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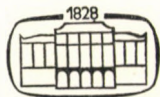
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ACADEMIAE SCIENTIARUM
HUNGARICAE

REDIGIT
E. SZÁDECZKY-KARDOSS

TOMUS XX

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AKADÉMIAI KIADÓ, BUDAPEST
1976

ACTA GEOL. HUNG.

ACTA GEOLOGICA

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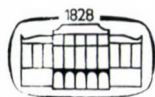
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CRYSTAL STRUCTURES AND GENETICAL STUDIES ON THE PALYGORSKITE-SEPIOLITE-SAPONITE (MONTMORILLONITE) GROUP*

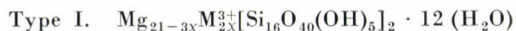
By

I. DÓDONY and J. KISS

MINERALOGICAL DEPARTMENT OF EÖTVÖS LORÁND UNIVERSITY, BUDAPEST

The present work deals with the study of the clay minerals-bearing hydrothermal veins of the Parádsasvár area (Mátra Mountains, North of Hungary). Two new crystal structures of clay mineral varieties have been revealed by X-ray powder diffraction and the electron microscopic investigations in a close agreement with the chemical analysis.

The two new clay mineral varieties have the following characteristics:



$$c_0 = 5.37 \text{ \AA}$$

$$b_0 = 68.74 \text{ \AA}$$

The symmetry of the (100) projection is p2gg possible space-groups:

$$\text{D}_{2h}^{16} - \text{Pnma} \quad \text{C}_{2v}^9 - \text{Pna2}_1$$

$$\text{D}_{2h}^9 - \text{Pham} \quad \text{C}_{2v}^8 - \text{Pba2}$$



$$c_0 = 10.56 \text{ \AA}$$

$$b_0 = 33.64 \text{ \AA}$$

The symmetry of the (100) projection is p2mg

Type I. (represents a) 2 : 1 palygorskite-sepiolite superstructure, in which the palygorskite-sepiolite and sepiolite-sepiolite chains show different geometry. The structure of Type II. consists of three-layer-chains of 16.82 Å width.

The presence of the chain-structures of different width and bond-type was confirmed by the infrared-absorption spectra.

As because of the lack of upper limit for the width of the three-layer chain, the superstructures provide with increase in the width of chains a gradual structural transition from palygorskite to saponite.

Thermal investigations have shown that in the course of crystallization of this group of clay minerals, an inverse relation between the energy-level of the environment and the b_0 values of the structure had existed. In this case the b_0 value of the saponite is considered to represent the largest width.

In hydrothermal veins and alteration zones, the sequence palygorskite — (palygorskite-sepiolite) — sepiolite — transitional members — saponite can be traced according the decreasing temperature gradient. In soils, the same sequence develops depthwards.

During ore-mineralogical studies in the Central-Mátra Mts in Hungary a white powder-like material was observed to fill the voids of ore-bearing veins

* A paper presented at a special meeting of the Mineralogy and Petrography Section of the Hungarian Geological Society on 1st December 1975 and at the 7th Conf. on Clay Min. and Petr., Karlovy Vary, 1976.

and occur as soft fibrous incrustations too. This material is abundantly found both in the fissures of the calcite-gangue of the veins and in the cavity altered country-rocks. Investigation of this material by various methods has given the following results.

1. Electron-microscopical Studies

The photomicrographs have been made on JEOL 100 U type electron-microscope, 80 kV accelerating voltage. In transmitted light the fibrous habit is recognizable. From the slightly curved outlines of the individual fibres, a gliding plane parallel to the elongation can be inferred (Fig. 1). Single-crystal electron-diffractograms were made from several crystals, leading to



Fig. 1. Transmission electron micrograph of the sample

the recognition of two coexisting crystal phases: Type I. (Fig. 2) and Type II. (Fig. 3). From the structural parameters, it was concluded that the micrographs were made in the b^* and c^* planes.

The two-dimensional space groups of the Type I. structure is $p2gg$, which excludes the possibility of any relationship among the space-groups

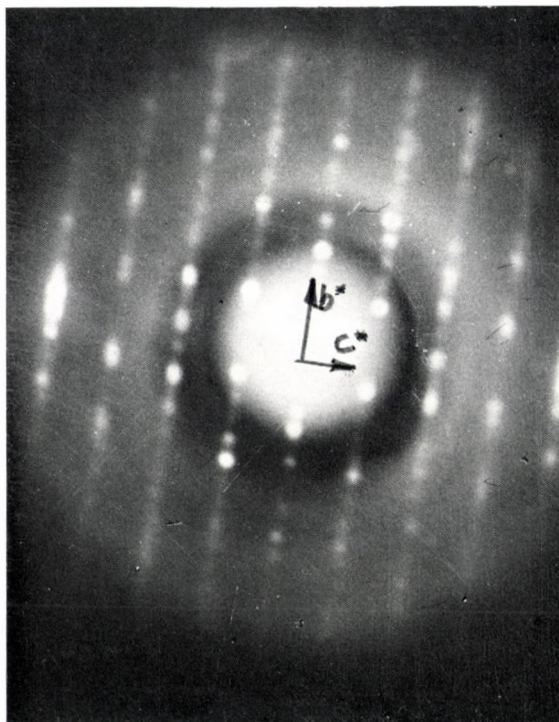


Fig. 2. Selected area electron diffraction pattern of Type I.

mentioned in earlier literature [1, 3, 12, 15, 17]. This type belongs either to $Pnma-Pna2_1$ or to $Pbam-Pba2$ space-groups. In these space groups, the markings of the axes are different or a better comparison, the traditional marking for the palygorskite-group is used.

The ED data of the first type are shown in Table 1. The intensities have been measured photometrically.

The two-dimensional space-group of Type II. is $P2mg$, the ED data of this variety are shown in Table 2.

The d_{hkl} values of the indexed reflections of Type I. and Type II. has been refined by a computer programme ODRA-1304 and so the following values have been obtained:

Type I. $c_0 = 5.37 \text{ \AA}$	Type II. $c_0 = 10.56 \text{ \AA}$
$b_0 = 68.74 \text{ \AA}$	$b_0 = 33.64 \text{ \AA}$

Table 1

d_{kl} (Å)	kl		d_{kl} (Å)	kl	
2.67	0	2	1.49	38	2
1.33	0	4	1.34	44	2
4.72	14	0	1.25	48	2
3.78	18	0	1.22	50	2
2.83	24	0	1.17	52	2
2.29	30	0	1.14	54	2
2.02	34	0	1.80	8	3
1.70	40	0	1.77	10	3
1.56	44	0	1.58	22	3
1.37	50	0	1.55	24	3
1.21	56	0	1.50	26	3
1.10	64	0	1.48	28	3
4.46	8	1	1.45	29	3
3.89	12	1	1.43	30	3
3.77	13	1	1.39	32	3
3.33	16	1	1.35	35	3
3.13	18	1	1.31	36	3
2.67	22	1	1.29	38	3
2.42	25	1	1.21	42	3
2.32	27	1	1.12	48	3
2.08	30	1	1.07	51	3
1.87	34	1	1.06	52	3
1.69	38	1	1.01	56	3
1.68	39	1	1.33	2	4
1.53	43	1	1.33	6	4
1.42	46	1	1.32	8	4
1.40	47	1	1.30	12	4
1.31	51	1	1.28	15	4
1.22	54	1	1.26	17	4
2.63	2	2	1.25	18	4
2.50	10	2	1.24	20	4
2.42	12	2	1.18	27	4
2.20	18	2	1.10	35	4
2.13	20	2	1.06	39	4
1.95	24	2	1.04	41	4
1.54	36	2			

Table 2

d_{kl} (Å)	kl		d_{kl} (Å)	kl	
3.62	0	3	2.04	5	5
2.7	0	4	1.99	6	5
2.24	0	5	1.71	12	5
1.83	0	6	1.78	2	6
1.53	0	7	1.74	4	6
1.34	0	8	1.7	6	6
1.18	0	9	1.57	8	6
1.06	0	10	1.56	10	6
0.95	0	11	1.44	14	6
0.86	0	12	1.28	18	6
5.6	6	0	1.23	20	6
4.31	8	0	1.52	3	7
3.4	10	0	1.51	4	7
4.47	4	2	1.49	5	7
3.77	6	2	1.48	6	7
3.33	8	2	1.46	7	7
2.51	12	2	1.41	9	7
2.21	14	2	1.4	10	7
1.93	16	2	1.33	2	8
3.54	1	3	1.32	4	8
3.43	3	3	1.3	6	8
2.77	8	3	1.27	8	8
2.63	9	3	1.24	10	8
2.44	10	3	1.22	11	8
2.25	12	3	1.21	12	8
2.1	13	3	1.2	13	8
2.01	14	3	1.16	14	8
2.64	2	4	1.19	1	9
2.55	4	4	1.182	3	9
2.33	7	4	1.178	4	9
2.23	8	4	1.1749	4	9
2.03	11	4	1.164	6	9
1.87	13	4	1.155	7	9
1.69	16	4	1.133	9	9
1.5	18	4			
1.42	20	4			
2.1	3	5			



Fig. 3. Selected area electron diffraction pattern of Type II.

2. Chemical Analyses

The chemical compositions of the analyzed samples are shown in Table 3. Comparing these data with the similar data of palygorskite, it is revealed that in the octahedral cation content there is 0.14 and 0.05 excess in the number of atoms as calculated for Mg^{2+} . Although a large error is presumably involved in our measurements but this shows an excess of cations as compared to palygorskite and a cation-deficiency as compared with sepiolite. The number of cations in octahedral position (as calculated for 8 tetrahedral cations per cell formula) is 0.33 and 1.0 more in the sepiolite and saponite respectively than in palygorskite.

Table 3

	a		b	
	W %	mol ratio	W %	mol ratio
SiO ₂	60.90	7.9992	61.9	7.993
TiO ₂	0.01	0.0008	0.07	0.007
Al ₂ O ₃	13.33	2.0629	13.09	1.9924
Fe ₂ O ₃	0.13	0.0126	0.41	0.0403
FeO	0.39	0.0426	0.47	0.0504
MnO	0.09	0.0103	0.1	0.0109
MgO	10.03	1.9635	9.99	1.9226
CaO	0.71	0.1002	—	—
Na ₂ O	0.04	0.0095	0.06	0.0155
K ₂ O	0.04	0.0063	0.06	0.0093
H ₂ O ⁻	—	—	—	—
H ₂ O ⁺	14.31	6.2685	12.12	5.22
P ₂ O ₅	0.006	—	0.01	0.0008
	99.986		99.88	

3. Infra-red Absorption Studies

The maxima of infra-red absorption measured in the sample after drying at 105 °C are summarized in Table 4. Following WILKINS [16] the presence of oxonium-ion in the sample must be possessed:

$$\begin{array}{llll}
 (\text{H}_3\text{O})^+ & \nu_1 (\text{A}_1) & 3300 \text{ cm}^{-1} & 3400 \text{ cm}^{-1} \text{ N}^\circ 28.29 \\
 & \nu_2 (\text{A}_1) & 1190 \text{ cm}^{-1} & \text{N}^\circ 21 \\
 & \nu_3 (\text{E}) & 3300 \text{ cm}^{-1} & 3400 \text{ cm}^{-1} \text{ N}^\circ 28.29 \\
 & \nu_4 (\text{E}) & 1740 \text{ cm}^{-1} & \text{N}^\circ 26
 \end{array}$$

These peaks are found in every measurement, although with different intensities. The peaks No 29—32 indicate strong bonding forces between the zeolitic-water and the surficial oxygen-atoms of the channels parallel to the *c*-axis. These peaks indicate at least three different positions of water characterized by different energy levels.

4. X-ray Diffraction Measurements

A Siemens Kristalloflex-4 type diffractometer has been used. Table 5. shows the (*hkl*) values in Columns 3 and 4 which have been computed from the refined parameters of the ED. The refinement of $a_0 = 12.75 \text{ \AA}$ value

Table 4

No	cm ⁻¹	Transmittance	No	cm ⁻¹	Transmittance
1.	415	31	17.	1000	1
2.	447	11	18.	1050	1
3.	490	4.5	19.	1095	3
4.	520	3	20.	1125	4.5
5.	570	22	21.	1190	25
6.	580	22.5	22.	1440	49
7.	650	34	23.	1660	39
8.	690	44	24.	1700	64
9.	705	46.5	25.	1720	73
10.	712	47	26.	1740	75
11.	735	49	27.	2365	83
12.	770	49	28.	3300	35
13.	795	50.5	29.	3400	25
14.	877	49.5	30.	3550	10
15.	912	28	31.	3620	22
16.	945	35	32.	3750	85

has been done by using literature data (2, 11). Cell parameters from the generated (hkl) and the relating d_{hkl} values are

$$\begin{array}{ll}
 \text{I. type } a_0 = 12.74(7) \text{ \AA} & \text{II. type } a_0 = 12.75(2) \text{ \AA} \\
 b_0 = 68.66(6) \text{ ,,} & b_0 = 33.60(4) \text{ ,,} \\
 c_0 = 5.36(9) \text{ ,,} & c_0 = 10.56(2) \text{ ,,} \\
 \alpha = \beta = \gamma = 90^\circ & 2 = \beta = \gamma = 90^\circ \\
 V = 4699.68 \pm 1.2 \text{ \AA}^3 & V = 4526.02 \pm 1.3 \text{ \AA}^3
 \end{array}$$

The set of d_{hkl} values corresponding to the generated (hkl) values calculated from the literature data (2, 11.) concerning the structural parameters of palygorskite is summarized in Columns 5. After the refinement of the data, the following lattice constants have been obtained:

$$\begin{array}{l}
 \text{orthorhombic } P \text{ cell } a_0 = 12.74(4) \text{ \AA}; b_0 = 17.95(2) \text{ \AA}; \\
 c_0 = 5.22(4) \text{ \AA}.
 \end{array}$$

The reflections which have given Column 5 are artificial, as different reflections are indicated by similar indices. The indices in Column 3 and 4 are not suitable for defining the space group, because the input of a given cell-type in the computer programme relates the measured d_{hkl} values to the nearest possible (hkl)-s. However, these calculations support the accuracy of lattice constants calculated from ED measurements.

Table 5

No	1.	2.	3.	4.	5.	No	1.	2.	3.	4.	5.
1.	15.781	2	040	020	—	25.	5.4081	1	—	211	130
2.	13.940	4	050	—	—	26.	5.3820	12	011	—	130
3.	13.056	3	—	100	—	27.	5.3021	1	021	002	001
4.	12.810	1	100	—	—	28.	5.2491	1	031	—	001
5.	12.592	1	110	—	—	29.	5.2093	1	1 12 0	012	001
6.	12.076	2	120	—	—	30.	5.1731	1	041	151	220
7.	11.834	1	—	110	—	31.	5.1198	1	280	160	220
8.	10.822	1	130	—	—	32.	4.5776	1	1 14 0	241	021
9.	10.526	5	—	001	—	33.	4.4681	17	2 11 0	—	040
10.	10.258	100	140	120	110	34.	4.4460	1	—	042	040
11.	10.025	5	—	011	110	35.	4.3766	1	091	071	230
12.	9.2898	1	150	—	020	36.	4.3343	1	1 15 0	—	230
13.	8.8362	2	—	021	020	37.	4.2765	12	181	300	121
14.	8.2525	1	—	101	—	38.	4.1966	1	330	080	140
15.	7.6480	2	090	031	120	39.	4.1386	12	340	171	310
16.	7.2723	1	—	121	120	40.	4.0730	1	0 11 1	202	201
17.	6.7632	2	—	050	—	41.	4.0275	2	1 10 1	212	201
18.	6.4075	2	200	200	200	42.	3.9917	4	241	180	201
19.	6.3391	11	210	—	200	43.	3.9480	2	251	152	211
20.	6.2065	1	—	210	200	44.	3.9222	1	0 12 1	311	031
21.	6.0918	1	1 10 0	—	210	45.	3.9052	1	370	081	031
22.	5.7276	2	0 12 0	051	—	46.	3.8884	1	2 14 0	—	031
23.	5.5667	1	260	230	—	47.	3.8273	1	0 18 0	232	320
24.	5.4577	1	—	201	130	48.	3.7603	1	0 13 1	181	131

Table 5 continued

No	1.	2.	3.	4.	5.	No	1.	2.	3.	4.	5.
49.	3.7308	1	2 15 0	162	131	73.	2.6921	1	3 15 1	422	411
50.	3.6896	1	281	242	221	74.	2.6842	2	012	461	251
51.	3.6597	9	1 18 0	—	050	75.	2.6420	2	1 22 1	004	341
52.	3.6086	1	1 13 1	271	050	76.	2.6337	2	3 16 1	014	—
53.	3.5519	1	2 16 0	072	050	77.	2.6210	3	122	—	002
54.	3.4836	1	1 14 1	252	330	78.	2.6077	5	132	024	002
55.	3.4570	1	—	023	150	79.	2.5909	6	072	2 10 2	440
56.	3.4308	2	2 11 1	172	150	80.	2.5807	2	152	343	012
57.	3.3759	1	—	113	041	81.	2.5714	1	3 17 1	034	070
58.	3.3399	12	2 12 1	281	231	82.	2.5601	2	2 21 1	093	102
59.	3.2238	2	0 17 1	290	311	83.	2.5460	10	4 10 1	510	500
60.	3.1910	1	400	400	400	84.	2.5259	3	2 25 0	452	510
61.	3.1743	7	420	410	400	85.	2.5170	7	3 22 0	044	510
62.	3.1168	54	3 15 0	053	250	86.	2.4846	5	1 9 2	283	431
63.	3.1061	7	381	1 10 1	321	87.	2.4682	2	5 7 0	481	—
64.	3.0956	5	450	272	321	88.	2.4487	1	2 4 2	521	520
65.	3.0480	1	391	371	—	89.	2.4282	3	—	1 12 2	351
66.	2.9717	2	2 16 1	233	051	90.	2.3612	2	5 11 0	373	132
67.	2.9602	2	—	192	051	91.	2.3330	2	1 26 1	2 13 1	222
68.	2.9364	1	3 11 1	0 11 1	—	92.	2.3105	3	4 16 1	433	071
69.	2.9233	3	1 19 1	282	160	93.	2.2626	5	5 14 0	4 10 1	171
70.	2.9029	5	1 23 0	163	160	94.	2.2357	2	0 17 2	084	232
71.	2.7483	1	411	292	350	95.	2.2157	7	2 14 2	542	361
72.	2.7119	1	2 19 1	303	260	96.	2.1849	2	4 19 1	0 14 2	460

Table 5 continued

No	1.	2.	3.	4.	5.	No	1.	2.	3.	4.	5.
97.	2.1664	4	0 29 1	344	451	124.	1.7151	2	422	574	113
98.	2.1560	1	5 17 0	094	322	125.	1.7077	3	5 27 1	1 17 3	571
99.	2.1279	4	4 24 0	354	242	126.	1.6822	1	792	2 19 1	1 10 1
100.	2.1241	3	3 28 0	562	600	127.	1.6765	3	0 32 2	2 11 5	203
101.	2.1160	4	1 32 0	2 15 0	280	128.	1.6709	2	4 35 0	780	033
102.	2.1020	1	4 21 1	0 16 0	610	129.	1.6623	1	7 11 1	5 12 3	—
103.	2.0255	3	—	650	371	130.	1.6364	2	0 17 3	693	670
104.	1.9709	2	6 13 0	602	190	131.	1.5849	1	3 12 3	3 15 4	810
105.	1.9677	2	631	612	062	132.	1.5721	1	1 35 2	5 11 4	243
106.	1.9633	2	641	2 15 2	601	133.	1.5633	3	3 14 3	6 11 3	4 10 0
107.	1.9601	2	—	2 16 1	281	134.	1.5609	3	0 44 0	831	671
108.	1.9521	1	3 18 2	0 16 2	611	135.	1.5569	2	433	—	562
109.	1.9438	1	4 28 0	670	162	136.	1.5533	2	6 16 2	2 20 2	760
110.	1.9255	2	2 34 0	5 11 1	551	137.	1.5420	1	1 12 3	6 15 0	680
111.	1.9194	3	0 25 2	642	621	138.	1.5334	1	6 31 0	860	—
112.	1.9160	3	3 32 0	4 14 0	640	139.	1.5232	1	831	6 14 2	3 11 0
113.	1.8942	1	4 12 2	2 14 3	352	140.	1.5209	1	841	1 21 2	253
114.	1.8901	1	0 34 1	0 15 3	352	141.	1.5097	3	6 32 0	783	—
115.	1.8708	1	1 34 1	3 13 3	631	142.	1.5075	3	702	017	423
116.	1.8163	2	572	5 13 0	561	143.	1.5042	3	881	—	063
117.	1.8112	3	0 28 2	720	172	144.	1.4926	1	2 23 3	625	702
118.	1.8029	4	3 32 1	4 14 2	390	145.	1.4914	1	8 10 1	5 18 1	702
119.	1.7962	2	592	633	0100	146.	1.4886	2	1 44 1	852	761
120.	1.7693	1	123	554	—	147.	1.4495	1	7 13 2	6 14 3	732
121.	1.7652	1	3 31 1	2 18 1	—	148.	1.4287	2	8 17 1	3 20 3	1 12 1
122.	1.7629	1	143	—	452	149.	1.4210	2	0 41 2	317	523
123.	1.7325	1	0 10 3	760	013	150.	1.4171	3	900	900	742

1.: d_{hkl} ; 2.: I/I_0 ; 3.: generated (hkl) from Type I; 4.: generated (hkl) from Type II; 5.: generated (hkl) from literature cell parameters.

5. Thermal Analysis

Untreated samples (Fig. 4) as well as HCl-treated samples (Fig. 5) were investigated by thermal analyses. The data of these measurements are summarized in Table 6.

Table 6

No	Transformation, T °C		Nature of process		Weight loss	
	untreated	HCl-treated	untreated	HCl-treated	untreated	HCl-treated
1.	150	150	endotherm	endotherm	10.60	10.82
2.	200	180	endotherm	endotherm	10.60	10.82
3.	260	260	endotherm	endotherm	2	2.57
4.	360	350	exotherm	exotherm	9.7	6.3
5.	420	400	exotherm	exotherm	9.7	6.3
6.	475	455	exotherm	exotherm	1.43	1.14
7.	520	480	endotherm	endotherm	0.86	0.86
8.	550	520	endotherm	endotherm	2	0.86
9.	655	550	endotherm	endotherm	2.57	1.14
10.						
11.	655—960	570—890	endotherm	endotherm	1.72	2.85
12.	—	920	—	exotherm	—	—

6. Discussion

Diffraction data exclude the relations of this material to palygorskite or sepiolite structures. The theory of random interlayering [18] does not give an adequate explanation (see Figs 2 and 3). Applying the results for ROTUREAU and TCHOUBAR [12] the following relationships can be drawn for the three-layer-chains (Figs 6 and 7). Type I shows 1 : 2 superstructure of palygorskite and sepiolite in which the bonds within the palygorskite-sepiolite and sepiolite-sepiolite chains are geometrically different. The application of distribution between the $(OH)_1$, $(OH)_2$ and $4C_1$ and $4C_c$ positions — is not possible because of the symmetry. The formula of Type I. is $Mg_{21-3x}M_{2x}^{3+}[Si_{16}O_{40}(OH_5)] \cdot 12(H_2O)$; $M^{3+} = Al^{3+}(Fe^{3+})$, supported by chemical analyses. The occupation of $4C_2$ positions [19] reduces the number of octahedral position in the models of both BRADLEY [1] and BRAUNER—PREISINGER.

Type II. cannot be related to pal-sep 2 : 1 superstructure, because of the $P2mg$ symmetry. The structure of Type II. consists of three-layer-chains of 16.82 Å width. The occupation of $4C_2$ positions, which should make the number of M^{2+} atoms equal to the ideal palygorskite, has not been considered.

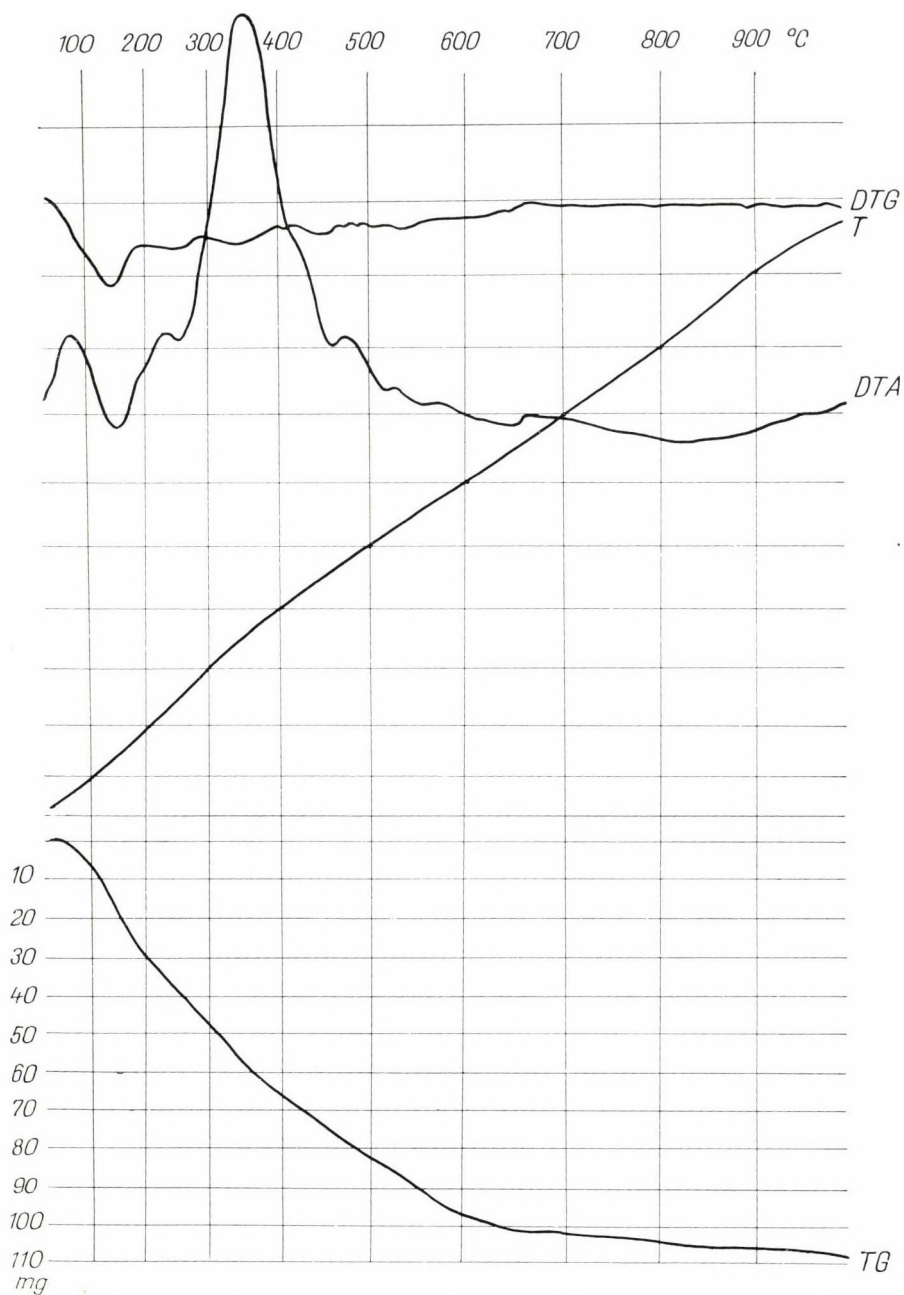


Fig. 4. DTG, DTA, TG, T, curves. Untreated sample. Sensibilities: DTA: 1/10; DTG: 1/15
TG: 100 mg. Quantity: 350 mg

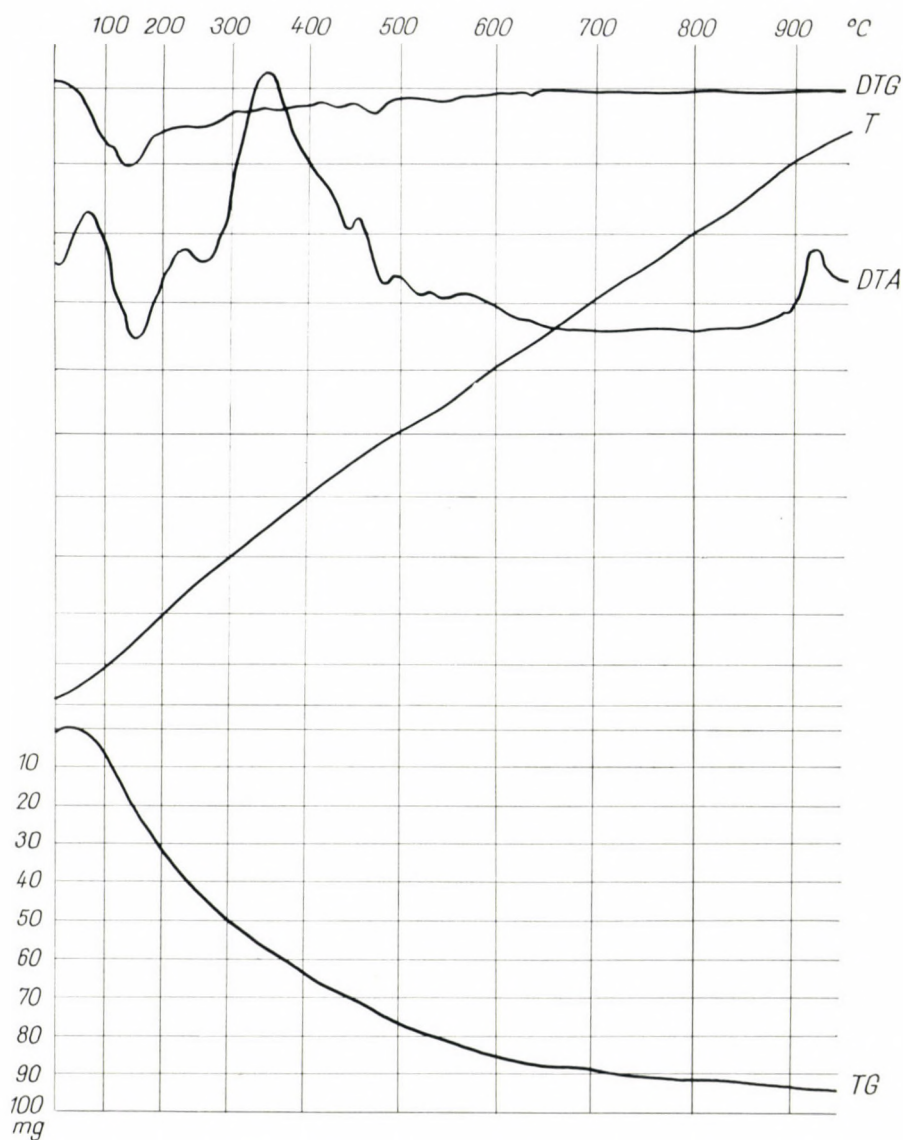


Fig. 5. DTG, DTA, TG, T, curves. Samples treated by HCl. Sensibilities: see on Fig. 4.

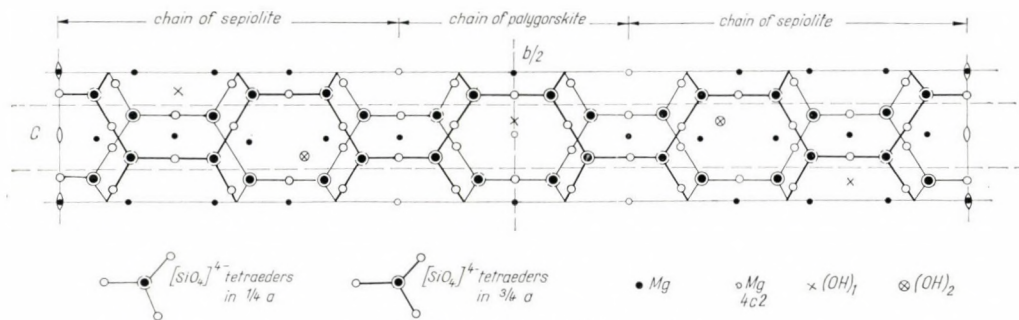


Fig. 6. Structure of Type I

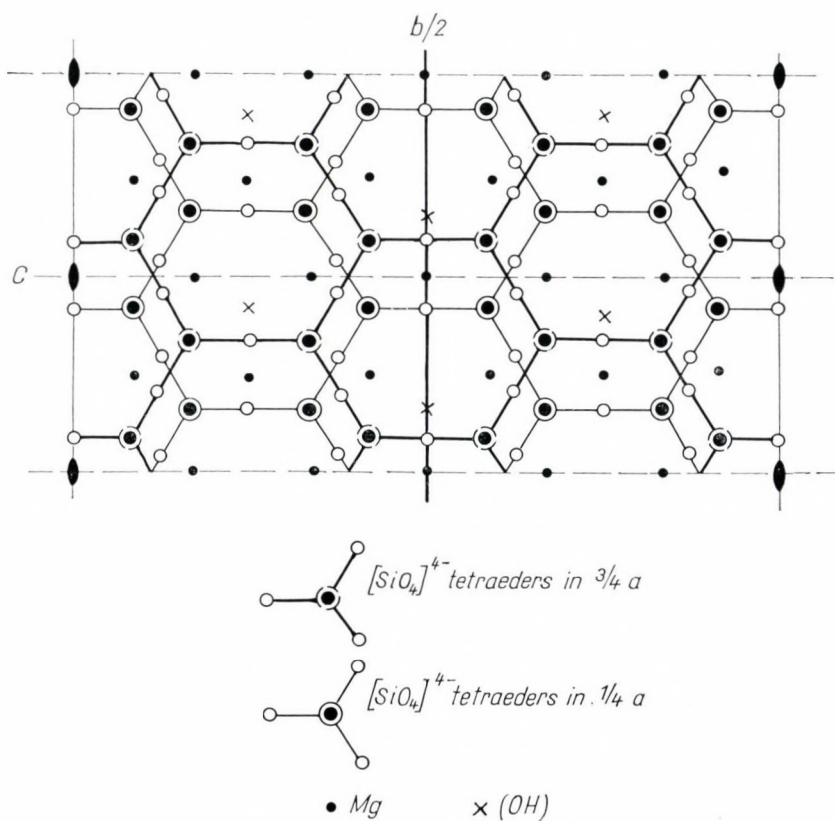


Fig. 7. Structure of Type II

The two varieties are apparently members of the genetical series of palygorskite-sepiolite-saponite (montmorillonite), from both chemical and structural aspects. There is no upper limit for the width of a three-layer-chain. A possibility exists for gradual transition between the two final members. The "interlayer" space is similar to that of saponite, but these minerals do not exhibit any swelling because of chessboard pattern of chain linkage. The potential swelling capacity in the zone of the "c"-axis produces a strain which prevents crystal growth in a direction normal to the c-axis, and this produces the fibre-like habit of the mineral.

The structural "channels" of the different structural types have constant size-distribution and chemically active surfaces, therefore they can be used as catalyzers or selective absorbents. N° 4—5—6 peaks in Table VI can be interpreted by oxidation or by polymerisation. The small quantity of $(\text{H}_3\text{O})^+$ is the only oxidable component. Since the intensities of these peaks have decreased after a treatment with diluted HCl, oxidation cannot be responsible for these peaks.

The three chain-bonds described in Type I. and Type II. can be polymerized to produce a chain-structure of larger width by shifting with $1/2$ "a" distance after the interlayer water partially releases. Exothermal alteration shows three maxima, which corresponds with the minimum number of the chain types and the different H_2O resonances processes (Table 6). Smaller inner energies of saponites and related minerals compared to palygorskite are indicated by N° 4—5—6 peaks in Table 6.

Structures of smaller chain-width have crystallized in a higher energy-level system, according to the descriptions of several mineral paragenesis [1, 2, 6, 14]. In hydrothermal zones the crystallization sequence of palygorskite-saponite follows the decrease of temperatures. In soils the thermodynamic energy level is a function of temperature and the change in the concentration of the solutions. In soil formation of desert zones such a development of clay minerals can be widely observed [2, 14]. In soil-formation the sequence downwards from the ground surface is palygorskite-saponite-(montmorillonite) [14].

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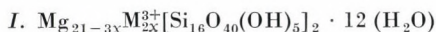
КРИСТАЛЛОСТРУКТУРНАЯ И ГЕНЕТИЧЕСКАЯ СВЯЗЬ РЯДА ПАЛИГОРСКИТ-СЕПИОЛИТ-САПОНИТ (МОНТМОРИЛЛОНИТ)

И. ДОДОНЬ—Я. КИШШ

Резюме

Были исследованы глинистые минералы гидротермальных жил у пос. Парадшавар (горы Матра).

Рентгеновские и электрондифракционные анализы до сих пор давали неизвестную кристаллическую структуру. При помощи электрондифракционного анализа монокристалла можно изучать проекцию структур (100). Применяя также данные химических анализов, два новых минерала можно охарактеризовать следующими формулами и кристаллографическими константами:



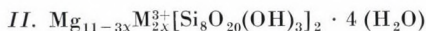
$$c_0 = 5.37 \text{ \AA}$$

$$b_0 = 68.74 \text{ \AA}$$

симметрия проекции (100): $p2gg$

Возможные пространственные группы:

$$\begin{array}{ll} D_{2h}^{16} - Pnma & C_{2v}^9 - Pna2_1 \\ D_{2h}^{24} - Pbam & C_{2v}^8 - Pba2 \end{array}$$



$$c_0 = 10.56 \text{ \AA}$$

$$b_0 = 33.64 \text{ \AA}$$

симметрия проекции (100) = $p2mg$

Пропорция 2 : 1 типа I. представляет собою сверхструктуру палигорскита и сепиолита, в которой цепные связи палигорскит-сепиолит и сепиолит-сепиолит геометрически разные.

II. структурный тип состоит из связей «широких» тройной слоистости лент 16,82 Å.

Для выделения различной ширины и связей цепей также кажется пригодным инфракрасный абсорбционный спектр.

По геометрии кристаллических решеток для отдельных тройной слоистости цепей верхнего предела не существует. Возрастание ширины цепей и сверхрешетки обеспечивают непрерывный переход от палигорскита до сапонита.

На основании данных термических исследований также можно установить, что при кристаллизации глинистых минералов, относящихся к этому ряду, энергетический уровень окружающей среды состоит в обратной пропорции с b_0 («шириною цепей»). В этом случае b_0 сапонита мы рассматриваем наибольшей величиной.

В зонах и жилах, претерпевших гидротермальное преобразование, в соответствии с градиентом снижения температур друг друга сменяют палигорскит — (палигорскит-сепиолит) — сепиолит — переходные члены — сапонит. В почвах данные минералы проявляются в этой очередности, считая от поверхности.

Доклад был зачитан на заседании Минералого-геохимического отделения Венгерского геологического общества 1 декабря 1975 г. и в Карловых Варах в 1976 г. на конференции «7th Conf. on Clay Min. and Petr.».

CONVERSION OF ILMENITE IN TO PEROVSKITE

By

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This research represents a study of the conditions of perovskite formation from ilmenite using a derivatograph. The experimental work was carried out with Abu Ghalqa ilmenite. The resulting phases of the reaction of ilmenite with calcite were studied by means of microscope and a Siemens Crystalloflex diffractometer. Complete conversion of ilmenite to perovskite was found to take place at 950 °C.

Introduction

In Egypt ilmenite is mainly found in two different modes of occurrences, one is the placer deposits along the Mediterranean coast, particularly between Rosetta and Damietta and the other is the magmatic ore body at Abu Ghalqa, Eastern Desert [2, 10, 11, 15]. The Abu Ghalqa ilmenite occurs in form of large lenticular bands and layers cutting the titaniferous gabbro.

The technological methods of processing of ilmenite include acid digestion, chlorination in the presence of coal, selective reduction with coal or hydrogen, smelting, fusion with sodium sulphate or caustic soda and alkaline hydrometallurgical digestion [5, 6, 17, 22].

Perovskite is one of the components of titanium slags, obtained from the smelting of ilmenite. Smelting is used for the enrichment of titanium in low grade titaniferous iron ores. Smelting of ilmenite with coke is carried out at 1100—1150 °C in the presence of fluxes. Fluxes used are lime or limestone, magnesium oxide or magnesite and soda. Smelting results in the production of cast iron and titanium rich slags (containing 70—80% TiO₂). At such smelting temperature, titanium dioxide within the ilmenite structure is not reduced. The addition of fluxes, e.g. limestone (in amount of 4—5%), facilitates the reduction of iron oxides by breakdown of the iron titanate molecule of ilmenite and its reaction with titanium dioxide and the perovskite formation [6, 17, 22].

Perovskite occurs as an accessory mineral in basic and alkaline igneous rocks. It also occurs in some contact-metamorphosed impure limestone [9].

It was previously reported that perovskite is formed at 970 °C from sintering of rutile with calcite [1]. The standard free energy (ΔF°) and the equilibrium constant of the reaction of perovskite formation (K) were esti-

mated to be -19.606 kcal/mol and 2.35×10^{14} . The equilibrium constant is very high and the reaction of perovskite formation may thus be practically considered to be irreversible. At $200-900^\circ\text{C}$, perovskite lattice remains orthorhombic, but it will be of higher symmetry. It fuses congruently at 1915°C [8, 14, 21].

Ilmenite melts at 1365°C . Rich in ferric oxide, ilmenite containing products of decomposition of solid solution, becomes homogeneous by heating in vacuum up to 1200°C . Ilmenite breaks down by heating in air to pseudobrookite and rutile or to hematite, rutile, pseudobrookite etc. [7, 12, 13, 18].

Less is known on the perovskite formation from ilmenite in titanium slags, obtained from smelting of ilmenite and also on its occurrence in contact—metamorphosed limestone. This work follows the conditions of perovskite formation from Abu Ghalqa ilmenite using a derivatograph. The resulting phases of sintering of ilmenite with calcite were identified both by microscope and by X-ray diffraction analysis.

Experimental Work

Investigations were carried out with Abu Ghalqa ilmenite, Eastern Desert.

Mineralogy of Ilmenite

The microscopic study of polished sections of Abu Ghalqa ilmenite shows that the ore is composed mostly of granular mosaic-like ilmenite grains, constituting 75% of the ore with silicate minerals. The gangue is composed of plagioclase feldspar, mostly labradorite, hornblende hypersthene together with their alteration products.

The ilmenite crystals often enclose minute exsolved bodies of hematite. Magnetite may also be present but in smaller amount than hematite. The hematite exsolution bodies are of different shapes, the most common of which are lamellae, blebs or irregular bodies. The hematite lamellae are of serial arrangement parallel to the (0001) direction of ilmenite.

Although ilmenite can take up about 6% Fe_2O_3 (Abu Ghalqa ilmenite contains 5.55% Fe_2O_3) into solid solution, exsolution lamellae of hematite or of titanohematite may also be present [3, 4, 9].

These ilmenite-hematite exsolution intergrowths may represent the progressive unmixing of mutual solid solutions, namely ferriiferous ilmenite and titaniferous hematite [3, 4, 5].

Chemical Composition and X-ray Diffraction Data of Ilmenite

The ilmenite in question was concentrated by gravity separation and by using isodynamic magnetic separator. The obtained ilmenite concentrate was chemically analyzed for major constituents and some trace elements and their data are given in Tables 1 and 2, respectively. The ferric oxide content shown by this analysis is probably in form of hematite.

The X-ray diffraction data of ilmenite given in Table 3 which are consistent with the ASTM standards. None of the contaminants was detected by X-ray analysis, therefore, no individual mineral may be present as major constituent.

Table 1*Chemical composition of ilmenite*

Chemical component	Content, %
SiO ₂	traces
TiO ₂	50.66
Al ₂ O ₃	traces
Fe ₂ O ₃	5.55
FeO	37.86
MnO	2.04
MgO	0.41
CaO	traces
Na ₂ O	0.17
K ₂ O	0.16
+H ₂ O	2.28
-H ₂ O	0.10

Table 2*Trace elements in ilmenite*

Element	Content, ppm
Nb	< 30
Ta	< 100
Sn	350
Mo	< 4
W	100
Ni	400
Co	200
Cr	500
V	1500

Table 3*X-ray diffraction data of Abu Ghalqa ilmenite*

d (Å) ASTM	d (Å) Observed	I ASTM	I Observed	h	k	l
3.73	3.741	50	35	1	0	2
2.75	2.753	100	100	1	0	4
2.55	2.550	70	70	1	1	0
2.23	2.244	70	35	1	1	3
1.87	1.872	50	30	2	0	4
1.73	1.731	80	70	1	1	6
1.63	1.635	50	30	1	0	8
1.50	1.508	85	60	2	1	4
1.47	1.471	85	70	3	0	0

Techniques of Work

Starting material consisted of ilmenite mixtures. Reagents used in mixtures are calcite and graphite in particular amounts. Mixtures were processed by repeated grinding in an automated agate mortar and sieving till all the powder passed through 200 mesh sieve and pestle for one hour in order to reach homogeneous material.

Apparatus

Sintering of ilmenite mixtures was carried out in ceramic pots heated in an electric furnace and removed the evolved gases. The character of the reaction of sintering of ilmenite with calcite was studied by thermal analysis by means of the derivatograph designed by PAULIK, F., PAULIK, J. and ERDEY, L. [19]. The apparatus records simultaneously four thermal curves: (T) the change of temperature of the sample (DTA), differential thermal analysis (TG), thermogravimetric (quantitatively in mg) and (DTG) derivative thermogravimetric, on a single sample under controlled conditions. DTA and temperature measuring thermocouples are Pt/Pt—Rh wires. Ceramic pots and a ceramic sample holder were used. Aluminium oxide, calcinated at 1000 °C was used as reference material. The parameters during tests were as follows: sample weight 1 g, T — 1200 °C, DTA — 1/10, DTG — 1/10, TG — 500 mg, and heating rate 10 °C per minute. All runs were carried out in air atmosphere and removing the evolved gases.

