horizontal gradients were calculated for each grid point. Those grid points which have higher gradient values than a preselected limit (in our case 2.5 E), compared to the neighbouring grid points, are marked on the map (*Fig. 4*). Most of the markings form lineaments. If it is supposed that lateral density contrasts are related to structural features then most of the lineaments outlined in the map reflect tectonic lines. The exceptions are those erosional structures where the effect of tectonic processes is negligible.

Given that our initial assumptions and data are correct the lineaments reflect the structures of the crust including the basement.

Table III. Relative weight of 250 m thick consecutive layers as a function of depth (h). Density function ($\sigma(h)$) from laboratory measurements, W=percentage of total effect for a basin of 1— H=500 m, 2 — H=2500 m, 3 — H=5000 m depth

III. táblázat. A 250 m vastag rétegek relatív súlya (W) a mélység (h) függvényében. Sűrűségfüggvény ($\sigma(h)$) laboratóriumi mérésekből. W a teljes hatás százalékában van megadva, 1 — H=500 m, 2 — H=2500 m, 3 — H=5000 m mély medencére

6. Contour map of the Moho based on gravity data

Correlation studies were carried out between the corrected gravity map and the Moho surface supposing that the corrected gravity map reflects mainly the undulations of the Moho, and that the effect of possible density inhomogeneities of the basement is negligible. The contours of the Hungarian part of the Moho are relatively well known. The seismic crustal investigations started in 1954. The first experimental measurements proved that the crust under the Pannonian basin is considerably thinner than the continental crust elsewhere [GÁLFI, STEGENA 1955]. Since that time many seismic profiles have been recorded to study the Moho; there is, however, variation in the quality of the sections mainly because of the long time span from the beginning. A synthesis of the deep seismic measurements was provided in 1991 by the publication of a map of the Mohorovičić discontinuity beneath Central Europe [POSGAY et al. 1991]. Our correlation studies were based partly on that contour map and partly on some new deep reflection seismic sections observed after the compilation of the map. The locations of the seismic profiles are presented as an insert in *Fig. 5*. The depth values and the corresponding gravity anomalies were read out from the two maps along seismic lines. The two data sets were correlated by linear regression. Having obtained the coefficients of the correlation function, the contours of Moho were determined from the gravity data. The resulting map is presented in Fig. 5.

It can be seen on the map that the depth of the Moho is between 23 and 32 km. The highest values are obtained in those regions where the basement is outcropping or lies at a shallow depth whereas the lowest values can be found over deep basins. This phenomenon indicates the striving for isostatic equilibrium. Generally speaking the south-eastern part of the country (the Great Hungarian Plain) — especially the Békés basin, the Dorozsma graben, the Danube–Rába Lowland and the Zala basin — is characterized by thin crust (< 30 km). The Mecsek Mts, the Transdanubian Central Range, and the Aggtelek–Rudabánya Hills have relatively thick crust (> 30 km). The highest discrepancies (5 km) between the seismic and gravity-derived maps can be found below the Mecsek Mts and the Dorozsma graben. It is interesting to note that in spite of the large depth of the

Makó trough, the crust below it is relatively thick suggesting a different structural evolution.

The Moho map derived from gravity data has many similarities with that derived from seismic data, but there are discrepancies as well. These originate from the essence of the two methods, viz. the better resolution but sparse net of the seismic profiles, and the smoothing effect of gravity. The main advantages of the latter are the good coverage and the elimination of local disturbing effects which are present in the seismic sections.

REFERENCES

BIELIK M. 1988: A preliminary stripped gravity map of the Pannonian Basin. Physics of the Earth and Planetary Interiors, **51**, pp. 185–189

GÁLFI J., STEGENA L. 1955: Deep reflections in NE Hungary (in Hungarian). Geophysical Transactions, **4**, 2, pp. 7–15

HAÁZ I. B. 1953: Relations between the potential of the attraction of the mass contained in a finite rectangular prism and its first and second derivatives (in Hungarian). Geophysical Transactions, **2**, 7, pp. 57–66

HAMMER S. 1963: Deep gravity interpretation by stripping. Geophysics, **28**, 3, pp. 369–378

KILÉNYI É. 1968: Geological-Geophysical considerations on the Great Hungarian Plain through a statistical analysis of density data (in Hungarian). Geophysical Transactions, **17**, 4, pp. 41–49

KILÉNYI É., RUMPLER J. 1984: Pre-Tertiary basement relief map of Hungary. Geophysical Transactions, **30**, 4, pp. 425–428

KILÉNYI É., KRÖLL A., OBERNAUER D., ŠEFARA J., STEINHAUSER P., SZABÓ Z.,WESSELY G. 1991: Pre-Tertiary basement contour map of the Carpathian Basin beneath Austria, Czechoslovakia and Hungary. Geophysical Transactions, **36**, 1–2, pp. 7–15

KOVÁCSVÖLGYI S. 1996: Density determination of near-surface formations based on gravity data — a case history (in Hungarian). Magyar Geofizika **37**, 1, pp. 38–45

LOWE C., DEHLER S.A. 1995: Crustal thickness beneath the Queen Charlotte Basin, Canada: Results of a seismically constrained gravity inversion. Journal of Geophysical Research, **100**, B12, pp. 24, 331–24, 345

MESKÓ A. 1984: Regional geophysical investigation of Hungary. A corrected gravity anomaly map of Hungary (in Hungarian). ELTE. Manuscript

PINTÉR A., ÁDÁM O., SZÉNÁS GY. 1964: Regional gravity interpretation in the Hungarian basin (in Hungarian). Geophysical Transactions, **13**, 3, pp. 316–328

POSGAY K., ALBU I., MAYEROVÁ M., NAKLÁDALOVÁ Z., IBRMAJER J, BLIŽKOVSKY M., ARIC K.,GUTDEUTSCH R. 1991: Contour map of the Mohorovičić discontinuity beneath Central Europe. Geophysical Transactions, **36**, 1–2, pp. 7–15

RENNER J., STEGENA L. 1965: Gravity investigation of the deep structure of Hungary (in Hungarian). Geophysical Transactions, **14**, pp. 103–114

ŠEFARA J. (editor) 1987: Struktúrno-Tektonicka Mapa: Vnútornych Západnych Karpát pre účely prognózovania lozísk geofyzikalne interpretácie (Structural and tectonic map of the inner West Carpathians for mineral exploration). Slovensky Geologicky Urád, Bratislava, 267 p.

SZABÓ Z., PÁNCSICS Z. 1999: Bouguer anomaly map of Hungary corrected with variable density. Geophysical Transactions, this issue

SZABÓ Z., PÁNCSICS Z. 1999: Rock densitites in the Pannonian basin. Geophysical Transactions, this issue

Magyarország medencehatástól mentesített gravitációs térképe

SZABÓ Zoltán és PÁNCSICS Zoltán

Magyarország területének 70 %-át fiatal üledékek fedik. Az üledékek vastagsága elérheti a 7000–8000 métert, ennek következtében az üledékek okozta tömeghiány hatása nagymértékben befolyásolja a Bouguer-anomália térképet. Az üledékek “torzító” hatásának meghatározására a szerzők modellszámításokat végeztek. Az üledékhatás meghatározása után megszerkesztették a korrigált gravitációs térképet, amely már csak a kéregben levő sűrűséganomáliák és a kéregvastagság hatását tükrözik. A szerkezeti vonalak kihangsúlyozására a szerzők megszerkesztették a korrigált térkép gravitációs lineamens változatát is. Korábbi szeizmikus adatok felhasználásával korrelációs számításokat végeztek a szeizmikus adatokból meghatározott Moho mélységek és a korrigált gravitációs adatok között. A kapott korrelációs együtthatók segítségével meghatározták a Moho szintvonalas térképét.

ABOUT THE AUTHORS

Zoltán Szabó, for a photogprah and biography, see this issue, p. 27.

Zoltán Páncsics, for a photogprah and biography, see this issue, p. 27.

GEOPHYSICAL TRANSACTIONS 1999
Vol. 42. No. 1 – 2. pp. 55 – 65

Horizontal gravity gradients and earthquake distribution in Hungary

Zoltán SZABÓ* and Zoltán PÁNCSICS*

The distribution of the maximum horizontal gravity gradients shows very often a linear pattern. If the topographic features of the basement were formed by horizontal and/or vertical block movements then this linear pattern marks faults. The structural lineaments based on geophysical data are, however, insufficient for making a proper judgement about the seismic hazard of a region: seismological data are indispensable. With the help of structural lineaments a relationship can be found between scattered earthquake epicentres, individual source regions can be delineated, and their relationships clarified. To study these relationships, a gravity lineament map of Hungary was constructed, based on the Bouguer anomaly map, and applying the maximum gradient method. The gradients were corrected for basin effect. This lineament map was correlated with the distribution of historical earthquakes and the earthquakes registered between 1995 and 1998 by the Microseismic Monitoring Network (MMN) of Hungary's Paks Nuclear Power Plant. Former investigations had indicated a correlation between high gradients and the epicentres of historical earthquakes. An even more striking correlation was found between the epicentre distribution of earthquakes recorded by the MMN and belts of high horizontal gravity gradients.

Keywords: gravity surveys, lineaments, earthquakes, Microseismic Monitoring Network, Hungary

* Eötvös Loránd Geophysical Institute of Hungary, H-1145 Budapest, Kolumbusz u. 17–23.

Manuscript (updated version) received: 31 January, 1999.

1. Introduction

Earthquakes originate as a consequence of stress and strain in the lithosphere. The source of these stresses is the movement of continental and oceanic plates, and their collision and subduction. High-magnitude lithospheric earthquakes concentrate in such collision zones, the so-called active plate margins (e.g. the western coast of America, Japan, etc.). Hungary, however, is far from any active plate margins. Her earthquakes with low-magnitude ($M_0 \leq 6$) and shallow focal depths (≈ 10 km), meaning they originate within the basement, belong to the so-called intra-plate category.

The composition and structure of the basement of the Pannonian basin is extremely complex, consisting of different plate fragments drifting and accreting during the late Mesozoic and early Tertiary. It is a difficult task to determine which of the different structural lines are linked with earthquakes: the intra-fragment ones, or those separating these fragments, or the faults transversing the present basement. Maybe this is why some earth scientists believe that there is no connection between the structural system of Hungary and its earthquake distribution. The truth is that even now the tectonism of Hungary is not known to the extent that an unambiguous relationship could be found between structure and earthquake occurrence.

Another difficulty in studying the correlation between earthquake distribution and geological structure is that there is no instrumental earthquake recording dating back further than the beginning of the 20th century. The records of historical earthquakes based on building damage can be regarded more-or-less exact since the time of the Komárom earthquake of 1763. Because of the moderate seismicity of Hungary this not much more than 200-year old history of earthquakes means that there is insufficient evidence to enable the exact regularity of the seismicity of Hungary to be identified.

In this respect, Hungary is not unique, as the eastern regions of the USA have similar problems: moderate seismicity, low-magnitude earthquakes not causing faults on the surface, etc. The increasing volume of data, however, confirms that generally speaking seismicity concentrates in the reactivated zones of Palaeozoic and Mesozoic deformations [HILL 1987]. In regions characterized by moderate and scattered earthquakes, the seismic hazard of an area can be assessed only by the seismologists' accumulated experience.

2. Analysis of data

Earthquake data for our analysis were taken from the catalogue compiled and updated by a team of the Geodetic and Geophysical Research Institute of the Hungarian Academy of Sciences [ZSÍROS et al. 1988]. This catalogue contains some 5000 items of data going as far back as 456 AD. The relative frequency of historical earthquakes for 20-year intervals are presented in *Fig. 1*. It can be seen that the number of recorded events grew

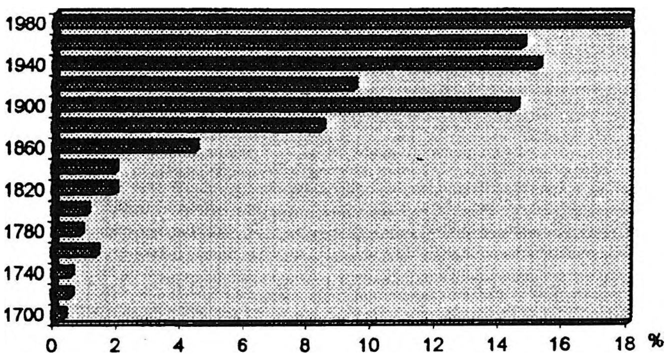


Fig. 1. Relative frequency of historical earthquakes for 20-year intervals
1. ábra. A történelmi rengések 20-éves idő intervallumokra számított relatív gyakorisága

considerably after 1860 which, however, does not mean that the seismic activity of Hungary increased from that time, but refers to the uncertainty concerning the characteristics of historical earthquakes. Before instrumental recording, an earthquake had to fulfil the following criteria to get recorded: population to observe the quake and regard the observed event to be so important that they would report it to the authorities who would then record it. Taking into account the above criteria it can be concluded that only a small proportion of the historical earthquakes were registered. It is evident that in a more advanced society more earthquakes are recorded (e.g. literacy is an essential factor). Since the beginning of the 20th century, owing to the starting of regular instrumental registration, the number of recorded earthquakes has been increasing significantly. The ever growing sensitivity of seismographs is also leading to more and more registered events.

There is a similar or even greater uncertainty in the locations of epicentres. Historical earthquakes are generally designated as being linked to the nearest large settlement, but it does not mean that the highest relative damage occurred in the given settlement. The quality of the buildings should also be taken into consideration. There is also no guarantee that the epicentre was near to a settlement. We can take into account as well that the effect of an earthquake is much less frightening in the country than in a town. Based on the above considerations it is evident that the reliability of epicentre determination depends on the spatial distribution of settlements.

The compilers of the above-mentioned catalogue tried to categorize the earthquakes according to the reliability of epicentre location. The relative distribution of the different categories is presented in *Fig. 2*. It can be

a	b	c
A	±5	5%
B	±10	35%
C	±20	33%
D	±50	5%
E	u.	22%

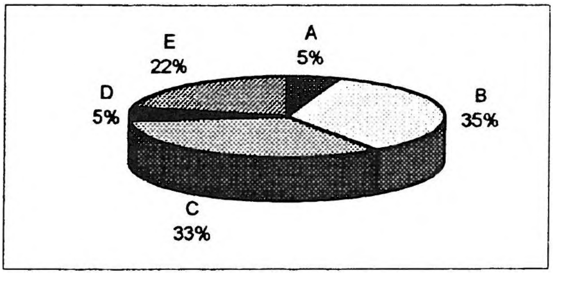


Fig. 2. Categorization and relative distribution of historical earthquakes based on the reliability of their epicentre location

a — category, b — accuracy of epicentre determination in km, c — relative distribution

2. ábra. A történelmi rengések kategorizálása az epicentrum meghatározási pontossága szerint és ezen kategóriák relatív gyakorisága

a — kategória, b — az epicentrum meghatározási pontossága km-ben, c — relatív gyakoriság seen that the uncertainty is rather high. That phenomenon has to be taken into consideration if we try to find an empirical correlation between geological structure and earthquake distribution, otherwise we may reach wrong conclusions. To avoid large errors we included into our investigations only those quakes the locations of which are categorized as A and B; in other words their epicentre locations are known better than ±10 km. Only 40% of the historical earthquakes fell in these categories.

Bouguer anomalies reflect the integrated effect of subsurface masses, thus they contain all information which can be deduced from mass effects. In basin areas, where the high-density basement is overlain by young, loose sediments, the topographical changes of the basement present themselves in a horizontal plane as density contrasts. If the topographic features of the basement were formed by horizontal and/or vertical block movements then the lineaments mark faults. Above a fault, where an abrupt density change exists, the horizontal gradient has a local maximum. This phenomenon is utilized in determining structural lineaments. The maximum gradients show very often a linear pattern. These gravity lineaments can be used for detecting structural lines taking part in the shaping of basement topography. Although they do not provide any direct information on the age of these structural lines, they are very useful pointers in seismotectonic studies.

3. Correlation

For studying the correlation between gravity lineaments seen in the map and the earthquake distribution we plotted all historical earthquakes whose epicentres are known to an accuracy at least ±10 km. The radii of the circles are proportional to the respective accuracy in the scale of the map (*Fig. 3*).

From this map we can conclude that at some places epicentres are concentrated along lineaments belonging to known structural lines (e.g. Kapos line, the south-western and the north-central part (Jász region) of the Central Hungarian dislocation zone). At other places the linear pattern can be recognized in the epicentre distribution (e.g. Komárom–Berhida, Rajka–Zirc) but one cannot see their equivalent in the gravity lineament map nor in the known geological structure. As gravity lineaments indicate the locations of sharp density contrasts, if earthquakes occur along such lineaments their cause must be attributed to some sort of fault even if it is not yet identified. Those earthquakes, however, which show a linear pattern but do not coincide with any gravity lineament are caused by such stress zones which are not linked with linear density contrast, proving that stress zones do not necessarily occur along geological boundaries.

As the epicentre determination of historical earthquakes is unreliable, in the next step we decided to use those data which were registered by the Microseismic Monitoring Network (MMN) of Paks Nuclear Power Plant, in 1995–98. The siting and the sensitivity of the recording seismographs allow us to recognize earthquakes of magnitude 1–2.5, and to determine the hypocentre with an accuracy of 1–5 km [TÓTH et al. 1996, 1997, 1998]. A preliminary analysis for 1995–96 has already been published [SZABÓ–PÁNCSICS 1997]. The reliability of this data set is comparable to that of the lineaments, thus their relationship is worth studying. The lineament map computed from the Bouguer anomalies and corrected for basin effect [SZABÓ–PÁNCSICS, a separate paper in this volume] seemed the best for comparison; in this map we plotted the epicentres of earthquakes recorded by the MMN, in 1995–98 (*Fig. 4*). *Figure 5* contains all locations and names used in the analysis below.

Moving from west to east on this map, we find the following:

— The earthquakes in the north-western corner of the map, on Austrian territory, confirm the activity of the Muhr–Mürz line, determined formerly by historical earthquakes.

— The earthquakes near Veszprémvarsány and Lébény lie in the Rajka–Zirc earthquake zone — delineated by historical earthquakes — along gravity lineaments intersecting perpendicularly that zone. This Rajka–Zirc zone has up till now evaded attention because it consists of small quakes ($I_0 \leq 5$ MSK-64) and its direction is nearly perpendicular to the known structural lines. Nevertheless its existence is supported by the latest results of the DANREG project [NEMESI et al. 1997]. One of the most outstanding results of its geophysical investigations was the detection by magnetotellurics of the Rába–Hurbanovo–Diósjenő dislocation zone.

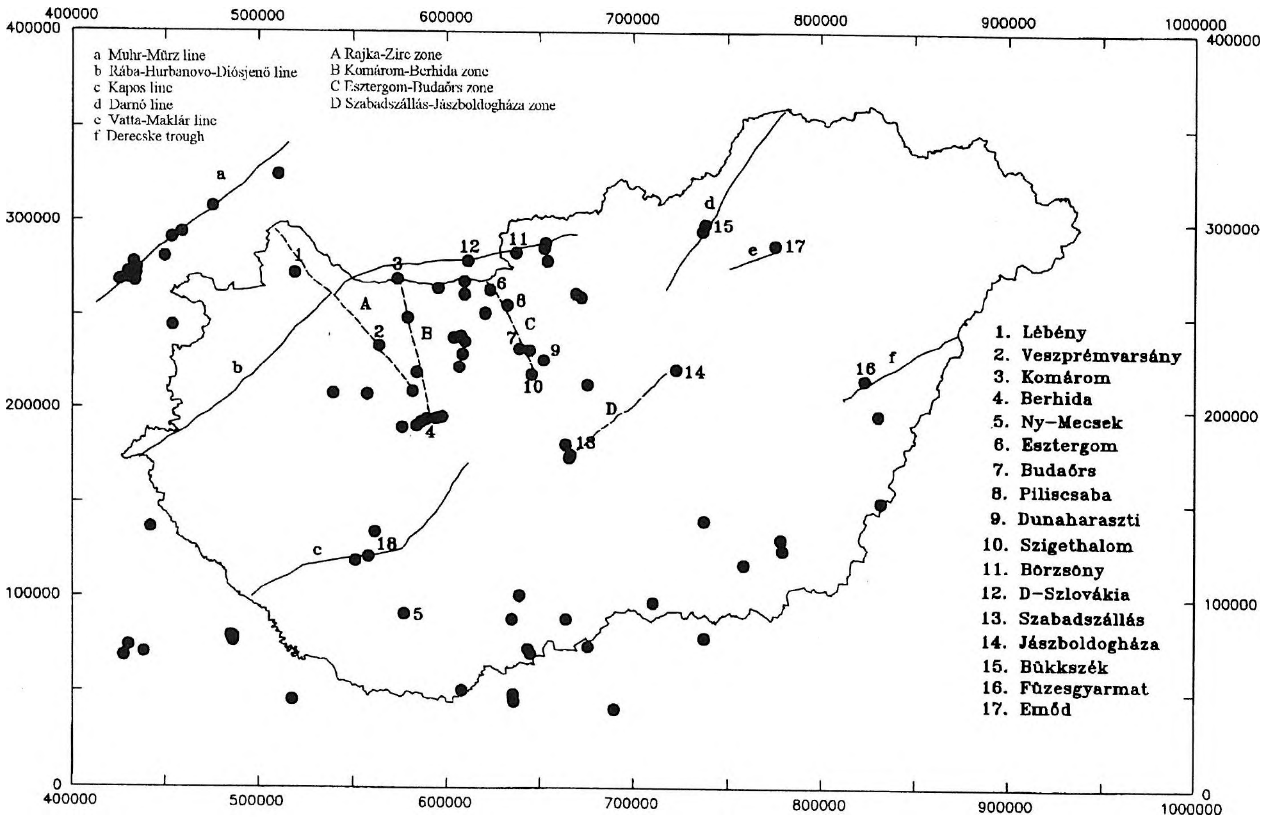


Fig. 5. Map showing the epicentres recorded by MMN from 1995 to 1998 and the structural lines mentioned in the text

5. ábra. Mikroszeizmológia Megfigyelő Hálózat által 1995–1998-ban regisztrált földrengések epicentrumai és a szövegben említett szerkezeti vonalak helyszínrajza

The magnetotelluric curves recorded on the N, and NW side of this zone sharply differ from those recorded on the S, and SE side, thus reflecting different geology. Based on the magnetotelluric survey, the tectonic zone can be located with high reliability. From our point of view it is of outstanding importance that in the Rába line, 5 km to the southwest from Győr, a strike-slip fault of about 6–7 km shift can be recognized which coincides with a part of the Rajka–Zirc earthquake zone.

