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EOCENE SEDIMENTARY FORMATIONS AND SEDIMENTATION IN THE BAKONY MOUNTAINS, TRANSDANUBIA, HUNGARY

By

E. DUDICH

LABORATORY FOR GEOCHEMICAL RESEARCH OF THE HUNGARIAN ACADEMY OF SCIENCES, BUDAPEST

The author has split up the Eocene of the Bakony Mountains into 9 formations (Fig.1) within which he has distinguished members. The formations have been characterized on the basis of the following analyses: grain size analysis, total carbonate and dolomite content, flame photometric determination of sodium and potassium oxide contents, wet chemical silicate analysis, emission UV spectrography of oligo- and micro-elements, DTA-DTG (derivatography), X-ray diffractometry of clay minerals, micro-mineralogy of the heavy and light mineral fractions, determinations of the C^{org} content, microfacies studies of thin sections.

The results are contained in Tables 1 to 3.

An evaluation of the facies types, with due regard, among other things, to data on fossils, is given in Tables 4—5.

The sedimentation history of the four distinct subareas has been sketched in Figs 2 to 5.

The Eocene stage of the Bakony Mountains on the whole has been illustrated in block-diagrams 6—7.

The Eocene history of evolution succeeded, as the third cycle, to two major, Mid- and Upper Cretaceous, sedimentary cycles, showing common features with these. Latest Eocene emergence, with the establishment of a terrestrial regime, marked the advent of an entirely new, different stage of evolution.

The tripartite Mid-Cretaceous to Eocene megacycle can be interpreted in terms of plate tectonics.

1. Introduction

The earliest papers on the Eocene series ("Nummulitenformation") of the Bakony area date back as far as 1862. The results of almost a century were synthetized in a monograph by E. SZŐTS in 1956 [19], which was of biostratigraphic character, based mainly on molluscs.

The years 1957—69 witnessed the first stage of a renewed, up-to-date geological surveying of the area (scale 1 : 20 000) by the Hungarian Geological Institute. The Eocene formations disclosed by several thousand boreholes exploring (Eocene) brown coal and (Cretaceous) bauxite deposits, were studied in a complex way by a team of Hungarian specialists headed by G. KOPEK. The present writer joined in 1962. Several foreign experts from Czechoslovakia, Italy, Rumania, USSR and Yugoslavia kindly contributed to the better understanding of the Bakony Mountains Eocene, mostly in the field of foraminiferal biostratigraphy.

In 1969, the latest stratigraphic and (partly) lithologic results were presented at the international Colloquium on Eocene Stratigraphy held in Budapest on the occasion of the centenary celebrations of the Hungarian Geological Institute. Along with several partial reports, a summarizing paper was published in the *Proceedings of the Colloquium* by G. KOPEK, E. DUDICH and T. KECSKEMÉTI [11]. Moreover, the *Proceedings* include a comprehensive bibliography of the Hungarian Eocene up to July 1969 (569 items).

Along with and following the prevailing stratigraphic investigations, important work has been done in the fields of lithology and geochemistry as well. E. g.: Eocene volcanism by V. SZÉKY-FUX—A. BARABÁS, 1953 [18]; geochemistry of rare elements in the brown coal basins of the NE-Bakony, by G. CSÁSZÁR, 1969 [2] and L. ÓDOR, 1969 and 1971 [13, 14]; glauconitic rocks by L. BONDOR, 1971 [1], micromineralogy by E. RADÓCZ-KOMÁROMY, 1971 [15]; lithology of a SW-Bakony key section by T. KECSKEMÉTI and A. VÖRÖS, 1975 [8]; coal petrography by I. IHAROS-LACZÓ, 1971 [7].

Several papers dealt with the comparison of the Hungarian Eocene with the Eocene of other countries, above all of the adjacent ones [17—23 and 42].

The present writer tried to collect and to synthesize all available lithological and geochemical evidence, in order to describe the discernable lithostratigraphic formations, following the recommendations of the *Hungarian Stratigraphic Committee* [48], to establish the sedimentation pattern and to decipher the paleogeographic evolution of the Bakony Mountains Eocene.

Unpublished mapping reports, laboratory data and the explanatory texts of geological maps (both published and unpublished) have also been consulted. The work could be done only with the collegial support obtained from the competent authorities of the *Hungarian Geological Institute*, the *Bauxite Prospecting Company* and the *Laboratory for Geochemical Research of the Hungarian Academy of Sciences* as well as the friendly co-operation of numerous colleagues. Help of any kind is to be acknowledged with sincere thanks.

2. Discussion

2.1. *Geological Setting*

The study area is a typical example of the "mobile shelf" region of the epicontinental facies province ("inner intermediary" IVth zone, see N. MÉSZÁROS, E. DUDICH [20].

The Eocene series of about 550 m maximum thickness is underlain by (mostly carbonate) rocks of Triassic, Cretaceous and subordinately of Jurassic

age. Where buried, it is overlain by Oligocene, Miocene and/or Pliocene sediments.

The Bakony Mountains Eocene is characterized by a great variety of detrital (clastic), chemo- and biogenic, and even of pyroclastic elements of sedimentation. Accordingly, rock types range from coarse conglomerates through more or less carbonatic sandstones, siltstones and shales to marls and eventually dolomitic, mostly biogenic limestones. Brown coal seams and bauxitic clays are of peculiar interest.

Biostratigraphy has been well established by numerous authors by means of benthonic larger Foraminifera, benthonic and planktonic smaller Foraminifera, Nannoplankton, Mollusca and spores-pollen.

2.2. *Methods of Study*

More than 80 selected boreholes and four key-sections outcrops have been investigated. The following methods have been used: grain size analysis, chemical determination of the total carbonate and dolomite contents, flame photometric determination of sodium and potassium oxide contents, wet chemical silicate analysis, emission UV spectrography of oligo- and micro-elements, DTA-DTG (derivatography) and X-ray diffractometry of clay minerals, micromineralogy of the heavy and light mineral fractions, determination of the C^{org} content, microfacies studies of thin sections. Hydrogeological and geophysical well-logging data have also been taken into consideration.

2.3. *Lithostratigraphic Formations*

Altogether nine formations have been proposed (Fig. 1).

1. *Darvastó Formation* (chiefly non-marine Lower Eocene)
   1. Bauxitic clay member
   2. Lower clastic member
   3. Limestone member
   4. Upper clastic member
2. *Devecser Limestone Formation* (marine Lower Eocene)
3. *Kisgyón Formation* (essentially non-marine Middle Eocene)
   1. Variegated underclay member
   2. Lower coal-bearing member
   3. Variegated interclay member
   4. Upper coal-bearing member
   5. Marl member

   K Fehérvárcsurgó member (of regional character)

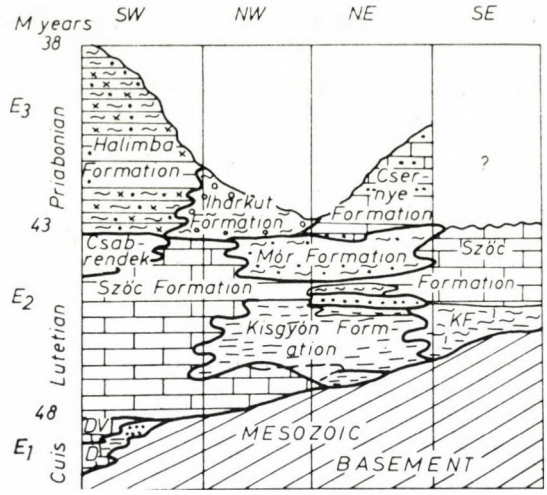

Fig. 1. Time-space interrelations of the formations in the Bakony Mountaines Eocene

4. ***Szőc Limestone Formation*** with four regional members:
   Balatonhegy (SW) member
   Weimpuszta (NW) member
   Dudar (NE) member
   Iszkaszentgyörgy (SE) member
5. ***Mór Shale Formation*** with three regional members:
   Olaszfalu (NW) member
   Dudar (NE-marginal) member
   Sur (NE-basin) member
6. ***Csabrendek Marl Formation*** (fore-reef Upper Middle Eocene)
7. ***Halimba Tuffitic Formation*** (pelagic Upper Eocene)
   (alternation of tuffitic sandstones and shales)
8. ***Csernye Formation*** (sublittoral Upper Eocene)
   (alternation of Lithothamnian limestones and tuffitic sandstones)
9. ***Iharkut Formation*** (littoral Upper Eocene)
   (conglomerate and red clay)

The characteristic parameters of the formations and members are summed up in Tables 1—3.

2.4. *Facies Types*

With due attention paid to the fossil record, the environmental factors of deposition have been analyzed (see Table 4) and a facies classification has been attempted (Table 5).

Table 1

Characteristic Values of Petrological and Geochemical Parameters I

Parameter	Darvastó Formation				Kisgyón Formation						Szőc Formation				Mór Formation			Csr	H	Cs
	1.	2.	3.	4.	1.	2.	3.	4.	5.	KF	SW	NW	NE	SE	NW	NE_{shore}	NE_{basin}			
Thickness (m)	0 5	5–25	4–8	0–4	1–39	3–11	4 32	1 39	2–39	24–75	34–200	17–56	8–100	13–90	10–39	16 128	200	0 87	(24) 275	4–78
Clay %	68–95	10–60	10–62	33–42	30–37	9–50	2–74	9–63	3–68	51–80	20–45	5–86	6–84	23–68	7–40	15–40	11–50	14–61	6–32	3–27
Silt	5–16	23–66	26–58	52–58	63–70	38–70	8–82	27 82	6–76	18–42	28–60	6–75	12–69	23–72	14–67	22–85	40–98	33–68	18–85	10–76
Sand	0–8	0–66	0–63	9–15	0.1–0.2	0.1–42	0.1 82	1–61	1–87	1–86	<10	0–94	1–82	2–30	1–76	0 66	0–30	1–38	1–76	1–74
Gravel	0	0–10	0	0	0–10	0	0–23	0–9	0	0	0–75	0–49	0	0	0	0	0	0	0–5	0
$Cycle_1$, m							25–40		30–40			30–40			30	40–68			200	
$Cycle_2$			6			11	8–10	10	7–15	10–15		10–15				8–16				7–16
$Cycle_3$			2				4–5	2–4	3–7			3–7	3–7		4–7			4–8		2–4
Calcite %	0–12	0–90	10–96	14–22	1–38	0–34	0–100	0–85	8–97	78–100	59–99	2–99	21–92	3–94	4–47	39–79	16–49	48–83	tr.–58	1–26 alt.
Dolomite	0	0–26	1–4	2–4			1–16	1–4			2–18	2	2	0–15		+	1–16			42–94 tr.
C/D ratio		0.4–100	6–100	5–8							4–38	53	39							
$Cycle_1$ in $Cycle_2$ metres $Cycle_3$												30–34 8–14 1–5	7–14 1.5–5			4–6	14–16	4–8	20–30	3
Bi + Amph.	0	0	0.6–1.6	0	0–35	0.4–3	0–13	0–30	0–45		0.2–2.5	0.1–0.7	0–96	3–100	0.5–79	15–85	5–46	62–92	0.4–100	3–98
Bi/Amph.			Amph. only		0.1–1	1.5–9	0/48–5/0	0/4–22			0 3–8	0/0 5–0.3	0.01–52/0	0/3–0/100	0.79–8	0.1–4.5	0/11 56	?	0.02	0.1–97
Ma/Me		1–12	0.1–2.5	2.3	3–10	0.2–24	0.1–100	0.1 7.0	0.6–4.0		1.4–50	1.2–37	3.4–74	4.0	3–11	1–89	0.7–15		3–5000	0.1–80
Pyrite	0	0–93	1–2	0–91	32–94	63–94	10–100	30–95	42–92									+++	+	alt.
Q/Feldspar	–	1–100	3–100	2–3	60	90	15–100	19–100	30–100		12–100	16 100	80–100		100	1 5–28	4–100	volc. glass		0.7–70
Glauconite	0	0	0–3	0	0–7	0	0–2	0–18	2–71		90	?	0–50		23–59	12 64	0–52	8–41	0–26	3–100
Other min.		rut.	gr, ep	szt, ilm	ap, cir	ep, tur	ep, magn	gr, ep, ort	ap, r, gr, ep		magn, ort	magn, ort	ep, disz	pirit		gr	gr, kl, ort	gr. kl, ilm	ap c kl gr	
			ort + plag		r, gdisz															

Bi = biotite; Amph. = hornblende; Ma = minerals of igneous origin; Me = minerals of metamorphic origin; Q = quartz.

2.5. *Paleogeographic History*

The main features of sedimentation are illustrated in Figs 2—5 for each of the four subareas of the Bakony Mountains separately. It should be noted that the most complete sequence is found in the SW subarea.

There are remarkable analogies ("situation recurrences") between the Mid- to Late Cretaceous and the Eocene paleogeography of the Bakony area.

During the Cretaceous, two continental areas consisting of igneous and slightly metamorphosed sedimentary formations of Paleozoic age were situated to the WNW and ESE, respectively, of the mighty Triassic and relatively thin Jurassic carbonate rocks which made up the bulk of what is now the Bakony Mountains. The igneous rocks of the southern, Pelso Range (Pelso = Latin for Lake Balaton) were of granitic type, while those of the northern, Little Plain, Promontory (of more restricted extension) included diabases, too. Geomorphologically, plateau-like elevations may have existed on both ranges. As a result of the subtropical climate (23—24 °C mean annual temperature, alternating wet and dry seasons) these were covered by forest and attacked by chemical weathering (laterization). Mechanically eroded lateritic materials were transported, mainly in form of suspension, onto the more or less karsted surface of the Mesozoic carbonate rocks.

Barremian-Cenomanian and Late Turonian-Senonian subsidences resulted in the deposition of two marine sedimentary cycles. The sediments of the first one covered the Perepuszta-Tés bauxite deposit, while those of the second one the Halimba bauxite deposit and the (in part overlying) Ajka-Padragkut coal-bearing formation.

Laramian uplift produced complete but unequal emergence. During the Paleocene, relief modelling by additional karstification and bauxite material deposition (respectively, reworking) were renewed. ("Infraeocene denudation".) Differential block faulting favoured a locally very important accumulation of piedmont type dolomite debris which were to bury even some bauxite deposits.

The Eocene sedimentary macrocycle comprises the time span E_1^2 (Late Cuisian) — E_3^3 (Late Priabonian).

Early Eocene subsidence was characterized by additional differential movements, along longitudinal (WSW-ENE) and transversal (NW-SE) axes. On the accentuated topography of the Mesozoic surface, paralic marshes and (partly hypersaline) lagoons developed in depressions of combined karstic-tectonic origin in the SW subarea. These were rapidly covered by characteristic brackish water and near-shore marine sediments of the Late Cuisian: the "Mediterranean" sea advanced from what is now northern Italy and Istria through southern Austria and the Zala Basin. (*Transgression* N°1, E_1^2, see Fig. 6, 1.)

Table 2

Characteristic Values of Petrological

Parameter / Member	Darvastó Formation				Kisgyón Formation					
	1.	2.	3.	4.	1.	2.	3.	4.	5.	KF
Kaolinite %	41—81	1—61	tr-9	2			7—21	21—49	6 (48)	0—37
Illite	0-tr	tr-6	0—4	1			3		2 (6)	0—10
Ill/Montm.	0	0—6	0—1	3			++	13—19	3 (8)	0—3
Montmorill.	0	0—3	0—1	0			0—16	11—12	0	0—37
Gibbsite, boehmite	0—40	0—8	tr	0			tr			4
Gypsum	0—2	0—56	0	0						
C^{org} % (aver.)	0.02	0.86	1.1			0.9—8.7	0.4—1.6	3.6	2.2	
O_{Fe} (aver.)	92	40	17		14.3	6.8	26.8	2.4	7.5	
Na_2O % (aver.)	0.10	0.17	0.22	0.20	0.20		0.06	0.3	0.3	0.2—0.6
K_2O % (aver.)	0.11	0.93	0.64	0.70	2.7		0.03		2.4	0.2—1.5
Na_2O/K_2O	0.9	0.2	0.3	0.7	0.1		2.0		0.1	0.1—3.4
SiO_2	27.8	31.5	4.8		37.8	42.6	22.6			
TiO_2	1.3	0.4	0.04		0.4	0.5	0.3			
Al_2O_3	32.7	11.2	1.7		12.9	30.5	5.9			
Fe_2O_3	10.6	5.3	0.9		4.7	5.1	3.9			
MnO	0.03	0.05	0.01		0.2	0	0.1			
P_2O_5	0.04	0.03	0.20		0.1		0.04			
SO_3										

tr = traces

During the second phase of marine conquest the sea reached the NE subarea and overflooded (partially) the Latest Cuisian to Earliest Lutetian marshy sediments, giving rise to the lower coal-bearing member. (*Transgression* N°2, E_2^1, Fig. 6, 2.)

The first pyroclastics — tuffites of dacitic composition — show up among the sediments of the SW subarea.

An "undulation" resulted in a Middle Lutetian *regression* in the

and Geochemical Parameters. II

Szőc Formation				Mór Formation			Csabrendek	Halimba	Csernye
SW	NW	NE	SE	NW	NE_{shore}	NE_{bas}			
tr-2	0—4		1—4		0—5		15—45	+	
0—3	2		tr-1				+++	++(+)	
0—6	6		0—47		3—8			?	
0	2		0—5		14—41		+	++(+)	++
tr	0	tr	0			tr			
0.1—0.6	0.05—0.16	0.1		0.17	0.63	0.37			
8—26	1—100	1—3			1.5—12.0		1.2—8.0	0.1—50.0	1.2—9.7
0.16	0.32—	0.06	0.55		0.5		0.2	1.9	0.5
0.26	0.60	0.99	1.16		1.2		0.6	1.1	1.2
0.6	0.5	0.06	0.47		0.4		0.3	1.7	0.4
4.4	42.0	40.0			14.8—49.6		15.2—44.2	9.7—60.0	5.3—52.7
0.2	1.0	0.1			0.2—0.4		0.1—0.3	0.1—1.1	0—0.7
1.3	5.1	2.1			3.0—14.3		3.7—15.5	2.8—22.8	2.0—22.5
0.8	2.9	0.2			1.6—6.7		4.6—6.1	2.1—5.9	0.6—12.6
0.01	0.05	0.07			0—0.1		0.04—0.05	0—0.44	0—0.2
0.02	0.02				0.1—0.3		0.1—0.1	0—0.3	0—0.2
					0—1.0		0.1—?	0—1.5	

NE subarea (Fig. 6, 3). This corresponds to the "intra-Lutetian denudation", with local material subply mainly from the Mesozoic dolomitic rocks.

A concomitant subsidence of the NW subarea gave birth to an archipelago characterized by micro- and meso-cycles of clastic and carbonate sedimentation combined. By means of the trace element data, double transport of sandy and silty materials could be recognized (from the N and from the S).

In the SW subarea marine carbonate sedimentation went on essentially undisturbed.

Table 3

Characteristic Values of Petrological

Element (ppm)	Darvastói Formáció				Kisgyón Formation			
	1.	2.	3.	4.	2.		3.	4.
					Coal	Shale		NW
Ag					0.1	—		0.6
As								
B	120	117	59	225	278	96	10	11—12
Ba	171	604	331	110	224	104	515	71—100
Be					52	50		
Co	8	16	9	9	20	9	12	3
Cr	156	128	89	90	80	80	115	75—167
Cu					110	93		
Ga	87	19	10	9	31	11	6	4—8
Ge					5	0		
In	3	5	2	4				
Li					131	42	25	49—66
Mn	54	331	84	125	645	202	128	112—120
Mo	4	5	3	4	8	1	1	2
Ni	24	15	15	20	77	50	145	74—148
Pb					18	7	14	5—18
Sn					4	5		1
Sr	38	288	241	200	1100	126	702	272—428
Ti	1500	2500	577	1800	4700	2000	3600	2100—6000
V	56	36	19	20	167	75	10	37—158
Zn					148	112		10—50
Zr	611	481	392	285	166	77		
Ratios								
Ti/Cr	10	20	6	20	59	25	32	12—82
V/Cr	0.4	0.3	0.2	0.2	2.1	0.9	0.1	0.2—2.1
Cr/Ni	6.5	8.5	5.9	4.5	1.0	1.6	0.8	0.5—2.2
Ni/Co	4.0	0.9	1.7	2.2	4.0	5.6	12	25—47
Mo/Ga	0.04	0.3	0.3	0.4	0.3	0.1	0.2	0.2—0.3
B/Ga	1.4	6.2	5.9	28.3	9	9	2	2—3

and Geochemical Parameters. III

				Szőc Formation				Mór Form.		Halimba
4.	5.		KF	SW	NW	NE	SE	NW	NE	
NE	NW	NE								
<0.4								0.5		
	40							165		
78–83	28	71	10–140	29	44	62	85	140	70	132
75–680	51	70	148–770	322	90	515	105	327	228	248
6						6				
2–5	1.7	0.4	4–22	4	7	99	7	6	4	11
75–77	35	31	4–150	17	549	118	15	82	86	149
21	22				48	31		75	67	
8–10	1.5	3	4–52	6	6	8	16	13	11	18
22	2	6.5	2–6.5	1	3	2				5
	54	27			49			120	71	
77–735	2000	200	42–97	178	429	575	18	1070	333	100
1–6	2.6	13	2–6.5	2	11	5	2	7		7
14–38	9	10	15–21	16	14	38	14	38	53	15
7–25	14	4			4	25		36	92	
	2.5				7			3	3	
100–622	1200	500	105–640	176	294	122	108	677	558	845
1400–5700	500	300	130–3900	50	3100	5000	1200	6300	657	210
107–126	11	13	8–60	10	27	339 (?)	13	59	89	18
				244	166	20		235	441	1767
350			255–622		483	274	418	1000		
20–74	44	10	4–62	3	6	42	83	77	8	1.4
1.4–1.7	0.3	0.4	0.4–1.8	0.6	0.05	2.9	0.9	0.7	1.0	0.1
2–5	4	3	0.3–8.1	1.1	39.0	3.1	1.1	2.2	1.6	9.9
7	6	25	0.8–3.8	3.6	2.0	4.2	2.0	6.4	13.7	1.4
0.1–0.5	2	4	0.1–0.5	0.7	1.8	0.6	0.1	0.5		0.4
8–10	19	24	2–3	8.0	7.3	7.8	6.6	10.3	6.4	7.3

Table 4
Environmental Factors

Formation	Darvastó Formation				Kisgyón Formation					
Member, Area	1.	2.	3.	4.	1.	2.	3.	4.	5.	KF
Factor										
Eh: reductive		+		+		+		+	+	
moderately oxidative			+		+		+			+
strongly oxidative	++						+			
pH: acidic		+		+		+		+	+	
alkaline	+	+	+	+	+		+		+	+
Light: euphotic	+		+		+		+		+	+
disphotic		+		+	+	+		+		(+)
Salinity: freshwater	+				(+)	(+)	+	(+)		
brackish water	(+)	+	+	+	+	+		+	+	+
normal sea water	(+)	+	(+)						+	+
hypersaline water		+								
Depth: 0—20 m	+	+	+	+	+	+	+	+	+	+
20—100 m										+
100—200 m										
Substratum: flat			+	+				+	+	
uneven	+	+			+	+	+			+
Nearshore										
Offshore	+	+	+	+	+	+	+	+	+	+
Agitation of water:										
slight	+		+			+		+	+	+
moderate		+	+	+	+	+	+			+
strong		+			+		+			
Components of sedimentation:										
terrigenous	+	+	(+)	+	+	+	+	+	+	+
chemical		(+)	+	(+)			+			
pyroclastic			ny		+	(+)	+	++	++	
biogenic										
allocht. plant subst.		++	+			+++		+++	+	+
coralline algae			(+)							
molluscs		+	+	+		+		+	+	+
benth. microforams		+	++	+		+		+	+	+
Nummulites				(+)					+	+
corals				(+)					+	+
echinoderms			fragm.	fragm.						+
planktonic microforams										
nannoplankton	+	+								
radiolaria										

of Deposition

Szőc Formation				Mór Formation			Csr	H	Cs	I
SW	NW	NE	SE	NW	NE_{sh}	NE_{bas}				
	+	+			+		+	+	+	
+	++		+	+		+		+		+
	+									+
							(+)	+	(+)	
+	+	+	+	+	+	+	+	+	+	+
+	+	+	+	+	+				+	+
	(+)					+	+	+		
	(+)									
	+		+							+
+	+	+	+	+	+	+	+	+	+	+
+	+	+	+		+					+
+			+	+	+		+	+	+	
						+	+	+		
+		(+)	+			+	+	+	+	
+		+	+	+	+					+
+	+	+	+	+	+			(+)		+
+						+	+	+		
		+	+			+	+	+		+
+	+	+		+	+				+	
+	++									+
(+)	++	(+)	+	++	++	++	+	++	+	++
++	++	++	++	+	+	+	+	+	++	(+)
+	+	++	++	++	++	++	++	+++	++	
	(+)				(+)					
+	+	+	+						++	
+	+	+	+	+	+	+	(+)	+	+	
+	+	+	+	+	+	+		+	+	
+	+	+	+	+	+	+	+	+		
+	+	+	+		(+)		+	(+)		
+	+	+	+							
+		+	+				+			
				+	+	+	+	+	+	(+)
	+			+	+	+		+		
						+		+		

Overall subsidence started in the Late Lutetian. (*Transgression* N°3, E_2^3, Fig. 7, 4). Marshy sediments were deposited in the NE and in the SE subareas, in several — tectonically controlled — stripes, giving rise to the upper coal-bearing member. They are overlain by (sub)littoral limestones and in the NE by pelagic shales. In the NW, cyclic sedimentation was going on.

Table 5

Facies

Formation	Darvastó Formation				Kisgyón Formation					
Facies \ Member, Area	1.	2.	3.	4.	1.	2.	3.	4.	5.	KF
Freshwater										
fluviatile (river)							(+)			
lacustrine (lake)	+				+		+			
palustrine (swamp)						+				
Brackish water										
delta (with mangrove)		+				+				
marsh (paralic)		+		+		+		+		+
open lagoon		+	+							+
Hypersaline										
closed lagoon		+								
Saline (normal marine)										
littoral									+	+
sublittoral										
algal meadow										
shell bank (coquina)									+	
patch reef										
neritic										
Variability										
(lithotope gradient)	1	5	2	4	1	3	1	1	2	2

By that time, marine communication was established with the "Central Carpathian Paleogene" region in Southern Slovakia. (The "Budinský vývoj" of the Slovak authors means "Buda Facies", alluding to a similarity to the Budapest Eocene.)

Pyroclastics occur in the SE and NE subareas, too. They differ in composition from the SW dacite tuffites, being of andesitic type. (With characteristic fluctuations of the biotite-amphibole ratio.) Consequently, the assumption

of two different eruption centres (one in the SW, i.e. in the Zala Basin, and another in the East — Velence Hills —, seems to be justified.

In the SW subarea, a remarkable facies differentiation occurred ranging from beach to outer shelf type sediments. Nummulites banks, patch reefs are typical over large areas. Glauconite abounds in the outer shelf sediments (it is

Distribution

Szőc Formation				Mór Formation			Csr	H	Cs	I
SW	NW	NE	SE	NW	NE_{sh}	NE_{bas}				
	+									
										+
										(?)
+	+	+	+	+	+					+
+			+			+	(+)	+		
+									+	
+		+								
+										
						+	+	+	+	
3	4	2	2	3	2	1	1	1	2	5

present also in the NE pelagic shales). At least partly it has been formed from pyroclastic sources. In the meantime, free alumina minerals disappeared and kaolinite was replaced by illitic minerals in the pelitic sediments (followed by montmorillonite).

Pre-Pyrenean-Balkanic (= Illyric) movements on the Lutetian Priabonian boundary (Fig. 7, 5) were manifested by the deposition of more detrital beds in the NE and in a "rejuvenation" of the NW archipelago characterized

by the coarse clastic to variegated-lagoonal Iharkut Formation. The SE subarea very likely became emergent at that time. In the SW "basin" several block islands emerged; nevertheless, the general trend of increasing subsidence was not interrupted.

The Priabonian sea was of great extension, connected even with the Transylvanian Basin (through NE-Hungary). (*Transgression* N°4, E_3^{1-2},

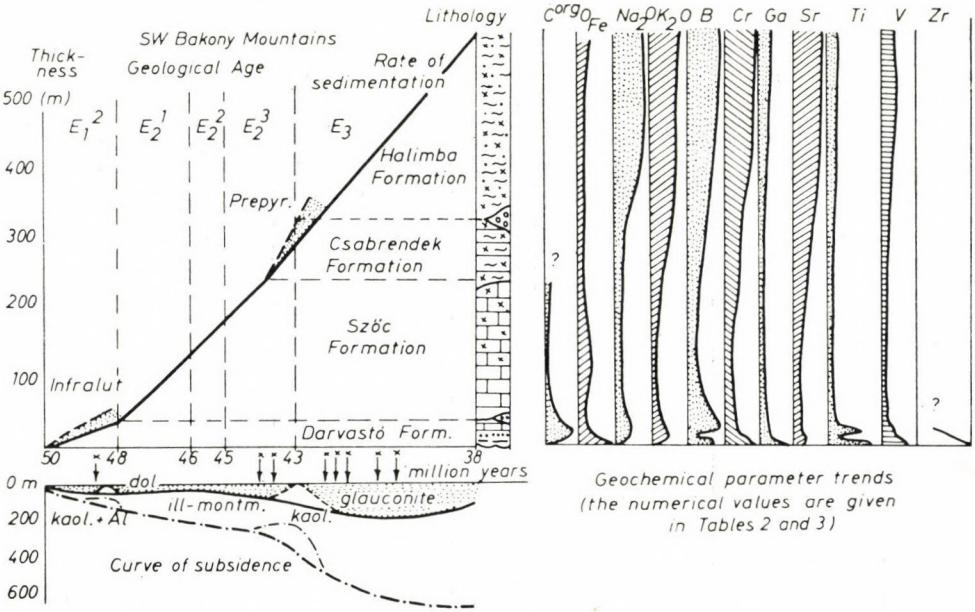

Fig. 2. Sedimentogram of the SW area of the Bakony Mountains Eocene

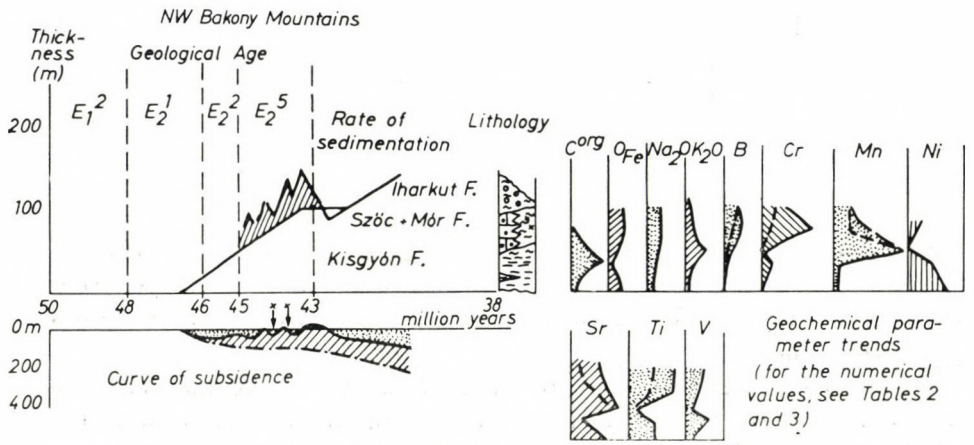

Fig. 3. Sedimentogram of the NW area of the Bakony Mountains Eocene

Fig. 7, 6.) The sea was of pelagic character in the SW and NE (the Halimba and Mór basins, respectively, were subsiding at a high rate), while archipelagic (slowly subsiding) in the NW. The alternating sands and biogenic limestones of the NE area may have been deposited on a submarine elevation.

Pyroclastics persist thoroughout the Late Eocene. They even attain a second, more important maximum of abundance indicating the paroxysm of

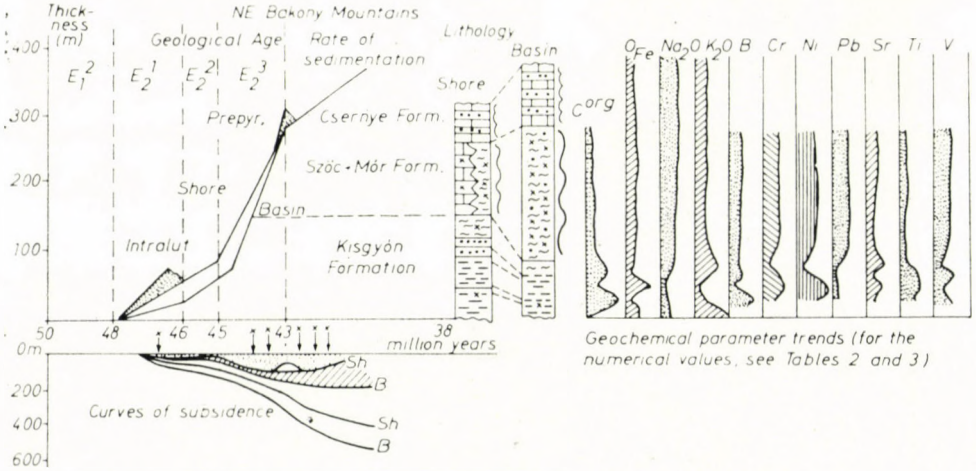

Fig. 4. Sedimentogram of the NE area of the Bakony Mountains Eocene

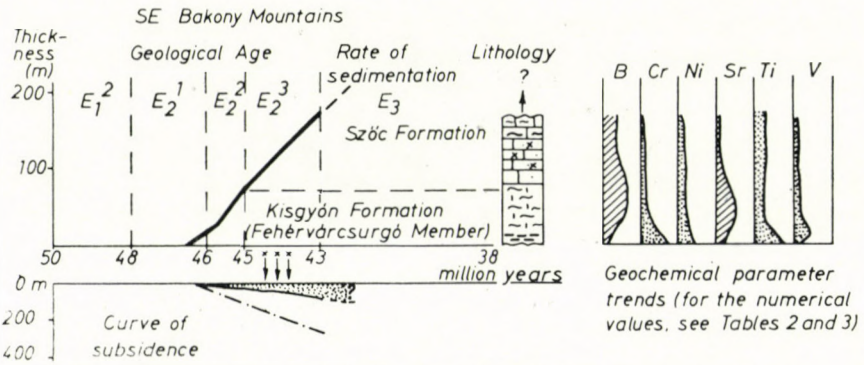

Fig. 5. Sedimentogram of the SE area of the Bakony Mountains Eocene

volcanic activity. The composition reveals repeated shifts of the biotite-amphibole ratio and a slight accentuation of the acidic character.

Priabonian near-shore sediments have been hit by drilling to the South of Lake Balaton, in the zone of the former Pelso Range. Accordingly, this must

have been completely planated and at least partly overflooded by that time. Pyreneic movements ones at the end of the Eocene, resulted in the complete emergence of the Bakony Mountains area, interrupting the Mediterranean marine communication. The ensuing "infra-Oligocene denudation" removed a considerable part of the (first of all Upper) Eocene sediments. Nevertheless,

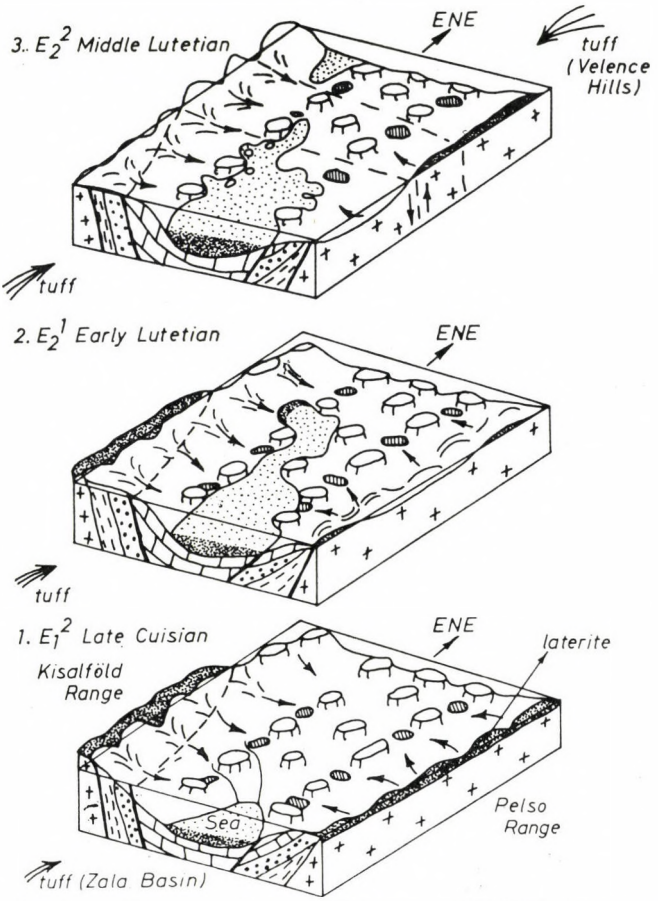

Fig. 6. Bakony Mountains Eocene Paleogeographic Block Diagrams

in some Upper Eocene sediments a "northern influence", i.e. boreal (more correct for those times, temperate) faunal elements, could be recognized by T. KECSKEMÉTI.

The Oligocene sediments in some areas contain big amounts of Eocene material, a rather misleading circumstance sometimes resulting in erroneous interpretations of age. However, in the NW border area of the Bakony Moun-

tains, the Oligocene and Miocene sediments provide no evidence of removed Eocene materials. Accordingly, the NW coast line of the Eocene sea remains an unsettled problem. In contrast to the Middle and Late Cretaceous and Eocene sedimentation cycles, the Oligocene sedimentation followed very different lines.

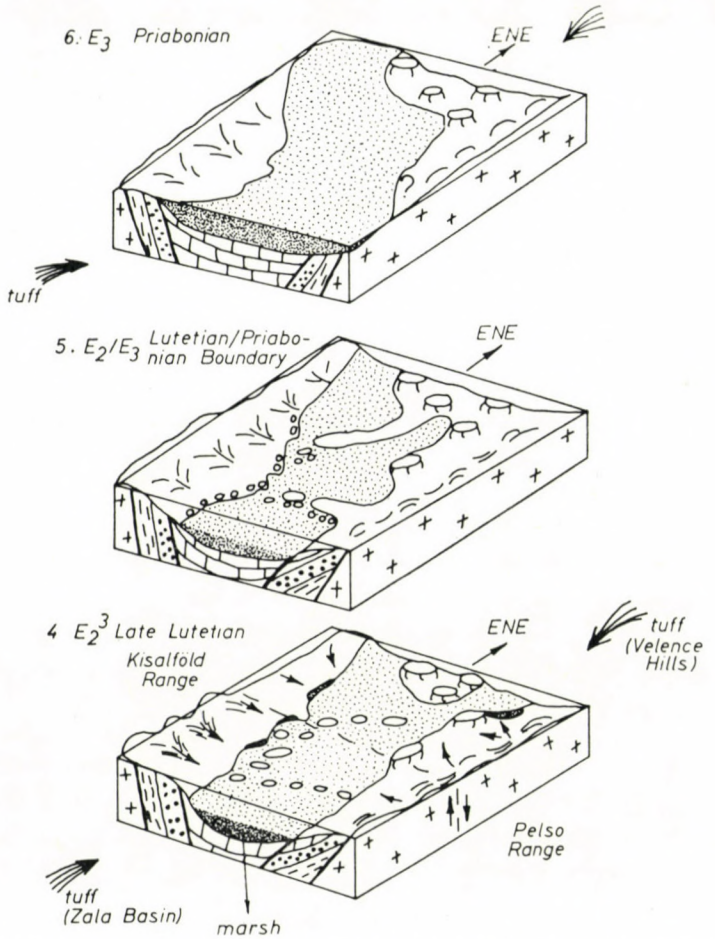

Fig. 7. Bakony Mountains Eocene Paleogeographic Block Diagrams

2.6. *Plate Tectonics Aspect*

E. Szádeczky-Kardoss [17] interpreted the Balaton lineament as a (second category) subduction suture.

The paleogeographic history sketched above complies with this concept, of one assumes a multi-phase subduction process starting in the Early through

Middle Cretaceous followed by "isostatic" compensation subsidences and uplifts, block faulting in the "second foreground" of subduction, and a Middle to Late Eocene subsequent dacite-andesite volcanism.

3. Conclusions and Work to be Done

The present contribution to G. KOPEK's monograph on the Eocene of the Bakony Mountains (in print, dealing principally with the stratigraphic aspects) permits to establish the main elements of a sedimentation model for the mobile shelf, as required by F. J. PETTIJOHN and P. E. POTTER in 1963 [52], and complemented by the present writer.

These would be (as evident from the antecedents):

— Geometry of the sedimentary basin (double bipartite, tectonically controlled by two axes).

— Basement of the sedimentary basin (Mesozoic carbonate rocks surrounded by older igneous and anchimetamorphic rocks).

— Relief of the sedimentary basin (highly dissected and rough due to differential block faulting and repeated karstification).

— Anisotropy axes, preferential directions (parallel and perpendicular to the coast line).

— Features of the sediments (clastic, chemogenic, biogenic and pyroclastic sediments combined in varying proportions).

— Rate, cyclicities and other parameters of sedimentation (to be established for the four subareas one by one).

— Synsedimentary movements (of both epeirogenic and "orogenetic" nature).

Further more, well premeditated and systematic sampling and laboratory work would make possible reasonable geomathematical data processing and even computer simulation of the sedimentation process.

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ОСАДОЧНЫЕ СВИТЫ И ОСАДОКОНАКОПЛЕНИЕ В ГОРАХ БАКОНЬ (ЗАДУНАЙСКИЙ КРАЙ, ВНР)

Э. ДУДИЧ

Резюме

Эоценовые отложения автор статьи делил на 9 свит (рис. 1), в пределах которых он выделил подразделения. Характеристика отдельных свит дается на основании следующих анализов: гранулометрический анализ, анализы общего содержания карбоната и доломита, определение содержания Na и K пламенным фотометром, силикатный анализ (влажный), эмиссионный спектральный анализ, дифференциально-термический и дифференциально-термогравиметрический анализы, анализ тяжелых минералов, определение содержания органического углерода, исследования микрофаций по дюншлифам рентгено—структурный анализ.

Результаты содержатся в таблицах 1—3.

Оценка типов фаций с учетом данных об ископаемых органических остатках дается в таблицах 4—5.

Ход осадконакопления в пределах четырёх, чётко выделяющихся участков показан на рис. 2—5.

История геологического развития гор Баконь в целом в эоценовое время проиллюстрирована блок-диаграммами 6—7.

Эоценовый этап развития района последовал — третим седиментационным циклом — после двух крупных осадочных циклов среднего и верхнего мела, соответственно, обнаруживав общие с ними черты. С поднятием рассматриваемого района в конце эоцена и наступлением террестрического режима начался совершенно новый этап геологического развития района.

Трехраздельный среднемелово-эоценовый мегацикл хорошо поддается интерпретации согласно теории о тектонике плит.

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SUBDIVISION ET DÉTERMINATION D'ÂGE DES FORMATIONS DE L'ÉOCENE INFÉRIEUR DU NORD-EST DE LA TRANSDANUBIE

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Pendant les années passées plus d'une centaine, plusieurs générations de géologues hongrois ont collaboré à l'évolution de la subdivision stratigraphique et détermination d'âge des formations éocènes de Hongrie. Cet article présente un aperçu sur le processus pendant lequel la connaissance stratigraphique est arrivée dès les premiers pas jusqu'à la subdivision et détermination d'âge basées sur l'étude complexe des coupes-repères.

M. HANTKEN (1872) a fait première subdivision des formations éocènes des environs de Dorog, juste encore aujourd'hui.

La distinction de l'Éocène inférieur en Hongrie commença par l'œuvre de H. TAEGER. Il a attribué le complexe lignitifère, situé à la base de l'Éocène dans la Montagne Vértes, à l'étage yprésien. L. LÓCZY sen. (1913) avança un peu plus, en y rangeant aussi la marne argileuse à Operculines et *Nummulites subplanulatus*. P. ROZLOZSNIK—Z. SCHRÉTER et K. TELEGDI-ROTH (1922) ont établi la première subdivision stratigraphique de l'Éocène de Hongrie, juste encore aujourd'hui dans ses traits fondamentaux. Ils ont attribué le complexe lignitifère inférieur, le complexe du mur et les formations saumâtres du toit, au Sparnacien.

Dans les dernières années passées presqu'une vingtaine, de nombreux spécialistes ont étudié les formations éocènes du Nord-Est de la Transdanubie, parmi eux on peut souligner les travaux faits par M. KEDVES, L. RÁKOSI, Mme L. VITÁLIS-ZILAHY et Mme M. JÁMBOR KNESS, L. GIDAI a rangé le complexe lignitifère inférieur, ses formations du mur et du toit jusqu'à la marne arguleuse à Operculines inclue, dans l'Éocène inférieur. Il a tracé la limite entre l'Éocène inférieur et moyen d'après l'apparition de l'association des grandes Foraminifères caractérise par *Nummulites perforatus*.

Dans la région du Bakony Méridional, *Nummulites perforatus* se présente en masse, dans la groupe de couches à *Nummulites laevigatus*.

1. Introduction

Pendant plus d'une centaine d'années passées plusieurs générations de géologues hongrois collaboraient à l'évolution de la subdivision stratigraphique et détermination d'âge des formations éocènes de Hongrie. Dans la suite, nous voudrions embrasser ce processus, où dès la première subdivision de l'Éocène du Nord-Est de la Transdanubie la connaissance stratigraphique est arrivée à la subdivision et détermination d'âge basées sur les études complexes des coupes-repères.

2. Évolution de la subdivision et détermination d'âge

M. HANTKEN (1872) était le grand pionnier de la recherche paléontologique et stratigraphique hongroise qui — sur la base des Mollusques et Foraminifères — a effectué, à manière juste encore aujourd'hui, la subdivision stratigraphique des formations éocènes aux environs de Dorog, l'un des régions à succession stratigraphique la plus complexe de l'Éocène du Nord-Est de la Transdanubie. Il n'employait pas encore une échelle stratigrafique plus fine, ne distinguant pas d'étages. La valeur fondamentale et impérissable de son œuvre est manifestée dans la caractérisation bonne et précise des groupes de couches.

K. HOFMANN (1871) a utilisé la subdivision établie par M. HANTKEN aussi aux formations éocènes de la Montagne de Buda-Kovácsi. K. HOFMANN était le premier à déterminer l'âge des formations éocènes de Hongrie. Il a attribué à l'étage parisien de Mayer ces formations que nous considérons appartenir, aujourd'hui, à l'Éocène inférieur et moyen.

Cette constatation faite par HÉBERT et MUNIER-CHALMAS, selon laquelle les plus anciennes formations éocènes appartiendrionts aussi à l'Éocène moyen, n'a pas été confirmée par les études postérieures.

En maintenant la subdivision établie par M. HANTKEN et K. HOFMANN, P. OPPENHEIM (1896) commence l'Éocène avec l'Éocène moyen, et ne distingue pas l'Éocène inférieur.

F. SCHAFARZIK (1902) a maintenu la subdivision établie par K. HOFMANN (Parisien, Lutétien) pour la partie inférieure de la succession stratigraphique et a divisé l'Éocène supérieur en deux parties: Bartonien (en bas), Priabonien (en haut). Dans son tableau stratigraphique l'Éocène inférieur est déjà présent, divisé en trois étages: Londonien (Yprésien), Suessonien et Thanétien.

H. TAEGER (1908—1911) suit aussi les pas de K. HOFMANN en attribuant l'Éocène de la Montagne Vértes à l'Éocène moyen et supérieur. Il divise les formations éocène moyen en deux étages, en bas c'est l'Yprésien (notoirement considéré comme Éocène inférieur) et en haut le Lutétien. Il ne justifie pas l'avis de HOFMANN selon lequel le complex lignitifère appartiendrait aussi au Parisien et il a inseré la formation lignitifère, située à la base du complexe éocène, à l'Yprésien.

Le travail de L. LÓCZY sen. (1913) a présenté un nouveau pas en avant dans le domaine de la subdivision et détermination de l'échelle statigraphique plus précise de l'Éocène inférieur de cette région. Il a constaté qu'on ne peut accepter l'opinion d'OPPENHEIM (1896), concernant la région transdanubienne. L. LÓCZY sen. était le premier à résoudre justement la détermination d'âge des formations éocène inférieur. Il a inseré le complexe lignitifère inférieur et les formations saumâtres de son toit au Landénien, et l'argile à *Operculines* et *Nummulites subplanulatus* à l'Yprésien.

En évoluant la subdivision stratigraphique établie par L. LÓCZY sen. (1913), P. ROZLOZSNIK—Z. SCHRÉTER et K. TELEGDI-ROTH (1922) ont établi la première échelle stratigraphique de l'Éocène de Hongrie, correcte encore aujourd'hui dans ses traits fondamentaux. Dans leur monographie, ils ont attribué le complexe lignitifère inférieur, les formations sous-jacentes et celles saumâtres du toit au Paleocène, resp. au Sparnacien. Selon eux la marne argileuse à Operculines représente l'Yprésien et la partie inférieure du Lutétien.

(Comme lithofaciès, la marne argileuse à Operculines, monte, en réalité jusqu'à la partie inférieure du Lutétien. Selon nos avis actuels, on peut désigner la limite entre le Cuisien et le Lutétien d'après l'apparition de l'association des grands Foraminifères caractérisée par Nummulites perforatus, alors la limite inférieure du groupe de couches à perforatus et celle du Lutétien coïncident.)

Deux des trois membres de la monographie présentaient des constatations précieuses supplémentaires sur l'Éocène inférieur.

K. TELEGDI-ROTH (1923) en traitant les conditions palégéographiques du Paléogène y compris les terrains de l'Éocène inférieur — a développé son opinion selon laquelle hors les régions de Tatabánya, Dorog et Tokod, ainsi que de Pilisvörösvár et Nagykovácsi, la formation lignitifère paléocène, les marnes argileuses saumâtres du toit, resp. à Operculines, — c'est à dire la partie la plus inférieure de l'Éocène, — seraient inconnues.

P. ROZLOZSNIK (1928) a employé la subdivision stratigraphique — développée à la région de Dorog — aussi au Basin de Tatabánya. Il a inseré les formations d'eau douce et saumâtres à gîtes de lignite dans Tableau n°2. le «Paléocène» et la marne argileuse à Foraminifères et Mollusques inférieure dans l'Éocène inférieur.

E. SZŐTS a pris position sur la subdivision stratigraphique des formations éocènes sur la première fois en 1938. à l'occasion de l'étude des formations paléogènes de la colline Antal-hegy de Mór. Il a attribué les argiles bariolées et les sables du mur au Paléocène, la formation lignitifère à l'Éocène inférieur, la marne à Mollusque du toit et la marne à *Nummulites perforatus* à l'Éocène moyen, enfin le calcaire à Discocyclines et Nummulites à l'Éocène supérieur. Contrairement à l'opinion de K. TELEGDI-ROTH sur la base d'une comparaison faunistitique, il a attribué les gîtes de lignite de Mór à l'Éocène inférieur.

La subdivision stratigraphique établi par lui en 1952 (in: E. VADÁSZ, 1953) a développé davantage l'opinion de P. ROZLOZSNIK, en divisant plus détaillément les terrains de l'Éocène inférieur. Dans son tableau, fait en 1956, entre le Lutétien inférieur et moyen au course de la discordance supposée — selon notre avis injuste — l'érosion aurait enlevé aussi une partie de l'Éocène inférieur.

La subdivision publiée par E. VADÁSZ (1960) est presqu'identique à celle établie par E. SZŐTS (1956).

Dans l'édition de 1956 du Fascicule «Hongrie» du Lexique Stratigraphique International (E. VADÁSZ et al., 1956) d'E. SZŐTS et E. VADÁSZ de jadis, et présentant selon notre opinion un bon aperçu aussi sur l'Éocène du Nord-Est de la Transdanubie.

G. KOPEK—T. KECSKEMÉTI et E. DUDICH (1966) ont attribué les formations — inserées par les auteurs anciens dans l'Éocène inférieur — au Lutétien inférieur.

Pendant les années passées, à peu près vingt, des nombreux spécialistes collaboraient aux études pétrographiques, micro-minéralogiques, micro-faunistiques et des grands Foraminifères et Mollusques de l'Éocène du Nord-Est de la Transdanubie.

Nombreuses descriptions monographiques et publications scientifiques ont été parues sur les résultats. Dans la suite, nous allons récapituler les résultats palynologiques, microfaunistiques et sur les grands Foraminifères, que l'on peut considéré, selon notre jugement, comme les plus importants, dans la reconnaissance et solution des problèmes de détermination d'âge et de corrélation des formations éocène inférieur.

D'après l'œuvre de M. KEDVES et L. RÁKOSI nous connaissons le contenu palynologique de l'Éocène inférieur du Nord-Est de la Transdanubie et leurs types de végétation. Les études sur les Sporomorphes ont livré des informations d'importance fondamentale sur la solution des problèmes de détermination d'âge et de corrélation de l'Éocène inférieur.

En 1960, M. KEDVES a achevé l'étude palynologique des séries éocène inférieur — sparnaciennes — des sondages approfondie dans le Bassin de Dorog. Il a constaté qu'on peut y démontrer en grande quantité le pollen de trois groupes végétaux: Taxodiaceae, Cupressaceae, Myricaceae et Palmae. Parmi ceux-ci, les palmiers — indiquant un climat tropique — sont sans discussion dominante. L'étude palynologique de l'Éocène inférieur du Bassin de Dorog a confirmé le caractère tropique du climat à cette époque, en Hongrie. Sur la base de ses études il a probabilisé les zones végétales suivantes: marais profond, marais peu profond, forêt marécageuse à Taxodiaceae et Cupressaceae, bosquet à Palmae, buissons à Myricaceae et forêts à Sequoia dans les régions montagneuses.

En 1964, il a palynologiquement justifié la présence du Sparnacien et de l'Yprésien dans la succession stratigraphique du sondage Lábatlan-4., approfondi à la région occidentale du Bassin de Dorog. Dans le groupe du couches du mur du complex lignitifère (à cette région déjà sans gîtes) éocène inférieur, il a montré la forme de *Tricolporopollenites liblarensis* (THOMS.) (= *quisguales* R. POT.) TH. et PF. en quantité dominante.

En 1962, il a distingué les différentes régions suivantes de la flore éocène inférieur — sparnacienne — en Hongrie:

— Dorog et Tatabánya: Palmae,

— Dudar, Halimba, Iszkaszentgyörgy: Cupuliferae, Castaneae.

La différence floristique a été causé premièrement par les facteurs écologiques, mais selon son opinion une différence d'âge n'est pas exclue, non plus.

En 1963, d'après l'analyse pollinique il a constaté qu'on peut distinguer plusieurs horizons (au moins deux) dans le Sparnacien de Hongrie.

1. Sparnacien

A) Dorog, caractérisé par la dominance des pollens de Palmiers et nombreuses spores de Fougères. Cette végétation de la forêt marécageuse est typique dans le Sparnacien de Hongrie.

B) Halimba, la forme typique est *Tricolporopollenites cingulum* (R. Pot.) Th. et Pf. 1953. Elle indique une végétation de marais plus humide que celle précédente: région marécageuse, forêt marécageuse semi-terrestre, etc., mais du point de vue stratigraphique à association de pollens et spores identique ou bien analogue.

2. Yprésien

Selon M. Kedves on peut le bien séparer du Sparnacien, car ici la dominance des pollens de Palmiers a cessé. La distinction du type de Halimba est plus difficile, car les pollens dominants sont identiques dans tous les deux cas. (formes de la forêt marécageuse semi-terrestre).

L. Rákosi (1968) a fait l'étude palynologique détaillée de l'argile lignitеuse et des couches de lignite argileuse, traversées dans les sondages Csehbánya-1. (Bakony Méridional) et Balinka-252. Les associations de Sporomorphes — trouvées dans les successions stratigraphiques des deux sondages — sont identiques et sembles convenables à la vérification de l'Éocène inférieur. En total, il a déterminé 47 Sporomorphes. Sur la base d'une étude comparative détaillée, il a correlé les couches étudiées à l'horizon supérieur de l'Éocène inférieur d'Allemagne et d'après la subdivision établie par W. Krutzsch il les a correlé au gîte inférieur de Helmstedt («HU-Bild»).

Dans sa publication, étant sous presse, il a rendu compte de l'étude du pollen du palmier Nipa appartenant au type de la végétation de mangrove. Selon ses études, le pollen du palmier Nipa se présente déjà dans les couches saumâtres du Cuisien à la région du Nord-Est de la Transdanubie, alors déjà dans les couches à lignite de la phase initiale de la transgression. Au Bakony Méridional, il les a retrouvé dans les couches à lignite en dessous et en dessus du groupe de couches à Alveolina oblonga. D'après son association, cette dernière végétation semble être un peu plus ancienne.

L. Vitális-Zilahy (1967, 1968) a reconnu six zones des Foraminifères planctoniques dans la succession stratigraphique éocène du Bassin de Dorog. (du haut en bas):

6. *Globorotalia cocoaensis*
5. *Globigerapsis semiinvoluta*
4. *Globigerina corpulenta, Globorotalia rugosoaculeata*
3. *Globorotalia rotundimarginata*
2. *Globigerapsis higginsi, Globigerapsis kugleri*
1. *Globorotalia pentacamerata*

On peut rangé la 1ère zone à la fin de l'Yprésien, caractérisée premièrement par Globorotalia pentacamerata. Il y a quelques espèces restreintes à cette zone:

Globanomalina pseudoiota (HORNIBROOK)
Globanomalina ovalis HAQUE
Globanomalina simplex HAQUE
Globigerina compressaformis CHALILOV
Globorotalia (A.) clara CHALILOV
Globorotalia (A.) subintermedia CHALILOV
Globorotalia (A.) pentacamerata SUBBOTINA *var. camerata* CHALILOV.

Quelques Foraminifères planctoniques présents dans cette zone montent aussi dans la 2e zone située à la base du Lutétien:

Globorotalia (A.) interposita SUBBOTINA
Globigerina inaequispira SUBBOTINA
Globigerina ex gr. eocaenica TERQUEM

M. JÁMBOR-KNESS effectuait l'étude des grands Foraminifères des nombreuses coupes de l'Éocène du Nord-Est de la Transdanubie. Elle a publié plusieurs ouvrages sur les résultats de ses études d'importance fondamentale dans la reconnaissance des positions stratigraphiques de l'Éocène du Nord-Est de la Transdanubie:

En 1967, elle a étudié les grands Foraminifères des successions stratigraphiques éocènes des sondages: Bajót 18, 23, 24, 31, 32 et Nyergesújfalu 19, 25, 26, 28. D'après ses études, parmi les espèces de Nummulites les formes «A» et «B» de Nummulites subplanulatus apparaissent premièrement, souvent constructeurs. On peut retrouvé *N. aff. planulatus* ensemble avec *N. subplanulatus*, mais son nombre d'individus est considérablement plus réduit. La détermination de *Nummulites planulatus* — faite par M. JÁMBOR KNESS — est aussi confirmée de la part du spécialiste français des Nummulites, A. BLONDEAU (L. GIDAI 1976/b. p. 146). Elle a constaté l'apparition systématique de Nummulites subramondi et *N. anomalus* dans la marne argileuse à Operculines de mer peu profonde. Dans les successions stratigraphiques des sondages étudiés, ces deux espèces ont été retrouvées toujours dans la marne argileuse

à Operculines. Outre les espèces de Nummulites ci-mentionnées, elle a encore déterminé les espèces suivantes dans les formations cuisiennes:

Nummulites globulus
Nummulites burdigalensis
Nummulites aff. pustulosus
Nummulites cf. atacicus
Nummulites nitidus
Nummulites aff. partschi.

Outre les Nummulites, dans l'étage cuisien elle a déterminé Assilina laxispira, beaucoup d'Operculines, Discocyclines et de Bryozoaires.

Dans sa publication de 1969/a, elle a fait des constatations importantes sur les formations éocène inférieur:

1. Au-dessous du niveau à *Nummulites subplanulatus* (complexe d'argiles bariolées, complexe lignitifère), elle n'a pas trouvé de Nummulites. On ne peut prouver l'âge paléocène avec Nummulites.

2. Dans la partie supérieure de l'Yprésien (Cuisien) les Operculines sont très fréquentes, pourtant à la caractérisation de cette zone de la succession stratigraphique les espèces Nummulites anomalus de la Harpe et *Nummulites subramondi* de la HARPE conviennent mieux, car les Operculines se présentent presqu'en toutes les zones de l'Éocène marin de Hongrie. Dans les couches les plus anciennes de l'Éocène marin de Hongrie, elle a démontré l'association de grands Foraminifères suivante:

Nummulites aff. planulatus (LAM.) A
Nummulites subplanulatus (HANTK. et MAD.) A B
Nummulites nitidus de la HARPE
Nummulites burdigalensis de la HARPE
Nummulites praelucasi DOUVILLÉ
Nummulites globulus (LEYM.) A
Nummulites rotularius
Assilina laxispira (de la HARPE)
Discocyclina douvillei (SCHLUMBERGER)

En 1972, elle a présenté le suivant aperçu sur l'étude des grands Foraminifères de l'Éocène inférieur du Nord-Est de la Transdanubie:

Détermination d'âge, lithofaciès et niveaux à Nummulites des successions stratigraphiques éocène inférieur du
Nord-Est de la Transdanubie.
D'après Mme JÁMBOR KNESS, m. 1972.

	Formations et niveaux à Nummulites	Oroszlány sond. 1838	Oroszlány sond. 1846.	Tatabánya sond. 1474.	Vértestolna sond. 9.
Étage yprésien	Niveaux a *N. anomalus* et *N. subramondi* (= argile à Operculines)	245,5 à 298,0 m	373,1 à 406,0 m	249,1 à 321,2 m	
	Complexe sans Nummulites (argile sableuse)		406,0 à 414,5 m	321,2 à 322,0 m	
	Niveau à *N. subplanulatus* (argile sableuse)	298,0 à 303,2 m	414,5 à 425,9 m	322,0 à 328,7 m	212,1 à 222,8 m
Étage sparnacien	Complexe de lignite et d'argiles bariolées	303,2 à 332,8 m	425,9 à 466,3 m	328,7 à 471,0 m substratum non atteint	222,8 à 255,6 m

Yprésien

Niveau à *N. subplanulatus*. Les générations A et B de *N. subplanulatus* (HANTKEN et MADARÁSZ) présentent l'espèce extrêmement fréquente ou fréquente — parfois presque unique — de la zone inférieure du complexe d'argile sableuse et d'argile à faciès marin. Dans certains cas elle est observable (sondages: Tatabánya 1474., Vértestolna 9.) ensemble avec l'espèce *N. globulus* LEYMERIE A. L'épaisseur du niveau varie entre 3,1 et 11,4 m.

Zone sans Nummulites. Elle se trouve entre les niveaux à *N. subplanulatus*, resp. à *N. anomalus* et *N. subramondi*. Lithofaciès: argile, argile sableuse. Par endroits on peut montrer la présence des fragments de coquille des Mollusques, quelques petits Foraminifères ou de Discocyclines; les Nummulites n'y sont parvenues que très rarement en fragments.

Niveau à *N. anomalus* et *N. subramondi* (= complexe d'argile à Operculines). Développé à partir du complexe à *N. subplanulatus* avec difference lithologiquement aussi observable, sous l'épaisseur de 32,0 à 72,0 m. Ici, outre les deux Nummulites marqueurs, *N. globulus* LEYMERIE forma A, N. *aff. globulus* LEYMERIE forma A, *N. pernotus* SCHAUB forma A, *N. praelucasi* DOUVILLÉ forma A, *N. aff. pustulosus* DOUVILLÉ forma A, *N. aff. rotularius* DESHAYES forma A B ont été aussi reconnues. Elle a identifiè l'espèce *Nummulites anomalus* de la HARPE *var. granulosus* (de CIZANCOURT) forma B jusqu'ici non connue en Hongrie.

Outre celles-ci, les formes d'Operculines — facilement reconnaissables même au terrain — sont particulièrement fréquentes: *O. ammonea* LAYMERIE, *O. granulosa* LEYMERIE, *O. parva* DOUVILLÉ. Même de ce point de vue, la coupe des deux sondages d'Oroszlány est aussi considérable, où ces espèces

se présentent en fréquence jusqu'ici jamais observée, même comme constructeurs. D'après sa détermination, parmi les grands Foraminifères ci-trouvés on peut encore mentionner les espèces *Assilina sp.*, *Discocyclina douvillei* (SCHLUMBERGER), *D. tenuis* DOUVILLÉ, *Actinocyclina radians* D'ARCHIAC. Elle a mentionné que parmi les deux Nummulites marqueurs ou niveau, la prédominance de *N. anomalus* est caractéristique par rapport à *N. subramondi.* Par contre, dans le Bassin de Dorog, les deux espèces se présentent en quantité pratiquement égale.

Dans sa récapitulation faite en 1973, elle a distingué deux niveaux à l'intérieur de l'Yprésien (Cuisien):

Niveau à *N. subplanulatus.* A la zone inférieure du complexe de marne argileuse de mer peu profonde, *N. subplanulatus* se présente en masse ou elle y est fréquente. *N. subplanulatus* ne se mélange pas avec les espèces *(N. anomalus, N. subramondi)* de la partie supérieure du complexe de marne argileuse, ainsi elle représente le niveau bien distinct de l'Yprésien.

Niveau à *N. anomalus* et *N. subramondi.* L'association de ces deux espèces ne caractérise que cette zone de l'Éocène. A ce niveau, elle a trouvé les grands Foraminifères suivants:

Nummulites anomalus de la HARPE *forma* A, B
Nummulites subramondi de la HARPE *forma* A
Nummulites nitidus de la HARPE *forma* A
Nummulites praelucasi DOUVILLÉ *forma* A
Nummulites partschi de la HARPE *forma* A
Nummulites aff. pernotus SCHAUB *forma* A
Nummulites aff. pustulosus DOUVILLÉ *forma* A
Operculina ammonea LEYM.
Operculina subammonea LEYM.
Operculina granulosa LEYM.
Operculina parva DOUV.
Assilina laxispira (de la HARPE)
Discocyclina douvillei (SCHLUMBERGER)

L. GIDAI (1964, 1969, 1971a, b, 1972a, b, 1976a, b, 1977a) s'occupait à plusieurs occasions des relations stratigraphiques de l'Éocène inférieur. Il a développé ses opinions en s'appuyant aux résultats des études stratigraphiques, durées depuis plus que 100 ans avec collaboration des spécialistes. Il a attribué le complexe lignitifère inférieur, les formations sous-jacentes et le surmontant jusqu'au complexe de marne argileuse à Operculines y inclu, à l'Éocène inférieur. Il s'occupait de la limite entre l'Éocène inférieur et moyen, à plusieurs occasions (1971a, b, 1972b). Il a tracé la limite entre l'Éocène inférieur et moyen d'après l'apparition de l'association faunistique des grands Foramini-

fères caractérisée par *Nummulites perforatus*. Son opinion — selon laquelle au territoire du Bakony Méridional *Nummulites perforatus* se présente en masse dans les couches à *Nummulites laevigatus* datant sans discussion du Lutétien inférieur — est confirmée par les études les plus récentes.

D'après ses études faites sur la coupe-repère géologique des formations éocènes des environs de Sümeg et Csabrendek — sur la succession stratigraphique du sondage Cn-850. (L. GIDAI, 1977b), *Nummulites perforatus* apparaît déjà à la base du Lutétien, à la partie inférieure du groupe des couches à *N. laevigatus*. La présence commune de *N. perforatus* et *N. laevigatus* confirme son opinion préalable (L. GIDAI, 1971a, b), selon laquelle il faut compter l'étage lutétien dès l'apparition de *N. perforatus*. Les informations obtenues dans le sondage Csabrendek-850. vérifient aussi l'apparition de *N. perforatus* à la base de l'étage lutétien, dans la Montagne Centrale de Transdanubie.

L'un des résultats essentiels des études, faites à la région de Tatabánya (L. GIDAI 1977b), était la subdivision plus détaillée des formations éocène inférieur, ainsi que la constatation selon laquelle il y a deux niveaux à *Nummulites subplanulatus*, l'un au-dessous du groupe de gîtes de lignite supérieur de l'Éocène inférieur et l'autre en dessus.

3. Récapitulation

Nous pourrions récapituler les résultats de nos travaux, en ce qui suit:

1. La constatation de la succession stratigraphique juste des formations éocènes et leur subdivision présentent le mérite de M. HANTKEN et K. HOFMANN.

2. H. TAEGER et L. LÓCZY sen. ont effectué la préparation de la subdivision stratigraphique et détermination d'âge bien fondées.

3. P. ROZLOZSNIK—Z. SCHRÉTER et K. TELEGDI-ROTH ont établi l'échelle stratigraphique, toujours juste même aujourd'hui en ses traits fondamentaux.

4. E. SZŐTS a employé aussi cette échelle, en composant la corrélation stratigraphique des formations éocènes de la Hongrie. Les géologues — en étudiant les formations éocènes — travaillaient sur la base de celle-ci, la perfectionnant sur différentes régions partielles.

5. G. KOPEK—T. KECSKEMÉTI et E. DUDICH dans leur publication parue en 1966, ont inséré le complexe lignitifère dans le sousétage lutétien supérieur et dans celle — parue en 1972 — l'ont attribué au Lutétien inférieur. On peut justifier ce dernier point de vue dans à certaines régions partielles de l'Éocène de Hongrie — par ex. à Várgesztes, — car la formation lignitifère d'ici est rattachée à la première phase transgressive du Lutétien. La situation est différente aux régions lignitifères de Mór, Pusztavám. Oroszlány, Tatabánya, Dorog, etc. où la marne argileuse à Operculines — à épaisseur de 120 m au max. — à faune de petits et grands Foraminifères de faciès de mer

Tableau N° 1.

Évolution de la subdivision stratigraphique et détermination d'âge des formations éocène inférieur du Nord-Est de la Transdanubie Construit par GIDAI L., 1977.

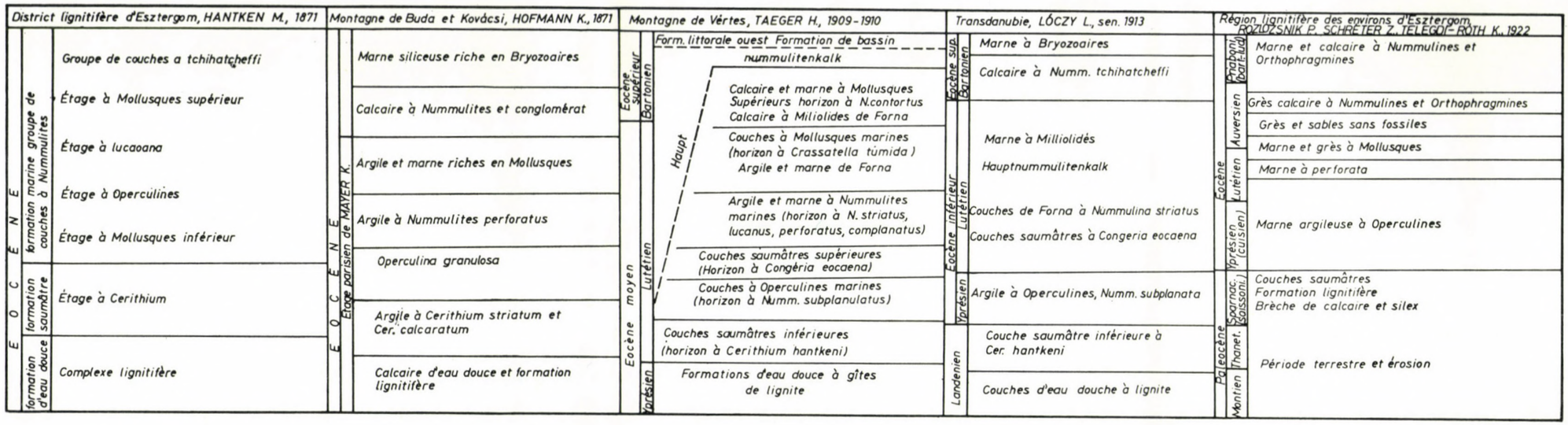
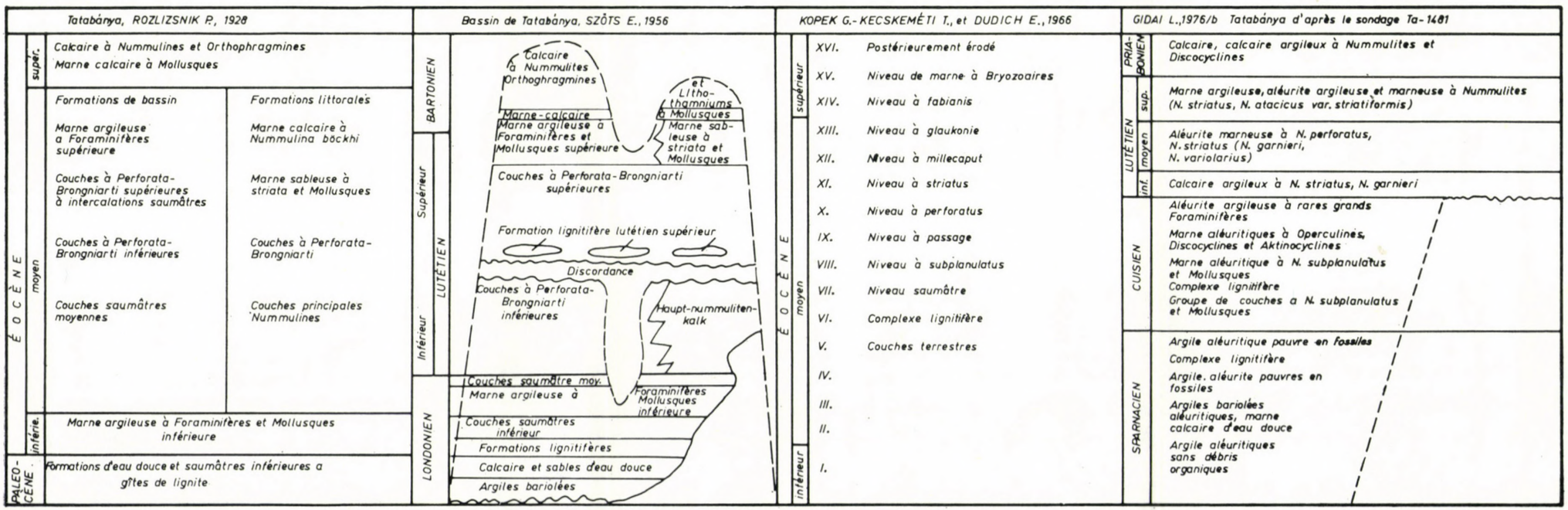


Tableau 2

Evolution de la subdivision stratigraphique et détermination d'âge des formations éocène inférieur du Nord-Est de la Transdanubie. Construit par GIDAI L. 1977



peu profonde, éocène inférieur surmonte la formation lignitifère inférieur, et seulement après se trouvent les formations éocène moyen qui recouvrent le complexe lignitifère, par ex. dans la région de Vértessomló et Várgesztes.

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СХЕМА СТРАТИГРАФИИ И ВОЗРАСТНАЯ ПРИНАДЛЕЖНОСТЬ НИЖНЕЭОЦЕНОВЫХ ОТЛОЖЕНИЙ НА СЕВЕРО-ВОСТОКЕ ЗАДУНАЙСКОГО КРАЯ

Л. ГИДАИ

Резюме

Результаты расмотренных в статье работ сводятся к следующему:

1. Определение правильной стратиграфической последовательности эоценовых отложений юго-востока Задунайского края является заслугой М. Ханткена и К. Гофманна.

2. Подготовку обоснованной стратиграфической схемы подразделения рассматриваемых отложений и определение их возрастной принадлежности выполнили Х. Тэгер и Л. Лоци старший.

3. Стратиграфическая схема, в основном действительная по сей день, была разработана П. Розложником, К. Телегди Ротом и З. Шретером.

4. Эта схема была применена Э. Сёчом, выполнившим дробное стратиграфическое расчленение эоценовых отложений Венгрии. Этой схемой и руководствовались и её усовершенствовали геологи, разработавшие стратиграфию отдельных районов страны.

5. В своем сообщении от 1966 г. Г. Копек, Т. Кечкемети и Э. Дудич отнесли нижнюю буроугольную толщу к верхнелютетскому подъярусу; в статье, вышедшей в 1972 г., к нижнелютетскому подъярусу. Правильность последнего отнесения оказывается обоснованной в некоторых участках развития эоценовых отложений на территории Венгрии — как например на участке Варгестеш —; дело в том, что угленосная толща на данном участке приурочена к первой, трансгрессионной фазе седиментационного цикла лютетского века. Однако, совсем другое положение наблюдается на угленосных участках Мор—Пуставам—Орослань, Татабанья, Дорог и т. д., где нижняя буроугольная толща подстилает мелководную нижнеэоценовую толщу мощностью до 120 м, представленную оперкулиновыми мергелями с фауной мелких и крупных фораминифер и только выше этих мергелей последуют среднеэоценовые отложения, перекрывающие в некоторых районах — например на участке Вертешшомло-Варгестеш — буроугольную толщу.

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SITUATION PALÉOGÉOGRAPHIQUE DES FORMATIONS ÉOCÈNES DU NORD-EST DE LA TRANSDANUBIE

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Après la période continentale à la fin du Mésozoïque et au début du Cénozoïque, dans la région de la Montagne Centrale de Transdanubie la première sédimentation aquatique a commencé à l'Éocène inférieur. En contournant au N le massif partiel de la Petite Plaine Hongroise un chenal marin épicontinental — en connexion au géosynclinal des Alpes occidentales — est arrivé à la région du Nord-Est de la Montagne Centrale de Transdanubie.

En direction vers le SW, la transgression de l'Éocène inférieur est arrivée jusqu'au Graben de Mór. La largeur de ce chenal marin était env. 30 km. dans la région de la Montagne Gerecse et Dorog, 10 à 15 km à l'avant—pays occidental de la Montagne Vértes et env. 10 km entre les Bassins de Dorog et de Pilis.

Entre les chenaux marins éocène inférieur du Bakony Méridional — à communication vers l'Istrie et les Dinarides — et du Nord-Est de la Transdanubie — à communication vers les Carpates occidentales — il n'y avait pas de liaison, entre eux se trouvait le seuil du massif paléozoïque et mésozoïque du Bakony Septentrional, Haut-Bakony et du Haut-plateau du Balaton. Je propose le nom du Seuil de Pápa et Veszprém pour le territoire élevé, séparant les deux chenaux marins éocène inférieur.

Vers la fin de l'Éocène inférieur la mer épicontinentale peu profonde — jusque là unie — s'est rétrécie. Certaines régions partielles ont été surélevées, et à plusieurs endroits des lagunes séparées ont été formées. Au début du Lutétien, au territoire de la trangression éocène inférieur une nouvelle trangression considérablement plus étendue s'est produite.

Au Lutétien la communication directe s'est installée entre les bassins sédimentaires du Bakony Méridional et du Nord-Est de la Transdanubie. Les contours du Seuil de TELEGDI-ROTH commençaient à développer déja à l'Éocène moyen. D'après les études faites par T. BÁLDI, le seuil de TELEGDI-ROTH existait encore à l'Oligocène (T. BÁLDI et al., 1976, T. BÁLDI, 1976).

La mer éocène supérieur a envahi un territoire plus grand que celui de la mer éocène moyen. Selon la probabilité supposée par nous, en direction vers l'W elle est avancée jusqu'à la limite d'extension de la mer éocène moyen environ. En direction vers l'E et le NE, elle a considérablement dépassé la limite d'extension de la mer éocène moyen. La mer éocène supérieur de la région du Nord-Est de la Transdanubie et celle de la Montagne Centrale du Nord et de la Grande Plaine Hongroise étaient en communication directe.

Par contre, les régions situées entre les Montagnes Vértes et Gerecse, ainsi que entre Vál, Tabajd et la Montagne de Buda présentaient une terre ferme déjà dans l'Éocène supérieur, ensemble avec le territoire des Montagnes Vértes et Gerecse. Le seuil de TELEGDI-ROTH a été développée déjà dans l'Éocène supérieur. Au SW du seuil de TELEGDI-ROTH, on peut con-

tourner un territoire long de 80 km. env. et large de 20 à 30 km — le Haut-plateau du Balaton actuel et la partie méridionale du Bakony — qui formait une île dans la mer éocène supérieur. Je propose à nommer ce terroitoire, comme seuil de Keszthely et Veszprém.

1. Introduction

La recherche de la situation paléogéographique du Nord-Est de la Transdanubie est rattachée aux intérêts importants de l'économie nationale et à la réalisation des objectifs scientifiques. La présence des gîtes de lignite à bonne qualité et le fait que les formations éocènes se trouvent dans le toit de la bauxite — l'une des plus importantes matières primaires de Hongrie — ont persuadé depuis longtemps les géologues hongrois à l'étude détaillée.

Plusieurs ouvrages traitaient aussi la situation paléogéographique. D'après l'évaluation des points de vue émis jusqu'ici, nous allons récapituler les changements plus importants de la situation paléogéographique de la région, à l'Éocène.

2. Évaluation des recherches faites jusqu'ici

C'est K. Telegdi-Roth (1923) qui a fait les premières constatations sur la situation paléogéographique des formations éocènes du Nord-Est de la Transdanubie. Selon lui la région lignitifère de Tatabánya, Dorog et Tokod et de Pilisvörösvár et Nagykovácsi a présenté le premier arrêt de la mer transgressant de la direction du NW. A partir de ces territoires-ci, dans l'Éocène les nouvelles transgressions ont pris leur départ radialement, presqu'en toutes directions. Selon K. Telegdi-Roth celles-ci formaient les golfes de la mer du Flysch, occupant le géosynclinal des Carpates, pénétrés entre les massifs, à côtes bien tailladées et à caractère d'archipel.

Nous considérons fondamentale et directive la constatation — faite par E. Szádeczky-Kardoss, (L. Ettre et al., 1952, p 160) d'après ses études pétrographiques sur les charbons — selon laquelle le dépôt de lignite des environs de Tatabánya continuait aussi au S du territoire lignitifère actuellement contournable, alors aussi à cette partie de la Montagne Vértes qui est aujourd'hui surélevée, comme substratum.

Selon E. Vadász (1953, 1960) la mer mésogéenne éocène a envahi le Bassin de Hongrie de la direction du SW et NW et l'a ateint le long du linéament géanticlinal, limitant à l'W le Bassin de Transylvanie.

Dans les esquisses paléogéographiques, publiées d'après E. Szőts, il indique les transgressions des différentes époques de l'Éocène d'être arrivées de la direction de l'Ouest. E. Szőts (1956) a émis une opinion pareille. Nous sou-

lignons la constatation faite par E. Szőts selon laquelle, à l'Éocène inférieur, un chenal marin — en partant de la région de la Dalmatie et Istrie — a pénétré au territoire de la Montagne Centrale de Transdanubie. Nous sommes d'accord sur cette constatation avec la seul réserve que le chenal marin — d'après nos connaissances actuelles — n'atteint que le territoire du Bakony Méridional.

Sur la base des sondages de recherches pétrolières, L. Kőrössy (1958) a montré l'absence des formations éocènes à la région de la Petite Plaine Hongroise.

Selon N. Mészáros et E. Dudich (1966), la mer éocène — à partir de l'Italie septentrionale — a progressivement transgressé vers le NE, à la bordure de la Montagne Centrale de Transdanubie, entre la Petite Plaine Hongroise et les îles de la Slovaquie.

D. Andrusov et E. Köhler (1963, p 156) ont mentionné qu'au milieu de la Petite Plaine Hongroise les formations paléogènes ne sont pas connues au-dessus de Néogène. Au N et au S de cette zone, le Paléogène commence par l'Yprésien. Ils ne considèrent pas exclu que l'absence du Paléogène au centre de la plaine soit le résultat de l'érosion pré-éocène et dès l'Yprésien ou le Lutétien la mer éocène des Petites Carpates a librement communiqué avec celle de Hongrie.

G. Kopek—E. Dudich, et T. Kecskeméti (1972) s'occupaient détaillément des relations paléogéographiques et de la situation paléogéographique de l'Éocène de la Transdanubie. Selon leur avis, la mer éocène inférieure est arrivée à la Hongrie de la direction du SW par suite de l'affaissement du substratum mésozoïque situé au SE de la ligne de la rivière Rába, prolongation du territoire actuel du Bakony Méridional. Ils ne considèrent pas la présence de l'Éocène inférieur acceptable à la région de Tatabánya et Dorog. Selon leur opinion, on pourrait supposer la communication entre cette région et le Cuisien du Bakony Méridional ainsi que la mer a contourné le Haut-Bakony et le Bakony du NE à travers du territoire de la Petite Plaine Hongroise actuelle. Ils ne considèrent pas probable la communication marine entre la région paléogène des Carpates Moyennes et la Montagne Centrale de Transdanubie, à l'Éocène inférieur. Ils considèrent l'origine des transgressions lutétien inférieur et moyen de la direction du SW. Au Lutétien supérieur, à l'époque de l'envahissement marin général, la communication a été établie envers le bassin paléogène intracarpatique, qui restait permanent aussi au Priabonien.

Dans la publication émise en 1971, L. Gidai a montré la présence d'un seuil (Seuil de Telegdi-Roth) à plusieurs centaines de km^2, où on ne connaît que les lambeaux plus ou moins petits ou grands séparés, épargnés de l'érosion (Zsámbék, Szomor, Bajna, Piliscsév, Piliscsaba, Pilisvörösvár, Pilisszentiván, Solymár, Nagykovácsi), entre les Montagne Vértes — Gerecse et Buda—Pilis. A ce territoire-ci, le substratum triasique est pour la plupart couvert de formations oligocènes. La direction de cette configuration paléogéographique est du NE au SW et elle sépare deux grands régions, resp. groupes de régions éocènes.

L. GIDAI (1974) a montré qu'ont peut considérer comme unités structurales, séparées les unes des autres, les régions éocènes de la colline Lencsehegy de Dorog, Buda-Pilis, Dunaszentmiklós, Vértestolna, Vértessomló et de Tatabánya. Il a considéré le seuil, situé entre la Montagne Gerecse, groupe des Horts de Tata, maximum gravimétrique de Dad et le Haut-Bakony, recouvert de formations post-éocènes, analogue au Seuil de TELEGDI-RÓTH. A celle-ci, la structure élevée de Környe, Vértessomló et de Kőhányáspuszta, érodée et couverte de formations oligocènes, est grosso modo perpendiculaire.

En 1976, L. GIDAI a constaté que l'extension territoriale de la sédimentation était plus restreinte à l'Éocène inférieur qu'elle ne l'était à l'Éocène moyen et supérieur. A l'avant-pays ouest de la Montagne Vértes, la sédimentation déroulait dans un bassin sédimentaire à direction du NE au SW. La limite nord-ouest de ce bassin sédimentaire éocène inférieur a été formée par la région de Dad, Kocs et de Tata surélevée à l'Éocène inférieur et la limite sud-est a été présentée par la Montagne Vértes. Les parties recouvertes de formations éocène moyen de ce dernier territoire-ci — bord est d'Oroszlány, Vértessomló. Várgesztes, Kálváriahegy à Tatabánya — étaient également en terre ferme, dans l'Éocène inférieur. L'espace de sédimentation éocène inférieur — dont la largeur était de 10 à 15 km à l'avant-pays ouest de la Montagne Vértes — est grosso modo contourné par l'extension de la marne argileuse à Operculines éocène à faciès de mer peu profonde. A l'époque des érosions postérieures — surtout à la limite entre l'Éocène et l'Oligocène — a été érodé le territoire entre les Bassins de Tatabánya et Oroszlány, le soi-disant seuil de Környe. A cette époque la région des environs de Mór, Pusztavám, Oroszlány et de Tatabánya avait une communication paléogéographique directe vers Dunaszentmiklós ainsi qu'à travers Héreg et Tarján vers la région de Dorog. Adhéré à la position de vue de D. ANDRUSOV et E. KÖHLER (1963), il a exprimé son opinion selon laquelle la mer éocène inférieur du Nord-Est de la Transdanubie communiquait avec celle des Carpates occidentales, à travers la partie nord de la Petite Plaine Hongroise. Les sédiments, prouvant les traces de cette communication, pouvaient être érodés ultérieurement. Après les phénomènes régressifs manifestés dans l'appauvrissement de la faune marine à la fin de l'Éocène inférieur, les transgressions, arrivées en plusieurs vagues à l'Éocène moyen, ont envahi aussi les territoires, étant jusque là en terre ferme, aux bordures nord-ouest (Dad, Kocs, Tata) et sud, sud-est (avant-pays ouest de la Montagne Vértes) de l'aire sédimentaire éocène inférieur. Le dépôt de lignite de Vértessomló, Várgesztes et du Bakony Septentrional est rattaché à ces transgressions.

3. Aperçu stratigraphique

Au territoire du Nord-Est de la Transdanubie l'extension des formations éocènes est grande. A l'avant-pays ouest de la Montagne Vértes, à Tatabánya et dans les environs de Héreg et Tarján leur épaisseur varie entre 50 et 350 m. Dans le Bassin de Dorog, au territoire de la dépression structurale de Nagysáp et Kenyérmező leur épaisseur maximale atteint 400 m, tandis que dans les bassins de Pilis cette valeur est de 50 à 250 m, et dans la Montagne de Buda 50 à 200 m. Aux territoires des volcanites, situés au Sud de Budapest, leur épaisseur maximale dépasse même 500 m.

Les Annexes n^{os} I, II, III, et IV présentent un aperçu sur les positions stratigraphiques des formations éocènes du Nord-Est de la Transdanubie, et sur leurs possibilités de subdivision, corrélation et de détermination d'âge. Nous avons construit ces esquisses d'après les successions stratigraphiques des sondages étudiées, en tant que coupes-repères, tout récemment. Ces esquisses de corrélation illustrent et approuvent aussi ce qui a été dit sur l'évolution de la situation paléogéographique. Faisant allusion aux Annexes n^{os} I. à IV., nous présentons l'aperçu suivant sur les positions stratigraphiques des formations éocènes du territoire.

Étage sparnacien

Groupe de couches du mur du complexe lignitifère

Composé de couches d'argiles bariolées — alternant surtout avec couches de sables — à faciès d'eau douce et continental et de calcaire d'eau douce. L'épaisseur varie, en général, entre 0 et 20 m, mais dans les dépressions géomorphologiques contemporaines à situation favorable, surtout à la partie occidentale du territoire de Dorog, elle atteint même 120 m (Annexes n^{os} I. à IV.)

Complexe lignitifère

Le complexe lignitifère — contenant des intercalations de calcaire d'eau douce et saumâtres — développe progressivement du complexe du mur, en relation étroite avec celui-ci. En considérant la répartition territoriale, elle est identique à celle du complexe du mur. La plus importante différence: à la partie occidentale du territoire de Dorog (grosso modo à l'Ouest de la ligne entre Bajót et Bajna), ainsi qu'au territoire de la Montagne Gerecse il n'y a plus de gîtes de lignite exploitable (Annexes n^{os} I. à IV.)

Étage cuisien

Marne argileuse à subplanulatus et Operculines

Dans le cas optimal, l'épaisseur du complexe à subplanulatus et Operculines, surmontant immédiatement le complexe lignitifère, atteint 120 m. Les

valeurs d'épaisseur les plus fréquentes varient entre 20 et 40 m. Le complexe est composé, en prédominance, de marnes argileuses aléuritiques, contenant dans la partie inférieure une couche de sable quartzifère et grès, brun, épaisse de quelques mètres. Au tiers inférieur les faciès saumâtres et marins s'altèrnent, ici on peut observer de nombreuses couches de Lumachelles, composées de coquilles et fragments de coquilles de Mollusques, épaisses de 10 à 20 cm. Les deux tiers supérieurs présentent le faciès de mer peu profonde. Il contient, en grande richesse en espèces et individus, des petits Foraminifères, Mollusques, par endroits des grands Foraminifères en grande abondance, même en masse. D'après la détermination faite par Mmes L. VITÁLIS-ZILAHY, et M. JÁMBOR-KNESS, tous les grands Foraminifères et ceux planctoniques prouvent l'âge éocène inférieur.

En dépendant des conditions faciologiques, aux différentes unités territoriales nous l'avons divisé — surtout aux environs de Dorog et de Tatabánya — en plusieurs groupes de couches moins épais. (Annexes n^{os} I. à IV.)

Étage lutétien

Dans l'étage lutétien, l'image paléogéographique du Nord-Est de la Transdanubie était beaucoup plus variée que dans les étages sparnacien et cuisien. Nous y pouvions distinguer trois secteurs faciologiques plus étendus:

A) Environs de Dorog, Tatabánya, Pilisvörösvár et Solymár

B) Avant-pays ouest et sud-est de la Montagne Vértes, bordure sud de la Montagne Gerecse

C) Environs de Dad, Kocs et de Tata.

A) *Environs de Dorog, Tatabánya, Pilisvörösvár et Solymár*

A ces territoires-ci, les formations du Lutétien surmontent les sédiments de l'Éocène inférieur. Le passage est continu à la partie importante du territoire (Annexe n° IV).

B) *Avant-pays ouest et sud-est de la Montagne Vértes, bordure sud de la Montagne Gerecse*

A cette unité territoriale appartiennent: la colline Antalhegy de Mór, bordure est d'Oroszlány, Várgesztes, Vértessomló, la colline Kálváriahegy de Tatabánya, Nagyegyháza—Csordakút—Mány, Zsámbék, la colline Őrhegy de Bajna, la colline Quadriburg de Sárisáp et les environs de Gánt et Csákberény. A cette unité faciologique — alignée grosso modo en direction du NE au SW — il est caractéristique que les formations éocène moyen surmontent directement le substratum mésozoïque. A la partie prédominante du territoire, il y avait de dépôt de lignite à la base de l'Éocène moyen. Une autre caractéristique bien importante: les formations de calcaire, marne calcaire et de marne prédominent dans l'Éocène moyen (Annexes n^{os} I. et II).

C) *Environs de Dad, Kocs et de Tata*

Ce territoire, situé au NW du territoire éocène inférieur de Pusztavám, Oroszlány et de Tatabánya, diffère de celui précédent en ce qu'en dessous de l'Éocène moyen à faciès de calcaire se trouve aussi l'Éocène inférieur mince prouvable aussi par faune marine. L'Éocène inférieur ne contient pas de gîtes de lignite exploitables (Annexe n° I).

Étage priabonien

Les deux territoires à plus grande extension de l'Éocène supérieur se trouvent entre les environs de la Montagne de Buda et le Lac Balaton ainsi qu'aux environs de Dorog. Dans la Montagne Vértes, l'Éocène supérieur présente un territoire continu plus vaste aux environs de la colline Antalhegy de Mór, et ses petits lambeaux sont connus à la bordure occidentale de la montagne, aussi.

Au S de la Montagne de Buda, se présente un complexe éocène moyen et supérieur, composé surtout de produits pyroclastiques de l'andésite — rattaché au volcanisme de la Montagne de Velence — épais de 500 m au max., indiqué par les mesures géophysiques et prouvé par plusieurs sondages.

4. Évolution de la situation paléogéographique

Après la période continentale à la fin du Mésozoïque et au début du Cénozoïque, à la région de la Montagne Centrale de Transdanubie la première sédimentation aquatique a commencé à l'Éocène inférieur. Un chenal marin — en connexion avec le géosynclinal des Alpes occidentales — a pénétré au territoire du Nord-Est de la Montagne Centrale de Transdanubie, en contournant au Nord le massif partiel de la Petite Plaine Hongroise. La mer, transgressant progressivement à la terre ferme, a envahi le territoire des Bassins de Dorog et de Pilis, la Montagne Gerecse, le Bassin de Tatabánya, et l'avant-pays ouest de la Montagne Vértes. Une vague probablement postérieure de la transgression éocène inférieur a envahi aussi une partie du territoire du bassin de Kocs. Cette transgression progressive de la mer éocène inférieur a résulté le dépôt de la séquence sédimentaire d'après un cycle sédimentaire continu. La transgression de l'Éocène inférieur est arrivée jusqu'au Graben de Mór en direction vers le SW. (Nous considérons la poursuite précise de limite de la transgression ici, comme un devoir suivant important.) La largeur de ce chenal marin est env. de 30 km au territoire de Dorog et de la Montagne Gerecse, 10 à 15 km à l'avant-pays ouest de la Montagne Vértes et env. 10 km au territoire situé entre les Bassins de Dorog et de Pilis. Selon notre opinion, dans l'Éocène inférieur le territoire situé entre les Montagne Vértes et de Buda,

la Montagne de Buda et Pilis s. s. et la Montagne de Szentendre et Visegrád entière formaient des terres fermes. Il n'y avait pas de connexion entre les chenaux marins du Bakony Méridional à communication vers l'Istrie et les Dinarides et du Nord-Est de la Transdanubie à communication vers les Carpates occidentales; entre eux, le massif paléozoïque et mésozoïque du Haut-Bakony

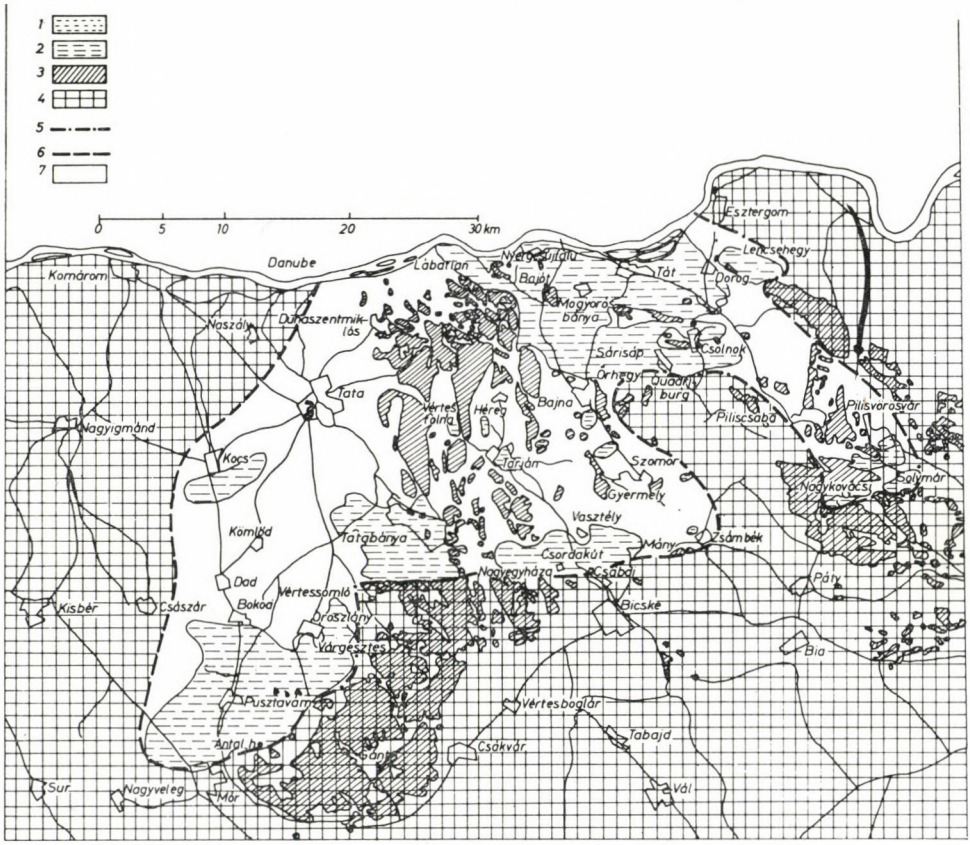


Fig. 1. Esquisse de l'extension des formations éocène inférieur du Nord-Est de la Transdanubie. Construite par L. GIDAI, 1977. 1. Formations éocène inferieur à la surface; 2. Formations éocène inférieur recouvertes; 3. Formations mésozoïques à la surface; 4. Terre ferme dans l'Éocène inférieur; 5. Limite probable de la sédimentation éocène inférieur; 6. Limite supposée de la sédimentation éocène inférieur; 7. Territoire recouvert jadis de formations éocène inférieur, actuellement de celles plus récentes

et du Haut-plateau du Balaton a formé un seuil. (Nous pourrions nommer le territoire surélevé, séparant les deux chenaux marins éocène inférieur, comme Seuil de Pápa et Veszprém.) Nous avons assemblé les formations calcaires éocène inférieur et moyen du Bakony Méridional. Selon CH. POMEROL, Le Paléocène durait env. 10 millions d'années. En ce sens, les parties du territoire

du Nord-Est de la Transdanubie recouvertes de sédiments éocène inférieur ont été émergées pendant 10 millions ans environ depuis le début du Cénozoïque.

Vers la fin de l'Éocène inférieur, la mer épicontinentale peu profonde, jusque-là unie, s'est rétrécie. Certains territoires partiels ont été surélevés, à plusieurs endroits des lagunes isolées se sont installées. Au début du Lutétien,

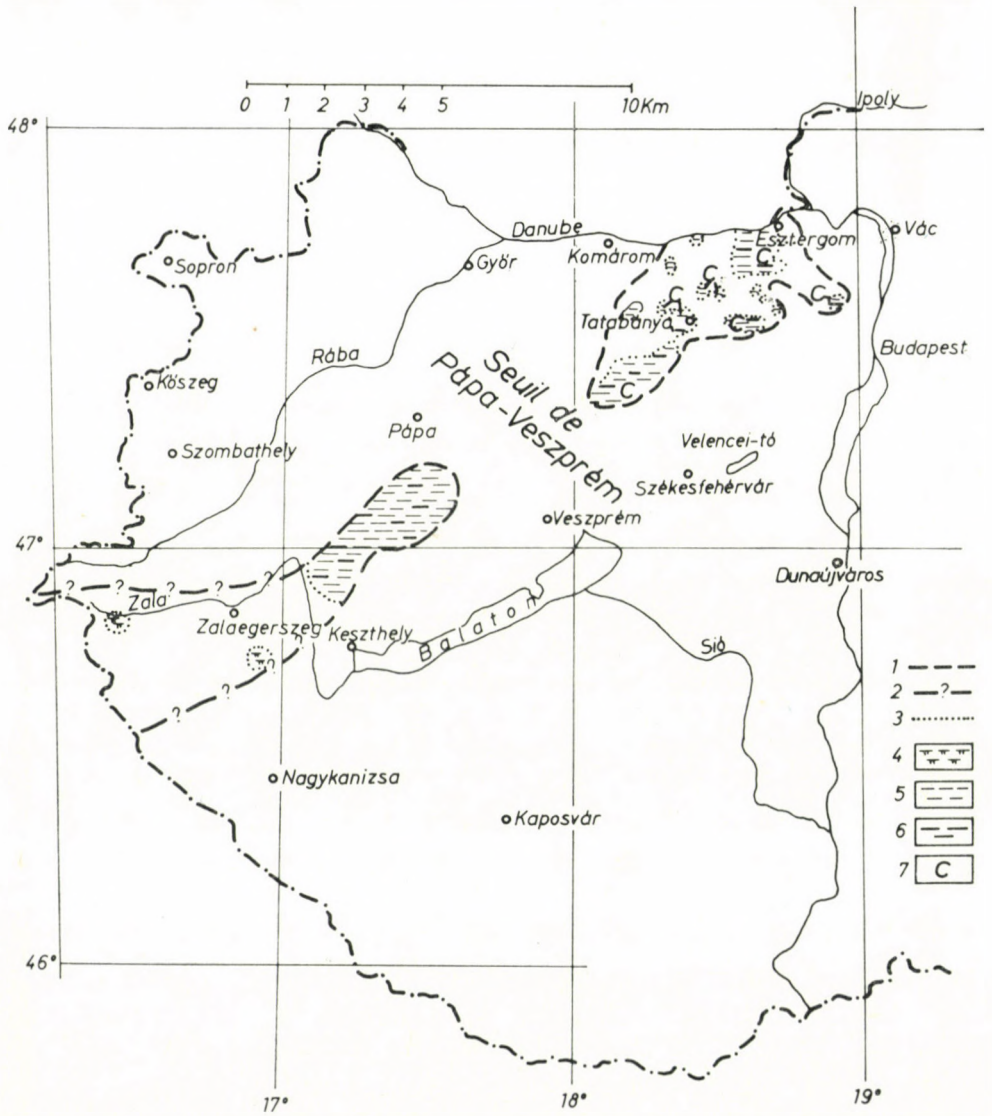


Fig. 2. Esquisse lithofaciologique des formations éocène inférieur de la Transdanubie. Construite par L. GIDAI, 1976. 1. Limite probable du bassin sédimentaire; 2. Limite incertaine du bassin sédimentaire; 3. Limite d'extension; 4. Marne; 5. Argile; 6. Argile ligniteuse; 7. Lignite

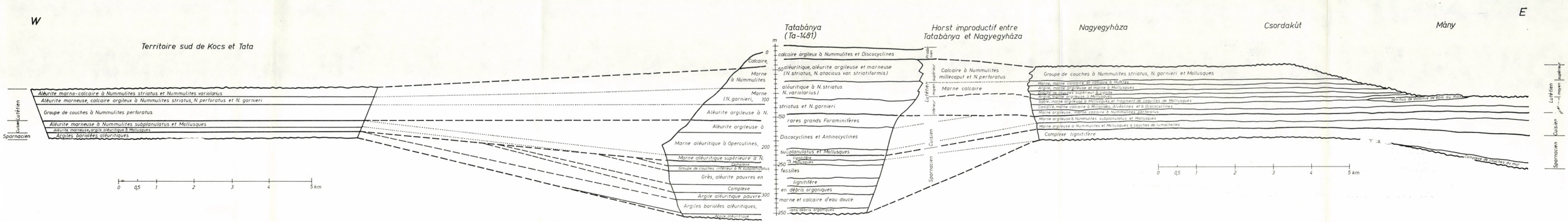
au territoire de la transgression éocène inférieur une nouvelle transgression bien plus étendue a commencé.

Au NW (Kocs, Dad, Tata), au SW (Bakony Septentrional, avant-pays du Bakony Septentrional), au SE (bordure ouest de la Montagne Vértes) et au S (Csákberény, Gánt, Tabajd, Páty) du territoire envahi à l'Éocène inférieur,

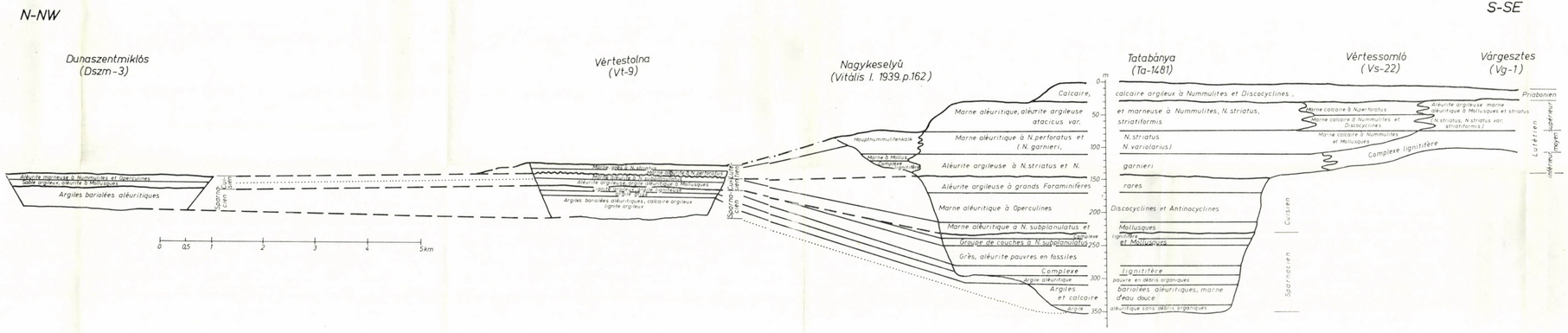


Fig. 3. Esquisse de l'extension des formations éocène moyen de la Transdanubie. Construite par L. GIDAI, 1977. 1. Formations éocène moyen à la surface; 2. Formations éocène moyen recouvertes; 3. Formations mésozoïques à la surface; 4. Terre ferme dans l'Éocène moyen; 5. Limite probable de la sédimentation éocène moyen; 6. Limite supposée de la sédimentation éocène moyen; 7. Andésites éocène moyen et ses produits pyroclastiques; 8. Territoires recouverts jadis de formations éocène moyen, actuellement de celles plus récentes

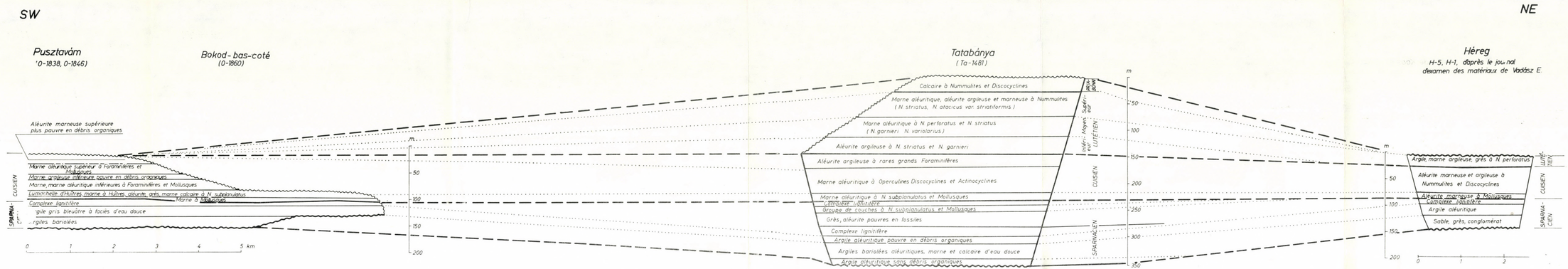
des nouveaux territoires, jusque-là étant terre ferme, ont subi l'envahissement marin. Le dépôt des gîtes de lignite du Bakony Septrentrional, de la bordure ouest de la Montagne Vértes (colline Antalhegy à Mór, Vértessomló, Várgesztes) ainsi qu'une partie des gîtes supérieurs de Pilisvörösvár et de la colline Lencsehegy à Dorog est rattaché à cette transgression. Au Lutétien, la communication directe a été établie entre les bassins sédimentaires du Bakony Méridional



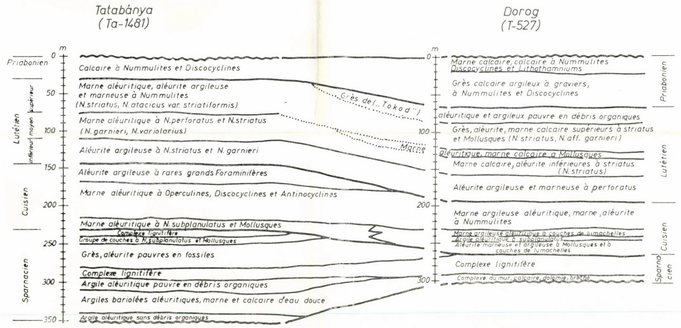
Annexe n° 1. Esquisse de corrélation entre les formations éocènes du territoire de Kocs, Tatabánya, Nagyegyháza, Csordakút et de Mány



Annexe n° 2. Esquisse de corrélation entre les formations éocènes des environs de Dunaszentmiklós, Vértestolna, Tatabánya, Vértessomló et de Várgesztes



Annexe n° 3. Esquisse de corrélation entre les formations éocènes de Pusztavám, côté-bas de Bokod, Tatabánya et de Héreg



Annexe n° 4. Esquisse de corrélation entre les formations éocènes de Tatabánya et Dorog

et du Nord-Est de la Transdanubie. A la fin du Lutétien des émersions locales (partie ouest du territoire de Dorog, Nagyegyháza—Csordakút—Mány, Kocs) et des remblayages (Dorog, Tokod, Csolnok) ont eu lieu. Dans l'Éocène moyen on peut considérer probable présence d'une terre ferme continuelle, située à l'Ouest de la ligne entre Kisbér, Nagyigmánd et le mont Naszály à Vác. A l'Éocène moyen, la Montagne de Pilis s. s. et la Montagne de Szentendre et Visegrád ainsi que la Montagne de Buda s. s., ensuite probablement la Montagne Vértes et le territoire situé à l'Est de celle-ci aux environs de Csákvár, Vértesboglár

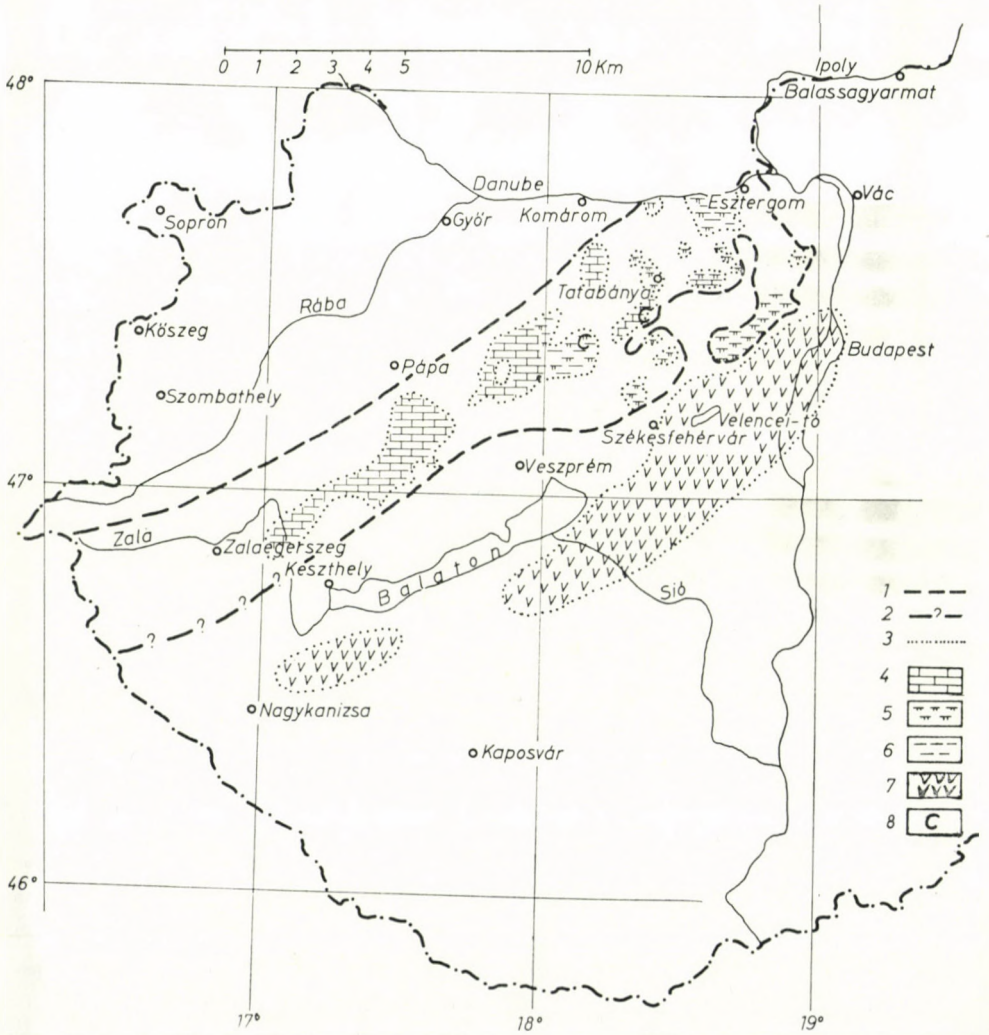


Fig. 4. Esquisse lithofaciologique des formations éocène moyen de la Transdanubie. Construite par L. Gidai, 1976. 1. Limite probable du bassin sédimentaire; 2. Limite incertaine du bassin sédimentaire; 3. Limite d'extension; 4. Calcaire; 5. Marne; 6. Argile 7. Volcanites (andésites et leurs tufs); 8. Lignite

et Bicske ont aussi été émergées. Les contours de Seuil de Telegdi-Róth commençaient à évoluer déjà à l'Éocène moyen. D'après les études faites par T. BÁLDI, le Seuil de TELEGDI-ROTH persistait encore dans l'Oligocène aussi (T. BÁLDI et al., 1976). A ce territoire en voie d'érosion a été rattaché à l'E le territoire situé entre la Montagne de Buda et la Montagne de Velence, où — selon nos connaissances actuelles — des sédiments aquatiques n'ont pas été déposés à l'Éocène moyen. Dans le golfe — pénétré entre les deux terres fermes ci-mentionnées — des tufs andésitiques ont été déposés en quantité considérable. Les territoires à substratum mésozoïque recouvert de formations éocène moyen étaient émergés pendant 10 millions ans environs depuis le début du Cénozoïque.

La mer de l'éocène supérieur a couvert un territoire bien plus vaste que celle éocène moyen. Selon notre supposition probable, en direction vers l'W elle s'est avancée jusqu'à la limite de l'extension de la mer de l'Éocène moyen. En direction vers l'E et le NE elle a considerablement dépassé la limite de l'extension de la mer éocène moyen. La mer éocène supérieur du Nord-Est de la Transdanubie a directement communiqué avec la mer éocène supérieure de la Montagne Centrale Septentrionale et de la Grande Plaine de Hongroise. Le territoire situé à l'W de la ligne entre Kisbér, Nagyigmánd et Naszály ainsi que le territoire à extension relativement plus réduite, situé à l'W d'Esztergom semblent avait été émergés.

Au seuil infraoligocène — situé entre les Montagnes Vértes et Gerecse, Tabajd et Vál et les Montagnes de Buda et de Pilis — les formations éocène supérieur manquent sur un territoire continu plus étendu. Il est nécessaire à l'avenir ce que l'absence des formations éocène supérieur a été causée entièrement par l'érosion infraoligocène ou peut-être certains territoires partiels ont formé des terres fermes, pareillement à la situation paléogéographique, à l'Éocène moyen. Selon notre avis, il y avait une communication directe entre la mer éocène supérieur du territoire de Dorog et celle de la Montagne de Buda. Par contre, le territoire situé entre les Montagnes Gerecse et Vértes, Vál et Tabajd et la Montagne de Buda a été déjà probablement en terre ferme, ensemble avec le territoire des Montagnes Vértes et Gerecse. Le seuil de TELEGDI-ROTH a été déjà évolué à l'Éocène supérieur. A SW du seuil de TELEGDI-ROTH — Haut-plateau du Balaton actuel et la partie méridionale du Bakony — on peut contourner un territoire long de 80 km environ et large de 20 à 30 km — qui a formé une île dans la mer éocène supérieur (Dorsale de Keszthely et Veszprém.)

En extrapolant les informations gagnées au territoire de Dorog, Montagne de Buda, Leányfalu et de Kósd et Vác, on peut supposer que les formations éocène supérieur sont présentes aussi à ce territoire-ci.

Les territoires mésozoïques recouverts de formations éocène supérieur étaient emergés pendant 22 millions ans depuis le début du Cénozoïque.

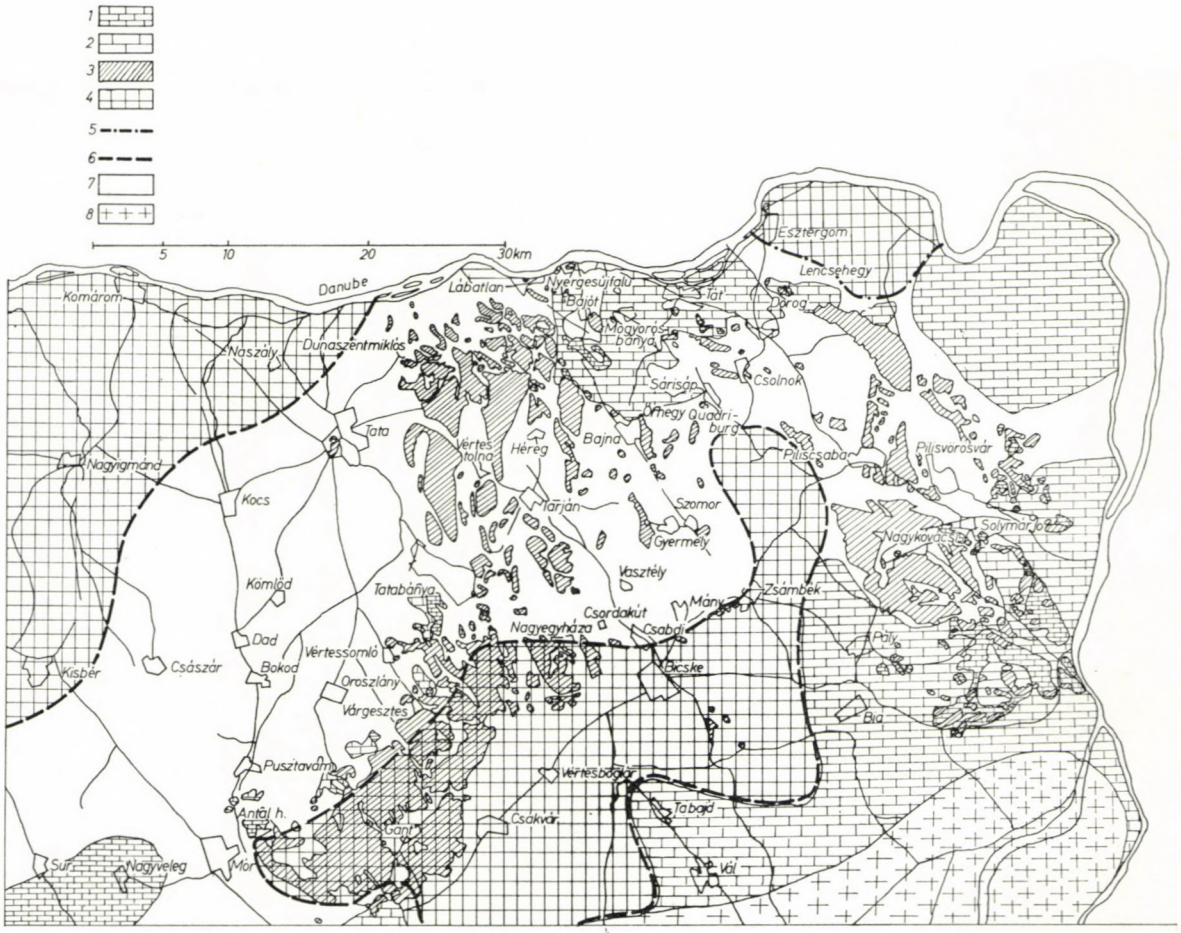


Fig. 5. Esquisse de l'extension des formations éocène supérieur du Nord-Est de la Transdanubie. Construite par L. GIDAI, 1977. 1. Formations éocène supérieur à la surface; 2. Formations éocène supérieur recouvertes; 3. Formations mésosoïques à la surface; 4. Terre ferme dans l'Éocène supérieur; 5. Limite probable da la sédimentation éocène supérieur; 6. Limite supposée de la sédimentation éocène supérieur; 7. Territoires recouverts jadis de formations éocène supérieur, actuellement de celles plus récentes; 8. Andésite éocène supérieur et ses produits pyroclastiques

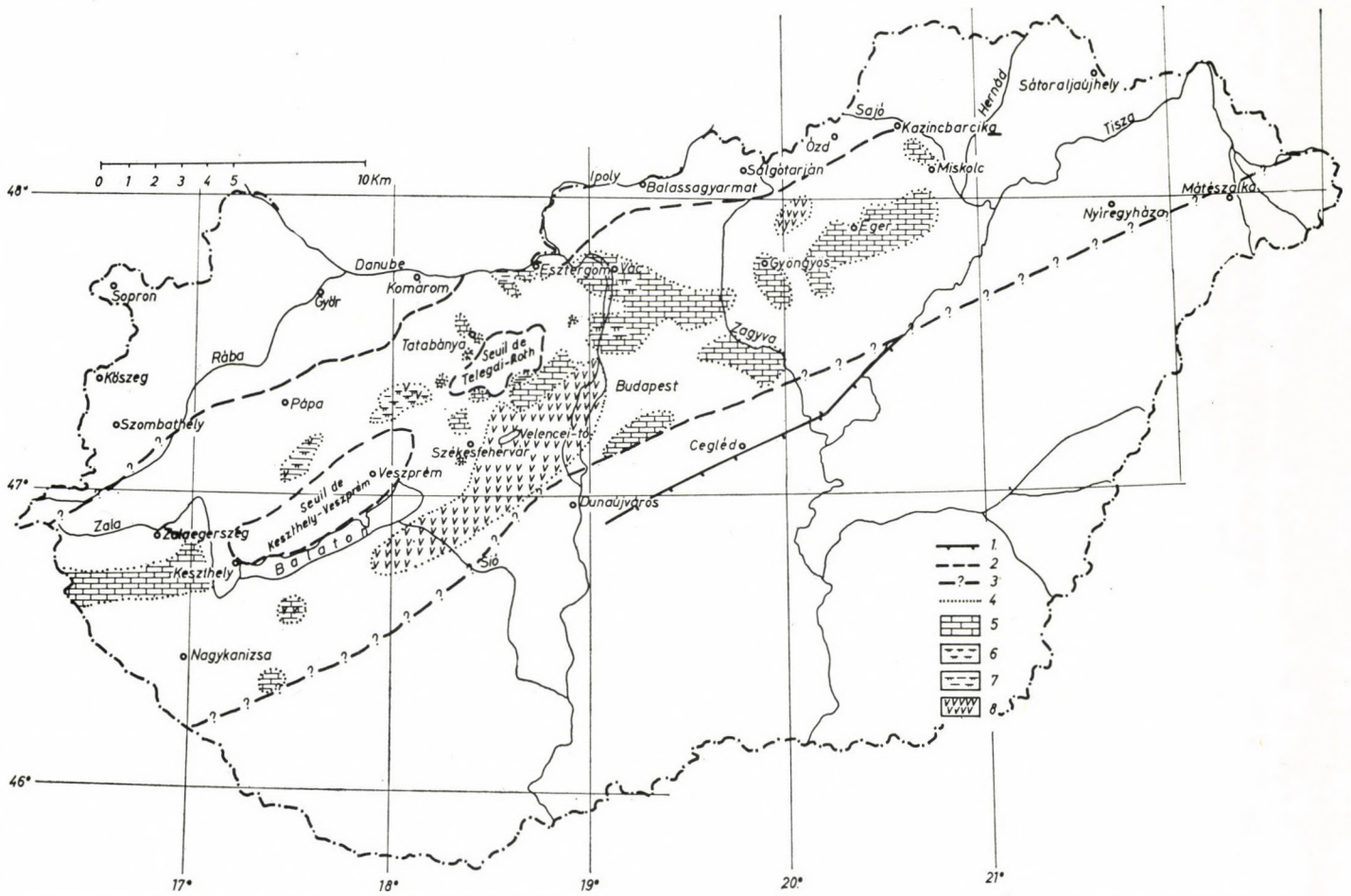


Fig. 6. Esquisse lithofaciologique des formations éocène supérieur de la Transdanubie et du Nord-Est de la Hongrie. Construite par L. GIDAI, 1976. 1. Ligne structurale principale; 2. Limite probable du bassin sédimentaire; 3. Limite incertaine du bassin sédimentaire; 4. Limite d'extension; 5. Calcaire; 6. Marne; 7. Marne argile; 8. Volcanites (andésites et leurs tufs)

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ПАЛЕОГЕОГРАФИЧЕСКИЕ УСЛОВИЯ ЭОЦЕНОВЫХ ОТЛОЖЕНИЙ НА СЕВЕРО-ВОСТОКЕ ЗАДУНАЙСКОГО КРАЯ

Л. ГИДАИ

Резюме

После континентального периода конца мезозоя — начала кайнозоя на территории Задунайского Среднегорья первая седиментация в водной среде началась в нижнеэоценовое время. На территорию северо-восточного участка Задунайского среднегорья втянулся рукав эпиконтинентального моря, обшедший субмассив Малой Венгерской (или Дунайской) низменности на севере и примыкавший к геосинклинали Западных Альп. Нижнеэоценовая трансгрессия в юго-западном направлении дошла до грабена Мор. Ширина этого морского рукава на Гереше-Дорогском участке составляла примерно 30 км, в западной предгорной зоне Вертеша — 10—15 км, а на участке между Дорогско-Пилишскими бассейнами примерно 10 км.

Сообщения между морским рукавом Южной Бакони, имевшим связь с Истрийско-Динарской областью и руковом СВ Задунайского края, связанным с территорией Западных Карпат, не было, так как в виде уступа между ними возвышалась палео-мезозойская масса Северной Бакони, Высокой Бакони и Балатонского нагорья. Приподнятый участок, разделивший рассматриваемые два нижнеэоценовых морских рукава, автор предлагает наименовать Папа—Веспремским уступом.

Под конец нижнего эоцена эпиконтинентальное мелководное море, до тех пор не разобщенное, стало значительно более узким. Некоторие участки при этом поднялись, в ряде мест возникли изолированные лагуны. В начале лютетского века наступила трансгрессия, значительно вышедшая за пределы территории, затронутой нижнемеловой трансгрессией.

Во время лютетского века создалось прямое сообщение между Южно-Баконским и Северо-Восточно-Задунайским осадочными бассейнами.

Очертания уступа им. Телегди Рота стали опознаваться уже в среднеэоценовой эпохе. По данным исследований, проведенных Т. Бальди, гребень им. Телегди Рота сушествовал даже в олигоценовой эпохе. (Т. Бальди и др. 1976, Т. Бальди 1976.)

Верхнеэоценовое море занимало большую территорию, чем среднеэоценовое. По предположению автора оно в западном направлении распространилось до пределов распространения среднеэоценского моря, но в восточном и севере- восточном направлениях оно значительно вышло за пределы распространения этого моря. Верхнеэоценовое море территории СВ Задунайского края при этом вступило в прямую связь с верхнеэоценовыми морями среднегорной зоны Северной Венгрии и Большой Венгерской низменности.

Однако, территория между районами Вертеш-Гереше, Вал-Табайд и Будайскими горами представляла собою, по-видимому, уже сушу в верхнеэоценовое время, вместе с территориями гор Вертеш и гор Гереше. Гребень им. Телегди Рота образовался уже в верхнеэоценовой эпохе. Юго-западнее гребня им. Телегди Рота — в южной части современного Балатонского нагорья и гор Баконь — оконтуривается зона шириной 20—30 км, выступавшая в виде острова из окружавшей ее морской среды. Её автор статьи предлагает наименовать Кестхейско—Веспремским гребнем.

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LOW-GRADE METAMORPHISM OF PALEOZOIC SEDIMENTARY FORMATIONS OF THE SZENDRŐ MOUNTAINS (NE-HUNGARY)

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The Silurian(?)-Devonian formations of the monoclinal folded structure of the Szendrő Mountains are built up by metamorphic rocks originated from varying and repeating mixtures of carbonatic (mostly limy) and pelitic-psammitic sediments. The averages of mineral composition of the formations and rock types determined by mineralogical-petrological investigations in the basic profiles of boreholes show cyclic and monotonous trends in the stratigraphic column. (The cyclic changes of calcite content as well as the monotonous change of the quartz, feldspar, sericite and dolomite contents support the geological model concerning the normal bedding of the monoclinal structure and the continuous sedimentation.)

In the vertical profiles the clay mineral parageneses (mixed-layer structure clay minerals, kaolinite ± montmorillonite) produced by secondary effects (weathering at the surface and along faults, resp. low temperature hydrotherms) and the goethite formed by oxidation can be fairly well distinguished from the metamorphic mineral assemblages (quartz, albite, (potash-feldspar), sericite, chlorite, calcite, ± dolomite, epidote, paragonite, pyrite, hematite, graphite).

The interconnected effects of the burial and dynamometamorphic recrystallization phases can be traced by means of rock textural and structural investigations.

On the basis of the illite crystallinity indices (KÜBLER- and WEBER indices) the regional metamorphism of the sequence of highly probably Hercynian age is assigned to the low temperature zone of low-grade metamorphism (greenschist facies, quartz--albite-muscovite-chlorite subfacies), resp. to the adjoining zone of low-grade and very-low-grade metamorphism (anchi-epi-zones).

The $b_0 - d_{002}$ diagram of sericites [11] can be fairly used to the approximate determination of the chemical composition of sericites. Parallel with the increasing carbonate content of the rocks the b_0 value (Fe + Mg content) of sericites increases, while the d_{002} value slightly decreases (the Na/Na + K ratio slightly increases).

The distribution of the b_0 parametres of the sericite-chlorite phyllites of low carbonate content ($\bar{x} = 9.003$, $s = 0.007$, $n = 64$) shows a pressure range corresponding to the boundary of the low- and medium-pressure (Barrow) ranges (about 2.5—3 kbar). This fact relates to the diversity of pressure conditions of the Hercynian metamorphism within the region of the Alps and Carpathians, or may be connected with Alpine metamorphic effect.

Introduction

In the northern part of Hungary the Early Paleozoic, predominantly pelitic-psammitic, subordinately carbonatic sedimentary and acidic-basic volcanic formations, outcropping in smaller spots form a continuous Hercynian low-pressure anchi-epimetamorphic belt. This belt shows connections with the Paleozoic of the Graz Basin to the west, with the Cambro-Silurian Gelnica Series and Devonian(?) Rakovec series of the Spissko-Gemerské Rudohorie

(W-Carpathians) to the northeast (SZÁDECZKY-KARDOSS [27, 28], SZEPESHÁZY [29], JANTSKY [12]).

Within this belt lithological and metamorphic grade differences are found depending on the stratigraphic and/or tectonic position and which have been hardly studied and evaluated. The metamorphic grade of the Mesozoic formations assigned also to this belt shows considerable differences (ÁRKAI and VICZIÁN [3]). From the spatial and temporal changes of the metamorphic grade conclusions can be drawn as to the regularities of intensity changes of different tectometamorphic cycles connected to the crustal structure.

The metamorphic-petrological investigation of the stratigraphic basic profiles of the North-East-Hungarian Szendrő, Uppony and Bükk Mountains chosen by FÜLÖP [9] aims to solve this problem and is sponsored by the Central Office of Geology.

The aim of this paper is to present the first results of this work, i.e. the metamorphic-petrological characterization and comparative evaluation of the Paleozoic formations of the Szendrő Mountains.

Geology

The geological sketch of the Szendrő Mountains (Fig. 1) was constructed after BALOGH [4] having applyed the data of JÁMBOR [13], RAINCSÁK-KOSÁRY [22] and PÉRÓ. The formations and members distinguished by FÜLÖP [9] are also demonstrated.

The Rakaca Crystalline Limestone Formation is of about 1500 m, the Szendrő Phyllite Formation is of about 1500 m thickness, while the Szendrőlád and Borsod Limestone Formations combined are of about 3400 m thickness (BALOGH [5]). The northern and southern part of the mountain are built up predominantly by carbonates, the middle part consists of pelitic-psammitic formations.