Phase Identification

The products of sintering of ilmenite with calcite were identified both by microscope and by X-ray diffraction analysis.

X-ray Procedure

A Siemens Crystalloflex diffractometer was used with nickel filtered copper radiation. Exposure was one hour and scanning speed was 1° 20 per minute, at 1 cm per minute chart speed. Intensities were collected to maximum 20 = 65°. The sensitivity of experiment was 1×10^4 impl/min, and the statistical error proved to be 1.5%.

Results and Discussion

The derivatogram of ilmenite mixed with calcite in 1 : 1 ratio is shown in Fig. 1. The DTA curve is characterized by a sharp and large endothermal peak at 890 °C, representing the intensive dissociation of calcite and the reaction between ilmenite and the resulted calcium oxide. This process is connected with a remarkable decrease of weight (TG) due to the release of carbon dioxide.

The X-ray diffraction pattern of the end product of sintering (Fig. 2) shows the formation of perovskite and calcium ferrite.

The derivatogram of ilmenite mixed with calcite and graphite in 1 : 1 : 0.1 ratio is shown in Fig. 3. The DTA curve was evaluated on the basis of literature data [1, 7, 12—14, 16, 21], trying to explain the reactions connected with certain peaks of DTA, and comparing them with that of calcite and other components in the mixture. The small endothermal peak at 100 °C represents the dehydration of the mixture. The exothermal peak at 560 °C may be due to the burning of graphite and the combustion of volatiles. This is accompanied by little decrease in weight (TG). The large endothermal peak at 860 °C represents the intense dissociation of calcite and probably the reaction between

the resulted calcium oxide and ilmenite. This is connected with a sharp decrease in weight (TG) due to the release of carbon dioxide. The presence of graphite leads to the decrease of the temperature of calcite decomposition. The beginning of dissociation of calcite follows directly the end of exothermal reaction. The sharp exothermal peak at 950 °C may be due to the intense

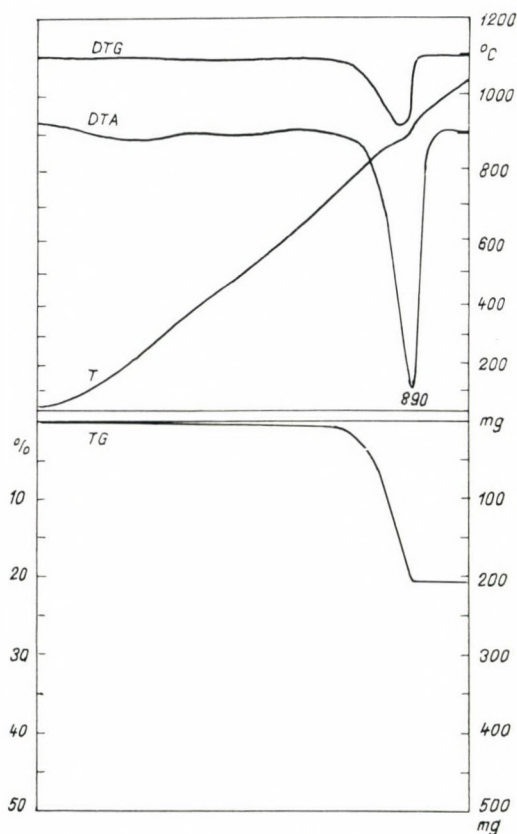


Fig. 1. Derivatogram of ilmenite sintering with calcite in 1 : 1 ratio. Weight of sample 1000 mg. Heating rate 10 °C/min

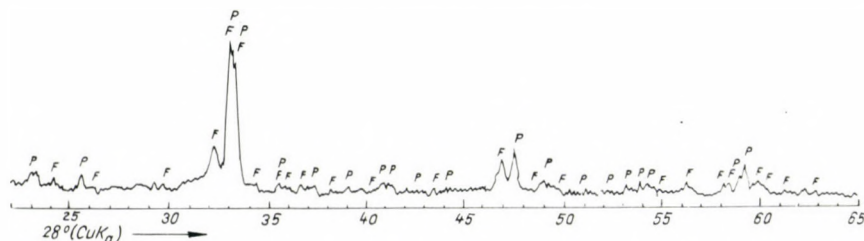


Fig. 2. X-ray diffraction pattern of product of ilmenite sintering with calcite. P: Perovskite; F: Calcium ferrite

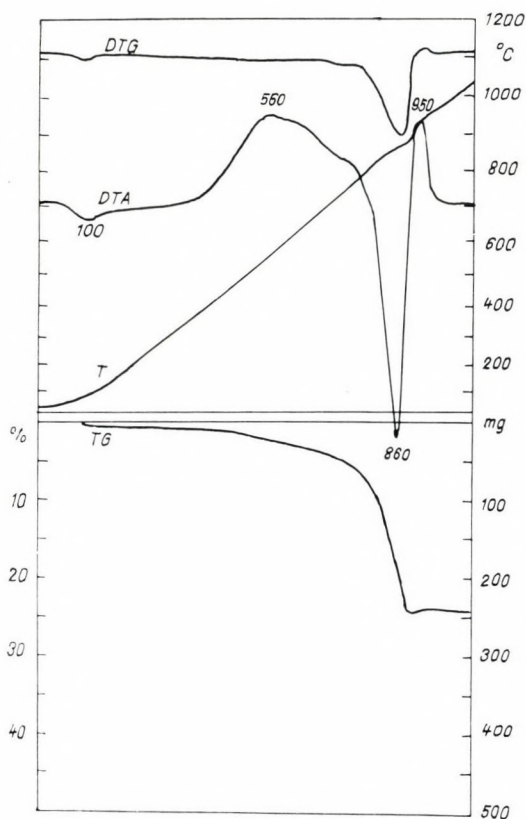


Fig. 3. Derivatogram of ilmenite sintering with calcite in presence of graphite 1 : 1 : 0.1 ratio. Weight of sample 1000 mg. Heating rate 10°/min

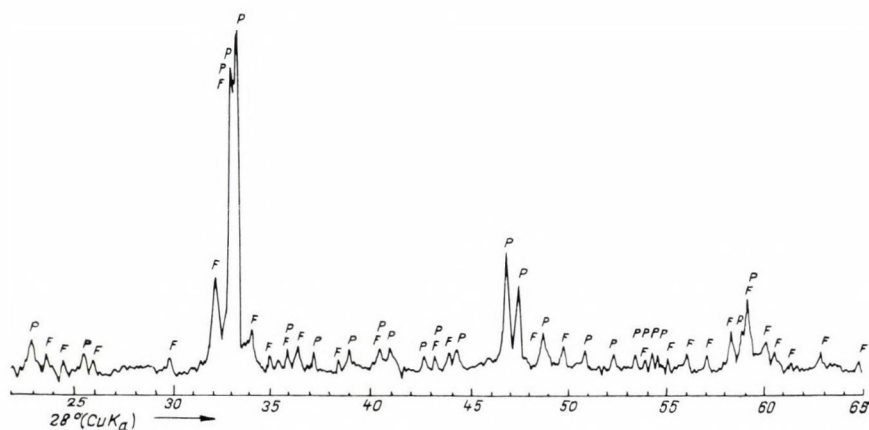


Fig. 4. X-ray diffraction pattern of the product of ilmenite sintering with calcite in presence of graphite at 950 °C. P: Perovskite; F: Calcium ferrite

Table 4
X-ray diffraction data of perovskite

d (Å) ASTM	d (Å) Observed	I ASTM	I Observed	h	k	l
3.824	3.833	14	12	1	0	1, 0 2 0
3.423	3.431	3	3	1	1	1
2.719	2.716	40	42	2	0	0
2.701	2.705	100	100	1	2	1, 0 0 2
2.563	2.565	1	2	2	1	0
2.428	2.434	1	2	2	0	1
2.413	2.409	2	5	1	0	2
2.313	2.314	4	5	2	1	1
2.303	2.303	7	7	0	3	1
2.217	2.219	6	7	2	2	0
2.201	2.207	4	5	0	2	2
2.121	2.125	2	4	1	3	1
2.050	2.061	2	4	2	2	1
2.040	2.042	1	2	1	2	2
1.911	1.913	50	38	0	4	0
1.860	1.864	2	4	2	3	0
1.856	1.858	3	4	2	1	2
1.757	1.758	1	1	2	3	1
1.752	1.753	1	1	1	3	2
1.746	1.747	1	1	0	1	3
1.719	1.719	22	4	3	0	1
1.710	1.713	3	4	2	2	2, 1 4 1
1.703	1.703	2	3	1	0	3
1.676	1.677	3	4	3	1	1
1.663	1.665	1	2	1	1	3
1.567	1.566	14	13	3	2	1
1.563	1.561	16	20	2	4	0
1.557	1.559	25	25	0	4	2

reaction between ilmenite and the dissociated calcite and also some oxidation of ferrous to ferric state. The reaction of perovskite formation follows immediately the endothermal reaction of calcite dissociation.

Microscopic investigation of the product of sintering shows that most ilmenite is converted in to perovskite with a few relict ilmenite grains. The X-ray diffraction pattern of the end product of sintering (Fig. 4) shows the perovskite formation and calcium ferrite formation of $\text{Ca}_2\text{Fe}_2\text{O}_5$ composition. The ilmenite peaks completely disappear, indicating its complete conversion

Table 5
X-ray diffraction data of calcium ferrite

d (Å) ASTM	d (Å) Observed	I ASTM	I Observed	h	k	l
33.89	3.885	10	17	1	0	1
3.76	3.752	5	3	1	1	1
3.69	3.693	20	15	1	3	0,
3.45	3.452	5	3	1	2	1
3.06	3.047	20	15	1	3	1
2.803	2.794	70	71	2	0	0
2.717	2.736	70	80	0	0	2
2.685	2.688	100	100	1	4	1
2.613	2.618	20	14	1	5	0,
2.543	2.522	5	7	0	2	2
2.488	2.489	10	8	2	0	1
2.453	2.454	5	7	2	1	1
2.352	2.355	20	15	2	2	1,
2.226	2.225	10	10	2	4	0
2.186	2.190	20	12	1	3	2,
2.080	2.079	40	30	1	6	1
2.062	2.063	40	32	2	4	1
1.950	1.948	70	64	2	0	2
1.883	1.869	20	24	1	5	2,
1.846	1.848	40	30	2	6	0,
1.821	1.819	10	10	0	6	2
1.743	1.747	20	14	3	3	0
1.665	1.672	10	8	1	8	1,
1.623	1.629	20	15	1	3	3
1.593	1.585	40	28	3	4	1
1.559	1.560	40	31	1	4	3
1.540	1.541	20	14	2	8	0

in to perovskite. The X-ray diffraction data of the resulted perovskite and calcium ferrite are given in Tables 4 and 5 respectively.

In thin section the obtained perovskite is brown, shows a very high relief, weak pleochroism and birefringence.

In favour to study the effect of temperature on the efficiency of perovskite formation from ilmenite, some runs were carried out using ilmenite mixtures with calcite and graphite in 1 : 1 : 0.1 ratio at different temperatures, ranging from 700 up to 950 °C during 1.5 hours. As it is evident from the ob-

Table 6*Effect of temperature on perovskite formation*

Temperature, °C	Efficiency of sintering, %
700	21.4
800	65.8
860	87.5
900	98.4
950	98.7

tained results (Table 6) the efficiency of perovskite formation sharply increases with temperature. The products of sintering of these runs were identified by microscope and by X-ray diffractometer.

Under microscope, the products of sintering at 700 and 800 °C contain considerable amount of unreacted ilmenite together with perovskite and calcium ferrite. The amount of perovskite and calcium ferrite increases parallel with increasing sintering temperature. At 900 and 950 °C, the reaction of sintering of ilmenite with calcite takes place completely. At these temperatures a few relict ilmenite grains were observed in a matrix of perovskite and calcium ferrite.

The X-ray diffraction pattern of the end product of sintering at 950 °C is shown in Fig. 4. No ilmenite peaks are observed in this product, indicating the complete sintering and the formation of perovskite.

To study the effect of graphite on the sintering of ilmenite with calcite, some runs were carried out using ilmenite mixtures of the following composition: ilmenite : calcite : graphite in 1 : 1 : 0.05, 1 : 1 : 0.1 and 1 : 1 : 0.2 ratios at 900 °C and during 1.5 hours.

The products of sintering of these runs were identified under microscope and by X-ray diffraction analysis. Under microscope in all sections a few relict ilmenite grains were detected in a matrix of perovskite and calcium ferrite. This indicates the complete sintering of ilmenite with calcite. The obtained products differ only in the crystallization degree. The coarser grains of perovskite are obtained using graphite in amount of 10% of ilmenite quantity.

The X-ray diffraction pattern of the sintering products of these runs are shown in Fig. 5. It can be observed that the ilmenite peaks completely disappear in all runs. This suggests the complete sintering reaction between ilmenite and calcite with the formation of perovskite and calcium ferrite. The peaks of perovskite and calcium ferrite are less intense in the products of runs A and C, while they are of high intensity in the product of run B, using graphite in 10% amount of the ilmenite. It is observed that the intensity

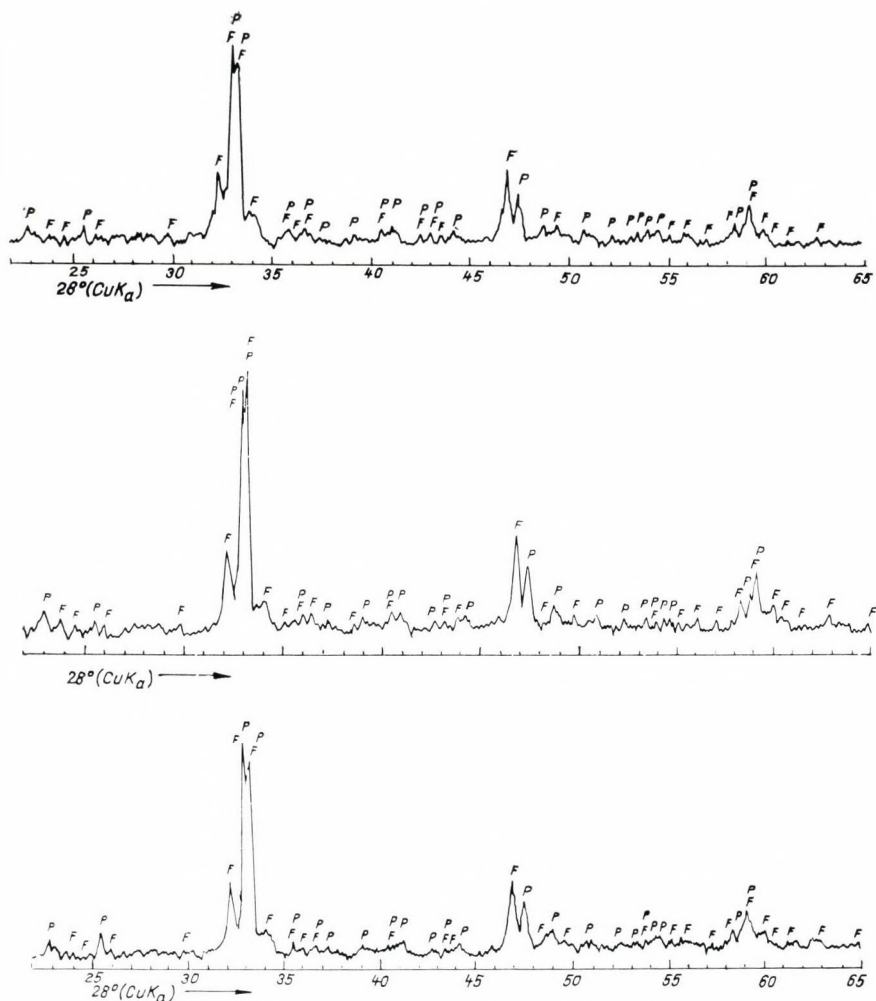


Fig. 5. X-ray diffraction pattern of the end product of ilmenite sintering with calcite in presence of different amounts of graphite at 900 °C. A: Using graphite in a quantity of 5% of that of ilmenite; B: Using graphite in a quantity of 10% of that of ilmenite; C: Using graphite in a quantity of 20% of that of ilmenite

of X-ray peaks of perovskite decreases when the amount of graphite becomes less or above 10% of ilmenite quantity and there is no change in the phases. The X-ray peaks of the obtained perovskite are narrow, sharp and intense, indicating well-crystallized products.

The microscopic study of the product phases of ilmenite sintering with calcite is consistent with the X-ray diffraction pattern.

From this study, it may be concluded that complete conversion of ilmenite in to perovskite takes place by its sintering with calcite at temperatures of 900—950 °C, depending on the time of sintering. Calcium ferrite is obtained

as a second phase with the orthorhombic perovskite. The obtained results are consistent with the previously reported data [1] of perovskite formation at 970 °C from the reaction of rutile with calcite.

Conclusions

Perovskite can be formed from ilmenite by its solid phase reaction with calcite. Complete conversion of ilmenite in to perovskite takes place at 950 °C. Calcium ferrite is resulted together with orthorhombic perovskite. The X-ray diffraction data of the obtained perovskite are in good agreement with the ASTM data of the natural mineral.

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К ВОПРОСУ ПРЕВРАЩЕНИЯ ИЛЬМЕНИТА В ПЕРОВСКИТ

А. М. АБДЕЛ РЕХИМ

Р е з ю м е

Перовскит представляет собой компонент титаносодержащих шлаков, получаемых в процессе расплавления ильменита с коксом в присутствии известняка-флюса. Он встречается как аксессуарный минерал в основных и щелочных по составу магматитах. Кроме того, он проявляется в некоторых контактово-метаморфизованных, нечистых известняках.

Рассматриваемые исследования касаются условий формирования перовскита из ильменита путем применения дериватографа. Эта экспериментальная работа была проведена над ильменитом Абу Гхалха. Вызванные при этом фазы реакции ильменита с кальцитом изучались микроскопически и путем применения диффрактометра Сименс Кристаллофлекс. При этом полное превращение ильменита в перовскит имело место при температуре 950 °С.

THERMAL TRANSFORMATION OF MONTMORILLONITE UNDER SIMULTANEOUS LOAD AND VOLATILE PRESSURE

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The transformation of the bentonite from Istenmezeje (Heves County) of about 90 per cent montmorillonite content was investigated as a function of temperature, load and volatile pressure. As contrasted to kaolinite, in case of montmorillonite the opposite effect of load and volatile pressures is of much smaller extent. At 550 °C and under 5.0 kilobar load pressure montmorillonite remains stable in closed system also during long experimental times. At 700 °C only the high temperature phases known also from literature can be demonstrated and out of them the identification of indialite bears great importance. In geological aspects the stability of montmorillonite in high p-t range indicates that if clayey sediments of montmorillonite content get greater depths, large quantity of water may get the same depth.

Introduction

As it has been stated (E. SZÁDECZKY-KARDOSS et al., 1971, 1972; L. PESTY—O. TOMSCHEY, 1976) the load and volatile pressures distinguished in the earth's crust in first approximation are of opposite effect. Among others and in addition to kaolinite the thermal decomposition of montmorillonite as a function of temperature, simultaneous load and volatile pressures was investigated. The results of the preliminary experiments resp. their data were reported earlier. In this paper the results unpublished till now and those obtained in a higher p-t range will be discussed. The applied experimental methods and tools were also demonstrated in foregoing papers (B. KLIBURSZKY et al., 1969; E. SZÁDECZKY-KARDOSS et al. 1969).

Experimental results and their discussion

Experiments carried out with montmorillonite starting sample are comprehended in Table 1, and the results of these experiments will be discussed as follows.

In case of experiments of lower temperature (200 and 400 °C) montmorillonite remains essentially unchanged. In opened system without pressure the change was indicated only by the appearance of dehydrated montmorillonite in small quantity. On the basis of the diffractograms of the samples treated with glycerine and endured 400 °C the swelling capacity of mont-

Table 1
Comprehensive table of the experiments carried out with montmorillonite

Number of experiment	Crystalline phases													
	Temperature (°C)	Load pressure (kilobar)	Volatile pressure (kilobar)	Time (hours)	Montmorillonite	Montmorillonite-chlorite mixed structure	Indialite	Mullite	Chlorite	Zoisite	Hematite	Cristobalite	Quartz	Quasi-amorphous
3288	200	—	—	36	++	—	—	—	—	—	—	+	(+)	—
3283	400	—	—	36	++	—	—	—	—	—	—	+	(+)	—
3303	400	—	0.8	48	++	—	—	—	ny	ny	—	+	(+)	—
3318	480	—	—	36	++	—	—	—	—	—	—	+	(+)	—
3324	200	1.0	—	36	++	—	—	—	—	—	—	+	(+)	—
3274	400	1.0	—	36	++	—	—	—	—	—	—	+	(+)	—
3281	400	1.0	—	36	++	—	—	—	—	—	—	+	(+)	—
3276	400	1.0	own	36	++	—	—	—	—	—	—	+	(+)	—
3280	400	1.0	own	36	++	—	—	—	—	—	—	+	(+)	—
3326	480	1.0	0.9	6	++	—	—	—	—	—	—	+	(+)	—
3313	480	1.0	—	36	++	—	—	—	ny	ny	—	+	(+)	+
3314	480	1.0	own	36	++	—	—	—	ny	ny	—	+	(+)	+
3690	450	2.0	0.95	51	+	—	—	—	—	—	—	+	+	—
3723	450	—	—	140	—	—	—	—	—	—	—	+	+	—
	550	2.0	1.3	96	(+)	+	ny	—	(+)	(+)	(+)	—	+	+
	600	—	—	4	—	—	—	—	—	—	—	—	—	—
3685	450	3.5	0.9	53	++	—	—	—	—	—	—	+	(+)	+
3749	450	3.5	own	48	(+)	+	—	—	+	+	(+)	—	+	+
	550	—	—	240	—	—	—	—	—	—	—	—	—	—
3682	450	5.0	own	48	(+)	+	—	—	(+)	(+)	(+)	—	(+)	+
	550	—	—	48	—	—	—	—	—	—	—	—	—	—
3861	450	5.0	own	48	(+)	+	—	—	(+)	(+)	(+)	—	(+)	+
	550	—	—	72	—	—	—	—	—	—	—	—	—	—
3949	450	5.0	1.0	24	(+)	+	—	—	(+)	(+)	(+)	—	(+)	+
	550	—	—	676	—	—	—	—	—	—	—	—	—	—
	450	—	—	44	—	—	—	—	—	—	—	—	—	—
4006	550	2.0	1.3	120	—	—	++	ny	—	—	—	—	+	mica: (+)
	700	—	—	100	—	—	—	—	—	—	—	—	—	anorthite: +
	450	—	—	48	—	—	—	—	—	—	—	—	—	—
4081	550	2.0	1.6	114	—	—	++	(+)	—	—	—	—	+	anorthite: +
	700	—	—	—	—	—	—	—	—	—	—	—	—	—

* Within the same experiment the temperature and time values denote the duration at the corresponding temperature, i.e. these do not mean total durations. ++ — very much; + — much; (+) — small quantity; ny — traces; — — lacking.

morillonite remained unchanged. As a result of the treatment with glycerine the base reflexion of the sample rose over 18 Å, independently of the load pressure applied in the course of the experiment. Transformation begins at 480 °C when as a result of load pressure small amount of bad-crystallized zoisite develops. On the effect of simultaneous volatile pressure improves the crystallinity of zoisite.

At 550 °C montmorillonite can be detected though subordinate in quantity and in bad crystalline state. At the same time, in addition to the independent chlorite phase a montmorillonite-chlorite mixed structure has also developed and this could be satisfactorily identified in the diffractograms. The change of the base reflexion of the mixed structure always produced the change of chlorite reflexion of this structure, but their ratio remained unchanged.

In the same temperature range, independently of the duration spent at 450 resp. at 550 °C, the same mineral assemblage developed regarding its composition, i.e. in addition to the afore-mentioned montmorillonite-chlorite mixed structure, chlorite and bad-crystallized montmorillonite small quantities of hematite, kaolinite, quartz, mullite, zoisite and quasi-amorphous phase of about 20 per cent consisting probably of SiO_2 appeared. As against this assemblage, during the experiments carried out with kaolinite (L. PESTY—O. TOMSCHEY, 1976) at higher temperatures (550—650 °C) cristobalite, keatite and finally quartz developed from the amorphous phase. In the experimental samples in question the amorphous SiO_2 released during the high p-t decomposition of montmorillonite developed as final product always in form of α -quartz (at 700 °C) and this phenomenon is proved also by the absence of keatite at 550 °C. In contrast to the thermal decomposition of kaolinite, in this case the order of crystallization is amorphous — cristobalite — quartz (without keatite). The scheme of the reaction is the same but the final product is dissimilar, i.e. during the thermal decomposition of kaolinite a SiO_2 modification develops which regarding its X-ray diffractometric parameters is intermediate between the α and β modifications, while in case of montmorillonite unambiguously α -quartz was determined.

Under 5.0 kilobar load pressure no change follows in the phase composition of the sample even after the duration of more than one month (see Table 1, runs No. 3682, 3681 and 3949).

The most considerable transformation was observed at 700 °C. At this temperature the montmorillonite and montmorillonite-chlorite mixed structure are absolutely lacking, moreover, the phases being considered to be transitional (zoisite, chlorite, hematite) are essentially also absent. The difference between the experiments of different duration is manifested only by the fact that in some traces relating to mica in the shorter one will be absent in the longer one.

At 700 °C, in accordance with the composition of the starting sample small quantity of anorthite develops. As against the heating experiments performed under 1 atmosphere the occurrence of indialite is worthy of mention. The formation of indialite was earlier reported (O. TOMSCHEY, 1974) as a result of the decomposition of montmorillonite under pressure. These data, however, were obtained in the experiments carried out at 550 °C and under only volatile pressure (i.e. in loose, closed system). In the experiments in question under

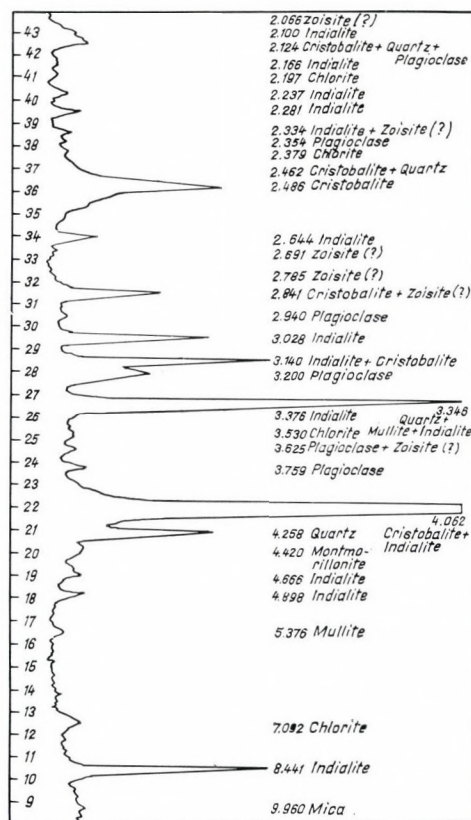


Fig. 1

simultaneous load and volatile pressure indialite could be determined neither in traces and appeared only in the experiments carried out at 700 °C and this proved to be well-crystallized. Its X-ray diffractometric data show good agreement with those known from literature (A. MIYASHIRO, 1957; Selected power diffraction data for minerals, ed. L. G. BERRY, 1974). Data are shown in Tables 2 and 3. The agreement of the data of three authors is obvious. The data of indialite obtained in our experiments (X-ray diffractogram in Fig. 1)

Table 2

X-ray diffractometric data of indialite (all the evaluable peaks with the relative intensities and height intensity values) on the basis of literature data and of the samples obtained at 700 °C

hkl	Literature data		Experimental samples					I _h
	(1974)		No. 4081			No. 4006		
	d (Å)	I/I ₀	d (Å)	I _{rel}	I _h	d (Å)	I _{rel}	
100	8.48	100	8.506	100	100	8.489	100	100
110	4.89	30	4.901	18	16	4.901	17	17
002	4.679	16	4.668	10	9	4.671	9	10
102	4.094	50	4.090	53	53	4.095	46	30
112	3.379	55	3.376	60	60	3.375	60	60
202	3.138	65	3.139	53	53	3.139	56	65
211	3.027	85	3.037	82	62	3.035	82	77
212	2.640	25	2.647	28	21	2.644	23	23
220	2.441	6	2.443	6	6	2.434	5	4
302	2.414	4	2.415	3	3	2.418	3	3
004	2.338	12	2.335	10	9	2.337	9	10
311	2.276	6	?	?	?	?	?	?
213	2.231	6	2.230	?	?	2.230	?	?
222	2.165	6	2.173	9	5	2.172	7	6
114	2.108	8	2.102	13	9	2.102	12	10
312	2.098	12						
204	2.046	4	2.046	2	2	2.047	2	1
320	1.941	8	1.946	6	6	1.946	6	5
402	1.927	6	1.931	5	5	1.931	6	5
321	1.901	4	1.904	3	2	1.901	3	2
313	1.875	16	1.876	16	10	1.880	16	13
410	1.846	6	1.850	5	5	1.851	5	5
411	1.811	8	1.802	11	10	1.803	10	10
304	1.800	10						
412	1.718	4	1.718	3	2	1.713	2	2
224	1.688	30	1.692	33	16	1.693	30	21
314	1.656	4	1.660	3	2	1.660	5	6
323	1.647	4	?	?	?	?	?	?
330	1.629	4	1.631	3	2	1.634	?	?
215	1.615	4	?	?	?	1.609	?	?
420	1.599	6	1.606	5	2	1.602	?	?
413	1.589	10	1.591	10	7	1.592	8	5
404	1.569	2	1.569	2	1	1.569	?	?
006	1.558	4	1.559	3	2	1.559	?	?
332	1.538	4	1.539	3	2	1.538	?	?
324	1.494	8	1.497	8	4	1.495	7	4
116	1.485	6				1.486	6	4
315	1.463	6	1.461	5	3	1.463	4	2
206								

show good agreement with those obtained by MIYASHIRO from the devitrification of glass at 1200 °C and this is in accordance with our former statement.

The fact that as against the experiments performed under volatile pressure, when applying load pressure indialite occurs only at much higher temperatures supports our former statement (E. SZÁDECZKY-KARDOSS et al., 1971), i.e. load pressure restrains the dehydration decomposition of clay minerals and increases the temperature of occurrence of the transitional and equilibrium phases, respectively.

Table 3

*Indialite peaks of more than 10 relative intensity
on the basis of literature data and of the experimental samples*

Literature data		Experimental sample			
		No. 4081		No. 4006	
d (Å)	I/I ₀	d (Å)	I/I ₀	d(Å)	I/I ₀
8.48	100	8.506	100	8.489	100
4.89	30	4.901	18	4.901	17
4.679	16	4.668	10	4.671	9
4.094	50	4.090	53	4.095	46
3.379	55	3.376	60	3.375	60
3.138	65	3.139	53	3.139	56
3.027	85	3.037	82	3.035	82
2.640	25	2.647	28	2.644	23
2.338	12	2.335	10	2.337	9
1.875	16	1.876	16	1.880	16
1.688	30	1.692	33	1.693	30
1.589	10	1.591	10	1.592	8

When summing up it can be stated that in case of simultaneous application of the load and volatile pressures montmorillonite is stable at lower temperatures, i.e. up to about 400 °C no difference can be observed between the starting and final samples. At 480 °C only the trace appearance of bad-crystallized zoisite indicates the start of reaction. In case of higher load pressure and relatively shorter duration (four days) significant change does not occur even when the water pressure caused by the water released by montmorillonite amounts to about the half of the applied load pressure.

In the experiments carried out at 550 °C and without load pressure the final crystalline phase assemblage is already developed but only the phases developed under 2.0 kilobar can be regarded to be of well-crystallized state. At the same time, applying simultaneous load pressure the measure resp. rate of montmorillonite decomposition decreases as a function of the load pressure itself and in case of 5.0 kilobar load pressure the montmorillonite-chlorite mixed structure can be considered to be stable even at 550 °C, further the amorphous phase is of the greatest quantity under this load pressure.

When applying simultaneous load and volatile pressures, an abrupt change follows above 550 °C, i.e. the quantity of the amorphous phase suddenly decreases (at 600 °C only the half of that of 550 °C can be observed), and all the other crystalline phases show good crystallinity, as well. The equilibrium phase system of indialite + anorthite + quartz obtained at 700 °C is accompanied by mullite of small quantity. The effect of duration is significant from

the point of view of reaching the equilibrium state and this is indicated by the gradual disappearance of mica traces and other accessory components as a function of time.

On the basis of these and from the geological point of view it is obvious in high p-t depths (i.e. 550 °C, 2—3—5 kilobar load and volatile pressures) the montmorillonite-chlorite crystal structure is viable. This bears great importance especially in the subduction zones: great quantities of crystalline water may get much greater depths than ever assumed. Being released in these depths this water highly lowers viscosity itself promoting in this way the subduction processes of great depths.

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ТЕРМИЧЕСКОЕ ПРЕОБРАЗОВАНИЕ МОНТМОРИЛЛОНИТА В СЛУЧАЕ ОДНОВРЕМЕННОЙ НАГРУЗКИ И ДАВЛЕНИЯ ЛЕТУЧИХ КОМПОНЕНТОВ

Л. ПЕШТИ—О. ТОМШЕИ

Резюме

Было исследовано преобразование иштенмезейских (область Хевеш) бентонитов, содержащих до 90% монтмориллонита, в зависимости от температуры, нагрузки и давления летучих компонентов. В отличие от каолинита в случае монтмориллонита противоположное влияние нагрузки и давления летучих компонентов значительно более меньшего размера. В случае закрытой системы монтмориллонит при температуре 550°C. и нагрузке в 5,0 килобар в течение длительного промежутка времени также стабилен. При температуре 700°C могут проявляться уже только известные и по литературным данным высокотемпературные фазы, среди которых наиболее важное значение имеет индиалит. Стабильность монтмориллонита, проявляющаяся в большом диапазоне давления-температуры, в геологическом отношении означает то, что во время попадания глинистых отложений, содержащих монтмориллонит, на большую глубину, туда также проникает в большом количестве и вода.

UNTERSUCHUNGEN ÜBER DIE TERRESTRISCHE KOMPONENTE NATÜRLICHER STRAHLUNG IN EINER ZONE RUMÄNIENS

Von

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Ungefähr 1000 Bodenproben aus einer Zone in Rumänien wurden hinsichtlich des Gehaltes an natürlichen Radioelementen: U, Th, ^{226}Ra und ^{40}K untersucht. Man stellte eine Übereinstimmung zwischen den Konzentrationen dieser Radioelemente im Boden mit der petrografischen Herkunft des Bodens fest. Im größten Teil des untersuchten Gebietes überschreiten die Gammadosen in der Atmosphäre, 1 m über dem Boden, den Wert von 50 mrad/Jahr, sie liegen aber innerhalb der in der Fachliteratur angegebenen Grenzen, so daß das untersuchte Gebiet nicht zu denen mit erhöhter natürlicher Radioaktivität gehört.

1. Einleitung

Die allerorts in der Welt unternommenen, umfassenden Studien zur Abschätzung der Intensität der natürlichen ionisierenden Strahlung verfolgen vor allem zwei Zwecke:

— Aufdeckung somatischer und genetischer Auswirkungen der natürlichen Strahlungskomponenten zwecks Festsetzung des Verhältnisses Dosis-Effekt im Bereiche kleiner Dosen von Kernstrahlungen und

— Auffinden von Beziehungswerten für ein bestimmtes Territorium welche später die Abschätzung der von der Verwendung von Kernstrahlungen herrührenden radioaktiven Verseuchung ermöglichen sollen.

Die natürliche externe Gamma-Strahlung terrestrischer Herkunft stammt von den natürlichen radioaktiven Elementenfamilien und vom Kalium-40, Radionukliden die in der ganzen Erdrinde verbreitet sind. Der Gehalt an radioaktiven Elementen in Böden hängt in erster Reihe von den chemisch-mineralogischen Zusammensetzung der Gesteine ab, auf denen sie sich bilden, wie auch von den physiko-chemischen Umwandlungen, die in den Gesteinen während der Bodenbildung stattfinden. So ist der Gehalt an radioaktiven Elementen in Böden, die sich auf sauren Gesteinen der Granitreihe bildeten, höher als in Böden, die sich auf Grund einer Umwandlung von basischen Gesteinen (Basalt) bildeten.

2. Materialien und Methode

Die Untersuchungen erstreckten sich auf ein Gebiet im Osten des Landes, welches eine bunte Vielfaltigkeit der mineralogischen Zusammensetzung des Bodens aufweist und wo im Zeitabschnitt 1973—1974 1000 Bodenproben eingebracht wurden. Gleichartige Proben von 350—500 g wurden aus dem Gesichtspunkte ihres Gehaltes an natürlichem Uranium, Thorium, ^{22}Ra und ^{40}K an einem einkanaligen Spektrometer VA-M-120 analysiert, wobei eine konstante Geometrie eingehalten wurde. Der Apparat wurde energetisch mit AMERSHAM-Quellen von ^{22}Na , ^{226}Ra , ^{137}Cs und mit labormäßig hergestellten Standardpräparaten von U, Th, Ra, K geeicht, um die Formeln zur Berechnung der Konzentrationen aufzustellen:

$$\text{mgU/kg Lös.} = 157 N_{\text{U}}/M \quad (1)$$

$$\text{mgTh/kg Lös.} = 171 N_{\text{Th}}/M \quad (2)$$

$$\text{mg}^{226}\text{Ra/kg Lös.} = 3,7 \cdot 10^{-4} N_{\text{Ra}}/M \quad (3)$$

$$\text{mg}^{40}\text{K/kg Lös.} = 187 N_{\text{K}}/M \quad (4)$$

wobei N_{U} , N_{K} , N_{Th} und N_{Ra} den Flächeninhalt der Kurvenspitze nach der Korrektur für den Nulleffekt und der Interferenz der entsprechenden Nukliden, M aber die Masse der Bodenprobe (Gramme) darstellen.

Die geringste nachweisbare Menge für die 4 Elemente beträgt: 0,62 mg ^{40}K /kg Boden, 1,08 mg nat. Th/kg Boden, 1,17 mg nat. U/kg Boden, 0,0014 μg ^{226}Ra /kg Boden.

Dem Gehalte an natürlichen Radioelementen des Bodens entsprechend wurde die Dosis externer Gamma-Strahlung im Luft 1m über dem Boden errechnet, wobei die von Ph. GUSTAFSON und SARMAKH S. BRAR vorgeschlagenen Formeln (3) verwendet wurden.

3. Ergebnisse und Diskussion

Die Analysenresultate (Tabelle 1) gestatteten die Einteilung des untersuchten Gebietes in sechs Zonen im Hinblick auf die Dosen der externen Gammastrahlung (Abb. 1).

Im allgemeinen stellte man eine innige Wechselbeziehung zwischen der mineralogischen Zusammensetzung des Bodens und der Konzentration der natürlichen Radioelemente fest.

So fand man die höchsten Konzentrationen an natürlichen Radioelementen bzw. Gamma-Dosen zwischen 80—180 mrad/Jahr im westlichen Teil der Zone A, wo sich an natürlichen Radioelementen reiche Formationen vom

Typus des Sandsteinkonglomeraten, Mergel-Sandstein-, Mergel-Ton-Flysches finden, die über kristalline Schiefer, metamorphes Gestein aufgelagert sind, welche gleichfalls einen erhöhten Gehalt an natürlichen radioaktiven Elementen aufweisen. Diese Gesteine zeigen ihrerseits Intrusionen magmatischen Gesteins, bei welchen der Urangehalt von den primären gegen die finalen Bildungsstadien hinansteigt. Im Osten der Zone A (A_3 — A_4) sind die terrestren Gamma-Dosen geringer (50—75 mrad/Jahr); dies ist von der Anwesenheit von sandsteinigen, bituminösen und kohlenhaltigen sedimentären Gesteinen bedingt.

Die Unterzone A_3 — A_4 bildet den Übergang zum Osten des untersuchten Gebietes (Zone C), wo sedimentäre Gesteine vorherrschen, die im allgemeinen einen geringern Gehalt an natürlichen Radioelementen als die Granitgesteine aufweisen.

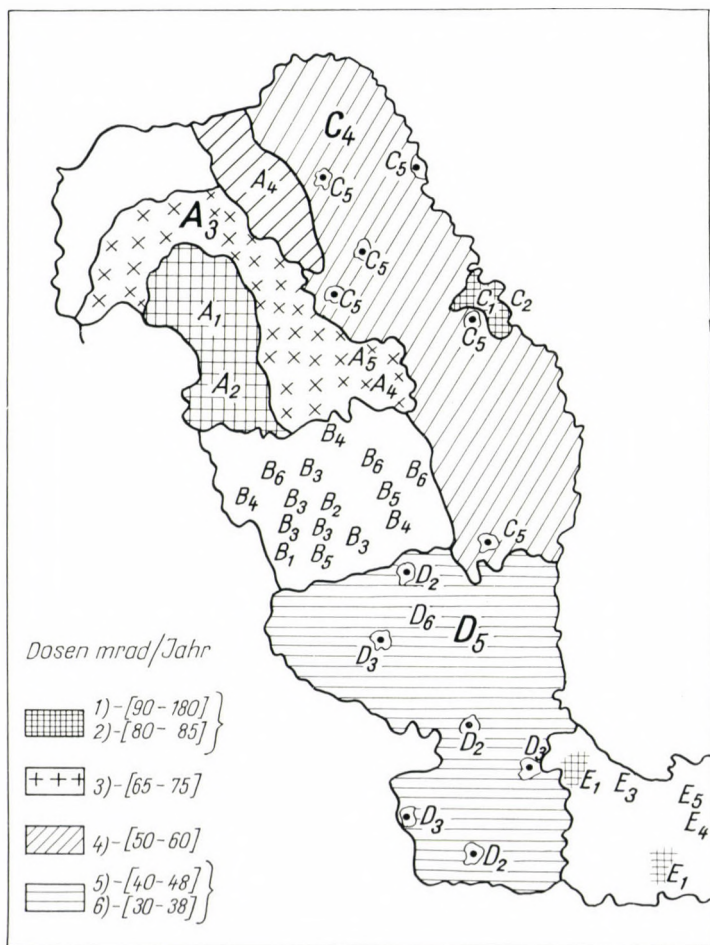


Abb. 1

Table 1

Die Konzentratione der natürlichen radioaktiven Stoffe

Zone		
A	A ₁ A ₂	Formationen vom Typus des Sandsteinkonglomerates, Mergel-Sandstein, Mergel-Ton-Flysches, die über kristalline Schiefer aufgelagert sind
	A ₃ , A ₄	sandsteinige bituminöse und kohlehaltige Sedimentgesteine
B	B ₁ , B ₂ B ₃ , B ₄ B ₅ , B ₆	Formationen vom Typus des Flysches, Tone
C	C ₁ , C ₂	Ton und Tonschiefer
	C ₄	Ton, Sand, Sandstein und Kalkstein
	C ₅	Anschwemmablagerungen und Löß
D	D ₂ , D ₃	Mergel-Sandstein
	D ₅ , D ₆	Alluvium, Löß und Sumpfablagerungen die über Mergel und Sand aufgelagert sind
E	E ₁ , E ₃	Magmatische Gesteine, Granite und metamorphe Gesteine
	E ₄ , E ₅	Kalkstein und Mergel

Die radioaktivsten sedimentäre Gesteine sind die Tone und die Tonmergel, die in der Unterzone C₁—C₂ verbreitet sind, wo die Gamma-Strahlungsdosis zwischen 71 bis 94 mrad/Jahr schwankt. Ärmer an natürlichen Radioelementen sind Steinsalz, Gips, Kalkstein, Dolomit (rein chemische und organogene Gesteine), die man im Rest der Zone C begegnet (Bild 1), wo die Gamma-Strahlungsdosis zwischen 50—60 mrad/Jahr schwankt (C₄). Geringere Strahlungspegel (42—46 mrad/Jahr) fand man in Ortschaften mit Anschwemm- und Lößablagerungen (C₅—C₆).

Im Süden des Gebietes (Zone D) mit Ausnahme einiger Ortschaften (D₂), ist der Gehalt an natürlichen Radioelementen der kleinste des ganzen studierten Territoriums; dies erklärt sich durch die Gegenwart quaternärer Sedimente, die aus Alluvien, Lössen und Sumpfablagerungen bestehen. Die Gammadosis in dieser Zone schwankt zwischen 30 und 48 mrad/Jahr.