— The earthquakes in the area of Berhida mark the activity of the southern termination of the Komárom–Berhida zone which is regarded as the most active seismic zone of Hungary. The gravity lineaments reflecting this zone are not very distinctive; they can be traced only intermittently.

— The focal depth of the earthquake in the western Mecsek Mts. is 1 km, thus it is most probably an event linked with the mining activity there.

— The two earthquakes at the two opposite ends of the Esztergom–Budaörs lineament — limiting the outcropping Mesozoic blocks of the Pilis–Buda hills in the west — are worthy of attention. The odd thing about it is that both ends set off on the same day, on 4 May 1995, calling attention to its activity. The earthquake of Piliscsaba in 1997, supports the activity of this zone.

— The two earthquakes at Dunaharaszti and Szigethalom are indications that the area is still under stress after the big earthquake of 1956. Another possibility, is that they are related to the Esztergom–Budaörs line.

— The earthquakes of the Börzsöny Mts. and of South Slovakia are most probably connected with the Hurbanovo–Diósjenő line.

— The earthquakes at Szabadszállás–Jászboldogháza occurred along a definite lineament of NE–SW direction, indicating its activity. Historical earthquakes suggest the continuation of this lineament further to the NE from Jászboldogháza.

— The earthquakes near Bükkszék refer to the activity of the Darnó line. The Darnó line is manifested by a strong gravity lineament, too.

— The 7 earthquakes occurring within two days in the area of Füzesgyarmat are marked with the mean value of their co-ordinates, because of their big errors. They coincide with the junction of two definite lineaments and mark, most probably, the activity of the Derecske trough.

— The earthquake in the neighbourhood of Emőd marks the activity of the Vatta–Maklár trough, well known from seismics.

— The two earthquakes along the Kapos line support the activity of this well-known tectonic line.

4. Relationship between earthquake distribution and gravity gradients

The geological-geophysical investigation of epicentral areas of historical earthquakes of $I_0 \geq 6°$ (MSK-64) intensity led to the finding of a correlation between the horizontal gradients of the gravity field and earthquake distribution [SZABÓ Z. 1990]. It seemed advisable to carry out a similar investigation for the earthquakes registered by the Microseismic Monitoring Network. In *Fig. 6* we have plotted those areas where horizontal gradients, computed from Bouguer anomalies, exceed 15 E (1.5 mGal/km), as well as the earthquake epicentres. We have found that 84% of earthquake epicentres coincide with high gradient belts. The importance of this correlation increases if we take into consideration that these high-gradient zones represent only 24 % of the area of Hungary.

5. Closing remarks

Because of the high number of gravity lineaments the probability of an earthquake coinciding with one of them is 50 %. In spite of this real problem we consider that the lineaments with their definite characteristics provide a sound basis for tracing the active zones. Good examples for this are the Esztergom–Budaörs and the Szabadszállás–Jászboldogháza lineaments discussed above. The earthquakes in both cases coincide with definite lineaments. The high correlation between high-gradient belts and epicentre distribution supports this statement.

Despite there being a degree of uncertainty in our findings, it should be taken into consideration that in areas of low to medium activity the possibility of identifying a correlation between seismic zones and geological structure is a world-wide problem. Doubtless the correlation of gravity lineaments and distribution of earthquakes has provided encouraging results in delineating active zones. The earthquakes to be registered in the future will increase the reliability of this method.

REFERENCES

HILL D. P. V. 1987: Seismotectonics. Reviews of Geophysics **25**, 6, pp. 1139–1148

NEMESI L., ŠEFARA J., VARGA G., KOVÁCSVÖLGYI S. 1997: Results of deep geophysical survey within the framework of the DANREG project. Geophysical Transactions **41**, 3–4, pp. 143–159

SZABÓ Z. 1990: Geological–geophysical parameters of epicentres of earthquakes of $I_0 \geq 6°$ (MSK-64) intensity of Hungary (in Hungarian). Unpublished report of ELGI

SZABÓ Z., PÁNCSICS Z. 1997: Gravity lineaments and earthquake distribution in Hungary (in Hungarian). Magyar Geofizika **38**, 1, pp. 16–24

TÓTH L., MÓNUS P., ZSÍROS T. 1996: Hungarian Earthquake Bulletin 1995. Budapest, Georisk

TÓTH L., MÓNUS P., ZSÍROS T. 1997: Hungarian Earthquake Bulletin 1996. Budapest, Georisk

TÓTH L., MÓNUS P., ZSÍROS T. 1998: Hungarian Earthquake Bulletin 1997. Budapest, Georisk

TÓTH L., MÓNUS P., ZSÍROS T. 1999: Hungarian Earthquake Bulletin 1998. Budapest, Georisk

ZSÍROS T., MÓNUS P., TÓTH L. 1988: Hungarian Earthquake Catalog (456–1986). Budapest, MTA GGKI

Horizontális gravitációs gradiensek és a földrengéseloszlás kapcsolata Magyarországon

SZABÓ Zoltán és PÁNCSICS Zoltán

A horizontális gravitációs gradiensek maximális értékei igen gyakran vonalas elrendeződést mutatnak. Amennyiben a medencealjzat domborzatát törések menti kiemelkedések, süllyedések

illetve elmozdulások alakították, akkor a gradiensek alapján kijelölhető lineamensek is a törések hatását tükrözik. A geofizikai adatokból szerkesztett lineamensek azonban önmagukban nem elegendők egy-egy terület földrengés veszélyességének megítéléséhez, ehhez szükség van szeizmológiai ismeretekre is. A lineamensek segítségével kapcsolatot találhatunk az egyes, látszólag szórtan elhelyezkedő földrengés epicentrumok között, ezáltal lehetővé téve az egyes forrásterületek körülhatárolását és a különböző forrásterületek közötti összefüggések tisztázását. A Bouguer-anomália térkép alapján, a maximális gradiens módszer segítségével, kijelöltük a gradienseket, melyeket javítottunk a medenceüledékek hatásával, majd megszerkesztettük a lineamens térképet. A lineamensek számos helyen korrelálnak a történelmi földrengéseloszlással, illetve a Paksi Atomerőmű Mikroszeizmológiai Megfigyelő Hálózata (MMH) által 1995–98-ban regisztrált földrengésekkel.

ABOUT THE AUTHORS

Zoltán Szabó, for a photogprah and biography, see this issue, p. 27.

Zoltán Páncsics, for a photogprah and biography, see this issue, p. 27.

GEOPHYSICAL TRANSACTIONS 1999
Vol. 42. No. 1 – 2. pp. 67– 81

Effect of vertical gravity gradient on the accuracy of gravimeter measurements based on Hungarian data

Géza CSAPÓ*

In the last decade significant developments have taken place in gravimetry as a result of the world-wide use of absolute gravimeters (at present about 25 items of equipment are in use). Their accuracy is 3–4 μGal and the results are obtained directly in SI units (ms^{-2}). With the application of absolute gravimeters standardized gravity networks for large regions can be established economically. The political changes of the 1990s in Eastern Europe have made it possible to extend over the eastern part of the continent the Unified European Gravity Network (UEGN) established earlier in Western Europe. The gravity datum of the network is based on absolute stations. The densification of the network between the absolute points is carried out with LaCoste–Romberg (LCR) gravimeters. The increased accuracy of the measurements made it necessary to investigate the effect of the applied normal vertical gradient (0.3086 mGal/m) for elevation reductions. The author presents the results of his investigations in this field. High precision gravity measurements (calibration lines, microgravimetric measurements) require the determination of the actual vertical gradient (VG) because the application of its normal value would lead to errors higher than the accuracy of the measurement. The gravity value measured with absolute gravimeters refers to a certain height (depending on the type of instrument) above ground level. The gravity value obtained is reduced to the benchmark of the station by means of relative gravimeter measurements. Based on the author's experience the accuracy of this reduction — when the vertical gradient is measured with 1 or 2 gravimeters in one observation series only — does not reach the necessary ± 10–20 E units.

Keywords: vertical gradient, gravimeters, calibration, Hungary

* Eötvös Loránd Geophysical Institute of Hungary, H-1145 Budapest, Kolumbusz u. 17–23.
Manuscript received: 17 August, 1998.

1. Introduction

The value of gravity (g) determined by absolute or relative gravimeters refers to the height of their sensing mass; it is called 'reference height'. To make them comparable to each other, the measured value should be reduced to a common reference point generally a benchmark on the ground level. The reduction is the 'reference height correction', which depends on the vertical gradient and the reference height. This latter one means the vertical distance between the sensing mass of the applied instrument and the benchmark.

2. Basic principles

Reference height correction (mentioned above) can be determined if one knows the local **vertical gradient** (*VG*) value at the observation site. The vertical gradient is the vertical derivative of gravity ($\partial g/\partial h$), i.e. the change in gravity acceleration referred to an infinitesimally small vertical distance.

The vertical gradient's theoretical value can be deduced from Faye reduction. Using certain assumptions (the equipotential surface corresponding to sea level is a sphere of $R_0 = 6372$ km radius, and the mass distribution in the Earth is homogeneous), acceleration due to gravity at sea level is:

$$g_0 = k^2M/R_0^2$$

where k^2 is the gravity constant and M is the mass of the Earth.

At a height 'h' above sea level it is given by the relation

$$g = k^2M/(R_0+h)^2$$

In this case — after series expansion of the expression $1/(R+h)^2$ — the reduction value (Δg_m) is:

$$\Delta g_m = g_0 - g = k^2M/R^2(2h/R_0 - 3h^2/R_0^2 + \dots)$$

Because 'h' is small compared with R_o, it can be further simplified

$$\Delta g_m = -0.3086 \text{ mGal/m}^{**}$$

Although the Faye reduction and the reference height correction are not identical, in both cases the observation point is 'shifted' vertically in free air.

In those tasks where the required accuracy of relative gravity field measurements (hereafter: gravity measurements) is a few tenths or hundredths of a mGal, this theoretical value can also be used for calculating reference height correction.

The vertical derivative of gravity can be determined by measurement as well. This has two versions: one of them is called direct, the other one indirect determination.

For **direct determination** vertical gradiometers were constructed. In actual fact, these devices have failed to fulfil the expectations that they would be able to determine the value of vertical gradients with an accuracy similar to that of horizontal ones. As was pointed out by YUZEPOVICH and OGORODOVA [1980], the limits are basically set by the instruments themselves. In practice, therefore, indirect determination is the commonly used method (in Hungary, indirect determination is the only method used).

Relative gravimeters are the tools for **indirect determination**. Here, it might be appropriate to emphasize the following:

— Relative gravimeters measure the difference in gravity between two points;

— The resolution of relative gravimeters does not allow any measurable change in gravity to be detected between two points with infinitesimally small separation;

— Because of the resolution constraints, the vertical gradient (defined earlier) is considered as the difference in gravity between two points separated by a distance of 1 m referred to the midpoint of the perpendicular line section between the two points.

Because the local perpendicular line is always a space curve, the vertical gradient belonging to a point on the Earth's surface changes depending

** 1 mGal = $1 \cdot 10^{-5}$ ms^{-2}, 1 μGal = $1 \cdot 10^{-8}$ ms^{-2}

on the position of the selected section to be measured along the given local perpendicular line.

Reference is important because experience has shown that, on the one hand, values of *VG* change in different sections of the perpendicular line at the given site and, on the other hand, the change is non linear [SZABÓ, CSAPÓ 1985].

Because the reference height for absolute gravimeters slightly changes even within one observation set, a reference height correction should always be applied referring to the height of the given drop [CHARLES, HIPKIN 1995]. The problem is made more complicated by the presence of the observation pillar and nearby masses in the laboratories (where the observations are made) and these cause various inhomogeneities in the gravity field [SAGITOV 1984]. (I would call attention here to the calculations performed in the Central Institute of Earth Physics in Potsdam [ELSTNER et al. 1986] as an example of how these effects can be taken into account). In order to make the observation results independent of the uncertainties resulting from the reduction to the benchmark, in the literature generally two *g* values, one for the reference level and one for the benchmark are given. The reliability of measurement refers to the value obtained on the reference level.

Mention should be made of the basic contradiction of gravimetry with regard to the reliability of measurements; the reliability is lower by an order of magnitude for vertical gradients measured in any way than that of horizontal gradients, the latter can be determined with a reliability of about 1E (0.1 μGal/m) using an Eötvös torsion balance.

The measurement accuracy for the new generation of absolute gravimeters (US made AXIS–F5 gravimeters) is 3–4 μGal [MARSON et al. 1995].

As a result of recent developments of the most up-to-date relative gravimeters (LaCoste–Romberg types, Scintrex CG–3) application of electronic levels, double thermostats, and feedback electronics, similar accuracy can be achieved under laboratory conditions.

The increasing number of absolute gravimeters, and the widening of international collaboration provide a ‘common mGal unit’ for gravity base networks for all over the world. In addition to other scientific and economic goals, the linking of these networks makes it possible to construct unified geophysical maps for large, contiguous areas [BOEDECKER et al. 1995].

Because of the improved measurement accuracy the theoretical value of free air reduction — 0.3086 mGal/m — does not provide satisfactory accuracy for the reduction of gravity from the reference height to the benchmark. The vertical gradient must be determined at each absolute station [CSAPÓ 1987]. Similarly, more accurate correction is required if the highest possible reliability is expected in surveys performed with relative gravimeters (establishment of calibration lines, laboratory or microgravimetric measurements).

In what follows the results of *VG* measurements carried out at the stations of the national calibration line, along the horizontal microbase established in the Mátyáshegyi Cave and at the other absolute stations in Hungary are presented. We want to show, on the one hand, possible deviations in *VG* values from the theoretical value, and, on the other hand, to provide an idea on the most cost-effective achievable reliability of *VG* measurements performed under different ambient conditions. Moreover, to show what kind of error might be caused by inaccurate knowledge of the *VG* value in gravity measurements.

3. *VG* measurements and processing methods

Based on domestic and international experience, measurements have been carried out at the absolute stations established in buildings and at field sites in accordance with identical principles [CSAPÓ 1997]. At the site to be measured two tripods are set up one above the other. The height of the higher tripod is 110 cm (P_u) and that of the lower one is 6 cm (P_l). The gravimeter type equipped with electronic level and electronic output can be set up on the tripods applying forced centring; the maximum linear eccentricity is 10 mm. After setting up the instrument the distance between the upper plate of the instrument (black lid) and the tripod is measured with mm accuracy. The vertical distance between the upper plate of the instrument and the beam is known. Three readings are taken at each level and in each series in the following order:

$$P_l - P_u - P_l - P_u - P_l - P_u$$

The null position is set by means of a digital voltmeter equipped with filter and connected to the output of the gravimeter. The reading is taken by means of the nulling dial, using the so called 'interpolation method'.

In this way five (not independent of each other) Δg values can be determined in each series; the value of *VG* in Eötvös units obtained from the series is the arithmetic mean of the corrected values interpolated to a height difference of 1000 mm. The observation results are then processed using our own (ELGI) software, applying corrections for air pressure (DIN 5450/1968.), earth tide, and instrumental drift.

The number of observation series and gravimeters is chosen so that the reliability of the measured *VG* value will be at least 30E in the case of field stations, and at least 20E in buildings.

4. Description of observation results

4.1. VG measurements performed along the national gravity calibration line

The stations of the national calibration line are generally in churchyards (some metres away from the church itself) or at airports, they are permanently marked with concrete blocks buried at ground level, and they have a benchmark. The gravimeter set was transported by car to the sites just before starting the observation. At each site, observations started after carrying out the daily instrument check — measuring a complete series with each gravimeter, one after another. The gravimeter set was selected from LCR Model G meters of the Eötvös Loránd Geophysical Institute of Hungary (G–type meters: No–220, No–821, No–963, No–1919). The time required to measure one series was generally 50–60 minutes per instrument. *Tables I* and *II* summarize the results of the VG measurements; all observations at the Pécs calibration site are listed (Table II) as an example.

name of point	location	H (m)	n_S	n_{GR}	VG (E)	m_i (E)	m_x (E)
Pécs	airport	200	8	4	3180	86	30
Mecseknádasd	church-yard	194	8	4	2960	80	28
Tolna	church-yard	100	7	4	3107	100	38
Dunaújváros	airport	122	12	3	3087	34	10
Ercsi	church-yard	124	12	3	3093	73	21
Budaörs	airport	126	13	4	3082	98	27
Mátyáshegy	quarry	201	9	3	2625	34	11
Dunakeszi	church-yard	126	6	2	3079	51	23
Rétság	church-yard	193	7	4	3028	48	18
Balassagyarmat	park	147	8	4	3208	88	31
Δ_{VG} = 583 E				mean:	3045	69	24

Table I. Results of vertical gradient (*VG*) measurements on the calibration line of Hungary
Legend: H — elevation above sea level; n_S— number of determinations (measurement series); n_{GR}—number of instruments in the given gravimeter set; m_i— r.m.s. error of one determination in E units: $m_i = \pm (\Sigma vv/n-1)^{½}$; m_x— r.m.s. error of the most probable error after adjustment: $m_x = \pm [\Sigma vv/n(n-1)]^{½}$; Δ_{VG} — $VG_{max}-VG_{min}$

I. táblázat. A magyarországi graviméter-kalibráló alapvonalon végzett *VG* mérések eredményei
Jelmagyarázat: H — a pont tengerszint feletti magassága; n_S— mérési sorozatok száma; n_{GR}—az alkalmazott graviméter csoportban szereplő műszerek száma; m_i— egy *VG* mérés középhibája eötvös egységben $m_i = \pm(\Sigma vv/n-1)^{½}$; m_x— a *VG* legvalószínűbb értékének kiegyenlítés utáni középhibája: $m_x = \pm [\Sigma vv/n(n-1)]^{½}$; Δ_{VG} — $VG_{max}-VG_{min}$

n	LCR-220 G	LCR-821 G	LCR-963 G	LCR-1919 G
1	3193	3095	3160	3308
2	3162	3079	3138	3306
		VG = 3180 E	30 E	

Table II. Results (in Eötvös units) of vertical gradient (*VG*) measurements at the Pécs point of the calibration line

II. táblázat. A kalibráló alapvonal Pécs pontján mért vertikális gradiens (*VG*) értékek eötvös egységben

4.2. VG measurements along the horizontal microbase

The fourteen stations of the microbase are set up in the man-made entrance gallery of the Mátyáshegyi Cave in Budapest, on a concrete floor spread directly over the natural limestone base. The gallery of the cave has a vaulted shape: it is about 2.5 m wide and 2.5–4 m high. The distance between the stations is 2–5 m, their elevations above sea level are equal to within a few cm. The rock face rises steeply above them (the rock thickness above station 1 is some 30–40 m; above station 14, it is about 5–6 m). The daily change in temperature in the gallery is negligible, the seasonal temperature variation is ±1°C, the average temperature is 15°C. (During the campaign the gravimeters were stored at the site.) Because of the extremely high horizontal gradients (2000–10,000 E) caused by the steeply rising hillside the stations are designated with markers ensuring forced centring when setting up the instrument; this means that the mass of the meters is taken as that at a height of 115 mm ±5 mm above the marker at each station, and the horizontal eccentricity of the mass is less than 1 mm. The *VG* values obtained for the stations along the microbase are shown in *Table III.*.

4.3. VG measurements at absolute stations

The absolute gravity stations in Hungary are generally to be found at the lowest level of long-standing historic buildings (castles, mansions, etc.).

The *VG* values determined at the 15 absolute stations are listed in *Table IV*; at the Budapest station (in Mátyáshegyi Cave) all *VG* observations are shown in *Fig. 1*.