The Paleozoic sequence is predominantly of southern, southeastern dip and forms a monoclinal folded structure. Disregarding the smaller displacements along bedding planes and faults depending on the different dynamic behaviour of the rock bodies of different mechanical features, the boundaries between the formations are atectonic, and continuous sedimentation can be assumed (JÁMBOR [13], BALOGH [4, 5], MIHÁLY [18]).

Assuming a normal stratigraphic sequence the northern Rakaca Crystalline Limestone Formation can be considered to be the oldest and the southern Borsod Limestone Formation the youngest part. Paleontological ages are available only from the 3. Szendrőlád Limestone Formation (MIHÁLY [19]): Middle Devonian (Eifelian-Givetian) Tabulata and Crinoidea. On the basis of continuous sedimentation the age of the Rakaca Crystalline Limestone and of the Szendrő Phyllite Formations can be assumed to be Early Devonian (probably Silurian), while that of the Borsod Limestone Formation to be Middle or Late Devonian.

On the basis of similarities between the northern and southern carbonate members RAINCSÁK-KOSÁRY and MIHÁLY [22, 23] considered the monoclinal structure to be an anticline overfolded in SE-NW direction. (Accordingly, the central clastic sequence could be the oldest one (Silurian-Early Devonian), the carbonate formations would be of Middle and Late Devonian age and the sedimentation would have been completed by the so-called Irota Schist Series being assigned in this paper to the Szendrő Phyllite Formation.) This tectonic concept as well as the views concerning the Silurian age of the Szendrő Phyllite Formation and the Ordovician age of the Rakaca Crystalline Limestone (BALOGH [5], NAGY and SZEPESHÁZY [20]) are insufficiently evidenced but may be significant when elaborating the tectonic-stratigraphic synthesis of the mountains.

The change of petrological parameters will be investigated on the basis of the theoretical stratigraphic sketch assuming a normal succession of beds. However, the data supposing, resp. denying the theory of symmetric structure (anticline) will be mentioned, as well.

From the metamorphic-petrological point of view the Paleozoic of the Szendrő Mountains is scarcely known. Only some references mention the anchi--epimetamorphic character of the sequence in general [27]. According to RAINCSÁK-KOSÁRY and MIHÁLY [23] the degree of transformation of the sequence as a whole falls below the epimetamorphic degree (though in the northwestern and southeastern margin of the mountains epimetamorphic rocks can also be found). The metamorphic degree of the central clastic sequence was believed to be lower just due to the false genetic interpretation of the clay minerals of mixed-layer structure. Paragonite had been described from the Irota Schist Member and it was interpreted as the criterion of the epi-zone.

The sequence of sedimentary origin does not contain critical facies-indicating metamorphic minerals. Statistical mineral-paragenetic and mineral--structural studies allowing the accurate determination of the metamorphic grade have not been carried out.

Sampling, methods of investigation

Nearly 400 samples of the uninterrupted core material of about 2100 m length of 9 bores (Fig. 1) disclosing the stratotypes and boundaries of the Paleozoic formations were investigated. By means of relatively uniform sampling (the number of samples were proportional to the thickness of rock types of each formation) the statistic processing of petrological data and the determination of the average parametres could be carried out.

The microscopic investigations aimed the determination of the rock structure, recrystallization, average grain size, and of the degree of transformation of clastic constituents, opaque minerals and disperse coalified plant remnants (the latter one was performed by Z. HORVÁTH).

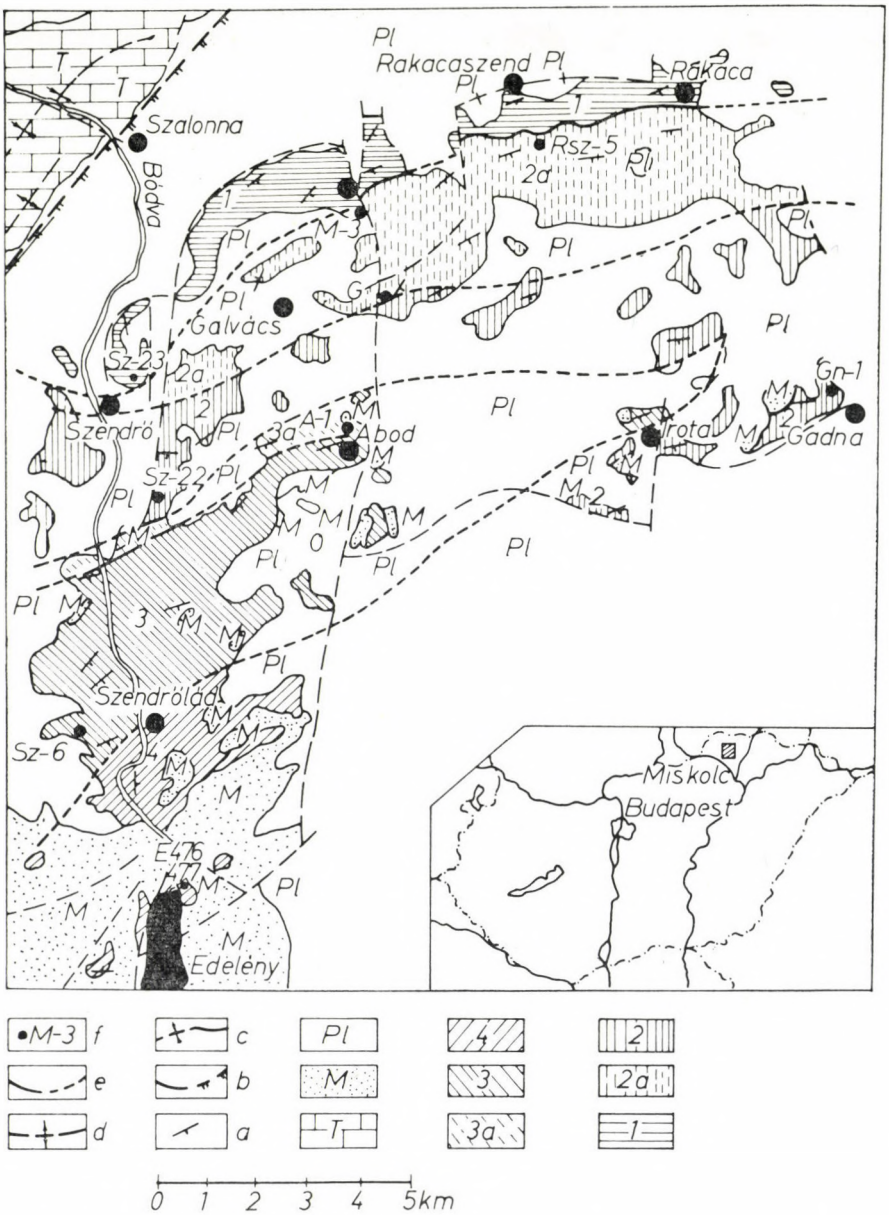


Fig. 1. Geological sketch of the Szendrő Mountains. 1—4: Paleozoic: 1. Rakaca Crystalline Limestone Formation; 2/a. Transitional Member (Authigenic-brecciated limestone, banded schist); 2. Szendrő Phyllite Formation; 3/a. Abod Limestone Member; 3. Szendrőlád Limestone Formation; 4. Borsod Limestone Formation. *T:* Middle Triassic carbonate sedimentary rocks of the Rudabánya Mountains; *M:* Miocene clastic sedimentary rocks and rhyolite tuff; *Pl:* Pliocene clastic sedimentary rocks with brown coal; a) Direction of strike/dip; b) Fault, overthrust; c) Axis of syncline: d) Axis of anticline; e) (Mostly atectonic) boundary of Paleozoic formations; f) Borehole investigated

The X-ray diffractometric analyses of the whole rock and carbonate-free powder samples (decarbonation by hydrochloric acid of 3%) were carried out by Philips Müller Micro 111 type X-ray diffractometer. (Recording conditions of the oriented and desoriented samples: CuK_α radiation, acceleration voltage 42 kV, amperage 32 mA, proportional counter, graphite monochromator, divergency and detector slit 1°, goniometer speed 2°/min resp. 0.5°/min, paper speed 1600 mm/h, time constant 1).

The semi-quantitative computation of the mineral composition was carried out according to the direct method of NÁRAY-SZABÓ—PÉTER [21] (this work was done in collaboration with M. TÓTH). To determine the constants the data of BÁRDOSSY [6], BÁRDOSSY et al. [7] and RISCHÁK and VICZIÁN [24] were also used. The results were controlled by the data of carbonate-free rocks, occasionally by DTG-analyses and mineralogical recalculation of wet chemical silicate analyses.

The structural ordering of illite-sericite-muscovite caused by temperature increase (decrease of water content, built-in of K) was measured by the half-width of the (002) base reflexion (= absolute half-width = KÜBLER-index, after KÜBLER [16]) in mm, resp. by the relative half-width method of

$$\left(\mathrm{Hb_{rel}} = \frac{\mathrm{Hb_{illite(002)}\ mm}}{\mathrm{Hb_{quartz(100)}\ mm}}\right) \times 100; \quad \text{Weber-index,}$$

WEBER [31] which rules out the effect of methodic differences. The relations of the values of the two methods obtained to samples of known metamorphic facies are shown in Fig. 2.

According to LAPHAM and JARON [17] and to the methodic experiments, the grinding time (i.e. the grain size) of the samples affects the half-width only in the grain size fraction of less than two microns. Since the average grain size of the studied metamorphites proved to be above 10 microns, no grain size separation has been applied in addition to coarse grinding of 5 minutes and fine grinding providing the grain size of 30 to 60 microns.

The standard deviations of half-width values at a goniometer speed of 2°/min: KÜBLER-index, oriented samples: $s = 0.18$, desoriented samples: $s = 0.16$; WEBER-index: oriented samples: $s = 6.3$, desoriented samples: $s = 9.4$ ($n = 12$).

On the basis of the diagram of GUIDOTTI and SASSI [11] the Na/Na + K and the RM value (RM = 1/2 Fe_2O_3 + FeO + MgO mole prop.) were determined from the d_{002} and b_0 lattice parametres of sericite corrected by means of quartz internal standard. The standard deviation of the b_0 values in case of: $\bar{x} = 9.003$ Å and $n = 64$: $s = 0.007$. Regarding its order of magnitude this agrees with literature data. The deviation of the d_{002} determination is greater ($s = 0.019$).

The pressure value of metamorphism was determined by means of the statistic investigation of the b_0 parameters after SASSI [25] and SASSI and SCOLARI [26].

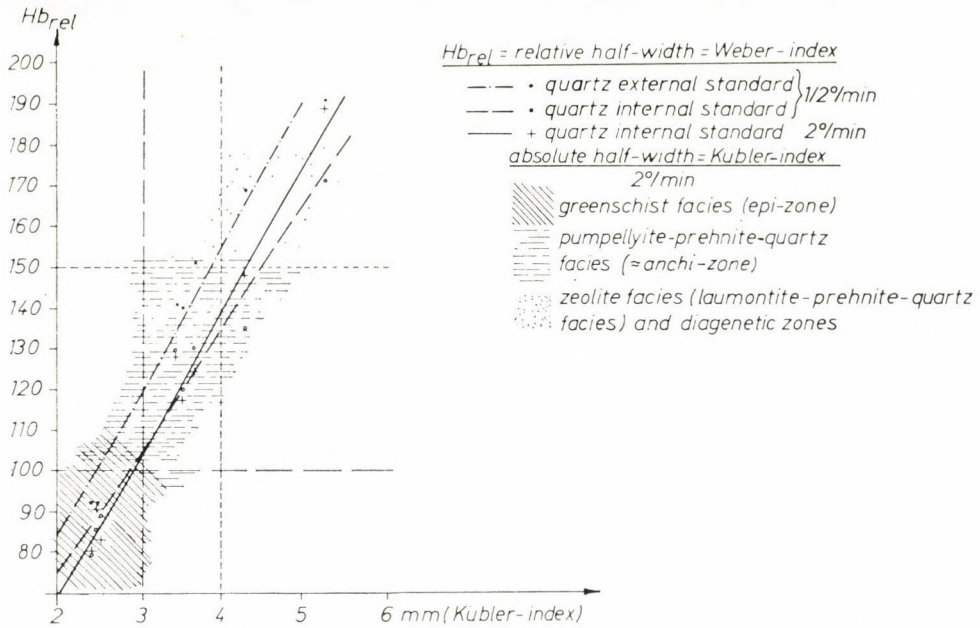


Fig. 2. Correlation between illite crystallinity indices and metamorphic mineral assemblages after KÜBLER (1968) and WEBER (1972). Correlation of illite crystallinity indices of formations of known facies at different X-ray diffractometric conditions

Lithology, changes in mineral composition

The Rakaca Crystalline Limestone Formation is characterized by the predominance of carbonate sediments (banded crystalline limestone), in the upper stratigraphic horizons by the mingling with submarine volcanic clastic material (cipolino) and in a fairly separable horizon by fine-clastic intercalation (sericite-chlorite phyllite).

The Transitional Member evolves gradually from this formation by continuous sedimentation and repetitions. The Transitional Member is characterized by the mixture and multiple alteration of, pelitic, psammitic and carbonate sediments reflecting the variegated-sedimentary conditions and which consist of the so-called authigenic-brecciated limestone, sandy crystalline limestone, limy meta-sandstone, sericite-chlorite phyllite, resp. of the banded schist built up by the rhythmic alteration of the stripes of mm to dm size of the two latter rock types.

The dark-grey sericite-chlorite phyllite, paragonitic phyllite of the Szendrő Phyllite Formation together with their graphite and lateral-secretion pyrite-chalcopyrite content indicate undisturbed, predominantly pelitic, subordinately psammitic sedimentation of reductive environment.

The Szendrőlád Limestone Formation is characterized again by the predominance of carbonate sediments. The undisturbed fine-clastic sedimentation was followed by the formation of unbalanced coarser-clastic-limy, subordinately fine-clastic sediments (Abod Member). In the upper horizons only intercalations of pelitic origin are found. The predominantly carbonate (limy) sedimentation of the Abod Member is diversified by the admixture of the material originating from the submarine decomposition of basic volcanic detritus (cipolino intercalations). These can also be found in the lower horizons of the Szendrőlád Formation of restricted sense. In the upper horizons the carbonate sedimentation became predominant (with corals and bryozoans) and this is interrupted several times by carbonate-pelitic intercalations.

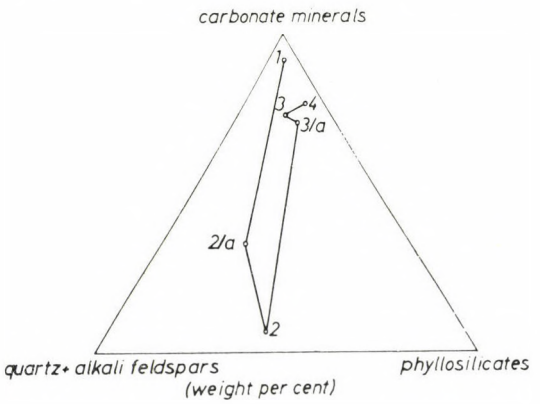
In the stratotype of the Borsod Limestone Formation the crystalline limestone sequence containing less material of clastic origin evolved by repeated alteration and gradually from the lime-schist derived predominantly from carbonate, subordinately from pelitic mixture in a reductive environment. This crystalline limestone sequence is characterized by the repeated alteration of the pure carbonate and tuffitic-carbonate (cipolino) strata.

In the diagram of mineral composition representing the three main type of sediments (carbonate, pelitic, psammitic, see Fig. 3/a) the averages of the Paleozoic formations and members of Szendrő Mountains show cyclic changes (carbonate — carbonate-psammitic-pelitic — pelitic — carbonate-pelitic — carbonate). When studying the elements of the cycle it is obvious that except the cyclic change of carbonate content in the stratigraphic profile there is a monotonous trend from north to the south: the quartz and feldspar contents relatively decrease, the quantity of phyllosilicates relatively increases. This can be explained by the decrease in grain size of the clastic part of the sediments and by the gradual maturity of the clastic material. The great sedimentation cycle is formed by the superposition of several (medium-sized) cycles (Fig. 3/b). In the cycles of medium size the systematic change of the quartz + feldspar/phyllosilicates ratio can also be observed. The cyclic character of sedimentation can be observed not only on the scale of the mountain or a formation but also within each strata groups and strata (e.g. banded schist).

In the distribution of the averages of mineral composition, in addition to the nearly symmetric changes of the calcite content, the dolomite content of the Szendrőlád and Borsod Limestone Formations is remarkable. This is a significant difference as compared to the other formations. The quantitative distributions of silicate minerals complement the distribution curves of carbonate

contents. Comparing the formation averages of each rock types the decrease of quartz and feldspar content is conspicuous, further the increase of the sericite content towards the upper stratigraphic horizons is remarkable (especially in the case of phyllites) (Fig. 4).

When studying the averages of mineral composition of the acid insoluble residue (= 100%) these trends are more conspicuous (Fig. 5). These trends



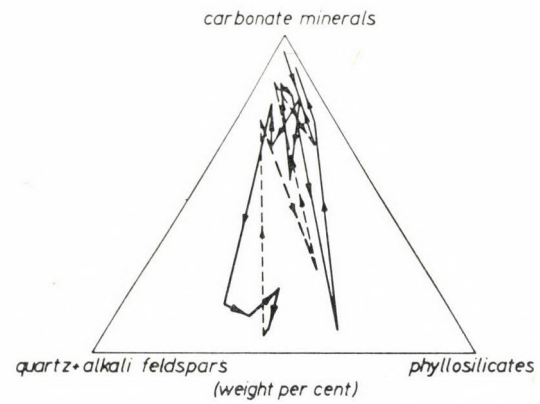


Fig. 3. Changes of the average mineral compositions of the Szendrő Paleozoic formations and members (a) and of the rock types (b) in the idealized stratigraphic profile
(Legend: 1—4: see in Fig. 1)

do not support the theory on the symmetric (anticlinal) structure of the mountain [22]. When assuming the identity of the northern and southern limestone members and the lateral facies changes, trends of opposite direction ought to be obtained since the axis of the Paleozoic geosyncline run north of the Szendrő Mountains.

Among the average mineral compositions of the main rock types (Table I) the following regularities can be established.

1. The quantity of carbonate minerals decreases in the following order: crystalline limestone — cipolino — sandy, authigenic-brecciated limestone — meta-sandstone — banded schist — sericite-chlorite phyllite.

2. Parallel with increasing quantity of carbonate minerals the chlorite/sericite and the feldspar/quartz ratios decrease.

3. In the case of rocks of detrital origin the quartz content is higher, the sericite-chlorite content is lower, the feldspar/quartz ratio is lower in the limy meta-sandstones than in the phyllites of clayey origin.

4. In carbonate rocks the quartz content is highest in the authigenic--brecciated limestone and lowest in the cipolino, the feldspar content decreases in the order of crystalline limestone — authigenic-brecciated limestone — cipolino. The quantities of sericite and chlorite are highest in the cipolino.

5. On the basis of low quartz and feldspar content, high sericite and chlorite content, of the locally high chlorite/sericite ratio and authigenic epidote content of the sericite-chlorite limestone (cipolino) as compared to the normal crystalline limestone, further of the hematite-siderite pseudomorphs after magnetite as well as of the oxidative character as against the general reductive type of the whole sequence assume the mingling of lime mud and basic volcanic clastics. The lack of igneous relict minerals indicates the halmyrolitic decomposition of the pyroclastics. The source of the basic volcanic material is geologically sufficiently proved by the basic volcanic sequence of great thickness of the Devonian(?) Rakovec Series in the Spissko-Gemerské Rudohorie (W-Carpathians).

Regional metamorphism

Constituents of detrital origin (mostly quartz, quartzite, subordinately albite-oligoclase, potash-feldspar, muscovite, muscovite-chlorite aggregates after biotite) can be recognized first of all in the rocks of coarser-detrital origin (sandy-aleuritic source). The metamorphic recrystallization eliminated the boundaries between the clastics and cementing material (phyllite- and quartzite-like structure, crystalline limestone, lime-schist were formed). The grain size of the crystalline limestones depends also on the degree of dispersion in addition to the quantity of the clastic and organic matter and to the pressure and temperature determining the recrystallization.

The metamorphic transversal schistosity is a common feature of the Paleozoic formations in the Szendrő Mountains. The diagenic-metamorphic recrystallization increasing parallel with increasing burial (p and t increase) generated a metamorphic orientation (I. schistosity plane) parallel with the sedimentary stratification. In the subsequent syntectonic recrystallization phase microfolds and from them shearing schistosity (II. schistosity) planes were

Table 1

Average values of the mineral composition and of the sericite parametres in the

Lithostratigraphic units, rock types	n	Semi-quantitative mineral composition (weight %)										
		quartz	albite-oligoclase	potash-feldspar	sericite	paragonite	chlorite	kaolinite	montmorillonite	mixed-layer clay minerals	calcite	dolomite
4. Borsod Limestone Formation												
average	46	5	1	tr	14		2	tr	tr	tr	76	tr
banded cryst. limestone	14	2	tr	tr	7		tr	tr	tr	(tr)	90	—
cipolino	8	6	tr	tr	19		2	1	(tr)	tr	71	—
phyllite	3	28	—	3	48		8	1	(tr)	—	2	4
lime-schist	21	4	2	1	11		2	(tr)	—	—	79	—
3. Szendrőlád Limestone Formation												
average	46	11	2	tr	10		2	1	tr	(tr)	68	5
coral-bearing cryst. limestone	25	8	1	tr	6		tr	1	tr	—	80	4
phyllite (banded schist)	16	18	3	1	18		4	2	—	—	46	7
cipolino	5	6	tr	tr	10		4	—	—	tr	80	—
3/a. Abod Limestone Member												
average	59	8	1	tr	12		5	tr	tr	tr	72	—
cipolino	16	9	1	tr	18		7	tr	tr	tr	64	—
cryst. limestone	33	4	tr	(tr)	7		2	(tr)	tr	tr	86	—
phyllite	7	23	4	1	25		20	tr	1	—	26	—
authigenic-brecciated limestone	3	19	1	(tr)	4		1	tr	(tr)	—	74	—
2. Szendrő Phyllite Formation												
average	24	33	16	2	21	1	15	1	1	3	6	—
paragonitic phyllite	12	40	10	2	23	1	10	3	1	3	5	—
sericite-chlorite phyllite	12	27	22	2	19		20	tr	—	2	8	—
2/a. Transitional Member												
average	111	27	15	1	12		10	tr	tr	tr	35	—
banded schist, phyllite	19	24	16	1	20		19	tr	tr	tr	20	—
banded schist, meta-sandstone	29	43	16	tr	7		7	tr	tr	(tr)	26	—
banded schist, average	30	29	23	1	17		17	(tr)	—	—	12	—
crystalline limestone	20	12	3	tr	5		2	(tr)	tr	—	78	—
authigenic-brecciated limestone	13	18	4	tr	8		4	tr	tr	1	65	—
1. Rakaca Crystalline Limestone Formation												
average	22	3	tr	tr	4		1	tr	tr	—	91	—
cipolino	5	5	tr	tr	11		3	tr	—	—	80	—
banded cryst. limestone	17	3	tr	tr	2		tr	tr	tr	—	94	—

tr = trace, oritented, desor. = desoriented sample preparation

generalized stratigraphic profile of the Paleozoic of the Szendrő Mountains

siderite	epidote	pyrite	hematite	goethite	Sericite parametres: crystallinity indices: Kübler (mm): oriented: n	Kübler (mm): oriented: $\bar{x}$	Kübler (mm): desor.: n	Kübler (mm): desor.: $\bar{x}$	Weber: oriented: n	Weber: oriented: $\bar{x}$	b_0(Å): n	b_0(Å): $\bar{x}$	d_{002}(Å): n	d_{002}(Å): $\bar{x}$
—	—	1	tr	tr	14	3.08	50	3.33	14	116	41	9.113	48	9.957
—	—	tr	—	(tr)	2	2.72	14	3.26	2	103	14	9.016	13	9.961
—	—	—	tr	tr	1	2.50	8	3.46	1	94	8	9.020	8	9.966
—	—	5	—	—	2	2.96	4	3.47	2	112	3	9.004	4	9.947
—	—	1	—	—	9	3.25	24	3.29	9	123	16	9.010	23	9.954
tr	—	1	(tr)	tr	25	2.51	51	2.81	25	95	39	9.008	47	9.948
tr	—	tr	—	tr	10	2.33	24	2.86	10	88	22	9.006	24	9.943
tr	—	1	—	—	11	2.66	22	2.77	11	100	12	9.008	18	9.952
—	—	—	tr	tr	4	2.51	5	2.77	4	95	5	9.016	5	9.962
—	(tr)	(tr)	tr	tr	22	3.02	64	3.27	22	114	55	9.010	64	9.961
—	tr	tr	tr	tr	9	2.94	19	3.38	9	111	17	9.015	19	9.958
(tr)	(tr)	(tr)	(tr)	(tr)	10	3.06	33	3.18	10	116	31	9.008	34	9.961
—	tr	(tr)	—	—	2	3.20	9	3.29	2	121	4	9.005	8	9.962
—	—	—	(tr)	—	1	3.00	3	3.57	1	113	3	9.012	3	9.966
—	—	(tr)	tr	tr	24	2.52	22	3.07	24	95	21	8.997	24	9.960
—	—	tr	tr	1	12	2.65	10	3.31	12	100	9	8.990	13	9.951
—	—	—	tr	—	12	2.39	12	2.86	12	90	12	9.002	11	9.971
tr	—	tr	tr	tr	61	2.73	101	3.03	61	103	75	9.006	11	9.954
—	—	tr	tr	1	12	2.74	19	3.11	12	104	16	9.004	20	9.971
tr	—	tr	—	tr	13	2.68	22	3.06	13	101	10	9.003	28	9.944
—	—	tr	tr	—	19	2.96	32	3.01	19	112	24	9.005	32	9.962
—	—	tr	(tr)	tr	11	2.47	15	2.83	11	93	15	9.011	19	9.944
—	—	tr	—	tr	6	2.56	13	3.11	6	97	10	9.005	12	9.947
—	—	tr	tr	tr	11	2.45	26	2.78	11	92	24	9.013	25	9.948
—	—	tr	tr	—	5	2.39	9	2.58	5	90	6	9.011	7	9.961
—	—	tr	—	tr	6	2.51	17	2.89	6	95	18	9.014	18	9.944

formed just due to the changes of oriented pressure. Along these planes oriented crystal growth and recrystallization proceeded. Since mineral paragenetic differences cannot be detected, these two recrystallization phases do not mean necessarily metamorphic phases of two different ages. This phenomenon more likely reflects the different stages of development of a tectonometamorphic

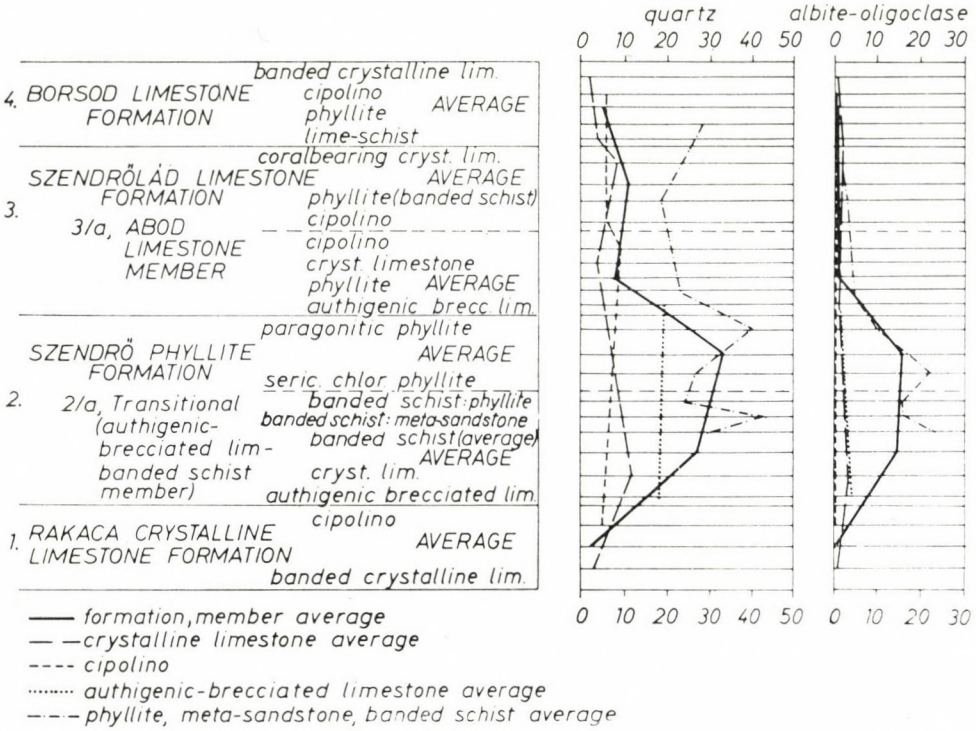


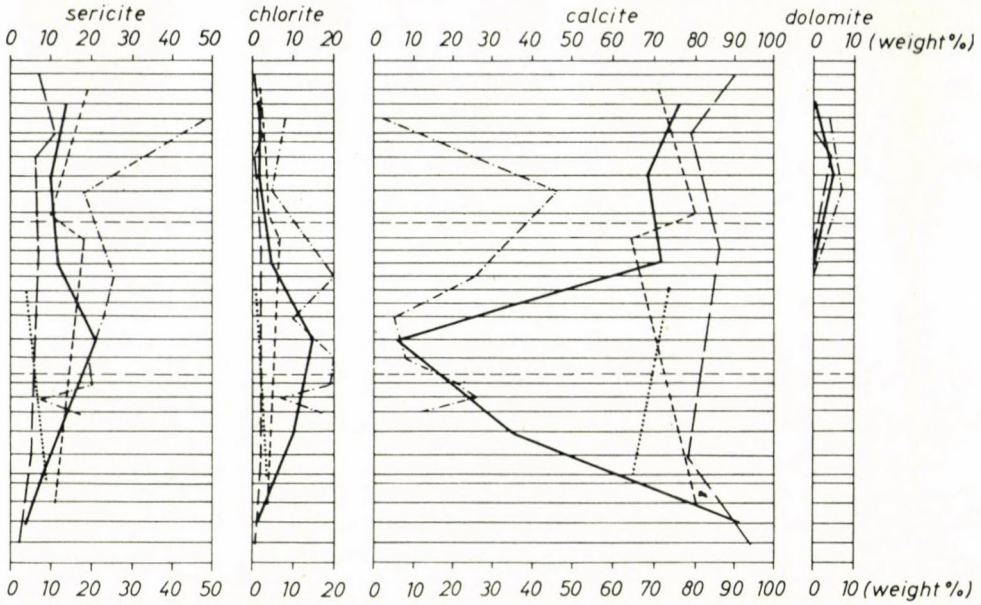
Fig. 4. Changes of average mineral compositions in the generalized stratigraphic profile (Whole-rock analyses)

cycle or phase. The transversal schistosity played an important role in the formation of several peculiar rock types (e.g. the authigenic-brecciated limestone formed from limy-sandy-clayey banded sediments by sediment sliding and metamorphic transversal schistosity, or the cipolino recrystallized from the limy and tuffitic banded sediments).

The *metamorphic mineral assemblage* of the Paleozoic of the Szendrő Mountains (quartz, low temperature (ordered) albite, (potash-feldspar), sericite, chlorite, calcite, ± paragonite, epidote, dolomite, siderite, pyrite, magnetite, hematite, graphite) is hardly suitable to the exact determination of the degree of metamorphism. Paragonite and epidote are stable also within the very-low-grade range (WINKLER [32]), the appearance of graphite, however, indicates

the boundary of the very-low-grade and low-grade metamorphism (anchi-epizone; FREY and NIGGLI [8]).

The formation of a weathering zone and along-fault weathering systems of clayey- and oxidative nature is common in the whole mountain. The thickness of this zone of weathering is varied depending on the rock-



mechanical, structural, geographic-paleogeographic and hydrological as well as climatic conditions. Its continuous thickness is max. 40 m but along faults goethite—clay mineral type decomposition was observed also in a depth of 375 m (in greater depths in addition to the descendent waters the effect of ascendent (telethermal?) waters should also be taken into account). The characteristic mineral assemblage of the weathering zones (in phyllites and crystalline limestones) is as follows: mixed-layer clay mineral showing base reflexions at 8—12—(14) Å, kaolinite $\pm$ montmorillonite, goethite. They may have been formed first of all from chlorite and iron-sulfide minerals, and to a smaller extent from sericite. In the phyllites the mixed-layer clay mineral, in the crystalline limestones the kaolinite is predominant.

Below the oxidation zone the kaolinitic decomposition (in presence of pyrite) can be observed down to varied depth.

The mineral assemblage consisting of montmorillonite, quartz, pyrite, siderite and subordinately of the illite/montmorillonite mixed-layer structure

reflects probably the transforming effect of ascendent (telethermal?) solutions.

When evaluating the regional metamorphism the minerals of secondary origin should be neglected. The fresh rocks contain only two phyllosilicates: chlorite and 2M-type sericite.

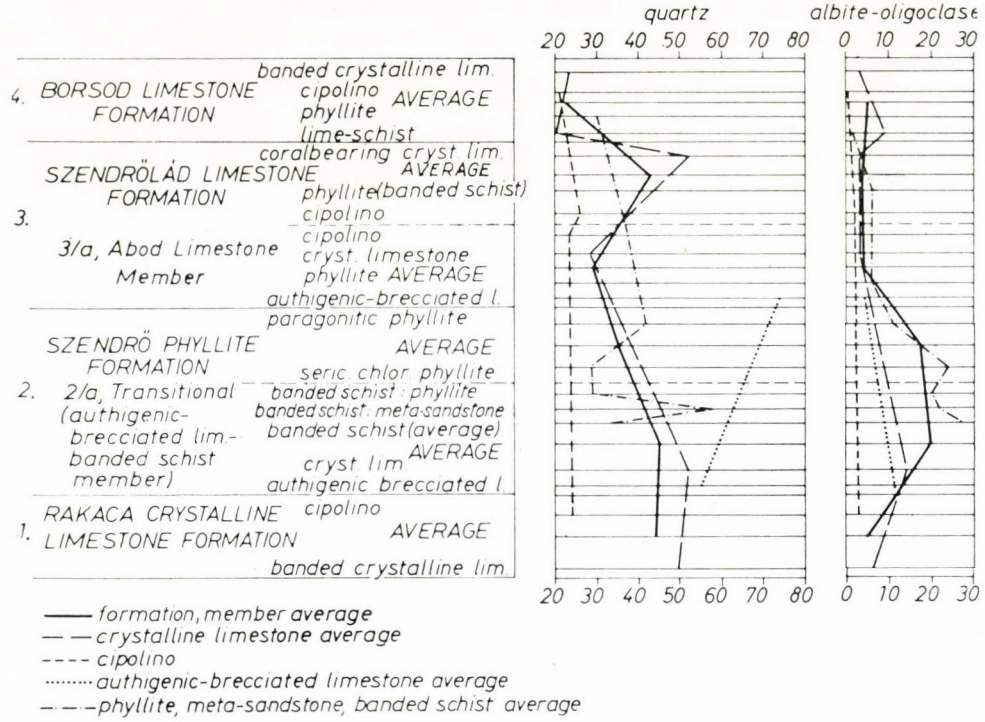
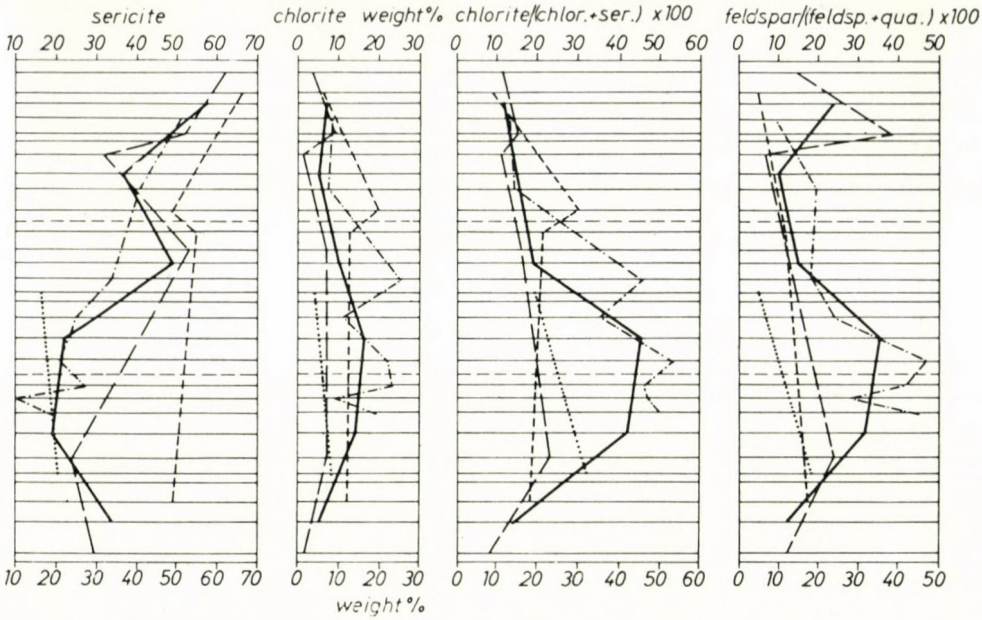


Fig. 5. Changes of mineral compositions of the acid insoluble residue in the generalized stratigraphic profile

On the basis of the frequency distribution of the illite-sericite crystallinity indices (Fig. 6/a—d) the metamorphic grade of the Paleozoic of the Szendrő Mountains corresponds to the initial, low-temperature phase of the low-grade metamorphism (greenschist facies, quartz-albite-muscovite-chlorite subfacies). Part of the parametres indicates the transitional conditions of the very-low-grade and low-grade metamorphism (the boundary of the anchi- and epi-zones).

In the idealized stratigraphic profile the degree of metamorphism slightly decrease upwards (from north to the south) regarding its main tendency (Fig. 7/a—b). The considerable fluctuations, however, point to the fact that in addition to the effect of lithostatic pressure also the intensity of dynamo-metamorphism changed within a relatively narrow strip band (10 to 20 km).

There is no systematic significant difference between the average illite crystallinite indices of different rock types containing clay minerals of originally presumably different degree of degradation. This can be explained by the homogenization effect of metamorphic recrystallization. In contrast to the



initial stages of diagenesis, under the greenschist facies conditions no delaying effect of the carbonate and organic matter content on the clay mineral recrystallization can be observed. Moreover, in certain formations, by means of mobilization producing the complete recrystallization of limestones the sericites of the carbonate rock types are of more ordered state as compared to those of the clastic rocks.

In the Szendrő Paleozoic the frequency distribution of the b_0 parametres of sericite (Fig. 8) is of one maximum with a slight positive asymmetry towards the greater values which is due to the change of carbonate content (see Figs 10/b—c). Most of the sericites lie in the muscovite field, their smaller part is of phengite composition (Fig. 9).

The applicability of the $b_0 - d_{002}$ diagram [11] elaborated to the determination of the Na/Na + K and RM values of muscovites (sericites) was proved by electron microprobe measurements in a phyllite sample of borehole Szendrő-22 (measurements on a JXA-type microprobe by G. Nagy). According to Fig. 10 the ranges of standard deviation of the two analytical methods

overlap each other: a relatively good agreement was obtained between the b_0 and RM values. The greater difference between the Na/Na + K and d_{002} values is probably due to the greater deviations of the X-ray diffractometric measurements.

The formation averages of the b_0 parametre of sericite do not show significant differences (Figs 10/c and 11). There are great differences, however, among the different rock types of the individual formations. As against the statements of GEYSSANT et al. [10] parallel with increasing carbonate quantity the value of the b_0 parametre of sericites also increases (determinations were made from acid insoluble residue, thus the disturbing effect of the overlapping

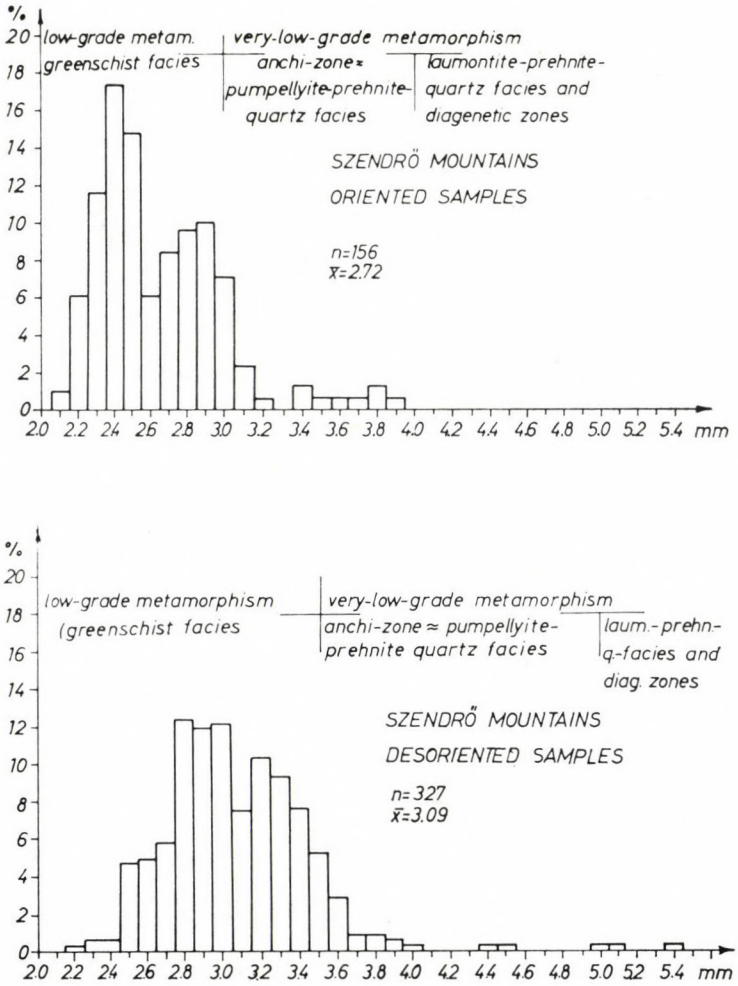
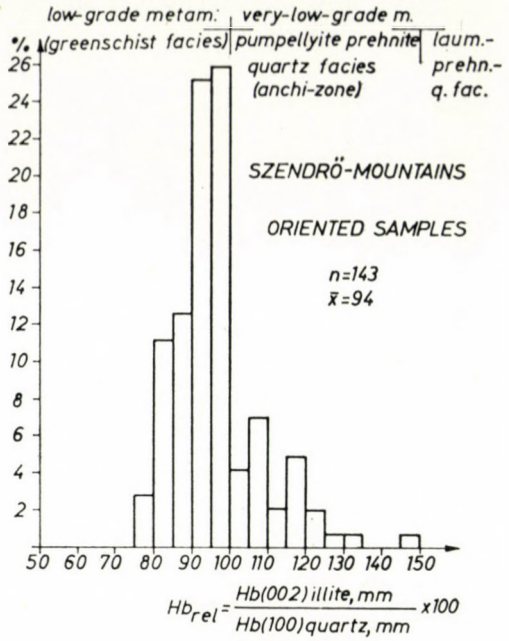
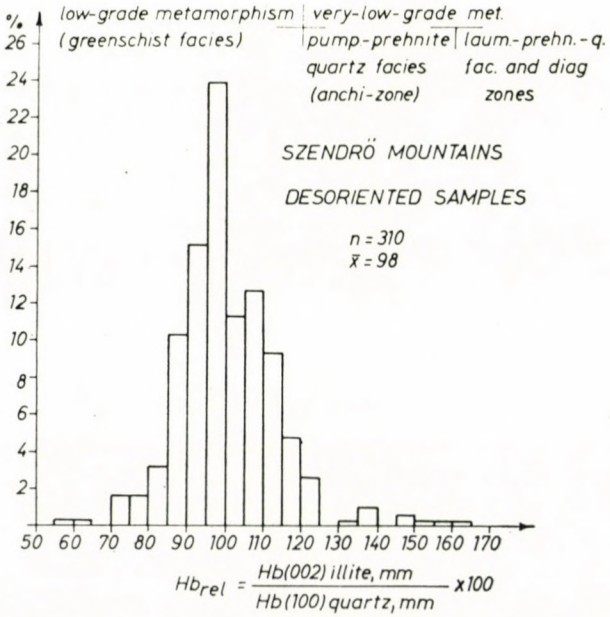


Fig. 6. Frequency distributions of illite crystallinity indices: (a—b): KÜBLER-indices; (c—d): WEBER-indices with internal quartz strandard


low-grade metam. (greenschist facies)
very-low-grade m.
pumpellyite prehnite quartz facies (anchi-zone)
laum.-prehn.-q. fac.
%
SZENDRÖ-MOUNTAINS
ORIENTED SAMPLES
n=143
x̄=94
50 60 70 80 90 100 110 120 130 140 150
Hb_rel = Hb(002) illite, mm / Hb(100) quartz, mm x100


low-grade metamorphism (greenschist facies)
very-low-grade met.
pump-prehnite quartz facies (anchi-zone)
laum.-prehn.-q. fac. and diag zones
%
SZENDRÖ MOUNTAINS
DESORIENTED SAMPLES
n = 310
x̄ = 98
50 60 70 80 90 100 110 120 130 140 150 160 170
Hb_rel = Hb(002) illite, mm / Hb(100) quartz, mm x100

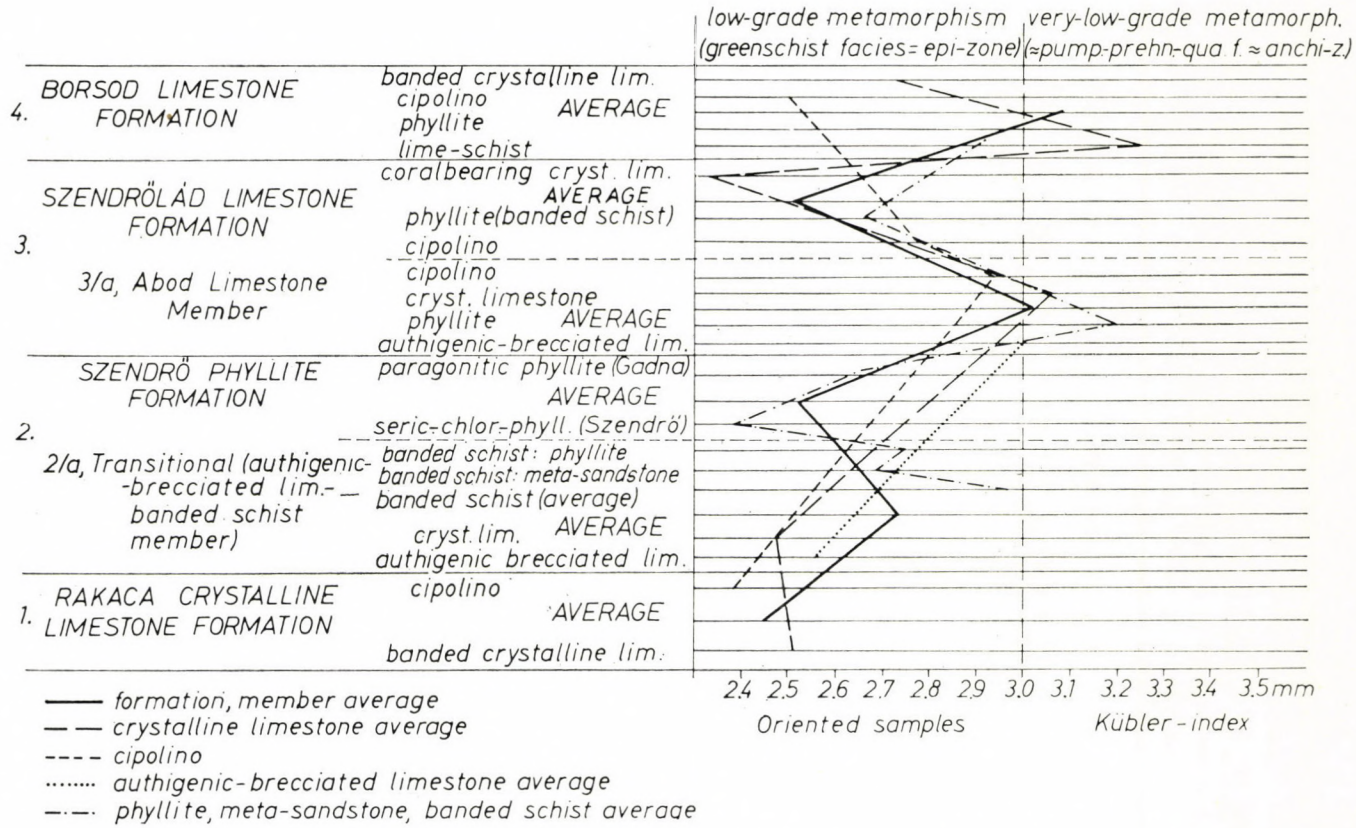
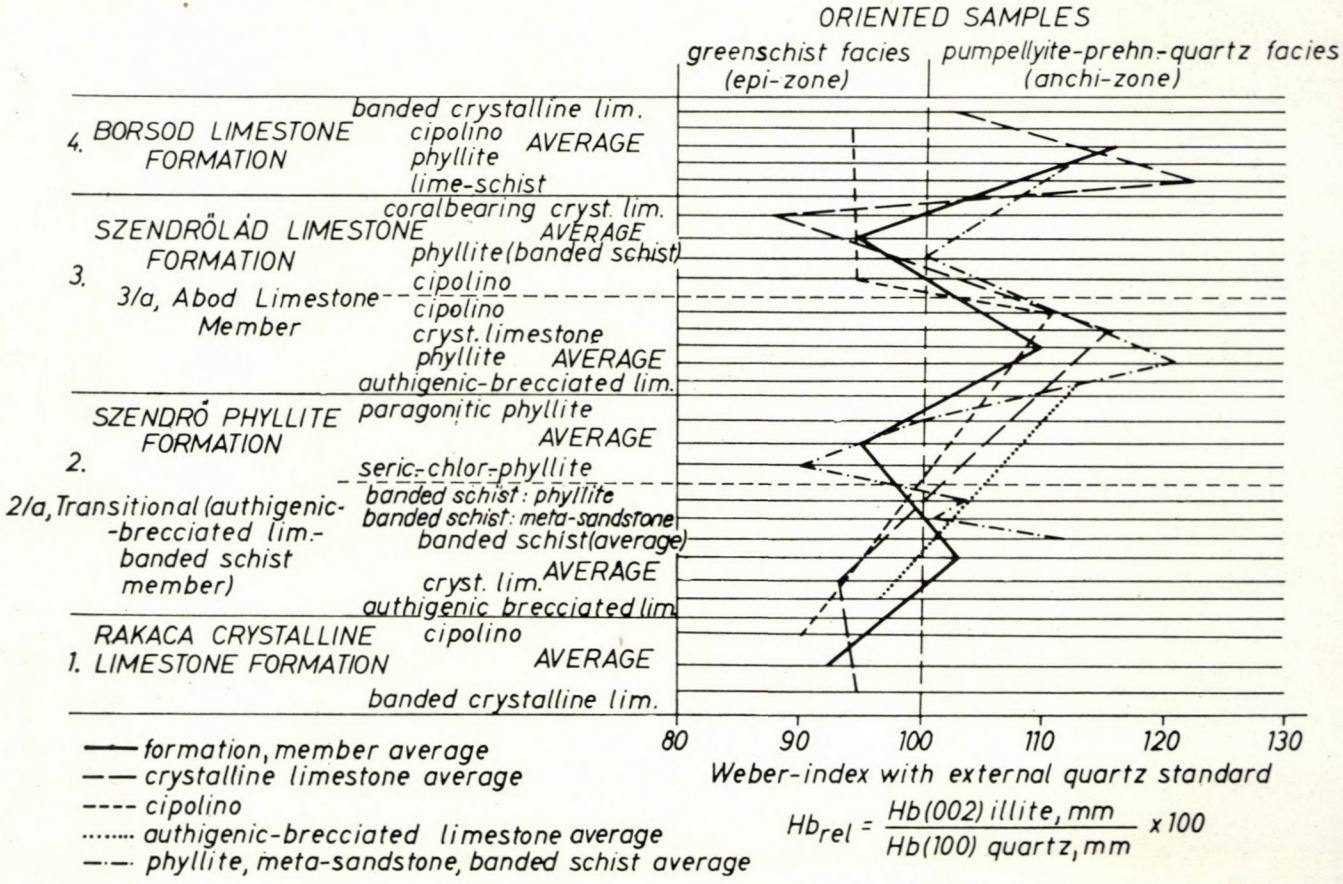


Fig. 7. **Changes of average values of illite crystallinity indices in the generalized stratigraphic profile of the Szendrő Mountains. a) KÜBLER-indices; b) WEBER-indices with external quartz standard**


ORIENTED SAMPLES
greenschist facies (epi-zone)
pumpellyite-prehn-quartz facies (anchi-zone)
4. BORSOD LIMESTONE FORMATION
banded crystalline lim.
cipolino
phyllite
lime-schist
AVERAGE
coralbearing cryst. lim.
SZENDRŐLÁD LIMESTONE FORMATION
AVERAGE
phyllite (banded schist)
3.
3/a, Abod Limestone Member
cipolino
cipolino
cryst. limestone
phyllite
AVERAGE
authigenic-brecciated lim.
SZENDRŐ PHYLLITE FORMATION
paragonitic phyllite
AVERAGE
2.
seric.-chlor.-phyllite
2/a, Transitional (authigenic--brecciated lim.-banded schist member)
banded schist : phyllite
banded schist : meta-sandstone
banded schist (average)
cryst. lim.
AVERAGE
authigenic brecciated lim.
RAKACA CRYSTALLINE LIMESTONE FORMATION
1.
cipolino
AVERAGE
banded crystalline lim.
80
90
100
110
120
130
Weber-index with external quartz standard
formation, member average
crystalline limestone average
cipolino
authigenic-brecciated limestone average
phyllite, meta-sandstone, banded schist average
$Hb_{rel} = \frac{Hb(002)\ illite, mm}{Hb(100)\ quartz, mm} \times 100$

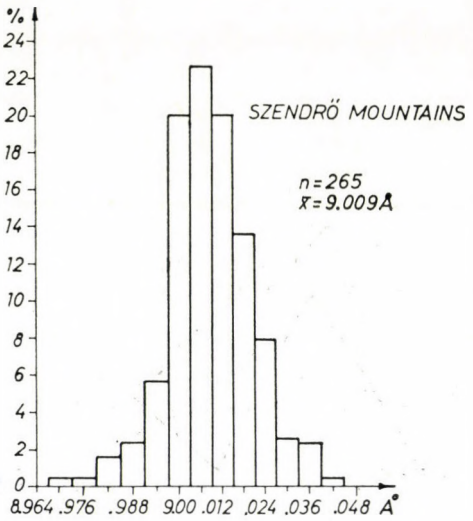


Fig. 8. Frequency distribution of the b_0 parametres of sericites in the Paleozoic of the Szendrő Mountains

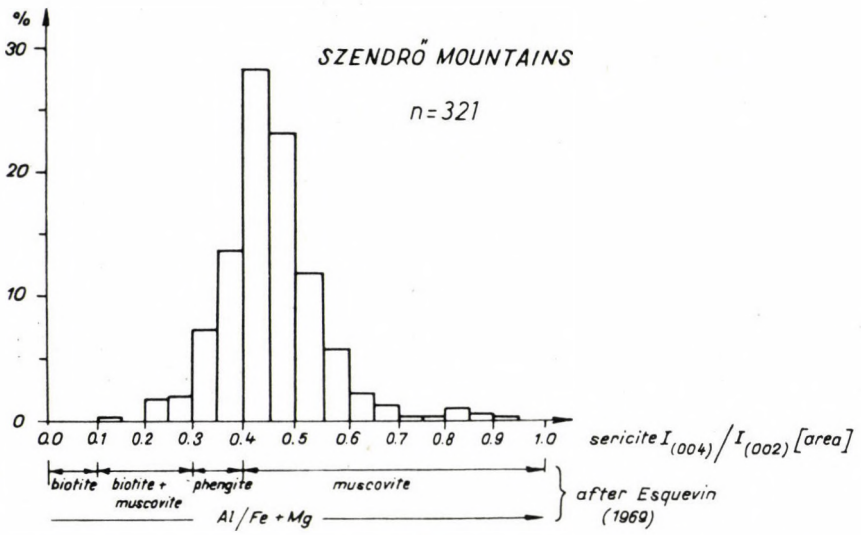


Fig. 9. Frequency distribution of intensity ratios of I_{004}/I_{002} and composition of sericites after ESQUEVIN (1969)

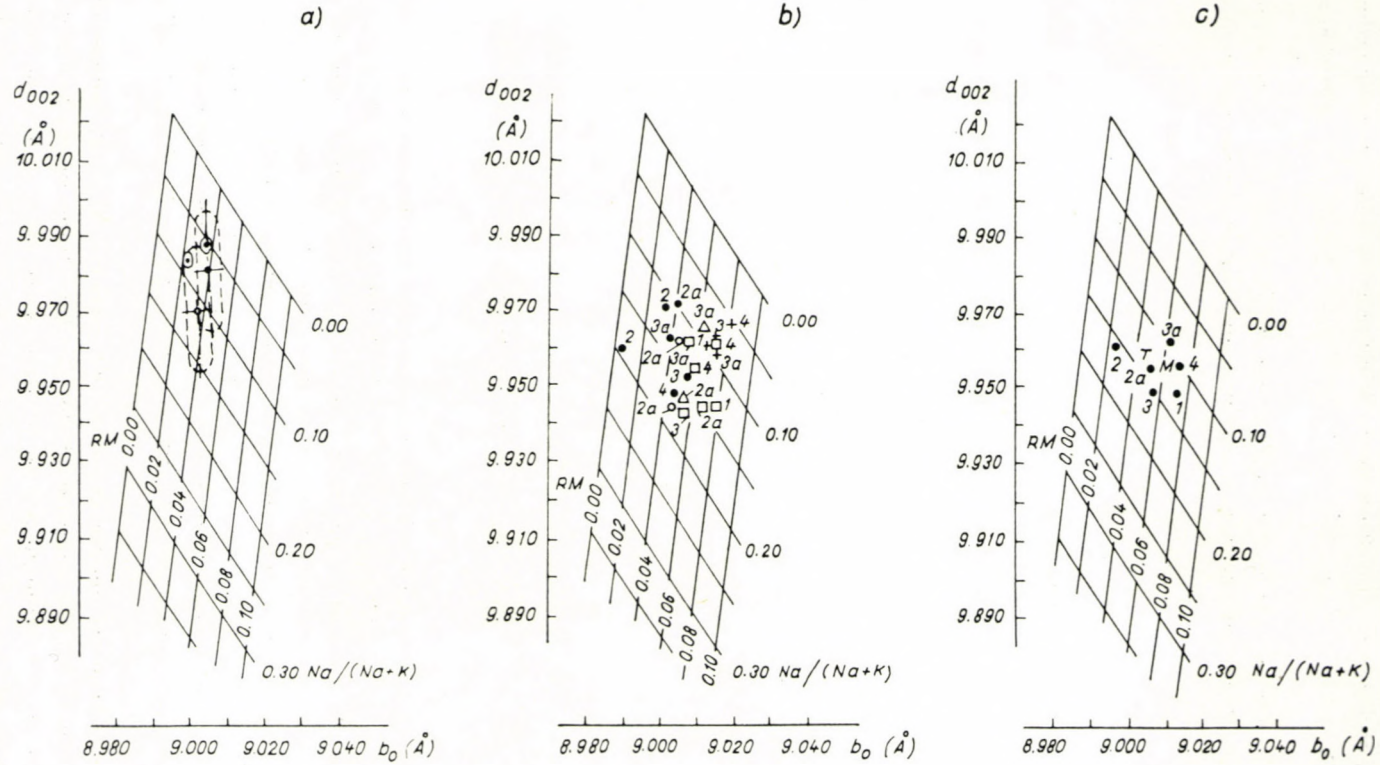


Fig. 10. Distribution of the investigated sericites in the b_0 vs d_{002} diagram of muscovites (after GUIDOTTI and SASSI, 1976). a) electron microprobe analyses (.), X-ray diffractometric analyses (○) and the average of analyses (⊙), with the areas of dispersion. b) average values of the main rock types of the Szendrő Paleozoic. — 1—4: formations, members; □ — crystalline limestone; + — cipolino; △ — authigenic-brecciated limestone; . — phyllite; ◐ — banded schist; O — limy meta-sandstone; c) average values of the formations and of the carbonate (*M*) and clastic (*T*) rock types

reflexions of carbonates was eliminated). With increasing quantity of carbonate minerals the value of d_{002} slightly decreases (the Na/Na + K ratio slightly increases).

Thus, in the geobarometric application of the sericite b_0 values suggested by Sassi [25], Sassi and Scolari [26] the samples of sericite-chlorite phyllite

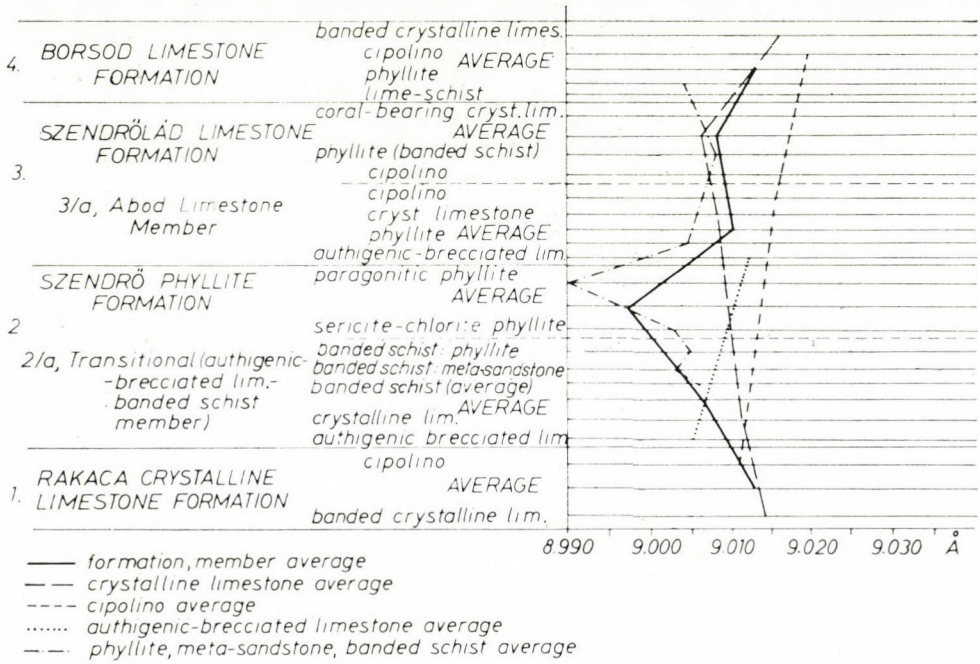


Fig. 11. Changes of average b_0 parametres in the idealized stratigraphic profile

were used which possessed the following mineralogical character: low, less than 20 per cent calcite + dolomite content, chlorite/chlorite + sericite ratio equals max. 1, and the mineral association: quartz, feldspar (albite, subordinately potash-feldspar), pyrite, graphite, rarely hematite. Thus, the conclusions are only informative and semi-quantitative since the carbonate-free rock type of low quartz and chlorite content suggested to pressure determination occur only rarely in the Szendrő Mountains. No interrelation was found between the change of the quartz, chlorite and hematite content and the b_0 cell parameter.

In the thermo-barometer diagram (Fig. 12) the d_{002}—b_0 points of the phyllites of low carbonate content of the Szendrő Mountains fall to the high-temperature area of the high- and medium-pressure series and to the low-temperature range of the low-pressure series. On the basis of the mineral paragenesis the high-temperature metamorphism can be excluded, thus the meta-

morphism of the Szendrő Mountains was of low-temperature and essentially of low-pressure (temperature < 450 °C, i.e. below the biotite isograd, pressure 2.5 to 3 kbar).

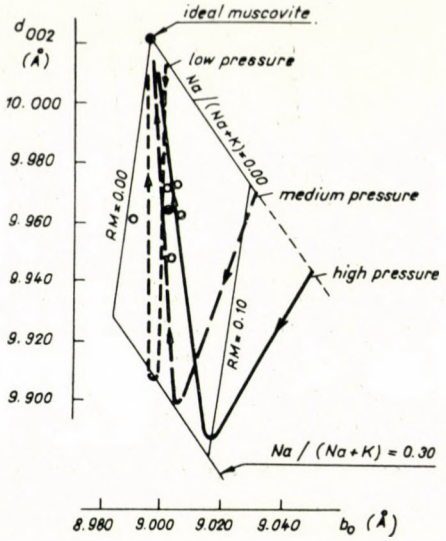


Fig. 12. Changes of the d_{002} and b_0 values (and composition) of muscovite as a function of temperature increase (arrows) in different pressure ranges (after GUIDOTTI and SASSI, 1976). Formation (○) and mountain (⊙) averages of the Szendrő phyllites

The b_0 cumulative frequency distribution of the sericites of Szendrő phyllites (Fig. 13) indicates pressure conditions corresponding to the low-pressure (Abukuma) and medium-pressure (Barrow) zones resp. to their transitions, the average value ($\bar{x} = 9.003$ Å, $s = 0.007$, $n = 64$) is higher than the average values of the typical low-pressure metamorphites free of chlorite zone [26]. The slope of the cumulative curve of Szendrő agrees with that of the curves of other typical locations. This relates partly to the sufficient accuracy of the measurements, partly to the homogeneous distribution of the b_0 values. The greater b_0 interval of the medium-pressure range is remarkable which needs further confined pressure classification (KRÄUTNER et al. [14]).

According to ZWART [33] the tectonometamorphic systems of orogenic belts of different ages were of different pressure range in Europe. The Dalslandian (Early Baikalian), and Assyntic (Late Baikalian), the Hercynian and the Alpine cycles are characterized successively by low-, medium-, low- and high pressure metamorphism.

In addition to the confirmation of these main regularities the up-to-date studies provide ever more pressure data [14, 15, 25, 26] which may be the elements of the synthesis approaching the real pressure-temperature conditions

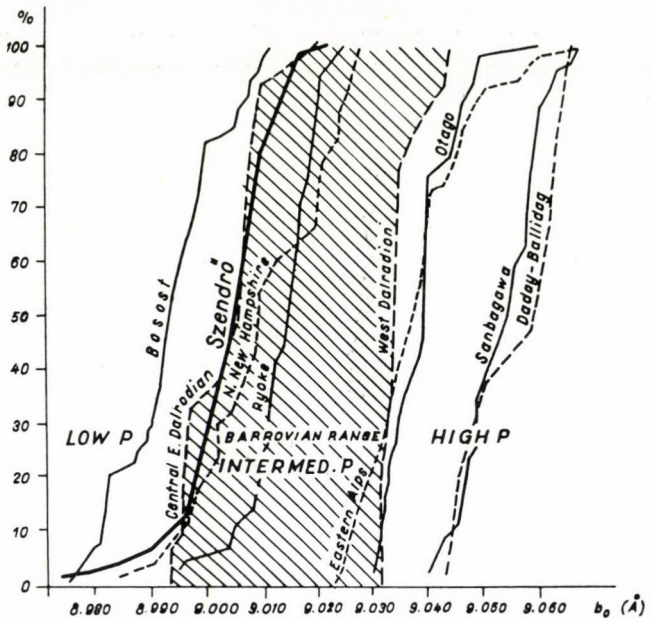


Fig. 13. Cumulative frequency curves of the muscovite b_0 parametres of low-temperature green-schist facies metamorphic sequences of different pressure range (after SASSI and SCOLARI, 1974)

of each tectonometamorphic period and zone (medium-pressure Dalslandian metamorphism in the Eastern Carpathians, medium-pressure Hercynian metamorphism in the Southern Carpathians, low-medium pressure Alpine metamorphism in the Bükk Mountains, etc.; Fig. 14).

In possession of these data it is less astonishing that the Hercynian Metamorphism in the (Silurian)-Devonian sequence of the Szendrő Mountains is not tipically low-pressure type but belongs to the boundary zone of the low- and medium-pressure range. On the possibility of superposition of a presumed later (Alpine) metamorphic effect conclusions may be drawn only after having investigated the complete Paleozoic-Mesozoic rock sequence of North Hungary.

Conclusions

The low-grade metamorphic zone ranging from the Graz Basin through the northern part of Hungary up to the Spissko-Gemerské Rudohorie (W-Carpathians) involves the metamorphic sequence of (Silurian?)-Devonian age and of carbonate, clastic origin of the Szendrő Mountains which forms a monoclinal structure of S, SE dip. The lithological conditions of the typical profiles of the Paleozoic formations and members suggest continuous sedimen-

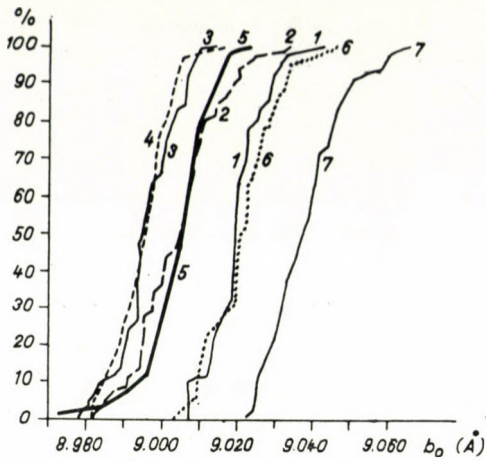


Fig. 14. Frequency distributions of the muscovite b_0 lattice parametre in the low-grade metamorphic sequences of different ages of the Alps and Carpathians (after KRÄUTNER, SASSI et al., 1975, 1976, SASSI and SCOLARI, 1974). 1. Eastern Carpathians, Dalslandian (= Early Baikalian) Rebra-Barnar series; 2. Eastern Carpathians, Assyntic (= Late Baikalian) Tulghes--series; 3. Eastern Carpathians, Hercynian, Repedea, Tibau, Rusaia series; 4. Eastern Alps, Hercynian phyillites; 5. Szendrő Mountains, Hercynian phyllites; 6. Southern Carpathians, Poiana Rusca Massif, Hercynian phyllites; 7. Eastern Alps, Alpine phyllites

tation between the formations which are characterized by repeated sediment alternations.

In the generalized stratigraphic profile the average mineral compositions of the formations and main rock types show cyclic (calcite) and within this monotonous (quartz, albite, sericite, dolomite) changes. The great sedimentation cycle is built up by the superposition of several different sedimentation cycle systems of different orders of magnitude. The monotonous changes of mineral compositions determined within the sedimentation cycle do not confirm the hypotheses concerning the overfolded anticlinal structure of the mountains but rather suggest the model of a simple monoclinal structure of normal position.

The metamorphic mineral assemblage of the formations (quartz, low--temperature albite, (potash-feldspar), sericite, chlorite, calcite $\pm$ paragonite, epidote, dolomite, siderite, pyrite, magnetite, hematite, graphite) is unsuitable to determine exactly the physical conditions of metamorphism. From the point of view of determining the grade of metamorphism it is significant to distinguish the clay mineral assemblages of secondary origin from the metamorphic mineral paragenesis.

On the basis of the statistic evaluation of the crystallinity indices of sericites determined by means of different methods the degree of metamorphism of the sequence corresponds to the initial stage of low-grade metamorphism (greenschist facies, quartz-albite-muscovite-chlorite subfacies). Its certain

upper (i.e. southern) horizons fall to the boundary of the very-low-grade and low-grade metamorphism (anchi-epi-zones). In the stratigraphic profile the grade of metamorphism slightly decreases upwards (from north to the south) but the fluctuations (as well as the rock-structural investigations) relate to the zonal changes of dynamometamorphic effects of folding tectonics, in addition to the effect of burial. There is no systematic difference in the average values of illite crystallinity indices of different rock types (homogenization effect of recrystallization).

The $b_0 - d_{002}$ diagram is fairly applicable to determine the chemical composition of sericites [11]. Among the b_0 averages of different formations there is no significant difference. There is a positive correlation between the b_0 values and the carbonate contents of the rocks while a slightly negative correlation is found between the d_{002} values and the carbonate content.

The b_0 average value and frequency distribution of the sericites of carbonate-poor phyllites chosen for geobarometric purposes refer to the conditions of the boundary between the low and medium-pressure ranges (about 2.5 to 3 kbar), the temperature of the metamorphism varied between 300 and 450 °C (above the boundary of the anchi-epi-zone and below the biotite isograde).

The pressure value of the metamorphites of the Szendrő Mountains possesses an intermediate site between the typical low-pressure Hercynian metamorphites of the Eastern Alps and Eastern Carpathians and the medium-pressure Hercynian metamorphites of the Southern Carpathians (Poiana Rusca) believed to be anomalous till now.

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МЕТАМОРФИЗМ НИЗКОГО СТЕПЕНИ ОСАДОЧНЫХ ОБРАЗОВАНИЙ ПАЛЕОЗОЙСКОГО ВОЗРАСТА В ГОРАХ СЕНДРЭ (СВ-ВЕНГРИЯ)

П. АРКАИ

Будапешт

Свиты силурского (?) — девонского возраста моноклинальной складчатой структуры гор Сендрэ состоят из метаморфических горных пород, представляющих собой неоднократно повторяющиеся смеси, по различным отношениям, первоначально карбонатных (преимущественно известковых) и пелито-псаммитовых отложений. Средние минеральные составы свит и типов горных пород, установленные путём минералогическо-петрографического исследования кернов опорных буровых скважин, характеризуются цикличным и также направленным изменением по обзорному геологическому разрезу. (Цикличное изменение содержания кальцита и монотонные изменения содержаний кварца, полевых шпатов, серицита и доломита подтверждают геологический модель полагающий нормальное положение моноклинальной структуры и непрерывное осадконакопление.)

В вертикальных профилях хорошо различаются метаморфические парагенезисы минералов (кварц, альбит, калиевый шпат, серицит, хлорит, кальцит ± доломит, эпидот, парагонит, пирит, гематит, графит) от вторичных парагенезисов глинистых минералов образованных в условиях выветривания и гидротермальных действий низкой температуры на поверхности и по сбросам (смешанно-слойные глинистые минералы, каолинит, монтмориллонит) и также от гетита окислительного происхождения.

Изучением структуры и текстуры горных пород удалось насследовать влияния различных фаз перекристаллизации связанных с погребанием и со складко-образующим динамометаморфизмом.

Региональный метаморфизм по всей вероятности герцинского возраста потерпевший исследованными образованиями по индексам кристаллизации иллита-серицита (по индексам Кюблера и Вебера) соответствует низкотемпературной стадии метаморфизма (низкого степени (кварц-альбит-мусковит-хлоритовой подфации) зеленосланцевой фации, т. е. к пограничной зоне анхи- и эпиметаморфизма). $b_0 - d_{002}$ диаграмма серицитов составленная по методу Гидотти и Сасси [11] оказалась очень полезной для приблизительного установления химического состава серицитов. С ростом содержание карбонатов в породе имеются более высокие значения b_0, (Fe + Mg) серицитов, а немножко понижаются величины d_{002} (с небольшим повышением отношения Na/Na + K)).

Распределение параметров b_0 малокарбонатных серицитово-хлоритовых филлитов ($\bar{x} = 9{,}003$; $s = 0{,}007$; $n = 64$) указывает на давление 2,5—3 кбаров, соответствующее границе низких и средних давлений. Эти данные свидетельствуют о том, что во время герцинского метаморфизма в пределах Альп-Карпатской области существовали значительные разници в давлениях.

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CHANGES OF MAGMA CHAMBER DEPTHS IN THE CARPATHIAN BASIN DURING THE TERTIARY AND QUATERNARY

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On the basis of nearly 2000 chemical analyses and of their statistic evaluation some features of magma genesis and evolution could be determined. Assuming the analogy between the recent and paleo volcanism the paleo magma chamber depths were determined by means of analyzing their magma chemical and seizmic conditions and applying the theory of plate tectonics.

The changes of magma chamber depths of the Intra-Carpathian Neogene and Plio-Pleistocene are given for each mountain and phases (geological cycles). Disregarding the acid "core" volcanism believed of crustal origin and developed mostly in the internal basin parts the chemistry of the andesites became gradually acid from the Late Helvetian through the Sarmatian. This indicates the ever higher position of the magma chambers. At the same time, volcanism was gradually displaced east- resp. northwards. In the Plio-Pleistocene the chemistry proved to be andesitic and andesito-basaltic, resp. basaltic but the magma chamber depth did not increase as compared to the Sarmatian. The final basalts of alkali type occurring in the internal part of the basin suggest much greater magma chamber depths.

The changes of magma chamber depths support the statements of Hungarian, Ukrainian, Romanian and Italian researchers who correlate the Carpathian Tertiary and Quaternary volcanism with subduction activities.

In this study the data of statistic evaluation will be presented which are of greatest significance from the magmagenetic point of view: average of SiO_2, average of alkalies, the SiO_2/K_2O ratio and the paleo magma chamber depths, the Na_2O/K_2O ratio characterizing the degree of contamination, computed by mountains and volcanic cycles (geological epochs). The depths of magma chambers have been determined by the NINKOVICH-HAYS diagram improving the HATHERTON-DICKINSON principle [8, 13, 28, 29, 30, 31].

To determine the depths of paleo magma chambers by applying the diagrams plotted on the basis of the magma chemistry of recent volcanic series and related BENIOFF-planes, this is a method used on a world scale. The accuracy of determination is believed to be of $\pm$ 10 km. As to my own opinion, in a megatectonic unit of given tectonic position this method is suitable to determine only the changes of the relative magma chamber depths, since the differences caused by the chemistry of the melting crust parts exclude the possibility of comparison of absolute validity.

The statistic values as well as the data of the comparative igneous associations are summarized in the Table. The numbers accompanying the names of mountains refer to the publications providing the chemical analyses. When

comparing the statistic values, the results of McBIRNAY concerning the andesite and those of Kutolin concerning the basalt associations, were used [3, 27]. In the NINKOVICH-HAYS diagram (see the Figure) the SiO_2/K_2O values of the volcanic cycles of each mountain are shown according to geological epochs.

To take a comparison the magma-chemical character of the Laramian magmatism of the Carpathian Basin (Upper Cretaceous — Paleogene: Vladeasa, Bihor; Upper Eocene: Eastern Mátra, Transdanubia) was also determined. On this basis it has been stated that no significant differences could exist in the depths of magma chambers (150 km in average), but the igneous formations in Transylvania are more acid than those of the Pannonian Basin.

Volcanism started in the internal zone of the Pannonian Basin at the boundary of the Burdigalian and Early Helvetian, represented by the so-called "Lower Rhyolite Tuff" explosion of enormous quantity. The Lower Helvetian rhyolite tuff and the whole acid volcanic series called "core-volcanism" by G. PANTÓ and the mass of which is about five times larger than that of the andesitic stratovolcanic sequence, differ in genesis from the andesitic one. This is proved both by the chemical character, homogeneity and by the tectonic position [33, 34]. (According to the idea of E. SZÁDECZKY-KARDOSS [54] this acid magma chamber was probably formed during the temperature rise in the acid crust by the ascending branch of the deep current system below the Great Hungarian Plain.)

The ungrounded data for magma chamber depths obtained from the chemistry of rhyolite tuffs (depths greater than 200 km) can be explained also by this fact. This extremely high value can be traced back to the high K_2O-content of the partially molten crust and cannot be compared with the genesis of the andesitic magmas.

In the North-Hungarian volcanic mountains the magmatism is started in the Late Helvetian by volcanic-sedimentary sequences of normal andesitic composition characterized by magma chamber depths of 150 to 160 km. In the Tortonian the intensity of volcanism increases, in this period the stratovolcanic and closing lava blanket sequences of the Dunazug—Börzsöny, Cserhát—Mátra and partly of the Zemplén—Tokaj and Štiavnica-Kremnica Mountains are formed. During the period between the formations of the stratovolcanic sequence and that of the closing lava blanket caldera collapses took place in most of the mountains and ore mineralization is connected to this phase. At the same time, this is accompanied by the increase in the magma chamber depths (150 to 200 km). In Transylvania the volcanism started at the same time, with volcano-sedimentary formations similar to those formed in the basin centre in the Late Helvetian. North and east of the mountains characterized by Tortonian paroxysm the intense magmatism started in the Sarmatian (Štiavnica-Kremnica, Tokaj Mountains, Beregovo environment, Vihorlat-Gutin, Apuseni, Calimani, Gurghiu, Hargita Mountains). The magma chambers

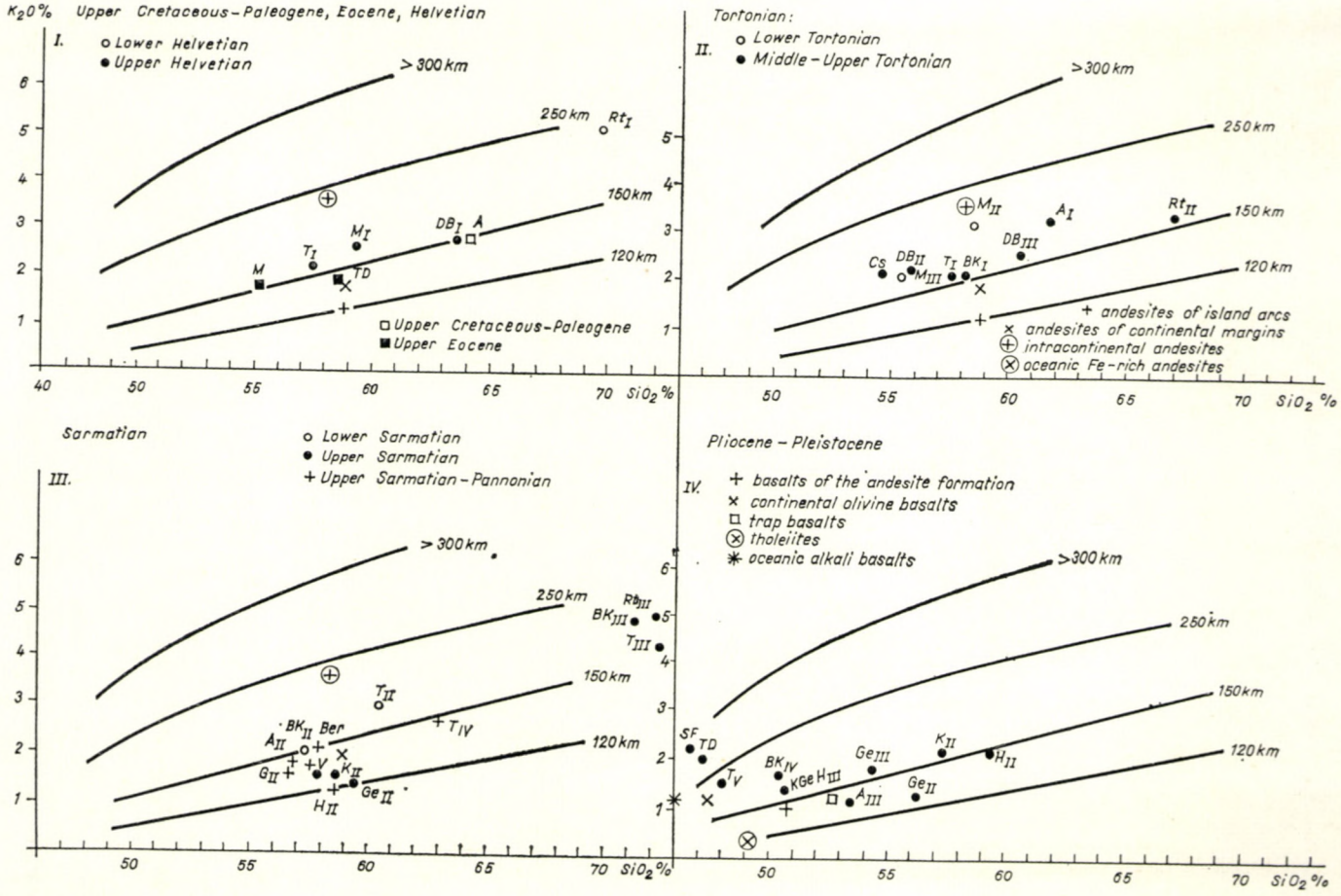
K2O%
Upper Cretaceous–Paleogene, Eocene, Helvetian
I.
Lower Helvetian
Upper Helvetian
>300 km
250 km
150 km
120 km
Rt I
M I
DB I
A
T I
M
TD
Upper Cretaceous–Paleogene
Upper Eocene
40
50
55
60
65
70
SiO2 %
Tortonian:
II.
Lower Tortonian
Middle–Upper Tortonian
M II
A I
Rt II
DB III
Cs
DB II
M III
T I
BK I
+ andesites of island arcs
× andesites of continental margins
⊕ intracontinental andesites
⊗ oceanic Fe-rich andesites
Sarmatian
III.
Lower Sarmatian
Upper Sarmatian
+ Upper Sarmatian–Pannonian
Rb III
BK III
T III
T II
BK II
Ber
A II
G II
V
K II
T IV
Ge II
H II
Pliocene – Pleistocene
IV.
+ basalts of the andesite formation
× continental olivine basalts
□ trap basalts
⊗ tholeiites
✳ oceanic alkali basalts
SF
TD
T V
BK IV
K Ge H III
A III
Ge III
Ge II
K II
H II

Fig. 1

Table 1

Mountains and volcanic cycles (age)	Number of analyses	SiO_2 %	Na_2O %	K_2O %	SiO_2/K_2O	Na_2O/K_2O	Depth of magma chamber (km)
Upper Cretaceous-Paleogene (4, 16) (Vladeasa-Bihor)	55	63.6	3.4	2.9	21.5	1.2	155
Eocene:							
Transdanubia (55, 56)	22	58.5	2.5	1.9	30.9	1.3	120
East Mátra Mts (26, 29)	98	55.0	2.2	1.8	30.6	1.2	160
"*Core*" *volcanism:* rhyolite tuff (dacite tuff) horizons (59)							
I. Lower Helvetian	7	69.6	1.7	5.1	13.6	0.3	200
II. Upper Tortonian	4	67.0	2.5	3.3	19.9	0.7	150
III. Sarmatian	5	72.1	1.1	4.9	14.5	0.2	200
Tertiary and Quaternary volcanic mts Dunazug—Börzsöny Mts (23, 26, 35, 62)	120						
I. Upper Helvetian	27	63.5	3.9	2.8	22.7	1.4	155
II. Lower Tortonian	72	55.5	3.2	2.1	26.4	1.5	170
III. Middle and Upper Tortonian	21	60.6	3.8	2.5	24.2	1.5	155
Cserhát Mts (26, 59) Middle and Upper Tortonian	50	54.5	3.3	2.2	24.8	1.5	180
Mátra	349						
I. Upper Helvetian	22	59.2	2.6	2.6	22.8	1.0	170
II. Tortonian (stratovolcanic unit)	233	58.3	5.8	3.2	18.2	0.7	210
III. Tortonian (closing lava blanket)	88	55.3	2.6	2.2	25.1	1.2	175
Presov—Tokaj Mts (12, 15, 44, 57)	518						
I. Upper Tortonian	92	57.5	2.9	2.1	27.4	1.4	155
II. Lower Sarmatian	134	60.5	3.9	3.0	20.2	1.3	180
III. Upper Sarmatian	153	72.5	2.9	4.2	17.3	0.7	160
IV. Upper Sarmatian and Pannonian boundary	132	63.0	3.1	2.6	24.2	1.2	150
V. Pannonian	7	48.0	2.7	1.5	32.0	1.8	180
Štavnica-Kremnica Mts (10, 11, 19, 22)	90						
I. Upper Tortonian	39	58.0	2.1	2.1	27.6	1.0	150
II. Lower Sarmatian	22	57.4	2.2	2.0	28.7	1.1	150
III. Upper Sarmatian	19	71.5	2.8	4.8	14.5	0.7	210
IV. Pliocene	10	50.4	2.9	1.6	31.5	1.8	160
Vihorlat-Beregovo Mts (2, 25, 32, 45. 58)							
Miocene	80	57.6	2.7	1.1	27.4	1.3	155
Pliocene		57.4	2.9	1.8	31.9	1.8	145
Gutii Mts (5)	87	—	—	—	—	—	—
I. Upper Tortonian-Lower Sarmatian							
II. Upper Sarmatian-Lower Pannonian	47	56.7	2.2	1.6	39.4	1.4	140
III. Pliocene (Pontian)	40	54.7	2.3	1.8	30.4	1.3	145

Mountains and volcanic cycles (age)	Number of analyses	SiO_2 %	Na_2O %	K_2O %	SiO_2/K_2O	Na_2O/K_2O	Depth of magma chamber (km)
Calimani-Gurghiu-Hargitha Mts (38, 40, 42)							
I. Upper Tortonian	—	—	—	—	—	—	—
II. *Calimani*							
Upper Sarmatian	50	58.5	3.2	1.5	37.9	2.1	130
Pannonian	17	57.1	3.8	2.1	26.9	1.8	155
Gurghiu							
Upper Sarmatian	6	59.4	3.5	1.4	42.4	2.5	125
Pannonian	31	56.3	3.6	1.3	43.3	2.8	130
Hargitha							
Upper Sarmatian	7	57.7	3.0	1.6	36.1	1.9	140
Pannonian	44	59.3	3.7	2.2	26.9	1.7	150
Pleistocene	6	50.5	2.8	1.4	35.1	2.0	160
Apuseni Mts (4, 16)	142						
I. Upper Tortonian	22	61.5	3.3	3.3	18.1	1.0	200
II. Sarmatian — Lower Pannonian	107	56.5	3.4	1.8	31.4	1.9	145
III. Pliocene	13	53.5	4.1	1.2	44.6	3.4	135
Final basalts (17) (Plio-Pleistocene)							
Transdanubia	194	47.1	3.4	1.9	24.1	1.7	250
Salgótarján-Filakovo	38	46.6	4.1	2.2	21.3	1.8	270
Andesite formations of the Earth (3)							
I. Andesites of island arcs	89	58.6	3.2	1.3	46.0	2.5	125
II. Andesites of continental margins	29	58.6	3.8	1.9	29.5	1.9	145
III. Intercontinental andesites	131	58.1	4.6	3.6	16.2	1.2	240
IV. Oceanic andesites (Fe-rich)	22	58.3	3.9	1.8	31.0	2.1	140
Basalt formations of the Earth (27)							
I. Basalts of the andesite formations	355	50.8	2.7	1.0	50.9	2.7	150
II. Continental olivine basalts	276	47.8	2.9	1.3	36.5	2.2	180
III. Trap-basalts	348	52.2	3.1	1.3	40.5	2.4	150
IV. Tholeiitic basalts	110	45.2	2.2	0.3	163.8	7.4	120
V. Oceanic alkali basalts	118	45.8	2.9	1.2	38.8	2.5	230

got higher positions (120 to 150 km), chemically the magma became more acid and abundant in K_2O (progressing contamination). In this period the type of ore mineralization also changed: the Au—Ag mineralization predominates as against the former Pb—Zn—(Cu) one. In the Štavnica-Kremnica and Prešov-Tokaj Mountains the appearance of wide-spread rhyolite tuffs points to the alternating occurrence of rhyolite tuffs deriving from the partial melting of regional extent of the upper crust and of andesitic magmas resulted by melting along subduction zones, resp. between these two kinds of volcanites strong interaction (hybridization) took place.

In the Pliocene and Pleistocene as against the Sarmatian predominantly acid magmas, mostly basic magmas developed (andesites — andesitobasalts, basalts). In this phase the normal calc-alkali basalts being the final members of the volcanic cycle of andesitic mountains (Hargita, Apuseni, Štavnica-Kremnica, Vihorlat-Gutin Mountains) deriving from 120 to 180 km depths, and the typical final alkali basalts characterized by 250 to 280 km magma chamber depths and formed in the Pannonian Basin (Transdanubia — Little Plain and the Salgótarján-Fiľakovo environment), should be distinguished. Slovakian researchers took similar genetic considerations though from other aspects [11].

When comparing the results with the data of the andesite and basalt associations of the Earth it can be stated that in the Helvetian and Tortonian the formations of mostly andesitic composition are akin to the intracontinental and continental-marginal andesite formations, while in the Sarmatian and Pliocene a deviation sets in towards the andesites of island arc type. The Pliocene and Pleistocene basalts are similar partly to the basalts of the andesite formations and partly to the continental olivine basalts. It is to be emphasized, however, that neither the andesites nor the basalts are uniform types as stated by K. KAROLUS [21].

Summing up: Disregarding the acid intra-basin crustal volcanism the chemical evolution of the Intra-Carpathian Tertiary and Quaternary magmatism is characterized by progressing acidity from the Helvetian up to the Sarmatian. This is counterbalanced by strong basification in the Pliocene and Pleistocene. The magmatism starting in the internal zone of the Pannonian Basin becomes younger when moving north-, northeast- and eastwards. In these directions the start and paroxysm as well as the final phase of magmatism is delayed by about 10 million years. The depth of magma chambers of the intra-basin volcanic mountains is somewhat greater (150 to 200 km) than the young mountains formed in the marginal parts (120 to 160 km).

The Carpathian-Ukrainian area as well as the Prešov-Tokaj Mountains developed under special tectonic conditions. They represent a transitional stage between the Transylvanian areas and the formations of the intra-basin mountains [45, 48]. The deviation of the western part of the Eastern Carpathians from the temporal regularity mentioned above, should be stressed, since in these areas the magmatism started in the Late Burdigalian and had been intesively active from the Tortonian through the Pliocene.

The Dunazug—Börzsöny and Štavnica-Kremnica Mountains as well as the Cserhát—Mátra Mountains represent individual structural units and can be traced back probably to two independent subduction zones, i.e. the Carpathian Klippe Zone and the Balaton Line [43]. The formation of the Vihorlat-Gutii and Calimani-Gurghiu-Hargitha Mountains can be deduced on the same way from the East-Carpathian Subduction zone. As to the opinion of

Hungarian (E. Szádeczky-Kardoss, 1972) and Rumanian (D. Radulescu, 1975) authors, however, the magmatism of the Apuşeni Mountains can be related to the subduction zone marked by the Mureş arc [6, 7, 14, 41, 46, 47, 52, 53, 54].

The Na_2O/K_2O ratio is strikingly low in general, i.e. the potassium enrichment is characteristic in the phase of most intense volcanic activity. At the same time, parallel with the rejuvenation of volcanism a decrease in the relative quantity of potassium can be observed. (Na_2O/K_2O: Upper Cretaceous — Eocene: 1.2—1.3; Helvetian: 1.0—1.4; Tortonian: 0.7—1.5; Sarmatian: 0.7—2.5; Pliocene-Pleistocene: 1.3—3.4).

It is remarkable that parallel with rejuvenation there are ever greater differences between the Na_2O/K_2O ratios of the individual mountains. This may be due to the structural differentiation of the mountains which produced considerable differences in contamination.

Further detailed evaluation of the compiled magma-chemical data will explain the formation of different rock types and the related ore mineralization of each mountain.

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ИЗМЕНЕНИЕ ГЛУБИНЫ ОЧАГОВ ВНУТРЕННО-КАРПАТСКОГО МАГМАТИЗМА В КАЙНОЗОЕ

Т. ПОКА

Резюме

По статистической разработке приблизительно 2000 петрохимических анализов установлены некоторые тенденции в генетических условиях и в развитии магм. По аналогии с современными вулканическими овластями, исходя из магматических анализов и сеисмического положения, на основе теории гловальной тектоники, делаются вывод о глувине палео-очагов.

Автором исследовано изменение глубины очагов внутренно-карпатского магматизма в неогене, по отдельным горным районам и по геологическим циклам (фазам). Не учитывая кислого вулканизма внутренней овласти бассейна к происходящего из коры в «андезитовых» горах отмечаются постепенное изменение химического состава к более кислому типу и также понижение глубины расположения очага, с позднего гельвета по сарматское время. Центр вулканической деятельности перемешался к В и С. В плиоцене и плейстоцене химизм стал андезитовым, андезито-вазальтовым и вазальтовым, но глубина очага не изменилась. Однако для очага типично финальных вазальтов щелогкого характера, развитых в то же время во внутренней овласти вассейна, рассчитывалась глубина 230—280 км.

Следовательно, происхождение магм кайнозойского возраста отдельных внутренно-карпатских гор связывается с различными зонами субдукции. Этот вывод совпадает с мнением некоторых венгерских, украинских, словацких, руминских и итальянских исследователей.

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CONTRIBUTION TO THE PROBLEM OF METASOMATIC DOLOMITIZATION

By

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In the Triassic formations of Hungary prymary dolomites as well as secondarily-metasomatically dolomitized limestones are known. The latter ones were generated along Middle Miocene (Styrian) faults and as a result of Pleistocene hot-spring activity. These rock types are demonstrated on the example of the Nagyszál region at Vác, an area explored also by boreholes.

Among the geological formations of Hungary dolomite sequences of large masses and mostly of Triassic age, are found. The geological, mineralogical-petrological and genetic features of these dolomites of primary character are known from the studies of E. JAKUCS, E. VÉGH-NEUBRANDT. Having surveyed the limestone areas secondary metasomatic dolomites were observed. The dolomitic limestone, limy dolomite and dolomite intercalations observed in open-cast limestone mining as well as in limestone exploring bores (first of all along major faults or in their close neighbourhood) are of irregular (veined, lenticular, stocky) appearance. These rock alterations observed in numerous locations in the Hungarian Central Mountains were generated by hot-spring, i.e. hydrothermal metasomatic effects.

The macroscopic observations will be demonstrated after the limestone area of Vác—Nagyszál. The geologic and tectonic setting of variegated occurrence of the Vác—Nagyszál limestone area as well as the percentual distribution of the rock types explored in the boreholes are demonstrated in the mining-geological map (Fig. 1), resp. in the cross-section pair of NNW-SSE direction (Fig. 2).

The dolomitic limestone, limy dolomite and dolomite intercalations found in the Norian limestone sequence show hydrothermal effects and dolomitization of different degree. They are connected with the dolomitization caused by strong hydrothermal metasomatism which acted on the Norian limestone simultaneously with or subsequently to the Miocene volcanism of the Dunazug andesite mountains.

Thus, the geological profiles of Fig. 2 were designed taking account of both primary stratification and subsequent metasomatism. In the profile "a"

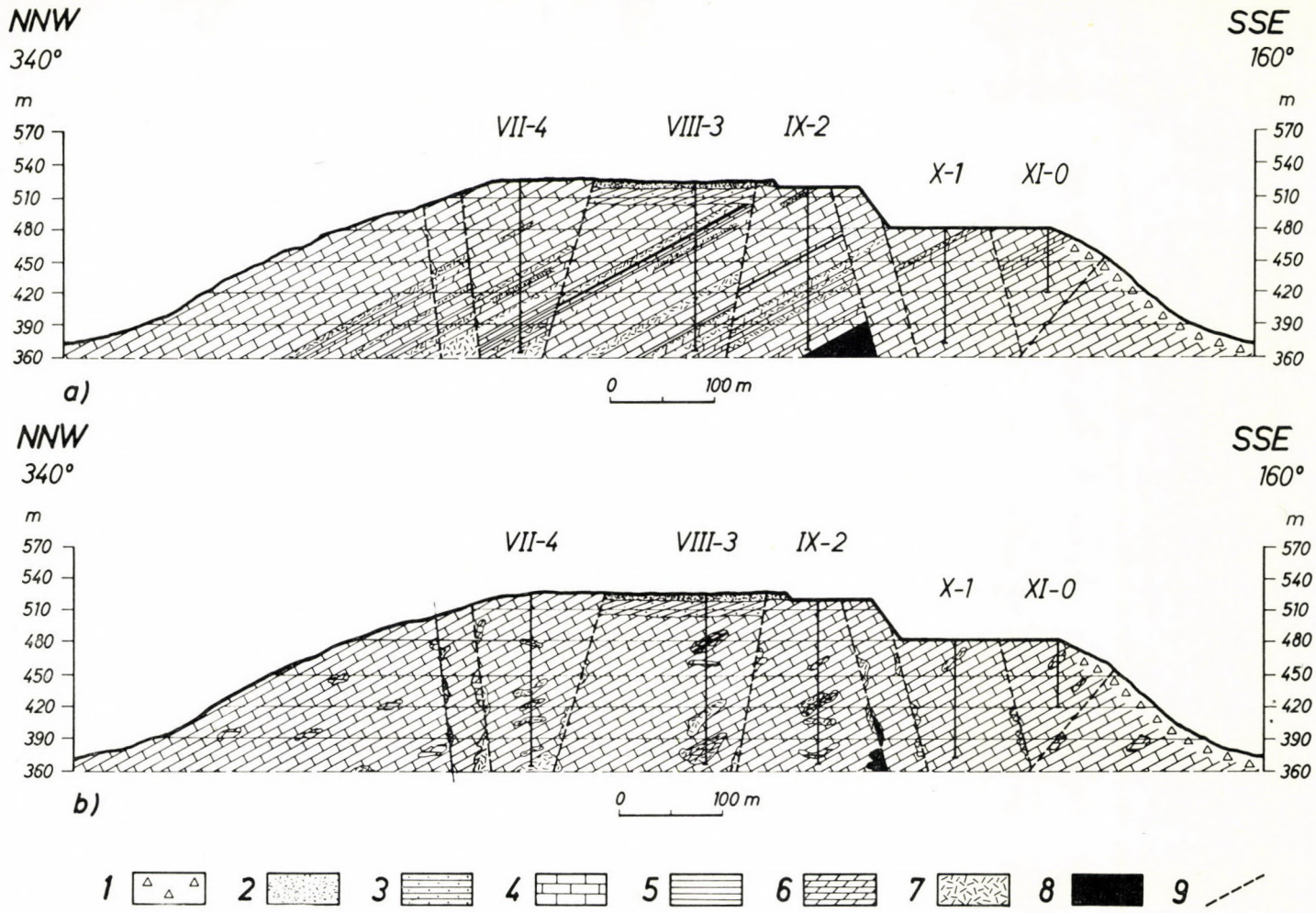

Fig. 2. Geological cross-section of NNE-SSE direction of the Vác-Nagyszál limestone area constructed by assuming: a) stratification, b) metasomatism. 1. Spoil-bank (Holocene); 2. clay, sand covering the surface or filling the caverns and issures (Pleistocene); 3. sandstone and conglomerate (Lower Oligocene); 4. limestone; 5. clayey and clay-intercalated limestone; 6. dolomitic limestone; 7. limy dolomite; 8. dolomite (Upper Triassic); 9. fault

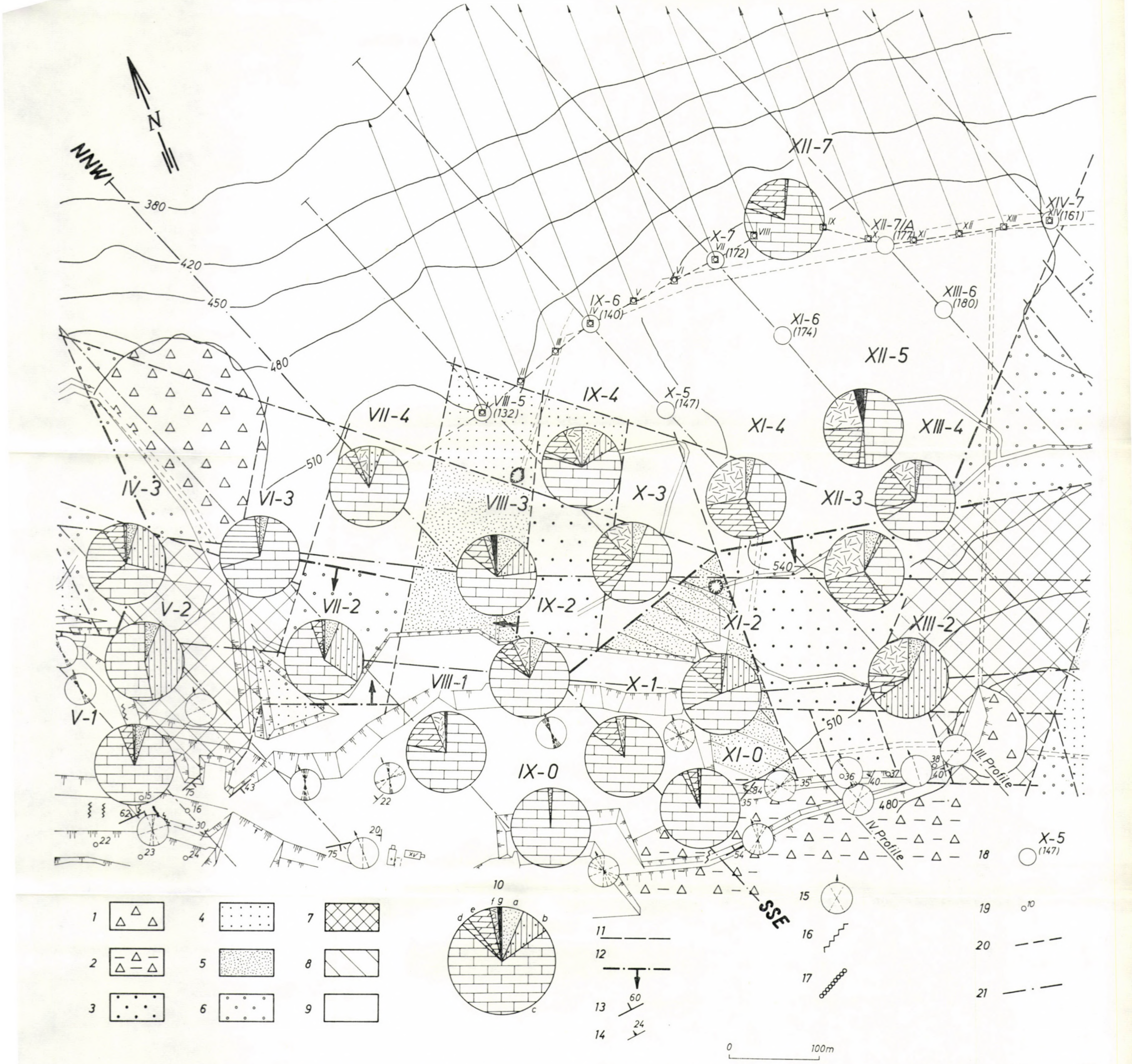

Fig. 1. Mining-geological map of the Vác-Nagyszál limestone area. 1. Spoil-bank (Holocene); 2. slope detritus (Holocene-Pleistocene); 3. 1—5, 4. 5—15, 5. 15—25, 6. 25—35, 7. 45—60 m thick sanstone and conglomerate (Lower Oligocene); 8. sanstone in the limestone sequence; 9. limestone, mainly with clayes limestone, dolomitic limestone and limy dolomite intercalation (Upper Triassic); 10. exploratory bore with the percentual distribution of the rocks crossed down to 420 m a.s.l.: a) clay and rock detritus (Pleistocene), b) sandstone and conglomerat (Lower Oligocene), c) limestone, d) clayey and clay-intercalated limestone, e) dolomitic limestone, f) limy dolomite, g) dolomite (Upper Triassic); 11. fault constructed after bores, structural and geolectric measurements and geomorphological patterns; 12. fault determined also by geoelectric measurement; 13. fault dip; 14 dip of strata; 15. lithoclase diagram; 16. calcite vein; 17. hot-spring cave entrance; 18. suggested exploratory bore (number and depth in metres); 19. sampling site at the mine-wall; 20. border of the area to be explored; 21. profile

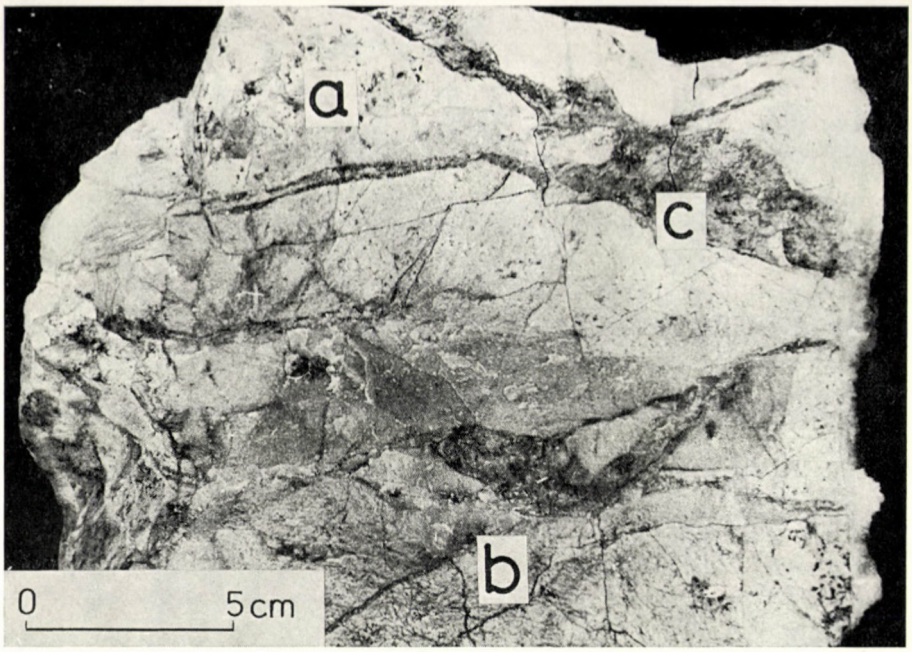

Fig. 3. Dolomitic limestone and limy dolomite in the metasomatically dolomitized Norian limestone. Sample 1—3. (Eastern part of the Vác-Nagyszál limestone quarry at the 480 m niveau). a) Limestone; b) dolomitic limestone; c) limy dolomite

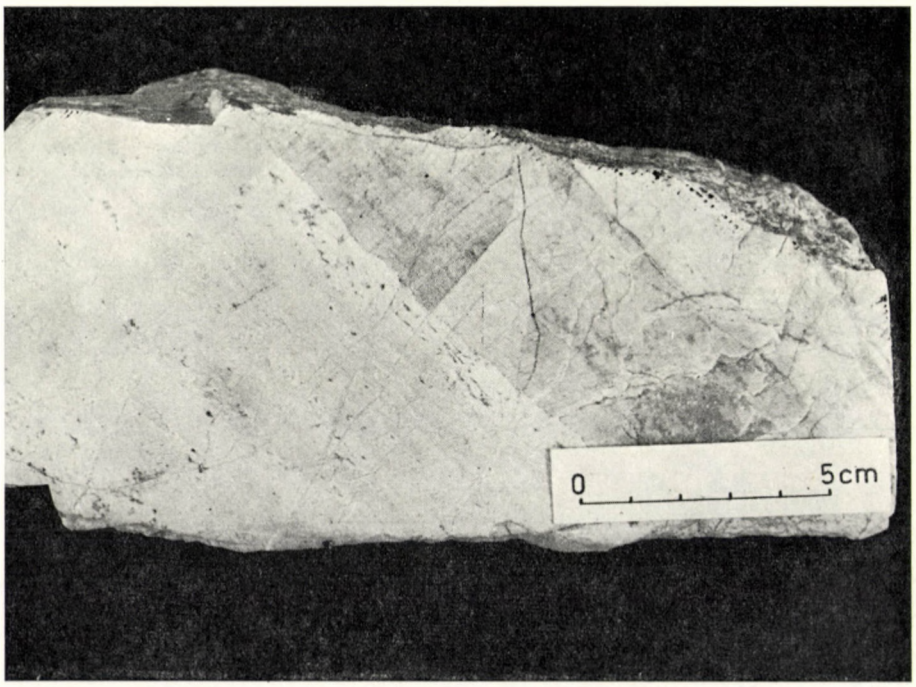

Fig. 4. Banded contact of limestone (dark band) and limy dolomite of metasomatic origin (left side) (Vác-Nagyszál limestone quarry, 480 m). Sample 4

of Fig. 2 the dolomitic formations crossed by boreholes were demonstrated as if they formed a continuous beds syngenetic with the limestone.

As a result of our observations, however, and based on the core matter as well as on the exploded rocks, it was stated that the dolomitic rock types found in the limestones do not form continuous strata but as it is shown in the profile "b" of Fig. 2 these are of veined, lenticular, stocky and alveolar appearance, in harmony with the hydrothermal activity. These profiles were plotted taking into account the respective thickness values revealed by the boreholes as well as on the basis of the forms found in surficial samples and mining walls, resp. of their statistic proportions.

As to our observations the fact of metasomatic dolomitization bears the same character both macro- and microscopically. Accordingly, the forms of appearance reflecting the geological processes can be regarded to be valid of the whole rock mass. These phenomena will be illustrated by several typical samples resp. by their investigation.

In the pictures of hand-specimens (Figs 3—5) and core samples (Figs 6—7) the forms of metasomatic dolomitization are well visible.

In Fig. 3 the veined metasomatic dolomitization of different degree of the limestone is shown. The chemical compositions of the dolomitic limestone (Fig. 3, field "b") and limy dolomite (Fig. 3, field "c") differing both in colour and in texture from the typical limestone (Fig. 3, field "a") are presented in Table 1.

Table 1

Chemical and mineral composition of the

No.	Term (geological age) of the rock	Location	Ignition loss	SiO_2	Al_2O_3
1.	Greyish-white, compact limestone (Norian)	Limestone quarry eastern side of the level of 480 m	43.90	0.60	0.09
2.	Yellowish-grey dolomitic limestone (Norian)	See sample 1.	44.54	0.78	0.17
3.	Yellowish-grey limy dolomite (Norian)	See sample 1.	45.66	0.35	0.20
4.	Yellowish-grey compact limy dolomite (Norian)	Limestone quarry level of 480 m	46.03	0.12	0.18
5.	Yellowish-grey dolomitic limestone (Norian)	See sample 4.	45.11	0.29	0.12
6.	Yellowish-grey limy dolomite (Norian)	Borehole No. XIII-4 depth: 138.6 m	46.23	0.08	0.05
7.	Yellowish-grey limy dolomite (Norian)	Borehole No. VIII-3 depth: 138.0 m	46.86	0.32	0.05

Fig. 4 shows the banded contact of limestone and limy dolomite, while Fig. 5 illustrates the irregular (stocky) contact of limestone and dolomitic limestone. In Fig. 6 a rock surface of characteristic impregnated limy dolomitic material can be seen.

The analytical results of the rock types in question deriving from the Nagyszál-area and shown in the figures are comprehended in Table 1. (The numbers of analyses are found also in the legends of Figs 3—7.)

The results of microscopic investigations concerning the metasomatic dolomitization were published in a previous paper.

In favour of local verification of the geological age of metasomatic dolomitization a core sample of bore VIII-3 drilled in the centre of the area is introduced (Fig. 7). The filling of hot-spring origin (freshwater clayey limestone + calcite) precipitated during the Pleistocene hot-spring activity in the Norian limy dolomite. The dolomitization of the Norian limestone, however, took place earlier, i.e. in the Middle Miocene during the Styrian orogenic movements and was due to the postvolcanic hydrothermal effects of the andesites of Visegrád- and Börzsöny Mountaines.

These observations provide a basis for the geochemical and geological investigations in progress. For instance, at the contact of the limestone and andesite (in greater depth) metasomatic, skarny and polymetallic, first of ass sulfide ore formation can be assumed.

rock types described in the paper

Chemical composition in per cent								Mineral Thermal	Composition X-ray
Fe_2O_3	TiO_2	CaO	MgO	Na_2O	K_2O	SO_3	$\frac{CaO}{MgO}$	analyses	
0.10	—	55.14	0.10	0.14	0.10	0.20	—	Calcite	Calcite
0.29	—	45.56	7.40	0.10	0.10	0.20	6.29	Calcite, dolomite	Calcite, dolomite
0.17	—	40.13	13.20	0.10	0.10	0.43	3.04	Dolomite, calcite	Dolomite, calcite
0.15	0.01	39.08	14.51	0.01	0.01	0.14	2.69	Dolomite, calcite	Dolomite, calcite
0.19	0.01	45.88	8.34	0.01	0.01	0.04	5.50	Calcite, dolomite	Calcite, dolomite
0.17	0.01	39.40	14.02	0.01	0.01	0.02	2.89	Dolomite	Dolomite, calcite
0.10	0.01	33.39	18.83	0.10	0.13	0.04	1.77	Dolomite	Dolomite

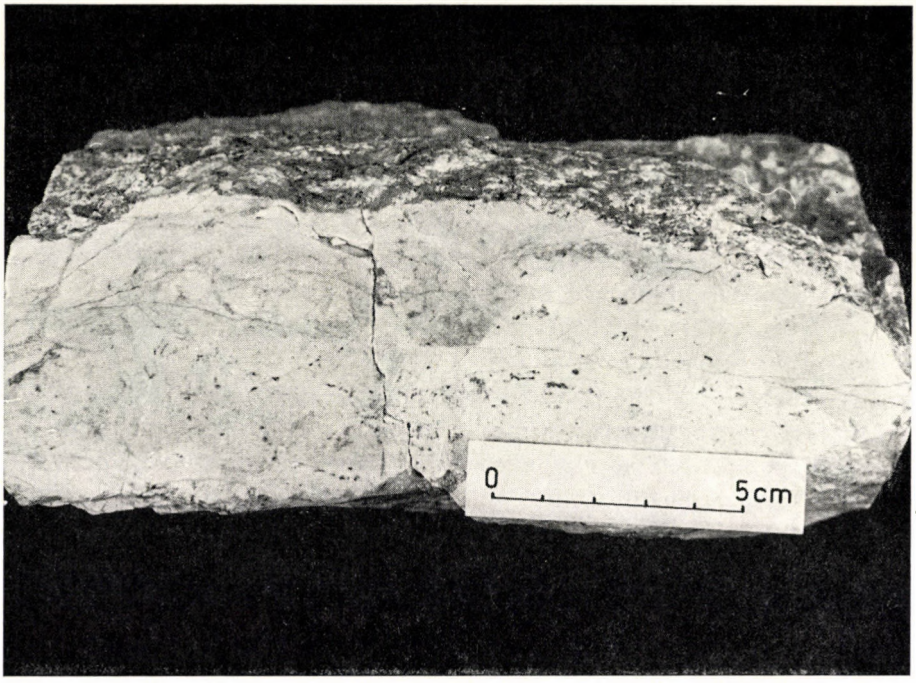

Fig. 5. Irregular contact of limestone (upper darker spots) and dolomitic limestone of metasomatic origin (stocky). (Vác-Nagyszál limestone quarry, 480 m). Sample 5



Fig. 6. Characteristic metasomatically dolomitized (impfegnation) rock surface (core sample, bottom view). (Vác-Nagyszál, borehole No. XIII-4. 138,6 m.) Sample 6

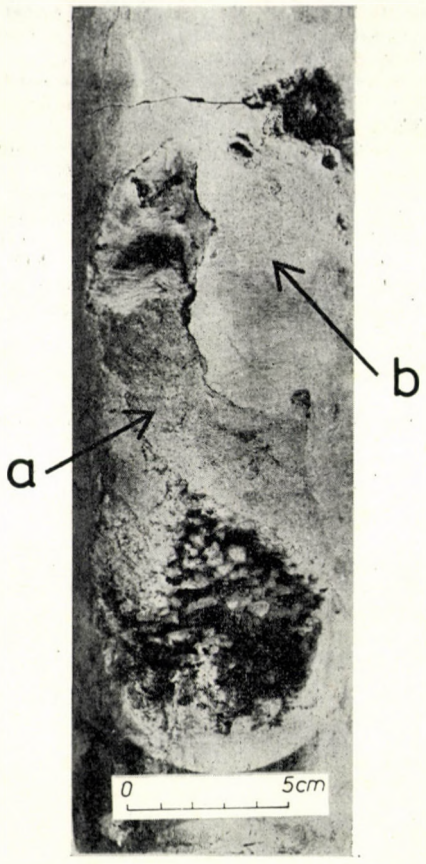

Fig. 7. Filling of hot-spring origin (freshwater clayey limestone and calcite (a) (in metasomatic limy dolomite; (b) (core sample, side-view). (Vác-Nagyszál borehole No. VIII-3. 138,0 m). Sample 7

In the Mesozoic basement areas those suitable to drill for thermal water can be designed by means of the former hot-spring and hydrothermal traces.

In case of the industrial utilization of the limestone because of its irregular pattern the metasomatic dolomitization requires special considerations from the points of view of utilization and prognostics.

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К ВОПРОСУ МЕТАСОМАТИЧЕСКОЙ ДОЛОМИТИЗАЦИИ

Д. ВИТАЛИШ—Й. ХЕДИ-ПАКО

Резюме

В триасовых отложениях Венгрии известны первичные доломиты, а также вторично метсоматически доломитизированные известняки. Последние образовались в процессе деятельности плейстоценовых термальных источников вдоль тектонических разломов среднемиоценового (штирийского) возраста. В работе они рассматриваются на примере участка Надьсал у г. Вац, хорошо изученного, разведенного даже глубоким бурением.

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METAMORPHISMUS, GRANITOIDBILDUNG UND PLATTENTEKTONIK IM GROßEN KAUKASUS

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Am Konvergenzrand der eintauchenden lithosphären Platte des Großen Kaukasus wurde ein Streifen der Hochdruckmetamorphiten gebildet und die regionale Hochtemperaturmetamorphose niedrigen Drucks erfolgte in der kontinentalen Kruste im Stadium der Faltungsdeformationen auf relativ geringer Tiefe.

Der Große Kaukasus ist der nördlichste gefaltete Gebirgsbau des Anatolisch-Kaukasisch-Iranischen Segmente des mediterranen Mittelmeergürtels (S. Schema). Nördlich davon liegt die epiherzynische Scythische Tafel, südlich das transkaukasische epipaläozoische Zwischengebirge. Die Bildung des mediterranen Geosynklinalengürtels erfolgte am Anfang des baikalischen Zyklus. Die maximale Dehnung der Erdkruste mit Trennung der einheitlichen Osteuropäischen und Afrikanisch-Arabischen Kratonen fand aber am Anfang des kaledonisch-herzynischen Zyklus statt. Im Großen Kaukasus ist sie in den Zonen der Vor- und Hauptkammes zu beobachten. Hier zerriß die epibaikalische Kontinentalkruste, eine ozeanische Kruste bildend (KHAIN, 1974; KHAIN, 1975; GAMKRELIDSE, 1976).

Die vormitteldevonische (Nordkaukaser Ophiolithgürtel) ist das Relikt der ozeanischen Kruste. Dieser Gürtel verläuft in der Tiefe weit nach Westen und Osten und bildet die Südgrenze der Osteuropäischen Platte (SMIRNOVA, 1974; KHAIN, 1965). Der Gesteinskomplex gehört zur Basis der Überschiebungsplatten, die insgesamt als tektonische Wurzelzone angesehen werden (SCH. A. ADAMIA). Nach tektonischer Stellung und Bau stellt dieser Komplex die dritte ozeanische Schichte dar und gehört zum oberen Mantel des eugeosynklinalen ozeanischen Beckens. Diese basit-hyperbasitischen Gesteine mit spätdermischer metasomatischen Granitisierung (360. Mio. J. nach AFANASJEW, 1968) wurden mit Gesteinen der 1. und 2. ozeanischen Schichten tektonisch vereinigt.

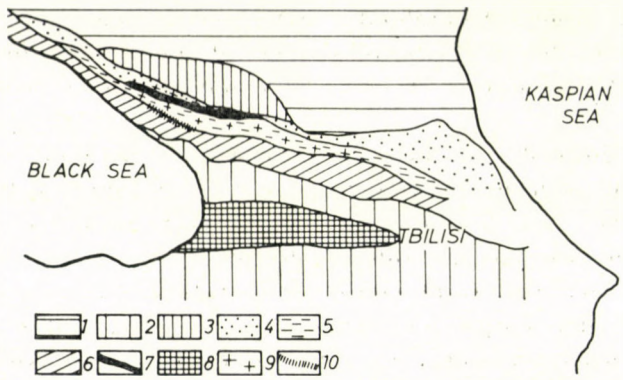
Neben basitischen und hyperbasitischen Gesteinen wurden auch kleinere Eklogitvorkommen festgestellt (TATRISCHWILI, 1969, 1970; AFANASJEW u. a. 1969). Die epibaikalische Scythische Tafel wurde am Ende des Ordovizium oder im Silur zersplittert. In der Zone des Vorkammes des Großen Kaukasus entstand die herzynische Eugeosynklinale. Die nördlich geneigte BENIOFF-Zone

wurde im Mittelpaläozoikum entlang der Südgrenze des Vorkammes angelegt (KHAIN, 1971).

Nachfolgende Kompression rief die bretonische Faltungsphase hervor (ADAMIA, 1968; GAMKRELIDSE, 1976), mit Bildung der Kordilleren-Inseln in der Haupt- und Vorkamm Zone. Um diese Zeit erfolgt der regionale (vorgranitoidische, genauer vorplagiogranodioritische) Metamorphismus höheren Grades, der als Relikte in jüngeren Gesteinen zu beobachten ist und mit dem Entstehen der anatektischen Plagiogranodiorite zusammenfällt. Die Gesteine des vorplagiogranitischen Metamorphismus entsprechen der Grünschiefer-, Epidot-Amphibolit und Niedrigtemperatur Subfazies der Amphibolitfazies, bewiesen durch: 1. paragenetische Assoziationen großer gefalteter Xenolithe und Nebengesteine, in die intrusiven (allochthonen) palingenen Granitoide der plagiogranodioritischen Formation eingeschlossen, 2. die Gesteine, der epidot-amphibolitischen und niedertemperaturen Subfazies der Amphibolitfazies sind älter (Argon-Methode) als die palingene Granitoiden und anatektische Migmatite hoher Stufe der Amphibolitfazies, 3. Vorhandensein anatektischer Migmatite der Hochtemperatur-Amphibolitfaziesen der Plagiogranodiorit-Formation.

Die sudetische Faltungsphase entspricht dem frühorogenen Entwicklungsstadium des Großen Kaukasus. Es treten intensive Faltungsprozesse, regionaler Niederdruckmetamorphismus der Amphibolit- (selten Granulit-) Fazies auf mit Plagiomigmatiten und wasserreichen Schmelzen von plagiogranitischer und plagiogranodioritischer Zusammensetzung, entsprechend dem vulkanischen Gürtel der Inselbogen der Kontinenträndер Doppelgürtel des regionalen Hochdruckmetamorphismus, wurde im Großen Kaukasus nicht festgestellt. Das kann möglicherweise durch die langsamere Senkungsgeschwindigkeit der lithosphären Platte erklärt werden. Nach einigen Forscher hatten hier Lithosphärenplatten bis zum Spätpaläozoikum geringere Mächtigkeit und bewegten sich langsamer, wobei regionaler Hoch- und Mitteldruckmetamorphismus vorherrschte und untypische Glaukophanschiefer nur zufällig gebildet werden (MIYASHIRO, 1973). Der metamorphe Hochdruckgürtel konnte auch wesentlich enger und von jüngeren Sedimenten bedeckter sein, als der Niederdruckgürtel mit Granitoiden. D. M. SCHENGELIA und D. N. KEZCHOWELI finden in den Metamorphiten des Südteiles des Hauptkammes (mittelpaläozoische Laschtrak-Folge) nicht nur Mitteldruck-, sondern auch Hochrruck Paragenesen (Cyanit-Sillimanit-Type), ferner Chloritoid-Biotit-Muskovit Reliktparagenese in Gesteinen der Disthenschiefer-Fazies (V. V. CHLESTOW, 1974, d. h. 4. basischer Subtyp, MIYASHIRO, 1961). Wir meinen daher, daß während der sudetischen Kompressionsphase ein schmaler Streifen von Hochdruckmetamorphiten gebildet wurde, der heute direkt mit dem metamorphischen Gürtel niedrigen Druckes in Berührung kommt. Zwischen dem vulkanischen Gürtel der Inselbogen kontinentalen Randes und der Rinne wurde hier eine nicht metamorphisierte Bresche nicht festgestellt. Die maximale Verringerung der

Entfernung erfolgte wahrscheinlich infolge der Verlegung der BENIOFF-Zone in Richtung des Bogens. Selektives Schmelzen unter sehr hohem partiellem Wasserdruck und hohem Ab/An-Verhältnis bildete in der sudetischen Faltungsphase überwiegend plagioalaskitische Produkte (SCHENGELIA, 1973). Das Substratum wird mit Anortit angereichert. Die Verminderung des Ab/An-Verhält-



Tektonisches Gliederungschema des Kaukasus: 1. Epiherzynische Scytische Tafel; 2. Transkaukasisches epipaläozoisches Zwischengebirge; 3. Betschasinzone; 4. Zone des Vorkammes des Großen Kaukasus; 5. Zone des Hauptkammes des Großen Kaukasus; 6. Zone der Südflanke des Großen Kaukasus; 7. Die »Wurzel« (= spreading) Zone; 8. Adsharo-Trialetzone; 9. Niedrigdruck Metamorphite und paläozoische Granitoide; 10. Mittel- und Hochdruckmetamorphite

nisses bei der wiederholten Anatexis während der asturischen Faltungsphase trug zur Bildung der mit Kalium angereicherten Granitoid-Schmelze (SCHENGELIA, 1973) bei.

Die meist intensive Mikroklinisation entspricht der asturischen Phase (SCHENGELIA, 1969). Porphyroblaste mit frischen Mikroklineinschlüssen enthalten kataklastische Quarz-, Plagioklaskörner und mylonitisierte Gesteinsrelikte. Die intensive Mikroklinisation der Kristallinenschiefer (Plagiogneise), sowie die Bildung der kataklastischen und mylonitisierten palingenen Plagiogranite und Plagiogranodiorite (ZARIDSE, TATRISCHWILI, 1953, 1959, u. a.; AFANASJEW, 1958; SCHENGELIA, 1969) erfolgte wesentlich später. Die Verminderung von P_{H_2O} führt auch zur Anreicherung der anatektischen Schmelze mit Kalifeldspat. Als Quelle des Kaliums kann das Anatexis-Gebiet betrachtet werden, das früher quarzalbitische Substanz mobilisierte. Gleichzeitig wurden die Metamorphite auch durch eine regressive regionale Metamorphose beeinflußt, wobei Lösungen entstanden, die die Granitisierung der Gesteine des kristallinen Substratums hervorriefen.

Berechnungen zeigen, daß Kalium in vielen Fällen nicht ausreichend ausgelaugt werden konnte. Wir meinen daher, daß Wärme und ein Teil der litophilen Elemente, darunter auch Kalium aus der BENIOFF-Zone stammt.

Zusammenfassend sei betont, daß im Paläozoikum des Großen Kaukasus der früheste Metamorphismus niedrigen Druckes (bis zur Grünschieferfazies) mit der Senkung der Geosynklinale in Verbindung steht.

Die bretonische Faltungsphase bedingt die Bildung der Kordilleren-Inselns in den tektonischen Zonen des Haupt- und Vorkammes. Da erfolgte auch der Vorgranitoide (vorplagiogranitoide) Hochtemperaturmetamorphismus (Grünschiefer-, Epidot-Amphibolit- und Niedrigtemperatur-Subfazies der Amphibolitfazies), dessen Produkte weitgehend von den nachfolgenden metamorphischen Prozessen vernichtet wurden und als kleine Relikte erhaltengeblieben sind.

Der nachfolgende regionale Metamorphismus fand in der sudetischen Phase statt, die auch die Weiterentwicklung der Kordilleren-Inselns hervorrief. Dies stellte einen Niedrigdruck -Hochtemperaturmetamorphismus dar, der (fast) gleichzeitig mit der Bildung der plagiogranodioritischen (anatektischen) Granitoiden bei Amphibolit-, seltener Granulit-Fazies-Bedingungen. Produkte dieses Metamorphismus entsprechen ihrem Typ und ihrer Zusammensetzung nach den vulkanischen Gürteln der Inselbogen der Kontinenträder.

Der terminale Metamorphismus im Großen Kaukasus ist die asturische Phase und stellt einen regressiven, regionalen Metamorphismus dar mit Bildung von metasomatischen Niedrigtemperatur- Mikroklingranitisation.

Die BENIOFF-Zone entstand hauptsächlich im herzynischen Zyklus, und liegt wahrscheinlich unter den Mittel- und Hochdruckmetamorphiten südlich vom Nordkaukasischen Ophiolit-Gürtel. Die Bildung dieser Metamorphiten erfolgte wahrscheinlich in Tief seerinnen im Zusammenhang mit der BENIOFF-Zone.

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МЕТАМОРФИЗМ, ОБРАЗОВАНИЕ ГРАНИТОИДОВ И ТЕКТОНИКА ПЛИТ НА БОЛЬШОМ КАВКАЗЕ

Г. М. ЗАРИДЗЕ—Д. М. ШЕНГЕЛИЯ

Резюме

На краю конвергенции погружавшейся литосферической плиты Большого Кавказа образовался поях метаморфитов в условиях повышенного давления, причем высокотемпературный региональный метаморфоз пониженного давления в континентальной земной коре произошел в стадии возникновения складчатых деформаций на относительно небольших глубинах.

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GENETIC SIGNIFICANCE OF RARE EARTH ELEMENTS IN THE GRANITOID ROCKS OF HUNGARY

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In Hungary two genetically different granitoid occurrences are found at the surface. The different conditions of formation of the anatectic (Mecsek Mountains) and igneous (Velence Mountains) granitoids are reflected both by the accessory trace mineral content and by the distribution of REE. The Mecsek Mountains is characterized by the primary and secondary forms of monazite as well as by the predominance resp. occurrence of the minerals of the bastnaesite group. The Velence Mountains represents a progressed state of igneous differentiation and is rich in xenotime. The change of element content is parallel with evolution of minerals regarding the monazite, bastnaesite and the Zr-minerals. Geochemically, the Velence Mountains shows higher abundances of heavy lantanides. In the normalized distribution of the REE indications pointing to genetic differences can be observed. Regarding the ratio of light and heavy lantanides the two occurrances completely differ from each other.

Introduction

Detailed petrological and geological description of the granitoid rocks of Hungary was given by JANTSKY (1957), SZÁDECZKY-KARDOSS (1968) and BUDA (1969, 1972, 1974). The two outcrop areas, i.e. the Mecsek Mountains and the Velence Mountains (Fig. 1), differ in genesis from each other.

In the Velence Mountains magmatism is represented by Hercynian (Carboniferous) synorogenic and postorogenic genuine igneous granites and quartz-diorites of batholite type and of discordant position. In the Mecsek Mountains deeper-level, synorogenic anatexitic (Caledonian) and late-orogenic potash--metasomatic (Hercynian) granitoids are found. These granitoids were divided by BUDA (1974) into three groups: 1. amphibolite, conformable with the anatexitic granite; 2. two kinds of anatexitic porphyroblastic granodiorite with white, resp. red porphyroblastic microline; 3. postorogenic potash-metasomatic, fine-grained, veined granite and aplite.