Im Südwesten des untersuchten Gebietes (Zone E) findet man Dosen

in der Erdrinde und die entsprechende Dosisleistungen

mgU/kg Löss	mgTh/kg Löss	mg ⁴⁰ K/kg Löss	μg ²²⁶ Ra/kg Löss	Dosisleistungen mrad/Jahr
6,43 (4,35—11,5)**	16,4 (13,88—21,3)	1,88 (1,37—2,65)	0,0095 (0,007—0,012)	121* (81—180)**
2,28 (1,45—3,18)	10,4 (9,6—16,7)	1,3 (1—1,7)	0,009	66 (50—75)
2,42 (0—9)	8,85 (4—13,3)	1,34 (1—3,42)	0,001 (0,0007—0,009)	62 (30—70)
4,66 (3,93—5,91)	11,6 (10,05—15,96)	1,2 (1,1—2,56)	0,0085 (0,005—0,011)	85 (71—94)
1,5 (1,2—1,9)	9,1 (6—16)	1,11 (1—2,56)	0,005 (0,002—0,009)	53 (50—60)
1 (0—1,9)	7,9 (7—10,8)	1,1 (1—1,4)	0,002 (0,001—0,0025)	45 (42—46)
3,82 (3,2—5,8)	11,5 (10,8—15,7)	1,25 (1,1—1,67)	0,01 (0,002—0,066)	79 (61—85)
1,4 (0—2,0)	7,05 (4—12)	1,25 (1—2,8)	0,0009 (0,0004—0,0048)	47 (30—48)
9 (2,9—17,8)	11,8 (9—15)	1,9 (1,5—2,5)	0,0073 (0,006—0,0088)	123 (66—161)
1,82 (1,37—2,4)	12,6 (7,8—16,3)	1,75 (1—2,8)	0,002 (0,001—0,003)	55 (40—65)

* — Mittelwerte

** — Grenzwerte

von 40—65 mrad/Jahr; dies entspricht dem Typus von Kalkstein- und Mergel-Schichtgesteinen. Die Dosen von 65—161 mrad/Jahr entsprechen magmatischen Gesteinen und im besondern den Graniten, mit metamorphen Gesteinen verbunden.

Eine besondere Lage bietet die Zone B (Bild 1), wo man die größte Veränderlichkeit des Gehaltes an natürlichen Radioelementen aus dem Boden wegen der sehr verschiedenen Gesteinsarten, aus welchen sich der Boden in dieser Zone gebildet hat, feststellte. So schwanken die Dosen im Bereiche von 30 bis 70 mrad/Jahr.

Aus unseren Resultaten ist zu ersehen (Tab. 1), daß das Verhältnis der Mittelwerte Th/U 1—5 beträgt; diese Angaben entsprechen den Werten aus der Fachliteratur. Was Ra und U betrifft, schwankt ihr Verhältnis im Bereiche

von $(1-10) \text{ Ra}/3 \times 10^6 \text{ gU}$. Diese Verlagerung des radioaktiven Gleichgewichtes erklärt sich durch die verschiedene Affinität dieser Radioelemente für das fein verteilte Material der Böden. Thorium und Radium werden von diesen verteilten Bodenteilchen besser fixiert, während das Uran mit Leichtigkeit verschlämmt und von den Huminsäuren fortgeführt wird.

4. Schlußfolgerungen

1. Der Gehalt an Uran, Thorium, Radium und Kalium im Boden schwankt in Abhängigkeit von der chemisch-mineralogischen Zusammensetzung der Gesteine, auf denen sich der Boden gebildet hat.

2. Die Gamma-Dosis in Luft, 1 m über dem Boden gemessen, die dem Gehalt an natürlichen Radioelementen entspricht, schwankt zwischen 38 und 177 mrad/Jahr.

3. Im größten Teil des untersuchten Territoriums wird der Wert von 50 mrad/Jahr überschritten; dieser Wert wird als typische Dosis für die tellurische Komponente des natürlichen Untergrunds betrachtet.

4. Die berechneten Höchstwerte der Gammadosen in der Luft liegen im Bereiche der Variationen der in der Fachliteratur zitierten Regeln natürlicher Strahlung, so daß das untersuchte Gebiet nicht an den Zonen eines erhöhten natürlichen radioaktiven Untergrunds gehört.

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ИССЛЕДОВАНИЯ ТЕРРЕСТРИЧЕСКИХ КОМПОНЕНТОВ ЕСТЕСТВЕННОГО РАДИОАКТИВНОГО ИЗЛУЧЕНИЯ В ОДНОЙ ИЗ ЗОН РУМЫНИИ

Э. БОТЕЗАТУ — Х. ВЕЙССБУХ — М. ГРЭДИНАРУ — А. САРАЙМАН

Резюме

Исследовалось около 1000 образцов почв, отобранных в одной из зон Румынии, для определения содержания в них естественных радиоэлементов U, Th, ^{226}Ra и ^{40}K . При этом наблюдалось, что концентрации этих элементов в почве обусловлены соответствующим геологическим строением окружающего района. На большей части исследованного района гаммадозы в атмосфере на высоте 1 м над почвой превосходят значение 50 мрад в год, но они все-же находятся в пределах указанных в литературе величин. Таким образом рассматриваемый район не относится к числу районов с повышенной естественной радиоактивностью.

BARROW TYPE METAMORPHISM IN THE CRYSTALLINE BASEMENT OF SOUTHEAST TRANSDANUBIA

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Below the Pannonian resp. Miocene strata mostly mesozonal crystalline schists were hit by a part of exploratory bores drilled in the area of South-Baranya. According to Slavonian evidences of similar features these are in direct connection with the metamorphic mass of the Papuk and Krndija Mountains. The crystalline schists covered by younger sequences and extending from the river Drava to the Mecsek Mountains, the so-called Barrow-type Crystalline Schist Formation of Görcsöny is a progressive serie of NW–SE strike and can be characterized by isogrades. The grades are increasing from southwest to northeast and are completed by granitization. The metamorphic character of the progressive serie can be well correlated with the classical Dalmatian sequence and it was produced by one significant deformation phase, except the basement of the southern foredeep of the Mecsek Mountains. The age of deformation is dated to be pre-Silurian, and is believed to be Upper Baikalian by B. JANTSKY. The basement of the southern foredeep 1.5 to 2 km wide mylonitized zone of the Mecsek Mountains is built-up by the Old-Paleozoic limestone-phyllite-greenschist series. The wedging of this sequence into the old metamorphite-granite mass was produced by Early Variscan transcurrent tectonics. The tectonic zone is bordered from the south by the retrogressively metamorphized margin of the Barrow type metamorphic sequence.

Introduction

In the area between the Western Mecsek Mountains and the Drava river fourteen water- and hydrocarbon prospecting resp. geophysical check bores explored crystalline schist formations of different metamorphic grade under Pannonian or Miocene strata (between 1960 and 1975). The major part of these bores supplied continuous core sampling and penetrated a considerable part of the metamorphic formations. The flat ridge is covered by the network of numerous geoelectric (resistance and telluric), seismic refraction profiles and large-scale, up-to-date aeromagnetic and aerogamma records provide considerable aid to evaluate the basement complex.

By means of an instrument designed by EGYED and SZEMERÉDY (1962) the orientation of the metamorphic mass has been determined from the aeromagnetic records of large scale (1 : 25 000), and this resulted in a definite WNW–ESE direction. From the point of view of regional classification this proved to be an important recognition (T. SZEDERKÉNYI, 1974), since this direction deviates from the general tectonic direction of the Pannonian massif

by right angle and seems to be identical with the orientation of island hills and buried basement in Slavonia.

The metamorphic zonal structure of the investigated area is demonstrated in Figs 1a, b.

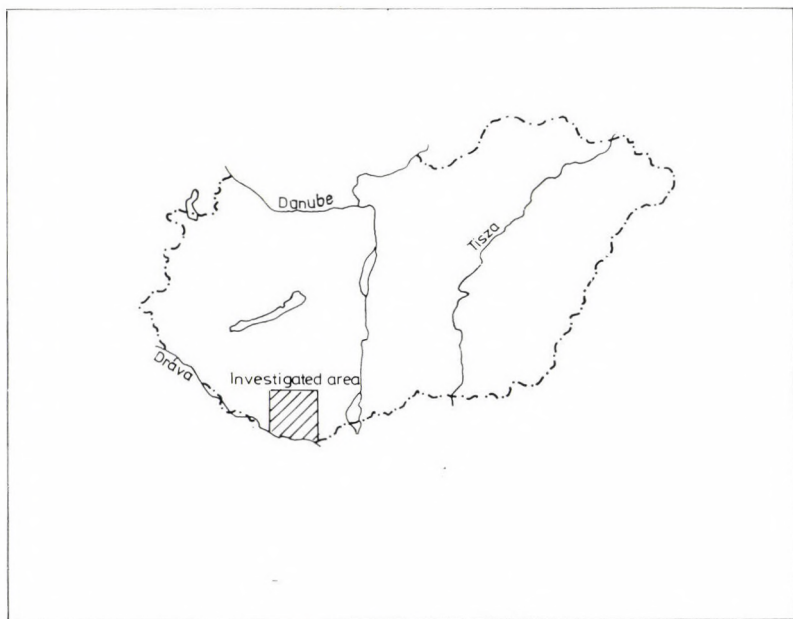
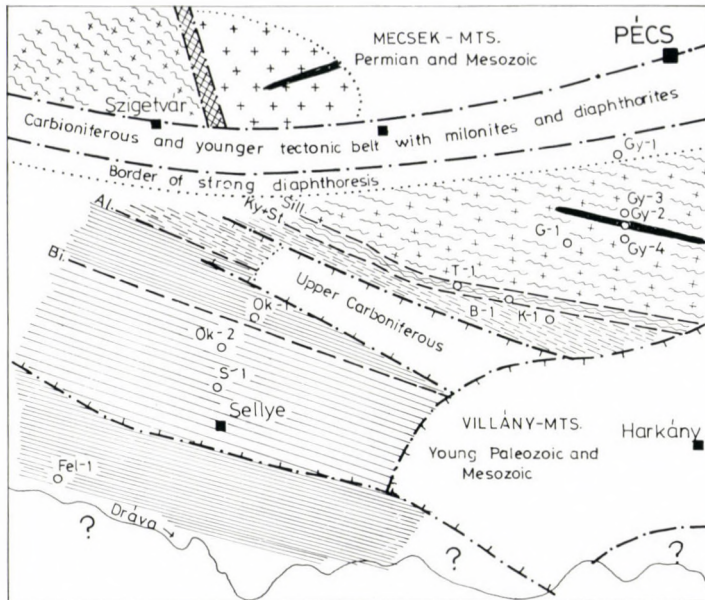


Fig. 1a. The investigated area

The rock material most suitable to investigation derives from the low emergence lying between the Western Mecsek and the Villány Mountains. This is characteristic of the whole area. These metamorphic rocks are called the Barrow Type Crystalline Schist Formation of Görcsöny. Their structure can be seen along a geological profile in Fig. 2.

I. Petrological Classification

On the basis of the metamorphic grades the formation can be divided into five groups. The boundaries between the members are primarily isogrades, secondarily faults. Their exact location cannot be determined because of their covered position. The boreholes mark their location by a precision corresponding to about 1 : 200,000 scale.



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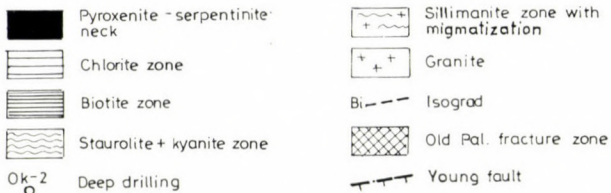


Fig. 1b. Zonal structure of metamorphites of the Görösöny Ridge — Drava Basin

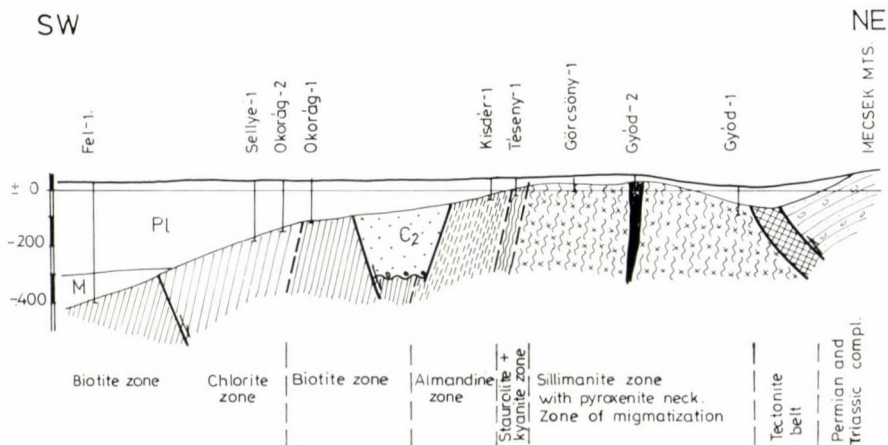


Fig. 2. Rough geological profile across the Görösöny Ridge

1. The Okorág-Sellye Epimetamorphic Member

The hydrocarbon prospecting bore Okorág No. 2. reached the crystalline basement at a depth of 1357.0 m under Pannonian sediments and was drilled down to 1398.0 m. According to the core samples deriving from the top of the basement complex and from the bore's bottom dark-grey chlorite schist was found. Based on the microscopic investigations this rock is not diaphthorite but a primary chlorite schist and contains acid plagioclase (albite) in addition to quartz and chlorite, as well as rather large quantity of tourmaline as accessory mineral.

South of the borehole Okorág No. 2. at a distance of about 3 km the oil prospecting bore Sellye No. 1. was drilled which explored the basement complex between 1932.0 and 1954.0 m. The core samples deriving from the depth intervals of 1932.0—1934.0 resp. 1953.0 and 1954.0 metres are slightly muscovitic, slightly chloritic metamorphic quartz sandstones. Their metamorphic grade is rather low.

2. The Felsőszentmárton Gneiss Member

The oil prospecting borehole Felsőszentmárton No. 1. was drilled in the deepest known part of the Drava Basin and explored the basement between 3919.5 and 3993.0 metres. The rock material is here a nearly vertically foliated, silicified light-grey gneiss. White quartzite veins of 2 to 3 mm thickness are imbedded into the planes of schistosity. This rock (investigated under microscope) might be a biotite gneiss chloritized and silicified during diaphthoresis. It contains relatively small quantity of feldspars which are oligoclase type, fresh twin-lamellated plagioclases. Potash feldspars are rarely found. Because of silification the rock seems to be fresh and the individual components are also fresh except biotite which has been transformed either to muscovite or to chlorite. Regarding the metamorphic grade the rock can be considered of similar facies as the Kisdér mica schists.

3. The Kisdér-Tésény Garnetiferous Mica Schist — Mica Gneiss Member

On the basis of thin section studies this Member can be divided into two Beds:

a) *Muscovite-biotite Schist Bed.* Its most peculiar occurrence was explored by the geophysical check well of Kisdér No. 1. between 128.7 and 136.3 metres under Upper Pannonian sandy sediments. Nearly the same rock was hit also by the water prospecting wells of Rádfapuszta No. 1 and Baksa No. 1. The crystalline schist is a uniformly muscovite-biotite schist characterized by the predominance of muscovite. It is medium-grey, white-stripped

strongly folded rock, usually in vertical position. The white strips and lenses consist of laminated quartz, their thickness varies between 0.5 and 2 cm. In pockets and especially in the strongly folded parts biotite nodules can also be found. The rock locally contains rust-brown, fractured almandine crystals of 0.7 cm size which seem to be of rotation structure. Along schistosity a characteristic pygma-like quartz vein of 5 cm thickness is also included.

In thin section the following two mineral associations of lepidoblastic-granoblastic texture can be observed:

quartz-oligoclase-muscovite-biotite-almandine;

quartz-oligoclase(orthoclase)-biotite-muscovite-almandine.

Tourmaline, apatite, magnetite and chlorite occur as accessory minerals.

On the basis of thin section measurements (6 pieces) the average mineral composition is as follows:

quartz	34.5%	muscovite	25.5%
plagioclase	7.4%	biotite	7.9%
orthoclase	18.0%	chlorite	3.5%
		almandine	5.3%

The *garnets* are strongly fractured, occasionally of rotatory structure and in the fractures chlorite precipitation is found.

The quartz grains are of undulatory extinction and of laminated, lense-like form sometimes with micromosaic texture. Subordinately recrystallized and rounded grains are also found, these, however, do not contain inclusions and are of slightly undulatory extinction.

Plagioclases are rarely twin-lamellated, they rather form large-sized (sometimes of 3 mm) rounded grains which are fresh in general, and can be separated from the quartz grains on the basis of their cleavage. The separation of the cleavage-free grains could be performed only in convergent light. The twin-lamellated forms are of oligoclase composition.

The *orthoclase* is represented by strongly rounded, often fractured laminated crystals of 2 to 3 mm size. The surfaces are strongly resorbed along the cleavage, suggesting former sericitization. Sericite, however, cannot be recognized, the lamellae along fractures are muscovite.

The presence of orthoclase in this rock type and also in the subsequent ones to be discussed, resp. its large quantity cannot be explained unambiguously. On the basis of the microscopic picture of the muscovite-biotite schists it can be stated that these feldspars formed not during metamorphism but these are the relics of the feldspars of the original rock. This rock did not reach the metamorphic grade so that granitization could develop.

Both muscovite and biotite form large-sized, often of 2 cm, bended plates. It is characteristic of biotite that it occurs in the surface schistose bends of the folded rock, i.e. where pressure maxima existed, following usually the

surface of some of the large-sized (quartz or orthoclase) grain. It is found always together often intergrown with muscovite. Muscovite lies between the quartz and feldspar "strips" forming practically strata among them (Fig. 3).

Chlorite is a rather frequent component. Sometimes it forms faces of about 100 micron, but more often occurs as pseudomorph after biotite. This two kinds of occurrence can be regarded primary and secondary chlorite.



Fig. 3. Muscovite-biotite schist; borehole of Kisdér-1. from 131.6 m; +N, M = 150 ×

Tourmaline is an other frequent accessory mineral. Its columns of sometimes 1 mm size are pleochroic, often of zonal structure and occasionally these are surrounded by a pleochroic aureole.

The less frequently occurring *epidote* forms small grains which are arranged in elongated heaps along schistosity.

b) *Biotite-schist, Mica-Gneiss, Biotite-Amphibole Gneiss Bed*

This Bed was separated practically only on the basis of one exploratory bore, i.e. Tésény No. 1. Its characteristic feature is that mica, non-mica ratio is decreased as compared to the foregoing Bed, and the rock becomes gneiss-like. In the gneiss both kinds of mica are found with the strong predominance of biotite on the account of muscovite. The dark-greyish-brown biotite schist strips of sharp boundaries and of 20 to 30 cm thickness are characteristic and these divide both types of gneiss into banks of 1.5 to 1.8 metres. Almandine is frequent both in the biotite schist and in the gneiss but the biotite

schist contains more almandine though its grain size is much smaller. When investigating the mica-gneiss rock perpendicularly of schistosity the alteration of the thin mica-rich and 2—3 mm thick quartz-feldspar strips can be well observed. The latter ones are of bluish shade caused by the kyanite columns occurring in large quantity and reaching occasionally the size of 2 cm. Less frequently, staurolite is also found.

In a depth of 155.0 m, at the boundary of the mica-gneiss and biotite schist a yellowish-white vein of unsettled run and 0.5 cm thickness can be found. The contacting mica-gneiss became also of lighter colour in a thickness of about 5 cm and contains fresh, unfractured, transparent bluish kyanite crystals of max. 4 cm size and 0.4 cm thickness crossing one another. Among the kyanite crystals also fresh, unfractured dark-brown idiomorphic staurolite crystals of max. 1 cm size can be observed, accompanied by a few garnet grains. These crystals are imbedded into the muscovite tissue, thus the whole rock strip is of light shade. In the side of biotite schist the contact strip is less expressed. This is accompanied only by a light (muscovitic) border of 2 mm thickness. The surface of the vein, however, towards the biotite schist is much more complicated (jagged). The embayments of several millimetres along schistosity relate to the fact that this vein intruded between the surfaces of the two rocks on the account of biotite schist. According to the investigations under microscope the yellowish-white vein consists of sillimanite, quartz and plagioclase.

A 36 metres interval of the crystalline basement was explored under the 154.0 m thick Upper Pannonian sediments. The upper part of 11 m of the explored basement consists of the above-mentioned biotite schist — mica gneiss of 25 to 30° schistosity dip.

The lower 25 m thick crystalline schist underlies the Gneiss Bed with an average dip of 60° and with sharp petrological discordance without any fractured boundary. Its rock is biotite-amphibole gneiss with 0.3 to 0.8 m thick biotite schist intercalations. In addition to the garnet kyanite are not found in this section. Its mica content decreases to minimum, except the biotite schist. As contrasted to the gneiss, the micro-folded quartz veins of unsettled run become frequent, moreover, feldspar-bearing veins also occur. Its folded structure and ptygmata reflect obvious migmatic effects.

The sillimanite of large quantity can be recognized also macroscopically. The mineral assemblages of the three rock types are as follows:

Mica gneiss:

quartz — potash-feldspar — oligoclase — biotite — muscovite — garnet;

quartz — potash-feldspar — oligoclase — biotite — (muscovite) — sillimanite — kyanite — garnet;

quartz — potash-feldspar — oligoclase — (biotite) — muscovite — staurolite — kyanite — garnet;
quartz — orthoclase — (oligoclase) — sillimanite.

Biotite-schist:

quartz — oligoclase — biotite — garnet;
quartz — oligoclase — biotite — sillimanite;
quartz — oligoclase — (potash-feldspar) — biotite — sillimanite.

Biotite-amphibole gneiss:

quartz — plagioclase — biotite — muscovite — garnet;
quartz — plagioclase — orthoclase — biotite — sillimanite — garnet;
quartz — plagioclase — hornblende — garnet — (epidote).

The mineral composition of the rocks based on thin section measurements are summarized in Table 1.

Table 1

Mineral	Mica gneiss	Sillimanitic vein	Biotite schist	Biotite-Amphibole gneiss
Quartz	38.3	35.0	12.7	32.5
Oligoclase	15.2	—	—	—
Plagioclase (andesine?)	—	28.0	13.9	7.2
Orthoclase	19.0	—	—	8.1
Biotite	13.7	—	59.6	5.0
Muscovite	5.8	—	1.1	0.5
Almandine	1.7	—	5.4	3.6
Kyanite	5.3	—	0.6	—
Staurolite	0.5	—	0.5	—
Epidote	—	—	—	1.2
Sillimanite	2.3	37.0	11.5	2.3
Hornblende	—	—	?	43.9
Number of investigations	10	1	7	8

The quartz grains are laminated, frequently spindle-shaped and of undulatory extinction. The grains are of jagged position according to the plane of schistosity and form strata. In the same plane secondary, mosaic-like quartz of slightly undulatory extinction, fractured potash feldspar and limpid plagioclase being discernable from quartz only in optical way, as well as a few twin-lamellated plagioclase can be found. The infrequent ptygma-like veins of unsettled run consist mostly of mosaic-like quartz.

In the biotite-amphibole gneiss veins of quartz + potash feldspar composition produced definitely by granitization are also found.

In the rocks of this type orthoclase is found in three modifications. In the mica-gneiss it occurs in form of fractured porphyroblasts together with quartz as primary mineral; in the quartz + potash feldspar veins intruded into the biotite-amphibole gneiss along the plane of schistosity. In the biotite-amphibole gneiss it occurs together with hornblende and quartz and consists of fine, unfractured grains and the fact cannot be decided whether these grains are primary or secondary ones, or secondarily regenerated original grains.

Out of the plagioclases some twin-lamellated grains of oligoclase composition and of 2 mm largest size can be recognized. Plagioclases occurring together with secondary quartz are found in large quantities and can be separated from quartz only by cleavage or on the basis of axial picture. In the biotite schist banks of several decimetres thickness and in the biotite-amphibole gneiss more basic and often twin-lamellated plagioclase of oligoclase-andesine composition is found. The feldspars of the sillimanite-rich vein are non-twin-lamellated plagioclases.

The quantity of muscovite becomes subordinate, moreover it is nearly absent in the biotite schist and biotite-amphibole gneiss, but it is the sole mica mineral in the contact of the sillimanite-rich vein.

In the biotite schist biotite becomes locally quantitatively determinant. The mica gneiss consists of two micas but biotite is predominant. Its oriented arrangement produces the lepidoblastic schistosity of the rock as contrasted to the mica gneiss of granoblastic texture. The formation of sillimanite from biotite can be well followed especially in the biotite schists.

Almandine occurs apparently in the same form in all the crystalline schists of the Görcsöny Ridge. The garnets of the biotite schist and biotite-amphibole gneiss, however, are of darker colour, their grain size is smaller but total quantity greater than those of the mica schist and mica gneiss. From the chemical composition of some prepared garnet grains it turned out that the darker varieties contain more iron.

In the crystalline schists of borehole Téseny No. 1. staurolite, kyanite, sillimanite and hornblende occur as new minerals.

Staurolite is irregularly fractured, strongly pleochroic (bright-yellow and colourless), and often of hexagonal cross-section (perpendicular of the *c* axis). In general, it is of columnar, prismatic appearance and the crystals of sometimes 0.5 cm lie in the plane of schistosity. Its fractures contain usually muscovite or grain is imbedded into muscovite. It occurs always with kyanite.

Kyanite is found in two kinds of occurrence, mostly in the mica gneiss. One form of occurrence is the columnar kyanite at distances of 15 to 25 cm above each other. These lie in nests resp. strips and the grains are fractured. The thickness of the strips does not exceed 3—4 cm and they do not occur

with sharp boundaries. In the planes of schistosity the light-blue kyanite columns are of desoriented position and often cross one another. The idiomorphic crystals amount to 3 mm in thickness and occasionally to 2 cm in length. The other form of occurrence is the kyanite in the contact of the sillimanite-rich vein in the mica gneiss similarly to that described above. The two kinds of kyanite of different origin, together with the columnar sillimanite and staurolite are definitely metamorphic facies indicators.



Fig. 4. Kyanite columns and staurolite from the mica gneiss of the borhole Téseny-1, from 156.2 m

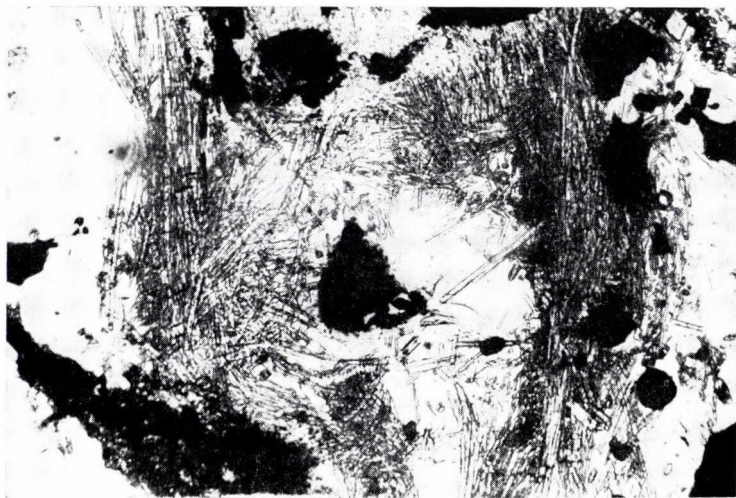


Fig. 5. Sillimanite-rich rock vein from the material of the borehole Téseny-1

Under microscope the sillimanite shows two kinds of appearance: in form of fine radiated fibrolites formed from biotite in the biotitic-muscovitic substance surrounding the kyanite-staurolite crystals, and in non-fibrous prismatic form in the yellowish-white vein of the mica gneiss.

Hornblende is the most frequent mineral of the biotite-amphibole gneiss. Its grass-green, fresh crystals elongated in the direction of a-axis and often of 1—2 mm size are oriented according to schistosity. Its characteristic cleavage and pleochroism can be well observed. In its greenish-blue — oil-green and light greenish-yellow pleochroism the blue shade as well as the axial angle somewhat less than the regular one (60°) suggest the somewhat higher alkali content of the mineral.

Epidote can be observed as small grains' heaps in certain phases of biotite-amphibole gneiss. It accompanies mainly the plagioclases and lies on their surface.

The wedge-shaped crystals of titanite growing sometimes up to 1 mm are the accessory components mostly of the biotite-amphibole gneiss. Magnetite and ilmenite are idiomorphic, their total quantity may occasionally amount to one per cent. Tourmaline occurs rarely in the mica gneiss, in form of oil-green columnar crystals. The rarely occurring chlorite is produced by biotite decomposition.

4. The Gyód Biotite-Amphibole Gneiss Member

This series was explored by the geophysical check boreholes of Gyód-3 and Gyód-4. The two boreholes were drilled in the northern and southern side of the serpentinized ultrabasite body explored by borehole Gyód-2. South of about 80 metres of the contact with the metamorphic rocks borehole Gyód-4 explored by continuous core sampling the progressively migmatitized, strongly biotite-rich rock containing large quantity of red feldspar (between 77.8 and 120.5 m). Borehole Gyód-3 located about 150 metres from the northern contact and also under Upper Pannonian strata explored the crystalline basement by continuous sampling and between 108.0 and 147.0 metres. This consists of less migmatitized biotite-amphibole gneiss, similarly to the rock material of borehole Gyód-4. The metamorphic orientation of the rock is practically vertical in both bores, and smaller schistose dip caused by folding can be observed only in a smaller section.

Within the metamorphic zone of the Göresöny Ridge the migmatization is most considerable in this area. The rock attains the grade of "stratified migmatite" and locally conceals the original features. The original rock proved to be two-mica gneiss and biotite schist in case of borehole Gyód-4 and garnetiferous biotite-amphibole gneiss and biotite schist in that of borehole Gyód-3. The material of borehole Gyód-4 is of more progressed stage. This migmatiza-

tion is correlated both in space and time by B. JANTSKY (1974) with the phenomena of the migmatite zone of the Mórág block.

The thin section study of the metamorphites of Gyód similar to those described in detail of the Tésény borehole but being of somewhat more "migmatic" character threw light upon two kinds of differences, i.e. the absence of sillimanite and kyanite, and the frequency of epidote and chlorite due to strongly regressive metamorphism. In several sections chlorite pseudomorphs after staurolite can also be found. In certain phases, biotite is considerably chloritized, especially where the rock is dissected by a young lithoclase system. Garnets, however, remained unchanged and have not been decomposed in the parts dissected by lithoclasses, either.

5. *The Pellérd-Keszü Actinolite Schist—Amphibolite Member*

In the northern foreground of the Görcsöny Ridge, east of the village Pellérd the basement sinks into great depths. The basement of the southern contacting strip of the narrow ENE—WSW-striking for deep of Upper Pannonian age consists practically of actinolite schist and amphibolite. This member was explored only by one borehole, i.e. the water prospecting well drilled between Pellérd and Keszü in 1958. Under the Pannonian and Sarmatian strata of 408.8 m thickness 13 metres of the basement was explored. The primary description lists the alternation of gneiss and amphibole schist but the core samples are lacking. A. BARABÁS (1959) studied some thin sections and described the rock as an actinolite schist.

It is assumed that this actinolite schist is the northward prolongation of the garnetiferous biotite-amphibole gneiss of migmatic character crossed by borehole Gyód-3. On the basis of thin section investigations the amphibole content of the garnetiferous biotite-amphibole gneisses of Tésény is of higher alkali-content than hornblende itself, in general. This higher alkali content of amphiboles was not observed in the borehole Gyód-3 and Gyód-4, probably due to the strong migmatic effect. Farther from the migmatic zone this higher alkali content becomes more and more evident and leads probably to the formation of actinolite schist.

With progressive metamorphism the migmatic phenomena are gradually increasing. In spite of the absolutely uncertain actinolite schist of Gyód, in the southern foreground of Pécs strongly granitized rocks are found (boreholes Szigeti-legelő-1, Ujmecekalja-1, etc.).

In the Barrow type Crystalline Schist Formation of Görcsöny a syngenetic eclogite (L. BARANYAI-RAVASZ, 1969) is intercalated probably similarly to that known from Pohorie (A. HINTERLECHNER-RAVNIK, 1972). On the basis of analogies the slightly serpentized pyroxenite neck explored by borehole Gyód-2 can be considered to be of Silurian age. These, however, do not dis-

turb to a considerable extent the zonal structure and progressive character of the Formation.

II. Pre-metamorphic Rock-type

The zones of the Dalradian series were produced by the metamorphism of pelitic sediments. The resistant fragments, especially in the lower-grade zones may be preserved where the original sediment had been psammitic.

The mineral association possibilities of the Barrow type metamorphism (Table 2) relate to the fact that the mineral associations of the Göresöny Ridge, i.e. from the almandine zone to the sillimanite zone are the same than those of the Barrow type series of pelitic origin, except the orthoclase and quartz relics. The biotite-muscovite schists, biotite schist and mica gneisses of Kisdér are of psammitic, while the amphibole gneisses of Tésény and Gyód are of igneous (basic) origin.

Table 2

Zonal index	Rock composition		
	pelitic	carbonaceous	basic (basaltic)
Chlorite	muscovite-chlorite-quartz	calcite-mica	chlorite-albite-epidote-titanite (calcite)
Biotite	biotite-muscovite-chlorite-quartz	quartz	hornblende plagioclase (with epidote, almandine or biotite)
Almandine	biotite-muscovite-almandine-quartz	garnet-epidote-hornblende	
Staurolite	biotite-muscovite-staurolite-almandine-quartz	garnet-anorthite-hornblende	
Kyanite	biotite-muscovite-almandine-kyanite-quartz		
Sillimanite	biotite-muscovite-almandine-sillimanite-quartz-(orthoclase)	garnet-anorthite-diopside	

III. Comparison with the Metamorphites of Papuk-Krndija

The crystalline basement of the Drava Basin is directly continued in the surficial higher-grade crystalline schists of the Papuk Mountains and Krndija. The comparison of SE-Transdanubia and of the Papuk-Krndija series is an evidence of the nearly identical paleohistory of the two areas as well as of the Barrow type metamorphism of them.

According to the Table 3 the regional metamorphic series of the Görcsöny Ridge—Drava Basin is similar to the Scotch Highland type, rather than the Papuk-Krndija series. In the former one all the index minerals occur (and major part of them can be arranged in zones) while in the latter the kyanite is lacking and sillimanite occurs not due to the high p-t values but just because of granitization. Thus it is problematic whether it is justifiable to extend the zones of biotite-muscovite gneisses and biotite gneisses up to the sillimanite zone. In the classical succession namely the sillimanite isograd follows that of kyanite and is not influenced by granitization. Nevertheless, in the Görcsöny Ridge granitization had been also active but sillimanite can be determined independently of this, thus the sillimanite zone exists independently of granitization.

Table 3

Dalradian series	Papuk-Krndija series	"Early Paleozoic" of SE-Transdanubia
Chlorite zone	Silurian quartz-schist muscovite-chlorite schist	Silurian quartz-schist (Szaltnak) muscovite-chlorite schist (Okorág-2; Sellye-1)
Biotite zone	muscovite-biotite schist biotite-amphibole gneiss	muscovite-biotite gneiss (Okorág-1)
Almandine zone	biotite-muscovite gneiss	muscovite-biotite schist (Kisdér-1)
Staurolite zone		mica-gneiss (Téseny-1) biotite-amphibole gneiss (Téseny— Gyód)
Kyanite zone		
Sillimanite zone	biotite-gneiss	biotite-schists

The rocks of the same name do not occur in the same zone in the Papuk and SE-Transdanubian series, since the Yugoslavian series consists of metamorphites originated only from pelitic source rock. In this case gneiss denotes a rock the feldspar of which developed during metamorphism while in the Hungarian series, just become of the pelitic-psammitic source rock the orthoclase of the gneisses is relict except the sillimanite zone, but it is classified as gneiss owing to the occurrence of orthoclase in large quantities (epigneiss problem).

Another difference is the fact that in the Papuk Mountains the muscovite-biotite schists form only a transitional zone highly above the almandine isograd, while in the Görcsöny Ridge the rocks are assigned to the almandine zone. The different metamorphic character of the amphibole-bearing rocks can be explained by the different ages of formation of the igneous source rock. These were in different levels in the original "rock column".

P. RAFFAELLI (1965) has drawn also the map of the metamorphic zones of Ravna Gora. In this map the isogrades are of E-W direction and the metamorphic grade increases from the south northwards. The kyanite isograde, however, is lacking here. The occurrence of sillimanite can be attributed also to granitization since without granitization the metamorphism of the Papuk Mountains would be completed by the staurolite zone.

The Görcsöny Ridge and its southwestern environs resp. their crystalline basement form a more complete Barrow type series. In the picture compiled on the basis of individual bores none of the Barrow zone are lacking although it is difficult to draw the boundaries exactly due to the lack of boreholes of required number.

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МЕТАМОРФИЗМ ТИПА «БАРРОУ» КРИСТАЛЛИЧЕСКОГО ФУНДАМЕНТА ГОР ЮГО-ВОСТОКА ЗАДУНАЙСКОГО КРАЯ

Т. СЕДЕРКЕНЬИ

Резюме

Часть разведочных скважин, пробуренных в южной части области Баранья, под паннонско-миоценовыми слоями вскрыла большей частью мезозонально кристаллические сланцы, которые согласно данных подобных славонских образований непосредственно связаны с метаморфической массой гор Папук-Кридня. От реки Драва до гор Мечек масса кристаллических сланцев, перекрытая более молодыми образованиями, т. е. так называемая Гёрчёнская формация кристаллических сланцев типа «барроу» юго-восточного простиранья, образует прогрессивную серию, характеризуемую изоградами, со степенью метаморфизма, увеличивающейся с юго-запада на северо-восток, и заканчивающуюся гранитизацией. Метаморфический характер прогрессивной серии хорошо сопоставляется с классической серией Далрадиан и, помимо основания южного передового прогиба, является результатом единственного значительного процесса деформации. Возраст деформации предположительно досилурский, по Б. Янчки — позднебайкальский. Основание южного передового прогиба гор Мечек образует пояс милонитизированных древнепалеозойских известково-филлитовых зеленых сланцев шириной 1,5—2 км, который вследствие ранневарисийских транскуррентных разрывных нарушений вклинился в древнюю метаморфитогранитную массу. Тектонический пояс с южной стороны соприкасается с претерпевшим ретроградно преобразование краем метаморфизованной серии типа «барроу».

CONTRIBUTION TO THE PETROGRAPHY, PETROCHEMISTRY AND PETROGENESIS OF METABASITES FROM ASWAN AREA, EGYPT

By

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Eight samples were investigated, ranging from ferromagnesian rich granite, quartz diorite to a more basic dioritic composition. The most basic dioritic samples are believed to have suffered early gravitative magmatic differentiation. These are enriched in some trace elements like V, Cr, Co and Ni. The most acidic ones show the progressive formation of porphyroblasts of feldspars that may grow up to one centimetre long. These rocks contain high values for Ba, Sr, Y, La and Sm except for the more basic samples where Ba, Rb, U and Th contents are even lower than that of the average basaltic rocks. However, all the analyzed samples are depleted in Zr.

From petrographic, petrochemical and geochemical points of view, it could be noted that the dark hornblendic rocks of the basement in the vicinity of Aswan derive initially from old geosynclinal igneous basaltic (doleritic and gabbroic) intrusions and not from impure calc-magnesium sediments, as believed former.

Introduction

Several authors worked on the geology of the basement rocks of Aswan, among others BALL (1907), RITTMANN (1953), GINDY (1952, 1954, 1956 and 1957), and HIGAZY and WASFY (1956).

Dark hornblendic rocks are very frequent both in the country rocks and as inclusions in the granite of Aswan. They particularly increase in the eastern borders of the map-area of the environs of Aswan town (Fig. 1).

GINDY (1952, 1954 and 1956) considered them to be originally of simatic igneous intrusions in the country rocks (of the eugeocyclinal stage). HIGAZY and WASFY (1956) considered them as metamorphosed and metasomatized marly and impure calcareous pelitic sediments. GINDY (1957) has disputed this last suggestion.

Despite the wide variety of rock types displayed by these dark hornblendic rocks and their migmatitic and feldspathized derivatives, only eight samples were selected for the present study and their locations are illustrated in Fig. 1. These are as follows:

Samples No. NA-197 and NA-198: from the dark grey porphyroblastic large migmatitic granodioritic patches apparently enclosed in the coarse Aswan Granite.

Samples No. NA-518 and NA-519: These are diorite—like rocks enclosed in the coarse Aswan-Granite which is cut by the new granite on Sehel Island. Close to them the so-called Cephren-Diorite occur which is of more feldspathic and more gneissic character than the samples selected for the present analyses.

Samples No. NA-300: Is dark diorite-like relatively less migmatized part in a large granodioritic patch similar to sample NA-197.

Samples No. NA-90, NA-91 and NA-92: These derive from the dark layer apparently differentiated from a large original gabbroid intrusion in the eastern part of the map-area.

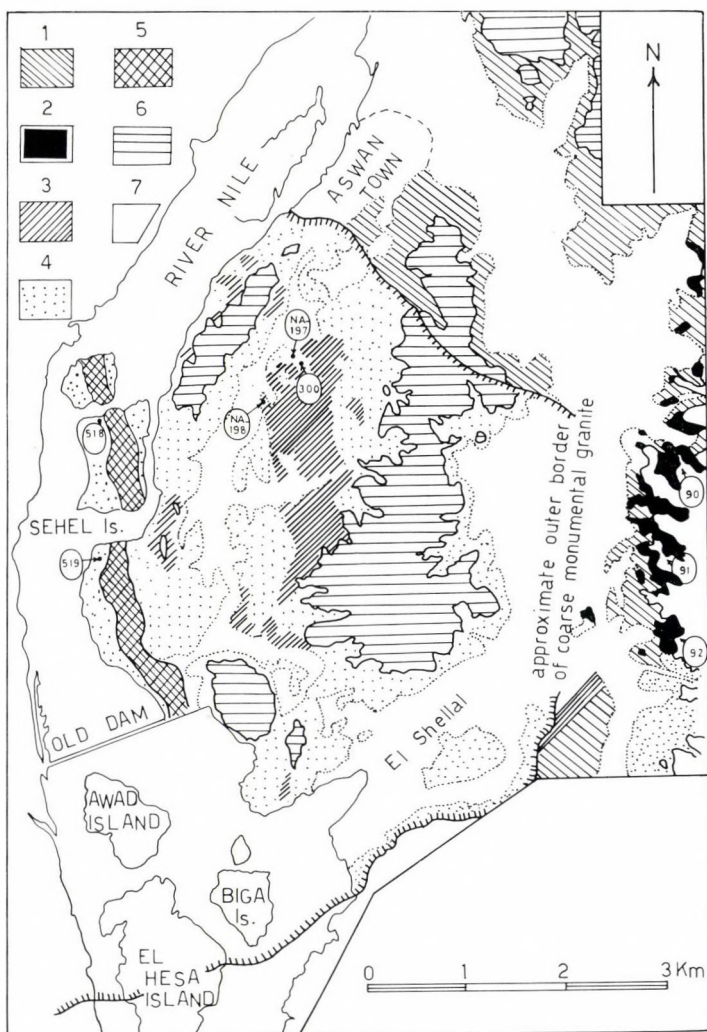


Fig. 1. 1. Semi-pelitic and pelitic biotite schist and fine grained biotite gneiss; 2. Metadolerites; 3. Granodiorite-like rock with large feldspar porphyroblasts; 4. Coarse Aswan Granite; 5. New Aswan Granite; 6. Nubian Sandstone; 7. Alluvial deposits

GINDY (1957) believed that some of the original gabbroic intrusions were rhythmically differentiated into lighter-coloured anorthositic and dark-coloured hornblendic layers.

Petrography

The studied samples are quite different in their petrographic characteristics and show some gradations in their mineralogical composition.

Sample No. NA-197 and NA-198 are medium to coarse grained dark mesocratic granodiorite-like rocks with different stages of formation of poikiloblasts of potash and plagioclase feldspars enclosing relics of the biotite of the matrix. These large feldspar poikiloblasts may grow up to two or three centimeters long. These are usually light-coloured and are surrounded by outer reddish zone. Sometimes these are less uniform and show irregular outlines. Under microscope these rocks are composed of zonal oligoclase, microcline, biotite and hornblende. Myrmekite is present as outer replacement of potash feldspar. It is worthy to mention that the hornblende in the granodiorites and diorites (No. NA-197, NA-518 and NA-300) seems to be of the same type and has the following pleochroic formula: α -brownish green, β -yellowish brown to brownish green, and γ -dark green. Opaques, apatite, sphene and zircon are the common accessories. Apatite occurs mainly as inclusions in feldspar and biotite. The modal analysis of these samples are given in Table 1, while the plot of their modal composition is shown in Fig. 2. According to STRECKEISEN's nomenclature and classification (1967), based on their light-coloured constituents, these samples plot in the field of granites.