5. Discussion of measurement results

From the numerous conclusions that can be derived from a comparison of the data in Table I, attention is drawn to the following. It can be seen that

Number of station	n_S	n_{GR}	VG (E)	m_l (E)	m_x (E)
1	4	2	2581	26	13
2	5	2	2591	8	3
3	3	1	2578	22	13
4	6	2	2447	38	15
5	3	1	2386	19	11
6	7	3	2556	46	18
7	4	2	2432	47	23
8	10	3	2358	46	14
9	3	1	2286	15	8
10	11	3	2356	35	10
11	6	3	2283	57	23
12	4	1	2281	28	14
13	5	2	2236	27	12
14	6	2	2102	66	27
Δ_{VG} = 489 E		mean:	2391	34	15

Table III. Results of vertical gradient (*VG*) measurements on the horizontal micro calibration line in Mátyáshegyi Cave, Budapest. Legend: See Table I.
III. táblázat. A Mátyáshegyi-barlang horizontális mikrobázis pontjain végzett *VG* mérések eredményei. Jelmagyarázatot lásd az I. táblázatnál

there is no correlation between the *VG* values (all the values presented were observed for a height of 620 mm ± 20 mm above the ground) and the elevation of the stations above sea level in the given range. But there is a close correlation between the reference heights and the corresponding *VG* values at the same site [see CSAPÓ 1987]. When using the instrument set consisting of the four given gravimeters and equal number of measurement series (n_s = 8) the r.m.s error of the most probable value after adjustment is nearly the same in all cases, viz. about 30 E.

Results of a *VG* determination are given in Table II. The maximum difference between the 8 determinations is 229 E (≈23 μGal/m), the r.m.s error of the most probable value obtained from the adjustment of *VG*s is ±30 E (≈3 μGal/m). Deviations might be larger in measurements carried out under extreme ambient conditions (strong wind, enhanced microseismic effects at vibration sensitive sites caused by intensive road traffic,

Name and location of station		Locality	n_s	n_{GR}	VG (E)	m_x (E)
81	Siklós	castle	22	11	3407	16
82	Budapest	Mátyáshegyi Cave	47	14	2519	7
85	K õszeg	town hall	22	4	2661	24
86	Szerencs	wine house	40	3	2968	7
88	Nagyvázsony	mansion	18	5	2565	12
89	Gyula	castle	29	2	2913	11
90	Szécsény	mansion	15	5	3059	18
91	Kenderes	mansion	12	5	2662	24
92	Madocsa	public building	8	4	2552	16
93	Iharosberény	castle	19	6	2805	10
94	Öttömös	wine cellar	16	6	2634	10
95	Tarpa	school	15	5	2710	21
96	Debrecen	garage	12	4	3075	13
97	Zalalöv	community house	9	3	2633	12
98	Penc	observatory	12	4	3098	15
		Δ_{VG} = 888 E		mean:	2817	14

Table IV. Results of vertical gradient (*VG*) measurements on the absolute points of Hungary. Legend: see Table I.
IV. táblázat A magyarországi abszolút állomásokon végzett *VG* mérések eredményei.

etc.); in our experience m_x may reach the value of 50–60 E. In favourable cases, on the other hand, higher reliability could be achieved by a lower number of observation series (e.g. Dunakeszi, Rétság). In repeat measurements carried out at the same site with one and the same gravimeter, only rarely do differences greater than 100 E occur. Observation results are values corrected for this long-term change in the calibration factor. Because long-term changes in the calibration factor (at least based on the 12 such meters used in the measurements up till now in Hungary) have values between 0.996 and 1.005, the maximum effect of their possible error on *VG* values is 10 E. To reduce the effect of systematic errors, measurements are always performed with more than one instrument.

In relative measurements of gravity the sensor mass of LCR gravimeters used is 60–125 mm above the benchmark of the site. As a consequence, the error due to the neglection of *VG* anomalies could reach 6 μGal (see Table I). The value might be even more, because at the other stations of the

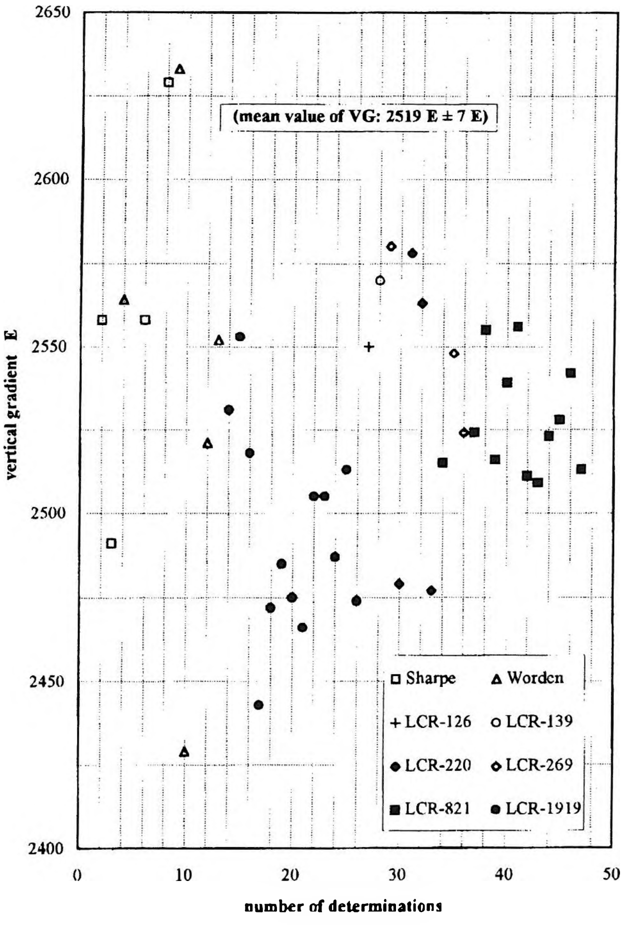


Fig. 1. Results of vertical gradient (*VG*) measurements at the Budapest absolute station
1. ábra. Budapest abszolút állomáson végzett vertikális gradiens (*VG*) mérések eredményei

national base network no *VG* measurements have been performed up till now. It is pointed out that the value of 6 μGal is higher than the measurement reliability achievable with these instruments! The magnitude of the effect is, of course, not proportional to the measured *g* value, it depends only on the surroundings of the observation site.

The results in Table III are examples of the effect of nearby large masses on the values of *VG*. The close correlation between the magnitude of the masses over the site and the *VG* values can be well demonstrated at the Budapest microbase. Between station 1 and station 14 the *VG* values continuously decrease (2591–2102 E) and all are much lower than those measured on the surface. (For comparison: site 1 called Mátyáshegyi Cave in Table I is in the open air but lies in a corrie-like strip pit about 50 m from site 14, which is in the gallery). The calculated average *VG* value (2391 E) is not more than 77% of the theoretical value. The negligence of local *VG* values would produce an error of 3 μGal (between sites 2 and 14). This value is comparable to the measurement reliability of the given LCR gravimeters under optimal observational conditions.

From Table III it is also evident that under optimal observational conditions (constant temperature, low level of vibration during transportation due to the closeness of sites and hand carrying of instruments) significantly higher reliability can be achieved in determining *VG* for microgravimetric purposes than at field sites where the frequently changing ambient conditions have an adverse effect on the observational results.

Table IV contains the most important parameters of *VG* measurements at the absolute stations in Hungary. To interpret the results Fig. 1 was also used where the results of all measurements at the Budapest absolute station are plotted. In addition to LCR instruments, observations were made with Sharpe and Worden gravimeter sets at this station. The scattering of the measurements performed with the latter is somewhat higher, especially if the scattering is compared with that of No. G–821 and G–1919 LCR gravimeters. One of the reasons for this is that *VG* determinations were carried out with these Sharpe and Worden instruments in connection with field measurements, generally after the daily field work. The maximum deviation between the obtained data is 250 E. The extremely high reliability of the *VG* value for the station is primarily due to the high number of series. It can also be seen that there is about 30 E difference of systematic character between the results of the two above mentioned LCR gravimeters; the probable reason for this is the determination problems of long-term changes in the calibration factor. The 47 measurement series were realized between 1980 and 1995. (Although the temporal stability of the vertical gradient of gravity has not yet been properly demonstrated, the possible change in time might not have a detectable influence on the results.) The disturbing effect of nearby masses can clearly be seen in the *VG* values of absolute stations; their scattering around the theoretical value is substantially higher than that of those measured in the open field.

Based on the results shown in Table IV the reliability of *VG* values at the absolute stations in Hungary lies between 7 and 24 E, which is in agreement with the results presented in the literature BECKER et al. [1995] on similar measurements.

6. Summary

It can be seen from the results described in the paper that knowledge of the local value of the vertical gradient is mandatory in high accuracy gravimeter measurements. The necessary (and satisfactory) number of measurements to achieve the accuracy required by the task can be planned based on the examples and method of determination presented here.

Acknowledgments

The author expresses his thanks to the U.S. – Hungarian Science & Technology Joint Fund for the opportunity it provided through grand JF No.369 to carry out the absolute and the associated relative measurements.

REFERENCES

BECKER M. et. al. 1995: Microgravimetric measurements at the 1994 International Comparison of Absolute Gravimeters, Metrologia **32**, 3, pp.145–152

BOEDECKER G., MARSON I., WENZEL H. G. 1995: The adjustment of the Unified European Gravity Network 1994 (UEGN 94). IAGS **113**, pp. 82–91

CHARLES K., HIPKIN R. 1995: Vertical gradient and datum height corrections to absolute gravimeter data and the effect of structured fringe residuals. Metrologia **32**, 3, pp. 193–200

CSAPÓ G. 1987: Some practical problems of the absolute gravity determination (in Hungarian). Geodézia és Kartográfia **2**, pp. 95-99

CSAPÓ G. 1997: Hungary's new gravity base network (MGH–2000). Geophysical Transactions **40**, 3–4, pp. 119–143

ELSTNER C., FALK R., KIVINIEMI A. 1986: Determination of the local gravity field by calculations and measurements Reports of the Finnish Geodetic Institute **85**:3

MARSON I., FALLER J. E. et al. 1995: Fourth International Comparison of Absolute Gravimeters. Metrologia **32**, 3, pp. 137–144

SZABÓ Z., CSAPÓ G. 1985: Microgravimetric survey in Mátyás cave. (presented at the 1st Absolute Gravimetric Intercomparison, Sévres).

SAGITOV M. U. 1984: A plausible interpretation of the deviations between the absolute gravity acceleration measurements performed at Sévres (in Russian). Reports of the Academy of Science of the Soviet Union, Series Geophysica **274**, 2, pp. 300–304

YUZEPOVICH A. P., OGORODOVA L. V. 1980: Gravimetry (in Russian). Nedra, 26 p.

A nehézségi erő vertikális gradiensének hatása nagypontosságú graviméteres mérésekre magyarországi meghatározások alapján

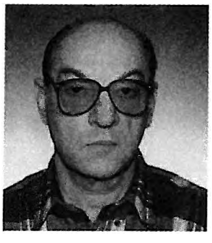
CSAPÓ Géza

Az elmúlt években nagyjelentőségű fejlődés következett be a gravimetria terén annak következtében, hogy világszerte elterjedtek az abszolút graviméterek (jelenleg mintegy 25 berendezést alkalmaznak különböző feladatokhoz), amelyek meghatározási pontossága ma már 3–4 mGal és az eredmények fizikai mértékegységben (ms^{-2}) adódnak. Velük nagy területekre vonatkozóan is gazdaságosan valósíthatók meg egységes gravimetriai alaphálózatok. A 90–es évek politikai változásai Kelet Európában lehetővé tették az eredetileg Nyugat Európa területére vonatkozó Egységes Európai Gravimetriai Hálózat (UEGN) kiterjesztését, amelynek referenciaszintjét az abszolút graviméterekkel mért állomások biztosítják. Ezen főpontok között a hálózat sűrítését LCR gyártmányú relatív graviméter csoportokkal végzik. A megnövekedett mérési pontosság és a homogén megbízhatóságra való törekvés szükségessé tette annak vizsgálatát, hogy a mérések magassági redukciójának számításához általánosan használt elméleti érték (0,3086 mGal/m) összhangban van-e az említett néhány mGal-os mérési pontossággal. A dolgozatban az ezirányú magyarországi mérések eredményeit ismerteti a szerző. Bemutatja, hogy nagy pontosságot igénylő relatív nehézségméré-

sekhez (kalibráló alapvonalak, mikrogravimertriai mérések, stb.) szükséges a nehézségi erő vertikális gradiense helyi értékének ismerete, mert az elméleti vertikális gradienssel számított műszermagassági korrekció a relatív nehézségi térerősség értékeket a meghatározási megbízhatóságot elérő, vagy meghaladó szabályos hibával terhelheti.

Az abszolút graviméterekkel — az adott műszerhez tartozó referenciamagasságra — meghatározott nehézséggyorsulási értéket ugyancsak relatív műszerekkel végzett mérésekkel vezetik le a földfelszínen elhelyezett magassági pontjelre. A szerző tapasztalatai szerint ezen levezetések reális pontossága — abban a gyakorlatban sokszor előforduló esetben, amikor a vertikális gradiens helyi értékét egy–két graviméterrel csupán egy mérési sorozattal határozzák meg — nem éri el a szükséges ± 10–20 E értéket.

ABOUT THE AUTHOR



Géza Csapó graduated as a geodesist from the Technical University of Budapest in 1966. He then joined ELGI and specialized in gravimetry. In 1975 he was awarded a postgraduate degree and, in 1984, his Ph. D. — both of them in gravimetry. From 1970 to 1990 he carried out gravity measurements in many European countries, in South-America and in Mongolia. At present he is principal scientific officer in the Earth Physics Department of ELGI. His main field of interest is high precision gravimetry, instrument development, and calibration.

GEOPHYSICAL TRANSACTIONS 1999
Vol. 42. No. 1 – 2. pp. 83–103

Relationship between reliability of the Hungarian part of the geoid and the spatial distribution of gravity data

Géza CSAPÓ*, Miklós GAZSÓ**, Ambrus KENYERES**

It was assumed that only one gravity field exists consequently the geoid is unique irrespective of the mathematical means utilized for its determination. The increase in the number of gravity measurements improves the resolution of the known gravity field consequently the calculated geoid will have more details and will more precisely approximate the real geoid. The errors of sampling (about the density of gravity measurements), of interpolation, and of different models used for the calculations were studied. Since the covariances representing the statistical nature of the gravity field are unknowns the effect of the changes in the following quantities were investigated: Faye-anomalies, geoid undulations, deflection of the vertical, gravimetric correction, the deflection–component of the gravimetric correction and the curvature. The gravity data were derived from two different data sets, viz. 1993 and 1995 of ELGI. The investigations of gravimetric correction as the potential error fulfilled the expectations because its variations caused ± 1–2 cm undulations of the geoid. Since this quantity is a measure of local deformations it represents the refinement of the geoid and offers better resolution. The regional errors of the global (long wave) components of the gravimetric geoid can be eliminated by evaluating the so called 'GPS elevations', which can be calculated from the ellipsoidal and orthometric heights derived from GPS measurements and high order levelling respectively. From the point of view of practical application, since the GPS reference points are between 30 and 70 km apart, the refinement of the local geoid is a necessity. The refinement can be based on detailed gravity data.

Keywords: geoid, Faye-anomalies, gravity survey maps, GPS, Hungary

* Eötvös Loránd Geophysical Institute of Hungary, H-1145 Budapest, Kolumbusz u. 17-23
** Satellite Geodetic Observatory, Institute of Geodesy, Cartography and Remote Sensing, H–1373 Budapest, P.O.B. 546, Hungary

Manuscript received: 20 September, 1996.

1. Introduction

One of geodesy's long-term scientific tasks is that of studying the geoid as a representation of the figure of the Earth determined by the gravity field. In view of the rapid expansion of the 3D positioning technique, the Global Positioning System (GPS) for everyday surveying and engineering tasks requires ever-increasing accuracy in determining the geoid.

Judjing from Hungarian experience the present level of accuracy of the local geoid* means that geometric levelling cannot be substituted by the GPS technique since the accuracy of the vertical coordinate determined by the GPS technique is significantly lower than that of the horizontal coordinates.

2. Aim of our studies

The aim of this study was to get an answer as to how the increasing number of gravimetric measurement sites in a certain area influences the accuracy of the local geoid. Our principal idea is that since there is only one gravity field, there can only be one geoid and it is completely another question as to what kind of technical–mathematical apparatus is used to produce this surface. Since the increasing number of gravity measurements (hereafter g–measurements) results in a better resolution of the gravity field the resulting change of geoid undulations due to this densification inevitably yields *increasing reliability of the geoid* because the same physical reality is involved. In fact this is what is needed in GPS applications, viz. more reliable undulations at certain points, not an overall increase of statistical accuracy of undulations (although there is a connection between the two).

* In this paper we regard the local geoid as that component of the geoid which is the effect of the irregular mass distribution of the Earth's crust and is reflected in gravity anomalies; in other words, that component which represents the short- and intermediate wavelength anomalies of the geoid.

Thus, our goal is to study the increase of accuracy. Here, we should like to show only certain general aspects of accuracy analysis of the theory of errors thereby making some preliminary steps towards the future task of its solution. The study of reliability and representation improvement is phenomenological and our mathematical analysis serves as a numerical support.

3. Main principles

The starting point of our study was the fact that today's accuracy of gravimetric measurements exceeds by orders of magnitude the accuracy required by geoid determinations. Therefore we are not investigating the reliability of each item of gravity data but the effect of gravimetric coverage: we are dealing with the error of representation and the accuracy of mathematical processing (model error).

— The reliability of representation is a measure of the goodness of the approximation of the real gravity field by using a certain set of gravity data over a given area; this simply means the distribution and density of measurement points and we are looking for the effects of this density variation. Bearing in mind that nowadays for any kind of study connected with the geoid regular grids are used exclusively, the so called interpolation error cannot be neglected. This problem has been thoroughly discussed in the literature. In the case of Hungary, SÁRHIDAI [1993] is cited and his results are referred to elsewhere in this paper.

— So far as mathematical processing (the above mentioned technical–mathematical apparatus) is concerned, two modells can be applied: one is the Stokes–integral, the other is the Vening–Meinesz integral. The first one directly yields undulations, the second one indirectly through the deflections of the vertical. (In recent years a new method has gained ground, viz. the stochastic model).

The geoids under discussion are 'gravimetric' ones, i.e. the undulations are computed by the Stokes–integral. This is done through two steps: the so-called long wavelength part — the *global geoid* as an overall picture

— is computed from spherical harmonic series; the medium and short wavelength part — the *local geoid*, which is characteristic of the local features — is computed from gravity anomalies. It is easy to estimate the accuracy of the global geoid because of its analytical form and abundant literature is available in this field e.g. [ÁDÁM 1993]. With regard to the local geoid the task is somewhat harder since in lack of a mathematical algorithm one would rely on numerically derived covariances only, but because the covariances reflecting the statistical behaviour of local gravity fields are not known, one is forced to adopt an empirical approach. For this reason the error study of local geoids is rather incomplete — though in recent years this field has witnessed promising investigations in Hungary. These facts were taken into account when the set of variables to be studied was constructed. As we will see later on, the development of the Hungarian gravity data bank (the more detailed representation of the gravity field) has led to changes of several centimetres in geoid determination. This is important since just the construction of a geoid of centimetre reliability is the most important task of today and of the near future. It will be shown that on the one hand these variations in the undulations are non-random (noise) but reflect a real reliability improvement; on the other hand the distribution of these variations show a very strong regularity — mostly as a function of the topography. It is generally known that Faye-anomalies and derived quantities also strongly correlate with topography, so they establish a regional representation of '*g*' (since 'noise' cannot have structure, it cannot be regional). To complete the study a gravity data bank and test areas of different morphology are required.

4. Variables to be studied

In Hungary two methods of geoid determination are applied: the first one (of a temporary nature) is *astro-gravimetric*, the second one is *gravimetric*. There is a principal difference between the two techniques. The astro-gravimetric method is an improvement over conventional astronomic levelling in a physical sense and so — indirectly — undulation differences are computed from deflections of the vertical. The improvement lies in the fact that the hypothetical linear change of deflections of the vertical inherent in astronomic levelling is resolved by the gravimetric correction (hereafter: *GK*), which can be computed from gravity anomalies [BÍRÓ 1985, GAZSÓ, TARASZOVA 1984, MOLODENSKII et al. 1962]. Although derived by simple algebraic considerations it carries a physical meaning therefore it may be used as a *quantity for comparison* even for the gravimetric geoid: it makes manifest by how much it would improve the results of astronomic levelling, which is geometrically correct but physically all the more unrealistic. In this sense it reflects the effects of the densification of *g-measurements.*

Gravimetric geoid determination is completely different: geoid heights are directly determined through the Stokes–integral, hence the result is a discrete set of undulations. In the astro-gravimetric process closed polygons can be assembled and r.m.s. errors can be determined by adjusting their closure errors — at least as regards the internal accuracy. Seemingly there is no such condition in the gravimetric geoid but through the *GK* one such a condition can be set up. It may seem that the *GK* is tied to astro-gravimetric levelling and it is really so for the geoid determination technique. On the other hand it also has a more general potential theoretical meaning; in other words to what extent this set of undulations represents — in linear terms — a level surface, the geoid.

A sufficient condition for two points and all points along their interconnecting curve to be on the same level surface (geoid) is that the tangents at each and every point of the curve must be perpendicular to the normal of the level surface (*Fig. 1*).

Notations:

i, k — two points in space

$\bar{g}$ — gravity

$\bar{n}$ — unit vector of the inner normal of the level surface

$\bar{l}$ — tangent vector at any point of the i and k curve section

s — projection of l onto the ellipsoid

N — undulation

Θ — projection of the deflection of the vertical onto the osculating plane of the interconnecting curve.