Fig. 1

Methodology

To shed light upon the genetic differences and upon the origin of the material of granitoids further information can be obtained by means of investigating the trace elements and accessory trace minerals of the rocks with special regard to the rare earth minerals and elements.

The investigation of rare earth minerals was carried out by means of the electron microprobe of JXA-5 type (PANTÓ, 1975). Routine trace element analyses were carried out by means of Q-24 UV quartz spectrograph, while the REE analyses were done by spark source mass spectrometry (JEOL JMS-OIBM-2 type) under the assistance of the Inter-Departmental Committee on Rare Metals in the Mining Research Institute (CORNIDES, 1973). During the evaluation of the analytical data correlation analyses were also carried out, in addition to the well-known graphic methods (variation diagram, distribution normalized against average chondrite).

Trace Mineralogy

1. Rutile associated to biotite is characterized by Nb $>$ Ta. Sc-content is characteristic of the granites of the Velence Mountains, being negligible in the Mecsek Mountains. Sn is known to occur only in the Velence Mountains, in rutile, and this suggests a pegmatitic origin.

2. In the Velence Mountains the commonly wide-spread zircon contains Y due to its iso-structural connection with xenotime, this is, however, not characteristic of the anatexitic material of the Mecsek Mountains. Y-bearing zircon contains also P; among the numerous heavy REE Dy is most abundant.

Zircon is often zoned in both mountains. Zoning is mostly caused by increasing Ca-content from the core towards the margin. On this basis an

early Ca-free and late Ca-bearing generation can be recognized. Th is enriched together with Ca in the Velence, while in the Ca-free parts in the Mecsek Mountains. Thus, on account of genetic differences Th is bound to Zr and Hf in the Mecsek Mountains and to the Y-earths accompanying Ca in the Velence Mountains.

3. Xenotime is an accessory mineral typical of the Velence Mountains. It is not known in the Mecsek Mountains. The occurrence in the properly igneous granitoids and the enrichment in aplite indicate the separation of the Ce- and Y-earths, the latter being gradually shifted towards the subsequent phases. The succession of heavy lantanides is as follows: Dy > Yb > Er > Ho; subordinately Gd, Tb, Sm, Tm and Nd should also be revealed.

4. Although monazite occurs also in the Velence Mountains, it is quantitatively more common in the Mecsek Mountains. It is absent in amphibolite. Contribution of Ca and Th during monazite formation results in varieties gradually approaching a cheralitic composition. Cheralite is younger than monazite. The bulk of their REE content consists of Ce, La and Nd. During progressive crystallization Ca will be gradually enriched, in extreme case the final member is apatite.

In the Mecsek Mountains, in addition to this primary monazite-cheralite assemblage metasomatic, star-shaped monazite, 10 to 15 microns in diameter, has been formed in thin calcite veins by carbonatic mobilization. Its REE probably derive from the content is Ce > Nd > La > Gd > Pm > Tb. In addition to these, it contains 1.8 per cent S as well.

5. Apatite, wide-spread in the granitoid rocks of both mountains is relatively free of trace elements, within the detection limit of the electron microprobe. It can be demonstrated, however, that Ce, Y, Si and Sr become significant in the Velence Mountains in general and the presence of F is unambiguous.

6. When investigating the distribution of the carbonates with substitute anions sharp differences are found between the two mountains. The minerals of the bastnaesite group are wide-spread in the Mecsek and unknown in the Velence Mountains.

In the Mecsek Mountains its occurrence does not coincide that of monazite. It favours amphibolite, amphibole-bearing granodiorites, resp. colourless porphyroblastic microcline granodiorite. Its formation is undoubtedly secondary and is connected with the re-arrangement of REE mobilized as a result of anatectic conditions. Monazite occurs in the Mecsek Mountains in the red porphyroblastic microclinic variety and is bound to biotite. Both this fact and other geochemical features show that the red microclinic granodiorite is more akin to real igneous granite than is the white one.

The members of the bastnaesite group contain first of all light REE, thus under the given genetic conditions these can be regarded as minerals

substituting monazite. The element content of the bastnaesite group and its variation during formation is worthy of mention. Y is the first to play a predominant role, i.e. the member close to doverite (Ca, Y, Ce, La, Nd, Pr, Gd, Sm, Eu, F, C). Later Y disappears and Ca, Ce, La, Nd, Pr, Gd, F and C become predominant. In the bastnaesites occurring in aplites Th also appears accompanied by a decreasing Ca-content. The REE elements of bastnaesite presumably derive from the rocks preceding granitization. Secondary bastnaesite can be found in the bauxites around the Velence Mountains (BÁRDOSSY et al., 1976). Part of the material of these bauxites is derived from granite, the REE from the feldspar of the granite.

Considering the REE content of the minerals it can be stated that, parallel with progressive differentiation, the light and heavy REE become separated. In the Velence Mountains the heavy REE are enriched in independent minerals. In the Mecsek Mountains the minerals accumulating Ce-earths prevail, the members of the final differentiation products accumulating the Y--earths have not been developed.

Geochemistry

The REE form a group of closely related elements but their distribution patterns change during formation. If in the given cases the *regularities of distribution* can be well interpreted, significant genetic conclusions can be drawn. The total REE content may vary within each group but curve normalized against average chondrite shows negligible changes. The configuration of the same distribution relates to similar origin and promotes the identification of the kindred rock groups.

The curves normalized against average chondrite shown in Figs 2 and 3 have been plotted on the basis of the values given in Table II and of the chondrite average by HERMANN (1970). Because of the criticism known from literature (BALASHOV, 1976) as to the applicability of mass spectrometry, certain values had to be corrected to some extent.

With due care it can be stated that:

1. In the granitoids of the Velence Mountains the heavy REE (Gd—Lu) show stronger enrichment, than in the Mecsek Mountains, resp. in this range their curves are a more marked upweard trend.

2. The Eu-anomaly is less expressed in the Mecsek Mountains, than in the igneous Velence Mountains.

3. The curves of aplites with highest SiO_2-content (No. 1 and No. 23) are irregular in the light REE (La—Eu) field for both mountains. Similar irregular behaviour can be observed in the granite strongly kaolinized due to the influence of Eocene andesite activity (No. 21), i.e. where posterior changes took place.

Table I

Chemical composition and trace element content of the rocks

	1	2	6	11	12	14	17	21	22	23
SiO_2	72.39	68.85	52.98	65.87	67.67	67.26	71.17	72.31	72.87	77.75
Al_2O_3	15.68	16.10	16.27	16.18	16.52	16.49	14.72	14.14	14.92	12.62
Fe_2O_3	0.79	1.09	1.43	0.36	0.50	0.23	0.35	2.25	0.17	0.08
FeO	0.39	1.22	4.34	2.28	1.75	1.80	1.39	0.17	1.06	0.13
MgO	0.39	1.05	4.95	1.45	1.39	0.86	0.46	0.42	0.72	0.26
CaO	1.28	2.57	3.86	3.58	2.56	1.92	1.28	0.36	0.55	0.55
MnO	0.01	0.03	0.04	0.02	0.01	0.01	0.03	0.01	0.01	—
Na_2O	2.75	2.68	4.05	3.27	3.24	4.25	4.12	2.02	2.52	2.95
K_2O	4.30	3.75	6.35	3.88	4.17	4.96	4.85	5.27	5.55	5.40
$+H_2O$	1.21	1.60	1.93	1.18	1.23	1.54	0.84	1.74	1.45	0.57
$-H_2O$	0.59	0.20	0.29	0.12	0.15	0.30	0.19	0.37	0.19	0.08
CO_2	0.09	0.05	1.30	1.36	0.04	0.27	0.59	1.27	0.12	0.17
Σ:	99.97	99.68	99.69	99.72	99.62	99.96	100.11	100.39	100.14	100.56
B	23	25	35	25	26	32	18	14	49	32
Sc	18	9.3	6	1	5.2	19	1.9	28	47	9.3
V	7	9	10	7	12	5	5	5	7	5
Cr	3	8	84	5	19	3	3	2	3	2
Co	10	17	7	10	12	12	11	7	14	11
Ni	8	10	7	8	10	7	7	6	8	5
Cu	20	23	24	14	20	16	16	15	16	17
Zn	8	10	15	8	10	8	10	8	9	8
Ga	3	5	5	3	7	4	5	2	5	2
Ge	2	5	2	3	2	2	2	2	1	1
Sr	11	69	10	64	17	3	14	6	17	6
Zr	160	89	170	160	180	140	160	110	160	230
Nb	4	5	2	8	14	5	8	4	6	3
Mo	2	5	2	4	7	2	2	2	1	2
Ag	—	—	—	—	—	—	—	—	—	—
Cd	8	9	—	9	—	8	8	—	8	—
In	2	4	—	2	5	2	—	—	—	—
Sn	—	14	10	10	11	10	—	—	15	—
Ba	150	280	130	170	240	125	120	110	170	73
Ta	10	13	10	10	13	11	10	10	10	10
W	12	30	7	42	19	5	28	20	42	50
Pb	4	25	20	40	30	25	71	39	24	93

Key and sample location

Mecsek Mountains

1. Granite aplite. Kismórágy.
2. Schistose, biotitic granodiorite. Kismórágy.
6. Biotitic amphibolite. Erdősmecske.
11. Red porphyroblastic microcline-bearing biotitic granodiorite. Erdősmecske.
12. Colourless, biotitic, amphibolic, porphyroblastic microcline granodiorite. Erdősmecske.

Velence Mountains

14. Grey biotite granite. Pátka.
17. Grey biotite granite. Szűzvár.
21. Kaolinized granite. Nadap.
22. Red feldspar-bearing biotite granite. Exposure at Motorway M.7
23. Aplite. Székesfehérvár.

Table II

Rare earth element

	1	2	6	s 11	12
Y	22	9.3	11	11	14
La	42	12	28	13	26
Ce	71	18	41	28	50
Pr	9.5*	2.5	5.5	4	5.9
Nd	49	8.5	28	13	25
Sm	5.5	2.2	7.1	4.3*	5.1
Eu	1.8	<0.12	0.41	0.65	0.32
Gd	20	1.4	2.5	2.6	2.8
Tb	3.3*	0.23*	0.47	0.43*	0.38
Dy	7.3	<0.43	<0.3	1.2	1.4
Ho	<0.3	<0.3	<0.25	0.53	0.33*
Er	<0.4	<0.4	<0.4	0.78	0.88
Tm	<0.3	<0.4	<0.3	0.12*	0.14*
Yb	<0.9	<1.2	<0.9	2	2
Lu	<0.22	<0.3	<0.22	0.5	0.32
$\frac{\text{La—Eu}}{\text{Gd—Lu Y}}$	3.4	3.9	7.9	3.3	5.0

* Corrected values.

4. The normalized curves of the schistose granodiorite (No. 2) and of the red porphyroblastic microclinic biotitic granodiorite (No. 11) of the Mecsek Mountains show greatest similarity to those of the genuine igneous granites of the Velence Mountains (Nos 14, 17, 22). The same holds true for the results of trace mineral investigations, too.

5. The irregular behaviour of Sm in the aplite and kaolinic granite (No. 23, resp. No. 21) of the Velence Mountains should be specially pointed out. It is interesting and is demonstrated in the tables, too, that Y reaches its maxima in these two cases.

6. It may be characteristic of the initial material of granitization that the normalized curve of amphibolite (No. 6) occurring in the granitoid series runs parallel with that of the granitoids.

The regularities of the *total REE content* of the two mountains also provide useful information.

1. The total REE contents of granitoids can be assigned to the smaller values given in the references cited. Average granitic values have been found

contents of the rocks

14	17	21	22	23	Average composition in chondrites (Hermann) 1970
29	14	99	16.8	180	1.96
20	7.4	21	7	31	0.32
36	12	46	13.3	34	0.94
4.6	1.7*	6.6*	1.8	8.1	0.12
20	6.2	37	6.7	25	0.60
4*	1.4	21	2.1	18	0.20
0.29	0.18	0.7	0.1	0.3	0.073
1.6	1.7*	2.3	1.1	2.2	0.31
0.82	0.28	2	0.45	0.4	0.050
<0.43	0.9*	8.3	1.7	<0.3	0.31
<0.33	0.23	1.8*	0.3*	<0.25	0.073
<0.5	0.43	9.3	1.6	<0.4	0.21
<0.42	0.07*	0.25*	0.25*	<0.3	0.033
<1.2	5.4	8	4.4	<0.9	0.19
<0.3	1.2	1.9	0.7	<0.22	0.031
2.9	1.2	1.0	1.1	0.6	

for aplites (No. 1 and No. 23), for the Mecsek Mountains amphibolite (No. 6) and for the related colourless porphyroblastic, microclinic amphibolic granodiorite (No. 12), as well as for the altered kaolinized granite of the Velence Mountains (No. 21).

2. The average Y-content of the granitoids of the Velence Mountains is five times that of the Mecsek Mountains.

3. This is confirmed by the light and heavy REE ratios (La—Eu/Gd—Lu, Y), though in certain cases only estimated values could be used owing to uncertainties of mass spectrometric measurements (Table II). This ratio is more or less characteristic of differentiation, i.e. parallel with its progression the light REE get relatively enriched as compared to the heavy ones. In our case, i.e. there is a sharp difference between the Mecsek and Velence Mountains in the two granitoid rocks of different genesis. In the Mecsek Mountains values close to the granitic average are found in the less differentiated anatectitic granitoids, while in the Velence Mountains surprisingly low values were obtained which are characteristic rather of the basic rocks. This is caused by the obvious Y abundance of the Velence Mountains, which made the ratio

similar to that of the less differentiated materials. This produces an extreme value, i.e. 0.6, for aplite. This phenomenon is in good correlation with the xenotime bearing and Y-rich pattern observed in the Velence Mountains.

The *correlation analyses* for which the values given in Tables I and II for the two mountains have been used both separately and jointly are unsuitable for drawing definite conclusions due to the small number of samples. It can be stated, however, that in the Mecsek Mountains Y shows a marked positive correlation both with the light and the heavy REE, while in the Velence

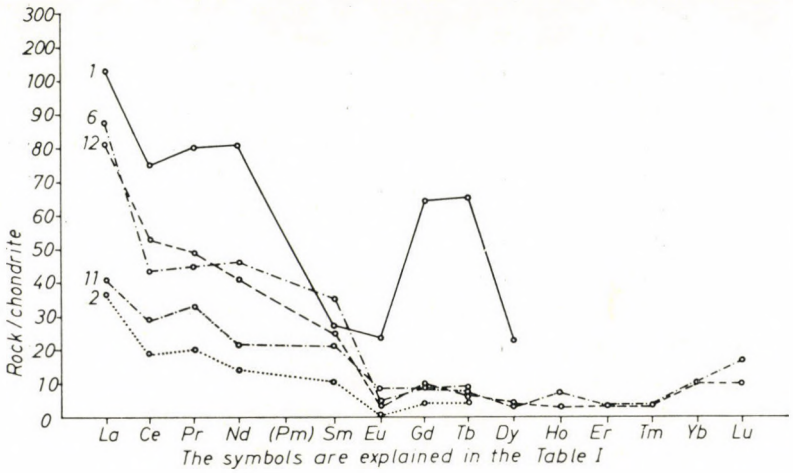

Fig. 2

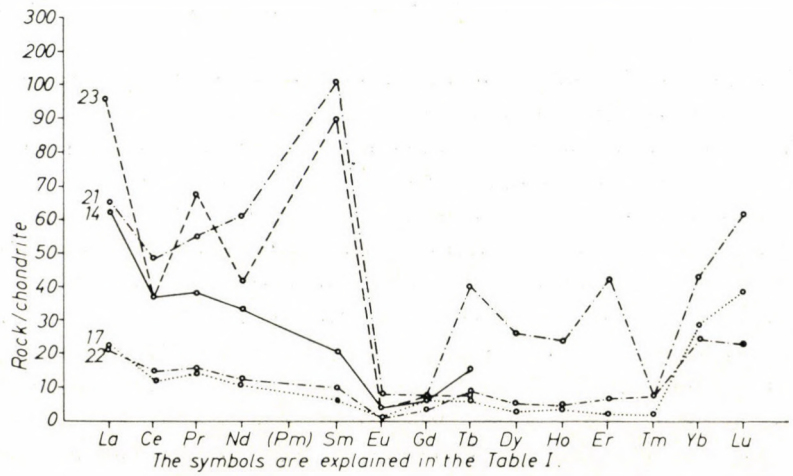

Fig. 3

Mountains it does so only with Sm. This also supports the significant and frequently independent role of Y in the granitoids of the Velence Mountains.

Summarizing the facts discussed above, the conclusion can be drawn that if the role of both the accessory minerals and of the REE is considered, there is a genetic difference between the two mountains. This bears witness to an anatectic and partly potash-metasomatic origin of the granitoids in the Mecsek and to a virtually igneous one in the case of the Velence Mountains.

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ГЕНЕТИЧЕСКОЕ ЗНАЧЕНИЕ РЕДКОЗЕМЕЛЬНЫХ ЭЛЕМЕНТОВ В ГРАНИТОИДНЫХ ПОРОДАХ ВЕНГРИИ

ДЬ. ПАНТО

Резюме

В Венгрии существуют две области появления на поверхности гранитоидных горных пород различного генетического типа. Различия условий генезиса анатектических гранитоидов (в горах Мечек) и магматических гранитоидов (в горах Веленце) установлены между прочим и по исследованию их акцессорных минералов и по распределению РЗЭ. Породы гор Мечек характеризуются первичными и вторичными формами проявления монацита, а также присутствием (преобладанием) минералов из группы бастнезита. Гранитоиды гор Вегенце являются продуктами развитой этапы магматической дифференцияции, они содержают относительно много ксенотима. Изменения состава РЗЭ в монаците, в бастнезите и в минералах циркония тоже параллельны с процессами магматической эволюции. В горах Веленце установлено более выраженное обогащение тяжелых РЗЭ. Различия отмечаются в нормализованном распределении РЗЭ и изьясняются различным генезисом. Отношения легких и тяжелых РЗЭ гранитоидных пород в двух областях оказались характерно различными.

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OLIGO- AND MICROELEMENTS IN SOME MESOZOIC AND CAINOZOIC ROCK SAMPLES OF HUNGARY

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Some geochemical characteristics of altogether 60 sedimentary rock samples of different age (from Carboniferous to Early Pannonian) deriving from boreholes drilled on the Great Hungarian Plain and in Transdanubia, as well as from shallow boreholes and outcrops of the Hungarian Central Mountains were investigated (Tables I—VII). Based on the synoptic evaluation of X-ray diffractometric, rock-chemical and emission UV spectrographic trace element data binary correlations have been established between some mineralogic-petrographic parameters and several trace elements as well as between some trace elements by pairs.

The positive correlations of the strontium content with the carbonate (preferentially calcite) percentage, of the boron and potassium contents with the sum of phyllosilicates, of the sodium and gallium content with the total feldspar contents and of the titanium content with the total percentage of terrigenous constituents could be stated.

The correlations between trace elements provided the possibility to probabilize different source areas for the Lower Pannonian. Different probable directions of transportation of the detrital materials can also be suspected.

Suggestions are made how to use the different geochemical indices and which of them are suitable for further geochemical studies in the given area.

Introduction

Sixty sedimentary rock samples were investigated from the geochemical point of view. The detailed study of these samples was made possible by the courtesy of the Hungarian Oil and Gas Trust.

The samples derive from different areas. Their areal distribution is as follows: 41 samples from the Great Hungarian Plain, all of them from boreholes; 9 samples from the Hungarian Central Mountains, some of them are surficial samples; and 10 samples from the S-Transdanubian Basin, also from boreholes.

The stratigraphic distribution of the samples (based essentially on the data of OGIL, i.e. National Gas-Industrial Laboratory) is as follows: 1 sample of Carboniferous age; 11 samples from the Permotriassic and Triassic, 9 samples from the Jurassic, 7 samples from the Cretaceous, 10 samples from the Oligocene, 5 samples from the Miocene and 17 samples of Lower Pannonian age.

The age of some of the samples is uncertain, this fact will be marked in the tables. Having investigated these samples some of them could be assigned with more probability to a given stratigraphic position.

Table I

Petrographically corrected list of the rock samples

No. of sample	Rock name	No. of sample	Rock name
1.	Shale, sandy and dolomitic-limy	32.	Siltite, limy and clayey
2.	Shale, silty	33.	Siltite, limy and clayey
3.	Claystone, silty and dolomitic	34.	Siltite, limy and clayey
4.	Dolomarl schist, silty	35.	Siltite, dolomitic and clayey
5.	Dolomite-breccia, (silty)	36.	Siltite, limy and clayey
6.	Dolomarl	37.	Siltite, dolomitic and clayey
7.	Dolomite-breccia	38.	Clay-marl, silty
8.	Clayey-limestone, dolomitic	39.	Marl, silty and dolomitic
9.	Dolomarl, limy	40.	Dolomarl, silty
10.	Limy-marl, dolomitic	41.	Marl, dolomitic and silty
11.	Dolomite conglomerate	42.	Marl, silty, slightly dolomitic
12.	Dolomite	43.	Siltite, limy and clayey, slightly dolomitic
13.	Clay-marl, silty and dolomitic		
14.	Marl, silty and dolomitic	44.	Siltite, sandy and clayey
15.	Clay-marl-schist, silty	45.	Siltite, sandy and clayey
16.	Marl, silty	46.	Clay-marl, silty, slightly dolomitic
17.	Marl, silty and dolomitic	47.	Siltite, clayey and limy-dolomitic
18.	Marl, silty and dolomitic	48.	Sandstone, clayey, dolomitic-limy
19.	Marl, silty	49.	Clay-marl, silty-sandy and dolomitic
20.	Limy-marl, silty and dolomitic	50.	Clay-marl, silty
21.	Dolomite- and limestone breccia	51.	Clay-marl, silty, slightly dolomitic
22.	Clay-marl, silty	52.	Clay-marl, silty, dolomitic
23.	Limestone-breccia	53.	Clay-marl, silty, dolomitic
24.	Limestone	54.	Marl, silty (dolomitic)
25.	Limestone	55.	Marl, silty, slightly dolomitic
26.	Limestone (brecciated)	56.	Limy-marl, dolomitic
27.	Marl, sandy and dolomitic	57.	Clayey-limestone
28.	Limy-marl, silty and dolomitic	58.	Clayey-limestone
29.	Dolomarl, sandy	59.	Clayey-limestone, dolomitic
30.	Siltite, clayey	60.	Limestone, clayey, slightly dolomitic
31.	Siltite, limy and clayey		

The number of samples proved to be too small to draw exact conclusions to the whole area in question. Nevertheless, certain trends as well as in some cases the directions of transportation and source areas could be fairly stated.

Methods of study

Wet silicate analyses and flame photometric analyses (performed by B. FÜZESY) were carried out only on a few samples. Data of the silicate analyses and flame photometry (Na_2O and K_2O) are demonstrated in Tables II, IV and VI.

To determine the semi-quantitative mineral composition of the samples the X-ray diffraction method was used. For the records and evaluations

Table II

Geological setting, mineral composition and alkali contents of the samples older than Cainozoic

Sample No./location	Depth (m)	Age	1	2	3	4	5	Na_2O	K_2O %
1. Dar-1/18	2219—2220	C_2—P	2	16	36	21	25	1.88	3.12
2. Csa-2/28	2347—2350	P—T_1	4	—	67	—	23	0.16	4.96
3. Makó-2/27	4924—4929	T_1	6	10	33	13	33	0.26	5.00
4. Makó-2/26	4898—4899	T_1	1	55	15	15	10		
5. Csa-5/10	1950—1968	T	—	82	7	—	11		
6. Csa-6/7	2184—2190	T	6	77	16	—	—		
7. Csa-6/9	2254—2259	T	—	96	4	—	—		
8. Kasz-D-7/12	1865—1881	T_3	78	7	9	—	5		
9. Szil-35/42	3319—3329	T_3	10	76	4	1	8	0.26	0.54
10. T-28/6	1892—1907	T?	87	7	3	—	3		
11. Bm-I/8	2952—2961	T?	—	98	—	—	2		
12. Kasz-D-3/11	1669—1687	T	3	93	—	—	—		
13. Psz-8/12	2138—2140	J_3	34	2	34	9	21	0.96	1.88
14. Psz-2/14	1780—1783	J_3	36	5	28	14	15		
15. Pf-129/6	1783—1800	J?	34	7	24	10	25	0.98	2.00
16. Med-2/3	2700—2717	J_3?	52	—	28	7	13	0.88	1.84
17. Psz-6/17	1828—1831	J	51	3	20	3	22		
18. Psz-2/21	2087—2091	J_3	57	3	24	7	9		
19. Psz-30/4	1782—1786	J_3?	66	2	15	3	12	0.44	0.84
20. TK-2/10	1659—1670	J_3	65	4	12	4	14	0.54	0.84
21. Psz-14/9	1743—1744	J_3?	41	50	7	—	2		
22. Sark-I/12	3731—3738	K_1	40	2	41	1	13	0.40	2.32
23. Sark-I/13	4034—4043	K_1	98	—	2	—	—		
24. Sark-I/9	3461—3470	K_1	98	—	2	—	—		
25. Sark-I/7	3431—3440	K_1	99	—	1	—	—		
26. Sark-I/11	3604—3613	K_1	99	—	1	—	—		
27. Felgyő-I/9	2790—2795	K_2	47	11	19	6	1	0.36	2.18
28. Szil-33/27	3350—3360	K_2	59	17	8	6	9		

1. Calcite
2. Dolomite
3. Phyllosilicate
4. Total feldspar
5. Quartz

Table III

Micro-elements of the samples older than Cainozoic (ppm)

Sample	B	Co	Cr	Mn	Ni	Sr	Ti	V	Ga
1.	110	6	145	390	29	55	>7000	29	20
2.	165	10	70	30	11	—	900	18	16
3.	120	10	150	30	13	15	150	18	9
4.	120	7	240	40	16	755	40	12	15
5.	36	5	3	25	29	175	20	8	18
6.	75	6	6	30	20	190	35	11	18
7.	24	6	3	—	30	140	110	10	20
8.	60	10	5	30	26	1350	20	13	24
9.	55	6	5	25	7	1200	1900	10	5
10.	30	8	4	25	34	>2000	20	13	28
11.	8	10	6	—	14	1300	—	20	5
12.	15	9	4	—	17	810	—	13	26
13.	175	10	140	260	19	230	320	21	24
14.	155	11	46	200	13	240	285	18	19
15.	170	16	57	550	16	1600	3150	36	18
16.	180	15	112	400	22	1175	1150	25	24
17.	180	12	115	240	15	890	700	26	17
18.	138	8	115	175	27	800	50	13	24
19.	160	33	73	240	15	830	1950	37	28
20.	100	7	210	120	24	1100	45	9	20
21.	63	6	4	50	18	1300	85	15	18
22.	190	11	130	195	17	715	280	20	22
23.	12	7	3	30	21	1200	—	14	15
24.	13	7	3	—	13	1300	—	15	12
25.	11	7	3	—	26	1160	—	13	24
26.	14	9	3	—	22	490	—	13	24
27.	160	6	23	60	21	85	1400	18	6
28.	150	10	35	70	61	1850	1900	44	8

Detection limits (in ppm): B: 2—1000, Co: 3—1000, Cr: 2—1000, Mn: 25—2500, Ni: 5— 500, Sr: 6—2000, Ti: 20—7000, V: 4—1000, Ga: 2—1000

Reference standards: Rock Standard Samples KH, GM, TB; Geologische Erkundung, Staatssekretariat für Geologie, Berlin.

Table IV

Geological setting, mineral composition and alkali contents of the Oligocene and Miocene samples

Sample No./location	Depth (m)	Age	1	2	3	4	5	Na_2O %	K_2O %
29. Fv-2/8	2300—2307	Pg ?	2	34	21	14	26		
30. R-8/7	83.4—90.9	0	4	7	35	25	21	0.80	2.86
31. Csillaghegy	—	0	8	11	35	11	27		
32. K-4	—	0	16	5	34	24	19		
33. Be-4	155—158	0	18	7	26	27	19		
34. R-8/7	77.0	0	15	7	28	22	20		
35. Rm-103	717.	0	28	1	16	26	25	1.34	1.92
36. Iliny-1	54.5—56.6	0	14	16	27	21	19		
37. Rm-103	1035	0	—	30	27	13	27	1.00	2.58
38. R-8/3	41.0	0	57	—	23	6	12		
39. L-1/19	3354—3369	M_2	30	14	9	12	34	1.10	1.40
40. Fv-4/5	2279—2282	M_2?	1	40	20	10	25		
41. Csapod-1/13	3180—3181	M_2	46	—	15	14	23		
42. Szil-35/6	2631—2639	M_2	38	10	12	11	28		
43. Csapod-1/11	2882—2884	M_3	18	10	27	15	27		

Table V

Micro-elements of the Oligocene and Miocene samples (ppm)

Sample	B	Co	Cr	Mn	Ni	Sr	Ti	V	Ga
29.	160	11	85	55	84	20	4000	46	9
30.	130	13	70	35	7	130	2700	25	25
31.	145	56	150	280	27	200	3600	98	40
32.	160	21	90	110	16	240	3900	30	35
33.	140	19	55	140	9	130	5200	23	20
34.	135	12	40	55	16	70	2700	36	20
35.	110	20	65	40	21	120	2100	33	13
36.	110	14	190	50	13	300	2500	50	35
37.	150	20	60	65	16	45	3000	28	11
38.	160	22	80	220	20	350	3500	13	9
39.	170	9	130	75	76	930	3500	39	14
40.	180	8	130	50	76	20	1500	49	9
41.	235	10	70	120	57	550	2300	39	8
42.	160	8	75	70	32	>2000	1500	26	6
43.	185	10	85	350	146	55	2700	27	15

Table VI

Geological setting, mineral composition and alkali contents of the Lower Pannonian samples

Sample No./location	Depth (m)	1	2	3	4	5	Na_2O %	K_2O %
44. Komádi-4/2	1913—1930	4	—	30	16	50		
45. Komádi-8/2	1849—1851	4	3	36	17	40	1.36	2.50
46. En-8/9	2469—2487	14	13	34	15	24		
47. Felgyő-I/6	2445—2463	16	11	32	11	30	0.86	2.60
48. Felgyő-I/4	2100—2114	10	20	12	10	48		
49. Or-Ny-1/1	1237—1240	15	16	21	17	32		
50. Komádi-8/5	1978—1984	23	2	34	11	23		
51. Felgyő-I/5	2282—2300	19	14	31	9	27		
52. En-18/4	2585—2601	21	15	24	11	29		
53. Or-Ny-1/7	2314—2315	40	15	18	5	21		
54. Csa-4/3	1880—1882	56	—	14	9	20		
55. En-7/12	2637—2655	49	7	24	7	12		
56. En-6	2668—2680	50	30	11	3	5	0.72	0.90
57. Csa-4/9	1940—1946	83	—	7	2	7		
58. Csa-4/6	1904—1920	83	4	7	—	6		
59. Felgyő-I/7	2724—2732	77	10	5	1	6		
60. Csa-4/8	1922—1940	89	—	5	—	6		

the authors are indebted to P. Árkai and M. Tóth. The X-ray recording conditions are as follows:

Records were made with the diffractometer of Philips Müller Micro 111 type. Parametres of the device are:

radiation: CuK_α
acceleration voltage: 42 kV
amperage: 32 mA
detector: proportional counter
monochromator: graphite
divergency and detector aperture: 1°
goniometer speed: 2°/min
paper speed: 1600 mm/h
time constant: 1.

Records were made from total rock samples. In the case of high calcite or dolomite content the records were made from the acid insoluble residue obtained after the treatment with hydrochloric acid of 3%. The semi-quantitative determination of mineral composition was carried out in desoriented records. In the case of qualitative determination some oriented and ethylene-glycol-treated records were also evaluated.

The semi-quantitative informative phase analysis was based on the direct method of NÁRAY-SZABÓ — PÉTER as well as on the method of BÁRDOSSY (1966, 1968) and, of course, the results of wet silicate analyses.

Table VII

Micro-elements of the Lower Pannonian samples (ppm)

Sample	B	Co	Cr	Mn	Ni	Sr	Ti	V	Ga
44.	70	25	55	60	15	30	940	15	22
45.	75	27	60	80	15	—	800	14	25
46.	80	14	85	60	15	50	240	15	8
47.	210	9	95	80	74	160	3050	37	15
48.	130	8	110	70	34	650	>7000	30	20
49.	160	8	45	55	45	120	1600	30	9
50.	70	16	100	40	12	40	430	15	10
51.	185	10	85	160	66	40	2850	32	16
52.	90	11	120	60	16	70	80	13	6
53.	115	10	30	75	61	700	3000	44	8
54.	155	17	75	280	100	1650	2700	65	16
55.	90	13	75	60	24	70	125	15	—
56.	70	12	6	45	54	140	50	10	—
57.	60	8	23	70	34	1850	1350	27	6
58.	55	7	16	55	22	1900	1000	28	5
59.	45	6	6	90	15	1750	130	15	4
60.	16	6	6	30	13	1800	130	21	4

The trace element determinations were made from UV emission spectrographic records, the results of which are shown in Tables III, V and VII. The recording conditions were as follows:

Q–24 UV spectrograph; plate: 23D56; split: 10 microns; blende: 3.2; amperage: 6 A; exp. time: 102 sec.; collimator: 2.75; electrodes: Al 9999; electrode distance: 2 mm.

In favour of exact denomination of rock types and to avoid the misconceptions the petrographically corrected names of each sample are given in Table I.

It is to be noted that if the number of the given set (subset) allowed, the relationships were considered separately for the stratigraphic units, as well.

The relevant parametres were selected and those which turned out to be irrelevant in this particular case were omitted. E.g. in Tables II, IV and VI

the occasional siderite contents are not demonstrated. Further, trace element analyses involved 16 elements but only those are mentioned which proved to be of interest for our study.

Two kinds of graphic examination were used. The first is to demonstrate the correlation of the individual chemical elements as a function of the main

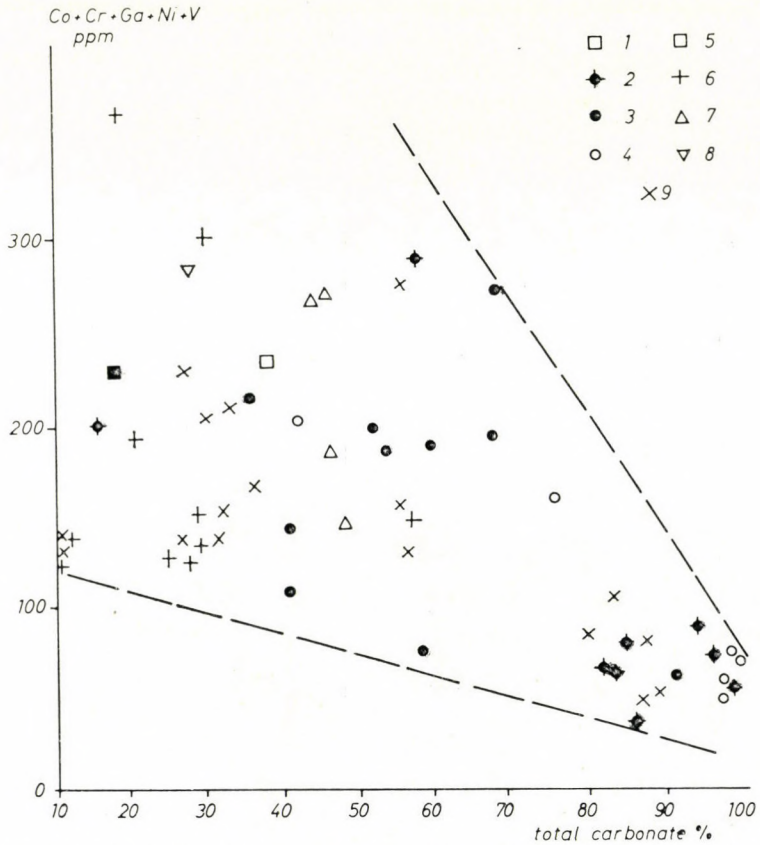

Fig. 1. Co + Cr + Ga + Ni + V versus total carbonate percentage. Legend: 1. Carboniferous; 2. Permotriassic and Triassic; 3. Jurassic; 4. Cretaceous; 5. Paleogene; 6. Oligocene; 7. Tortonian; 8. Sarmatian; 9. Lower Pannonian

rock constituents. This type of correlation is seen in Figs 1—7. The graphic evaluation of the trace elements by pairs is shown in Figs 8—14.

In the tables the rock samples are grouped by ages, i.e. Table II and III contain the geological setting, mineral composition and alkali percentage as well as the minor element contents of the samples older than Cainozoic, Tables IV and V contain the same parametres of the Oligocene and Miocene samples, while Tables VI and VII show those of the Lower Pannonian samples.

Discussion

Having investigated the element abundances as a function of rock composition the following proved to be of general validity.

The combined Co + Cr + Ga + Ni + V content decreases with increasing carbonate content (Fig. 1). The pure limestones and dolomites are of rather low trace element content as compared to the rocks of medium carbonate contents. First of all the change of the chromium content is responsible for this though the other elements show the same tendency.

As compared to the former relationship, the change of the Sr-content as a function of total carbonate content shows an opposite trend, i.e. it posivitely correlates with the carbonate content (Fig. 2). The correlation is preferential

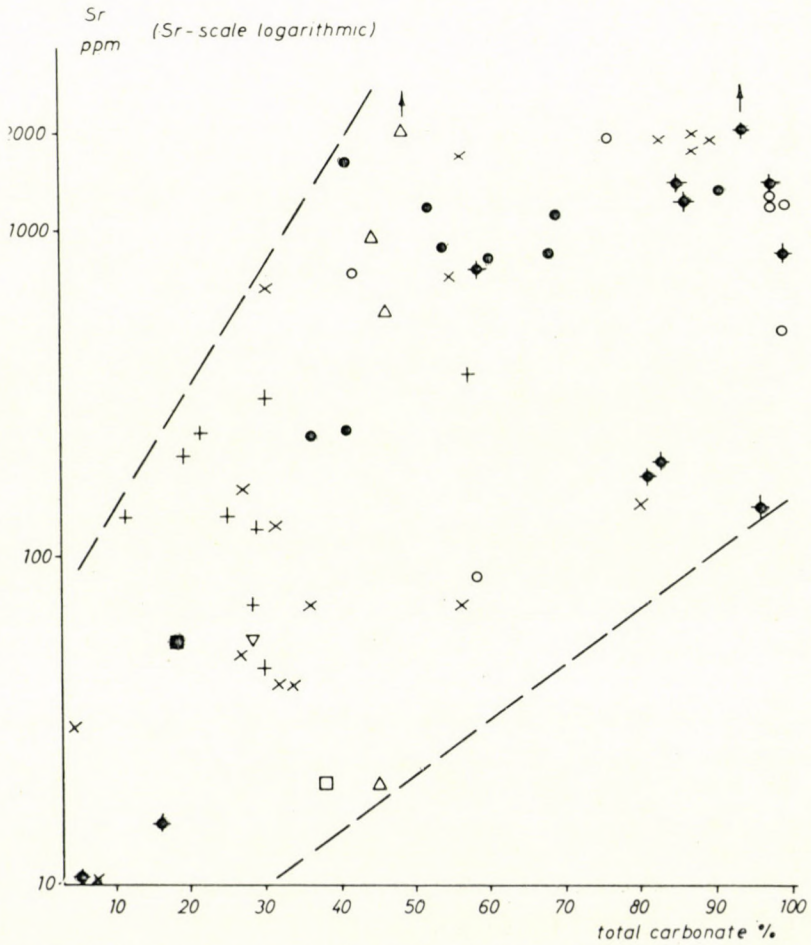

Fig. 2. Sr-content versus total carbonate (calcite + dolomite) percentage

with calcite, i.e. in dolomites and dolomite breccias the Sr-content is rather low. It is worth of mention that the Sr-content reaches the maximum value (more than 2000 ppm) in the organogenic limestones.

The boron content shows a steady positive correlation with the quantity of phyllosilicates (Fig. 3). As a matter of fact, there are different directions

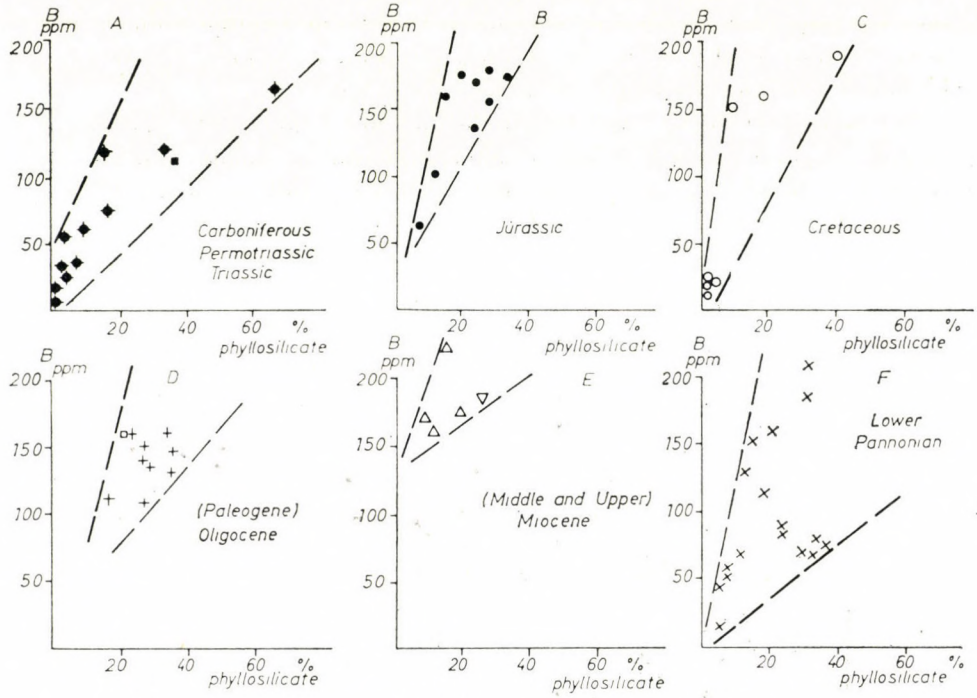

Fig. 3. B-content versus total clay mineral (phyllosilicate) percentage

but the trend is the same in case of all the samples of different ages. Similar correlation was found between the K_2O-content and the quantity of phyllosilicates (Fig. 4). The only exception is, however, characterized by high potash-feldspar content, thus the deviation of this sample is less astonishing.

The sodium content co-variates with the total feldspar content (Fig. 5). The only exception shows rather high potash-feldspar content (see above) thus the deviation can be readily explained by this fact. Similarly, in the Lower Pannonian samples Ga shows positive correlation with the total feldspar content (Fig. 6).

Titanium is concomitant with the total percentage of terrigenous constituents (Fig. 7). When distinguishing the different areas (lying rather far from one another) the same tendency can be determined only the rise of the given curves differs. This phenomenon is suitable to make distinctions between

the areas where different source areas, resp. different directions of transportation can be assumed.

No figure is presented showing the distribution of Mn in the rock samples studied. It is to be emphasized, however, that the Jurassic samples have systematically higher Mn-contents than any other group of samples (this can

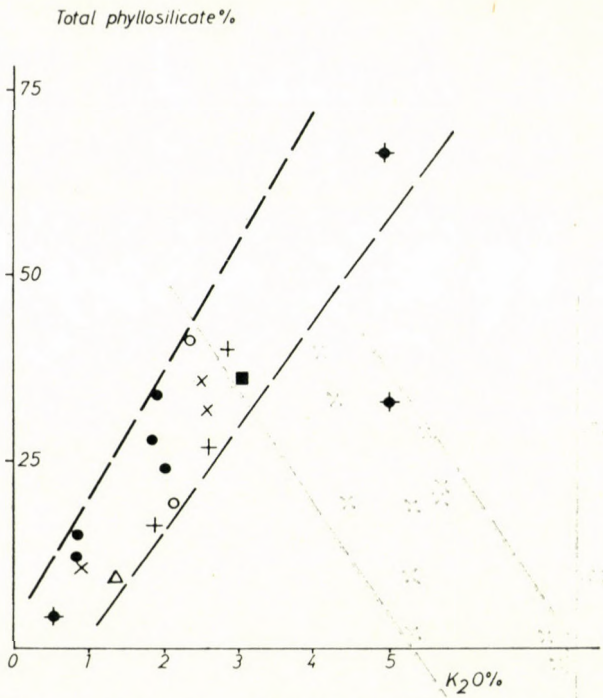

Fig. 4. K_2O percentage versus total clay mineral (phyllosilicate) percentage

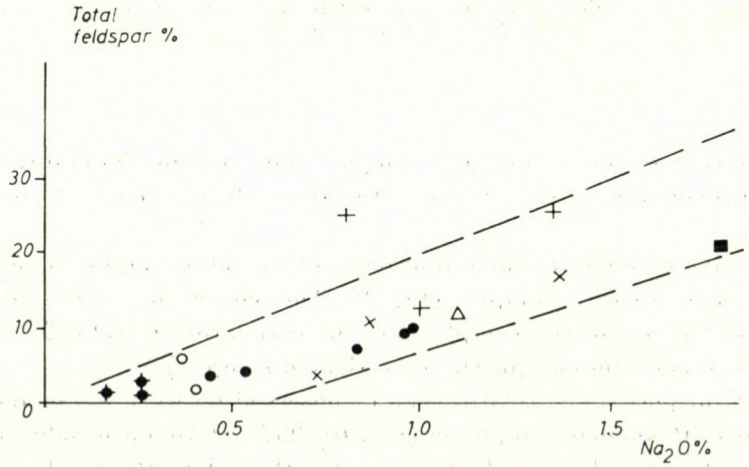

Fig. 5. Na_2O percentage versus total feldspar percentage

be read also from the tables). This high Mn-content is in perfect accordance with the common geological features of the Hungarian Jurassic sediments.

The binary correlations between trace elements have provided also some relevant information. In the Mesozoic sample set the chromium is negatively correlated with titanium (Fig. 8). Only the sample of Carboniferous age falls

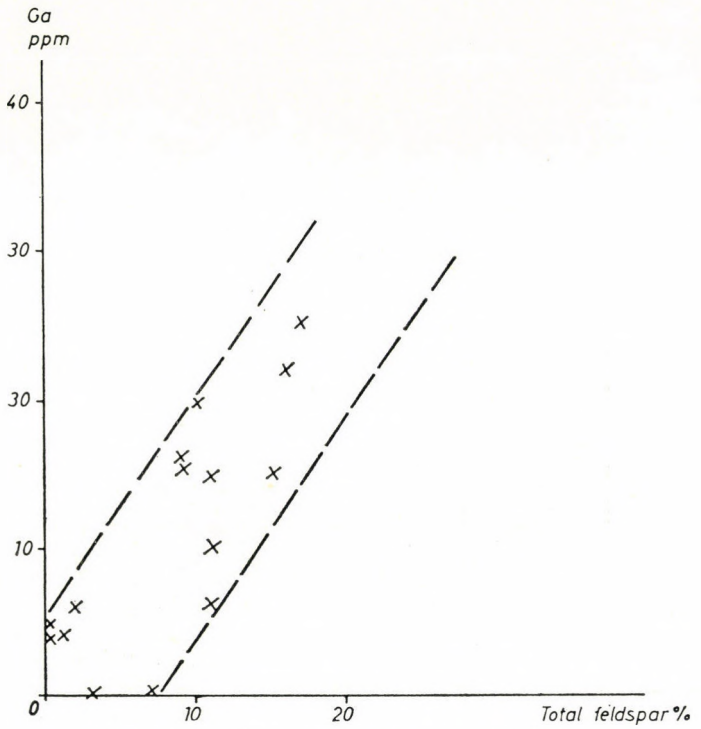

Fig. 6. Positive correlation of the Ga-content with the total feldspar percentage for the Lower Pannonian samples

out of the field of correlation. Ten samples show very low Ti and Cr contents, these, however, are pure carbonates the trace element contents of which is very low.

Positive correlation was found between vanadium and chromium in the Oligocene and Miocene samples (Fig. 9). This correlation seems to be independent of the age of the samples, i.e. the correlation is valid for the whole of Oligocene and Miocene (in the investigated samples).

A slight positive correlation can be observed between boron and gallium in the Lower Pannonian sample set (Fig. 10). This relation is usually considered to be a salinity indicator. Unfortunately, the relatively small number of

samples does not allow to draw reliable conclusions on the degree of salinity of the medium of sedimentation.

Similarly to the Mesozoic sample set, there is a rather complicated correlation between chromium and titanium also in the Lower Pannonian samples (Fig. 11). When considering only the data, no correlation is seen in the figure. Nevertheless, in possession of the data the trends can be determined for each area. Accordingly, there is correlation between Ti and Cr, but the rise of the correlation curve is characteristic of the individual areas. In the figure the lines indicate the regions in question. It is to be noted that the region

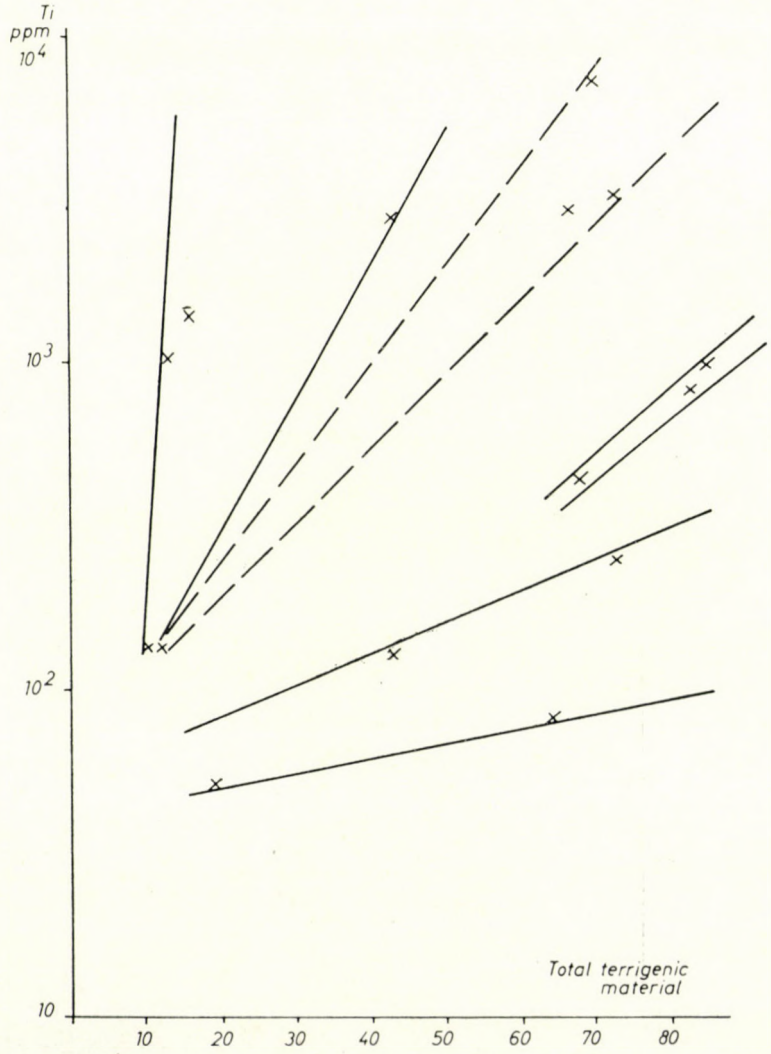

Fig. 7. Positive correlation of Ti with the total percentage of terrigenous materials for the Lower Pannonian samples

pairs, i.e. sample sets 1—2, resp. 3—4 lie close to each other, while the two sample groups represent two structurally different regions.

The same phenomenon can be observed in the V vs Cr diagram (Fig. 12). The same areas can be distinguished, i.e. the V/Cr relation confirms the previous statement on the distinction of certain areas.

The peculiarity of the Ni vs Cr relation is even more conspicuous (Fig. 13). In this case not only the rise of the curves but also their signs are different. Positive correlation is found in the Csanádapáca—Felgyő area while negative

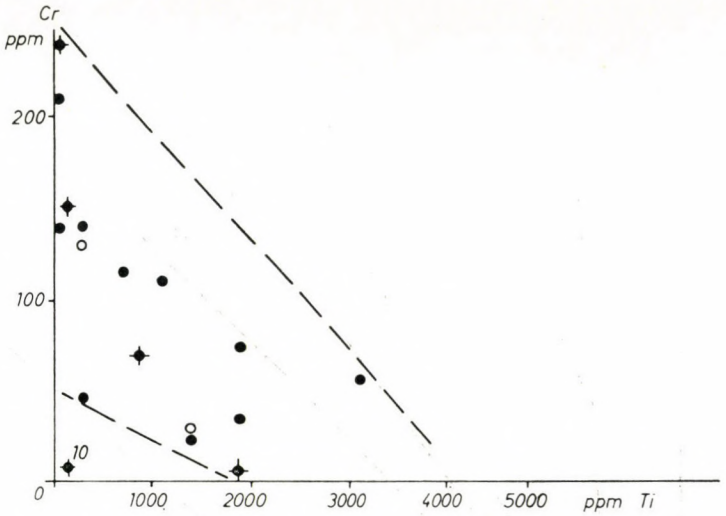

Fig. 8. Cr versus Ti-content for the Mesozoic samples

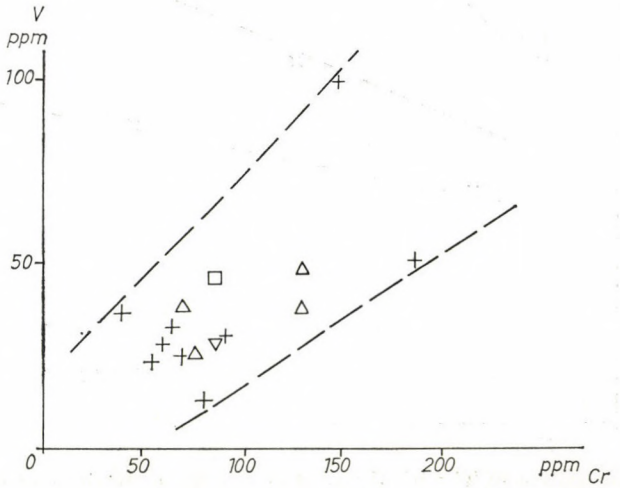

Fig. 9. V versus Cr-content for the Oligocene and Miocene samples

one in the Endrőd—Komádi region. The two trends are nearly perpendicular of each other, thus the two regions should be different in other aspects, as well. This is proved also by the relations between Cr and Ti, resp. V and Cr.

Similar opposite trends can be observed in the diagram of Co vs Ni (Fig. 14). The rather flat rise but opposite trend of the two trends support the facts mentioned above.

Figures 11—14. give rise to some suspicions, i.e. within the Lower Pannonian samples there are two areas at least being of different trace element distribu-

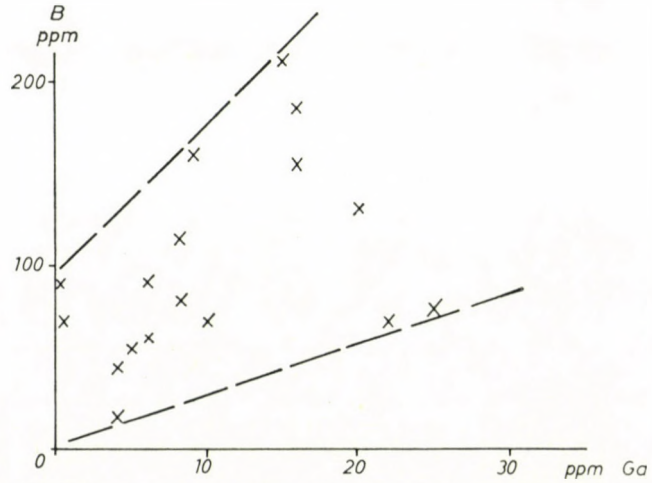

Fig. 10. Positive correlation of B and Ga for the Lower Pannonian samples

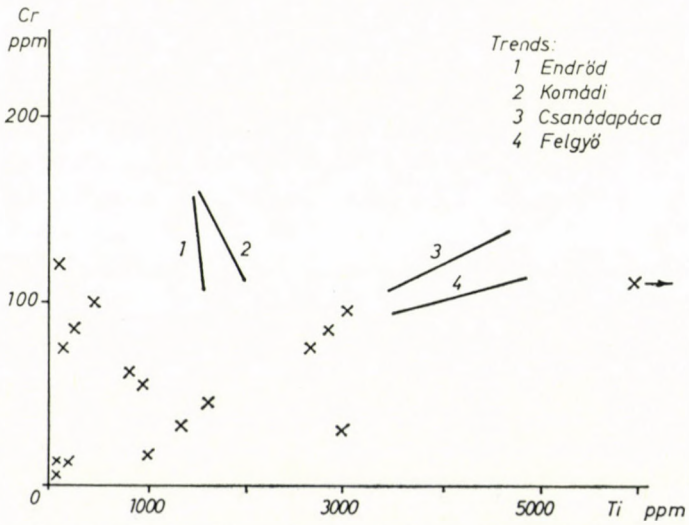

Fig. 11. Cr versus Ti-content for the Lower Pannonian samples

tion. The Endrőd—Komádi area unambiguously differs from the Csanádapáca—Felgyő region and suggests the assumption that the two areas can be characterized by different source area and at the same time by different directions of transportation.

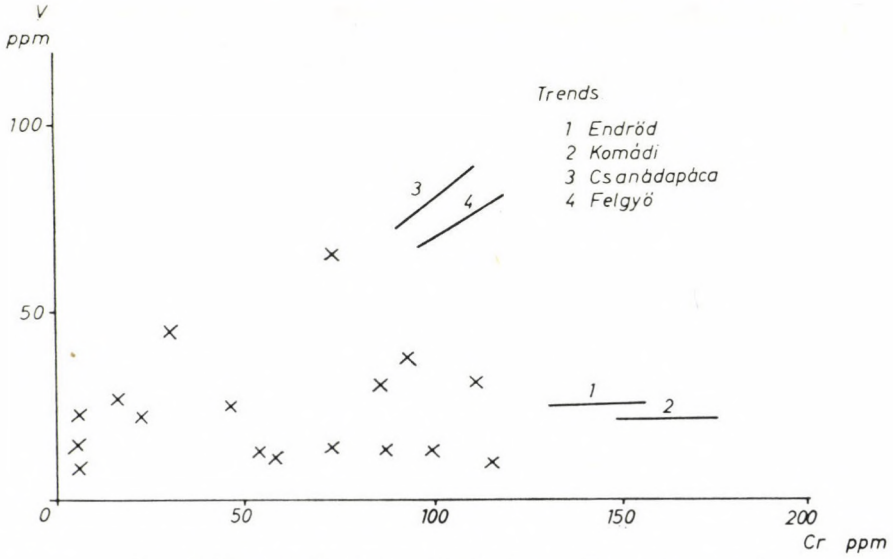

Fig. 12. V versus Cr-content for the Lower Pannonian samples

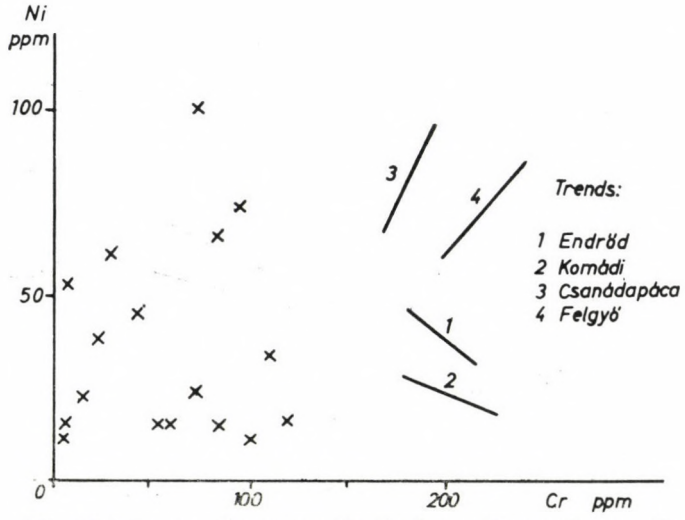

Fig. 13. Ni versus Cr-content for the Lower Pannonian samples

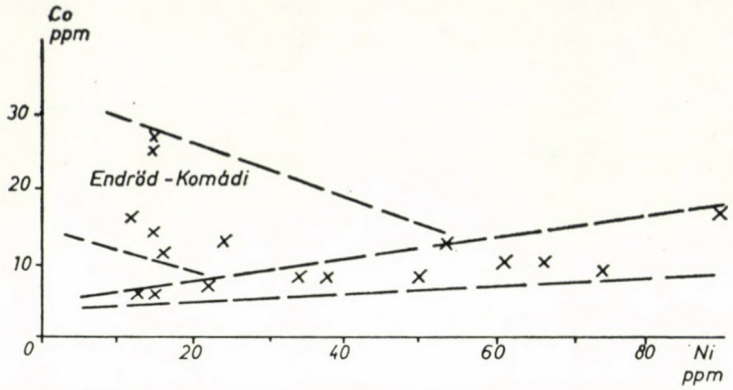

Fig. 14. Co versus Ni-content for the Lower Pannonian samples

Conclusions

Even in the case of a rather restricted number of samples and sampling oriented to completely different purpose from that of microelement geochemistry, some relationships could be determined:

Also in very different sedimentary rocks "carbonatophile", "pelitophile" and "psammitophile" elements can be well distinguished in good accordance with the ionic radii, adsorption potential and solubility data.

There seem to be some differences in case of some elements (e.g. Ti, Mn) in formations of different geological age (concerning only the area of this study!).

Regional differences are possibly due to transportation from source areas differing in their (paleo)geographic position and petrological-geological setting.

As a result of these investigations it is recommended that out of the trace elements only those showing some significant change as a function of some of other geological-petrological parametre are to be investigated, e.g. Cr, Co, V, Ti, Mn, Ni, B, Ga and Sr. In this relation the necessity of K_2O and Na_2O determinations is also emphasized. If possible, it is worthy to perform the determination of the mineral composition since this may also provide significant information regarding both the relationships between trace elements and minerals and the nature of the sedimentary rocks in question.

Acknowledgements

Special acknowledgement is said to the competent authorities of the Hungarian Oil and Gas Trust for the submittal of the samples as well as for the permission to publish the aforementioned results; to the Scientific Director and staff of the Laboratory for Geochemical Research of the Hungarian Academy of Sciences, for his interest and for their active co-operation and consent to incorporate their analytical results.

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МЕЛЫЕ ПРИМЕСИ В НЕКОТОРЫХ ОБРАЗЦАХ ГОРНЫХ ПОРОД МЕЗОЗОЙСКОГО И КАЙНОЗОЙСКОГО ВОЗРАСТА ВЕНГРИИ

Э. ДУДИЧ, О. ТОМШЕЙ

Резюме

Исследовались некоторые геохимические параметры 60 образцов осадочных пород различного возраста (от карбона до раннего плиоцена) взятых из глубоких буровых скважин прибуренных на Большой венгерской нижменности и в Задунайском краю и также из обнажений и скважин Венгерского среднегорья. (Табл. 1—7). Исходя из комплексной интерпретации рентгено-структурных, петрографических и эмиссионно-спектральных анализов были установлены корреляционные связи некоторых минералогическо-петрографических параметров с отдельными элементами-следами и также некоторых микроэлементов по парам.

Установлены положительние связи содержания стронция со содержанием карбонатов (точнее кальцита), содержаний бора и калия с общим коичеством филлосиликатов, содержаний натрия и галлия с общим количеством полевых шпатов, и содержания титания с общим количеством терригенных компонентов.

Корреляционные связи существующие между микроэлементами делают возможным выделить разные области и направления сноса раннеплиоценового осадконакопления. Даются рекоммендации к целесообразному выбору и применению геохимических параметров для дальнейшего геохимического изучения данных областей и образований.

Acta Geologica Academiae Scientiarum Hungaricae, Tomus 21 (1—3), pp. 133—141 (1977)

TECTONIC SIGNIFICANCE OF SOME GEOCHEMICAL DATA ASSOCIATED WITH THE OPHIOLITIC COMPLEXES OF THE DARNÓ MEGATECTONIC LINE, NE- HUNGARY

By

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The original tectonic setting of the mafic and ultramafic rocks of the Darnó megatectonic line are investigated using evidence provided by their major element and trace element contents. The geochemistry of these igneous rocks supports an ocean floor origin for them, and their differentiation and fractionation trends agree very well with those of abyssal tholeiites. The TiO_2—FeO^t/MgO and Ti—Cr relations show that the volcanic rocks of the Darnó-line are not related to island arc magmatism, while the Y/Nb ratios confirm that they are not generated by intra-plate volcanism. The implications of these results are discussed in light of the plate tectonic evolution of the Pannonian region.

Introduction

With the advent of plate tectonics in the last decade, studies of ophiolite complexes (radiolarites — pillow basalts — dolorite dyke swarms — diabase — gabbros — ultramafics) have been intensified in many parts of the world. The magmatic members of present-day ophiolite suites have a significant part to play in reconstructing plate movements in the past. It has been shown (JAKES and WHITE, 1972; PEARCE and CANN, 1973; NISBET and PEARCE, 1973; PEARCE, 1975) that they can provide information on the location of converging and accreting plate margins, of "hot-spots" within plates, and of suture zones where oceanic lithosphere has been consumed during collision of continental blocks.

It appears that geochemical methods provide us with the most effective ways of identifying and separating the various magma types. PEARCE and CANN (1973) have shown that the trace elements Ti, Y, Sr, Nb and Zr could distinguish the magma types in most cases. This paper presents the results of the study made of the major oxide chemistry and selected trace element geochemistry of the ophiolitic complexes of the Darnó megatectonic line.

Geological and tectonic background

The Darnó megatectonic line is perhaps the most important tectonic feature in the northern half of Hungary. This line trending NNE-SSW (Fig. 1) divides the formations of the Bükk Mts from those of the Central Hungarian

Mts. Towards the SW, the Darnó-line joins the Balaton-line which is regarded as the continuation in the Pannonian region of the Periadriatic Lineament, the major dividing line between the Northern and Southern Alps (WEIN, 1969; SZEPESHÁZY, 1975; SZÁDECZKY-KARDOSS, 1974).

Outcrops of Mesozoic (Upper Jurassic — Lower Cretaceous?) mafic and utramafic rocks have been reported from various places along the Balaton—

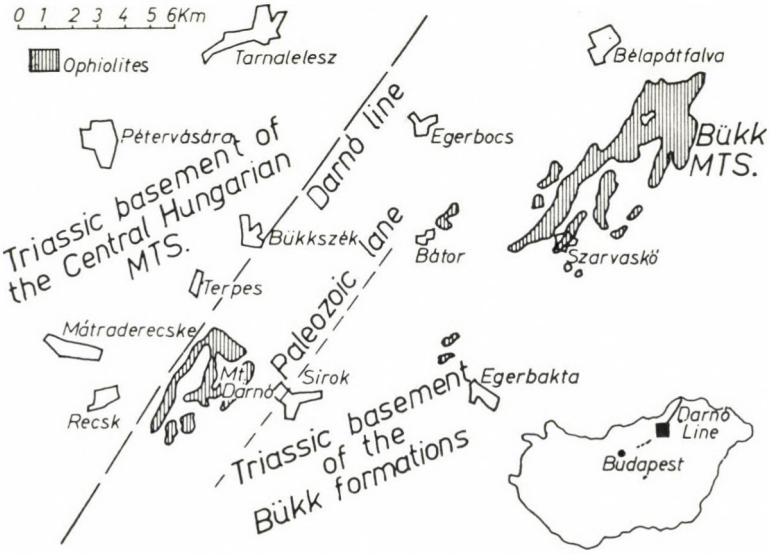

Fig. 1. Position of the investigated ophiolites of the Darnó tectonic zone

Darnó-line (DANK and BODZAY, 1970; SZEPESHÁZY, 1975). The samples used in this study were from the area around Mt. Darnó, Bátor and Szarvaskő (Fig. 1). A complete ophiolite sequence (pelagic sediments, pillow lavas, diabase, gabbros, peridotites) occurs in the area of Szarvaskő (Fig. 2b). The complex is dismembered in other places in the megatectonic zone. Near Mt. Darnó for example, the ophiolites are represented from the cumulate units upwards; ultramafics do not occur (Fig. 2a). A detailed tectonic, stratigraphic and mineralogical description of the volcanic rocks together with the associated sedimentary sequences have been given by LENGYEL (1957), FÖLDESSY (1975) and ZELENKA (1975).

SZÁDECZKY-KARDOSS (1973, 1976) interpreted the whole Balaton—Darnó-line as an ancient subduction belt. From his calculations he gives a value of 210—220 km for the width of the original ocean whose obducted fragments occur along the line. Despite the work of these authors mentioned earlier, little is still known at present about the genetic relationship between the ophiolites of the Darnó-line and those found in other parts of the Carpathian—

Balkan region. The consensus of opinion among Hungarian geologists seem to be that the Darnó-line represents a very deep fracture in the crust, which got periodically re-activated in the past (LENGYEL, 1957; ZELENKA, 1973, 1975). The mafic and ultramafic rocks are thought to be remnants of a depleted mantle intruded in situ during a re-activation of the fracture. The present author had earlier (ONUOHA, 1977) stressed the need for re-interpreting the

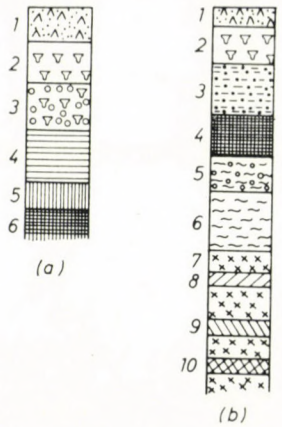

Fig. 2 a. Idealized composite section of the ophiolites near Mt. Darnó. 1. radiolarite beds and tuffs; 2. spilitic pillow lavas; 3. brecciated pillow lavas; 4. sill-like spilitic diabases; 5. diabase dyke complex; 6. gabbro-diabase. b. Idealized composite section of the ophiolites at the Várbérc gorge near Szarvaskő. 1. marine sediments, radiolarites; 2. pillow lavas; 3. diabase sheets with minor dykes; 4. gabbro-diabase; 5. Quartz diorite; 6. gabbros; 7. peridotites; 8. hornblendites; 9. pyroxenites; 10. wehrlite

igneous complex as being typical ophiolites and probably remnants of a Mesosoic ocean floor. This observation was made on the basis of stratigraphy, petrophysics and on the magnetic anomalies over the complex. The present geochemical study strengthens this proposition.

Data collection and analysis

Analyses of major element composition of the igneous rocks were taken from existing published material (SZENTPÉTERY, 1953; LENGYEL, 1957; FÖLDESSY, 1975; MÁFI data bank). Only those samples for which detailed petrological and mineralogical descriptions were available were considered. Analyses for spilites and brecciated pillows were not included because of the possibility that they may contain certain foreign materials that alter their bulk chemical composition. Data for all other members of the igneous complex were included. In addition, ten samples of basalt (four from outcrops near

Mt. Darnó and the rest from the area near Szarvaskő) were analysed for their Ti, Y, Nb and Cr contents. These trace elements have been used in this study because it has been demonstrated, that they are insensitive to alteration and other secondary processes (PEARCE and CANN, 1973; PEARCE, 1975).

Results of the chemical analyses

Major elements: The results are presented here in a series of variation diagrams. Nearly all the samples considered plot in the field of the tholeiite series on the total alkali versus silica diagram (Fig. 3). Since all the rocks are chemically similar, it would appear that those which plot outside this field must have been subjected to some post-emplacement alteration.

Figs 4 and 5 appear to remove any further doubts about the abyssal tholeiitic affinity of the rocks. On the $FeO^t - FeO^t/MgO$ diagram (Fig. 5) the ultramafics show a faster enrichment in FeO^t, than the gabbros and diabases, while on the $SiO_2 - FeO^t/MgO$ plot (Fig. 4) these same rocks show a slower enrichment in silica. This is typical of a tholeiite differentiation trend (MIYASHIRO, 1973, 1975). Alkaline tholeiites with low K_2O contents occur in most

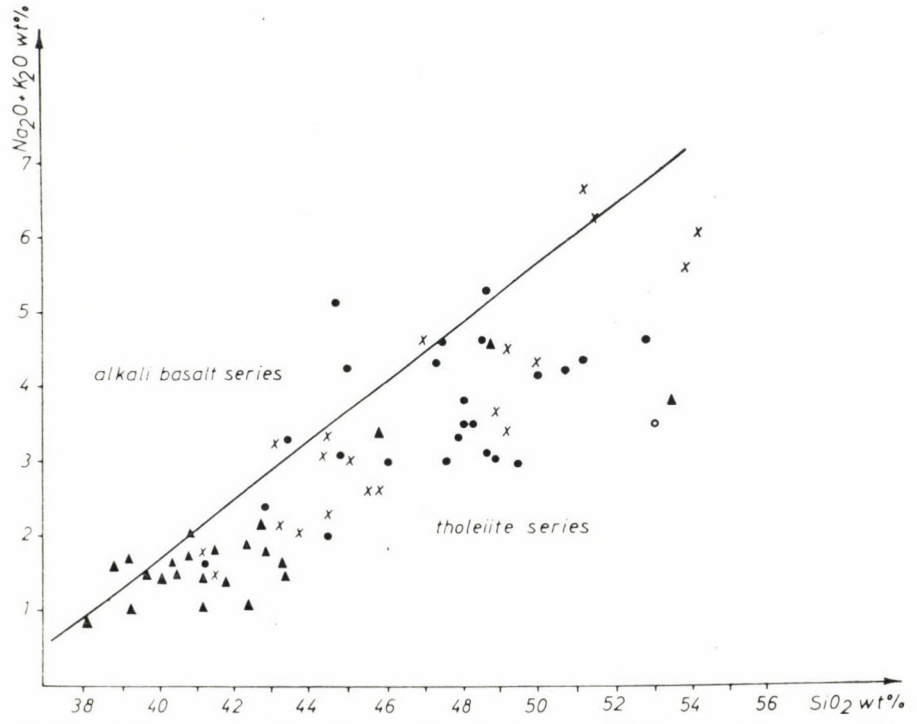

Fig. 3. Total alkali versus silica diagram for the Darnó ophiolites. ● diabases and lavas; × gabbros; ▲ ultramafics. The dividing line is according to MACDONALD and KATSURA (1964)

island arc environments, so *the* TiO_2—FeO^t/MgO *diagram (Fig. 6) shows that the volcanic rocks of the Darnó complex are not related to island arc magmatism.* The differentiation trend for the complex rather resembles that for basalts from the ocean floors as confirmed by the Deep Sea Drilling Project (DSDP).

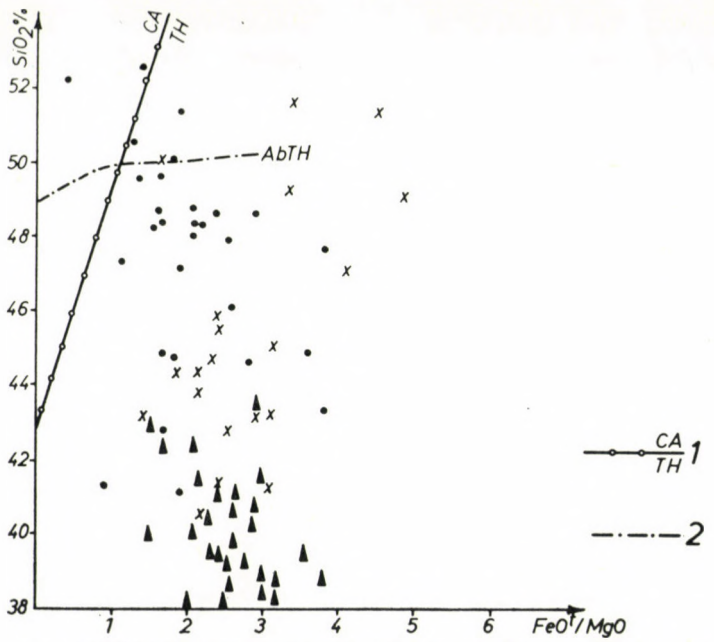

Fig. 4. SiO_2 versus FeO^t/MgO diagram for the Darnó complex. ● Diabases, pillow lavas; × gabbros; ▲ ultramafics. 1. Separation line between calcalkali and tholeiite series of MIYASHIRO (1973, 1975). 2. Abyssal tholeiites; ($FeO^t = FeO + Fe_2O_3$)

Trace elements: PEARCE and CANN (1973) have shown that the Y/Nb ratio in basalts can throw light on their tectonic origin. Alkaline basalts generally have Y/Nb ratios less than 1.5 for within-plate basalts, and less than 2.0 for ocean floor basalts (BICKLE and PEARCE, 1975).

The samples from the Darnó complex classify as tholeiitic with Y/Nb ratios between 2.4 and 8.75. The Ti—Cr plot of PEARCE (1975) is shown in Fig. 7. This diagram is useful in distinguishing ocean floor and island arc basalts, and shows that the Darnó samples belong to the ocean floor basalt series.

Implications of the results

The major element and trace element geochemistry of the Darnó igneous rocks favour an ocean floor origin for them. The similarity in fractionation trends observed between rocks of the complex and those of other ophiolites

and modern oceanic crust, coupled with stratigraphical resemblance cannot be attributed to chance. The cautiousness in regarding the complex as true ophiolites and in comparing them with similar occurences elsewhere in the Alpine-Mediterranean region seems unfounded.

Furthermore, the usual interpretation given to the Darnó-line needs to be reviewed. The volcanic rocks are not in situ occurences, but were tectonically emplaced (obducted) together with the associated pelagic sediments (cherts and radiolarites) into their present positions probably during the Late Creta-

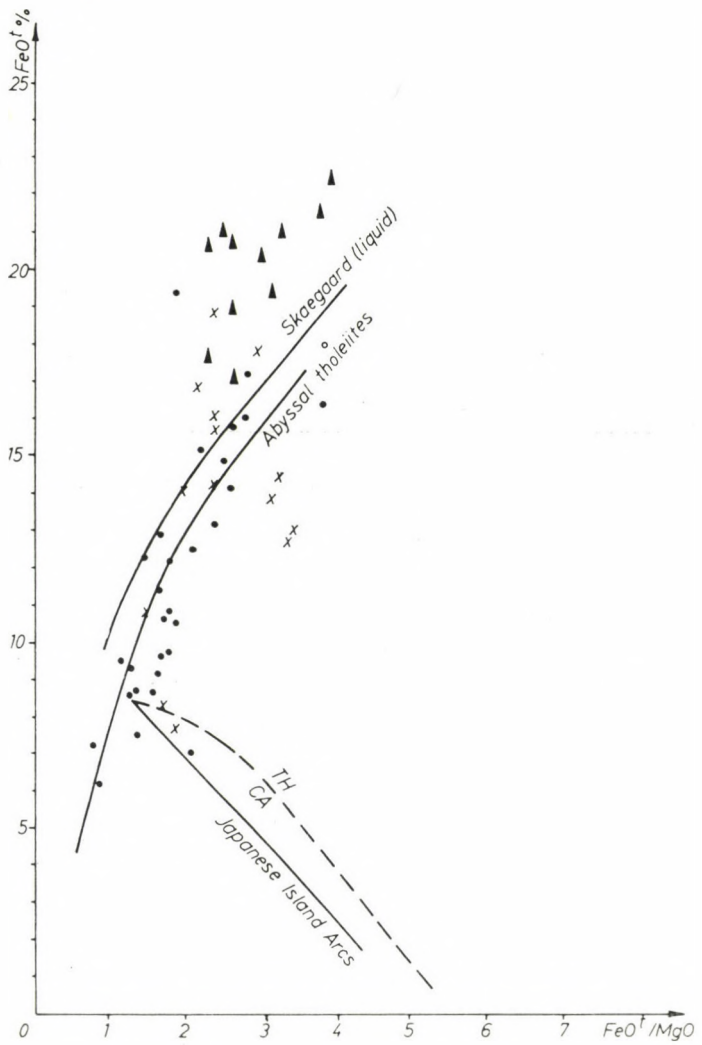

Fig. 5. FeO^t versus FeO^t/MgO diagram. ● Diabases, pillow lavas; × gabbros; ▲ Ultramafics. The differentiation lines shown for comparison are according to MIYASHIRO (1973)

ceous-Early Tertiary. The identification of the Darnó-line as an ancient subduction zone (SZÁDECZKY-KARDOSS, 1973, 1974) is supported by the presence of the ophiolites, but the problem of determining the actual boundaries of the microplate units involved remains to be solved. An alternative although yet

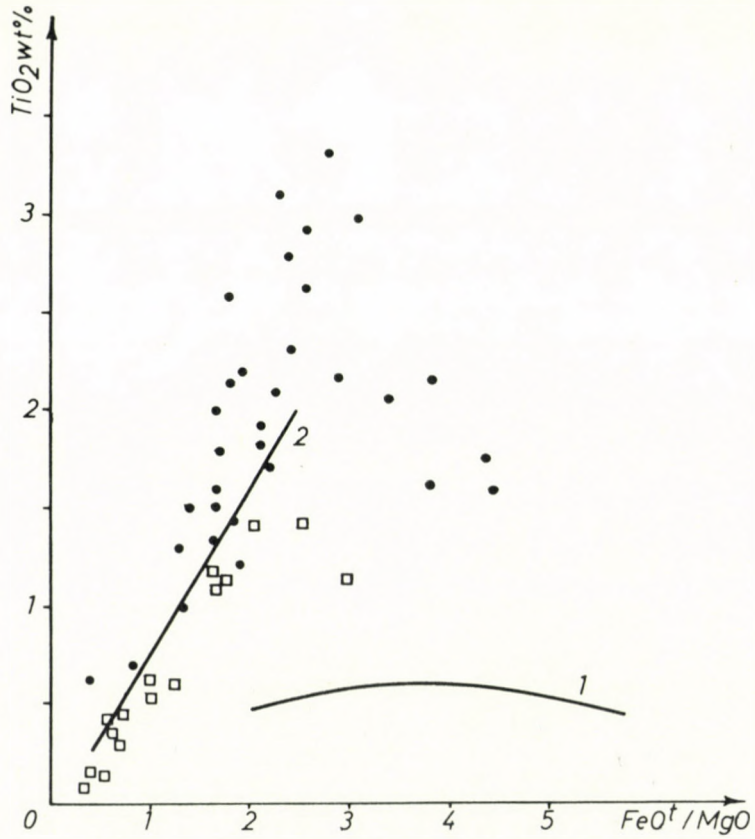

Fig. 6. TiO_2 versus FeO^t/MgO diagram. ● Diabases, gabbros and peridotites from the Darnó-line complex; □ Dolerites, gabbros and peridotites from the Kizil Hatay ophiolite complex, Turkey (COGULU et. al., 1975) 1. Low K_2O tholeiites from Tonga (EWART and BRYAN, 1972); 2. Basalts from Leg 37 of DSDP (BLANCHARD et al., 1976)

unproved solution might be to regard the ophiolites as occurences located at tectonic windows.

A suitable mode of emplacement and interpretation would still have to be found for the Darnó ophiolites in order to fit them into the existing plate tectonic model given for the Pannonian basin (STEGENA, GÉCZY and HORVÁTH, 1975; SZÁDECZKY-KARDOSS, 1974, 1976), and into the plaeogeographical reconstruction of CHANNEL and HORVÁTH (1976).

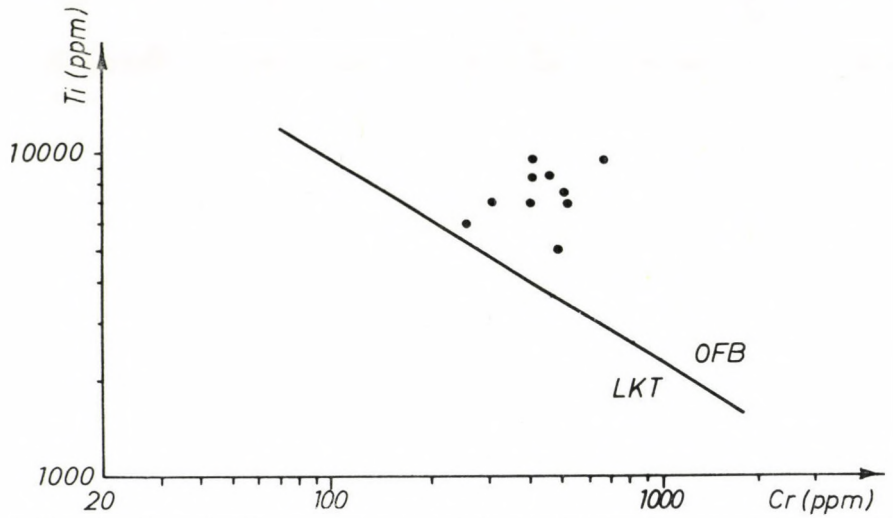

Fig. 7. Samples from the Darnó complex plotted on a Ti—Cr diagram of PEARCE (1975). OFB = Ocean Floor Basalts; LKT = Island Arc Low-Potassium Tholeiite suite

Acknowledgements

I am very much indebted to F. HORVÁTH for fruitful discussions and help while carrying out this study. K. SZEPESHÁZY and T. ZELENKA kindly allowed me access to unpublished material. I am also grateful to P. ZENTAI for spectrographic determination of the trace elements.

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ТЕКТОНИЧЕСКОЕ ЗНАЧЕНИЕ НЕКОТОРЫХ ГЕОХИМИЧЕСКИХ ДАННЫХ, СВЯЗАННЫХ С ОФИОЛИТОВЫМИ КОМПЛЕКСАМИ МЕГАТЕКТОНИЧЕСКОЙ ЛИНИИ ДАРНО (СВ ВЕНГРИЯ)

К. М. ОНУОХА

Резюме

Первоначальные тектонические условия мафических и ультрамафических пород, развитых вдоль мегатектонической линии Дарно, исследуются путем использования фактического материала, представленного анализами содержаний главных и рассеянных элементов в этих породах. Геохимические условиях этих магматических пород дают основани предположить приуроченность их образования к океаническим днам, причем их тенденции к дифференциации и фракционированию весьма хорошо совпадают с соответствующими тенденциями абиссальных толеитов. Закономерности, вытекающие из этих результатов, рассматриваются в свете эволюфии Паннонского региона, обусловленной тектоникой плит.

Acta Geologica Academiae Scientiarum Hungaricae, Tomus 21 (1—3), pp. 143—167 (1977)

THE ROLE OF GEOTHERMAL CONDITIONS AND HYDROCARBON PROGNOSTICS IN HUNGARY

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In general, the state of thermal equilibrium is dynamic and favourable to hydrocarbon migration. There is, however, a thermodynamic limit beyond which only organic substances unsuitable for migration occur.

The significance of paleo-temperatur and of the "heating time" in geohistorical sense are emphasized. The critical "temperature threshold values" influencing generation, migration, preservation and decomposition of hydrocarbons, as well as their determination and problems are also dealt with.

The changes of thermodynamic parametres in function of depth are analiyzed relying upon 1400 data from 86 prospecting areas of the Great Hungarian Plain. Acutally, the value of the average geothermal gradient independent of the location (area, depth) is 5.5 °C/100 m, i.e. 18 m/°C. The real values, however, show considerable deviations, i.e. between 26 and 10 m/°C. In different areas of the Great Hungarian Plain all the possible trends of changes have been established, i.e. constant, increasing, decreasing, repeating, etc. tendencies. In different basin parts the same isotherm value with several thousand metres of difference. According to the shape of the isotherms the oil and gas deposits are assigned to different temperature-depth horizons. These mean at the same time a classification according to composition, as well. The comparison of the Hungarian data with the world data proves the peculiar geothermal position of the hydrocarbon deposits in Hungary.

According to the adaptation of the "accumulation limiting value" introduced in the Polish Plain, the hydrodynamic state where the organic matter is unable to migrate can be drawn in a depth about 5 to 6 thousand metres. Consequently, below this depth no hydrocarbon deposits of industrial value can be expected.

The analysis of the thermal history of an area is of crucial importance since without knowing the generation and accumulation zones no up-to-date hydrocarbon prognostics can be carried out.

Introduction

In the ranks of energy resources hydrocarbons have a distinguished position. Thus, the study of the genesis, migration and accumulation of hydrocarbons had to be scientifically re-evaluated. Among the problems to be studied in connection with hydrocarbon prognostics the pressure and temperature conditions of fluids, their role in the generation, accumulation and preservation are of particular importance. Both in Hungary and in the world there is an increased demand for the exploration of greater depths. This involves higher temperatures putting the role of temperature into the focus of hydrocarbon prognostics.

The role of heat in the generation and decomposition of hydrocarbons
Change in the internal energy of sediments

The volume of subsurface materials changes considerably during sedimentation. Porosity rapidly decreases in the early state of burial followed by a normal trend of compaction. In this case the pore pressure is nearly hydrostatic and the squeeze of fluids is steady.

The pressure originating as a function of the prevailing conditions is close by related to the volume. The measure of compaction of sedimentary rocks being known, the paleo pore pressure resp. its change during geological ages can be estimated (K. MAGARA, 1976).

On the basis of sonic well logging the normal trend of marl compaction can be determined for each area and formation. Under-compacted pelites point to overpressure zone. In this case the marl compaction is not a simple function of depth. The relative over-compaction observed in the hydrostatic zone suggests former erosion the numeric value of which (the thickness of denudated sediments) can be computed. In the geological model area of the West-Canadian Basin (Pacific Amoco Ricinus, borehole 16—29—34—8—W5) the compaction as well as the measure of fluid squeezing proved to be normal down to 2530 m depth. Farther down to the maximal burial depth of 3290 m no more compaction could be verified, i.e. the fluids were enclosed in the pore volume. This is a state when neither the pressure- nor the temperature-increase can produce fluid migration. The consequences are the increase of internal energy and thermal effect.

The conductive thermal energy transmission of solid rocks is specific in general within the geological formations of an area and depends on the heat conductivity of the rocks in question. These are of the same order of magnitude, i.e. $2—10 \times 10^{-3}$ cal/cm. s. °C. The heat flow of fluids is a much more intense and rapid process since the displaced molecule groups carry their heat source. Thus, the convection flows are of great significance for hydrocarbon geology. From the point of view of hydrocyrbons the temperature increasing with the depth of burial is of double effect it is a prerequisite to reach the temperature threshold value of hydrocarbon generation and a change of pressure is caused by the change of water temperature. In subsiding areas the simultaneous increase of temperature and lithostatic pressure is fairly reflected by the behaviour of clay minerals of the pelitic rocks forming the closing rocks of hydrocarbons (E. SZÁDECZKY-KARDOSS et al., 1971). Last but not least, in the preservation and decomposition of hydrocarbon deposits the heat plays a more important role than other factors, due to the high-grade thermal instability of hydrocarbons.

In the case of thermal equilibrium

$$dp = dv = d\,T = 0,$$

i.e. the form of the determining function is

$$A = f(p, v, T).$$

A proves to be constant, when p, v, T are also constant.

Most of the hydrocarbon deposits are in state of equilibrium on the time scale of human history but, in fact, this is a dynamic (V. DANK, 1970).

From the point of view of hydrocarbons and at a geohistorical time scale the state equilibrium is represented only by a site where the organic materials are unable to migrate during several millions or ten million years.

Interpretation and application of the temperature gradient

The geothermal gradient denotes the temperature as a function of location, i.e.

$$G_t = \frac{\partial T}{\partial X}$$

The description of its relative change is possible only in a spatial polar-coordinate system. On the contrary, in the practice the geothermal gradient is demonstrated only in projection to a plane.

In the case of a study demanding the knowledge of local features the changes as a function of depth and area are to be determined, as well.

In the case of detailed studies the correction to a "neutral zone" should not be neglected. In the neutral zone a compensation between the atmosphere and lithosphere takes place. The heat of endogenic origin is not necessarily zero at the surface but its value is represented by the annual mean temperature determined by the climate. In Hungary the average of 30 years is 10.9 °C, i.e. nearly 11 °C. Thus, when computing the gradient the basic of computation is the measured temperature minus 11, and instead of zero meter about 20—30 m should be taken into account. In addition to the climatic factors the local features are to be taken into consideration, as well. E.g. in the case of lakes and seas the effect of surficial irradiation may prevail occasionally down to several hundred metres. Nevertheless, in the Dobsina Ice Cave or within the local geothermal anomaly of Tiszakécske (L. ALFÖLDI, et al., 1973) other ways of computation are to be used. The same is true for the neighbourhood of fault zones, too. Occasionally, the heat producing effects of chemical processes are to be dealt with as well.

A simple practical example: in 2600 m the measured temperature is 116 °C (static).

Without correction: 116/2600 = 0.0461 °C/m = 21.69 m/°C
With correction: 105/2570 = 0.0408 °C/m = 24.50 m/°C.

The gradient figure (Fig. 1) plotted according to the previous data shows the incorrect t_a line and the correct t_b line resp. their difference. (In case of extrapolation deviatiens of +6 and −13 °C may occur.)

The dependence of the geothermal gradient on the geographic location and depth is shown in Fig. 2. with the average curves of temperature gradients of 14 areas of the Great Hungarian Plain. The average extreme values of the gradient are as follows:

minimum: 3.8 °C/100 m = 26.3 m/°C.
maximum: 9.7 °C/100 m = 10.3 m/°C.

A constant gradient independent of depth was found only in one area. The gradient of all the other areas change as a function of depth in all possible

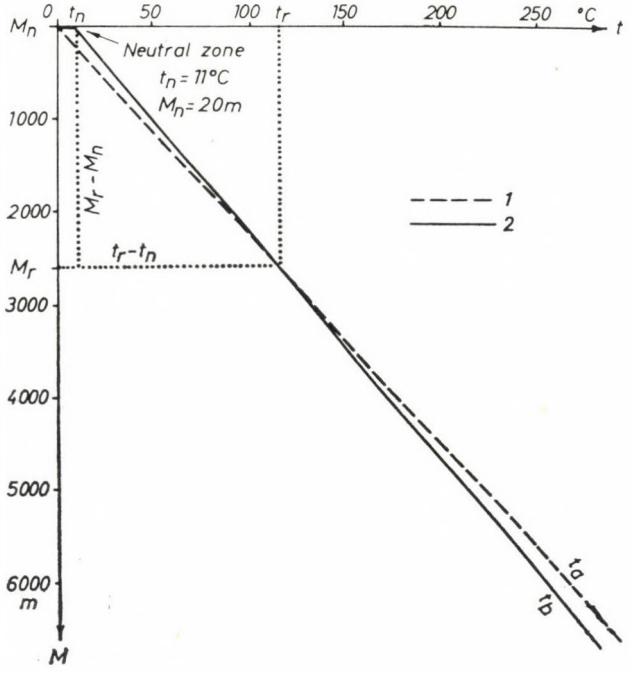

Fig. 1. Application of the computation of geothermal gradient. — 1: GG_a; 2: GG_b; Theoretically: $GG_a = t_r/W_r$; practically: $GG_b = \frac{t_r - t_n}{M_r - M_n} = \frac{t_r - 11}{M_r - 20}$ (°C/m) Measured data: $M_r = 2600$ m; $t_r = 116$ °C. — Temperature difference computed to M_a depth:

$$t_b - t_a = \frac{M_a - M_r}{M_r - (M_r - 20)} [20t_r - M_r 11]$$

According to the example:

M_a = 1000 m = + 6.27 °C
2000 m = + 2.35
3000 m = − 1.57
4000 m = − 5.48
5000 m = − 9.40
6000 m = − 13.38

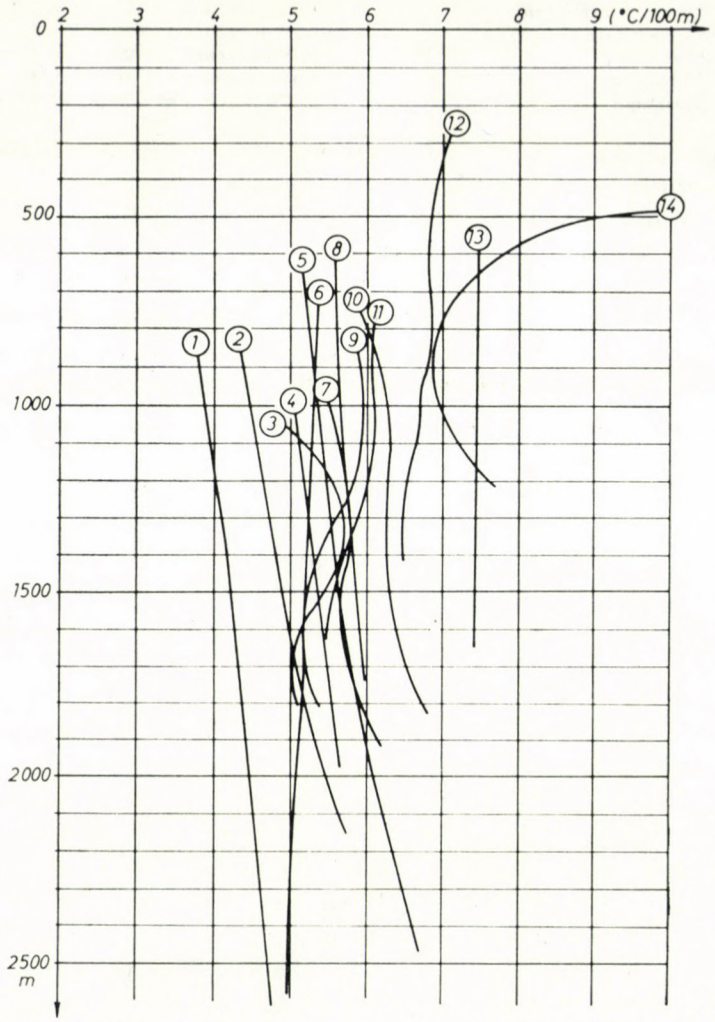

Fig. 2. Local changes of the geothermal gradient. 1. Algyő; 2. Kiskunhalas; 3. Kisújszállás-Ny; 4. Szank; 5 Üllés; 6. Endrőd; 7. Kunmadaras; 8. Kaszaper-D; 9. Szandaszöllős; 10. Pusztaföldvár; 11. Hajdúszoboszló; 12. Ebes; 13. Tótkomlós; 14. Mezőhegyes

variations. There are gradients increasing and decreasing with depths, and repeating or reversed tendencies are also found.

Evaluation of the conditions of thermal history

The concept of "heating time" (CONNAN, 1974, and HUNT, 1975) was born out of the consideration of the time effect on the geological scale. According to CONNAN the temperatures required for the start of oil generation are as follows (by geological formations and taking into account the geological ages):

Upper Pannonian	4 million years	= 140—150 °C (estimated)
Lower Pannonian	10 million years	= 120—130 °C
Miocene	25 million years	= 105—110 °C
Paleogene	50 million years	= 90— 95 °C
Mesozoic	185 million years	= 90— 70 °C.

The values (Fig. 3) concern the oil generation. Regarding the methane opinions and results differ considerably.

Hungarian scientists dealing with geochemical hydrocarbon prognostics have analyzed directly or indirectly the paleo-temperature conditions determining the maturity of organic matter (GY. CRASSELLY, 1970, 1975; A. BALÁZS—I. KONCZ, 1975; I. BODZAY, 1975; J. TÓTH, 1975). The generalized result of the investigations on diagenesis coefficient is that the value of CR/CT increases with depth in the range of 0.55 to 0.90 (BALÁZS—KONCZ, 1973). It is to be noted, however, that by studies concerning large areas, e.g. the Great Hungarian Plain, or Transdanubia, the real geological reasons of the changes can not be detected. The investigations restricted to one research area, e.g. in Endrőd found first increase later decrease in function of depth. These results correspond to the local changes of the geothermal gradient. In the Algyő—Makó region a fairly well fitting vitrinite reflectance gradient has been determined, i.e. 0.013—0.017 R_0% per 100 m in the Laboratory for Geochemical Research of the Hungarian Academy of Sciences. In accordance with the depth difference between the Algyő and Makó area 0.58—1.34 resp. 1.10—1.51 R_0% were determined for the Lower Pannonian resp. for the Miocene. The ancient temperature turned out to have been lower than the present one. This proves the continuous subsidence of the area. In the region of considerably different geological setting, e.g. in Kömlő, the contact effect of Miocene volcanites on the Pannonian formations caused an advanced coalification to the anthracite-graphite state already. The paleogeothermal gradient computed to the Pannonian floor (about 2300 m) is 6.47 °C/100 m as against the recent 5.17 °C/100 m. Consequently, the reciprocal gradient has changed from the Lower Pannonian 15.45 to the recent 19.34 m/°C.

The thermal history of the Algyő—Makó region displays a warming tendency while that of the Kömlő area a relative cooling.

In addition to the nearly traditional vitrinite reflectance investigations (J. KARWEIL, 1956) further methods are to be applied in the studies of thermal history. WALTER (1973) used the ESR-method (electron spin resonance) to determine paleotemperatures. The method of "homogeneization temperature" elaborated to estimate the paleo-temperature (CURRIE and NWACHUKWU, 1974) provides new possibilities to study the paleo-temperature. (In thin sections the fluid inclusions of the quarz material filling the fissures were heated till the bubbles disappeared and this temperature was called homogeneization temperature. This temperature value coincides with the paleo-temperature at which the inclusions were formed.)

The cumulative heat impulse computation of the temperature-time index was developed by N. V. LOPATIN (1971). This is based on the assumption that a temperature increase of 10 °C doubles the rate of coalification describing the degree of diagenesis of the initial organic matter. When knowing the depth, temperature (or paleo-temperature) and the absolute geological age of the

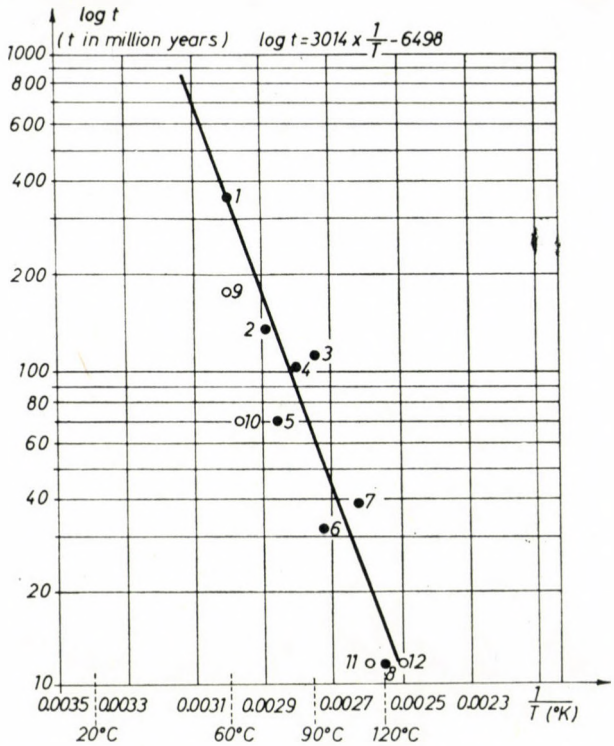

Fig. 3. Temperature threshold value of the main phase of oil formation as a function of the absolute geological age (CONNAN, 1974). 1. Amazonas Basin; 2—3. Aquitanian Basin; 4. Rio de Oro, El Aduin area; 5—6. New Zealand, Taranaki Basin; 7. Camargue Basin (France); 8. Sulu-area (Sabah); 9. LOUIS and TISSOT, 1967; 10. ALBRECHT, 1969; 11—12. PHILIPPI, 1965

formations, the staying in the intervals of 10 °C in question can be graphically determined. The multiplication of this by the coefficient of reaction rate gives the cumulative heat impulse which can be parallelized with the scale of coalification as well as with the phases of oil and gas formation.

Temperature "threshold-values" and the hydrocarbons

In the geological cycle of hydrocarbon formation, migration, accumulation and decomposition there are several threshold values. These are differentiated according to the hydrocarbons and other gas-state fluids.

At low temperatures (10 to 20 °C, i.e. the same as in the surface) only hydrocarbons of no industrial amount may develop. It is characteristic of the methane originating from organic matters on a microbiological way that it occurs only dissolved in water. The formation of fluid hydrocarbons starts above 60—65 °C from kerogen. TISSOT and WELTE, however, assume a "lateral chain" of CO_2, H_2S and N_2 end-products in the lateral branch of the reaction. According to G. T. PHILIPPI, (1965) a temperature of at least 115 °C is needed for the formation and primary migration of oils. The pro and contra discussion has been continuous since 10 to 12 years (pro: KARTSEV et al., TISSOT, PELET, NIXON; contra: CORDELL, WELTE, POWERS, BURST). The followers of the most up-to-date concept have recognized that the "maturity" of the organic matters depends on the joint effect of temperature and geological time, thus the "threshold" value of oil formation may change between 60 and 120 °C.

The Hungarian experts draw the lower and upper limit of oil formation at 50 to 60, resp. 135—153 °C, the corresponding values of diagenesis coefficient and vitrinite reflectance are 0.6 to 0.8 resp. 0.4 to 1.5 $R_0\%$. In the Hungarian Mesozoic — Pannonian interval (i.e. 340 to 4 million years) the temperature intervals of the start and end of oil formation can be estimated according to I. KONCZ, as follows:

	Start	End
oil	51—106 °C	88—153 °C
light oil	113—160 °C	145—198 °C
gas condensate and wet gas	180—218 °C	207—251 °C
dry gas	>218	>251

Less significance has been attributed to the role of temperature in the migration. Papers of the 60-s have investigated the temperature conditions first of all in relation with the solubility of oil and gas in water. This was followed by the investigation of the role of dehydration level (POWERS, BURST, etc.) which dealt with the transformation of montmorillonite into illite was bound to a relatively narrow temperature interval (80 to 120 °C). In connection with the migration of aquaeous fluids K. MAGARA proves (1974) that the volume of water increases with increasing burial even in case of constant geothermal gradient and this expansion is favourable to the migration in great depths. Barker (1972) constructed a diagram to compute the overpressure in a closed system (water pressure-temperature density) and verified the fact that under such conditions the high-temperature is nearly solely responsible for the overpressure. The thermal expansion of rocks is negligible as compared to that of water.

In the samples of the Gulf Coast area of Louisiana volume increase of 3 to 15 per cent was determined (K. MAGARA, 1974). The geothermal gradients proved to be 1.8—2.5—3.6 °C/100 m. This is worthy of mention since the

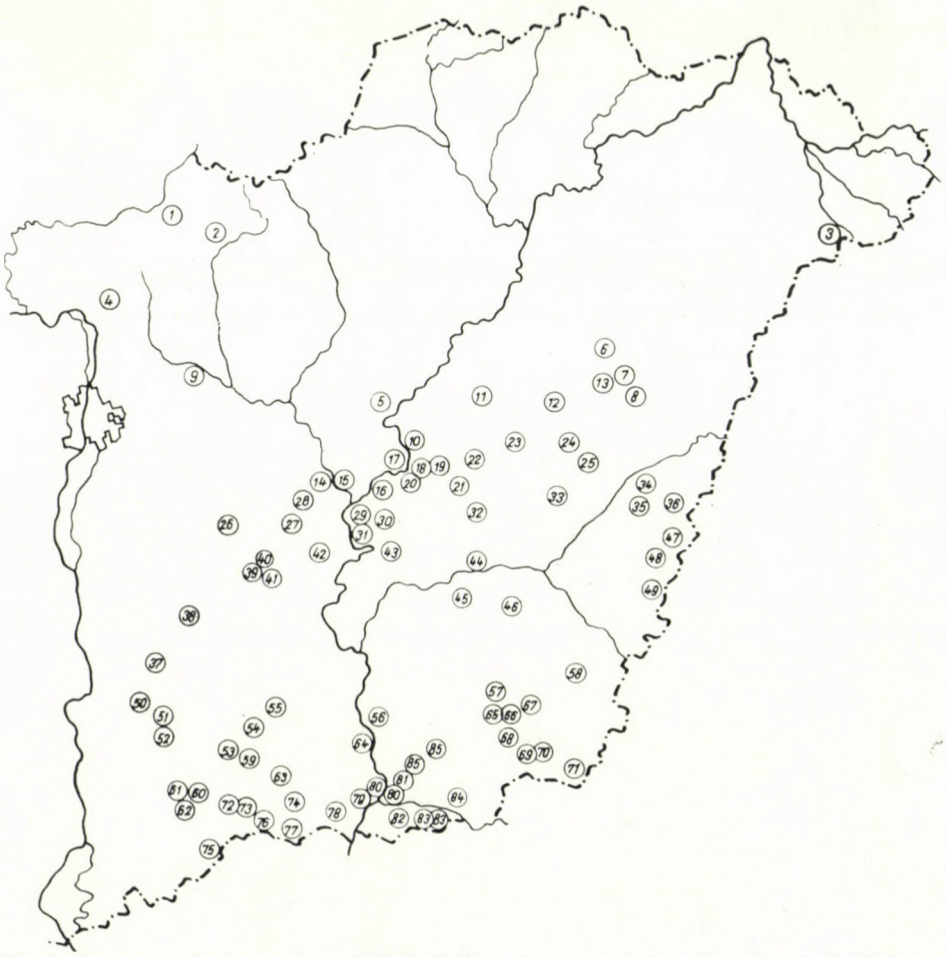

Fig. 4. Locations of thermal sampling in hydrocarbon prospecting boresholes. 1. Szécsény; 2. Sóshartyán; 3. Nagyecsed; 4. Őrszentmiklós; 5. Komló; 6. Kaba-É; 7. Hajdúszoboszló; 8. Ebes; 9. Farmos; 10. Tiszagyenda; 11. Kunmadaras-Tülés; 12. Nádudvar; 13. Kaba; 14. Zagyvarékás-É; 15. Zagyvarékás; 16. Tiszapüspöki; 17. Nagykörü; 18. Fegyvernek; 19. Fegyvernek-K; 20. Surján; 21. Turgony; 22. Kisújszállás-Ny; 23. Karcag-Bucsa; 24. Püspökladány; 25. Biharnagybajom; 26. Cegléd; 27. Törtel; 28. Abony; 29. Szandaszőllős; 30. Kengyel; 31. Rákóczifalva; 32. Túrkeve; 33. Füzesgyarmat; 34. Furta; 35. Mezősas; 36. Kismarja; 37. Izsák; 38. Kecskemét-Ny; 39. Nagykőrös; 40. Nagykőrös-K.hegy; 41. Nagykőrös-D; 42. Jászkarajenő; 43. Martfü; 44. Endrőd; 45. Szarvas; 46. Hunya; 47. Biharugra-Ny; 48. Komádi; 49. Sarkadkeresztur-I; 50. Kiskőrös-ÉK; 51. Soltvadkert; 52. Kecel; 53. Tázlár; 54. Szank; 55. Pálmonostora; 56. Mindszent; 57. Pusztaföldvár; 58. Medgyesbodzás; 59. Harka; 60. Kiskunhalas; 61. Kiskunhalas-ÉK; 62. Kiskunhalas-ÉNy; 63. Forráskút; 64. Felgyő; 65. Pusztaszőllős; 66. Pusztaszőllős-K; 67. Kaszaper-D; 68. Tótkomlós; 69. Mezőhegyes; 70. Mezőhegyes-K; 71. Battonya-K; 72. Öttömös; 73. Öttömös, Ny; 74. Üllés; 75. Tompa; 76. Kelebia; 77. Ásotthalom; 78. Dorozsma, 79. Szeged; 80. Algyő; 81. Algyő-vízkút; 82. Ferencszállás; 83. F-K-Kiszombor; 84. Makó; 85. Maroslele; 86. Hódmezővásárhely

maximum of this area coincides with the minimum measured in the Great Hungarian Plain. Consequently, under the Hungarian conditions the role of thermal expansion of water should be much more important than in the original data of KENNEDY and HOLSTER (1966) used by Magara. The role and significance of the intra-basin water flows have been evaluated also by Hungarian researchers (E. SZÁDECZKY-KARDOSS, 1970; L. STEGENA, 1972; L. STEGENA et al., 1975). From the point of view of hydrocarbon migration and accumulation the water flows should be considered as a thermal impulse. The geological evolution of each basin and basin part should be taken into account. (L. KŐRÖSSY, 1971).

When investigating the free energy of hydrocarbons as a function of temperature increase it can be concluded that except the methane, ethane and propane all the hydrocarbons become instable above 100 °C at normal pressure. The decomposition temperature of liquid hydrocarbons is believed to be around 145 °C (ADAMS, KIRKBY). However, in the borehole Makó-1 in a depth of 4145 m light oil is known at 173 °C. According to Hunt above about 180 °C only by-products deriving mostly from decomposition, may exist.

Changes of the thermodynamic parameters in function of depth

476 × 3 = 1428 basic data of 86 surveying areas of the Great Hungarian Plain were used. The related basic parametres are depth, temperature and pressure. The deficient or erroneous data were neglected. Further 1428 data were obtained by computation, i.e. temperature gradient, pressure gradient and the accumulation value (T/p). The areas of thermal sampling are shown in a sketch map (Fig. 4).

Change of the temperature gradient in function of depth

Most of the data fall into the region of 4 to 8 °C/100 m (25—12.5 m/°C). In Fig. 5 the distribution of geothermal gradients as a function of depth is shown. On the basis of about 480 data of special accuracy the geothermal gradient resp. its reciprocal in the Great Hungarian Plain are 5.5 °C/100 m, i.e. 18 m/°C.

In the deepest borehole of the world (Berta Rogers-1, Louisiana) 232 °C temperature was measured in a depth of 9582 m. Without the surficial correction this provides a gradient of 2.42 °C/100 m. According to the geothermal gradients of the Great Hungarian Plain the individual isotherms occur in the following depth intervals:

90 °C = 2600— 800 m
120 °C = 3600—1100 m
150 °C = 4600—1400 m

Recently, no systematic hydrocarbon production is carried out above 150 °C in the Great Hungarian Plain. This temperature may occur, however, in the depth interval of 1400 to 4600 m, our productive beds lie above 3000 m.

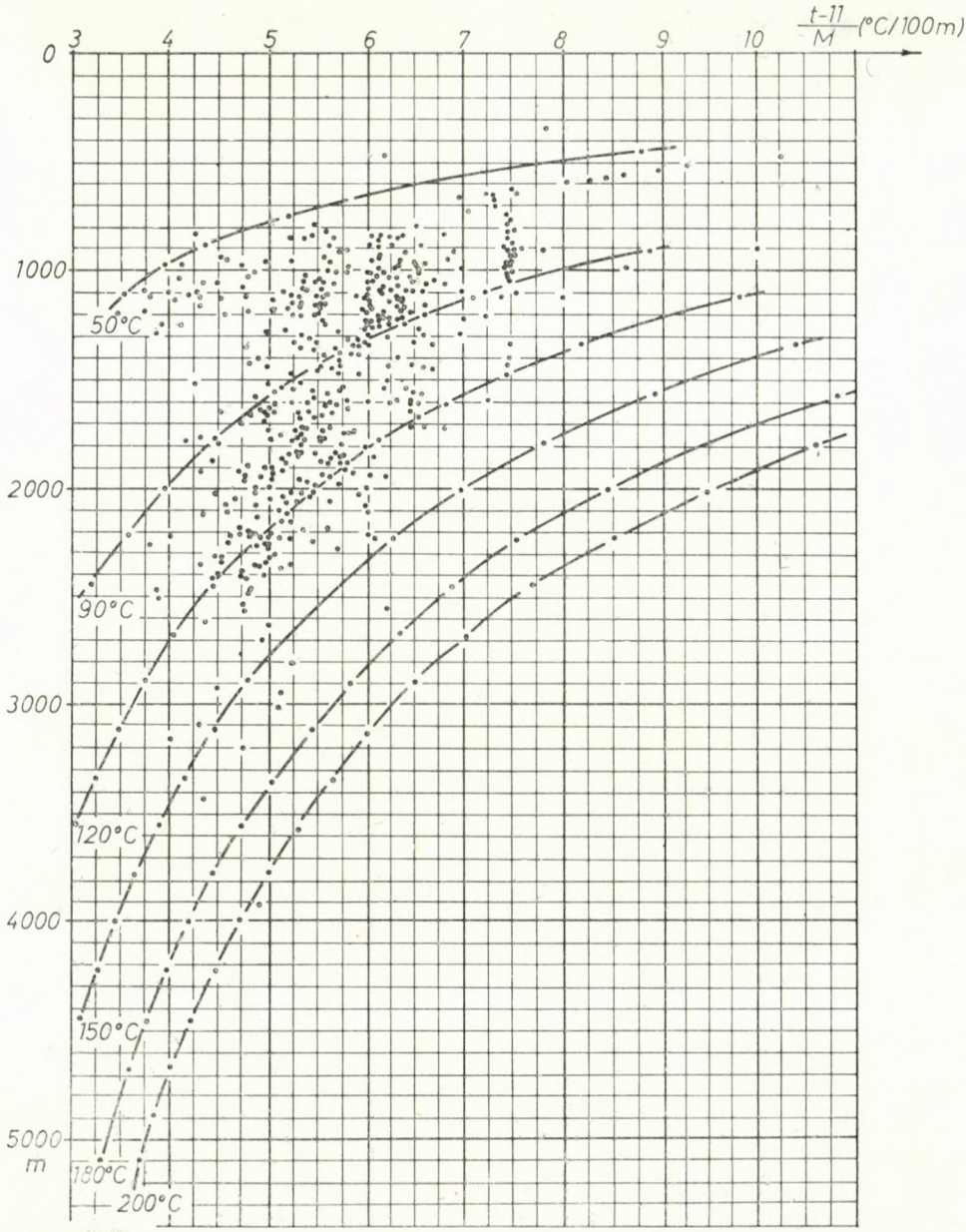

Fig. 5. Change of the geothermal gradient as a function of depth (data from the Great Hungarian Plain). I. isotherms decisive for hydrocarbons

The remaining depth interval of 1600 m (between 3000 and 4600 m) is the main region of our perspectives.

Change of the geothermal gradient in overpressure areas

Only 110 data of the overpressure areas could be used. The geothermal gradients show more concentrated position in the "focus" at about 2000 m depth (Fig. 6). The average geothermal gradient, however, is less than the total average of the Great Hungarian Plain since it is characterized by 5 °C/100 m (20 m/°C). This is probably caused by the fact that these types of hydrocarbon deposits occur in the areas where the geothermal gradient is of decreasing trend.

In a closed system the progressing thermal energy transmission is realized in pressure increase due to the expansion of water.

Arrangement of overpressure zones

Numerous authors dealt with the pressure gradient conditions of overpressure areas. In Hungary the most comprehensive study on this topic was given by A. SOMFAI, (1976). Attention is called here only to several tendencies which can be seen in the pressure gradient — depth diagram (Fig. 7). The isobar curves may run in a rather wide depth range. E.g. 300 att may be between 1300 and 4000 m but practically it is found only below 1900 m. Between 2000 and 5000 m the overpressure gradually increases up to 16 att/100 m. This is the maximal value of the normal trend known so far and is represented by 785 att pressure in the borehole Hódmezővásárhely-I.

Oil and gas horizons in function of temperature and depth

J. J. KORTCHAGINA distinguished genetic zones which essentially agree with the results of American experts. J. M. HUNT (1974, 1975) and K. K. LANDES (1966) distinguished oil and gas zones of different composition according to the depth and geothermal steps and as a function of formation temperature.

Based on about 450 sufficiently accurate data the change of the geothermal depth-steps was also constructed (Fig. 8). The upper value of the main zone of oil generation was taken to 93 °C (Landes), 80 °C (Vassoevich), 75 °C (Visotzki); E. SZÁDECZKY-KARDOSS (1973) put the upper limit of the oil zone of intermediary type to 80 °C, the start of oil formation was believed to be in average at 84 °C, (KONCZ) etc.

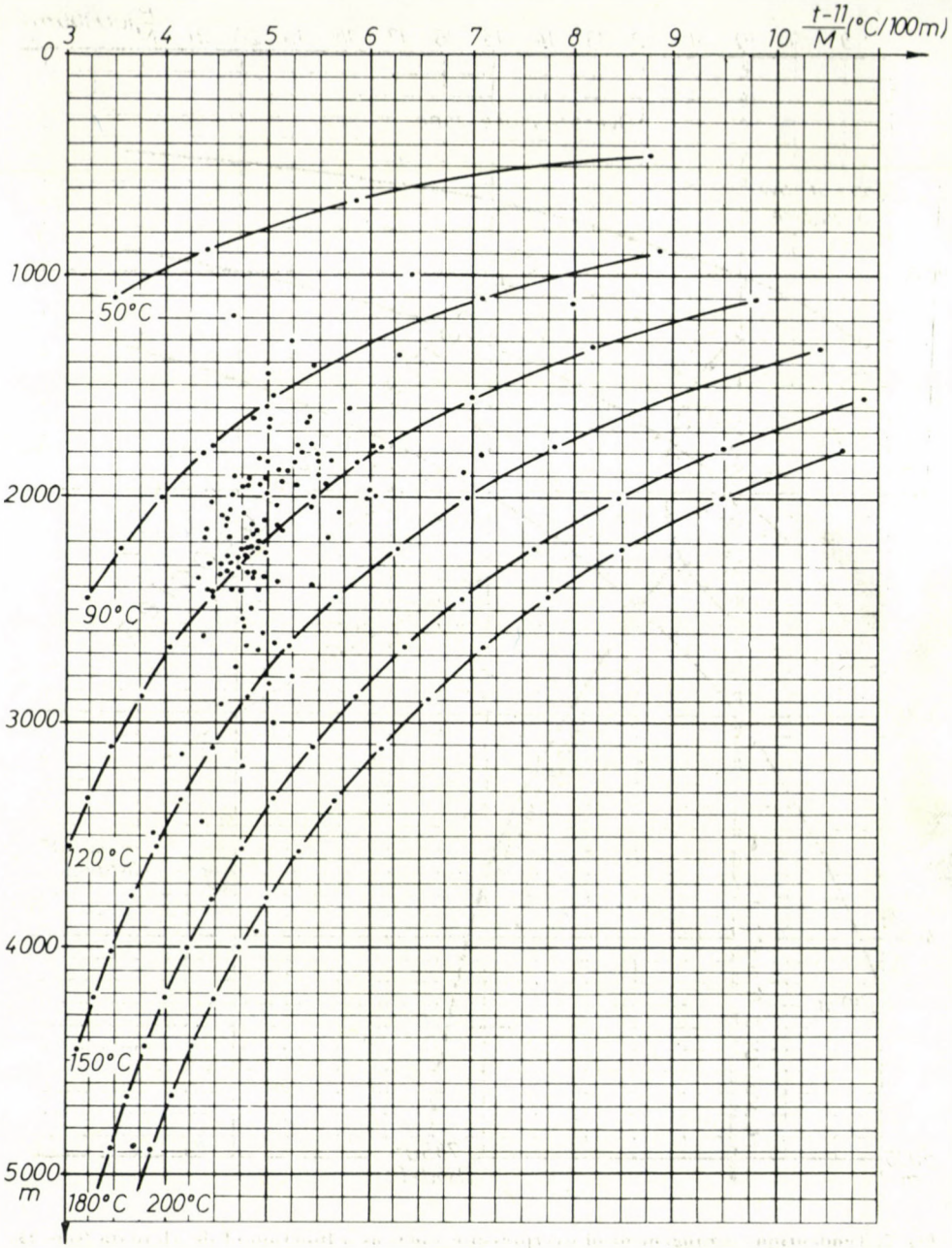

Fig. 6. Change of the geothermal gradient in overpressure areas (data from the Great Hungarian Plain) 1: isotherms decisive for hydrocarbons

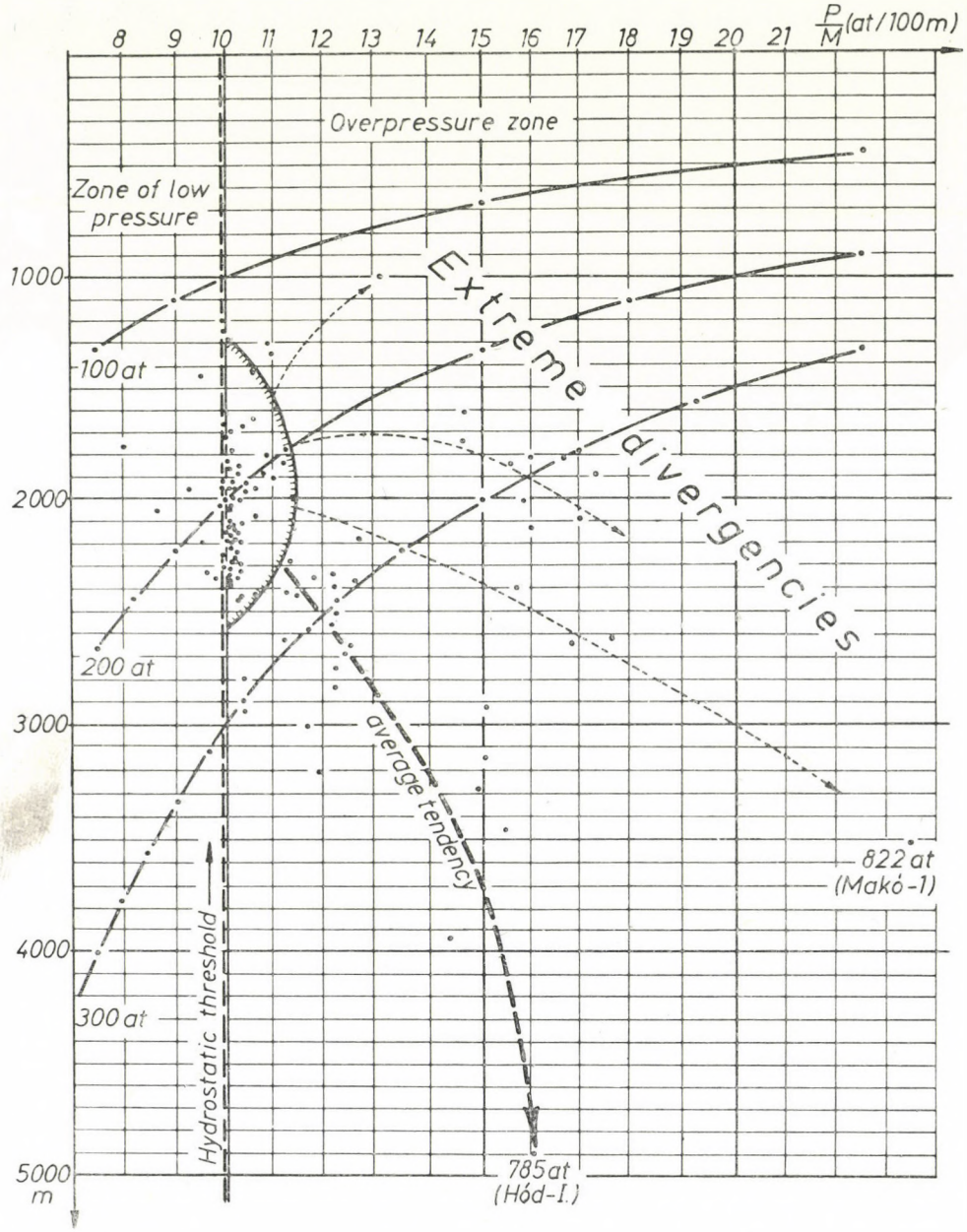

Fig. 7. Tendentious arrangement of overpressure zones as a function of depth (data from the Great Hungarian Plain)

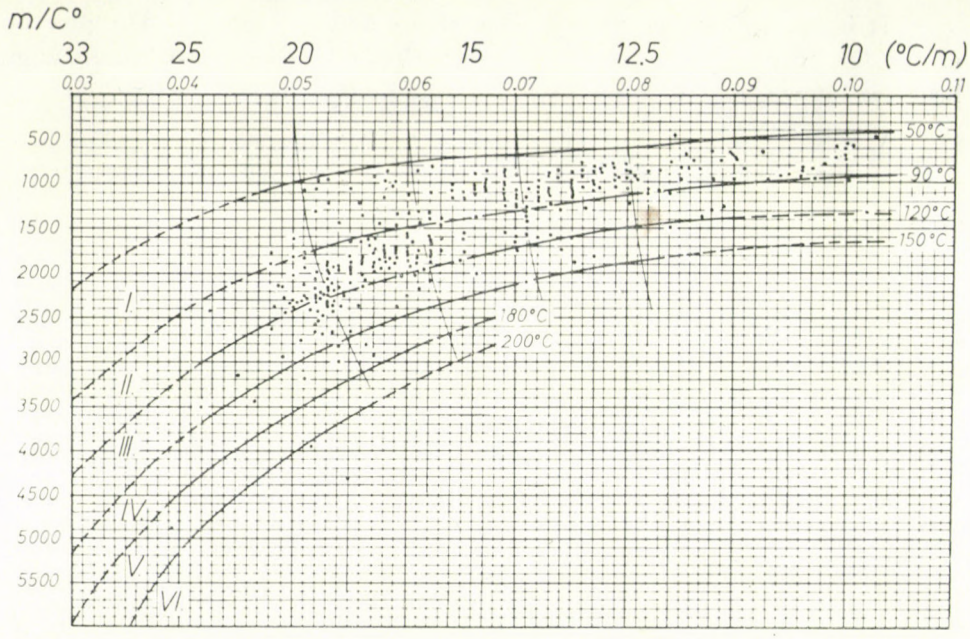

Fig. 8. Geothermal depth-step. I. Oil and gas; II. light oil and gas; III. mainly gas; IV. ceasing of oil; V. only gas; VI. carbon dioxide and hydrogen sulfide

Consequently, the limiting temperature should be between 70 and 90 °C. From the surface downwards the following temperature limits were constructed as the isotherms of the depth — geothermal gradient curve:

20, 50, 90, 120, 150, 180, 200 °C.

The goethermal depth-step and the boundaries of hydrocarbon zones characterized by isotherms are shown in Fig. 8.

Near-surface zone uninteresting to hydrocarbons

Temperature interval: 11 to 20 °C
Depth interval: 180 to 500 m.

Its most characteristic term is biochemical-gas-genetic zone. Not illustrated here this is the horizon of potable water.

Zone of artesian wells

Temperature interval: 20 to 50 °C
Depth interval: 450 to 1250 m.

It is open upwards, thus hydrocarbons may "dissipate". The presence of gas dissolved in water is characteristic. Upper Pliocene local lenticular gas beds are known, e.g. at Pusztaföldvár, Hajdúszoboszló.

Upper oil and gas zone

Temperature interval: 50 to 90 °C
Depth interval: 900 to 2200 m.

In the Great Hungarian Plain this is the zone of "heavy" oils and free gas beds. Some of the occurrences (in order of 0.1—0.05 °C/m): Mezőhegyes (Lower and Upper Pannonian gas beds); Tótkomlós (Upper Pannonian gas beds); all the gas beds of Hajdúszoboszló and Ebes; Battonya (Upper Pannonian gas and Lower Pannonian oil beds); Biharnagybajom (Pannonian — Miocene oil and gas bed); Pusztaszőllős (Upper Pannonian gas beds); Kelebia (Miocene oil beds); Kunmadaras (Lower Pannonian gas beds); Kisújszállás (Lower Pannonian gas beds); Szandaszőllős (Pannonian gas beds); Farmos (Pannonian — Miocene gas beds); Nagykőrös (Miocene oil beds with carbon dioxide, Pannonian CO_2-beds); etc.

Main zone of light oil and gas

Temperature interval: 90 to 120 °C
Depth interval: 1200 to 3000 m.

This is the zone of thermocatalytic formation and accumulation of oil and gas beds. The most important mainly gas-capped beds of the Great Hungarian plain are assigned to this group. The reservoir rocks are of Neogene age.

Characteristic occurrences (between the values of 0.05 to 0.08 °C/m): Algyő (Upper Pannonian gas-capped oil beds and free gas beds); Ferencszállás (Lower Pannonian gas — gas-condensate and oil beds); Szank—Tázlár—Kiskunhalas (Miocene oil and gas beds); Endrőd (Lower Pannonian gas beds); Püspökladány—Füzesgyarmat (Miocene oil beds with carbon dioxide); Pusztaföldvár (Lower Pannonian gas and oil beds); Nagykörü—Fegyvernek—Kisújszállás (Lower Pannonian beds with carbon dioxide and mixed gas).

Predominantly gas zone

Temperature interval: 120 to 150 °C
Depth interval: (1500)—1900 to 3800 m.

The gas priority of this zone is uncertain in Hungary. The Miocene-Mesozoic oil bed of Szeged where the gas is negligible should be assigned to this zone, but the Miocene-Mesozoic gas and gas-condensate beds of Üllés are

assigned zonally also to this group. This inconsistency suggests that this is the thermal limiting zone of formation and decomposition but without "heating time" the temperature points only approximately to the thermodynamic state. If the organic matter is of Lower Pannonian age, then according to Connan the oil formation only starts at 120 to 130 °C; according to others (ADAMS, KIRKBY) the liquid hydrocarbons decompose already at 145 °C, but this warming is effectless to the methane gas.

Zone of oil ceasing

Temperature interval: 150 to 180 °C
Depth interval: (1800)—3000 to 4500 m.

The Early Paleozoic oil bed of Dorozsma and the light oil explored in the Lower Pannonian reservoir of the borehole Makó-1 are assigned to this group. According to Vassoevich and Kortchagina this is the upper limit of the "secondary migration" horizon. As to the concordant Soviet and American references, 180 °C is the upper stability limit of hydrocarbons.

The zone containing only natural gas

Temperature interval: 180 to 200 °C
Depth interval: (only theoretical probability:) about 4000 to 6000 m.

No gas bed has been found in the Great Hungarian Plain in this group. According to Soviet data this zone is qualified as the zone of "tertiary migration" and this is the zone of gas condensates.

The ultra-deep and super-deep zone (carbon dioxide and hydrogen sulfide)

Temperature: 200—250—350 °C.

On the basis of Hungarian data the gradient — depth relation cannot be determined (the theoretical construction is shown in Fig. 8).

In the borehole Hód-1 230 to 240 °C was measured in the depth interval of 4500 to 5800 m. The temperature of 238 °C measured in the Budafa-II and deriving from the carbon dioxide zone is nearly of the same value.

In these temperature ranges the hydrocarbons decompose into their elements. Hydrogen sulfide, carbon dioxide and occasionally methane traces develop from kerogen. Hydrogen sulfide and carbon dioxide are formed from the gas, sulfur and water of the reservoir. In super-depths and at about 290 °C the precipitation of original sulfur was observed. On the basis of the depth and temperature data of Soviet and American ultra- and super-deep drills it has been stated that the gradients of 1.8—3.1 °C/100 will be hardly found in Hungary. Our data approaching the 6000 m depth are only of 3.8—4.0 °C/

100 m. Thus, it is unexpected that our 240 °C of 5800 m depth (Hód,I) will decrease to 232 °C of the Louisiana (Berta Rogers-1) value in the same depth of Hungary.