Table 1
Modal composition of the studied samples

Constituent minerals	NA-197	NA-198	NA-518	NA-519	NA-300	NA-90	NA-91	NA-92
Quartz	19.5	18.3	15.3	16.5	31.0	—	—	—
K-feldspar	27.3	29.2	—	—	1.0	—	—	—
Plagioclase	29.7	31.6	45.9	46.2	33.0	57.4	56.7	55.7
Biotite	17.4	16.1	12.5	11.1	19.8	—	—	—
Hornblende	3.9	2.5	17.6	18.4	8.7	41.2	42.1	42.9
Opaques	1.2	1.2	2.5	2.7	2.6	0.2	0.3	0.4
Apatite	1.0	1.1	4.4	3.6	3.7	0.2	0.2	0.3
Sphene	X	X	0.8	0.7	X	0.9	0.6	0.6
Zircon	X	X	1.0	0.8	0.2	0.1	0.1	0.1
Allanite	—	—	—	—	—	—	—	—

X — Indicating that the mineral is present but not counted

Samples No. NA-518 and NA-519 are diorite rocks occurring as inclusions in the old coarse grained granite which is cut by the new granite on the northern part of Sehel Island. Close to them the so-called Cephren Diorite occurs which has a maximum feldspathic and definitely more gneissic character than the samples selected for analysis. In thin section, these are composed of plagioclase

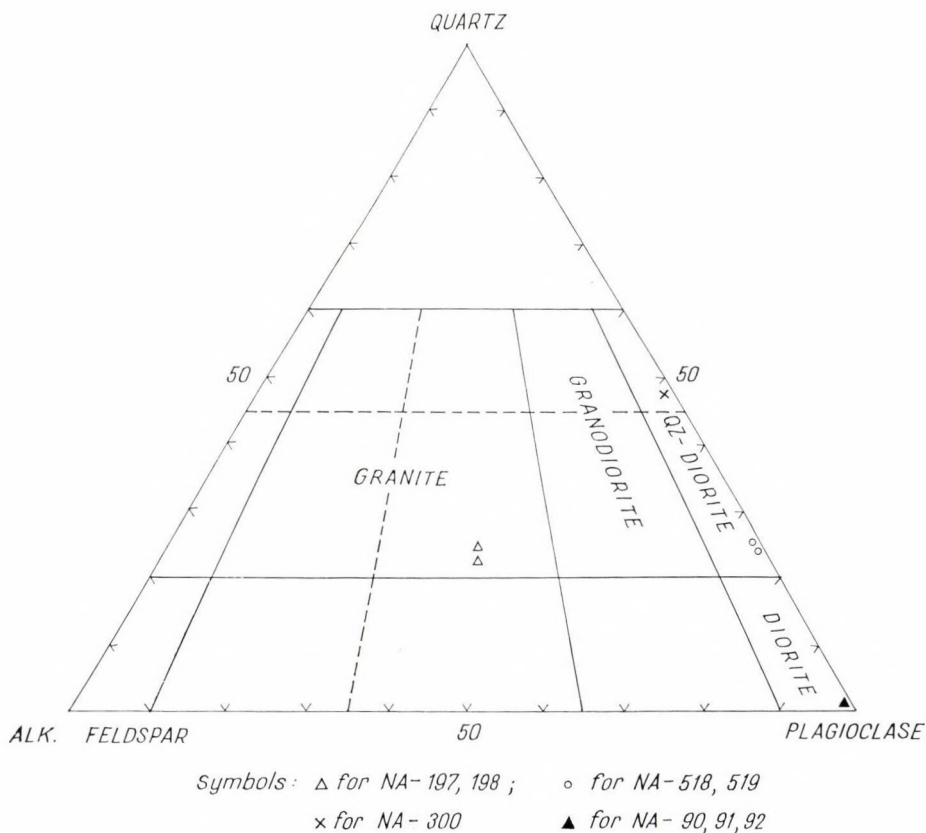


Fig. 2

class (andesine), quartz, hornblende and biotite. Apatite, opaques, zircon and sphene are found as accessory minerals. The plagioclase is zoned with saussuritized cores of more anorthitic composition. Quartz shows undulatory extinction suggesting post-crystallization stresses. According to STRECKEISEN's classification and nomenclature (1967), they plot in the field of quartz-diorites.

Sample No. NA-300 of Aswan diorite is a medium grained melanocratic quartz diorite. It contains more quartz than the preceding ones but is of darker colour suggesting higher content of ferro-magnesian minerals. In this

section, this rock is composed of plagioclase (zoned acid andesine), quartz showing undulatory extinction, biotite and hornblende. No good foliation is apparent in the studied sample, but incipient stages of formation of potash feldspar is revealed by the appearance and growth of minor, small interstitial anhedral microcline crystals. Both biotite and hornblende seem to be adjacent to each other as if in perfect equilibrium state. Apatite is a common accessory mineral and is present as inclusions in biotite and plagioclase and as intergranular crystals. Opaques, sphene and zircon are other accessories.

Samples No. NA-90, NA-91 and NA-92 are medium to fine grained melanocratic diorites composed of plagioclase and dark ferromagnesian minerals. In thin section these show more or less equal amounts of andesine and hornblende which tend to be segregated into different bands. The hornblende crystals are subparallel forming a crude gneissic structure. The hornblende bands could form after original pyroxene crystals which were differentiated gravitatively in the initial simatic eugeosynclinal intrusion. Hornblende is moderately pleochroic in the following manner, α -light-green, β -green and γ -pale bluish-green. Small anhedral crystals of sphene, apatite and minor opaques are accessories.

The plot of ternary diagram in STRECKEISEN's classification and nomenclature (1967) falls in the field of diorite and gabbro (based only on the light-coloured constituents). The mode of the lighter "anorthositic" band opposite to this dark band was given by GINDY (1957, Table 1, p. 681) consisting of acid andesine 85.36 and hornblende 14.64 per cent, with minor amounts of sphene and opaques. When compared to each other, the quantity of hornblende in the dark band is about three times higher than in the adjacent light band. It is worthy of note that the accessory opaques are sparse and too small in size as compared to those of the other studied dioritic varieties.

Petrochemistry

Complete chemical analyses of eight hornblende rocks are given in Table 2. Rock powders were chemically analyzed using the method described by RILEY (1958) for the determination of Si, Ti, Fe (as total iron), Na, K and P. Aluminium and ferrous iron were determined following the method described by SHAPIRO and BRANNOCK (1962). Mn, Mg and Ca were determined separately by a spectrochemical method. Rock analyses were carried out in duplicate with good agreement.

The samples range in composition from the "ferromagnesian-rich granite", quartz-diorite to diorite. Beside the eight analyses of this paper, three previously analyzed samples (G, H and I) are also presented in Table 2. These three analyses are quoted from HIGAZY and WASFY (1956) and are incorporated in some of the respective diagrams. Analyses of the coarse-grained por-

Table 2
Complete chemical analysis (Wt%) for the studied metabasites

Oxides	NA-197	NA-198	NA-518	NA-519	NA-300	NA-90	NA-91	NA-92	G	H	I
SiO ₂	60.80	59.21	54.95	54.40	58.84	52.49	53.82	51.95	54.19	60.56	46.04
TiO ₂	2.00	2.41	2.70	2.95	2.60	0.35	0.27	0.28	2.49	1.58	2.10
Al ₂ O ₃	14.45	16.25	15.65	15.84	16.53	18.56	18.82	17.98	14.30	14.15	22.36
Fe ₂ O ₃	1.10	1.11	1.66	1.86	1.23	0.30	0.27	0.34	3.22	2.23	0.99
FeO	7.57	5.98	8.67	8.91	6.80	7.67	7.80	8.54	8.26	6.24	7.47
MnO	0.16	0.18	0.23	0.15	0.22	0.12	0.11	0.14	0.20	0.16	0.16
MgO	3.25	3.64	3.25	3.14	3.22	3.70	4.35	4.22	2.91	1.78	4.94
CaO	3.43	4.56	5.20	5.68	4.15	7.80	8.15	7.98	5.93	4.33	11.64
Na ₂ O	2.40	1.95	2.80	2.42	2.05	3.00	2.78	2.85	3.51	3.83	2.39
K ₂ O	3.45	3.64	3.38	3.20	3.45	0.50	0.42	0.36	2.45	3.33	0.62
P ₂ O ₅	0.70	0.72	1.90	1.75	0.80	1.85	1.56	2.05	1.38	0.85	0.09
H ₂ O ⁺	0.60	0.46	0.44	0.34	0.52	2.12	2.00	2.51	0.95	0.64	0.79
H ₂ O ⁻	0.10	0.38	0.14	0.12	0.45	0.28	0.22	0.32	0.12	0.09	0.13
CO ₂	—	—	—	—	—	—	—	—	—	0.29	0.27
Total	100.01	100.49	100.97	100.76	100.84	98.74	99.57	99.52	99.91	100.06	99.99

G — Coarse-grained porphyritic granodiorite of Gabal Tegog (HIGAZY and WASFY, 1956).

H — Coarse-grained porphyritic granodiorite, north of the English cemetery, western side of the road to Aswan dam (HIGAZY and WASFY, 1956).

I — Diorite from the eastern portion of the district (HIGAZY and WASFY, 1956).

Table 3

C.I.P.W. norms of the analyzed metabasites

Norm	NA-197	NA-198	NA-518	NA-519	NA-300	NA-90	NA-91	NA-92
Q	18.82	17.81	10.24	11.51	18.82	9.98	10.21	9.50
Or	20.41	21.52	20.02	18.90	20.41	2.95	2.50	2.11
Ab	20.28	16.53	23.68	20.41	17.34	25.36	23.51	24.11
An	12.54	17.90	13.54	16.72	15.46	26.69	30.32	26.32
C	2.16	2.51	2.40	2.21	3.73	3.29	2.70	3.21
Di	—	—	—	—	—	—	—	—
Hy	18.10	15.52	18.62	18.11	15.56	22.70	24.73	25.80
Mt	1.60	1.61	2.41	2.70	1.79	0.44	0.41	0.51
Hm	—	—	—	—	—	—	—	—
Il	3.80	4.61	5.14	5.62	4.94	0.67	0.55	0.52
Ap	1.65	1.70	4.50	4.11	1.88	4.36	3.73	4.80

Table 4

NIGGLI values of the analyzed metabasites

NIGGLI	NA-197	NA-198	NA-518	NA-519	NA-300	NA-90	NA-91	NA-92
al	29.46	31.70	27.73	27.90	32.45	31.40	30.71	29.42
fm	42.11	38.01	40.79	40.82	38.65	35.29	36.90	38.51
c	12.75	16.32	16.79	18.21	14.87	24.04	24.22	33.80
alk	15.67	14.01	14.68	13.11	14.03	9.27	8.21	8.31
+qz	+48.06	+41.00	+6.82	+10.32	+40.20	+14.78	+16.21	+11.50
ti	5.19	6.02	6.10	6.60	6.50	0.75	0.60	0.61
p	1.01	1.02	2.42	2.21	1.12	2.21	1.81	2.42
k	0.48	0.63	0.44	0.53	0.53	0.10	0.11	0.12
mg	0.40	0.54	0.36	0.31	0.41	0.45	0.52	0.50

phyritic granodiorite of HIGAZY and WASFY (samples No. G and H) are very similar to the present analysis of the "ferromagnesian-rich granite" (samples No. NA-197 and NA-198).

It had been stated that samples No. NA-90, NA-91 and NA-92 have some peculiar features, i.e. to be the basic component of a differentiated layer in an old initial gabbroic intrusion. In spite of their more basic composition among the granodiorites and quartz diorites, they have lower Ti and Mn contents than the other hornblendic rocks. Other values such as those of Ca and Mg are in accordance with their more basic composition.

The analyzed hornblendic rocks are generally comparable with DALI's

Table 5
Trace element contents (ppm) in the analyzed metabasites

Element	NA-197	NA-198	NA-518	NA-519	NA-300	NA-90	NA-91	NA-92
B	7	5	4	3	7	>15	>15	>15
Ba	1665	1372	640	550	1169	45	32	23
Be	>10	7	8	5	4	<1	<1	<1
Co	12	17	21	29	14	74	83	64
Cr	2	4	3	5	2	77	90	53
Cu	20	24	20	16	28	6	4	9
La	135	79	134	162	42	20	17	29
Ni	17	12	11	14	8	39	51	32
Sc	24	21	30	27	19	16	12	8
Sr	606	430	585	470	372	184	141	160
V	57	72	106	118	86	230	190	175
Y	67	51	80	92	62	34	47	28
Yb	7	9	10	13	6	<1	2	<1
Zr	10	12	10	12	10	18	27	14
Cs*	1	1	0.7	0.4	0.9	1.1	1	1.3
Hf*	11.8	9	9.7	7.4	7.6	1.0	<1	<1
Rb*	65	61	47	37	54	N.D.	N.D.	N.D.
Sm*	16.9	12	14	11	8	1.2	2	1.6
Ta*	3.5	2	3	2.6	2	0.22	1	0.5
Th*	12.2	10	10.5	9.8	8.6	N.D.	N.D.	N.D.
U*	2.3	2	2	1.3	1.8	0.04	0.2	0.1

* Values determined by neutron activation technique.
N.D. = not detected.

average of diorites (DALY, 1933). These are relatively enriched in TiO_2 , total Fe_2O_3 , K_2O and have lower Na_2O than DALY's average of diorites especially for samples No. NA-518, NA-519 and NA-300.

The C.I.P.W. norms of the migmatitic hornblendic rocks are characterized by the appearance of quartz (Table 3). Actually, these all have quartz in their mode except Nos. NA-90, NA-91 and NA-92. This is best explained by the presence of some later diffuse migmatizing quartzo-feldspathic permeations in the chemically analyzed portion of these samples. These are relatively more enriched in anorthite, hypersthene, ilmenite and apatite than the other granitic samples. It must be remarked, however, that these always contain orthoclase in their norm inspite of the scanty of potash feldspar in their modal composition, except Nos. NA-197 and NA-198.

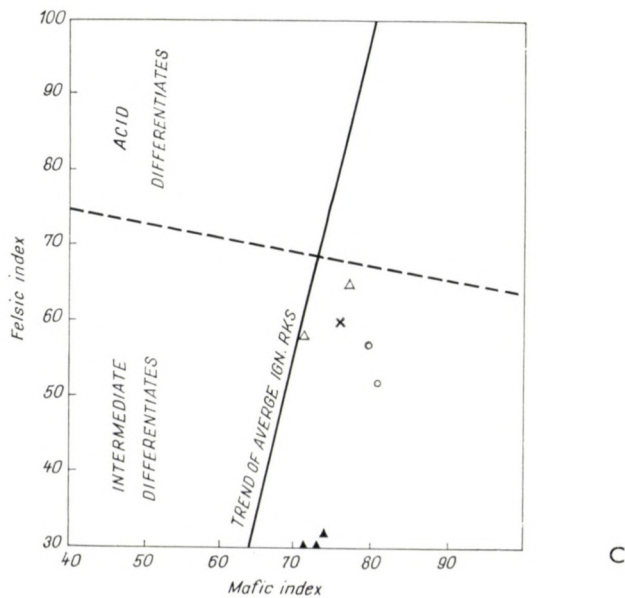
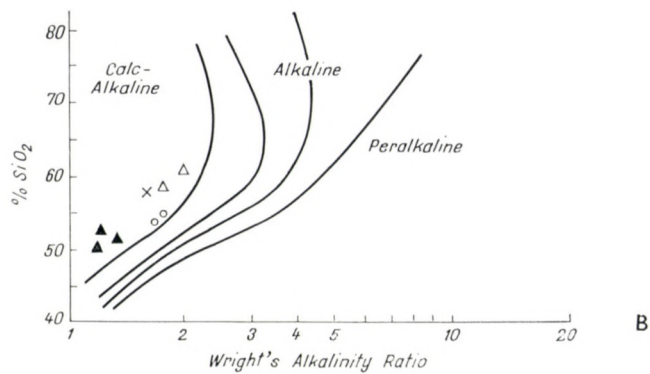
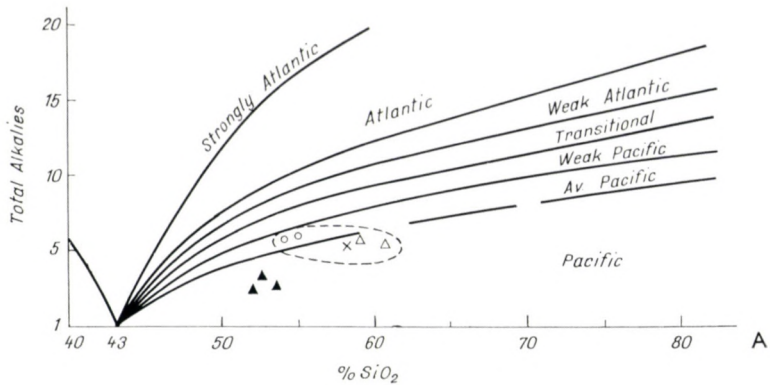


Fig. 3

NIGGLI values given in Table 4 apparently reflect some of the observed mineralogical variations in these hornblendic rocks. Fig. 3-A clearly shows that the samples are of Pacific character according to RITTMANN (1957). With respect to WRIGHT's alkalinity ratio (1969), the Aswan hornblendic rocks fall in the calc-alkaline field as seen from Fig. 3-B. Regarding their mafic

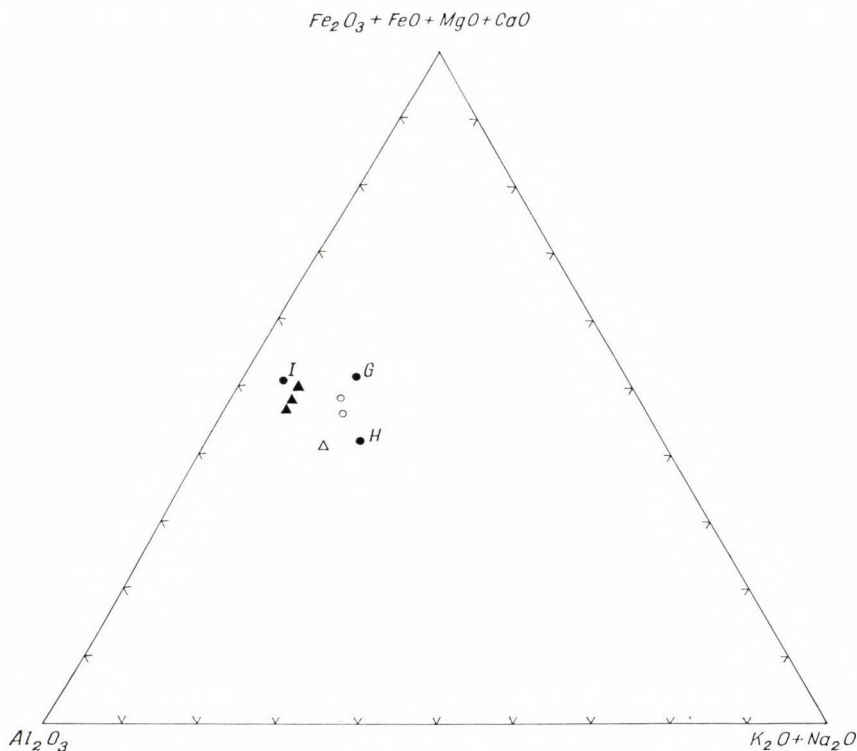


Fig. 4

and felsic "SIMPSON (1954)" indices they fall in the region of intermediate differentiate, Fig. 3-C with a trend of differentiation towards the most acidic, migmatized samples (NA-197 and NA-198).

As for the ternary ($\text{Fe}_2\text{O}_3 + \text{FeO} + \text{Mg} + \text{CaO}$); Al_2O_3 ; total alkalis diagram (Fig. 4), it is to be noted that the most basic samples in the hornblendic rocks (NA-90, NA-91 and NA-92) are comparable with the analyses of HIGAZY and WASFY for Aswan diorites (sample No. I in this diagram), while the rest of the hornblendic samples are shifted towards the coarse-grained porphyritic granodiorites (samples No. G and H) of the same authors.

Trace elements:

Trace elements were determined quantitatively by two different methods:

1. Spectrochemical method for determining B, Ba, Be, Co, Cr, Cu, La, Ni, Sc, Sr, V, Y, Yb and Zr.

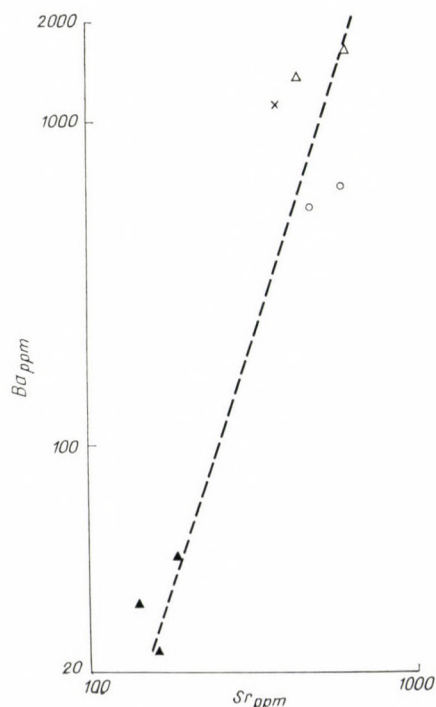
2. Neutron activation technique for determining Cs, Hf, Rb, Sm, Ta, Th and U.

In case of the first method a Q-24 Carl Zeiss medium quartz spectrograph supplied with an intermittent D.C. source was used. The generating power was a Carl Zeiss ABR3 source with a maximum voltage of 28 kV at 220 V. Operating conditions for arc and technique of work were usual as described elsewhere (e.g. ANWAR et al., 1971).

For the second method, the neutron activation scheme used in this work is based on the different irradiation (A.C.) carried out at a thermal neutron flux of 1.5×10^{13} h/cm² · sec involving the independent 0.1 g portions of the sample, as described by ANDERS, 1969 and BRUNFELT et al., 1969 and 1974.

Trace element contents in the hornblendic rocks of Aswan, Table 5 cannot be compared with any proposed average for normal igneous rocks, since these have suffered a series of complex geological processes including possible simatic intrusion, differentiation, deformation metamorphism, metasomatism and migmatization.

Geochemical evidence completely agrees with the petrographic evidence which indicates the increase of the microcline content in passing from samples Nos. NA-90, NA-91 and NA-92 (the least migmatized samples) containing no modal potash feldspar and successively to the samples Nos. NA-518, NA-519 and NA-300 where incipient stages of microcline formation are observed;



and finally to samples Nos. NA-197, NA-198 characterized by the highest microcline and Sr contents. Sr decreases with the increase of Ca, reflecting the increase of Sr/Ca ratio as migmatization and feldspathization proceeds. Thus it can be seen that the progressive migmatization and feldspathization bears similar effects as magmatic differentiation because the Sr/Ca ratio increases as differentiation proceeds in many calc-alkali igneous rock series (NOCKOLDS and ALLEN, 1953). Both Ba and Sr show a sympathetic increase with each other as well as with progressive migmatization and feldspathization of the hornblendic rocks (Fig. 5).

V, Cr, Co and Ni are enriched in the most basic samples Nos. NA-90, NA-91 and NA-92. It may be noticed that Cr content in migmatized samples is less than that of the Aswan Granite, while Co and Ni values are higher than any of the fresh Aswan basement rocks.

Y, La and Sm values increase with the increasing trend of feldspathization in the samples. However, the basic samples are much depleted in La and Sm together with Ta, Th, U, Hf and Rb.

In general, the rock varieties mentioned above agree roughly with those encountered in the usual differentiation of igneous rocks except for the lower Cr content in the feldspathized ferromagnesian-rich granite and quartz-diorite.

Speculations on the Origin of Metabasites

GINDY (1954) considered the country rocks of Aswan to have initially consisted of a geosynclinal pile made up chiefly of two kinds of rocks, semipelites (greywackes) and (subgreywackes), doleritic and gabbroic igneous intrusions. These were later metamorphized, metasomatized and migmatized to take the different varieties forming the country rocks, the migmatized aureole around the coarse granite and of its inclusions. The semipelitic and metasedimentary members were much more altered and granitized than the metagabbros and metadolerites.

HIGAZY and WASFY (1956) considered all the hornblendic rocks of Aswan and their derivatives to have been initially of sedimentary origin (impure calcareous sediments). GINDY (1957), has disputed the sedimentary origin of the hornblendic rocks and showed that some of them are actually branching discordant intrusions into the meta-semipelites and the core of the largest, least migmatized hornblendic bodies retains a typical doleritic to gabbroic fabric.

Recently MISRA (1971) has reviewed the different geochemical parameters suggested earlier to prove the origin of the metabasites whether they were of sedimentary or igneous origin. MISRA suggested the geochemical relations between TiO_2 and OSBORN factor-F (1959, 1962), and between

TiO_2 and MnO as forming the best available geochemical tools to prove the igneous or sedimentary origin of the metabasites.

The eight analyzed samples are shown according to MISRA's diagram. The three analyses given by HICAZY and WASFY, Nos. G, H and I are also plotted. In Fig. 6-A the data fall in the field of ortho-metabasites except for the differentiated dark hornblende samples and the coarse-grained migma-

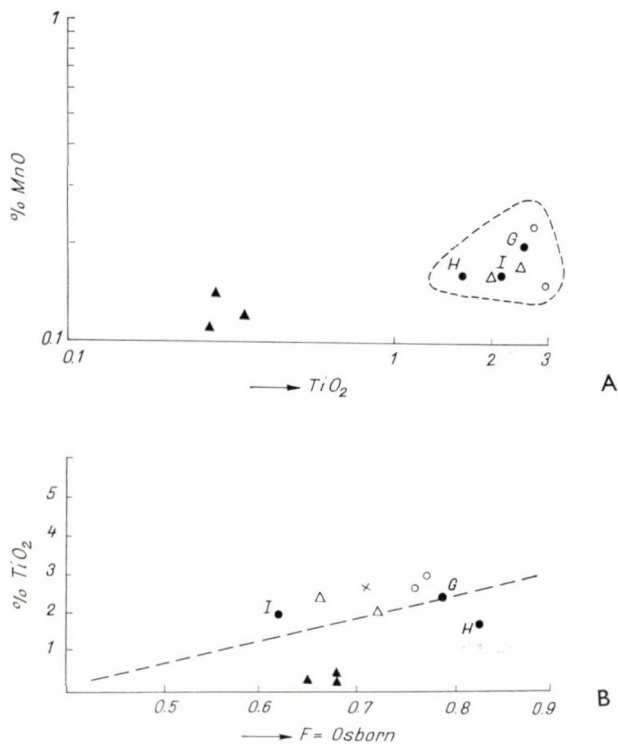


Fig. 6

tite porphyritic granodiorite No. H. In Fig. 6-B all samples except Nos. NA-90, NA-91 and NA-92, fall in the ortho-metabasite field. Actually, these samples were supposed by GINDY to represent one of the dark differentiated layers of old rhythmically banded structures in the initial gabbroic intrusion.

TAYLOR (1965) quoted the distinction of FAUST et al. (1956), that serpentines of ultrabasic igneous rocks contain more than 100 ppm Cr, while serpentine derived from the metamorphism of calcareous and dolomitic sediments "had little or no detectable Cr". TAYLOR (1965) has also quoted EVANS and LEAKE (1960) findings that in stripped amphibolites of basic igneous percentage, the Ni (and Cr) content exceeded by a factor of two its maximum content in the amphibolites derived from shales or dolerites. The Ni as well

as Cr contents of samples Nos NA-90, NA-91 and NA-92 are more than double of the other hornblendic samples. On the other hand, the enrichment of such samples in B could be due to their high plagioclase content which is much higher than those of the remaining samples.

According to RANKAMA and SAHAMA (1952), Ti belongs to the Fe family and shows a distinct tendency to become separated from the crystallizing magma at an earlier stage, whereby Ti becomes enriched in the early crystallates of ilmenite and titanomagnetite. Inspection of thin sections of the basic hornblendic samples shows that opaques are very rare relative to the other hornblendic rocks. The only Ti-mineral observed was very rare and tiny sphene. The absence of these two opaque minerals from the analyzed dark band would suggest that within the magma chamber ilmenite and titaniferous magnetite might have originally gravitatively differentiated and sunk in an earlier stage at the bottom of the magma chamber just before the rhythmic banded structure occurred.

It follows that somewhere in the Aswan basement one would expect to discover in the future, some dark melanocratic metabasites (hornblendites) abnormally rich in ilmenite and titanomagnetite or their other metamorphic derivatives.

Conclusions

The metabasites in the vicinity of Aswan range from ferro-magnesian rich granite, quartz diorite to a more basic dioritic composition. The most basic dioritic samples are believed to have suffered early gravitative igneous differentiation. They are enriched in some trace elements like V, Cr, Co and Ni. The most acidic rocks contain high values for Ba, Sr, Y, La and Sm.

From petrographic, petrochemical and geochemical discussions it could be noted that the dark hornblendic rocks derived initially from old geosynclinal igneous basaltic (doleritic and gabbroic) intrusions and not of impure calc-magnesium sediments. Some of these initial intrusions were even differentiated and subsequently rhythmically layered as suggested by the impoverishment in Cr and Ti in only certain dark bands of these rhythmically banded differentiates.

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К ПЕТРОГРАФИИ, ПЕТРОХИМИИ И ПЕТРОГЕНЕЗИСУ МЕТАБАЗИТОВ РАЙОНА АСУАНА В ЕГИПТЕ

С. О. КХАЛИЛ—А. Р. ГИНДИ—А. И. АРСЛАН

Резюме

Изучено было восемь образцов, изменявшихся по составу от гранитов, богатых ферромагнемом, через кварцевые диориты вплоть до пород диоритового, более основного состава. Самые основные по составу образцы диорита были подвержены, по мнению авторов, гравитативной магматической дифференциации ранней фазы. Эти образцы обогащены некоторыми рассеянными элементами, а именно — V, Cr, Co и Ni. Самые по составу образцы свидетельствуют о постепенном формировании порфиробластов полевых шпатов, которые могут достигнуть 1 см в длине. Эти породы содержат большие концентрации Ba, Sr, Y, La и Sm, за исключением более основных по составу образцов, где содержание Ba, Rb, U и Th еще ниже чем средние содержания их в базальтоидных породах. Однако, все анализированные образцы почти или полностью лишены Zr.

С петрографической, петрохимической и геохимической точек зрения можно было отметить, что темные амфиболовые породы фундамента в районе г. Асуан происходят от древних геосинклинальных, магматических базальтовых (долериты и габбро) интрузивов, а не от нечистых кальциево-магниевых осадков, как это предполагалось раньше.

CHEMICAL COMPOSITION AND GEOCHEMICAL EVALUATION OF THE NUBIAN SANDSTONES OF KALABSHA AREA, EGYPT

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Ten representative samples covering all the possible varieties of Nubian Sandstones in Kalabsha area (Egypt) were chemically analyzed by author. The chemical analyses as well as the petrographic modal analyses allowed the classification of these sandstones as protoquartzites (sublithic arenites). Also, it was revealed that these sandstones derived most probably from preexisting underlying basement rocks.

Introduction

Nubian sandstone is a term given to the brownish, buff-brown and grey-brown beds of sandstones widely occurring in Sinai, Upper Egypt and Nubia. The “Nubian Sandstone Formation”, “Nubian Sandstone Group”, “Nubia Sandstone” and “Nubia Formation” are frequent technical nomenclatures for similar sandstone in North Africa and in the Middle East.

POMEYROL (1968), gave a short review of the geology of the Nubian sandstones in which he summarized the views of some previous works concerning the stratigraphical and age relationships of these sandstones. He came to the conclusion that the term “Nubian Sandstone” must be abandoned, since it has lost its original meaning and has been used very loosely in the stratigraphy of North Africa. The work of POMEYROL (1968), was subjected to several criticism and discussions, e.g. HASSAN (1971); ISSAWI (1971); POMEYROL (1971); PRASAD (1971); RIGASSI (1969 and 1970); SHAWA (1969, 1970 and 1971); WEISSBROD (1970 and 1971) and WHITEMAN (1960 and 1971). The main subject of these discussions aimed either to retain or abandon the term “Nubian Sandstone”. POMEYROL (1968), and WEISSBROD (1970), suggest that the term should be abandoned, while SHAWA (1969 and 1970), RIGASSI (1969) and HASSAN (1971) call for retaining the term. ABDEL WAHAB (1972) and EL-HINNAWI (1973) think, however, that the present knowledge of the geology, mineralogy, geochemistry, . . . etc. of the “Nubian Sandstones” is still in its infancy, and such a discussion of nomenclature and classification can only be made if we have adequate and sufficient data.

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Although the Nubian Sandstones cover about 25% of the surface area of Egypt, it is surprising that very few works have been published on the mineralogy and chemistry of these sandstones. A few grain size analyses were carried out by some authors on some surface and subsurface samples from different parts of Egypt. Of these works, those of SHUKRI and SAID (1944 and 1945), SHUKRI (1945 and 1953), SHUKRI and EL-AYOUTI (1953), EL-SHAZLY and SHATA (1960), ZAHRAH (1970) and Abdel WAHAB (1972) are to be mentioned.

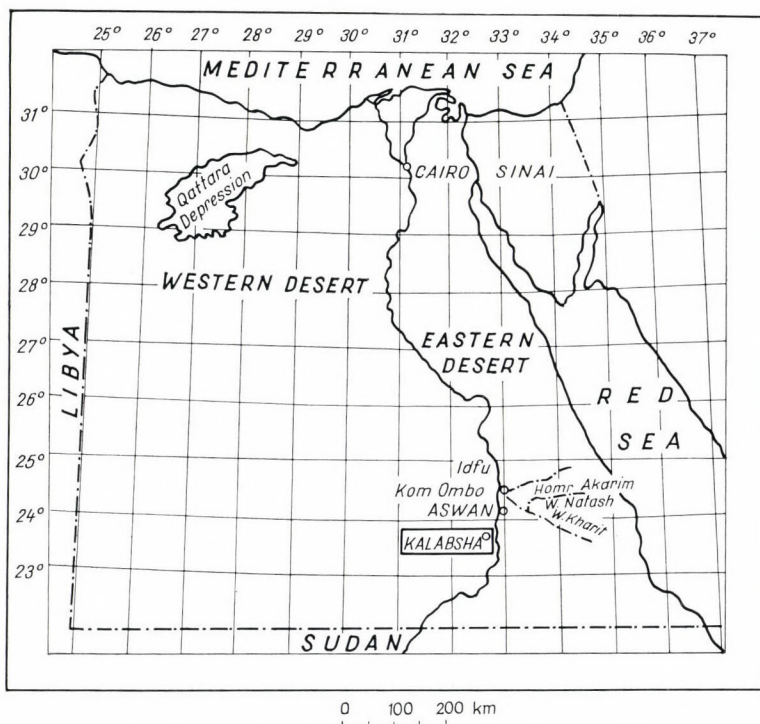


Fig. 1. Location map of the investigated area

As far as the writer is aware, the first chemical analyses of Nubian Sandstones in Egypt, have been introduced by ZAHRAH (1970), who gave 10 analyses of Nubian Sandstones from Barramiya area, central Eastern Desert of Egypt. ZAHRAH (1970) discussed the relationship between the chemistry of the Nubian Sandstones and their distance from the basement complex. Abdel WAHAB (1972), gave 14 complete analyses of Nubian Sandstones from the area, east to Kom Ombo (Wadi Kharit, Wadi Natash and Hamr Akarim areas; south Eastern Desert of Egypt, Fig. 1), and he came to the conclusion that these sandstones are to be classified as "Quartzite clan".

Sampling and Methods of Analyses

The present work deals with the study of the chemistry and geochemical evaluation of the Nubian Sandstones in Kalabsha area, Southwest of Aswan, Upper Egypt, approximate latitude 23°33' N and longitude 32°52' E. (Fig. 1). According to SAID (1962), the Nubian Sandstones of Kalabsha area are of Upper Cretaceous age. This area was chosen because no mineralogical or geochemical studies have been carried out on the Nubian Sandstones in this region. Another reason for this choice is the occurrence of Nubian Sandstones with all the possible varieties in this area.

Ten samples representing all the different varieties of Nubian Sandstones were chemically analyzed by author, using a modified rapid method for silicate analyses (EL-HINNAWI, 1970). In this method, silica, alumina, iron, titanium, phosphorous and manganese were determined colorimetrically. Sodium and potassium were determined by flame photometry, while calcium and magnesium were determined complexometrically.

The results of these analyses are given in Table (1), together with the average analyses of Nubian Sandstone samples from the area east of Kom Ombo (Wadi Kharit, Wadi Natash and Homr Akarim areas) as given by Abdel WAHAB (1972). The average analyses of orthoquartzite, lithic arenite, protoquartzite (sublithic arenite) and average "sandstone" as given by PETTIJOHN (1963) are also tabulated.

Classification and Terminology

The classification and terminology used in this paper follow that of PETTIJOHN (1963) and PETTIJOHN et al. (1972).

Geochemical Evaluation of Nubian Sandstones of Kalabsha Area

Before discussing the chemistry and the geochemical evaluation of the Nubian Sandstones of Kalabsha area, it must be mentioned that a number of problems best any geochemical study of sandstones, as the chemical characteristics of sandstones depend upon the geological processes acting after the deposition of the sand (including compaction, cementation and removal of minerals by interstratal solution). The diagenetic factors are subject to erratic variation, so that it is generally agreed that sandstones should be primarily classified on the basis of characteristics which reveal the provenance or transportational and depositional history of the rock.

In the present work, discussion is made with reference to the "present-day" sandstone, i.e. the sandstones which had been diagenetically changed. No attempt has been made to re-calculate the chemical composition, so as

Table 1
Chemical analyses of Nubian

	K ₁	K ₂	K ₃	K ₄	K ₅	K ₆	K ₇	K ₈
SiO ₂	84.54	82.20	87.16	82.67	83.84	86.15	84.26	81.18
Al ₂ O ₃	3.16	4.18	2.21	3.15	4.00	3.99	4.01	4.66
Fe ₂ O ₃ **	2.96	3.00	2.00	3.95	2.87	2.02	3.16	3.14
MgO	1.68	2.14	1.98	1.66	2.00	1.90	2.04	2.16
CaO	3.01	2.87	2.04	3.15	2.14	1.53	1.10	2.14
Na ₂ O	0.40	1.20	0.80	1.00	0.84	0.70	0.36	0.81
K ₂ O	0.27	0.67	0.36	0.64	0.71	0.38	0.20	0.57
H ₂ O ⁺	0.68	0.40	1.20	0.94	1.02	0.36	1.27	0.53
TiO ₂	0.53	0.91	0.84	0.27	0.14	0.36	0.51	0.36
P ₂ O ₅	0.10	0.09	0.13	0.03	0.09	0.12	0.08	0.14
MnO	0.10	n: 1	n: 1	n: 1	0.03	n: 1	0.05	0.10
CO ₂	2.98	3.01	2.16	1.99	3.18	2.93	2.40	4.96
Total	100.41	100.67	100.16	99.45	100.86	100.44	99.44	100.75

* Analyst: Hafez S. Abdel WAHAB.

** Fe₂O₃ represents total iron.

K₁—K₁₀ Are representative samples from Kalabsha area.

Av. K Represents the average chemical analyses of the Nubian Sandstones of Kalabsha area.

Av. Kh Represents the average chemical analyses of Nubian Sandstones of Wadi Kharit (Abdel WAHAB, 1972).

Av. N Represents the average chemical analyses of Nubian Sandstones of Wadi Natash (Abdel WAHAB, 1972).

to represent the "original" rock, since such calculations are subject to a great number of factors, and would never give the exact picture (MIDDLETON, 1960).

From the analyses given in Table 1, the bulk chemical composition of the Nubian Sandstones of Kalabsha area may reflect their modal composition and the following observations can be outlined:

1. The percentage of silica (SiO₂) ranges from 82.20 to 87.16 percent with an average of 84.27 percent. It is agreed that the chemical composition of the different varieties of sandstones are highly modified by the introduction of various void-filling cements, as the silica content may be augmented by the introduction of silica. Hence, the abundance of silica is practically valueless for any indication of the prediagenetic history of the rock.

2. Alumina (Al₂O₃) ranges from 2.21 to 4.66 percent with an average of 3.55 percent. According to MIDDLETON (1960) all sandstones with less than 5 percent alumina are considered to be "Quartzite clan". This is the case in the Nubian Sandstone samples of Kalabsha area.

*Sandstones of Kalabsha Area**

K ₀	K ₁₀	Av.K	Av.Kh	Av.N	Av.H	A	B	C	D
85.70	85.00	84.27	83.78	84.39	77.11	95.4	66.1	84.01	77.6
2.99	3.18	3.55	2.90	4.54	9.27	1.1	8.1	2.57	7.1
1.98	2.68	2.78	3.22	2.09	2.92	0.6	5.2	0.43	3.2
1.90	2.31	1.98	3.04	2.68	3.14	0.1	2.4	0.67	1.2
2.30	1.84	2.21	2.13	1.66	1.09	1.6	6.2	5.41	3.1
0.78	1.00	0.72	0.58	0.40	1.88	0.1	0.9	0.17	1.2
0.47	0.43	0.47	0.27	0.38	0.64	0.2	1.3	0.86	1.3
0.40	0.84	0.76	0.42	0.63	1.72	0.3	3.6	0.54	1.7
0.21	0.35	0.45	0.41	0.56	0.91	0.2	0.3	0.05	0.3
0.15	0.08	0.10	0.10	0.07	0.07	—	0.1	0.04	0.1
0.02	n: 1	0.03	0.10	0.05	n: 1	—	0.1	0.04	0.1
2.87	2.80	2.93	3.29	2.62	1.98	1.1	5.0	4.65	2.5
99.77	100.51	100.25	100.24	100.07	100.73	100.7	99.3	99.44	99.4

Av. H Represents the average chemical analyses of Nubian Sandstones of Homr Akarim area (Abdel WAHAB, 1972).

A Represents the average chemical analyses of orthoquartzite (PETTIJOHN, 1962).

B Represents the average chemical analyses of lithic arenite (PETTIJOHN, 1963).

C Represents the average chemical analyses of protoquartzite "sublithic arenite" (PETTIJOHN, 1963).

D Represents the average chemical analyses of "sandstone" (PETTIJOHN, 1963).

3. It is shown in Table 1 that Na₂O considerably exceeds K₂O, this is, however, a singular chemical attribute of graywacke (PETTIJOHN et al., 1972), but the present author explains the high Na₂O/K₂O ratio in the samples under investigation (which are evidently not graywacke; see the average chemical analyses and the modal analyses data) by the following:

a) The source rocks of these Nubian Sandstones must be soda-rich. The most abundant possible source rocks for these sandstones are the underlying basement rocks (granites, granodiorites and gneisses). Modal analyses showed that the Nubian Sandstones of Kalabsha area contain appreciable amount of rock fragments (7.8%), which are composed mainly of the following types of fragments arranged in order of decreasing abundance: granites, granodiorites, gneisses and chert. The presence of these types of fragments within the Nubian Sandstones of Kalabsha area may be a fair indication of their origin from the preexisting underlying basement rocks.

b) The high Na₂O/K₂O ratio may have been produced by the albitization of the feld spars in the Nubian Sandstones under consideration during dia-

genesis. Petrographic investigation revealed that the major part of the plagioclase is highly albitized. The distribution diagram for Na_2O and K_2O is graphically represented in Fig. 2.

As alumina (Al_2O_3) and alkalis (Na_2O and K_2O) occur essentially in sandstones in two main mineral groups: (i) clay minerals, micas and chlorites and (ii) feldspars, therefore, alumina and alkalis may show good positive correlation (Fig. 3).

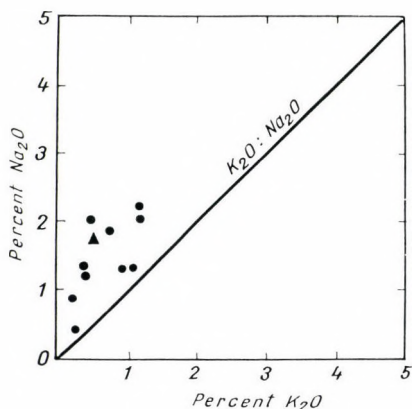


Fig. 2. $\text{K}_2\text{O}/\text{Na}_2\text{O}$ ratio in Nubian Sandstones of Kalabsha area. M = Represents the average

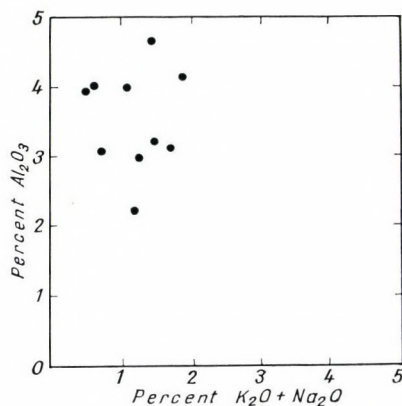


Fig. 3. Content of alkalis versus Al_2O_3 in Nubian Sandstones of Kalabsha area

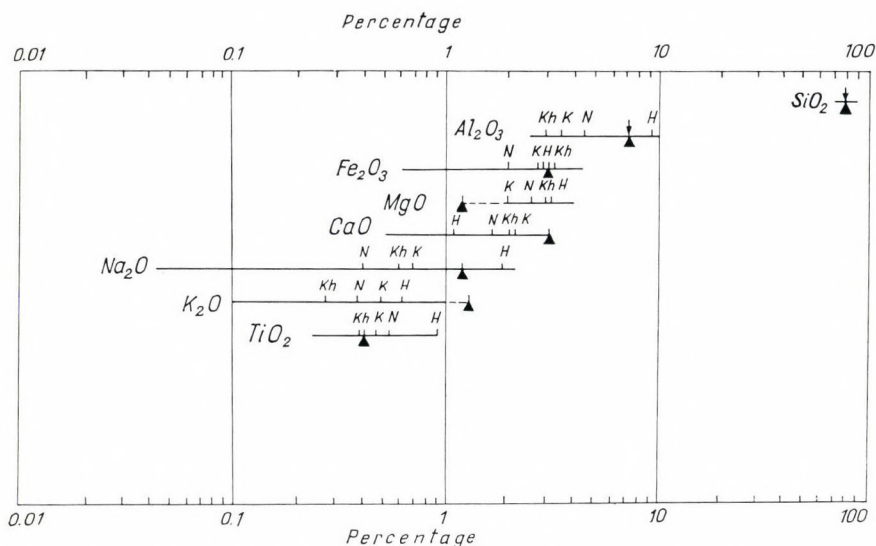


Fig. 4. Major oxide composition of Nubian Sandstones. K = Represents the average of Kalabsha Nubian Sandstones; Kh , N and H = Represents the average Nubian Sandstones of Wadi Kharit and Hamr Akarim, resp. (Abdel WAHAB, 1972); $\blacktriangle$ = Represents average sandstones (PETTJOHN, 1963)

4. The Fe_2O_3 content (total iron) ranges from 1.98 to 3.95 percent with an average content of 2.78 percent. Here, the iron content seems to be in significant since its presence is not only controlled by its provenance, but also by the chemical conditions which prevailed in the basin of deposition and in the sediments during diagenesis. A further factor is the intensity and duration of weathering of the source materials prior to deposition. MIDDLETON (1960), pointed out that "the more intense and prolonged the weathering, the more iron will pass into solution and therefore will not be deposited in a sandstone".