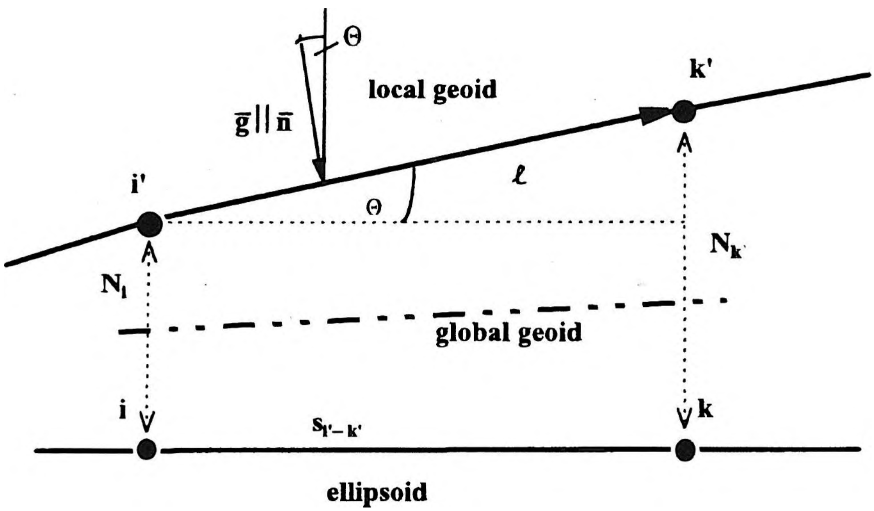


Fig. 1. Gravimetric correction scheme

1. ábra. A gravimetriai korrekció fizikai értelmezése

The condition of perpendicularity at any point of the curve is fulfilled when the scalar product $n \cdot l$ vanishes ($n \cdot l = 0$). In geodesy the astronomical coordinate minus ellipsoidal coordinate gives the sign of vertical deflection, consequently:

$$-1/\rho'' \cdot g \cdot \Theta \cdot s - g(N_k - N_i) = 0$$

By assuming linearity this pointwise condition can be transformed into a sidewise one:

$$(N_k - N_i) + 1/2\, \rho'' \; (\Theta_k - \Theta_i)\, s_{i,k} = (GK) = (PE)/g$$

The left side of this equation is precisely the formula of *GK*, and *PE* represents the 'potential error', thus referring to the difference with respect to the condition. By dividing the potential error between points '*i*' and '*k*' by '*g*' and making it equal to zero means the same as reducing *GK* to zero.

We would mention that the same result will be obtained from potential theory if work is done along the direct path between two points or it is summed up from work done along the horizontal and vertical paths. After all these preliminaries, changes of our two variables to be studied, viz. undulation 'N' and gravimetric correction GK, will be determined and compared. If the letter D is used to denote differences then the expression:

$$DX = X_{95} - X_{93}$$

denotes the difference of a certain quantity X based on two data archives belonging to the years indicated by the lower indices [KOVÁCSVÖLGYI 1994].

It is mentioned that since values of N are computed by the Stokes–integral, and the Vening–Meinesz formula is also needed to determine the GK values, these latter values also contain information on modelling errors. In addition to the quantities DN and $D(GK)$, the components of the latter were also studied:

Quantities at points:

F — Faye- (free-air) anomaly

N — undulation

Θ'' — $(\xi^2 + \eta^2)^{1/2}$ magnitude of deflection of the vertical

Quantities along sides:

GK — gravimetric correction

$\Theta^+ = 1/2\,\rho''(\Theta_k + \Theta_i)\, s_{i,k}$ — deflectional component of GK
where $s_{i,k} = 2.83$ is the diagonal of the grid square

$\Theta^- = 1/2\,\rho''(\Theta_k - \Theta_i)\, s_{i,k}$ —measure of curvature (which is the limit of the differential rotation of the tangent of a superficies curve)

Quantities along sides were computed for both SW–NE and SE–NW directions, the positive direction always being the direction from S to N. These above-mentioned directions were denoted by α and β in the computations.

5. Some considerations on the variables

Many qualitative statements can be drawn from the different values of the variables:

DF — this can be regarded physically as mass surplus ($DF > 0$) or mass deficit ($DF < 0$), from this follows its effect on the shape of the geoid.

DN — if $DN > 0$ then the computed quantity indicates a 'mass surplus"; if $DN < 0$ the reverse is true. Generally speaking, DN indicates the local rising/sinking of the geoid surface.

Θ^- — we have only qualitatively studied sign changes of curvature and inflections; if $\Theta^- > 0$ then the geoid surface is concave from below, if $\Theta^- < 0$ then it is convex.

Since GK is proportional to PE, if $GK > 0$ then the 'tilt' of the tangent to the geoid means increasing N_k and $GK < 0$ means its decrease with respect to N_i, if the positive direction is $i \rightarrow k$.

6. Preparation and visualization of variables for study

The area of evaluation of gravity anomalies was reduced to 50x50 km from 100x100 km on the 2x2 km grid to make the effect of neighbouring areas more pronounced.

Quantities defined at points were represented at grid points and those defined along sides were relocated to the NE and NW corners of grid squares.

Plots were made for each study, in some cases the isolines were also provided.

7. Regions under investigation

For our present studies two morphologically different regions were selected where there has recently been a considerable increase of gravity data. The first one (A) is the region where the Transdanubian Hills and the Great Hungarian Plain meet and where a characteristic topographic feature, the Tolna Ridge, rises in an approximately north–west direction; this region is 4500 km^2 and the topography varies in height between 90 and 270 meters. This region has 1170 grid points. The second region (B) is the region of the Mátra and Bükk Mountains with their southern surroundings, where the topography of these mountains of medium height gradually becomes a plane area almost without any transition zone. This region covers 8520 km^2, the number of grid points is 2231, the height range is between 90 and 1000 m. These regions can be seen in *Figs. 2* and *3*.

The above mentioned regions A and B were subdivided into areas as follows:

region 'A': External part of Somogy (A1), Tolna Ridge (A2), Mező-föld (A3), Kalocsa Depression (A4), see Fig. 4.

region 'B': Mátra Mountains (B1), Bükk Mountains (B2), Aggtelek–Szendrő–Cserehát region (B3), Heves–Borsod Plain (B4), see Fig. 9.

The reason for subdividing A and B into smaller areas was partly to reveal more accurately the effects of interval densification [CSAPÓ 1994], partly since there are pronounced morphological differences within the given areas. This latter factor is of importance because the investigations concerning the whole region produce average results, but the area subdivision demonstrated the effects of local changes as well.

8. Gravimetric databases used for the study

There are two ways of selecting gravimetric databases: the first is to select a database containing all available data from which the second one could be derived by neglecting certain items following a certain rule (fictitious database); the second way is to use two real different databases. We chose the second way and this was made possible because the gravimetric database of ELGI was increased significantly during the years 1993 and 1995. The first gravimetric database (1993) consists of approximately 250 000 items, the second one (1995) contains more than 320 000 gravimetric data. Faye-anomalies were interpolated from both databases, on a 2x2 km grid by using a digital terrain model (the accuracy of interpolation of Faye-anomalies can be increased significantly if the heights of the points are taken into consideration). A detailed description of the interpolation method can be found in the literature [SÁRHIDAI 1994]. We considered it extremely important that the same technique be applied to both databases and two identical grids because different interpolation methods may yield different interpolated values even from the same data sets.

9. Results

Numerical studies

Before introducing our analyses we thought to mention that in a rigorous sense we can hardly speak of a statistical analysis but based on the integral limit theorem of probability theory the sample distribution may be normal if the corresponding value is the sum of small effects. Since this certainly true for *DF* values and the other variables of the study are a linear transform of these (Stokes and Vening–Meinesz integrals), asymptotically the normality holds true also for these with special regard to their very large number of elements. But when statistical characteristics and their tests are regarded as an algebraic fitting problem, the above mentioned problem will not even arise.

DX quantities

The results of numerical studies are summarized in *Table I*. Our analysis was based on the results shown by this table and by comparing them with the computer drawings. Among the results of Table I, the quantities m_x and K need further explanation, the others being known mathematical statistical quantities. From the theory of errors it is known that mean square errors can also be determined from the actual errors if the difference of two measurements, made for the same quantity, is known [DETREKŐI 1991].

If d is the difference of two measurements, n is the number of measured differences, L_1 and L_2 are the 1st and 2nd measurements for the same quantity then $m_d = (\Sigma\, d^2/n)^{1/2}$ is the mean square error of one difference, $m_L = (m^2_d\,/\,2)^{1/2}$ is the mean square error of one measurement.

Although our DX differences in the rigorous sense of error theory cannot be treated strictly as 'independent and equally accurate measurements for the same variable', bearing in mind what was said at the beginning of this section this formalism can be used. If DX is substituted for d and X for L, then $m_{DX} = \text{r.m.s.}(DX)$ is the 'root mean square error' (r.m.s.) of the one difference DX, and $m_X = \text{r.m.s.}(DX)\,/\,\sqrt{2}$ is that of the one variable X. Moreover, the effects of systematic errors can be neglected when the ine-

Fig.	X	a	s	max	min	n	r.m.s.	m_x	K	unit
					Region 'A'					
6	$D\Theta$	−0.01	±0.11	0.62	−0.82	1170	±0.11	±0.08	0.07	arcsec
5	DN	−0.2	±0.46	0.4	−2	1170	±0.5	±0.35	0.4	cm
4	DF	−0.13	±1.43	9.04	−12.8	1170	±1.44	±1.02	0.1	mGal
7	$D(GK)_\beta$	0	±0.1	0.7	−0.8	1100	±0.1	±0.07	0	cm
8	$(GK)95_\beta$	0	±0.1	0.6	−1.1	1100	±0.1			cm
					Region 'B'					
9	DF	−0.84	±2.23	9.87	−25.3	2601	±2.39	±1.7	0.35	mGal
10	DN	−1.6	±1.5	0.3	−6.5	2601	±2.2	±1.56	0.73	cm
11	$D\Theta$	−0.06	±0.24	1.16	−1.82	2601	±0.2	±0.17	0.24	arcsec
12	$D\Theta^{+}_{\alpha}$	0.1	±0.2	1.2	−1.1	2500	±40.3	±0.21	0.33	cm
13	$D\Theta^{+}_{\beta}$	0	±0.3	1.6	−1.3	2500	±0.3	±0.21	0	cm
14	$D\Theta^{-}_{\alpha}$	0	±0.2	0.9	−1.3	2500	±0.2	±0.14	0	cm
15	$D\Theta^{-}_{\beta}$	0	±0.2	0.9	−1.6	2500	±0.2	±0.14	0	cm
16	$D(GK)_\alpha$	0.1	±0.2	1.2	−1.1	2500	±0.2	±0.14	0.5	cm
17	$D(GK)_\beta$	0	±0.2	1.2	−1.4	2500	±0.2	±0.14	0	cm
18	$(GK)95_\alpha$	−1.8	±1.1	0.8	−4.9	2500	±2.1			cm
19	$(GK)95_\beta$	0.5	±.07	3.7	−2.6	2500	±0.8			cm

Table 1. Statistical characteristics of DX quantities

1. táblázat. A DX mennyiségekkel végzett számszerű vizsgálatok eredményei

quality $|a| \leq 0.2\ m_d$ holds. From this follows the upper limit of 'neglection':

$$K = |\,a\,| \;/\; \text{r.m.s.}(DX) \leq 0.2$$

where $|\,a\,|$ denotes the average of DX differences (sample mean).

t–tests and F–tests

The two fundamental parameters of mathematical statistics are the sample mean (a) and the empirical variance (s), therefore first we performed two sample t–tests and F–tests with a very high number of elements [DETREKŐI 1991]. Using the notations of Table I. (these notations were also used later) we compared the following samples and variables:

— between the regions 'A' and 'B': DF, DN, $D\Theta$, $D(GK)_\beta$ and $(GK)95_\beta$

— inside region 'B' the corresponding side quantities taken along α (strike direction) and β (transverse direction): $D\Theta^+$, $D\Theta^-$, $D(GK)$, $(GK)95$, and differences of $D(GK)$ and $(GK)95$ along both directions. The numbering of plotted figures based on the results of the above-mentioned tests is also presented in Table I.

The null-hypothesis of t–tests (H_0) was the identity of compared sample means. Since the null-hypothesis (with two exceptions) was not fulfilled either for the confidence level $p = 90\%$ or for $p = 95\%$, the conclusion was drawn that for the major part of the studied variables the means of these samples are different for the two regions. The two exceptions are:

— the mean values of $D(GK)_\beta$ of regions 'A' and 'B'. The corresponding data of Table I and the relevant figures plainly show that sample means are zero but the variances are significantly different. Over region 'A', $D(GK)$ may be treated as 'noise'; over region 'B' two regional anomalies of different sign can be seen.

— for the $D\Theta^-$ values of region 'B' neither means nor variances differ along the given directions. The corresponding figures show positive and negative bands of approximately identical distribution rotated by 90°. These bands (axes of inflexion) are always perpendicular to the positive SW–NE and SE–NW directions.

In both of these cases this structure is due to a regional phenomenon, foreshadowing the necessity for a regional analysis of the investigated regions.

The null-hypothesis of the F-tests was the identity of compared variances. This had to be rejected on all possible confidence levels because of the very high number of samples. Also here two exceptions were found where the identity of variances cannot be rejected statistically (F=1). Both of these cases arose when comparison was made inside region 'B'. The first case was for Θ^- values, the second one at *D(GK)* values, where the same kind of 'averaging' occurred as was the case for *t*–tests made for *D(GK)* variables between the two regions.

Both of these studies have shown that the changes arising as a consequence of the increased number of gravimetric measurement points are of a physical nature (not noise), since they have structure. This statement is true for the regions and for the *DN* and *D(GK)* quantities, and also for the directions.

General conclusions on numerical studies

The number of samples studied is very large (Table I, column '*n*') by which the regionalities are smoothed out (because of the great number of elements they become partly randomized). This fact is clearly indicated by the ratio of '*a*' to '*s*'. From this it follows that the 'max' and 'min' values of Table I are very characteristic and here even a difference of two orders of magnitude is not so infrequent between the sample average and the extreme values.

Among *DF* quantities negative values are dominant for both regions, i.e. the database extension induced a general decrease of *F* values.

With respect to *DN*: a decrease of anomalies indicates physically a 'mass deficit' which causes geoid anomalies to be flatter. The increase in the number of measurements resulted in a decrease of geoid undulations — depending on the topography. The value of *DN* lies in the range of reliability requirements of the 'cm geoid' and it shows a sinking of the geoid surface.

With respect to *D(GK)* and *D(PE)*, our study of gravimetric correction *GK* — which was also interpreted physically as a potential error — has fulfilled our expectations since its effect produced fluctuations of ± 1–2 cm magnitude. This quantity being the measure of local variations it reflects the refinement of geoid and resolution increase, therefore with respect to the 'GPS geoid' it makes its practical realization more realistic. In other words, by processing the so called 'GPS geoid heights', i.e. the differences between ellipsoidal heights obtained from GPS measurements and orthometric heights obtained from first-order levelling at reference stations 30–70 km apart, it is possible to filter out regional errors of the global (regional) part of the gravimetric geoid. In this way the accuracy of the local (short- and medium wavelength) part, which is based on detailed gravity measurements, will be important from the point of view of applications.

Regression and correlation computations

These studies were performed only for region 'B', which has highly variable topography along two profiles of characteristic direction [HAJTMAN 1968, MUNDRUCZÓ 1981]. The first direction is the strike direction denoted by α, the second is the transverse direction β. Our analysis aimed at clarifying the relations of the basic quantities *DN*, *D(GK)*, and *Kh* (in the differences $Kh = h_k - h_i$, 'h' denotes heights of grid points above sea level and indices k and i are the same as for the previously treated *GK*. In *Table II*, the usual statistical parameters are presented in centimetres.

X	*a*	*s*	max	min	r.m.s.	unit
profile SW–NE (α)						
DN	− 2.6	±1.8			±3.1	cm
*(GK)*95	− 1.9	±0.9			± 2.1	cm
profile SE–NW (β)						
DN	− 1.8	±2.1	+0.24	− 6.07	±2.8	cm
*(GK)*95	0.5	±0.7			±0.8	cm
D(GK)			+ 0.86	− 1.42		cm
(Kh)			+ 21	− 10		m

Table II. Statistical characteristics of fundamental quantities

II. táblázat. Az alapvető vizsgálati mennyiségek statisztikai jellemzői

Results of the correlation study of the SE–NW transverse–directional profile:

a) For the two–variable study:

$DN = -2.4 + 1.19\, D(GK)$ $\quad R = +0.75 \quad m_0 = \pm 2.19$

$DN = -2.4 + 0.02\, (Kh)$ $\quad R = +0.74 \quad m_0 = \pm 2.22$

b) For the three–variable study:

$DN = -2.37 + 1.13\,D(GK) + 0.01\,(Kh)$ $\qquad m_0 = \pm 2.23$

$R_T = 0.75 \qquad R/D(GK) = -0.17 \qquad R/(Kh) = -0.03$

In these equations R_T denotes the total correlation coefficient, R/X the partial correlation coefficient of X, and m_0 is the r.m.s. variance computed from the corrections. It can be concluded from the results that the effect of $D(GK)$ is almost identically transferred to DN—there is a definite relationship. The correlation of Kh is also high but its effect is rather small. Finally, R_T stands unaltered with respect to R but the partial correlation coefficients (R/X) are smaller than the previous ones by one order of magnitude so the variables $D(GK)$ and (Kh) do not give independent information with respect to each other. Judging from the equation for DN, (Kh) seems redundant. The m_0 errors are practically constant for the three equations in the order of magnitude of DN.

From what is said above, it follows that the effect of the densification of gravity data is outstanding in DN itself but this is affected to the same degree by variation of GK. This indicates refinement of the geoid, it combines the sinking of the surface with a 'tilt'.

Results of merged data sets of SE–NW and SW–NE directions:

a) For the two-variable study:

$DN = -2.62 - 0.24D(GK) \qquad R = +0.80 \qquad m_0 = \pm 1.9$

$DN = -2.51 - 0.004(Kh) \qquad R = +0.79 \qquad m_0 = \pm 1.9$

b) For the three-variable study:

$DN = -2.62 - 0.24\,D(GK) - 0.003\,(Kh) \qquad m_0 = \pm 1.9$

$R_T = +0.80 \qquad R/D(GK) = -0.13 \qquad R/(Kh) = -0.04$

From these studies it can be stated that the regression coefficient changes by an order of magnitude (from +1.1 to −0.2) with respect to the 1D study along profiles, and this is caused by the 2D region treatment that combines areas of different morphology. The m_0 values did not change. To back up the necessity for subdividing the test regions into smaller areas we compiled the correlation coefficients of all six variable data sets (*Table III*). It can be seen that the partial correlation coefficients of $D(GK)$ and (Kh) are unchanged.

partial correlation coefficients	
R_T	+ 0.63
$R'_{D(GK)}$	– 0.19
R'_{Kh}	– 0.02
R'_X	+ 0.21
R'_Y	– 0.30
R'_t	– 0.30

Table III. Partial correlation coefficients
III. táblázat. Parciális korrelációs együtthatók

Attention is drawn to the fact that all variables appear on the previously mentioned 2x2 km grid, the diagonal of which is 2.8 km, so variations of DX *refer to these distances; hence they are indeed significant!*

Discussion of computer plots

Region 'A':

DF (*Fig. 4*): The Faye-anomalies are predominantly negative. It can be seen that detailed survey of the Tamási area (A1), resulted differences, as high as –5, –10 mGal. The same holds true in areas A2 and A3, whereas the Kalocsa Depression (area A4) densification has altered the previous picture very little.

DN *(Fig. 5)*: Undulation changes are all negative (their magnitudes are about 1.0–1.5 cm) and correlate with the decrease of *F* values.

DΘ and ***D(GK)*** (*Figs. 6, 7*): Changes are without any structure (noise).

***(GK)*95** (*Fig. 8*): In the NW corner of the region (area A1) there are changes of max. +2 cm; the other parts of the region are homogeneous, (cca. –1 cm changes can be seen).

It can be stated for region 'A' that densification of *g*-measurements was effective, the distribution pattern of improvement is in correlation with the topographic features. *The Tamási region needs further densification*, whereas other parts of region 'A' seem to have adequate gravity coverage.