The peculiar geothermal conditions of Hungary are reflected also by the temperature- and depth-dependent zones of the oil and gas beds (Table I). The isotherms denoting the limiting values lie higher by 0.6 to 3.0 km than indicated by Landes. This, however, means that in Hungary the successful hydrocarbons prospecting is worthy to carry out down to 6000 m resp. up to the corresponding temperature.

Table I

Zones according to temperature and depth of the oil and gas beds

Characteristic composition of the bed (quality)	Temperature interval (°C)	Depth of occurrence (km)		Deviation
		according to K. K. LANDES	in the boreholes of the Great Hungarian Plain	
I. Oil and gas	50—90	0.6—4.0	0.4—3.4	—0.6
II. Light oil and gas	90—120	1.2—5.5	0.9—4.3	—1.2
III. Mainly gas	120—150	1.8—7.0	1.3—5.2	—1.8
IV. Ceasing of oil	150—180	2.5—8.5	1.6—6.0	— 2.5
V. Only gas	180—200	2.8—10.0	2.0—(7.0?)	—3.0
VI. Carbondioxide and hydrogen sulfide	200—felett	?	—	

The limits of depth categories and terms of the drills are given in Table II.

Table II

Borehole categories

1.	0—1500 m	Shallow borehole
2.	1500—3000 m	Medium-deep borehole
3.	3000—4500 m	Deep borehole
4.	4500—6000 m	Very deep borehole
5.	6000—7000 m	Ultradeep borehole
6.	7000—8000 m	Superdeep borehole

Depth world record: 9.583 m (in 1974)
Deepest Hungarian borehole (Hód-I): 5 842.5 m
Technically realizable borehole: 12 000 m
Debated but realizable possible depth: 15 000 m

On the lower limiting value of hydrocarbon accumulation

The basic formula of the relationship determined by L. CIMASZEWSKI (1976) is

$$Y = \log a\ X, \text{ where}$$

$Y = (-1)\ (-H)$ absolute depth,
$X = T/p$ the quotient of the temperature in °K and of the pressure in ata.

$$a = \sqrt[Y]{X}$$

The formula is based on empiric data neglecting the mathematical deduction only the basic principle will be presented.

Under natural geological conditions the hydrocarbon beds occur in the state of thermodynamic equilibrium, i.e. the common thermodynamic equation of state is as follows:

$$ds - \frac{du - W_{el}}{T} = 0, \text{ where}$$

ds = change of the enthropy
du = change of the internal energy
W_{el} = internal energy of hydrocarbons (at absolute zero)
T = absolute temperature (°K).

In case of accumulation the values of p, v and T remains unchanged, ie. $dp = 0$, $dv = 0$, $dT = 0$, thus the accumulation function is

$$A = f(p, v, T)$$

In the state of equilibrium $A = \text{const.}$ if $p, v, T = \text{const.}$

In possession of the data of p and T the partial derivative of the accumulation function of p and T resp. its quotient can be determined.

$$\frac{T}{P} = -\left(\frac{\partial u}{\partial s}\right)_v \left(\frac{\partial v}{\partial u}\right)_s$$

Consequently, the T/p value is a special function which connects the volume, internal energy and entropy of the hydrocarbon accumulation.

Applying about 1400 data the T/p function has been constructed in two variations, i.e. for the entire region of study and for only the overpressure areas, both as a function of depth.

Figure 9 shows the change of accumulation value in function of depth. The average curve shows a power function and its run based on the applied data is as follows:

$$T/p = 722123 \cdot M^{-0.783}$$

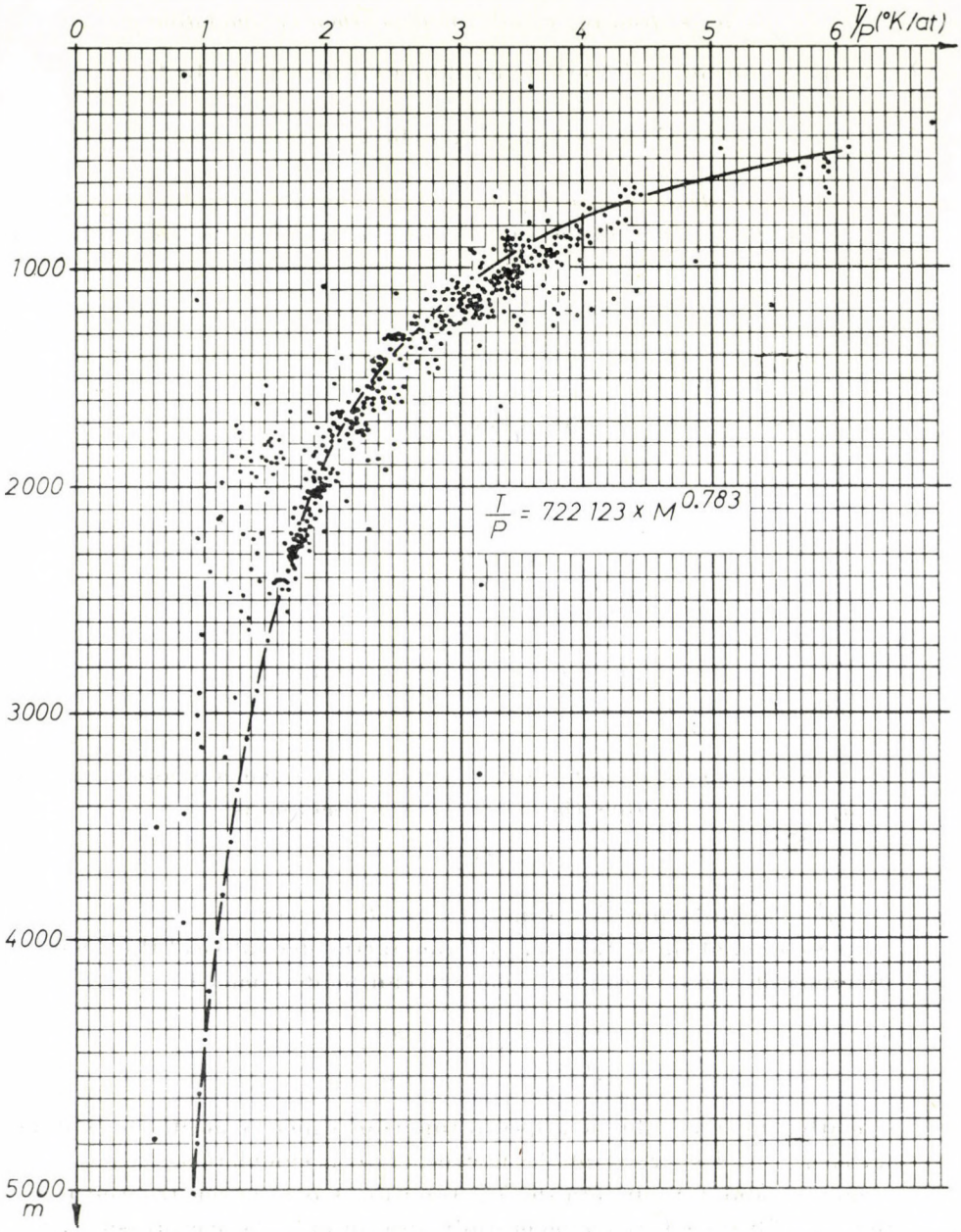

Fig. 9. Change of the accumulation values (T/p) as a function of depth (data from the Great Hungarian Plain)

Its value is about 6.0 in a depth of 450 m and reaches the critical value of 1.0 at about 4450 m. Below this depth, i.e. between 5000 and 6000 m it is dependent practically on the depth only. This limiting value indicates the change of the thermodynamic system.

The curve of accumulation values has also been constructed to overpressure areas. The run of the function is nearly the same, and can be described by the following equation (Fig. 10):

$$T/p = 5028027 \cdot M^{-1.05}$$

The two curves intersect in a depth of about 1400 m. In the lower section, below 3000 m their rise is the same, but the value of the overpressure curve becomes depth-dependent at 5000 m depth, with a value smaller by 2—2.5 tenth than that of the T/p curve.

Except one locality (Makó) the accumulation values are everywhere greater than one. (Where this is less than one, no productivity of industrial value will be found.) The accumulation values of Algyő, i.e. of the most productive area are as follows (by geological formations, T/p):

Upper Pannonian	2.1—1.3
Lower Pannonian	1.53 (average)
	1.65 (without the base horizon)
Miocene	1.45
Mesozoic	1.14
Paleozoic	1.3 (Deszk-horizon: 1.320).

The parallel rise below 3000 m relates to the fact that below this depth most of the reservoirs are overpressure ones. The T/p values of the reservoirs being most satisfactory for practical purposes lies between 1.5 and 3.5, this means a depth interval of 900 to 2700 m.

According to CIMASZEWSKI in the Polish Plain $T/p = 0.6$ is the lowest value at which accumulation can be expected. In Hungary this value has to be between 1 and 0.5 which could be expected between 5000 and 6000 m in the Great Hungarian Plain. In the thermodynamic sense the quotient of the T/p accumulation value less than one means that the possibility of volume change being only the highly restricted energy available derives from the change of entropy. This is to small to maintain the migration even at a geological timescale. The concepts of open and closed systems should be interpreted differently for deep-geological conditions that for the surface. As to my own opinion the deep-lying open system provides an unlimited heat transmission at minimal pressure difference, while the closed system is a state unsuitable for migration

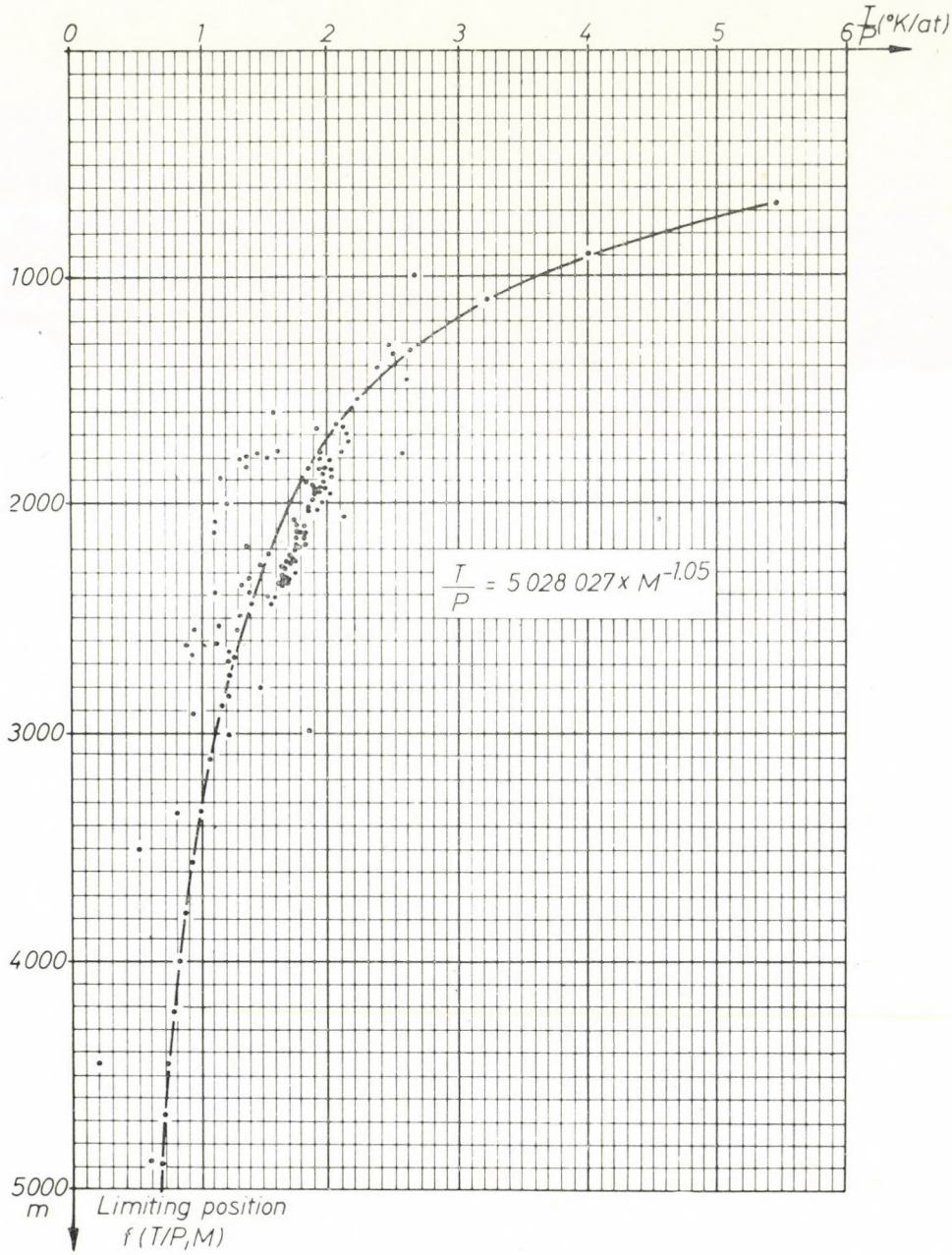

Fig. 10. Arrangement of the accumulation value (T/p) in overpressure areas (data from the Great Hungarian Plain)

and accumulates only the internal energies. The value of T/p around 0.5 to 0.6 indicates most probably a closed system where in lack of migration capacity no accumulation can be expected.

Geothermal data in hydrocarbon prognostics

In the geological recurrence the threshold value of formation temperature is of primordial importance. From the point of view of preservation or decomposition the thermal expansion of water and its pressure increasing effect in migration as well as the thermal history are the final and decisive factors.

The following basic data are to be collected and systematically processed:

— the interdependent values of depth, temperature and pressure,

— temperature and pressure gradients and the accumulation value as derived data,

— determination of paleo-temperature and evaluation of thermal history by geological formations.

These data should be processed in a generalized form for individual occurrences and basin units since in this way the question arising during prognostics from the geothermal point of view can be onswered, i.e.

— whether the temperature and heating time were sufficient to hydrocarbon formation or not,

— whether the thermal history made possible the preservation of hydrocarbons or not,

— what is the lower and upper depth limit of the recent accumulation zone.

The hydrocarbon occurences to be expected are to be calculated to this latter interval. This statement is valid, of course, only of the Neogene sequence. To evaluate the Paleogene and Mesozoic is more complicated since in these formations the role of geological changes should be evaluated by structural units and stage separately.

The peculiar geothermal position of Hungary is favourable from the point of view of hydrocarbon formation and migration expecially as for the Neogene. The heating time of these formations is relatively short, thus the major part of hydrocarbons could not be decomposed even at higher temperatures. Considerable hydrocarbon resources of older sequences can be expected only in the ares where the long heating time was accompanied by relatively low temperatures. This, however, does not rule out the possibility that Mesozoic and Paleozoic formations may be reservoir rocks.

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РОЛЬ ГЕОХИМИЧЕСИХ УСЛОВИЙ В РАЗРАБОТКЕ ПРОГНОЗОВ УГЛЕВОДОРОДОВ

Л. ВЁЛЬДИ

Р е з ю м е

Состояние термического равновесия является как правило динамичным и благоприятным для миграции углеводородов. Есть, однако, точка предела, где имеются только лишь органические вещества, неспособные к миграции.

Подчеркивается значение палеотемператур и так называемого «времени нагрева», воспринимаемого в геоисторическом смысле. Кроме того, рассматриваются критические «предельные величины температуры», влияющие на образование, миграцию, сохранность и разложение водородов, а также определение этих величин и связанные с ними проблемы.

Изменения термодинамических параметров в зависимости глубины анализируются с привлечением около 1400 данных, полученных для 86 поисковоразведочных участков на территории Большой Венгерской Низменности. В настоящее время величина среднего геотермического градиента независимая от местоположения (район, глубина), составляет 5,5°C/100 м, то есть 18 м/°C. Однако, реальные величины значительно колеблются, находясь между 26 и 10 м/°C. В различных районах Большой Венгерской Низменности наблюдаются всевозможные направления и тенденции изменений, например тенденции к увеличению, уменьшению, повторению и т. д. В разных частях бассейна те же самые значения изотерм встречаются в пределах нескольких тысяч метров по вертикали. В зависимости от формы изотермальных линий нефтегазоносные тела относят к различным температурно-глубинным горзонтам. Вместе с тем, при этом имеется в виду также и классификация по вещественному составу соответствующих толщ. Сравнение полученных для Венгрии данных свидетельствует о специальном геотермическом положении углеводородов в Венгрии.

Если приспособить к условиям Венгрии «предельную величину аккумуляции», предложенную для Польской Низменности, то нижний предел гидродинамических условий, позволяющих миграцию органического вещества, должен находиться на глубине 5—6 тысяч метров от земной поверхности. Следовательно, ниже этой глубины нельзя ожидать присутствие промышленных залежей нефти и газа.

Анализ истории развития геотермического режима любой территории имеет специфическое значение, так как при неизвестности зон образования и накопления углеводородов немыслимо разработка прогнозов нефтегазоносности, основанная на совеременных принципах.

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GEOTHERMICS AND METAMORPHISM IN THE MEDITERRANEAN TYPE OF PLATE-TECTONICS

By

E. SZÁDECZKY-KARDOSS

LABORATORY FOR GEOCHEMICAL RESEARCH OF THE HUNGARIAN ACADEMY OF SCIENCES, BUDAPEST

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The temperature of the oceanic plates is much higher than that of the continental plates even 10 or more million years after their formation by accretion. "Heat sink" around subduction is formed only at the great old oceanic plates ("Circumpacific type"). The surroundings of narrow oceanic plates are, however, heated from almost the beginning of the subduction ("Mediterranean type"). Since the lithospheric temperature of the Mediterranean Tertiary microplate area is increased by numerous quasi-oceanic hot subducted slabs, the depth values of magma chambers found by the K : Si petrochemical method are there systematically too high. The right values are about the half of the former calculated ones, enabling more realistic profiles for microplate areas.

Anhydrous reactions without volatile pressure require usually hypercrustal (> 10 kilobar) load pressure (e.g. the basalt → eclogite). Intracrustal reactions of the usual rock metamorphism are enabled mainly by the "effective" volatile pressure. Three kinds of volatile pressure depending on the confining load may be discerned: 1. Virtual or suppressed volatile pressure *(v.p.)* at high load pressure in closed system, without intracrustal metamorphism according to the mass action law. 2. Effective *v.p. at medium* load pressure in semi-closed systems, enabling (intracrustal) metamorphism. 3. Subnormal *v.p. at* low load pressure in open systems by continuous release of volatiles, without metamorphism.

Thus metamorphic reactions requiring some days or weeks in experimental vessels are delayed many million years in the closed sedimentary systems of the rapidly and deeply subducted slabs. Due to this "pressure paradox" experimental petrology neglecting porosity does not reflect crustal metamorphism.

The usual three kinds of metamorphism and their metamorphic facies give geological information on the rate of subsidence, depth (: load pressure), and temperature of recrystallization as well as on structural relations of the area: 1. In *burial metamorphism* the rate of subsidence is about 0.001—0.1 mm pro year and indicates low temperature (<500 °C), and low load pressure (i.e. depths less than about 18 km) as well as area outside of the subduction belts. 2. *Dynamometamorphism* indicates a rate of subsidence of 10—100 mm pro year, and recrystallization within the belts of subduction at medium to high pressure and medium temperature (fig. 3). 3. *Contact metamorphism* indicates rapid rise of magma in the order of km pro year mostly above subductions and recrystallization in shallow depth (<10 km) at about 300—650 °C.

Succeeding the high pressure metamorphism in subducted slabs, the rock is driven mostly into the virtual v.p. realm, hence dynamometamorphism is interrupted. The "low-pressure" part of the paired metamorphic belts along subduction is formed by obduction or by a special kind of structural relaxation (described in the text) which both return the slab in the realm of the effective v.p.

Examples of evaluation of Alpine, Hercynian, Caledonian, Late and Early Baikalian metamorphism for the Alps, Carpathians, Balkan and Dinarides are given.

Tectonic structure, plate tectonics, magmatism, metamorphism and heat flux in the East Mediterranean may be mostly interpreted by the clockwise rotation of the Moesian plate. Metamorphic basement of the Transylvanian Basin may represent a horizontally outpressed and NNW-drifted section between the East and South Carpathians of the former continuous and more straight Baikalian metamorphic belt around the Ukrainian shield and Moesian plate.

1. Geothermics of the Microplates. Principles of Mediterranean Plate-Tectonics

It is not by chance that the two main principles of the first comprehensive geodynamic model — the plate-tectonics — have been discovered at the eastern border of the Alps, where the most intensely folded mountain range of the Earth is joined to the Carpathians, the prototype of the basin arcs. Vienna, where the arch-type model of the subduction, the "Verschluckungs"-theory by AMPFERER 1906, and Graz, where the theory of the continental drift by WEGENER 1912 were first elaborated, lie at this border. About half a century later, these first, mostly disregarded theories were unified to a general geodynamic model, the plate-tectonics by the comprehensive geological-geophysical investigation of the great and relatively simple oceans. The intensive geodynamical interpretation of the Mediterranean on this background began, however, much later (DEWEY and BIRD, 1970; LAUBSCHER et al., 1971; SZÁDECZKY-KARDOSS, 1971). Soon it was discovered that the Mediterranean continent–continent collision differs much more from the Circumpacific ocean–continent collision, than it had been supposed at first. Contradictions due to the complexity of the area hampered the synthesis of the Mediterranean geodynamics.

Some new principles of the Mediterranean plate-tectonics described in this paper show that not the adaptation of Circumpacific plate-tectonics, but a new concept of Mediterranean drift is needed.

We start from the fact that Circumpacific Alpine age subductions controlled by the relatively unimpeded spreading are about perpendicular to the main strike of accretion, whereas the Mediterranean drift is ruled mainly by parallel translation and the succeeding collision of Africa and Eurasia, producing strike-parallel motion and rotation of the fragmented bordering microplates (Fig. 1). Before the Upper Cretaceous the Mediterranean represented a great ocean differing from the present ones by the quasi-parallel movement of the bordering continents, hence resulting the fracturation into microplates. Following the Middle Cretaceous the perpendicular collision of both great continents produce the mosaic motion and increasing rotation of the microplates.

The main consequence of this difference is the heating of the surrounding lithosphere by the subducted slab in the Mediterranean, opposite to the cooling

"heat sinks" by the Circumpacific subductions. Moreover, the scars of the Mediterranean subductions are permanently heated by the friction of the side by side longitudinal or rotary movement of the microplates, too, creating abundance in high-temperature ultrabasic melt products; opposite to the scars of the great oceanic plates cooled by the drifting away from the original

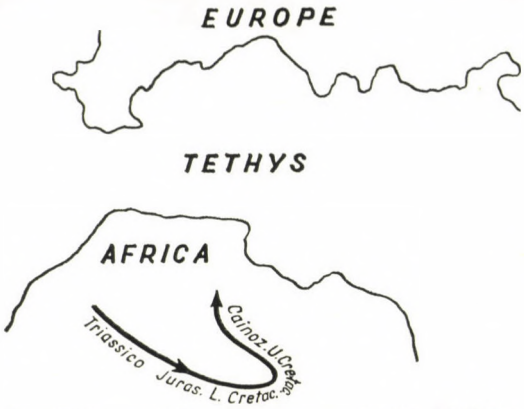

Fig. 1. Changes in the relative position of Africa to the "stable" Europe since the Triassic. Contours of Africa for the Upper Cretaceous

heat source hence producing mainly basalt crust with lower melting alkaline rocks.

In the Mediterranean, therefore, 1. a fourth type of plate border, the longitudinal one — beside those of accretion, subduction and transform — prevails. 2. The related microplates are less rigid than the great Circumpacific plates. Thus, Euler geometry concerning the motion of plates at the surface of spheres is less valid for Mediterranean microplates. 3. The width of the Mediterranean "oceanic" eugeosynclines, hence the depth of the subduction is mostly about 80 to 300 km, they represent "quasi-oceans" characterized by more intense sedimentation and often by some sialic remainder at the basement. 4. Accordingly, the volcanism related to the subduction also differs from the mighty uniform Circumpacific andesite–dacite belts. Petrochemical investigations elucidated that andesite magma is formed chiefly by the partial melting of sediments subducted to the depth of about 130 to 160 km, or less (see below). Since the subducted Mediterranean microplates often do not reach this depth, the andesite suite is there sometimes even missing. Reaching, however, this depth (e.g. in the Carpathian Basins) the andesite suite appears even as mighty masses, according to the great quantity of sediments accumulated by the erosion at both sides of the narrow eugeosyncline. 5. Different irregularities, e.g. deviation and bending backwards of the less rigid microplates in the depth are also more frequent in the Mediterranean.

The *mobility* as prerequisite for the drifting is a complex heat problem. It depends mainly on heat conduction and convection, on the one hand, and on friction, viscosity and volatile content, on the other.[1]

The drift and the starting accretion seems to be always preformed by crustal fracturing influenced mainly by surface factors. In the Circumpacific characterized by dilatation, fracture is deepened swiftly to the mantle provoking the basaltic partial melt of the oceanic crust. In the compressed Mediterranean, however, the expansion of fracture lines is mostly slow, the microcontinents being impeded by horizontal compression. Therefore, first a granitic partial melt along the lineaments is formed here ascending slowly as a belt of thermal domes. In this way the Alpine drift in the Mediterranean is preceded by the formation of Hercynian granite belts extending parallel to the borders of the microplates. Synchronously narrow quasi-oceanic (Prealpine) grabens may appear along theses borders, e.g. the eugeosyncline of Igal–Bükk.

Generally, the sinking lithospheric plate is regarded as a "very efficient heat sink", since "the plate should induce considerable cooling of the surrounding mantle" (LE PICHON, 1973) and only some million years later the surrounding mantle should be heated by the magma ascending from the depth. This statement is based on computations of the heat distribution of average continental plates on both sides of the subductions (HASEBE, FUJI et UYEDA, 1970; MINEAR and TOKSÖZ 1970, TOKSÖZ et al, 1971; SZÁDECZKY-KARDOSS, 1971; OXBURGH and TOKSÖZ, 1971, etc.)

When laborating with co-workers the Map of Metamorphites of the Carpatho–Balkan–Dinaride area, I found, however, that this concept of us is inconsistent with the temperature data of metamorphism. If early subduction were characterized by a heat sink, the depth of metamorphism of the subducted slab would be greater than in the undisturbed "normal" platform. In reality, however, the opposite is true. The characteristic temperature of about 500 °C for the amphibolite facies lies, e.g. in a depth of about 29 km in the normal continental plates. Thus, in the belt of subduction amphibolite facies should crystallize in a depth greater than 29 km. Experiments reveal, however, (Fig. 3) that amphibolite facies is formed already at 3 to 9 kilobar pressure, i.e. in a depth of about 12 to 24 km. (The depth of 29 km is according to some authors already the realm of eclogites.)

[1] In general, the drift mobility of plates is due to the lubricating vapor-rich melt film (abbreviated "vapor pillow") at the bottom of the lithosphere. The volatiles of this lubricant derive from the subducted sediments, the silicate melt, however, mainly ascends from the astenosphere (SZÁDECZKY-KARDOSS, 1971).

At the sides of the drifted plates an other lubricant, the serpentinite is formed from the ultrabasic rocks by the mainly descending water along the lineaments. Serpentinite becomes, however, thermally dissociated in a certain depth of the subductions and is no more stable at the bottom of the lithosphere, the temperature being there more than 1100 °C. The lubricant of the obductions and overthrusts are mostly clay slates and salts.

It is usually admitted that some mobility in other rocks, too begins at about half of the (Kelvin) melt temperature.

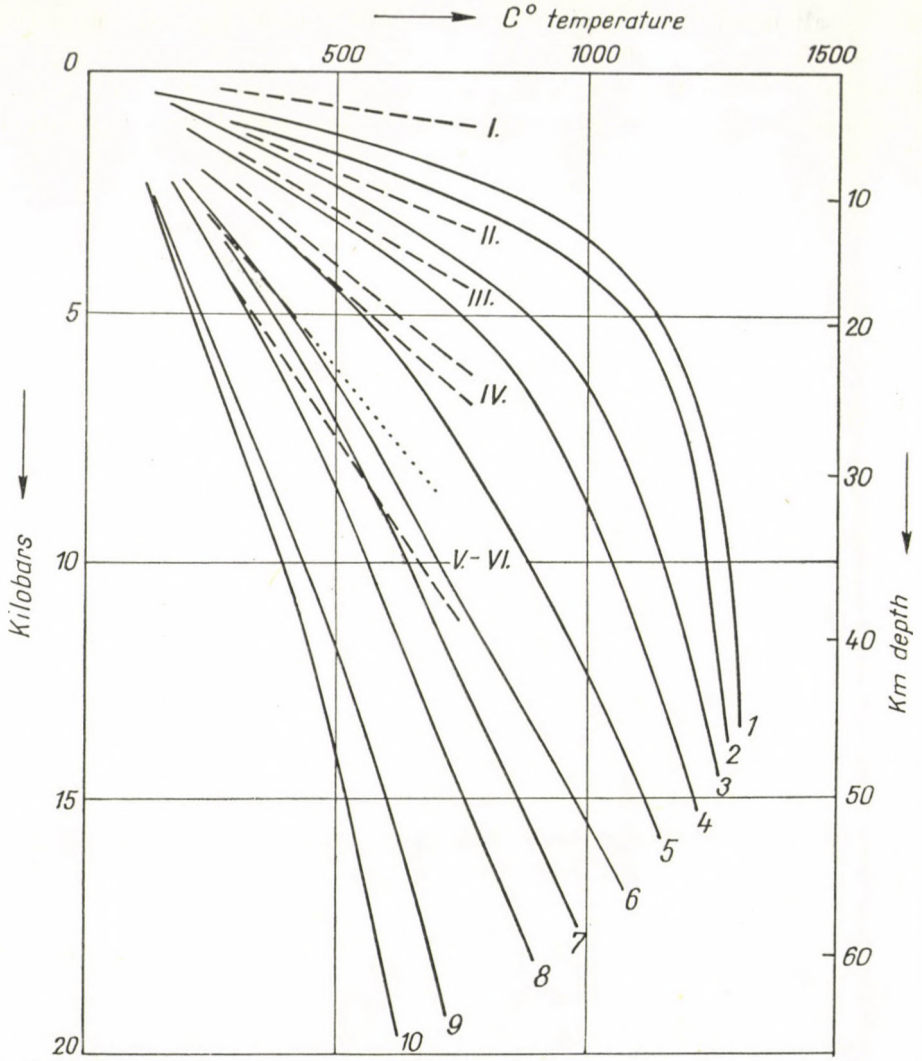

Fig. 2. Heat distribution in the subducted slabs and relation between metamorphic facies and rate of subduction. I. Hornfels facies. II. Low-pressure amphibolite facies. III. Medium-pressure amphibolite facies. IV. High-pressure amphibolite facies. V. Glaucophane schist facies. VI. Eclogite

1. 1.25 m.y.-old oceanic plate, 100 km width, drifting rate 8 cm/y; 2. 2.5 m.y.-old oceanic plate, 100 km width, drifting rate 4 cm/y; 3. 10 m.y.-old oceanic plate, 100 km width, drifting rate 1 cm/y; 4. 12.5 m.y.-old oceanic plate, 1000 km width, drifting rate 8 cm/y; 5. 25 m.y.-old oceanic plate, 1000 km width, drifting rate 4 cm/y; 6. 150 m.y.-old oceanic plate, 2000 km width, drifting rate 4 cm/y; 7. "Oceanic plate" according to WYLLIE (1977); 8. "Oceanic plate" according to GREEN and RINGWOOD (1967); 9. continental plate (shield) according to WYLLIE (1977); 10. continental plate (shield) according to GREEN and RINGWOOD (1967)

The solution of this paradox lies in the fact that the temperature distribution on both sides of the subduction is different. The temperature of the subducted oceanic plate is much higer that that of the "normal average''

continental plate. Oceanic plates formed by the accretional magmatism at about 1300 °C, cool slowly according to the small heat conductivity of rocks. Especially hot are the narrow Mediterranean (quasi)-oceanic belts. Hence, the estimation of oceanic geothermal gradient by RITTMANN, RINGWOOD and

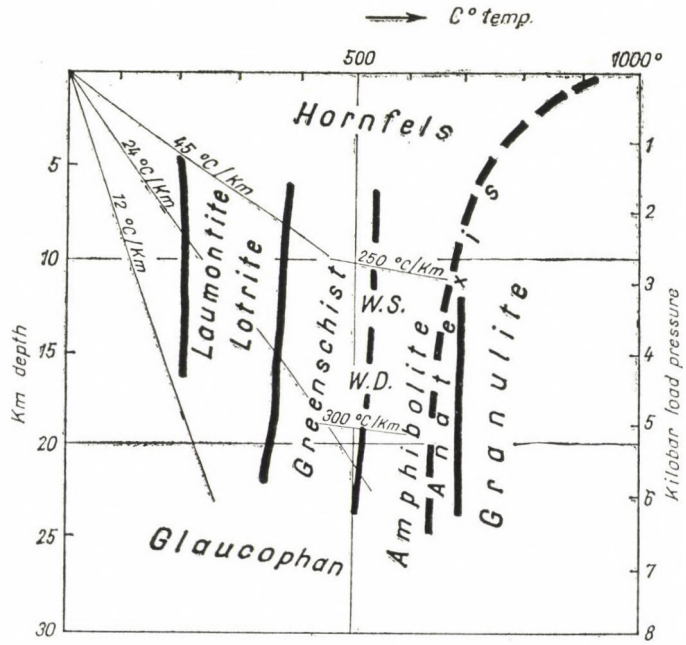

Fig. 3. Distributions of the metamorphic facies in the *p-t* field, models 1950 and 1978, and geothermal gradients for thermally mantles gneiss domes (e.g. Agout dome; W. S. Water-saturated systems, W. D. Water-deficient system after DEN TEX 1975

GREEN, WYLLIE, etc. are roughly valid for the great oceans, but not for the narrow Mediterranean quasi-oceans. E.g. the temperature in a depth of 25 km of an oceanic plate is about 1000 °C after 10 million years and 100 km far from the original belt of accretion (i.e. at a drift velocity of 1 cm/year). It is even 1160 °C, 1000 km far from the accretion after 1.25 million years, i.e. by a drift velocity of 8 cm/year (BOTTINGA, 1972). In a depth of 10 km the cooling velocity of a 100 km broad oceanic plate is 20,000 years/°C at 1 cm/year drift velocity, and 2500 y/°C for a 1000 km broad ocean by 8 cm/year drift velocity.[2] — Summing up, the narrower is the (quasi)-oceanic plate, the higher is its

[2] In the Central Alps radiometric determination resulted in even still smaller cooling rates (WAGNER, RIEMER, JÄGER, 1977). E.g. 23,000 years/°C for the last 12 m. years and 55,000 y./°C for the time interval 12 to 15 m. y. in the Bergell Massif and 31,000 y./°C for the Oberwallis, even 78,000 y./°C for the Monte Rosa.

average temperature (Fig. 2). The border between asthenosphere, i.e. the partially melted mantle and lithosphere lies the higher, the smaller and warmer the oceanic plate is

The Mediterranean represents a network of microplates and subductions formed by relatively narrow and thin hence hot quasi-oceanic plates. The temperature of partial melting in the quasi-oceanic plates is reduced also by

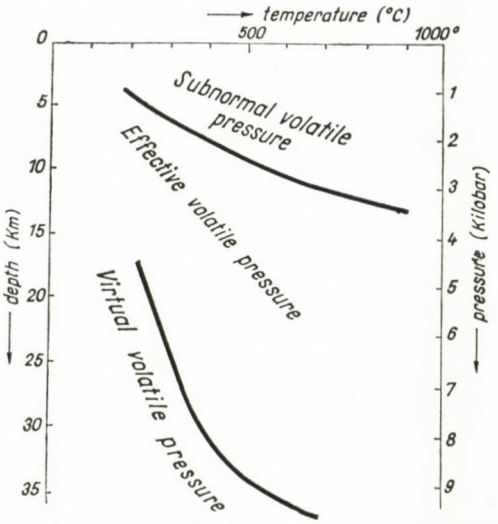

Fig. 4. The realms of volatile pressure in the *p-t* field

the great volatile content of the descending slab deriving from the relatively high amount of the (quasi)-oceanic sediments.

Thus, in the Mediterranean area the lithosphere is often predominantly heated by the subducted slab from the very beginning — opposite to the general "heat sink" conception. Cooling is restricted in space and time for the 1—2 km small strip along the sedimentary cover of the about 50—70 km thick slab. Since the strip is heated from both sides, the temperature reaches there 200 to 300 °C in some million years. If in this time interval the depth of about 25 to 30 km, i.e. the pressure of 7 to 8 kilobar is attained, the high-pressure low-temperature (blue schist) metamorphism takes place, already in about 1.5 m.y. The crystallization of amphibolite facies in a depth of 29 km instead of glaucophane facies is not likely.

According to these extremely high quasi-oceanic plate temperatures, the scars of subduction are partly characterized by positive geothermal anomalies. At the Kapos line section of Transdanubia, Hungary geothermal gradient of 13 to 18 m/°C occurs (NÉMEDI, VARGA, 1977), and similar values are known along the Balaton line, too (BOLDIZSÁR, 1968).

The *prospection for hydrocarbons* around the scars of the subduction is not sigificantly influenced by this new geothermal conception. We posses already a wealth of exact data concerning carbon and hydrocarbon contents of old metamorphized rocks (SIDORENKO and SIDORENKO, 1975), indicating the continuous release of hydrocarbons from subducted slabs and from the metamorphic basement. This is especially true for the mantle domes ("mantle diapirs", e.g. in the Pannonian basin) where the temperature increase is very significant and longlived. Hydrocarbons deriving from the metamorphic basement and accumulated in the young roof sediments also represents sources of the present oil-field.

The evaluation of the *depth of magma* chambers by the Si : K ratio must be significantly modified for the Tertiary Mediterranean. The potassium content of the volcanites related to subducted slabs roughly increases with the increasing depth of the magma chamber (NINKOVICH and HAYS, 1969). This is demonstrated on the main median Mediterranean and on different Circumpacific subductions. The increase of the potassium content with increasing depth of magma chamber may be interpreted mainly by the contamination of the ascending magma by the most mobile component potassium of the overlying lithosphere, being (roughly) proportionate to the thickness of the underthrusted lithosphere around the volcanic vent. (The mobility of the elements increases with their increasing ionic potential.)

The mobility of the elements, thus the intensity of the contamination depends, however, also on the temperature of the lithosphere. The lithospheric temperature is increased by the small, hot, young, subducted slabs. Especially high is the temperature increase — hence the potassium content — in the Tertiary Mediterranean surrounding area characterized by many subducted quasi-oceanic hot narrow basaltic slabs. The contamination increases especially with the decreasing rate of the last actual subduction (table next page) because the surrounding lithosphere becomes heated by the slow subduction already at the time of volcanism. In the Mediterranean, especially in its Tertiary surroundings both phenomena prevail, thus its volcanic rocks are high in potassium as recognized already in 1920 by P. NIGGLI ("potassium or Mediterranean magma province"). The relation between rate of subduction and volcanic potassium content was recognized by MYASHIRO 1972, 1973.

In this way the former reckonings concerning depths of Tertiary magma chambers in the Tertiary Mediterranean are too high. Values about of the half of the former ones are much nearer to reality. The profiles of subductions, e.g. in the Carpathian Basins becomes so much plausible.

In extremely heated crusts even metasomatic alkali magmatism occurs. The nephelinsyenite mass of Ditró appears at the border of the longterm drifted Baikalian basement of the Transylvanian basin (Chapter 4).

K_2O contents (%) according to MYASHIRO, *1972—73*

	Rate of subduction		
	2 cm/year	3—6 cm/year	8—9 cm/year
at SiO_2 50%	1.6—2.6	0.5—1.1	0.2—0.7
at SiO_2 60%	3.6 <	1.1—1.9	0.6—1.1
Examples	Calabrian Macquerie	Aleutian, Ryu-Kyu	Tonga, Bonin, NE Japan

For *island arcs* a more complex explanation for the increasing volcanic alkaline content toward the continent was elaborated by KUSHIRO, in accordance with his experiments. He found that in the possible upper mantle material the liquidus boundary shifts from the sialic-rich to the silica-poor side with the increase of the confining pressure. The pressures under which tholeitic and alkaline basaltic magma are produced shall be, according to these experiments, ca. 10 kb (corresponding to 30—40 km depth) and 20—30 kb (corresponding to 70—100 km depth), respectively. This is about one third of the depth ranges of the mantle earthquake zone, where the primary magma should be formed. Therefore, KUSHIRO proposes that during the partial melting volatile and alkaline components tend to stay in the liquid, so that the volatile concentration in the magma increases through the upward displacement, but finally in a "critical depth of magma rise" the pressure of the volatile components in the magma exceedes the ambient pressure, so that fissures are formed, enabling a quick ascension of the magma. This critical depth is the mentioned 30—40 km for tholeiite magmas and 70—100 km for the alkaline basalt magma, whereas the depths of their initial melting are 130—150 km and 250—300 km, respectively (see also SUGIMURA and UYEDA, 1973). This model may represent perhaps a secondary factor of the potassium enrichment. The experimental findings of KUSHIRO are, however, questionable.

2. The Evaluation of Metamorphites for Geothermics and Plate-tectonics

The exact theoretical determination of the *p-t* values of the metamorphites — already before the concerning experimental works (SZÁDECZKY-KARDOSS, 1950[3]) — was intended to elucidate geological and geothermal knowledges of the depth. Now, after almost 30 years the *p-t* values of meta-

[3] The theoretical deduction before the instrumental possibility of high-pressure, high-temperature experiments was based on geological data received by the sensitive peat coal graphite sequence up to about 3 kilobar pressure, i.e. 10 km depth and then by dividing of metamorphic facies up to the temperature of anatexis according to their relative frequency.

morphic facies isograds (Fig. 3) and subgrades are thoroughly known and the petrographic nature of metamorphites of almost all continents is mostly exhaustively described, too, even many sidedly synthetized by new maps of metamorphites. The informations contained by the metamorphites are, however, predominantly still unexplored. This is mainly due to the fact, that the paradox is not elucidated, why the metamorphic reactions connected to the critical temperature and pressure at the borders of the equilibrium fields of different metamorphic facies require only few days in experimental vessels but are in the nature postponed often more million years after reaching the same critical values. By the discovery of the paired metamorphic belts along the subductions (MIYASHIRO et al., 1965, 1973) this paradox became even more embarrassing.

The possible solution given in this chapter enables the junction between results in experimental and structural geology. (For better intelligibility metamorphic subgrades are omitted in the followings.)

Let us to start by the statement that a given metamorphic grade indicates *p-t* values only in the presence of volatiles. Dry lithostatic pressure (i.e. confining or load pressure[4]) alone does not produce true polyphase recrystallization in the lithosphere. Without some volatiles no crustal metamorphism is possible even in the "dry" eclogite facies. Metamorphism takes place only if the critical temperature and effective volatile pressure characterizing the given metamorphic grade are realized *synchronously*, a condition which is often not realized in the orogenic belts. Thus below the calcite dissociation temperature the great limestone bodies (e.g. of the Southern Alps) are mostly almost indifferent to the changes in depth. A metamorphosed and already devolatilized "dry" rock becomes overprinted by an other metamorphic grade when sufficient new volatiles are released or introduced in the rock mass *and* the temperature–pressure conditions of the new metamorphic grade appear synchronously. Without these conditions only weathering takes place by the introduction of volatiles.

The simple presence of volatiles in the lattices of the minerals and the elevated temperature and pressure are, however, still not sufficient conditions for metamorphism. The volatiles must be *released* from the lattices and this

[4] Load pressure is mostly hydrostatic since it is approximately equal in all directions. In schistose rocks this condition is, however, not realized. Load pressure represents a limit value which cannot be substantially exceeded by volatile pressure, since rock becomes fractured by volatiles which slip away. So volatile pressure may suddenly drop. In contrast, the temperature values mostly are changing slowly due to the small heat conductivity of the rocks. Horizontal tectonic forces do not significantly increase the lithostatic pressure, i.e. it does not alter the depth values of the metamorphic grade since it is largely consumed by the plastic flow of rocks. The compressive strength of rocks is about 3000 kg/cm^2 (mostly 1000 to 5000 kg/cm^2) and decreases with increasing hydrostatic pressure, as provided first by KÁRMÁN (1911) and then investigated thoroughly for special cases by many other scientists (see e.g. KNOPF and INGERSON, 1938).

is not the case at a too great confining (load) pressure. By high load pressure volatiles are compressed more and more, their relative surface and "effective" pressure decrease, too. In this case a self-stopping closed system during progressive metamorphism (i.e. for reactions producing volatile by-products) according to the mass action law is formed. Thus, the increase of load pressure beyond a certain limit suppress progressive metamorphism, i.e. delays recrystallization. Hence, metastable states and pseudo-equilibria are much more frequent during the *p-t* changes of rocks as admitted usually.[5]

Hence, the role of the load (or lithostatic) pressure is twofold. It maintains the volatiles in the rock, assuring the development of volatile pressure required for metamorphism but — at too high values — hampers the thermal release and the migration of volatiles. This *pressure paradox* means that the conditions for metamorphism are tight, they shall unify two opposite requirements; a certain accumulation of volatiles to produce the activating volatile pressure on the one hand, and the release of the surplus volatile reaction products enabling the reaction according to the mass action law, on the other. In small-volume experiments this is mostly not conspicious since during the high-pressure state the volatile by-products (especially the hydrogen) are transiently absorbed by the pore remaining in the probe and possibly by the wall of the experimental vessel, too.[6]

On the contrary, at low load pressure (in shallow depth) rock pores form often continuous drainage system which enables the complete release of the volatiles. Hence, metamorphism is again impossible lacking sufficient volatile pressure. In this case metamorphism is impeded because the rock represents a completely open system contrary to the former case, when it is impeded by the complete closeness of the system. Thus metamorphism requires an optimal (minimal) flow of volatile realized in great rock masses only by special conditions.

Thus, three kinds of volatile pressure are to be discerned: 1. *Virtual volatile pressure* bound in the lattices thus without meaning for metamorphism.[7] It occurs in closed system under high load pressures. 2. *Effective volatile pressure* sufficient to activate also inert compounds thus involving meta-

[5] Before our experiments the role of the pressure concerning diffusion was contradictorily judged. According to READ (1957) pressure promotes, according to GIRIFALCO (1964) it decreases the ionic diffusion.

[6] Due to the pressure paradox the thesis that "the pressure of the fluid phase is approximately equal to the load pressure" led mostly to erroneous conclusions. In experimental petrology the role of rock porosity in metamorphism was not duly considered.

[7] It may be admitted that the limiting load pressure at which effective volatile pressure is completely suppressed (i.e. virtual volatile pressure is zero, reactions between water-bearing minerals are impossible, difference between load and effective volatile pressure is maximum) is mostly attained at about 10 kilobar and 400 °C, where the boundary between stability field of glaucophane facies and eclogite become indefinite, hence interpreted until now in very different ways.

morphic reactions.[8] It occurs in semi-closed systems. 3. *Subnormal volatile pressure* insufficient to activate metamorphic reactions. It appears in open systems, mostly at low load pressures.

In this way only a relatively narrow realm for burial and dynamometamorphism exists in the temperature–pressure field between the virtual and the subnormal pressure (Fig. 4 p. 175).

The field of the true *burial metamorphism* is the low-pressure, low-temperature realm represented by sedimentary basins. The rate of sedimentation is (also in the basins) mostly 0.003—0.3 mm/year. Sediments represent here a semi-closed system, where the original volatiles of the sediments are relatively well preserved. Metamorphism gradually progresses from the zeolite facies partly to the greenschist facies, but does not surpass the latter. In the depth at about 10 km (2—3 kilobar load pressure) the border of the zeolite and greenschist facies may be reached only in about 10^8 years. The age even of the greater sedimentary basins is, however, limited by the tectonophase of $2—3 \times 10^8$ years. Zeolite facies may often pass over in glaucophane facies, due to lineaments or to subduction within the basin.

Burial metamorphism occurs in subsiding continental and postorogenic areas (cratogeosynclines), too. Sedimentary basins of the great old continental plates, such as Swaziland, are examples of this type of weak metamorphism.

Burial metamorphism appears in the basins of the mantle domes, too (e.g. Pannonian Basin, Chapter 4) surrounded by the orogenic belt. The high temperature is there attained in relatively shallow depth. Thus, rocks must be metamorphosed in the low-pressure (Abukuma) facies series and relatively thin prehnite-lotrite, greenschist, even amphibolite facies may follow in tight sequence towards the depth.

Dynamometamorphism occurs mostly in the subducted slabs. The increase of the depth is there about 1—10 cm/year, i.e. greater by about 2—3 orders of magnitude than in the burial metamorphism. This provokes an intense devolatilization of the subducted sediments maintaining a coherent open-pore system near the surfaces. The volatile pressure remains therefore often subnormal. Metamorphism is realized only by a rapid subduction into great depth. Mostly glaucophane-schist facies metamorphism occurs in a depth of 25—30 km (load pressure 9—10 kilobar) at 200—300 °C within about a million year. A steeper and still more rapid subduction lies in the realm of the virtual volatile pressure without metamorphism.

By continued subduction the sediments are finally partially melted, thus the magmatism deriving from the subducted slab begins. This produces

[8] Theoretically, the effective volatile pressures increases abruptly at the bivariant equilibrium boundaries of the facies stability fields which are characterized by release of volatiles. In geology this is, however, practically imperceptible since the critical temperature changes gradually in the successive parts of the rock.

the *contact metamorphism* in the overlying lithosphere and in its sedimentary cover.

By special tectonic events, however, the subducted sediments are metamorphized in greenschist and amphibolite facies, too. Theoretically, the greenschist and amphibolite facies metamorphism may appear in moderately steep slab by medium-load pressure. This metamorphism by "direct" subduction, however, seems to be still questionable.

A sure kind of formation of these facies is the *relaxation* of the subducted slab which may occur also in the slabs producing previously glaucophane metamorphism. Relaxation is created by a fine fracture system suitable for carrying away the surplus volatiles but preserving the required amount for the effective volatile pressure. Folding, overthrusting, isostatic subsidence or uplift are mostly unsuitable for this condition. More suitable is a small change in the direction of the (main) stress field producing a moderate release of volatiles. This may produce even an about synchronous dynamometamorphism in different metamorphic facies in a thick rock sequence. An example of this process concerning the Alps is described in the next chapter. After the relaxation the slab, in which previously glaucophane facies was produced, is transformed into a paired metamorphic belt.

According to the interpretation of MIYASHIRO, only the high-pressure metamorphism of the parallel belts is formed in the subducted slab, usually on the oceanic side. The low-pressure belt is assumed to have been crystallized on the continental lithosphere below the rock complex of the island arc volcanism. The two belts are often in direct contact separated by a main tectonic line (Japan, New Zealand). The direct contact is interpreted by MIYASHIRO as possible result of the oceanwards shift of the descending slab, or of large strike-slip movement along this tectonic line. The two belts are formed either about synchronously (e.g. 100 to 120 m.y. in Japàn) or separated by a greater time distance. E.g., in the Alps the age of the high-pressure belt is about 54 m.y., that of the low-pressure belt is 38—40 m.y. (FREY et al. 1974, 1976). In this sense low-pressure metamorphism is connected to the island are volcanism.

But how forms a low-pressure belt at plates, where no island arc exists? One possibility, the relaxation was discussed above. An other possibility is the *metamorphism by obduction.*

According to radiometric age determination low-pressure belt metamorphism does not take place during active subduction. The minimum volatile pressure required for the lowest-grade dyanamometamorphism is 2 to 3 kilobar, meaning a depth of about 10 km for the load pressure, but much more depth for volatile pressure, as volatiles continuously escape in the disturbed belt of subduction. In flat lying and slow subductions the relatively rapid increase of temperature by the sluggish increase of depth promotes the escape of volatiles. In this case the depth required for medium- and high-pressure meta-

morphism 20 km is at least. Erosion alone does not suffice to bring metamorphites to the surface from this depth. Uplift — overthrusting — especially obduction or some kind of faults are also required for this.

Steep-angle subductions, where the soon attained high load pressure supresses the effective volatile pressure are also unfavourable for low-pressure dynamometamorphism.

Thus paired belts of metamorphism may be formed in the way too, that the high-pressure belt crystallizes during the subduction along the trench, the low-pressure belt, however later, during the obduction in a branch of the same slab, which produced the high-pressure metamorphism, too. Since obduction takes place often nearer the continent on a new pathway opened by the increasing mechanical effect of the subduction. It may dip even opposite to the main subduction.[9]

The *micro-texture (fabrics)* of the metamorphite may give information concerning crystallization by subduction or obduction. The porphyroblasts in the metamorphites of subducted rocks are rotated counter-clockwise. On the contrary, clockwise rotation of the porphyroblast indicates obduction. This is the theoretical forecast but many observations are required for the verification and deepening on this concept related to the mostly questionable problem of the syntectonic ("rotational") or pretectonic ("rotated") origin of metamorphites (SPRY, 1969).

Regional metamorphic belts are often characterized by one, sometimes more thermal axes, the temperature of metamorphism increasing toward these axes from both sides. Thermal axes are accompanied sometimes by syntectonic granite masses. This thermal structure seems to reflect the original basin structure of the "geosyncline" characterized by deepest hence most heated median axis. This type may represent even the ***simple geosyncline*** metamorphism without significant subduction.

In this way metamorphites inform us always on the temperature and depth of the active stage of recrystallization. But some indications concerning the ***rate of subduction and the width of the subducted oceanic plate*** may also be obtained by the metamorphites. The rate of subduction is not equal to the rate of drifting. It may be different, even opposite, (i.e. obduction instead of subduction) in the different parts of the same slab. The rate of the topmost sequence of the slab is the greatest by entering the depth when it travels the

[9] Thus in the steep slabs two stages of the "low-pressure belt", i.e. of the greenschist-amphibolite facies dynamometamorphism must be discerned: *1.* A long inactive closed and metastable stage characterized by the increase of load pressure and temperature but without the development of the sufficient effective volatile pressure (volatile pressure "virtual"). *2.* The mostly short active stage characterized by the rapid recrystallization of the transitionally opened system realising the surplus volatiles (volatile pressure "effective" and approximately equal to the load pressure). A new stable equilibrium is approached within some weeks according to the instant *p-t* conditions.

greatest exterior distance. Later this topmost plastic sedimentary sequence reaching a greater depth may be pressed out from the slab, i.e. becomes obduced in the opposite direction due to the accumulation of the increasing pressure in the slab.

At a small rate of subduction the increase of the depth and pressure is slow relative to a given increase of temperature. The specific increase of load pressure and depth (relative to that of the temperature) grows with increasing rate of subduction. Thus the pressure–temperature *difference* of the subducted rock represents — at a given angle of subduction and a given original heat distribution — a measure of the rate of subduction:

$$V_s = f(p - t)$$

If α is the angle of the subducted plane to the horizontal and Δ the original heat distribution measured roughly by the original temperature difference between the metamorphized rock sequence and the surroundings (slab and underthrusted lithosphere) during the subduction, the rate of subduction may be expressed by the approximate formula:

$$V_s = f\frac{(p - t)}{\cos \alpha \Delta}$$

At present six ranks of the $(p - t)$ differences may be distinguished. In the sequence of increasing $(p - t)$ differences these ranks are: 1. hornfels facies, contact metamorphism; 2. low-pressure grade amphibolite facies; 3. medium-grade amphibolite facies; 4. high-grade amphibolite facies; 5. glaucophane schist facies; 6. eclogite metamorphism (Fig. 3).

Before an exhaustive determination by mathematical demonstration of the $V_s = f(p - t)$ relation the quantitative relation between pressure grades of metamorphism and rates of subduction may be empirically estimated. It is known from the measurements on Pacific and Atlantic Tertiary plate-tectonics that drift velocities lie mostly between about 1 cm/year to 10 cm/year. In first approach the rate of subduction — as average of the whole slab — is considered to be roughly equal to drift velocities. Thus, the drift velocities of the 6 ranks may be estimated as follows:

— below 1 cm/year: hornfels facies, contact metamorphism;
— 1 to 3 cm/year: low-pressure and paracrystalline "Abukuma" series;
— 3 to 6 cm/year: "paracrystalline" medium-pressure facies, Barrovian series;
— 6 to 8 cm/year: "paracrystalline" high-pressure amphibolite facies;
— 8 and more cm/year: blueschist facies ((with or without eclogite)
— obduction at different rate: "pre- and postcrystalline" greenschist and amphibolite facies.

Obviously, this method refers to the recrystallization by straight pathway subduction and is not valid for metamorphic facies formed, e.g. by relaxation.

3. Metamorphism in the East Mediterranean

According to our map of the Carpatho–Balkan–Dinaride metamorphites the pure *Alpine-age monometamorphism* occurs mainly in the Bükk Mts, and is especially widespread in the Dinarides. In the Bükk Mts Alpine-age metamorphism appears in prehnite-lotrite grade (ÁRKAI, 1973); in the Dinarides presumably also in the laumontite-prehnite grade, both indicating a some km deep shallow crystallization. In the Dinarides (e.g. Novi Sad) also glaucophane-schist facies occurs, indicating high-pressure crystallization, too. The same is valid for the Southern Alps. The Bükk Mts may be interpreted as the part of the Igal-Bükk eugeosyncline outpressed in the maximum pressure zone between Moesia corner-Bohemian Massif. The Bükk-Igal eugeosyncline (WEIN, 1969) contains rather well preserved Carboniferous-Triassic sedimentary rocks and basic volcanics, partly also ophiolites showing that this is a remnant of the old Tethys, now subducted and horizontally translated to NE, but mostly not obduced. According to WEIN 1978, the Pennine Zone may extend even to the Zagreb—Hernád line.

On the contrary, in the (Northern) Alps — especially in the intensely investigated Swiss Geotraverse — the grade of the 35—40 m.y.-old lepontine metamorphism increases from North to South, reaching amphibolite grade at the Periadriatic (Insubric) Line (NIGGLI 1970; FREY et al. 1974, 1976). There it falls suddenly to the laumontite-lotrite-prehnite grade being predominant in the Southern Alps. At the Insubric Line the maximum metamorphic grade indicates more than 7 kilobar pressure and about 650—700 °C temperature. This corresponds to a crystallization in a depth of about 27 km (Fig. 3) indicating an orogenic rock accumulation by overthrusts, or in the term of plate tectonics a subduction. It may be admitted, that succeeding the Jurassic accretion of both Penninic oceans (FRISCH, 1977), the subduction of these oceanic plates begins in the Upper Cretaceous (i.e. before about 100 m.y.) due to the northward drift of Africa. The oceanic volcanic sedimentary sequence is metamorphized in the glaucophane facies 65 m.y. ago getting in a depth of 25 to 30 km (7 to 9 kilobar confining pressure) and heated synchronously to 200 to 300 °C. Further subduction in the next 25 m. years does not provoke low-pressure metamorphism, (since the migration of volatiles is inhibited by the increasing pressure). Volatiles become mobilized by the 35—40 m.y. "lepontine" structural events, producing a general low-pressure metamorphism. The structural event causing the overall lepontine metamorphism in the Alps is presumably due to the clockwise rotation in the Rheingraben

domes at about 38—45 m.y. (LIPPELT et al., 1974) loosening thus the subducted slab of the opposite counter-clockwise rotated Central and Western Alps. The loosening results in fracturing of the surrounding rocks, hence activates the suppressed volatiles and produces low-pressure metamorphism. The grade of the lepontine metamorphism depends on the depth, i.e. on the temperature within the subducted slab: at the top laumontite (?) prehnite-lotrite facies, in the middle greenschist and at the bottom amphibolite facies appear. The further northward drift of Africa (relative to Europe) squeezes out these metamorphites obduced and northward overthrusted in the sequence of the decreasing grade of metamorphism.

Alpine plate-tectonics, at least in the East Mediterranean, is preformed and introduced by the Hercynian lineaments characterized by the Hercynian granite belts (see below). Hence, the "Alpine" tectonic cycle in the East Mediterranean may be characterized — considering also the investigations of JANKOVIČ 1976 by 5 main phases:

1. Formation of a Late Hercynian lineament system reaching a depth down to about 70 km and producing *granitic thermal domes*.

2. Beginning of the accretion mainly along the Hercynian lineaments by slow ascension of basaltic partial melt, contaminated by the sial, hence transformed to *spilite, keratophyre*, etc. They contain hydrothermal polymetallic ores, Hg, As, and at the surface grabens with sedimentary massive sulphide ores and iron oxide deposits.

3. The fracture system widened into a quasi-oceanic eugeosyncline with faster ascending thus uncontaminated *tholeiitic* volcanism and at the longitudinal drifts with relatively frequent *ultrabasites* and Cr, Pt, Ni, Cu ores.

4. Early subduction characterized by the mixing of oceanic sediments with basalts thus by *calcalkaline and basalt magmatism* and Cu—Mo ores.

5. Main subduction producing *alkaline and acid magmatism* and Hg, As, Sb, Te, Tl, later Pb and Zn ores.

Concerning the *Hercynian and older plate-tectonics*, first the microcontinents must be replaced in their original Prealpine positions.

In the Mediterranean the *Hercynian* structural cycle and metamorphism differ fundamentally from the Alpine one first by the granitoid lineament system representing germs of the Alpine belts of accretion. Opposite to the belts of accretion of the great oceans forming mid-oceanic ridges, in the Mediterranean (without wide distance spreading) the Hercynian granitoids represent the future borders of microcontinental plates. The granitoid masses of about 10—30 km diameter and often in a similar distance from one another are related with the "*mantled gneiss dome*".[10] According to petrochemical data

[10] Mantled gneiss domes were first described in 1945 by Eskola in the Karelidic belt on the Baltic Shield and then reported from the Variscan Pyrenees, from Central France, from the Alps, from the Appalachians, in the Basin and Range Province and in other continents,

the granite melting appears first mostly in a depth of about 30—70 km. Because of the low melt temperature of granites they form long-lived lineaments with granitoid masses of highly different (Baikalian to Alpine) age. The Hercynian melting in the Pannonian Basin begins already in the Early Paleozoic. The main alkali mobilization is about Early Carboniferous and a third mobilization, related to the formation of quartz-porphyries is about Early Permian (400, 330 and 270 m.y. at least in the Mecsek Mts, according to KOVÁCH et al.). The quarz-porphyries seem to appear already perpendicular to the granitoid belts, along a transversal fault system (e.g. Villány line, KASSAI, 1977). It is even likely that these granites represent partly remelted products of older, Proterozoic Baikalian granites (JANTSKY). In this way in the Pannonian Area at least 3 SW—NE granite belts appear (Velence Mts, Balaton line, with a young Alpine age continuation to the West mainly along the Insubric Line; the Mecsek—Kecskemét belt; and the Battonya belt (SZALAY, 1977).

Hercynian metamorphic belts connected to these lineaments form also narrow and shallow low-pressure belts.

The metamorphic basement of the Pannonian Basin consists chiefly of Hercynian (Variscan) metamorphites in greenschist facies mainly in the Barrovian and Abukuma (i.e. medium- and low-) pressure series indicating a shallow crystallization of no more than 10 to 18 km depth. It is presumably formed by regional burial metamorphism in a thick sedimentary sequence of a long-living inland sea.

The premetamorphic sedimentary rocks of the Hercynian metamorphic belts in the East Mediterranean were mainly pelites and psammites, in the Eastern Alps also carbonate rocks and in the Drocea also psephites. The scarcity of metabasalts indicates that the Hercynian belts mostly do not represent a spreading-type (semi)-oceanic crust. Metabasalts are frequent only in the polymetamorphic (Baikalian+Hercynian+partly Alpine) area of Southern Hungary and Drocea Mts, indicating here a presumably Early Baikalian oceanic crust and subduction.

In the Alps the Hercynian metamorphites appear mainly in the amphibolite facies corresponding to a depth of crystallization of about 18—25 km,

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too. According to ESKOLA, the high-metamorphic-grade core of the mantled gneiss dome consists of recycled gneiss and granite and the lower-metamorphic-grade mantle of monocyclic supracrystal rocks. Thermal domes form linear or grid-patterns centred on the axial zone of the orogenic belts. Their index minerals suggest an abnormal steep local thermal gradient of 200 to 300 °C/km (Fig. 3, SCHUILING, 1960; MIYASHIRO, 1961; DEN TEX, 1975).

According to the computation of DEN TEX neither the normal conductive heat transfer from the general radioactive source nor that from additive magmatic sources are alone sufficient to explain these extreme gradients around the thermal domes. Convection as described first by TALBOT 1971, is also necessary for their realization. Viscosity in the order of 10^{17} stokes for felsic gneisses at about half of their melting temperature and a thickness at least of 10 km of source layer, i.e. the crustal thickening by nappe accumulation are prerequisites for these convections (DEN TEX, 1975).

and indicating a Hercynian subduction before the beginning of the great Alpine drifts, which preformed the Alpine plate-tectonics of the area. In regions where this metamorphic basement is covered by younger sedimentary rocks, the subsidence was succeeded by an emergence and erosion of the same extent. Thus, Variscan greenschist and amphibolite facies basement covered by Carboniferous and Permian sediments (e.g. Alps, Igal-Bükk eugeosyncline) indicate Prealpine orogenic event with subduction but now in the ancient original position of these microplates some hundred km towards South and East.

Between the Moesian platform and the Balkanides a *Caledonian metamorphic belt* of greenschist facies appears. Mostly the diabase-phyllite formation representing presumably an accreting ("spreading type") oceanic series forms the premetamorphic rocks of this belt. It indicates that in the Early Paleozoic the Moesian platform was separated from the Ancient Rhodope Massif. The Rhodope is represented mostly by Baikalian metamorphites of amphibolite facies. It is noteworthy that along the Caledonian belt the surface of the metamorphic basement reaches the greatest depth (12 km) known until now in SE Europe representing a subsided area. It may be interpreted as a protected belt between the Moesian plate and the Rhodope Massif drifted and rotated together. It was not considerably stressed also during the following Hercynian metamorphism.

The conservation of this Caledonian metamorphite belt is probably due to the fact that the subsequent Hercynian metamorphism reaches only the lotrite prehnite grade in this area: In the more loaded area north of the Moesian platform the Hercynian metamorphism appears also in greenschist-grade, thus Caledonian metamorphism there — if it occured altogether — may be overprinted by the Hercynian one.

In the Eastern Alps relatively few examples of Caledonian metamorphism even in amphibolite facies are known similar to the Hercynian one.

Late Baikalian metamorphism occurs mainly as the crystalline basement of the Carpathians and their continuations parallel to the Dinarides, as well as the Apuseni bordering the Early Baikalian Ghilau Mts (East Apuseni). The metamorphites of the Late Baikalian appear mostly in greenschist and epidote-amphibolite facies. They represent mainly metapelites and metapelitic metapsammites, rarely also metacarbonatites, in the East Carpathians partly metavolcanites, in the South Carpathians (where they appear as high mountains) also metagranites, even granites, and along the Dinarides also metabasalts.

The *Early Baikalian* occurs in the West and South Carpathians, subordinately in the East Carpathians and in the Apuseni, as well as in the crystalline basement of the Transdanubian Basin. Early Baikalian metamorphites are frequent in the Outer Dinarides as well as in the Rhodope, too. All these metamorphites appear chiefly in amphibolite facies. The basement

of South Hungary is represented by polymetamorphic amphibolite and epidote-amphibolite facies metamorphized during the Early Baikalian and Hercynian and even in the Alpine phase. They form mostly metapelites and metapsammites in the West Carpathian, and in the Ghilau Massif. In South Hungary, beside metapelites and metapsammites, (meta)granites are frequent too, even with metaarcoses and metabasites. In the East Carpathians the migmatites and in the South Carpathians metabasites (including metaophiolites) are frequent, as well. Metapelites-metapsammites and metabasalts are the main rocks of the Early Baikalian in the Outer Dinarides. Especially different is the Early Baikalian of the Rhodope, where metamigmatites, metaarcose, metapsammite prevail.

In this way in Eastern Europe the events may be followed up to 1200 m.y. Around the Prebaikalian of the Ukrainian shield and of the Moesian platform forming its southern thorn in the area of the Black Sea a sedimentary and igneous border of at least 200 km width was formed. These border rocks were mostly metamorphized during the Baikalian phase, in the south by subduction, in the north by simple geosyncline basement crystallization. In this way, the increase of the old Ukrainian platform begins. The rest of the Baikalian metamorphites became elevated in the front region of the much later succeeding Alpine drift accompanied by the Moesian rotation (: West and South Carpathians, Rhodope, partly Outer Dinarides). In the corner of the rotation, however, the Baikalian is laterally outpressed (basement of the Transylvanian Basin.) In the external western part (i.e. in the South of the Pannonian Basin) the Baikalian subsides and becomes polymetamorphic once more recrystallized, partly during the Hercynian and then covered by younger sediments.

During the Early and Late Paleozoic the increase of the old platform is continued by a new external belt of sedimentation metamorphized in the simple geosynclinal basement type.

In this area the Alpine phase is presumbaly controlled mainly by the clockwise rotation of the Moesian plate.

4. The Moesian Rotation Model of the East Mediterranean Plate-Tectonics

In contrast to the abundance of papers dealing with the West Mediterranean plate-tectonics (see also this volume), the East Mediterranean is less investigated. A summing of opinions on East Mediterranean plate-tectonics in relation to ore formation is given by the Belgrade UNESCO Symposium 1976 (JANKOVIČ et al., 1977). The plate-tectonic models of the East Mediterranean roughly agree in the main respect. Now the further elaboration of these models in the light of new paleomagnetic and other data is needed.

The model of the present author is based on the concept of the Moesian rotation, i.e. on the supposition that the Moesian plate separated now from the Ukrainian shield but agreeing with it in age and main petrographic particularities represented before the Triassic the south continuation of the Ukrainian shield placed in the West of the Black Sea. The mainly clockwise rotation and west- and northward drift of Moesia created the bend of the Carpathians with all its consequences described below.

In this sense products of the fracturation related to the separation of the Moesian and Ukrainian plates are, e.g. the Permian granites of the Dobrudja wedge intruded in more steps in the enlarging fracture. A next product of fracturation are, e.g. the non-differentiated tholeiitic Triassic ophiolites of the Persányi Mts and then the very coarse Cretaceous conglomerates of Bucegi indicating the pushing out of a section of the Baikalian metamorphic belt around the Ukrainian and Moesian Prebaikalian mass (as shown by the Map of Metamorphites of the Carpatho—Balkan—Dinaride region).

This section of the Baikalian belt forms now the crystalline basement of the Transylvanian Basin. By its NNW drift room was made for the bending of the Carpathians. The Apuseni and the North Transylvanian Baikalian metamorphites were also outpushed together with the basement of the Transylvanian Basin. Their strange SW—NE strikes fit well to the original position between the East and South Carpathians previously forming of straight belt.

The NNW drift is related to the *pressure minimum belt along the (North)-East Carpathians* characterized by 1. the gap of the Baikalian crystalline belt in the Vrancea region (the seismically most active area of the Mediterranean), 2. the Barót—Barault depression and on the other end 3. by the Nyírség depression. The pressure minimum belt represents a relaxation zone, with the Transylvanian Basin in its centre. The half dish-like crystalline basement of the Transylvanian Basin reflects the pincers motion of the Moesian plate relative to the Ukrainian shield by the inner horse-shoe-like pattern of its metamorphic and Permo-Triassic crests along overthrust lines. Perpendicular to the NW drift the Transylvanian Cretaceous-Paleogene is three times underthrusted to the metamorphic sequence (Meszes Mts (Fig. 5), Balázsfalva-Parajd crest, the two latter being covered by Neogene sediments.

The East-Inner Carpathian relaxation belt terminates wedge-like in the less disturbed Zemplenides (Szepesházy). Zemplenides are bordered by the Pienninian Klippen-belt to the East, and by the Bükk—Tokaj continuation of the Balaton line to the West.

Parallel to the East-Inner Carpathian relaxation belt the *maximum-pressure zone* along the minimum distance between the (most rotated) western corner of the Moesian Plate and the Bohemian Massif appears (Szádeczky-Kardoss, 1971). This maximum-pressure zone is characterized by WSW—ENE striking overthrusts: in the South Carpathians, in the Apuseni, in the Cordillera

of the Szolnok—Debrecen—Máramaros flysch belt and in the Slovakian nappes.

Due to the NNW drift of the crystalline basement of the Transylvanian Basin some NNE striking side fractures and bends of fracture system appear on the border of the maximum-pressure belt in regular distances of 60—70 km from one another: 1. the NNE continuation of the Cretaceous Maros-Mures

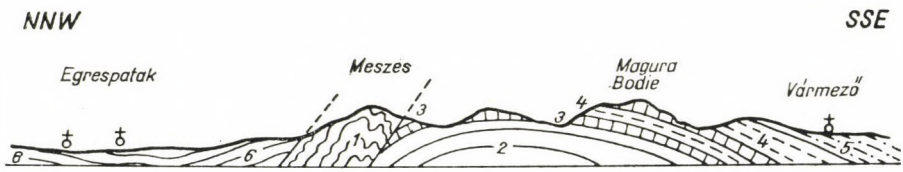

Fig. 5. The overthrust of the crystalline basement on the Eocene in the Meszes Mts, Transylvania, (according to E. SZÁDECZKY-KARDOSS, 1930)

flysch belt; 2. the Uppermost Cretaceous fracture system of the Vlădease magmatism; 3. the Senonian flysch belt of Békés (SZEPESHÁZY 1977); 4. the Upper Cretaceous Paleogene Szolnok—Máramaros flysch belt.

Due to the Moesian rotation, the tectonic units of the Flysch Carpathians are thrusted successively from the interior (i.e. from the Basin System side) toward the exterior. Hence the Carpathian tectonic units appear as relatively underthrusted by the Carpathian Foreland.

The width of the overthrusting in the Northeast and East Carpathians decreases from north toward the south according to the decreasing rotation distance toward the rotation centre.

The compression by the Moesian rotation north of it elucidate why a continuation of the Carpathian Flysch belt along the South Carpathians does not exist.

The subducted quasi-oceanic plates in the Pannonian Transylvanian Basin system are relatively thin (— roughly about 50 km —) and flexible due to the high temperature during the subduction of these small quasi-oceanic plates (see Chapter 1). Thus, they dip partly very steeply especially at the front of the collision of Europe and Africa in the centre of the basin system (e.g. the Oligocene in the bore hole of Buzsák).

The Balaton—Darnó subductions dip already opposite to the Pienninian Klippen-belt subduction.[11]

The Tertiary volcanism related to the subduction in the Carpathian Basin system is shifted roughly clockwise according to the Moesian rotation.

[11] Similarly to the "more subduction" model within the Carpathian Basin system (SZÁDECZKY-KARDOSS, 1971) substituting the one subduction model (DEWEY et al., 1970), also in the Alps now more subductions are supposed (FRISCH, 1976 and in this volume; BOGEL—SCHMIDT, 1976) instead of the older subduction models (ERNST, 1973; DIETRICH 1976).

The rotation and the subsiding of the Neogene basins result together in a screw-like movement at the border of the basin, transporting the rocks of the Bükk–Igal (Balaton—Darnó line)) and the Middle Great Plain eugeosyncline (longitudinal) obliquely to the depth toward northeast. Due to the high heat flow in the Pannonian Basin these rocks become soon partially melted, resulting in two central accumulations of volcanism in the area of the Middle Tisza Region south of Tokaj and in the Gutin Mts.

Moesia rotation explains the whole Alpine-age magmatism of the area. *a)* The Triassic, Jurassic and Lower Cretaceous ophiolites are products of the quasi-oceanic spreading (partial melts of the mantle in the 4 subduction belts Carpathian, Vardar—Maros, Middle Great Plain and Bükk—Igal belts); *b)* the Senonian andesitoides and the Laramian granodiorites ("banatites") at the most loaded edges of the rotated microcontinents due to the additive heat of rock friction by the rotation represent the partial melt products of the lithosphere (crust upper mantle); *c)* the Tertiary andesitoides along the subducted slabs, deriving mainly from subducted oceanic sediments, the volatile content of which decreases the melt temperature. The heat effect by the ascension of the melt mobilizes the most mobile alkali and silica complexes of the lithosphere along the volcanic vent, thus the amount of these components gradually increases in the magma, forming finally even alkali-trachytoids (Szádeczky-Kardoss, 1968); *d)* the Neogene ignimbrites mostly in the central part of the Great Plain deriving from secondary magma chambers are formed by the high heat flow and hydrothermal ascension in the centre of the mantle dome; *e)* the Late Neogene-Quaternary "final" basalts appear mainly along the borders of the thin crust of the basin area where the ramified "hydrothermal" flow system along the Moho begins to turn downwards and is heated partially melting the mantle.

The counter-clockwise rotation prevailing in the West Mediterranean opposite to the clockwise rotation prevailing in the East Mediterranean may depend partly on the original position of the microplates by the N—S collision of Africa and Europe. Microplates elongated in about SW—NE direction are rotated counter-clockwise, whereas microplates elongated SE—NW become rotated clockwise. On the other hand, transversal drift of the great continents rotates the microplates between them in the direction of their own relative motion. Hence the east drift of Africa relative to Europe promotes counter-clockwise rotation. The Turkey plate standing in the way of the eastward drift of Africa is northward drifted, hence promotes the clockwise rotation of the Moesian plate.

Paleomagnetic measurements synthetized by Krs and Roth in this volume as well as measurements by Petrescu (W-Apuseni), Nojarov (Sredna Gora and Timok), Pucher et al. 1974 (Argolis relative to the Pindos Mts) seem to corroborate the mostly clockwise rotation of the Moesian plate. The rotation

may be influenced by different secondary factors, too, producing deviation from the main trends described.

Consequences of the Moesian rotation control the Tertiary-Quaternary and present dynamism of the Carpathian Basin system mostly by the formation of the mantle dome[12] structure, The Pannonian mantle dome is younger than the subductions of the area and not older than Miocene. The characteristic deep flow system of the mantle dome consists of more branches. The descending and cooling wing along the Carpathians and the Dinarides consists of the volatiles of the subducted sediments. The high-temperature central branch consists of alkali-rich "hydrothermal" solution, solving and transporting the bottom part of the crust to the peripheries and producing the subsidence of the basin, its high heat flow and the intense Tertiary volcanism. Thus, it is thinning out the crust in the central part and thicking it at the peripheries (see also the paper by GUTDEUTSCH, STEGENA et al. and BODRI—BODRI in this volume).

This form of the Moesian rotation is obviously still a rough approximation, but suitable for influencing further investigations.

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[12] Mantle dome is named also "mantle diapir". Disregarded of the priority of the name "mantle dome" elaborated by the author (1966), the name of "mantle diapir" is erraneous. Diapir means piercing fold, i.e. anticline in which a mobile core such as salt has been broken through the more brittle overlying rock. Such broken through structure does not appear in the mantle dome, hence the name mantle diapir is misleading.

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ГЕОТЕРМИКА И МЕТАМОРФИЗМ В СРЕДИЗЕМНОМОРСКОМ ТИПЕ ТЕКТОНИКИ ПЛИТ

Е. САДЕЦКИ-КАРДОШ, Будапешт

Р е з ю м е

Температура океанских плит является более высокой континентальных плит даже 10 миллионов лет или больше после их образования путем акреции. «Мешок теплоты» вокруг субдукции возникает только на крупных старих океанских плитах («Циркумпацифический тип»). Однако, окрестность тонкой океанской плиты подогревается с начала субдукции («средиземноморский тип»). Ведь температура литосферы средиземно-морской третидной микроплитовой области нарашивается многочисленными квази-океанскими теплыми субдуционными плитами, данные глубины очагов вычисленные по петрохимическому методу отношений K/Sl систематически высше ожиданных. Правильные данные соответствуют только лишь половине данных вычисленных ранее. Таким образом можно составлять более реалистические разрезы микроплитовых областей.

Для сухих реакций без летущего давления попажабится сверх корное нагрузочное давление (> 10 кбар, напр. базалт — эклогит). Внутрикорные реакции обыкновенного метаморфизма осуществляются в большинстве случаев при помощи «эффактивного летущего» давления. Различаются три варианта летущего давления зависимого от нагрузочного давления: 1. Виртуальное или затаённое летущее давление под высоким нагрузочным давлением в закрытой системе безвнутрикорного метаморфизма в соответствии со законом действия масс. 2. Эффективное летущее давление под средним нагрузочным двалением в полузакрытой системе разрешающим метаморфизм. 3. Субнормальное летущее давление под низким нагрузочным давлением в открытой сисстеме сопровождающим беспрерывное освобождение летучих, без метаморфизма.

Таким образом реакции метаморфизма требовающие несколько дней или неделей экспериментальных бомбах задерживаются много-миллионными летами в закрытых осадочных системах скоро и глубоко субдуцированных плит. Из-за этого «парадокса давления» результаты экспериментальной петрологии запускающей пористость не отражают корного метаморфизма.

Обыкновенные три варианта метаморфизма и их метаморфические фации дают геологическую информацию о скорости снижения, о глубине (т. е. нагрузочном давлении), о температуре рекристаллизации и о структурных отношениях области. 1. В метаморфизме погружением скорость снижения 0,001—0,1 мм в год и указывать низкую температуру (< 500 C°) и низкое нагрузочное давление (т. е. глубины менше чем 18 км). 2. Динамометаморфизму соответствует скорость снижения 10—100 мм в год; рекристаллизация происходит в зонах субдукции в условиях средних до высоких давлений и средних температур. 3. Контактный метаморфизм демонстрирует скорый подьём магмы в порядке километра в год, главным образом над субдукцией и также рекристаллизацию в небольшой глубине (< 10 км), у температуры 300—650 C°.

После метаморфизма субдукцированных плит порода обычно попадает в области виртуального давления, поэтому динамометаморфизм прекращается. Низко-давленная часть парных метаморфических зон вдоль субдукции возникает путём обдукция или специальным вариантом структурной реляксация (написанной в английском тексте) которые вохвращают плиту в район эффективного летцщего давления.

Даются примеры оценки альпийского, герценского, каледонского, ранне- и позднебайкалского метаморфизма Альп, Карпат, Балкана и Динарид.

Тектоническую структуру, тектоника плит, магматизм, метаморфизм и термическое течение в восточной средиземноморской области можно истолковать ротацией по часовой стрелке Мезийской плиты. Метаморфический фундамент Трансильванского бассейна наверно представляет собой горизонтально выжатую и перемещанную к ССЗ часть, между Восточными и Южными Карпатами прежней постоянной и более прямой байкалского метаморфического пояса вокруг Украинского шита и Мезийской плиты.

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RECONNAISSANCE GEOPHYSICAL AND GEOCHEMICAL INVESTIGATIONS OF THE OPHIOLITES OF THE DARNÓ TECTONIC-LINE (NE HUNGARY)

By

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The results obtained from a series of petrophysical, geochemical and field magnetic investigations of the mafic and ultramafic complexes along the Darnó-line suggest a strong genetic affiliation to similar occurrences elsewhere in the Alpine—Mediterranean region. Geochemical evidence suggests that the original basaltic magma was tholeitic and the differentiation and fractionation trends agree well with those for abyssal tholeiites from the ocean floor. Y/Nb ratios, Ti—Cr relations and the major oxide chemistry all support this view and rule out the possibility of the igneous series being intra-plate occurrences created *in situ* by magmatic activity. Negative magnetic anomalies (ΔZ, ΔT) were measured over the complex and the interpretative models seem to bear out the allochthonous position of these rock suites. The magnetic properties of two rock units of the complex (diabases and gabbros) are compared with those from similar units of other ophiolites, but the preliminary results cast doubts on the suitability of these rocks for detailed paleomagnetic studies because of the effects of tectonic overprinting.

Introduction

A lot of attention is being paid at present to the study of ophiolite complexes all over the world. This has come about mainly with the identification of ophiolites as slices of the suboceanic lithosphere tectonically emplaced on the continental margin during convergence of two continental blocks. These rock suites mark the suture zones where oceanic lithosphere was consumed by subduction. A detailed study of ophiolite complexes contributes a lot to the paleographical and paleotectonic reconstruction of the area where they occur.

The Darnó-line is perhaps the most important tectonic feature in the northern half of Hungary. Trending NNE—SSW, it divides the formations of the Bükk Mts from those of the Central Hungarian Mts (Fig. 1). The line towards the SE joins the Balaton-line which is supposed (WEIN, 1969; DANK and BODZAY, 1971) to be the continuation in the Pannonian region of the Periadriatic Lineament, the major dividing feature between the Southern and Northern Alps. The Darnó-line appears distinctly on Bouguer anomaly, magnetic and aerial photo-geological maps. Mafic and ultramafic rocks occur in several places along the Balaton—Darnó-line, but the complexes investigated in detail in this study are those around Mt. Darnó (between Sirok and Recsk), Bátor, Szarvaskő and Bélapátfalva (Fig. 2).

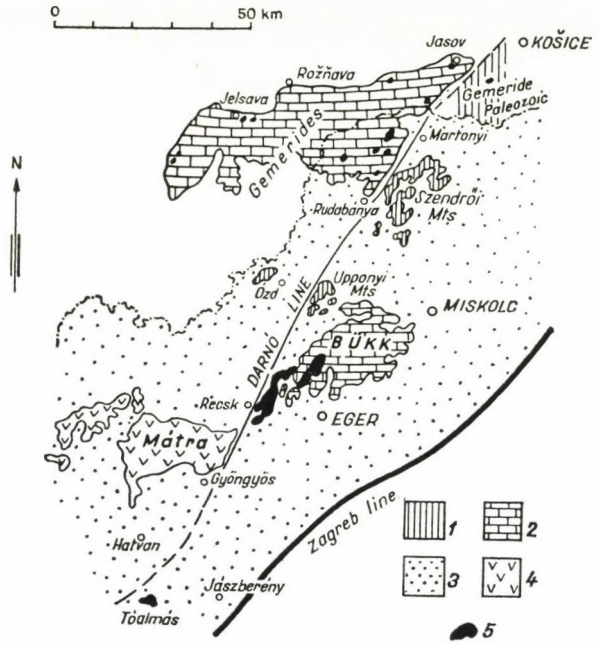

Fig. 1. Schematic geological map of NE Hungary. 1. Paleozoic; 2. Mesozoic; 3. Neogene; 4. Neogene volcanites; 5. Ophiolites

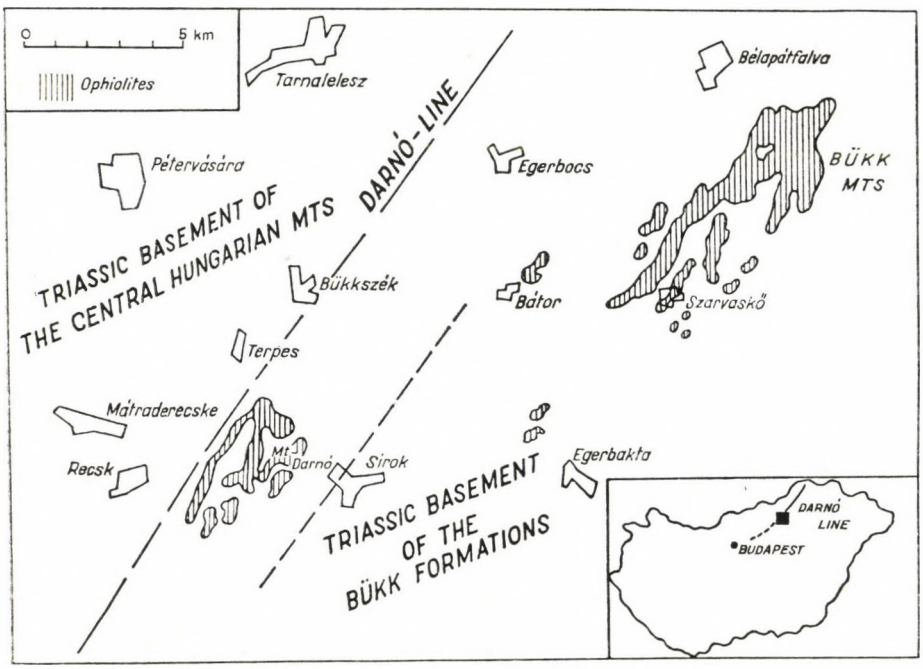

Fig. 2. Position of the investigated ophiolites of the Darnó tectonic zone

Structural and Tectonic Background

Near Szarvaskő the ophiolite sequence is complete (radiolarites and pelagic sediments — pillow lavas and diabases — gabbros — quartz diorites — peridotites, in a more or less pseudo-stratigraphical order, from the top downwards). Depending on the pyroxene contents, the peridotites which occur at the deeper levels grade into pyroxenite. Hornblendite and wehrlite occur at the core of the ultramafic complex. Complete petrographic and stratigraphic descriptions of the complexes have been given by: SZENTPÉTERY (1953) and LENGYEL (1957) for the Szarvaskő area, KISS (1958) and FÖLDESSY (1975) for the Mt. Darnó—Sirok area, and by PANTÓ and FÖLDVÁRI (1950) for the occurrences along the Bódva valley in the northeastern-most corner close to the Czechoslovakian border. SZEPESHÁZY (1977) has also discussed the structural positions of these rock complexes.

Early work done on these mafic and ultramafic rocks (like those quoted above) were mainly in connection with mineral prospecting. The first attempt to incorporate them into the plate tectonic hypothesis was by SZÁDECZKY-KARDOSS (1971) when he identified the Balaton—Darnó-line as a U. Cretaceous—Paleogene subduction zone. In subsequent works (SZÁDECZKY-KARDOSS, 1975, 1976) he pointed out the possible connection between the ophiolite occurrences and the Eocene andesite volcanism west of the Darnó-line, and gave a model for the plate tectonic evolution of the Pannonian region. The paleogeographical reconstructions of CHANNELL and HORVÁTH (1976) and HORVÁTH and CHANNELL (1977) also took into account the ophiolite occurrences of the Balaton—Darnó-line.

In this paper results of a series of geochemical, field magnetic and petrophysical investigations are discussed. The geochemical studies have helped to clarify the original tectonic setting of the igneous rock suites, while field studies support the view that the rocks are in allochthonous positions.

While most of the problems pertaining to these rocks are gradually being solved, one particular problem, however, still remains an engima. This is the problem of ascertaining the right ages of these rocks. Ages suggested for the ophiolites include Upper Paleozoic, Triassic (BALOGH, 1964; ZELENKA, 1975; FÖLDESSY, 1975), Cretaceous (PANTÓ—FÖLDVÁRI, 1950). Such diversities in opinion are caused by the fact that no diagnostic fossils have been found in the sedimentary volcanic assemblage itself. On the basis of the ages of similar occurences in the Alpine—Mediterranean area an Upper Jurassic—Lower Cretaceous age has also been suggested (SZEPESHÁZY, 1977; ONUOHA, 1977a). Radiometric age determination of the rocks as well as geochronological work based on the available radiolarias should be an urgent matter for the immediate future. The work of KOZUR and MOCK (1977) on the Meliata series of the SW

Bükk Mts. has helped to solve part of the problems mentioned earlier. On the basis of conodonts they have ascertained a Middle Triassic — Upper Carnian age for the sedimentary series associated with the ophiolites.

Geochemical Investigations

The major oxide contents of the rock units of the ophiolites formed the basis of the geochemical study. Ten samples were also analysed for their Ti, Sr, Y, Nb and Cr contents. A separate and detailed account of the geochemical investigations has been published by the author (ONUOHA, 1977b). Making use of several variational parameters ($Na_2O + K_2O$ versus SiO_2; SiO_2 versus FeO^t/MgO; FeO^t versus FeO^t/MgO; TiO_2 versus FeO^t/MgO; Y/Nb ratio; Ti—Cr relation of PEARCE (1975); $FeO^t = Fe_2O_3 + FeO$), it was found that the differentiation and fractionation trends for the complex agree very well with those of other ophiolite complexes, and with that of abyssal tholeiites from the ocean floor. Y/Nb ratios for the complex are high (2.75—8.75), thus supporting the tholeitic nature of the rocks. The TiO_2—FeO^t/MgO and the Ti—Cr relations not only confirm their similarity to ocean floor basalts, but also bear out the fact that their original tectonic settings had nothing to do with island arc magmatism.

Magnetic Anomalies and Rock Magnetic Studies

Detailed field magnetic work with a proton precession magnetometer was carried out along several profiles that cut across the Darnó-line. As reported earlier by the author (ONUOHA, 1977), the ophiolite outcrops are associated with negative ΔT anomalies. The most intensive minimum values (—400 gammas) were recorded over the largest, thickest outcrops. The magnetic profiles celarly reflected the sub-surface distribution of the mafic rocks (Fig. 3). An interpretative model using a series of steeply inclined dykes of finite lengths appears to explain the observed anomalies over Mt. Darnó. On the basis of a regional magnetic survey carried out earlier in the area a 15 km long, 1—2 km wide anomaly showed up between Sirok and Bátor. This anomaly goes right up to the western margin of Mt. Darnó in a branch that is about 500 m wide and 3 km long. According to the estimates of the specialists at the State Geophysical Institute (Annual Report, 1973), the interpretation gave a body 70 m deep, 150 m wide with a NE—SW strike, and dipping deeply under Mt. Darnó. The susceptibility contrast between the volcanic rocks and the country sedimentary rocks is appreciable enough to give the type of correlation that is reflected in Fig. 3.

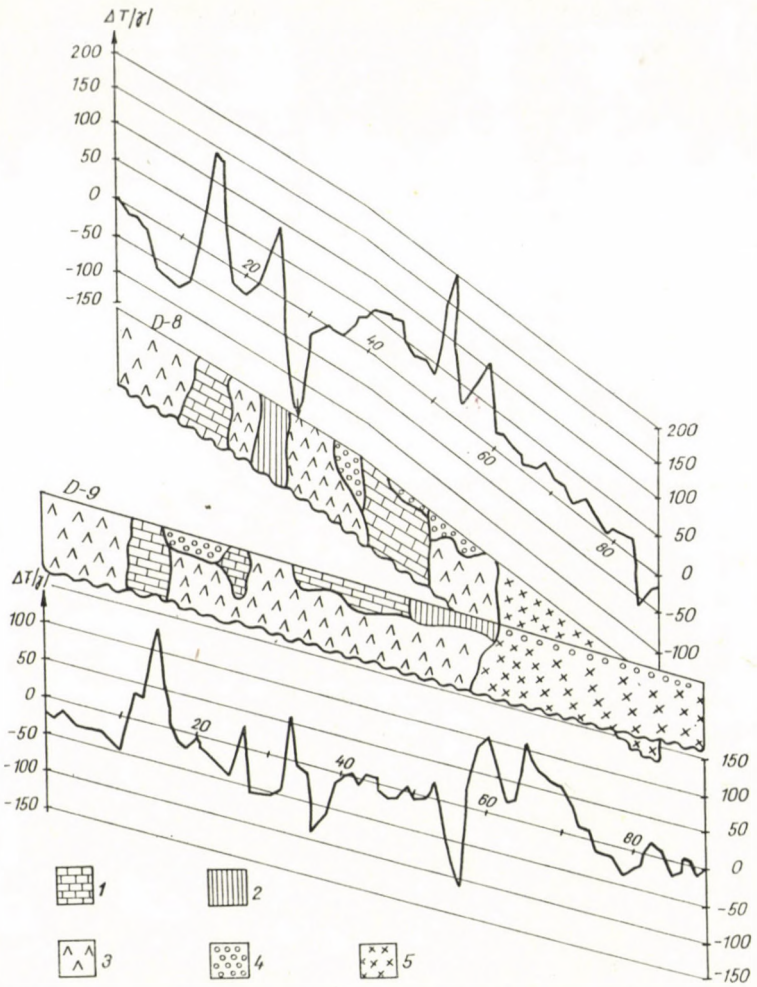

Fig. 3. Magnetic profiles over Mt Darnó and their correlation with sub-surface features: 1. Upper Paleozoic (Permian) limestones; 2. (Triassic?) marls and shales; 3. (Jurassic-Cretaceous?) spilite diabase-pillow lava formations; 4. Terrigenous and 5. Marine clastic sediments (Neogene)

It was thought necessary to carry out preliminary rock magnetic investigations of certain units of the complex with a view of determining the suitability of those units for detailed and reliable paleomagnetic work. The results to the magnetic susceptibility K, initial N.R.M. intensity J_n and KOENIGSBERGER ratio Q of two of the rock units (diabases and gabbros) are shown in Table 1 and in Figs 4 and 5. The gabbros and diabases have susceptibilities of the same order of magnitude (10^{-3} G/Oe), and cannot be separated from each other on the basis of susceptibility alone. The initial NRM

intensities are significantly different, being nearly three times greater for the diabases. The KOENIGSBERGER ratio is almost three times greater for the diabases, than for the gabbros, but both average values are of the order of 10^{-3}, values not characteristic of basic rocks. A comparison of the rock magnetic

Table 1

Magnetic properties of two rock units from the Darnó ophiolite complex

Rock type	Susceptibility ($K \cdot 10^{-6}$ G/Oe)		NMR intensity ($J_n \cdot 10^{-6}$ G)		Koenigsberger ratio $Q \cdot 10^{-3}$ ($F = 0.45$ Oe)	
	range	mean	range	mean	range	mean
Gabbro	(13)* 74—151	112	(29) 0.36—18.62	4.30	(17) 1.18—35.09	6.64
Diabase	(39) 35—471	116	(21) 1.22—94.37	11.39	(11) 6.84—44.54	18.81

* Numbers in parentheses is the number of samples considered for the investigation. All mean values were computed from the log–normal distribution

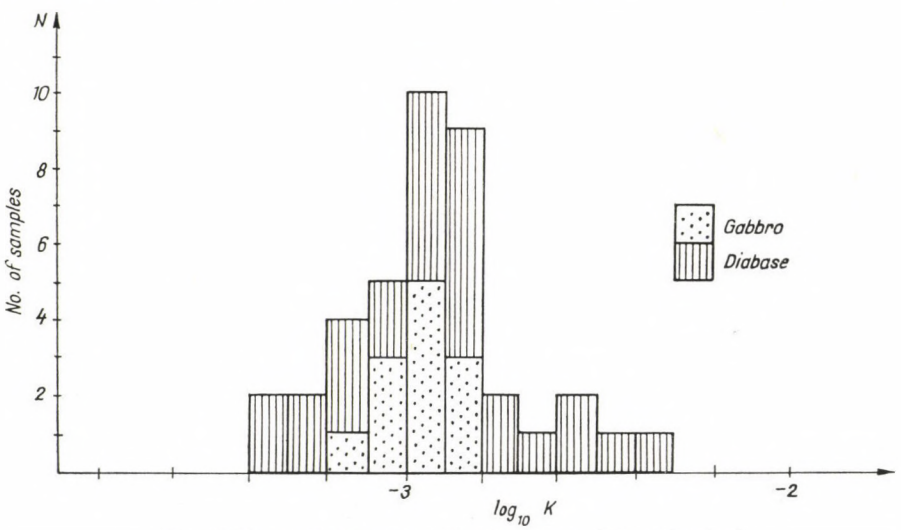

Fig. 4. Magnetic susceptibility (log–normal distribution)

properties of theses rock units with those for similar units of other ophiolite complexes from other parts of the world is made in Fig. 6.

The low J_n values and the extraordinarily low Q values deserve some few comments. Since there is no doubt about the basicity of the rocks, and since their ocean floor affinities have been ascertained geochemically, it appears that the low J_n and Q values are due to some or all of the following factors:

a) the effect of weathering (all the samples considered for the rock magnetic studies were from surface outcrops);

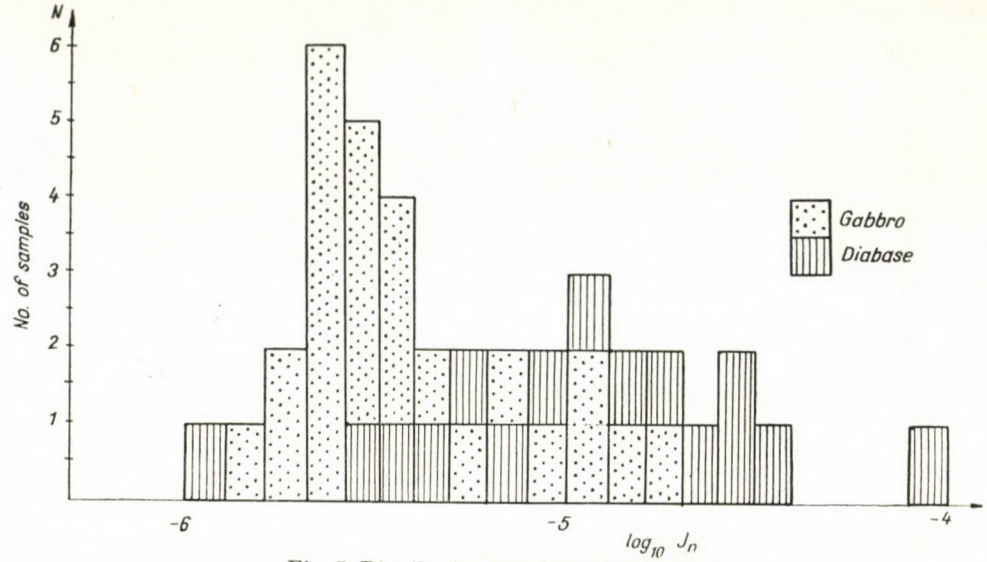

Fig. 5. Distribution of initial NRM values

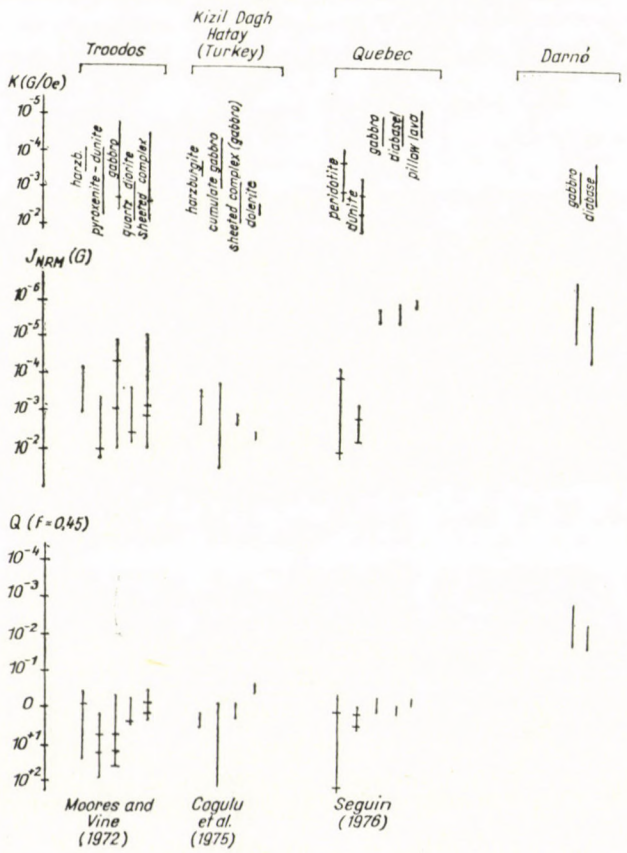

Fig. 6. Comparison of rock magnetic properties

b) low-temperature oxidation of the remanence carrying titanomagnetic minerals;

c) tectonic overprinting. At present we do not yet know for sure what possible changes in magnetic properties occur during the obduction of ophiolites. Large-scale tectonic interactions and whole-sale deformation have occurred all along the Darnó-line. The effects of these tectonic processes on the original magnetic properties are difficult to ascertain, but presumably they must have been destructive.

Discussion of the Results

The stratigraphy and geochemistry of the Darnó mafic and ultramafic rocks favour an ocean floor origin for them. The fractionation and differentiation trends resemble those of abyssal tholeiites from the ocean floors. The detected magnetic anomalies seem to suggest that the rocks are in allochthonous positions and are not deeply rooted. Induced polarization investigations carried out in the area confirmed that the whole morphological area of Mt. Darnó is a foreign body in the Mátra surroundings (Annual Report, 1973). Most probably, the Darnó zone marks an oceanic suture; thus the usual interpretation given to the complexes as *in situ* volcanic occurrences has to be reviewed.

The chemical characteristics of the Darnó ophiolites are very similar to those of the Upper Schieferhülle of the Penninic windows of the Tauern and Strobl as described by BICKLE and PEARCE (1975), OBERHÄNSLI and DIETRICH (1975), and RAITH et al. (1977). The reconnaissance rock magnetic studies reveal that paleomagnetic work on the ophiolites may not be reliable unless the samples were subjected to the most complicated and sophisticated processes of magnetic cleaning to remove the imprints of past tectonic processes. There are also other views (e.g. P. MÁRTON; personal comm.) that because of the paucity of data the present rock magnetic results may not be diagnostic of the whole mafic-ultramafic rocks of the Darnó zone. However, at present it appears that if paleotectonic reconstructions for the area have to make use of paleomagnetic data, then such data would have to be for rocks from the less disturbed (autochtonous?) sedimentary units, and not for the ophiolites.

The Darnó ophiolites are genetically related to similar occurrences found in the Mesozoic of the Gemerides. Any explanations for the origin and tectonic positions of the Darnó complex would thus have to be valid for the areally dispersed complexes of the Southern Gemerides. The Darnó-line has been interpreted as a subduction zone (SZÁDECZKY-KARDOSS, 1971, 1975). Stratigraphic considerations and other tectonic features of the area (including the

area of the S. Gemerides) allow for the alternative interpretation that the ophiolites are located on (Tauern-type?) tectonic windows. This interpretation, like the one of SZÁDECZKY-KARDOSS (1971, 1975, 1976) naturally assumes that the complexes were created by plate accretion at a mid-ocean ridge, and later transported to a subduction zone, but does not necessarily identify the present positions of the ophiolites along the Darnó megatectonic zone as marking the site of the subduction. The plate-tectonic significance of the Darnó ophiolites and the associated Meliata series in the reconstruction of the Mesozoic — Cainozoic evolution of the Western Carpathian — Pannonian region is discussed in details somewhere else (see: HORVÁTH, VÖRÖS, ONUOHA, this volume).

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ГЕОФИЗИЧЕСКИЕ И ГЕОХИМИЧЕСКИЕ ИССЛЕДОВАНИЯ ОФИОЛИТОВ ДЕРНОВСКОЙ ГЕОТЕКТОНИЧЕСКОЙ ЛИНИИ

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Резюме

Результаты петрофизических, геохимических и полевомагнитных исследований проведённых на мафических и ультрамафических образованиях находящися вдоль Дарновского линеамента показывают на их тесное генетическое сходство с подобными образованиями, известными в других районах Альпийско-средниземноморской оюласти. По геохимическим данным, первоначальная базальтовая магма выла толеитого типа; её тренды дифференцирования, фракционирования хорошо состветствуют тем абиссальных толеитов океанского дна. Это мнение подтверждается также отношениями У/Nb, Ti—Cr и ющим химическим составом. Таким образом эти магматические породы не могут выть межплитовыми образованиями, возникшими на месте вследствии магматизма. Над этой толщи обнаружились отрицательные магнитные аномалии (ΔZ, ΔT); интерпретационные модели говорят за аллохтонные положение этих свит. Магнитные особенности двух единиц изучаемой толщи (диабазов и габбро) сравниваются с подобными единицами других офиолитов, но по предворительным данным эти породы едва ли подходят для подробных палеомагнитных исследований, из-за эффектов действия молодого тектонизма.

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PLATE-TECTONICS OF THE WESTERN CARPATHO-PANNONIAN REGION: A WORKING HYPOTHESIS

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The Alpine belt is made up of the southern and northern marginal complexes of the Mesozoic Tethys, which were juxtaposed during the Cretaceous-Tertiary as a consequence of the convergence of the African and European plates. Facies and faunal considerations as well as paleomagnetic data suggest that the Inner Western Carpathians and the northwestern part of the Pannonian Basin-were situated to the south of the main Tethys ocean, probably as a part of the Adriatic promontory. This region was further divided by the Meliata ocean into two parts with Austroalpine (Pieniny Klippen belt, Tatrides, Križna, Choč, Gemeride units) and South Alpine (Volovec, Bükk) affinities, respectively. The Veporic unit originally part of the European margin (and later separated from it by the Magura flysch trough) was overridden by the Austroalpine units during the Middle to Late Cretaceous. To the south of this zone of continent–continent collision the blueschist metamorphism of the Meliata series (ophiolites and deep-water sediments) and their emplacement to the Volovec block were brought about. The Tertiary saw the consumption of the Magura flysch trough, the formation of the Outer Carpathian nappes and their overthrusting unto the Molasse foredeep.

Introduction

The Alpine belt separates into the Western Carpathians and the Dinarides at the western margin of the Pannonian region. As the major part of the junction is covered by thick Neogene-Quaternary sediments, the tectonic reason of separation has been debated for a long time.

One group of authors accept the idea of KOBER (1921), who considered the Pannonian region as a median mass (Zwischengebirge) lying between the oppositely verging nappe-complexes of the Carpathians and Dinarides. An extreme version of this idea suggests that the area has been a rigid nucleus (Pannonian "massif") which was either not, or was just slightly affected by Alpine orogeny. On the other hand, UHLIG (1907) suggested that the Hungarian Central Range was an integral part of the Western Carpathians and formed one of their nappes. The boldest hypothesis was put forward by ARGAND (1922), who visualized the whole Inner Carpathian—Pannonian—Dinaric

area as a huge nappe of African origin which overthrust the European margin.

The majority of Hungarian geologists took a stand inbetween the two extreme alternatives as concerns the tectonic evolution of the Pannonian region. They recognized that the basement of the Pannonian Basin is made up of nearly parallel belts of Mesozoic and older troughs and ridges. It was supposed that the present zonal arrangement reflected the Mesozoic paleogeography, i.e. primarily vertical movements controlled the tectonic evolution of the region (VADÁSZ, 1961; WEIN, 1969).

The advent of plate-tectonics favoured the more mobilistic interpretations (DANK and BODZAY, 1971). First of all, SZÁDECZKY-KARDOSS (1971, 1973) argued that in the Carpatho-Balkan region the Pieniny, Klippen belt, the Vardar zone, the Maros zone and an Intrapannonian belt can be interpreted as sutures of Cretaceous-Tertiary subduction zones. Later on, other authors suggested regional or more local plate-tectonic models for the Carpatho-Pannonian region (STEGENA et al., 1975; NEY, 1975; CHANNELL and HORVÁTH, 1976; BIRKENMAJER, 1976; WEIN, 1978). Very recently the new results of Carpathian geology relevant to the plate-tectonics of the region has been reviewed by MAHEL (1977, 1978). Moreover, he suggested a plate-tectonic model for the Alpine evolution of the Western Carpathians. In this article, we put forward an alternative model for the Western Carpatho-Pannonian region. We use this name for the area made up of the Western Carpathians and that part of the Pannonian basin which lies to the NW of the Zagreb-Zemplén line. Our model attempts to place more emphasis on the Mesozoic paleogeography, the presence of ophiolites in the Gemeric-Darnó-line region, and on the Alpine analogies.

Basic Data for Paleogeographic Reconstruction of the Western Carpatho—Pannonian Region

One of the basic results of plate-tectonics is the recognition that mountain belts are made up of sedimentary and volcanic assemblages of Atlantic-type continental margins. These assemblages are transformed into orogenic belts in the Alpine—Himalayan region by continent-continent or continent-island arc (microcontinent) collisions (DEWEY and BIRD, 1970). It is widely accepted that in the Alps (s.s) the Helveticum, the Flysch belt, the northern and central Penninicum were associated with the northern (European) marginal complex of the oceanic Tethys. On the other hand, the Southern Alps—Dinarides and the Austroalpine nappes are considered to be derived from the southern (African) marginal complex (LAUBSCHER, 1971; ANGENHEISTER et al., 1972, 1975; BERNOULLI and JENKYNS, 1974; DIETRICH and FRANZ 1976).

The association of the southern marginal complex with the Periadriatic belt can be explained in terms of Adria as promontory of Africa (CHANNELL and HORVÁTH, 1976).

In the following we summarize arguments which suggest that the Inner Western Carpathians and the western Pannonian region constitute the eastward extension of the Austroalpine and partly the South Alpine realms, (Fig. 1), i.e. the region also belonged to the southern marginal complex of the oceanic Tethys.

A) *Structural and Facies Connections*

The main tectonic zones of the Western Carpathians and the Alps can be correlated on facies and structural ground. The Carpathian molasse zone continues uninterruptedly towards the Alps. The connection of the Western Carpathian and Alpine flysch belts cannot be seen on the surface but deep-drillings have demonstrated the connection of the two belts. The Pieniny Klippen-belt also appears to be continuous below the Vienna basin (ROTH, 1967). Its westernmost outcrop can be found near Vienna (PREY, 1974). More to the west the belt probably disappears below the Northern Calcareous Alps (BECK-MANNAGETTA, 1977). As the Pieniny Klippen-belt Mesozoic shows Austroalpine affinities (MIŠIK et al., 1977), it can be considered the most northernly transported parts of the southern marginal complex.

The Inner Western Carpathians are composed of the core mountains and extensive north-vergent nappes, which were formed during the mid-Cretaceous times. It is widely accepted that the pre-orogenic position of the main tectonic units is the following: Tatric, Fatric, Veporic, Hronic and Gemeric unit (BIELY and FUSAN, 1967; ANDRUSOV et al., 1973). Although there are differences in details of correlation, there is general agreement that the Inner Western Carpathians can be considered the eastern continuation of the Austroalpine realm (ANDRUSOV, 1968; TOLLMANN, 1969; MAHEL, 1974). Deep-drillings have demonstrated that the Tatric and Veporic units continue towards the Eastern Alps in the basement of the northwestern Pannonian Basin (FUSAN et al., 1971; WEIN, 1972). The correlation of the Upper Austroalpine nappes with the Hronic—Gemeric unit is particularly good as they exhibit nearly identical formations. The Hungarian Central Range is considered the connecting link between the Gemericum and Drauzug (VADÁSZ, 1954; TOLLMANN, 1969). The Mesozoic of the Southern Gemerides represents a transitional type between the Austroalpine type and the South Alpine-Dinaric type, which latter is well developed in the Bükk Mts (BALOGH, 1964; MAHEL, 1974). Below the Neogene sediments of the Pannonian Basin Bükk-type Permian and Triassic has been found by deep-drillings in the Igal-Bükk belt (WEIN, 1969). The southern border of the Igal—Bükk belt is a deep fault zone and can be considered the eastern limit of the Austroalpine-South Alpine realm (Fig. 1). We call it Zagreb-Zemplén line. The Intrapannonian mobile belt lies on

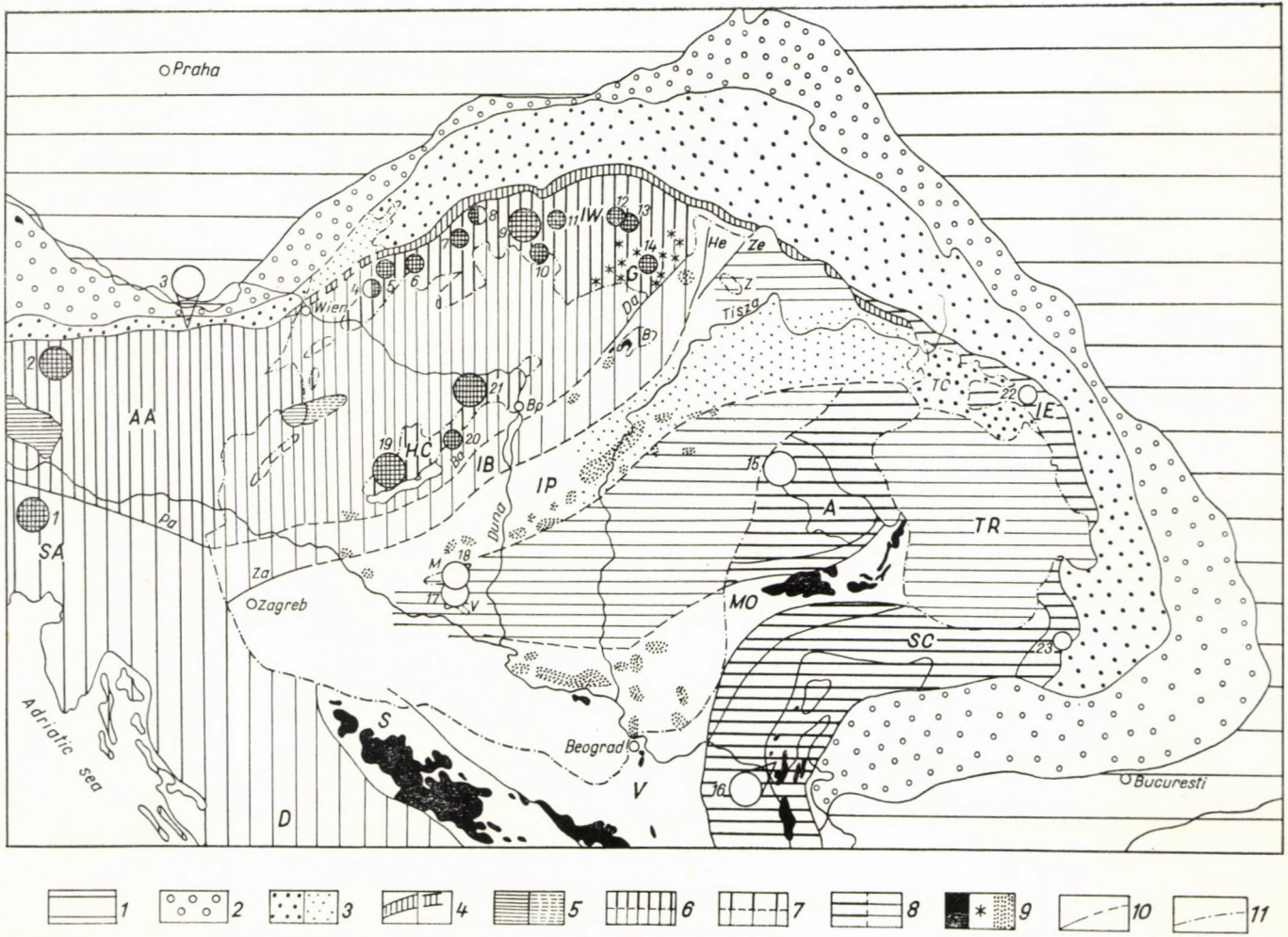
its other side and then the Mecsek—Villány basement of the Great Hungarian Plain—Apuseni Mts unit follows, which shows "European" affinity (SZEPESHÁZY, 1975, 1977).

B) *Faunal Affinities*

It has been known long ago that some of the Mesozoic marine fauna in the European—Mediterranean region exhibits characteristic associations, called provinces. Although there is no consensus in the cause of provinciality, it is widely accepted that Mesozoic paleogeography and other environmental factors were mainly responsible for the provinciality. This gives a good possibility to use faunal data for the delineation of Mesozoic paleogeography. On the basis of distribution of the "Mediterranean" and "NW European" faunal provinces of Lower Jurassic ammonites GÉCZY (1973) suggested that the Hungarian Central Range belonged to the southern margin of the Mesozoic Tethys, while the Mecsek—Villány complex were part of the northern margin.

The distribution and character of Pliensbachian brachiopods in the western Tethys has recently been studied by VÖRÖS (1977). He considered the South Alpine and South German fauna as type-fauna of the Mediterranean and NW European provinces, respectively. By comparing the fauna of a given locality with these type-faunas, the Mediterranean or NW European character of the locality can be determined by quantitative methods (e.g. by counting the Jaccard coefficient). In the present work 23 brachiopod localities were considered from the area under investigation. The results are shown on Fig. 1. Circles with cross-hatching and open circles show localities with Mediterranean and NW European faunal character, respectively. The larger circles represent faunas with large number of species, the character of which could have been determined quantitatively. The smaller circles represent occurrences with a few but characteristic brachiopod species. Figure 1 clearly shows that the European margin and the area to the southeast of the Intrapannonian mobile belt is

Fig. 1. The main paleogeographic units of the Carpatho–Pannonian area (mainly after MAHEL, 1973; Geological map of the Paleozoic and Mesozoic basement of Hungary, 1967). The circles show the character of the Lower Jurassic brachiopod fauna of a given locality. Cross-hatched circles: Mediterranean province, open circles: NW European province. Localities: 1. Southern Alps; 2. Northern Calcareous Alps; 3. Gresten; 4. Pristodolok; 5. Jablonicke Hills; 6. Čahticke Hills; 7. Zliechov; 8. Kostelec; 9. Belanska valley; 10. Drienkyna valley; 11. Lupcianska valley; 12. Vernar; 13. Geravy; 14. Bleskovy pramen; 15. Padurea craiului; 16. Sviniţa; 17. Villány; 18. Mecsek; 19. Bakony; 20. Vértes; 21. Gerecse; 22. Rarau; 23. Brasov. *Keys:* 1. Stable Europe; 2. Molasse foredeep; 3. Flysch on the surface and below Neogene cover; 4. Pieniny Klippen belt on the surface and below Neogene cover; 5. Penninicum and its possible extension below the Neogene cover; 6—7. Austroalpine and South Alpine realm on the surface and below Neogene cover, respectively (Southern marginal complex); 8. Tectonic units associated with the northern continental margin; 9. Ophiolites: large outcrops, minor bodies and ophiolitic bodies below thick sedimentary cover, respectively; 10. Tectonic boundary on the surface and below Neogene sedimentary cover; 11. Boundary of outcrops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characterized by NW European fauna. On the other hand, the distributions of the Mediterranean fauna clearly outlines the regions which were far from the European margin in Early Jurassic times.

C) *Paleomagnetic Data*

In the recent years instrumental development has made it possible to measure the low magnetic remanence of sedimentary rocks with good accuracy. The main interest of paleomagnetic research in Alpine Europe has been focused on the Mesozoic (mainly carbonate) sediments of the Periadriatic region (Sicily, Italy and Southern Alps). All the available data have recently been reviewed by CHANNELL and HORVÁTH (1976) and CHANNELL et al. (1978). Paleomagnetic data suggest that Adria has been part of the African plate since the Early Mesozoic, and the Periadriatic mountain belt was produced by the deformation of its marginal complex, due to the movement of the African plate.

Paleomagnetic measurements on the Jurassic limestones of the Transdanubian Range have given also African directions, i.e. the region haves probably formed part of the African plate in the Adriatic region (Márton and Márton, 1978). On the other hand, paleomagnetic data from the Upper Jurassic—Lower Cretaceous beds of the Villány Mts indicate their European affinity (MÁRTON, pers. comm.).

D) *Suture Zones*

The western Carpatho-Pannonian region (excluding the Flysch Belt) appears to be associated with the margin of the oceanic Tethys, which got juxtaposed with the European marginal complex in the course of Cretaceous-Tertiary convergence. If this interpretation is correct then sites of subduction, i.e. suture zones and/or transform boundaries should be associated with the unit. The Pieniny Klippen belt and the inner part of the Magura flysch zone can be considered as a suture zone. The Klippen belt is an extremely compressed zone. It is made up mainly of Mesozoic limestone blocks which are generally in tectonic contact with the surrounding marl and flysch envelope of Albian to Paleogene age (ANDRUSOV et al., 1973). The steeply dipping thrust planes, the presence of ophiolitic fragments in the Lower and Middle Cretaceous beds, and hP/IT metamorphic rocks in the Cretaceous conglomerates lead to the suggestion that the belt is a suture zone (SZÁDECZKY-KARDOSS, 1973; NEY, 1975; BIRKENMAJER, 1976; MAHEL, 1977). Crustal structure and the interpretation of the Carpathian conductivity anomaly support this conclusion (ÁDÁM et al., this volume).

Traces of an other possible plate margin can be found in the Intrapannonian belt which separates the western Carpathian-Pannonian unit from the east-

ern part of the Pannonian Basin (Fig. 1). This zone is marked with ophiolites and flysch deposits (JUHÁSZ and VASS, 1974; SZEPESHÁZY, 1975, 1977). In the Duna—Tisza interfluve region the ophiolites were found in boreholes in constantly changing levels between 1100 m and 2300 m, a fact that is attributed to younger Tertiary movements which disrupted and truncated the once continuous sheets. The complex comprises mainly of volcanic glass, spilites and diabases, and has a sedimentary cap of radiolarites and cherts. Dolerites occur and there is an occasional gabbro, but ultramafics are generally absent. The associated flysch deposits are of Upper Cretaceous-Paleogene age. In most places the Intrapannonian ophiolites overlie Upper Triassic beds, while their top cover is provided by Miocene and Lower Pannonian sediments on the west and by the flysch on the east. Jurassic-Lower Cretaceous ages of the ophiolites have been suggested (SZEPESHÁZY, 1977). There is evidence that the belt was compressed during the Late Cretaceous and at the beginning of Miocene. The location of ophiolitic bodies below the sedimentary cover of the Pannonian Basin suggests that the Intrapannonian belt is presumably connected with the Vardar zone on the southwest (Fig. 1).

It had been known for a long time that mafic and ultramafic rocks outcrop in some places close to the Darnó line (e.g. at Szarvaskő), yet the identification of these rocks as members of the ophiolite suite, and the subsequent recognition of their tectonic significance has come about only in the last couple of years. The best outcrops are near Mt. Darnó, Egerbakta, Bátor, Szarvaskő, Bélapátfalva and in the Bódva valley (LENGYEL, 1957; ZELENKA, 1975). The complex is mainly made up of radiolarites and cherty sediments, spilites, pillow lavas and diabase dykes, diorites, and ultramafics (in a more or less pseudostratigraphical order, from top to bottom). The rocks have been metamorphosed only to the prehnite-pumpellyite stage. The major and trace element geochemistry of the volcanic rocks suggests an ocean floor origin for them, and the differentiation and fractionation trends agree well with those for abyssal tholeiites from modern ocean floors (ONUOHA, 1977). The chemical composition of these rocks point to the fact that they could not have been created in their present positions by intraplate volcanism.

This is supported by observations in the Gemeric region where other ophiolites occur in several places (Fig. 1). These ophiolitic bodies which vary in dimension from small fragments to blocks of several kilometers in diameter, are considered rootless, having been emplaced into their present positions by tectonic processes (HOVORKA and ZLOCHA, 1974). The Gemeric ophiolites show higher degrees of serpentinization and metamorphism, than those of the Darnó zone. They are usually associated with the Meliata series which is dominantly made up of radiolarites, radiolarian shales and schistose carbonates with bodies of glaucophanic rocks. It has been recently realized that the Meliata series

(including the ophiolites) can be interpreted as remnants of oceanic crust (MAHEL, 1975). Their age is, however, a matter of debate. Until recently it was thought that the age of the Meliata series is Late Permian, but KOZUR and MOCK (1973) demonstrated its Middle and Late Triassic age on the basis of conodonts. This recognition made it possible to consider the Triassic shale complex of the Bükk Mts which is characterized by similar facies and fauna, to belong to the same development as the Meliata series (KOZUR and MOCK, 1977). There is no reliable age data for the Darnó and Gemeric ophiolites which are associated with the Meliata series development. It is widely believed that they are of Triassic age (HOVORKA and ZLOCHA, 1974), but ANDRUSOV et al., (1973), postulated that they are of post-Triassic age. MAHEL (1975) thinks that the ophiolitic magmatism started in the Triassic but some of the ophiolites are probably of Late Jurassic age. The Meliata series rest on the North Gemeric Paleozoic (Volovec block) probably in tectonic contact. ROZLOŽNIK (1978) called attention to the fact that these two complexes constitute a particular "paired metamorphic belt". The Volovec block shows hT metamorphism which culminated during the Late Cretaceous with the intrusion of granitic masses. The overlying Meliata series is characterized by hP/lT metamorphism which predated the hT metamorphism. The presence of the paired metamorphic belt also suggests that subduction took place here. However, the age and spatial relationships of the two complexes make their interpretation in a simple subduction model very difficult. We call it the Mesozoic ocean branch which was consumed here "the Meliata ocean" and suggests that the site of this subduction was associated with the Lubenik suture (now a fault zone where the Volovec overthrust the Veporides).

Plate-Tectonic Model

Our working hypothesis relies upon analogies with two adjacent segments of the European Alpides: the larger part of the Western Carpatho-Pannonian region (lying to the northwest of the Lubeník — Margečany line) shows facies and structural similarities to the Eastern Alps while South Alpine — Dinaric analogies can be recognized in the southeastern zones.

It is generally held that a fundamental difference between the Eastern Alps and the Western Carpathians is the lack of Penninic windows in the latter. Here we put forward the hypothesis that the Veporides were in similar paleogeographic position and played a similar tectonic role as the Briançonnais — Central Penninic unit of the Alps. The evidences are mainly circumstantial. The Mid Cretaceous orogenic events in the Inner Western Carpathians are hardly understandable ortherwise than in terms of continent — continent (or island arc) collision. In our model the Veporic unit serves as a northern microcontinent which was overridden by a part of the southern marginal complex

(Pieniny Klippen belt, Tatrides, Križna and Choč units). The strong Middle to Late Creataceous metamorphism of the Veporic basement and its sedimentary cover (Föderata and Struženik series) (BAGDASARIAN et al., 1977; KRÁL, 1977) is due to a burial by the thick basement nappe of the Tatrides.

We think that this deviation from the classical concept is a fruitful one and helps to explain some important features of the Carpathian orogeny. Using

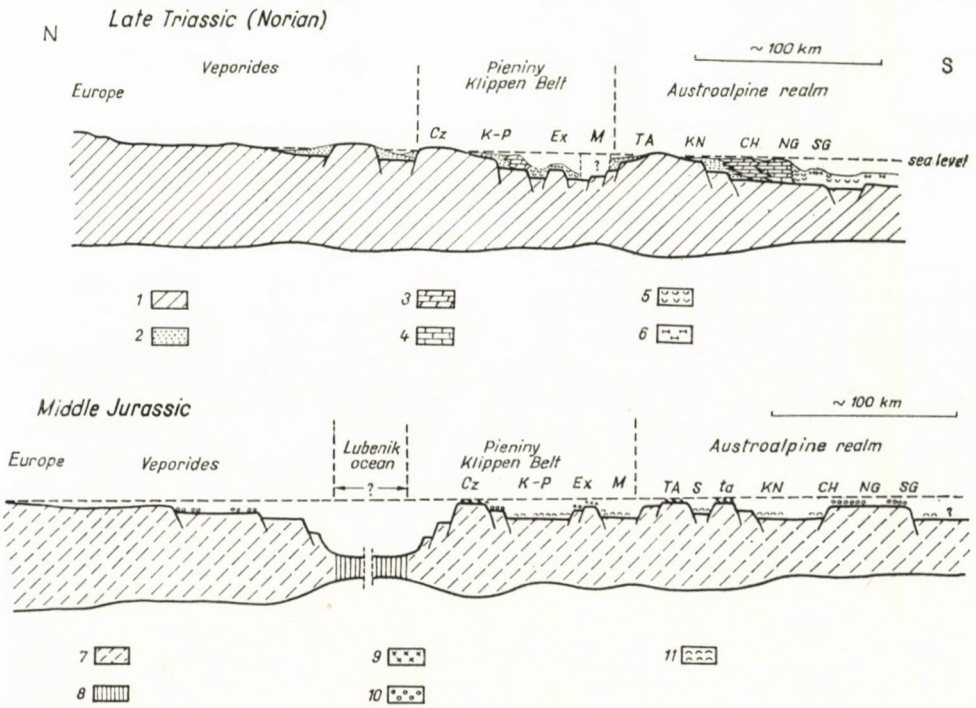
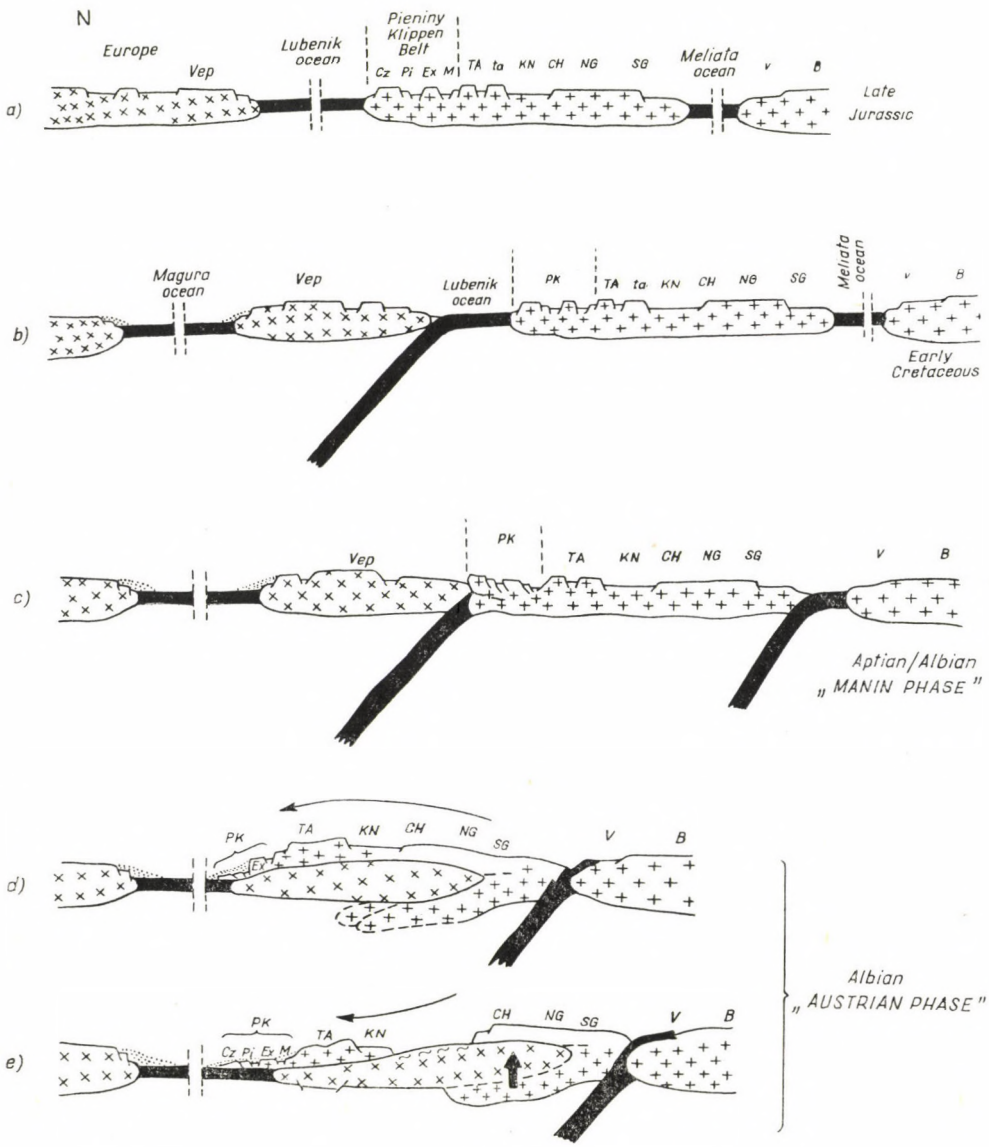


Fig. 2a—b. Schematic palinspastic sections through the Western Carpathians in the Late Triassic and Middle Jurassic. Cz = Czorsztyn; K—P = Kysuca-Pieniny; Ex = Exotic (Pieninic) ridge; M = Manín; TA = High Tatra, or Tatric in general; KN = Križna; CH = Choč; NG = North Gemeric; SG = South Gemeric; S = Šipruň; ta = Lower Tatra; 1. Basement including pre-Late Triassic formations; 2. Carpathian Keuper; 3. "Hauptdolomit"; 4. Dachstein limestone; 5. Cherty limestones; 6. Hallstatt limestone; 7. Basement including pre-Middle Jurassic formations; 8. Oceanic crust; 9. Crinoidal limestones; 10. Ammonitic facies („Ammonitico rosso", "Fleckenmergel"); 11. Radiolarites

our hypothesis as a premise we suggest a plate-tectonic model for the tectogenesis of the Western Carpatho-Pannonian region. In the course of the construction of this model we relied mainly on the following works: BIELY and FUSÁN, 1967; ANDRUSOV, 1968; ANDRUSOV et al., 1973; ANDRUSOV, 1975; MICHALIK, 1977; BIRKENMAJER, 1977; MIŠIK et al., 1977; MAHEL, 1977.