5. The calcite cement present in these sandstones led to a comparatively high CaO and CO_2 content in some samples. Further the high MgO obtained in some of these samples is attributed to the presence of dolomitic carbonate in the cement (Petrographic studies confirm these explanations).

6. Titanium (TiO_2) is found with an average content of 0.14 percent, this percentage is mostly found in clays and in some heavy minerals recorded, such as anatase, rutile and ilmenite.

Figure 4. illustrates the major oxide composition of the Nubian Sandstone samples of Kalabsha area, together with those of the area east of Kom Ombo (ABDEL WAHDB, 1972). The average "sandstone" given by PETTIJOHN (1963) is also illustrated.

Chemical Composition as a Function of Sandstone Type

MIDDLETON (1960) made the first important attempt in evaluating geochemically different groups or "clans" of sandstones. He pointed out that, if a chemical investigation of sandstones is to be independent of diagenetic factors, it may be based upon the ratio between Al_2O_3 and alkalis or between some other individual pairs of oxides. Diagenetic processes during compaction and in the later history of a sandstone, play an important role in its chemical characteristics.

PETTIJOHN et al. (1972) pointed out that the differentiation between mature and immature sandstone is possible by using the ratio of $\text{SiO}_2/\text{Al}_2\text{O}_3$, rather than by using either of them. The mature quartz rich sandstones are of high $\text{SiO}_2/\text{Al}_2\text{O}_3$ ratio because of the absence of aluminosilicates, clays (a textural attribute) or primary silicates (a mineralogical attribute). PETTIJOHN et al. (1972) also mentioned that differences between alkali metal-rich and alkali metal-poor sandstones can almost equally be taken as corresponding to a maturity index defined in chemical composition terms. This was expressed by the ratio of $\text{Na}_2\text{O}/\text{K}_2\text{O}$ and the ratio of $\text{SiO}_2/\text{Al}_2\text{O}_3$ (Fig. 5). Applying the chemical analysis data of the Nubian Sandstones of Kalabsha area to the diagrammatic maturity index given by PETTIJOHN et al. (1972), the sandstone of Kalabsha area are accordingly classified as protoquartzite "sublithic

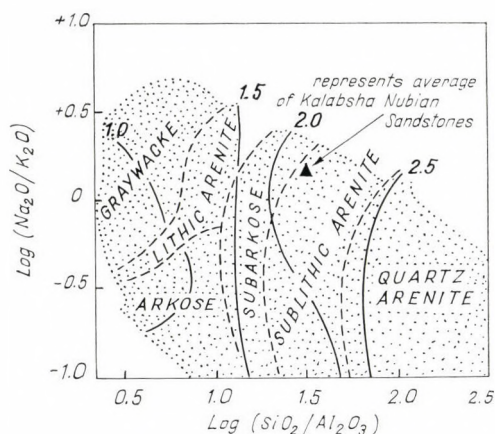


Fig. 5. $\text{SiO}_2/\text{Al}_2\text{O}_3$ vs. $\text{Na}_2\text{O}/\text{K}_2\text{O}$ ratios of sandstones. Stipple outlines area of most analyses. Contours (heavy) lines show values of $\log \frac{\text{SiO}_2 - \text{Al}_2\text{O}_3}{\text{Na}_2\text{O} - \text{K}_2\text{O}}$, PETTIJOHN, 1963; $\blacktriangle$ = Represents average of Kalabsha Nubian Sandstones

arenites²⁴. This is in close agreement with the petrographic modal analyses of these sandstones (Table 2).

As shown from Table 2, the Nubian Sandstones of Kalabsha area are characterized, by the dominance of quartz, abundance of rock fragments (granite, granodiorite, gneiss and chert), subordinate content of feldspars (mostly plagioclase which is highly albitized, but K-feldspars are found to a smaller extent). The dominant cement is silica, while carbonate cement is second in abundance and is composed mainly of calcite and dolomite. Clayey matrix is present as patches and is represented by kaolinite and illite (as revealed from the X-ray diffraction study).

Petrographically, these sandstones are generally moderately sorted and medium in size (i.e. they are of 0.25 to 0.50 mm size). Microscopic investigations

Table 2

Modal analysis of Nubian Sandstones of Kalabsha Area

Constituent	Average percentage
Quartz	75.2
Feldspars	2.5
Mica	0.3
Rock fragments	7.8
Clay or matrix	3.6
Silica cement	7.6
Carbonate cement "calcite and dolomite"	3.0

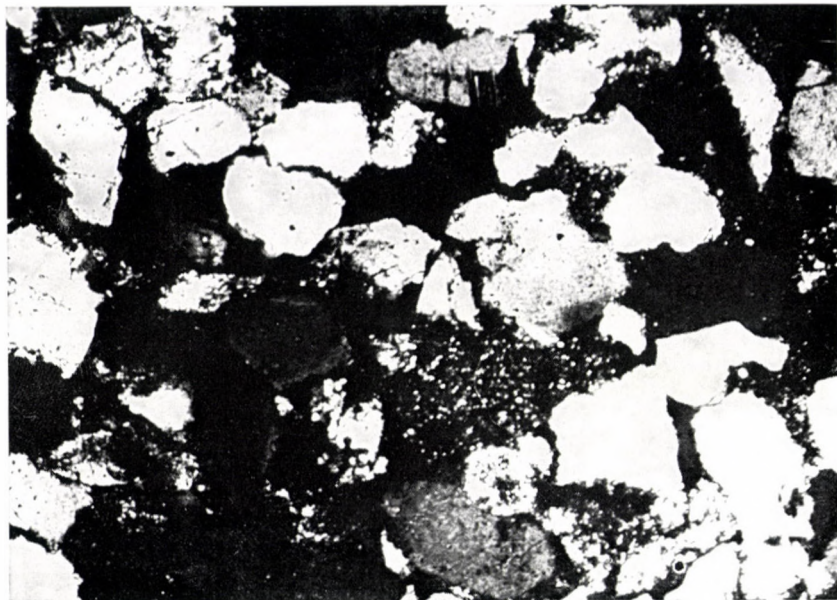


Fig. 6. A photomicrograph showing a medium grained, moderately sorted and subrounded Nubian Sandstone. Single quartz grains, some of which exhibit undulatory extinction. Crossed nicols, $\times 64$

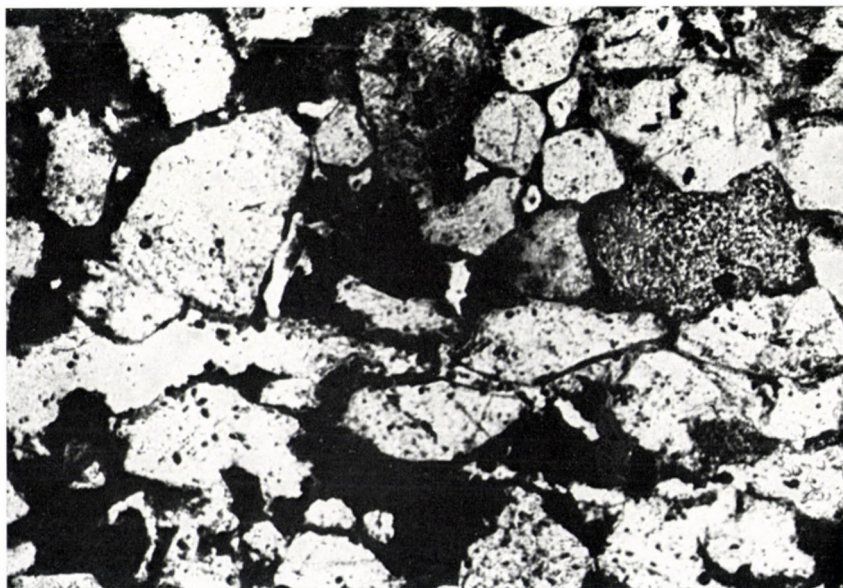


Fig. 7. A photomicrograph showing moderately sorted Nubian Sandstone composed of quartz and rock particles, the latter being in part granits and gneiss. Quartz grains are generally subrounded. Crossed nicols, $\times 64$

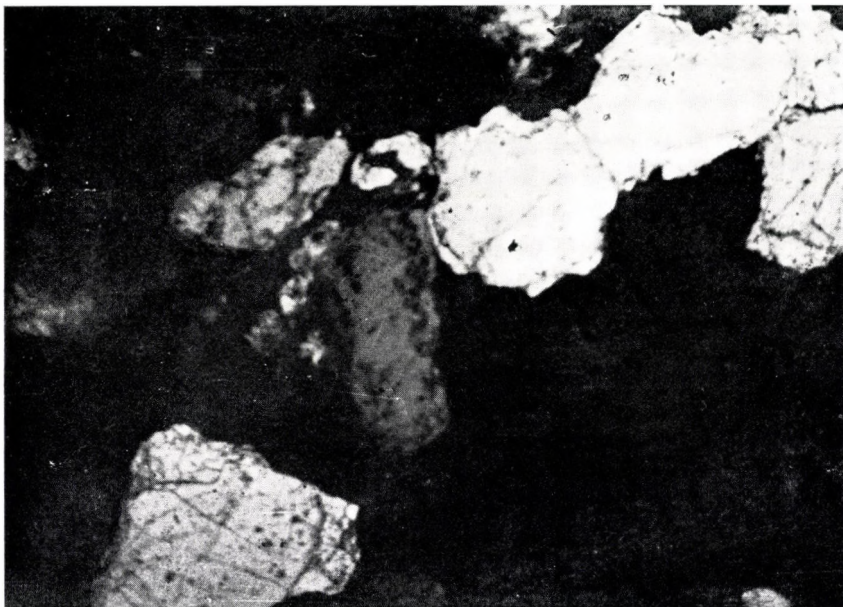


Fig. 8. A photomicrograph showing a considerable secondary overgrowth of silica. The original quartz grain boundaries are well shown. Crossed nicols, $\times 64$

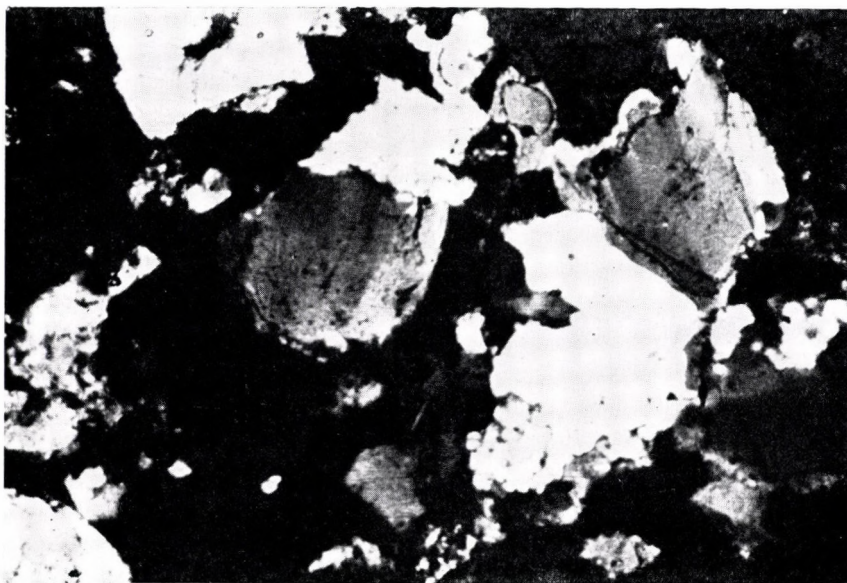


Fig. 9. A photomicrograph showing undulatory composite quartz grains with an overgrowth attaining crystal form in a void space. The detrital grains are mottled with iron oxides (hematite and magnetite) and the grains margins are delineated by iron oxides. Crossed nicols, $\times 64$

revealed that these sandstones are composed essentially of quartz grains; these are mostly monocrystalline (single grains) and polycrystalline to a smaller extent (composite). The quartz grains are highly subrounded and subangular, usually of homogeneous extinction and to a smaller extent of undulatory extinction. Silica is the most common cementing agent, it is almost quartz deposited in optical and crystallographic continuity with the rounded detrital quartz, thus forming the silica overgrowths (Figs 6, 7, 8 and 9).

Summary and Conclusion

The chemical and modal analyses of the Nubian Sandstones of Kalabsha area led the author to designate these sandstones as protoquartzites "sublithic arenites". These sandstones of Kalabsha area were found to be similar in chemical composition to those sandstones of Salt Wash Member of Morrision Formation which were considered to be protoquartzites "sublithic arenites" (PETTIJOHN et al., 1972).

The Nubian Sandstones of Kalabsha area (Western Desert of Egypt) were found to be similar to the Nubian Sandstones of Wadi Natash and Wadi Kharit (Eastern Desert of Egypt). The slight variations — if any — in chemical and/or mineralogical composition are mainly attributed to the diagenetic factors acting on these sandstones after deposition.

Nubian Sandstones recorded from the Barramiya area (hundred kilometers north of Wadi Kharit, Eastern Desert) were designated as quartz arenites "orthoquartzites" (ZAHRAH, 1970). It is to be noted that these Nubian Sandstones of Barramiya Area are far from the basement rocks as compared with those of Wadi Kharit and Wadi Natash. Therefore, the consistency in nomenclature of the Nubian Sandstones as attained by most methods of classification leaves no room for doubt in designating the sandstones, those which are far from the basement rocks are "quartz arenites" or "orthoquartzites", and those near the basement rocks are "sublithic arenites".

The presence of a considerable proportion of rock fragments (granites, granodiorites and gneisses) as well as the field observation, i.e. the Nubian Sandstones cap the basement rocks in Kalabsha area, are indications of the origin of the Nubian Sandstones of Kalabsha area from the preexisting underlying basement rocks.*

* Detailed discussions regarding the source and mode of formation of the Nubian Sandstones in Egypt are now in preparation for publication by the author (see references).

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ХИМИЧЕСКИЙ СОСТАВ И ГЕОХИМИЧЕСКАЯ ОЦЕНКА НУБИЙСКИХ ПЕСЧАНИКОВ РАЙОНА КАЛАБШИ В ЕГИПТЕ

Т. С. А. ВАХАБ

Р е з ю м е

Автор статьи выполнил химический анализ десяти показательных образцов, представляющих собою все возможные разновидности нубийских песчаников района Калабши (Египет). Результаты химических и модальных петрографических анализов позволили отнести эти песчаники к протокварцитам (сублитическим аренитам). Кроме того, было выявлено, что эти песчаники происходят, по-видимому, от ранее существовавших подстилающих пород фундамента.

STRATIGRAPHIC AND EVOLUTIONARY ANALOGIES OF THE EOCENE OF THE PARISIAN BASIN AND OF NE-TRANSDANUBIA

By

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On the basis of personal experiences and literature data the author gives a review on the analogies of the Eocene stratigraphy and evolution of the Parisian Basin and of the NE region of the Transdanubian Central Mountains.

The formations corresponding to the Danian-Montian, Thanetian of the Parisian Basin are unknown in NE Transdanubia.

The Lower Eocene browncoal sequence of NE-Transdanubia as well as its marine cover can and should be correlated with the Sparnacian and with the Cuisian or Ypresian stratotype of the Parisian Basin, respectively.

During the Sparnacian the formation and evolution of the two areas proved to be similar.

Further significant mutual characteristics of the geological evolution are reflected by the events at the Cuisian-Lutetian boundary: local uplift and erosion took place in both areas.

In the Lower Lutetian the transgressive marine formations (with *Nummulites perforatus*) are of general extension also in the area of the Transdanubian Central Mountains showing relatively uniform facies. As contrasted to the Parisian Basin showing more uniform facies pattern in the Middle Lutetian, in NE-Transdanubia the facies fluctuations are frequent (marshy, freshwater, brackish water and marine facies).

Parallely with the enclosing of the Upper Lutetian in the Parisian Basin, in the Dorog Basin the so-called Tokod sandstone was deposited indicating separation from the sea. The lack of the fauna characterized by *Nummulites perforatus* in the Parisian Basin is significant.

In the Upper Eocene analogies are hardly found, in the Parisian Basin the effect of the North Sea predominates with English (Bartonian) and Belgian relations.

In France the Paleogene is tripartite. The Paleocene includes the combined Danian-Montian and Thanetian stages. To classify the Eocene formations the Ypresian (Lower), the Lutetian (Middle) and the Bartonian (Upper) stages are used, in case of the Ypresian two, in that of the Lutetian and Bartonian three substages are distinguished.

Just like in the Transdanubian Central Mountains, before the Paleogene sedimentation the Mesozoic formations were considerably evolved. Considerable difference exists, however, in the make up of the basement: in the Parisian Basin the basement of the Paleogene is built up by Cretaceous formations. The youngest Cretaceous formation can be assigned to the Campanian (chalk with *Belemnitella mucronata*). In the area of the Transdanubian Central Mountains the Paleogene formations are for the most part underlain by Triassic dolomite and Dachstein limestone, to some extent by Cretaceous and

subordinately by Jurassic formations testifying to the more considerable denudation of the Transdanubian Central Mountains at the end of the Mesozoic.

During the Danian and Montian the sea returned from the West through the Channel into the Parisian Basin, the sediments of which, i.e. the marl of Meudon (southwest of Paris), and the "pisolitic limestone" up to Melun (southeast of Paris along the banks of the Seine) are known in the depression preformed structurally in NW-SE direction following the Seine-line, and further in SW-NE direction.

The formations marked formerly as "pisolitic" limestone are really coarse limestones or biocalcarenites which were developed from calcareous algae fields. E. SZÁDECZKY-KARDOSS (1925) described similar pseudo-oolitic freshwater algal limestone from the Transylvanian Paleocene. On the age of the formation series in question discussions were carried out during more than fifty years in the French literature. E.g. the fauna, and particularly the corals include several Mesozoic elements. By means of ostracod and foraminiferal studies has been unambiguously evidenced that their age is undoubtedly Paleogene, i.e. Danian-Montian.

When summing up the results of the investigations carried out till now, according to Ch. POMEROL (1973) this gulf drawing from the West, i.e. from

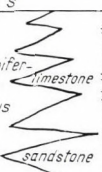
PARISIAN BASIN (after Ch. Pomerol) NE- TRANSDANUBIA							
epoch	stage	substage	main facies		stage	substage	
E O C E N E	Bartonian	Ludian	white marls of Pontin blue marls of Argenteuil gypsum and marl intercalations marl with <i>Pholadomya ludensis</i>	 limestone sandstone with Nummulites and Discocyclina	Upper	Priabonian	
		Marinesian	Marinesian sands sands of Cresnes		Middle		
		Auverasian	sand and sandstone of Beauchamp Auversian sands		Lower		
	Lutetian	Upper	sandstone of Faulgoues "pelecypod field"	Takod" sandstone poor in fossils	Upper	Lutetian	
		Middle	limestone of Millolima	sequence with Nummulites striatus and molluscs	Middle		
		Lower	Coarse limestone, Nummulites laevigatus	sequence with Nummulites perforatus	Lower		
	Ypresian	Cuisian	sand of Belleu Cuisian sand	clay-marl with Nummulites and Operculina clay-marl with molluscs, coprina sequence	Upper	Cuisian	
		Spannacian	clay and lignite of Soissons plastic clay	brown-coal sequence variegated-clay, quartz-sand- stone, freshwater limestone	Upper Lower		Spannacian
	PALEOCENE	Thanetian	sand of Bracheux, tuff of La Fere	terrestrial period			
		Danian - Montian	marl of Meudon, "pisolitic" limestone				

Fig. 1. Stratigraphic comparison of the Paleocene and Eocene of the Parisian Basin (after CH. POMEROL, L. FEUGUEUR, 1968) and of NE Transdanubia

the Atlantic through the Channel had no connection with the Montian Basin of Belgium.

Except the opinion of F. HORUSITZKY (1964) and according to the recent Hungarian stratigraphic views the area of the Transdanubian Central Mountains was emerged land at that time. With regard to the opinion of F. HORUSITZKY and taking into account the variegated clay sequence which amounts locally to more than 120 metres thickness and which lies between the Sparnacian lacustrine-lagoonal facies and the Mesozoic basement, the stratigraphic investigation of the variegated clays at the bottom of the Eocene rises as a stressed stratigraphic problem. After the regression of the Danian-Montian sea and the erosion of the Danian-Montian formations the new transgression arrived from the North during the Thanetian and only its completing phase reached the Parisian Basin. The most widespread formation of the Thanetian is the Bracheux sand (up to 30 metres thick) which contains numerous molluscs in addition to *Cyprina scutellaria*. At Coye (NE of Paris) the littoral facies of the Thanetian sands is represented by a chert-conglomerate which is equivalent with the Bracheux sands. The Bracheux sands are overlain by lacustrine limestones in the eastern part of the Parisian Basin. These are of less extent than the Bracheux sands.

As to our recent knowledge in the Transdanubian Central Mountains no formations being analogous both in age and facies with the Thanetian of the Parisian Basin, are known. The problem of the Lower Eocene terrestrial discussed above in connection with the Danian-Montian concerns also the problem of the Thanetian (Fig. 1).

In the Parisian Basin the Thanetian-Sparnacian boundary is well marked (Fig. 2). East of Reims, on Mont Berru in the Cernay conglomerate a rich vertebrate fauna of Thanetian age was found which has been studied by RUSSEL (1964). The fauna consisting of crocodiles, large turtles, of a gigantic bird (*Gastornis*) as well as of mammals of Mesozoic affinity has been correlated with the Upper Paleocene fauna of North America. Nevertheless, this fauna indicates the end of evolution of the Mesozoic mammals and differs from the Sparnacian mammals initiating the Tertiary (*Coryphodon*). According to Ch. POMEROL this fact is an important argument in favour of the assignment of the Sparnacian to the Ypresian cycle, and of the drawing of the Paleocene-Eocene boundary between the Thanetian and Sparnacian.

As to the connection of the Sparnacian Ypresian, similar conditions are known in the northeastern part of the Transdanubian Central Mountains: the transition is continuous from the Sparnacian lower brown coal sequence through brackish water formations towards its higher-seated cover being assigned to the Cuisian, i.e. the neritic *Operculina*-bearing clay-marl. An unsettled question of the Hungarian Eocene stratigraphy is connected to this, i.e. the more precise correlation with the stratotypes.

The experiences obtained in France support the authors former opinion that the Lower Eocene brown coal sequence of NE-Transdanubia and its marine cover can and should be correlated with the Sparnacian resp. with the Cuisian or Ypresian stratotypes of the Parisian Basin. The thick, (locally 10 metres) Sparnacian of ours is correlable in duration with the Sparnacian clayey brown coal sequence and plastic clay (several metres thick) of the Parisian Basin.

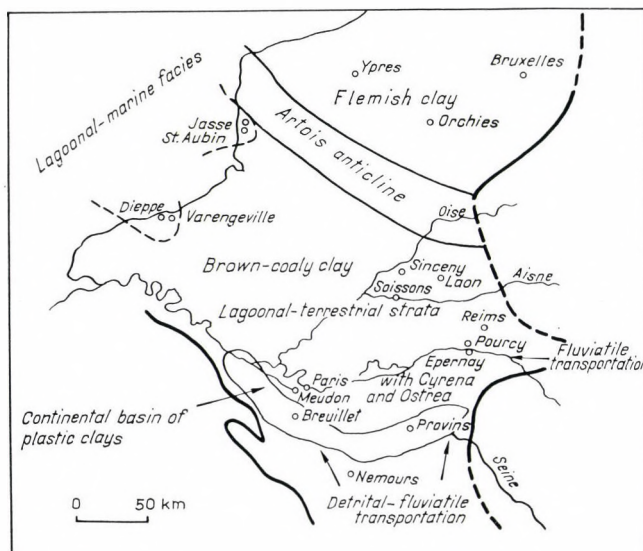


Fig. 2. Extension and main facies of the Sparnacian in the Parisian Basin (after CH. POMEROL, 1973)

Before the general Cuisian transgression the area of the Parisian Basin was covered by lagoons. The area of there was separated from the Belgian Ypresian sea by the Artois barrier (anticline). The lagoonal Sparnacian and the marine Cuisian are equivalent to the completely marine Belgian Ypresian. The Sparnacian has several lateral facies. The most complete outcrop of the Sparnacian is found in the Guitrancourt quarry (west of Paris, in the right bank of the Seine), where its total thickness amounts to 10.85 metres.

In spite of the fact that the Hungarian Eocene belonged to the Mediterranean and that of the Parisian Basin to the Northern facies zone, it can be established that the facies and evolution of the two areas were similar during the Sparnacian.

In the Cuisian substage the Artois barrier (anticline) disappeared. Its area was inundated by the sea together with that of Belgium and the northern and north-western parts of the Parisian Basin (Fig. 3). The area of the Parisian Basin became opened north- and westwards, i.e. towards the Atlantic. Nummulites planulatus, the highly abundant index fossil of the Cuisian sand got into the

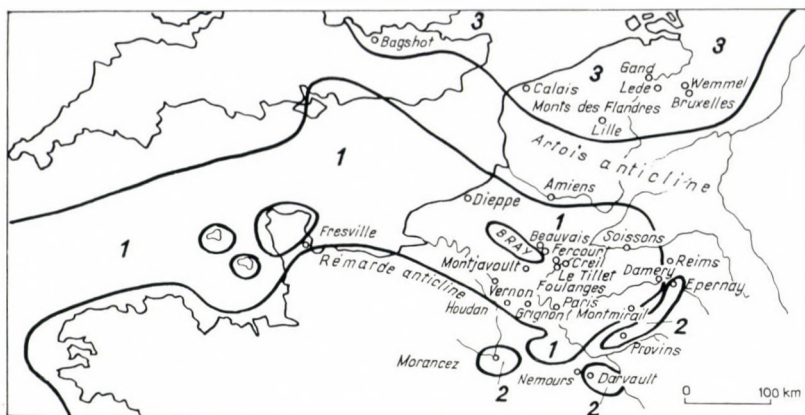


Fig. 3. Extension of the Middle and Upper Lutetian separated by the Artois anticline in the Parisian and Belgian basins (after CH. POMEROL, 1973). 1. Marine limestones and sandstones of lagoonal-marine facies (caillases); 2. Lacustrine limestone; 3. Calcareous sandstones of Lede and Wommel

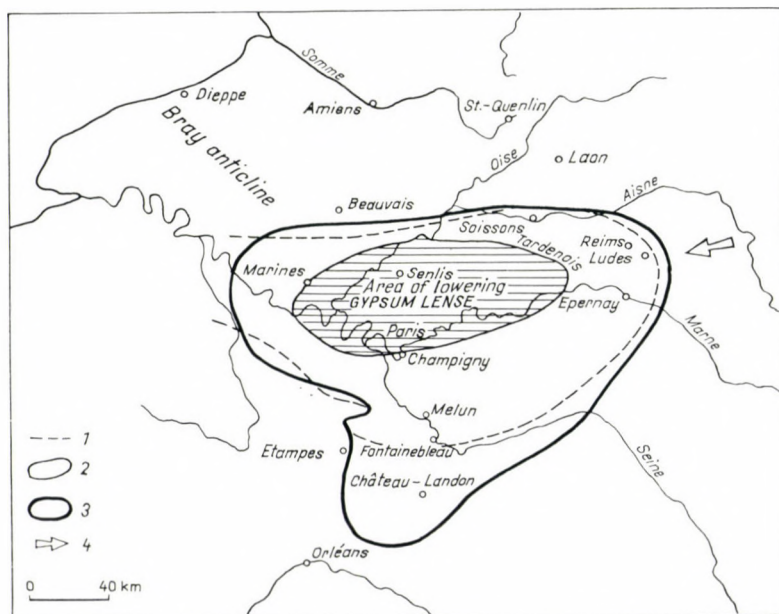


Fig. 4. Ludian in the Parisian Basin (after CH. POMEROL, 1973). 1. Extension of the marls with *Pholadomya ludensis*; 2. Extension of the gypsum lense; 3. Lacustrine limestones and marls; limestone of Champigny, limestone of Château-Landon marls of Tardenois; 4. Direction of fluvial transportation into the Ludian sea

Parisian Basin from the Atlantic. This species is found also in the clay marls with *Operculina* of the Transdanubian Central Mountains as it has been stated both by M. JÁMBOR-KNESS and by A. BLONDEAU. In addition to the same age an indirect communication between the Eocene seas of the Transdanubian Central Mountains and of the Parisian Basin can also be assumed. Further significant common features of the geological evolution are reflected by the events at the Cuisian-Lutetian boundary. In the Guitrancourt profile only the lower 50 cm of the Cuisian have remained untouched, all the strata from the Cuisian-Lutetian boundary up to the Middle Lutetian being eroded. The Cuisian profile at Fosses is more complete but both uplift and denudation can be demonstrated.

Similar phenomena can be observed in the NE area of the Transdanubian Central Mountains: uplift at the Cuisian-Lutetian boundary, erosion of Lower Eocene formations within the Eocene as well as the fauna redeposited in reversed order from the Lower Eocene formations into the Middle Eocene ones have repeatedly been ported on (M. JÁMBOR-KNESS 1968; L. GIDAI, 1968). Taking into account the distance of about one and a half thousand kilometres, one can conclude that the uplift and separation at the Cuisian-Lutetian boundary proved to be of regional scale.

The Lutetian sequence of the Parisian Basin of about up to 40 metres thickness shows the classical scheme of one sedimentary cycle: the sequence consisting of coarse detrital facies (in lowermost part coarse glauconite) at the bottom is followed by coarse limestone in its middle part. At the end of the cycle the organogenic-detrital sedimentation is replaced by evaporitic-chemical sedimentation.

The evolution of the Parisian Basin and of the Dorog Basin (the most complete Lutetian sequence in the Transdanubian Central Mountains) shows certain analogies: In the Lower Lutetian the transgressive marine formations (with *Nummulites perforatus*) are of regional extension also in the area of the Transdanubian Central Mountains showing a relatively uniform facies. In the Middle Lutetian, the Parisian Basin shows a more uniform facies pattern, while in the Transdanubian Central Mountains the frequent change of facies (marshy, freshwater, brackish water and marine), viz. the archipelagic character predominates. During the Upper Lutetian similar analogies can be outlined: simultaneously with the Upper Lutetian separation of the Parisian Basin, the coarse fauna-poor Upper Lutetian Tokod sandstone of the Dorog Basin suggests also disconnection with the sea.

In the Lower Lutetian of the Parisian Basin the "en masse" presume of *Nummulites levigatus* indicating to La Manche — Atlantic as well as to Mediterranean connections makes possible a direct correlation with the *Nummulites laevigatus* bearing formations of the Southern Bakony Mountains.

A significant difference is the lack of the *Nummulites perforatus* fauna in the Parisian Basin. On this basis it is obvious that the Parisian Basin

belonged to the northern facies zone and had no direct connection with the Mediterranean.

After the drying of the Upper Lutetian lagoons the Parisian Basin had been inundated again, i.e. the Bartonian transgression set in. After the deposition of the Auversian and Marinesian marine sands (Lower and Middle Bartonian) — their lateral facies consist of clay and limestone — in the Ludian the Upper Bartonian marls with *Pholadomya* indicate a new but short transgression.

CUVIER described *Paleotherium magnum*, *Anoplotherium commune* and *Xiphodon gracile* from the Montmartre gypsum. These proved to be index species of the Eocene. Only these mammal into the Eocene and to the fact that the Oligocene should be started with the Romainville clays in the Parisian Basin.

In the Upper Eocene only a few analogies are found, e.g. that the characteristic Nummulites of the Auversian, i.e. *Nummulites variolarius* occurs also in Hungary. After the Auversian the indirect connections of the Parisian Basin with the Mediterranean were stopped: in the Upper Eocene of the Parisian Basin the influence of the North Sea predominated with English (Bartonian stage) and Belgian relations.

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АНАЛОГИЯ СТРАТИГРАФИИ И ИСТОРИИ ГЕОЛОГИЧЕСКОГО РАЗВИТИЯ ЭОЦЕНОВЫХ ОБРАЗОВАНИЙ ПАРИЖСКОГО БАССЕЙНА И ЗАДУНАЙСКОГО СРЕДНЕГОРЬЯ (ВЕНГРИИ)

Л. ГИДАИ

Резюме

Автор на основании личных наблюдений и важнейших литературных данных дает обзор аналогичных явлений стратиграфии и истории геологического развития эоценовых образований Парижского бассейна и Задунайского Среднегорья.

Образования, соответствующие датско-монскому и танетскому ярусам Парижского бассейна, в северовосточной части Задунайского Среднегорья не известны.

На северо-востоке Задунайского края нижнеэоценовую толщу бурых углей можно и следует сопоставить со спарнакским, кровлю морского происхождения — с кюизским или ипрским стратотипами Парижского бассейна.

Путь развития этих двух территорий на протяжении спарнакского века был схожим.

Дальнейшие значительные общие черты пути развития отражают процессы, происшедшие на границе кюизского и лютецкого веков: локальные поднятия и размыв имели место на обеих территориях.

Морские (с *Nummulites perforatus*) трансгрессивные образования нижнелютецкого яруса на территории Задунайского Среднегорья имеют также общее распространение, показывая при этом сравнительно единую фациальную картину. В среднем лютеции, по сравнению с Парижским бассейном, имеющим более единое фациальное развитие, на северо-востоке Задунайского края часто наблюдаются фациальные колебания (среди болотных, пресноводных, слегка солоноводных и морских фаций).

Одновременно с процессом изоляции в позднем лютеции Парижского бассейна в Дорогском бассейне отложились токодские песчаники, указывающие также на изоляцию от моря.

Значительное отличие заключается в отсутствии в Парижском бассейне фауны с *Nummulites perforatus*.

В верхнем эоцене уже мало можно встретить примеров аналогии, в Парижском бассейне преобладает влияние Северного моря с английскими (бартонский ярус) и бельгийскими связями.

BIOSTRATIGRAPHIC SIGNIFICANCE
OF THE MEANDROSPIRA PUSILLA (Ho)
(FORAMINIFERA) AND ITS DISTRIBUTION
IN TRIASSIC SEDIMENTARY ROCKS IN HUNGARY,
AS REVEALED BY HYDROCARBON EXPLORING
BOREHOLES

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On the basis of a number of references the distribution of the Foraminifera-species *Meandrospira pusilla* (Ho) in the Lower Triassic sedimentary rocks of Europe, Asia and Africa is shortly discussed, referring to its scarce occurrence in Middle Triassic rocks, as well. The *Meandrospira pusilla* (Ho) shows a general occurrence in shallow marine formations and is a species of wide ecological toleration by adapting itself easily to the changes in depositional environment and to the sea-floor lithology. This species is of considerable biostratigraphic significance, for, in addition to its narrow vertical occurrence (Campilian only), it could be traced in wide areas, in almost all Lower Triassic marine formations.

The stratigraphy of the *Meandrospira pusilla* (Ho)-bearing Lower and Middle Triassic formations discovered by hydrocarbon exploring boreholes in Hungary are discussed and a short description of the species *Meandrospira pusilla* (Ho) as well as that of its rather poor but characteristic microfaunal assemblages are reported. In the Lower Triassic formations this latter assemblage is characterized by few individuals of the following Foraminifera-species: *Ammodiscus* sp., *Earlandia tintiniformis* (MISIK), *Cyclogyra* cf. *mahajeri* BRÖNNIMANN, ZANINETTI, BOZORGNIA, *Spirorbis phlyctaena* BRÖNNIMANN, ZANINETTI, *Glomospirella* sp. The individuals of the *Meandrospira pusilla* (Ho) in the Middle Triassic (Anisian) formations are accompanied by some *Trochammina* cf. *almutensis* KOEHN-ZANINETTI and *Meandrospiranella* sp. only.

Introduction

Lower Triassic (Werfenian-Scythian) marine formations can be easily traced throughout the Tethys-area (Alps, Carpathians, Dinarids, Balkan-paeninsula), in the marginal areas to Eurasia (Asia Minor, Iran), throughout the area of the Triassic Middle Sea in Asia (Himalaya, India, Southern China, Archipelago of Malaya, Pacific Coast of Siberia) and in the environ of the Pacific Sea (California). In these formations, the littoral nearshore and shallow-water facies and the lagoonal sediments play most important role.

Some horizons within the Campilian (Lower Triassic) are characterized by a massive occurrence of the Foraminifera species *Meandrospira pusilla* (Ho), forming the earliest Foraminifera zone of the Triassic. On the basis of characteristic Foraminiferal assemblages the Triassic has been split into 11 Foraminifera-zones (J. SALAJ, 1969a), the lowermost of which is the "*Meandrospira pusilla*"-zone.

*The "Meandrospira pusilla"-zone in Europe,
Asia and Africa*

The lower boundary of occurrence of the species *Meandrospira pusilla* (HO) is unknown. The Seisian sandstone of NW-Montenegro, Yugoslavia (S. PANTIĆ, 1967) is the earliest formation in which it has been observed. The microfaunal assemblage known from the formation involved is rather poor: i.e. it consists of some individuals of *Meandrospira pusilla* (HO) and *Fronicularia woodwardi* HOWCHIN.

The species *Meandrospira pusilla* (HO) described from the Triassic of China (Y. HO, 1959) is characteristic of the Campilian locations of the Alpine Triassic (Italian-Alps, Western-Carpathians, Hungarian Central Mountains, Dinarids, Eastern Serbia, Eastern Balkan, Greece, Turkey). In the stratigraphic literature data are found referring to its additional occurrence in the Triassic sediments of the Asian Middle Sea (Caucasus, Iran, Pakistan) as well as in the northern part of the African continent (Southern Tunisia) (Fig. 1).

In the Campilian sediments it is a common phenomenon that the faunal assemblage is only composed of the individuals of species *Meandrospira pusilla* (HO) only. (J. SALAJ—A. BIELY—J. BYSTRICKY, 1967b; A. RAMOVŠ,

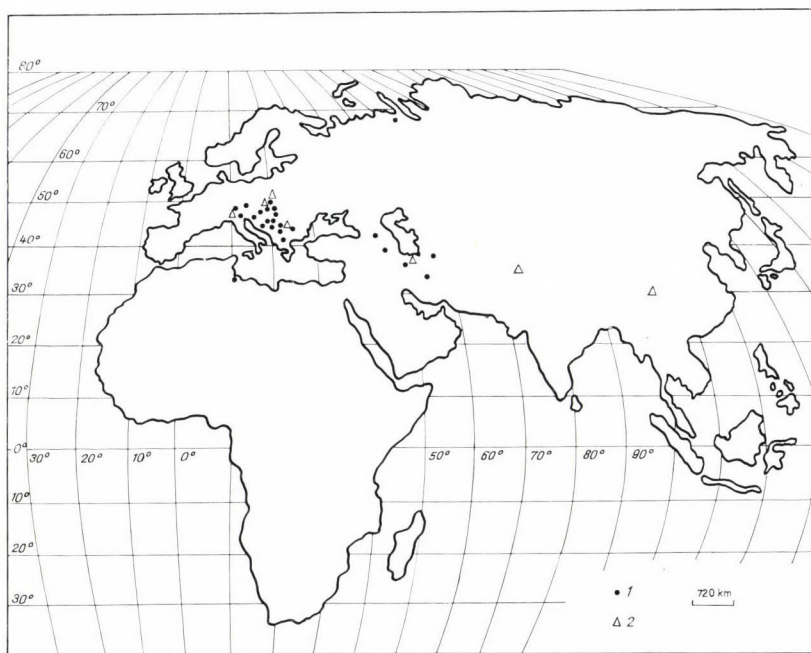


Fig. 1. Occurrence of the species *Meandrospira pusilla* (HO) in Europe, Asia and Africa. 1. Individuals of *Meandrospira pusilla* (HO) in Lower Triassic sedimentary rocks; 2. Individuals of *Meandrospira pusilla* (HO) in Middle Triassic sedimentary rocks

1968; M. ROKSANDIĆ—M. CANOVIĆ, 1970; E. TRIFONOVA, 1972; A. BÉRCZI-MAKK, 1972; G. CHRISTODOULOU—S. TSAILA MONOPOLIS, 1972). The accompanying fauna — if any — is rather poor and is represented by several *Ammodiscus* sp., *Glomospira* sp., *Glomospirella* sp., *Earlandia tintinniformis* (MISIĆ), *Cyclogyra* ? *mahajeri* BRÖNNIMANN, ZANINETTI, BOZORGNIA, *Spirorbis phlyctaena* BRÖNNIMANN, ZANINETTI with some Mollusca- and Ostracoda shell fragments. The characteristic fact is worthy of note that in certain cases a massive occurrence of Echinoid-fragments can be found together with the individuals of the *Meandrospira pusilla* (HO) (e.g. P. BRÖNNIMANN—L. ZANINETTI—A. MOSHTAGHIN—H. HUBER, 1973; A. BÉRCZI-MAKK in this paper). In these instances the accompanying faunal assemblage — if any — is not characteristic. In the Lower Triassic of the Alps (L. KOEHN—ZANINETTI, 1969) and in the Campilian sediments of the Western Carpathians (J. SALAJ, in press) the *Meandrospira pusilla* (HO) can be found together with the Foraminifera-species *Hemigordius chialingchiangensis* (HO).

In the Middle Triassic Anisian sedimentary rocks the species *Meandrospira pusilla* (HO) is represented by few individuals only, accompanied, however, by a number of other microfossils (Y. HO, 1959; J. SALAJ—A. B. NIELY—J. BYSTRICKY, 1967a, 1967b; T. BECHSTÄDT—R. BRANDNER, 1970; P. BRÖNNIMANN—L. ZANINETTI—F. BOZORGNIA—H. HUBER, 1972; P. BRÖNNIMANN—J. CADET—L. ZANINETTI, 1973a; A. GAZDZICZKI—J. TRAMMER—K. ZAWIDZKA, 1975; L. ZANINETTI—P. BRÖNNIMANN, 1975).

In the younger Middle Triassic (Lower Ladinian) occurrence of *Meandrospira pusilla* (HO) has been reported in one location only: in the Krizna-Basin of the Western Carpathians i.e. in the northern foreground of the Low Tatra Mts. It can be found in Lower Ladinian Diplopora-bearing dolomite accompanied by a number of other microfossils (J. SALAJ—A. BIELY—J. BYSTRICKY, 1976).

Individuals of the species *Meandrospira pusilla* (HO) are most frequently found in limestones (L. KOEHN—ZANINETTI, 1969; A. RAMOVŠ, 1968, 1974; A. BÉRCZI-MAKK, 1972; J. SALAJ, 1976), in dolomites (Y. HO, 1959; Z. DURDANOVIĆ, 1967; B. SCAVNICAR—A. SUSNJARA, 1967; S. PANTIĆ, 1970; P. BRÖNNIMANN—L. ZANINETTI—F. BOZORGNIA—H. HUBER, 1972; P. BRÖNNIMANN—L. ZANINETTI—A. MOSHTAGHIAN—H. HUBER, 1973; A. BAUD—P. BRÖNNIMANN—L. ZANINETTI, 1974); in dolomitic limestones (Y. HO, 1959; P. BRÖNNIMANN—L. ZANINETTI—F. BOZORGNIA, 1972; S. PANTIĆ—J. P. RAMPNOUX, 1972) rarely in marly limestones (S. PANTIĆ—S. MOJSILOVIĆ, 1967); in sandy limestones (M. CANOVIĆ—R. KEMENCI, 1972); in marl slates (J. SALAJ—A. BIELY—J. BYSTRICKY, 1967b); in oolitic limestones (H. HAGN, 1955; J. P. RAMPNOUX, 1968); in sandstones (S. PANTIĆ, 1967); in marine clastic series (J. SALAJ, 1969a).

Biostratigraphic Significance of the Meandrospira pusilla (HO)

The *Meandrospira pusilla* (HO) has a great biostratigraphic significance, since, it is considered as a zone-marking Foraminifera-species of the Campilian substage just because of its massive appearance. More definitely; the species involved should be a zone-marking one, because — in addition to its fairly narrow vertical occurrence (practically Campilian only) — it could be horizontally traced in a large area — according to biostratigraphic references.

Having identified the species *Meandrospira pusilla* (HO), Campilian formations of different regions (for instance Alps, Carpathians, Dinarids, Balkan Mts., North Africa, Asia-Minor, China) could be easily correlated. The species involved adapts itself well to the changing environments. The fact that a massive occurrence of individuals of the *Meandrospira pusilla* (HO) without any other significant fossil can be equally found in dolomites with gypsum-intercalations (Y. HO, 1959), in anhydritic marls (A. BÉRCZI-MAKK, 1972) allows to conclude that this species adapts itself well to the salinity increase of sea-water. At the same time, the normal marine conditions provide also favourable conditions for the *Meandrospira pusilla* (HO), since it shows a widespread occurrence accompanied by a number of macrofossils as *Tiroites cassianus* (QUENSTEDT), *Naticella costata* (MÜNSTER), *Myophoria costata* (ZENKER), *Gervilleia* sp. etc. (A. BÉRCZI-MAKK, 1972; K. BUDUROV—E. TRIFONOVA, 1974; A. RAMOVŠ, 1968).