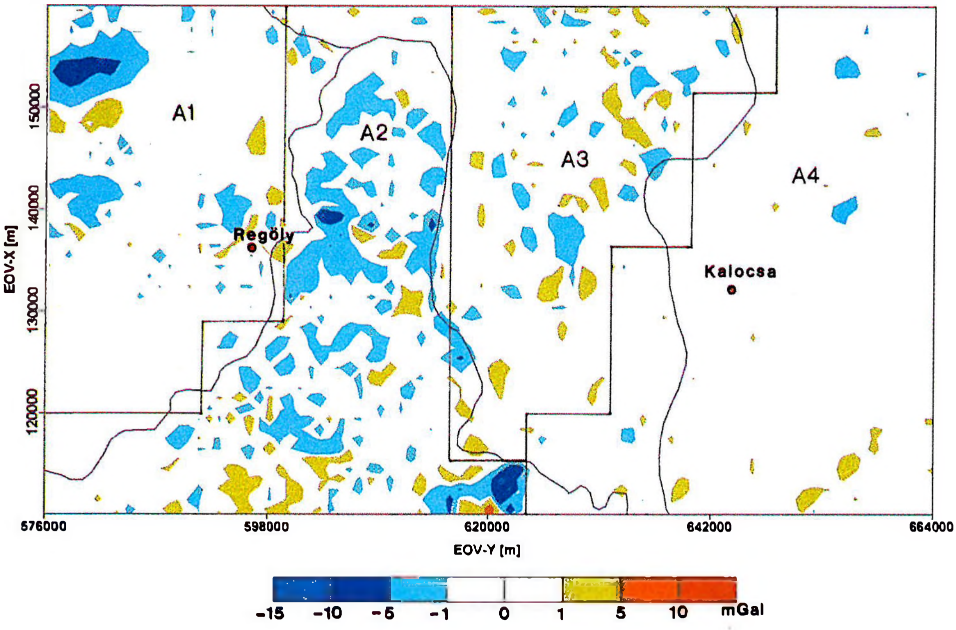


Fig. 4. Region 'A': Map of Faye-anomaly differences (*DF*). $DF = F_{95}–F_{93}$

4. ábra. "A" terület: A Faye-anomáliák változásának (*DF*) térképe. $DF = F_{95}–F_{93}$

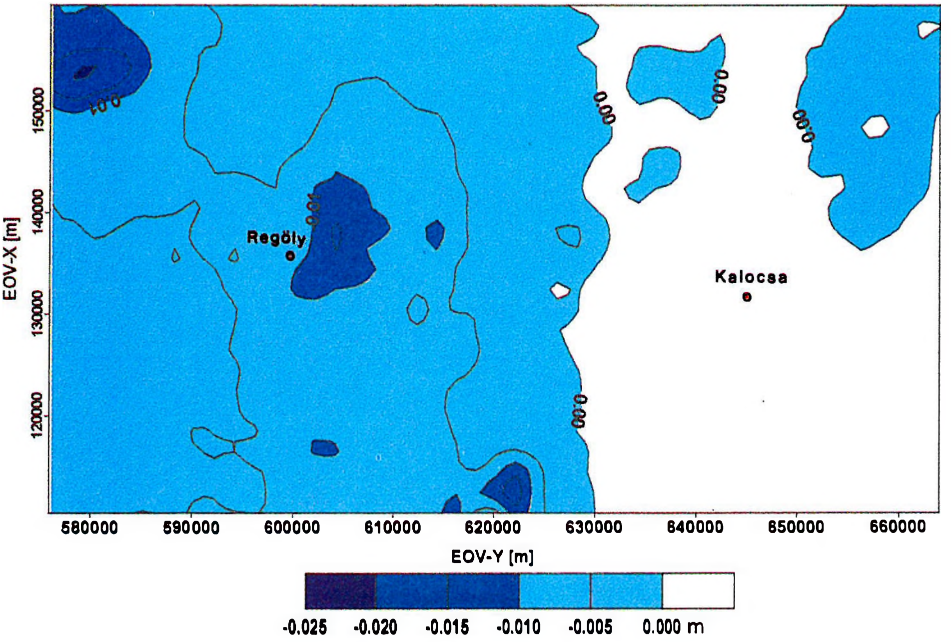


Fig. 5. Region 'A': Effect of *DF* in geoid undulations (*N*)

5. ábra. "A" terület: A *DF* hatása a geoid–undulációkra (*N*)

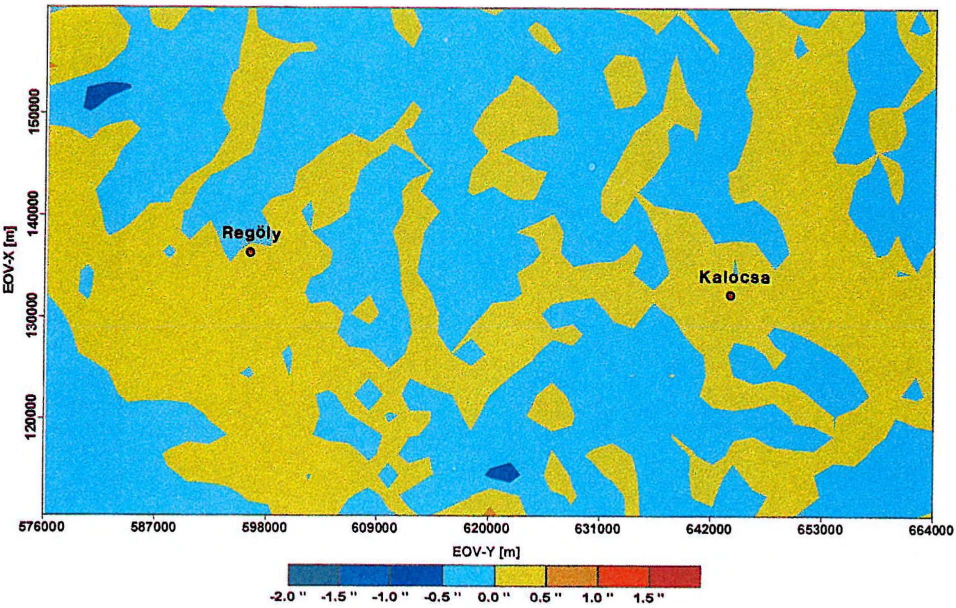


Fig. 6. Region 'A': Effect of *DF* in the absolute values of vertical deflections (Θ)
6. ábra. "A" terület: A *DF* hatása a függővonalelhajlások abszolút értékére(Θ)

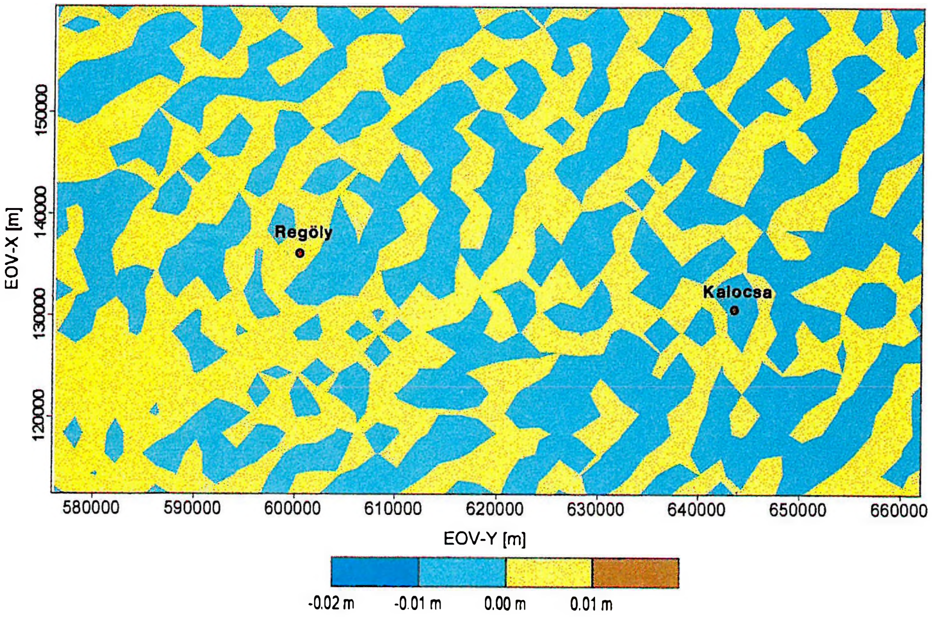


Fig. 7. Region 'A': Effect of *DF* in the β component of gravimetric correction
7. ábra. "A" terület: A *DF* hatása a gravimetriai korrekció β irányú összetevőjére

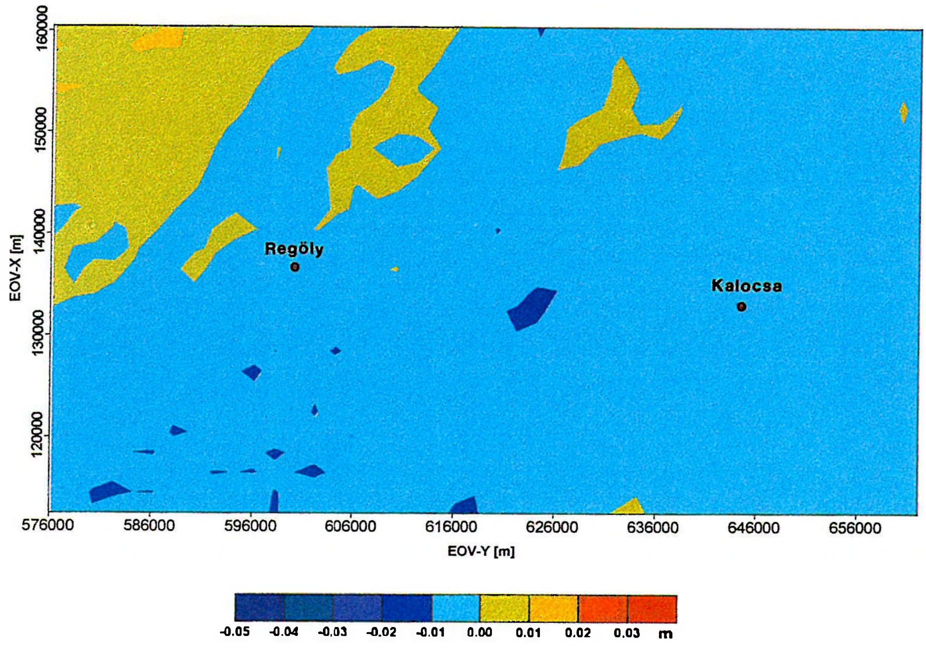


Fig. 8. Region "A": Map of β components of gravimetric corrections computed from 1995 database

8. ábra. "A" terület: A gravimetriai korrekciók β irányú összetevőjének térképe (1995. évi adatbázisból)

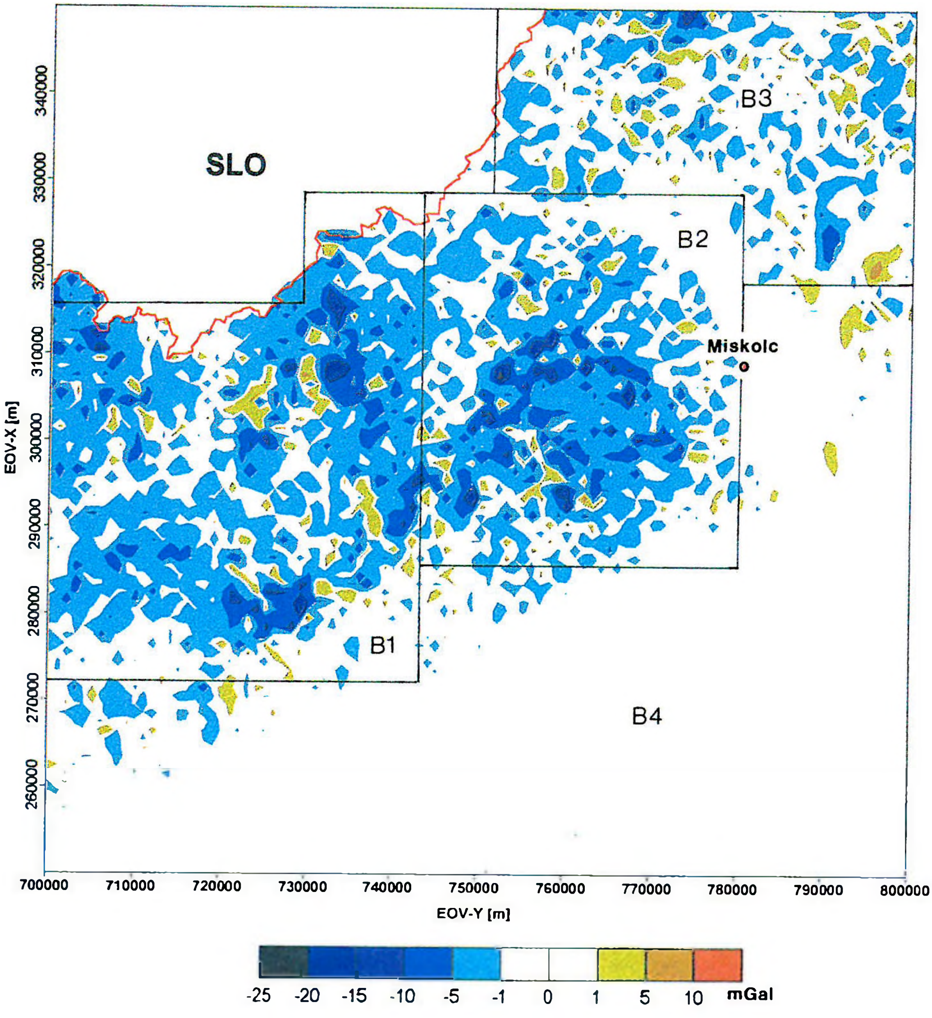


Fig. 9. Region 'B': Map of Faye-anomaly differences (*DF*). $DF = F_{95}-F_{93}$

9. ábra. "B" terület: A Faye-anomáliák változásának (*DF*) térképe. $DF = F_{95}-F_{93}$

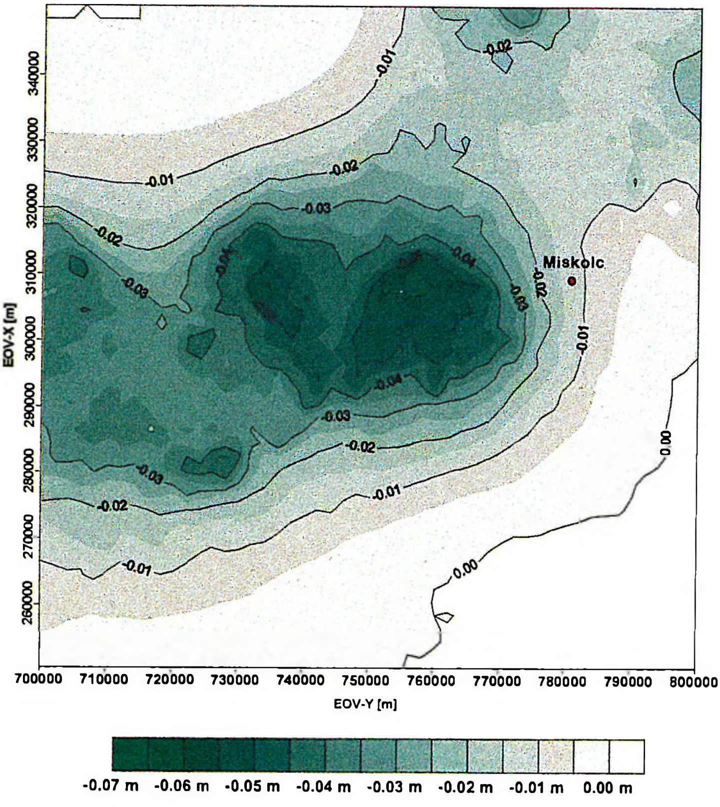


Fig. 10. Region 'B': Effect of *DF* in geoid undulations *(N)*

10. ábra: "B" terület: A *DF* hatása a geoid-undulációkra *(N)*

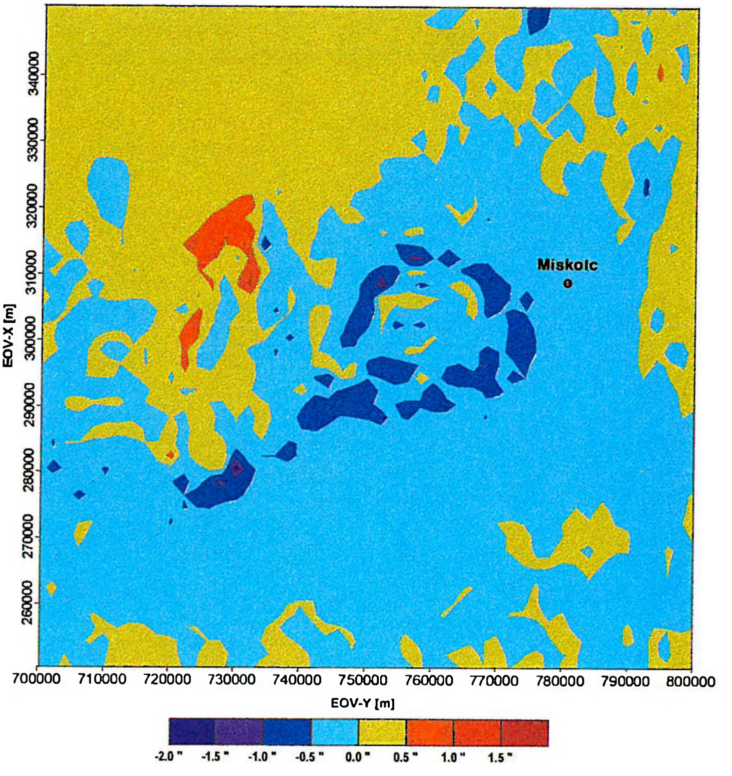


Fig. 11. Region 'B': Effect of *DF* in the absolute values of vertical deflections (Θ)

11. ábra: "B" terület: A *DF* hatása a függővonalelhajlások abszolút értékére(Θ)

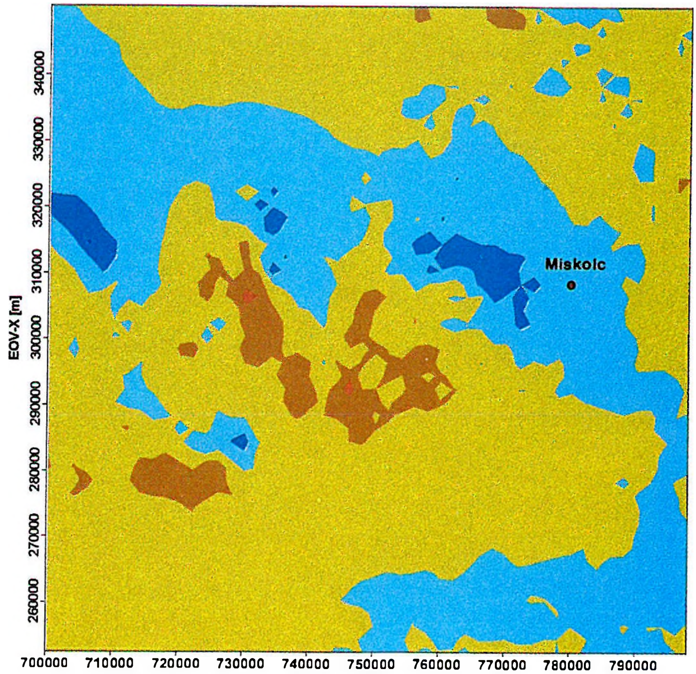


Fig. 12. Region 'B': Effect of *DF* in the α component of vertical deflections

12. ábra. "B" terület: A *DF* hatása a gravimetriai korrekció α irányú függővonalelhajlási összetevőjének változására

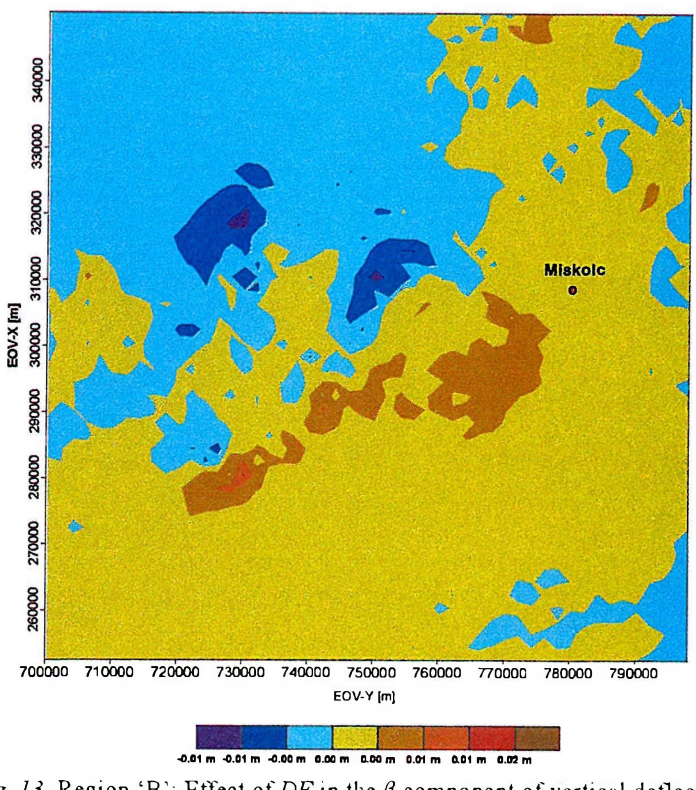


Fig. 13. Region 'B': Effect of *DF* in the β component of vertical deflections

13. ábra. "B" terület: A *DF* hatása a gravimetriai korrekciók β irányú függővonalelhajlási összstevőjének változására

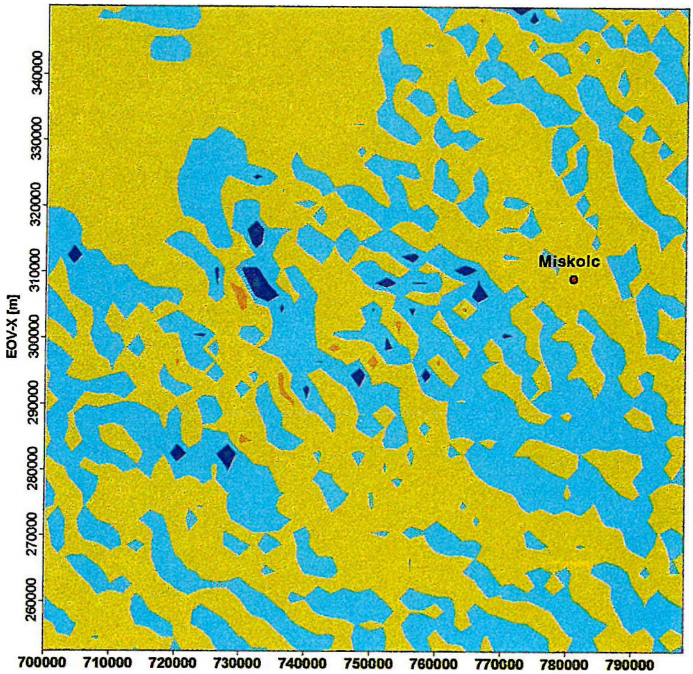


Fig. 14. Region 'B': Effect of *DF* in the α component of curvature values
14. ábra. "B" terület: A *DF* hatása az α irányú görbületi értékek változására