Figure 2a shows the Late Triassic paleogeographic section. The Tethys is characterized by platform — trough morphology with the appropriate sedi-

mentation. Southward from the S. Gemerides deep basin sedimentation occurred and spreading of the Meliata ocean might have started. The axis of rifting migrated toward the north and by about the Middle Jurassic a real oceanic area was formed by spreading between the Veporides and the Pieniny Klippen belt. We call it the Lubeník ocean. During the Late Jurassic — Early Cretaceous the Veporic unit was separated from stable Europe by the opening of the Magura trough (Fig. 3a, b).



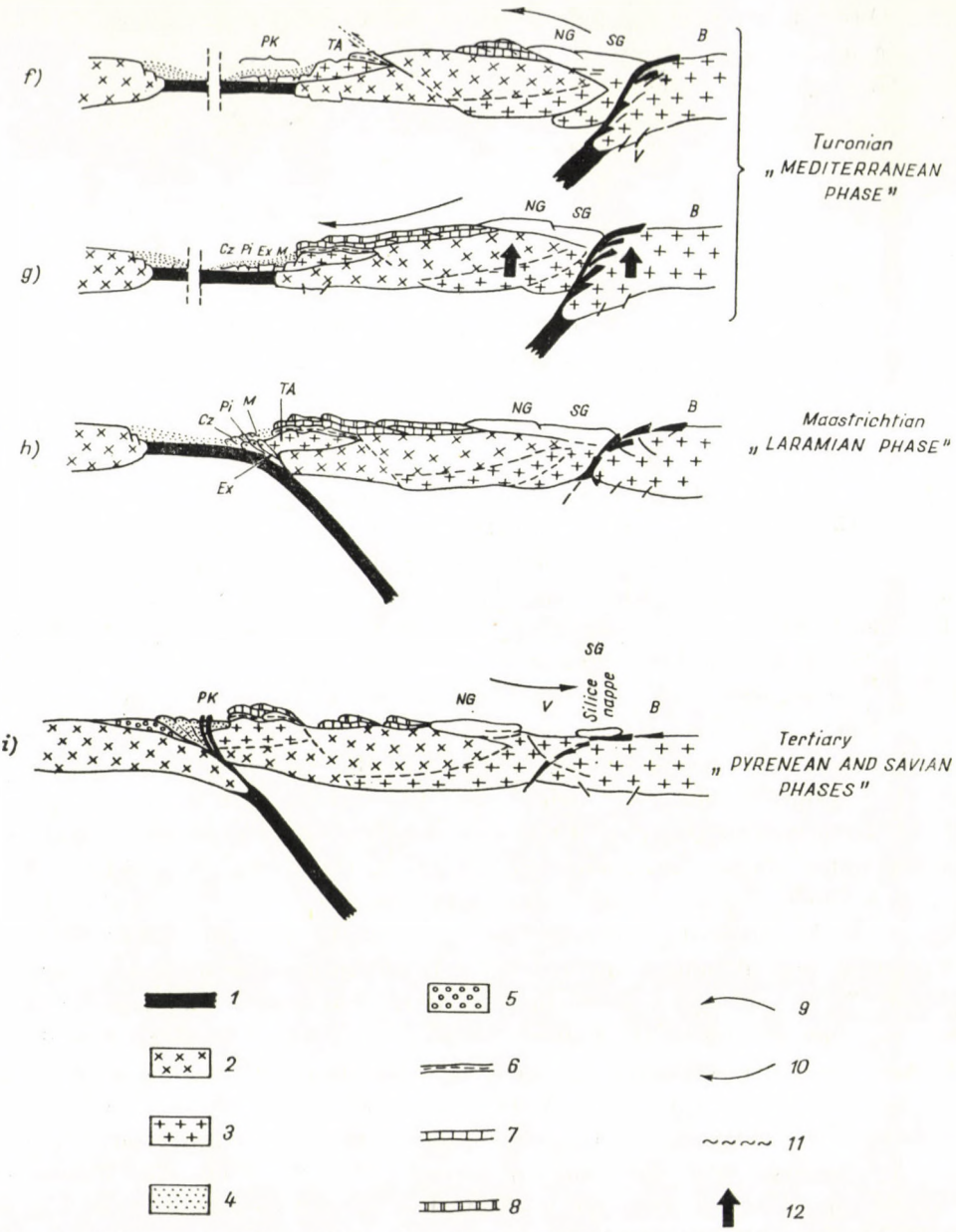


Fig. 3a—i. Schematic palinspastic sections through the Western Carpathians from the Late Jurassic to Tertiary. Keys: 1. Oceanic crust; 2. European continental crust; 3. Continental crust of the southern margin; 4. Flysch deposits; 5. Molasse; 6—7—8. Middle cretaceous nappes (Križna, Hronic and Gemeric nappes, resp.); 9. Overthrusting; 10. Gravity sliding; 11. Metamorphism; 12. Isostatic; uplift; Vep = Veporides; Pi = Pieniny zone (s. s.); PK = Pieniny Klippen belt; V = Volovec; B = Bükk

Our Early Cretaceous palinspastic section fits into the paleogeographic picture of the adjacent areas with the following analogies:

Magura flysch trough ~ Valais — North Penninic trough

Veporides ~ Briançonnais — Central Penninic ridge.

Lubeník ocean ~ Liguride — Piemontais ocean (i.e. the main Tethys branch)

Inner Carpathian zones

(From the Klippen belt to the S. Gemerides) ~ Austroalpine units

On the other hand, the Meliata ocean could be the northwestward continuation of the Subpelagonian — Serbian trough while the Volovec — Bükk zone shows South-Alpine — Dinaride affinity.

The subduction of the Lubeník ocean beneath the Veporides commenced during the Early Cretaceous and further spreading in the Magura trough (now possibly of marginal basin character) occurred. By the end of the Aptian the Lubeník ocean closed and as a consequence of this collision the first phase of deformation in the Klippen belt and the northward subduction of the Meliata oceanic crust took place (Fig. 3c). During the Albian the upper crustal layers of the whole southern marginal complex — a relatively thin flake (Klippen belt) and a thick basement sliver (Tatrides) — overthrust the Veporides giving rise to their Middle Cretaceous metamorphism. At the same time the Meliata ocean was closed and the Meliata series acquired their hP/1T metamorphism along the subduction zone; some ophiolitic and sedimentary masses were obducted to the Volovec block (Fig. 3d). By the end of the Albian the Pieniny Klippen belt, the Tatrides and the Križna unit were transported by gravity sliding to the rear of the isostatically uplifted Veporides and glided partly to the Magura marginal basin (Fig. 3e). During the Turonian further compression and gravity sliding caused the emplacement of the Križna, Choč and Gemeric nappes; the Volovec block was deeply underthrusted in the form of limited crustal subduction which led to the generation of the "Gemeride granites" (Fig. 3f—g). The subduction of the Magura ocean started in the Senonian and resulted in the Latest Cretaceous — Paleocene deformation of the Klippen belt. It was a time of isostatic adjustments and relaxation in the more southern zones, e.g. uplift of the Volovec block together with its allochthonous Meliata cover (Fig. 3h). It is very probable that the emplacement of the south Gemeride unit (Silica nappe) occurred after this time. The Paleogene subduction consumed much of the oceanic basement of the Magura trough, and, probably towards the end of the Oligocene, continent — continent collision started. Continued compression caused the formation of the nappe structure of the Outer Carpathians and led to their overthrusting unto the molasse (Fig. 3i). It is possible that not only the Outer but also the Inner Western Carpathians were overthrust unto the molasse during this compressional period (TOMEK pers. comm.). It is however certain that these tectonic events together

with subsequent younger compressions helped to modify and in some cases completely obscured the pre-Tertiary structural patterns of the Inner Western Carpathians.

*

In conclusion, we would like to stress that our model which is a working hypothesis may not claim to provide solutions to all the main geological and geophysical problems of the Western Carpatho-Pannonian region. We are, however, convinced that it will provoke serious discussions and further investigations which will ultimately lead to a better understanding of the Alpine evolution of this bit of the Earth's crust.

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ТЕКТОНИКА ПЛИТ В ЗАПАДНОЙ ЧАСТИ КАРПАТО-ПАННОНСКОЙ ОБЛАСТИ: РАБОЧАЯ ГИПОТЕЗА

Ф. ХОРВАТ, А. ВЕРЕШ, К. М. ОНУОХА

Резюме

Альпийский пояс состоит из южных и северных прибрежных образований мезозойского Тетиса, современное положение которых создалось в течениим мела и третичного периода конвергенцией африканской и европейской плит. По фациологическим палеонтологическим и палеомагнитным данным Внутренные Западные Карпаты и северозападная часть Паннонского бассейна принадлежили к южному сектору, как часть африканско-адриатического носа. Остатки оцеанского Тетиса обнаружены в Пьеннинской зоне и вдоль Гемерско-дарновского линеамента. Предполагается что серия фёдэрата Вепорского кристаллического массива бриянсоннейского типа. Они появляются в виде тектонических окон. Исходя из этой гипотезы можно разрабатывать достоверную плитово-тектони ческую модель западной части Карпатопаннонской области.

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PLATE-TECTONIC EVOLUTION OF THE EASTERN ALPS[1]

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The plate-tectonic evolution of the Eastern Alps can be well explained by the process of suture progradation in the sense of ROEDER. The tectonic activity (subduction of the South Penninic ocean, subduction of the North Penninic ocean, underthrusting in the molasse foredeep) shifts from internal zones in the south to external zones in the north. The South Penninic ocean which represents the continuation of the Central Atlantic in the Jurassic, begins to be consumed along a southward-dipping subduction zone in the Lower Cretaceous and is closed by the collision of the Austroalpine and the Middle Penninic continental realms in the Upper Cretaceous. A paired metamorphic belt develops during this process. With this first collision subduction shifts to the meanwhile formed small North Penninic ocean which is, in turn, closed in the Eocene. Subsequently, the axis of deposition and tectonic activity shifts again to the north into the molasse foredeep.

This concept also applies to the Western Alps and is in accordance with the suggested plate movements in the western Mediterranean region.

The mutual action between rigid continental masses and mobile oceanic zones which are enabled to subduction into the asthenosphere, plays the key role in the organic process. In general, continent—continent collisions are the climax of orogenesis and ultimately lead to a gradual cessation of the orogenic movements due to the enormous friction between the colliding continental masses which cannot be subducted. In most orogens, however, after a first continental collision, orogenic activity shifts to other zones of weakness: either into young oceanic basins or remaining basins between the irregularly shaped continents, or into a continental foredeep. If the shift of orogenic activity is in the direction of vergency, i.e. into external parts of the orogen, we are speaking of "suture progradation", and of "suture retrogradation" if it is in the opposite direction (ROEDER, 1977). The Alps are an excellent example of suture progradation.

The Alpidic Mediterranean belt is a large, complex continent—continent collision. Between the great blocks of Eurasia and Africa smaller continental masses split off during orogenic evolution. These are the so-called microcontinents which may move quickly and rotate around large angles. For the Alps, the Adriatic plate plays an important role.

[1] This paper is an abstract of a lecture held at the KAPG symposium in Sopron in April 15, 1978. Fore more detailed information the reader is referred to FRISCH, 1977, 1978, and in press.

There are three domains in the nappe pile of the Alps underlain by continental crust: these are, from south (internal part) to north (external part) the Austroalpine-South Alpine domain, the Middle Penninic or Briançonnais domain and the Helvetic domain. These continent-based zones become separated during Alpidic evolution by oceanic basins being active in succession: the South Penninic or Piedmont (Piémontais) zone, and the North Penninic or Valais (Valaisan) zone. These zones are bare of continental crust except of some small splinters, contain remnants of ophiolites, pass partly into mélange structures and are made up of marly to clastic, often turbiditic series which represent deep-water basin and trench-fill sediments. In the South Penninic domain there are, in addition, high-pressure metamorphic zones.

So we have good evidence for the oceanic nature of the two Penninic basins, and for their subduction along southward dipping subduction zones. The constraints are stronger for the Piedmont basin than for the Valais basin. It must be emphasized, however, that although we can fairly well reconstruct fossil subduction activity, we miss any hint for volcanic arcs or marginal basins formed during the subduction phases.

The evolution of the Alps can be divided into five periods which are in part identical with the five epochs of CLAR (1973). In the Permo-triassic, the *first period*, all continent-based zones of the Alps were united in Pangaea which was bordered to the east by the Tethys ocean (Fig. 1a). Extensive shelf seas spread from this ocean to the west causing progressive transgression onto the European-African continent displaying the features of a subsiding Atlantic type continental margin.

Paleomagnetic data (ZIJDERVELD et al., 1970; SOFFEL, 1972; CHANNELL and HORVÁTH, 1976) show that the Adriatic region was similarly oriented as the African continent which caused CHANNELL and HORVÁTH (1976) to consider the Adriatic spur as an African promontory in the sense of ARGAND (1924). New paleomagnetic evidence shows the Northern Calcareous Alps (Upper Austroalpine nappe) to be rotated clockwise about 90—100° relative to the Adriatic region (MAURITSCH and FRISCH, 1978).

The *second period* begins in the Liassic with the opening of the South Penninic or Piedmont ocean which separates the Austroalpine-South Alpine domain from Europe (Fig. 1b). The Austroalpine realm remains connected with Africa by the Adriatic region forming a spur of the Gondwana continent, the African promontory of ARGAND (1924) and CHANNELL and HORVÁTH (1976). The consistency in the evolution of the central Atlantic and the Piedmont ocean and the fact that there is no other realistic possibility of continuation for the central Atlantic leads to the conclusion that the Piedmont ocean is the continuation (that means: the *only* continuation) of the early central Atlantic, connected with it by a major ridge-ridge transform fault. This concept deviates from that of BIJU-DUVAL et al. (1977), who already separate the Adriatic

plate in the Jurassic during our second period by the creation of an ocean, the Mesogea, between the Adriatic region and the African continent.

Constructions using a rotation pole between Laurasia and Gondwana in northern Europe show that the width of the Piedmont ocean remains limited. Only in its southwestern part the width will have exceeded 500 km in the Lower Cretaceous, the time of presumed maximum extent of this ocean.

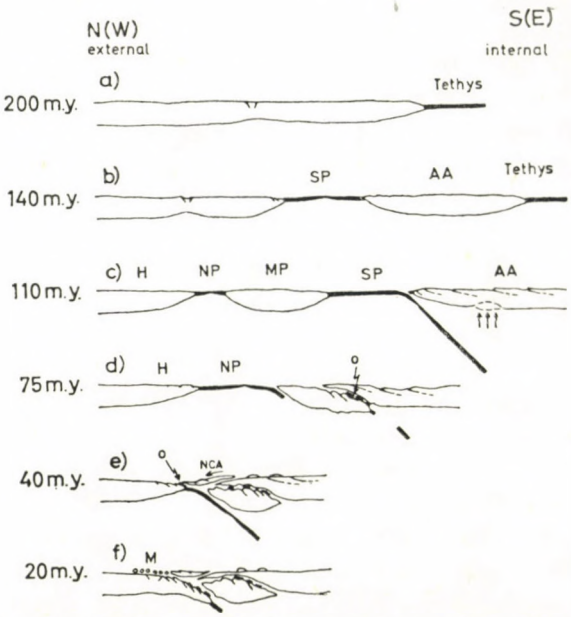

Fig. 1a—f. Evolution of the Eastern Alps in cross sections through the Tauern window: *a)* during the first period; *b)* during the second period; *c)* during the third period; *d)* at the beginning of the fourth period; *e)* at the end of the fourth period; and *f)* during the fifth period. AA — Austroalpine (—South Alpine) zone [N.C.A.: Northern Calcareous Alps]; SP — South Penninic (Piedmont) zone; MP — Middle Penninic (Briançonnais) zone; NP — North Penninic (Valais) zone; H — Helvetic zone; M — Molasse zone (small circles). Shown is continental crust (blank) and oceanic crust (black). o = site of ophiolite obduction. Wavy arrows in *c):* site of increased heat flow. Slightly exaggerated

During the second period, the Austroalpine-South Alpine zone was surrounded by oceanic realms on two sides (Fig. 1b). This is reflected in a sudden facies change in the Liassic from carbonate platform to pelagic conditions.

At some time in the Lower Cretaceous, the Piedmont ocean begins to subduct along a southward dipping subduction zone under the Austroalpine-South Alpine domain (Fig. 1c). This event opens the *third period* and may be marked by the appearance of lime-free black shales and turbidites in the southern part of the Piedmont ocean, so for instance the Fusch facies series of the

Tauern window. They are connected with coarse clastic material and olistolites and are considered to represent, at least in part, trench-fill sediments.

Contemporaneous with the subduction, a sort of paired metamorphic belt develops with high-pressure assemblages at the ophiolite-bearing base of the South Penninic Glockner nappe of the Tauern window. Corresponding mineral assemblages in the same position yielded radiometric cooling ages of about 80 m.y. in the Western Alps. In the Austroalpine realm, there is a metamorphic event with about normal or a little lowered P/T gradients which yielded cooling ages in the same range, of about 80 m.y.

The reason why there is no pronounced high-temperature—low-pressure metamorphism is most likely the same as why there is no volcanic arc: the Piedmont ocean was too small so that only a short subduction zone could form which was able to pierce into the low-velocity layer of the upper mantle for a geologically short period only (Fig. 1c). Subduction in the Piedmont ocean terminated with the collision of the Austroalpine and Middle Penninic (Briançonnais) zones and the north-vergent overthrusting of the former over the latter (Fig. 1d) which probably occurred at some time in the early Upper Cretaceous in the Eastern Alps but later (early Tertiary) in the Western Alps. An oblique overthrusting of the Austroalpine over the Penninic domain has been already postulated by CLAR (1971).

Contemporaneous with the subduction in the Piedmont ocean the North Penninic or Valais ocean opens by the separation of the Briançonian plate from the European continent and its southward drift (cp. BÖGEL and CLOSS, 1976). The opening of the Valais ocean is believed to correlate with the break-through of the Atlantic to the north and the opening of the Bay of Biscay.

The Adriatic region is considered to separate from the African continent as an independent plate at some time in the Lower Cretaceous which is supported by two reasons: *a)* the fast eastward movement of Africa relative to Eurasia in the Cretaceous would imply large areas of continental overlap in the Dinaric region which contradicts geological evidence; *b)* the Piedmont ocean would widen enormously in the Upper Cretaceous although we know that this ocean already closes up at this time and that there are important compressional events in the southern Penninic zones and the Austroalpine domain.

The collision of the Adriatic plate with the Briançonian plate causes the disintegration of the counter-clockwisely rotating Adriatic plate. The Northern Calcareous Alps are considered to commence with their clockwise rotation of about 40—50° relative to Eurasia at the time of the first collisional contact (MAURITSCH and FRISCH, 1978). The behaviour of their basement is yet unknown.

The Valais ocean did not reach a width exceeding a few hundred kilometres. Subduction may have begun at the moment when subduction ceased in the Piedmont ocean, or somewhat later (Fig. 1d). Anyhow, the first conti-

nental collision in the South Penninic zone caused the compressional events to shift to the north into more external parts of the orogen, the Valais basin, thus opening the *fourth period*. The principal axis of deposition has already shifted into this basin some time before.

During the fourth period the oblique front of the Austroalpine nappe system over the Penninic zones persists and parts of the Piedmont zone of the Western Alps remain uncovered. This may explain the similarity in facies between the Rheno-Danubian flysch of the Eastern Alps which is North Penninic and the Helminthoid flysch of the Western Alps which is South Penninic. The Valais ocean is subducted in this period and its closure occurs in the Eocene by the underthrusting of the foreland, the Helvetic zone, under the southern continental block in which Briançonnais, Piedmont and Austroalpine are already welded together and are part of the upper plate (Fig. 1e). This is the second continent—continent collision. With this event, the rocks of the Middle and South Penninic zones exposed today in the Tauern window, become an allochthonous pile and suffer regional metamorphism. The Northern Calcareous Alps glide down from the updoming central axis into the trench in the north, thus accelerating the isostatic upward movement of the central axis of the range (Fig. 1e).

With this second continental collision in the Eocene, the *fifth period* begins (Fig. 1f). The depositional trough again shifts to the north. For the lack of another oceanic zone, a foredeep forms on the foreland by downbuckling of the continental crust. The foredeep catches up the molasse sediments coming from the isostatically raising Alpine range. Continuing compressional forces are responsible for the underthrusting of the molasse foreland under the nappes of the Alpine range in the Neogene, a process which is responsible for the overall allochthony of the Helvetic zone.

The twice prograding shift of the orogenic activity is a characteristic feature of the Alps and is geologically well documented (e.g. CLAR, 1973). The progressive disappearance of mobile, subductible oceanic lithosphere with simulateneously increasing involvation of continental masses into the thrusting processes is responsible for the gradual cessation of the compressional activity in the region concerned.

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РАЗВИТИЕ ВОСТОЧНЫХ АЛЬП ПО ТЕКТОНИКЕ ПЛИТ

В. ФРИШ

Резюме

Развитие Восточных Альп хорошо объясняется процессом «проградации сутуры» в смысле Родера. Тектоническая активность (субдукция дна Южно-паннонского океана, субдукция дна Северно-пеннинского океана, поддвиг в молассовом прогибе) переходит из внутренных, южных поясов к внешним, северным поясам. В нижнем мелу начинается консумпция южно-пеннинского океана, бывщего продолжения Центральной Атлантики в юре, вдоль зоны субдукции склоняющей к югу. Южно-пеннинский океан кончается коллизией Австроальпийской и Средне-пенинской континентальных плит во верхем мелу. В течение этого процесса развивается парный метаморфический пояс. После этой первой коллизии, субдукция переходит к только что возникшему маленькому Северно-пеннинскому океану, который заключается в Эоцене. Потом ось осадконакопления и тектонизма переходит ещё севернее, к молассовому прогибу.

Эта теория применимая и к Западным Альпам; она соответствует предпологвемым движениям плит в Западно-средиземноморской области.

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PALAEOMAGNETIC RESULTS OF THE NORTHERN CALCAREOUS ALPS, AUSTRIA

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In 1974 two profiles from Liassic to Lower Malm were sampled and a variety of rockmagnetic experiments were carried out. These experiments established that magnetite is the carrier of the stable part of the NRM in the red limestones of the Liassic and the Dogger. With this information the investigations were continued in 1975/76 using only the red Liassic limestone (Adneter Kalk) throughout the Osterhorn massif from Adnet south of Salzburg to Moosbergalm south of Strobl (Wolfgangsee). A clockwise rotation of the whole unit of at least 55° were found. Furthermore, the result supports the idea of the northward shift of the NCA since the Jurassic. For the comparison with stable Europe the compilation of paleopoles of KRS (1977) was used. Because this rotation and shift fit very well the trend found in other parts of the NCA and Graywacke zone, a clockwise rotation of the whole Upper East Alpine is assumed.

Introduction

Since the late fifties many investigations have been carried out on the counter-clockwise rotation of the Apennine peninsula (VAN HILTEN, 1960; MANZONI, 1970; SOFFEL, 1972; ZIJDERFELD—VAN DER VOO, 1973; LOWRIE—ALVAREZ, 1974; CHANNELL—TARLING, 1975; CHANNEL—HORVÁTH, 1976; VAN DEN BERG—WONDERS, 1976). As it is impossible to see the evolution of this area without considering its consequences in the Alpine belt and *vice versa* it was obviously necessary to check this rotation in the Eastern Alps. Based on the early results from Jurassic sediments in the area of Lofer (Austria) by HARGRAVES—FISCHER, 1959, and the fact that there is flat-lying Jurassic in the Osterhorn mountain groups, this area was chosen (Fig. 1), for the paleomagnetic investigations in the NCA. As in all the palinspastic reconstructions the NCA has to be placed much further to the south than its present-day position, much lower paleoinclinations could also be expected.

With these facts in mind, rockmagnetic investigations started in 1974 in two profiles (Schafbach- and Saubachgraben). In the Schafbachgraben (locality 2) 20 sites and in Saubachgraben (locality 3) 14 sites covered the Jurassic from the Lowermost Liassic to Lower Malm. (On a number of pilot samples high- and low-temperature, saturation magnetization and Ac-field demagnetization experiments were carried out.)

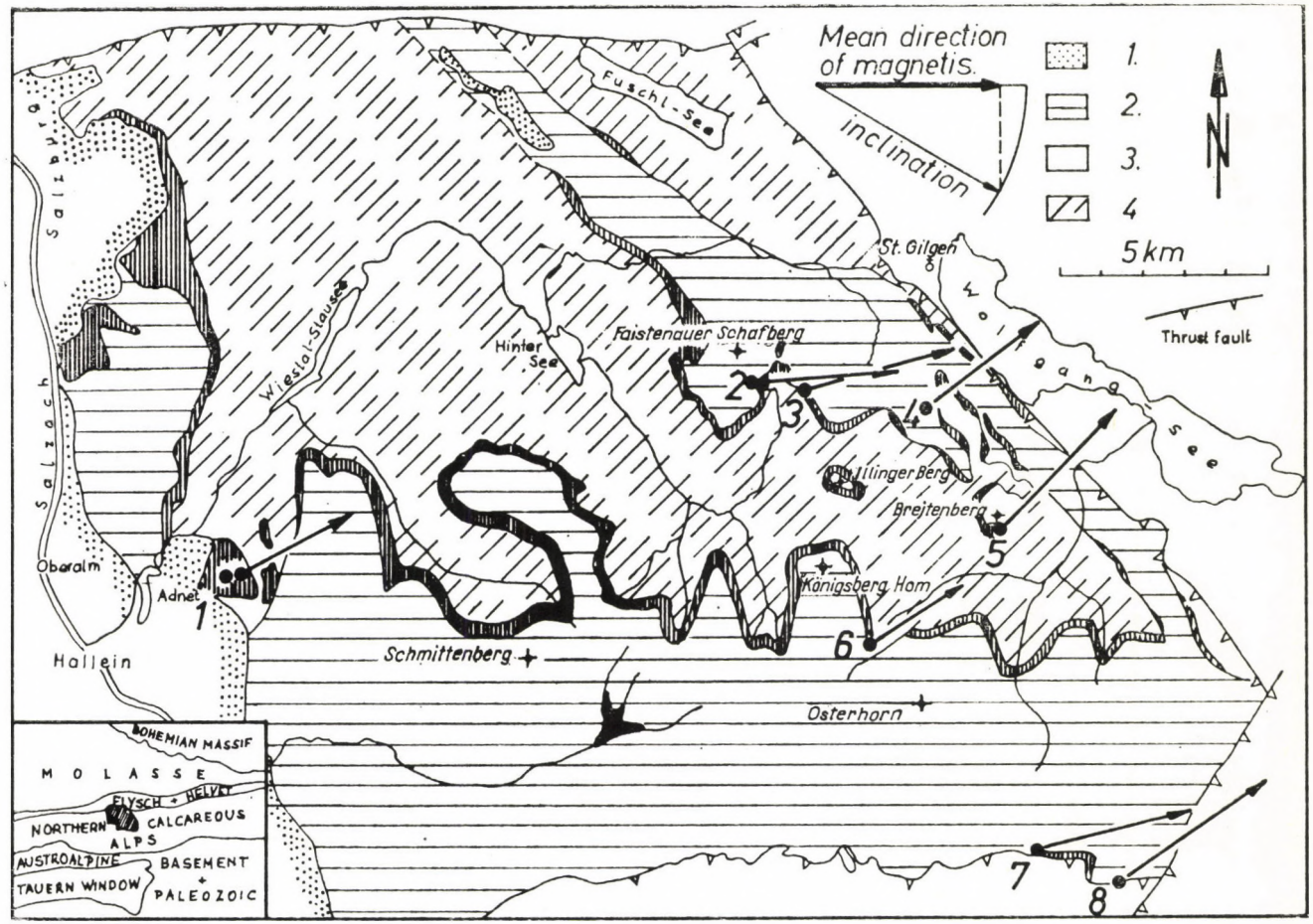

Fig. 1. Scetch map of the sampling area with the localities (numbers), mean directions and geological units: 1. Upper Cretaceous (Gosau Formation and Quaternary); 2. Malm and Lower Cretaceous; 3. Liassic and Dogger; 4. Upper Triassic

As it is shown in Fig. 3, there are differences in the behaviour of the Malm layers represented by S 3, S 4 in Fig. 3a and the Liassic silicified limestones represented by S 7 in Fig. 3a and the red Adneter limestones represented by S 8 in Fig. 3b. In both cases magnetite is the dominant carrier of the magnetic

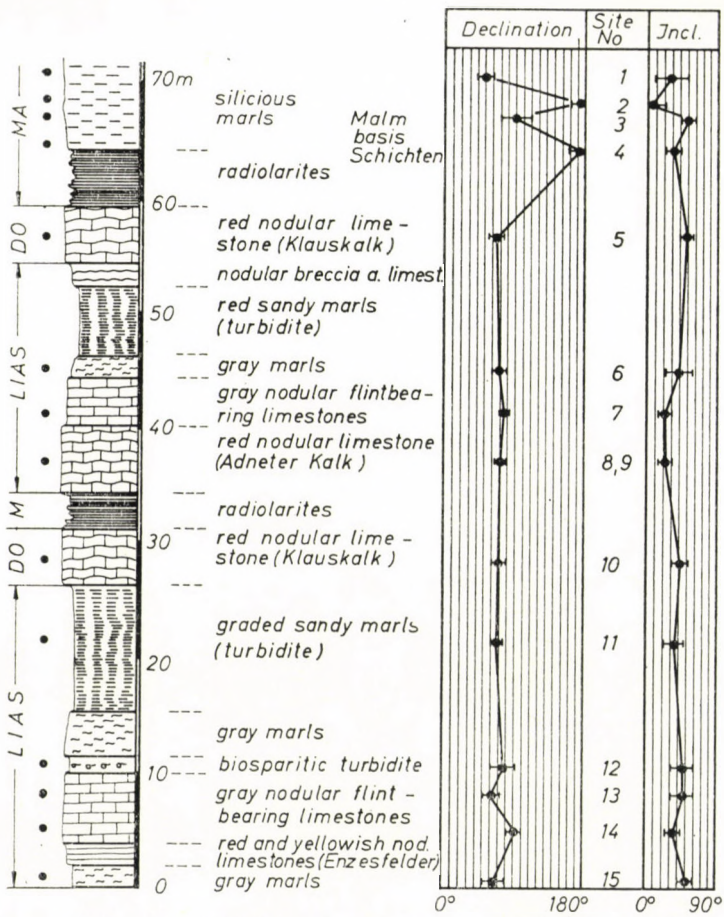

Fig. 2. Rockmagnetic result. *a)* High-temperature behaviour of Malm and Liassic (silicified) samples; *b)* high-temperature behaviour of a typical red-coloured Liassic limestone (Adneter Kalk); *c)* saturation magnetisation and Coercive force (H_{cr}) initial and after 200°C heating; *d)* K_1-transition for magnetite in red-coloured Liassic samples

properties but the susceptibility in Fig. 3a indicates a second mineralization which could be due to iron hydroxide. This assumption is supported by microscope analysis (limonite) and by Fig. 3c which indicates a chemical change in the initial mineralizations between 0 and 200 °C. The fact that a softer magnetization occurs after heating up to 200 °C can be explained by the dehydra-

tion of iron hydroxide (α and, or γ — FeOOH) and the formation of iron oxide (Fe_3O_4 and, or Fe_2O_3). Fig. 3d shows the low-temperature results which also indicate magnetite.

As the material was cleaned at 300 °C the distribution of the characteristic remanent magnetization (ChRM) shown in Fig. 2 is very satisfying.

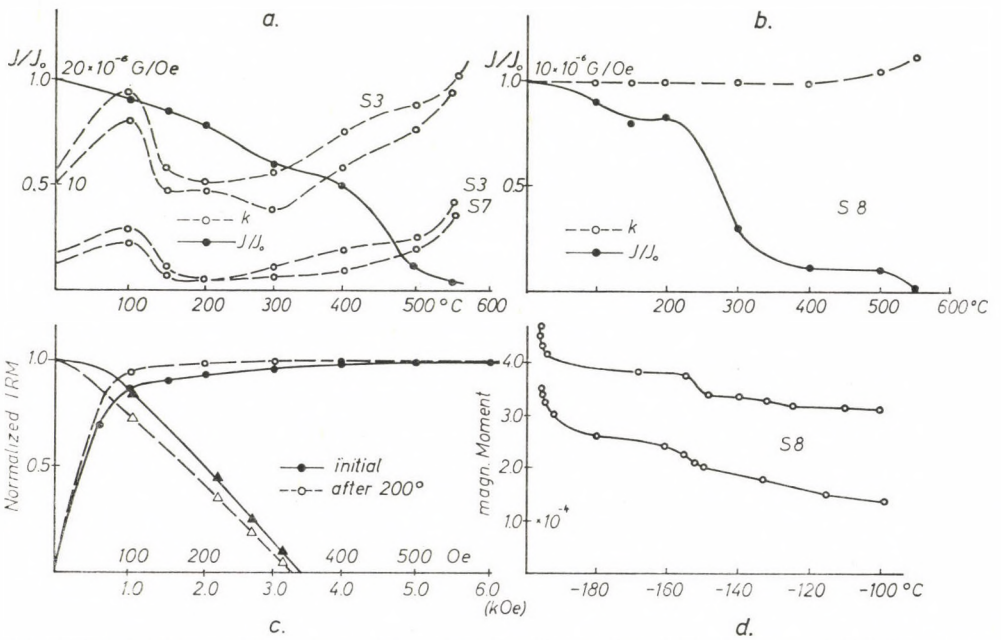

Fig. 3. Geological profile in comparison with the distribution of Declinations and Inclinations for the sampled sequence in the Saubachgraben

The larger scatter in the upper part of the profile is caused by internal movements during the gliding of the olistolite repetition (PLÖCHINGER, (1973).

Summarizing these investigations, it was decided to continue to use only the most stable material, the Adneter limestones, for the paleomagnetic investigations.

In 8 sampling areas (Fig. 1) samples were drilled in the field and oriented with sun or magnetic compass. The diameter and the length of the cores were 1 inch. The magnetic moment was measured with a spinner magnetometer (MOLYNEUX, 1971). The statistical analysis was made after FISHER (1953). As it is shown in Table 1, the mean directions and statistical parameter were satisfactory and a mean value for the whole mountain group was calculated.

Table 1

Paleomagnetic results from the Northern Calcareous Alps (Austria)

Sampling area	LOC.	Paleo data (Bedding corrected) DEC	INC	N(Sites)	k	COFC.	sampled members
Adnet	1	63.7	56.1	6	26.8	13.1	Adneter Kalk
Schafbachgraben	2	56.4	48.9	6	15.3	17.6	Liassic
Saubachgraben	3	71.1	42.3	8	34.1	9.6	Liassic
Breitenberg	5	59.3	24.1	6 cores	255.73	4.2	Adneter Kalk
Königsbachtal	6	57.9	63.8	2	417.3	12.2	Adneter Kalk
Ausserlimbach	7	75.7	49.1	2	688.6	9.5	Adneter Kalk
Moosbergalm	8	55.2	30.0	2	228.0	16.6	Adneter Kalk
Mean values Osterhorn Gruppe (7 locations)		62.5	45.1		29.1	11.3	
Lofer		47.9	50.6	30 cores	70.7	6.5	Liassic

Paleopole position for Lofer and Osterhorngruppe

	Sampling area position	Paleopole position	
Lofer	47.37°N; 12.34°E	LAT.: 53° N;	LONG.: = 112° E
Osterhorngruppe	47.7 °N; 13.3 °E	LAT.: 37,5°N;	LONG.: = 103.4°E

Results

Based on the mean directions in the 8 localities (Fig. 1), the mean direction for the whole mountain group was calculated and a 55° clockwise rotation was found. A comparison was made with the pole data for Lower Jurassic after VAN DER VOO—FRENCH (1974). As a pure rotation decribes a movement along a small circle the virtual pole for NCA and the pole after VAN DER VOO—FRENCH must lie on the same small circle if the rotation occurred in the present position. As it is easely seen in Fig. 4, this is not the case and it is necessary to shift the NCA from the present position (LAT.: 47.7; LONG.: 13 °E) to the paleoposition (LAT.: 33 °N; LONG.: 22 °E) over approximately 1200 to 1500 km within the statistical accuracy. The comparison with the compilation of KRS (1977) makes a shift of approximately 3000 km necessary. In both cases the direction of shifting is in good agreement with the geological idea of overthrusting of the Austroalpine system and fits also very well in very recent plate-tectonic models for this area (FRISCH, 1975).

This clockwise rotation was surprizing because the dominant trend of rotation in the Apennines, Southern Alps and even north of the Insubric line in the basement to the Permo-Triassic sequence of the Drauzug (Carboniferous of Nötsch) HEINZ (1978) was found to be counter-clockwise.

From the fact that clockwise rotations are prooved in a Jurassic sequence near Lofer (HARGRAVES—FISCHER, 1959), in Triassic material near Wörgl (SOF-

FEL, pers. comm.), Osterhorn massif (MAURITSCH, this paper), Triassic limestones of the Dachstein area (MAURITSCH, unpublished) and in the Western Graywake Zone near Saalfelden (MAURITSCH, 1978), one can postulate this clockwise rotation for the whole northern part of the Upper East Alpine. Furthermore, this rotation is in agreement with clockwise rotations in the Carpathians (Choč map) found by KRS (pers. comm).

Additional to the general trend of clockwise rotation there seems to be an influence by the Bohemian Massif. This idea is based on Triassic results by SOFFEL from an area near Wörgl (LONG.: 12.8 °E; LAT.: 47.5 °N) with a virtual pole position of LONG.: 92 °E; LAT.: 54 °N and by MAURITSCH from the

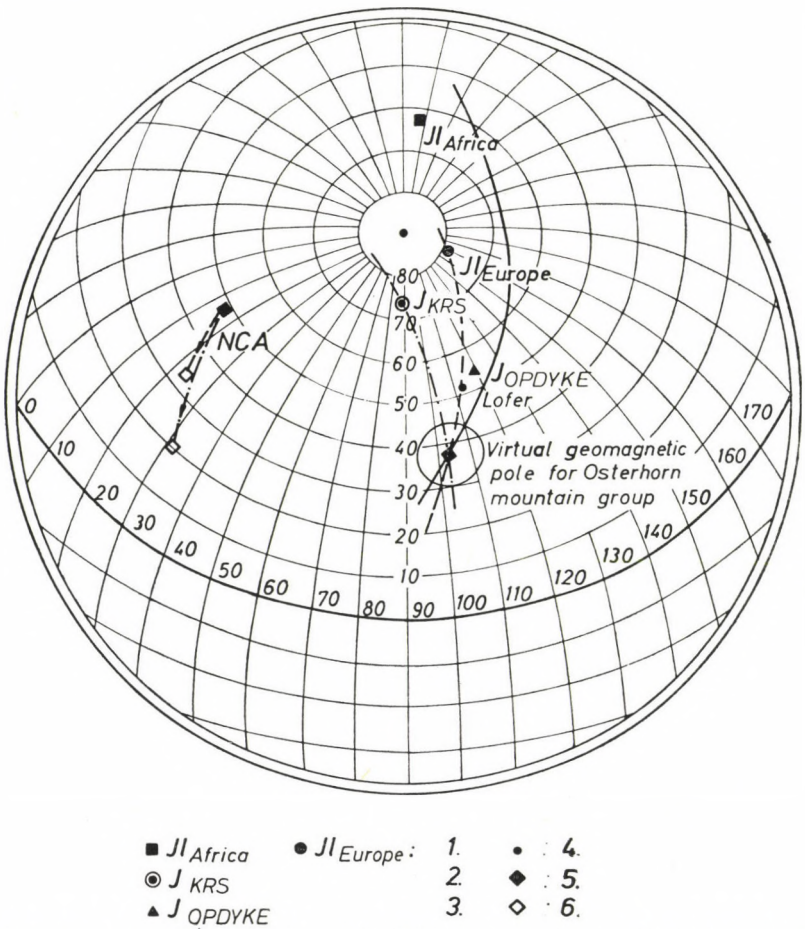

Fig. 4. Reconstruction of the rotation of the NCA in comparison with the Jurassic pole positions after OPDYKE, VAN DER VOO—FRENCH and KRS; 1. Stable Europe and Africa poles after VAN DER VOO—FRENCH; 2. Stable Europe pole after KRS; 3. Stable Europe pole after OPDYKE; 4. Virtual geomagnetic pole for Lofer after HARGRAVES—FISCHER; 5. NCA present position and virtual geomagnetic pole; 6. Shifted NCA positions after comparison with the Jurassic pole after VAN DER VOO—FRENCH—KRS for stable Europe

Dachstein area (LONG.: 13.6 °E; LAT.: 47.5 °N) and with a virtual pole position of LONG.: 168 °E; LAT.: 23 °N. As one can see there is an increase of clockwise rotation to the east.

The second comparison is for the Jurassic where near Lofer (LONG.: 12.34 °E; LAT.: 47.37 °N) HARGRAVES—FISCHER found a virtual pole position of LONG.: 112 °E; LAT.: 53 °N and MAURITSCH (this paper) a pole position of LONG.: 103.4 °E; LAT.: 37.5 °N. Again, there is an increase of rotation to the east. If we compare this increase of rotation with the form of the southern border of the Bohemian Massif then it makes sence that the additional rotation occurred under the influence of the Bohemian Massif during the gravity gliding processes of the NCA. Future work in selected profiles in the east of the NCA should check this idea.

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РЕЗУЛЬТАТЫ ПАЛЕОМАГНИТНЫЙ ИССЛЕДОВАНИЙ СЕВЕРНЫХ ИЗВЕСТНАКОВЫХ АЛЬП, АВСТРИЯ

Х. Й. МАУРИЧ

Резюме

В 1974-ом году образцы выли взяты в двух разрезах от лиаса до нижнего мальма и выли исследованы с точки зрения магнитности горных пород. Обнаружено что магнети является носителем стабильной части ЯМР в красных известняках лиаса и доггера. Получив эту информацию эксперименты продолжались в 1975—76 гг. только на красных известняках лиаса (Adneter Kalk) во всей территории массива Остерхорн от с. Аднет (на юг от г. Сальцбург) до с. Мосбергальм (на юг от Шторбля у озера Вольфгангзе). Установлено правая ротеция всей единици не менее за 55 градусов. Эти данные поддерживают предстановление о движении СИА в северном направлении после юрского периода. Для сравнения со стабильной Европой, данные о палеополюсах выли использованы (KRS, 1977).

Так как эта ротация и это горизонтальное движение соответствуют трендам установленным в других областях Северных известняковых албп и зоны Грейвакке, предполагается правая ротация всей Верхней восточно-альпийской единицы.

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A HYPOTHESIS OF THE DEVELOPMENT OF THE INSUBRIC—CARPATHIAN TERTIARY BLOCK SYSTEM

By

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The material necessary for a study of tectonic deformations in the Alpine belt has been obtained by synthesizing all hitherto published paleomagnetic data for Eurasia and Africa as well as by analyzing newly introduced geophysical parameters — paleorotation, paleoinclination difference, and paleodivergence of paleomagnetic directions. Interpretation of the paleomagnetic data supports the plate-tectonics concept with regard to the Tertiary development of the Alps and the Carpathians. The proposed model of development of the Insubric—Carpathian block system with respect to the movement of the Adriatic microplate is briefly discussed.

Introduction

Paleotectonic deformations which are particularly prominent in the Alpine belt can be reconstructed by using a geophysical model based on mathematical treatment of paleomagnetic data. The use of paleomagnetic data in deciphering paleotectonic history can best be illustrated by their statistical treatment. Comparison must be made between sets of values derived from *a)* stable platforms (plates, paleoplates), and *b)* regions exposed to the Alpine tectonics (block mosaics, sub-blocks, microplates). By statistically treating the newly introduced terms of paleorotation ΔD, paleoinclination difference ΔI and paleodivergence Φ, the resultant data clearly demonstrate a dominant dipolar nature of the mean paleomagnetic field for the entire Phanerozoic, except for the insufficiently studied Silurian (KRS, 1978, in press). At the same time, basic data are available for a quantitative evaluation of the paleotectonic deformations which are characterized by conspicuous horizontal rotations in the Alpine belt. On the basis of these data so analyzed, and with regard to both geological and drilling results, it is possible to present a rough picture of the development of the Alpine—Carpathian mountain system and its single blocks (sub-blocks) as far as the Adriatic microplate.

Paleotectonic Aspects of Interpretation of Paleomagnetic Data

This paper is successive to the statistical treatment of global paleomagnetic data (KRS, 1978) aimed at confronting them with the plate-tectonics concept even for the geological past. The Eurasian continent, as compared with the African plate, is suitable for such studies on account of the numerous published data available as well as the existence of the tectonically markedly mobile Alpine belt in which intense paleomagnetic studies have been carried out.

The Eurasian continent was divided into plates and zones comprising (i) the North European Platform, the so-called stable Europe, extending from the Ural Mts as far as Great Britain and north of the Alpine belt, (ii) the area east of the Ural Mts as far as the Lena river drainage basin, incl. a broader region of the Siberian Platform, and (iii) the Alpine belt including the regions south of it, the so-called mobile Europe. Eurasia was thus divided into larger plates (paleoplates) and the belt representing a mosaic of blocks (microblocks, microplates).

At an early stage of the paleomagnetic data study, paleomagnetic pole positions were treated statistically. All pole positions were given a unit weight of observation; that is, paleomagnetic dipoles were identical with unit vectors. FISHER's (1953) statistical method was used. A scatter of the mean paleomagnetic pole position (l, m, n) was determined for the N positions of those paleomagnetic poles showing a certain scatter of dipole axes as defined by the directional cosines l_i, m_i, n_i. The mean paleomagnetic pole position was determined by calculating the semi-vertical angle of the cone of confidence $\alpha_{(1-P)}$ at the probability level of $P = 0.05$. Stereographic projections of the paleomagnetic pole positions derived from Triassic and Permian rocks of the North European Platform (stable Europe), the Alpine belt and the regions south of it (mobile Europe) are shown in Figs 1 and 2. The numbers pertinent to the single poles correspond to the references in catalogues of IRVING et al. (1976a, b). Letters A, B and C indicate data added here but omitted in the mentioned catalogues (cf. KRS, 1978). At first sight, it is apparent that the poles of stable Europe are homogeneously grouped, whereas the mobile European pole positions display a remarkable scatter characterizing rotational tectonic deformations. This striking difference between stable and mobile Europe is very well illustrated by the orientation of paleomeridians. Figures 3 and 4 show the orientation of paleomeridians derived from the paleomagnetic data relating to the particular localities and areas. The paleolatitudes are also illustrated, calculated from the mean position of the corresponding pole, whether Triassic or Permian in age, of the North European Platform.

These results provide not only geological information, but also clearly demonstrate that the mean paleomagnetic field can be approximated by the geocentric magnetic dipole field. This is evidenced by rocks of Quaternary to

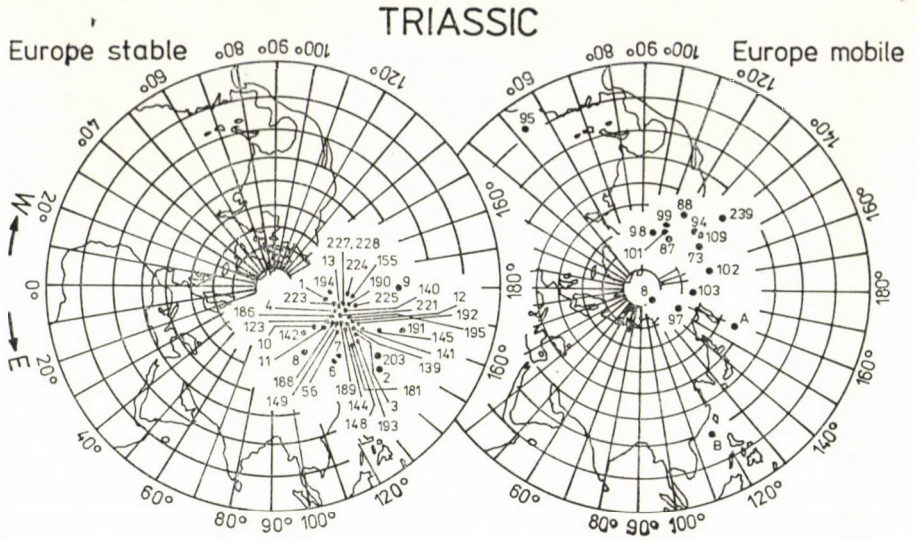


Fig. 1. Triassic paleomagnetic pole positions. References cf. IRVING et al. (1976a) and KRS (1978)

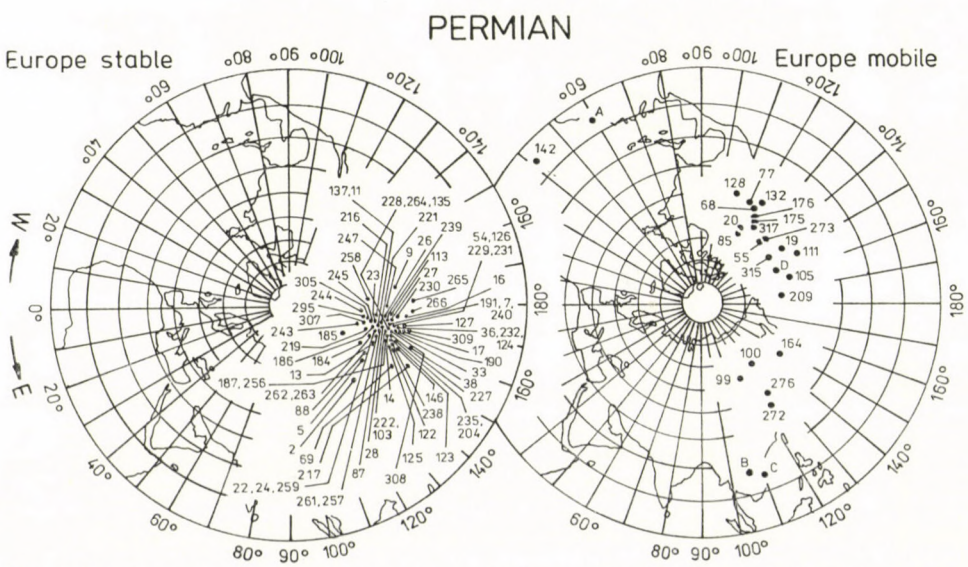


Fig. 2. Permian paleomagnetic pole positions. References cf. IRVING et al. (1976b) and KRS (1978)

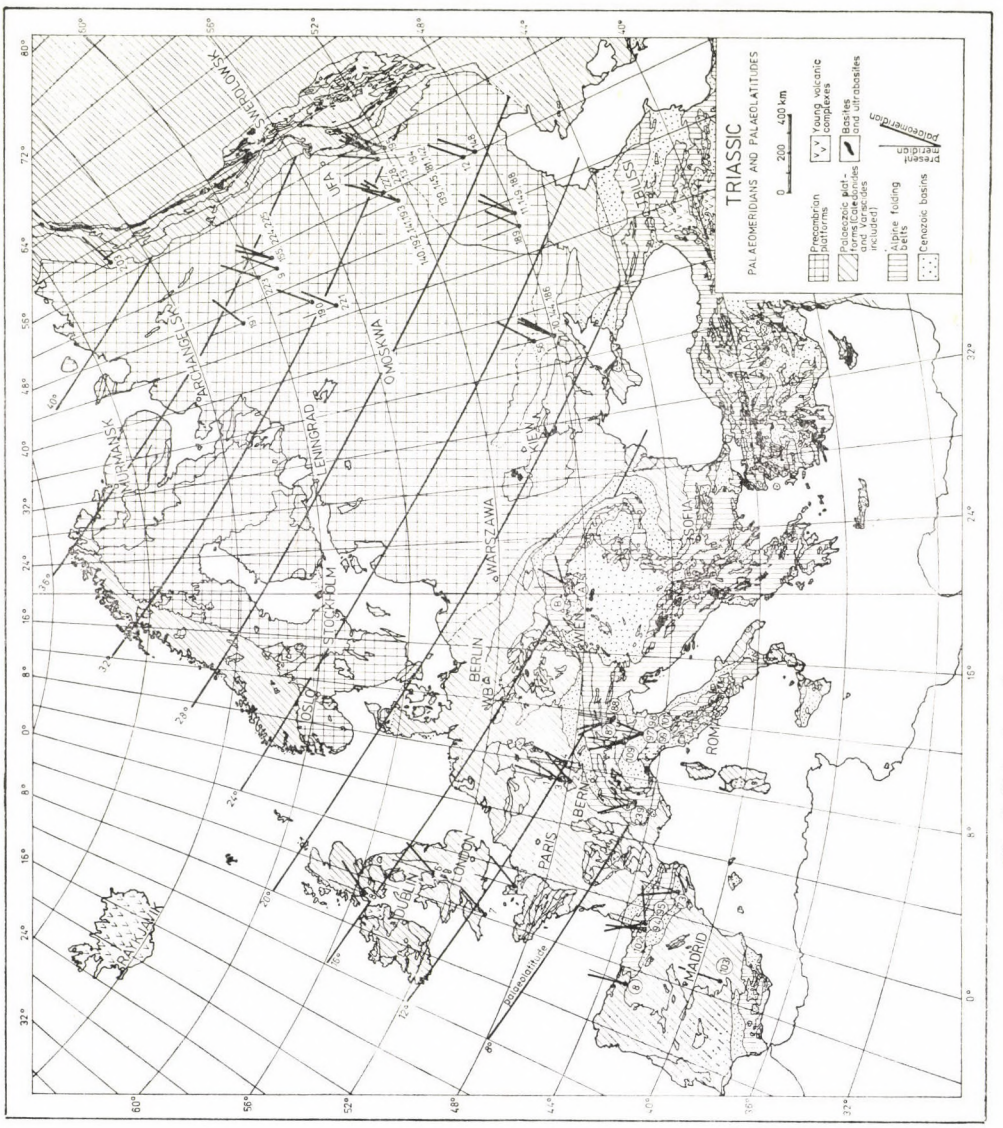


Fig. 3. Triassic paleomeridians and paleolatitudes

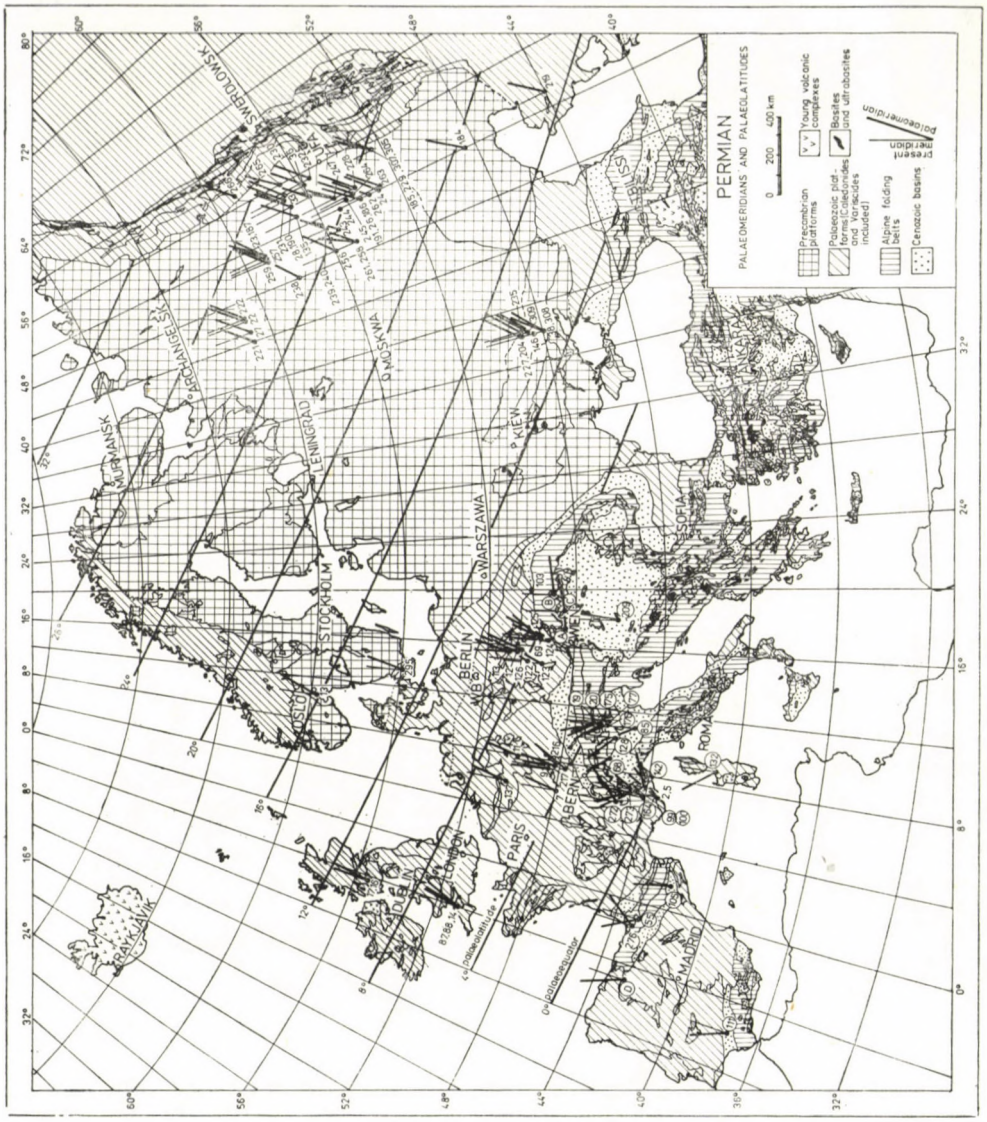


Fig. 4. Permian paleomeridians and paleolatitudes

Permian and in part Devonian age from the North European Platform and by rocks of the Siberian Platform for the whole Phanerozoic, except the insufficiently investigated Silurian, cf. KRS (in press). Thus it can be demonstrated that the scatters of paleomagnetic data inferred for rocks of the Alpine belt are caused by paleotectonic deformations, not by the non-dipolarity of the paleomagnetic field.

A study of paleotectonic deformations related to the history of the geomagnetic field can be best illustrated by introducing new parametres suitable for an analysis of paleomagnetic data, namely paleorotation ΔD, paleoinclination difference ΔI and paleodivergence Φ. Contrary to the terms of rotation, inclination difference and divergence, which were introduced into geophysics by BLACKETT et al. (1960), paleorotation ΔD, paleoinclination difference ΔI and paleodivergence Φ of paleomagnetic directions are related to the paleomagnetic direction derived for a particular locality from the mean paleomagnetic pole position within a stable block (plate, paleoplate, in this particular case within the North European Platform). The histograms of ΔD, ΔI and Φ for Permian and Triassic rocks from the North European Platform (stable Europe) and the Alpine belt (mobile Europe) are given in Figs 5 and 6. Grubbs' test was used to check anomalous values for the parametres derived from stable Europe; GAUSSian distribution of the calculated deviations was proved. The validity of this distribution was verified by the x^2-test at the probability level of $P = 0.01$, cf. KRS (in press). It is apparent, therefore, that the mean paleomagnetic field can be approximated by the magnetic dipole field within the limits of standard deviations attaining a few degrees. The striking deviations in the ΔD, ΔI and Φ values derived for rocks of the Alpine belt are due to systematic effects resulting from the paleotectonic deformations. Table 1 summarizes paleomagnetic data for the Permo-Triassic rocks of the Insubric—Carpathian block system to be used as a basic material for a study of its origin and development.

The Development of the Insubric—Carpathian Block System

From the point of view of Tertiary tectonics, the Alpine—Carpathian system has two longitudinal zones. The northern zone is characterized by a Tertiary nappe structure, whereas the southern zone displays a block structure. The nappe structure of the northern zone reflects Tertiary horizontal movements of the southern zone; the latter is collectively referred to as the Insubric-Carpathian block system (KRS—ROTH, in press).

The Insubric—Carpathian system is a Tertiary unit bounded on the north by the Canavese, Tonale, Gailtal, and peri-Pieninian faults, and in the East Carpathians by a boundary between the Dacides and Moldavides (SĂNDULESCU, 1975). On the south and southwest, it is bounded by the Dinarides and a hypo-

thetical fault zone beneath the Po Lowland in its western continuation (TRÜMPY, 1975a).

In the Alps, the system comprises the western (A) and eastern (B) blocks of the Southern Alps; in the Carpathians, it covers the Carpathian—Transylvanian block of Tertiary age (C) which is bounded in the south by the Moesian Plat-

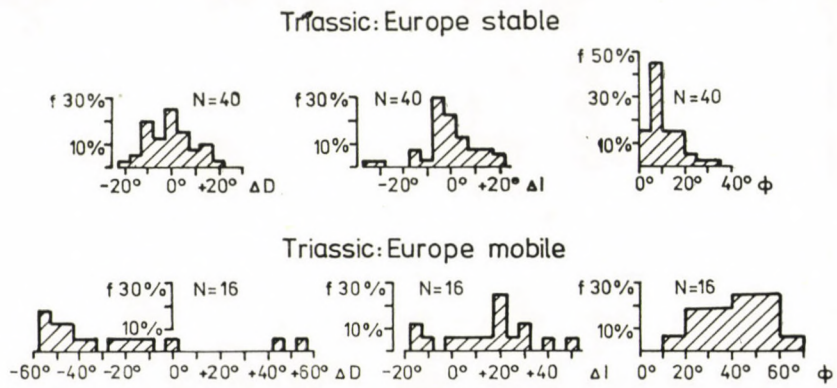


Fig. 5. Triassic: paleorotation ΔD, paleoinclination difference ΔI and paleodivergence Φ

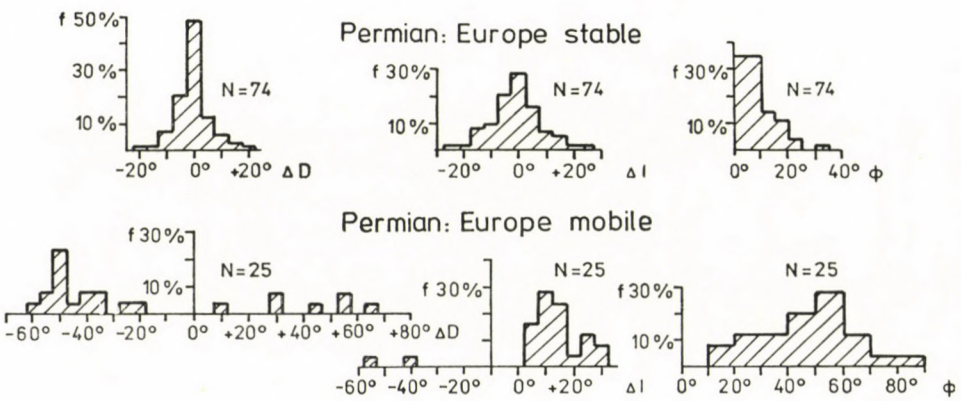


Fig. 6. Permian: paleorotation ΔD, paleoinclination difference ΔI and paleodivergence Φ

form and the Dinarides. Although all blocks of the Insubric—Carpathian system are equivalent to each other with regard to late Styrian movements (10—14 m.y.), the Carpathian—Transylvanian block, unlike South Alpine blocks, exhibits an Upper Cretaceous nappe structure in its northern part. The Mesozoic is essentially autochthonous in the blocks of the Southern Alps.

From the viewpoint of Tertiary tectonics, the Carpathian-Transylvanian block (Fig. 7) can be subdivided into the western and eastern parts. The for-

Table 1

Review of paleomagnetic parameters related to the Insubric–Carpathian block system

Reference	Age	Paleomagnetic direction			ΔD	ΔI	Φ	Category	Rocks, stability tests
		D	I	α_{95}					
99[1]	Triassic	330.0°	41.0°	5.0°	−53.2°	22.1°	50.2°	A**	Volcanics + sediments, A. C.
101[1]	Triassic	333.0°	41.9°	—	−50.2°	23.0°	48.2°	A*	Igneous rocks, Italy, A. C.
88[1]	Triassic	329.0°	25.0°	9.0°	−54.8°	3.3°	50.1°	A**	Porphyrites + dykes + tuffites, A. C.
A[2]	Triassic	29.2°	16.9°	5.3°	0.0°	−13.0°	13.0°	—	Werfenian shales, Košice region, A. C. + thermal tests
B[2]	Triassic	71.8°	19.8°	6.4°	42.9°	−10.1°	40.0°	—	Werfenian shales, Choč nappe, A. C. + thermal tests
68[3]	Permian	143.5°	−17.0°	8.7°	−52.9°	11.7°	53.0°	A**	Porphyries, Lugano, A. C.
128[3]	Permian	135.0°	−21.0°	2.0°	−62.3°	15.4°	62.2°	A**	Volcanics, Collio Auccio, A. C.
85[3]	Permian	332.0°	32.0°	6.0°	−45.8°	27.6°	50.9°	A*	Igneous rocks, Camparno Staro, A. C.
176[3]	Permian	148.0°	−20.0°	4.0°	−49.9°	13.1°	50.1°	A	Igneous rocks, S. Alps, A. C. + thermal fields
77[3]	Permian	143.0°	−15.0°	11.0°	−55.1°	8.1°	54.6°	A**	Quartz porphyry, Lagorai, A. C.
20[3]	Permian	150.0°	−31.0°	4.0°	−48.1°	23.6°	50.7°	A	Quartz porphyry, Bolzano, A. C.
175[3]	Permian	150.0°	−19.5°	4.9°	−48.2°	12.4°	48.3°	A**	Quartz porphyry, Bolzano
19[3]	Permian	164.0°	−11.0°	8.0°	−34.0°	3.4°	33.7°	A	Quartz porphyry, Bolzano, A. C.
A[4]	Permian	264.8°	27.2°	15.9°	62.6°	−40.9°	81.7°	—	Melaphyres, Little Carpathians, Choč unit, A. C. + thermal tests
B[4]	Permian	261.2°	−28.7°	11.8°	57.4°	12.9°	69.9°	—	Melaphyres, Low Tatras, Choč unit, A. C. + thermal tests
C[4]	Permian	258.3°	−20.7°	10.3°	55.3°	6.4°	74.6°	—	Melaphyres, Tríbeč, Choč unit, A. C. + thermal tests
209[3]	Permian	186.9°	−18.4°	5.8°	−15.7°	8.6°	17.5°	A**	Red beds, Mecsek, thermal tests

[1] *cf.* IRVING et al. (1976a); [2] *cf.* KRS (1966); [3] *cf.* IRVING et al. (1976b); [4] *cf.* MUŠKA (1977)

mer includes the Carpathian—Pannonian sub-block C_1, the common feature of which is the post-orogenic Intracarpathian Paleogene and the upper Cretaceous. The latter, i.e. the eastern part covers the Apuseni—Transylvanian sub-block C_2 including the East Carpathian crystalline massif. The two sub-blocks are divided by a reversed fault. This fault issues from the inner margin of the



Fig. 7. Carpathian—Transylvanian Tertiary block and its subdivision. C_1 — Carpathian—Pannonian sub-block; C_{1a} — Tatride partial sub-block; C_{1b} — Gemeride—Pannonian partial sub-block; C_2 — Apuseni—Transylvanian sub-block. A, B, 209 — references to areas and sites of paleomagnetic investigation. Arrows and cones denote paleomagnetic directions and cones of confidence at 95% probability level, respectively (cf. Table 1). α — Schrattenberg—Bulhary fault; β — Jakubov fault; γ — Smolenice—Boleráz fault; δ — Rába—Lubeník—Margecany fault zone; η — Baia Mare—Debrecen—Szolnok—Szeged fault zone. 1. Block and sub-block boundaries; 2. assumed block and sub-block boundaries; 3. surface block and sub-block boundaries (if different from their deep-seated position); 4. outlines of the Outer Carpathian nappes

klippen belt bending near Poiana Botizei to the southwest (SĂNDULESCU, 1975) where it seems to join the line Baia Mare—Debrecen—Szolnok—Szeged (η).

Published paleomagnetic data indicate that the South Alpine blocks are characterized by a counter-clockwise rotation of the Permo-Triassic paleomeridians by about 50°. This rotation is believed to be mainly Pyreneean in age (38—40 m.y.; TRÜMPY, 1975b). Few paleomagnetic data are yet available on the Carpathian—Transylvanian block of Tertiary age, except for its northwestern part. These data suggest that counter-clockwise rotation can be recognized, apart from South Alpine blocks, only in the Gemeride—Pannonian partial sub-block (C_{1b}) forming a southeastern part of the Carpathian—Pannonian sub-block that is separated from the Tatride partial sub-block of Tertiary age (C_{1a}) by the Rába—Lubeník—Margecany fault zone δ shown in Fig. 7.

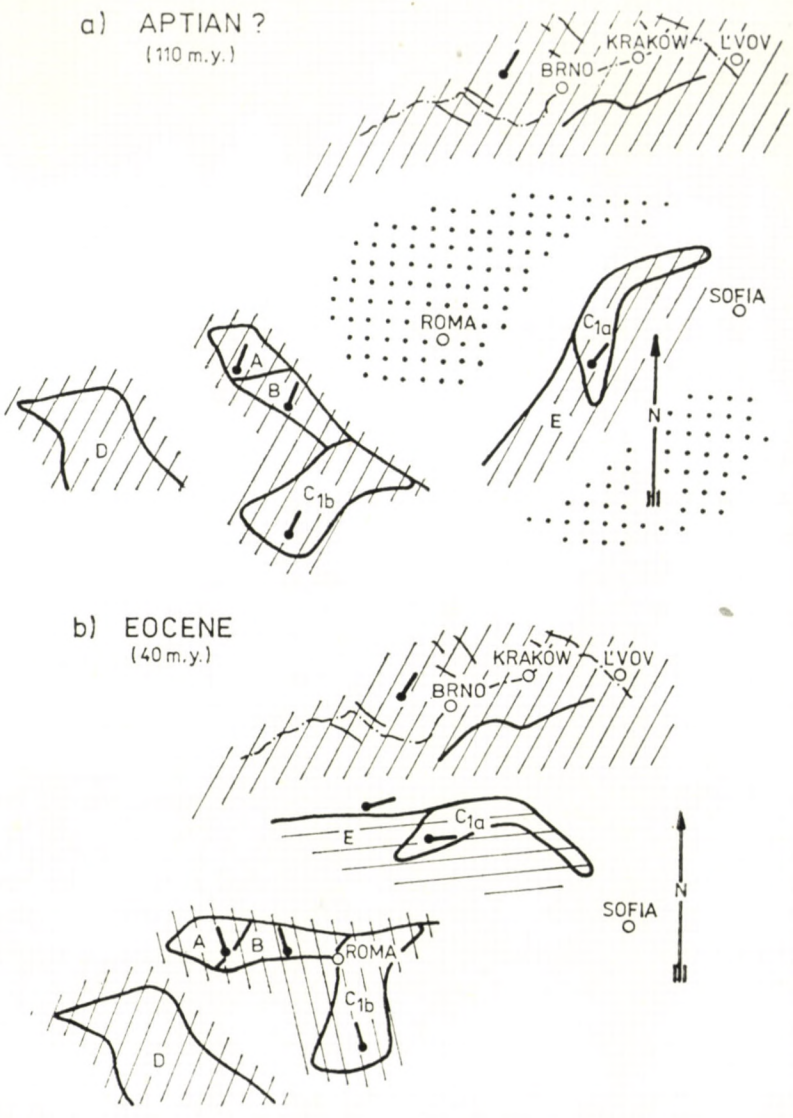
Contrary to this, the Tatride partial sub-block of Tertiary age and the Northern Calcareous Alps (MAURITSCH—FRISCH, in press) display a clockwise rotation by about 45°.

The clockwise and counter-clockwise rotations of Permo-Triassic paleomeridians, which occur simultaneously in the Alpine–Carpathian system, are demonstrated in Fig. 8 (a—d). It is explained in term of rotational and simultaneous north-vergent translational movements of the Insubric–Carpathian (A, B, C_{1b}) and Briançonnais–Tatride (E, C_{1a}) zones of essentially continental lithosphere toward the North European Platform. These two zones (placed between the Eurasian and African continents) were originally oriented diagonally, but in mutually opposite directions, to the east-westward trending Tethyde belt (Fig. 8a). In accordance with our hypothesis, which is based on CHANNELL—HORVÁTH's (1976) regarding the Adriatic microplate (D) as a promontory of Africa, the Pieninian sedimentation basin containing a suboceanic crust of Upper Jurassic to Lower Cretaceous age is an eastern continuation of the North Penninic (Valais) eugeosyncline and the Tatride partial sub-block (C_{1a}) extends as an eastern continuation of the Alpine Briançonnais zone (E).

The shift of the initially compact, NW—SE trending Insubric–Carpathian block and the Briançonnais–Tatride block trending in NE—SW direction toward the southern margin of the Platform was caused by a translation of the Adriatic microplate north-eastward in the Tertiary.

Once the contact with the North European Platform began to set limits to the northward movement of both isolated blocks, the Insubric–Carpathian block system (with dominant counter-clockwise rotation) and the Briançonnais–Tatride block zone (with dominant clockwise rotation) adhered to the platform's irregular margin, started to break into blocks under the pressure of the Adriatic microplate. The blocks ultimately stopped after an adaptive revolution at the southern margin of the stable Europe. Minor shifts and insignificant clockwise and counter-clockwise adaptive rotations were secondary phenomena, yet they played a considerable role in the final placing of the blocks.

To these adaptive movements probably also belong: the Late Pyreneean (33—38. m.y.) clockwise rotation of the eastern block of the Southern Alps (by about 5°), the Late Styrian (10—14 m.y.) counter-clockwise rotation of the Carpathian–Transylvanian block (by about 14°), and the Early Styrian (17 m.y.) sinistral shift of the last block (relative to South Alpine blocks) northward by about 300 km, most evident along the Schrattenberg—Bulhary (α) and Smolenice—Boleráz (γ) faults (KRS—ROTH, in press). This hypothesis is supported by paleomagnetic evidence of an counter-clockwise Early Styrian rotation of the Lower and Upper Cretaceous rocks of the Silesian unit in northern Moravia (KRS et al., in press).


a) APTIAN ?
(110 m.y.)
KRAKÓW
L'VOV
BRNO
ROMA
SOFIA
A
B
C1a
C1b
D
E
N
b) EOCENE
(40 m.y.)
KRAKÓW
L'VOV
BRNO
E
C1a
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A
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D

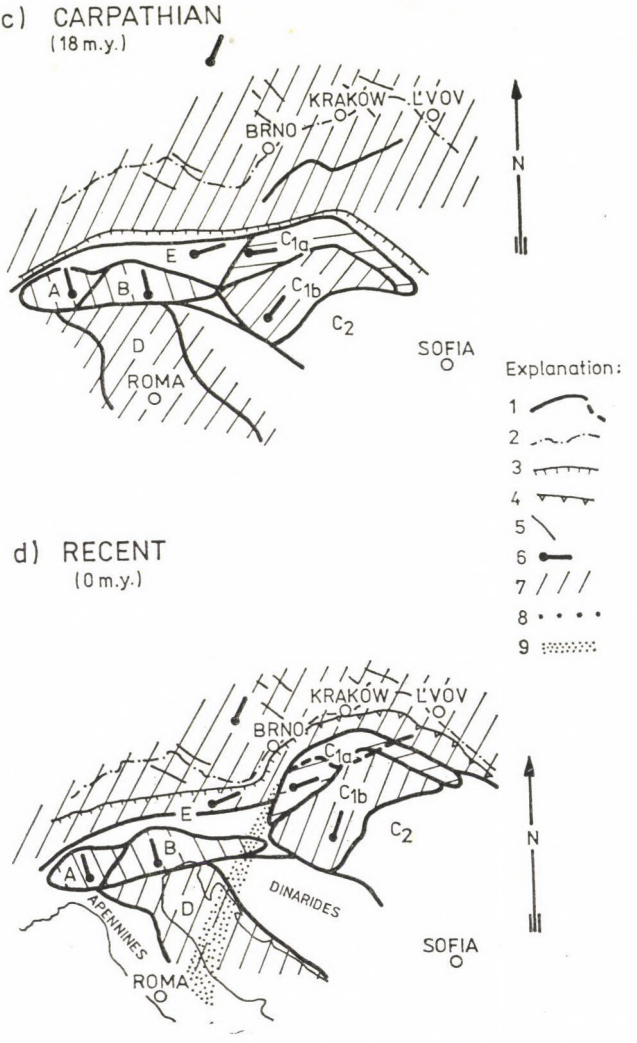

Fig. 8. Scheme of paleotectonic development of the Alpine territory of Central Europe during the Tertiary. 1. Present-day surface block, sub-block and platform boundaries; 2. present-day margin of the Neogene foredeep; 3. outer boundary of the Alpine nappe structure; 4. Late Styrian compression outer margin of the Carpathians; 5. minor fractures and present-day coast lines; 6. Permo–Triassic paleomeridian orientations; 7. continental-type lithosphere; 8. (sub)oceanic-type lithosphere; 9. zone of Late Styrian crust extension. A, B — South Alpine blocks; C — Carpathian—Transylvanian block; D — Adriatic microplate; E — Briançonnais (Middle Penninic zone)

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ГИПОТЕЗА О РАЗВИТИИ ТРЕТИЧНЕЙ ИНСУБРИЧЕСКО-КАРПАТСКОЙ СИСТЕМЫ БЛОКОВ

М. КРС, З. РОТ

Резюме

Материалы для изучения тектонических деформаций Альпийского пояса получились путём сопоставления всех до сих пор опубликованных палеомагнитных данных о Евразии и Африке, и также анализа новых геофизических параметров как палеоротация, разница палеонаклонений, и палеодивергенция палеомагнитных направлений. Интерпретация палеомагнитных данных поддерживает представления тектоники плит, что касается третичнего развития Альп и Карпат. Рассмотривается предлагаемая модель развития Инсубрическо-карпатской системы блоков в связи с движением Адриатической микроплиты.

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GEODYNAMICS OF THE PANNONIAN BASIN: GEOTHERMAL AND ELECTROMAGNETIC ASPECTS

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The Pannonian Basin is characterized by geothermal highs; the temperatures at 1 km depth vary from 50 to 70 °C, the surface heat flow values are in the range of 80—130 mW/m². The heat flow and temperature values computed to the depth of the MOHO discontinuity are also high; 50—70 mW/m² and 700—1000 °C, respectively.

Magnetotelluric deep soundings show that the high conductivity zone of the upper mantle is in an elevated position (45—70 km) below the Pannonian Basin. Deep seismic soundings demonstrate that the low-velocity zone is also in the same elevated position, i.e. the lithosphere is thinner than the normal value for stable continental region.

The correlation of geothermal and magnetotelluric data suggests that the thin lithosphere (i.e. the updoming of the astenosphere) is the result of deep geothermal processes. The age *versus* thickness curve of the oceanic lithosphere, if applied for the Pannonian Basin, suggests that this geothermal process occurred about 30 m.y. ago. Eocene—Early Miocene subduction along the Carpathian arc probably induced convection in the asthenosphere, which led to the formation of an active mantle-diapir. The Neogene calc-alkaline volcanism and the subsequent Late Miocene—Quaternary subsidence of the region is also related to the mantle-diapir.

The Carpathian conductivity anomaly is interpreted in terms of underthrusting of the Outer Carpathian flysch complex in the course of subduction.

Introduction

In previous publications (SZÁDECZKY-KARDOSS, 1967; STEGENA et al., 1975; HORVÁTH—STEGENA, 1977) a geodynamic model has been proposed for the tectonic evolution of the Pannonian Basin (Fig. 1). According to the model, the Pannonian Basin is a back-arc basin, lying at the inner (concave) side of the Carpathian arc. The subsidence of the region (i.e. the formation of the basin) has occured since the Late Miocene. The point of the model is that the basin evolution is related to the Eocene—Early Miocene southward subduction of the Carpathian flysch basin's lithosphere beneath the Inner Carpathian/Pannonian region.

In the asthenosphere of the region, secondary convection was induced by the downgoing slab. Because of viscous processes the convection gave rise

to the formation of a thermal mantle-diapir. The upwelling mantle-diapir progressively melted the lower lithosphere (including the bottom of the crust) and the bulk of the molten material was transported away laterally. Some amount could break through the crust and produced intensive calc-alkaline

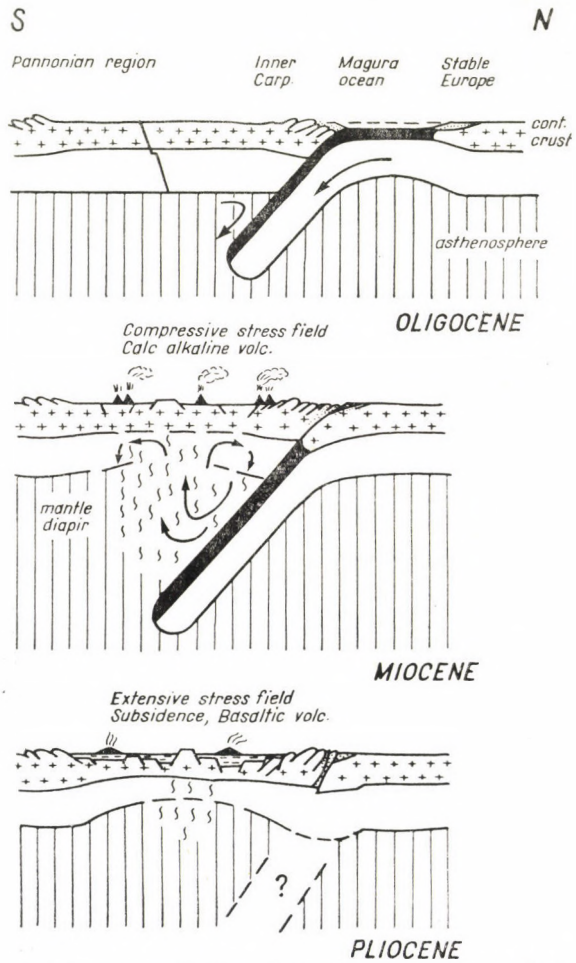

Fig. 1. Scheme of the Neogene evolution of the Pannonian Basin (not scaled vertically). After STEGENA et al. (1975), modified

volcanism. Because of subsequent conductive cooling and isostatic forces, blocks of the thin continental crust started to subside along normal faults after the termination of subduction.

The goal of the present paper is to review geothermal and electromagnetic aspects of the model.

Geothermics of the Pannonian Basin

The area of the Pannonian Basin is characterized by pronounced geothermal highs. Several thousand temperature measurements carried out in boreholes demonstrated that the temperature at 1 km depth is in the 50—70 °C interval (STEGENA, 1976). Heat flow determinations in Hungary and in the Czechoslovakian, Soviet and Roumanian parts of the Pannonian Basin have given values from 80 to 130 mW/m². Figure 2 shows that the Pannonian Basin is indeed

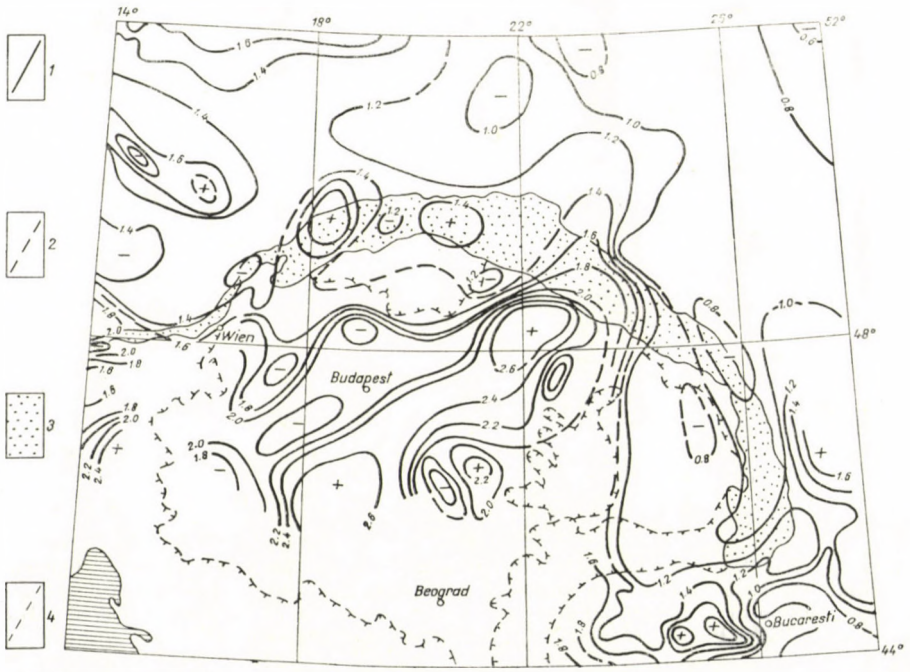

Fig. 2. Contour map of heat flow in the Carpatho–Pannonian region and in its surroundings, in HFU (after ČERMÁK and HURTIG, 1977). 1. Isoline of heat flow; 2. uncertain isoline; 3. flysch complex of the Carpathians and Eastern Alps; 4. approximate boundary of Eastern Alps, Inner Carpathians, Apuseni Mts and Dinarides

a "red spot" on the heat flow map (ČERMÁK and HURTIG, 1977). The question can arise whether the Pannonian Basin's geothermal high is the result of deep-seated processes or it is caused by near-surface (e.g. hydrothermal) disturbances. Recently, all the Hungarian heat flow determination have critically been reviewed and the near-surface disturbances evaluated (HORVÁTH et al., 1979). It has been concluded that although the thermal field is locally disturbed, the Pannonian thermal high is valid and can be continued downward.

Using the method suggested by ROY et al. (1972), high "reduced" heat flow (50—60 mW/m²) was obtained for the Pannonian Basin (STEGENA, 1974).

Two dimensional temperature model calculations along seismic crustal sections (BUNTEBARTH, 1976; BODRI, 1976) gave concordant results; the Moho heat flow and temperature are in the range 60—70 mW/m² and 700—1000 °C, respectively, for the Pannonian Basin (Fig. 3).

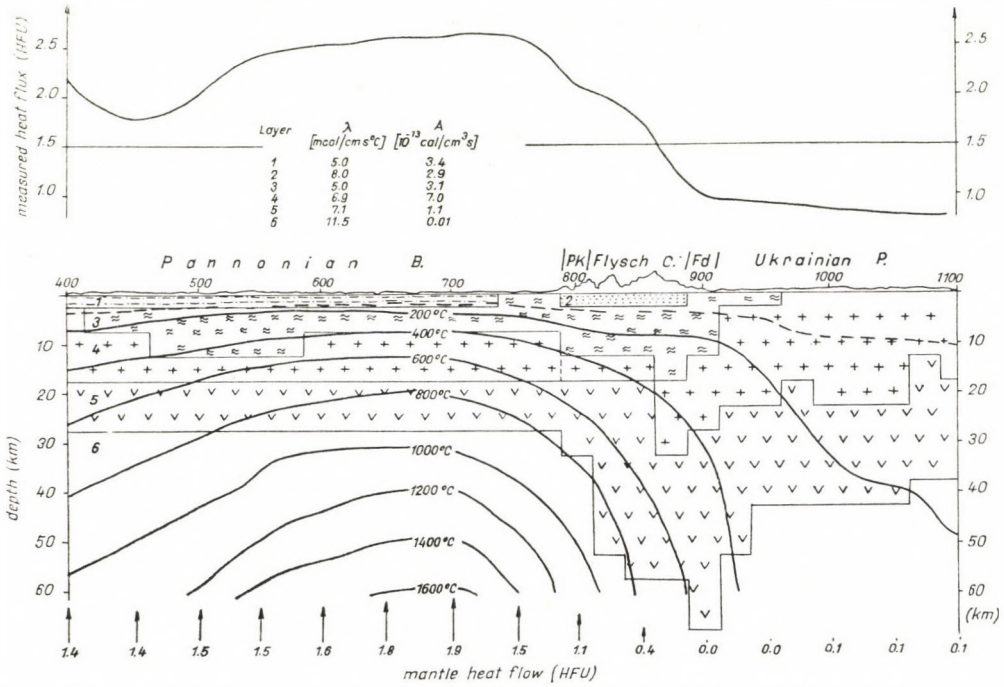

Fig. 3. **Measured heat flow and calculated geoisotherms, and heat flow at 60 km depth along the seismic profile IP-III. In the course of calculation the crustal model given in the Figure was used, supposing exponential decrease of heat production (A) in each layer (after BODRI, 1976). PK = Piening Klippen belt; Fd = Mollasse foredeep; HFV = 1 μcal/cm²s**

Magnetotelluric and Geomagnetic Deep Soundings

Numerous magnetotelluric deep soundings have been carried out in the Pannonian Basin and its surroundings (ÁDÁM, 1976, 1978b). Figure 4 summarizes the depths of highly conductive layers and/or bodies and their specific resistivities determined by magnetotelluric and geomagnetic soundings. The near-surface highly conductive zone (2—3) was determined in the Transdanubian Central Range (ÁDÁM, 1974). It probably consists of pre-Mesozoic graphite or graphitic schists. The relatively high amount of electrolyte in the fracture zones can also be one of the causes of the great anisotropy in the region. This conductivity anomaly is not related to the basinal evolution of Pannonian region but appears to be very important from the point of view of the autochthony or allochthony of the Transdanubian Mesozoic.

Geomagnetic deep soundings (Wiese-arrows Fig. 5) supplemented with magnetotelluric soundings show that a highly conductive body (4—5, Fig. 4), is associated with the inner boundary of the Carpathian Flysch belt all along its strike from the Western Carpathians to the Southern Carpathians (ÁDÁM, 1974; ROKITYANSKY et al., 1975). On the basis of model calculations JANKOWSKY et al. (1977) suggested that the slab-like body is about 16—18 km wide and

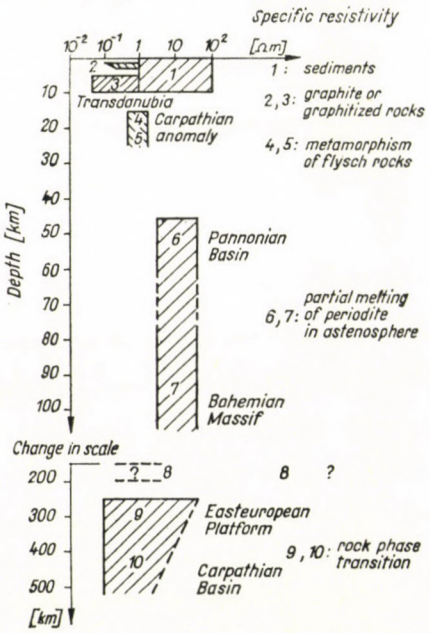

Fig. 4. **Highly conducting zones of the crust and upper mantle in the Carpathian system and in the East European platform**

is situated in 22—45 km depth interval. ROKITYANSKY et al. (1975, 1976), interpreted the Carpathian conductivity anomaly in terms of regional metamorphism of the flysch complex below 10 km depth. They suggested that at the lower part of the flysch wedge water release and interstitial melting took place due to the effect of pressure and temperature. In the presence of water vapour and interstitial melt the electrical conductivity of flysch rocks can reach 1 $\Omega^{-1}\,m^{-1}$ of order of magnitude. This interpretation looks reasonable but offers no explanation for the presence of flysch rocks at depth of 20 km and more. We think that the only plausible explanation is the transportation of these rocks to that depth by the southward subduction of the flysch-basin crust. This interpretation is in agreement with the deep structure and tectonic evolution of the Carpathians (NEY, 1975) and with the recent refinement of electric model calculations (Fig. 6, PETR et al., 1975).

Figure 4 shows that the top of the first highly conducting layer of the upper mantle is in an elevated position (45—70 km) in the Pannonian Basin relative to its surroundings. We try to demonstrate in the next section that this elevated position is the result of thermal activization. It is an interesting feature of the magnetotelluric sounding curves of the Pannonian Basin that they

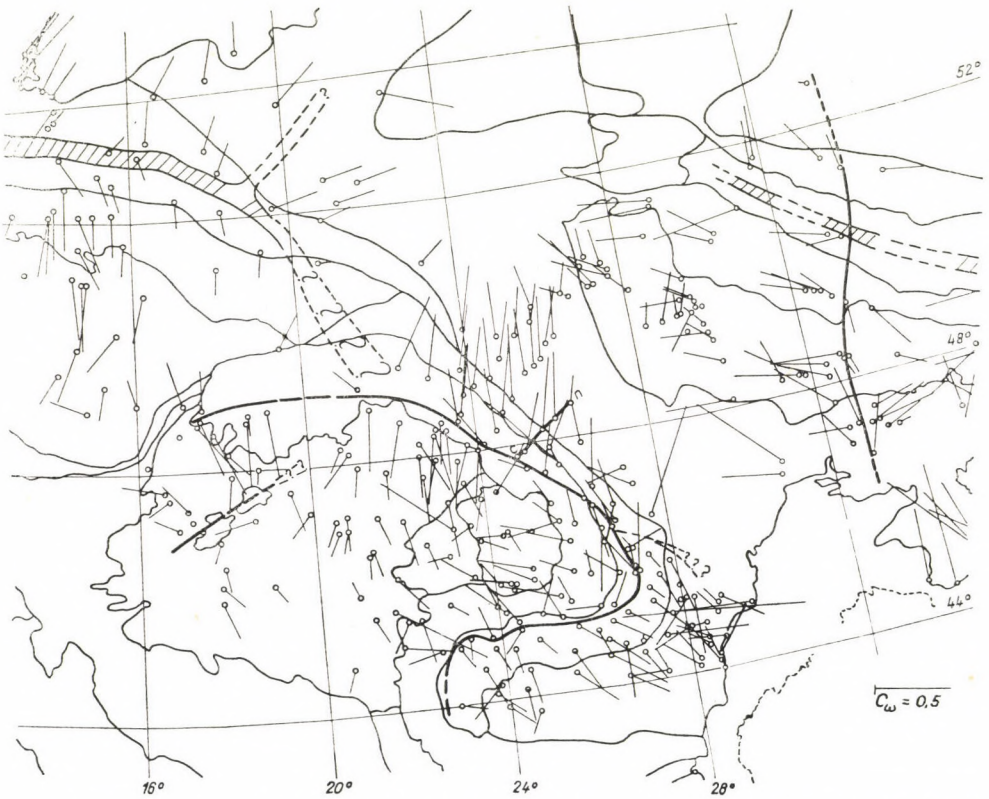

Fig. 5. Wiese vectors in the Carpathian area and in its surroundings (after ROKITYANSKY et al., 1976)

exhibit a pronounced regional anisotropy. The ϱ_x curves usually show higher apparent specific resistivities than the ϱ_y curves in the 10^2—10^4 s period interval. It is probably related to the "linear" structure of the Pannonian basement (STEGENA, ÁDÁM, HORVÁTH, 1971). Namely, the basement is composed of several NNE—SSW striking belts, the arrangement of which reflects the paleogeographic conditions and the prevailing compression during the Alpine orogeny.

The second highly conductive layer of the upper mantle (9—10, Fig. 4) can probably be explained in terms of olivine–spinell phase transition (ÁDÁM 1978a).

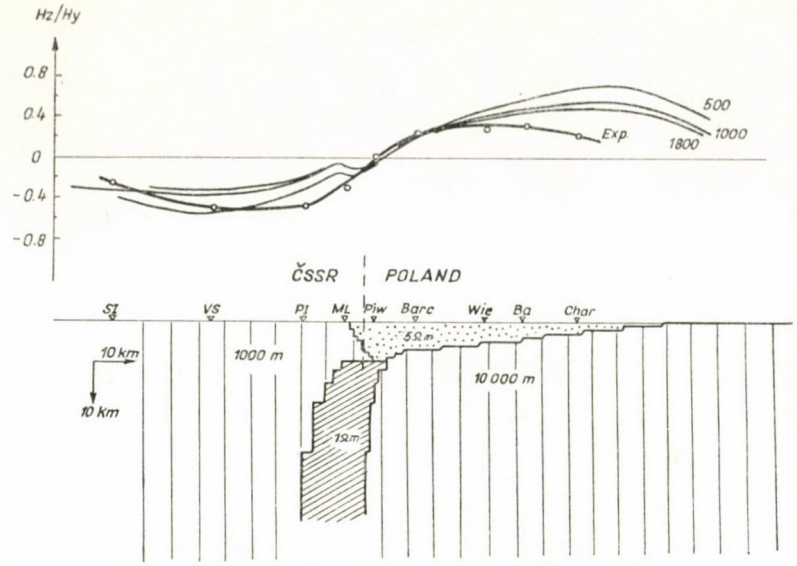

Fig. 6. Calculated conductivity model for section perpendicular to the Western Carpathian arc (after PETR et al., 1975). Upper part: Theoretical H_z/H_y anomalies for periods of 500, 1000, 1800 s and the measured anomaly (EXP)

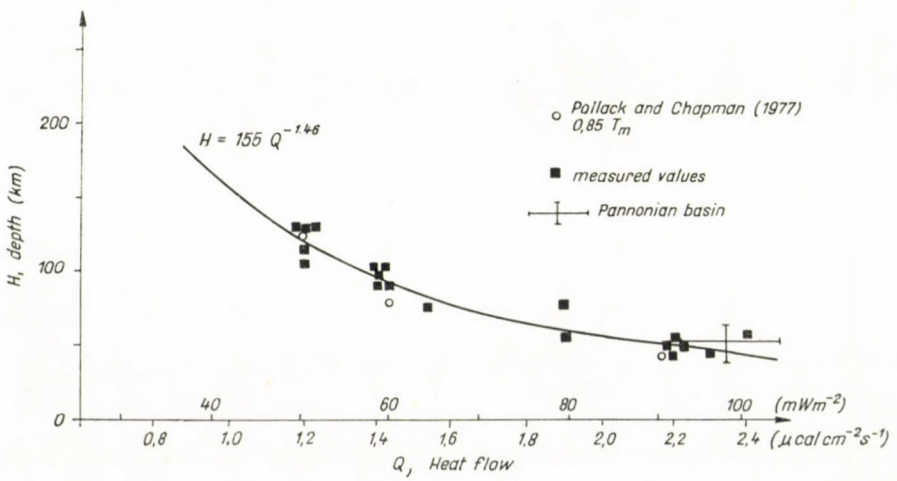

Fig. 7. Plot of the regional heat flow versus the depth of the upper mantle conducting layer, for Eurasian areas (ÁDÁM, 1976)

Geothermal and Magnetotelluric Connections

There appears to be a good correlation between the position of the first high-conductivity layer of the upper mantle and the average heat flow. Based first of all on the data of KAPG monograph "Geoelectric and geothermal studies", ÁDÁM (1978a) quantitatively described this relationship. In Fig. 7 the average surface heat flows and the depths of the highly conducting layers are plotted for Eurasia. The measured data can be approximated with the following function:

$$H = 155\, Q^{-1{,}46}$$

where the depth H is in km and the average heat flow Q is in HFU. The Figure also shows for different heat flow values those depths where the temperature of the upper mantle equals 0.85 T_m, where T_m is the melting point of peridotite. These depths can be obtained from the intersections of the 0.85 T_m curve with the temperature–depth curves, computed for different surface heat flow regions (POLLACK and CHAPMAN, 1977; Fig. 8). The good fit of the measured depth of highly conducting layer with the computed depth of the partial melting temperature (0.85 T_m) of peridotite (Fig. 7) gives evidence that the position of the highly conducting layer is controlled by the temperature of the upper mantle. As the data characteristic of the Pannonian Basin fits the diagram, it can be concluded that the elevated position of the highly conducting layer below the Pannonian Basin is the result of increased upper mantle temperature.

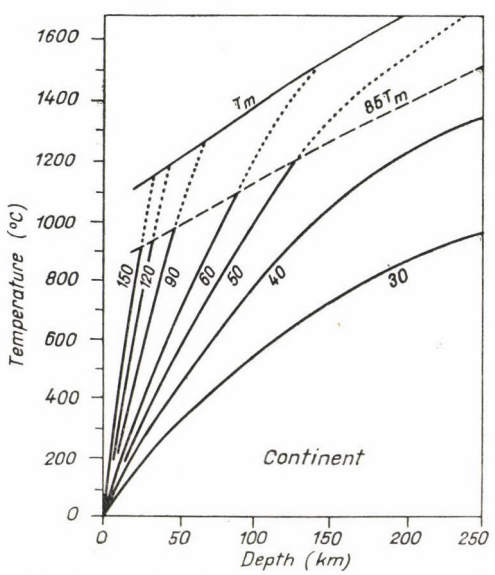

Fig. 8. The temperature *vs* depth curves of POLLACK and CHAPMAN (1977), for different heat flow values (in mWm^{-2}), and the T_m and 0.85 T_m melting curves for periodite

Conclusions

It is suggested that the heating up of the upper mantle is a consequence of the Late Eocene Miocene subduction along the outer margin of the Inner Carpathians. Thermal mantle-diapir was generated below the Pannonian region (Fig. 1). The upwelling mantle-diapir melted the lower part of the lithosphere (including the lowermost crust) during the Miocene times. The present

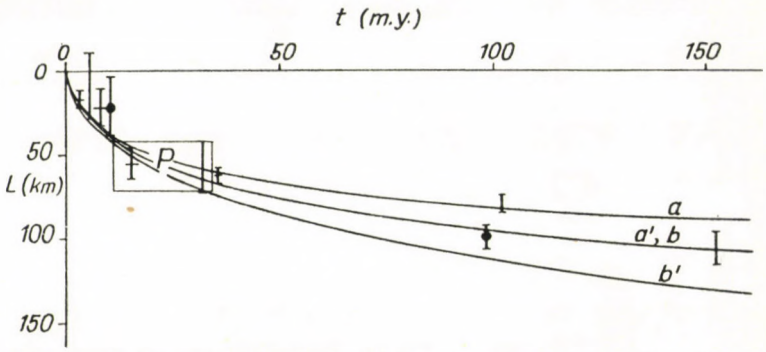

Fig. 9. Predicted and observed lithospheric thickness in the oceans as a function of age (after CROUGH, 1977). Solid lines: theoretical predictions for the thermal models a—a′: Q = 0.80 HFU; b—b′: Q = 0.56 HFU. a′ and b′: 0.25 HFU is caused by a temperature gradient in the astenosphere. Vertical bars: Pacific. Bars with seash: East Pacific. Bars with dot: Atlantic. P: the Pannonian Basin

thin lithosphere of the Pannonian Basin has been formed since then by relaxation of the thermal anomaly. This process was somewhat similar to the formation of lithosphere along mid-oceanic ridges. It appears to be not unreasonable, therefore, to apply the age vs. thickness relationship of the oceanic lithosphere for the Pannonian Basin. Figure 9 shows that 15—30 m. year is the necessary cooling time for lithosphere to attain 45—70 km thickness. This time fairly well agrees with the supposed Middle Miocene termination of subduction (i.e. the beginning of conductive cooling of the mantle-diapir).

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ГЕОДИНАМИКА ПАННОНСКОГО БАССЕЙНА: ГЕОТЕРМИЧЕСКИЕ И ЭЛЕКТРОМАГНИТНЫЕ АСПЕКТЫ

А. АДАМ, Ф. ХОРВАТ, Л. ШТЕГЕНА

Резюме

Паннонский бассейн характеризуется высокими геотермическими значениями: температуры в глубине 1 км находятся в интервале 50—70 C°, и термическое течение на поверхности — 80—130 мW/m². Соответствующие величины расчитанные для глубины поверхности Мохо тоже высокме: 700—1000 C° и 50—70 mW/m².

По глубинному магнитотеллурическому зондированию обнаружено что зона высокой проводимости верхней мантии находится в воздвигнутом положении под Паннонским бассейном (45—70 км). Глубинные сейсмические зондирования указывают на то, что зона низкой скорости расположена тоже высоко, т. е. литосфера является менее толстой чем нормальное значение в стабильных континентальных облатях.

Сопоставление геотермических и магнитотеллурических данных позволяет нам делать вывод что воздвигнутое положение астеносферы получилось вследствии глубинных геотермических процессов.

Кривая возраст/мощность оцеанской литосферы, составленная для Паннонского бассейна, указывает на то, что этот геотермический процесс происходил 30 миллионов лет тому назад. Субдукция в позднем олигоцене и в раннем миоцене индуцировала конвекцию в астеносфере, что проводило к возникновению активного диапира мантии. С этим диапиром мантии связаны также известко-щелощный вулканизм в неогене и субсеквентное погружение этой области в плиоцене и в плейстоцене.

Карпатская аномалия проводимости изталкивается как следствие поддвига флишовой толщи Внешних Карпат в течении субдукции.

Acta Geologica Academiae Scientiarum Hungaricae, Tomus 21 (4), pp. 261—263 (1977)

RELATION BETWEEN HEAT FLOW AND CRUSTAL THICKNESS

By

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Areas of high heat flow are usually characterized by weak or thinned crust. The relation between the heat flow and crustal thickness is negative, i.e. the thinner the crust, the higher heat flow can be expected. Radioactivity does not significantly influence this relationship.

Correlation studies between crustal structure and observed surface geothermal activity can be a worthwhile approach to explain the role of deep-seated processes in the geological history. Since heat-producing radioactive elements are mostly concentrated in the upper parts of the continental crust, higher heat flow values should be expected in the regions of thick crust. However, when analyzing the regional heat flow pattern in Europe, one may distinguish many areas of high heat flow, which are, on the contrary, characterized by considerably weakened crust. If this feature is a general phenomenon, the older ideas of relative constancy of the upper mantle heat flow contribution must be considerably re-evaluated.

The recently obtained heat flow map of Europe (ČERMÁK—HURTIG, 1977) was for the above correlation completed with the provisional map of the crustal thickness in Europe (ČERMÁK, 1977) based on numerous published data on the Mohorovičić depth. Figure 1 shows the preliminary results of such correlation. Solid dots represent the mean heat flow values versus the mean crustal thicknesses for specific large tectonic units in Europe. Both above means were calculated for respective data taken from corresponding maps in all $1^\circ \times 1^\circ$ area elements covering the particular tectonic province. The length of the arms of each individual cross equals the standard deviation of both means. The hatched areas in Fig. 1 represent the characteristic heat flow versus crustal thickness of some smaller, but typical tectonic or geological structures for which the number of particular $1^\circ \times 1^\circ$ grid elements would not be sufficient, which may represent the possible extreme cases. Four curves represent the direct observed relation between heat flow and Mohorovičić depth found in the Western Carpathians (curve AA) and in the Bohemian Massif (curve BB) in Czechoslovakia (ČERMÁK, 1976); in Roumania (DEMETRESCU—VELICIU, 1978) and in Ukraine (KUTAS, 1978).

The results of heat flow versus crustal thickness correlation thus obviously suggest a prevailing negative character, higher heat flow being observed over locally thinner crust and low heat flow being typical for areas of thick to very thick crust. It is not likely that the increased surface geothermal activity over

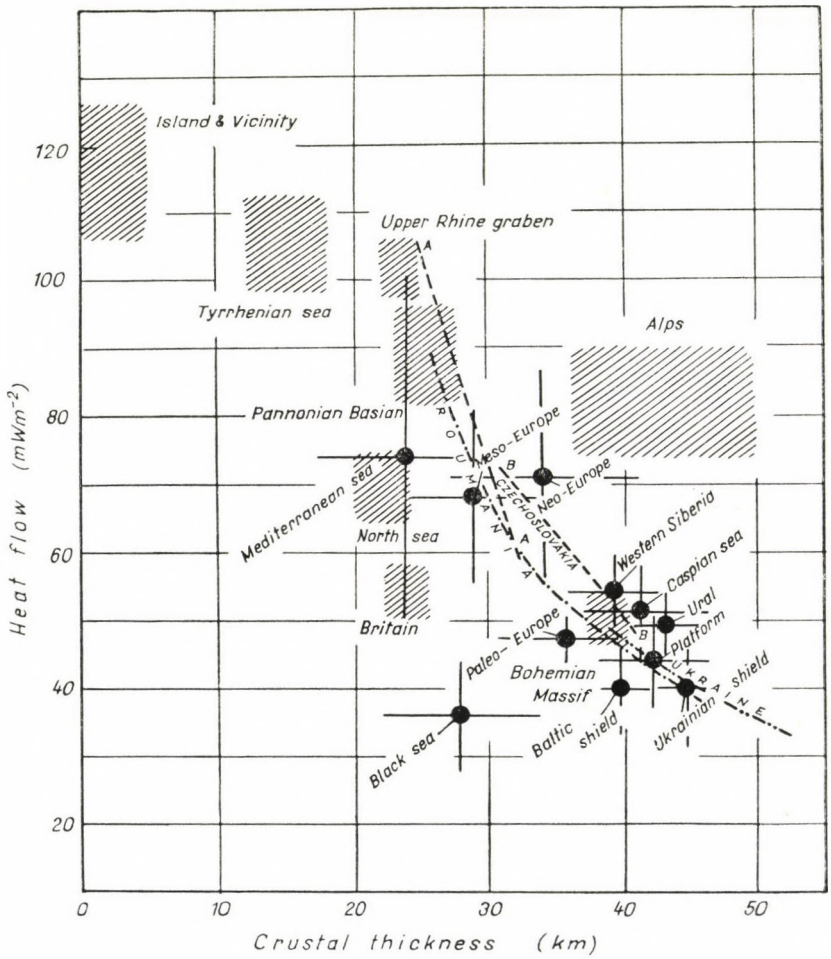

Fig. 1. Correlation between crustal thickness and surface heat flow, for Eurasian areas

local elevations of the Moho discontinuity can be explained by crustal sources, especially when in areas like the Pannonian Basin or in the Upper Rhine graben the radioactivity of the crustal rocks is rather low or normal. The origin of enhanced heat flow must be therefore connected with the anomalous conditions in the upper mantle and increased energy inflow from considerable depths.

Relatively low temperatures (350—500 °C) at the Mohorovičić discontinuity at the depth of 45—50 km are to be expected beneath the shields and/or the East European platform. Higher Moho temperatures of 500—600 °C were calculated for the Paleozoic folded units such as the Bohemian Massif or the Ural Mts at the depth of 35—40 km. Local temperature extremes of 600—700 °C seem to exist in zones of a weakened earth's crust and/or in areas of deep faults, which manifest themselves by relatively high heat flow of up to 70—80 mWm^{-2}. Very high crustal temperatures ($T_M \sim$ 800—1000 °C) are probable in hyperthermal regions, such as the Pannonian Basin, Tyrrhenian Sea, Upper Rhine graben, characterized by very thin crust and very high heat flow. Calculated crustal temperatures are very closely connected with the extrapolated Moho heat flow. While beneath the shields, even when the crustal rocks are considerably depleated in radioactivity, the heat flow from the upper mantle cannot exceed 15—20 mWm^{-2}; areas such as the Pannonian Basin, even when the crustal rocks are very rich in radioactivity (which is, however, not probable), must have high Moho heat flow of about 50—55 mWm^{-2}. There must be considerable regional variations in the energy transport from the upper mantle to the crust attaining up to 30—40 mWm^{-2}, which may represent the driving force for the geological evolution and the critical parameter for the geophysical understanding of deep-seated processes.

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СВЯЗЬ МЕЖДУ ТЕПЛОВЫМ ПОТОКОМ И ТОЛЩИНОЙ ЗЕМНОЙ КОРЫ

В. ЧЕРМАК

Резюме

Территории сильного теплового потока характеризуются слабой или тонкой корей. Связь между тепловым потоком и толщиной земной коры является отрицательной, т. е. более высокые тепловые потоки характеризуется более тонкой корей. Радиоактивность не оказывает значительного влияния на эту связь.

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SOME RESULTS ON THE MANTLE HEAT FLOW INVESTIGATION IN ROMANIA

By

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This contribution is a review of the present data concerning the temperature distribution at the level of the Moho discontinuity and heat flow from the upper mantle for the territory of Romania.

From a geothermal point of view the Romanian territory is especially interesting, owing to the fact that distinctive changes in the thermal field are interrelated with the peculiar crustal structure and dynamics in a region where the Alpine orogenic belt comes in direct local contact with the old East European platform.

Some correlation has been established between geographic distribution of the terrestrial heat flow values and major tectonic units within Romania (VELICIU et al., 1977). According to these data, the Moldavian Platform is characterized by low heat flow, the Moesian Platform by low to normal and the Carpathian folded chain and its depressions by normal to high heat flows, except the Carpathian foredeep which is associated with a low heat flow.

In Fig. 2 after RADULESCU et al., (1976) a schematic cross section is presented with the earth's crust structure revealed by DSS on the International Profile XI. The location of the profile is given in Fig. 1.

On the cross section after VISARION et al., (1977) the calculated temperature distribution in the crust as well as the mantle heat flow values were plotted. The calculation has been performed assuming a stepmodel (ROY et al., 1968) and steady-state conditions (POLLACK, 1965; ČERMÁK, 1975).

For the uppermost few kilometres the heat productivity was obtained from gamma ray spectrometry, whereas a correlation with seismic velocity (RYBACH, 1975) was used for the lower crust (0.6 μW/m^3) and upper mantle (0.04 μW/m^3). Some heat conductivity values were quoted from KAPPELMEYER and HAENEL (1974) for the deep-seated layers (granitic layer: 2.4 W/mK°; basaltic layer: 3.1 W/mK°; upper mantle 4.0 W/mK°).

The calculated mantle heat flow (q_M) was obtained according to:

$$q_M = q_0 - \int_0^{z_M} A(z)\, dz$$

where: q_0 = measured surface heat flow; $A(z)$ = vertical distribution of the heat productivity; z_M = depth of the MOHO discontinuity.

Regarding the temperatures at the MOHO discontinuity and mantle heat flow, the studies performed so far allowed the following remarks:

Within the territory of Romania the MOHO discontinuity is clearly an unisothermal surface. There is an important horizontal temperature gradient



Fig. 1. Sketch showing the location of the profile (thick bar)

(may be of 200—300 °C/100 km) at the level of MOHO under the Eastern Carpathians.

The temperatures at the base of the crust are lower under platforms in front of the Carpathians (350—450 °C) than under the Alpine orogenic area (800—900 °C for the Eastern Carpathians, 600—700 °C for the Transylavanian Basin and over 800 °C for the eastern border of the Pannonian Basin).

The heat flow from the upper mantle increases westward. Under the foreland tectonic units of the Carpathian chain a mantle heat flow may exist being 2—4 times lower as compared to the Alpine orogenic area (11—16 mW/m², resp. 24—61 mW/m²). At the eastern border of the Pannonian Basin the contribution of the mantle to the surface heat flow exceeds probably 50%, while for the Moldavian Platform and the eastern part of the Moesian Platform this contribution may be less than 25 per cent.

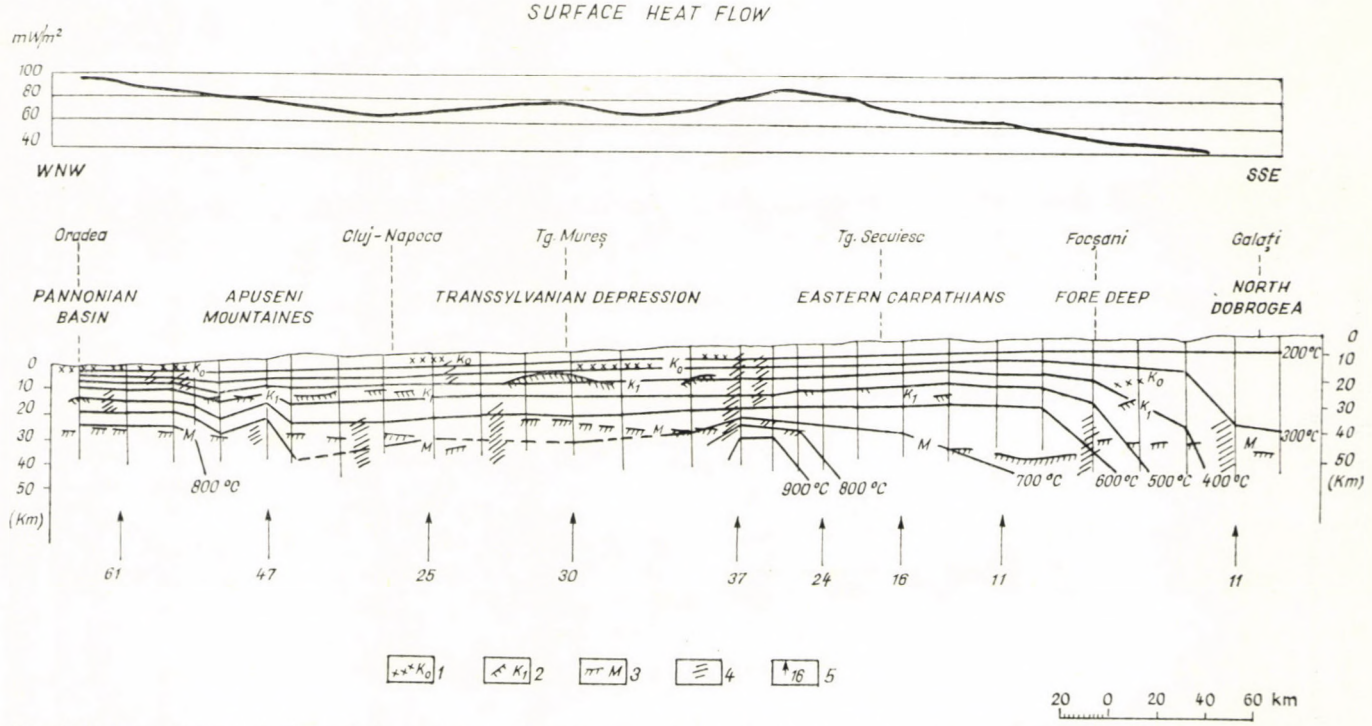

Fig. 2. Temperatures and the earth's crust structure along the deep seismic sounding profile XI. 1. Surface of the crystalline basement; 2. CONRAD discontinuity; 3. MOHO discontinuity 4. Faulted zone; 5. Calculated heat flow for the upper mantle in mW/m²

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НЕКОТОРЫЕ РЕЗУЛЬТАТЫ ИССЛЕДОВАНИЙ ТЕРМИЧЕСКОГО ТЕЧЕНИЯ МАНТИИ В РУМЫНИИ

СЕРБАН ВЕЛИЧЮ

Р е з ю м е

Даётся обзор имеющихся данных о распределении температур на уровне слоя Мохоровичича и о термическом течении происходящемся из верхней мантии на территории Румынии.

Acta Geologica Academiae Scientiarum Hungaricae, Tomus 21 (4), pp. 269—275 (1977)

CONTRIBUTION OF THE MANTLE TO THE HEAT FLOW IN NORTHERN CENTRAL EUROPE

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Heat generation models for the crust were carried out, based on the results of seismic measurements along the deep seismic sounding profile VII and of terrestrial heat flow measurements. The calculated models suggest that the heat flow contribution originating from crustal radioactivity is almost the same for different tectonic units (from 33.5 mWm^{-2} to 37.6 mW^{-2}).

The Paleozoic Platform region, the Sudetic Mountains (Variscan Internides) and the European Precambrian Platform, however, show considerable differences in mantle heat flow and temperatures.

In Central Europe there are about 812 heat flow determinations. In Poland heat flow investigations were made by PLEWA (1967, 1976), MAJOROWICZ (1973a, b, 1978) and WESIERSKA (1973).

A heat flow map of Poland and adjacent areas (Fig. 1) was constructed using published heat flow data, new geothermal measurements in deep boreholes of the Geological Institute and thermal conductivity measurements for sedimentary rocks. For Poland 110 data were used. Heat flow measurements for Czechoslovakia, USSR and the German Democratic Republic were published by ČERMÁK (1975), KUTAS et al., (1976), and HURTIG (1975).

To calculate mantle heat flow, definite vertical distribution of radiogenic heat generation must be specified and the surface heat flow must be known. As it was shown by LACHENBRUCH (1970) and ROY et al., (1971), heat generation in the crust is probably decreasing with depth. The relationship between the heat flow and heat generation obtained for different tectonic areas of the world (SWANBERG et al., 1974; RYBACH, 1976; RAO and JESSOP, 1975; ROY et al., 1971; LACHENBRUCH, 1970; ČERMÁK and JESSOP, 1975; MAJOROWICZ, 1978) lead to advance in understanding of the vertical distribution of heat sources in the crust. The model indicating decrease of heat flow sources with depth is consistent with the observed linear relationship between the heat flow and heat generation. The hypothesis seems to hold for all continents. The relationship between heat flow Q and surface heat production of magmatic rocks A for pre-Variscan areas including Precambrian shield, Precambrian Platforms and Moldanubicum of Czechoslovakia is shown in Fig. 2. Heat flow and radiogenic heat production values for different geotectonic continental provinces form a linear relationship (reduced heat flow — $Q_0 = 23$ mWm^{-2}).

These data agree very well with the result based on heat flow and heat generation data collected by JESSOP et al. (1976), for all Precambrian sites $Q_0 = 25$ mWm^{-2}.

It is surprising that the reduced heat flow values obtained for all Precambrian sites and pre-Variscan areas are lower than for Precambrian shields ($Q_0 = 29.3$ mW/m^2 after RAO and JESSOP 1975). For Precambrian shields and

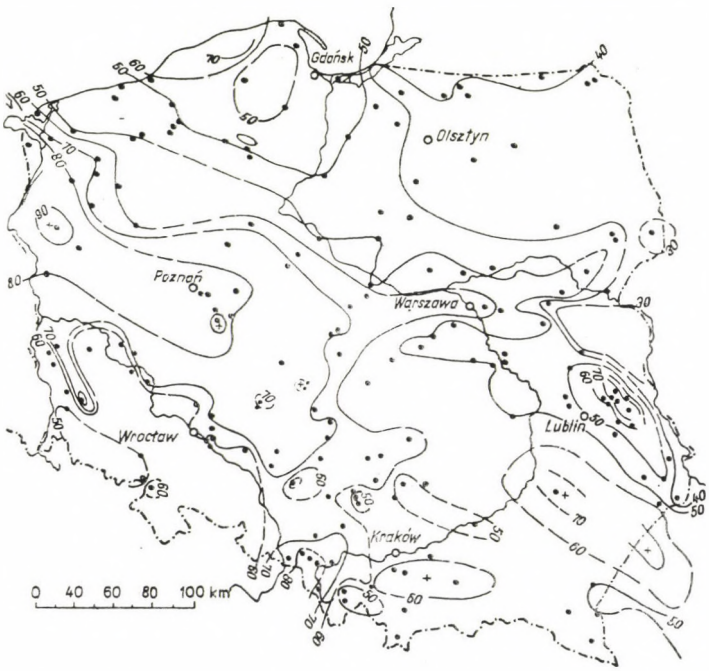

Fig. 1. Heat flow map of Poland, and adjacent areas in mWm^{-2}. High heat flow values can be observed in NW—SE direction parallel to the western border of the Precambrian Platform. The Precambrian Platform is characterized by relatively low heat flow around 45 ± 9 mWm^{-2} and low mantle heat How around 8—13 mWm^{-2}

craton (platform) areas a Q_0 of 28.5 was given by SLACK (1974). The differences, however, between 23 and 29.3 mW/m^2 are not significant in view of the error of the method that were used to estimate heat generation and Q values.

The relation between Q and A for the younger orogenic province of Basin and Range was defined by ROY, BLACKWELL and DECKER (1971) ($Q_0 = 58.7$ mWm^{-2}). This line is also shown in Fig. 2. The other published values from Meso-Cainozoic orogenic areas fall very close to this relation.

The model of heat generation decreasing with depth in the crust is supported by seismic data. There exists a strong correlation between the heat generation and compressional wave velocity V_p in the range from 5 to 8 km/s. Heat generation decreases with increasing V_p which reflects changes in rock

type, as it was shown on the basis of empirical data by RYBACH (1976). Heat generation shows also relation to density (HORAI and SIMMONS, 1968).

Terrestrial heat flow on crustal blocks along the international profile VII is well known. The velocity – depth relations are also well recognized (GUTERCH et al., 1975). From the relationship between the heat generation and compres-

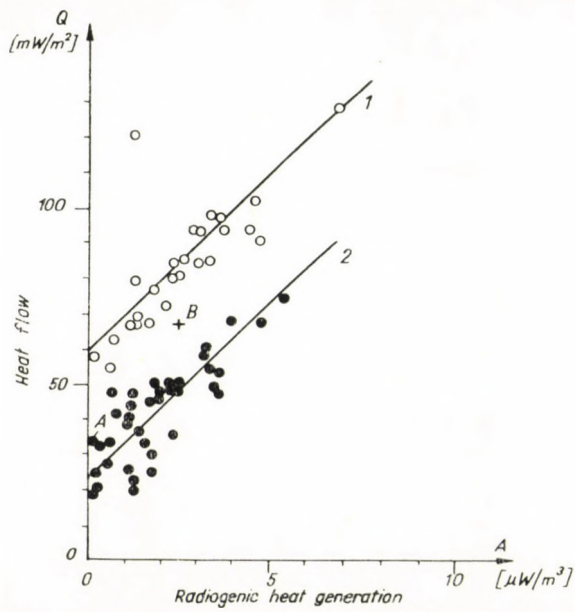

Fig. 2. Heat flow—heat production relationship. 1. The straight line corresponds to continental Meso-Cainozoic orogenic regions (Q and A data according to ROY et al., 1971; DUCZKOW SOKOLOWA, 1974; RYBACH, 1973; STEGENA et al., 1975); 2. the straight line corresponds to values for continental pre-Variscan tectonic regions (data after ARSHAVSKAYA et al., 1972; POLLACK, CHAPMAN, 1977; ČERMÁK, 1975; ČERMÁK, JESSOP, 1975; HYNDMAN, 1968; KUTAS, 1972; SASS, 1968; SWANBERG et al., 1974). The point marked with "B" ($Q = 67$ mW/m², $A = 2.5$) μWm^{-3} corresponds to mean heat flow and mean heat generation for the Variscides of SW Poland, the point marked with "A" ($Q = 34$ μWm^{-2}, $A = 0.1$ mWm^{-3}) represents heat flow and heat generation in an anorthosite intrusion in Northeast Poland (MAJOROWICZ, 1978)

sional wave velocity (RYBACH, 1976) and from the velocity model of the crust along the international profile VII, the heat generation models were calculated (Fig. 3). The heat generation models were established for the Sudets, the Fore-Sudetic Monocline (block *C*, block *D*), and the Old European Platform (block F_1, block F_2), where sufficient heat flow data were available. It can be seen that in general radiogenic heat production is decreasing with depth. The highest A is for the "granitic layer" and much lower A is characteristic of the lower crust. The mean heat generation in the sedimentary layer for the Precambrian Platform is 1.25 μWm^{-3} and 1.46 μWm^{-3} for Paleozoic Platform on the basis of laboratory determinations.

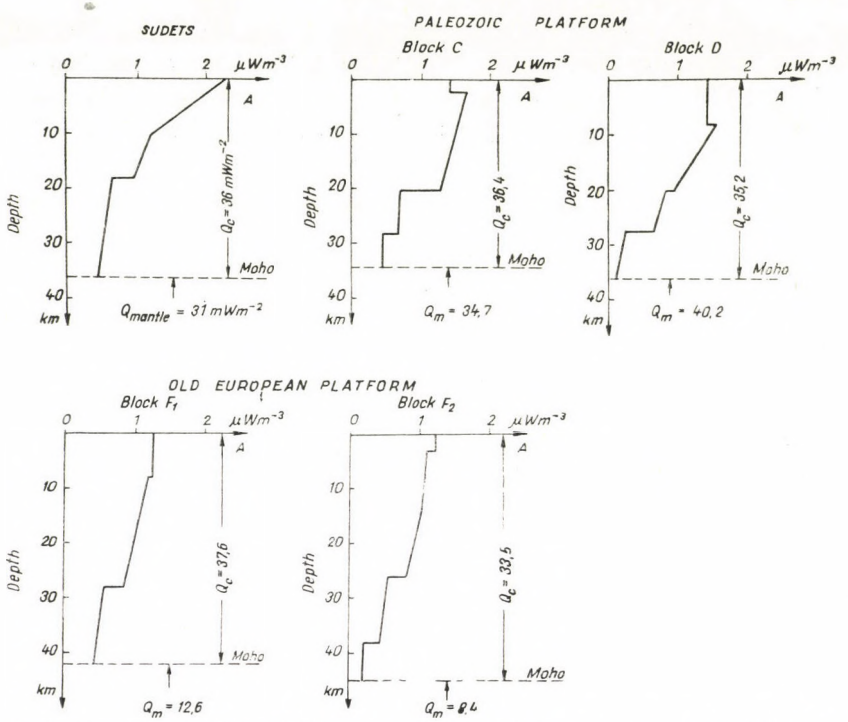

Fig. 3. Heat generation models A(z) for crustal blocks along the international deep seismic profile VII (Q_C is the heat flow contribution from crustal radioactivity, Q_M is the mantle heat flow)

From the heat generation $A(z)$ in the crust and the surface heat flow Q the mantle heat flow Q_M can be calculated

$$Q_M = Q - \int_0^{z_{Moho}} A(z)\,dz \tag{1}$$

The calculated crustal heat flow contribution and subcrustal heat flow (mantle heat flow) for Sudets, Paleozoic Platform (block *C*, block *D*) and Procambrian Platform (block F_1, block F_2) are shown in Fig. 4.

It can be seen that the crustal heat flow contribution originating from radiogenic heat generation is almost the same for the different blocks (from 33.5 mWm^{-2} to 37.6 mWm^{-2}). The observed differences in heat flow between the blocks of Precambrian Platform, Paleozoic Platform and Sudets are mainly caused by different mantle heat flow contributions. Mantle heat flow varies from 8.4 mWm^{-2} (block F_2) to 40.2 mWm^{-2} (block *D*). The mantle heat flow values for Paleozoic Platform and Sudets varies from 31 mWm^{-2} to 40.2 mWm^{-2} and are much higher than Q_M for the blocks of Precambrian Platform (Q_M = 8.4 — 12.6 mWm^{-2}).

The mean reduced heat flow for pre-Variscan sites ($Q_0 = 23$ mWm^{-2}) is considerably greater than mantle heat flow values obtained for blocks F_1 and F_2. (Q mantle = 8.4—12.6 mWm^{-2}.)

A surface layer of high heat generation interpreted from heat flow *vs.* heat generation relationship is 6.3 km for Precambrian shields after RAO and JESSOP (1975) which is considerably thinner than the thickness of "granitic

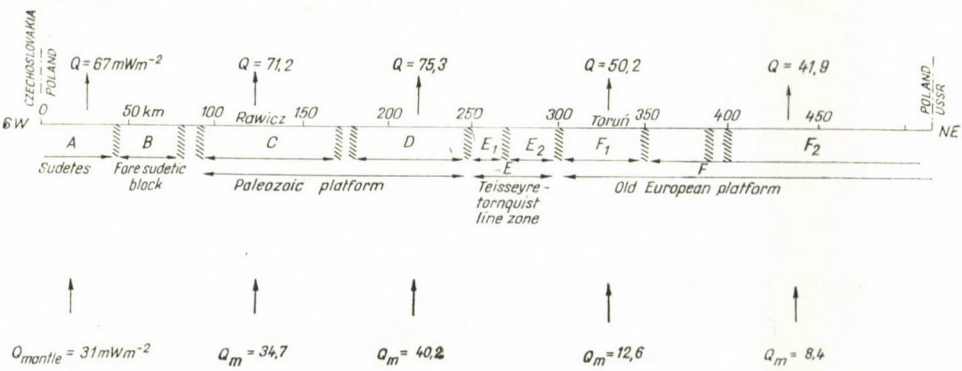

Fig. 4. Mantle heat flow values Q_M for crustal blocks along the international deep seismic profile VII

layer" in the area of European Precambrian Platform (about 20 km). The upper crust is of high heat generation as it follows from compressional wave velocity *vs.* heat generation relationship. The highest mantle heat flow value is obtained for block *D* of Paleozoic Platform, characterized by the highest terrestrial heat flow values ($Q = 75$ mWm^{-2}).

Considerable physical differences in the lower crust and upper mantle beneath the European Precambrian Platform region, Sudets and Paleozoic Platform region with Variscan basement should be expected, produced by lateral temperature variations. Petrophysical properties, especially electric conductivity are temperature-dependent. Magnetic properties of rocks change also with increasing temperature. Electrical conductivity differences of one or two orders of magnitude could exist in the lower crust and upper mantle along the international deep seismic profile VII between the blocks of Precambrian Platform and Paleozoic Platform.

The interpretation of further geophysical and geological data confirms that different subcrustal heat flow contributions can be expected in areas of different age.

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ВКЛАД МАНТИИ К ТЕПЛОВОМУ ПОТОКУ В СЕВЕРНОЙ ЧАСТИ СРЕДНЕЙ ЕВРОПЫ

Я. МАЙОРОВИЧ

Р е з ю м е

Установлены модели возникновения тепловому потоку для земной коры, на основании результатов сейсмических измерений вдоль рахреза глубинного сейсмического зондирования № VII и также измерений земского теплового потока. Вычисленная модель указывает на то, что доль радиоактивности земной коры в тепловым потоком почти одинакова в различных структурных единицах (от 33,5 mWm^{-2} до 37,6 mWm^{-2}). Однако тепловой поток происходящееся из мантии, и также температуры, оказались значительно различными в областях Палеозойской платформы, Судетских гор (Варисцийских интернид) и Европейской докембрийской платформы.