The lithological composition of the sea-floor, the agitation of the sea-water, as physical-environmental factors are significant influencing the living conditions of a benthonic Foraminifera-species. The individuals of the *Meandrospira pusilla* (HO) can be generally found in clay-free, pure, recrystallized limestones and dolomites, although its occurrence is characteristic of clayey dolomites (P. BRÖNNIMANN—L. ZANINETTI—F. BOZORGNI, 1972), dolosparite containing quartz grains (P. BRÖNNIMANN—L. ZANINETTI—A. MOSHTAGHIAN—H. HUBER, 1973) as well. Individuals of this species are known from clayey marls and from silty limestones, too (A. BÉRCZI-MAKK, this paper). Consequently, this Foraminifera-species has a considerable capacity to accommodate itself to the ecological factors and could exist not only in a sea-floor composed of clay-free lime mud (A. RAMOVŠ, 1968; S. PANTIĆ, 1967; L. KOEHN—ZANINETTI, 1969; A. BÉRCZI-MAKK, 1972), but could accommodate itself easily to the physical changes of the sea-floor. The individuals of the species *Meandrospira pusilla* (HO) recovered from stratified sandstone of NW-Montenegro (Yugoslavia) (S. PANTIĆ, 1967) and described from the Lower Triassic shallow-water, clastic sediments revealed by boreholes in Vojvodina (N-Yugoslavia) (M. CANOVIĆ—R. KEMENCI, 1972) could be considered as additional examples of its fairly considerable self-accommodating capacity.

The species *Meandrospira pusilla* (HO) shows a general occurrence in sediments deposited in poorly agitated, shallow-water (neritic) environments of low energy level (Z. DURDANOVIĆ, 1967; B. SCAVNICAR—A. SUSNJARA, 1967; P. BRÖNNIMANN—L. ZANINETTI—F. BOZORGNIA—H. HUBER, 1972; P. BRÖNNIMANN—L. ZANINETTI—A. MOSHTAGHIAN—H. HUBER, 1973). In the formations where individuals of the species involved appear with Echinoid-fragments, the microfacies investigations revealed quartz and rock-fragment grains of silt size referring to some kind of water-agitation during sedimentation. The Echinoid-fragments and the fine-grained clastic material were carried by this water-movement to the sea-floor covered by lime-mud. The occurrence of this species in oolitic limestones (J. P. RAMPNOUX, 1968; L. KOEHN-ZANINETTI, 1969; S. PANTIĆ—J. P. RAMPNOUX, 1972) refers also to the agitation of sea-water.

The general appearance of the species *Meandrospira pusilla* (HO) in the Lower-Triassic marine formations is considered as an additional evidence for the idea, that the marine faunas of the Lower-Triassic were uniform on a world-scale. This conclusion has been drawn by other geologists-paleontologists when studying other taxons as Bivalves and Ammonites (B. GÉCZY, 1974; B. KUMMEL, 1973).

*The Occurrence of the Species Meandrospira
pusilla (HO) in the Lower Triassic Sedimentary
Formations of Hungary as Revealed by Hydrocarbon
Exploring Boreholes*

A number of hydrocarbon exploring boreholes drilled in different regions of Hungary revealed Lower and Middle Triassic sedimentary rocks containing sections of *Meandrospira pusilla* (HO) (Fig. 2). Largest number of individuals was found in Lower Triassic Campilian formations being rather poor in other microfossils.

A massive occurrence of Echinoid and Mollusc-fragments as well as that of some Foraminifera remnants of rather poor quality (*Ammodiscus* sp., *Glomospira* sp., *Meandrospira* cf. *deformata* SALAJ, *Frondicularia* sp.) and Holothuroidea-fragments (*Binoculites* sp.) are characteristic of the accompanying fauna of the *Meandrospira pusilla* (HO) found in large number in dark-grey schistose, anhydritic marl and in reddish-brown schistose, fine silt-bearing clayey marl as well as in coarsely crystallized limestone (cores 22—25 in borehole Ortaháza-11; core 34 in borehole Ortaháza-West-1) recovered from a depth of 2300 resp. 3700 m b.s.l. in the Ortaháza area, Zala County, W-Hungary.

In thin sections of a brownish-gray massive limestone with calcite-veins, anhydrite and gypsum patches as well as with quartz and feldspar grains of

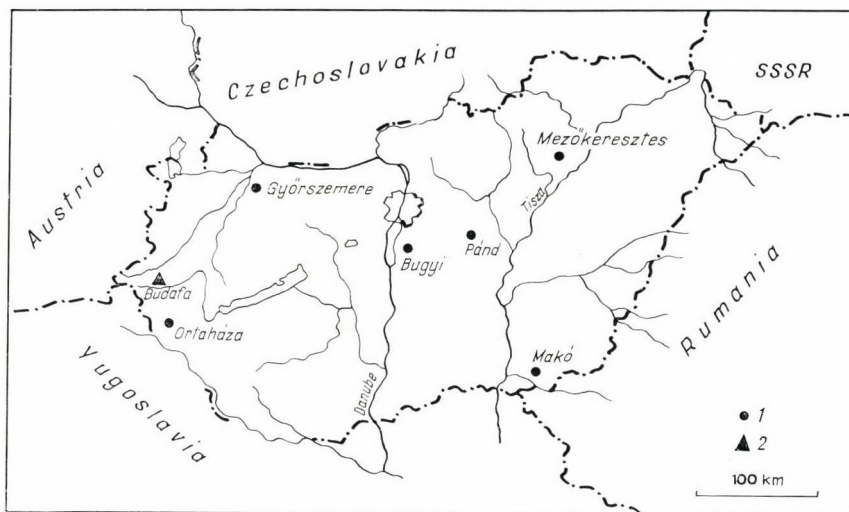


Fig. 2. Occurrence of the species *Meandrospira pusilla* (HO) in Triassic sedimentary rocks of Hungary as revealed by hydrocarbon exploring boreholes. 1. Individuals of *Meandrospira pusilla* (HO) in Lower Triassic (Campilian) sedimentary rocks; 2. Individuals of *Meandrospira pusilla* (HO) in Middle Triassic (Lower Anisian) sedimentary rocks

silt size from a depth of 3000 m b.s.l. in the borehole Győrszemere-2 (south of the town Győr, W-Hungary) a massive occurrence of *Meandrospira pusilla* (HO) accompanied by fragments of Crinoids and Echinoids can be observed. (Core 31/1.) In addition, several individuals of *Ammodiscus* sp. *Earlandia tintinniformis* (MISIK), *Glomospirella* sp., *Spirorbis phlyctaena* BRÖNNIMANN—ZANINETTI and Ostracode shell fragments can be found in the thin sections. In the fine-crystallized silty marl of the core 31/2 numerous sections of *Cyclogyra* ? *mahajeri* BRÖNNIMANN, ZANINETTI, BOZORGNIA accompanied by several *Meandrospira pusilla* (HO) were observed. The association of microfossils recovered from the borehole Győrszemere-2 refers to the Lower Triassic (Campilian) age of the sequence (P. BRÖNNIMANN—L. ZANINETTI, 1972; P. BRÖNNIMANN—L. ZANINETTI—F. BOZORGNIA, 1972).

In numerous places of E-Hungary the hydrocarbon prospecting boreholes discovered Lower Triassic (Campilian) sedimentary rocks containing *Meandrospira pusilla* (HO). The reddish-brown, hard, slightly laminated, breccia-like, fractured, dolomitic, sandy, locally sericitic shale recovered by the core 27 (4924—4929 m) of the ultradeep well Makó-2 (SE-Hungary) contains sections of a number of *Meandrospira pusilla* (HO) beside some Ostracode shell fragments. The age of the sample is undoubtedly Lower Triassic.

In the Mezőkeresztes area, in the southern foreground of the Bükk Mts. (NE-Hungary), *Meandrospira pusilla* (HO)-bearing Lower Triassic formations have been intersected by several boreholes (Mezőkeresztes-8—29—50) in a

depth of 1400—1500 m b.s.l. This Campilian sequence consists of a reddish-brown, slightly sandy, pressed, limonitic limestone with calcite veins and reddish-brown, locally sandy shale intercalations of definitely schistose character. Thin sections from both the limestone and the shale intercalations contain sections of *Meandrospira pusilla* (HO). In addition, shell fragments of Echinoids, Molluscs and Ostracods as well as sections of *Glomospira* sp. were observed.

In the Northern part of the Danube-Tisza Interfluve Lower Triassic (Campilian) sedimentary rocks containing a number of *Meandrospira pusilla* (HO) sections were recovered from two boreholes (Bugyi-5, Pánd-1).

The thin section (borrowed by the courtesy of K. SZEPESHÁZY, Hungarian State Geological Institute) prepared from the dark-grey, pressed, silty schistose dolomitic marl recovered from the core 5 of the hydrocarbon prospecting borehole Bugyi-5 (680.5—692 m) contains a number of *Meandrospira pusilla* (HO) sections. The Lower Triassic (Campilian) age of the sample is beyond doubt. As for the autochthon character and tectonic position of the rock, problems arise because thin sections from the rocks of the core 3 taken in a structurally position higher by about 200 m (i.e. in 486—488 m) show an extremely impressive Upper Permian marine microfauna and algal remnants with the following species (S. PANTIĆ, 1969; oral communication): *Hemigordius harltoni* (CUSHMAN—WATERS), *Agathammina pusilla* (GEINITZ), *Glomospira* sp. *Gymnocodium bellerophontis* (ROTHPLETZ), *Permoloculus fragilis* (PIA). The rocks underlying the Lower Triassic and recovered by core 6 (818—821 m) give a fairly heterogeneous sequence containing unfossiliferous dolomite pieces, Upper Permian, *Gymnocodium bellerophontis* (ROTHPLETZ)-bearing silty dolomite fragments, as well as well rounded pebbles of Upper Permian *Miliolina*-limestones.

In the borehole Pánd-1, a 532 m thick Mesozoic sequence of uncertain age was intersected from a depth of 1000 m b.s.l. downward, under Miocene layers. The upper 40 m of this sequence (1068—1108 m) consists of strongly tectonized, grey, hard, foliated, stressed, silty clayey marl, silty marl, limy marl and polymict breccia. This breccia cemented by clayey, dolomitic material is composed of grey, less rounded particles of marl, limy marl and of limestones with calcite veins. The diameter of the particles varies between 5—80 mm. The sandy clayey marl layers are marked by a massive occurrence of microfossils characteristic of the Lower Triassic: i.e. individuals of species *Meandrospira pusilla* (HO). Additionally, in thin sections the sections of Echinoid and Ostracode shell fragments as well as fragments of *Crinoid* *pelms* are characteristic. In spite of the Lower Triassic microfauna, the presence of the polymict breccia casts some doubts to the autochthon character of the sequence. Nevertheless, it seems to be sure that a heavily tectonized area was intersected by this borehole. The Lower Triassic (Campilian) formations

underlying the series discussed above in a thickness of 490 m are fairly various. Layers of dark-grey, compact, slightly sandy, foliated clayey marl, limy marl and oolitic limestone (core 9: 1148—1150.5 m) all of them intersected by bright slip surfaces and calcite veins are interbedded by dark-gray, compact, steeply dipping (dip is as high as 45°) limestones with calcite veins. Fossil content of this series is as follows: *Meandrospira pusilla* (HO), *Dentalina* sp., *Earlandia* sp. Mollusca, Echinoid and Ostracoda shell fragments.

Individuals of *Meandrospira pusilla* (HO) from Anisian sedimentary rocks are known in one borehole only. In the borehole Budafa-502 (Zala-County, W-Hungary) several sections of *Meandrospira pusilla* (HO) accompanied by several sections of *Meandrospiranella* sp. and *Trochammina* cf. *almtalensis* KOEHN—ZANINETTI were observed as characteristic additional components of the microfauna in the gray, massive, finely crystallized oolitic limestone recovered by the core 33 in a depth of 3476—3478 m b.s.l. In any case, the Anisian age of this interval is evidenced by the less frequent occurrence of the *Meandrospira pusilla* (HO) and by the characteristic accompanying microfauna as well.

Paleontological description

Phylum: RHIZOPODA

Classis: Foraminifera

In classifying the species the system of R. A. LOEBLICH,—H. TAPPAN, (1964) has been used.

Family: TROCHAMMINIDAE

Genus: TROCHAMMINA PARKER—JONES, 1859

Trochammina cf. *almtalensis* KOEHN—ZANINETTI

Pl. III, Fig. 11

- 1968 *Trochammina almtalensis* — KOEHN—L. ZANINETTI, p. 3. nomen nudum
- 1969 *Trochammina almtalensis* — KOEHN—L. ZANINETTI, p. 38, pl. 5, fig. E, F; Textfig. 6/A-P.
- 1971 *Trochammina almtalensis* — BAUD, A.—L. ZANINETTI—P. BRÖNNMANN, p. 86, pl. 2, fig. 9; textfig. 3p, q.
- 1971 *Trochammina almtalensis* — PREMOLI—I. SILVA, p. 333, pl. 26, fig. 1—4.
- 1972 *Trochammina almtalensis* — PANTIĆ, S.—J. P. RAMPNOUX (no illustrated)
- 1972 *Trochammina* sp. — UROSEVIĆ, D.—Z. RADOVANOVIĆ, p. 30, pl. 1, fig. 8—9.

- 1972 *Trochammina almtalensis* — SAMUEL, O.—K. BORZA—E. KÖHLER, pl. 17, fig. 1—3.
- 1973 *Trochammina almtalensis* — GAZDZICKI, A.—K. ZAWIDZKA, pl. 1, fig. 5—7.
- 1973a *Trochammina almtalensis* — BRÖNNIMANN, P.—J. P. CADET—L. ZANINETTI, p. 308, pl. 23, fig. 3.
- 1973b *Trochammina almtalensis* — BRÖNNIMANN, P.—J. P. CADET—L. ZANINETTI, p. 467, pl. 48, fig. 14—15.
- 1973 *Trochammina almtalensis* — JENDREJÁKOVA, O. (No illustrated)
- 1974 *Trochammina almtalensis* — EFIMOVA, N. A. (No illustrated)
- 1975 *Trochammina almtalensis* — GAZDZICKI, A.—J. TRAMMER—K. ZAWIDZKA, pl. 7, fig. 7.
- 1976 *Trochammina almtalensis* — ZANINETTI, L. p. 113, pl. 14, fig. 1—9, pl. 23, fig. 8.

An individual of the species *Trochammina* cf. *almtalensis* KOEHN—ZANINETTI was found in the Middle-Triassic, Anisian limestone of the core 33/4 (3476.0—3478.0 m) of the hydrocarbon exploring borehole Budafa-502 (Zala County, W-Hungary).

The high of the oblique longitudinal section is 0.3 mm, the diameter of the basis equals to 0.2 mm, the testwall is as thick as 0.012 mm. The coiled-up test consists of 4 whorls, the chambers are inflated and grow quickly during the evolution. The wall is calcareous, strongly recrystallized, microgranulated.

Its accompanying faunal assemblage is characterized by few individuals of the *Meandrospira pusilla* (HO) and by an *Meandrospiranella* sp. section found in the oolitic limestone.

This species was described by L. KOEHN—ZANINETTI, (1969) from the Upper Anisian Reifling Limestone of the Northern Limestone Alps.

F a m i l y : MORAVAMMINIDAE

G e n u s : EARLANDIA PLUMMER, 1930

Earlandia tintinniformis (MISIK)

Pl. III, fig. 2

- 1971 *Aeolisaccus tintinniformis* — MISIK, M. p. 169, textfig. 1—7.
- 1972 *Aeolisaccus* sp. — SAMUEL, O.—K. BORZA—E. KÖHLER, pl. 107, fig. 1—3.
- 1972 *Earlandia tintinniformis* — BRÖNNIMANN, P.—L. ZANINETTI—F. BOZORGIA, p. 871, pl. 3, fig. 1—5, 8—9, 12; pl. 4, fig. 16—17.
- 1972 *Earlandia tintinniformis* — PANTIĆ, S.—J. P. RAMPNOUX (No illustrated)

- 1972a *Earlandia tintinniformis* — ZANINETTI, L.—P. BRÖNNIMANN—
A. BAUD, (No illustrated)
- 1972b *Earlandia tintinniformis* — ZANINETTI, L.—P. BRÖNNIMANN—
A. BAUD, p. 471, pl. 6, fig. 20—26, 28—
31; pl. 7, fig. 12—14; pl. 11, fig. 1—2.
- 1972 *Earlandia tintinniformis* — ZANINETTI, L.—P. BRÖNNIMANN—F.
BOZORGNI—H. HUBER, p. 228, pl. 1,
fig. 16—17.
- 1972 *Earlandia tintinniformis* — BRÖNNIMANN, P.—L. ZANINETTI—F.
BOZORGNI—H. HUBER (No illus-
trated)
- 1973b *Earlandia tintinniformis* — BRÖNNIMANN, P.—J. P. CADET—L. ZA-
NINETTI, p. 467, pl. 47, fig. 6.
- 1974 *Earlandia tintinniformis* — BRÖNNIMANN P.—L. ZANINETTI—
A. MOSTAGHIAN—H. HUBER, p. 22, pl.
1, fig. 1—2, 11—12.
- 1974 *Earlandia tintinniformis* — ZANINETTI, L.—P. BRÖNNIMANN (No
illustrated)
- 1975 *Earlandia tintinniformis* — GAZDZICKI, A.—J. TRAMMER—L. ZA-
WIDZKA, pl. 8, fig. 6.
- 1975 *Earlandia tintinniformis* — BRÖNNIMANN, P.—J. E. WHITTAKER—
ZANINETTI, L. p. 13, pl. 3, fig. 3—4.
- 1976 *Earlandia tintinniformis* — ZANINETTI, L. p. 120, pl. 3, fig. 20.

Several examples of this species were observed from the Lower Triassic (Upper Campilian) silty limestone recovered by the core 31/1 (3059.5—3062.0 m) of the borehole Győrszemere-2 and by the core 9 (1145.0—1150.5 m) of the borehole Pánd-1.

The average outer diameter of the coniform tube amounts to 0.072 mm, its length is 0.3 mm. The wall consisting of fine-grained calcite is as thick as 0.007 mm.

In the borehole Győrszemere-2 its faunal assemblage is characterized by a massive occurrence of Echinoid fragments as well as by the presence of several sections of the species *Meandrospira pusilla* (HO), *Glomospirella* sp., *Ammodiscus* sp., *Spirorbis phlyctaena* BRÖNNIMANN, ZANINETTI.

This species was described by M. MISIK, (1971) from the Ladinian carbonaceous formations of the Low Tatra Mountains.

F a m i l y : FISCHERINIDAE

G e n u s : CYCLOGYRA WOOD, 1842

Cyclogyra cf. mahajeri BRÖNNIMANN, ZANINETTI, BOZORGNIA

Pl. II, fig. 2—3

- 1972 *Cyclogyra ? mahajeri* — BRÖNNIMANN, P.—L. ZANINETTI—F. BOZORGNIA, p. 869, pl. 3, fig. 8—9, 11—15; pl. 4, fig. 2, 4?, 8, 10—11, 18—19; textfig. 2 A-I.
- 1973 *Cyclogyra mahajeri* — BRÖNNIMANN, P.—L. ZANINETTI—A. MOSHTAGHIAN —H. HUBER, p. 15, pl. 1, fig. 1—3, 19; pl. 2, fig. 19—21.
- 1975 *Cyclogyra mahajeri* — DAGER, Z.—L. ZANINETTI (No illustrated)
- 1976 *Cyclogyra nahajeri* — ZANINETTI, L. p. 133, pl. 1, fig. 3—4.

A considerable number of the individuals of this species was found in the Lower Triassic (Upper Campilian) fine-silty marl recovered by the core 31/2 (3059.0—3062.0 m) of the borehole Győrszemere-2.

The diameter of the test coiled up planispirally varies between 0.1—0.17 mm, with an axial width of 0.04—0.05 mm. The wall (accompanied by small, dark brown patches of probable organic origin) is as thick as 0.007—0.01 mm, recrystallized, calcareous, imperforata and strikingly different with its dark colours from the rock material. The height of the whorls increases during evolution. The diameter of proloculum is 0.026 mm; the height of the first whorl amounts to 0.015 mm, the second does to 0.025 mm; while that of the lastone is 0.036 mm including the wall.

Its faunal assemblage is characterized by a frequent occurrence of the species of the genus *Ammodiscus* as well as by some sections of the species *Meandrospira pusilla* (HO) in thin sections.

This species was described from the Lower Erika Formation (Scythian = Lower Triassic) in Iran (P. BRÖNNIMANN,—L. ZANINETTI,—F. BOZORGNIA, 1972).

G e n u s : MEANDROSPIRA LOEBLICH—TAPPAN, 1946

Meandrospira pusilla (HO)

Pl. I, fig. 2—6, 8;

Pl. II, fig. 5—8;

Pl. III, fig. 3—9.

- 1955 *Small Foraminifera* — HAGN, H. pl. 1, fig. 2; pl. 2, fig. 1.
- 1959 *Trochamminoides pusillus* — HO, Y. p. 416, pl. 7, fig. 18—29; pl. 8, fig. 1—5.
- 1959 *Trochamminoides flosculiformis* — HO, Y. p. 416, pl. 8, fig. 6—10.
- 1959 *Trochamminoides cheni* — HO, Y. p. 416, pl. 8, fig. 16—19.

- 1959 *Trochamminoides insolitus* — HO, Y. p. 416, pl. 8, fig. 11—15.
- 1964 *Citaella iulia* — PREMOLI—I. SILVIA, p. 66, pl. 48, fig. 1—20; pl. 49, fig. 1—20; pl. 50, fig. 1—7; pl. 51, fig. 5.
- 1965 *Meandrospira iulia* — KOCHANSKY—V. DEVIDE—S. PANTIĆ, p. 20, pl. 2, fig. 1—15; pl. 3, fig. 1, 4—8.
- 1966 *Meandrospira iulia* — MASOLI, M. pl. 2—3.
- 1967 *Meandrospira iulia* — PANTIĆ, S. p. 89, pl. 1, fig. 1.
- 1967 *Meandrospira iulia* — PANTIĆ, S.—S. MOJSILOVIĆ (No illustrated)
- 1967a *Meandrospira iulia* — SALAJ, J.—A. BIELY—J. BYSTRICKY, p. 129, pl. 2, fig. 1.
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—L. ZANINETTI, p. 209, pl.
30, fig. 1—3, 7. |
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VA, (No illustrated) |
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MANN, (No illustrated) |
| 1974 | <i>Meandrospira iulia</i> | — EFIMOVA, N. A. pl. 3, fig.
11—14. |
| 1974 | <i>Meandrospira iulia</i> | — RAMOV, A. (No illustrated) |
| 1975 | <i>Meandrospira pusilla</i> | — GAZDZICKI, A.—J. TRAMMER
—K. ZAWIDZKA, (No il-
lustrated) |
| 1975 | <i>Meandrospira pusilla</i> | — ZANINETTI, L.—P. BRÖNNI-
MANN, p. 261—262, pl.
34, fig. 1—15. |
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TO—M. PASINI, p. 431, pl.
49, fig. 3. |
| 1975 | <i>Meandrospira pusilla</i> | — DAGER, Z.—L. ZANINETTI,
(No illustrated) |
| 1976 | <i>Meandrospira pusilla</i> | — ZANINETTI, L. p. 135, pl. 1,
fig. 5—11. |
| 1976 | <i>Meandrospira cheni</i> | — SALAJ, J. (in press) |
| 1976 | <i>Meandrospira pusilla</i> | — TOLLMANN, A. textfig. 19. |

Individuals of *Meandrospira pusilla* (HO) are known from Lower Triassic and Middle Triassic formations revealed by hydrocarbon exploring boreholes (boreholes Ortaháza-11; Ortaháza West-1; Győrszemere-2; Makó-2; Mezőkeresztes-8, -29, -50; Bugyi-5; Pánd-1 and Budafa-502 respectively) (in this publication).

The small nearly spherical test with a diameter of 0.080—0.150 mm consists of a proloculum and an undivided tubelike chamber coiling up in a zig-zagging way. As a result of this rather intricate mode of coiling the individual sections are in most cases totally incomparable. The most frequent diameter values are 0.11—0.12 mm. The wall is 0.01 mm thick, totally recrystallized, calcareous, imperforata and strikingly different with its dark, dark-brown, almost black colours in transparent light from the rock material.

The small Foraminifers described and illustrated by H. HAGN, (1955) from the Werfenian oolitic limestones of the Bavarian Alps belong also to the species *Meandrospira pusilla* (HO).

From 1966 a number of studies have been successively published de-

scribing new and new *Meandrospira pusilla* (HO) bearing layers in most different areas of Europe, Asia, Africa, emphasizing at the same time the horizon-marking character and biostratigraphic significance of this species.

G e n u s : MEANDROSPIRANELLA SALAJ, 1967

Meandrospiranella sp.

Pl. III, fig. 10.

Single individual was found in Middle Anisian limestone recovered by the core 33 (3476.0—3478.0 m) of the borehole Budafa-502, Zala-County, W-Hungary.

The individual coiling up in a Meandrospira-like zig-zagging way tends to uncoil after 2.5 whorls. The test wall is totally recrystallized and is strikingly different as compared to the host rock. Its length is 0.25 mm, width is 0.17 mm.

The microfaunal assemblage is characterized by few individuals of *Meandrospira pusilla* (HO) and by a *Trochammina* cf. *almtalensis* KOEHN—ZANINETTI section.

The only individual recovered here resembles mostly to the species *Meandrospiranella samueli* SALAJ, but the species characteristics of it could not be unambiguously identified and its measures are also less as compared to those of the species *Meandrospiranella samueli* SALAJ.

P h y l u m : ANNELIDA

C l a s s i s : Polychaetia

O r d o : Sedentarida

F a m i l y : SERPULIDAE

G e n u s : SPIORBIS DAUDIN, 1800

Spirorbis phlyctaena BRÖNNIMANN, ZANINETTI

pl. II, fig. 1

- | | |
|----------------------------------|---|
| 1959 <i>Gasteropode</i> ? | — <i>Microfacies italiane</i> , pl. 7, fig. GP. 291, sez. 12588. |
| 1972 <i>Spirorbis phlyctaena</i> | — BRÖNNIMANN, P.—L. ZANINETTI, p. 75, pl. 10, fig. 1—9; pl. 11, fig. 1—15; pl. 12, fig. 2, 4—6, 8—13. |
| 1972 <i>Spirorbis phlyctaena</i> | — ZANINETTI, L.—P. BRÖNNIMANN—F. BOZORGIA—H. HUBER, p. 239, pl. 4, fig. 4, 10. |
| 1973 <i>Spirorbis phlyctaena</i> | — BRÖNNIMANN, P.—L. ZANINETTI—A. MOSHTAGHIAN—H. HUBER, p. 6, pl. 1, fig. 4—5, 9, 16—17. |

- 1974 *Spirorbis phlyctaena* —BAUD, A.—P. BRÖNNIMANN—L. ZANINETTI; p. 207, pl. 30, fig. 13.
 1975 *Spirorbis phlyctaena* —GAZDZICKI, A.—J. TRAMMER—K. ZAWIDZKA, pl. 11, fig. 12.

A *Spirorbis phlyctaena* BRÖNNIMANN, ZANINETTI was found in the thin section of Lower Triassic (Campilian) limestone recovered by the core 31/1 (3059.5—3062.0 m) of the borehole Győrszemere-2, south of the town Győr, NW-Hungary. The diameter of the flat dorsal side of the oblique axial section is 0.795 mm, the axial section is as high as 0.396 mm. The thickness of the test-wall varies between 0.007—0.015 mm. The thin calcareous wall consists of alternating light and dark platesets.

In oblique section the tubular test coiled up in a slightly Trochospira-like way, the outline of the diagonal section is rounded in the ventral side and eclipsed in the dorsal side. The sediment filling the tubular test is the same as the host rock. In the sediment filling the tubular test of the *Spirorbis phlyctaena* BRÖNNIMANN, ZANINETTI shown in Fig. 1 of the Plate II a section of a fragmentary *Meandrosira pusilla* (HO) can be observed.

Its microfaunal assemblage is characterized by the massive occurrence of Echinoid fragments as well as by the presence of some sections of *Meandrosira pusilla* (HO), *Glomospirella* sp., *Earlandia tintinniformis* (MISIK), *Ammodiscus* sp. This microfaunal assemblage is known from the Dolomites and from the Alborz-Mts., Iran.

This species was described by P. BRÖNNIMANN—L. ZANINETTI (1972) from the Lower Triassic of the Dolomites (Castalunga).

P h y l u m : ECHINODERMATA
 C l a s s i s : *Holothuroidea*

In classifying this fossil the system of L. FRIZZEL—H. EXLINE, (1966) was used.

F a m i l y : STICHOPITIDAE

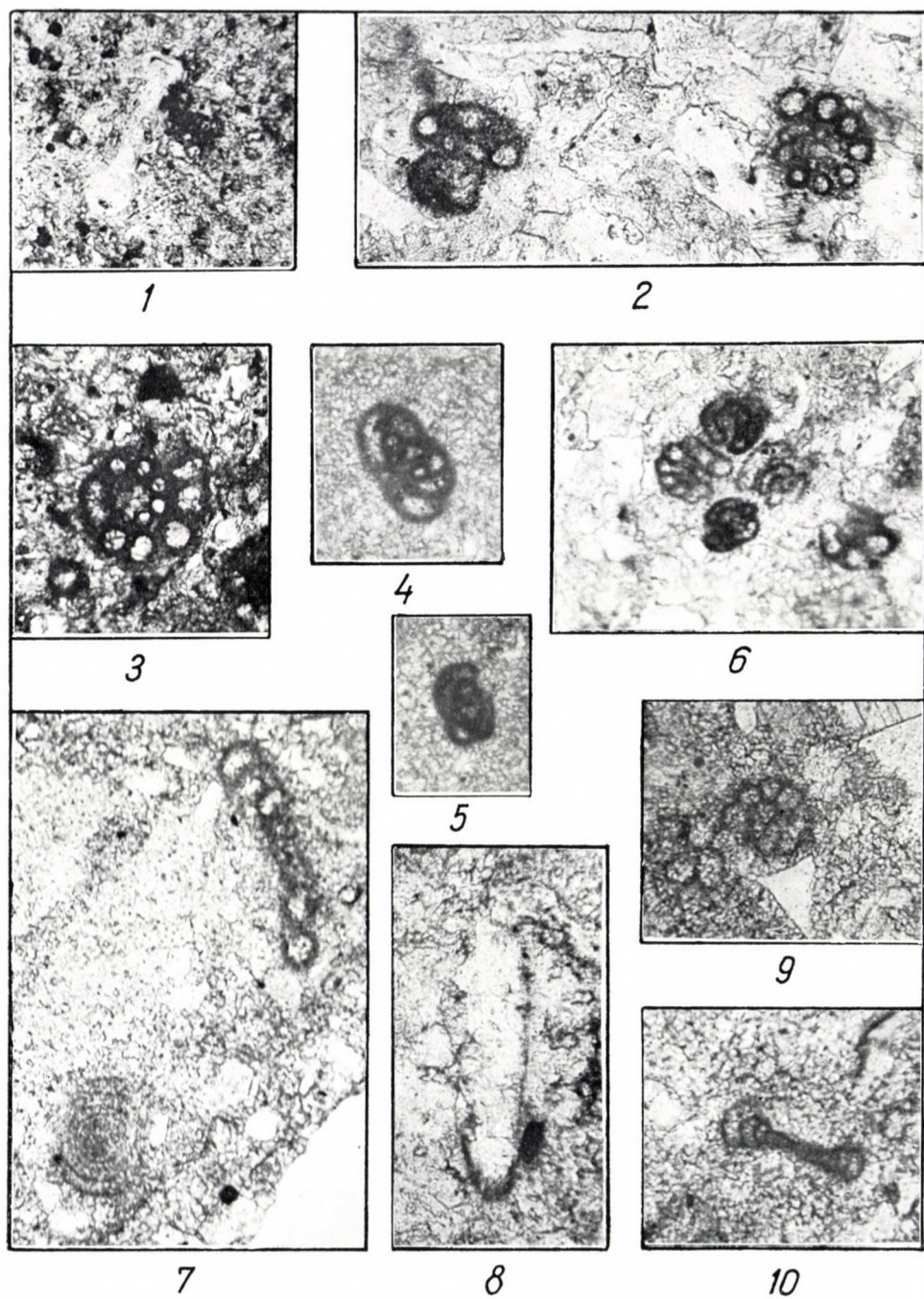
G e n u s : BINOCULITES DEFLANDRE-RIGAUD, 1952

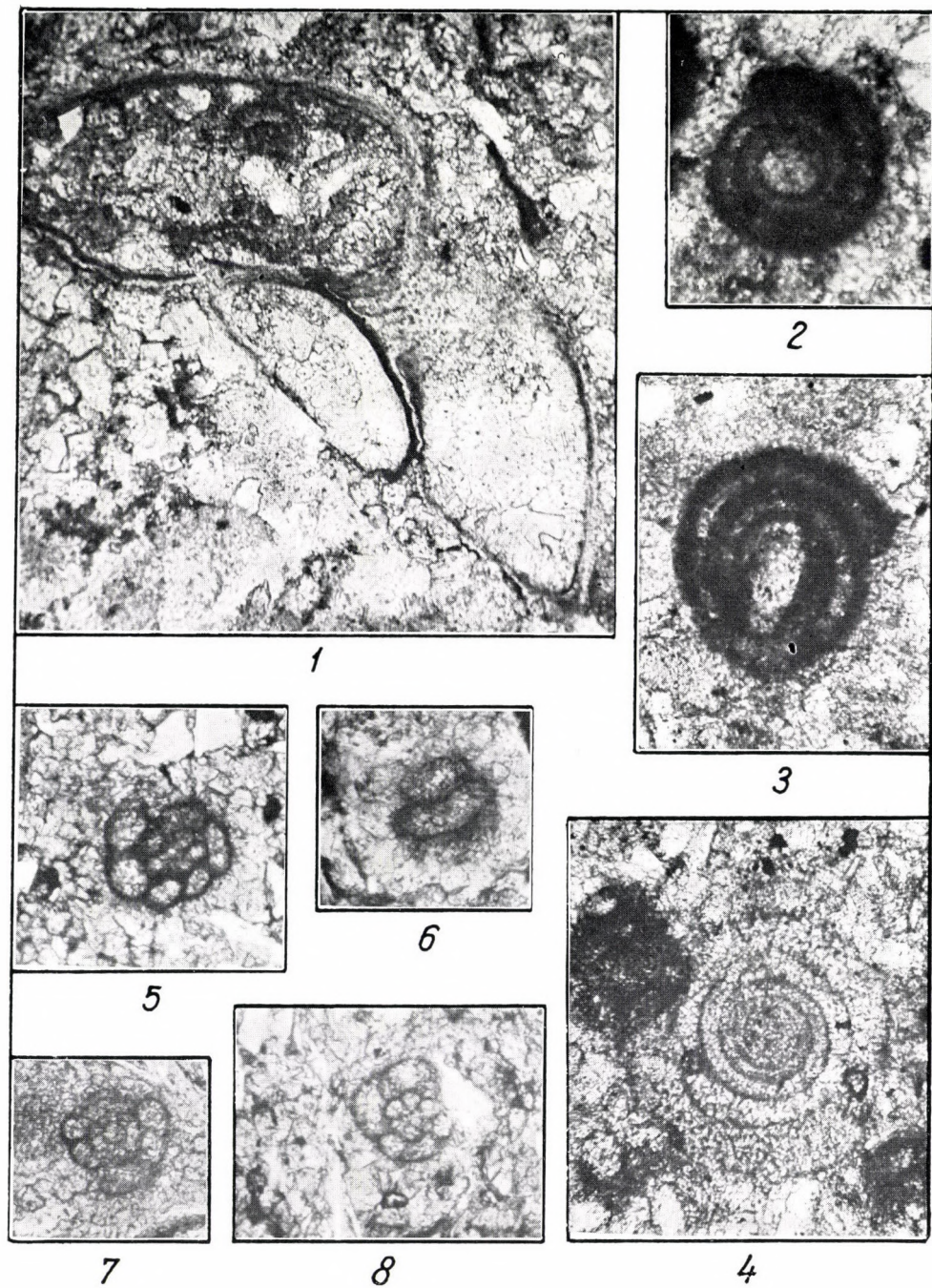
Binoculites sp.

Pl. I, fig. 1

One complete and several fragmentary individuals were found in the Lower Triassic (Campilian) cherty, silty claystone recovered by the core 33/2 (3806.0—3812.0 m) of the borehole Ortaháza West-1, W-Hungary.

It is a rod-like form with a small disc at both ends. One of these discs shows a rather compressed irregular shape. At the end of the axis of the rod a trace of a gap-like hole can be observed. The disc at the opposite end of the

*Plate I.*

*Plate II.*

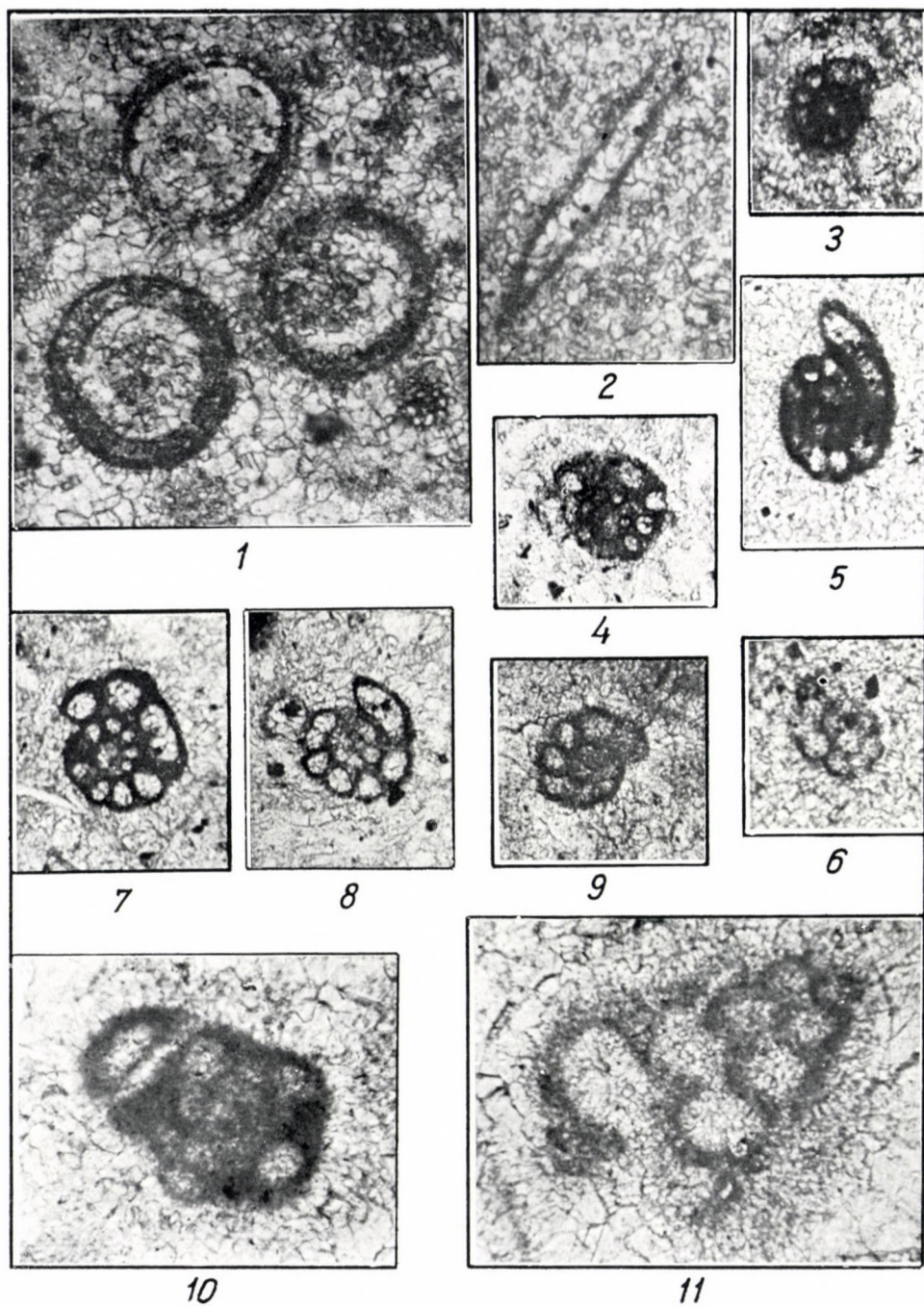


Plate III.

Plate I

1. Section of a *Binoculites* sp. (Holothuroidea) in cherty silty clayey marl recovered by the core 33/2 (3806.0—3812.0 m) of the borehole Ortaháza West 1. Lower Triassic (Campilian). Length: 0.2 mm
2. Sections of *Meandrosira pusilla* (HO) in calcareous marl recovered by the core 25/a (2477.0—2478.5 m) of the borehole Ortaháza-11. The diameter of the individual in the right sides is 0.11 mm
3. Section of a *Meandrosira pusilla* (HO) in cherty clayey marl recovered by the core 34/d (3806.0—3812.0 m) of the borehole Ortaháza West-1. Lower Triassic (Campilian). Diameter = 0.2 mm
- 4—5. Sections of *Meandrosira* ? sp. in dolomitic marl recovered by the core 34/c (3806.0—3812.0 m) of the borehole Ortaháza West-1. Lower Triassic (Campilian). Diameter: 0.22 mm (Fig. 4); and/or 0.12 mm (Fig. 5)
6. Sections of *Meandrosira pusilla* (HO) in calcareous marl recovered by the core 25/2 (2477.0—2478.5 m) of the borehole Ortaháza-11. Lower Triassic (Campilian). The diameter of the section in the left side: 0.11 mm
7. Sections of *Ammodiscus* cf. *parapriscus* HO; *Glomospirella* ? sp. and *Echinoid* fragments in anhydritic limestone recovered by the core 31/1 (3059.5—3062.0 m) of the borehole Győrszemere-2. Lower Triassic (Campilian). The diameter of the *Ammodiscus* cf. *parapriscus* HO is 0.18 mm
8. Section of *Earlandia* ? sp. in anhydritic limestone recovered by the core 31/1 (3059.5—3062.0 m) of the borehole Győrszemere-2. Lower Triassic (Campilian). Length: 0.25 mm
9. Section of a *Meandrosira pusilla* (HO) in anhydritic limestone recovered by the core 31/1 (3059.5—3062.0 m) of the borehole Győrszemere-2. Lower Triassic (Campilian). Diameter: 0.09 mm
10. Section of an *Ammodiscus* cf. *parapriscus* HO in anhydritic limestone recovered by the core 31/1 (3059.5—3062.0 m) of the borehole Győrszemere-2. Diameter: 0.15 mm

Plate II

1. Section of a *Spirorbis phlyctaena* BRÖNNIMANN, ZANINETTI in anhydritic limestone recovered by the core 31/1 (3059.5—3062.0 m) of the borehole Győrszemere-2. Lower Triassic (Campilian). The height of the axial section is 0.396 mm
- 2—3. Sections of *Cyclogyra* cf. *mahajeri* BRÖNNIMANN, ZANINETTI, BOZORGIA in the silty marl recovered by the core 31/2 (3059.5—3062.0 m) of the borehole Győrszemere-2. Lower Triassic (Campilian). Diameter: 0.17 mm (Fig. 2); and/or 0.22 mm (Fig. 3)
4. Section of an *Ammodiscus* sp. in the silty marl recovered by the core 31/2 (3059.5—3062.0 m) of the borehole Győrszemere-2. Lower Triassic (Campilian). Diameter: 0.26 mm
- 5—8. Sections of *Meandrosira pusilla* (HO) in dolomitic shale recovered by the core 27/a (4924.0—4929.0 m) of the borehole Makó-2. Lower Triassic (Campilian). Diameters: 0.1 mm (Fig. 5); 0.07 mm (Fig. 6); 0.09 mm (Fig. 7); 0.08 mm (Fig. 8)

Plate III

1. Thin section of oolitic limestone recovered by the core 9/a (1148.0—1150.5 m) of the borehole Pánd-1. Lower Triassic. Diameter: 0.26 mm
2. A section of an *Earlandia tintinniformis* (MISK) in calcareous marl recovered by the core 9/f (1148.0—1150.5 m) of the borehole Pánd-1. Lower Triassic (Campilian). Length: 0.275 mm
3. A section of a *Meandrosira pusilla* (HO) in limestone recovered by the core 7/3 (1100.0—1105.0 m) of the borehole Pánd-1. Lower Triassic (Campilian). Diameter: 0.1 mm
4. A section of a *Meandrosira pusilla* (HO) in limestone recovered by the core 6/2 (1066.0—1070.0 m) of the borehole Pánd-1. Lower Triassic (Campilian). Diameter: 0.11 mm
- 5—6. Sections of *Meandrosira pusilla* (HO) individuals in the shale recovered by the core 4 of the borehole Mezőkeresztes-29. Lower Triassic (Campilian). Diameters: 0.12 mm (Fig. 5); and/or 0.11 mm (Fig. 6)
- 7—8. Sections of *Meandrosira pusilla* (HO) individuals in the dolomitic schistose marl recovered by the core 5 (688.5—692.5 m) of the borehole Bugyi-5. Lower Triassic (Campilian). Diameters: 0.135 mm (Fig. 7); and/or 0.11 mm (Fig. 8)
9. A section of a *Meandrosira pusilla* (HO) in the limestone recovered by the core 33/a (3476.0—3478.0 m) of the borehole Budafa-502. Middle Triassic (Lower Anisian). Diameter: 0.095 mm
10. A section of a *Meandrosiranella* sp. in the limestone recovered by the core 33/a (3476.0—3478.0 m) of the borehole Budafa-502. Middle Triassic (Lower Anisian). Length: 0.25 mm
11. A section of a *Trochammina* cf. *almtalensis* KOEHN—ZANINETTI in the limestone recovered by the core 33/a (3476.0—3478.0 m) of the borehole Budafa-502. Height: 0.3 mm; basal diameter: 0.2 mm

rod is rounded, fairly circular, with a point-like hole at the end of the axis of the rod, close to its margin. Its length amounts to 0.2 mm. The thinnest part of the rod is its central part, where its width is 0.02 mm.