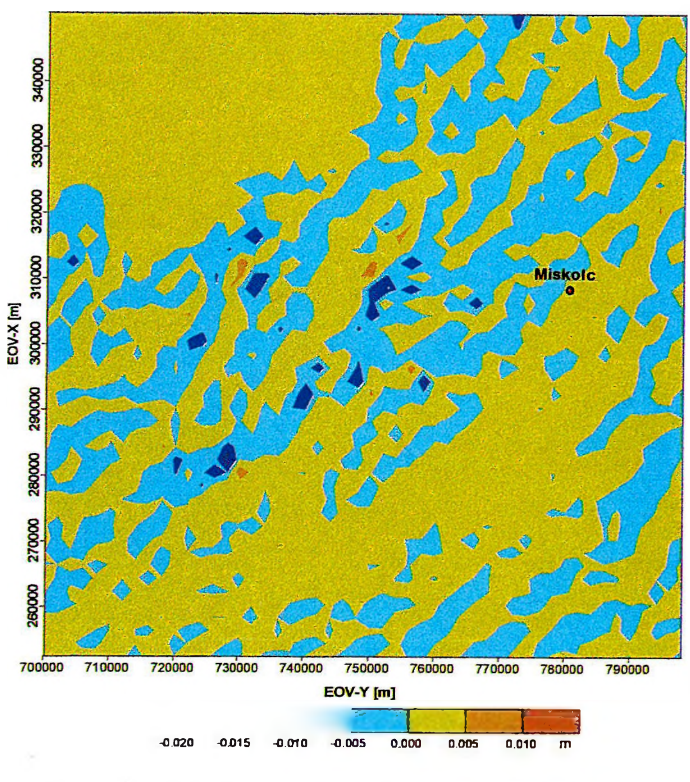


Fig. 15. Region 'B': Effect of *DF* in the β component of curvature values
15. ábra. "B" terület: A *DF* hatása a β irányú görbületi értékek változására

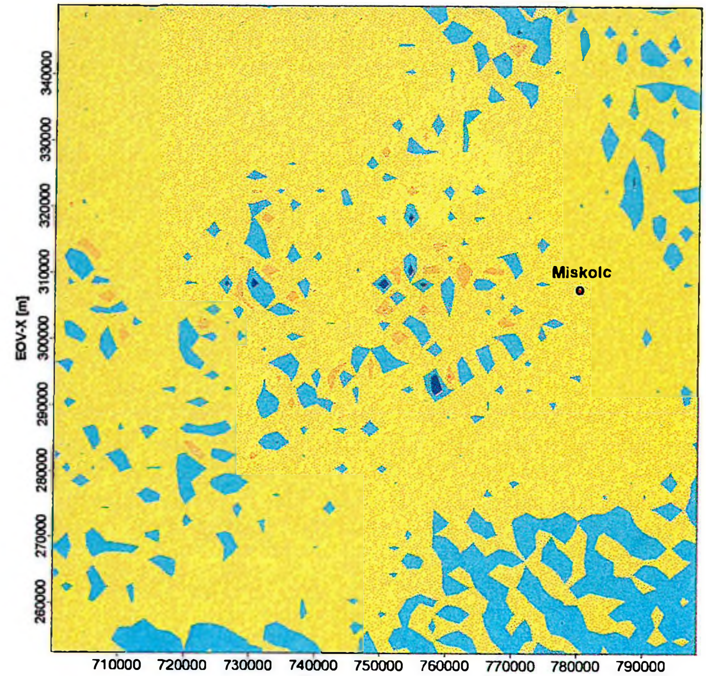


Fig. 16. Region 'B': Effect of *DF* in the α component of gravimetric corrections
16. ábra. "B" terület: A *DF* hatása a gravimetriai korrekciók α irányú összetevőjének változására

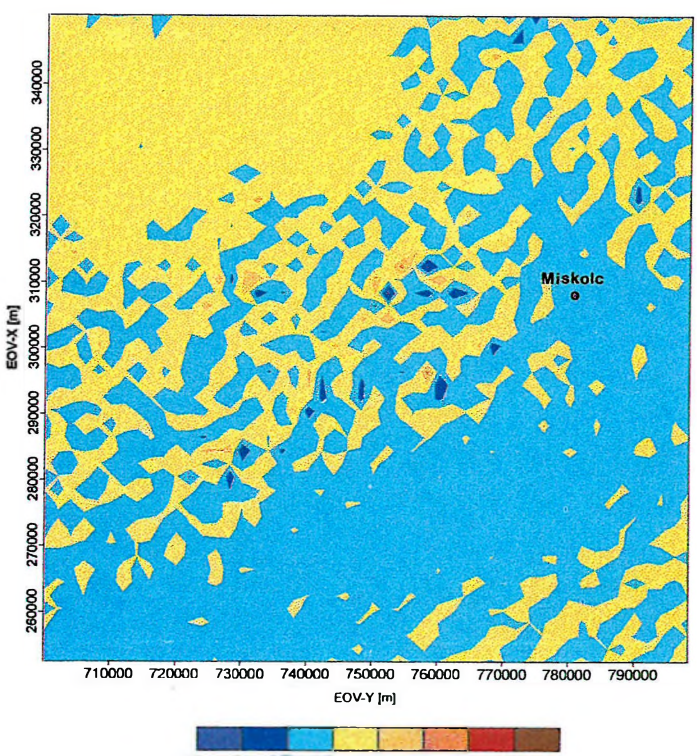


Fig. 17. Region 'B': Effect of *DF* in the β component of gravimetric correstions
17. ábra. "B" terület: A *DF* hatása a gravimetriai korrekciók β irányú összetevőjének változására

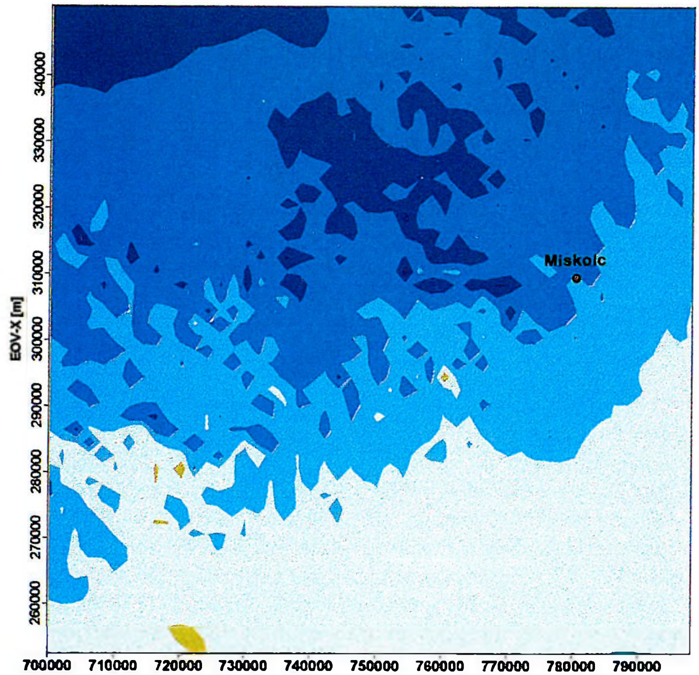


Fig. 18. Region 'B': Map of α component of gravimetric correstions calculated from 1995 database

18. ábra. "B" terület: A gravimetriai korrekciók α irányú összetevőjének térképe az 1995. évi adatbázisból történt meghatározásából

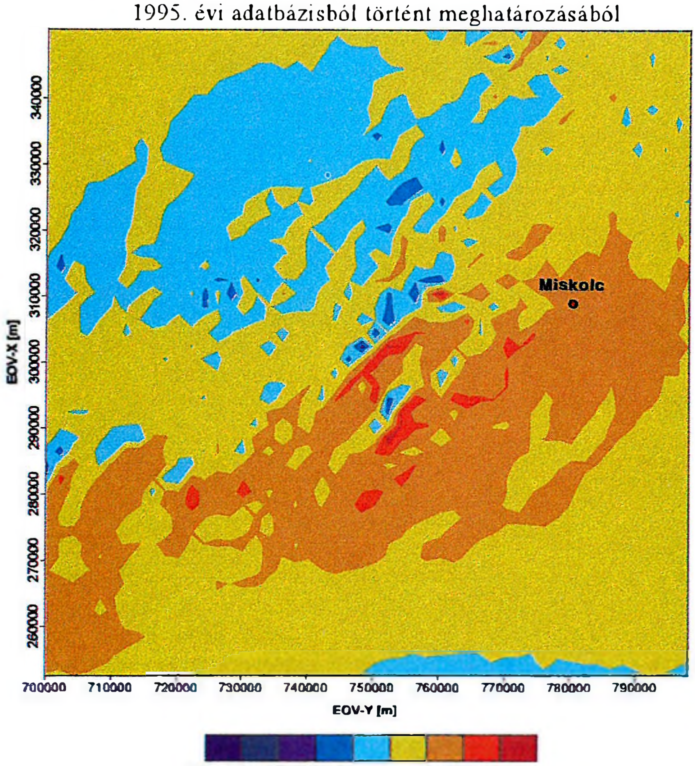


Fig. 19. Region 'B': Map of β component of gravimetric correstions calculated from 1995 database

19. ábra. "B" terület: A gravimetriai korrekciók β irányú összetevőjének térképe az 1995. évi adatbázisból történt meghatározásából

Region 'B':

DF (*Fig. 9*): In the mountainous part of the region the Faye–anomalies are negative with minimum values of even −15 mGal, but in area B4, which is of a lowland character, there are only a few low level isolated anomalies.

DN (*Fig. 10*): Geoid undulation variation strictly follows the DF < 0 distribution pattern and it is negative everywhere in correlation with the decrease of the gravity field. *The effect of data densification here is the most significant: undulation changes amount to as much as −6 cm.*

DΘ″ (*Fig. 11*): This is one of the most characteristic of our figures: a definite negative anomaly ring encircles the mass of Bükk Mountain. At the southern base of the Mátra there seems to be only an elongated negative anomaly. This is the result of mass deficit and it is compensated by the great northern masses that lie north of Bükk and mainly north of the Mátra Mountains.

DΘ^{+} (*Figures 12, 13*): The decrease of gravity anomalies causes positive changes in the deflection of the vertical south of the mass deficit and negative ones north of it. In agreement with this rule the Mátra–Bükk (areas of B1, B2) and the Szendrő–Cserehát (area B3) also have their positive regions of influence of approximately 1.5 cm magnitude whereas between them in the Sajó valley there seems to be a negative band of −1.0 cm. The approximately 90° rotation of the variations along the SW–NE strike direction and the SE–NW transverse direction underlines this fact.

DΘ^{-} (*Figs. 14, 15*): Curvature changes in both α and β directions are within the noise level, but their stripped pattern is very characteristic. In the case of the α component the axes of the strips are perpendicular to the strike of the mountains (Fig. 14) whereas the axes of the β components are parallel to it. Since their line of contact is an axis of inflection, the geoid surface of region 'B' is composed of convex and concave pieces arranged in a chessboard–like manner.

D(GK) and ***(GK)95*** (*Figs. 16–19*): The figures of these quantities have to be examined in pairs (Figs. 16, 18 and Figs. 17, 19). It is noteworthy that areas of *(GK)*95 > 0 in Fig. 16 correspond to spots of *D(GK)* < 0 in Fig. 18 and similar but not so dominant characteristics can be seen in Figs. 17 and 19. A detailed examination of this phenomenon should provide further information about the local features of the geoid, but it must be kept in mind that this phenomenon is within the noise level.

It can be concluded that the densification of the gravity stations provided lower gravimetric corrections, i.e. decreased the potential errors, thereby supporting the improving effect of detailed gravity surveys in geoid determination.

11. Conclusions and further tasks

Our studies have shown that increase of the number of gravity data during the years 1993–95 has made possible to refine the local part of the gravimetric geoid providing a tool to promote the practical application of GPS technique. The improving effect of data-densification is shown by the general conformity of the quantities examined — compatibility of the sets of *DF*, *DN* and *D(GK)* values (inner consistency), correlation of these data sets with the topography (outer consistency). The work described in this paper was only the first step towards applying the regression method to analyse the links between complex interdependent quantities of physical geodesy. Based on all the above mentioned facts the following can be concluded:

It was seen that at different test areas — mainly because of different geomorphological features — the influence of densification may differ by orders of magnitude. Joint study of variance and regression–correlation analysis is suggested to make progress from variance analysis towards factor analysis [HAJTMAN 1968].

The Faye-anomaly maps represent not only the effect of topography but the effect of subterrain masses as well. Therefore it seems advisable to extend the investigations to such relatively flat regions where high gravity anomalies of subterrain origin exist.

The stochastic investigations carried out recently in the Geodetical and Geophysical Research Institute at Sopron provided a sound basis for the numerical study of the relationship between the density of gravity data, the related statistical parameters, and the gravimetric correction in the field of geoid determination [PAPP 1993].

Acknowledgements

The authors express their thanks to Mr. S. Marczinus for the computer program of correlation and regression analysis and for making the computations. Thanks are also due to the Hungarian–American Research Fund for providing financial support under contract No. JF–369, thereby enabling extra gravimetric measurements to be carried out in the Tamási region.

REFERENCES

ÁDÁM J. 1993: Global Geopotential Models in the Region of Hungary. Periodica Polytechnica, Serie: Civil Engineering **37**, 2, pp. 69–90, Techn.Univ. Budapest

BÍRÓ P. 1985: Geodesy. (in Hungarian) Tankönyvkiadó, Budapest,. pp. 171–174

CSAPÓ G. 1994: Report for 1994 on the year's activity in the field 'Geodetic gravimetry' (in Hungarian). ELGI unpublished report, Budapest

DETREKŐI Á. 1991: Adjustment computations. (in Hungarian) Tankönyvkiadó, Budapest, pp. 65–80, 123–130

GAZSÓ M., TARASZOVA G. 1984: Astro-gravimetric determination of the quasigeoid in Hungary. (in Hungarian). FÖMI publication, V. évf., Budapest

HAJTMAN B. 1968: Introduction to mathematical statistics – for psychologists (in Hungarian). Akadémiai Kiadó, Budapest, pp. 172–232

KOVÁCSVÖLGYI S. 1994: ELGI's gravimetric database (in Hungarian). Magyar Geofizika **35**, 1, p. 44–46

MOLODENSKI M.S., EREMEEV V.F., YURKINA M.I. 1962: Methods for study of the External Gravitational Field and Figure of the Earth. Works of the Central Research Institute of Geodesy, Aerial Photogr. and Cartography, Jerusalem, No.131, pp. 125–142

MUNDRUCZÓ Gy. 1981: Applied regression analysis (in Hungarian). Akadémiai Kiadó, Budapest, pp. 13–62

PAPP G. 1993: Trend models in the least-squares prediction of free-air gravity anomalies. Periodica Polytechnica, Serie: Civil Engineering **37**, 2, pp. 109–130

SÁRHIDAI A. 1993: Comparison of different methods for interpolating Faye-anomalies (in Hungarian). ELGI, unpublished report, Budapest

SÁRHIDAI A. 1994: Compilation and study of interpolated gravity data set of Hungary and its surroundings (in Hungarian). ELGI unpublished report, Budapest

Összefüggés a geoid magyarországi felületdarabjának megbízhatósága és a gravitációs adatok területi sűrűsége között

CSAPÓ Géza, GAZSÓ Miklós, KENYERES Ambrus

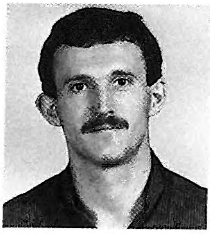
A dolgozatban abból az alapelvből indultak ki a szerzők, hogy miután egyetlen földi nehézségi erőtér létezik, így geoid is csak egy lehetséges, függetlenül attól, hogy ezt a felületet milyen matematikai–technikai apparátussal állítják elő. Tekintettel arra, hogy a gravitációs mérések számának növelése a nehézségi erőtér jobb felbontását eredményezi az adott helyen, ezért ott a geoidundulációknak a sűrítés következtében előállt változásainak eleve a megbízhatóság növekedésével kell járnia, mivel ugyanarról a fizikai valóságról van szó. A vizsgálatok a reprezentációs hibára és a modell hibára terjedtek ki. Jelenleg nem ismeretesek a helyi nehézségi erőtér statisztikai természetét leíró kovarianciák, ezért vizsgálati mennyiségeknek a Faye–anomáliák, a geoid undulációk, a függővonalelhajlás abszolút értékei, a gravimetriai korrekciók, a gravimetriai korrekció függővonalelhajlási összetevői és a görbületi értékek változását választották — az ELGI 1993-as és 1995-ös gravimetriai adatbankjára támaszkodva. A gravimetriai korrekció potenciál hibaként értelmezett vizsgálata beváltotta a hozzá fűzött várakozást, mert hatása a geoid ± 1–2 cm nagyságú ingadozásait eredményezte. Lévén ez a mennyiség a lokális alakváltozások mértéke, egyben a geoidmegjelenítés finomodását, a felbontóképesség növekedését mutatja, tehát a "GPS geoid" vonatkozásában a gyakorlati alkalmazhatóság reálisabbá válásának mértékét is. Ugyanis a 30–70 km sűrűségben elosztott referenciapontokon a GPS mérésekből származó ellipszoidi magasságok és a felsőrendű szintezés szolgáltatta ortométeres magasságok különbségéből adódó, ún. "GPS–geoidmagasságok" feldolgozásával ki lehet szűrni a gravimetriai geoid globális (hosszúhullámú) összetevőjének regionális hibáit. Így éppen a földi gravimetriai méréseken alapuló lokális (rövid- és középhullámú) rész pontossága lesz az alkalmazások szempontjából a meghatározó!

ABOUT THE AUTHORS

Géza Csapó, for a photograph and biography, see this issue, p. 81.

Miklós Gazsó graduated as a civil engineer from the Technical University of Budapest in 1950. From 1950 to 1956 he was a surveyor. He then joined the Institute of Geodesy, Cartography and Remote Sensing as a research engineer, working there from 1956-1974. From 1974 to 1997 he was a senior research engineer.

He completed his Ph. D. in geodetic gravimetry in 1986. The main topics of his research and other acitvities embrace the local geoid, recent crustal movements and network adjustment.



Ambrus Kenyeres graduated as a geophysicist from the Lorand Eötvös University in 1986. After graduation he joined the Institute of Geodesy, Cartography and Remote Sensing as a scientific officer. His field of activity is connected with the study of the local geoid, recent crustal movements including the related field work, especially the application of the GPS technique in geodynamic research.