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INDUCED CONVECTION — A POSSIBLE SOURCE MECHANISM OF HEAT ANOMALY OF THE PANNONIAN BASIN

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According to the results of our numerical model calculations the Moho temperatures beneath the Pannonian Basin are between 600°—1000 °C and the heat flow at the Moho surface is 40—60 mWm^{-2}. All these values are much higher than those in the surrounding areas, so it can be concluded that the heat anomaly of the Pannonian Basin is well-developed also at the Moho surface. Such a thermal state can be explained probably only by taking into account also the convective heat transport. Numerical modelling of induced convection has been carried out. These latter calculations give a theoretical possibility for quantitative estimation of convective heating of the upper mantle.

It is well established that the Pannonian Basin is the area of an extremely developed heat anomaly. According to experimental evidences, the surface heat flow on this territory varies in general between 70 and 100 mWm^{-2} (Čermák and Hurtig, 1977), in some cases it even reaches the values of 130—135 mWm^{-2} (Boldizsár, 1975). In the surrounding areas, however, the mean surface heat flow values are not more than 50—60 mWm^{-2} (Fig. 1). Other geothermal data which prove the highly heated state of the Pannonian Basin are the results of measurements of temperature and the temperature gradient in water wells (VITUKI special edition, Budapest, 1965). In the territory of Hungary the mean temperature gradient is about 50—60 °C/km and in some particular areas (NE Hungary, SW part of the country) it reaches or exceeds 80—90 °C/km.

Similar results are obtained from results of electric carrotage in oil wells (e.g. Juhász, 1977), according to which the mean temperature gradient in SE Hungary is about 45—50 °C/km, moreover several points were found where this gradient has values more than 70 °C/km. In spite of the fact that the existence of the heat anomaly is clearly recognized, the causes of it are far from being understood. Most of the Hungarian authors are adherents of the plate tectonic model of the Pannonian Basin, i.e. it is assumed that early subduction took place in the neighbouring regions of this territory and this subduction led to partial melting of the upper mantle above subducting slabs and also induced anomalously high heat flow over it. Then the upwelling hot mantle material spread away horizontally at the base of the continental crust and slowly thinned it. Apart from detailed discussion of different works on this

subject we note only that the subduction hypothesis seems to be a very realistic one, a generally accepted plate tectonic model of the region, however, does not exist. In most of the related papers any exact dynamic — or at least qualitative — treatment of the whole phenomenon and of physical mechanisms of heating and melting of material in subduction zones is not given. Essential

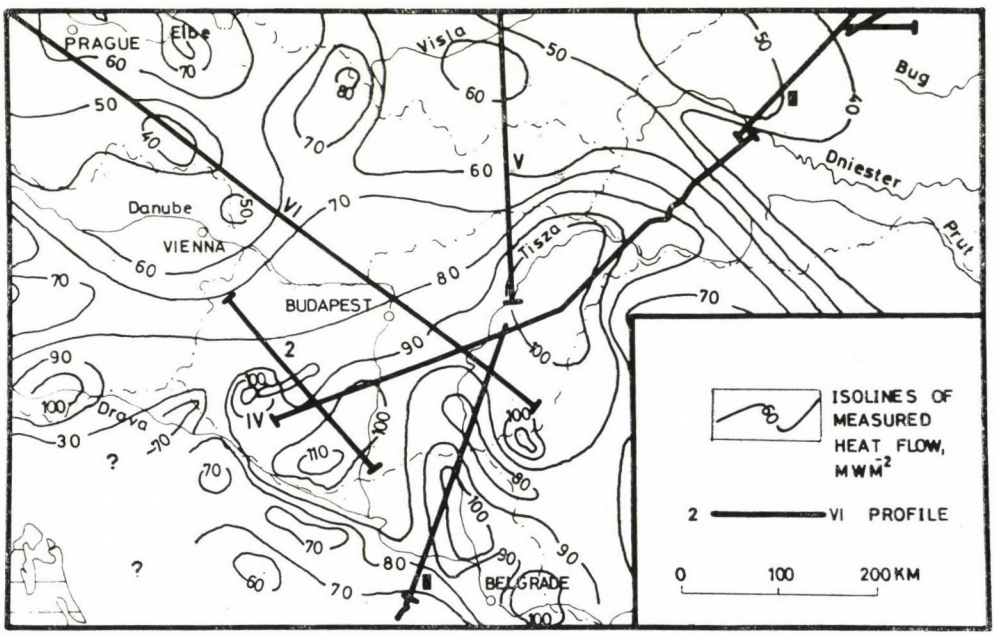

Fig. 1. Surface flow data in the investigated area

differences exist how the different authors approach the problem (sharp differences in the location of supposed subduction zones, in the number of subduction zones, e.t.c.). For the verification and refinement of the subduction model of the area different geological and geophysical data were used. Some authors (STEGENA et al., 1975; HORVÁTH and STEGENA, 1976) made attempts to use also surface heat flow and temperature data as evidences in favour of the plate tectonic model of the region, however, the surface heat flow pattern (Fig. 1) as well as other geothermal data obtained by surface observations show an extremely complicated behaviour in the investigated area and therefore these data cannot be of great use from this point of view. The existence of the heat anomaly itself can be hardly considered as such an evidence.

One of the main points of the present work was to obtain to what an extent the heat flow data (since these are the basic geothermal data) in the Carpathian Basin can be made use of for further evidence and refinement of the plate tectonic model of the region. For this purpose, we made first of all some

special reduction of surface heat flow data, i.e. the influence of the crust was excluded.

It was realized by model calculations of deep temperature along DSS profiles intersecting the investigated area.

It can be seen in Fig. 1 that five seismic profiles (four of them are international) cross the Pannonian Basin and these profiles cover the investigated area densely enough. Geological interpretation of these profiles is given by MITUCH and POSGAY (1972). The area in question was considered as a multilayered media with well-known thermal characteristics taken from ROY (1963), LUBIMOVA (1968), KAPPELMEYER and HAENEL (1974) and ČERMÁK (1975).

The thermal state of the investigated area was described by the following equation

$$\operatorname{div}(K \operatorname{grad} T) + A = 0\,, \tag{1}$$

where T is temperature, K is thermal conductivity and A is heat generation rate per unit volume. All of these variables are depending on the depth and horizontal extension and K depends also on temperature.

The following boundary conditions have been taken in present calculations:

a) There is no horizontal heat flow at the two sides of the area in question,

b) there is no heat flow in the direction perpendicular to the plane of the profile,

c) there is a given surface temperature distribution along the profile,

d) there is a given heat flow distribution through the lower boundary of the area in question.

This last heat flow distribution was obtained as a result of an additional iterative procedure.

Figure 2 shows part of the results of the above mentioned calculations, i.e. isotherms and deep heat flow along three of the investigated DSS profiles crossing the Pannonian Basin. For the sake of brevity we do not present here in detail the results obtained for all of the investigated profiles, instead of this only the short description of the results will be presented.

The main and general feature for all profiles is that the temperature distribution (or the position of the isotherms) out of the Pannonian Basin is completely determined by the crustal structure, while in the Pannonian Basin itself the heat flow from the upper mantle dominates. In other words, the thermal state of the Pannonian Basin is practically completely determined by the thermal state of the upper mantle.

In all cases the isotherms beneath the Pannonian Basin are much heightened as compared with those out of this territory, i.e. the Pannonian Basin is in a highly heated state in comparison with the surrounding area. If the melting temperature is taken to be of 1300 °C (pyrolite solidus curve from

CLARK and RINGWOOD, 1964) then part of the upper mantle beneath the Pannonian Basin should be in partially molten state. Due to the highly heated state of the territory, the problem arises: whether the stationarity condition used in equation (1) may be applied to the task of modelling of deep tempera-

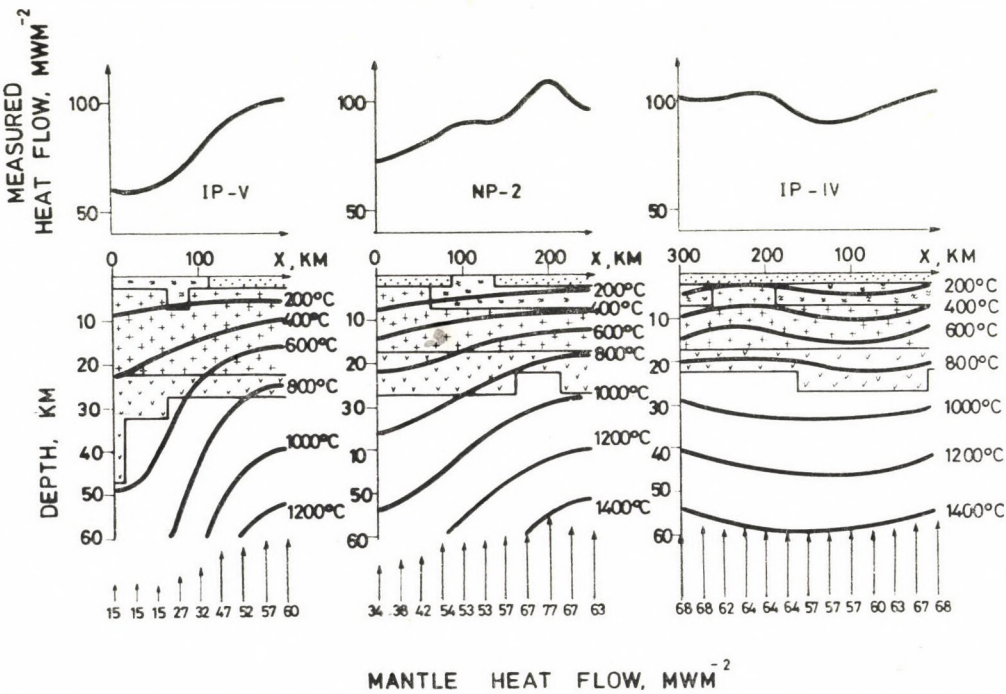

Fig. 2. Observed heat flow, the calculated temperature distribution and mantle heat flow along DSS profiles

tures or not. While this problem is not solved the calculated temperatures should be regarded as the first approximation to the real deep temperatures.

On the basis of numerical modelling of deep temperatures the map of heat flow through the Moho also has been constructed (Fig. 3). It can be seen that the heat flow pattern extrapolated for Moho depth is much more simple than the surface heat flow distribution shown in Fig. 1 and therefore it can be probably interpreted with much better accuracy than the other one.

After the part due to heat generation in the crust was separated from the surface heat flow, a long narrow zone of maximal Moho heat flow with NE—SW orientation was obtained instead of two local maximums of surface heat flow observed in SW and NE Hungary. In the peripheral regions of the Pannonian Basin, however, the isolines of the Moho heat flow repeat the contours of the main geological structures.

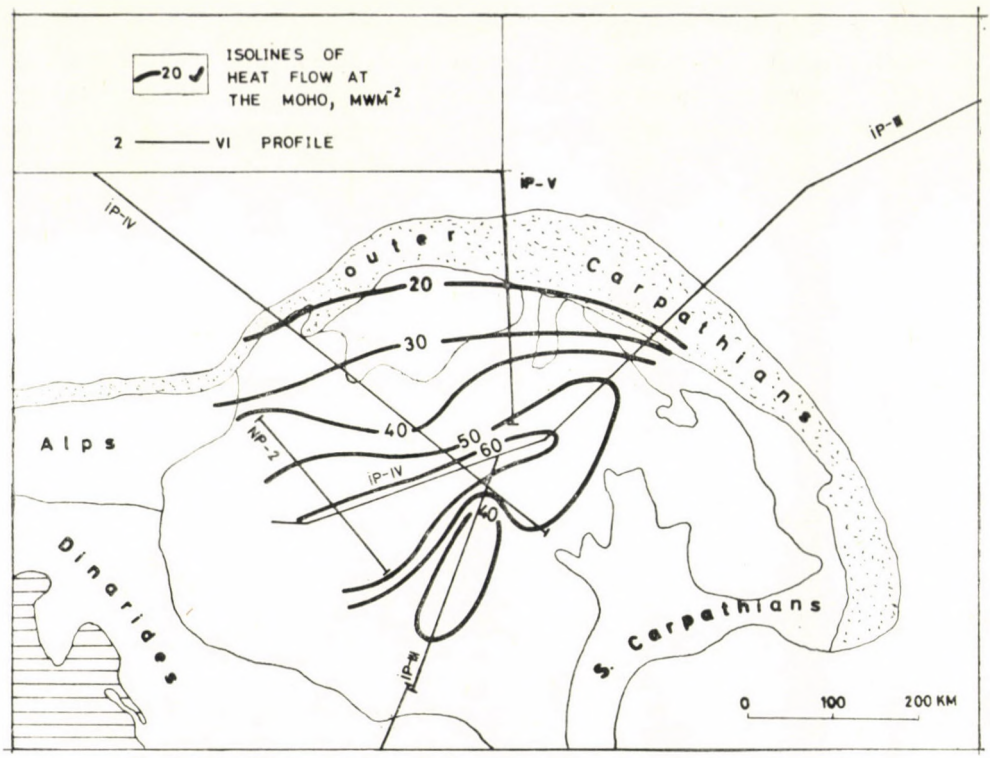

Fig. 3. Distribution of the calculated heat flow through the Moho surface

Similar variation of temperature can be seen on the map of the Moho temperatures which was also constructed. The zone of high Moho temperatures (900—1000 °C) is also very narrow and lies along the same direction as the zone of the maximal Moho heat flow.

Model calculations described below have shown that remarkable temperature changes at these depths cannot take place during the time of about 15 m.y. presumably passed from the stop of the subduction. Therefore, it can be concluded that upwelling and horizontal spreading of molten material (if it took place at all) could not be a general phenomenon for the whole Pannonian Basin but it could take place only along its NE—SW axis. This conclusion is supported also by the map of depths of the Moho surface (MITUCH and POSGAY, 1972), where the isolines of minimal depths have the same NE—SW orientation.

It is obvious that the anomalously heated state of the Pannonian Basin can be produced only by some special mechanism, since the radioactive heating alone is insufficient for it. The results of our calculations show that such a special mechanism of heating can be the frictional heating of material arising in areas of induced secondary convection.

In order to decide whether this mechanism can be applied to the Pannonian Basin or not, we made for this region a model calculation of subduction according to the algorhithm given in our previous papers (BODRI and BODRI, 1976; BODRI and BODRI, 1978). The subducting slab was given a constant velocity of 2 cm/yr, the presumable subduction started 30 m.y. ago and stopped

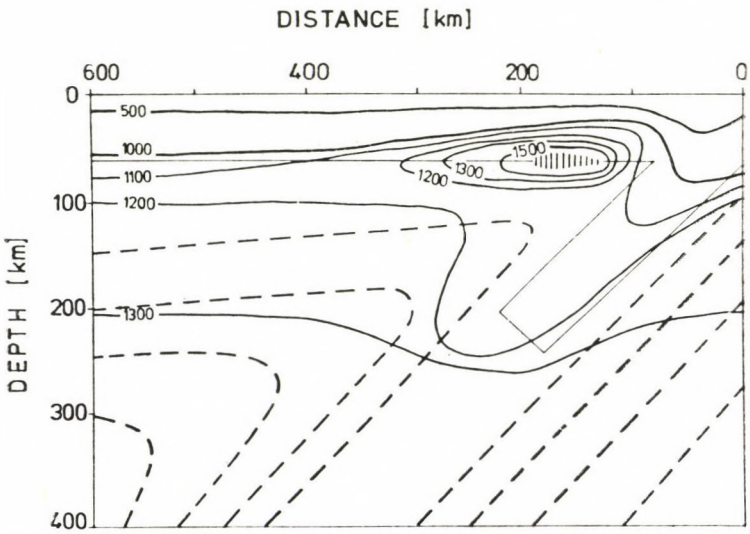

Fig. 4. Stream lines and isotherms at 10 m.y. from start of subduction

15 m.y. ago. Figure 4 shows a well-developed flow and temperature field at 10 m.y. after the subduction started.

The presence of a hot, molten zone at the base of the lithosphere is remarkable. The cause of heating and melting the material near the base of the lithosphere is the frictional heating arising at horizontal motion of the material induced by the subducting slab. The streamlines (indicated by dashed lines in Fig. 4) and flow velocities show that there are no upwelling and horizontal spreading of material near the corner of the wedge-shaped region bounded by the two plates, but on the contrary, the hot material is drawn down by the subducting slab. Figure 5 shows the isotherms of the model at 15 m.y. after the subduction stopped. As it has been referred to earlier, the temperature field (especially near the base of the lithosphere) has insignificant changes, only small smoothing of the isotherms can be observed.

Comparing the heat flow map given in Fig. 3 with the typical heat flow profile over island arcs (OXBURGH and TURCOTTE, 1968) and also with the surface heat flow profile calculated in the above mentioned modelling of subduction, one can see only one small local area (namely the southern periphery of the Pannonian Basin) where the similarity exists. The distribution of deep heat

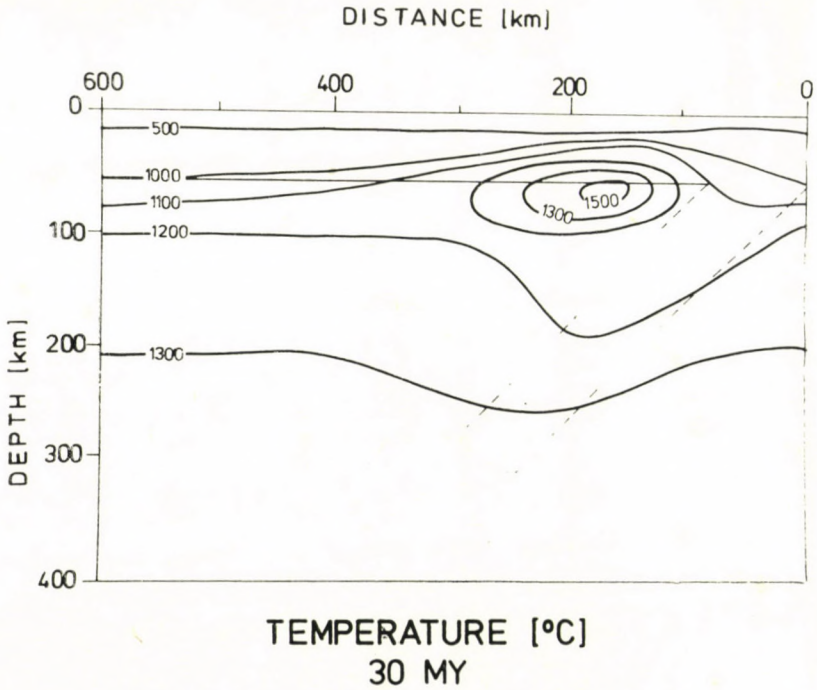

Fig. 5. Isotherms at 15 m.y. after subduction stopped

flow presented in Fig. 3 cannot be in accordance with the concept that subduction of lithospheric plates took place at the NE and SW boundary regions of the investigated area and it suggests that the observed surface heat flow distribution in the Pannonian Basin itself cannot be a simple reflection of processes occurring near the corners of the fixed and subducting lithosphere plates.

The different geothermal data presented or referred to in the present work suggest to consider the dynamic hypothesis elaborated qualitatively by SZÁDECZKY-KARDOSS (1966, 1971) as the most realistic one. According to this hypothesis, the Carpatho-Dinaric double subduction created a double cell of induced convection, the upwelling lines of which are approximately beneath the central part of the Pannonian Basin. So, the NE—SW axis of the investigated area along which the maximal deep heat flow and the minimal depth of the Moho surface propagated and along which some other geological and geo-

physical characteristics also present special behaviour, should probably be considered as a rift axis (provided that term "rift" is used in general sence, i.e. for localization of upwelling lines of convection cells of any kind). The numerical modelling of the stationary thermal state of usual rift systems was carried out by ANDREWS (1972). However, the model used in this work is not the same as the presumed one for the Pannonian Basin, the results of ANDREWS may be probably applied to our case (at least near the rift axis) without big errors. According to ANDREWS (1972) the hot molten or partially molten material is upwelling from a depth of about 200 km and is spreading away horizontally with a velocity of about 1 cm/yr. During this process the cooling of material is not significant since the temperature at the axis of the ridge reaches 1300 °C even at depth of about 25 km. No doubt that the calculations of Andrews (opened convection cells) cannot be applied completely to the Pannonian Basin and therefore, it is necessary to carry out the modelling of double subduction itself just for the Pannonian Basin. However, it can be already seen that the mechanism of heating proposed in the present work (i.e. frictional heating near the base of the lithosphere induced by horizontal mass flow) will take place also in this case and will produce anomalously high temperature and heat flow in the zone of upwelling of material.

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ВЫНУЖДЕННАЯ КОНВЕКЦИЯ — КАК ВОЗМОЖНЫЙ МЕХАНИЗМ ВОЗНИКНОВЕНИЯ ТЕПЛОВОЙ АНОМАЛИИ В ПАННОНСКОМ БАССЕЙНЕ

Л. БОДРИ—Б. БОДРИ

Резюме

Согхасно результатам нашил чисхенных модельных расчетов, температура поверхности Мохо под Паннонским бассейном составляет 600—1000°С, а тепловой поток на поверхности Мохо — 40—600 mWm^{-2}. Все эти значения гораздо выше, чем в окружающих районах, то есть на основании наших модельных расчетов можно сделать вывод, что наблюдаемая в Паннонском бассейне тепловая поверхностная аномахия хорошо развита и на поверхности Мохо. Такое термическое состояние можно объяснить, только приняв во внимание конвективный перенос тепла. В работе использовались также результаты по проведенной авторами ранее численной оценке вынужденной конвекции. Эти последние вычисления указывают теоретическую возможность для количественной оценки конвективного нагрева верхнией мантии.

Acta Geologica Academiae Scientiarum Hungaricae, Tomus 21 (4), pp. 287—296 (1977)

EINE DISKUSSION GEOPHYSIKALISCHER MODELLE DES GRENZBEREICHES ZWISCHEN OSTALPEN UND PANNONISCHEM BECKEN

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Das Modell der »Manteldiapire«,[1] welches L. STEGENA, B. GÉCZY und F. HORVÁTH 1975 für die Entstehungsgeschichte des pannonischen Beckens untersucht haben, wird in Hinblick auf geophysikalische Daten aus dem Ostalpenraum diskutiert.

Hebungsmessungen im Ostalpenraum bestätigen das relative Absinken des Pannonikums im Vergleich zum Alpenkörper. Herdlösungen von 3 Erdbeben im Wiener Becken zeigen eine Zerrung in E—W-Richtung an, was gleichfalls gut mit dem vorgeschlagenen Modell im Einklang steht. Der Frage nach dem Verbleib der durch den »Manteldiapir« erodierten und vermutlich wegtransportierten Unterkruste wird besonders nachgegangen. Schweremessungen zeigen eine vom Burgenland bis in die Gurktaler Alpen reichende schmale Zone mit Massenüberschuß an. Diese Zone könnte durch eine Intrusion von schwerem Unterkrustenmaterial in die Oberkruste erklärt werden. Das refraktionsseismische Alpenlängsprofil zwischen dem SPE und SPF überdeckt dieses Gebiet zwar, liegt jedoch ungünstig zum Aufschluß von Details. Es gibt lediglich Hinweise auf eine Überschuppung der mittleren Erdkruste vom E nach W über Teile der Oberkruste.

Einleitung

Der Bereich, von dem hier die Rede sein wird, ist im Westen durch das »Austroalpin« der Ostalpen, im Norden und Osten durch die Karpaten und im Süden durch die Dinariden begrenzt. L. STEGENA, B. GÉCZY und F. HORVÁTH (1975) haben für die Entstehung des pannonischen Beckens folgendes Konzept vorgeschlagen. Die Kollision der pannonischen Mikroplatte mit dem europäischen Kontinent im Miozän erzeugt Subduktion der europäischen Platte im Norden und der afrikanischen im Süden. Partielle Aufschmelzung der subduzierten Kruste und konvektives Aufsteigen der leichten Massen bewirkt eine Erosion der Kruste von unten her (Manteldiapir). Das geschmolzene Krustenmaterial wird seitlich abtransportiert. Nach Abschluß der Pressungsphase und Subduktionszonenbildung sinkt die Kruste isostatisch ab. Es kommt zu örtlichen Zerrungen und dem Aufsteigen basaltischer Laven. Beim Absinken füllt sich das Becken mit leichten Sedimenten. Dieses Modell läßt jedoch die Frage offen, wie der Alpenostrand bei der Bildung der Subduktionszonen und bei der

[1] Der Begriff "Manteldiapir" wird so benutzt wie bei STEGENA et al. (1975)
Die wesentlichen Züge des hiermit dargestellten Prozesses der Krusten- und Lithosphärenbildung findet man bereits bei SZÁDECZKY-KARDOSS (1966)

Beckenbildung mitgewirkt hat. Es ist daher das Ziel dieser Arbeit, durch Hinzufügen geophysikalischer Informationen Beiträge zur Beantwortung dieser Frage zu erbringen.

Seismizität der Ostalpen

Die Vorstellung, daß Erdbebenzonen die Berührungszonen in sich starrer Krustenplatten darstellen, ist auch auf den Alpenostrand anzuwenden. Es gibt z. B. eine von den *Kleinen Karpaten* bis ins *Lavanttal* nach WSW streichende Bebenzone, die eine beachtliche Länge von 300 km erreicht (Abb. 1). Die Bebenaktivität ist im östlichen Teil der Zone an den Ostrand des Wiener Beckens gebunden, das man zu den inneralpinen Einbruchsbecken rechnet. TOLLMANN (1970) hat nachgewiesen, daß es im Aufbau alle Kennzeichen der großen Grabenrüche der Erde, wie z. B. des Rheintalgrabens aufweist. Die

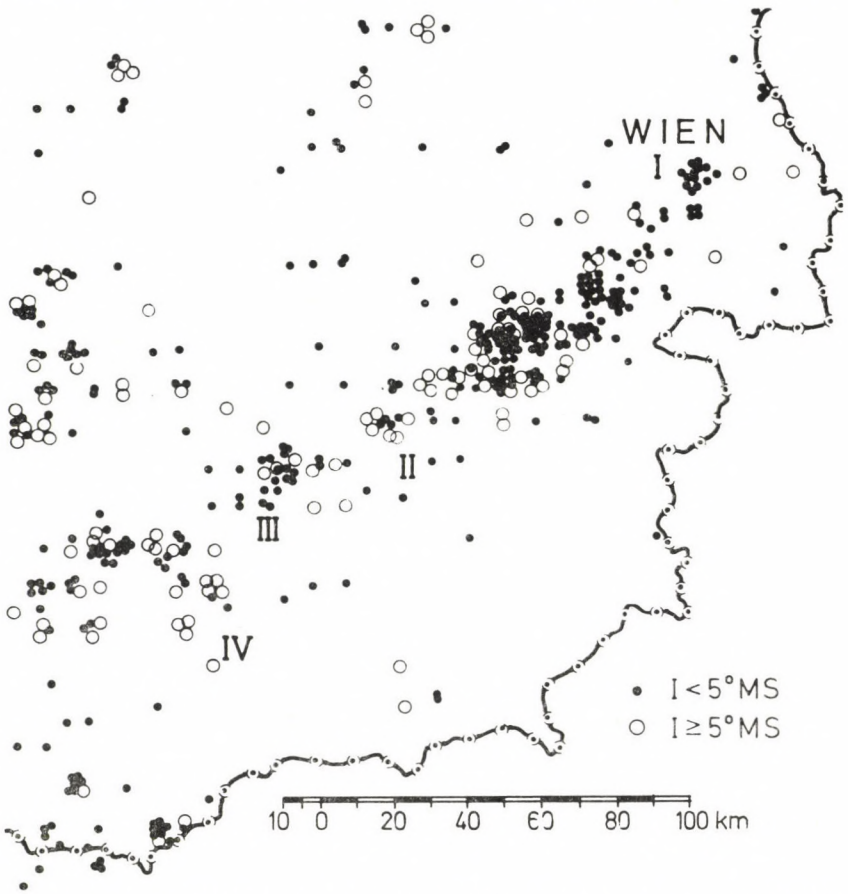

Abb. 1. Erdbeben in Österreich. I_0 5° MS, I_0 5° MS. I: Wiener Becken, II: Mürztal, III: Murtal, IV: Lavanttal;

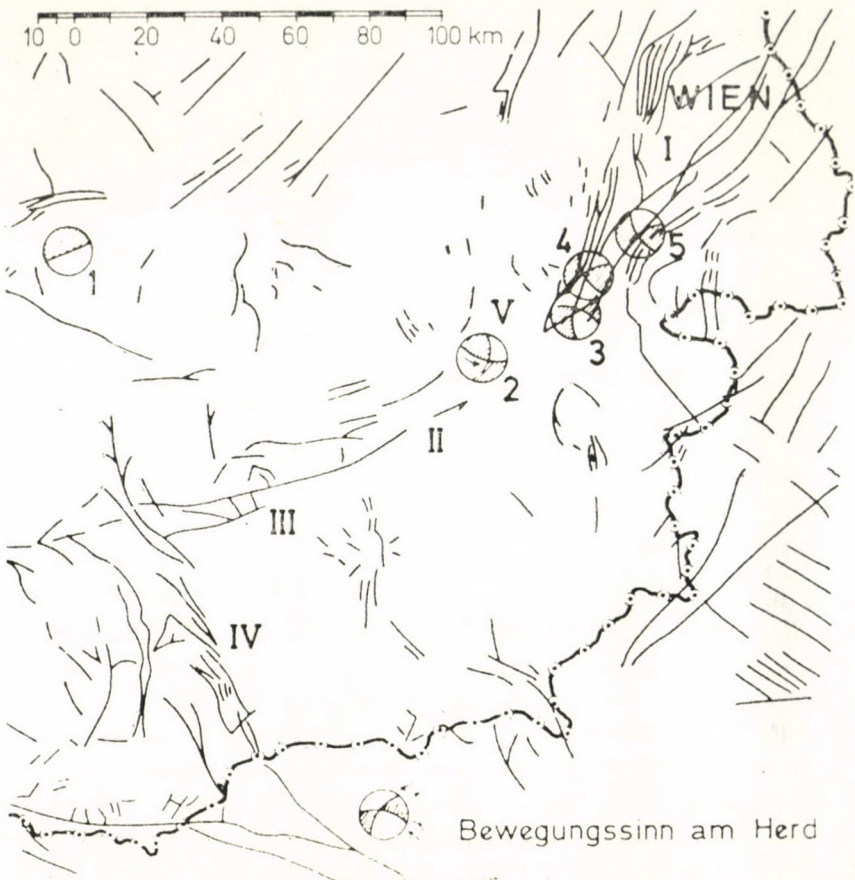

Abb. 2. Geologische Störungen nach A. TOLLMANN (1970) und Herdlösungen. I: Wiener Becken, II: Mürztal, III: Murtal, IV: Lavanttal, V: Semmering. Herdlösungen der Beben: 1. Molln 29. Jan. 1967 (J. DRIMMEL und E. TRAPP 1975); 2. Semmering 27. Okt. 1964 (G. GANGL 1975); 3. Seebenstein 16. Apr. 1972 (G. GANGL, 1975); 4. Neukirchen 16. Apr. 1972 (G. GANGL, 1975); 5. Wr. Neustadt 1. Mai 1972 (G. GANGL, 1975)

Bebenzone setzt sich aber nach SW hin über den Gebirgsrand des SEMMERING fort und verläuft westlich davon im Tal der Mürz und Mur. Dieses Tal ist durch ein Verwerfungssystem vorgezeichnet (Abb. 2). Die Bebenzone verläuft demnach längs verschiedener Störungszonen, die untereinander keine offensichtliche Verbindung haben. Demnach muß das tektonische Spannungsfeld, welches die Ursache für die Seismizität darstellt, von wesentlich größeren räumlichen Dimensionen sein als die besprochenen Einheiten.

Auf Grund der geologischen Befunde ist zunächst nur für das Wiener Becken eine E—W-gerichte Zerrung vorstellbar. Für diesen und den SW anschließenden Raum liegen heute bereits einige Bearbeitungen von Erdbeben vor, die stark genug sind, um eine gesicherte Aussage über den Herdmechanismus zu liefern. Abbildung 2 zeigt die Herdlösungen einiger Beben nach

G. GANGL (1975) aus dem Gebiet des Wiener Beckens und dem Mürztal. Drei der vier dargestellten Herdlösungen weisen auf eine normal zum Alpenstreichen verlaufende Zerrung hin und sind daher gut mit dem Konzept des absinkenden pannonischen Beckens verträglich. Der weiterreichende naheliegende Schluß, daß die gesamte beschriebene Erdbebenzone, von den Kleinen Karpaten bis zum Lavanttal, als ein Zerr-Riß zu interpretieren sei, der das absinkende Pannonikum im Westen abgrenzt, dürfte jedoch noch verfrüht sein. Allerdings wäre es auch denkbar, daß die Beben des Mur- und Lavanttales bereits im Einflußbereich des südalpinen Spannungsfeldes mit vorwiegend NS gerichteter Stauchung liegen (A. R. RITSEMA (1974), G. MÜLLER (1976). Um darüber eine Entscheidung zu treffen, wären aber noch viel mehr Daten über Herdlösungen erforderlich.

Präzisionsnivellements

Ein vollständigeres Bild der rezenten Bewegungen erhält man, wenn man die bei Präzisionsnivellements erhaltenen Hebungraten beachtet. Eine solche Karte existiert für die Ostblockländer, in der Ungarn als Senkungs- und die Südalpen und Dinariden als Hebungsgebiete aufscheinen. Der maximale Unterschied zwischen den Hebungen und Senkungen liegt bei 10 mm pro Jahr. Wenn auch der Einwand der Geodäsie, daß die Meßgenauigkeit zu klein für solche Aussagen ist, besteht, so wird doch die Tendenz sicher richtig wiedergegeben. So ist z. B. nachgewiesen worden, daß das Wiener Becken sich gegen den Alpenostrand (Hohe Wand) absenkt und daß die Südalpen sich im Vergleich zu den Nördlichen Kalkalpen heben (G. GANGL, 1975, E. SENFTL und CH. EXNER, 1973). Demnach bestätigen auch die bisherigen Hebungsmessungen das oben erwähnte Konzept.

Die Frage nach dem Verbleib der erodierten Unterkruste

Gegen das Konzept der Manteldiapire gibt es einen wesentlichen Einwand. Es stellt sich die Frage, wohin das weggeschmolzene Material der Unterkruste gelangt ist. Beim Wegschmelzen von 8 km Unterkurste müßte unter ganz Ungarn mehr als 1 Million Kubikmeter aufgeschmolzene Masse abgewandert sein. Diese Menge läßt sich durch den Vulkanismus allein nicht wegschaffen. Es ist auch schwer vorstellbar, daß die Massen Wanderwege von kontinentalem Ausmaß zurücklegen, weil die Wurzeln der umgebenden Gebirge einem ungestörten Abfließen eher hinderlich wären. Die Aufgabe wird darin bestehen, mit geophysikalischen Methoden das verschwundene Unterkrustenmaterial in den Randzonen des Pannonikums zu suchen. Zur Klärung dieser Frage können Schweremessungen und die Refraktionsseismik beitragen.

Das abgeschmolzene Unterkrustenmaterial hat eine Dichte von ca. 3,0 g cm^{-3}. Wenn es irgendwo in die leichtere Oberkruste mit ca. 2,7 g cm^{-3} eingedrungen sein sollte, müßte man dort eine positive Schwereanomalie feststellen. Solche Zonen wären dort zu erwarten, wo zur Zeit der Schmelzenwanderung noch eine tektonische Aktivität herrschte, z. B. am Alpenostrand. Abbildung 3 zeigt die Bouguer-Karte des ostalpin-karpatischen Raumes (E. SENFTL, 1965, Bureau Gravimétrique International, 1965). Hier ist folgendes festzustellen:

Die Schwereanomalie im Wiener Becken liegt wie im Pannonikum um Null. Das Wiener Becken ist ein geologisch sehr junges Senkungsfeld. Außerdem haben Bohrungen jeden Zweifel ausgeräumt, daß die Kalkalpen sich unter dem Wiener Becken nach Nordosten fortsetzen und die Verbindung zu den Karpaten herstellen. Aus diesen Fakten ist u. E. der mögliche Schluß zu ziehen, daß die Alpenwurzel die Absenkung nicht mitgemacht hat. Hierfür sprechen auch Fernbebenbeobachtungen (R. GUTDEUTSCH, A. FRANKE, W. SEIBERL, K. ARIC, 1976). Wäre dieses nämlich der Fall, müßte man zwangsläufig im Wiener Becken, ähnlich wie im Alpenhauptkamm, eine negative Schwereanomalie

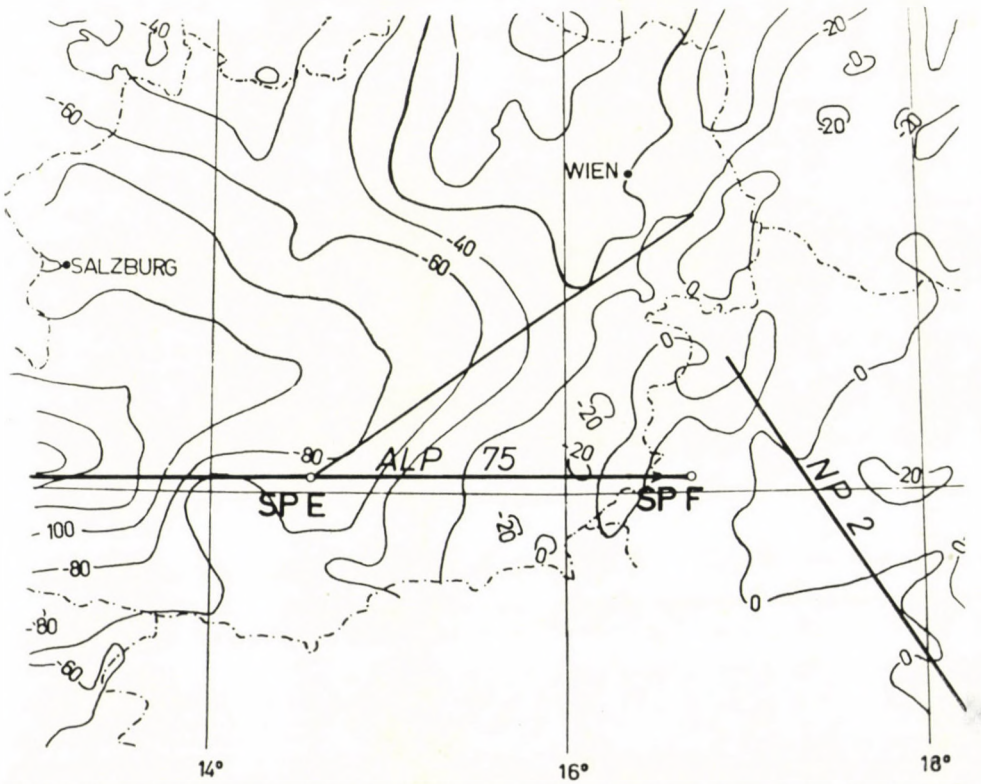

Abb. 3. Karte der BOUGUER-Anomalie des Grenzbereiches. Ostalpen-Pannonisches Becken; Die refraktionsseismischen Profile ALP mit den Schußpunkten SPE und SPF und das Ungarische Nationale Profil NP12 sind eingetragen

beobachten. Eine weitere Möglichkeit bestünde darin, daß der offensichtliche Massenüberschuß durch das aus dem Pannonikum abgewanderte Unterkrustenmaterial hervorgerufen wird.

Eine langgestreckte positive Anomalie von ca. 40 mgal reicht vom Gurktal über das Drautal von NW nach EW einschwenkend in den jugoslawischen Raum, aus dem z. Zt. keine Schweredaten zur Verfügung stehen. Es hat aber den Anschein, daß sie sich weiter östlich ins Burgenland nach NE fortsetzt. Dort wird sie vom Alpenlängsprofil zwischen SPE und SPF begrenzt. Wenn diese Verbindung wirklich besteht, hat die Anomalie die Länge von fast 300 km und könnte mit der Wanderung von Schmelzen in der Unterkruste im Zusammenhang stehen. W. SEIBERL (1978) hat durch Methoden der Feldfortsetzungen gefunden, daß der Störkörper im Osten nahe der Erdoberfläche liegen muß. Im Westen dagegen taucht er in größere Tiefen ab. Jenseits des Knickes dürfte er in Tiefen über 10 km liegen.

Abbildung 4 zeigt einen Ausschnitt des letzten Bildes und stellt im wesentlichen den westlichen Ausläufer der Anomalie dar. Neben den Isogrammen

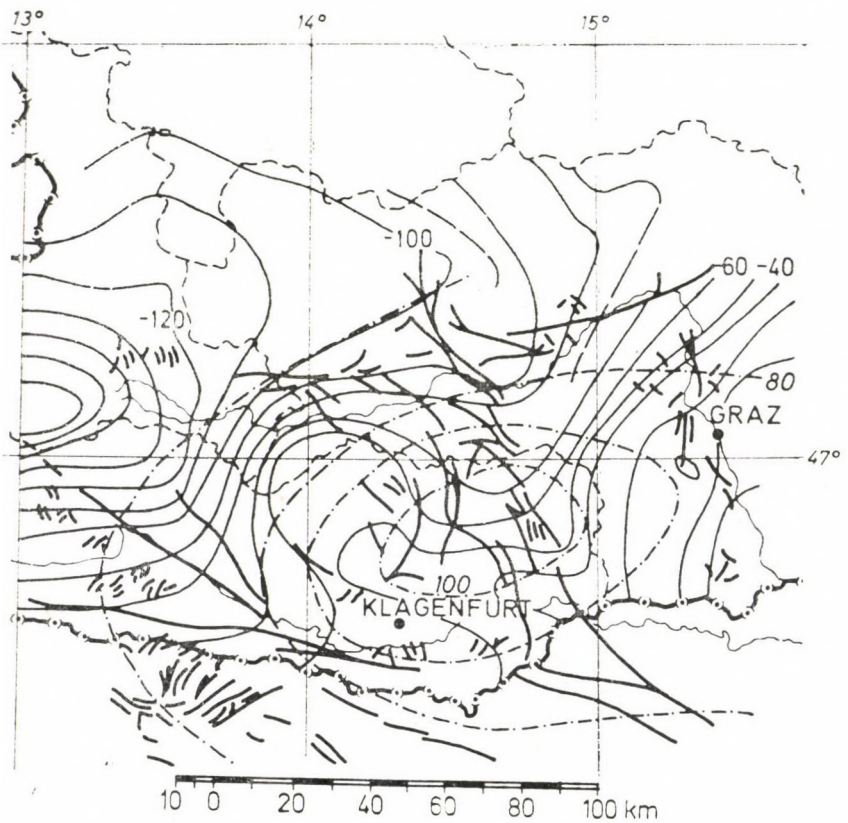

Abb. 4. Schwereanomalie im Lavanttal. —— Schwereanomalie in mgal; — . — . Wärmestrom in mWm^{-2}; Störungszonen nach A. TOLLMANN (1970)

sind auch die Hauptstörungslinien nach A. TOLLMANN (1970) und die Wärmestromdaten von V. ČERMÁK und E. HURTIG (1977) eingetragen. Eine positive Wärmestromanomalie deutet sich an. Es fällt auf, daß die Störungszonen des *Lavanttales* die Schwereanomalie seitlich um die Bouguer-Anomalie herumlegen. Die Störungszonen sind seismisch aktiv. In diesem Jahrhundert hat es in diesem Raum über 40 Erdbeben gegeben. Die Flächen der Verwerfungen deuten auf ein Absinken der westlichen Partien gegen die östlichen (P. BECK-MANNAGETTA, 1952). Demnach liegt das Schwerehoch in einem rezenten Senkungsraum. Möglicherweise stehen diese beiden Fakten in einer ursächlichen Beziehung zueinander, wie es — wenn auch in anderen Dimensionen — im pannonischen Becken der Fall ist.

Refraktionsseismik

Auch die Seismik kann Beiträge zu der erwähnten Frage liefern. Man wäre sicher durch geeignet angelegte Profile in der Lage, Zonen mit Unterkrustenmaterial auf Grund seiner abweichenden akustischen Parameter seismisch zu lokalisieren. Der Grenzbereich zwischen Ostalpen und pannonischem Becken ist durch das 1975 durchgeführte Alpenlängsprofil refraktionsseismisch erfaßt worden [*Alpine explosion seismology group* (21 Autoren), 1976]. Die Lage des Profils wird durch die Abbildung 3 dargestellt. Eine Seismogrammontage des östlichen Profilstückes zwischen SPE und SPF zeigt Abbildung 5. Bei $x = 90$ km scheint die seismische Energie stark nachzulassen. Man kann nicht entscheiden, ob die Laufzeitkurve abbricht, d. h. eine "Schattenzone" vorhanden ist. Das Abbrechen der Laufzeitkurve wäre das einzig sichere Indiz für das Vorhandensein einer Geschwindigkeitsinversion. Im Westen erschien uns der Einbau eines Kanals wegen gewisser Einsätze, die man als Reflexionen deuten könnte, sinnvoll. Die Reflexionen sind zwar berechnet, jedoch hier noch nicht eingetragen. Es ist ein 2-dimensionales Modell berechnet worden, das die von K. ARIC, I. ALBU, R. GUTDEUTSCH, E. MITUCH und K. POSGAY, 1978 erarbeiteten Resultate verwendet. Man könnte die Gestalt des Modells in der mittleren Kruste als Überschuppung von Osten nach Westen deuten. Das Modell ist noch nicht bestens angepaßt und steht auch noch nicht mit den Schweremessungen in Einklang. Der Bereich des burgenländischen Schwerehochs (Abb. 3) konnte durch die Anordnung des Alpenlängsprofiles nicht ausreichend erfaßt werden. Daher bleibt auch die Frage noch offen, ob unter dem Alpenostrand ein abgeschmolzener Subduktionsrest in Form einer Moho-Einsenkung vorhanden ist.

Die hier diskutierten geophysikalischen Ergebnisse liefern gemeinsam einen Beitrag zur Beantwortung der Frage nach der Entstehung des pannonischen Beckens und zeigen ihrerseits auch eine methodische Grenze auf.

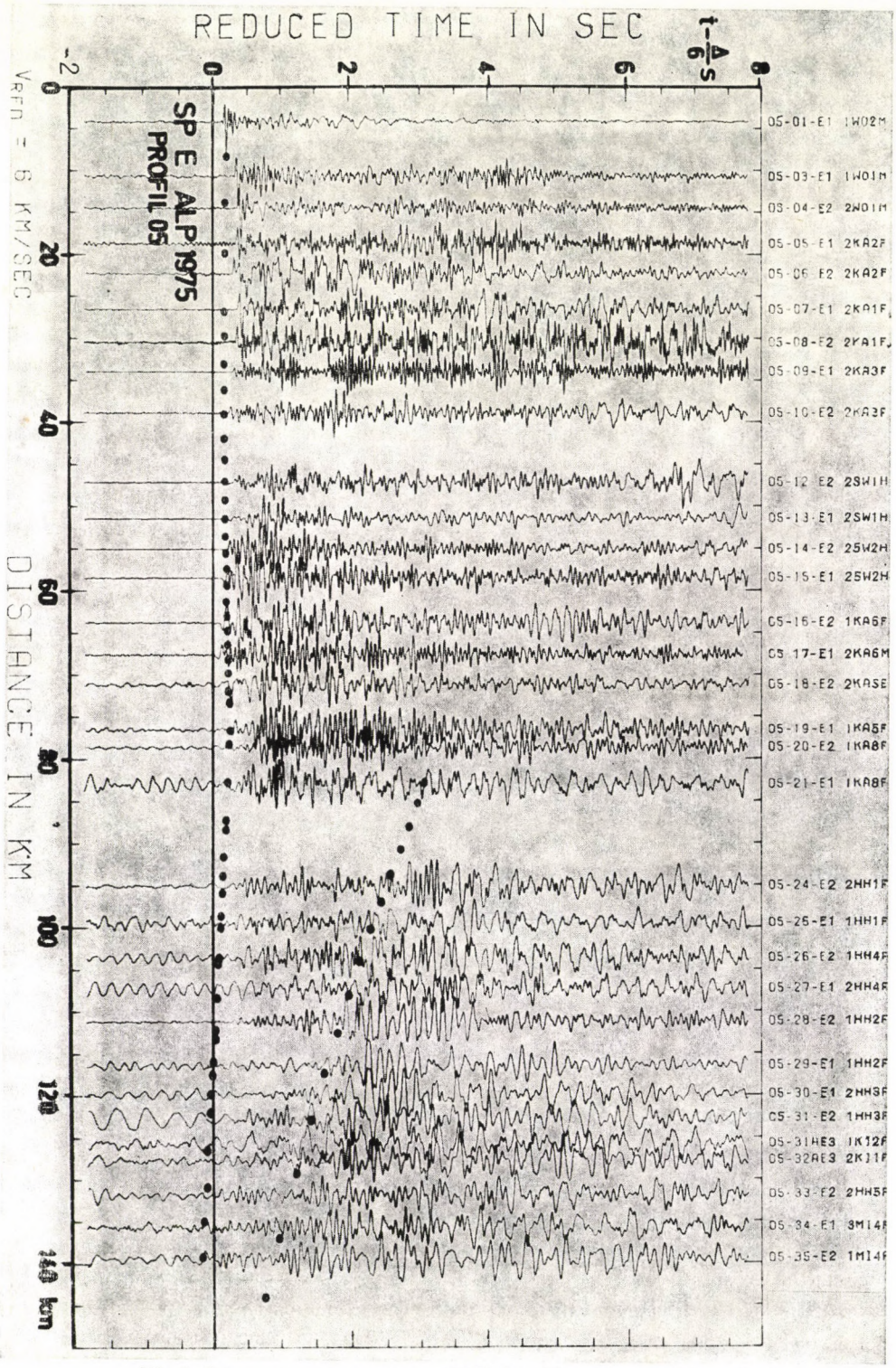

Abb. 5. Seismogrammontage von SPE des Alpenlängsprofils nach E

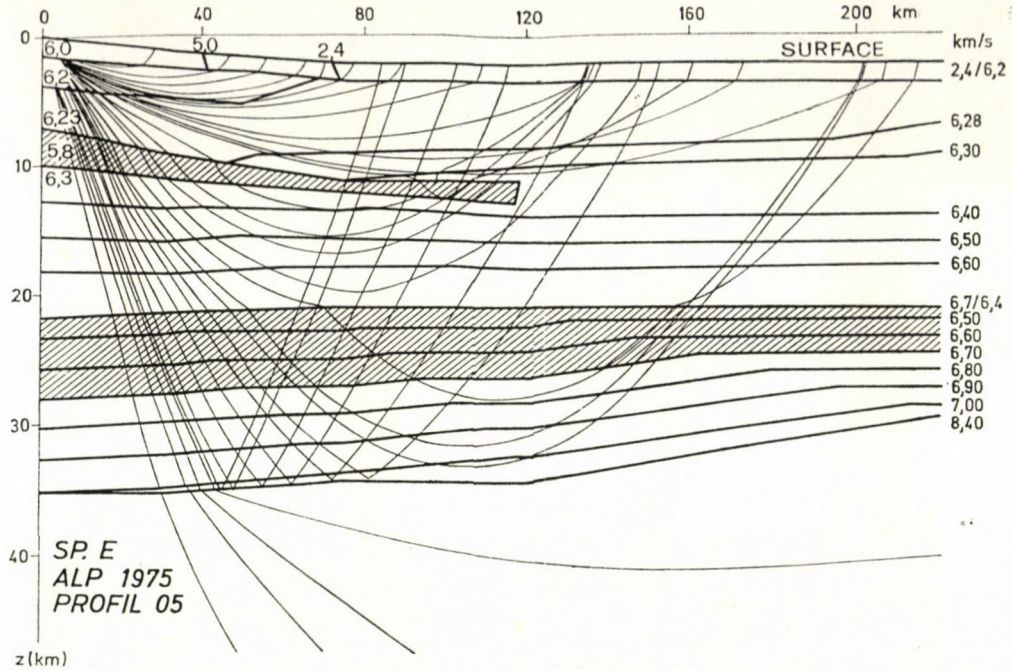

Abb. 6. Erster Entwurf eines Modells des Alpenostrandes, Verteilung der P-Wellengeschwindigkeiten nach K. ARIC, A. FRANKE und R. GUTDEUTSCH (1978)

DANKSAGUNG

Diese Arbeit stellt einen Beitrag zu dem von der Österreichischen und der Ungarischen Akademie der Wissenschaften geförderten Forschungsschwerpunkt »Das Übergangsgebiet zwischen Ostalpen und Pannonikum« dar. Das Österreichische Nationalkommitee des Geodynamischen Projektes hat diese Arbeit finanziell unterstützt.

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ДИСКУССИЯ ГЕОФИЗИЧЕСКИХ ПОГРАНИЧНОЙ ОБЛАСТИ МЕЖДУ ВОСТОЧНЫМИ АЛЬПАМИ И ПАННОНСКИМ БАССЕЙНОМ

Р. ГУТДАЙЧ и К. АРИЦ

Резюме

Рассматривается модель «диапиров мантии» предложения Л. Штегена, Б. Геци и Ф. Хорват (1975) для Паннонского бассейна, при помощи геофизических данных о Восточных Альпах.

Изменения нивеллировки проведённые в оюласти Восточных Альп поддерживают идею о поднятии массы Альп в отношении к Паннонскому бассейну. Фокусные решения трех землетрясений в Вьенском бассейне указывают на горизонтальное напряжение в З-В направлении. Путём подкорной эрозией получились огромные каличества расплавлённого вещества. Возникает вопрос, где они находятся в настоящее время. На карте Бугер-аномалий установлено положительная аномалия от Бургенланда до Гуркталъских Альп, избясняемая диапиром вещества высокой плотности походящимся во верхней мантии. Данные рефракционной сейсмики не подтверждают этой гипотезы так как разрезы «Альп 1975» неудачно расположены с этой точки зрения. Они соответствуют модели флексуры средней коры над верхней коры направленной от В на З.

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ПЕРВЫЙ ВАРИАНТ КАРТЫ СЕЙСМИЧЕСКОЙ АКТИВНОСТИ НА ТЕРРИТОРИИ ЧЕХОСЛОВАКИИ ВЕНГРИИ И АВСТРИИ

И. БРОУЧЕК,
ГЕОФИЗИЧЕСКИЙ ИНСТИТУТ СЛОВАЦКОЙ АН, БРАТИСЛАВА

Д. ЧОМОР
ГГИИ ВАН, СЕЙСМОЛОГИЧЕСКИЙ ИНСТИТУТ БУДАПЕШТ

Д. ПРОХАЗКОВА
ГЕОФИЗИЧЕСКИЙ ИНСТИТУТ ЧСАН, ПРАГА

Эта карта даёт объективную картину распределения эпицентров соответствующих одному уровню энергии и создавает возможность для определения что какие области являются высокой сейсмической активностей связанной с тектоникой. Изолинии $A_4 = 0{,}1$ 0,3, 1,0 и 3,0 видные на карте. Для окончательного варианта карты надо ещё решить некоторые открытые вопросы, напр. изображение активности на переход из одной геологической системы в другую, определение оптимальных параметров для расчёта активности и продолжительности репрезентативного временного интервала.

В рамках подготовки материалов для конструкции карты сейсмического районирования КАПГ подготовливается карта сейсмической активности стран КАПГ. Эта карта дает объективное представление о распределении плотности эпицентров по отношению к одному энергетическому уровню и позволяет лучше сопоставить области повышенной сейсмической активности с тектоническим строением чем простая карта эпицентров. Для определения сейсмической активности были использованы следующие знакомые формулы:

$$\log N = a - bM = a - bM - bM_0 + bM_0 =$$

$$= a - bM_0 - b(M - M_0) = a' - b(M - M_0)\,, \qquad (1)$$

откуда

$$N = 10^{a' - b(M - M_0)}\,,$$

Где

N-число землетрясений с магнитудой M, Mo — избранная магнитуда, a, b, — численные параметры.

$$N_c(M_j) = \sum_{i=j}^{S} N(M_i) = \sum_{i=j}^{S} 10^{a' - b(M - M_0)}\,,$$

Где S число магнитудных классов.

Т. е., для числа землетрясений N_Σ с магнитудой $M \geq M_{\text{мин}}$ имеем:

$$N_\Sigma = N_c(M_{\min}) = 10^{a'} \cdot 10^{-b(M_{\min} - M_0)}[1 + 10^{-b} + 10^{-2b} + \ldots + 10^{-sb}] \qquad (2)$$



Фиг. 1. Сейсмическая активность 1900—1970. $M_{min} = 3{,}0$; $M_{0C} = 4{,}0$; $\tau = 13{,}22$ km

Выражение в скобках является геометрической прогрессией со знаменателем 10^{-b}. После доставки $A_0 = 10^{a'}$ в управление (2) получими:

$$N_\Sigma = A_0\, 10^{-b(M_{\min} - M_0)} \frac{1 - 10^{-(s+1)b}}{1 - 10^{-b}},$$

откуда

$$A_0 = \frac{N_\Sigma(1 - 10^{-b})}{10^{-b(M_{\min} - M_0)}} \frac{1}{1 - 10^{-(s+1)b}} = \alpha \frac{N_\Sigma}{1 - 10^{-(s+1)b}} \qquad (3)$$

$A_0 = 10^{a'}$ — сейсмическая активность для магнитуды Мо. N_Σ-общее число землетрясений с магнитудой $M \geq M_{\min}$ (при предположении постоянного шага магнитудных классов). Численный расчет показывает, что выражнием $1/(1-10^{-(s+1)}$, можно перенебречь при $b > 0{,}45$ и произвольном S.

Сейсмическая активность нормирована для времени $S_{норм}$ и площади $T_{норм}$. Карта сейсмической активности была составлена на основе параметров из следующей таблицы:

$Mo = 4$ $T_{norm} = 100$ лет			$S_{norm} = 1000$ км²	
	T_{obs}	b	M_{min}	$\alpha \frac{T_{norm}}{T_{obs}}$
Австрия	70	0,92	3,4	0,343
Венгрия	70	0,83	3,1	0,220
Чехия и Моравия	100	1,30	3,1	0,064
Словакия	70	0,75	3,1	0,248
Граничные области	56	0,92	4,1	1,942 (Альпы)
		0,83		1,732 (Карпаты)

На карте изображено несколько отдельных геологических систем: чешский массив, альпийская и карпатская системы и венгерская равнина. Этот факт вносит некоторые методические трудности. По этой причине для составления карты был использован метод «многократной» круговой палетки:

Предполагаем, что для расчета сейсмической активности в точке $P(\varphi, \lambda)$ влияние эпицентров тем больше, чем меньше эпицентральное расстояние. Отдельные эпицентры рассматриваются с весом:

$$W_i(\varphi, \lambda) = k \exp\left[-\Delta^2/(2\sigma^2)\right]$$

где i — порядковый номер землетрясения, k — постоянная, Δ — эпиц. расстояние, σ — произвольный параметр.

Тогда выражение (3) примет следующий вид:

$$A_0(\varphi, \lambda) = \frac{1 - 10^{-b}}{10^{-b(M_{min} - M_0)}} \frac{T_{norm}}{T_{obs}} \sum_{i=1}^{n} W_i(\varphi, \lambda)$$

n — число всех землетрясений в данной области.

Для практического расчета была применена шестикратная палетка с параметрами, приведенными в следующей таблице:

j	r_j(km)	s_j(km²)	w_j
0	0		
1	12,62	500	.8028
2	17,84	500	.5340
3	25,23	1000	.1964
4	30.90	1000	.0600
5	39,84	2000	.0247
6	56,42	5000	.0042

$$W_j = k \int_{r_{j-1}}^{r_j} e^{-\frac{\Delta^2}{2\sigma^2}} \, d_\Delta$$

$$\sigma = 13.22 \text{ km}$$

$$k = 1.61$$

Активность была рассчитана для квадратной сети 20 × 20 км взятой в каждой из рассмотренных областей. Активность в переходных зонах представлена суммированным ефектом всех соседних областей. Изолинии были построены при помощи квадратической интерполяции. На карте изображены изолинии $A_4 = 0.1, 0.3, 1.0, 3.0$. Изолинии окружают эпицентры и хорошо совпадают с главными тектоническими зонами.

Для окончательного варианта карты надо еще решить некоторые открытые вопросы, напр. изображение активности на переходе из одной геологической системы в другую, определение оптимальных параметров для расчета активности. Очень серьезным вопросом является также и вопрос продолжительности репрезентативного временного интервала.

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FIRST VERSION OF THE SEISMIC ACTIVITY MAP OF CZECHOSLOVAKIA, HUNGARY AND AUSTRIA

I. Broucek, D. Csomor, D. Prochaskowa

Abstract

This map provides an objective picture of the distribution of epicentres corresponding to one energy level and allows to establish which areas are of high seismicity, in relationship with the tectonic setting. The isolines $A_4 = 0.1, 0.3, 1.0, 3.0$ figure on the map. For the definitive version it is needed to solve some problems, e.g. how to represent the activity passing from one geological unit to another, which are the optimum parameters for computing the activity, and which would be the best representative time interval to be chosen.

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P-WAVE TELESEISMIC SPECTRA OF THE VRANCEA EARTHQUAKE OF MARCH 4, 1977

By

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Amplitude FOURIER spectra of deep-focus Vrancea earthquake ($MPV = 7.1$, $H = 120$ km) were calculated using P-wave records at teleseismic distances. The examination of P-wave forms especially recorded by short-period instruments and the shape of long-period part of spectra reveal that this earthquake is a complicated multiple event consisted of 3 shocks at least. Source parameters estimations obtained from P-wave spectra are: the seismic moment $Mo = 2.2 \cdot 10^{27}$ dyne · cm, fault length $-L = 228$ km.

Investigations of teleseismic spectra are treated for determining of dynamic source parameters: seismic moment Mo, source dimension L, stress drop $\Delta\sigma$ and average dislocation $\bar{u}$. For this aim we first examined the P-wave records of short-, middle- and long-period seismographs (A-, B- and C-types of standard instruments, respectively) at some seismic stations, at epicentral distances = 14—75°. The typical instrumental responses are shown in Fig. 1. The examples of P-wave seismogramms recorded by the vertical component of A-type instruments at stations MAK, ANR, FRU, ILT located at different distances are shown in Fig. 2. This complex wave form of the Vrancea earthquake is very different from typical deep events. There are complex multiple arrivals here which are not predicted by the travel-time curves, while most earthquakes have very simple wave forms. Usually the deep-focus earthquake has a sharp first short-period swing. Here the first swing is of a very little amplitude, then amplitudes of oscillations increase and reach the maximum value in about 15 sec after the first arrival. Thus, it is possible to pick three clear groups of P-waves at least within the mentioned time interval. The criteria for picking a later P-phase are an abrupt change in phase and a sudden substantial increase in amplitude.

The first group of P-waves (P_1) is characterized by a little amplitude. Only the short-period instrument with a large magnification was able to record it clearly (for example ILT, FRU with magnification factor V of about $2.5 \cdot 10^4$).

This group of P-waves was recorded more weakly by instruments with a lesser magnification (as ANR with $V = 10^4$ and MAK with $V = 0.25 \cdot 10^4$).

The second group of P-waves (P_2) arrived in 4—5 sec after the first one. It is picked on the base of amplitude and frequency criteria. The first oscilla-

tions of P_2 group have interference character with predominant period of about 2—3 sec. Initial amplitudes of P_2 are three times more than the P_1 one, then they increase up to 20—30 times more than the P_1 one.

The third group of P-waves (P_3) arrived abruptly in 9—13 sec. after P_2 and its amplitudes are 2—3 times more than the latter with periods of 1.5—2 sec.

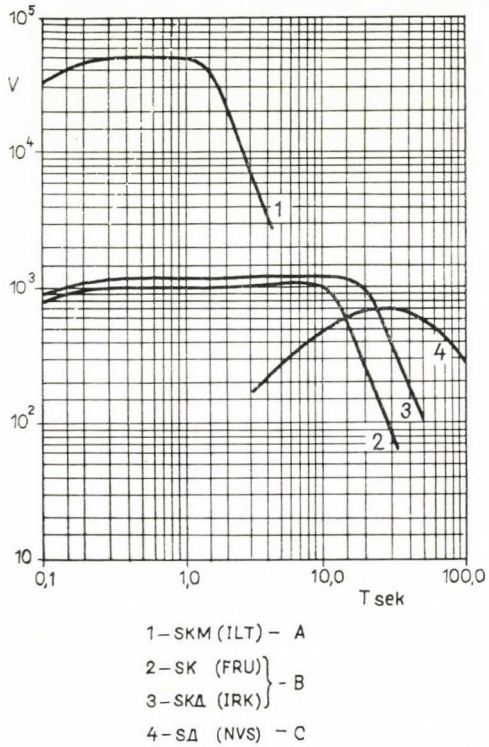

Fig. 1. Magnification curves of different seismographs: short-period (SKM), middle-period (SK, SKD), long-period (SD)

In about 24—28 sec and 32—43 sec after the P_1-arrival there are some other P-phases on seismogramms. Two phases of them can be interpreted as reflections from the earth's surface of the type pP and sP for a deep-focus earthquake ($H = 120$ km). Their periods are near to the P_3 one but amplitudes are lesser than those of P_3. Thus the amplitudes of P_3 are maximum in all P-wave groups.

In Fig. 3 P-wave seismogramms of B-type instruments (FRU, ILT) are presented. Here the first weak arrival of P, the P_1-phase, is usually absent and the P_2-phase is a very beginning of record. Its period is about 4 sec. In 9—13 sec after P_2 the phase — P_3 arrives. This phase has maximum amplitudes and periods of 18 sec and superimposed with periods of 2—3 sec. It is difficult to pick sP- and pP-phases approximately as they are located within oscillation

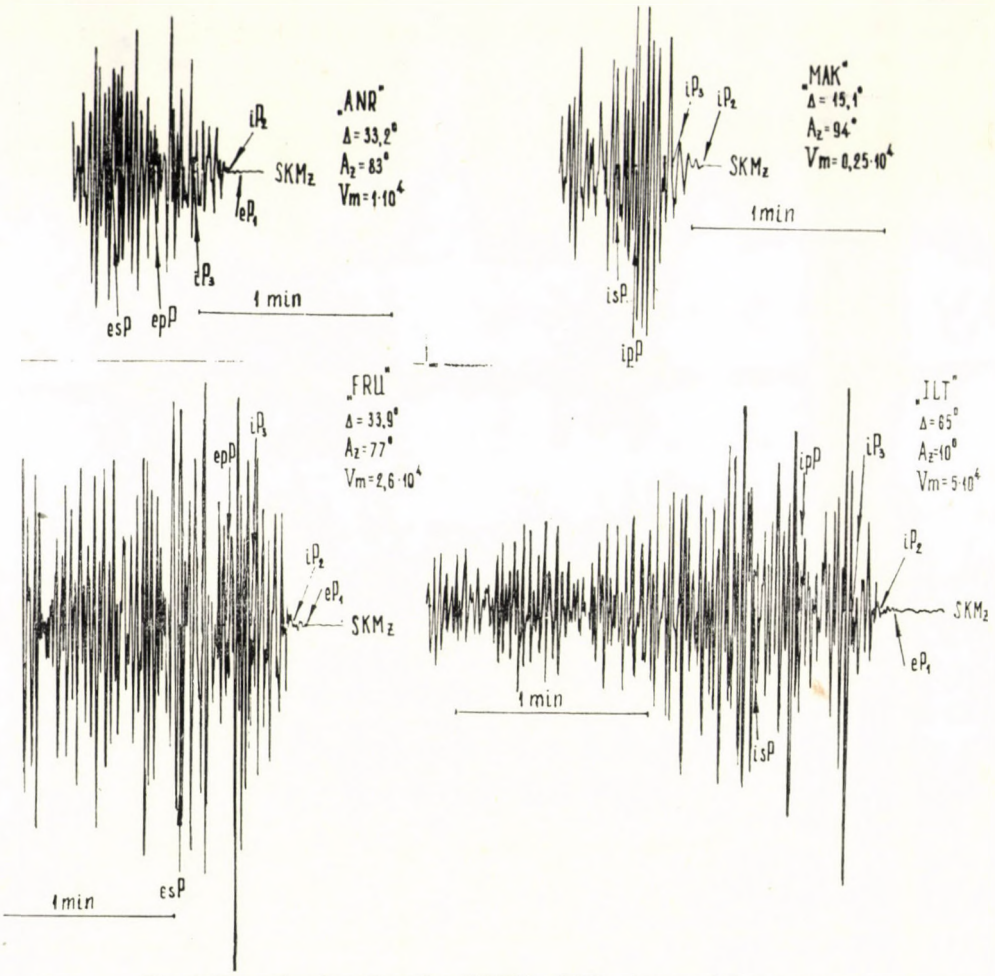

Fig. 2. Records of earthquake of III/4—1977; short-period instruments

intervals of maximum amplitudes. On *P*-wave records of C-type instruments (Fig. 3, NVS) the P_1-arrivals are also not picked. The first sharp arrival is P_2-phase, which interferes with P_3-phase. The initial part of this interference wave has a period of 25 sec, *sP*- and *pP*-phases are not approximately picked. As in the case of B-type instrument records these phases correspond to a time interval of maximum amplitudes of P_3-group with a period of 50 sec.

Considering the noted 3 groups of *P*-waves as distinct shocks, we calculated their magnitudes MPV on the basis of short-period records. These values are 5.0, 6.4 and 7.1, respectively.

To estimate the seismic moment and fault dimensions of earthquake with MPV = 7.0 it is needed to use the *P*-wave spectra in their long-period part. For this reason the *P*-records of longest period instruments of 8 stations were

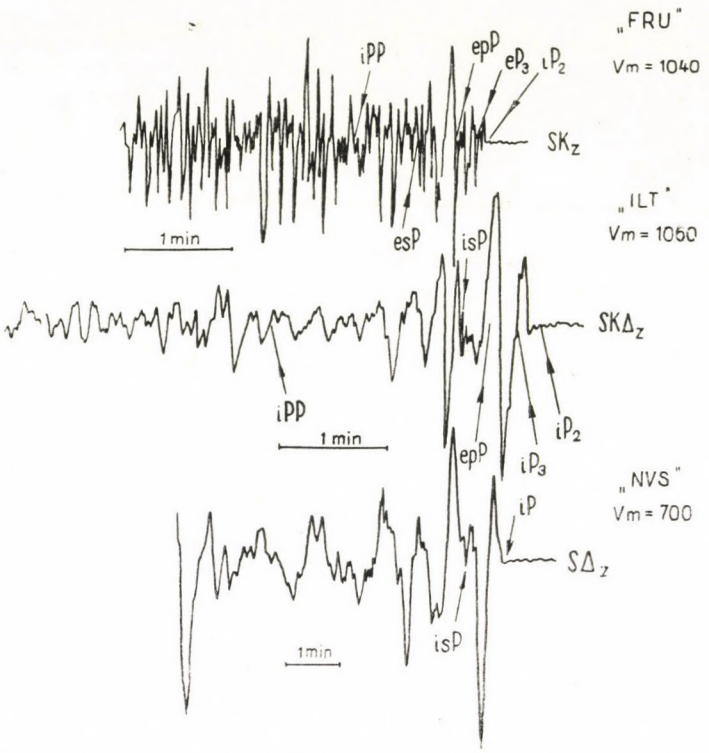

Fig. 3. Records of earthquake of III/4—1977; middle- and long-period instruments

chosen. These records from the first arrival of *P* to those of *PP* these records as a rule were digitized at 0.2—0.4 sec sample intervals and then Fourier-analyzed. The lengths of digitized intervals of *P*-waves were 27—151 sec and corresponded to epicentral distances. There are not only *P*-waves within these intervals, but *pP*- and *sP*-phases mentioned above at distances less than 60 degrees and *PP*-, *sP*- and *PcP*-phases at more distances, are here. These phases can change the form and level of spectra, as compared to those of clear *P*-waves and we must take this into account during the source parameter determinations.

The data of seismic stations, instruments, epicentral distances, azimuths and digitized intervals of *P*-records are presented in Table 1.

Experimental *P*-wave spectra corrected for instruments and attenuation of *P*-waves were approximated in accordance with the spectral low by AKI (1967) to find the spectral parameters Ω_0 (the long-period level) and f_0 (the corner frequency), shown in Table 2.

Determination of seismic moment was made by KEILIS-BOROK (1960), HANKS, WYSS (1972):

$$M_0 = \frac{\Omega_0 \cdot 4\pi\varrho\alpha^3 \cdot R}{R_{\theta\varphi} \cdot c(\omega)} \tag{1}$$

where α is the P-wave velocity in focal region (in this case 8.0 km/sec); ϱ is density near the source ($\varrho = 3{,}37$ g/cm^3); $R_{\theta\varphi}$ is the radiation pattern of P-wave ($R_{\theta\varphi} = 0.4$); $c(\omega)$ is the crust response under station ($c(\omega) = 2$) and R—the geometrical spreading factor as a function of epicentral distance, focal depth and earth structure—was calculated with a help of tables by SHIMSHONI and BEN-MENAHEM (1970) and KOGAN (1975).

Table 1

Station date

No.	Station		Distance (deg)	Station azimuth (deg)	Instrument	$\delta\tau$ sec
1.	Apatity	APA	22.1	8	SKD	27
2.	Novosibirsk	NVS	36.2	54	SD	100
3.	Chagan-Usun	CUR	40.3	60	SKD	84
4.	Ust-Elegest	UEI	43.3	57	SKD	76
5.	Irkutsk	IRK	48.9	53	SKD	74
6.	Zakamensk	ZAK	49.2	56	SKD	118
7.	Iultin	ILT	65.0	10	SKD	148
8.	Montreal	MNT	66.1	310	COL	151
9.	Kurilsk	KUR	75.5	39	SKD	97

Table 2

Spectral and source parameters

Stations	APA	NVS	CUR	UEI	IRK	ZAK	ILT	MNT	KUR
$\Omega_{01} \cdot 10^{-2}$cm · sec		12	8.5	8.0	9.1	5.6	5.8	7.2	3.5
$\Omega_{02} \cdot 10^{-2}$cm · sec	12	5.6	3.3	5.0	5.0	6.3	3.8	3.3	3.5
$f_{01} \cdot 10^{-2}$ Hz		2.5	2.5	2.2	2.2	2.7	3.4	3.3	2.5
$f_{02} \cdot 10^{-2}$ Hz	7.3	5.6	5.3	5.0	5.0	5.2	7.4	7.4	7.3
L, km		238	238	270	270	220	176	180	238
$M_0 \cdot 10^{27}$dyne · cm		3.2	2.5	2.3	2.7	1.7	2.1	2.6	1.5

The fault length L was determined in accordance with the model by BRUNE (1970).

$$L = 2r = \frac{2.34 \cdot \alpha}{\pi f_0} \tag{2}$$

where r is a radius of circular dislocation in BRUNE's model.

The stress drop $\Delta\sigma$ and average displacement $\bar{u}$ are calculated from the expressions:

$$\Delta\sigma = \frac{7}{16} \cdot \frac{M_0}{r^3} \tag{3}$$

$$\bar{u} = \frac{M_0}{\mu \cdot A} \tag{4}$$

where μ is a shear modulus, $A = \pi \cdot r^2$ is the fault surface.

The spectra are calculated on the basis of records of 9 stations. Corresponding values of Ω_0, f_0, Mo, L, are presented in Table 2.

The examples of experimental spectra are shown in Figs 4 and 5. All these spectra resemble much each other. In the long-period part ($T > > 15$ sec) they have two maxima of cupolashape, in the short-period part spectral amplitudes decrease sharply with increasing frequency. The periods of these two noted maxima are near to 50 sec and 25 sec, respectively. As a rule, an amplitude of the first maximum is more than that of the second (NVS, CUR, UEI, IRK, MNT, ILT), sometimes these amplitudes are equal (KUR) or the second amplitude is more than the first one (ZAK).

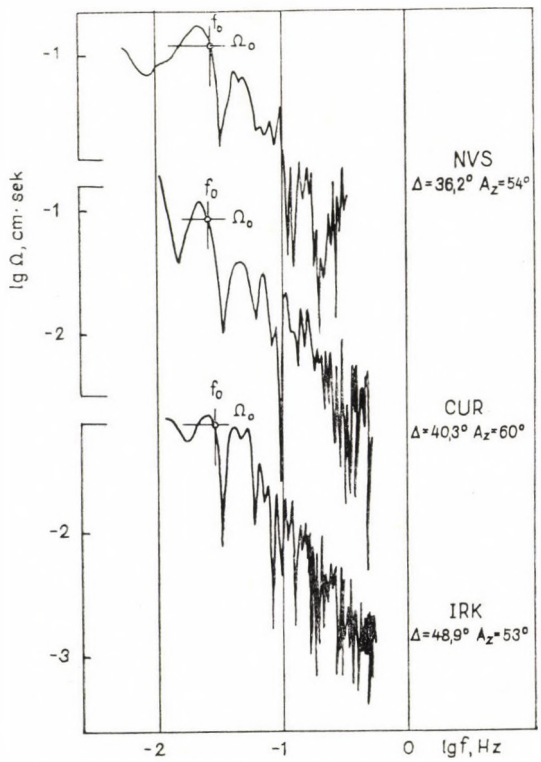

Fig. 4. P-wave spectra corrected for instrumental response

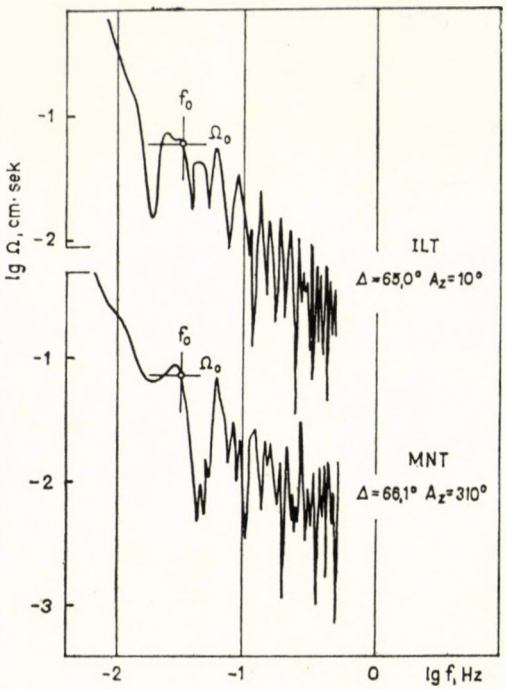

Fig. 5. **P-wave spectra corrected for instrumental response**

Spectrum of APA has only second long-period maximum ($T \sim 20$ sec); the first maximum ($T \sim 50$ sec) is not distinguished. It is connected with a little length of digitized interval of *P*-record ($\delta t = 27$ sec). The positions of these maxima don't depend on the analyzed interval length and epicentral distance, but on the epicenter-station azimuth.

The presence of two long-period maxima of *P*-wave spectra according to the model investigations of VINOGRADOV (1978) can be considered as evidence of complicated rupture propagation of stick-slip type. Thus, the results of records and spectra of *P*-waves examination give a possibility to suggest that the Romanian earthquake of March 4, 1977, was a multiple event.

Spectral parameters (Ω_{01}, Ω_{02}, f_{01}, f_{02}) were determined for all spectra except of spectrum of APA for which only Ω_{02}, f_{02} were obtained. Estimations of source parameters were made by using only values Ω_{01}, f_{01} (Table 2). Individual values of source parameters ***Mo***, ***L*** are different very little, their average values are: $Mo = (2.3 \pm 0.6)\ 10^{27}$ dyn. cm; $L = 228 \pm 36$ km. Using average values of ***Mo*** and ***L*** parameters, $\Delta\sigma$ and u were calculated by (3.4), $\Delta\sigma = 0.7$ bar; $u = 20$ cm.

It is interesting to compare these parameters with those for earthquakes of analogous magnitude and focal depth. For the Kamchatka earthquake of

November 24, 1971 with $M = 7.2$ and H = 125 km, similar data were obtained by ZAKHAROVA, STAROVOIT, CHEPKUNAS (1977) who used observations of a single station: $Mo = 2.6 \cdot 10^{27}$ dyne · cm, $L = 152$ km. It is shown that the values of seismic moments and fault dimensions for both earthquakes compared are approximately the same.

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ТЕЛЕСЕЙСМИЧЕСКИЕ СПЕКТРЫ ВОЛН «P» ЗЕМЛЕТРЯСЕНИЯ ВРАНЧА, 4-ОГО МАРТА 1977 Г.

А. И. ЗАХАРОВА, Л. С. ЧЕПКУНАС

Резюме

Амплитудные фурье-спектры глубоко-фокусного землетрясения Вранча (MPV = = 7,1, H = 120 км) были рассчитаны по воли «P» регистрированных на телесейсмических расстояниях. Изучение формы волн «P» специально регистрированных коротко-периодными приборами и также облик длинно-периодными части спектра указывают на то, что данное землетрясение P-воли было сложным событнем состоявшим из не менее 3 ударов. Оценки 5 параметров источника, полученные при помощи спектров воли «P»: сейсмический момент Mo = $2{,}3 \cdot 10^{27}$ dyne. см, длина разлома L = 228 км.

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EARTHQUAKE EPICENTRES AND MACROSEISMIC INTENSITY IN THE CARPATHO-BALKAN REGION

By

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The investigated region covers the territories of Bulgaria, Romania, SW Ukraine, Moldavia, Hungary, Czechoslovakia and parts of Austria, Yugoslavia and Greece. Maps of epicentres and of the largest observed intensities demonstrate the geographical distribution of activity and the corresponding macroseismic effects. Earthquakes occur in several seismogenic belts (provinces) related to neotectonic features of the region. A special phenomenon is the intermediate-depth focus of Vrancea releasing in the region the largest amounts of energy together with a few crustal foci in Bulgaria.

1. Introduction

This contribution is the result of activities of the Working Group 4.3.1.3 of the Commission of Academies of Sciences of Socialist Countries for Planetary Geophysical Research (K.A.P.G.) in preparing basic seismological information aimed at the compilation of a seismic zoning map.

According to the recommendation of the Jena session of the Working Group 4.3 on April 3, 1974, the Geophysical Institute of the Czechoslovak Academy of Sciences in Prague, was entrusted with the preparation of the Catalogue of Earthquakes and of the Atlas of Isoseismals for Central and Eastern Europe. The investigated region includes the territories of Bulgaria, Czechoslovakia, the German Democratic Republic, Hungary, Poland, Romania and the SW part of the Soviet Union, defined by the geographic coordinates (Fig. 1): 55 °N — 30 °E, 41° — 30°, 41° — 22°, 44° — 22°, 44° — 20°, 45 — 20°, 45° — 16°, 48° — 10°, 53 °N — 10 °E. This region includes also parts of Austria, of the German Federal Republic, of Yugoslavia and Greece.

2. Catalogue of Earthquakes

It was recommended that all earthquakes which occurred prior to January 1, 1973 should be included in the catalogue. The first version of the *Catalogue of Earthquakes for Central and Eastern Europe* (1979) was compiled by the coordinators Schenková and Kárník using the following sources:

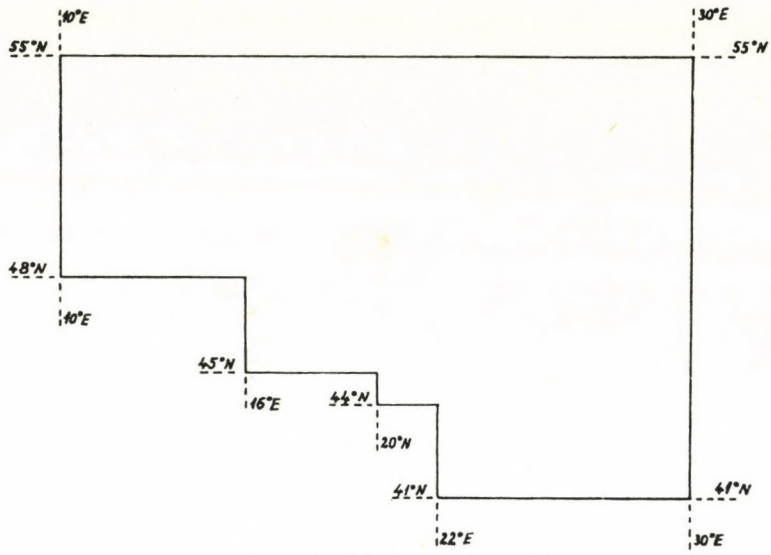


Fig. 1. Map of the investigated area

1. Catalogue of European earthquakes 1801—1955 (KÁRNÍK, 1968 and 1971) and KÁRNÍK's unpublished list of parameters for the period 1956—1969 based particularly on B.C.I.S. data and regular I.S.C. bulletins from 1970—1972.
2. National catalogues compiled by GRIGOROVA for Bulgaria, BROUČEK, PROCHÁZKOVÁ and SCHENKOVÁ for Czechoslovakia, KUNZE and MAAZ for the German Democratic Republic, CSOMOR for Hungary, GUTERCH and LEWANDOWSKA—MARCINIAK for Poland, RADU for Romania, EVSEEV, KONDORSKAYA, KOSTYUK, MOSKALENKO, ROMAN, SAGALOVA, SHEBALIN for the SW part of the Soviet Union, DRIMMEL for Austria, FÖRTSCH and SCHMEDES for the German Federal Republic.
3. the *Unesco Catalogue of Earthquakes of the Balkan Region* (1974)
4. I.S.C. and B.C.I.S. summaries.
5. Station bulletins.

The list of the basic earthquake parameters in the Catalogue follows the form of the KÁRNÍK's catalogue of European earthquakes (1968). In order to fulfil the condition of homogeneity in space, time and classification of the seismological information, the following lower limits of intensity (scale MSK-64) and magnitude were adopted:

$I_0 \geq$ VIII for all historical earthquakes prior to the year 1800
$I_0 \geq$ VII for the period 1801—1900
$I_0 \geq$ V or $M \geq 4$ for the period 1901—1972.

The Catalogue has also an Appendix listing weak shocks which, though being incomplete, may provide additional information for seismological analysis.

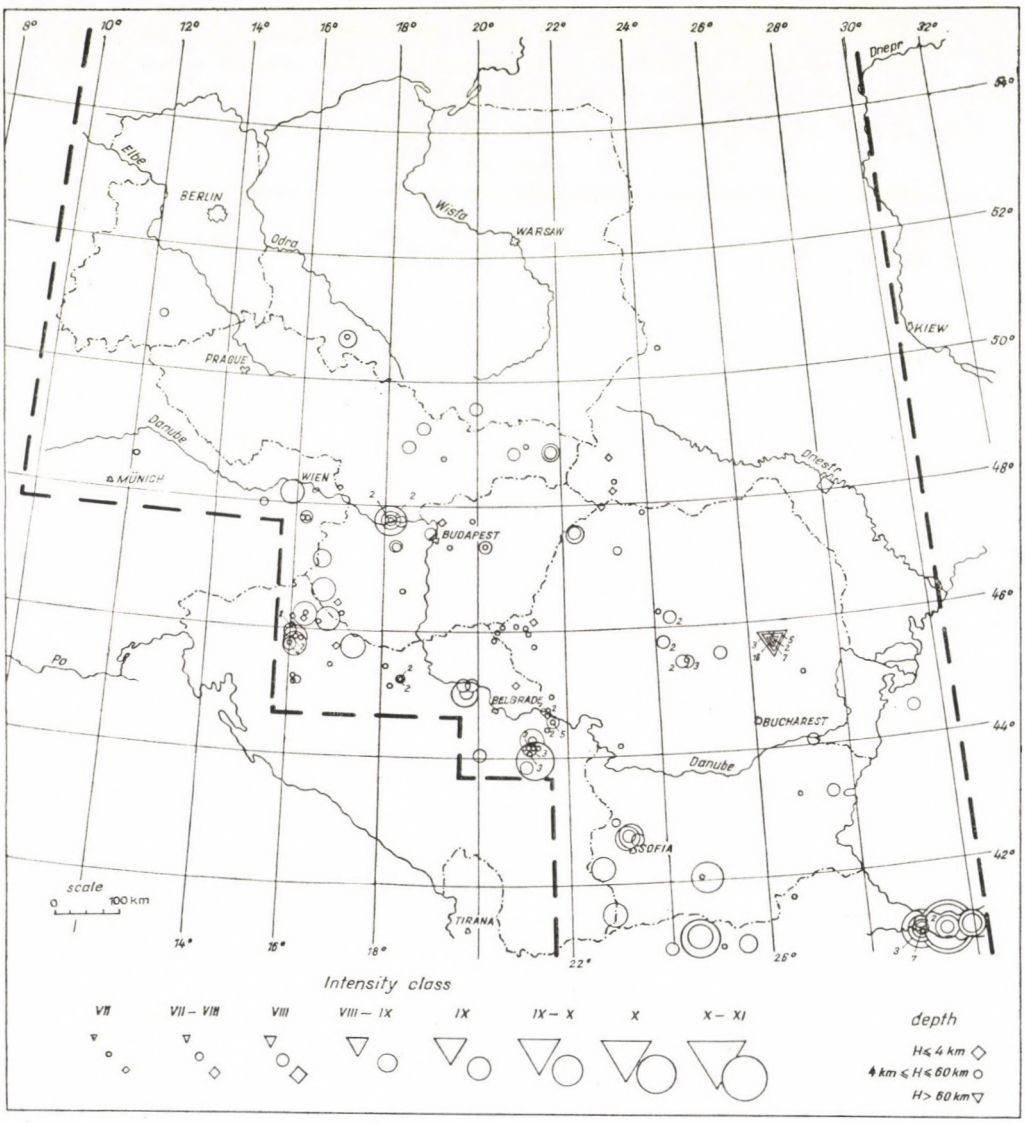
All the dates in the Catalogue follow the Gregorian calendar. The time of the origin is given in the Greenwich Mean Time. Date, time of origin, epicentre co-ordinates and intensity given by the national contributors were accepted as final parameters. Macroseismic epicentres defined on the basis of available maps of isoseismals were preferred to the instrumental ones. The classification of the accuracy of the coordinates was performed by national specialists. Radii of isoseismals were determined whenever possible. Focal depths were estimated using the formula of Kövesligethy for all earthquakes for which at least 3 isoseismals are available. The formula of Blake was applied if only one or two radii of isoseismals were determined. At present, earthquake magnitudes are being unified using amplitudes and periods of S_g and LR reported by stations. If these data are not available the magnitudes are determined by conversion from macroseismic parameters. The Catalogue will be handed over to the printers in November 1979.

3. Epicentre Maps

The epicentre map is the first basic product resulting from the Catalogue; its preparation is also recommended by the K.A.P.G. project. Two draft maps of epicentres on a scale of 1 : 2.5 million, based on the data from the main part of the *Catalogue of Earthquakes for Central and Eastern Europe* (1979) were compiled by SCHENKOVÁ and SCHENK to display the distribution of earthquake activity in the investigated area.

The first map (Fig. 2) contains epicentres of historical earthquakes with $I \geq$ VIII prior to 1800 and $I \geq$ VII for the period 1801—1900 in accordance with the limits of homogeneity of the Catalogue. The map distinguishes intensity classes (from intensity VII to X—XI) expressed in terms of the MSK-64 scale and three classes of the focal depth. Each intensity and depth class is appropriately marked in the map. Circles denote shallow foci within the Earth's crust, $4 < h \leq 60$ km, triangles intermediate depth foci between 60 and 300 km and squares denote very shallow foci located presumably in sedimentary layers ($1 < h \leq 4$ km). The total number of earthquakes in one point, belonging to a certain magnitude class, is indicated by a number at the corresponding mark.

There are large differences in the level of earthquake activity over the investigated area. It is also evident that the 20th century map cannot give a full idea about the position of earthquake foci, particularly in the northwestern part of the area. There were damaging earthquakes in some sections of the Carpathian system prior to 1900 and very low or no activity in the 20th cen-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VIII – IX
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IX – X
X
X – XI
depth
H ≤ 4 km
4 km ≤ H ≤ 60 km
H > 60 km

Fig. 2

tury. Only the combination of both maps can provide a better insight into the problem of delineation of active earthquake regions. For the historical period there are some events of doubtful epicentral location; e.g. the reports on damaging shocks during 1441—1443 indicate the possibility of two foci, one in Lower Silesia and one in Central Slovakia. However, only the focus in Silesia is marked in the map because it was determined by the Polish contributors, in Slovakia the epicentre has been left out because of large uncertainty about the position of the epicentre, the number of reports is too low. Reliable location of these strong earthquakes during 1441—1443 remains therefore open. Another difficulty is presented by large single earthquakes, e.g. west of Vienna or near Komárno which demonstrate one outburst of seismic energy in the history and do not permit to estimate any recurrence interval. By comparing both maps one observes a relative lack of historical epicentres in Bulgaria compared with high activity of the 20th century. This absence of data may be partly caused by incomplete reports.

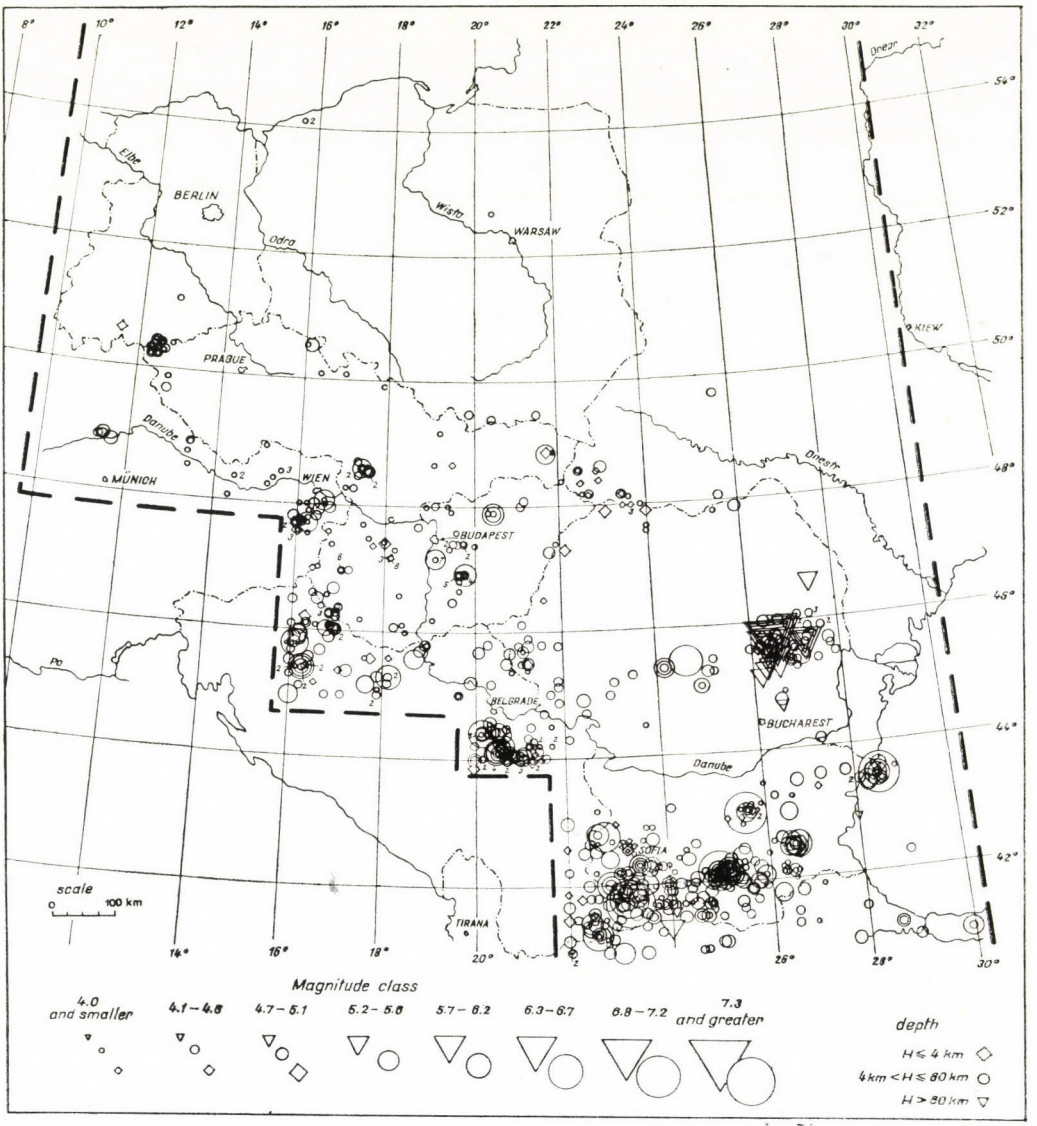
The second map (Fig. 3) contains epicentres plotted for the period 1901—1972 and $M \geq 4$ or $I \geq$ V; the quantity magnitude was chosen as the basic measure of the size of earthquakes in the map. The following magnitude classes with a step of 0.5 were used:

$$\geq 4.0,\ 4.1 - 4.6,\ 4.7 - 5.1,\ \ldots,\ 7.8 - 8.2\ .$$

Each class is differently marked in the map, as well as class of the depth of focus. The magnitudes based on surface waves M_{LH} were taken as representatives for shallow earthquakes and those based on body waves ($M_B = m$) for intermediate ones, respectively. It must be noted that in regions of high density of epicentres some low-magnitude shocks (e.g. in Vrancea and Bulgaria) cannot be simply plotted on the map, these omissions, however, do not influence the pattern.

Combining the distribution of epicentres in both maps first, of all the striking difference in the activity between the Hercynian Bohemian Massif and the Alpine Carpathian and Rodope systems is evident. In the investigated area two source regions of the highest seismic energy release exist: the Struma valley with the largest European shallow shock of the 20th century (April 4, 1904, $M = 7\,3/4$) and the Vrancea intermediate depth focus with several earthquakes of $m = 7$—7.5 occurring in average twice during a century, the last one being that of March 4, 1977. Large shallow earthquakes occurred also in Central and Eastern Bulgaria during the first 30 years of the 20th century and at the boundary of the Pannonian Basin, if we consider the whole observation period of six to seven centuries.

Some epicentres follow distinct earthquake belts (Fig. 4) which are well observed if we include also epicentres from neighbouring areas. One of the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Danube
SOFIA
TIRANA
scale
100 km
Magnitude class
4.0 and smaller
4.1—4.6
4.7—5.1
5.2—5.6
5.7—6.2
6.3—6.7
6.8—7.2
7.3 and greater
depth
H ≤ 4 km
4 km < H ≤ 80 km
H > 80 km

Fig. 3

belts coincides with to very distinct zone extending from Friuli to Vienna and following further the trend of the West Carpathians. This belt can be followed also to the western section of the Eastern Carpathians. There is, however, an activity gap in the central part of the Carpathians north of Vrancea. The Vardar zone as a seismotectonic belt of first order can be extended with some imagination in the direction to Budapest, a transversal belt enters our area from Yugoslavia towards the Balaton lake. The relation of earthquake activity with main tectonic lineaments is evident. We marked in Figure 4 the observed earthquake belts and it is now the next task to investigate the seismogeological links more closely. The first study of this kind has been carried out by BORISSOFF, REISNER and SHOLPO (1978) within the programme K.A.P.G. 4.3.1.2. It should be followed by similar studies in the participating countries.

4. Atlas of Isoseismal Maps

The *Atlas of Isoseismal Maps for Central and Eastern Europe* (editors PROCHÁZKOVÁ, KÁRNÍK, 1978) was prepared as a data basis for a uniform evaluation of seismic activity. The Atlas contains all available isoseismal maps compiled by GRIGOROVA and GLAVCHEVA for earthquakes with epicentres on the territories of Bulgaria, BROUČEK, PROCHÁZKOVÁ and ZÁTOPEK for Czechoslovakia, KUNZE for the German Democratic Republic, CSOMOR for Hungary, GUTERCH and LEWANDOWSKA-MARCINIAK for Poland, KOSTYUK and SHEBALIN for the SW part of the Soviet Union. In addition, the isoseismal maps of earthquakes with epicentres in the neighbouring countries are included if their macroseismic effects were observed also in the investigated region, e.g. maps compiled by DRIMMEL, TRAPP, LUKESCHITZ and PROCHÁZKOVÁ for Austria and the German Federal Republic have been inserted into the volume. The maps for Bulgaria and Romania already published by the UNESCO Balkan Project in the *Atlas of Isoseismal Maps—Balkan Region* (1974) were not included. The K.A.P.G. Atlas covers the historical period until 1973. The collection is not complete, isoseismal maps are missing for the events for which sufficient data are not available; the Atlas contains 316 isoseismal maps.

5. Depths of Foci and Intensity Attenuation

The isoseismal maps from the K.A.P.G. and UNESCO Atlases were used for the determination of depths of foci h and absorption coefficients α using the formula by KÖVESLIGETHY (SPONHEUER, 1960). The results are summarized in the Figs 5 and 6, the numbers are based on at least five depth determinations in one place, isolated or small numbers of determinations are not shown

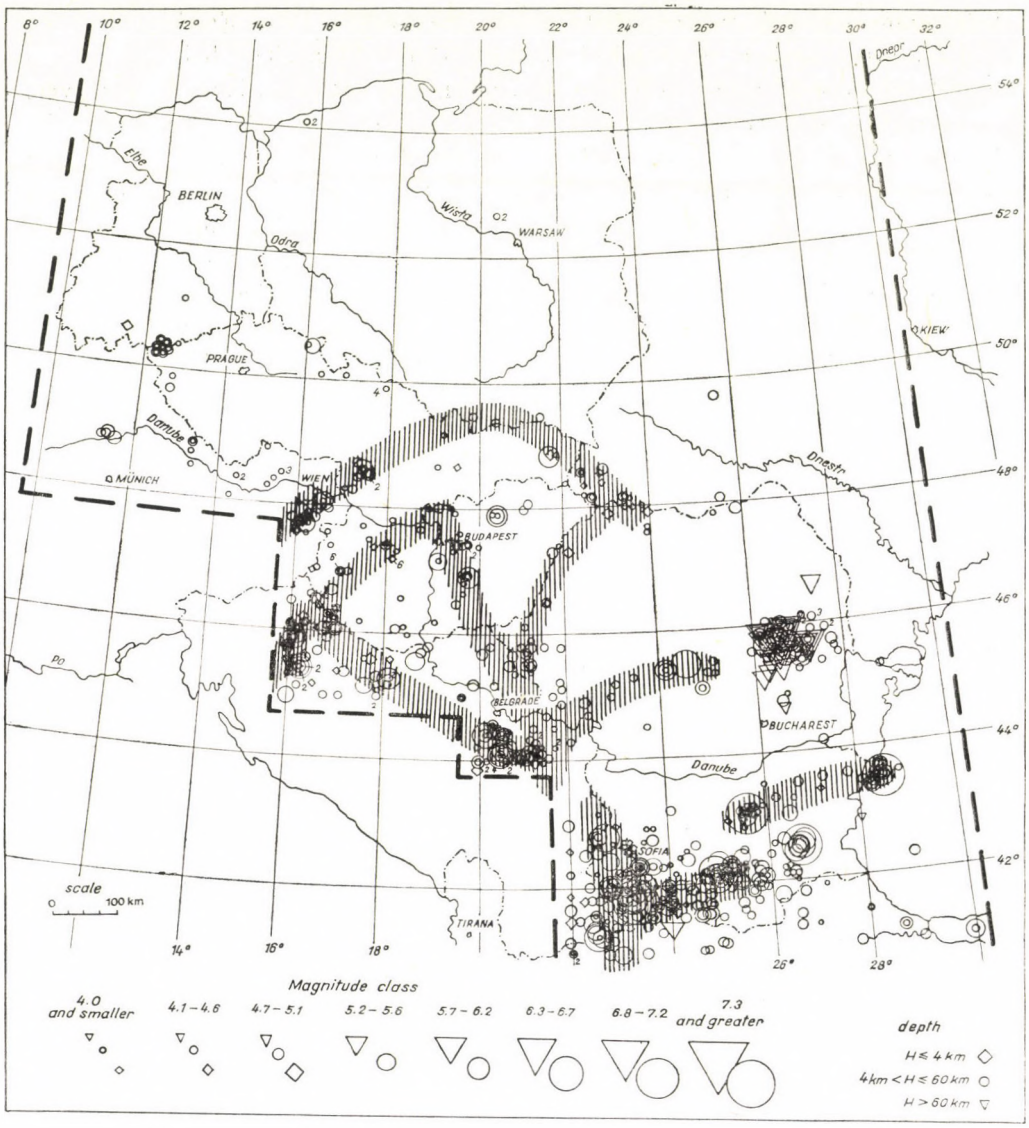

Fig. 4. Scheme of earthquake belts

in the maps. It is seen from Fig. 5 that all investigated foci, except those of Vrancea, are within the Earth crust, the majority of them is located in the upper crust. Very shallow foci are typical for Vogtland (swarms, 4—14 km), the Little Carpathians (4—15 km), Eastern Slovakia (4—12 km), Ruthenia

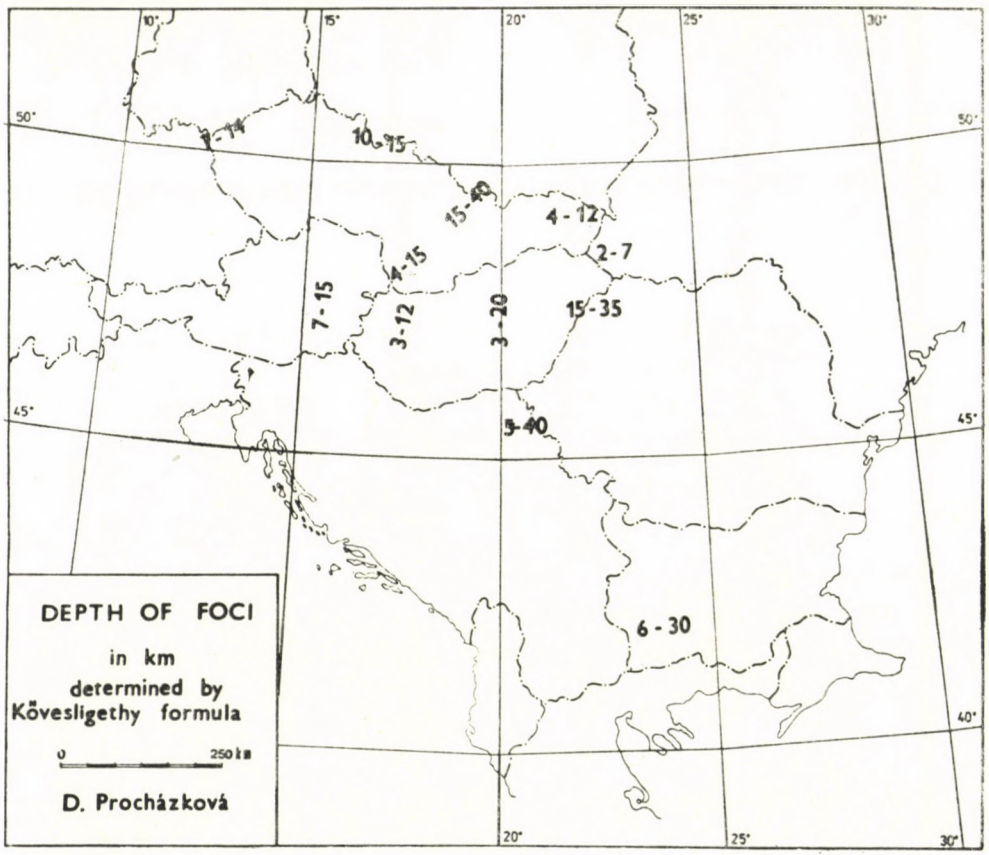

Fig. 5

(2—7 km), some of them are also found in the Pannonian Basin (3—12 km), in the Giant Mts (10—15 km) and in the Orlické hory Mts (10—14 km). The "seismic active layer", i.e. the layer comprising earthquake foci, is thicker in the northern part of the Western Carpathians, in the eastern boundary of the Pannonian Basin and in the Rodope Mts.

Figure 6 demonstrates first of all the difference between the small absorption in the Bohemian Massif and much larger absorption of seismic energy propagating into the Carpathian system. This fact is well known from the asymmetric shape of isoseismal maps of all East-Alpine earthquakes (*Atlas of*

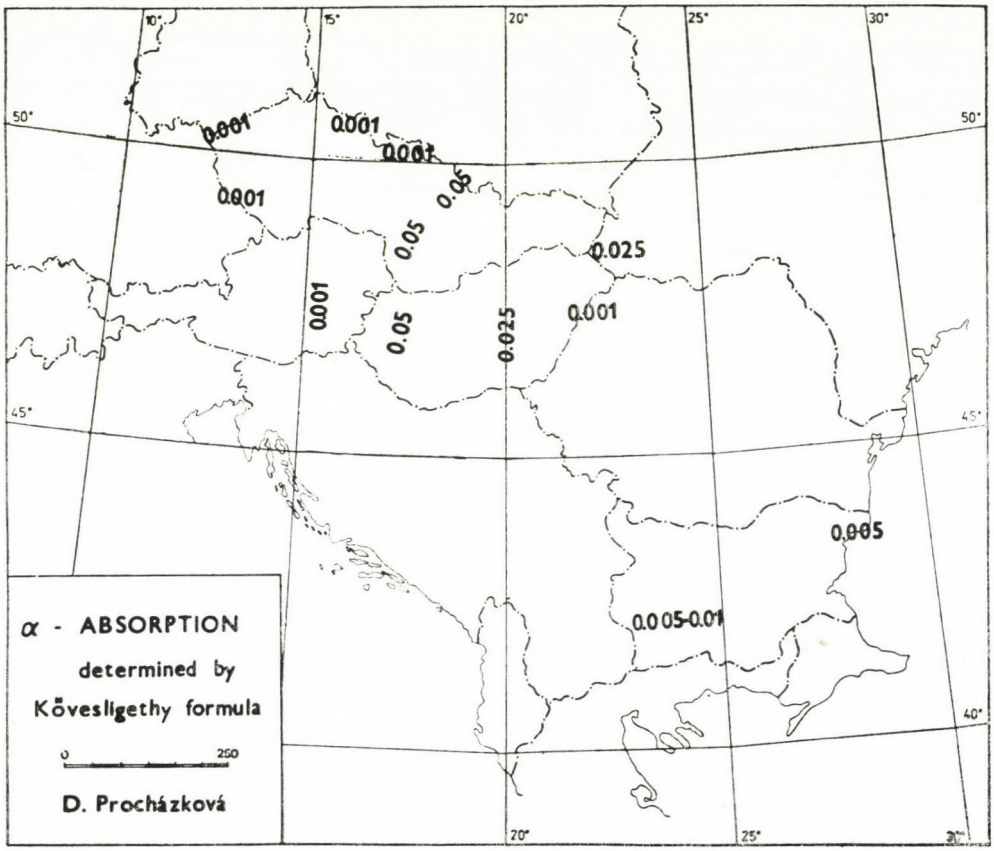

Fig. 6

Isoseismal Maps for Central and Eastern Europe, 1978). The following numerical values were obtained: the Bohemian Massif $\alpha = 0.001$, the Carpathian Mts $\alpha = 0.05$, SW part of the USSR $\alpha = 0.025$, western part of Hungary $\alpha = 0.05$ and central part of Hungary $\alpha = 0.025$.

6. Map of Maximum Observed Intensity

The draft map of maximum observed intensity (Fig. 7) has been compiled by Procházková on the scale of 1 : 2.5 million. Again, as it was the case with the epicentre maps, the *International Tectonic Map of Europe and Adjacent Areas*, 1/ 2 500 000 (1973) was used as a base in order to facilitate the comparison between geological and seismological phenomena. In our definition the isoseismal of a certain degree involves the area where the corresponding intensity


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Wisła
WARSAW
KIEW
PRAGUE
MUNICH
WIEN
BUDAPEST
Po
BELGRADE
BUCHAREST
Danube
SOFIA
scale
0 100 km
Intensity
10 9 8 7 6 5 4

Fig. 7

prevails. Isolated higher- or lower-intensity values are also marked, single isolated values were neglected. Relative increase in intensity values might indicate differences in the geological structure.

The presented map of maximum observed intensity was compiled from the following maps:

1. Map of maximum observed intensity of Czechoslovakia, Poland Hungary and Austria; scale 1 : 1 million (PROCHÁZKOVÁ, BROUČEK, GUTERCH, LEWANDOWSKA, CSOMOR and DRIMMEL, 1977).
2. Map of maximum observed intensities of Yugoslavia; scale 1 : 1 million, period 1801—1900, 1901—1970 (BAŠOVIĆ, CVIJANOVIĆ, HADŽIEVSKI, JANKOVIĆ, METOVIĆ, RIBARIĆ, SKOKO and VUČINIĆ, 1974). Reliable macroseismic data for the period prior to 1900 were also included. This map is the result of the synthesis of isoseismal maps of strong earthquakes with $I_0 \geq$ VII.
3. Map of maximum observed intensity in the Balkan region (SULSTAROVA, KOCIAJ, GRIGOROVA, CHRISTOSKOV, GALANOPOULOS, DRAKOPOULOS, DELIBASIS, CONSTANTINESCU, CORNEA, LĂZĂRESCU, RADU, SIPAHIOĞLU, BAŠOVIĆ, CVIJANOVIĆ, HADŽIEVSKI, JANKOVIĆ, METOVIĆ, RIBARIĆ, and VUČINIĆ, 1974). The map was originally prepared within the UNESCO Balkan Project in the scale 1 : 1 million, for the period 1600—1970. It was compiled on the basis of three main sources: *a)* generalized isoseismals from 1901—1970, *b)* theoretical isoseismals based on magnitude, focal depth and epicentral intensity for the large shocks lacking isoseismal maps, *c)* isoseismals of strong historical earthquakes from 1600—1900.
4. Map of maximum observed intensity in Albania during 1800—1970 (SULSTAROVA, KOÇIAJ, 1974); scale 1 : 1 million.
5. Map of maximum observed intensities in Bulgaria in the period 1800—1970 (GRIGOROVA, CHRISTOSKOV, 1974); scale 1 : 1 million.

There were no serious difficulties in drawing isoseismals of the degrees VII and VIII and higher. However, the isoseismals of the degrees V and VI are marked by a dashed line in regions with insufficient number of data. There were problems in connecting some isoseismals at the frontiers of Austria, Yugoslavia, Hungary, Romania, the Soviet Union, the German Democratic Republic and the German Federal Republic. Thirty individual isoseismal maps (Tab. 1) of the largest shocks were used to resolve the discrepancies. For the course of isoseismals on the boundaries of Austria, Italy and Yugoslavia the strongest quake of May 6, 1976, in the region of Friuli was decisive (KÁRNÍK et al., 1977).

Before discussing the macroseismic field it must be emphasized that the map contains only macroseismic intensities observed until the end of 1973. At this stage no attempt was made to complete the data by constructing theoretical isoseismals for epicentres without observed intensities. This remark relates mainly to medium or even large earthquakes prior to 1900. It can

Table 1

Earthquake	Author	Reference
1903 Mar. 5, 20^h37^m GMT, 50.4°N, 12.4°E, $I_0 = 7$	PROCHÁZKOVÁ, KUNZE	KAPG Atlas
1915 June 2, 02^h33^m GMT, 48.9°N, 11.4°E, $I_0 = 6$	PROCHÁZKOVÁ	KAPG Atlas
1915 Oct. 10, 03^h50^m GMT, 48.8°N, 11.6°E, $I_0 = 7$	PROCHÁZKOVÁ	KAPG Atlas
1872 Mar. 6, 14^h55^m GMT, 50.8°N, 10.3°E, $I_0 = 8$	KUNZE	KAPG Atlas
1875 Aug. 15, 15^h45^m GMT, 50.3°N, 24.2°E, $I_0 = 7$	KOSTYUK, SHEBALIN	KAPG Atlas
1926 June 28, 20^h00^m GMT, 48.02°N, 23.11°E, $I_0 = 6$	KOSTYUK, SHEBALIN	KAPG Atlas
1926 Aug. 10, 01^h10^m GMT, 48.02°N, 23.7°E, $I_0 = 7$	KOSTYUK, SHEBALIN	KAPG Atlas
1931 Apr. 7, 01^h42^m GMT, 48.22°N, 22.69°E, $I_0 = 6$	KOSTYUK, SHEBALIN	KAPG Atlas
1935 Aug. 22, 02^h10^m GMT, 48.2°N, 23.52°E, $I_0 = 5.5$	KOSTYUK, SHEBALIN	KAPG Atlas
1893 Mar. 11, 09^h25^m GMT, 47.98°N, 23.05°E, $I_0 = 7$	KOSTYUK, SHEBALIN	KAPG Atlas
1936 July 29, 06^h00^m GMT, 48.62°N, 22.46°E, $I_0 = 5$	KOSTYUK, SHEBALIN	KAPG Atlas
1936 Aug. 2, 20^h27^m GMT, 48.61°N, 23.53°E, $I_0 = 7$	KOSTYUK, SHEBALIN	KAPG Atlas
1937 Sept. 14, 08^h58^m GMT, 48.21°N, 23.54°E, $I_0 = 6.5$	KOSTYUK, SHEBALIN	KAPG Atlas
1965 Oct. 24, 06^h26^m GMT, 48.22°N, 22.65°E, $I_0 = 7$	KOSTYUK, SHEBALIN	KAPG Atlas
1829 July 1, 19^h30^m GMT, 47.5°N, 22.2°E, $I_0 = 7$	KOSTYUK, SHEBALIN	KAPG Atlas
1834 Oct. 15, 06^h30^m GMT, 47.6°N, 22.3°E, $I_0 = 8.5$	KOSTYUK, SHEBALIN	KAPG Atlas
1902 Nov. 26, 12^h15^m GMT, 49.7°N, 12.8°E, $I_0 = 6$	PROCHÁZKOVÁ	KAPG Atlas
1908 Nov. 3, $17^h21^m42^s$ GMT, 50.3°N, 12.4°E, $I_0 = 6.5$	PROCHÁZKOVÁ, KUNZE	KAPG Atlas
1908 Nov. 6, $04^h36^m11^s$ GMT, 50.4°N, 12.4°E, $I_0 = 6.5$	PROCHÁZKOVÁ, KUNZE	KAPG Atlas
1738 June 11, 09^h00^m GMT, 47.5°N, 26.6°E, $I_0 = 8.5$	SHEBALIN, DRUMEA, MOSKALENKO, ROMAN	KAPG Atlas
1790 Apr. 6, 19^h26^m GMT, 47.5°N, 26.6°E, $I_0 = 8$	SHEBALIN, DRUMEA, MOSKALENKO, ROMAN	KAPG Atlas
1802 Oct. 26, 10^h55^m GMT, 45.7°N, 26.0°E, $I_0 = 9$	SHEBALIN, DRUMEA, MOSKALENKO, ROMAN	KAPG Atlas
1940 Nov. 10, $01^h39^m07^s$ GMT, 47.8°N, 26.8°E, $I_0 = 9$	SHEBALIN, DRUMEA, MOSKALENKO, ROMAN	KAPG Atlas
1921 Jan. 15, 10^h55^m GMT, 45.7°N, 20.1°E, $I_0 = 7$	METOVIĆ	UNESCO Atlas
1922 Nov. 24, 02^h15^m GMT, 45.6°N, 18.7°E, $I_0 = 7$	CVIJANOVIĆ	UNESCO Atlas
1926 Aug. 10, 01^h10^m GMT, 48.0°N, 23.7°E, $I_0 = 7$	EVSEJEV	UNESCO Atlas
1927 Feb. 20, 06^h47^m GMT, 46.3°N, 16.9°E, $I_0 = 7$	CVIJANOVIĆ	UNESCO Atlas
1938 Mar. 27, 11^h06^m GMT, 46.0°N, 16.9°E, $I_0 = 8$	CVIJANOVIĆ	UNESCO Atlas
1964 Apr. 13, 08^h30^m GMT, 45.3°N, 18.1°E, $I_0 = 8$	CVIJANOVIĆ	UNESCO Atlas

be said that the map reflects the macroseismic data for not more than the last three or four centuries, earlier isolated observations are not included, such as the reports on destructive shocks during 1441—1443 in Central Europe. The resulting map of maximum observed intensity shows the following general features:

1. The whole Bohemian Massif has been shaken with intensity I = III—IV and locally also with intensities ranging from V to VII because of local shocks in the western and northern mountainous regions; shocks from the Eastern Alps are felt with intensities up to V. Maximum intensities from VI to VIII were observed in the Western Carpathians, in the Little Carpathians, in the vicinity of Žilina and in Central and Northeastern Slovakia. The strongest shock originated in 1763 near Komárno with the intensity IX.
2. The outer part of the Western Carpathians has been shaken mainly with intensities III to IV and only locally with intensities V to VII. These intensities have been caused by local shocks in the region of the Carpathian foredeep and the fore-Sudetic monocline.
3. The SW part of the Hungarian Basin has been shaken with intensities up to V, the central and the northern part with maximum intensities VI to IX. Regions of the largest intensities are near the towns Komárom, Mór, Jászberény, Kecskemét, Eger, Dunaharaszti, Pincehely and Tereske-Tolmács.
4. The territory of Eastern Austria has been shaken with maximum intensities V to IX because of many foci along the active belt of the Eastern Alps and of Carinthia.
5. The general pattern of the maximum isoseismal lines shows characteristic features for the various zones in Bulgaria. There are five zones in which the maximum intensity was IX or even higher. Rodope Mts and the Struma Valley are the regions of the largest observed intensity in our investigated area.
6. On the territory of the Eastern Carpathians the macroseismic field is mainly determined by the effects of intermediate depth shocks in the Vrancea region. A relative increase of intensity was recorded in areas of recent subsidence in parts of the Pannonian Basin (I = VII—VIII) and in the front of the Eastern Carpathian arc (I = VIII—IX). Intermediate values (I = VI—VII) were observed in Dobrogea and in the Getic depression (western zone of the Carpathian fore-deep).
7. The isolines of the largest observed intensity in Ruthenia were drawn using only the isoseismal maps available in the UNESCO and K.A.P.G. Atlases; the resulting isolines deviate from those published in *Seismic Zoning of the USSR* (1968).
8. For comparison the isolines of maximum observed intensity are added also for regions outside our area of investigation, i.e. for Yugoslavia, Northern Greece and Albania.

In general, the map demonstrates well the influence of focal depth on the size and shape of isoseismals, see e.g. very small surface of isoseismals of intensity VI in Western Carpathians and large surface in the Vrancea region. By comparing the map of maximum observed intensity with the tectonic map it follows that the observed intensities are relatively higher along the boundaries of tectonic structures. This phenomenon reflects a higher mobility of young faults. Examples are faults along the Giant Mts, the Ore Mts, the faults belonging to the Peripienian lineament, the faults of EW direction in Bulgaria, the deep rejuvenated Hercynian faults in Romania, etc. In some regions local increase of intensity can be related to thick alluvial deposits (South Bohemia, Pannonian Basin, etc.).

7. Conclusion

The presented basic seismological maps were prepared for a general review of seismicity of the area investigated within the K.A.P.G. programme. The work will be continued in compiling further maps of statistical character showing the distribution of earthquake hazard for practical aims. Before reaching this goal statistical and seismological analyses must be completed. Most tasks will be again accomplished in mutual collaboration of experts from countries involved in the KAPG programme.

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ЭПИЦЕНТРЫ ЗЕМЛЕТРЯСНИЙ И МАКРОСЕЙСМИЧЕСКАЯ БАЛЛЬНОСТЬ В КАРПАТО-БАЛКАНСКОЙ ОБЛАСТИ

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Резюме

Изучённая область охватывает территории Болгарии, Румынии, югозападной Украины, Мольдавии, Венгрии, Чехословакии и части Австрии, Югославии и Греции. Карты эпицентров и самые большие наблюдённые балльности показывают географическое распределение активности и соответствующие макросейсмические действия. Землетрясения появляются в нескольких сейсмогенных зонах (провинциях) связанны с неотектоническими особенностями данной области. Очат средней глубины в Вранцеа является особенным явлением освободившим наибольшее количество энергии вместе с несколькими очагами вБолгарии.

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DERIVATION OF STRUCTURAL TRENDS FROM GRAVITY AND MAGNETIC DATA WITH APPLICATIONS IN THE PANNONIAN BASIN

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The studies of large-scale structural and tectonic features are based on correlation functions derived from potential field (gravity and magnetic) data. Directivity diagrams obtained from auto-correlations reveal clearly the main trends and offer a quantitative measure, free of personal bias, for an objective evaluation of the significance of the trends. Good correspondence has been found between boundaries of major structural units and major faults and directions derived by the proposed method from regional gravity data.

By changing the input parameters (block size, maximum shift, shift between adjacent blocks, etc.) the method could be made sensitive to variations of different scale. Therefore, it may become a useful tool for tectonic studies of both local and regional features.

Introduction

This paper reports a part of the experimental studies on the determination of trends and direction statistics from gravity and magnetic data. Preliminary results of the application to the Bouguer anomaly field in Hungary are also discussed.

The method can be generalized and improved in many ways. Some of the possible modifications are summarized in the conclusions. Good correlation has been found, however, between the boundaries of major structural units and major faults and directions derived by the proposed method in its present version from gravity and magnetic data. Thus, it may be of interest to report the investigations even at this early stage and direct attention to possible applications in tectonic studies.

A number of authors proposed methods for determining directivities on potential field maps.

Affleck (1963) e.g. described a graphical analysis which consists of drawing straight lines along closed highs and lows as well as along noses and nuxures in contour lines. The lengths and azimuths of all lines are tabulated and then summed in 5° azimuths increments. The total length in a given azimuth interval characterizes the directivity. Main directions are shown by large sums.

A similar approach is to substitute each contour line by polygons and to determine again the azimuths and lengths of the sides of the polygons. Both methods obviously require time-consuming preparations and cannot be easily

implemented on computers. On the other hand, both methods virtually disregard amplitude, therefore the effect of near-surface geology and even that of the topography becomes significant. In regional studies, however, it is advisable to use amplitude-sensitive methods because high-amplitude anomalies are related to deeper sources.

HORTON et al. (1964) determined tectonic trends with a two-dimensional auto-correlation. AGARWAL and KANASEWICH (1971) described an empirical method based on cross-correlations computed between larger and smaller blocks of data and then on fitting a least-squares third degree polynomial to the maximum coefficients in the correlation matrices. Their methods have the drawback that trend lines may disappear or intersect each other at the boundaries of so-called sample maps and may have rather significant curvatures.

The approach reported in this paper attempts to retain the advantages of the two latter approaches as well as to get rid of their drawbacks. Auto-correlations are determined for shifted blocks of data and main directions are computed from the auto-correlations. Both the dimensions of the data blocks as well as the shifts could be changed, thus allowing for an optimal choice of parameters corresponding to the problem at hand.

Main Features of the Proposed Method

The determination of the main direction (or directions) consists of three steps. First the auto-correlation is computed from the data, then directivity diagrams are derived and, in the third step, interpretation follows. These three steps are discussed and illustrated separately.

Figure 1. illustrates the definition and computation of the auto-correlation. The auto-correlation at the argument jl is defined by the expected value of the products $(g_{i,k} \cdot g_{i+j,k+e})$

$$\varphi_{j,l} = \varepsilon\{g_{i,k}\, g_{i+j,k+l}\}$$

where $g_{i,k}$ $(i = 1, 2, \ldots, N;\ k = 1, 2, \ldots, N)$ denote gravity (or magnetic data) in point ik.

The expected value is estimated by the average of the products. When we could use $N \times N$ data (called data block in the following)

$$\varphi_{j,l} = \frac{1}{(n-j)\,(n-l)} \sum_{i=1}^{n-j} \sum_{k=1}^{n-l} g_{i,k}\, g_{i+j,k+l}$$

The auto-correlation can be normalized by its central value

$$\varphi_{j,l}^{norm} = \frac{\varphi_{j,l}}{\varphi_{0,0}}\,.$$

Other estimates and some modifications of the data have also been suggested but now we do not go into more computational details. It suffices to observe that auto-correlation is estimated from data in the dashed area. When the shifted portion of data describes a field similar to the field represented by the data, which remain in the original position, the average of the

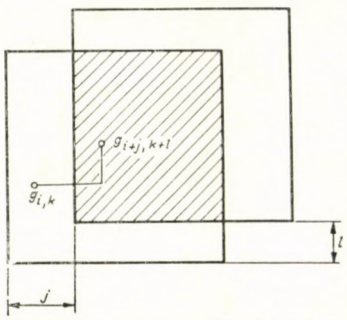

Fig. 1. The auto-correlation at the argument jl is the average of products of coincident data in the dashed area

products is large. Thus, the auto-correlation at point (j, l) provides a measure of similarity between original and shifted portions of the field. When long linear features are present in the map, shifts parallel to the linear trend of the map do not essentially descrease similarity.

Figure 2 shows a part of the BOUGUER anomaly field in Hungary over an area of 120×120 km^2. Auto-correlations computed from the data are shown in Fig. 3. Each auto-correlation corresponds to a 80×80 km^2 data block, i.e. block centers are 40 km apart.

The auto-correlation has the property

$$\varphi_{j,\pm l} = \varphi_{-j,\mp l}$$

because shifts $(j, \pm l)$ and $(-j, \mp l)$ obviously result in the same situations. Therefore, negative values of shifts in the North–South direction are shown, only.

The auto-correlation is determined by the Fast Fourier Transform algorithm.

Directivity diagrams are obtained by summing values of the auto-correlation along given azimuths. The main directions can be determined from the positions of the maxima. Four directivity diagrams, computed from the four auto-correlations of Fig. 3 are shown in Fig. 4. Theoretically $I_R(\alpha)$ is defined by

$$I_R(\alpha) = \int_0^R \varphi(r \cos \alpha, r \sin \alpha)\, dr$$

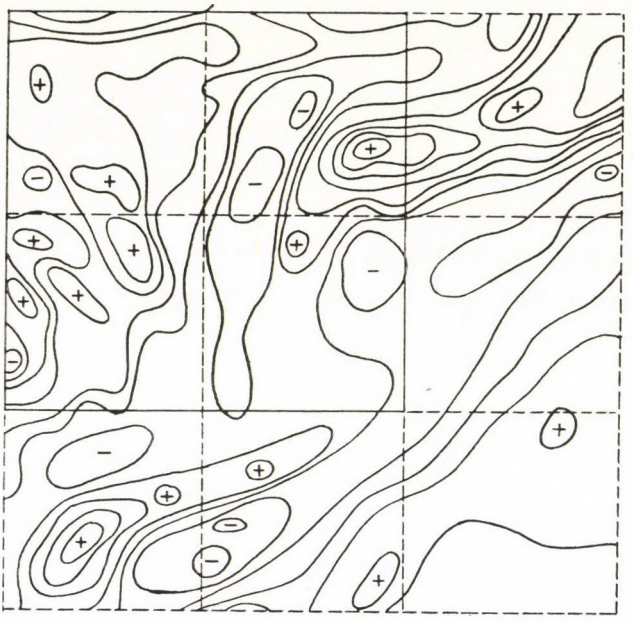

Fig. 2. A part of the BOUGUER anomaly field in Hungary. 80 km × 80 km data blocks are indicated by dashed lines

where φ denotes the continuous auto-correlation with arguments expressed by polar coordinates. Instead of a continuous function, however, we have at hand the digital auto-correlation φ_{jl}. Therefore, $I_R(\alpha)$ is obtained by

$$I_R(\alpha) = \sum_{i=1}^{n} \varphi(id, \alpha)$$

where $\varphi(id, \alpha)$ denotes a value of the auto-correlation at the azimuth α, at a distance of id from the origin. $\varphi(id, \alpha)$ is determined by interpolation.

Small n corresponds to a close neighbourhood of the origin, i.e. then similarity between original and shifted data has been evaluated for small shifts, only. Since similarity can be expected and usually it is observed for small shifts, $I_R(\alpha)$ is approximately circularly symmetric for small values of n.

For statistic stability reasons, on the other hand, n can not be very large, $n = \mathrm{L}/5\mathrm{d}$ is about the upper limit of n (where L is the length of the data block). E.g. for $L = 80$ km, $\mathrm{d} = 2.0$ km, $n \leq 8$ should be chosen.

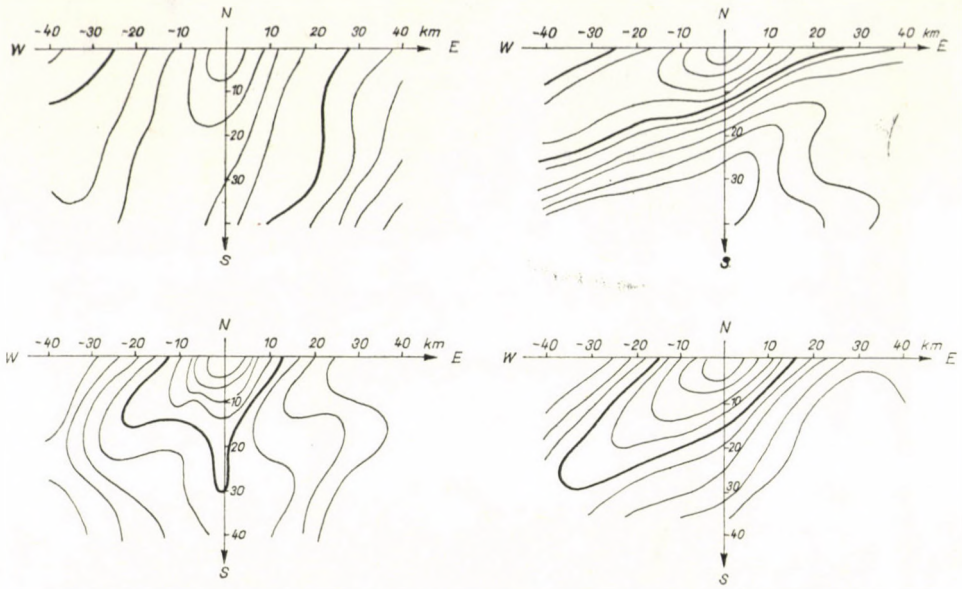

Fig. 3. The auto-correlation functions of the data blocks shown in Fig. 2

For larger shifts similarity between shifted and original data disappears, except when the anomaly field possesses long linear features. Shifts along the directions of the linear features do not destroy similarity, while shifts perpendicular to it make the coincident data blocks dissimilar and thus give small auto-correlations. Therefore, long linear features correspond to large peaks in I_R when n is sufficiently large. The diagram refers to the whole block of data and the peak becomes largest when the axis of the long linear feature goes through the center of the block. (Compare Fig. 2 with Fig. 4.) Therefore, by searching for the maximum of the peak amplitudes one could determine the position of the axis of the dominating linear trends in the data block.