Probably this remnant should be considered as a new species since it could not be identified with the Holothuroidea species presently known. The fact, however, that only one complete individual was observed in thin section, does not allow a detailed and reliable description of the species characteristics.

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БИОСТРАТИГРАФИЧЕСКОЕ ЗНАЧЕНИЕ И РАСПРОСТРАНЕНИЕ В ВЕНГРИИ
ВИДА MEANDROSPIRA PUSILLA (HO) (FORAMINIFERA) В ТРИАСОВЫХ
ОТЛОЖЕНИЯХ, ВСКРЫТЫХ РАЗВЕДОЧНЫМИ СКВАЖИНАМИ НА НЕФТЬ И ГАЗ

А. БЭРЦИ-МАКС

Резюме

Работа на основании литературных данных кратко знакомит с распространением в нижнетриасовых осадочных породах Европы, Азии и Африки вида *Meandrosira pusilla* (HO), а также упоминает о находках данного вида и в среднем триасе. Представители *Meandrosira pusilla* (HO) обычно распространены в мелководно-морских образованиях и являются видом, хорошо приспособленным к изменениям физических условий морского дна, экологически выносливым. Биостратиграфическое значение вида большое,

так как при заданном вертикальном распространении (кампильский ярус) можно проследить на большой территории почти всюду в морских нижнетриасовых формациях.

Рассматриваются стратиграфические условия ниже- и среднетриасовых образований, содержащих *Meandrospira pusilla* (HO) и вскрытых разведочными скважинами на нефть и газ Венгрии, а также дается краткое описание вида *Meandrospira pusilla* (HO) и скудной, но характерной сопровождающей микрофауны. Для сопровождающей фауны в нижнетриасовых образованиях характерно наличие в небольшом количестве экземпляров видов *Ammodiscus* sp.; *Earlandia tintinniformis* (MISIK); *Cyclogyra* cf. *mahajeri* BRÖNNIMANN, ZANINETTI, BOZORGNIA; *Spirorbis phlyctaena* BRÖNNIMANN, ZANINETTI; *Glomospirella* sp. Автор из одной скважины запада Задунайского края описала новый вид под именем *Binoculites* nov. sp. (Holothuroidea). В среднетриасовых (анизийских) образованиях находящийся в небольшом количестве экземпляров вид *Meandrospira pusilla* (HO) сопровождает всего несколько *Trochammina* cf. *almtalensis* KOEHN-ZANINETTI; *Meandrospiranella* sp.

ÜBER BAU UND ENTWICKLUNG DER OSTALPEN MIT AUSBLICK AUF WESTUNGARN (Vortragsbericht)

Von

P. BECK-MANNAGETTA

GEOLOGISCHE BUNDESANSTALT, WIEN

Die neuen Untersuchungsergebnisse in den Ostalpen weisen auf eine wesentliche Erstreckung der Molasse unter dem Flysch mit Helvetikum und im Osten noch unter die Nördlichen Kalkalpen hin. Die tektonische Abkunft des Flysches aus der penninischen Zone der Westalpen und Hohen Tauern weicht einer näher gelegenen Herkunft am Nordrande der Kalkalpen im Raume von Wien ostwärts. An der Grenze nach Ungarn erscheint die Rechnitzer Schieferinsel mit penninischen Bündner Schiefern, die von der unterostalpinen Grobgneisserie überlagert wird. Das Wechselfenster ist mit unterostalpinem Semmeringmesozoikum verbunden und eine Fortsetzung in die Veporiden der Westkarpaten ist denkbar.

Von dem oberostalpinen Deckenstapel wird ein tieferes mittelostalpines Stockwerk abgetrennt, das eine zentralalpine Trias trägt. Das variszisch geprägte Altkristallin der Sau- und Koralpe wird normale Basis der oberostalpinen Grauwackendecken aufgefaßt. Der variszische Deckenbau der südlichen Grauwackenzzone wurde teils tektonisch-metamorph überprägt und noch in der variazischen Ära mit dem anscheinend kaledonischen Gleinalmkristallin verschmolzen. Jüngere alpidische Aufquetschungen ließen fensterartige Aufbrüche in Kor- und Saualpenkristallin erscheinen (Wolfsberg, Klienig etc.). Durch Fossilfunde ist der variszische (vor Unterkarbon) Decken- bis Schuppenbau in der südlichen Saualpe und damit auch in dem östlichen Gurktal (Glantaler Schuppenzone) erwiesen.

Die vereinzelt radiometrischen Altersbestimmungen lassen noch keine endgültige Aussage der gewonnenen Daten zu.

Die Ausgangslage der Lineation des Plattengneises der Koralpe weist in die im Norden der Alpen gelegene Moldanubisch-Morawischen Störungszone und der Deformationszustand ähnelt dem des Bitteschen Gneises. Die Torsionen dieser Lineation als Umstellungen sind der Ausdruck des Auseinanderweichens der Alpen in die Karpaten und Dinariden. Dem tektonischen Element des Deckenbaues der Alpen und Karpaten steht die magmatische Mobilisation im Inneren der Faltenbögen gegenüber.

Der Stoff des Themas könnte ein dickes Buch füllen; daher sollen in diesem Zusammenhang nur die neuesten Ergebnisse und auszugsweise einzelne Probleme berührt werden. Ansich schließt das Thema das Problem des Ostendes der Alpen mit ein, das allein aus Verständigungsgründen geographischerseits scharf abgegrenzt wird. Geologisch sind jedoch alle Übergänge in die Karpaten und Dinariden vorhanden.

Die subalpine Molasse zeigt in ihrem oligozänen Schichtanteil wesentliche fazielle Abweichung von dem Teil der Molasse, den auf der Böhmischen Masse transgrediert. Im Fenster von Rogatsboden (S. PREY, 1957) tritt dies deutlich in Erscheinung und in der Bohrungen Texing und Urmannsau (A. RUTTNER, 1963) die Überschiebungen der beiden Molasse-

zonen klar erkennbar geworden. Dadurch weiß man, daß die Molasse sogar unter die Kalkalpen hereinstreicht.

Die Flyschzone der Ostalpen wird im Westen von der penninischen Zone an der West-Ostalpengrenze abgeleitet. Im Osten jedoch hat S. PREY (1975) die Abkunft der Kahlenbergerschichten von der St. Veiter Klippenzone wahrscheinlich gemacht. Damit ändert sich die Vorstellung, wonach der penninische Trog von den Westalpen bis in die Südkarpaten ziehen sollte, ohne unter den Karpaten als Fenster aufzutauchen. Trotz der weiterhin durchwegs allochthonen Lagerung wäre der Flysch eher von einem Raum etwa nördlich der penninischen Klippenzone der Karpaten abzuleiten.

Die Entdeckung von Spongiennadeln in kieselingen Kalkschiefern der Rechnitzer Schieferinsel weisen auf ein mittelmäozoisches Alter dieser Schichten hin (H. P. SCHÖNLAUB, 1973). Damit ist das mesozoische Alter dieser Bündnerschieferfolge erwiesen. A. PAHR (1960) konnte die Überlagerung der Grobgneisserie auf den Rechnitzer Schiefer allseits nachweisen.

Weiterhin problematisch bleibt die kristalline Unterlage der Rechnitzer Schiefer: A. PAHR (1973) sieht in dem Wechselkristallin die Basis der Rechnitzer Schiefer. Im Wechselgebiet selbst werden die Schieferserien von einem unterostalpinen Semmeringmesozoikum transgressiv überlagert, dem jegliche Beziehung zu einer Bündnerschieferfazies fehlt. Unter einer Porphyroidserie (P. FAUPL, 1971) folgen immer höher metamorphe Gesteine einer »Metadiabasserie«, die ohne scharfe Grenze in die Wechselschiefer und -gneise übergehen. Im Süden ist eine Fensternatur der Wechselgesteine gegenüber der hangenden Grobgneisserie nicht nachweisbar, da dort die gleichen Gesteine, wie in der Grobgneisserie auftreten. Dieselben kristallinen Schiefer findet man auch im Liegenden der Rechnitzer Schieferserie. Das bedeutet, daß der kristalline Untergrund der Grobgneisserie (mit Semmeringmesozoikum) und des Wechsel Fensters (mit Altpaläozoikum und Semmeringmesozoikum) als Unterostalpin, im Gegensatz zur Rechnitzer Schieferinsel (mit Bündner Schiefer) die als Penninikum anzusehen ist.

Ähnliche Vorstellungen entwickelte S. PREY (1963) in den Gurktaler Alpen und dieselbe Situation scheint auch in den Westalpen zwischen der Basis des Helvetikums und des Penninikums im wesentlichen zu bestehen. Bei dieser Art der Betrachtung könnte man die Fortsetzung der Grobgneisserie und des Wechsels gegen Nordosten in den Veporiden der Karpaten sehen (siehe Kartenskizze).

Die zentralen Teile des Ostalpins wurde von A. TOLLMANN (1959) in ein Mittel- und Oberostalpin unterteilt, wobei letzterem die kristalline Basis vollkommen fehlen soll. Betrachtet man aber die Abfolgen genauer, so ist im Bereich der Saualpe trotz aller sekundärer Störungen ein schrittweiser Übergang des Altpaläozoikums in das Hochkristallin zu beobachten. Das Altpaläozoikum ist transgressiv mit den jüngeren Schichten verbunden. Dabei ist in diesem

Raum die Kontinuität der Deckschichten mit der kristallinen Basis sicher gegeben. Außerdem verlieren sich die alpidischen Deckengrenzen im voralpidischen Kristallin des Ostalpins und die Unterschiede der kristallinen Einheiten folgen anderen Grenzen, die nicht durch mesozoische Einschaltungen gekennzeichnet sind. Aus diesem Grund spricht E. CLAR (1975) von einer Digitation (Zweiteilung) des Oberostalpins in eine tiefere Einheit im zentralalpinen Raum und eine höhere mit dem gesamten oberostalpinen Deckenstapel. Da die Gliederung des Oberostalpins nicht allein auf die erkennbaren Lagerungsverhältnisse zurückgeführt werden darf, sondern die faziellen Unterschiede und Anklänge der mesozoischen Deckschichten eine wesentliche Rolle spielen, ist zu beachten, daß die tiefere zentralalpine Einheit stets mit einem zentralalpinen Mesozoikum verbunden ist, das faziell dem Unterostalpin eher verwandt ist, als den oberostalpinen Schichtfolgen (H. FLÜGEL, 1960). Deshalb möchte P. BECK-MANNAGETTA (1969) dieser Teilung in ein Ober- und Mittelostalpin aus letzteren Gründen allein im Norden zustimmen und die oberostalpinen Deckschichten mit der kristallinen Basis der Kor- und Gleinalpe verbinden. Als weiteren Hinweis hierfür könnte das Kristallin der Kaintaleckscholle an der Basis der Norischen Decke (A. HAUSER und O. FELSER, 1939; H. P. CORNELIUS, 1952) herangezogen werden, in dem sowohl Anteile des Kor- sowie des Gleinalmkristallins stecken und beide Teile von ihrem Ausgangsort tektonisch abgeschuppt sind. Der Veitscher Decke kommt dadurch eine generelle Lage unter der Norischen Decke zu, die wahrscheinlich bereits in jungvariszischer Zeit angelegt wurde.

Nach der altersmäßigen Fixierung des Blasseneckporphyroides als Untersilur durch G. FLAJS (1967) sind weitere wesentliche Altersbestimmungen durch Conodontenfunde im Altpaläozoikum der Zentralalpen zu erwarten. Durch Frau Dr. H. PRIEWALDER wurde der Ennstaler Quarzphyllit als Silur (H. PRIEWALDER und R. SCHUHMACHER, 1976) eingestuft, was für die Deutung der Tektonik der Niederen Tauern wichtig ist. Besondere Bedeutung kommt den später stark erweiterten Crinoidenfunden in der Phyllitzone der südlichen Saualpe zu, die ein Mittel-Devon bis Siluralter der Kalke ergeben haben (G. KLEINSCHMIDT, 1966; J. NEUGEBAUER, 1971). Dadurch wurde die Ansicht von P. BECK-MANNAGETTA (1957) bestätigt. Aus der Wiederholung der kalkigen Schichtfolgen mit Keratophyren und Graphitschiefern konnten G. KLEINSCHMIDT und J. NEUGEBAUER (1975) einen Decken-(Schuppen-)bau im epizonalen Bereich der Saualpe nachweisen, der vor der Metamorphose entstanden ist. Ähnliches hat P. BECK-MANNAGETTA (1959, 1960) im oberen Glantal (Mittelkärnten) in analogen Schichtfolgen angenommen (»Glantaler-Schuppenzone« 1960), was wiederum von H. HAJEK (1962) bezweifelt wurde.

Diese Ergebnisse in der südlichen Grauwackenzone könnte man dahingehend zusammenfassen: Die kalkigen altpaläozoischen Schichtfolgen des Devon und Silur der südlichen Grauwackenzone wurden durch eine Decken-

überschiebung von der ordovizischen Metadiabasserie (Magdalensbergserie) in variszischer Zeit vor der Metamorphose generell überschoben, wobei stellenweise im Liegenden innerhalb des kalkigen Altpaläozoikums ein Schuppen- bis Deckenbau entstand. Dieser prämetamorphe Deckenbau wurde in die drei (A. PILGER, 1975) Metamorphoseabschnitte des Saualpenkristallins verschiedenstark einbezogen und die Schichtfolgen weiterhin verstellt. Daher ist dieser Deckenbau im Hochkristallin als Überschiebung der Eklogitserie auf die Marmorserie (E. HABERFELLNER, 1937) nur undeutlich in der Saualpe erkennbar. Vor allem diese Überschiebungsfläche dürfte in der Koralpe zur Ausbildung des Plattengneisstockwerkes geführt haben. Die synmetamorphe Plattengneistektonik bewirkte eine Wiederholung (Vervielfältigung und Komplizierung) des prämetamorphen Deckenbaues, wie dies in der südwestlichen Koralpe am deutlichsten zum Ausdruck kommt. (A. KIESLINGER, 1928). Solche tektonische Wiederholungen nehmen jetzt A. PILGER und N. WEISENBACH (1975) in der Saualpe an. Die kataklastische Zerschering der Plattengneisgesteine (P. BECK-MANNAGETTA, 1956) geriet gegen Westen zu in den zunehmenden Wirkungsbereich einer postkataklastischen thermischen Metamorphose, die die Spannungen der zerriebenen Mineralien auslöschte und zur Umkristallisation in Glimmergneise etc. führte, ohne eine wesentliche fazielle Änderung der Mineralassoziation herbeizuführen.

Durch die Arbeiten in der Saualpe wurde das altbekannte Problem (F. HERITSCH, 1932) der Überlagerung des Koralm- (als »Kata-«) auf das Gleinalmkristallin (Meso-) mit anderen Worten wieder aufgerührt. Diese Deutung als Überschiebung eines »heißen Bügeleisens« befriedigt nicht, da zu den zahlreichen Unsicherheiten noch die Fragen nach Herkunft und Ausbildung des Kor- bzw. Saualmkristallins hinkommen. Z. B. Unberührt bleibt der Umstand, weshalb allein saure Ganggesteine in diesem Kristallin auftreten und gar keine granitischen Gneise; inwiefern eine Deutung als wurzellose venitische Metamorphose (P. BECK-MANNAGETTA, 1949) den Tatsachen gerecht wird, ist hiedurch nicht geklärt. Die Herkunft der pegmatioden Lösungen aus dem Sediment wird anscheinend auch angenommen (O. HOMANN, 1962); auch über eine mögliche mehrfache Mobilisation der leichtflüchtigen Stoffe als Pegmatoide wird hiedurch keine Auskunft erteilt.

Betrachtet man ein Querprofil der Ostalpen, so erscheint unweigerlich das Problem des Südendes der Böhmisches Masse. Das gleiche Problem ergibt sich bezüglich der Westkarpaten. Nirgends ist östlich der Westalpen zu sehen, daß das Kristallin innerhab der Ostalpen, oder Karpaten in Zusammenhang mit dem im Norden vorgelagerten Vorland stehe. Das Untertauchen des Vorlandes weist aber darauf hin, daß die Grenze innerhalb unter den nichtmetamorphen Teilen der Alpen liegt. Die Nordgrenze des karpatischen Kristallins ist in den Karpaten scharf durch den Südrand der Flyschzone markiert, der durch die pieninische Klippenzone gebildet wird: Im Wiener Becken ver-

schwindet jedoch dieser Streifen vermutlich unter die Kalkalpen. Würde man ihn etwa an der Grenze Kalkvoralpen-Kalkhochalpen gegen Westen weiterziehen, so könnte eine solche Linie die Grenze zwischen der Böhmisches Masse und der kristallinen Basis des alpinen Altkristallins darstellen.

In diesem Zusammenhang ist das mögliche Ostende der penninischen Hohen Tauern zu besprechen (Ch. EXNER, 1966). Die Strukturen des Oberostalpins (Gurktaler Decke) zeigen ein hakenförmiges Vorspringen des Ostteiles gegen Norden. Dadurch werden bis zur Pölslinie (als Forsetzung der Lavantaler Störungszone J. STINY, 1931) die tieferen Teile immer mehr gegen Nordosten gedrängt. Damit könnte ein Ausspitzen des penninischen Troges verbunden werden und diese Nordoststörung könnte in die verdeckte westliche Fortsetzung der pieninischen Klippenzone einmünden. Für die Rechnitzer Schiefer am Ostrande der Alpen müßte ein eigener Trog mit SE—NW-Streichen angenommen werden, der mit dem Penninikum der Hohen Tauern nicht in Verbindung stehen braucht.

Im Südosten dieser Tiefenlinie reihen sich die verschiedenen Kristallineinheiten an, die südwärts immer weniger von den alpidischen Trennfugen berührt werden, bis im zentralen Teil des steirischen Randgebirges (H. FLÜGEL, 1963) ein hauptsächlich unberührtes voralpidisches kristallines »Rückgrat« besonderer Stellung auftritt, die Westpannonische Masse (WPM), die das alpidische Orogen zum Auseinanderweichen nötigte (P. BECK-MANNAGETTA, 1967). Das fensterartige Auftauchen der Granat-Glimmerschiefer mit Granitgneisen unter dem Koralmkristallin (P. BECK-MANNAGETTA, 1951; A. PILGER und N. WEISSENBACH, 1965) ist mit partiellen diaphthoritischen Vorgängen gepaart, die auf eine alpidische Aufquetschung einer voralpidischen Struktur hinweisen.

Der Versuch die Altersdeutung der Gesteine und ihrer Metamorphose mit Hilfe der radiometrischen Altersbestimmung zu erfassen, hat im Ostalpin noch keine eindeutigen Ergebnisse gezeigt. Die Bestimmungen an Glimmern der Koralpe mit 80—115 Mill. J. (H. FLÜGEL, 1964) bedürfen weitere Hypothesen, die sich zu einer befriedigenden Einordnung entwickeln sollen (Abkühlungsalter? Mischalter?). Die Gesamtgesteinsanalyse des Pegmatites von Oberwölz ist besser mit den gegebenen geologischen Befunden in Übereinstimmung zu bringen (E. JÄGER und K. METZ, 1973). Die Alterseinstufung saurer Anteile des Gleinalmkomplexes (W. FRANK, 1975) zeigt prävariszische Anteile im Altkristallin der Ostalpen auf. Weitere Untersuchungen an Gesteinen der östlichen Zentralzone stehen bevor (Villacher Granitgneis, Wolfsberger Granitgneis, Plattengneis etc.). Im Kristallin südlich der Hohen Tauern ist nach Untersuchungen von R. OXBURGH und Mitarbeitern (1966) ein deutlicher Alterssprung zwischen der Metamorphose des Tauernkristallins (ca. 20—30 Mill. J.) und der altkristallinen Hüllen (Biotit: 80 Mill. J.) ersichtlich, wobei bisher noch größere Unterschiede angenommen wurden. Weiter

im Südwesten ergaben Untersuchungen im Altkristallin Werte, die mit einer kaledonischen Orogenese in Beziehung gebracht werden (F. SASSI, A. ZANFERRARI und G. ZIRPOLI, 1974). Ein sicheres Präkambrium ist im Kristallin der Ostalpen bisher noch nicht gefunden worden (N. GRÖGLER; M. GRÜNENFELDER und E. SCHROLL, 1965).

Der Deformationsstil der Plattengneise erinnert sehr an die teils diaphthoritische Deformation des morawischen Bittescher Gneises (P. BECK-MANNAGETTA, 1947). Die tordierten Lagen der Lineationen des Plattengneises weisen im Norden gegen Nordosten, im Süden gegen Südosten; insofern wird der Bau des Ostendes der Ostalpen vorgezeichnet. Diese Torsionen der Lineationen aus der Nord-Südrichtung stehen in Zusammenhang mit einem bis einige kilometerbreiten Muldenbau der Plattengneiskomplexe (P. BECK-MANNAGETTA, 1970). Mit den Torsionen ist kein mineralfazieller Wechsel (retrograder Metamorphosesprung) verbunden; daher haben die alpidischen Beanspruchungen bereits diesen voralpidischen Bau angetroffen. Diese Stellung des variszisch geprägten Koralmkristallins erzeugte daher das eigenartige Auseinanderweichen der Ostalpen gegen Nordosten und Südosten. Solcherart wurden alte Anlagen in verschiedenen Stücken der Ostalpen bis in das Jungtertiär an Störungen wiederbelebt, wodurch der morphologisch bedeutsame Bogen des steirischen Randgebirges entstand (H. FLÜGEL, 1963). Tiefe jungtertiäre Kohlenmulden grenzen im Westen und Nordwesten diesen ab.

Verläuft gegen Nordosten der Gebirgszug ins Deckengebirge der Karpaten, so zerschlägt sich diese Klammer gegen Süden und Osten in einzelne, tektonisch begrenzte, Ost-Westverlaufende Riegel und Streifen, die stets gegen Nordosten zu umbiegen. Bedeutende Störungslinien (Raab-, Balaton- und Zagreblinie) zeigen den Verlauf dieser Schollen im Pannonischen Becken an, die im Südwesten in die südöstliche Fortsetzung der Periadriatischen Naht einmünden (Gy. WEIN, 1973).

Der durch Einengung erzwungene alpidische Deckenbau der Ostalpen weicht somit eher einer auf Zerrung beruhenden Schollenbildung im Untergrund des Pannonischen Beckens. Die tangentialen Momente der alpidischen (Kreide bis Mittel-Eozän) Deckenbildungen geben den Innenraum der Karpaten für ein Aufdringen magmatischer Schmelzen frei, deren Zyklen von Westen (Lavanttal) gegen Osten jeweils jünger werden (Ostkarpaten) und sich im Zentrum auch überschneiden. Die klare zeitliche und stoffliche Trennung der Magmastämme im Westen überschneidet sich im Osten und die Intensität vulkanischer Tätigkeit nimmt gegen Osten (Hargitagebirge) zu. Die Verlagerung der dynamischen Komponente der Gebirgsbildung an die Flysch-Molassengrenze erzeugte gleichzeitig in der inneren Zone der Gebirgsketten eine vertikale Aufsteigungstendenz zum Hochgebirge, deren ausgleichendes Gegenstück in den weiten Beckenmulden des Pannonischen und Siebenbürgischen Beckens mit der außergewöhnlichen Mächtigkeit der jungen Sedimente zu

finden ist. Entgegengesetzt den internen Absenkungsvorgängen erscheinen gegenwärtig die thermischen Aufstiege, die am Rande der ungarischen Tiefebene am bedeutendsten sind (F. RONNER, 1974).

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О СТРОЕНИИ И РАЗВИТИИ ВОСТОЧНЫХ АЛЬП С ТОЧКИ ЗРЕНИЯ ПРИВЯЗКИ К ЗАПАДНОЙ ВЕНГРИИ

П. БЕК-МАНАГЕТТА

Резюме

Новые результаты научных исследований в Восточных Альпах указывают на значительное проникновение моласс под флиш с гельветикомом, а на востоке — также и на их проникновение даже под Северные Известняковые Альпы. В отличие от ранее существовавшего мнения о тектонической структуре, обусловившей флишеобразование, в настоящее время возникает новое представление, согласно которому данная структура начинается не в пеннинской зоне Западных Альп и не в горах Хохе-Тауерн, а где-то вблизи северного края Известняковых Альп, причем, в районе Вены она принимает восточное направление. На границе с Венгрией появляется Рехницкий сланцевый остров с пеннинскими сланцами типа «Бюнднершифер», на которые налагает нижнеостальпийская серия грубо-зернистых гнейсов. Вехсельское окно связано с нижнеостальпийским семмерингским мезозоем, причем представляется возможным продолжение в Вепориды Западных Карпат.

От верхнеостальпийской пачки покровов отделяется среднеостальпийский структурный ярус с центрально-альпийским триасом. Обусловленный варисским орогенезом

древний кристалликум Зау-Альп и Кор-Альп рассматривается как нормальная база верхнеостальпийских покровов, сложенных граувакками. На вариасское покровное строение южной зоны и метаморфические образования, и еще в варисское время они сливались с Глейнальмским кристалликумом, имеющим, по-видимому, каледонский возраст. В результате более молодых альпидских воздыманий образовались оконобразные поднятия в кристаллических породах Кор-Альп и Зау-Альп (Вольфсберг, Клининг и т. п.). На основании находок ископаемых органических остатков доказана приуроченность к варисским движениям (до раннего карбона) бассейнового и покровного строения южных Зау-Альп и тем самым и восточного Гуркталь (Глантальская зона чешуй).

Единичные результаты радиометрического определения возраста пород еще не дают основания для какого-либо окончательного высказывания по поводу абсолютного геологического возраста соответствующих образований.

Исходное положение линеамента плитняковых гнейсов Кор-Альп свидетельствует об ориентации данной структуры в направлении Молданубийской—Моравской зоны нарушений, расположенной в северной части Альп, причем степень и состояние деформации похожи на состояние деформации гнейсов «Битте».

Проявления скручивания данного линеамента являются выражением разветвления Альп в Карпаты и Динариды. Тектоническому элементу покровного строения Альп и Карпат противостоит магматическая мобилизация, проявляющаяся внутри отдельных складчатых дуг.

DIE ENTWICKLUNGSGESCHICHTE DES BUDAER GEBIRGES

Von

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Aufgrund seiner strukturgeologischen Reambulation unterscheidet der Verfasser vier Phasen in der tektonischen Entwicklung des Budaer Gebirges: I. Paläoalpine, II. Mesoalpine, III. Neoalpine, IV. zum neoalpinen Zyklus angehörende, aber im Charakter abweichende Phase der pleistozänen Bewegungen.

Im paläoalpinen Tektonozyklus kann man eine Geosynklinale-Epoche (vom Ladin bis Ende der Trias) und eine orogene Epoche alp-karpater Typus unterscheiden. Die austrisch-mediterranen Bewegungen brachten zuerst breite Falten zustande, später Querbrüche und Schuppungen, endlich die sog. »budaer Knickung« mit 20–30 km langen, SO-gerichteten transversalen Verschiebungen. Infolge dieser Verschiebung setzt sich die Balaton-Linie in der Darnó-Linie fort.

Der mesoalpine Tektonozyklus begann mit epirogenen Bewegungen während des Eozän. Der vorhandene Andesitvulkanismus könnte mit einer Subduktion in Zusammenhang stehen. Nach der letzten kontraktiven Phase am Ende des Eozän begann die disjunktive Zerstückelung, der neoalpine Tektonozyklus. (Verwerfungssysteme, Entstehung von Vortiefen.)

Die pleistozäne Phase ist von pseudodiapirartigen Bewegungen und Thermalwassertätigkeit gekennzeichnet (insgesamt 8 Travertin-Niveaus).

Der auch jetzt dauernde Aufstieg des Budaer Gebirges ist bewiesen.

Viele Forscher beschäftigten sich mit dem geologischen Aufbau und der Entwicklungsgeschichte des Budaer Gebirges. Es ist vielleicht diesem Umstand zuzuschreiben, daß wir über dessen tektonischen Aufbau kein zusammenhängendes Bild erhalten haben. Der älteren Auffassung entgegen, gemäß welcher das Gebirge ein sich längs Brüche geformtes Schollengebirge wäre, wies zuerst F. SZENTES in 1934 darauf hin, daß der Bruchstrukturbildung vorangehend im Budaer Gebirge auch durch tangentielle Krafteinwirkungen entstandene Faltungsformen zustande gekommen waren. Er hat zuerst die einzelnen orogenen Phasen voneinander abgesondert, und ebenso die dazu geknüpften tektonischen Elemente. Er hat all dies auf Grund seiner am Nagykevélyberg ausgeführten Beobachtungen zustande gebracht. Er hat leider seine Arbeit im ganzen Gebiet des Gebirges nicht fortgesetzt. Gemäß Beobachtungen von F. PÁVAI-VAJNA (1934) sind die »Raibler« Schichten auf den Hauptdolomit schuppig aufgeschoben; er erwähnt also als erster Bewegungen solchen Charakters im Budaer Gebirge.

Gemäß Beobachtungen von S. JASKÓ (1948) in der Pálvölgyer Höhle, sind die »Raibler« Schichten gleichfalls auf den jüngeren Feuersteindolomit und auf die obereozänen Schichten geschoben. A. FÖLDVÁRI (1934) lenkte

die Aufmerksamkeit auf die jungen pannonischen Bewegungen. Schließlich F. HORUSITZKY meinte (1943), daß die durch ihn abgesonderte Faziesseinheiten der Trias, und zwar die Budaer und Pilis-Kovácsier Einheiten sind durch Einwirkung der obereozänen pyrenäischen Phase aufeinander geschoben worden. Das Maß der Aufschiebung war seiner Ansicht nach so groß, daß die Budaer Einheit taucht von unter der Pilis-Kovácsier Einheit in Halbfenster auf.

Wie wir sehen, die Ansichten über den Aufbau des Gebirges haben sich vom Schollenstruktur bis zur »Decke« Lösung entwickelt. Der Hauptgrund dafür, daß so extreme tektonische Ansichten entstanden sind, liegt darin, daß die einheitliche Karte des Budaer Gebirges ist auf Grund der von F. SCHAFARZIK, M. PÁLFY, Z. SCHRÉTER in 1921 fertiggestellten, doch nicht erschienen, später in 1938 durch F. HORUSITZKY herausgegebenen und auch noch jetzt gebrauchten Karte gezeichnet worden. Diese Karte ist gemäß der damaligen Ansichten in atektonischer Auffassung entworfen. Die Teilbeobachtungen der oben erwähnten Autoren waren unzulänglich zur Formung eines zusammenhängenden tektonischen Bildes.

Die Erkenntnis von diesen Mängeln ließ uns die tektonische Reambulation des sich in Schlüsselposition des Ungarischen Mittelgebirges befindenden Budaer Gebirges vollzuziehen.

Das Ziel der Reambulation war nicht bloß eine den heutigen stratigraphischen und tektonischen Auffassungen angemessene Kartierung des Gebirges, sondern mittels neuer Beobachtungen die selektive Interpretation und Festsetzung des tektonischen Entwicklungsganges. Es ist uns gelungen in der Entwicklung des Budaer Gebirges vier Phasen zu unterscheiden: I. Paläoalpine, II. Mesoalpine, III. Neoalpine, IV. zum neoalpinen Zyklus angehörende, aber im Charakter abweichende Phase der pleistozänen Bewegungen.

Untergrund der Triasbildungen

Die älteste der im Aufbau des Budaer Gebirges teilnehmenden Bildungen ist der ladinische Diploporendolomit. Über älteren Gesteinen geben bloß entferntere Ausbisse (Velenceer Geb., Balaton-Hochland, Vepor Geb.) bzw. etliche Angaben von Tiefbohrungen außerhalb des Budaer Gebirges liegend, und die Einschlüsse der eruptiven Gesteine des Visegráder Gebirges Auskunft (A. KOCH, 1887; F. SCHAFARZIK, 1889; Gy. SZÁDECZKY, 1895; M. SZÚCS, 1937; E. LENGYEL, 1951; T. ZELENKA, 1960). Demgemäß ist unter dem ladinischen Diploporendolomit des Budaer Gebirges die ältere Trias in Mittelgebirgsentwicklung zu erwarten. Im Nordteil des Gebirges ist wahrscheinlich Oberperm, in den südlichen Gebieten, unter der übergeschobenen Mittelgebirgs-Trias Schichtfolge sind Trias—Perm Gebildungen Bükker Typus. Gemäß den

Xenoliten sind die altpaläozoischen Gesteine des Balaton-Hochlandes und des Velenceer Gebirges auch hier entwickelt. Die Granit- und Diorit-Einschlüsse sind mit den intrusiven karbonischen Gesteinen des Velenceer Gebirges gleichzusetzen. Kristalline Schiefer mesozonalen Charakters könnten auf präkambrische Bildungen weisen. Der kontakt Hornsteinfelseinschluß weist auf die kontakt Wirkung von Magmatiten unbekannten Alters hin. Die Amphibolite können altpaläozoisch und präkambrisch sein. Diese kargen Angaben ermächtigen uns nicht über den vortriadischen Entwicklungsgang des Budaer Gebirges mehr zu sagen.

I. Der paläoalpine Tektonozyklus

A) Geosynklinale Epoche

Der paläoalpine Zyklus kann im Budaer Gebirge in zwei Zeitabschnitten geteilt werden. Der erste ist der geosynklinale Abschnitt im Laufe dessen, vermutlich ebenso, wie im ähnlich entwickelten, aus Triasbildungen aufgebauten Vértés Gebirge, die Sedimentation im sinkenden Mittelgebirgstrog eine mächtige komplette Schichtenfolge zustandebrachte. Dieser Zeitabschnitt wurde — wie wir wissen — im Vértésgebirge von einer durch vertikale Bewegungen verursachter Emersionszeit unterbrochen, bis Ende des Cenomans. Im Budaer Gebirge über den norischen Dachsteinkalk sind weder Jura, noch Kreide-Bildungen bekannt. Die Spuren von Juraschichten sind nur am Pilisberg und in der Umgebung von Dorog bekannt. Es liegt nahe, auch im Budaer Gebirge von Ende des Nor bis zum Eozän auf eine zusammenhängende Emersionsphase zu denken, während welcher diese südliche Randregion des ungarischen Mittelgebirgstroges schon Festland war, dagegen während in der Achsenzone das Jura und Niederkreidemeer sich befand (Abb. 1).

Die Ereignisse der geosynklimalen Epoche im Budaer Gebirge können wir nur vom Ladin an verfolgen. Zu dieser Zeit entstand eine karbonatische Schichtfolge eines seichten, offenen, tropischen Meeres. Die 1.500 m übertreffende ladinische Schichtfolge (Diploporiten und Übergangsdolomit) weist auf gleichmäßige Sedimentation. Das bedeutet, daß der tektonische Vorgang, das heißt das gleichmäßige, ziemlich schnelle Sinken im ganzen Gebiet gleichen Charakters war. Die ungefähr 150 m mächtige unterkarnische »Raibler« Schichtfolge, welche im ganzen Gebiet des Gebirges in Aufschlüssen und in Bohrungen sich als märgelig und abwechslungsreich erweist, deutet darauf hin, daß das gleichmäßige Sinken durch Oszillation gestört wurde. Diese vertikale Bewegung, welche sich nur in veränderter Fazies äußert, kommt auf die Rechnung der Labaer Phase. Hiernach, ganz bis zum am Ende des Nor erfolgten totalen Emporhebung haben sich die Obertriasbildungen im Budaer Gebirge in zwei voneinander gut unterscheidbaren Fazies ausgebildet. Die eine ist die »feuer-

stein-dolomitische Ausbildung«, welche 6—700 m mächtig ist und aus Feuersteinsdolomit, Dolomit und am oberen Teil gestreiften Feuersteinsdolomit besteht. In dieser Ausbildung kennen wir keinen Dachsteinkalk. F. HORUSITZKY (1943) hielt diese Ausbildung für das Budaer Gebirge charakteristisch. Die andere ist die »kalkige dolomitische Ausbildung«, in welcher wir keine Feuersteinsdolomite finden. Den unteren Teil der Schichtfolge bilden Dolomite vom Typ »Hauptdolomit«, den oberen Teil schließen die sog. »Übergangskalkstein-Dolomit«-Schichtfolge und der mächtige Dachsteinkalkkomplex ab. F. HORUSITZKY (1943) trachtete diese Ausbildung auf das Gebiet der »Pilis-Nagykovácsier« tektonische Einheit zu beschränken. Diese Ausbildung ist wesentlich mächtiger, als die vorige: etwa 1.500 m.

Im Laufe unserer tektonischen Reambulation wurde festgestellt, daß die zwei Obertrias-Ausbildungen in synsedimentärer Weise, parallel mit der Küstenlinie des einstigen Mittelgebirgstroges, d. h. in SW—NO Richtung entstanden. Die Streifrichtung der Faziesgürtel — als es später behandelt wird — hat auf Einwirkung der in der oberen Kreide vorangehenden orogenen Phasen sich in der jetzigen NW—SO Richtung aufgestaut. Das ursprüngliche Nebeneinandersein ist, abgesehen von den infolge der Raumverengung erfolgten horizontalen Abschiebungen und Schuppungen unverändert geblieben. Von F. HORUSITZKY erwähnte übereinandergeschobene tektonische Einheiten sind nicht zu beobachten. Das verfehlte Deuten mag auf Kosten der sich auf Einzelheiten nicht erstreckenden Kartierung zuzuschreiben sein. Immerhin die vier miteinander parallelen Faziesgürtel (Irhásárok—Sas-Berg, János-Berg, Hármashatár-Berg, Nagykevény) sind westlich von der Solymár-Nagykovácsi Störungszone schon nicht zu folgen. Hier in dem Westteil des Budaer Gebirges kommt nur die kalkig-dolomitische Ausbildung vor. In östlicher Richtung kann man auf Grund der kargen Bohrungsangaben der Pester Tiefebene die Ausbildung der oberen Trias nicht festsetzen (Abb. 2). Das Entstehen der Faziesgürtel können wir auch folgendermaßen erklären: parallel mit der Richtung der einstigen Geosynklinalachse sind synsedimentäre Brüche entstanden. Entlang dieser kamen schneller, bzw. langsamer sinkende Streifen zustande. Auf ersteren haben sich die das schnellere Sinken zeigenden kalkig-dolomitischen Schichten ausgebildet, auf letzteren die auf gleichmäßigere Bewegung hinweisende und tiefere Lage zeigenden feuersteinig-dolomitischen Ausbildungen abgelagert.

Das in der oberen Trias stattfindende völlige Emporhebung, die sich — wie schon erwähnt — in der Achsenlinie des Mittelgebirgstroges nur während den kurz dauernden Emersionsabschnitten äußerte (vom Lias bis zum Aptien), ist gewöhnlich entlang der einstigen Küstenlinie zu beobachten. Das in der oberen Trias solcherart entstandenes Festland nannte T. SZALAI (1969) den »Rücken von Pelso«. Diese tektonische Einheit ist mit dem spätvariszischen Granitmagmatismus gleichzeitig entstandenen Balaton-Velenceer Granitzug

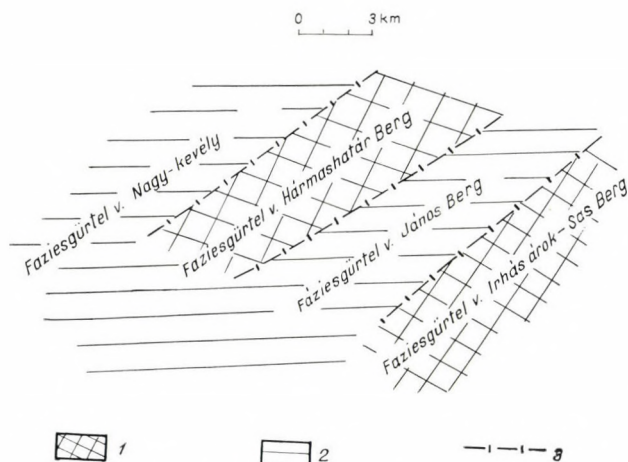


Abb. 2. Die ursprüngliche Lagerung der geosynklinalen Trias-Bildungen des Budaer Gebirges (vor den Austrischen-Mediterranen Bewegungen). 1. Karn.-Nor.: Feuerstein-Dolomit Ausbildung; 2. Karn.-Nor.: Dolomit-Kalkstein Ausbildung; 3. Grenze von Faziesgürtel

identisch. Diese Struktureinheit hat von der oberen Trias an den ungarischen Mittelgebirgstrog von dem Igal-Bükk Trog endgültig getrennt (G. WEIN, 1969). Diese Bewegung kann also mit der altkimmerischen epirogenen Phase verbunden werden. Zu dieser Zeit unterkamen die Bildungen der geosynklinalen Zeitspanne ungestört in ursprünglicher SW—NO Streifrichtung nordwestlich von dem aus altpaläozoischen Bildungen aufgebauten Rahmen der Balaton-Velenceer Granitzug.

B) Orogene Epoche

Den Zeitpunkt der orogenen Phasen, die im altalpinen Zyklus die grundlegende Struktur des Ungarischen Mittelgebirges ausgestalteten, können wir nur dort feststellen, wo die vollständige mesozoische Schichtfolge sich entwickelt hat. In dem Budaer Gebirge können wir diesbezüglich bloß soviel festsetzen, daß sie zwischen der oberen Trias und dem unteren Eozän, bzw. inwiefern die pilisvörösvärer Bauxiten sich als oberkretazisch erweisen, in der oberen Kreide sich abgespielt haben. Im Bakony, Vértes und Gerece begannen diese Bewegungen am Ende der unteren Kreide, im Aptien (Tisiaer Phase von K. TELEGDI-ROTH, 1934), und endeten vor dem Senon. Die Hauptphase spielte sich zwischen Cenoman und Senon ab. Jene außerordentlich starke Bewegungen kompressiven Charakters, welche die grundlegende Struktur des Budaer Gebirges zustandebrachten und welche es an seinen jetzigen Platz schoben, haben sich in den Austrischen und Mediterranen Phasen abgespielt. Auf Grund von Analogien aus dem Ungarischen Mittelgebirge und

den Nordwestkarpaten (F. SZENTES, 1961; D. ANDRUSOV, 1968; Gy. WEIN, 1969) ist feststellbar, daß der Hauptparoxismus der Bewegungen in der Infragosau-Mediterranen Phase, zwischen Cenoman und Senon stattfand (TOLLMANN, 1966). In dieser Zeit entstand das Deckensystem der Interniden der Alpen und Karpaten. Die Bewegungen äußerten sich am Balatonhochland, im Bakony- und in geringerem Maße im Vértesgebirge in Form von horizontalen Verschiebungen, Schuppungen, selten in Form von Faltungen. Im Turon Emportauschen, Konsolidation, dann Transgression des senonischen Meeres schließen den bewegten Zeitabschnitt des altpaläozoischen Zyklus ab.

Der ursprünglich SW—NO gerichtete ungebrochene Mittelgebirgstrog stand in gerader Linie in Verbindung mit dem südgömörer Mesozoikum ähnlicher Ausbildung. Die einzige ungestörte Lage hat sich während der austrisch-mediterranen Bewegungen geändert. Unter Einwirkung des starken Druckes sind zuerst breite Falten entstanden, parallel mit der Geosynklinalachse (Abb. 3). Deren Spuren hat schon F. SZENTES (1934) im Solymárer Tal erkannt. Während unserer Reambulation gelang uns insgesamt vier, heute schon NW—SO gerichtete Spuren von glatten Falten zu rekonstruieren. Sie sind die folgenden: 1. Pátyer Schloßberg—Kiskopasz, 2. Julianna Meierei—Hunyad Horst, 3. Ördöggraben, 4. Solymárer Tal. Die milden Falten sind ziemlich schwer zu erkennen, da dieselbe durch horizontale Verschiebungen und Schuppungen, welche die später kulminierenden kompressiven Bewegungen hervorriefen, dann durch das tertiäre Verwerfungssystem vollkommen zerstückelt wurden. In Stärke weiter zunehmende Bewegungen haben nachher Brüche, Überschiebungen und horizontale Verschiebungen verursacht. Während dieser Vorgang sich vollzog erhielt der ganze Mittelgebirgstrog ungefähr vom Váler Graben bis zum Romhányer Gebirge entlang Brüche eine transversale Verschiebung. Inzwischen ist die Streifrichtung der mesozoischen Gebilden um 90° von der ursprünglichen Mittelgebirgsrichtung abgebogen. Demzufolge hat sich die Streifrichtung des Mesozoikums des Budaer Gebirges NW—SO gestaltet. Das wird »Budaer Streifbruchzone« oder »Budaer Knickung« genannt. Das Maß der SO gerichteten transversalen Verschiebungen mag 20—30 km gewesen sein.