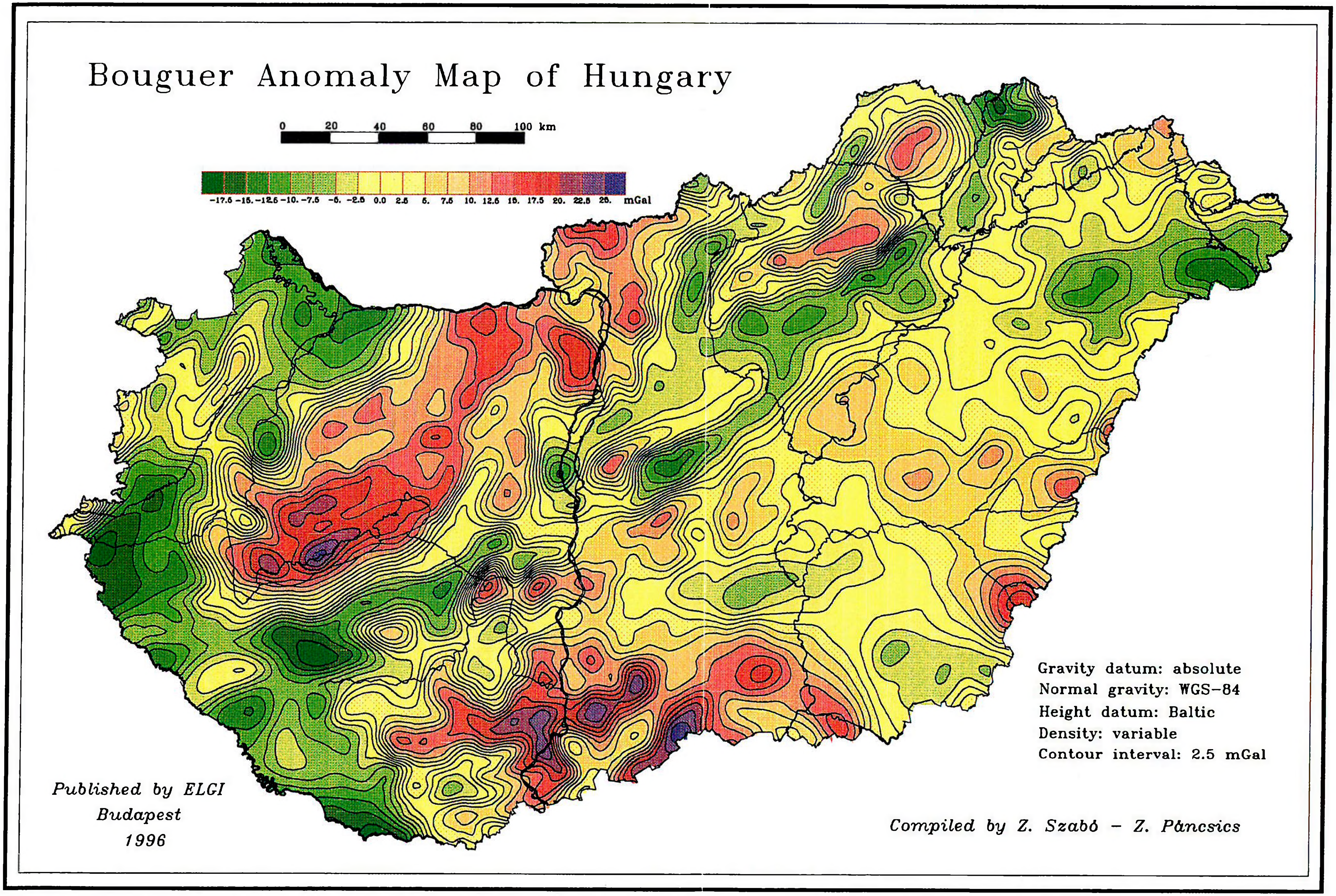

Bouguer Anomaly Map of Hungary
0 20 40 60 80 100 km
-17.5 -15. -12.5 -10. -7.5 -5. -2.5 0.0 2.5 5. 7.5 10. 12.5 15. 17.5 20. 22.5 25. mGal
Gravity datum: absolute
Normal gravity: WGS-84
Height datum: Baltic
Density: variable
Contour interval: 2.5 mGal
Published by ELGI
Budapest
1996
Compiled by Z. Szabó - Z. Páncsics

Fig. 2. Bouguer anomaly map of Hungary
2. ábra. Magyarország Bouguer anomália térképe

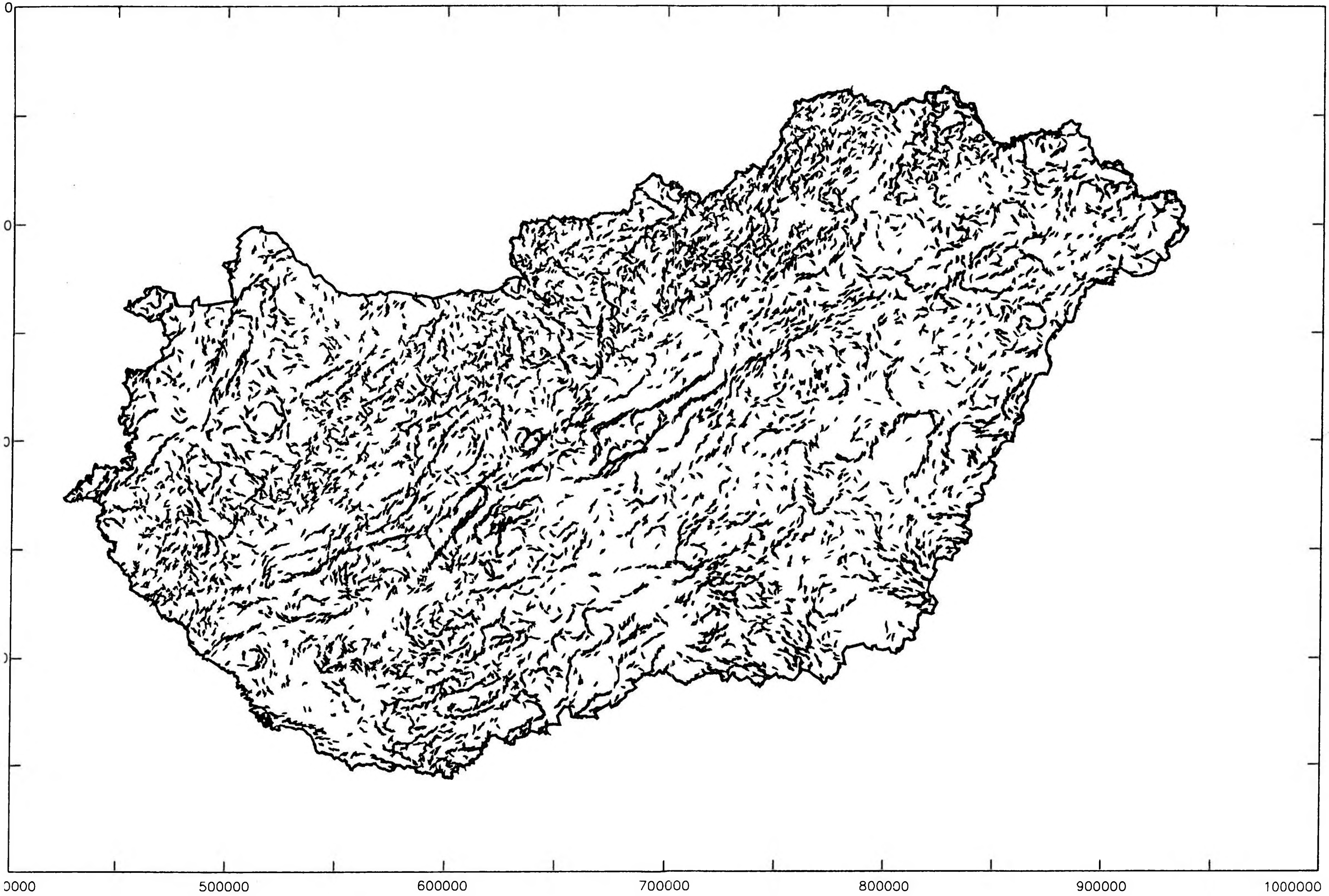

500000
600000
700000
800000
900000
1000000

Fig. 4. Gravity lineaments, hatched version
4. ábra. Gravitációs lineamensek vonalkázott változata

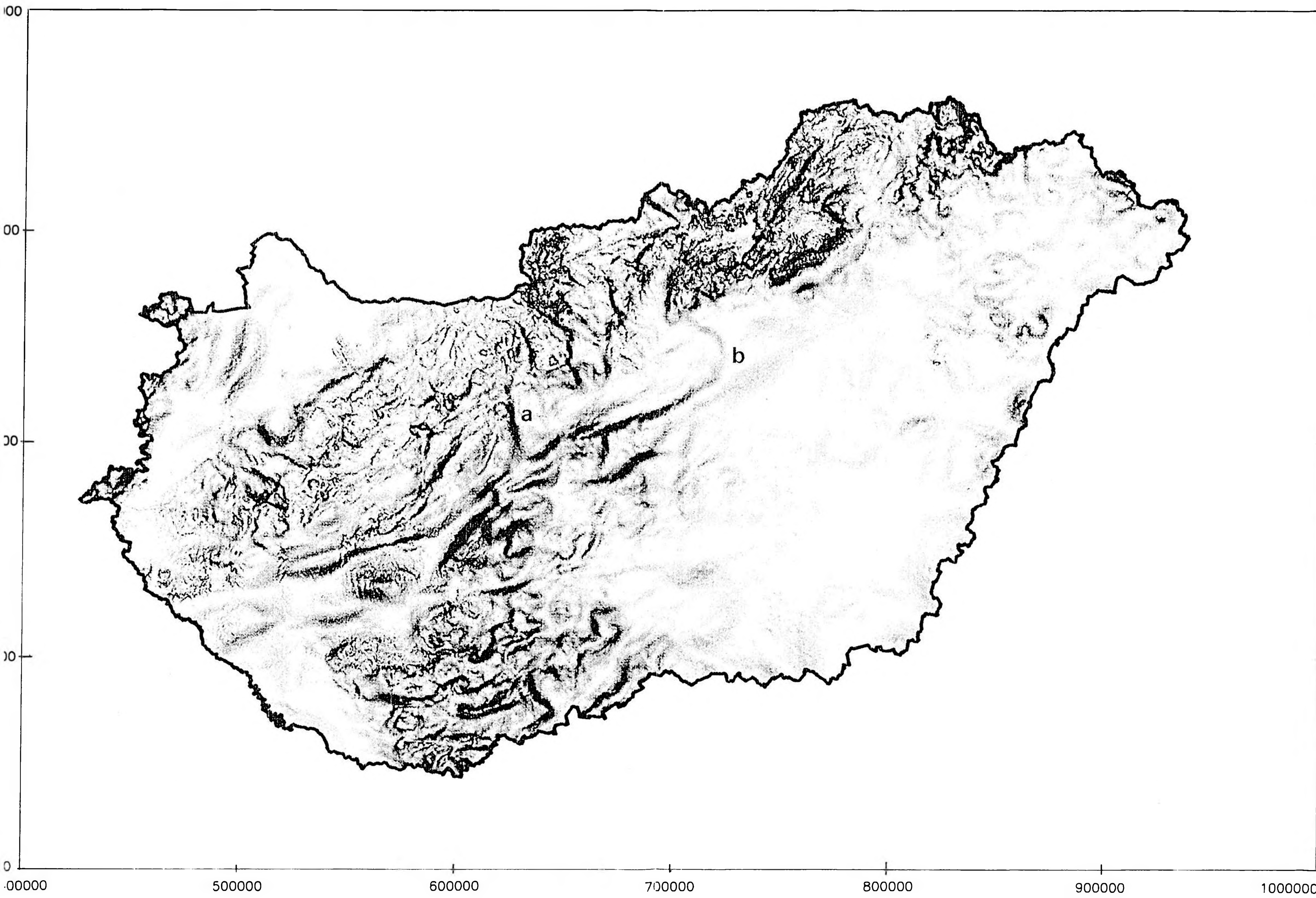

a
b
500000
600000
700000
800000
900000
1000000

Fig. 5. Gravity lineaments, shaded version

5. *ábra*. Gravitációs lineamensek árnyékolt változata

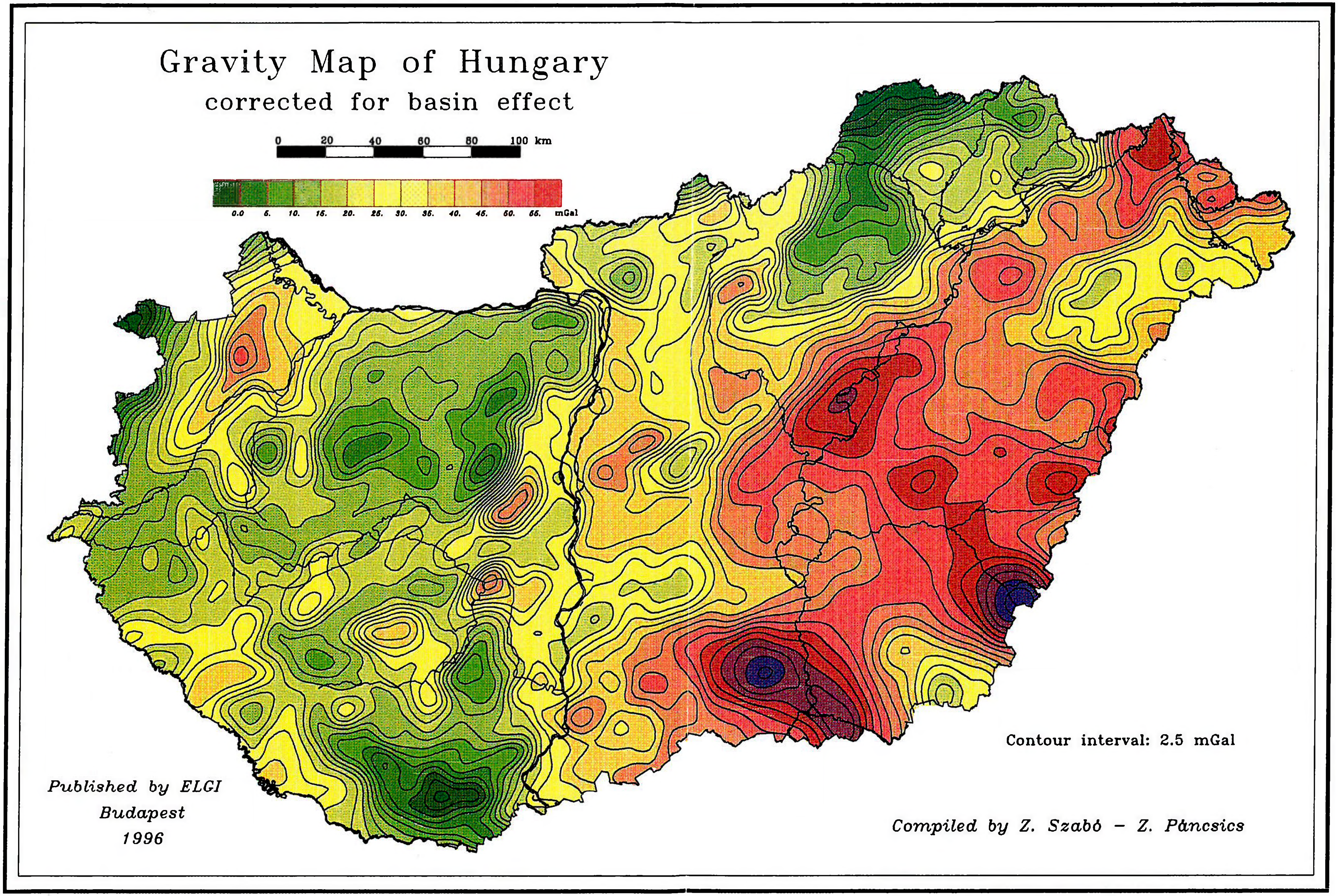
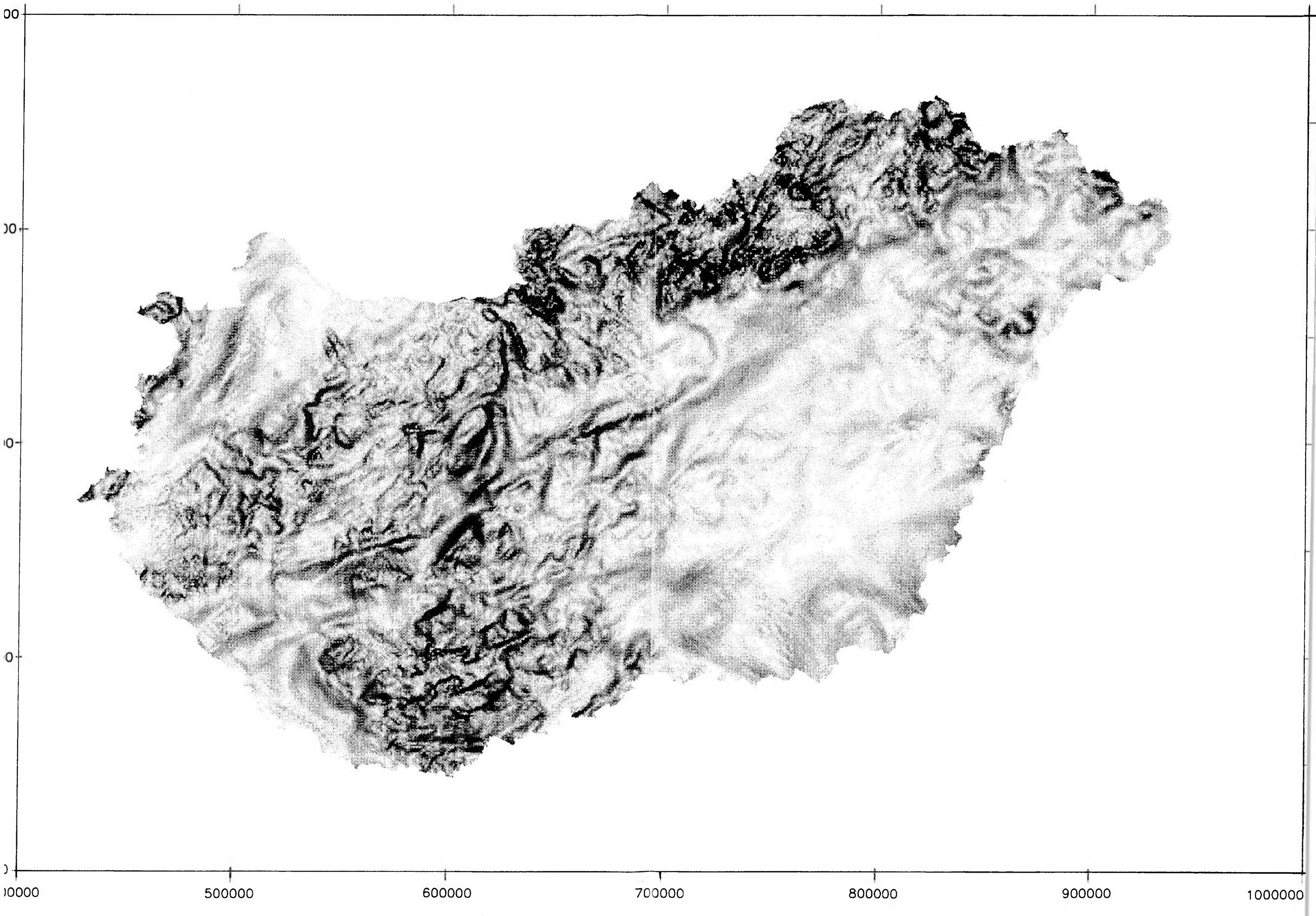

Gravity Map of Hungary
corrected for basin effect
0 20 40 60 80 100 km
0.0 5. 10. 15. 20. 25. 30. 35. 40. 45. 50. 55. mGal
Contour interval: 2.5 mGal
Published by ELGI
Budapest
1996
Compiled by Z. Szabó – Z. Páncsics

Fig.3. Gravity map of Hungary corrected for basin effect

3. ábra. Magyarország medencehatástól mentesitett gravitációs térképe



500000
600000
700000
800000
900000
1000000

5

Fig. 4. Gravity lineaments determined from corrected map

4. ábra. Medencehatástól mentesített térképből meghatározott gravitációs lineamensek