The role of parameter n is illustrated by Fig. 5. Each directivity diagram in group (*a*) as well as in group (*b*) is derived from the same auto-correlation. The maximum shifts (R) are 10 km, 20 km and 40 km, respectively. The data block which supplied the auto-correlation for the computation of the diagrams in group (*a*) contains a single dominating trend direction. The larger is the maximum shift R, the clearer is the main direction.

The diagrams in group (*b*) depict a different situation. The data block now contains two dominant directions. With maximum shift $R = 10$ km the directions can not be discerned. Maximum shift $R = 20$ km shows that there are two directions, which are about of the same significance, but the picture is still too complex, $R = 40$ km, at last, clearly resolves the directions and shows their relative weights.

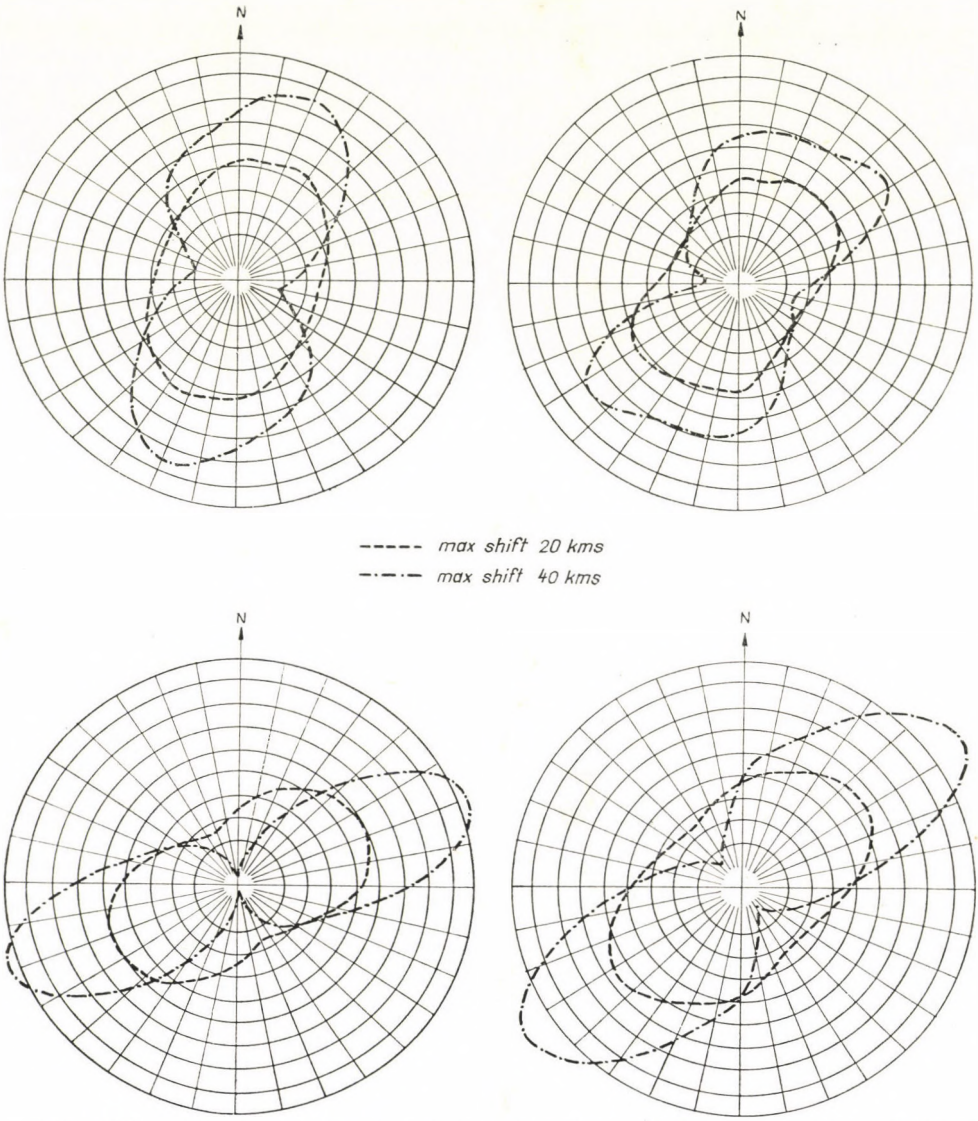

Fig. 4. Directivity diagrams computed from the auto-correlation functions shown in Fig. 3

In the interpretational stage the significance of the direction is characterized by

$$Q_R = \frac{I_{\max} - I_{\perp}}{I_{\max}}$$

where $I_{\max}$ denotes the value in the diagram, measured along the main direction and $I_{\perp}$ denotes the value in the diagram perpendicular to the main direc-

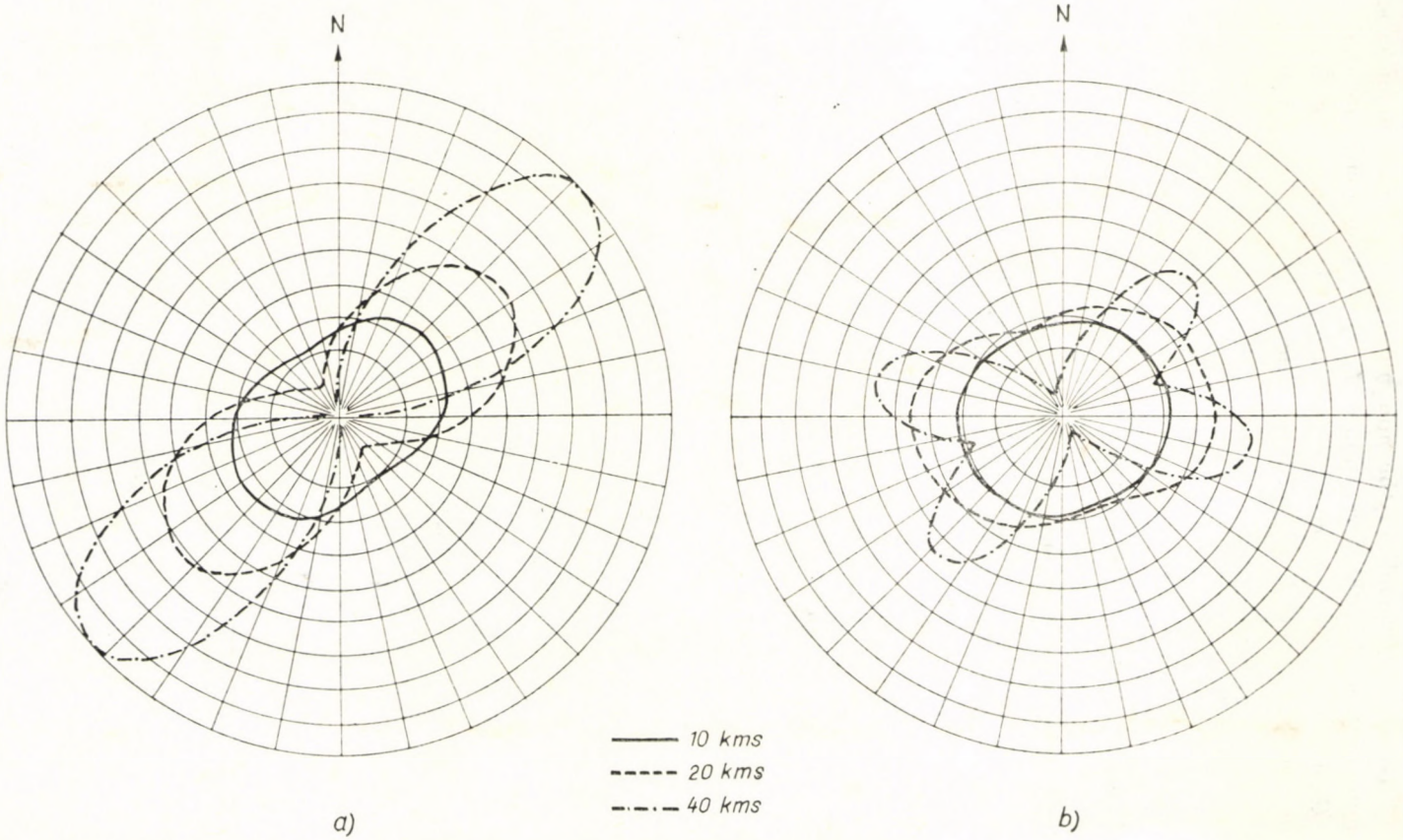

Fig. 5. The effect of the parameter R. Group (a) illustrates one main direction, group (b) is an example for data blocks where two directions interfere

tion. $I_{\perp}$ usually corresponds to the minimum. Q_R is again a function of the maximum shift, used in the derivation of $I_R(\alpha)$. It is obvious, that the definition of Q_R can be used only when the diagram is similar to an ellipse and it does not show more than one direction. For complex diagrams individual interpretations are necessary.

Results

Figure 6 summarizes the results obtained from the anomalies of the vertical intensity of the magnetic field in a part of Transdanubia. The contours are drawn at 100 γ intervals wherever it is possible. Due to the large variation, however, in the neighbourhood of maxima and minima not all 100 γ contours could be represented. 32×32 km^2 large data blocks have been used. Main directions are given by arrows whose lengths are proportional to Q_R ($R = 8$ kms).

Figure 7 shows directions derived from smoothed Bouguer gravity data. 80 km $\times$ 80 km blocks have been used with 40 kms shifts between adjacent blocks. Directions are shown by straight lines whose lengths are proportional to $Q_{20\ \mathrm{km}}$. The lines go through the block centers. *C* denotes blocks where more than one dominant trends interfere with each other and the complexity of the directivity diagrams does not allow the determination of separate directions.

Figure 8 shows a tectonic map of Hungary after Kőrössy (1971). The good correlation is obvious. The block size 80 km $\times$ 80 km could not show small faults but major faults with lengths comparable with block size boundaries of large tectonic units are clearly revealed in most places.

Conclusions

The first experiments aiming at the detection of directivities from gravity and magnetic data proved that the proposed method is sound in principles and it merits further investigations.

The derived directions correlate well with the boundaries of major tectonic units and with faults, if the fault is long enough and the throw is sufficiently large. Resolving power and accuracy can be increased by the following modifications:

a) Different depth regions and different structures could be analyzed if input data are passed through high-pass or low-pass filters. High-pass filters enhance local features and low-pass filters indicate deep-seated regional features.

b) Block size could be changed and adapted to the type of data to be analyzed. For local studies residual fields and smaller block sizes should be used. For regional studies larger blocks including regional field values are

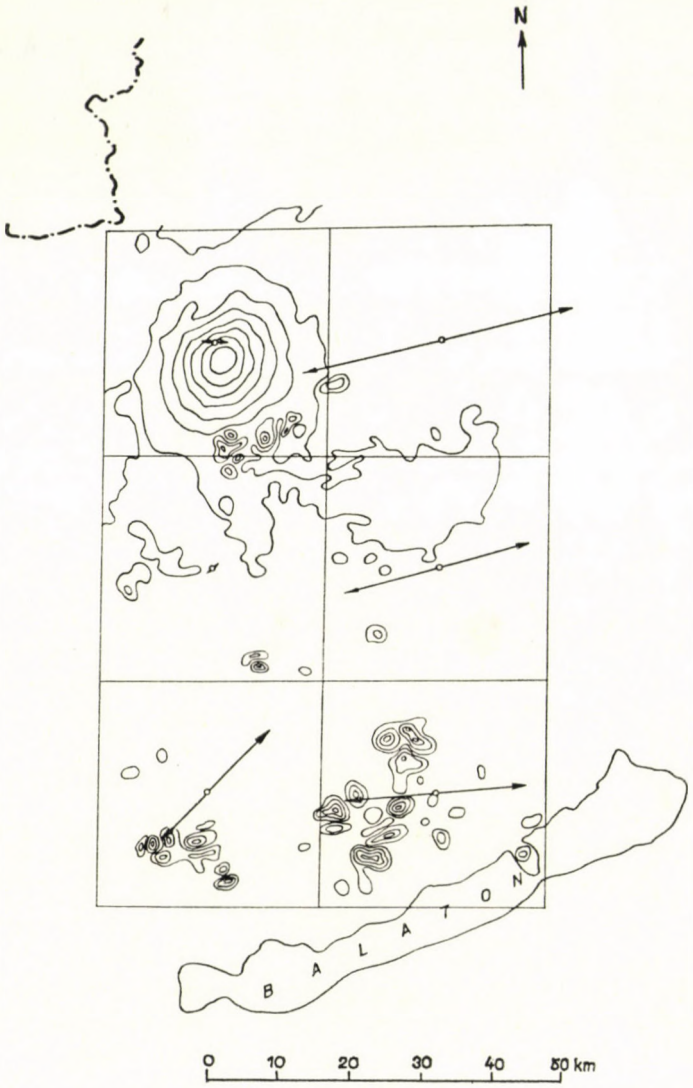

Fig. 6. Anomalies of vertical intensity of the magnetic field in a part of Transdanubia, boundaries of data blocks and main directions are superimposed

suitable. By changing filter parameters and block sizes, the main direction and secondary directions could be separated.

c) By using smaller shifts between adjacent blocks the positions of tectonic lines could be better localized. In Fig. 8 the accuracy of the position is ± 20 kms.

The potential use of the method in tectonic studies is twofold. First, it may support the existence and significance of already known tectonic lines

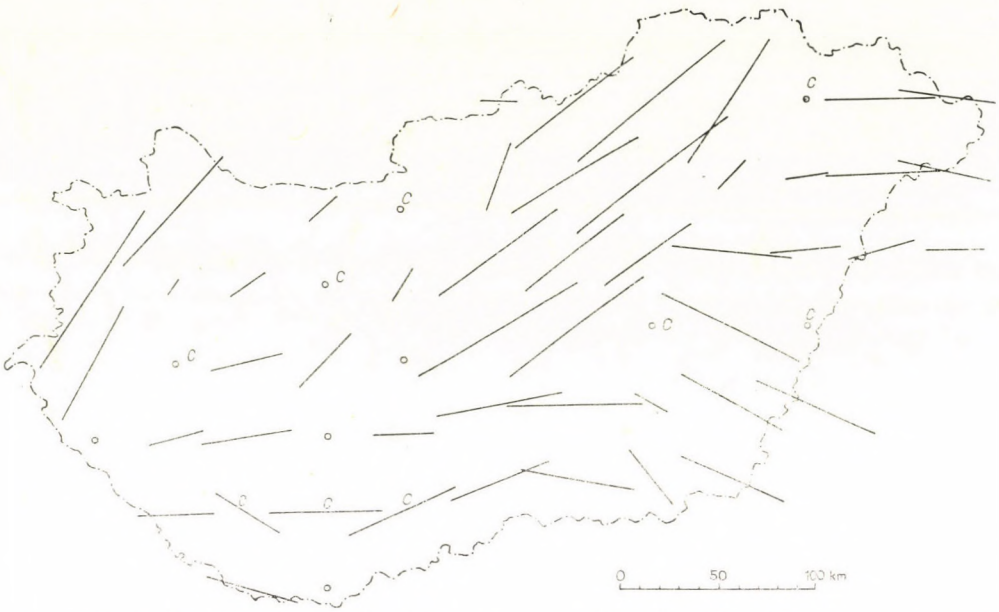

Fig. 7. Main directions, derived from Bouguer gravity data. Lengths of lines are proportional to Q_{20km}. Centers of lines coincidence with block centers, *C* denotes blocks where interference of directions does not allow the computation of *Q*

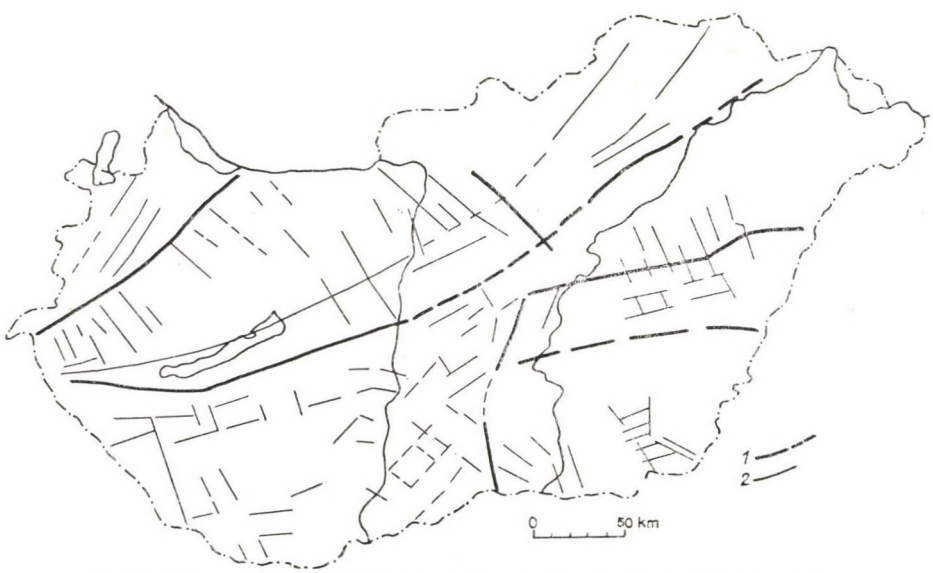

Fig. 8. Main structural units and faults in Hungary (after KŐRÖSSY, 1971)

derived from geological and geophysical data. On the other hand, it may be instrumental in establishing of such features, where data are scarce or lacking or contradict each other.

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УСТАНОВЛЕНИЕ СТРУКТУРНЫХ ТРЕНДОВ ПО ГРАВИТАЦИОННЫМ И МАГНИТНЫМ ДАННЫМ И ЕГО ПРИМЕНЕНИЕ К ПАННОНСКОМУ БАССЕЙНУ

К. КИШ, А. МЕШКО

Резюме

Изучение крупомасштабных структурных и тектонических явлений ведётся с помощью корреляционных функций установленных по гравитационным и магнитным данным. Диаграммы направленности полученных из автокорреляций четко показывают главные тренды и дают возможность оюбективной оценки значений трендов. Границы крупных структурных единиц и также главные разломы хорошо совпадают с направлениями полученными предлагаемым методом, по региональным лравитационным данным.

Путём изменения входных параметров метод легко прослособляется к различным (местным и региональным) масштабам.

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THE CARPATHIAN GEOELECTRIC ANOMALY AND ITS RELATION TO INDEPENDENT GEOPHYSICAL INFORMATION

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The geoelectric conduction anomaly was located near the boundary of the Outer and Central Carpathians by investigating the anomalous behaviour of geomagnetic variations along several profiles crossing both the West and East Carpathians on the territories of Poland and Czechoslovakia. The depth of the anomalous induction field source was estimated from geomagnetic data to be between 15 and 25 km.

The nature and significance of the anomaly are discussed in correlation with available geological information and independent geophysical (geothermal and seismological) data.

Introduction

For several years the Polish and Czechoslovak authors of this report have dealt with research into the geoelectric conduction anomaly detected in the transition zone between the Carpathians and adjacent older geological units such as the Bohemian Massif and the Polish Paleozoic platform.

During recent years certain number of profiles starting on the Polish Paleozoic platform, crossing the Outer and Central Carpathians and terminating in the inner Carpathian region have been covered with one-dimensional arrays of 5 to 12 simultaneously operating geomagnetic field observatories. The geomagnetic variation data collected during these field experiments were analyzed to obtain

(i) the induction vectors at each field site and subsequent localization of zero-variation line of vertical geomagnetic component marking the zone of anomalous induction under which an electrically well conductive channel can be assumed;

(ii) the transfer functions in the range of periods from about 1000 to 8000 *s* and characteristics of the period dependence of induction effect;

(iii) the internal and external parts of variation fields with subsequent localization and depth estimation of internal geoelectric inhomogeneity producing the anomalous induction effect.

Partial results obtained by detailed analysis of the geomagnetic data from DSS profile No. V were published in papers by JANKOWSKI et al. (1975, 1977). In this paper new data on the geoelectric anomaly and their interpretation are presented.

Induction Vectors

The results of induction vector estimates over the Czechoslovak and Polish section of the Carpathians are plotted in the geological map (Fig. 1). The zone of anomalous induction is spotted with high accuracy of about 1—2 km in the eastern part of Czechoslovakia and with an accuracy of about 5 km in the West Carpathians. As there are no measurements between the town Žilina and the High Tatra region so far, the northern part of the anomaly is located only approximately by relating it to the Klippen Belt by analogy from other regions. Certain correction, however, seems to be suggested by Polish measurements in the High Tatra region, as will be shown later.

In correlating the position of the geoelectric induction anomaly with important geological and geophysical lines, we have to mention first changing relationship to both the Klippen Belt and the Peripieninian lineament. In the western part of the Carpathians a shift of about 10—12 km from the Klippen Belt into the Outer Flysch Carpathians seems to characterize the position of the induction anomaly. Further northeastward the anomaly approaches the Klippen Belt, merges with it and between the town of Žilina and the High Tatra region it seems to penetrate relatively deep onto the inner Carpathian side of the Klippen Belt, instead of following it to the region of Nowy Targ, as generally anticipated. This new position on the inner side of the Klippen Belt is suggested by Polish measurements in the foreland of High Tatras. Specifically, by a small induction vector the Polish field observatory at Brzegi seems to be very close, but still on the northern side of the geoelectric anomaly which must, thus, be anticipated approximately under Belanské Tatry Mountains (Fig. 1).

Further to the east the anomaly returns back to the Klippen Belt, probably in the area of the Pieniny Mountains. Starting from the DSS profile No. V and continuing to the south-east Eastern Slovakia, the anomalous zone separates substantially from the Klippen Belt and shifts gradually further into the Flysch Carpathians. On our profile P-76 the separation of both belts amounts to about 30 km and in this relative position they go on into the area of the Western Ukraine. Here, however, the anomaly converges to and finaly merges again with the Marmaros zone, the geological equivalence of the Klippen Belt.

With the exception of the West Carpathians, the Klippen Belt coincides over greater part of the Czechoslovak territory with the Peripieninian lineament which is marked by anomalous behaviour of geophysical fields. The most prominent feature of this lineament, however, is its seismic activity. Figure 2 represents the earthquake epicenters over the period from 1400 to 1970 as compiled by BROUČEK (1973). Similarly, Fig. 3 displays schematically the epicenters on the territory of West Ukraine compiled by KOSTYUK (1974). For comparing these phenomena, we marked schematically the position of geo-

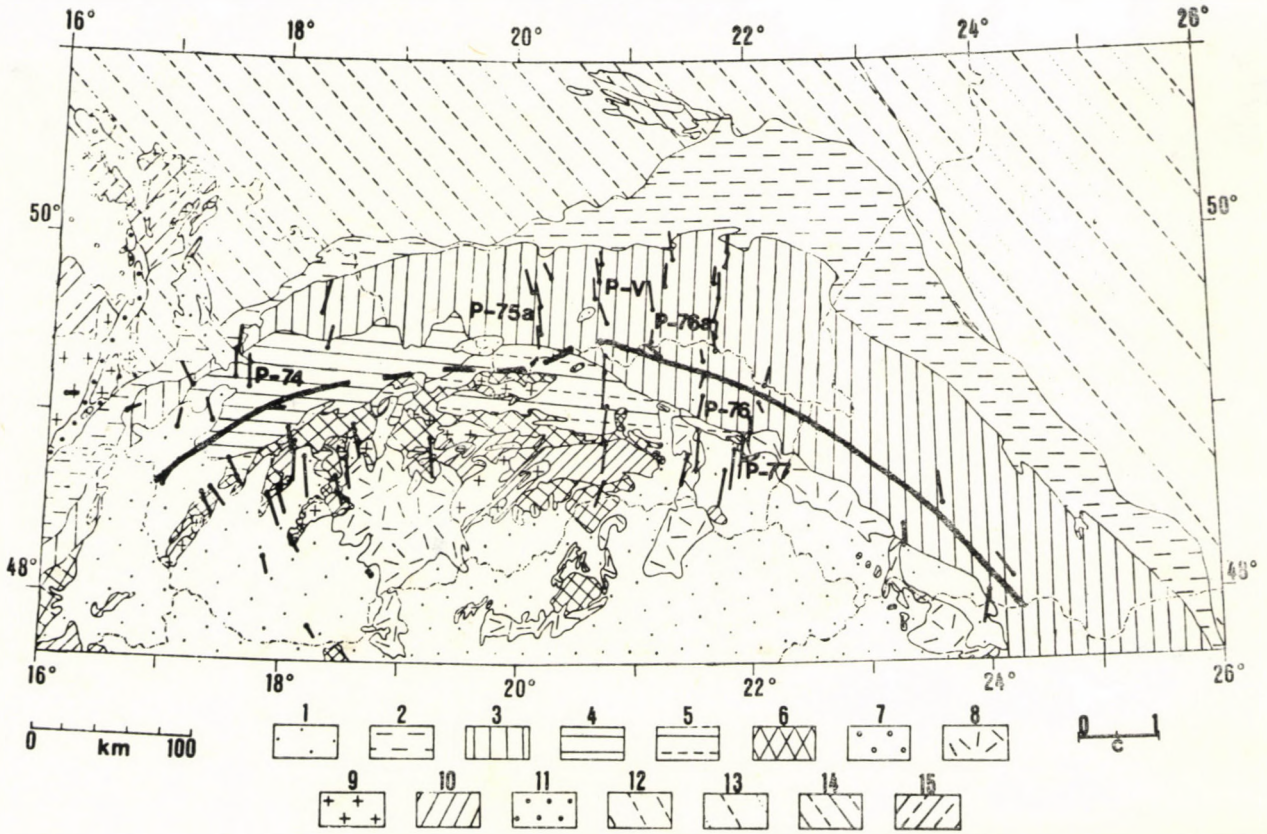

Fig. 1. Schematized geological map of the Western and Polish Czechoslovak part of the East Carpathians (compiled by M. Tobyášová according to MAHEL (1973)) and the induction vector estimates. Geoelectric anomaly is marked by inversion of induction vector azimuths. One-dimensional arrays of geomagnetic field observatories are labelled as P-V, P-74, P-76, etc. Explanation of geological symbols: 1. the Tertiary covers, filling of basins and intramontane depressions; 2. sediments of the Carpathian foredeep. The Outer Carpathians: 3. tectonic units of the Outer Flysch Group; 4. tectonic partial unit of the Magura nappe; 5. the Inner Carpathians; 6. the Carpathian nappes (Lower Triassic—Cretaceous); 7. the Cretaceous sediments of the Bohemian Massif; 8. neovulcanics and pyroclasts; 9. late synkinematic and postkinematic magmatites; 10. older structures reworked during Neoidic orogeny; 11. Permian-Carboniferous of the Boskovice Furrow; 12. platform Palaeozoic units; 13. the Russian platform; 14. Paleocaledonian units of the Carpathian foredeep; 15. Assyntian and Moldanubian structures of the Bohemian Massif

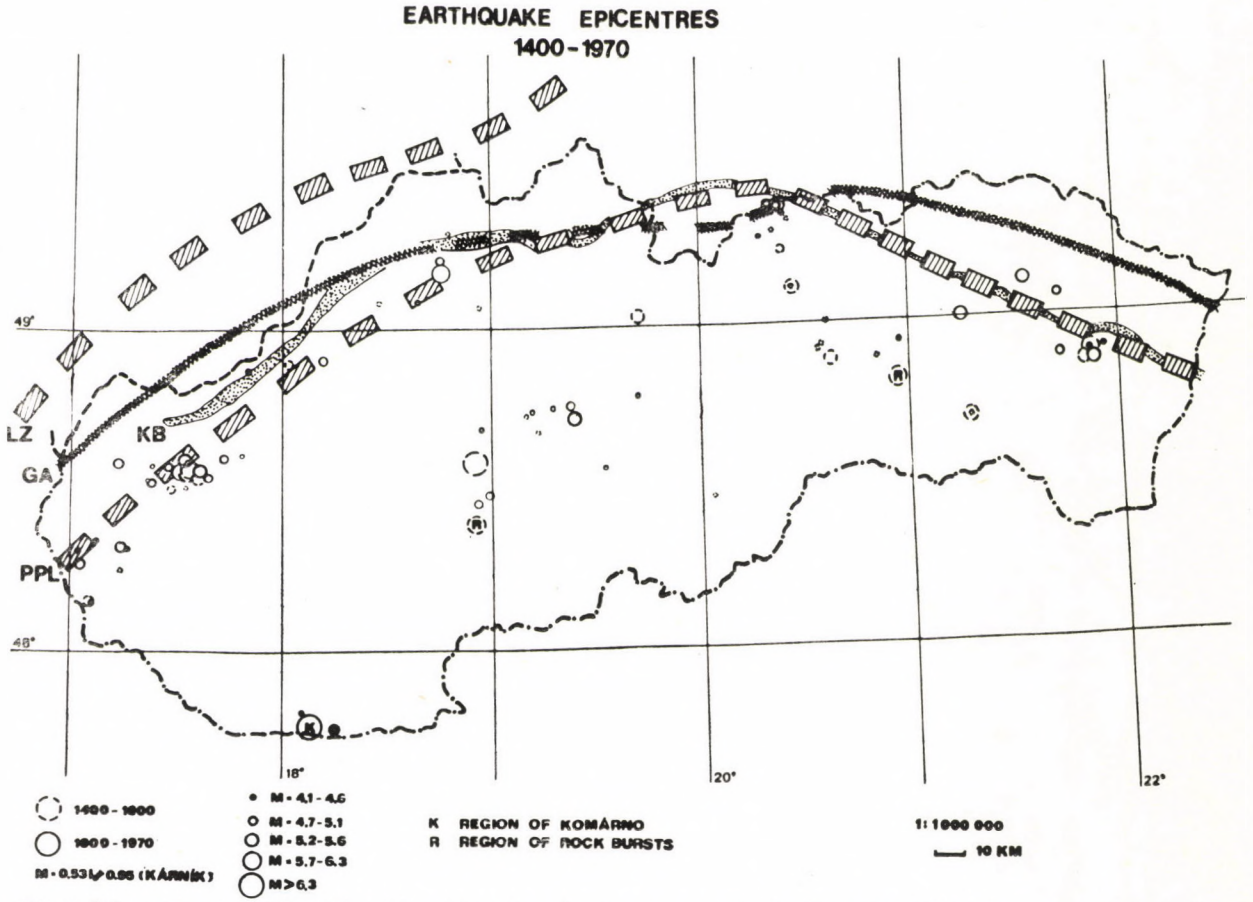

Fig. 2. Schematic map of Slovakia with earthquake epicenters as compiled by BROUČEK (1973). Important geological and geophysical lineaments are shown and labelled by following symbols: LZ — the Lednice Zone; GA — geoelectric anomaly marked by inversion of induction vectors; KB — the Klippen Belt; PPL — the Peripieninian Lineament

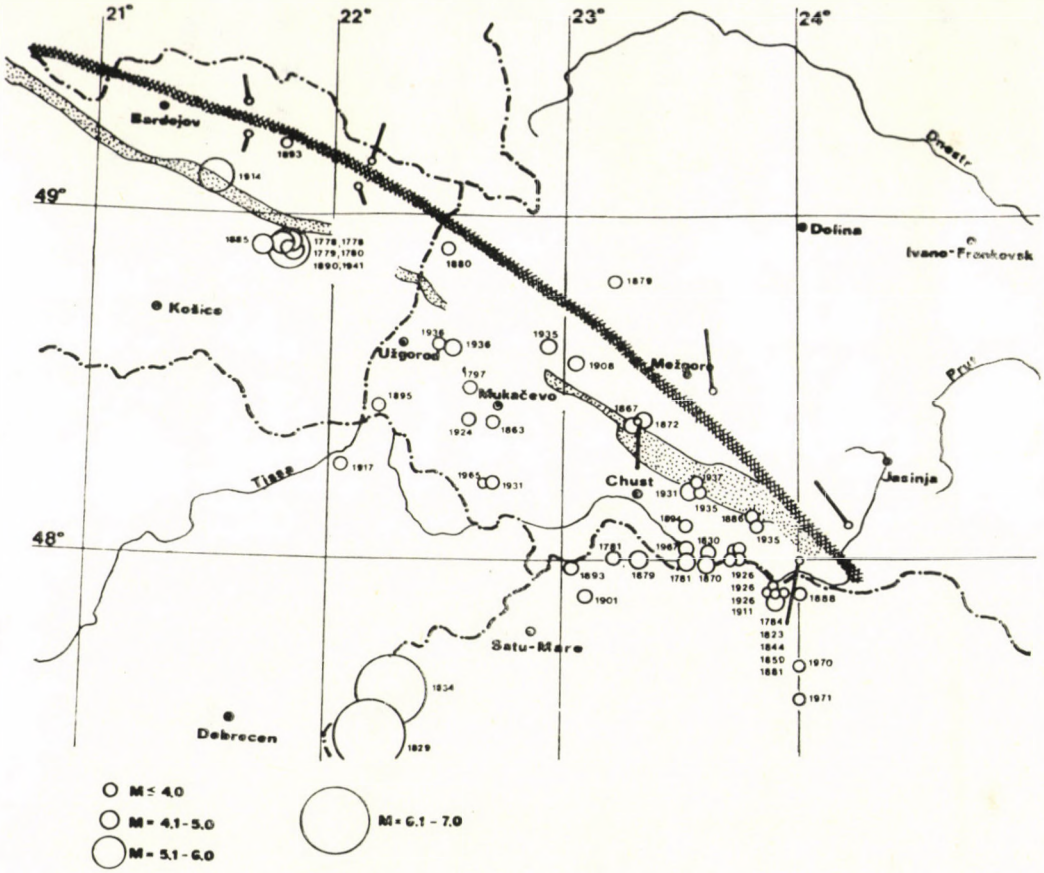

Fig. 3. Schematic map of Western Ukraine with earthquake epicenters compiled by KOSTYUK (1974). The geoelectric anomaly is shown schematically by cross-hatched zone

electric anomaly in both maps. It seems remarkable that majority of epicenters is located close to the anomaly, usually enclosed in the inner side of the geoelectric anomaly on regional scale. On the other hand, there are epicenters with no relation to the anomaly.

It is possible that under some specific conditions both phenomena are related. In support to this indication a short communication by LILLEY (1975) is referred to. Here, the electric conductivity, anomalies and the continental seismicity in Australia are mentioned (Fig. 4). Two-dimensional magnetometer arrays over two wide areas of Southern Australia have detected anomalous induction effects. Their interpretation has located deep-seated good conductors. Figure 4, reproduced from LILLEY's communication, displays the earthquakes that occured between 1897 and 1972. Several major earthquake zones are apparent, though on a world scale the seismicity of Australian continent is

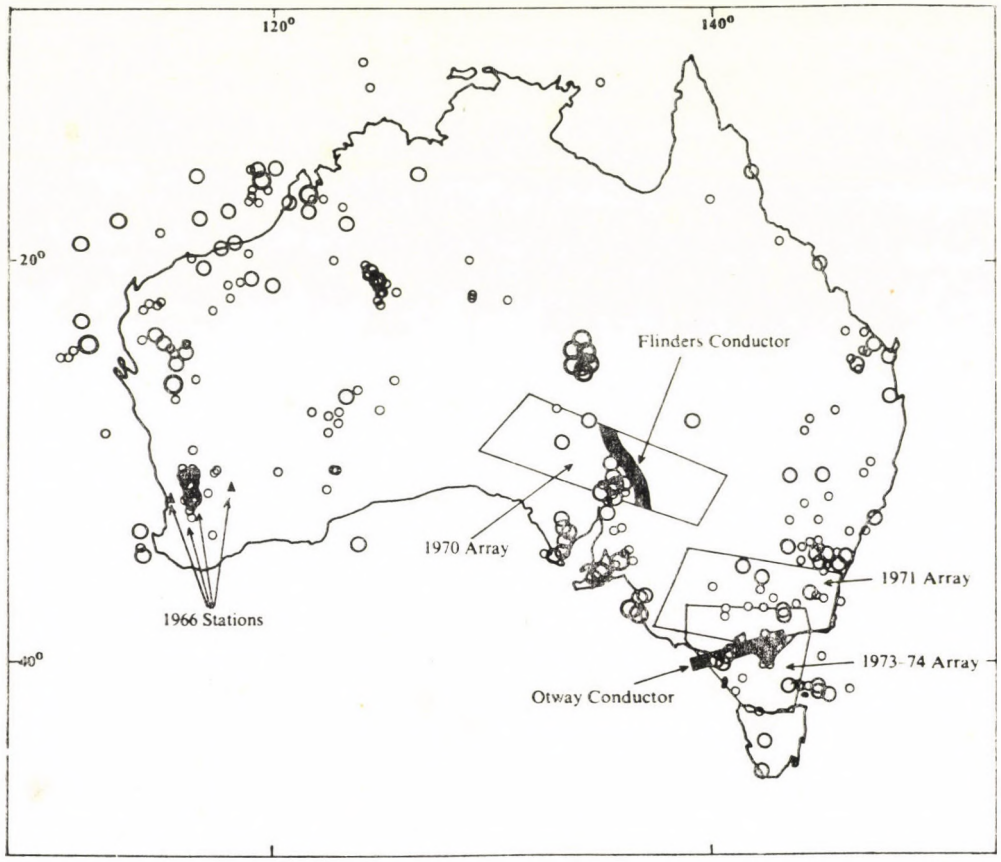

Fig. 4. Electric conductivity anomalies and continental seismicity in Australia as reproduced from LILLEY (1975)

low. Both continental conductivity anomalies, so far discovered, are in regions of seismic activity.

General relationship between seismic activity and electric conductivity anomalies has not been established yet. Speculations about possible correlation can involve, according to LILLEY's communication, processes of ductile flow along fault planes and increase of electric conductivity by ductile shearing of rocks. The rheological process of ductile flow beneath continental earthquake zone may involve the conversion of considerable amount of mechanical energy to heat.

The Depth Estimation of Induction Anomaly

When investigating other important characteristics of the geoelectrical anomaly, we analyzed the spatial distribution of variation events to estimate both the horizontal position and the depth of anomalous field source. Assuming a linear concentrated electric current as an idealized field source, we applied the methods of separating the variations into their internal and external parts and estimated the source depth from the geometry of internal field. In virtue of source idealization, the estimate, thus obtained, must be considered as the *maximum depth* assessment.

Applying this method, we also accounted for certain regional part involved in the geomagnetic horizontal component. It must be subtracted from the internal component to get a proper estimate. Omitting the details, the method of estimating the source depth by field separation is summarized in Fig. 5.

Separating all variation events recorded simultaneously by one-dimensional magnetometer arrays, we were able to assess the depth of anomaly source in different regions of the Czechoslovak—Polish section of the Carpathians. As an example, a summary of all depth estimates (and also the horizontal location of the anomaly) from East Slovakian profile P-76 is shown in Fig. 6. Similar procedure was applied to variation data from other profiles. All estimates scatter within the limits between 16—25 km on different profiles. Due to this scatter, we have not been able to decide, whether there is any regional tendency for the anomaly to dip or rise in certain specific areas of the Carpathians.

Relation to Geothermal Activity

From the results of field separation and from independent geophysical and geological information certain geoelectric cross-section along profiles was suggested and investigated by numerical modelling in paper by JANKOWSKI et al. (1977). Summarizing the results of calculations for models fitting in the best way the anomaly, we can estimate the electric resistivity of the channel to about 1—2 Ωm, while the resistivity of adjacent crystalline blocks is of the order of 500—1000 Ωm. The electric resistivity of superficial flysch formations was assumed to be about 8—10 Ωm.

Possible existence of such highly conductive material at crustal depth on the Carpathian scale gives rise to a natural question on its possible nature. It is a well-known fact that increase of electric conductivity can, among others, be an indication of increased internal temperatures. Under certain conditions the temperature can be sufficient to cause even partial melting of rocks increasing, consequently, their electric conductivity by 2—3 decimal orders. This can

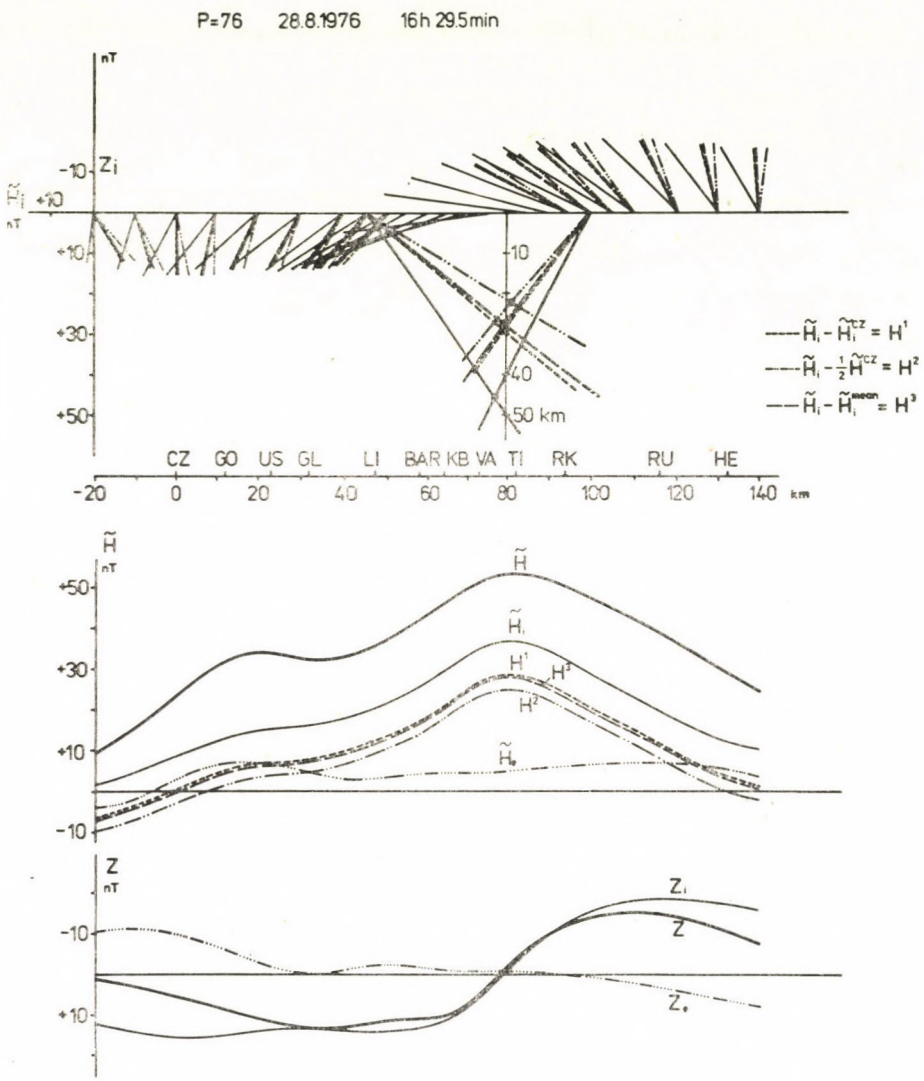

Fig. 5. Geoelectric anomaly source depth estimated by separating the variation field into external and internal parts. Hilbert transform was applied to input data of $\tilde{H}$ (horizontal geomagnetic field component projected into the azimuth of the profile) and Z (vertical component of the geomagnetic field) plotted as functions of location on the profile in the lower part of the figure. As a result of transform, separated internal (Z_i, $\tilde{H}_i$) and external (Z_e, $\tilde{H}_e$) fields were obtained (lower part of the figure). From the geometry of the total internal field (vector summation of Z and $\tilde{H}_i$) the anomaly source depth was obtained according to Biot-Savart law by intersecting the perpendiculars to internal field vectors on both sides of inversion zone (the upper section of the figure). Correction for a regional normal component of the horizontal field was introduced in different ways (corrected fields H^1, H^2, H^3 in the upper part of the Figure). Mention a substantial reduction of anomaly source depth from corrected fields as compared to results, where no corrections were introduced (thick lines in the upper part of the figure intersecting between the depths of 40 to 50 km)

take place especially in presence of water which substantially lowers the initial melting temperature. This mechanism was discussed by ČERMÁK et al. (1976) in connection with layers of increased electric conductivity suggested by the results of magnetotelluric soundings.

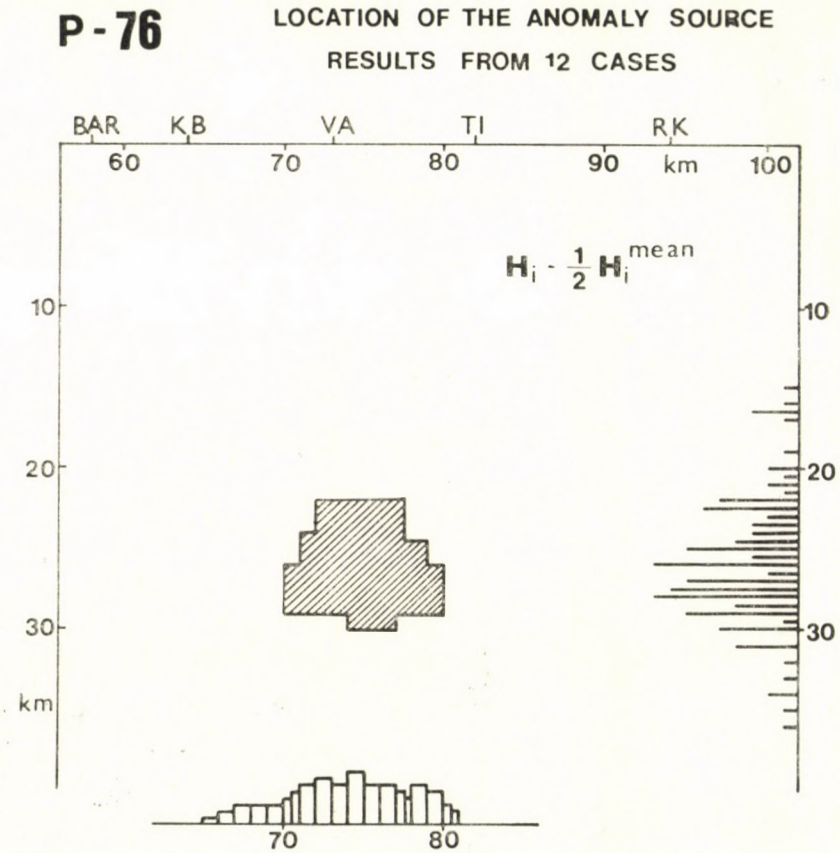

Fig. 6. Summary of depth estimates and location of anomaly source on the profile P-76 (Fig. 1) as obtained by separating 12 simultaneously recorded variation events. Horizontal geomagnetic field corrected for normal regional component was used in estimating the depths. Hatched area in the figure represents a scatter domain of individual estimates. There is no connection between the pattern of the area and the cross-section of assumed anomaly source

Assuming this mechanism as one possibility, we would expect relatively close correlation of the geoelectric anomaly with the results of geothermal activity. For this purpose, we made use of the heat flow map compiled for the European area by ČERMÁK and HURTIG (1977) with a large group of co-authors. Confining ourselves to the Carpathian area, we marked the geoelectric anomaly into a schematized version of the heat flow map as shown in Fig. 7.

In the southern part of the contact zone between the Bohemian Massif and the West Carpathians the geoelectric anomaly coincides with an area of relatively low heat flow values between 50—60 mW/m². In the northern section of this contact zone a marginal part of increased heat flows of the order of 70 mW/m² is crossed by induction anomaly. Further eastwards the induction

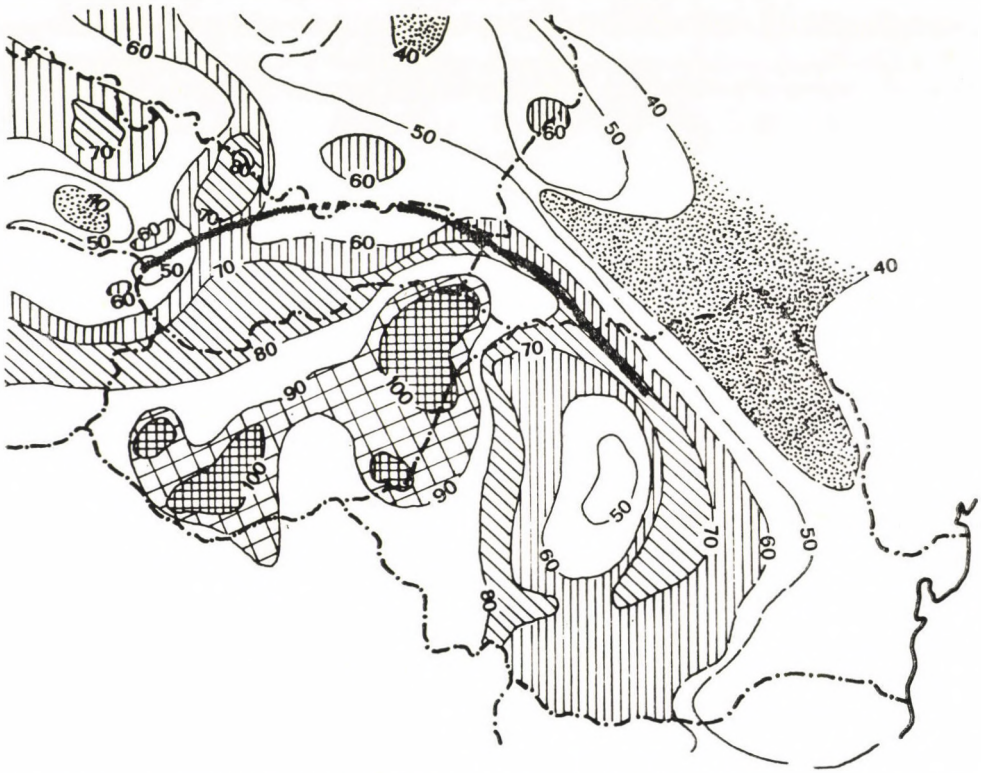

Fig. 7. Schematized heat flow map (isolines in mW/m²) of the Carpathian region according to ČERMÁK and HURTIG (1977). The geoelectric anomaly is shown as a cross-hatched zone

anomaly traverses again an area of relatively low heat flow values (50—60 mW/m²). Only in the East Slovakian and Carpatho–Ukrainian region the anomaly merges with a belt of increased heat flow values (between 70—80 mW/m²), and its course fits into the general pattern suggested by heat flow isolines. This comparison, though being only qualitative, does not suggest unique and very close correlation of both phenomena.

The surface heat flow, however, need not be a decisive indication of anomalous temperature conditions at crustal depths. Structural pecularities can cause local temperature enhancements.

Extrapolation of the superficial heat flow observations together with some models of the thermal conductivity and heat production allow the evaluation of crustal temperatures as a function of depth. Such model calculations for certain areas with typical internal structure in Czechoslovakia were performed by ČERMÁK (1975). Recently, ČERMÁK (1978) has solved the temperature—depth distribution along two profiles traversing the transition zone between the Bohemian Massif and the West Carpathians. The geological cross-section and the structure of the Earth's crust was suggested by CHMELÍK (1976) on the basis of extensive geological and geophysical information.

We have used some of these results as an independent information for speculating about another mechanism which may have produced a zone of increased electric conductivity at crustal depths. Main results are summarized in Fig. 8. It represents the geological cross-section by CHMELÍK (1976), isotherms calculated by ČERMÁK (1978) and our estimate of the geoelectric anomaly location in the cross-section. By analyzing this graph, we can conclude that

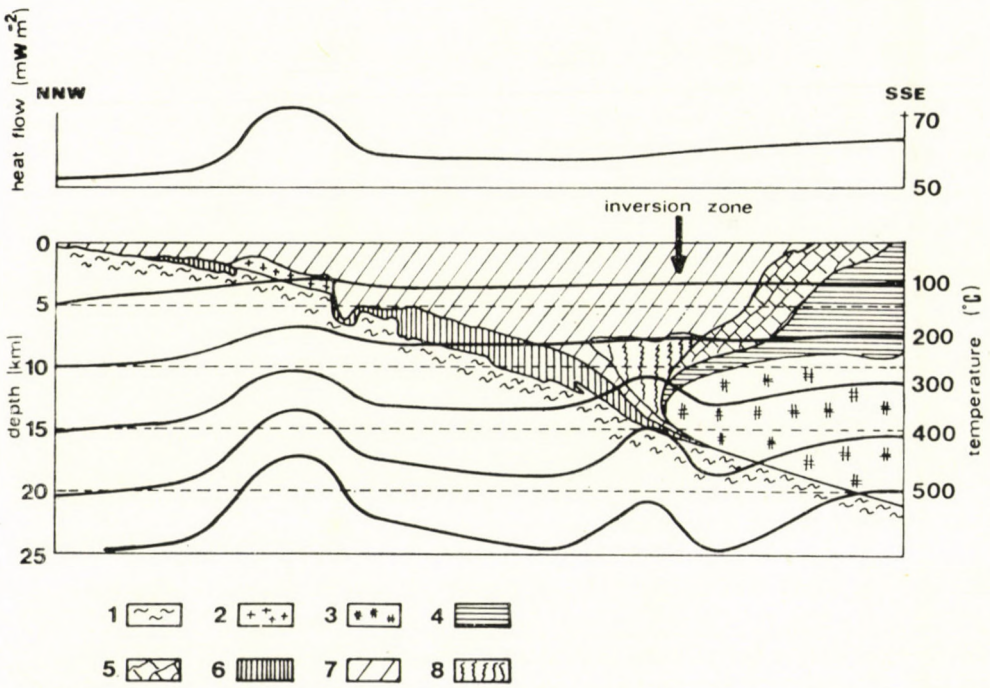

Fig. 8. Schematized geological cross-section of the contact zone between the Bohemian Massif and the Carpathians by CHMELÍK (1976); isotherms are by ČERMÁK (1978). The position of geoelectric anomaly (inversion zone) is shown by arrow. 1—2. The crystalline platform basement of the southeastern slopes of the Bohemian Massif, metamorphites + granitoids; 3. the crystalline of the Carpatho–Pannonian block (the Slovakian block); 4. sedimentary formations of the Central Carpathians, mainly carbonates; 5. Klippen Belt, mainly carbonates; 6. sedimentary cover of the platform of the southeastern slopes of the Bohemian Massif (clays, schists, etc.); 7. flysch formations of the Outer Carpathians; 8. extremely crushed crystalline (mylonites, clays)

(i) the geoelectric induction anomaly seems to mark the extreme northwest position of the Slovakian block overthrusted to the southeastern slopes of the Bohemian Massif and covered by flysch sediments (in the West-Carpathians);

(ii) at depths between 10—15 km on geological basis CHMELÍK assumes the existence of an extremely crushed crystalline block formed by mylonites and clays; due to low thermal conductivity of mylonites and clays, local enhancement of temperature can be expected, as seen from the isotherms according to ČERMÁK's calculations; this local thermal field cannot penetrate as deep as 25 or 30 km.

The enhancement of temperatures in the left-hand side of Fig. 8 is belived to be not realistic. It is a consequence of anomalous heat flow values which are caused by surface effects and no deep-seated source is involved.

Conclusion

Regional location of the geoelectric induction anomaly has been refined on the basis of recent geomagnetic field observations on profiles traversing the Czechoslovak—Polish section of the Carpathians. Certain corrections in geoelectric anomaly location are suggested in the High Tatra region (Fig. 1).

Relation of the anomaly to seismic activity and heat flow values distribution over the Carpathian region is investigated qualitatively (see Figs 2, 3, 4 and 7) and nature of the geoelectric anomaly is discussed in correlation with geothermal results.

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КАРПАТСКАЯ ГЕОЭЛЕКТРИЧЕСКАЯ АНОМАЛИЯ И ЕЕ ОТНОШЕНИЕ К НЕЗАВИСИМОЙ ГЕОФИЗИЧЕСКОЙ ИНФОРМАЦИИ

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Р е з ю м е

Исследуется аномалия элечтропроводности обнаруженная в окрестности границы между Внешними и Центральными Карпатами по особом поведению вариаций геомагнитного поля вдоль профилей пересекающих Западные и Восточные Карпаты на территориях Польши и Чехословакии.

По имеющимся геомагнитным данным глубина залегания источника аномального поля оценивается на 15—25 км.

Обсуждается природа, геофизическое дначение аномального геоэлектрического эффекта и его возмофная связь с другими геофизическими параметрами (сейсмической активностью, геотермическими данными).

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THE INVERSE MAGNETOTELLURIC PROBLEM FOR SOUNDING CURVES IN THE CARPATHIAN SECTION OF DSS PROFILE No. VI

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The inverse magnetotelluric problem has been solved for sounding curves obtained by analyzing the electromagnetic variations at field sites approximately on DSS profile No. VI. We confined our attention to curves from localities covering the inner Carpathian part of the profile and the transition zone between the Bohemian Massif and the Carpathians.

In fitting certain geoelectric model to sounding curve at each locality, we applied the "Hedgehog" method. It uses, in the first step, the Monte Carlo procedure to find certain successful model. Then, it goes systematically through the values of both the electric conductivities and layer thicknesses varying within given limits in the neighbourhood of the previously found successful model. The families of the best fitting models at individual sounding sites are compared and differences in the geoelectric structure of both geological regions are investigated.

Introduction

In the preceding years the magnetotelluric soundings at 12 field stations along DSS profile No. VI were carried out on the Czechoslovak territory. The results of measurements and their interpretation were summarized in the paper by PĚČOVÁ et al. (1976). In our further work in this direction we aim at re-interpretation of results by using the Hedgehog inversion method.

We concentrated our attention to magnetotelluric sounding curves (MTS) from the contact zone between the Bohemian Massif and the Carpathian system and its surrounding. This area limited by the Lednice fault zone and the Klippen Belt has an interesting, but complicated geological structure. It is assumed from the results of magnetovariational profiling that a high-conductivity zone at a depth of about 20 km extends to the outer side of the Carpathian arc approximately parallel to the Klippen Belt. We attempted to show the difference in the structure of basic geological units which this specific area consists of. In a preliminary interpretation, we could estimate only the depths of layers without pointing out their conductivities. In this new approach we are able to follow the layers with similar conductivities in individual geological units.

The localization of magnetotelluric field stations Šrobárová, Gabčíkovo, N. Blatnica, Prievaly, Hodonín-Pánov and Šemíkovice chosen for new interpretation is shown in Fig. 1.

The Hedgehog Inversion for MTS Curves

The inversion problem for the above mentioned MTS curves was solved by the assumption that a model consisting of N horizontal homogeneous layers would be a good approximation of internal geoelectric structure. The bottom layer with subscript N was assumed to be a half-space of infinitely high conductivity. As parameters of the model the resistivity ratios ϱ_i/ϱ_1, i.e. resistivities normalized to those of the surface layer, and layer thickness ratios h_i/h_1 were chosen. The resistivity ϱ_1 and the thickness h_1 of the surface layer were estimated independently.

Under these assumptions, our model is described by $2(N-2)$ parameters $h_2, h_3, \ldots h_{N-1}, \varrho_2, \varrho_3, \ldots \varrho_{N-1}$. Procedure for solving the direct problem was available so that we might calculate directly theoretical master curves for a given set of parameters.

The problem consists in choosing the parameters so that the solution of the direct problem may be as close as possible to the measured resistivity curves. For estimating the closeness of the fit, we defined a difference function $D(h_i, \varrho_i)$. This difference function is non-negative. It is zero if perfect coincidence of both the theoretical and experimental curves occurs. It increases if the theoretical and experimental curves are getting different. In our case, the mean square deviation was chosen for the difference function. Using this function as a criterion, we may divide calculated models into the class of successful models which meet the condition $D(h_i, \varrho_i) \leq A$ $A > 0$) and into a class of rejected models which do not meet this condition.

To solve the inversion problem, we were obliged, first, to specify roughly a region in which the parameters of successful models can lie. It means that for each parameter $p_i (p_1 = h_2,\ p_2 = h_3, \ldots, p_{N-2} = h_{N-1},\ p_{N-1} = \varrho_2, \ldots, p_{2(N-2)} = \varrho_{N-1})$ we define the lower and upper bounds, i.e.

$$B_i \leq p_i \leq C_i \quad (i = 1, 2, \ldots, M) \quad (M = 2(N-2)).$$

The constant A was assessed intuitively at the start of the procedure. In the course of the calculations, which were performed by separate blocks consisting roughly of 500 successful models, the constant A was made smaller from one step to the other to set the parameters of successful models closer to local minimum of difference function D. The initial values of parameter bounds B_i, C_i were estimated sooner by trial and error method. These bounds were

also changed during the process of calculation to satisfy better the conditions of the problem.

For the inversion the Hedgehog method was used. It was first referred by VALJUS (1968) who applied it to inversion of seismic data. Each interval B_i, C_i of individual parameters was divided by a grid of points sufficiently

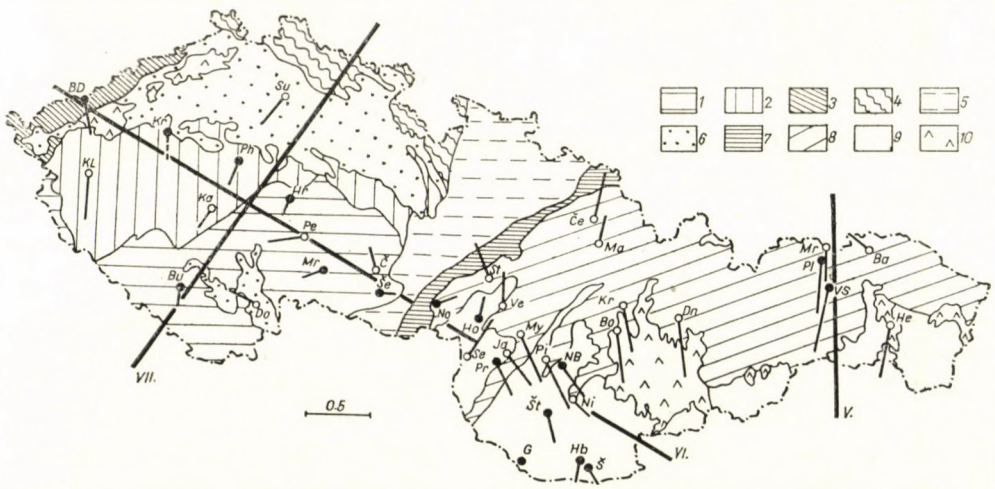

Fig. 1. Geological map of Czechoslovakia with DSS profiles Nos V, VI and VII. Full circles indicate the Czechoslovak field stations, open circles specify observation sites of the GDR groups (H. WIESE and E. RITTER). Induction vectors as characteristics of electromagnetic induction in deep parts of different areas are displayed. The symbols for the stations the sounding curves of which were analyzed by Hedgehog inversion: Š — Šrobárová; G — Gabčíkovo; NB — Nitrianská Blatnica; Pr — Prievaly; Ho — Hodonín (Pánov); No — Nosislav; Še — Šemíkovice. Geological symbols: Bohemian Massif: 1. Moldanubian block; 2. Assyntian block; 3. Saxo–Thüringian block; 4. West Sudetan block; 5. Moravian–Silesian zone; 6. Cretaceous and Tertiary platform deposits; 7. Alpine–Carpathian foredeep. Carpathian region: 8. Pre-Neogene folding of the Carpathian unit; 9. Neogene basin of the Carpathian region; 10. Tertiary eruptive rocks

dense so that each parameter p_i ($B_i \leq p_i \leq C_i$) might get only values corresponding to the grid points called the knots.

The Hedgehog inversion starts with the application of the Monte Carlo method. The parameters p_i are randomly chosen within the intervals (B_i, C_i). The direct problem is calculated for each set of randomly chosen parameters and the result is tested by the difference function criterion. If a set of parameters corresponds to an unsuccessful model, the procedure goes on by the Monte Carlo method. As soon as a successful model is found, the control of the calculations passes from the Monte Carlo subroutine to Hedgehog method itself. Contrary to the Monte Carlo method which is blind in the sense that the search for the following model is independent of the preceding step, the Hedgehog method performs a systematic search for successful models in the neigh-

bourhood of the first good knot. In other words, the Hedgehog method examines systematically the neighbour knots and tests whether the solution is also successful.

Let $P = (p_i)$ be a knot of a given grid. Then the neighbour knot can be expressed as

$$P' = P + \sum_{i=1}^{M} \varepsilon_i \, \delta p_i, \quad M = 2(N - 2),$$

(M . . . dimension of the parameter space)

where δp_i is grid spacing and ε_i are numbers which can get values $+1$, 0, -1. It is easy to see that there is a lot of knots in a more dimensional parameter space ($3^M - 1$ neighbours for M dimensions). From the point of calculation efficiency, it is preferable to reduce the number of neighbour knots to a given rank. We define the rank $r (r \leq M)$ which describes the number of ε_i different from zero. The neighbours of the rank r include all neighbours of the lower rank. During our calculations we restricted ourselves to the neighbours of rank 1. In this case all neighbour knots lie on the co-ordinate's axes.

In the process of calculation each examined model was put on the list of models. By systematic examination of neighbour knots, each new one is checked against the list and decided whether it was already calculated. If all successful neighbour knots are exhausted, the control of the program is taken over by the Monte Carlo method again.

Application of the Hedgehog Inversion Method

The normalized sounding curves for four stations of Šrobárová, Gabčíkovo, Prievaly and Pánov-Hodonín are shown in Figs 2—5. The first two curves belong to the Pannonian Basin on the inner Carpathian side. The other ones characterize the northern part of the Vienna Basin on the outer side of the Carpathians.

The next four Figs 6—9 show the conductivity models for stations mentioned above. Each model represents the result from a block of computation for given constant A. The full line model shows the most frequent case in the last block of calculations, the dashed line refers to the best fitting model. The dot-and-dashed line sets the bounds of depths and conductivity values. The width is different for individual models. It depends not only on the value of the constant A, but also on the dispersion of experimental data and, finally, on the compatibility of realistic conditions in the basement with theoretical assumptions. From Fig. 8 for the Prievaly station, we can get an idea about the depths and conductivity limits for which all calculations were performed.

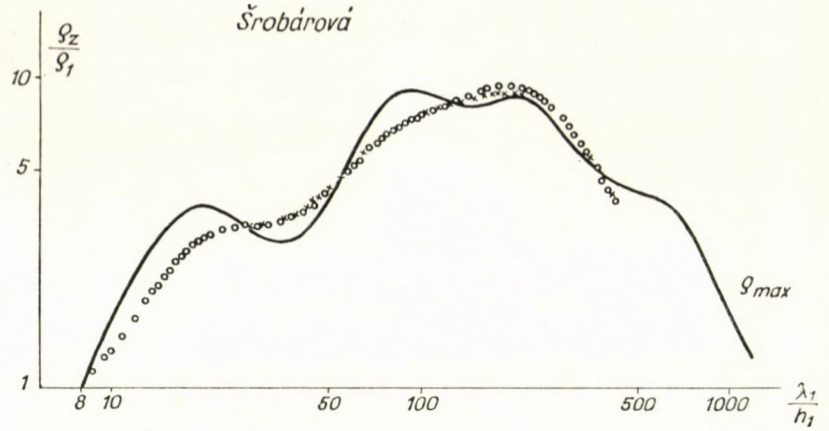

Fig. 2. Normalized MTS curve ϱ_{max} in the azimuth of maximum resistivity for the station at Šrobárová (full line). The curve marked by open circles represents the resulting curve for the most frequent theoretical model. The curve with crosses represents the best fitting theoretical model

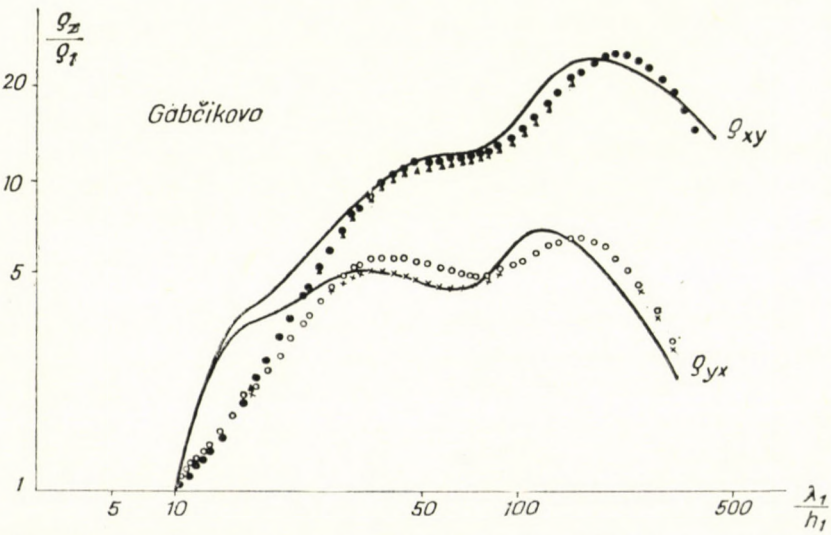

Fig. 3. Normalized MTS curves ϱ_{xy} and ϱ_{yx} for the station of Gabčíkovo. Theoretical curves marked with open (full) circles, or with crosses (triangles) represent the same types of models as in Fig. 2

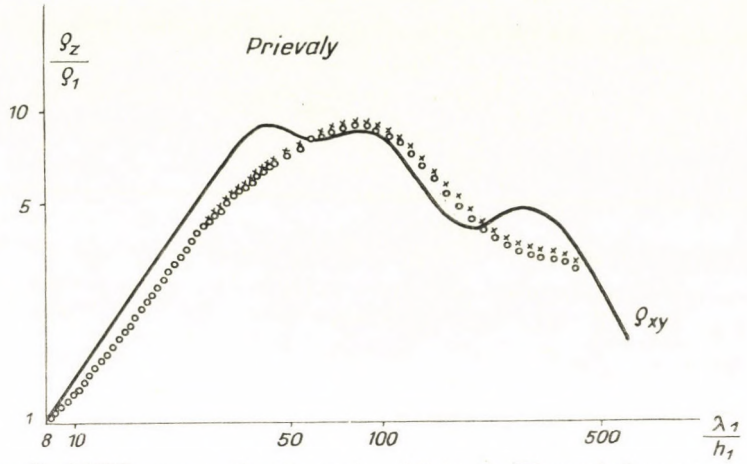

Fig. 4. Normalized MTS curve ϱ_{xy} for the station of Prievaly. Theoretical curves are marked in the same way as in Fig. 2

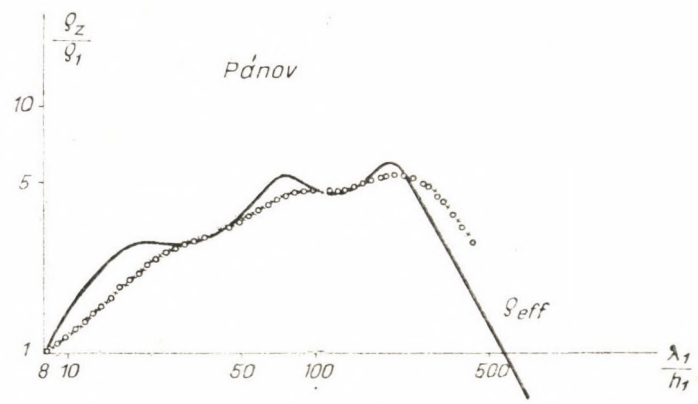

Fig. 5. Normalized MTS curve ϱ_{eff} for the station of Pánov (Hodonín). Theoretical curves are marked in the same way as in Fig. 2

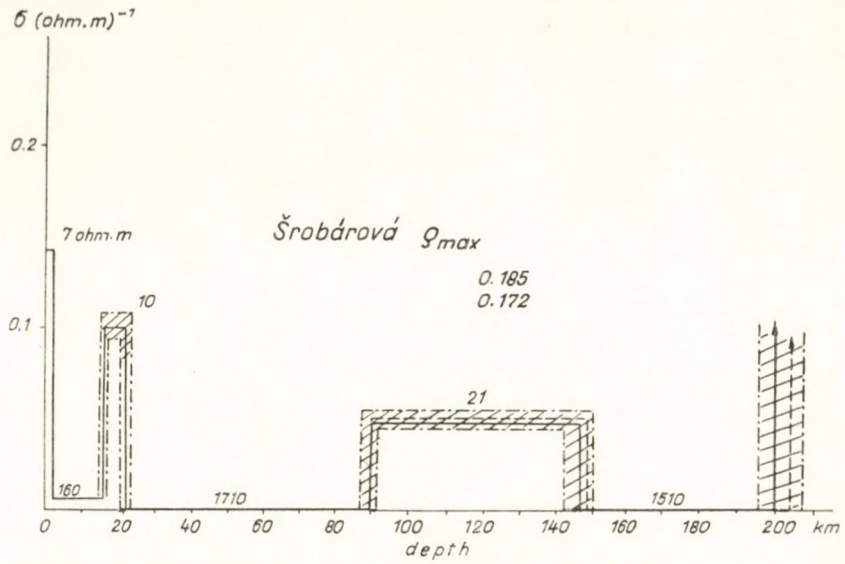

Fig. 6. Conductivity model for the station of Šrobárová. Two figures below the station name represent the sum square deviations for a given constant A. The upper one relates to the whole block of computation, the lower one corresponds to the best fitting model. Specific resistivities (in ohm · m) are shown for each layer

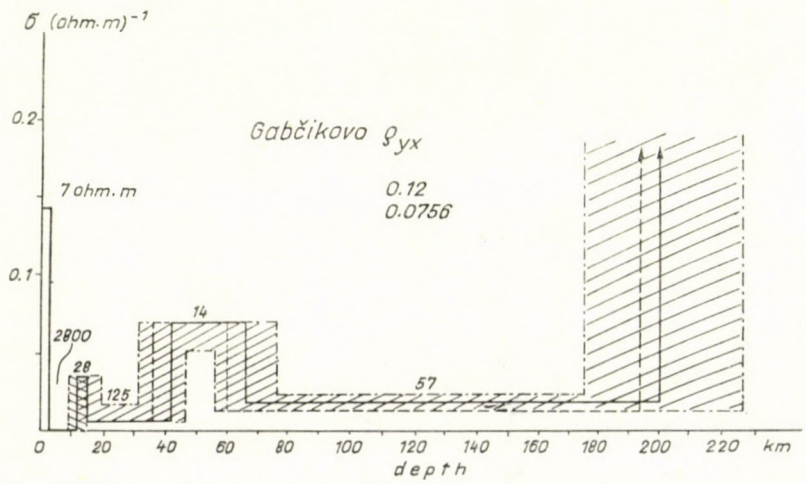

Fig. 7. Model of electric conductivity distribution for the station of Gabčíkovo

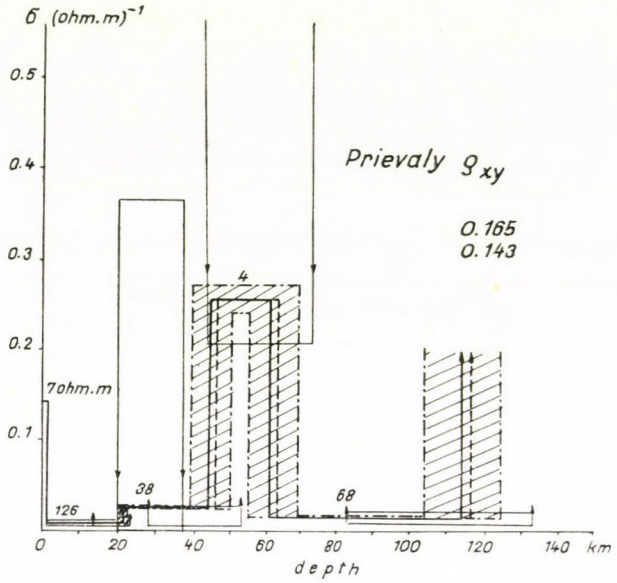

Fig. 8. Model of electric conductivity distribution for the station of Prievaly

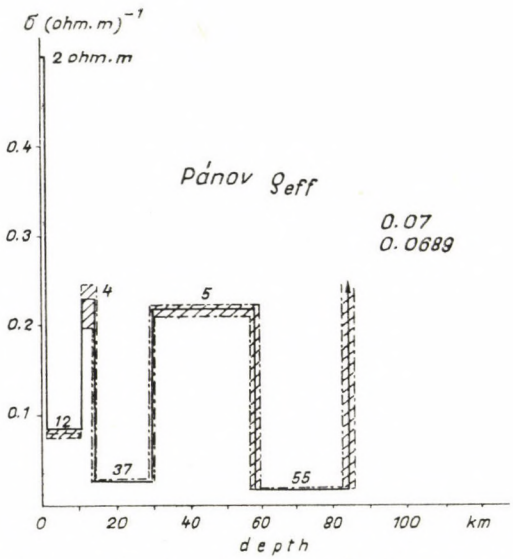

Fig. 9. Model of electric conductivity distribution for the station of Pánov (Hodonín)

Station	Type of curve	Number of layers	$\Sigma(\ )^2$	Number of models
Šrobárová	ϱ_{max}	7	0.76—0.185	2 880
Gabčíkovo	ϱ_{xy}	7	1.0—0.130	2 635
Gabčíkovo	ϱ_{yx}	7	0.222—0.120	2 145
N. Blatnica	ϱ_{eff}	5	0.600—0.046	21 320
Prievaly	ϱ_{xy}	7	0.500—0.165	4 455
Pánov	ϱ_{eff}	7	0.110—0.070	3 285
Nosislav	ϱ_{eff}	5	0.400—0.092	15 150
Šemíkovice	ϱ_{xy}	7	0.550—0.160	2 000
Šemíkovice	ϱ_{yx}	7	0.180—0.041	3 455

The table displays the summary of all calculated successful models. Only those cases are included which lie within the limits of the mean-square deviation (constant A). Majority of these curves is represented by 2000 to 5000 models, except for N. Blatnica and Nosislav.

Sounding curves calculated according to direct problem formula for parameters obtained by Hedgehog inversion are shown in Figs 2—5 with circles and crosses for the most frequent and the best fitting models, respectively. We can conclude from comparing all three curves that the inverse method led to resulting models and curves which partly agree with input curves, partly are smoothed. A very important step is the estimation of surface layer parameters. For their assessing we used the results of vertical electric *dc.* soundings and bore-hole measurements. In several cases these measurements were carried out far from the field stations or we had no geophysical information. In the latter case assessment was made on the basis of geological conditions. Parameters of the surface layer can be improved by using proper results of inverse method. Comparison of experimental and theoretical curves can give an additional relation between ϱ_1 and h_1.

In performing the calculations, we have not respected the importance of individual period ranges as yet. In the future work, we intend to put different weights on different period ranges of sounding curves in accordance with frequency of observations.

Interpretation of Inversion Results

In Fig. 10 models for the area of contact zone are summarized. We shall concentrate our attention to two important geological units. The Pannonian Basin is represented by curves from Šrobárová, Nagycenk and Gabčíkovo. The

Vienna Basin is characterized by curves from stations at Prievaly, Pánov and Nosislav. The curve from N. Blatnica, which was situated at one of the belts of the Carpathian mountains, can be considered for a certain transition curve between both units.

We can see a good similarity between the models from Šrobárová and Nagycenk, both obtained from sounding curves in the azimuth of maximum resistivity. The first conductive layer at the depth of 16 km with the same specific resistivity can be realistic. The distance between both stations is 125 km and the effect of anisotropy in the basement is different. Therefore, this layer cannot be a fictitious one caused by influence of a superficial structure. Existence of a similar layer at small depth was confirmed by authors from the Soviet Union in several areas.

The second conductive layer at depths ranging from 70 to 100 km was suggested by ÁDÁM and VERŐ (1970) at many curves from the Pannonian Basin. The depth of the third conductive layer varies within large range of depths from 200 to 500 km.

The sounding curves from Gabčíkovo station show the same anisotropy effect as it has been known in Nagycenk and other stations in Hungary according to ÁDÁM et al. (1968). The curve ϱ_{xy} gives similar parameters for the second conductive layer. The first layer is suppressed and the third layer is shifted to greater depths. Model for Gabčíkovo ϱ_{yx} curve yields the depths of layers shifted to shallower depths probably due to anisotropy. Here, the second layer is at the depth of 42 km. It is interesting that Soviet authors, BERDICHEVSKY et al. (1973) have found in three different areas of the Soviet Union a similar structure with high conductive layers at depths of 20 km and of about 50—60 km and have compared their results with those from the Pannonian Basin. It is probable that in the latter case the second layer lies somewhere between 40 and 90 km. In connection with the results from the Danube Basin we call the attention to conductive layers in the Vienna Basin. In models for Nadrlenisko (Prievaly) and Pánov (Hodonín) we can see a very pronounced conductive layer at the depths of 42 and of 30 km practically with the same conductivity. Conductive basement is suggested at depths of 115 and 85 km. The model from Nosislav shows two pronounced layers at 22 and 130 km depths too, but this station is situated already near the margin of the basin.

Question can be raised about the existence of two conductive layers in the crust and upper mantle in the Vienna Basin similarly as in the Danube Basin. This problem is important also in relation to the zone of anomalous induction marked between the stations of Pánov and Nadrlenisko by inversion of induction arrows as shown by PĚČOVÁ et al. (1976). The source of the conductivity anomaly which can be connected with these layers was found at depth of about 20 km and it could extend below the whole Carpathian belt.

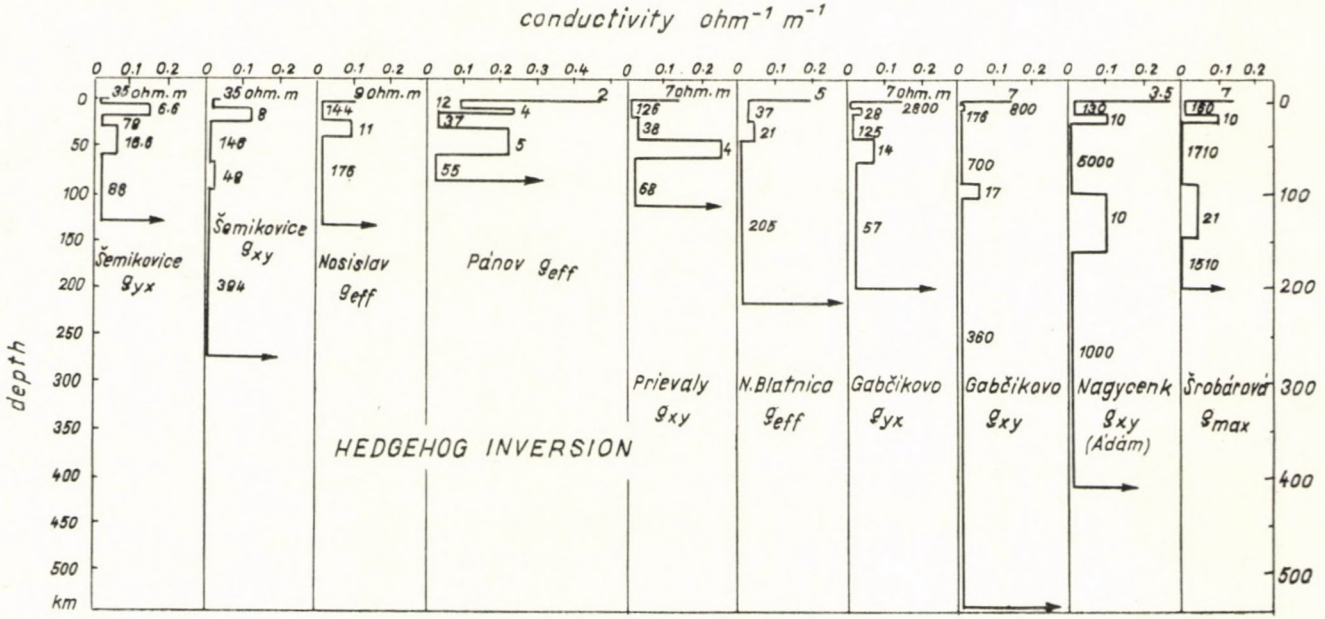

Fig. 10. Summary of geoelectric models for the area of the contact zone

Conclusion

The magnetotelluric sounding curves measured at field stations situated along the Czechoslovak part of DSS profile No. VI were subject to Hedgehog inversion procedure by VALJUS (1968) and KENNETT (1972). A theoretical model consisted of a sequence of horizontal layers conductivities and thicknesses of which were looked for. The main aim of new and detailed interpretation of MTS curves from the contact zone of the Bohemian Massif and the Carpathians consisted of investigating and comparing the resulting models for both the Pannonian and Vienna Basins on both sides of the contact. Basic features of geoelectric structure in the basement were compared with the results from similar areas in Hungary and in the Soviet Union.

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ОБРАТНАЯ МАГНИТОТЕЛЛУРИЧЕСКАЯ ЗАДАЧА ДЛЯ КРИВЫХ ЗОНДИРОВАНИЯ В КАРПАТСКОЙ ЧАСТИ ПРОФИЛЯ ГСЗ № VI

К. ПЕЧ, Й. ПЕЧОВА и О. ПРАУС

Резюме

Авторы решают обратную магнитотеллурическую задачу для кривых зондирования полученных обработкой записей колебаний естественного электромагнитного поля на полевых станциях расположенных приблизительно вдоль профиля ГСЗ № VI, именно в его внутренно-карпатской части и в переходной зоне между Чешским массивом и Карпатами.

Используется метод Hedgehog, который процессом Монте Карло избирает в начальном шагу параметры первой заданное условие удовлетворяющей модели. Потом, метод систематически подбирает в окрестности этой первой модели в заданных пределах меняюшиеся значения электрических проводимостей и мощночтей слоев и проверает, которая из этих моделей удовлетворяет заданное условие найлучшим образом.

В результате расчетов полученные для отдельных точек зондирования совокупности самых подходящих моделей сопоставляются и обсуждаются таким образом обпаруженные разницы между геоэлектрическим строением обеих геологических областей.

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A NEW PROCESSING METHOD OF TELLURIC CURRENT DATA

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The processing of telluric current data involves concepts from electromagnetic theory, time series analysis and linear systems theory for reducing natural electric field variations recorded at the earth's surface to forms suitable for studying the electric properties of the earth's interior.

This paper presents a processing method based on the spectral analysis which enabled the estimation of the complex spectral coefficients for the various field quantities.

Introduction

The telluric method involves the comparison of horizontal electric field measurements recorded simultaneously at a base site and a remote site. Let us suppose that at an instant t the orthogonal components of the electric field at a point B, called base site, are E_x, E_y and at the same instant, the components at another point R, called remote site, measured in another orthogonal coordinate system, are E_X, E_Y. It is assumed that these components are related to each other as follows:

$$\begin{vmatrix} E_X \\ E_Y \end{vmatrix} = \begin{vmatrix} a & b \\ c & d \end{vmatrix} \cdot \begin{vmatrix} E_x \\ E_y \end{vmatrix} \tag{1}$$

where: a, b, c and d are correlation coefficients which depend only on the directions of the measuring axes and on the electric properties of the subsurface; $|ad - bc| = J$ is the JACOBIAN of the transformation from the x, y system to the X, Y system and depends only on the electrical properties of the earth's interior, containing valuable geologic information.

The telluric method has evolved over a number of years through the notable effort of SCHLUMBERGER (1939), KUNETZ (1952), BONDARENCO (1955), BERDICHEVSKIY (1960), Y-SHU (1963) and YUNGUL (1966), but considerations concerning cost *versus* usefulness relationship have generally been unfavourable for telluric surveys.

One possible way of progress in the telluric method is to find the mode of lowering the cost without going into specialized and sophisticated processes.

The vectogram method (YUNGUL, 1968), which involves the polar diagram recording of the telluric field vectors by means of the so-called "X—Y" records is the first example in this direction.

Let us refer to Figures 1 and 2 which show one pair of simultaneous vectograms, representing real data, they would be recorded in about ten

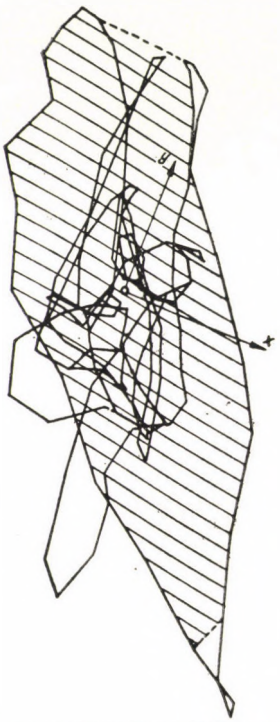

Fig. 1. The polar diagram of the telluric field vectors in the base site

minutes. The coordinate systems x, y and X, Y are oriented north—south. Such vectograms are traced with Hewlett—Packard "X—Y" plotter.

After YUNGUL, the parameter J at the remote site is numerically equal to the ratio of the hatched area at R (Fig. 2) to that at B (Fig. 1).

By means of this method, however, it is difficult to get the parameter J in areas where the telluric field is somewhat linearly polarized; this means that the majority of the field vectors is confined within a small-angle beam.

This paper presents a processing method based on the spectral analysis which has dominant advantage over classical methods.

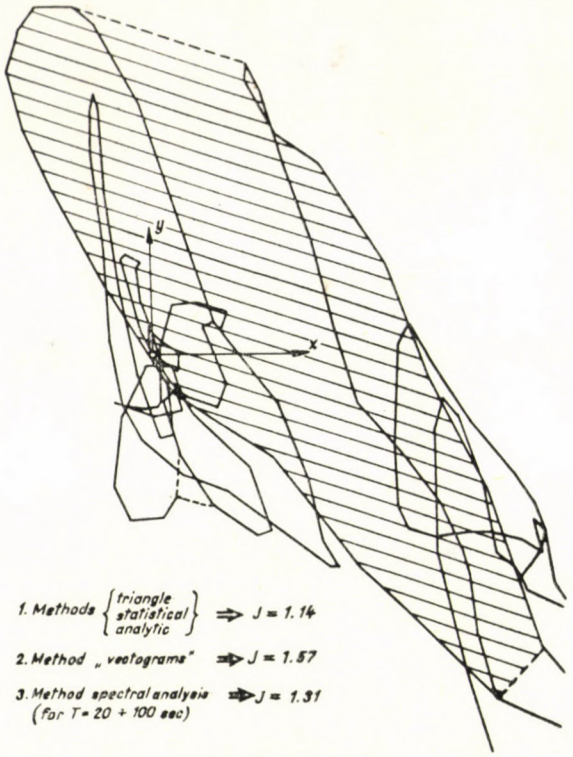

Fig. 2. The polar diagram of the telluric field vectors in the remote site

Data Processing

Spectral Analysis of Time Series

The interpretation and processing of telluric data are usually done in the time domain, but spectral analysis of the raw data represents an important aspect of data processing in the frequency domain. This involves creating a time series X_i, by sampling a signal $X(t)$, at equal intervals of time Δt, from $i = 1$ to N, where $T_0 = (N - 1)\ \Delta t$ is the duration of the signal.

In a more rigorous sense, an absolute requirement is that the signal to be sampled should be of negligible spectral energy at periods shorter than twice the sampling interval. There is a number of techniques for transforming the time series into spectral information in the frequency domain. We are going to present only FOURIER harmonic analysis.

In FOURIER harmonic analysis the fundamental assumption is that the time series of length T_0 is cyclic or periodic with fundamental period T_0, such that $X(t) = X(t + T_0)$. This assumption places strict constraints on the

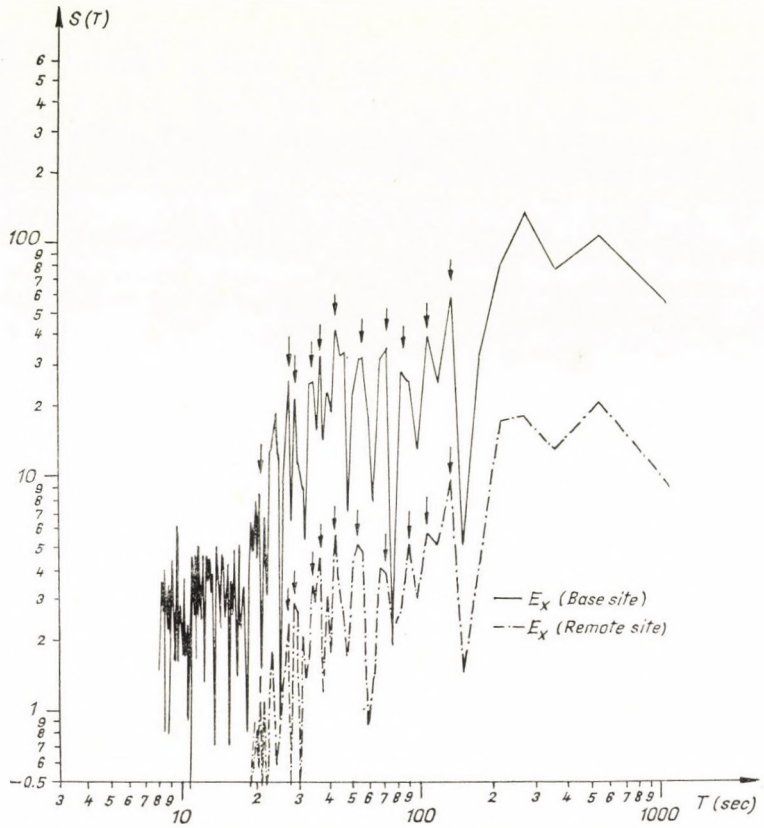

Fig. 3. Amplitude spectra of telluric field components an x direction for the base and remote sites

spacing between harmonics as well as implying that the record section can be optimally approximated over its length T_0 by a finite number of harmonics.

The output signal $U(t)$ of a telluric recording set is composed by a useful signal $X(t)$ always present together with the noise signal $Y(t)$:

$$U(t) = X(t) + Y(t) \tag{2}$$

The noise has several sources of instrumental, cultural and natural origin. One of the main target of data analysis is the extraction of the useful signal $X(t)$ from the recorded one $U(t)$.

Over an interval of time $(-T_0/2, T_0/2)$ a function $X(t)$ can be expressed as the sum of a FOURIER series:

$$X(t) = \sum_{K=-N}^{K=+N} \tilde{X}_k \, e^{ik\omega_0 t}, \quad \omega_0 = \frac{2\pi}{T_0} \tag{3}$$

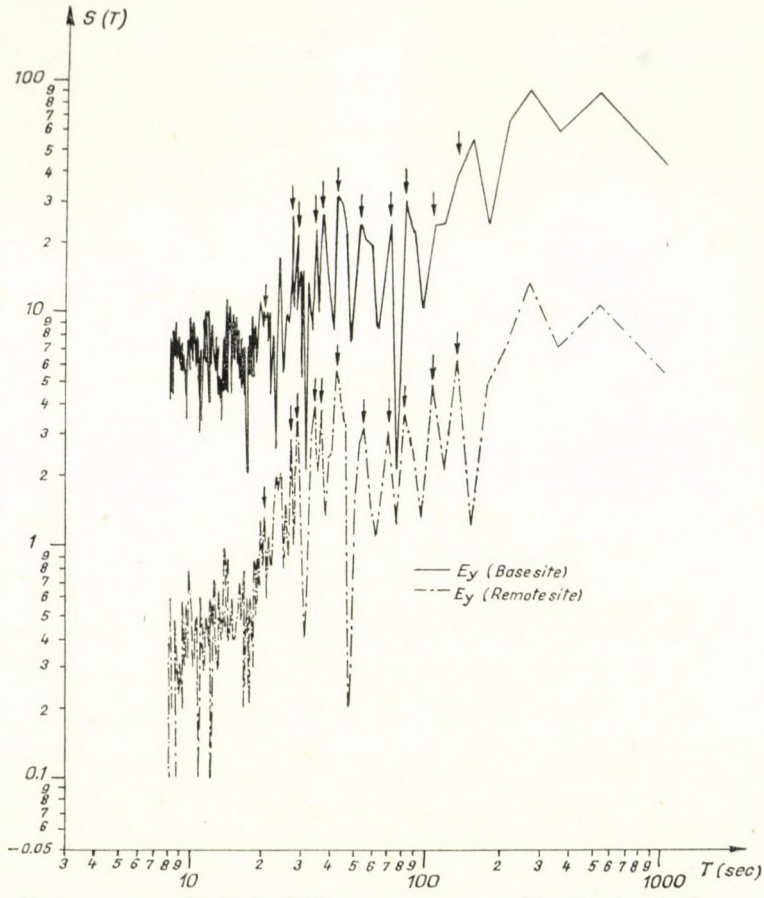

Fig. 4. Amplitude spectra of telluric field components on y direction for the base and remote sites

where:

$\widetilde{X}_k$ is given by:

$$\widetilde{X}_k = \frac{1}{T_0} \int_{-T_0/2}^{+T_0/2} X(t)\, e^{-ik\omega_0 t}\, dt \tag{4}$$

The amplitude spectral lines $\widetilde{X}_k$ are spaced at intervals k/T_0 along the frequency axis, which are harmonics of the record length T_0. Since we have required that the signal $X(t)$ should be band-limited to frequencies less than the folding frequency $(1/2\,\Delta t)$, and since $T_0 = (N-1)\Delta t$, the maximum harmonic is $k_{max} = (N-1)/2$.

After the linear effect elimination due to the electrode polarization and temperature differences between electrodes, the processing consisted in the spectral analysis of sampled data ($\Delta t = 4$ sec) using the FOURIER transform of a time function.

The Telluric Transfer Tensor

Using the FAST—FOURIER Transform (F.F.T.) algorythm the digital estimation of spectral components was realized and the results are shown in the Figs 3 and 4, where the amplitude spectra for the data sets E_x, E_X, E_y, E_Y are plotted versus period in seconds. The Figs obtained show that between amplitude spectra of the components E_x, E_X and E_y, E_Y for the base and remote sites there is a high degree of correlation and coherency.

By means of the amplitude spectra, it is allowed to rewrite the relation (1) as:

$$\begin{bmatrix} \widetilde{E}_X \\ \widetilde{E}_Y \end{bmatrix} = \begin{bmatrix} T_{Xx} & T_{xY} \\ T_{Yx} & T_{Yy} \end{bmatrix} \cdot \begin{bmatrix} \widetilde{E}_x \\ \widetilde{E}_y \end{bmatrix} \tag{5}$$

or in a more condensed form

$$\widetilde{E}^R = [T] \cdot \widetilde{E}^B$$

where: $\widetilde{E}_X$, $\widetilde{E}_Y$, $\widetilde{E}_x$, $\widetilde{E}_y$ are the amplitude spectra of the orthogonal electric field components in the base and in the remote sites; $[T]$ is the telluric transfer tensor. Except for the *dc* limit, the tensor elements in $[T]$ will be complex functions of frequency as well as position.

$$\begin{aligned} T_{Xx} &= \frac{\widetilde{E}_X \sin(\tilde{\varphi}_{E_y} - \tilde{\varphi}_{E_X})}{\widetilde{E}_x \sin(\tilde{\varphi}_{E_y} - \tilde{\varphi}_{E_x})} \\ T_{xY} &= \frac{\widetilde{E}_X \sin(\tilde{\varphi}_{E_X} - \tilde{\varphi}_{E_x})}{\widetilde{E}_y \sin(\tilde{\varphi}_{E_y} - \tilde{\varphi}_{E_x})} \\ T_{Yx} &= \frac{\widetilde{E}_Y \sin(\tilde{\varphi}_{E_y} - \tilde{\varphi}_{E_Y})}{\widetilde{E}_x \sin(\tilde{\varphi}_{E_y} - \tilde{\varphi}_{E_x})} \\ T_{Yy} &= \frac{\widetilde{E}_Y \sin(\tilde{\varphi}_{E_Y} - \tilde{\varphi}_{E_x})}{\widetilde{E}_y \sin(\tilde{\varphi}_{E_y} - \tilde{\varphi}_{E_x})} \end{aligned} \tag{6}$$

and

$$\tilde{\varphi}_{E_X}, \tilde{\varphi}_{E_Y}, \tilde{\varphi}_{E_x}, \tilde{\varphi}_{E_y}$$

are the phase spectra of the orthogonal electric field components in the base and in the remote sites.

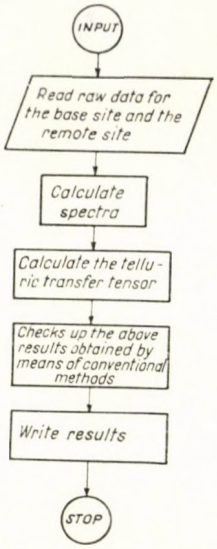

Fig. 5. Schematic of an algorithm for estimating parameters J and μ

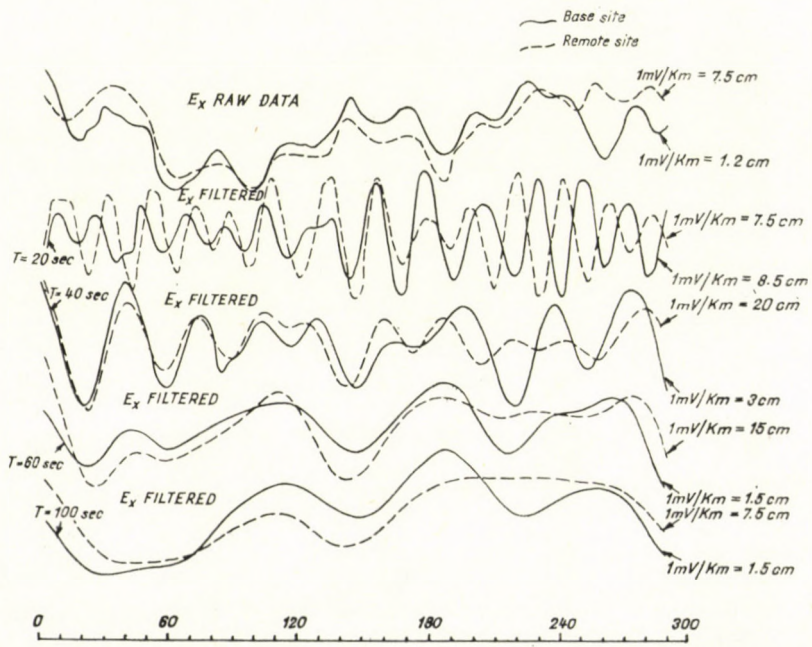

Fig. 6. Telluric components band-pass filtered at 20; 40; 60; 100 sec. periods. The comparison of E_x electric component recorded simultaneously at a base site and a remote site. Time, seconds

Estimation of the Telluric Parameters

On the basis of such considerations the parameter J may be rewritten as function of frequency:

$$J(f) = T_{Xx} \cdot T_{Y_y} - T_{Y_x} \cdot T_{xY} \tag{7}$$

where:

f — frequency [hertz]

For the telluroparameter μ we have used the following relation:

$$\mu(f) = \frac{\tilde{E}^R(f)}{\tilde{E}^B(f)} \tag{8}$$

where:

$$\tilde{E}^R(f) = \sqrt{\tilde{E}_X^2(f) + \tilde{E}_Y^2(f)}$$

$$\tilde{E}^B(f) = \sqrt{\tilde{E}_x^2(f) + \tilde{E}_y^2(f)}$$

Having described in the last section the method for converting time series into spectral amplitude information, we now present in Fig. 5a schematic of an algorithm that we have found useful for estimating the tensor elements of equation 6 and the parameters J and μ from spectral amplitudes of the telluric field components.

A second estimation way of the telluric parameters consists in the computation from filtered traces at prevailing periods that appear in the spectra.

For calculating digital band-pass filtred data we have used the convolution between the input time series and a filter impulse response.

Convolution filtering is an algorythm based on the convolution integral:

$$U(t) = \int_{-\infty}^{+\infty} h(t - b) \cdot X(b)\, db \tag{9}$$

where $X(t)$ is the original time series, $h(t)$ is the impulse response function of the filter function and $U(t)$ is the output of the filtering operation.

Figure 6 shows the comparison between E_X and E_x components recorded simultaneously in the remote and in the base sites, band-pass filtered at 20, 40, 60, 100 sec · periods.

The correlation between the output filtered series E_X and E_x for 40, 60 and 100 sec · periods is quite obvious. We have eliminated the high-frequency fluctuations as well as long-period trends and the output is not only smoothed, but it is band-limited.

The whole processing work was done by an I.B.M. 370 computer, using a Fortran IV programme whose schematic flow-chart is shown in Fig 5. A num-

ber of subroutines are realized for the linear detrending, for processing of spectra, of the telluric transfer tensor, of parameters J and μ, as well as for the filtering process.

Conclusions

The dominant advantages of the processing method based on the spectral analysis and outlined here are the following.

First, the time required for a telluric recording in each station is reduced to ten minutes or less.

Second, the telluric data are analyzed at different periods and, consequently, it is possible to eliminate the screen horizons effect.

Third, there is the possibility to apply the telluric-magnetotelluric method, which supposes magnetotelluric measurements at the base site, but only telluric measurements at remote sites.

Finally, this method allows an interpretation in quantitative terms of the telluric data.

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НОВЫЙ МЕТОД РАЗРАБОТКИ ДАННЫХ ТЕЛЛУРИЧЕСКОГО ТОКА

Д. СТАНИКА, М. СТАНИКА и Р. ПЛАВИЦА

Резюме

Разработка данных теллурического тока охватывает понятия теории электромагнетизма, анализа временных серий и линеарной теории систем. Таким образом изменения природного электрического поля загеристрированные на поверхности земли превращаются в коивые поджодящие для изучения электрических особенностей недр.

Излагается метод основанный на спектральном анализе временных серий, позволяющий оценку комплексных спектральных коэффициентов для различных полевых значений.

Acta Geologica Academiae Scientiarum Hungaricae, Tomus 21 (4), pp. 373—384 (1977)

DEVELOPMENT OF THE DESTRUCTIVE PROCESSES IN A LARGE EARTHQUAKE SOURCE USING THE *P*- AND *S*-WAVE KINEMATIC AND DYNAMIC PARAMETERS

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The destructive process in an earthquake source has been analysed using the Daghestan earthquake of May 14, 1970 as an example. The seismic wave picture from the large earthquake has been interpreted in terms of the concept on discrete development of the destructive source process. Source mechanism parameters for some consistent in time displacements have been used.

The Wave Picture Analysis

Basing on the seismogram analysis of the Daghestan earthquake recorded by short-period (BENIOFF and CKM-3 seismographs) and intermediate-period (Kirnos-CK and CKD seismographs) intruments, (GORBUNOVA, 1977; GORBUNOVA, SHATORNAYA, 1977) provided a hypothesis for discrete development of the destructive process in an earthquake source. According to this hypothesis, the source process seemed to develop impulsively as discrete shocks dynamically related to each other in time and in space. The onset of destruction corresponded to a small rupture gradually increasing in strength to maximum when the maximum energy release had taken place. In the initial part of a record *P*-wave could be partitioned into a few groups of different intensity increasing with time and provided a picture of destruction like that given by REID (1933). "The surface where the rupture started was small in area, then it was enlarged very soon. These relative displacements did not occur instantaneously but reached their maximum gradually during a more or less prolonged period of time."

One can clearly distinguish on the Daghestan earthquake records three or four arrivals in the *P*-wave group from the beginning of record to its maximum phase not corresponding to generally accepted hodographs. These arrivals were well seen on records of short-period and intermediate-period instruments because short-period and intermediate-period radiation (0.5—15 sec) reflects best of all the discrete development of the main displacement in the earth-

quake source of such a magnitude. Long-period instruments integrated these swings and one could see on a record one long-period swing with a period $T = 12—15$ sec.

Seismograms of the Daghestan earthquake show that the first P-wave arrivals are very weak, hardly noticable and distinguishable. A few seconds later, they are followed by more distinct arrivals of great intensity which, according to the accepted hypothesis, correspond to a new point of disturbance on the way of rupture. The character of records changes in arrivals similar to the onset of a new disturbance, namely the amplitude of a new arrival increases abruptly after the noticable decay of a previous swing. At some seismic stations equipped with high-magnification instruments one could see in the initial part of records phases with earlier arrivals which are not available on records of low-magnification instruments.

Figure 1 (a, b, d, d) shows the Daghestan earthquake records from various instruments set up at different azimuths from the earthquake source (quarters I—IV). A step-like increase of P-wave amplitudes could be seen first of all on records of short-period and intermediate-period instruments at different azimuths and irrespectively of distance. The "Yakutsk" and "Vladivostok" records show that CK instruments (magnification 1000) detect only the maximum phase in P-wave. One could see well P_{max} arrival of reverse sign in respect to P on a low-magnification component at the "Moskva" seismic station. Afterwards, in order to determine source mechanism, we used only data which had the first and ensuing arrivals distinctly expressed in P and S waves on a seismogram (Fig. 2a, b).

However, it should be noted that in this interpretation we always deal with the interference picture which in each separate case may break the harmony of the above description.

In order to distinguish P arrivals from different points of disturbance in a large source area to make sure of their source generation, different criteria were applied to distinguish them from pP or sP or any other waves whose generation was related to the structure of the earth's interiors (PcP, etc.).

The following kinematic and dynamic indications served as a criterion for the fact that waves in the initial part of records (P wave) were the source but not secondary ones due to the medium structure:

1. The intensity of reflected and refracted waves is smaller than that of the main wave (BULLEN, 1966). In the case of the Daghestan earthquake each ensuing arrival had amplitudes considerably exceeding the previous ones. From this point of view only the sP wave is questionable. Its hodograph varies by 1—2 sec at a distance from 20 to 90°. One can see in our case the azimuthal dependence between times of P_i-wave arrivals and times of P_i-phase delay reaching 5—13 sec.

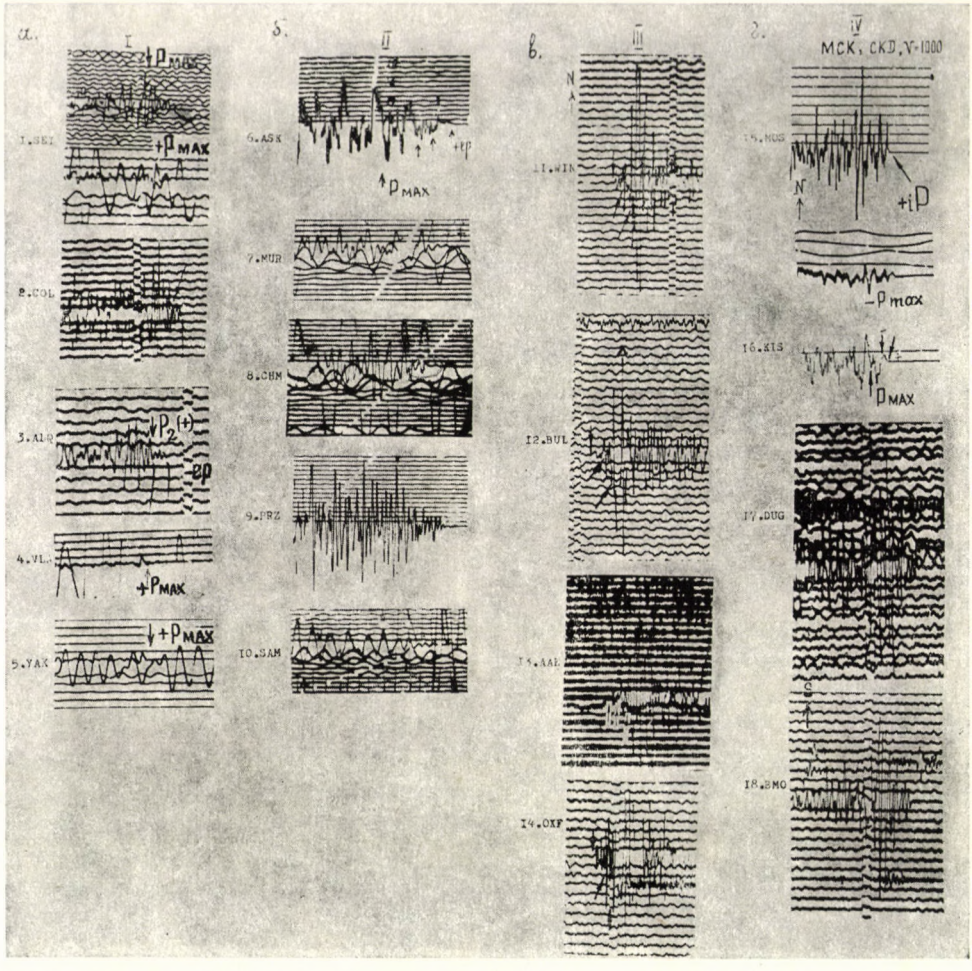

Fig. 1. Examples of P-wave records of the Daghestan earthquake. ECCH stations, USSR, CK and CKD instruments, evolvent of record 60 mm/sec and 30 mm/sec. World network stations, BENIOFF instruments, and evolvent of record 75 mm/sec are designated with asterisks. *a)* Quarter I: 1. Seimchan (SEI), $\Delta = 59°$, CK, V = 1000; CKM — 34,000; 2. Colden (COL), $\Delta = 71°$, V = 100,000; 3. Albuquerque (ALQ), $\Delta = 97°$, V = 100,000. 4. VLADIVOSTOK (VLA), $\Delta = 59.1°$ CK, V = 1000. 5. Yakutsk (YAK), $\Delta = 49.7°$, CK, V = 1000. *b)* Quarter II: 6. Ashkhabad (ASK), $\Delta = 10.2°$, CK, V = 970. 7. Murgab (MUR), $\Delta = 21°$, CK, V = 1000. 8. Chimkent (CHM), $\Delta = 16.7°$, CK, V = 1000. 9. Przhevalsk (PRZ), $\Delta = 23°$, CKM, V = 15,000. 10. Samarkand (CAM), $\Delta = 15.4°$, CK, V = 980. *c)* Quarter III: 11. Windhoek (WIN), $\Delta = 71°$, V = 50,000. 12. Bulawayo (BUL), $\Delta = 66°$, V = 100,000. 13. Addis-Ababa (AAE), $\Delta = 35°$, V = 50,000. 14. Oxford (OXF), $\Delta = 93°$, V = 50,000. *d)* Quarter IV: 15. Moskva (MOS), $\Delta = 13.9°$, CKD, V = 1000, below low-magnification channel V = 100. 16. Kishinev (KIS), $\Delta = 13.4°$, CK, V = 860. 17. Dugwey (DUG), $\Delta = 91°$, V = 400,000. 18. Blue Mountains (BLU), $\Delta = 97°$, V = 100,000

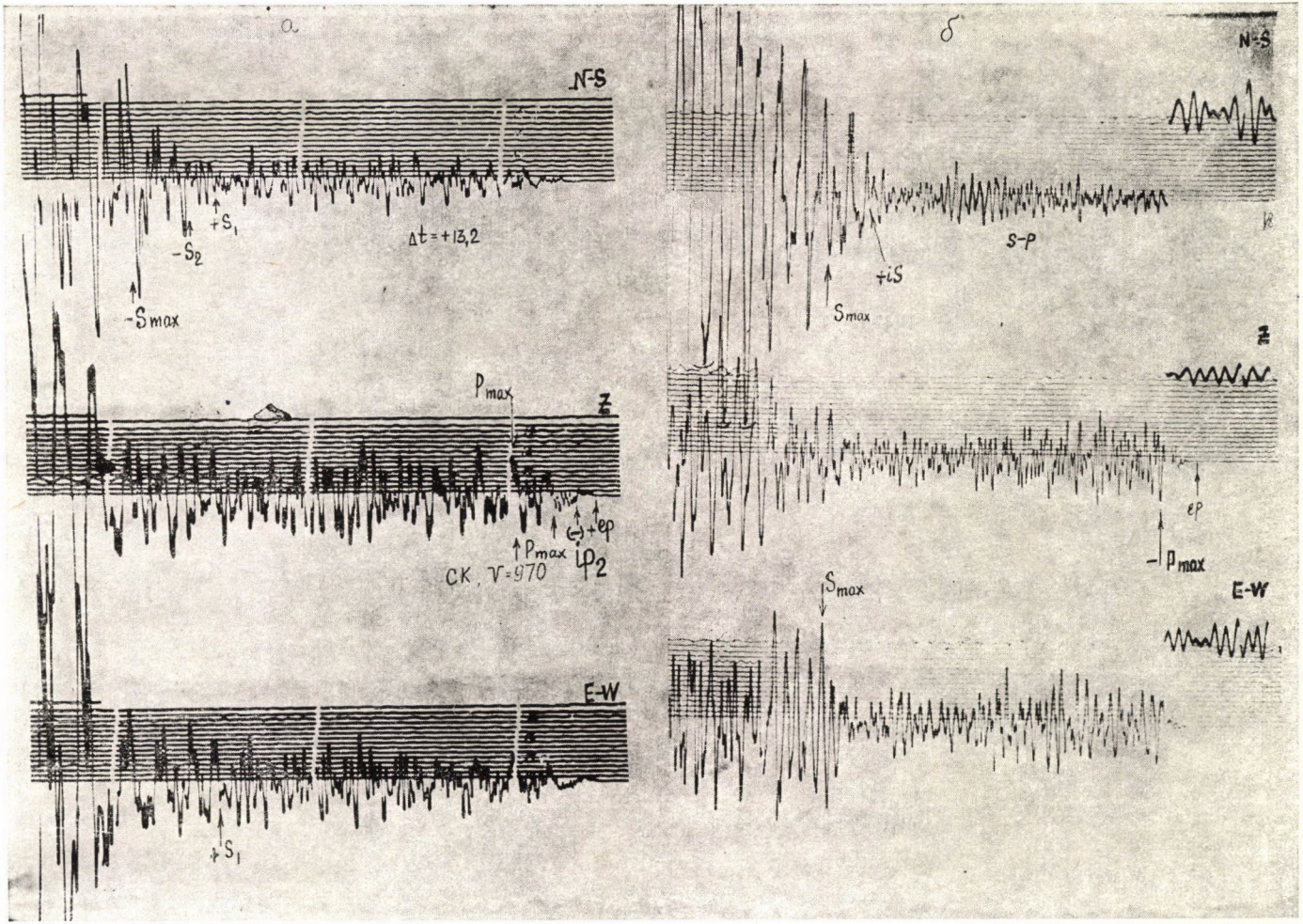

Fig. 2. Examples of P_i and S_i records and distinguishing $S_i - P_i$ on a seismogram. Seismic stations: a. Ashkhabad, b. Tashkent

2. P_i phases are seen at all distances and azimuths. All other wave types are usually observed at definite distances and azimuths where conditions are available for their generation. Times of P_i phase delay follow a certain azimuthal regularity which depends upon the length of a source area and rupture velocity and correspond to a ratio:

$$\Delta t = \frac{l}{c} - \frac{l \cos \varphi}{V}$$

where φ is the angle between the direction to a station and the rupture trend; c is the rupture velocity; V is the wave velocity and l is the rupture length.

In order to estimate P_i phase delay at different stations depending on azimuth and distance, phase identification has been made, i.e. arrivals which corresponded to the same disturbance point in the source are were distinguished at various seismic stations. The following procedure was used: the maximum amplitudes, time of wave emergence on a record and corresponding periods were determined in each wave group recognized on a seismogram, then A/T value and magnitude were calculated. To reach the better phase correlation, the influence of distance and azimuth peculiarities of the station was rejected by calculating coefficients at each station: $K_i = \left(\frac{A}{T}\right)_{\max} : \left(\frac{A}{T}\right)_i$. Phases for which K coefficients were close, were considered as identical.

The azimuthal curve was constructed for identical phases. Times were matched, using the Jeffreys—Bullen travel-time table, to one epicentral distance taking into account residuals $f_K = t_H - t_{J-B}$ (where t_H is the wave travel-time from a source to the station, t_{J-B} is the time from the Jeffreys—Bullen travel-time table). The experimental curve was compared with the data calculated for a source region of corresponding dimensions (from published data) and various rupture velocities.

The azimuthal diagram of $P_{\max}$ wave travel, clearly distinguisable on seismograms, is shown in Fig. 3. The travel time for the Middle Asia stations "Dushanbe", "Murgab" and "Namangan" proves to be longer than that at stations "Lvov", "Uzhgorod" and "Simferopol" located approximately at the same distances in the opposite direction. Using this curve one can determine the direction and rupture spreading. It occurred in the northwest direction at the azimuth 300°. Choosing various rupture velocities, one can estimate the source zone extension with maximum energy release and source dimensions (GORBUNOVA, 1977). The authors made a phase identification not only for P waves but also for S waves. This identification represented the most important part of investigations because all the main conclusions depended on the validity of this identification.

The wave group identification according to maximum phase was followed by determining the direction of displacement in the initial phase for

every wave group. Records with distinct arrivals were used to analyse the destructive processes in an earthquake source. S waves were interpreted using the Jeffreys—Bullen travel-time table. Phases of transverse wave S_1, S_2 . . . S_{max} were assigned according to the same $S-P$ value respective to phases P_1, P_2 . . . P_{max}. Examples of distinguishing S_1 . . . S_{max} are shown in Fig. 2. All values were plotted on the Wolf net.

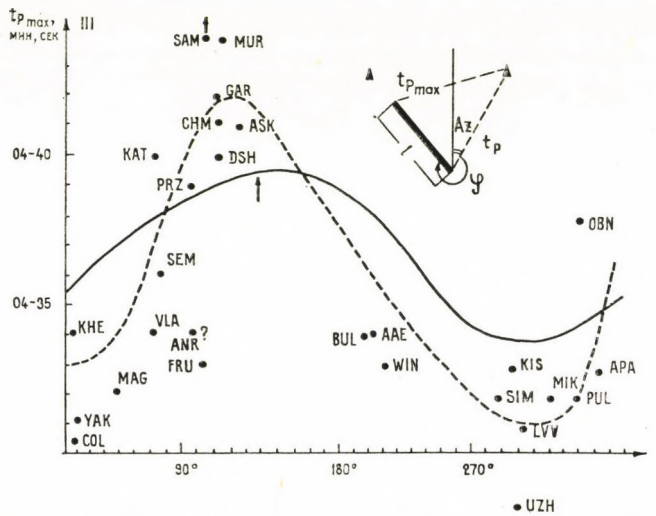

Fig. 3. Travel time of maximum phase in P wave at stations located at different azimuths (Times are cited from the Jeffreys—Bullen travel-time table at $\Delta = 20°$). Instruments CK. 2. CKM. 3. Theoretical curve with parameters: $\Delta = 40$ km, Az = 310°, $\Delta = 20°$, $V_p = 8.1$ km/sec. In the right upper corner a scheme of seismic wave propagation is shown from a source to seismic stations

Mechanism of the Destructive Process in an Earthquake Source

Based on discrete development of the destructive source process, an attempt has been made to reproduce mechanism of the destructive process in an earthquake source by using the generally accepted A. V. VVEDENSKAYA technique (BALAKINA et al., 1972) for determining source mechanism parameters by using P_2 and P_{max} phases. The results are portrayed on stereograms and in the Table (Fig. 4).

In order to determine source mechanism data were used on signs of longitudinal and transverse wave displacement on seismograms from ECCH network with records from CK and CKD instruments and microfilms from the American WWSSN stations with records from Press—Young and Benioff instruments. Signs of displacements in the first arrivals of longitudinal waves were taken from bulletins.

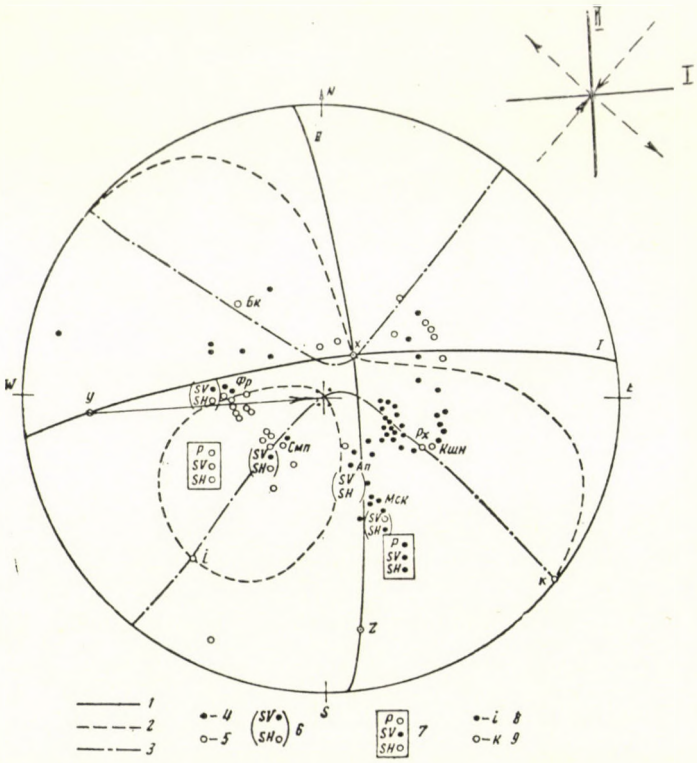
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Fig. 4a

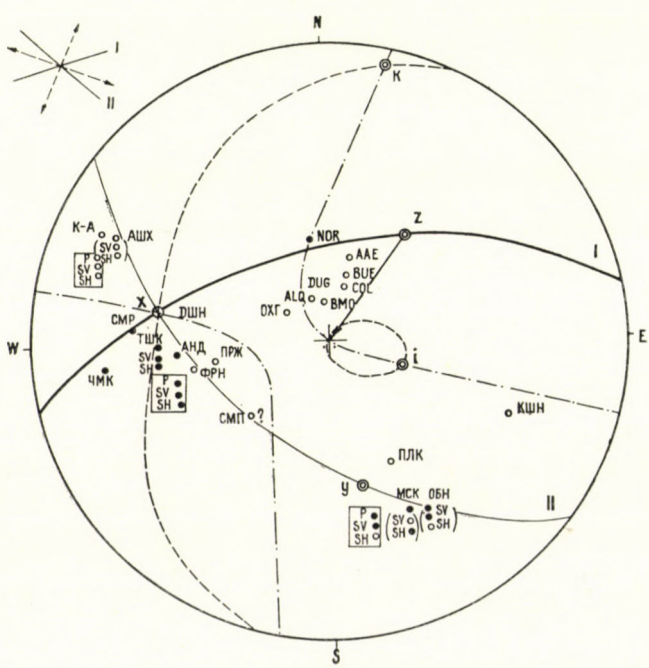
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Fig. 4b

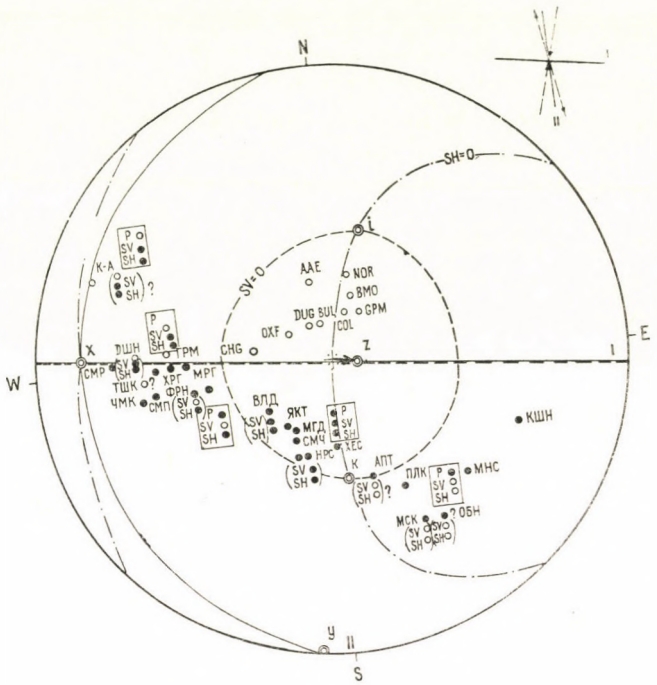

Fig. 4c

Fig. 4. Examples of determining source mechanism according to data of first arrival (SHIROKOVA, see Fig. 4a) and second arrivals (SIMBIREVA, see Fig. 4b, c). 1. Nodal planes of P waves. 2. Nodal planes SV. 3. Nodal planes SH. 4. Compression waves. 5. Dilatational waves. 6. Observed signs of displacement of SV and SH waves. 7. Calculated signs of body wave displacements. 8. Compression axis. 9. Tension axis

Table 1

Source mechanism

Date, origin time, GMT h. min.	Source coordinates		M	H	Fault plane I			
					Dip direction		Upper side movement	
	φ°N	λ°E			Az°	e°	strike component	dip component
14.V								
P_1								
18 12	42.95	47.17	6.8	12	173	76	−0.97	−0.24
P_2								
18 12	43.09	47.11		30	170	54	−0.57	+0.82
P_{max}	43.25	46.98		36	186	88	+0.17	−0.98

Note: Az° = azimuth; e° = the angle with horizontal surface; dip-component: (+) reverse type of faulting; (—) normal type of faulting; strike component: (+) right-lateral strike-slip; (—) left-lateral strike-slip

Hypocenter co-ordinates of the first and ensuing P_i phases were taken in (GORBUNOVA, 1977). The hypocenter corresponding to the moment of first arrival of P_1 main shock was located at a depth of 12 km. Hypocenters of second arrival of P_2 and P_{max} phases corresponding to the moment of maximum energy release in an earthquake source were considered to be at depths of 30 km and 36 km, respectively. Co-ordinates were determined with an accuracy of ± 10 km. In ground plan epicenters of subsequent rupture phases, not including the first arrival, formed a broken chain elongated in the northwest direction. The distance in plan between epicenters P_1, P_2 and P_{max} was 15—20 km. The total extent of bursts in the source area up to the moment of maximum energy release was about 40 km.

It is noteworthy that, if the accuracy of determining the moments of brittle destruction in a source and of P_i phase identification is insufficient, the precise correspondence between the hypocenter surfaces determined from P_i (P_1, $P_2 \ldots P_{max}$) arrivals often cannot be reached.

Angles i with vertical surface were defined from graphs i versus epicentral distance Δ constructed by KOGAN (GOTSADZE et al., 1957). The accuracy of the nodal plane position is different.

Nodal planes of main shock mechanism according to the first arrivals (focal mechanism according to the first arrivals was determined by SHIROKOVA (STEINBERG, LEVSHIN et al., 1974; SHEBALIN et al., 1973; STEINBERG, APTEKMAN et al., 1974) were constructed reliably and unanimously with an accuracy of $\pm 5°$.

The position of nodal lines according to the second phase arrivals was determined rather reliably from data on signs of longitudinal and transverse wave displacement. The displacement signs in SV and SH waves agree well with theoretically expected for the "Ashkhabad", "Tashkent" and "Obninsk"

parameters

Fault plane II				Compression		Intermediate stress		Tension	
Dip direction		Upper side movement							
Az°	e°	strike component	dip component	Az°	e°	Az°	e°	Az°	e°
265	76	+0.97	−0.24	218	20	38	69	129	0
40	50	−0.64	+0.77	110	60	284	30	14	0.3
93	10	−0.99	+0.09	16	43	275	10	177	45

seismic stations. At "Moskva" seismic station SV and SH signs were obtained according to less reliable data and, as it turned out, they did not correspond to the theoretical ones. Therefore, we attach less importance to them, owing to indistinct signs of P_2 arrivals on seismograms of horizontal components.

Source mechanism solution obtained from P_{max} (maximum arrivals) phases agrees with sign distribution of P waves and SV and SH at "Ashkhabad", "Dushanbe", "Frunze", "Vladivostok", "Norilsk", "Apatity", "Moskva" and "Obninsk" stations. Only sign P at "Tashkent" seismic station does not correspond to the results obtained. The first nodal plane was drawn reliably. The accuracy of construction of the second one is smaller. The scatter in azimuth is ± 10—15°. However, the change in its azimuthal position does not entail an essential change in stress axis position. The azimuthal change of stress, compression and tension axes does not exceed 10°.

The analysis of the results obtained and their comparison with geologic and seismological data permit to suggest that the source destructive process occurred in all the phases considered along planes of near-latitudinal trend with rather steep dip angles (in an interval of 60—90°). From shock to shock the rupture trend change from ENE to WNW and the azimuth of the rupture plane dip changed from southeastern for P_1 and P_2 to northwestern for P_{max} (Fig. 4a, b, c). The movement along P_1 and P_2 ruptures had a strikeslip fault character, the hanging rupture sides moving downward and in opposite azimuths. Such a movement is possible when there occurs an entire uplift of the lying rupture plane sides which are surfaces of a single block. In the moment of maximum energy (P_{max}) release, almost vertical upthrust of the hanging side with a small fault component was observed. The lying side of the southwestern face of the removed block, as it could be expected, has sunk. This interpretation of movement mechanism in the source is reasonable, as the real mosaics and block structure of the Paleozoic basement in the source zone favours the key uplift and subsidence of the neighbouring blocks (Fig. 5).

We now consider the geologic–tectonic environment in the source area of the Daghestan earthquake.

According to data of the South geologic expedition and from GOTSADZE et al. (1957), STEINBERG et al. (1974) and SHEBALIN et al. (1973), the territory under consideration in tectonic respect lies within the Daghestan wedge which has divided at the end of Miocene time the eastern North Caucasus geosynclinal trough into Tersk-Sunzhen and South Daghestan parts (BROD, 1958). Two large protrusions are located within the Daghestan wedge, namely Sulaksky in the west and Talginsky in the east separated by Kapchugaiskaya depression filled with limestone and carbonate sediments 5—10 km thick of Cenozoic, Cretaceous and Jurassic age which are superimposed on the consolidated Paleozoic basement. The earthquake focus emerged within the Kapchugaiskaya depression. On the basis of geologic and geophysical investi-

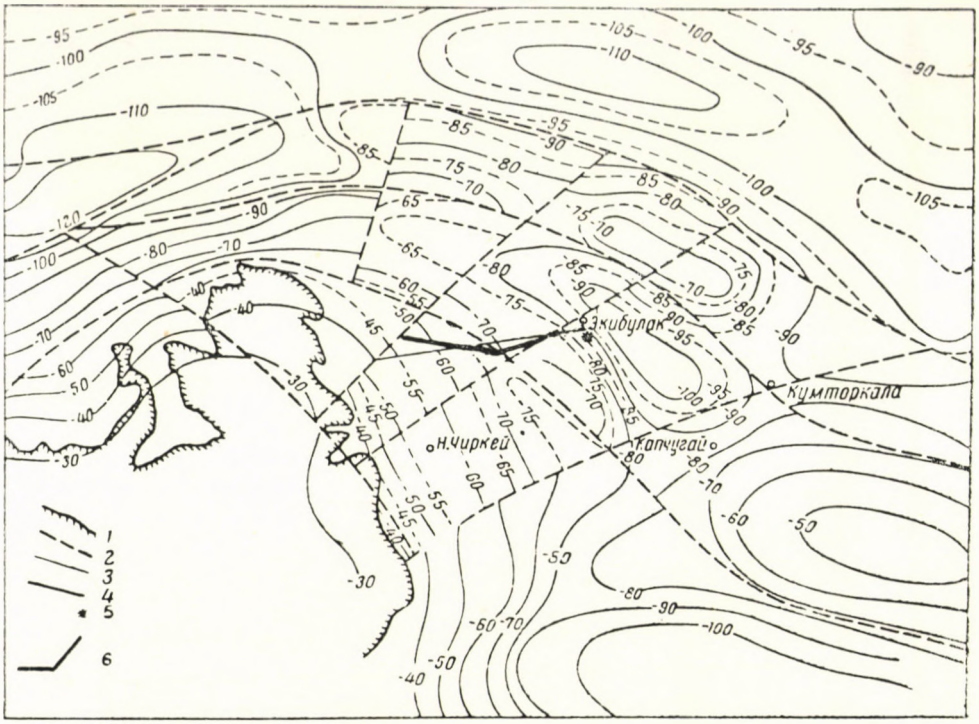

Fig. 5. Comparison of source mechanism with geologic environment of the region (Geologic—tectonic scheme was constructed by A. M. MAGOMEDOV, R. A. LEVKOVICH, M. N. SMIRNOVA and E. M. BRAZHNIK). 1. Outcrops of Cretaceous rocks; 2. lines of tectonic dislocation; 3. contour lines of the Pre-Jurassic basement; 4. direction of rupture plane trend at the onset of rupture; 5. epicenter of the main shock; 6. projections of rupture planes originating in a source in a moment of P_2 and P_{max} phase arrivals

gations the faults of near-latitudinal, meridional as well as northeastern (presumably) strikes have been distinguished there (GEOLOGIC STRUCTURE . . . , 1960).

It is assumed that the *P* basement is dissected by faults into single blocks sinking, in passing from west to east. Then one can suggest that the destructive process in a source is due to upthrust displacement of a large block of the earth's crust west of Ekibulak (Fig. 5).

The general views that the earthquake could be caused by submeridional uplift of the Daghestan wedge were suggested earlier by STEINBERG, LEVSHIN et al. (1974).

The stress field in a source of the main shock varies during each ensuing impulse. One can judge of the character of these variations according to the orientation of main normal axes in a source of each phase (Fig. 4a, b, c).

The entire uplift in a source zone appears to occur under conditions of highest compression in the northeast—southwest direction.

Conclusions

Based on the analysis of wave picture from the large Daghestan earthquake (*P* and *S* waves) from records by CK, CKM and Benioff instruments, a hypothesis has been accepted for the discrete development of the destructive source process. The record of these waves on seismograms looked like individual wave groups of increasing intensity. The wave group with maximum amplitude on a record corresponded to the greatest destructions in a source and to the moment of maximum energy release.

2. A procedure is suggested for identification of these wave groups on earthquake records from various seismic stations.

3. A method of azimuthal curve construction is suggested to distinguish source waves and to determine the predominant direction of rupture in a source.

4. The first attempt has been made to reproduce the source destructive process according to the first impulses of various wave groups which were distinctly expressed on seismograms. Source mechanism parameters for several subsequent-in-time displacements have been used.

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ОПРЕДЕЛЕНИЕ РАЗВИТИЯ ПРОЦЕССОВ РАЗРУШЕНИЯ В КРУПНОМ ОЧАГЕ ЗЕМЛЕТРЯСЕНИЯ С ПОМОЩЬЮ КИНЕМАТИЧЕСКИХ И ДИНАМИЧЕСКИХ ПАРАМЕТРОВ ВОЛН P И

И. Б. ГОРБУНОВА и И. Г. СИМБИРЕВА

Р е з ю м е

Процессы разрушения, приуроченные к очагу землетрясения, были зучены на примере землетрясения, имевшего место в Дагестане 14 мая 1970 г. Картина сейсмических волн, полученная для данного крупного землетрясения, интерпретировалась на основе концепции о дискретном развитии из очага процессов разрушения. Для некоторых смещений, выдержанных во времени, применялись специальные параметры механизма очага землетрясения.

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О РАЗМЕЩЕНИИ НЕОГЕНОВЫХ ВУЛКАНОВ КАРПАТСКОГО РЕГИОНА

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ВСЕВЕНГЕРСКИЙ ТРЕСТ НЕФТЯНОЙ И ГАЗОВОЙ ПРОМЫШЛЕННОСТИ, БУДАПЕШТ

Основное методическое следствие гипотезы тектоники плит заключается в необходимости анализа латеральной структурно-магматической зональности для обоснованных палеотектонических реконструкций. В этом отношении весьма важную роль могут играть выводы по просранственному распределению известково-щелочного вулканизма. Без ясного понимания закономерностей распределения неогенового вулканииама Карпатсково региона вряд ли можно создать правильную картину его геодинамики.

Учитывая состояние изученности состава карпатских вулканитов, нами виделены всего две ассоцнацни: риолитово—андезитовая (нередко без риолитов) и андезитово—риолитовая (большей частью с андезитами). Выявление латеральной магматической зональности должно базироваться на данных только по вулканам, поэтому туфово-туффитовые прослои—пачки среди осадочных толщ нами полностью были исключены из рассмотрения, и на карте (фиг. 2.) показаны лишь сами вулканы, в том числе и погребенные более мододыми отложенями.

По характеру соотношений с тектоническими злементами Карпат нами выделены: Карпатская вулканическая дуга и Внутрикарпатский вулканический ареал. Первая по своему положению вполне соответствует островным дугам, если Предкарпатский прогиб считать аналогом глубоководных океанических желобов. В строении второй выделены три зоны, в общем поперечные предыдущим структурам. Централъная зона почти непрерывна и сложена андезитово—риолитовыми вулканами, краевые прерывисты и состоят из (риолитово) — андезитовых вулканов. Централъная зона отчетливо тяготеет к Палеогеновому флишевому поясу Большой Венгерской впадины.

Общая картина третичного вулканизма Карпатского региона сильно напоминает таковую мезозойского магматизма Монголо—Охотского пояса. При эгом раннемезозойскому этапу там (фиг. 3а) соответствует эоценовый этап здесь (фиг. 4), позднемезозойскому там (фиг. 3б) — миоценовый здесь (фиг. 2. Внутрикарпатский ареал), наконец, позднемеловому—раннепалеогеновому там (фиг. 3в) — плиоценовый здесь (фиг. 2. Карпатская вулканическая дуга). Для более глубокой интерпретации выявленной латеральной матматической зональности и намеченного сходства с Монголо—Охотским поясом необходимы дальнейшие исследования.

Геодинамика кайнозойской истории Карпатского региона ясна лишь в самых общих чертах, и то в основном касательно самой Карпатской дуги. По внутренним областям существуют лишь разнообразные гипотезы, ни одна из которых не учитывает и не объясняет в достаточной мере основной фактический материал.

Пожалуй, основным методическим следствием теории тектоники плит является необходимость изучения латеральной тектоно-магматической зональности обширных пространств для сравнительно узких промежутков времени, вместо изучения длительной последовательности образований в отдель-

ных структурно-формационных зонах. Исходя из этих соображений, основную задачу дальнейших исследований мы видим в создании серии карт, отражающих латеральную тектоно-магматическую зональность определенных этапов развития Карпатского региона. Шагом к этому является составление карт распределения неогеновых вулканических образований, что и выполнено нами в порядке опыта на основе многочисленных локальных схем.

Затруднения в создании региональной схемы очень велики. Значительная часть внутрикарпатского региона скрыта плиоценово-четвертичными отложениями, и представления о размещении вулканических образований до последнего времени основывались главным образом на изучении выходов, составляющих небольшую часть всей территории. За последние десятилетия буровые работы шагнули далеко вперед, вместе с тем сильно продвинулась и методика геофизических измерений и интерпретации их результатов. Оказалось, что в Большой Венгерской впадине и в Задунайщине неогеновые вулканиты распространены весьма широко, и мы уже вполне можем их учитывать при создании схем вулканизма.

Существенную проблему означает обстоятельство, что внутрикарпатские неогеновые вулканиты разбросаны на территории четырех стран, исследователи в которых не только говорят и пишут на разных языках, но и изучают эти образования по-разному. Именно поэтому мы считаем необходимым в первую очередь осветить принятые нами методические принципы.

Основой анализа латеральной магматической зональности является расчленение на формации. Состав неогеновых вулканитов варьирует в широких пределах, от риолитов до базальтов. При этом намечаются контуры, с одной стороны, известково-щелочной серии (риолит—андезито-базальт), а с другой, — щелочно-базальтовой. В данной работе мы ограничимся рассмотрением только известково-щелочной серии. Ее расчленение на формации, необходимое для составления региональных карт, можно провести по-разному.

1. Можно выделить относительно узкие интервалы составов, соответствующие которым породы образуют устойчивые естественные ассоциации, сравнительно однозначно различающиеся друг от друга. Следуя этому пути, можно было бы говорить, например, об андезито-базальтово—андезитовой формации, но практически с не меньшим основанием — об андезито-базальтово—андезито-дацитовой и др. Все подобные варианты расчленения характеризуются значительной произвольностью в выборе категорий вследствие слабой изученности объективных характеристик такого расчленения вообще и распределения составов карпатских вулканитов по массивам, областям и т. д. — в частности. Далее, при таком подходе в большинстве вулканических областей приходится выделять ряд формаций, которые трудно изображать в масштабе, годном для составления региональных карт (очевидно, не крупнее 1 : 500 000).

2. Можно выделить формации, отличающиеся в основном количественным соотношением наиболее характерных компонент, например, андезитово—риолитовую (с преобладением риолитов) и риолитово—андезитовую (с преобладанием андезитов); их крайними представителями будут чисто риолитовые и чисто андезитовые комплексы. Такое расчленение, очень удобное с практической точки зрения, с теоретических позиций может быть принята лишь при условии тесного генетического родства компонент. Если подкоровое происхождение андезитов (с глубин порядка 150 км) можно считать безусловным во всех случаях, в отношении генезиса риолитов допускаются два толкования: они, повидимому, могут быть как мантийными, так и коровыми. Объединить же в одну формацию с андезитами можно было бы лишь риолиты мантийного происхождения. Состояние изученности риолитов Карпатского региона не позволяет отделить эти риолиты от коровых, так что данный путь можно выбрать, только заранее отказываясь от возможностей исследования соотношений мантийного и корового магматизма.

Нам представляется, что при современном уровне изученности более оправдан второй путь, как более простой и более однозначный. Учитывая, однако, что термину «формация» обычно стараются придавать и генетический смысл, мы вместо него будем пользоваться более нейтральным термином «ассоциация». Итак, мы выделяем всего две ассоциации: риолитово—андезитовую (с преобладанием андезитов) и андезитово—риолитовую (с преобладанием риолитов). При этом отметим, что если имеются многочисленные примеры комплексов с полным отсутствием риолитов, то безандезитовых комплексов почти неизвестно (в Карпатском регионе).

При изучении латеральной магматической зональности вулканогенных образований возникает особая проблема учета дальности переноса материала. Ввиду того, что в прежних сводках по вулканизму Карпатского региона этот вопрос был обойден молчанием, считаем необходимым остановиться на нем детально.

Совершенно естественно, что латеральная магматическая зональность вулканогенных образований должна основываться на изучении и сопоставлении только тех участков, где магма прорывалась сквозь толщу земной коры, то-есть только вулканических центров. Проблема, следовательно, сводится к диагностике и оконтуриванию этих центров.

При региональных работах вряд ли будет большой ошибкой выделение и рассмотрение вулканических построек вместо центров. Эта задача решается сравнительно просто лишь для вулканов центрального типа, продуктами которых в Карпатском регионе являются, повидимому, все вулканиты андезитового состава и часть — дацитового и риолитового (экструзии и связанная с ними вулканокластика).

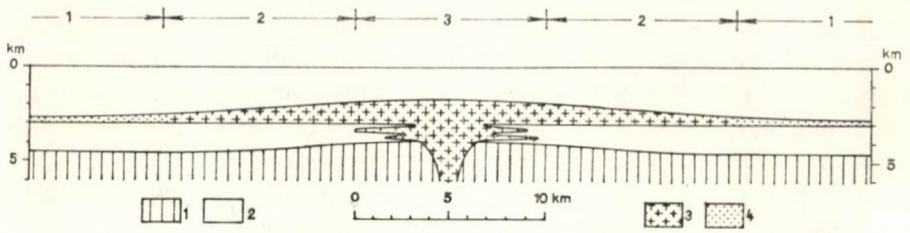
Относительно характера источников материала кислых вулканитов типа игнимбритов, широко распространенных на СВ-е Венгрии, существуют

лишь гипотезы, из которых наиболее широкой известностью пользуется та, что считает эти образования продуктами «трещинных» или «ареальных» извержений — практически безо всякой надежды на выявление и оконтуривание конкретных вулканических аппаратов или центров. Нам представляется, что независимо от взглядов на генезис игнимбритов и на характер соответствующих аппаратов, необходимо стремиться оконтурить хотя бы приблизительно области наиболее вероятной локализации центров соответствующих извержений, причем это можно сделать на основе тех же общих принципов фациального анализа, которые успешно применяются при изученим вулканов центрального типа.

Диагностическими признаками вулканических построек можно считать наличие в разрезе собственно магматических (лавовых) и грубых вулканокластических пород (агломератов, брекчий и т. п.), а также повышенную мощность вулканических толщ (сотни метров и более); очевидно, дополнительным признаком для игнимбритов (какой-то мере эквивалентным грубости вулканокластики центральных вулканов) будет интенсивное спекание и особенно сваривание обломков. Пользуясь этими признаками, из рассмотрения нами полностью исключались прослои, пачки и даже толщи туфов, туфитов и т. п. пород мощностью до 100 (200) м, залегающие внутри осадочных серий. Правомерность такого подхода подтверждается тем, что материал современных извержений литологическими методами улавливается до расстояний в 300—600 км (Лучицкий И. В., 1971; Рухин Л. Б., 1962; Страхов Н. М., 1962), следовательно, продукты длительно действовавших вулканов могут прослеживаться до сотен км, а это уже расстояние, которым нельзя пренебречь при региональных исследованиях.

Выделение и оконтуривание вулканических построек, находящихся и ныне на поверхности, не слишком проблематично, тем более, что все они центрального типа. В случае захороненных вулканов диагностика по скважинам также осуществима более-менее однозначно с использованием указанных признаков. Наибольшие затруднения возникают при попытках оконтурить площади, занятые захороненными вулканами. Неоценимую помощь — в этом оказывают региональные геофизические иследования (**Bodoky T. et al., 1972**).

Захороненные вулканы искажают магнитное и гравитационное поля, создавая соответствующие аномалии. На профилях сейсморазведки методом отраженных волн они обозначаются куполовидными структурами с элементами концентрического внутреннего строения. Но наиболее надежно они выявляются и оконтуриваются методом магнитотеллурических зондирований при наличии в районе вулканизма более древней третичной осадочной толщи. В этом случае наблюдаются три основных типа геоэлектрических разерзов (фиг. 1):



Фиг. 1. Схема строения погребенных вулканов. 1. Докайнозойский фундамент (высокого сопротивления). 2. Кайнозойские отложения (низкого сопротивления). 3. Вулканический аппарат (высокого сопротивления). 4. Вулканиты удаленных зон (низкого сопротивления); Цифры в верхнем ряду обозначают типы геоэлектрических разрезов (см. в тексте).

1. На удалении от вулканических построек соответствующие им туфово-туффитовые пачки—толщи не выделяются на фоне осадочной серии, имеющей низкое сопротивление и четко обособляющейся от дотретичного фундамента высокого сопротивления. Геоэлектрический разрез нормальный, двуслойный.

2. На окраинах построек мощность вулканического комплекса возрастает, в его составе появляются грубообломочные и лавовые образования, повыщающие его сопротивление. Начиная с некоторого момента, вулканический комплекс обосабливается в геоэлектрическом разрезе в качестве промежуточного слоя высокого сопротивления, так что разрез становится четырехслойным.

3. По мере приближения к центрам построек изменяются соотношения вулканического комплекса с подстилающей осадочной толщей: мощность вулканического комплекса возрастает, а мощность осадочной либо остается постоянной, либо даже уменьшается, как это мы нередко видим в почве вулканов на поверхности. Это изменение все более затрудняет выделение нижнего геоэлектрического слоя низкого сопротивления. В том же направлении действует и появление все большего количества субвулканических тел внутри подстилающей осадочной толщ, увеличивая ее результирующее сопротивление и приближая его к таковому вулканического комплекса. Вследствие всего этого в центральных частях вулканических построек геоэлектрический разрез снова становится двуслойным, но на этот раз подошва слоя низкого сопротивления проходит не вдоль поверхности дотретичного фундамента, а по кровле вулканического комплекса, что отчетливо обнаруживается при латеральной корреляции геоэлектрических разрезов.

Надежность выделения вулканических центров этим способом подтверждается приуроченностью к ним магнитных и/или гравитационных аномалий и/или куполовидных структур (по сейсморазведке). Нами за вулканические центры принимались только объекты, для которых не менее трех геофизических методов (из четырех) давали сходный результат.

Полный комплекс указанных геофизических методов поставлен пока только в СВ-ной Венгрии. В других районах для оконтуривания захороненных вулканов остается использовать одну лишь карту геомагнитного поля. На базе опыта, полученного в СВ-ной Венгрии, а также изучением характера поля неогеновых вулканов на поверхности, основными признаком захороненных неогеновых вулканов мы считаем неоднородность магнитых аномалий с относительно большой ролью отрицательных значений.

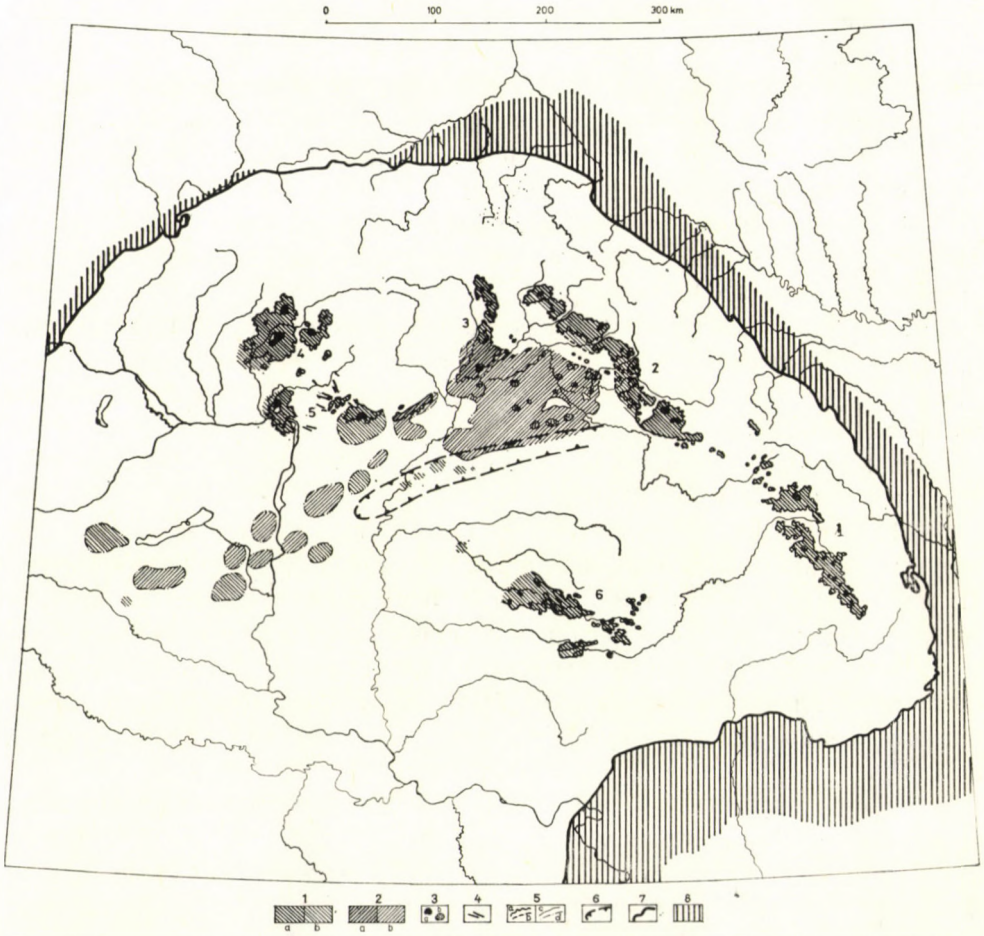
В свете изучения магнитных свойств вулканитов при палеомагнитных исследованиях указанный признак объясняется следующим. В неогеновых вулканитах остаточная намагниченность примерно в половине случаев обратна, причем она часто сильнее наведенной намагниченности; поэтому эти вулканиты часто имеют обратную результирующую намагниченность, отсюда относительная частота отрицательных значений магнитного поля. Но так как во-первых вулканиты прямой и обратной остаточной намагниченности слагают небольшие геологические тела неправильной формы, чередующиеся в плане и в разрезе, а во-вторых соотношения остаточной и наведенной намагниченности довольно сильно колеблются даже в пределах одного и того же тела, то магнитное поле над неогеновыми вулканитами характеризуется прихотливостью, лишь частично сглаживаемой глубиной залегания.

В более древних вулканитах, уже начиная с эоценовых, остаточная намагниченность играет подчиненную роль по сравнению с наведенной, поэтому магнитные аномалии над ними проще, и они легче увязываются с геометрией возмущающих тел. Туфово—туффитовые пачки и толщи как правило не дают ощутимых магнитных аномалий.

Несмотря на четкость данного признака, при конкретном его применении в оконтуривании захороненных вулканических построек всегда остается та или иная доля субъективности, поэтому проводимые контуры следует считать предварительными. Как правило, по магнитным картам оконтуривались лишь вулканы, подтвержденные бурением. В некоторых случаях характер магнитного поля позволял и выделять вулканические постройки, однако, естественно, их существование нельзя считать доказанным.

Обращает на себя внимание, что всюду, где установлены признаки одних лишь риолитовых вулканов, имеются магнитные аномалии такого же типа, как и над андезитовыми. Ввиду слабой изученности магнитных свойств риолитов конкретных районов неясно, о чем этот факт свидетельствует: то ли об обязательном присутствии андезитов в ассоциации с риолитами (с отсутствием их лишь в отдельных случайных пересечениях), то ли о часто значительной намагниченности риолитов. На основе общих соображений и результатов изучения магнитных свойств вулканитов Закарпатья (Михайлова Н. П. и др., 1974) мы склоняемся к первому мнению, но не считаем его достаточно аргументированным.

На основании приведенных данных нами составлена карта неогеновых вулканов Карпатского региона с подразделением их на андезитово—риолитовые и риолитово—андезитовые (фиг. 2). Сопоставим распределение вулканов с тектоникой Карпат. Согласно современной геодинамической интерпретации (Зоненшайн Л. П. и др., 1976), Предкарпатский прогиб можно считать аналогом глубоководных океанических желобов, то-есть выходом зоны Беньофа. Судя по форме этого прогиба в плане, зона Беньофа могла быть на-



Фиг. 2. Размещение неогеновых вулканов Карпатского региона. 1. Вулканы риолитово—андезитовой ассоциации, *a* — на поверхности, *b* — погребенные. 2. Вулканы андезитово—риолитовой ассоциации, *a* — на поверхности, *b* — погребенные. 3. Вулканические центры, *a* — на поверхности, *b* — погребенные. 4. Дайки андезитов. 5. Контуры вулканов и вулканических областей, *a* — по данным геологической съемки, *b* — по данным бурения, *c* — по комплексу геофизических данных, *d* — по магнитным аномалиям. 6. Палеогеновый флишевый пояс Большой Венгерской впадины. 7. Контур Карпатской складчатой системы. 8. Предкарпатский прогиб. Цифры на карте: 1. цепь Кэлиман—Харгита; 2. цепь Выхорлат—Гутин; 3. Токайско-Прешовская гряда; 4. вулканическая область Средней Словакии; 5. вулканическая область Северной Венгрии; 6. вулканическая область гор Апусени

клонной и иметь падение внутрь Карпат. С этой точки зрения цепи Выхорлат—Гутин и Кэлиман—Харгита по своему положению в точности соответствуют островным дугам, прослеживаясь параллельно Предкарпатскому прогибу, на расстоянии в 100—150 км от него. Их мы выделяем в качестве Карпатской вулканической дуги; отметим, что перобладающий андезитобазальтовый состав вулканитов цепи Кэгиман—Харгита и во всяком случае наиболее молодых вулканитов (бужорской свиты) цепи Выхорлат—Гутин хорошо согласуется с представлением о том, что Карпатская вулканическая дуга является аналогом настоящих островных дуг.

Прочие неогеновые вулканы не обнаруживают очевидной связи с тектоническими элементами самих Карпат, следовательно, они не могут являться аналогами современных островных дуг. По этому признаку они объединяются нами во Внутрикарпатский вулканический ареал, общими особенностями которого являются следующие:

1. Диффузный характер вулканизма, проявляющийся в разрозненности вулканов, лишь местами образующих короткие цепи (Токайско—Прешовская гряда, зоны гор Апусени).

2. Четкая латеральная магматическая зональность ЗЮЗ—ВСВ направления с центральной зоной андезитово—риолитового и периферическими зонами (риолитово) — андезитового вулканизма; при этом последние сильно прерывисты по простиранию.

3. Увеличение масштабов андезитово—риолитового вулканизма центральной зоны с ЗЮЗ на ВСВ.

Единственным элементом тектоники Внутрикарпатского региона, с которым можно связать внутрикарпатский неогеновый вулканизм, является Палеогеновый флишевый пояс Большой Венгерской впадины: она прослеживается вдоль южной окраины восточной половины центральной андезитово—риолитовой зоны, причем западная половина этой последней как бы продолжает флишевый пояс на ЗЮЗ. Таким образом, как флишевый пояс, так и центральная андезитово—риолитовая зона отмечают один и тот же глубинный разлом, в общем поперечный к зоне субдукции.

Современных аналогов внутрикарпатского известково-щелочного вулканизма неизвестно, и это сильно затрудняет актуалистическую интерпретацию. Претендентом на роль ископаемого аналога может являться Монголо—Охотский пояс. Общими для обеих областей можно считать следующие особенности:

1. Диффузный характер вулканизма со слабой связью распределения вулканов с тектоникой фундамента.

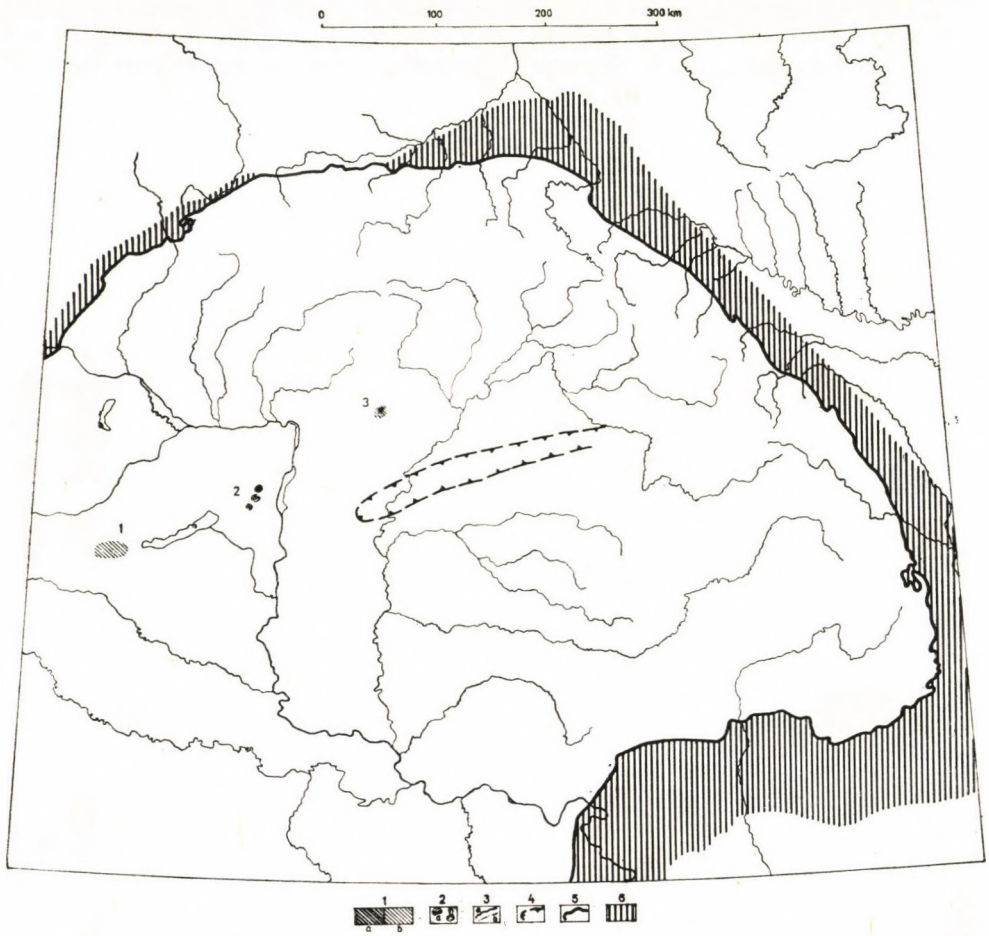
2. Контроль магматизма глубинным разломом, поперечным к зоне субдукции и отмечаемым узким трогом флишевого осадконаколения.

3. Четкая латеральная зональность магматизма вкрест глубинного разлома и некоторое изменение его характера вдоль этого разлоа.



Фиг. 3. Стуркгурно-магматическая зональность Монголо-Охотского пояса (по ЗОНЕНШАЙНУ Л. П. и др., 1973, без щелочного магматизма), а — ранний мезозой (триас — ранняя юра); б — поздний мезозой (поздняя юра — ранний мел); в — поздний мел — ранний палеоген (палеоцен). 1. Эвгеосинклинальные зоны. 2. Поперечный разлом (Монголо-Охотский). 3. Контур тылового прогиба. 4. Известково-щелочной вулканизм. 5. Гранит-гранодиоритовые интрузии. 6. Диорит-монцонитовые интрузии. 7. Граниты стандартного и литий-фтористого типов

Распределение неогенового вулканизма Внутрикарпатского ареала ближе всего соответствует позднеюрско—раннемеловому и позднемеловому—раннепалеогеновому этапам Монголо—Охотской области (Зоненшайн Л. П. и др., 1973; см. фиг. 3). При этом раннемезозойскому этапу его развития соответствовала бы картина Внутрикарпатского региона в позднем эоцене: с флишевой зоной и с вулканами (фиг. 4.) На восточной окраине Азиатского континента аналогичный по своему положению островным дугам современности вулканический пояс формируется примерно к окончанию магматизма в Монголо—Охотском поясе (поздний мел—ранний палеоген). В свою очередь



Фиг. 4. Размещение позднеэоценовых вулканов Карпатского региона. 1. Вулканы (андезитовой ассоциации), *a* — на поверхности, *b* — погребенные. 2. Вулканические центры, *a* — на поверхности, *b* — погребенные. 3. Контуры вулканов и вулканических центров, *a* — по данным геологической съемки, *b* — по магнитным аномалиям. 4. Палеогеновый флишевый пояс Большой Венгерской впадины. 5. Контур Карпатской складчатой системы. 6. Предкарпатский прогиб. Цифры на карте: 1. Хахотский район; 2. район гор Веленце; 3. Речкский район

наиболее молодые известково-щелочные вулканиты Карпатского региона (паннон—плиоцен) тоже отчетливо тяготеют к Карпатской вулканической дуге. Таким образом, миоценовый этап развития Карпатского региона примерно соответствует позднеюрско—раннемеловому этапу Монголо-Охотской области, а плиоценовый — позднемеловому—раннепалеогеновому. Вследствие того, что корреляция карпатских вулканов по возрасту разработана слабо, данное заключение следует рассматривать только в качестве сугубо предварительного.

Наряду с отмеченными признаками безусловного сходства могут быть еще и другие, но существуют также и несомненные отличия. Нам кажется, что разбор сходств и отличий увел бы нас в область почти беспочвенных фантазий из-за совершенной недостаточности сведений о развитии неогенового вулканизма Карпатского региона во времени. Поэтому мы прервемся на констатации определенной степени сходства между Внутрикарпатским вулканическим ареалом и Монголо-Охотским поясом и на постановке вопроса об острой необходимости обоснованных корреляций вулканов Карпатского региона по возрасту — основы любых заключений об истории развития.

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ARRANGEMENT OF THE NEOGENE VOLCANOES OF THE CARPATHIAN REGION

By

Z. BALLA, T. ZELENKA, E. BALÁZS

Abstract

It can be deduced from the theory of plate tectonics that for a well-founded paleotectonic reconstruction the analysis of the lateral structural-igneous zonation is needed. The conclusions drawn from the spatial distribution of calc-alkalic volcanism can play an important role. Without elucidating the distribution regularities of the Neogene volcanism no real picture can be drawn on the gedynamics of the Carpathian Region.

Taking into account the fact that the composition of the Carpathian volcanites have been widely studied, only two associations were chosen: a rhyolitic-andesitic one (often with no rhyolite) and an andesitic-rhyolitic one (mostly with andesite). To make clear the lateral igneous zonation, only the data concerning the volcanoes are to be used, thus the tuffaceous-tuffitic strata and strata groups occurring in sedimentary sequences were completely neglected. In the map (Fig. 2) only the volcanoes are demonstrated, even those covered by younger sediments.

Based on the nature of relationship to the tectonic elements of the Carpathians the Carpathian Volcanic Arc and the Intra-Carpathian Volcanic Area were distinguished. Regarding its position, the former completely corresponds to the island arcs if the Carpathian foredeep is considered to be an analogon of the oceanic abyssal trenches. In the structure of the latter one three belts were distinguished which are globally perpendicular to the former structures. The central belt is nearly continuous and consists of andesitic-rhyolitic volcanoes, the marginal ones are interrupted and consist of (rhyolitic)-andesitic volcanoes. The central zone is obviously connected to the Paleogene flysch zone of the Great Hungarian Basin.

The general picture of the Tertiary volcanism of the Carpathian Region is rather similar to that of the Mesozoic magmatism of the Mongolian—Ochot Belt. The Early Mesozoic phase of that region (Fig. 3a) corresponds to the Eocene Phase of this region (Fig. 4), while the Late Mesozoic as well as Late Cretaceous — Early Paleogene phases can be paralleled with the Miocene resp. Pliocene phases (Fig. 2) Intra-Carpathian volcanic area resp. Carpathian volcanic belt (Fig. 3b and 3c). The lateral igneous zonations determined as well as the similarity to the Mongolian—Ochot Belt, resp. their interpretation in detail need further investigation.

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ИСТОРИЯ ФОРМИРОВАНИЯ И ВОЗРАСТ БЁРЖЁНЬСКОГО И ДУНАЗУГСКОГО ПАЛЕОВУЛКАНОВ

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Центр Бёржёньского палеовулкана возник на кристаллическом фундаменте, Дуназугского — на мезозойском карбонатном. Разницы в составе продуктов практически нет; к пограничной между вепоридами и мезозойскими образованиями Венгерского Среднегорья зоне дислокаций не приурочено даже второстепенных, мелких центров.

Вулканизму предшествовало накопление оилгоцен—миоценовых моласс мощностью до 1300 м. Бёржёньский палеовулкан имел более сложную и более длительную историю со следующими этапами: крупная кальдера обрушения (20 км) с внешним кольцом мелких (4—6 км) вулканов вдоль бортов, купол (10—12 км) в ее центральной части, присводовая кальдера обрушения (5—6 км) с внутренним кольцом мелких вулканов вдоль бортов, крупный эксцентричный стратовулкан (14 км), кальдера обрушения в нем (10 км), меньший стратовулкан в ней, латеральная кальдера обрушения на его склоне, на месте прежней присводовой. Основные этапы хода развития Дуназугского палеовулкана: крупный стратовулкан (13—14 км), кальдера обрушения (7—8 км) в его центре, меньший стратовулкан в ней.

Возраст вулканизма, по биостратиграфическим данным, — начало раннего бадена, по радиологическим, — 14,7 млн. лет (35 проб с относительной дисперсией в 12,0—13,5% и соответственно с разбросом значений от 11,6 до 19,4 млн. лет). Продолжительность формирования Бёржёньского палеовулкана, по биостратиграфическим данным, — не более 1—2 млн. лет (часть зоны NN 5 по наннофлоре), а по палеомагнитным, — около 0,5—0,7 млн. лет (две полные и две неполные зоны). По палеомагнитным данным, формирование Дуназугского палеовулкана закончилось близодновременно с возникновением присводовой кальдеры в Бёржёни; по геологическим данным, вулканизм начался в обоих центрах практически одновременно.

Сопоставление трех независимых схем расчленения вулканитов: по петрографо-петрохимическому составу, по структурно-морфологическим признакам и по палеомагнитным характеристикам — показывает, что хорошо увязываются лишь последние две.

Бёржёньский и Дуназугский палеовулканы расположены севернее г. Будапешт, в районе коленообразного изгиба р. Дунай. Они изучаются совместными усилиями Венгерского Геологического и Венгерского Геофизического институтов; здесь впервые в стране разработана методика комплексного геолого-геофизического изучения древних вулканических аппаратов (Balla Z., 1977) и впервые в стране реконструированы палеовулканы (Balla Z., 1978).

Докайнозойские образования района четко обособляются на вепориды (на севере) и пермско-мезозойские отложения Задунайского Среднегорья — Гемерского массива (на юге), которые отделяются мощной зоной дислокаций,

четко прослеживаемой в геофизических полях. Центр Бёржёньского палеовулкана возник над вепорским кристаллическим фундаментом, Дуназугского — над пермско-мезозойским терригенно-карбонатным (фиг. 1). Состав продуктов и возраст вулканов практически одинаковы, к пограничной зоне дислокаций не приурочено даже второстепенных вулканических центров.

Вулканической деятельности предшествовало накопление *молассовых отложений*. Время их формирования охватывает, повидимому, средний и поздний олигоцен, ранний миоцен и начало среднего миоцена (**Báldi T.—Csepreghyné Meznerics I.—Nyirő R., 1965; Nagy G.—Zsille A., 1971; Hámor G., 1974; Nagy G., 1976**). В целом район неогенового вулканизма расположен в области довольно резкого изменения мощностей подстилающей молассы, что свидетельствует о возникновении вулканов где-то на окраине бассейна осадконакопления, центр которого был расположен восточнее.

Вулканизм начался при подолжающемся осадконакоплении, и это обстоятельство позволяет считать уровень появления первых прослоев с вулканогенным материалом хроностратиграфическим горизонтом. Сопоставляя списки фауны бентосных фораминифер и моллюсков в разных разрезах и анализируя их статистическими методами, можно показать, что состав фаунистических комплексов в почве этого горизонта и выше него постепенно меняется от эгерского (хаттского) до карпатского (верхнегельветского); на этом основании еще недавно выделяли «гельветские», «позднеолигоценовые» и даже «более древние палеогеновые »вулканиты (**Nagy G. — Zsille A., 1971; Hámor G. — Nagy B. — Nagy G., 1976; Böjtösné Varrók K., 1977. Czakó T. — Nagy B., 1977**), хотя при геологическом картировании и бурении ясно прослеживается один и тот же горизонт, просто он датируется в разных пунктах по-разному. Поэтому мы отказались от прослеживания фаунистических горизонтов, выделяемых способом комплексов по бентосной фауне.

Начало вулканизма по наннофлоре однозначно датируется зоной **NN** 5, что грубо соответствует нижнебаденскому подъярусу (**Steininger F. — Rögl F. — Martini E., 1976**); при этом наннофлора зоны **NN** 5 прослеживается вниз до уровней на 150—160 м ниже первых прослоев с вулканогенным материалом в непрерывной осадочной толще. Этот уровень примерно на 40 м ниже основания конгломератовой пачки, непосредственно подстилающей вулканический комплекс, проходит в пачке песчаников—алевролитов (скв. **Dp-2, Kem-1**, определения: Балдине Беке М). Подстилающие отложения немые, так что базис зоны **NN** 5 может находиться еще глубже.

Характер начальных этапов развития *Бёржёньского палеовулкана* (фиг. 2—3) пока не установлен. Первая фиксируемая структура — это крупная кальдера поперечником в 18—20 км; возможно, она возникла в заключительную стадию развития крупного раннего стратовулкана, остатки которого пока нигде не выделены. Вдоль края кальдеры возникло кольцо мелких вул-

канов поперечниками в 4—6 км; они заметно сгущаются над пермско-мезозойским фундаментом. Эти вулканы извергали разный материал: одни — почти дацитовый, другие — андезито-дацитовый, третьи — андезитовый. Помимо общности структурного положения (внешнее кольцо) их объединяет одинаковая прямая намагниченность.

Следующий этап фиксируется возникновением в средней части кальдеры купола поперечником в 10—12 км; его свод обрушился с образованием присводовой кальдеры с внутренним кольцом мелких вулканов вдоль ее края. Состав пород, их слагающих, также колеблется (примерно в тех же пределах), в то время как намагниченность их всюду обратна.

На СВ склоне внутрикальдерного купола впоследствии возник центр крупного стратовулкана, развившегося в две стадии: сначала сформировался конус поперечником в 14—16 км, затем в его центре возникла кальдера обрушения, и от прежнего конуса осталась лишь сомма (на севере); после этого в кальдере вырос новый стратовулкан поперечником в 10—12 км. Его строение отчетливо выявляется морфологическим анализом, а также геолого-геофизическим картированием (**Balla Z.**, 1978); в центре — эрозионная кальдера поперечником в 4—5 км, на дне которой ныне вскрыто основание этого наиболее молодого стратовулкана: по данным бурения (скв. **P-7**, **P-18** и **Nb-10=17**) мощность подстилающих вулканитов здесь свыше 1200 м. На ЮЗ склоне — латеральная кальдера, возникшая возобновлением прежней, присводовой (на куполе). Состав пород — от андезитов до андезито-базальтов; их намагниченность — по всему разрезу прямая.

В заключение развития Бёржёнского палеовулкана внедрились субвулканические интрузии андезито-дацитового состава и обратной намагниченности, и сформировалась гидротермальная минерализация, отчетливо наложенная на все вулкано-структуры.

Окончание вулканизма датируется по наннофлоре той же зоной **NN 5**: в скважинах около села Сокоя (**Szokolya**) нижние 80 м осадочной толщи, перекрывающей вулканический комплекс, относятся к этой же зоне (определения Надьмароши А.); выше с постепенным переходом следуют пресноводные отложения, так что истинное положение кровли зоны **NN 5** не установлено.

Продолжительность вулканизма, следовательно, охватывает лишь какую-то внутреннюю часть временного интервала, соответствующего зоне **NN 5** и равняющегося 2,4 млн. лет (по **Steininger F.—Rögl F.—Martini E.**, 1976). Это вполне согласуется с более ранней оценкой длительности вулканизма (0,5—1 млн. лет) на базе иных биостратиграфических данных (**Báldi T. — Kókay J.**, 1970); Если принять во внимание, что формирование Бёржёньского палеовулкана происходило в течение четырех геомагнитных эпох (**Balla Z. — Mártonné Szalay E.**, 1978) по 200 000 лет в среднем (**McElhinny M. W.**, 1973), из которых крайние вовсе не обязательно охватываются полностью, то вулканическая деятельность могла продолжаться

0,5—0,7 млн. лет. Таким образом, биостратиграфический и палеомагнитный методы дают практически одинаковый результат, наиболее вероятное значение которого нами оценивается в 0,6 млн. лет.

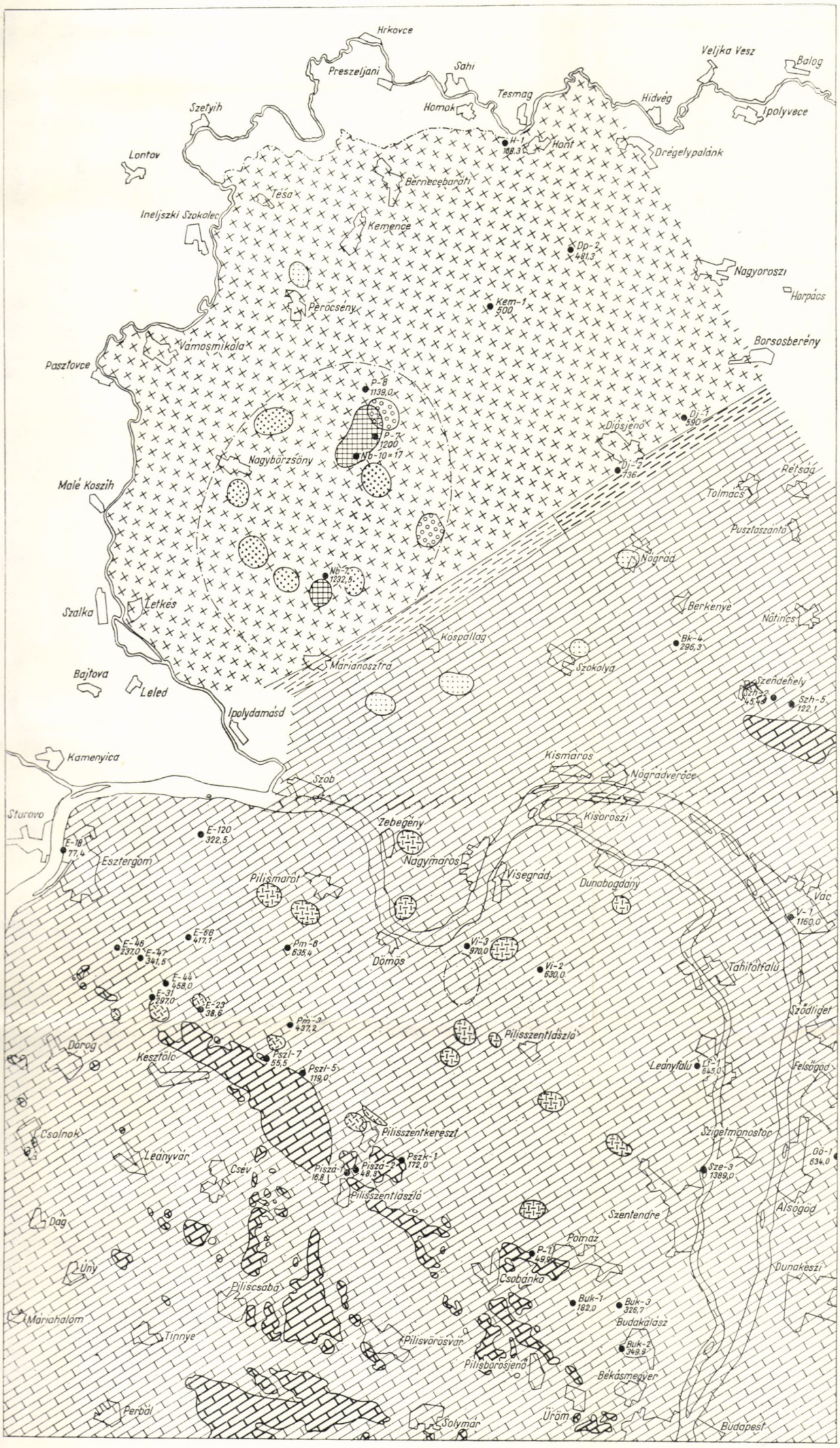
По вулканитам в 1977 г. сделано 35 определений *абсолютного возраста* калий-аргоновым методом (аналитик Балог К.). Поскольку геологические данные свидетельствуют о том, что все вулканиты относятся к одному и тому же комплексу, а палеонтологические и палеомагнитные данные — о весьма ограниченной продолжительности формирования этой толщи (менее 10% вероятного возраста), можно было все определения абсолютного возраста рассматривать в качестве единой совокупности и поставить вопрос о характере распределения в ней данных.

Оказалось (фиг. 4), что эта совокупность с высокой точностью описывается логнормальным распределением (коэффициент регрессии прямой Генри — $r^2 = 0{,}987$). Отсюда следует, что в качестве наиболее вероятного значения возраста нельзя пользоваться среднеарифметическим (это было бы допуситмо лишь в случае нормального распределения), а необходимо брать модальное значение, которое проще всего определить по пересечению прямой Генри с нулевой линией стандартных вероятностей. Это значение для бёржёньских вулканитов составляет 14,7 млн. лет и попадает в самую верхнюю часть зоны NN 5 (по Steininger F. — Rögl F. — Martini E., 1976), не противоречив таким образом биостратиграфической датировке.

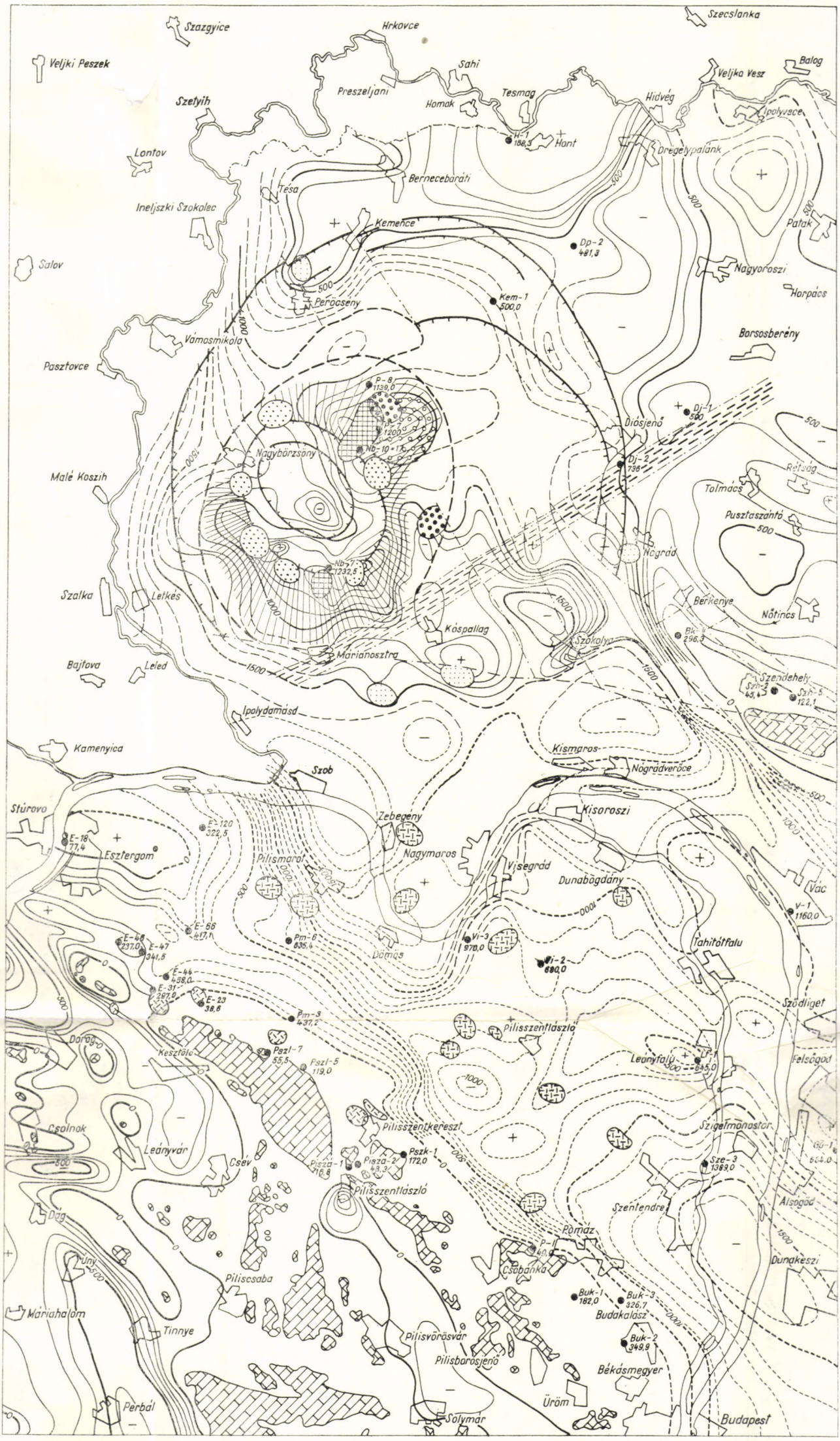
По наклону прямой Генри можно определить дисперсию модального значения; она равна 0,054 в абсолютном выражении (логарифм отношения возрастов в млн. лет) и 4,6% — в относительном. Переходя к годам, можно получить лишь приближенное значение дисперсии; оно равно 1,7—2,0 млн лет в абсолютном и 12—13,5% в относительном выражении. Учитывая, что аналитическая ошибка, имеющая подобный математический смысл, составляет 7%, на долю всех прочих факторов остаются 5—6,5%.

Логнормальная модель подразумевает, что все пробы имеют точно одинаковый истинный возраст, определение которого производится с какой-то ошибкой, характеризуемой дисперсией. На самом же деле действительный возраст проб разбросан в каком-то интервале, и это обстоятельство, естественно, увеличивает дисперсию модального значения. Поэтому встает вопрос о влиянии ширины интервала на величину дисперсии.

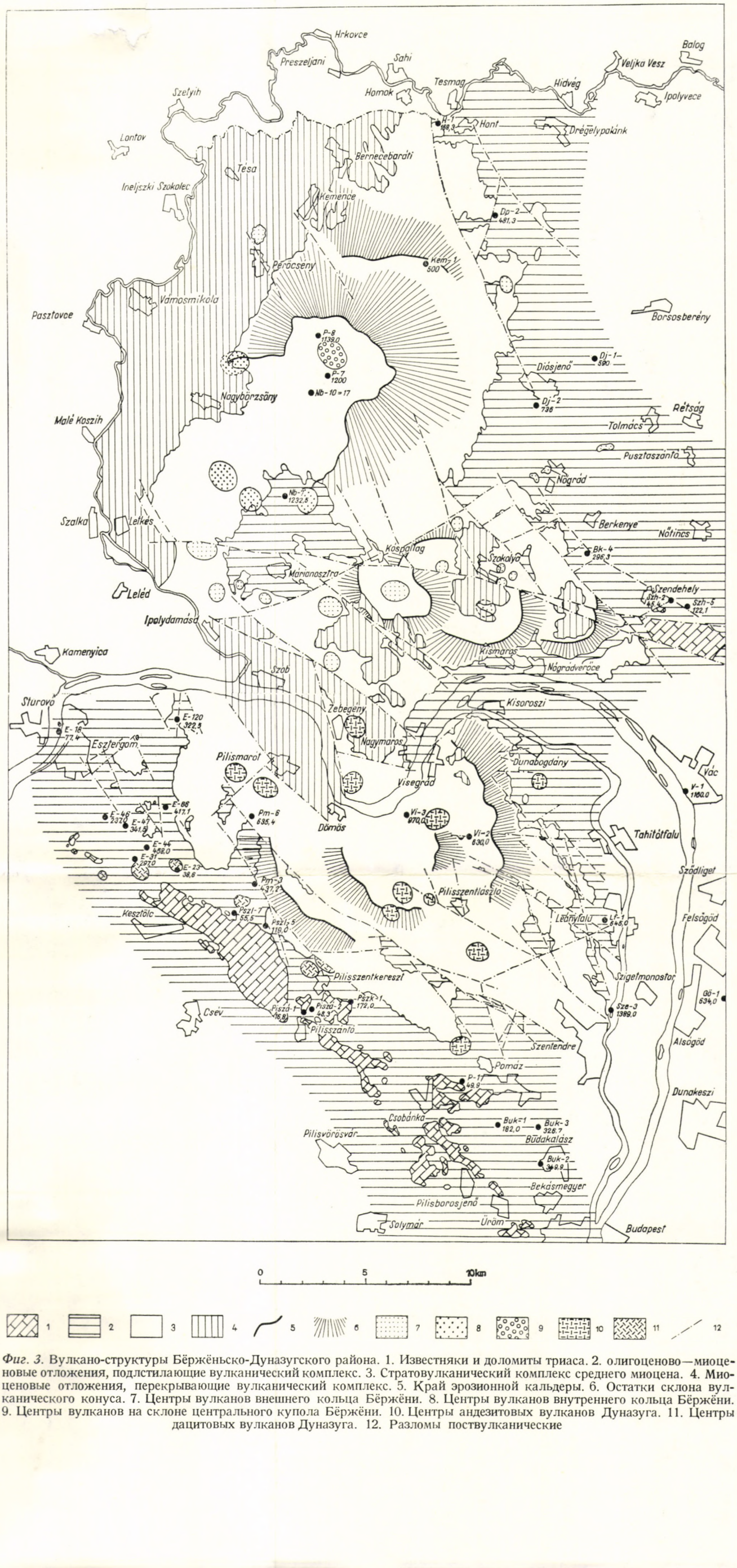
Мы исследовали простейшую модель, предположив, что истинные возраста проб разбросаны по всему интервалу равномерно. Для этого случая нами определена зависимость увеличения дисперсии от ширины интервала, нормированной по дисперсии точечных значений истинного возраста (фиг. 5). Оказалось, что при дисперсии порядка 1,5—2,0 млн. лет ширина интервала в 0,5—1,0 млн. лет (соответствует нормированной ширине 0,3—0,5) увеличивает дисперсию максимум на 1%, следовательно, этим явлением можно пренебречь.



Фиг. 1. Докайнозойские образования иннеогеновые вулканические центры Бёржёньско-Дуназугского района. 1. Кристаллические породы венорид, скрытые третичными образованиями. 2. Пермско-мезозойские образования Задунайского Среднегорья и Гемерского массива, скрытые третичными образованиями. 3. То же в выходах на поверхность. 4. Пограничная зона дислокаций в фундаменте. 5. То же с офолитами (?). 6. Контур центральных областей третичных вулканов (на севере — Бёржёньского, на юге — Дуназугского). 7. Жерловины вулканов внешнего кольца Бёржёни. 8. Жерловины вулканов внутреннего кольца Бёржёни. 9. Жерловины вулканов на склонах центрального купола Бёржёни. 10. Участки интенсивной гидротермальной проработки. 11. Жерловины андезитовых вулканов Дуназуга. 12. Жерловины дацитовых вулканов Дуназуга. Кружками обозначены глубокие скважины: черными — достигшие фундамента (цифры — глубина его залегания), пустыми — не достигшие фундамента (цифры — глубина забоя).

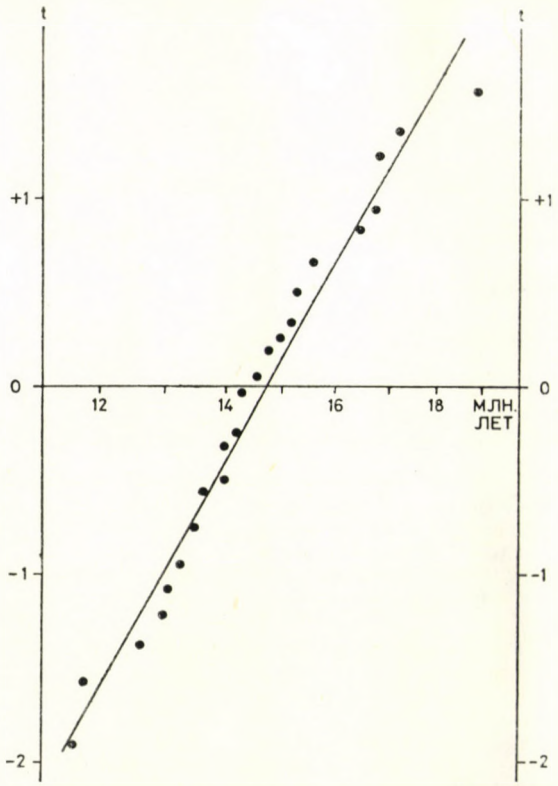


Фиг. 2. Морфология поверхности докайнозойских образований Бёржёньско-Дуназугского района и ее вулкано-тектонические особенности. 1. Мезозойские образования в выходах (известняки и доломиты триаса). 2. Пограничная зона дисло каций. 3. То же с офиолитами (?). 4. Краевые кольцевые разломы кальдер (по сейсморазведке). 5. Контур основания центрального купола Бёржёни (по электроразведке). 6. Склоны центрального купола Бёржёни. 7. Вторичный купол в Бёржёни (по гравиразведке и сейсморазведке). 8. Центры вулканов внешнего кольца Бёржёни. 9. Центры вулканов внутреннего кольца Бёржёни. 10. Центры вулканов на склоне центрального купола Бёржёни. 11. Участки интенсивной гидротермальной проработки (по сейсморазведке). 12. Центры андезитовых вулканов Дуназуга (по магниторазведке). 13. Центры дацитовых вулканов Дуназуга (по геологической съемке). 14. Разломы, в основном поствулканические. 15. Стратоизогипсы поверхности докайнозойского фундамента через 500 м; а — по полному комплексу геофизических измерений (сейсморазведка, электроразведка, гравиразведка), б — по неполному комплексу геофизических измерений, в — по гравиразведке. 16. То же, через 100 м



Фиг. 3. Вулкано-структуры Бёржёньско-Дуназугского района. 1. Известняки и доломиты триаса. 2. олигоценово—миоценовые отложения, подлстилающие вулканический комплекс. 3. Стратовулканический комплекс среднего миоцена. 4. Миоценовые отложения, перекрывающие вулканический комплекс. 5. Край эрозионной кальдеры. 6. Остатки склона вулканического конуса. 7. Центры вулканов внешнего кольца Бёржёни. 8. Центры вулканов внутреннего кольца Бёржёни. 9. Центры вулканов на склоне центрального купола Бёржёни. 10. Центры андезитовых вулканов Дуназуга. 11. Центры дацитовых вулканов Дуназуга. 12. Разломы поствулканические

Таким образом, можно думать, что ошибка в 5,0—6,5% целиком обусловлена факторами, влияющими на потерю или захват аргона. Эта величина вполне правдоподобна в качестве суммарной геологической ошибки. Она возникает в первую очередь вследствие гипогенных и гипергенных изменений, поэтому истинную ошибку можно считать скорее положительной, что дает



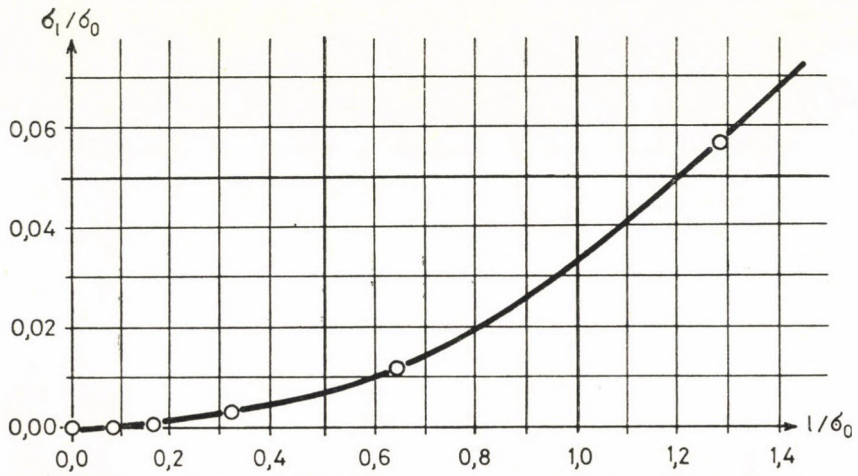
Фиг. 4. Кумулятивный график распределения результатов 35-и определений абсолютного возраста бёржёньских вулканитов калий-аргоновым методом ($\lambda = 0{,}584$). По горизонтали — абсолютный возраст (в логарифмическом масштабе); по вертикали — переменная в интеграле Гаусса (стандартная вероятность); точки — результаты определений абсолютного возраста; линия — прямая Генри, определенная по точкам аналитически (по стандартной программе миникалькулятора HP-67)

лучшее согласование с биостратиграфической датировкой, сдвигая возраст внутрь зоны NN 5.

Общим выводом по анализу характера распределения определений абсолютнбого возраста является невозможность надежных датировок (в миоцене!) с точностью до 5 млн. лет по единичным пробам и необходимость в наличии не менее 30-и проб для обеспечения точности в 2 млн. лет. Без этого на основе радиологических датировок можно придти к совершенно необосно-

ванным выводам вроде того, что бёржёньский вулканизм продолжался 8 млн. лет (разница между крайними значениями) — на целый порядок больше действительного (порядка 0,6 млн. лет).

Стратиграфическое положение *Дуназугского палеовулкана* таково же, что и Бёржёньского: в почве и кровле одни и те же отложения, базальные



Фиг. 5. Зависимость суммарной дисперсии от ширины интервала, в котором истинные значения рассеяны равномерно и имеют постоянную дисперсию. σ_0 — дисперсия точечного значения; σ_l — суммарная дисперсия интервала; l — длина интервала, выраженная в тех же единицах, что и дисперсия; *o* — рассчитанные точки. Регрессия рассчитана по стандартной программе миникалькулятора HP-67.

Расчетные данные

l/σ_0	σ_l/σ_0	a	b	r^2
0,00	1,0000	0,0000	1,0000	1,0000
0,04	1,0000	0,0000	1,0000	1,0000
0,08	1,0000	0,0001	0,9999	1,0000
0,16	1,0001	0,0003	0,9994	1,0000
0,32	1,0003	0,0010	0,9979	1,0000
0,64	1,0012	0,0041	0,9918	1,0000
1,28	1,0057	0,0210	0,9579	0,9999

a и b — коэффициенты в уравнении прямой регрессии типа

$$y = a + bx, \qquad \text{где} \qquad y = \sigma_l/\sigma_0 \quad a \quad x = l/\sigma_0.$$

r^2 — коэффициент регрессии

горизонты прослеживаются из одного массива в другой непрерывно. В отличие от Бёржёньского, в — составе Дуназугского палеовулкана четко обособляются дациты — андезито-дациты от андезитов. Первые образуют полосу экструзивных куполов и их корневых частей протяженностью в 18 км вдоль ЮЗ края вулканической области.

В результате андезитового вулканизма возник двойной сомма-вулкан: поперечник раннего стратовулкана — около 13 км, кальдера обрушения и выросший в ней более поздний конус имеют в диаметре 7 км. Вулканизм начался в эпоху прямого геомагнитного поля, в которой сформировались вулканы внешнего кольца Бёржёни, но протекал в основном в эпоху обратного поля близодновременно с вулканами внутреннего кольца Бёржёни и в ней и закончился. Время жизни Дуназугского андезитового вулкана, следовательно, порядка 200—300 тысяч лет.

Методический опыт изучения Бёржёньского и Дуназугского палеовулканов заключается в следующем:

1. Общая направленность изменения составов антидромна: от более кислых к более основным разностям при вообще-то узком пределе колебаний. В этих пределах петрохимический или петрографический состав является весьма ненадежным критерием сопоставлений, в то время как прямые стратиграфические соотношения (налегание — прорывание), структурная позиция и намагниченность легко увязываются в единую картину.

2. На основе выделения этапов в довольно сложном ходе развития этих палеовулканов и широкого интервала разброса значений абсолютного возраста легко можно было бы сделать вывод о значительной длительности вулканизма. В то же время биостратиграфические и палеомагнитные данные равным образом свидетельствуют о том, что продолжительность формирования этих палеовулканов не превышала 1 млн. лет. Это вполне согласуется с оценками возраста современных крупных и сложных стратовулканов (Лучицкий И. В., 1971; (MacDonald G. A., 1972). Следовательно, сложность и многоэтапность (цикличность) развития, равно как и сам по себе разброс результатов определений абсолютного возраста, ни в коей мере не могут считаться признаками значительной длительности вулканизма.

3. Биостратиграфические датировки на базе способа характерных комплексов бентосной фауны весьма ненадежды, даже в пределах значительных возрастных диапазонов: в случае Бёржёньско—Дуназугского района — от позднего олигоцена до среднего миоцена. К этому можно добавить, что это относится и к среднему олигоцену: нами встречены случаи, когда типично рупельская (кищцельская) фауна формаминифер наблюдалась выше уровня распространения эгерских (хаттских). Поэтому не следует слепо доверять биостратиграфическим датировкам на базе характерных комплексов и основывать на них далеко идущие выводы.

Общим заключением, касающимся всех внутрикарпатских вулканологических обощений, является следующее: необходимо в деталях разобраться в основах датировок, в их надежности и обоснованности и не следует растягивать вулканическую деятельность конкретных центров во времени на базе данных о ее сложности или цикличности и на базе отождествления результатов определений абсолютного возраста с действительным возрастом пород, из

которых взята проба. Надежная схема корреляций внутрикарпатских вулканов может быть создана только при одновременном и массовом использовании трех независимых методов:

1. Биостратиграфического, способом руководящих видов — выделением и прослеживанием планктонных зон.

2. Радиологического, способом статистического анализа — определением наиболее вероятных значений возраста совокупностей заведомо одновозрастных образований.

3. Геологического, способом непрерывного прослеживания маркирующих горизонтов — базальных поверхностей вулканических комплексов и туфово-туффитовых пачек среди осадочных толщ.

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EVOLUTION AND AGE OF THE PALEOVOLCANOES OF THE BÖRZSÖNY AND DUNAZUG MOUNTAINS

By

Z. BALLA, L. KORPÁS, J. CSONGRÁDI

Abstract

The centres of the paleovolcanoes in the Börzsöny and Dunazug Mountains developed on crystalline resp. carbonate basement. In the composition of their products there is practically no difference; in the dislocation zone lying at the boundary of the Veporides and of the Mesozoic formations of the Hungarian Central Mountains not even a secondary centre could be found.

Volcanism was preceded by the accumulation of an Oligocene—Miocene molasse sequence of about 1300 m thickness. The evolution of the paleovolcano of Börzsöny was more complicated and longer. It can be characterized by the following phases: big collapse caldera (20 km) with an external ring of smaller (4 to 6 km) marginal volcanoes, dome (10 to 12 km) in the central part, dome-top collapse caldera (5 to 6 km) with an internal ring of smaller marginal volcanoes; huge stratovolcano (14 km) with an inner collapse caldera (10 km). Within this latter there was a smaller stratovolcano with a lateral collapse caldera on its slope replacing the former one at the dometope. The main phases of evolution of the Dunazug paleovolcano are as follows: huge stratovolcano (13 to 14 km) with a collapse caldera (7 to 8 km) in its centre and within this a smaller stratovolcano.

The age of volcanism is Earliest Badenian according to biostratigraphic data and 14.7 million years according to radiological data (35 samples with 12.0 to 13.5 per cent relative dispersion, i.e. with data varying between 11.6 and 19.4 m.y.). The duration of evolution of the Börzsöny paleovolcano is max. 1 to 2 m.y. according to biostratigraphic data (part of the NN 5 nannoplankton zone), 0.5 to 0.7 m.y. according to paleomagnetic data (two complete and two partial zones). As to the paleomagnetic data the formation of the Dunazug paleovolcano was terminated simultaneously with the formation of the dome-top caldera of Börzsöny; according to geological data, the two centres became active nearly at the same time.

When comparing the three classification sketches of the volcanites, i.e. those based on petrographic–petrochemical, structural–morphological and paleomagnetic characteristics, resp., it can be stated that only the latter two fit each other.

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ВУЛКАНОТЕКТОНИЧЕСКАЯ СХЕМА ТОКАЙСКИХ ГОР

ДЬЯРМАТИ П.

ВЕНГЕРСКИЙ ГЕОЛОГИЧЕСКИЙ ИНСТИТУТ

Линия Самош северозападного-юговосточного направления является самой значительной из структурных линий Токайских гор. Дальше на запад находится другая параллельная линия следующая течению речки Радвань. Геофизические измерения и глубокие бурения вдоль берега реки указывают на то, что фундамент опустился больше чем на 1500 м. С запада Токайские гори ограничиваются более молодой альпийской структурой с морфологической точки зрения совпадающей с долиной реки Хернад и легко заметной тоже геофизическими методами. Существование этой структуры было непосредственно доказано бурениями в районах Черехат и Токай. Эти структурные системы пересекающие друг друга под острым углом играли значительную роль в вулканотектоническом развитии в миоцене. По размерам отдельных формаций и по геофизическим данным можно определить главные вулканотектонические линии. Одна из них проходит параллельно реке Хернад, и две пересекают долину река под острым углом. В основном все характеризуются андезитовым вулканизмом и были активны с позднего тортона по поздний сармат. Вулканотектонические линии Шаторальяуйхей — Фельшерегмец и Телкиваня — Маккошхотька, параллельные реке Хернад характеризуются преобладанием риолотовых пород. Несколько из эруптивных центров только гипотетических из-за значительной денудации первоначальных форм, лежат в точке пересечения этих линий, а другие находятся на самих вулканотектонических линиях. Тесное соотношение обнаружено между линейным вулканизмом (типа «грабен») и вулканотектоническими линиями. Места центров поствулканических гейзерных действий и места гидротермальных жил силикатного и карбонатного состава приурочены главным образом к структурно-вулканотектоническим линиям.

Токайские горы являются одним из районов развития внутрикарпатского третичного вулканизма. Группа геологосъемщиков Венгерского геологического института под руководством Габора Панто с 1958 по 1968 гг. составила геологическую карту этого района в масштабе 1 : 25 000. Первое обобщение научных результатов этих работ представляли собой карта «Шатораляуйхей» и объяснительная записка к ней, опубликованные внутри серии геологической карты Венгрии масштаба 1 : 200 000. Составлено описание в виде монографии вулканитов среднего и кислого состава и осадочных отложений, опубликование этих работ осуществляется в настоящее время. В части геофизических исследований были проведены обзорные по масштабу геомагнитные съемки, а также геоэлектрические, геомагнитные и сейсмические измерения методом преломленных волн небольших участков в пределах рассматриваемого района. В результате проведенных геологических и

геофизических работ была достигнута соответствующая степень изученности геологического строения, структурных особенностей и истории геологического развития рассматриваемого района.

1. *Геологическое строение*

В отношении стратиграфических условий и строения Токайских гор опубликованы современные по своему уровню обобщающие работы (Г. Панто в работе Б. Боцан и др., 1966; Э. Илкеи-Перлаки, 1972; Л. Пентеленьи, 1972; П. Дьярмати, 1977). На основании этих работ схема геологического строения района сводится к следующему: протерозой представлен выходами гнейсов и слюдяных сланцев в районе Фельшёрегмец—Вийвитань у государственной границы. Ордовикская система представлена порфироидами-филлитами в линии Фюзеркаята—Фельшёрегмец, вскрытыми глубоким бурением. Аналогично им, также только из глубоких скважин, известны силурско-девонские глинистые сланцы, подстилающие, по всей вероятности, вулканиты в западной части Токайских гор. Пермокарбоновые образования, представленные песчаниками и конгломератами с кварц—порфировым вулканизмом, известны как в виде выходов, так и из глубоких скважин. В основании северо-восточной части Токайских гор глубокими скважинами вскрыты красные, пестроцветные песчаники пермо-триаса, а также известняки триаса и юры. Первое образование, залегающее после крупного перерыва на палеозойских и мезозойских отложениях, это нижнетортонские пирокластические потоки риодацитовых туф («средние риолитовые туфы»), встречающиеся на довольно большой площади и в пределах Паннонского бассейна. В результате последовавшей за этим образованием трансгрессии началась седиментация глинистых, тонкопесчанистых морских осадков, которой сопутствовала ритмичная, андезитово-риолитовая, подавляющей частью подводная вулканическая деятельность. Она достигла наибольшей интенсивности в сарматском веке. Впоследствии выдержанное, беспрерывное морское покрытие района разрывалось в систему лагун, а подводный вулканизм сменился вулканической деятельностью в наземных условиях. К продуктам субсеквентного вулканизма относим оливиновые базальты, встречающиеся на границе миоцена и плиоцена в районе г. Шарошпатак.

2. *Палеогеографические условия*

Палеогеографические условия, вырисовывающиеся на основании вышесказанного, от протерозоя до пермо-карбона остаются довольно неясными. Исходные породы гнейсов и слюдяных сланцев были созданы протерозойским магматизмом и осадконакоплением, точно так же как и порфироиды и филлиты, указывающие на вулканизм и седиментацию, имевшие место в ордовикское время. В условиях прибрежного осадконакопления образова-

лись силурско девонские, темновато-серые глинистые сланцы с небольшими прослоями известняка и прожилками антрацито-графита. В пермо-каменноугольное время в результате размыва кристаллического фундамента в районе Вийвитань—Фельшёрегмец образовались, преимущественно в наземных и в меньшей степени в прибрежных условиях, песчаниковые конгломераты с углем; залегающие внутри этой толщи известняковые линзы свидетельствуют о небольшой трансгрессии. Вулканизм представлен кварцевыми порфирами и их туфами. Переход из перми в триас характеризуется терригенным осадконакоплением с красными, пестроцветными песчаниками с приуроченным к ним небольшим по объему вулканизмом среднего состава (порфириты). С наступлением трансгрессии к концу раннего триаса началось карбонатное осадконакопление с образованием пластинчатых, а затем массивных известняков. Для стратиграфического интервала от нижнего триаса до верхней юры у нас не имеется данных, а верхняя юра характеризуется, опять-таки, седиментацией известняков в морских условиях. В конце мезозойской эры и в палеогеновой эпохе рассматриваемый район, судя по имеющимся данным, представлял собой сушу, подвергшуюся денудации. За вулканической деятельностью, начавшейся в миоцене, последовала трансгрессия с северо-восточного направления, со стороны неогеновой депрессии Восточной Словакии. Также с востока наступила, на наш взгляд, позднетортонская трансгрессия, охватившая всю территорию Токайских гор, охарактеризованная накоплением в прибрежных условиях глинисто-тонкопесчанистых осадков, параллельно которому происходил многократно повторявшийся вулканизм, от кислого до среднего состава. Опускание района, происходившее вместе с вулканической деятельностью в подводно-морских условиях, достигло своего максимума около современной оси Токайских гор, с глубиной примерно 2000 м; таким образом подавляющая масса быстрорастущего вулканического сооружения всё время оставалась под уровнем моря. Начало сарматского века характеризовалось регрессией, постепенным — но неравномерным — опреснением тортонского моря и его отступлением назад, а также солоноватоводным осадконакоплением и разобщением сплошного водоема в систему изолированных лагун. Основным фактором морфогенеза осталась по-прежнему вулканическая деятельность, в перерывах вулканизма происходила интенсивная гидротермальная, кремнисто-карбонатная поствулканическая деятельность. В заливах, окружающих позднесарматские лагуны, материал последней разнообразно перемешивался с материалом переотложенных вулканитов, в основном — пирокластитов, но чисто биогенное осадконакопление здесь также имело место. От тортона, через сарматский век до нижнепаннонского времени, при постепенном, но неравномерном опреснении водоема, происходила литорально-неритическая седиментация, которая только временами прерывалась небольшим по объему вулканизмом. В плейстоцене присутствие красных глин, саманов и лессов свидетельствует уже о континентальном осадконакоплении.

3. *Тектонические условия*

Реконструированная на основании геологического строения, история геологического развития района тесно связана с его тектогенезом.

Гнейсы и слюдяные сланцы образовались после протерозойского осадконакопления и магматизма, под влиянием ассинтской складчатости рифейского периода (962 мин лет). Ордовикские порфироиды-филлиты подверглись каледонскому метаморфизму (450 мил лет). В герцинскую, судетскую и австрийскую фазы складчатости (262 мин лет) серия гнейсов и слюдяных сланцев вдоль линии Сомеш были чешуеобразно вдвинуты в ордовикскую порфировидно-филлитовую толшу. Проявления динамометаморфизма, обнаруживаемые в более древних членах пермо-каменноугольной толщи, указывают на заальскую фазу складчатости. В период континентальной денудации, длившейся до середины миоцена, были большей частью выравнены геоморфологические различия, обусловленные расхождениями по геологическому строению и ходу развития между тремя структурными единицами, приуроченными к герцинскому и раннеальпийскому орогенезу, а именно — между северо-восточной, сложенной гнейсами, слюдяными сланцами, порфироидами-филлитами, песчаниками, конгломератами и кварцевыми порфирами; восточной, состоящей из триасово-юрских известняков и западной, сложенной глинистыми сланцами.

Ход образования вулканитов Токайских гор показан на структурно-вулканотектонической схеме рис. 1.

Из числа структурных линий безусловно наиболее значительной является линия реки Сомеш, протягивающаяся в юго-восточном направлении у г. Шатораляуйхей, вдоль которой произошёл сброс палеозойско-мезозойского блока, представленного на земной поверхности Земплинскими островными горами. Сброшенный в юго-западном направлении блок в настоящее время находится — за исключением Шарошпатакской триасовой глыбы приподнятого положения — на глубине от 500 до 700 м. Данная линия сыграла, повидимому, значительную роль также и в извержении продуктов вулканизма, имевшего свой центр тяжести в северо-восточной части района Тисантул (то есть территории Венгрии, расположенной к востоку от реки Тисса). В результате оживления этой линии произошло извержение так называемых «средних риолитовых туфов», охарактеризовавшееся грабенным эксплозивно-эффузивым механизмом. Субпараллельно ей, несколько западнее, тянется структурная линия, следующая направлению речки Радвань. Согласно геофизическим и буровым данным вдоль этой линии дотретичный фундамент опущен, опять-таки в юго-западном направлении, до глубины более 1500 м. Токайские горы с запада ограничены более молодой альпийской тектонической линией, морфологически совпадающей с долиной реки Хернад. Она отчетливо выявляется также и на основании геофизических материалов, при-

чем ее существование непосредственно подтверждено скважинами, пробуренными на территории Черехата и Токайских гор. В создании вулканотектонической депрессии, формировавшейся в тортонское время, определенную роль

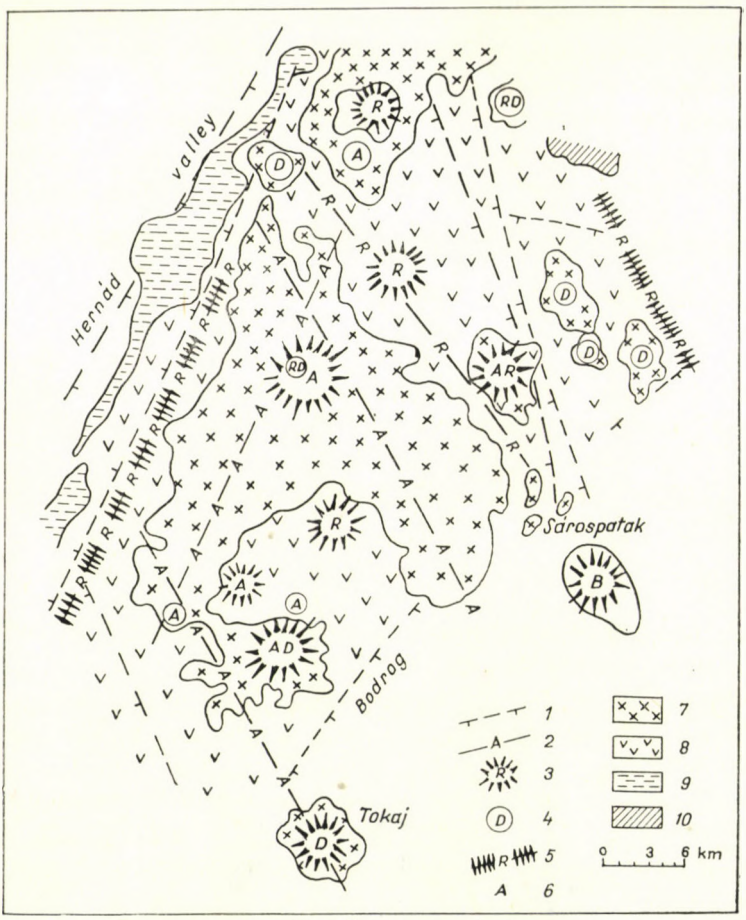

Рис. 1. Вулканотектоническая схема Токайских гор. 1. Структурная линия, 2. вулканотектоническая линия, 3. центр ерупции (гипотетический), 4. субвулканическая тела, 5. линейный (трещинный) вулканизм, 6. R = риолитовые породы, RD = риодацитовые породы, D = дацитовые породы, A = андезитовые породы, B = базальтовые породы, 7. вулканнческая формация (среднего состава), 8. вулканческая формация (кислого состава), 9. осадочная формация, 10. кристаллические породы

сыграла, повидимому, и эта линия, а более глубокие разломы открыли путь для извержения вулканических материалов среднего состава. На основании распространения отдельных образований и по геофизическим материалам можно выделить главные вулканотектонические линии. Преимущественно андезитовые по составу продукты образовались вдоль линии,

тянущейся параллельно реке Хернад и двух других линий, образующих с ней острый угол, которые были в действии от позднего тортона до позднего сармата. Продукты преимущественно риолитового состава приурочены к линиям Шатораляуйхей-Фельшёрегмец (с верхним тортоном в его центре) и Телкибанья-Маккошхотька (с сарматом в центре), тянущимся параллельно линии Сомеш и к вулканотектонической линии Гёнц-Абауйсанто, параллельной реке Хернад (верхний сармат-нижний паннон).

Состав поднимавшейся магмы, палеогеографическая обстановка, фундамент и уже упомянутые структурные — вулканотектонические линии были факторами, определившими форму вулканизма, а также и способ его проявления. Часть предполагаемых центров извержений находится как раз в точке пересечения этих линий, а другие располагаются на самих линиях. Еще теснее связь между трещинным вулканизмом (грабенные эксплозии) и вулканотектоническими линиями. Аналогичная зависимость обнаруживается между центрами поствулканической деятельности и тектоническими линиями.

Продолжающаяся вулканическая деятельность проявлялась — видимо вследствие дизъюнктивных движений, последовавших за II фазой аттической складчатости — в виде вулканизма центрального типа в вулканитах риолитового состава горы Сокоя у с. Эрдёхорвати и в виде линейного вулканизма молодых раскалённых лавин риолитовых туфов, извергшихся вдоль линии Хернад в результате эксплозивно-эффузивных проявлений. К субсеквентному вулканизму относим оливиновые базальты, встречающиеся при г. Шарошпатак.

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VOLCANOTECTONIC SKETCH OF THE TOKAJ MOUNTAINS

P. GYARMATI

Abstract

Among the structure lines of the Tokaj Mountains the NW—SE trending Szamos line is the most significant. By and large parallelly, farther to the west, another fault stretches according to the course of the Radvány brook along, which as shown by both geophysicale measurements and deeper drills, the basement has subsided down to more than 1500 m depth. The

Tokaj Mountains are bounded in the west by a younger Alpine structure line coinciding morphologically with the valley of the river Hernád, being easily detectable by geophysical methods too, and its existence has been proved in a direct way by boreholes put down in the Cserehát and within the Tokaj Mountains area. These two structure systems intersecting each other in an acute angle were crucial for the volcanotectonic evolution having taken place in the Miocene. On the basis of the extent of the individual formations and of geophysical information, the main volcanotectonic lines can be delineated. One of this runs parallel to the Hernád river, two form an acute angle with it. All three are predominantly andesitic in composition, having been active from Late Tortonian to Late Sarmatian times. An overwhelmingly rhyolitic composition is characteristic of the zones of the volcanotectonic lines of Sátoraljaújhely—Felsőregmec and of Telkibánya—Makkoshotyka running parallel to the Szamos Line and of the Gönc—Abaújszántó line parallel to the Hernád river. Some of the eruption centres, only hypothetical because of the large scale denudation of the original landforms, are situated at the intersections of these lines, while others lie on the volcanotectonic lines. An even closer relationship exists between linear-fissure volcanism (graben explosions) and volcanotectonic lines. The locations of the centres of postvolcanic geyser activites and the sites of hydrothermal veins of siliceous and carbonate composition, have been largely controlled by the structural–volcanotectonic lines already mentioned.

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BEITRAG DER GEODÄSIE ZU DER UNTERSUCHUNG DER REZENTEN VERTIKALEN ERDKRUSTENBEWEGUNGEN IN DER KARPATEN—BALKAN-REGION

Von

I. Joó

LANDESAMT FÜR BODEN UND KARTENWESEN, BUDAPEST

Im Rahmen des KAPG wurde die rezente vertikale Krustenbewegungskarte Osteuropas hergestellt und im Jahre 1971 in Moskau vorgelegt. Auf der Grundlage der kritischen Bewertung der Karte und der benutzten Daten hat eine ausführlichere Untersuchung der Fläche der Karpaten—Balkan-Region stattgefunden. In die Untersuchung haben wir (außer der Tschechoslowakei und Bulgarien) auch Jugoslawiens und Rumäniens geodätisches Material, sowie auch die Daten der an den Küsten des Adriatischen Meeres funktionierenden Mareographen einbezogen.

Der Ausgleich der geodätischen und der ozeanographischen Daten kam in mehreren Variationen zur Durchführung. Auf der Grundlage des Ausgleichs hat sich die Geschwindigkeit und die Richtung der Krustenbewegung an vielen Stellen geändert. Auf Grundlage der ausgeglichenen Daten wird gegenwärtig die Redigierung einer Karte der rezenten vertikalen Krustenbewegungen des Maßstabes 1 : 1 000 000 durchgeführt. Die Vervielfältigung der Karte findet im zweiten Halbjahr 1979 statt. An dem aus zwe Kartenausschnitten bestehenden Kartenwerk werden wir — außer den üblichen Elementen und Isolinien — auch die wichtigsten tektonischen und seismologischen Daten anführen.

Im Rahmen der Abhandlung findet auch eine skizzenhafte Vorlegung des Kartenwerkes statt und weiter werden die Daten der neuesten Untersuchungen den Daten der rezenten vertikalen Krustenbewegungskarte Osteuropas gegenübergestellt.

Bekanntlich fand im Rahmen der Zusammenarbeit der Akademien der Wissenschaften und der Geodätischen Dienste osteuropäischer sozialistischer Länder die ausführliche Untersuchung der rezenten vertikalen Erdkrustenbewegungen Osteuropas statt, und auf deren Grundlage wurde auch eine zusammenfassende Karte im 2,5 Millionen- Maßstab hergestellt. Diese Karte ist der plenaren Sitzung der GGIU im Jahre 1971 in Moskau vorgelegt worden. Diese Untersuchung hat sich natürlich auch auf den größten Teil der Karpaten—Balkan-Region (KBR) erstreckt. Am gemeinsamen Ausgleich des zur Verfügung stehenden Untersuchungsmaterials hat unter anderen auch das Material Ungarns, der Tschechoslowakei und Bulgariens teilgenommen.

Eine Schwierigkeit verursachte der Umstand, daß sich Jugoslawien und Rumänien dem Programm nur später angeschlossen hatten, und aus diesem Grunde

— konnte das bulgarische Material der übrigen Länder unmittelbar nicht angeschlossen werden, und weiter

— konnten die Daten der an den rumänischen, bulgarischen und jugoslawischen Küsten befindlichen Mareographen bei dem gemeinsamen Ausgleich nicht angewandt werden.

Später, als Jugoslawien und Rumänien mit der Zusammenstellung des eigenen Materials fertig geworden sind, hat sich die Möglichkeit geboten, dieses Material dem gemeinsamen Ausgleichsmaterial anzuschließen. Diese Lösung konnte natürlich nur annähernden Wertes sein, und löste dabei das grundlegende Problem nicht auf, nämlich, daß das Material der Länder der Karpaten—Balkan-Region (Tschechoslowakei, Ungarn) an die südwestliche Ecke des geprüften Netzes geraten ist, ohne, daß sich der Anschluss zu den in diesem Raum funktionierenden Mareographen ereignet hätte.

Auf diese Weise sind die auf geodätische Daten gestützten abgeleiteten Geschwindigkeitswerte, beziehungsweise die auf dieser Grundlage konstruierte Krustenbewegungs-Karte stellenweise mit sonstige geologischen und geomorphologischen Kenntnissen in Widerspruch geraten.

Diese Umstände in Betracht genommen, haben wir seitens Ungarns vorgeschlagen, das zur Verfügung stehende Untersuchungsmaterial der Karpaten—Balkan-Region zusammen mit dem nachträglich bereits hergestellten rumänischen und jugoslawischen Material aufs neue zu ordnen, die notwendigen Ergänzungen durchzuführen und den—jetzt schon gemeinsame—Ausgleich des Materials von neuem vollzuziehen. Diese Aufgabe wurde im Rahmen des KAPG akzeptiert, und mit der Koordinierung ist Ungarn beauftragt worden.

Seitens Ungarns haben wir die Ordnungsaufgabe des Materials der berührten Länder, dessen Ergänzung mit Hilfe der Teilnehmer durchgeführt, und die Ausgleichung des KBR-Untersuchungsnetzes wurde in mehreren Variationen von neuem vollzogen.

Über diese Ausgleichsarbeit haben wir im Rahmen des im Laufe der in Grenoble stattgefundenen Beratung der GGIU veranstalteten Symposiums schon Bericht erstattet. Hier möchte ich bloß die wesentlichsten Charakteristika des Ausgleichs skizzieren.

Zum Untersuchungsnetz gehörten acht Mareographen: am Baltischen Meer, am Schwarzen Meer und am Adriatischen Meer. Das Netz bestand aus 133 geschlossenen Polygonen, die 389 Linien und 249 Knotenpunkte enthielten. Zwischen den im Untersuchungsnetz benutzten beiden Nivellements war die durchschnittliche Zeitdauer 32 Jahre. Die minimale Zeitdauer hat 6 Jahre, die maximale 76 Jahre betragen.

Der Ausgleich wurde mit Korrelatenverfahren durchgeführt. Auf diese Weise lieferte der Ausgleich die widerspruchfreien Geschwindigkeitsänderungen zwischen den Knotenpunkten. Darauffolgend ist die Berechnung der Knotenpunkt-Geschwindigkeiten — von den bekannten Geschwindigkeitswerten der Mareographen ausgehend — möglich geworden.

$\Delta V_I = V_{Allg.-KBR} - V_{Osteuropäisch}$

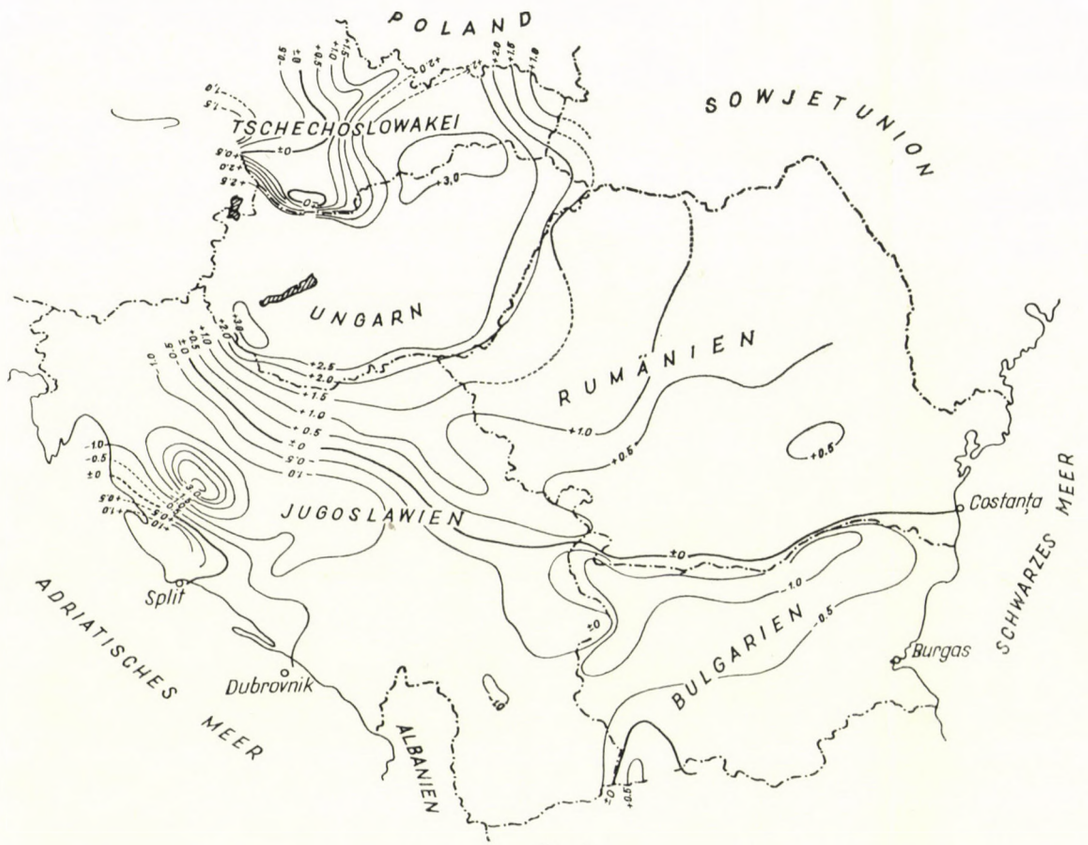

Abb. 1

$\Delta V_{II} = V_{Spez.-KBR} - V_{Allg.-KBR}$

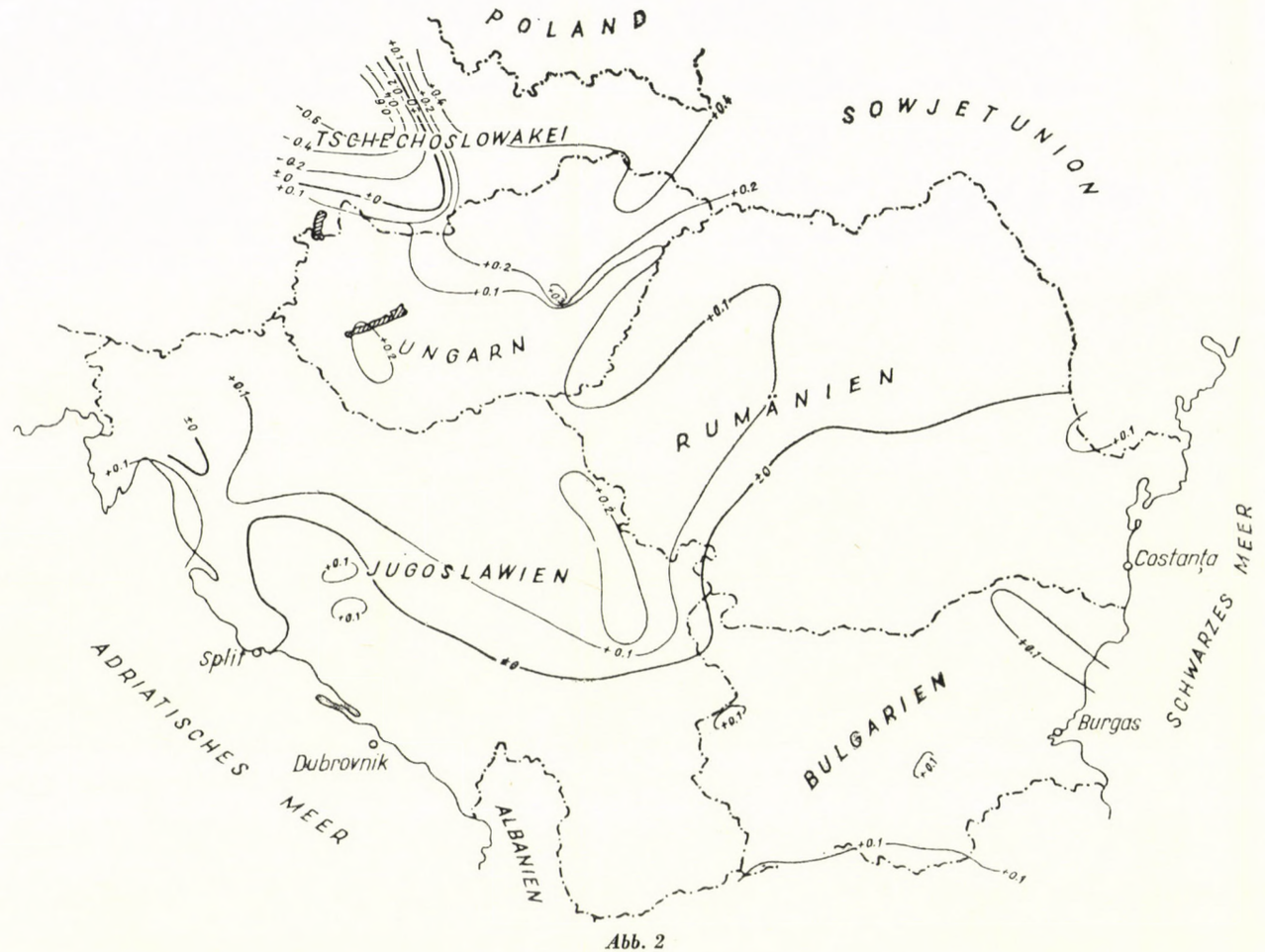

Abb. 2

Eigentümlichkeit der Ausgleichung war, daß sie nicht bloß einmal, sondern in sieben verschiedenen Variationen durchgeführt wurde. Laut dieser wurden bei dem ersten sogenannten "allgemeinen KBR-Ausgleich" sämtliche zur Verfügung stehenden geodätischen und ozeanographischen Daten benutzt.

Bei der zweiten Ausgleichsvariation — bei dem sogenannten "beschränkten KBR-Ausgleich" — kamen bloß die geodätischen und ozeanographsischen Daten der Karpaten—Balkan-Region zur Anwendung. Wesentlich ist also, daß wir aus diesem zweiten Ausgleich die Beobachtungsdaten der an den Küsten des Baltischen Meeres befindlichen Mareographen ausgeschlossen haben.

Das Ziel der letzten fünf Variationen war wesentlich, daß wir bezüglich des Höhenhauptfestpunktes Nadap die absoluten Geschwindigkeitswerte gesondert ableiten sollen, immer nur von den Mareographdaten eines einzigen Landes ausgehend. Dementsprechend haben wir für Nadap fünf verschiedene Werte erhalten, auf Grund der Mareographen der DDR, Polens, Rumäniens Bulgariens und Jugoslawiens. Die aus den sieben Ausgleich-Variationen für Punkt Nadap abgeleiteten absoluten Geschwindigkeitswerte sind in Tabelle I zu finden.

Die zu den in Tabelle I angeführten absoluten Geschwindigkeitswerten berechneten mittleren Fehler befinden sich zwischen $\pm 0{,}78$ mm/Jahr und $\pm 1{,}59$ mm/Jahr.

Von den, zu den abgeleiteten Geschwindigkeitswerten gehörenden mittleren Fehlern werden ihre eigentlichen Werte selbst ziemlich angenähert, daher müssen die abgeleiteten Werte mit gewissem Vorbehalt behandelt werden.

Tabelle I

Ausgleich-Variation	Absoluter Geschwindigkeitswert des Punktes Nadap (V_{Nadap} mm/Jahr)
1. Allgemeiner KBR-Ausgleich (Karpaten—Balkan-Region)	+0,44
2. Beschränkter KBR-Ausgleich	+0.57
3. DDR-Polygon	+1,94
4. Polnisches Polygon	−1,76
5. Rumänisches Polygon	−1,99
6. Bulgarisches Polygon	+1,92
7. Jugoslawisches Polygon	+1,55

Auf Grund von Tabelle I dürfen wir im Zusammenhang mit der Bewegung vom Punkt Nadap folgendes sagen:

— Die auf Grund der polnischen und rumänischen Polygone durchgeführte Ausgleichung zeigt Senkung,

— sämtliche anderen Ausgleich-Variationen zeigen aber eine Erhöhung.

Es soll aber erwähnt werden, daß wir bei zwei Untersuchungsmaterialien von mehreren solchen Voraussetzungen Gebrauch machen mußten, die deren Zuverlässigkeit herabsetzen.

Wir müssen auch darauf aufmerksam machen, daß sämtliche bisherigen geologischen, geophysikalischen Untersuchungen für die Umgebung des Punktes Nadap eine geringere Erhöhung abgeleitet haben. Dies wird auch durch die Angaben der Tabelle I — mit Ausnahme von zwei Werten — unterstützt.

Auf diese Weise dürfen wir auf Grund der Untersuchung aussprechen, daß Punkt Nadap eine sanft ansteigende Bewegung zeigt, deren Wert (auf Grund der allgemeinen Ausgleich-Variation) + 0,4 mm/Jahr beträgt.

In Anbetracht dessen, daß bezüglich der rezenten vertikalen Erdkrustenbewegung der Karpaten—Balkan-Region zwei — verhältnismäßig selbständige — Ausgleiche stattgefunden haben (namentlich die osteuropäische und die von uns durchgeführte, sogenannte KBR-Untersuchung), so ergibt sich für uns die Möglichkeit, die Daten der zweierlei Untersuchungen vergleichen zu können.

Im weiteren machen wir kurz die Ergebnisse des osteuropäischen Ausgleichs bezüglich der KBR bekannt, dann zeigen wir das sich ausgestaltende Bild der von uns durchgeführten Untersuchungen; zum Abschluß machen wir das Ergebnis der Gegenüberstellung der Daten der zweierlei Untersuchungen bekannt.

Über die osteuropäische Erdkrustenbewegungs-Karte im 1 : 2,5 Millionen-Maßstab können wir folgendes sagen:

Die sich in der KBR meldenden größeren Senkungsgebiete sind also:

— im allgemeinen das Karpatenbecken,
— ein bedeutender Teil des Transsylvanischen Beckens,
— Rumäniens südlicher Teil und östlicher Küstenteil,
— Einige, unmittelbar an die Küste grenzenden Gebiete Jugoslawiens.

Sich erhöhende Gebiete sind:

— der größte Teil Jugoslawiens und Bulgariens,
— die Süd-, Ost- und Nordostkarpaten eindeutig,
— weiters, ein bedeutender Teil der Nordkarpaten.

Hingegen zeigt sich auf Grund der in der Karpaten—Balkan-Region durchgeführten neueren Untersuchungen und Ausgleich über die vertikalen Erdkrustenbewegungen folgendes Bild:

— an den, zu den Karpaten gehörenden Landesgebieten der Tschechoslowakei zeigt sich größtenteils eine Erhöhung, deren Maximalwert + 2 mm/Jahr beträgt, es zeigen sich aber auch bedeutende Senkungszonen bis zum Maximalwert — 2 mm/Jahr (entscheidend in der Westslowakei),

— im Landesgebiet Ungarns zeigt sich grösstenteils eine Erhöhung, deren Geschwindigkeitswerte von — 0,5 mm/Jahr (Westungarn) bis + 1,7 mm/Jahr (Nordostungarn) betragen,

— im Landesgebiet Rumäniens zeigt sich im allgemeinen Erhöhung, deren Maximalwert + 3 mm/Jahr beträgt, dabei sind nördöstlich von Brasov, sowie entlang der ungarisch-rumänischen Grenze Senkungszonen zu finden,

— in Landesgebiet Jugoslawiens wird die im allgemeinen zur Geltung kommende Erhöhungstendenz auch durch die neueren Untersuchungen bestätigt, zwar hat sich das Maß der Erhöhung etwas gemäßigt. Die Bewegungswerte bilden im nördlichen Teil Jugoslawiens einen guten Übergang zu den im Landesgebiet Ungarns ausgewiesenen Bewegungen. Die an den Küsten des Adriatischen Meeres bereits früher ausgewiesene Senkung ist auch jetzt vorhanden und hat sich sogar etwas verstärkt. In den mittleren Teilen Jugoslawiens hat sich die Senkungstendenz gemäßigt,

— in der Beziehung Bulgariens hat sich die Erhöhungstendenz gleichfalls gemäßigt, beziehungsweise ist sie nicht so eindeutig. Die Tendenz der Senkungszone in der Umgebung von Burgas ist ausdrücklicher geworden.

Der Vergleich des für die Karpaten—Balkan-Region abgeleiteten "Osteuropäischen Untersuchungsmaterials" und des von Ungarn koordinierten Karpaten—Balkan-Regions-Untersuchungsmaterials und dessen Ergebnisse sind in Tabelle II zu sehen.

In der zweiten Spalte der Tabelle haben wir die Anzahl der identischen Knotenpunkte angegeben. In der dritten Spalte sind die Differenzen der Geschwindigkeitswerte zu sehen. Hier hat sich ΔV derart ergeben, daß wir von dem in dem allgemeinen KBR-Ausgleich erhaltenen Wert ($V_{\text{allg.—KBR}}$) den aus dem osteuropäischen Ausgleich erhaltenen Wert ($V_{\text{osteuropäisch}}$) subtrahierten.

Tabelle II

Land	Anzahl der identischen Knotenpunkte	Verteilung der ΔV-Werte nach Vorzeichen		ΔV mm/Jahr	Extreme Werte von $\lvert \Delta V \rvert$	
		+	—		$\lvert \Delta V \rvert_{max}$	$\lvert \Delta V \rvert_{min}$
		St.	St.		mm/Jahr	
1	2	3		4	5	
Tschechoslowakei	11	11	—	+2,6	3,2	0,3
Ungarn	31	31	—	+2,7	3,2	2,0
Jugoslawien*	46	15	31	—0,4	3,6	0,2
Rumänien*	19	18	1	+0,7	1,9	0,3
Bulgarien	31	1	30	—0,6	1,5	0,1
Zusammen	138	76	62	+0,6	3,6	0,1

* Bei der Zusammenstellung der hier mitgeteilten Daten stammen die früheren absoluten Geschwindigkeitswerte (V) nicht aus dem gemeinsamen osteuropäischen Ausgleich, sondern aus den Zusammenstellungen der einzelnen Länder

$\Delta V = V_{KBR} - V_{\text{osteuropäisch}}$

V = Wert der absoluten Geschwindigkeit, mm/Jahr

In der vierten Spalte der Tabelle haben wir nach Ländern und auch zusammen den vorzeichenrichtigen mittleren Wert der ΔV-Werte angegeben. In der letzten Spalte sind die extremen Werte der absoluten Werte von $|\Delta V|$ zu sehen.

Die Abweichungen der aus zweierlei Ausgleichen erhaltenen Geschwindigkeitswerte werden auch in Form von Isolinien dargestellt. Als ersten zeigen wir auf Abb. 1 die Darstellung der in Tabelle II vorkommenden ΔV. Bei dieser beträgt der Grundwert der Isolinien 0,5 mm/Jahr. In der Abbildung wird die Gestaltung der Geschwindigkeitdifferenzen in den einzelnen Ländern gut veranschaulicht. Hier können wir einerseits auf die West-Slowakei hinweisen, wo sich ziemlich bedeutende Abweichungen ergeben haben, andererseits darauf, daß auf der Landesfläche Ungarns die Abweichungen vermutlich eindeutig sind.

In der Abbildung 2 zeigen wir wesentlich das, in welchem Maße bei der im Rahmen des KBR-Programmes durchgeführten zwei Ausgleichsvariationen die aus der ersten und zweiten Variation erhaltenen Werte voneinander abweichen. Die hier angewandten Grundwerte betragen 0,1, beziehungsweise 0,2 mm/Jahr. Es ist zu sehen, daß die Ergebnisse der ersten und der zweiten Ausgleichungsvariation nur in geringen Maße voneinander abweichen und daß sich diese Abweichung in der ganzen KBR kaum ändert.

Im Laufe der Untersuchungen haben wir auch die Häufigkeit der ΔV-Werte in Funktion ihrer Werte gesammelt. Das Ergebnis dessen stellen wir — nach Ländern zusammengestellt — mit Hilfe von 5 Diagrammen dar (Abb. 3—7). Die Diagramme zeigen sowohl bei Ungarn, wie auch bei Bulgarien die nahezu eindeutige regelmäßige Änderung der Geschwindigkeitswerte. Im Fall der anderen drei Länder ist die Änderung der Geschwindigkeiten schon weniger eindeutig.

Auf Grund des Vergleiches der zweierlei Untersuchungsmaterialien können wir folgendes aussprechen.

1. Auf dem ganzen untersuchten Gebiet zeigten 70% der Knotenpunkte eine Senkungstendenz, die sich im Laufe der neueren Untersuchung jetzt beachtenswert gemäßigt hat.

2. Auf der zu dem Karpatenbecken, beziehungsweise zu den Karpaten gehörenden Landesfläche der Tschechoslowakei haben die früheren Untersuchungen für etwa zwei Drittel Senkungen abgeleitet. Dies hat sich im Laufe der KBR-Untersuchungen verringert. In Tabelle II haben wir gezeigt, daß die Differenzen der für die Knotenpunkte erhaltenen zweierlei Geschwindigkeiten in alles Fällen positiven Vorzeichens sind. Der vorzeichenrichtige Durchschnitt der Geschwindigkeitswert-Änderungen beträgt +2,6 mm/Jahr.

3. Auf der Landesfläche Ungarns haben sich die absoluten Geschwindigkeitswerte durchschnittlich um 2,7 mm/Jahr erhöht; aus fünf Ländern ist dies die bedeutendste Änderung. Auf diese Weise zeigt sich anstatt der frühere

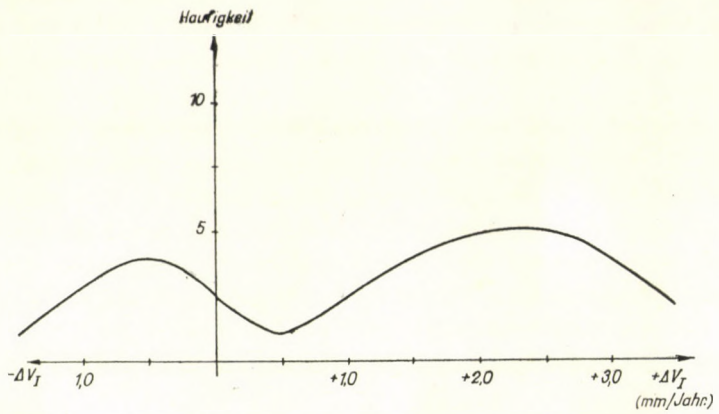

Abb. 3. Tschechoslowakei. $\Delta v_I = v_{\text{alg.-KBR}} - v_{\text{osteuropäisch}}$; Intervall: 10 mm/Jahr; $\Delta v_{\text{I-Durchschnitt}} = +2{,}6$ mm/Jahr

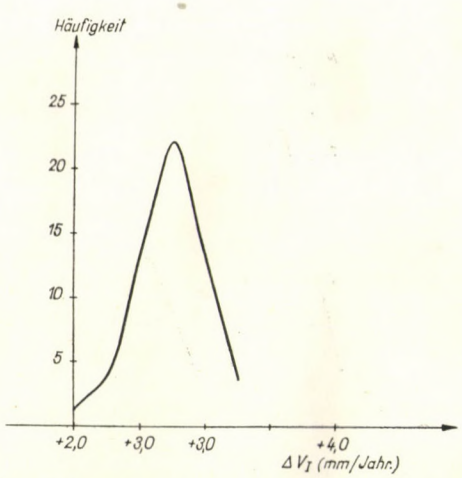

Abb. 4. Ungarn. $\Delta v_I = v_{\text{allg.-KBR}} - v_{\text{osteuropäisch}}$
Intervall: 0,5 mm/Jahr; $\Delta v_{\text{I-Durchschnitt}} = +2.7$ mm/Jahr

für die Landesfläche Ungarns abgeleiteten 100prozentigen Senkungstendenz jetzt größtenteils eine Erhöhung. Eine Ausnahme bilden Westungarn und zum Teil Ost-, beziehungsweise Südost-Ungarn.

4. In der Beziehung Jugoslawiens sind die für die Knotenpunkte abgeleiteten Geschwindigkeits-Differenzen größtenteils negativen Vorzeichens, Der vorzeichenrichtige Durchschnitt der Geschwindigkeitsdifferenzen beträgt —0,4 mm/Jahr, das heißt auf Grund des neueren Untertuchungsmaterials der Karpaten—Balkan-Region wird für die Landesfläche Jugoslawiens die Senkungstendenz einen größeren Platz finden, als durch die osteuropäische gemeinsame Untersuchung.

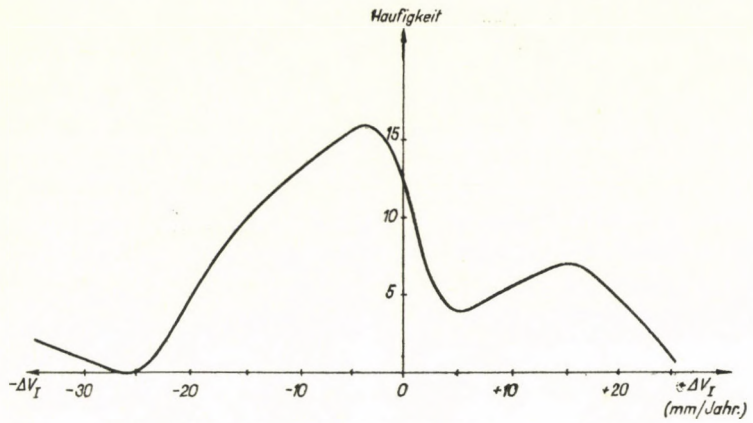

Abb. 5. Jugoslawien. $\Delta v_I = v_{\text{allg.–KBR}} - v_{\text{osteuropäisch}}$; Intervall: 10 mm/Jahr; $\Delta v_{I\text{–Durchschnitt}} = -0.4$ mm/Jahr

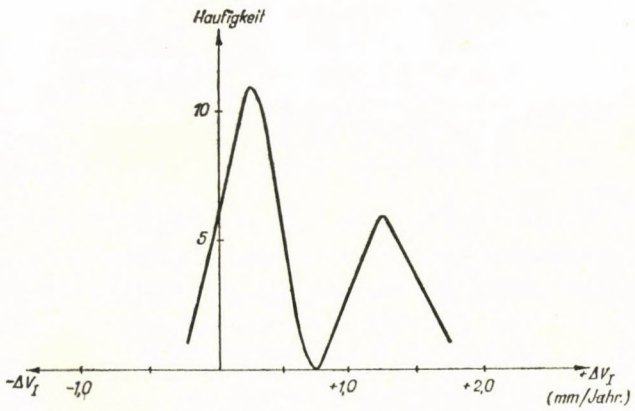

Abb. 6. Rumänien. $\Delta v_I = v_{\text{allg.–KBR}} - v_{\text{osteuropäisch}}$; Intervall: 0,5 mm/Jahr; $\Delta v_{I\text{–Durchschnitt}} = 0{,}7$ mm/Jahr

5. Anf der Landesfläche Rumäniens änderten sich die absoluten Geschwindigkeitswerte — mit Ausnahme eines Knotenpunktes — im positiven Sinne, und der vorzeichenrichtige mittlere Wert beträgt +0,7 mm/Jahr; das heißt, die Untersuchung der Karpaten—Balkan-Region förderte — wie in Ungarn — auch auf der Landesfläche Rumäniens die Erhöhungstendenz.

6. Auf der Bulgariens Landesfläche wurden durch die neuere Untersuchung der Karpaten—Balkan-Region die Geschwindigkeitswerte durchschnittlich um —0,6 mm/Jahr verändert, das heißt, die für die Landesfläche Bulgariens abgeleitete, ziemlich allgemeine Erhöhungstendenz wurde gemäßigt.

Auf Grundlage der durchgeführten Untersuchungen und der aus dem Ausgleichen erhaltenen Geschwindigkeitswerte wird die Redigierung einer

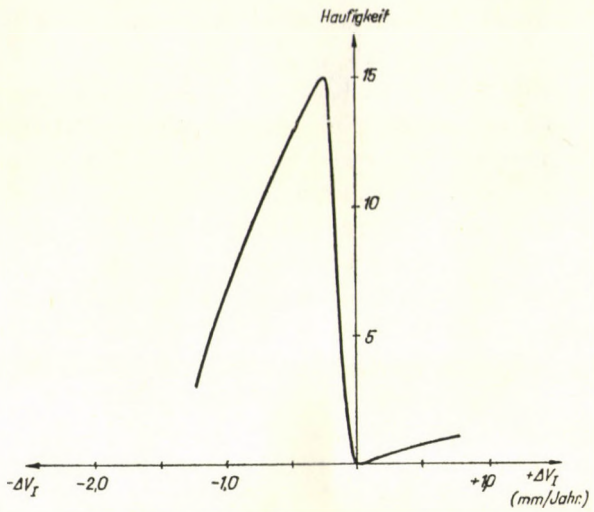

Abb. 7. Bulgarien. $\Delta v_I = v_{allg.-KBR} - v_{osteuropäisch}$; Intervall: 0,5 mm/Jahr; $\Delta v_{I-Durchschnitt} = -0{,}6$ mm/Jahr

Karte im 1 : 1 Million-Maßstab der rezenten vertikalen Erdkrustenbewegungen an die Reihe kommen. Die Karte wird folgendes enthalten:

— die Landesgrenzen,

— einige charakteristische Siedlungen,

— die wichtigeren tektonischen Linien (mit informativem Charakter),

— die charakteristischen Erdbebenzonen (die Epizentren der stärkeren Erdbeben, 6sten und noch höheren Stärkegrades nach der Mercali-Skala),

— die für den Ausgleich benutzten Mareographstationen,

— die Untersuchungslinien und Knotenpunkte,

— die Geschwindigkeitswerte der Knotenpunkte und Mareographen,

— die Isolinien, im allgemeinen mit dem Grundwert von 0,5 mm/Jahr.

Bisher ist die kartographische Grundlage der Karte bereits hergestellt worden. Die Teilnehmer haben an ihrer eigenen Landesfläche die Isolinien konstruiert und diese dem internationalen Koordinator zugesandt. Seitens Ungarns ist das zusammenfassende Konstruieren der Isolinien im Geodätischen Institut FÖMI bereits im Gange.

Die Klarlegung der Bemerkungen im Zusammenhang mit der kartographischen Grundlage hat bereits stattgefunden. Wiederholt haben wir uns schon mit den Anschlüssen der Isolinien entlang der Landesgrenzen beschäftigt.

Es tauchte auch ein Vorschlag auf: eine Übersichtskarte im 1 : 2 Millionen-Maßstab sollte auch hergestellt werden. (Diese sollte die gesamte Karpaten—Balkan-Region auf einem einzigen Kartenausschnitt darstellen.) Gegenwärtig ist die Redigierung der kartographischen Grundlage der Übersichtskarte schon

im Gange. Es scheint uns, zum vorgesehenen Termin wird es uns möglich sein, die Karten in beiden Maßstäben herzustellen.

Die farbigen Probedrucke der Karte der rezenten vertikalen Erdkrustenbewegungen der KBR werden bis Ende 1978 fertiggestellt. Diese werden wir den nationalen Verantwortlichen zusenden. Nach der Durchführung der nötigen Korrekturen wird die Vervielfältigung der Karte bis Herbst 1979 fertig.

Zusammenfassend darf festgestellt werden, daß im Rahmen des KAPG die Ergänzung des Untersuchungsmaterials der Karpaten—Balkan-Region und dessen erneuterter Ausgleich stattgefunden haben. Als Ergebnis des Ausgleichs hat sich die über die Krustenbewegung des Raumes früher ausgestaltete Meinung geändert. Der Vergleich der Ergebnisse der zweierlei Untersuchungen hat stattgefunden und die Redigierung der rezenten vertikalen Krustenbewegungskarte der KBR ist im Gange.

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ВНОС ГЕОДЕЗИИ В ИССЛЕДОВАНИЯ СОВРЕМЕННЫХ ВЕРТИКАЛЬНЫХ ДВИЖЕНИЙ ЗЕМНОЙ КОРЫ КАРПАТО-БАЛКАНСКОГО-РЕГИОНА (КБР)

И. ЙОО

Резюме

В рамках КАПГ была составлена и представлена в Москве в 1971 году карта современных вертикальных движений земной коры восточной Европы. На основании исследования результатов, критической оценки использованных при уравнивании данных было проведено более подробное исследование территории КБР. К исследованиям были привлечены и югославские и румынские геодезические данные, а также данные мареографов, расположенных на югославском побережье Адриатического моря. Уравнивание геодезических и океанографических данных было проведено в нескольких вариантах. В результате уравнивания во многих местах изменились скорости и направления движения земной коры. В настоящее время проводится составление карты современных вертикальных движений земной коры КБР на основании уравненных данных в м-бе 1 : 1000000. Издание карты планируется на второе полугодие 1979 года. На картографическом произведении, состоящем из 2-х листов, кроме привычных элементов и изолиний будут обозначены еще и наиболее важные тектонические и сейсмические элементы.

Это картографическое произведение будет показано в рамках доклада, и далее будет проведено сравнение данных новейших исследований Карпатского бассейна с данными карты восточноевропейских современных вертикальных движений земной коры.

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EINIGE PROBLEME DER BESTIMMUNG DER VERTIKALEN ERDKRUSTENBEWEGUNGEN MIT BESONDERER BERÜCKSICHTIGUNG DES REFRAKTIONSFEHLERS

Von

F. HALMOS

GEODÄTISCHES UND GEOPHYSIKALISCHES FORSCHUNGSINSTITUT DER UNGARISCHEN AKADEMIE DER WISSENSCHAFTEN, SOPRON

Es werden die durch Refraktion entstehenden Fehler der Erdkrustenbewegungs-Untersuchungen analysiert. Auf Grund der Arbeiten von Remmer wird die Bestimmung des Maßes der Refraktionswirkung erörtert. Es wird auf die Wichtigkeit der Errichtung von Gezeitenstationen in der Nähe von Mareographen hingewiesen.

Das einfache stochastische Modell der Nivellierungen ist

$$h = h' + x + y \qquad \begin{matrix} x \in N(O,\ \sigma x^2 L) \\ y \in N(O,\ \sigma y^2 L) \end{matrix} \tag{1}$$

wobei h' den gemessenen Wert des tatsächlichen Höhenunterschiedes h, x den zufällige Fehler und y den regelmäßigen Fehler bezeichnen. Zur Bestimmung des tatsächlichen Fehlers wurden zahlreiche Versuche durchgeführt. Wir beziehen uns auf Formel II. 38 der im Jahre 1975 veröffentlichten Arbeit von REMMER (S. 42) und wollen in erster Linie bezüglich der Bestimmung des Refraktionsfehlers einiges beitragen. REMMER gab für die von der Refraktion herrührenden regelmäßigen Fehler — ausgegangen von den Kukkamäkischen Untersuchungen, das Fermatsche Prinzip angewendet — folgende Formel an:

$$y_r = \frac{s^2}{6}\left[\frac{\mathrm{d}^2 n}{\mathrm{d}t^2}\left(\frac{\mathrm{d}t}{\mathrm{d}h}\right)^2 + \frac{\mathrm{d}n}{\mathrm{d}t}\left(\frac{\mathrm{d}^2 t}{\mathrm{d}h^2}\right)\right](h_1' - h_2') = \frac{s^2}{6}(h_1' - h_2') \cdot f \tag{2}$$

wobei s die zweifache Entfernung der Visierung (die Entfernung zwischen den Nivellierlatten), n der Refraktionsindex, $\mathrm{d}t$ der Temperaturunterschied, $\mathrm{d}h$ die Änderung der Höhe, $h_1' - h_2'$ der abgelesene Höhenunterschied (Differenz der Lattenablesung) sind.

Aus der Analyse von Präzisions-nivellierungen kann auch bestimmt werden, daß mit der Zunahme der Höhenunterschiede innerhalb der Nivellierlinie der mittlere Fehler der Bestimmung σy ebenfalls größer wird. Wären die in Gleichung (2) vorkommenden Glieder bekannt, so könnte die Wirkung der Refraktion bestimmt werden.

Man kann einfach vorgehen, wenn z.B. in einer Nivellierlinie oder in einem Polygon die Werte s und $\Delta h' = h'_1 - h'_2$ bekannt sind und auf Grund des gleichen Vorzeichens des Schlußfehlers, sowie auf Grund von gewissen statistischen Tests bestimmt wird, daß ein regelmäßiger Fehler vorhanden ist, und anhand einer Ausgleichung ein für real annehmbarer Refraktionskoeffizient f ermittelt wird. Ausgegangen von Gleichung (2) entsteht

$$\Delta h_i = \Delta h'_i + \frac{s_i^2}{b} (\Delta h'_i) f + v \tag{3}$$

worin v die dem zufälligen Fehler entsprechene Verbesserung bezeichnet. Man kann f als konstant annehmen, falls dn/dt und dt/dh = const. angenommen wird. So ist im Nivellierungspolygon

$$\sum_{i=1}^{m} \Delta h_i = \sum_{i=1}^{m} \Delta h'_i + \sum_{i=1}^{m} \frac{s_i^2}{b} \Delta h_i f + \sum_{i=1}^{m} v \tag{4}$$

$\Sigma \Delta h$ muß in einem geschlossenen Polygon 0 sein. Aus der Größe des Schlußfehlers, dessen Vorzeichen, sowie auf Grund der statistischen Tests und der im voraus berechenbaren Funktion $\sum \frac{s_i^2}{6} \Delta h'_i$ kann man folgern, ob ein regelmäßiger Refraktionsfehler dieser Art vorhanden ist oder nicht, bzw. man kann dessen Wert schätzen. Da die Werte $\Delta h'_i$ vornehmlich an reliefmäßig gegliedertem Gelände bedeutend sind (z. B. in der Karpaten—Balkan-Region), ist die eingehende Untersuchung dieser Frage unbedingt nötig. Auf Grund der Gleichung (4) ist bei $\Sigma \Delta h = 0$

$$f = - \frac{\sum_{j=1}^{n} \sum_{i=1}^{m} \Delta h'_{ij} + \sum_{j=1}^{n} \sum_{i=1}^{m} v}{\sum_{j=1}^{n} \sum_{i=1}^{m} \frac{s_{ij}^2}{6} \Delta h'_{ij}} \tag{5}$$

Werden verschiedene geschlossene Polygone vereint (z. B. jene von ähnlichen Geländeverhältnissen) und der Wert $\Sigma \Delta h'_{ij}$ (i-ter Abschnitt, j-te Linie), so ist der Schlußfehler und die Annahme $\Sigma v = 0$ zugelassen und der Koeffizient f ist modellierbar. Wird auf Grund des Zusammenhanges (5) das Vorhandensein des regelmäßigen Refraktionsfehlers y_r angenommen, so ist es nicht zweckmäßig, sämtliche Nivellierungswerte mit diesem Wert zu verbessern. An deren Stelle ist es ratsamer, in die Verbesserungsgleichung je einer Linie den unbekannten Koeffizienten f laut Gleichung (4) einzubauen. Hier kommt an Stelle von v_i eine einzige Verbesserung v und eine einzige Höhendifferenz Δh vor,

und der Koeffizient von f kann auf Grund der Geländeverhältnisse der Nivellierlinie gebildet werden:

$$\Delta h_i = \Delta h_i' + v_i + \Sigma \frac{s_i^2}{6} \Delta h_i' f \tag{6}$$

Der Zusammenhang kann in den Ausgleichung der zur Bestimmung der vertikalen Erdkrustenbewegungen dienenden wiederholten Nivellierungen eingegliedert werden. Bezüglich f kann man auch nach der Ausgleichung z.B. eine Hypothesenprüfung durchführen, wie dies auch in anderen Fällen geschieht.

Der geschätzte Wert von f ist $-0{,}05 - 0{,}1 \cdot 10^{-6}$ $1/m^2$. Die Wirkung auf den Schlußfehler kann im Falle einer 100 km langen geschlossenen Nivellierlinie, falls die Lattenhöhendifferenzen 1—2 m sind, 1—2 cm betragen, welcher bedeutende Wert unbedingt in Betracht gezogen werden muß.

Naturgemäß können andere der Funktionsformel (4) entsprechende regelmäßige Fehlerwerte (wie z. B. Kompensationsfehler, Komparierungsfehler, Fehler der Lattensenkung) einbezogen werden. Es ist zweckmäßig, diese mit Eichung bzw. sorgfältigem Messen zu minimalisieren.

Da man zu der Bestimmung vertikaler Erdkrustenbewegungen außer wiederholten Messungen und Messungen der Schwerebeschleunigung die zeitlichen Änderungen des, das Bezugsystem bildenden Schwerefeldes kennen muß, worüber z.Z. keine genügend genauen Daten zur Verfügung stehen, so können einstweilen absolute Geschwindigkeitswerte nicht bestimmt werden (Biró, 1977; Fajtelson—Jurkina, 1974; Totomanov, 1977). Dies bedeutet in der Zukunft eine enge theoretische und praktische Zusammenarbeit der Geodäsie, der Geophysik und der Geologie, und falls auch die beiden anderen horizontalen Komponenten in Betracht gezogen werden, so die der Astronomie. Einstweilen scheint die Zukunft der Entwicklung der Satellitengeodäsie gesichert zu sein, bei welcher die auf das Massenzentrum der Erde bezogenen präzisen räumlichen Koordinaten und deren absolute Änderungen abgeleitet werden können. Zur verläßlichen Bestimmung der Erdkrustenbewegungen muß auch die Kontinuität der Satellitenbeobachtungen gesichert werden. Es wäre zweckmäßig neben den Daten der präzisen Nivellierungen und der gravimetrischen Messungen auch die Gezeiten- und klinometrischen Beobachtungen bei einigen Hauptgrundpunkten und besonders in der Nähe von Mareographen in Betracht zu ziehen, wo die plattentektonischen Bewegungen mit den Änderungen der Neigungen beobachtet werden können unabhängig davon, ob sie durch Gezeiten oder Meeresströmungen verursacht worden sind.

Es ist leicht möglich, daß die Niveauänderung des Meeresspiegels eine gewisse Neigung der Erdkruste hervorruft. Außer den durch die Erde und den Mond verursachten Erdgezeiten entstehen durch Niveauänderungen des

Meeresspiegels (Ebbe und Flut) auch gewisse Krustenbewegungen. Da in diesem Falle oft Niveauänderungen von einigen Metern entstehen, können dadurch in der Kruste auch Neigungsänderungen vorkommen hervorgerufen werden.

Das hier kurz Geschilderte beweist bereits, daß es bei den Erdkrustenbewegungs-Untersuchungen zweckmäßig, ist die Ergebnisse anderer Beobachtungsverfahren in Betracht zu ziehen. Man muß auch der Entwicklung des Ausgleichungsmodells größere Sorgfalt widmen. Das Geodätische und Geophysikalische Forschungsinstitut der Ungarische Akademie der Wissenschaften sowie das Zentrallaboratorium für höhere Geodäsie der Bulgarischen Akademie der Wissenschaften planten und begannen gemeinsame Forschungen bezüglich der mathematisch-rechentechnischen Modelluntersuchung der Erdkrustenbewegungen um ein besseres und wirksameres Berechnungs- und Ausgleichsverfahren zu entwickeln.

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НЕКОТОРЫЕ ПРОБЛЕМ ОПРЕДЕЛЕНИЯ ВЕРТИКАЛЬНЫХ ДВИЖЕНИЙ ЗЕМНОЙ КОРЫ, С ОСОБЕННЫМ ВНИМАНИЕМ НА ОШИБКИ ИЗЗА ЭФФЕКТА РЕФРАКЦИИ

Ф. ХАЛМОШ

Резюме

Анализируются ошибки исследований движений земной коры возникшие вследствии рефракции. Исходя из работ РЕММЕРа излагается метод определения величины вляния рефракции. Подчеркивается особое значение установления сезональных станций вблизи мареографов.

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