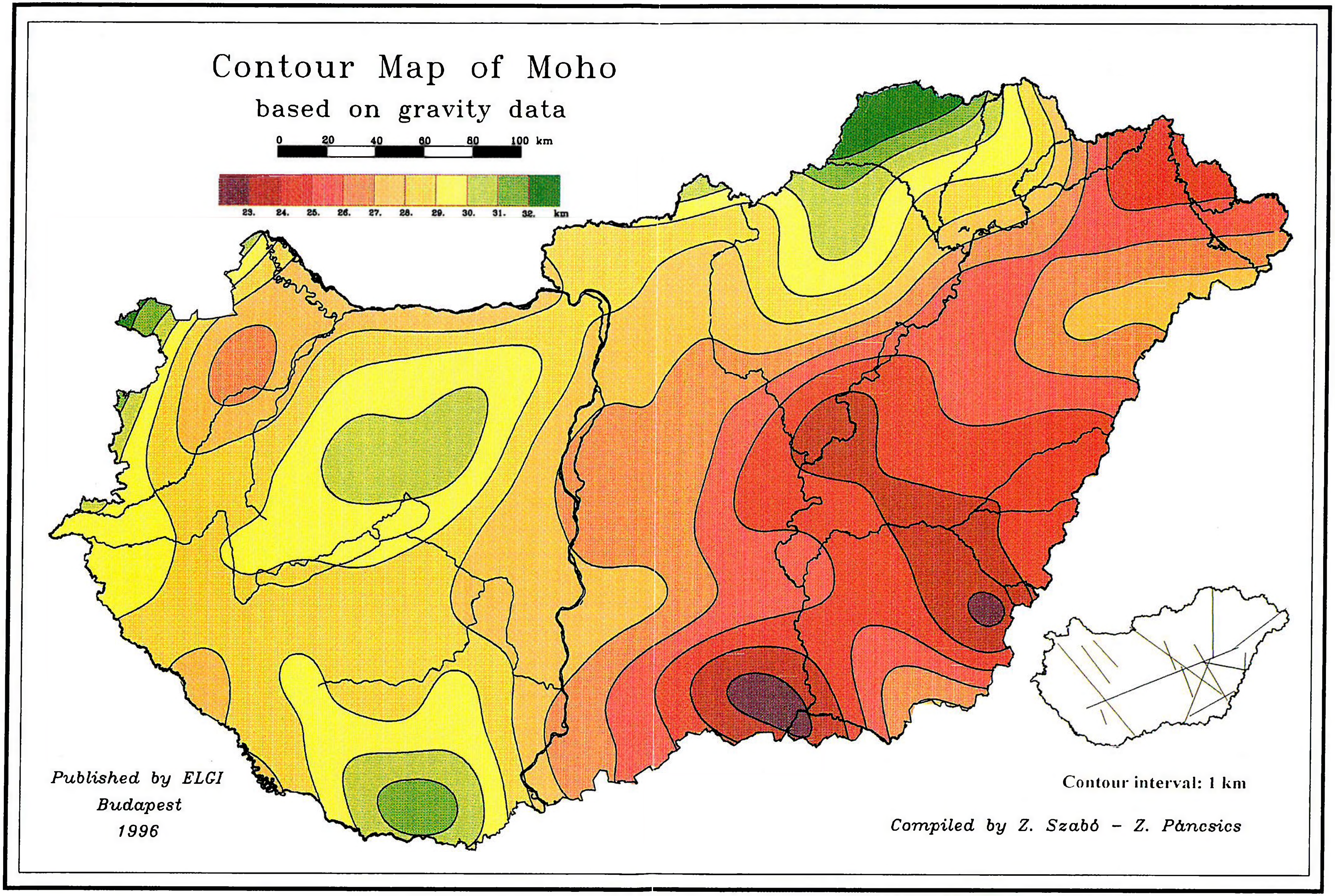
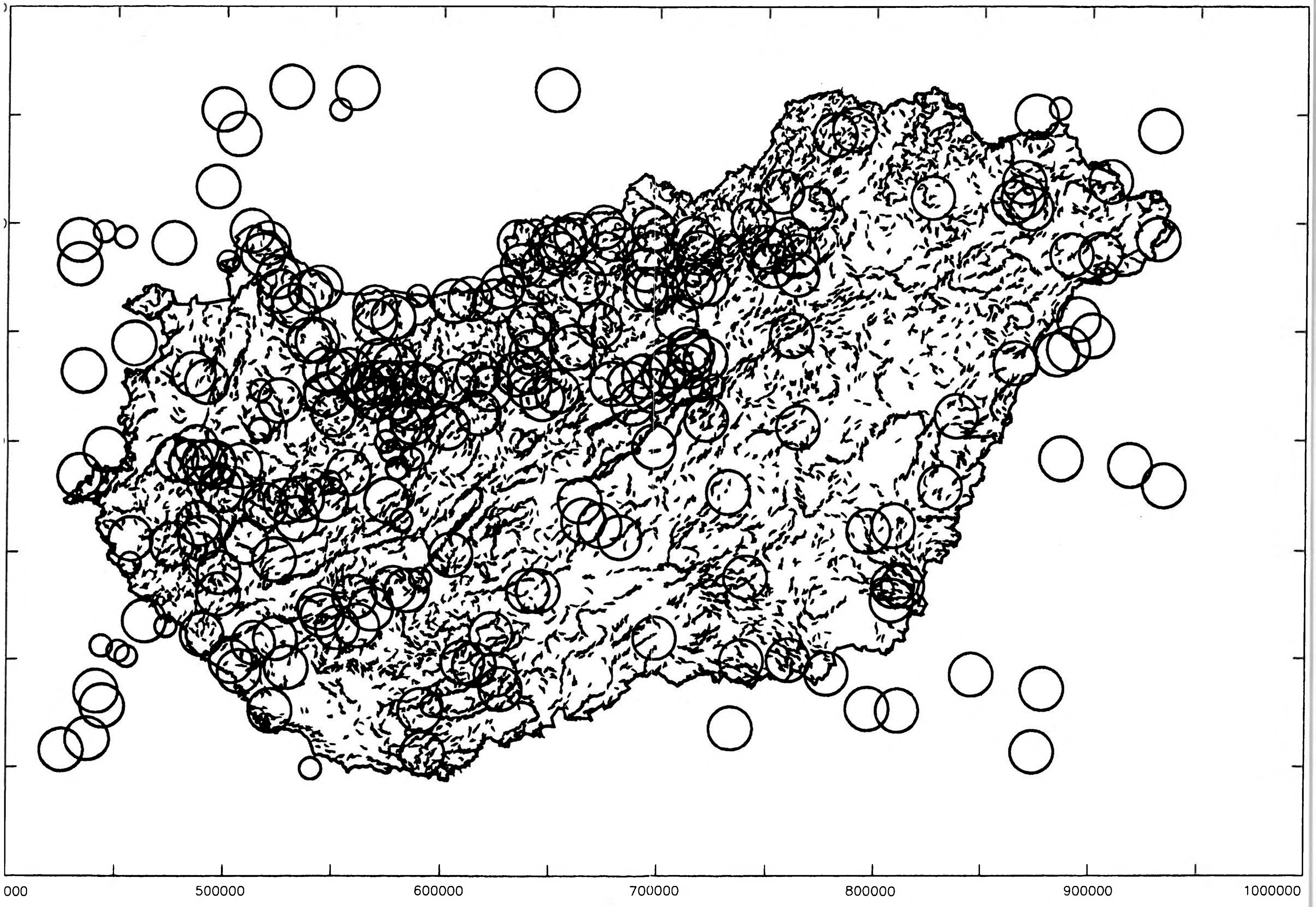

Contour Map of Moho
based on gravity data
0 20 40 60 80 100 km
23. 24. 25. 26. 27. 28. 29. 30. 31. 32. km
Published by ELGI
Budapest
1996
Contour interval: 1 km
Compiled by Z. Szabó – Z. Páncsics

Fig. 5. Contour map of the Moho based on gravity data

5. ábra. Gravitációs adatokból számított Moho térkép


000
500000
600000
700000
800000
900000
1000000

1

Fig. 3. Gravity lineaments and the epicentres of historical earthquakes

3. ábra. Gravitációs lineamensek a történelmi rengések epicentrumainak feltüntetésével

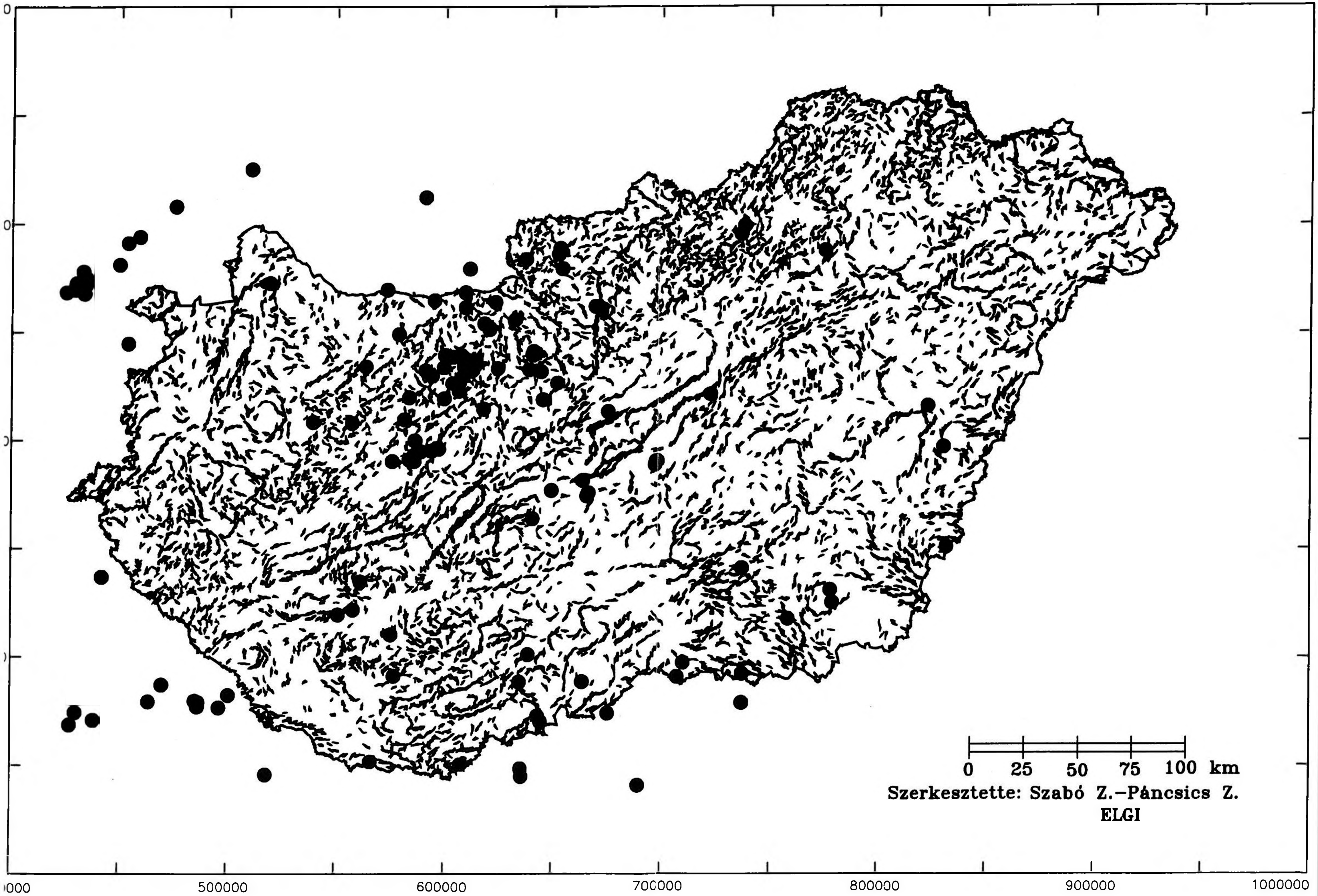

0 25 50 75 100 km
Szerkesztette: Szabó Z.–Páncsics Z.
ELGI
500000
600000
700000
800000
900000
1000000

Fig. 4. Gravity lineaments and the epicentres of earthquakes registered by the Microseimic Monitoring Network in 1995–1998

4. ábra. A gravitációs lineamensek térképe a Mikroszeizmológia Megfigyelő Hálózat által 1995–1998-ban regisztrált földrengések feltüntetésével

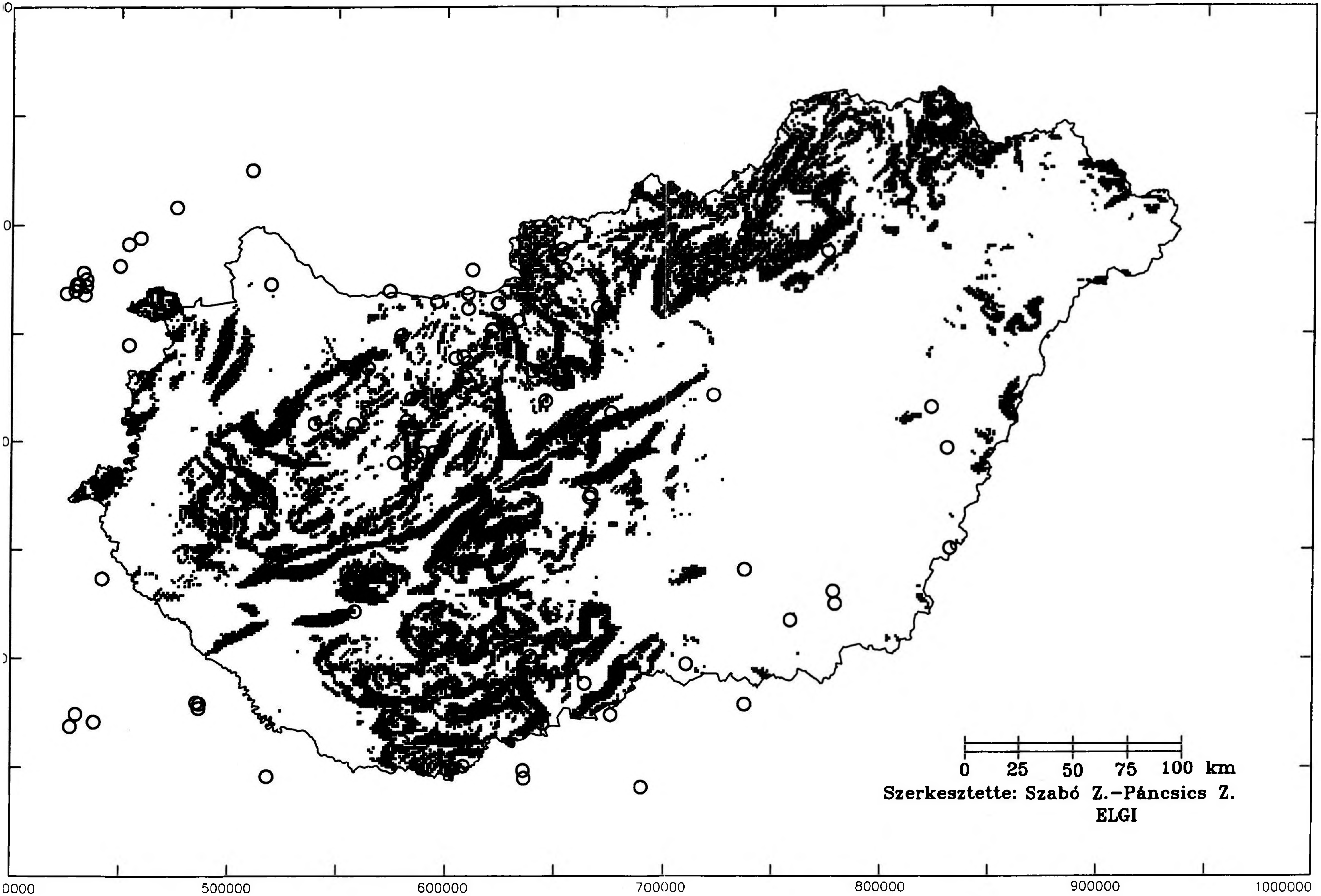

0 25 50 75 100 km
Szerkesztette: Szabó Z.–Páncsics Z.
ELGI
0000
500000
600000
700000
800000
900000
1000000

Fig. 6. Belt of horizontal gravity gradients higher than 15 E unit and instrumentally located epicentres of earthquakes recorded from 1995 to 1998

6. ábra. 15 E egységnél nagyobb horizontális gravitációs gradiensű zónák és az 1995–1998 között regisztrált földrengés epicentrumok

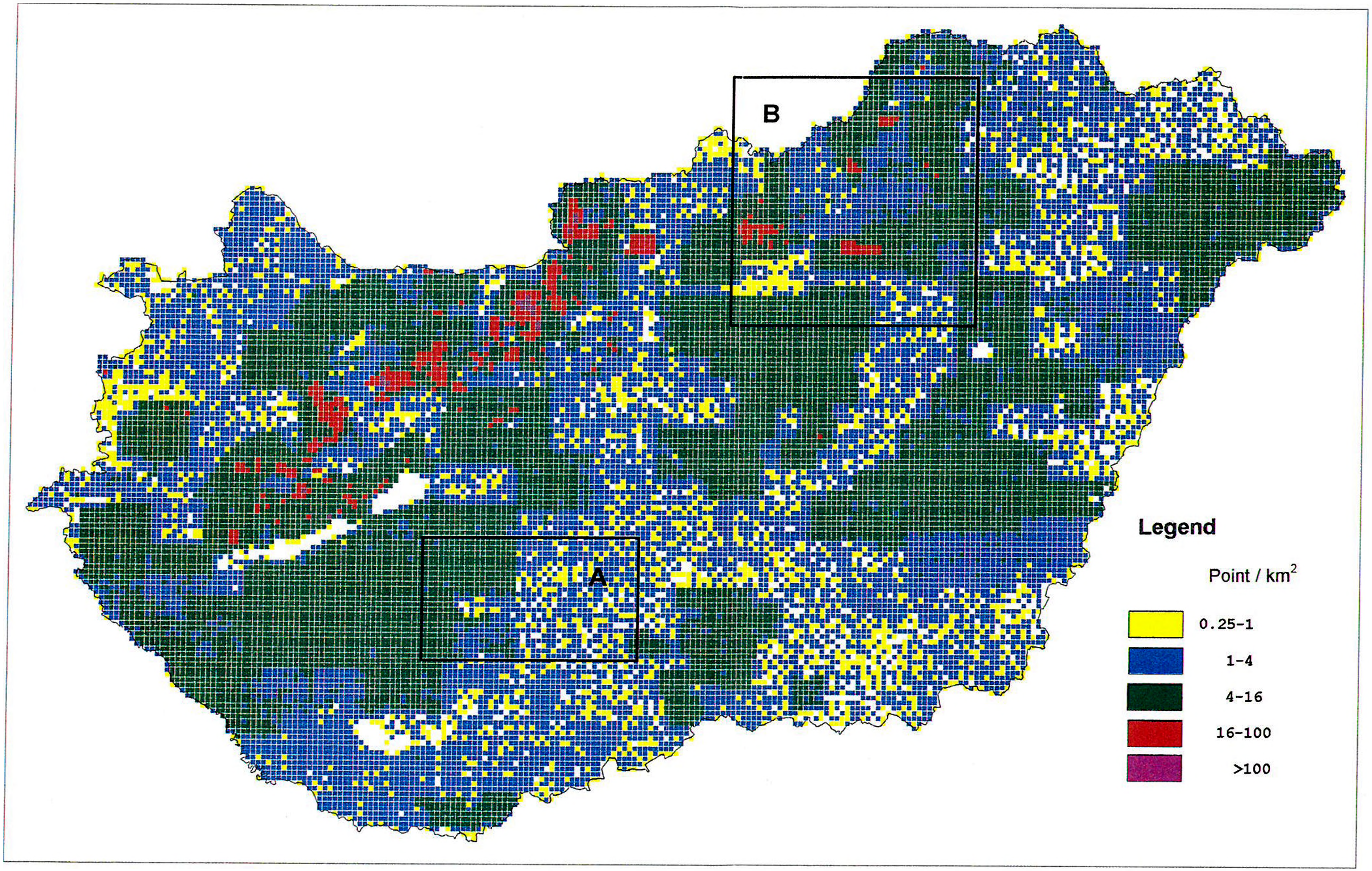

B
A
Legend
Point / km2
0.25-1
1-4
4-16
16-100
>100

Fig. 2. Spatial distribution of gravity stations in Hungary and location of test regions

2. ábra. Magyarország gravimetriai felmértségi térképe a vizsgálati területekkel

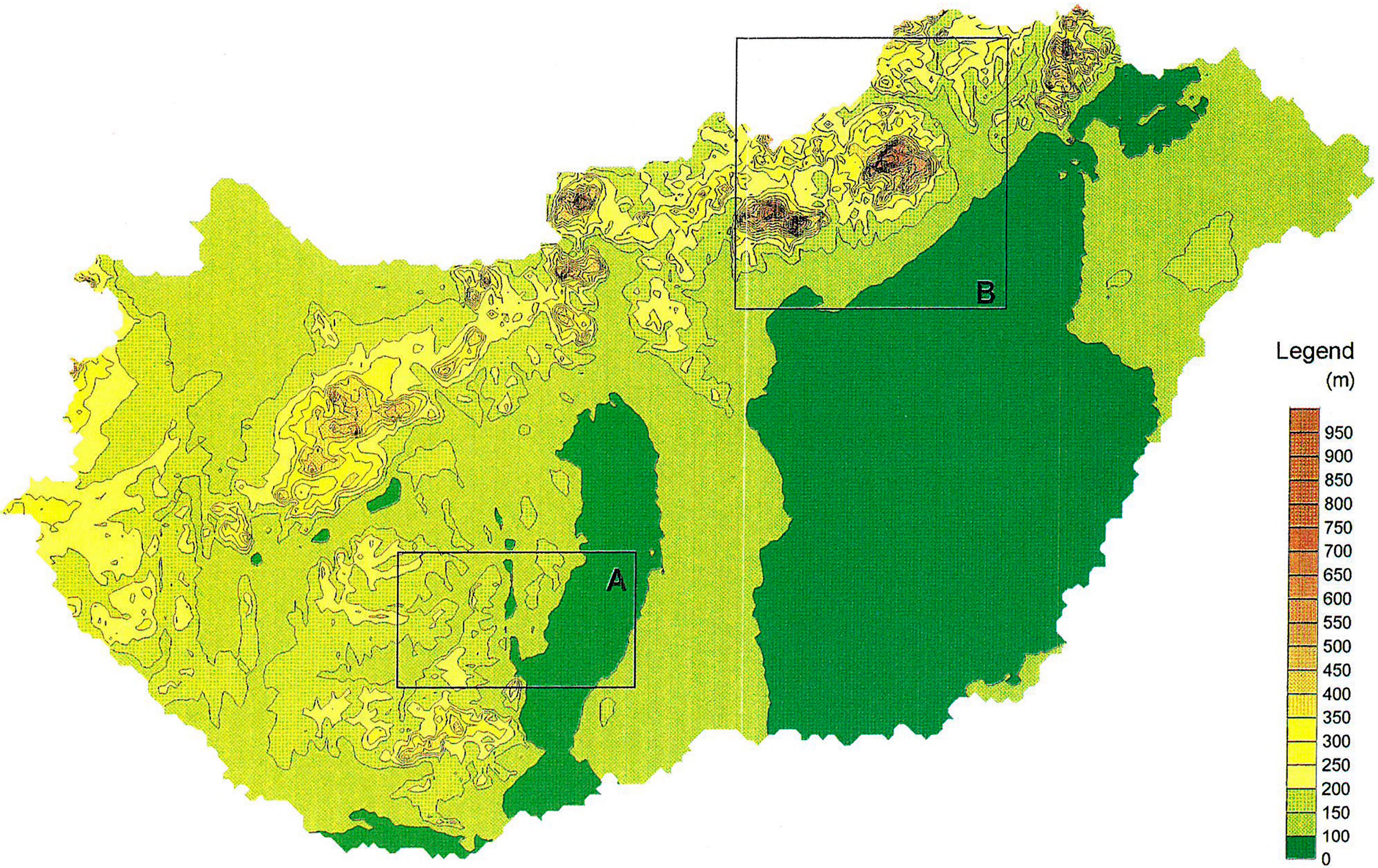

A
B
Legend
(m)
950
900
850
800
750
700
650
600
550
500
450
400
350
300
250
200
150
100
0

Fig. 3. Topographic map of Hungary — and location of test regions
3. ábra. Magyarország domborzati térképe a vizsgálati területekkel