Dergestalt hat sich der Mittelgebirgstrog nicht nur senkrecht auf ihre Streifrichtung verengt, sondern infolge des Streifbruches hat sich auch seine Streiflänge vermindert. Diese Erscheinung läßt sich in verschiedener Weise erklären. Man kann einerseits daran denken, daß die konkave innere Seite des Karpatenbogens sich verkürzte, andererseits mag auch eine entlang der Zággráb—Kulcs Hauptstrukturlinie stattfindende Verschleppung als Ursache dienen. Während der altpaläozoischen orogenen Epoche jene Strukturelemente welche die jetzige Lage und Gestalt des Budaer Gebirges bestimmten, sind in folgender Reihenfolge entstanden: 1. große Wölbungen, 2. Querbrüche, 3. Überschiebungen-Schuppungen, inzwischen beginnt der Vorgang des Streif-

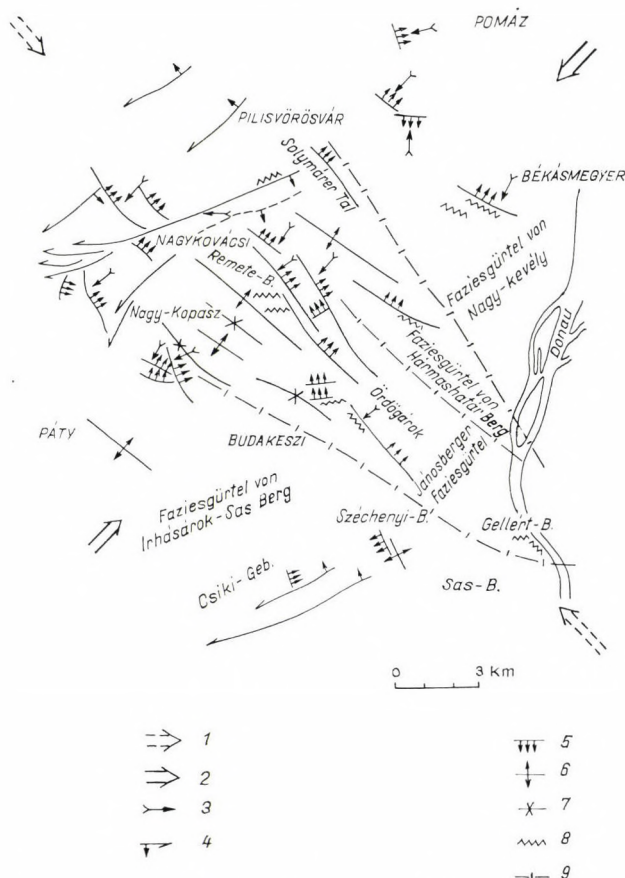


Abb. 3. Das Bruchsystem der Austrischen-Mediterranen Phasen des Budaer Gebirges (Entw.: GY. WEIN, 1974). 1. Druckrichtung der ersten-zweiten Phasen (noch in ursprünglicher Lage); 2. Druckrichtung der dritten-vierten Phasen (während des Bruches und nachher); 3. Richtung der Bewegung, beobachtet an Schollen und Gesteinen; 4. Platte Verwerfung mit horizontaler Verschiebung; 5. Überschiebung; 6. Faltensattel; 7. Faltenmulde; 8. Zertrümmerte Zone; 9. Grenze von Faziesgürteln

bruches (Knickung), 4. horizontale Verschiebungen, teils entlang der Querbrüche (Abb. 4).

Es war eine sehr wichtige großstrukturelle Folge des »Budaer Streifbruches«, daß die vom Váler Bruch NO gelegene mesozoische Schichtfolge des Ungarischen Mittelgebirgstroges sich über die altpaläozoischen Gebilden des Balaton-Velence Granitzuges schob und sich dem Igal-Bükk Mesozoikum anstaute. Zwar beobachteten wir früher an dieser Stelle den Streifbruch (Gy. WEIN, 1969), aber fanden wir dafür keine entsprechende Erklärung. Jetzt ist es schon klar, daß die Balaton Linie nach infolge des Budaer Streifbruches erlittenen 20—30 km Verschiebung sich in der Darnó-Linie fortsetzt. Die altpaläozoischen Bildungen der Balaton—Velence Linie sind unter der sich nach

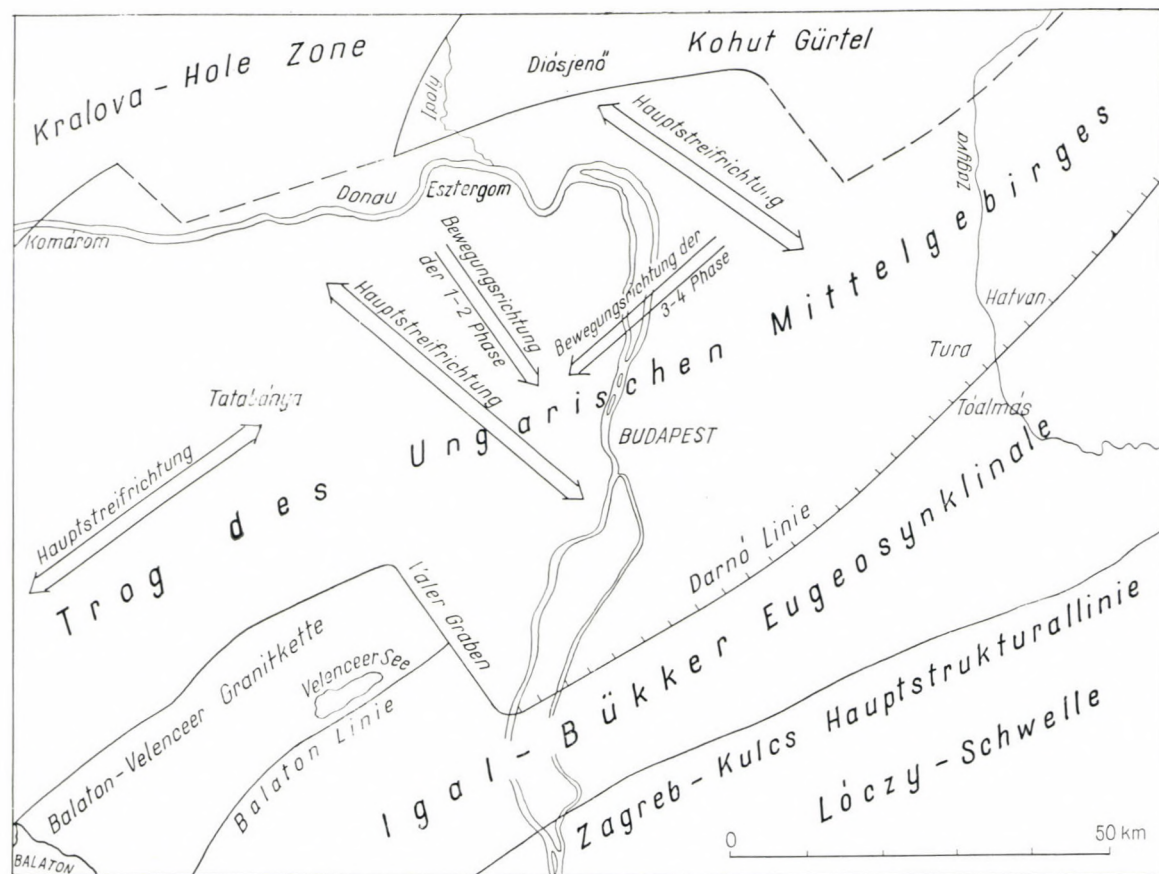


Abb. 4. Großstrukturentwurf des während der Austrischen-Mediterranen Phasen erfolgten »Streifbruches« des Ungarischen Mittelgebirges (Entw.: GY. WEIN, 1974)

SO verschobenen Trias-Schichtfolge zu vermuten. Im Sinne der aufgeführten Erwägungen, können wir ruhig die zusammengezogene Benennung »Balaton-Darnó Linie« benutzen.

Die intensive, kompressive Bewegungen haben die schon konsolidierte, steif gewordene karbonatische Triasgesteinsmassen zerbrochen, zermalmt und größtenteils in kataklastische Gebilden verwandelt. Besonders die Dolomiten litten infolge der Bewegungen, aber auch der Dachsteinkalk ist in vielen Gegenden zerquetscht, dann mit Kalzit neu zementiert worden, auf hohe Alter der Deformation weisend. Das entlang junger Brüche zerbrochene Gestein zementiert sich in dieser Form nicht wieder zusammen. Die Tatsache, daß im »Gürtel des Budaer Streifbruches« unter der juvenilen, überhaupt nicht tektonisierten Schichtfolge des Bauxits, bzw. des unteren Eozäns überall die stark in Anspruch genommenen mesozoischen Gesteine zu finden sind, ist ein sicheres Zeichen dafür, wie mächtig die orogene Phase des altalpinen Zyklus war. Nur die starken, lang dauernden und sich wiederholenden kompressiven Phasen können eine solche Inanspruchnahme von starren Gesteinen verursachen. Dieselbe Erscheinung ist im mesozoischen Deckensystem der Karpaten zu beobachten, wo die mesozoischen Gesteine der übereinandergehäuften Subtatradecken während einer Bewegung von mehreren 100 km-n zerbröckelt, dann wieder zusammenzementiert worden ist, gleich wie die Triaßgesteine des Budaer Gebirges. Ähnliche Erscheinungen sind in den verhältnismäßig ruhigen Gebieten (Balatonhochland, Bakony, Vértes und Gerecse) nur ab und zu zu erfahren. Dagegen im Gürtel des »Budaer Streifbruches«, welcher sich vom Váler Graben bis zu den Schollen von Romhány, vielleicht noch weiter ausdehnt, kann als charakteristisch bezeichnet werden.

Im Laufe der Klärung der eigentümlichen tektonischen Lage des Budaer Gebirges können wir uns nicht mit Feststellung der bloßen Tatsachen begnügen, sondern es ist auch notwendig deren großstrukturelle Bedeutung zu entwerfen. Die Piliser Bruchlinie, welche das Budaer Gebirge und das Donauzuggebirge (Visegráder Gebirge) voneinander trennt — und welche schon bis jetzt als eine neogene Bruchlinie von großer Bedeutung verzeichnet war — hat V. SCHEFFER (1963) inspiriert, jene nach E. BONČEV (1958) als Fortsetzung des Kraistiden-Lineaments zu interpretieren. In der Richtung der Slowakei die Spuren des »Budaer Streifbruches«, nachforschend, sind im östlichen, an der Slowakei fallenden Teile des Kisalföld im Untergrund des Neogen zu finden (O. FUSAN et al., 1971). Die hier festgestellten älteren kristallinen Bildungen, u.a. Granite — ähnlich zum Mesozoikum des Budaer Gebirges — ändern nämlich die Streifrichtung. Nach Auffassung von V. SCHEFFER (1963) und L. BENDEFY (1967) die Piliser Bruchlinie kann durch die Kleine Ungarische Tiefebene mit der den nördlichen Teil der Kleinen-Karpaten abgrenzenden Ölvéder Linie verbunden werden.

Wir können uns zu dieser Konzeption nicht anschließen, da wir zwar

eine NW Fortsetzung des Budaer Streifbruches beobachtet haben, aber eine SO Verlängerung nirgends zu entdecken vermochten (Gy. WEIN, 1969). Wenn wir zwar die Kraistiden-Theorie nicht annehmen, soviel können wir schon jetzt feststellen, daß der Budaer Streifbruchgürtel ein Strukturgürtel von größter Bedeutung und vermutlich lineamentartig ist (Gy. WEIN, 1974). Die andere sehr wichtige Feststellung, welcher wir Nachdruck geben müssen, ist, daß ein so großes Ausmaß der Bewegungen, welches schon den Bedingungen der Deckenbildung nahe kommt, aber keinesfalls Schollenstrukturbildung zu nennen ist, weist darauf hin, daß der Ungarische Mittelgebirgstrog im altalpinen Zyklus zum Alpen-Karpaten System angehörte, damit organisch verbunden war, seine Entwicklung sich unter gleichen Bedingungen und zu gleicher Zeit vollzog. Demzufolge können wir hier bis zur neogenen Zerstückerungs-Strukturbildung von einer Zwischenmaße nicht reden. Wenigstens die bis zur Zagreb—Kulcs Hauptstrukturlinie sich entwickelte Struktureinheit (G. WEIN, 1969) — die Eugeosynklinale von Igal—Bükk inbegriffen — müssen wir als Fortsetzung der Alpen betrachten. Das von der Zagreb—Kulcs Hauptstrukturlinie südwestlich gelegenes Gebiet ist durch andere geologische Bildungen aufgebaut, was auf die exotische Lage desselben verweisend, wirft das Problem der Abstammung auf. Zugleich ist annehmbar, daß hier schon während des altalpinen Zyklus sogar viel früher eine »Urtisia« existierte, welche mit dem »Orientalischen Festland« von K. PETERS (1857) und E. MOJŠISOVIĆ (1880) identifiziert werden kann. Wir wollen uns hier mit dieser Frage nicht weiter befassen, bloß darauf hinweisen, welche ungeheuerere Bedeutung der altalpinen Tektonozyklus im Karpatenbecken beizumessen ist, und dessen äußerst wichtige Gegend — sozusagen in Schlüsselstellung — der »Budaer Streifbruchgürtel« war.

II. Der mesoalpine Tektonozyklus

a) Zeitabschnitt der Bewegungen epirogenen Charakters

Der mesoalpine Tektonozyklus ist nach Konsolidation der altalpinen Strukturetage im SW Teil des Ungarischen Mittelgebirges (Bakony) durch die Transgression des Senon, während im Vértes, Gerece und Budaer Gebirge nur durch das Vordringen des eozänen Meeres den Beginn des Neuen Zeitabschnittes angezeigt. Im Budaer Gebirge fand, während der von der oberen Trias bis zum unteren Eozän (bzw. mittleren Eozän) dauernden Festlandzeit eine Verkarstung, Bauxitbildung und Erosion statt. Die Abtragung, wie das von Gy. BÁRDOSSY (1961) festgestellt wurde, geschah von Süden, aus der Richtung eines aus kristallinen Gesteinen aufgebauten Festlandes von größerem Umfang. Dieses Festland war der nach der oberen Trias wieder aufgetauchte Balaton-Velence Granitzug (der »Pelsoer Rücken« von T. SZALAI). Die am

ganzen Gebiet des Ungarischen Mittelgebirges wahrnehmbar, sehr stark entwickelte Paläokarstformen lassen darauf folgern, daß das Festland sich hoch über den Meeresspiegel hob, ein entwickeltes Wassersystem besaß und von großer Ausdehnung war (Abb. 5). Gemäß bisherigen Tiefbohrungsangaben sind weder im Alföld, noch im südlichen Dunántúl — ausgenommen den Alfölder Flyschtrog und die Linie des Igal-Bükk Trog — eozäne Bildungen bekannt. Demgemäß, nach dem alpalpinen Tektonozyklus beginnt die Herausbildung jenes konsolidierten Blockes von großem Umfang, welchen J. PRINZ (1914, 1958) »Tisia« nannte.

Im Budaer Gebirge drang die untereozäne Transgression — wahrscheinlich ähnlich, wie bei Tatabánya (F. SÓLYOM, 1960) — in entlang Brüche absinkende littorale Becken ein. Annehmbarerweise — zwar bis Heute noch nicht nachgewiesen — sind die Kohlenbecken von Solymár, Pilisszentiván und Nagykovácsi dieserart entstanden. Das Absinken von den Becken ist jedenfalls zur laramischen Phase zu knüpfen, und es setzt die erste Bewegungsphase dilatierenden Charakters des mesoalpinen Zyklus fest.

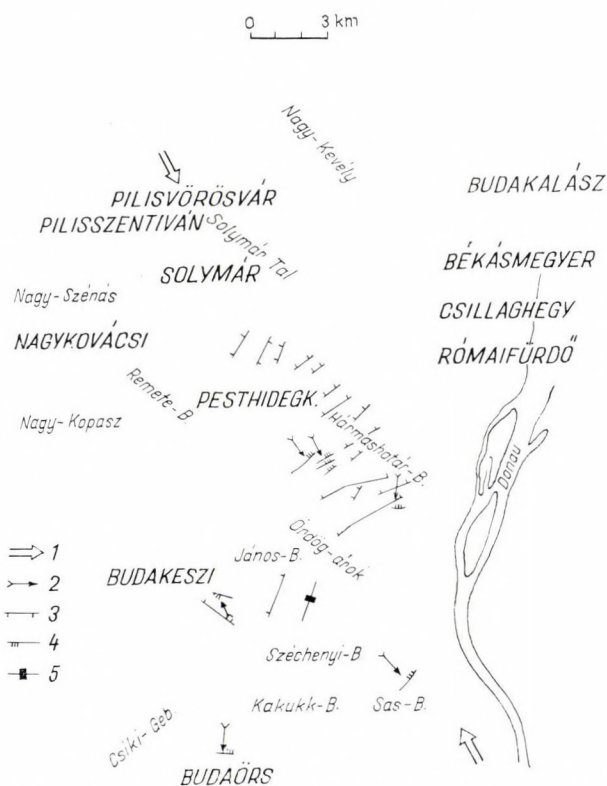


Abb. 5. Das nachweisbare Bruchsystem der Eozänen (hauptsächlich Pyrenäische) Phase im Budaer Gebirge (Entw.: GY. WEIN, 1974). 1. Richtung der Druckes; 2. An Schollen und Gesteinen beobachtete Richtung der Bewegung; 3. Verwerfungslinien; 4. Überschiebungslinien; 5. Faltenachse

Wir müssen bemerken, daß gemäß der neuen Auffassung (G. KOPEK 1968, G. KOPEK — T. KECSKEMÉTI — E. DUDICH, 1966) der bisher in das untere Eozän gereichte Braunkohlenkomplex des Budaer Gebirges (M. HANTKEN, 1884; K. TELEGDI-ROTH, 1923; P. ROZLOZNIK, 1935; E. SZŐTS, 1952; L. GIDAI, 1970) wird in das mittlere Eozän eingereiht. Sie behaupten, daß die eozäne Transgression den nordöstlichen Teil des Ungarischen Mittelgebirges erst im Lutetien erreichte. Wir haben — bis endgültiger Entscheidung der Diskussion — den unteren Komplex der eozänen Schichtenfolge des Budaer Gebirges mit den Kohlenlagern, der alten Auffassung gemäß in das untere Eozän eingereiht.

Das mitteleozäne Meer drang nach einer kurzen Emersion in der Richtung des Budaer Gebirges vor und überschwamm es beinahe vollkommen. Diese epirogene Bewegung deutet die illyrische Phase an. Zu dieser Phase knüpfen sich in dem Ungarischen Mittelgebirge und so auch im Budaer Gebirge die Anfänge des charakteristischen neutralen Vulkanismus an. Seine Spuren fanden wir auch im Budaer Gebirge. Die, in den lutetischen und priabonischen Basiskonglomeraten gefundenen zahlreichen Andesit- und saure eruptive Kiesel (F. SCHAFARZIK, 1913; A. VENDL, 1929; Gy. WEIN, 1974) und die zwischenlagernden, sich nach oben vermindernden Andesittuffeinlagerungen deuten die bis zum unteren Oligozän dauernde vulkanische Tätigkeit an. Bohrungen haben biotitreiches Ganggestein (B. WEBER, 1962) und Andesitgänge aufgeschlossen. Weiterhin ein Gravitationsminimum läßt darauf folgen, daß Gesteine von minderem spezifischen Gewicht als die karbonatischen — voraussichtlich saure, neutrale Vulkanite — sind in größeren Tiefen vorhanden (Sz. OSZLACKY, in: F. HORUSITZKY—Gy. WEIN, 1962). Infolge der pleistozänen Thermalwassertätigkeit haben telethermale Mineralien und in Zusammenhang damit Elementanreicherungen stattgefunden. Aus all dies müssen wir daran denken, daß in der Tiefe unter der Triasbildungen des Budaer Gebirges, bzw. in diese hineingedrungen sich eozäne Vulkanite befinden, mit welchen auch eine primäre Mineralisierung verbunden ist. Die Spuren der mitteleozänen vulkanischen Tätigkeit können wir in der Linie der Balaton—Velence Granitkette-Struktureinheit durch das Budaer Gebirge bis Reesk folgen (V. SZÉKY-FUX, 1957; Gy. WEIN, 1969). Wenn wir an die Möglichkeiten der Plattentektonik an theoretischen Folgerungen denken, so können wir aus der Lage und dem Alter der eozänen Vulkanite die Folge ziehen, daß es um einen subsequenter Vulkanismus handelt, welcher mit dem Entstehen des subtrischen Deckensystems, bzw. Horstgürtels untergeschobenen Klippenzuges in Zusammenhang steht. Er ist aber auch erklärbar durch entlang der Balaton-Darnó-Linie erfolgten Subduktion, oder durch eine Subduktion von entgegengesetzter Richtung entlang der Zagreb—Kulcs Linie (E. SZÁDECZKY-KARDOSS, 1971). Das eine scheint zweifellos zu sein, daß die vulkanische Tätigkeit, welche mit gleichen Merkmalen, in derselben Streifrichtung, in gleichen Zeitabschnitt, entlang Lineamente sich entfaltete, tiefgreifende tektonische Ursachen hatte.

Vorläufig können wir noch nicht entscheiden, zu welchem tektonischen Gürtel sie sich knüpft. Nach der illyrischen Phase leitet die weitere Transgression des oberen Eozän schon die pyrenäische Phase ein.

b) Pyrenäische Phase

Die pyrenäischen Bewegungen sind zweiphasig. In der ersten erfolgt epirogenes Sinken, welches die vollkommene Überschwemmung des Budaer Gebirges verursacht. In dieser Zeit geraten die höchsten Gipfel und der südliche Teil des Gebietes unter das Wasser des Obereozänen Meeres.

Die zweite Phase spielte sich am Ende des oberen Eozän, aber nicht vor dem Unteren Oligozän ab, und äußerte sich in starken kontraktiven Bewegungen. Zu Folge des in NW—SO Richtung wirkenden Druckes sind vorwiegend SO gelegene, steil gerichtete Schuppungen entstanden. Mit den Schuppungen gleichzeitig an der Neigeseite sind auch mit der Überschiebungsfläche parallele Verwerfungen entstanden, welche zu den von J. KÓKAY (1956) aus dem Bakony beschriebenen »assymmetrischen Keilstrukturen« ähnlich sind. Es ist kennzeichnend für diese Bewegungen, daß Gestein nur entlang der Brüche sich zerstückelte, in regionalen Ausmaßen, was bei den Triasbildungen nirgends zu beobachten ist. Diese Erscheinung ist im allgemeinen zu Gürteln geknüpft, oder isoliert beobachtet werden. Das schönste Beispiel der Schuppentektonik der pyrenäischen Phase können wir im nördlichen Steinbruch bei der Pálvölgyer Grotte beobachten, wo die karner »Raibler« Schichten und Feuerstein-dolomit sich auf den obereozänen nummuliten-discocyclinen führenden Kalkstein anstauten. Besonders beim Bau der U-Bahn, an der Strecke Südbahnhof-Batthyány-Platz kann man klar sehen, daß im Budaer Mergel noch auf kompressive Bewegungen weisende horizontale Verschiebungen stattfanden (GY. WEIN, 1973), während in den unteroligozänen Tarder Schichten bloß einige Verwerfungen zeigen den geänderten Stil an. Diese war die zweite, und damit die letzte Gebirgsbildungsphase kompressiven Charakters in der Entwicklungsgeschichte des Budaer Gebirges. Nachher ändert sich grundsätzlich der tektonische Stil.

Beginn des Zerstückelungs-Zeitabschnittes. Das Oligozän gehört altersmäßig noch zur paläogenen Strukturetage, im Sinne der grundsätzlichen Umwandlung des tektonischen Stils aber müßte man sie in die neogene Etage einreihen. Nach der pyrenäischen kontraktiven Phase der mittlere, nördliche und westliche Teil des Budaer Gebirges wurde für eine kurze Zeit Festland, dann beginnt die oligozäne Transgression mit dem »Hárshegy«-er Sandstein. Am östlichen Gebiet kündet das seicht werdende und sich versüßende Meer mit Faziesänderung (Tarder Schichten) die positive Bewegung an. Nach dem kurz dauernden infraoligozänen Emporhebung (Infraoligozäne Denudation, K. TELEGI-ROTH, 1928) überschwemmte die oligozäne Transgression das ganze Budaer Gebirge. Die Transgression drang entlang synsedimentärer Bruch-

linien vor. In dieser Zeit begannen jene beträchtliche — hauptsächlich NW—SO in kleinerem Masse SW—NO gerichtete — Bruchsysteme sich auszubilden, welche auch im Neogen weiterlebend jene tiefe Grabensysteme und Horste zustandebrachten, welche die markantesten tektonischen Elemente des Budaer Gebirges sind. Solche sind in erster Linie der Bruch von Pilis, der Graben von Solymár und Ördögárok. Im Graben steht hoch der Kisceller Ton, mancherorts mehrere 100 m mächtig, während er an den Horsten sich verdünnt und von sandiger Entwicklung ist. Im oberen Oligozän beginnt das Gebirge sich von neuem zu erheben; der mittlere Teil ist zu dieser Zeit schon Festland. Die erste Phase der charakteristischen Zerstüklungs-Strukturgestaltung können wir mit den in unteren-mittleren Oligozän stattfindenden »helvetischen Bewegungen« (A. TOLLMANN, 1966) synsedimentären Charakters verknüpfen. Hier tritt zuerst jener tiefgreifender Unterschied zwischen der tektonischen Entwicklung des Karpatenbogens und der »Pannonischen Zwischenmasse« auf, welche die Herausbildung der jetzigen Struktur der »Tisia« ankündet. Diese Frage kurz berührend müssen wir zuerst darauf hinweisen, daß die Herausbildung des Flysch-Deckensystems der Nordkarpaten im Oligozän beginnt und beendet sich bis Ende des Tortonien. Diese Epoche ist im Gebiet der Pannonischen Zwischenmasse die Zeit der sog. »Zerstüklungs-Struktur-bildung« (Gy. WEIN, 1969). Dazu knüpft sich der im Oligozän beginnender und im Neogen seinen Höhepunkt erreichender zweiter subsequenter Vulkanismus. Die im mittleren Eozän stattgefundenen magmatektonischen Vorgänge eines subsequenter Vulkanismus wiederholen sich also aber nicht am gleichen Ort. Während das Deckensystem der äußeren Karpaten sich herausbildet, im pannonischen Becken vollzieht sich gleichzeitig die Zerstükelung und zugleich fängt an auf eine neuere Subduktion hinweisender subsequenter Vulkanismus an. Hier können wir zwischen zwei Alternativen wählen. Wir können den neogenen Vulkanismus zur Subduktion der äußeren Karpaten knüpfen, oder nach E. SZÁDECZKY-KARDOSS mit der schon erwähnten Zagreb—Kulcs, oder Balaton—Darnó-Linie in Verbindung bringen.

III. Der neoalpine Tektonozyklus

a) Zeitabschnitt der Zerstükelung und des neogenen Vulkanismus

Das Budaer Gebirge hat sich am Ende des Oligozän ganz erhoben. Im Laufe des Miozän war es Insel, dann Halbinsel zwischen den meeresüberschwämmten neogenen Senken. In dem Gebirge stehen uns über die Phasen der neogenen Struktur-bildung wenig Beobachtungen zu Verfügung. Jenes können wir eher in den Vortiefen und an entfernteren neogenen Gebieten beobachten. Die miozäne Schichtfolge des Randes des Tétényer Hügelgebietes zeigt die sawische und steierische Bewegungen durch Verschiebung der Küsten-

linie und durch Diskordanzen an. Demgemäß ist an der Grenze Oberoligozän-Burdigalien (Egerien-Eggenburgien) die savische Phase, zwischen dem unteren und oberen Helvet (Otnangien-Karpatien) die altsteierische Phase, an der Grenze vom oberen Helvet-Torton (Badenien) die jungsteierische Phase und endlich zwischen Torton-(Badien)-Sarmat die Moldauer Phase feststellbar auf Grund der in den Ingressionen sich äußernden epirogenen Bewegungen. Diese Bewegungen, welche gemäß der Strukturformen die neogene Gebilde der Pester Ebene und der Dunazúg-(Visegráder-)Gebirges anrühren, haben sich nicht nur in Bewegungen epirogenen Charakters, sondern auch in Erneuerung der alten Bruchsysteme und in Entstehen von neuen geäußert (Abb. 6). Die Bruchlinien sind vorwiegend von NW—SO, manchmal von SW—NO Streifrichtung. Die dilatationsartige Krustenbewegung (Zerstückelungs-Strukturgestaltung) formte weiter das teils schon herausgebildete Verwerfungssystem des Budaer Gebirges und gestaltete die Vortiefen (Becken von Bicske, Dunazúg Gebirge, Pester Ebene, Südbudaer Gesänke). Es war ein charakteristischer und wichtiger Vorgang dieser Epoche das Wiederaufleben des Piliser Bruchsystems. Der sich zu diesem System knüpfender neogene neutrale Vulkanismus muß eine wichtige Rolle bei den magmatischen Vorgängen gespielt haben. Wir ordnen den neogenen subsequenten Vulkanismus — welcher im Sinne der Plattentektonik die Ursache des Entstehens des neutralen Magmas ist — dem zweiten starken karpatischen Subduktionsprozeß zu, aber beim Emporkommen des Magmas hat das Piliser Bruchsystem die Hauptrolle gespielt. Die Tatsache, daß seine Richtung mit dem altalpinen Bruchsystem zusammenfällt, und westlich vom Bruch die neogene vulkanische Tätigkeit plötzlich aufhört, scheint zu bestätigen, daß dieses im neogenen sich erneuerndes Bruchsystem sehr tief drang und sich zur jene lineamentartige Zone des Budaer Streifbruchs (Budaer Knickung) knüpft, welche am Ende des altalpinen Tektonozyklus entstand.

Die Transgression des unterpannonischen Zwischenmeeres hält die attische Phase fest, während das noch stärkere Vordringen der oberpannonischen See die Slavonische Phase kennzeichnet. Während dieser Phasen sank schnell der Becken von Bicske; in ihm und am westlichen Rand des Budaer Gebirges sind NNW—SSO gerichtete Verwerfungssysteme und mit denen parallele milde Faltungen entstanden. (A. FÖLDVÁRI, 1932; S. JASKÓ, 1939; K. MIKE, 1963; L. SZABADVÁRI, 1966). Wir betrachten die im Bicskeer Becken beschriebenen Wölbungen als »falsche Wölbungen«, welche durch Erhebung der den Untergrund bildenden und selektive Bewegung ausführenden Trias-Schollen und durch Kompaktion entstanden.

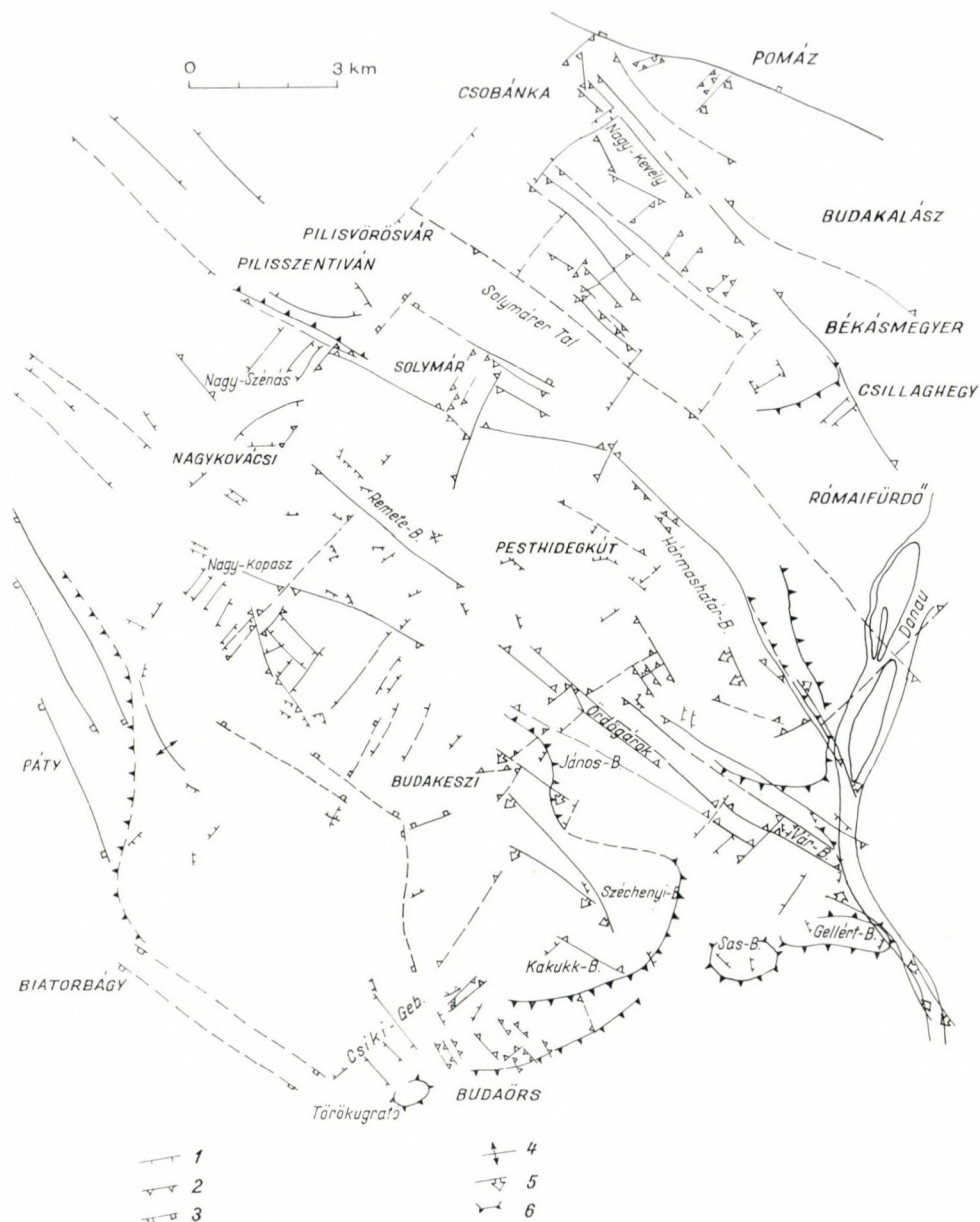


Abb. 6. Das Oligozän-pleistozäne Bruchsystem des Budaer Gebirges (Zeitabschnitt der Zerstückelung) (Entw.: GY. WEIN, 1974). 1. Bruchlinie unbestimmten Alters; 2. Oligozäne Bruchlinie; 3. Miozäne, pliozäne Bruchlinie; 4. Miozäne, pliozäne Faltensattel; 5. Pleistozäne Bruchlinie; 6. Pleistozäne selektiv sich positiv bewegend Scholle

IV. Der Zeitabschnitt der pleistozänen Bewegungen

Die am Ende des oberen Pannon beginnende Regression führte zum endgültigen Emporatauchen des Budaer Gebirges. Dieses epirogenes Ereignis ist vermutlich mit der ostkaukasischen Phase gleichzustellen und kündigt die Fortsetzung der neogenen Zerstückelungsperiode an (Abb. 7). Die quartären Bewegungen haben wir hauptsächlich deshalb abgesondert, weil sie mit stürmisch schnellen selektierten pseudodiapirartigen Bewegungen die Prägung der schnell in Gebirge verwandelnden Epoche des nealpinen Tektonozyklus innehaben. Am Ende des Pliozän, Anfang des Pleistozän beginnt die Halbinsel der Tertiärende sich rasch zu erheben und während der 2. Mill. Jahre des Pleistozäns verwandelt sich in das Budaer Gebirge. Die Erhebung erfolgt stätig, mit sich wiederholenden aktiven Abschnitten. Das Gebirge entwickelte

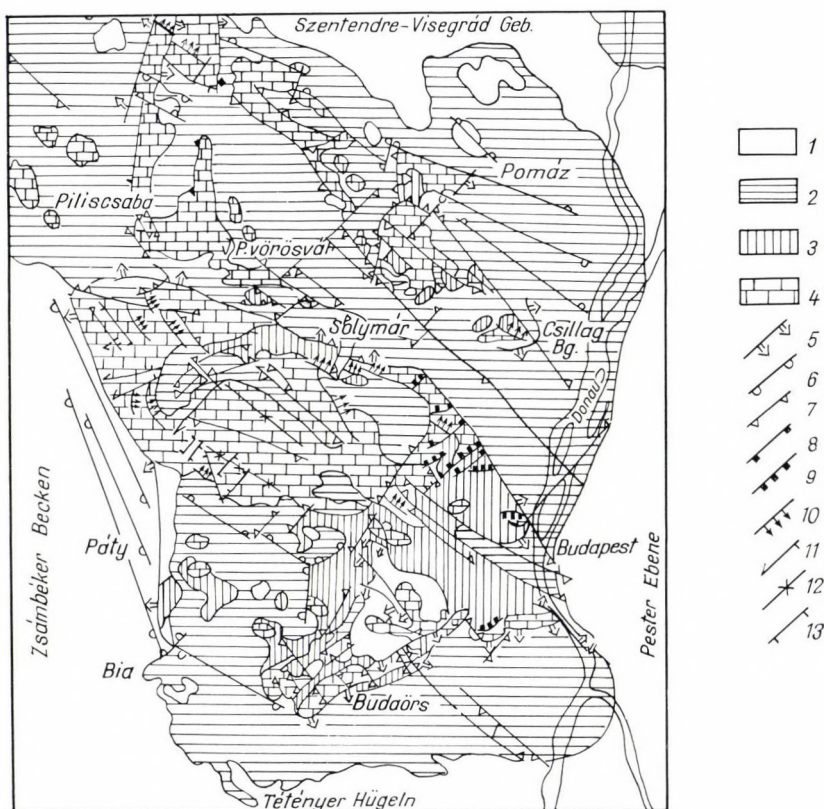


Abb. 7. Geologische Kartenskizze des Budaer Gebirges (Entw.: GY. WEIN, 1974). 1. Neogen; 2. Oligozän; 3. Eozän; 4. Trias; 5. Pleistozäne Verwerfungs- und »Pseudodiapir«strukturen; 6. Miozän-pliozäne Verwerfung; 7. Oligozän-miozäne Verwerfung; 8. Obere-eozäne Verwerfung; 9. Obere-eozäne Überschiebung; 10. Kreide (austrisch-mediterrane Überschiebung); 11. Kreide (austrische-mediterrane horizontale Verschiebung); 12. Kreide (austrische-mediterrane horizontale Faltentrog); 13. Verwerfung unbestimmten Alters

sich nicht einheitlich, die voneinander mehr oder weniger abgesonderten Schollen erhoben sich mit verschiedener Geschwindigkeit. Die Absonderung der Vortiefen hat sich fortgesetzt (K. MOLDVAY, 1966). Besonders bezeichnend sind für den Stil der pleistozänen Gebirgsbildung die pseudodiapir (schollendiapire, diapiroide) Strukturen (Gy. WEIN, 1974). Sie sind charakteristische, kleinere Trias-Schollen, welche die über ihnen gleichmäßig liegende eozäne-oligozäne Schichtfolge sozusagen durchbohrten und heute abgesonderte, sich steil emporhebende morphotektonische Formen bilden. Die schönsten Beispiele dafür sind der Gellért-Berg, Törökugrató, Sas-Berg, Róka-Berg. Das mindere Alter der Emporhebung wird nicht bloß durch die steile paläogene Mantelbildungen, sondern auch im pleistozänen Löß und Travertin gemessenen 20° Kippungen bekräftigt. Es sind auch nicht selten auf horizontale Verschiebungen verweisende Verwerfungsspanzern. Jene sind aber keine Zeichen für regionalartige horizontale Bewegungen, sondern stehen mit den vertikalen in Verbindung. Bezeichnend ist das Entstehen von offenen Brüchen, welche bunte lehmige Breccia, oder mit pleistozänen Thermalwassertätigkeit verbundene telethermale Mineralien, so hauptsächlich Aragonit und Kalzit ausfüllen. Es ist auch gelungen etliche zusammenhängende Bruchlinien am Szabadság-Berg (Schwabenberg) und am Ostrand des Gellért-Berges abzusondern. Letztere ist bis zur Petőfibrücke zu verfolgen.

Wir müssen darauf hinweisen, daß die pleistozäne Thermalwassertätigkeit, deren Spuren im ganzen Gebiet des Budaer Streifbruches zu beobachten sind, auch eine Eigenart dieses sich schnell emporhebenden Gebirges ist. Nur in dieser Zeit konnten jene hydrodynamische Bedingungen zustandekommen, welche ermöglichten, daß sich das senkende Karstwasser in Form von Thermalwasser emporströme (M. VENDEL, P. KISHÁZY, 1962).

Wir beabsichtigen mit Hilfe der an einzelnen Bach- und Donauterrassen entstandenen Travertinniveaus — welche nach einer schneller steigenden Phase eintretende Ruhe spiegeln — die steigenden Phasen des pleistozänen Abschnittes des Budaer Gebirges festzustellen. Es gelang uns — den am Gipfel des Szabadság-(Schwaben-)berges vorgefundenen levantiner Süßwasserkalkstein inbegriffen — 8 Erhebungsstufen zu unterschneiden. Von oben nach unten:

- VIII. Travertinniveau. 500—420 m Meereshöhe. Süßwasserkalkstein. Szabadság-Berg, Kakukk-Berg.
- VII. Travertinniveau. 240—235 m Meereshöhe. Ostseite Széchenyi-Berg, Máriaremete, höchste Niveau des Travertinzuges von Budakalász Üröm.
- VI. 200—195 m Meereshöhe.
- V. 220—195 m Meereshöhe.
- IV. 180—170 m Meereshöhe.

In den VI—IV. Stufen, teils in das ältere Pleistozän, teils ins Günz gehörende Travertinniveaus sind oft schwer voneinander zu trennen. Dort, wo

an einer Stelle, wie z. B. im Budakalász—Ürömer Abschnitt sich alle drei, bzw. vier Niveaus ausgebildet haben, ist es doch lösbar. Das V. Niveau hat M. PÉCSI (1973) mit dem den vértesszöllösch Menschen erhaltenden Niveau identisch genommen, auf Grund radiometrischer Altersbestimmung als 350.000 Jahre alt erklärt und es in das Günz-Mindel Interglacial eingereiht. Niveau IV. muß folglich unter dem Günz versetzt werden.

Das III. Travertinniveau, 160—150 m Meereshöhe, kann man von Pomáz durch Óbuda bis zum Mária-Berg folgen. Neulich denkt man daran, daß der Schloß-Berg (I. Bezirk Buda) und das kisceller Travertinniveau enthält teils das VI. Niveau (Gy. SCHEUER—F. SCHWEITZER, 1973). Das Alter wird auf 190.000 Jahre geschätzt.

II. 120 m Meereshöhe. Am Csúcs-Berg ist gut zu beobachten. Alter 70.000 Jahre.

I. 107—105 m Meereshöhe. Der junge Quellenkalkstein des Rómaifürdő gehört hierher.

Die endgültige Erhebung am Ende des Levantins können wir zum ersten Abschnitt der Walacher Bewegungen knüpfen. In dieser Zeit begann die Thermalwassertätigkeit, der wir das Entstehen von weiteren Travertinniveaus verdanken. Nachher entstanden die am Szabadság-Berg beobachteten NW—SO gerichteten, auch den Levantiner Kalkstein streifenden Verwerfungen. Das Entstehen von diesen können wir mit dem Walacher II. Abschnitt in Zusammenhang bringen. Der Bruch des Ördögárok hat sich wahrscheinlich in dieser Zeit, oder nachher in Laufe des mittleren Pleistozän, aber jedenfalls nach der Herausbildung des IV. Travertinniveaus erneuert, welchen wir im Metroschacht unter Vérmező beobachteten (Gy. WEIN, 1973). Die Verwerfungslinie berührt schon den Travertin des budaer Schloß-Berges nicht, dagegen das jánosberger oberpannone Abrasionsniveau rückt von 450 m Meereshöhe in Húvös-Tal und Rózsa-Hügel zu 275 m Meereshöhe ab. Die gestaffelte Emporhebungen, die in der Zeit des Entstehens der mittleren pleistozänen Terrassen erfolgten, knüpfen sich zur Pasadena-Phase. Schließlich die zwei letzten Emporhebungsphasen — zu welchen die I—II. Travertinniveaus gehören — knüpfen wir zu den Bakuer Phasen. Ich muß bemerken, daß bei Absonderung der Travertinniveaus ebenso, wie bei ihrer Einreihung in Bewegungsphasen Schwierigkeiten und Unsicherheiten vorhanden sind. Dieser vorgreifender Versuch, welcher sich bemüht die pleistozäne Bewegungen des Budaer Gebirges zu registrieren, kann keinesfalls als endgültig betrachtet werden.

Wenn wir als Vergleichsniveau zur Pester Ebene das obere Pannon als Basis nehmen, dann hat sich das Budaer Gebirge auf Einwirkung der Walacher Bewegung 120 m, im mittleren Pleistozän infolge der Pasadenaer Bewegungen 100 m und im oberen Pleistozän infolge der Baltischen Bewegungen 60 m erhoben. Wenn wir den jetzigen Wasserspiegel der Donau und die höchstgelegenen Thermalwassertrichter zueinander messen, so ist das Maß der ganzen Erhebung 370 m.

Gleichzeitig mit der gestaffelten Erhebung des Ungarischen Mittelgebirges und innerhalb diesen des Budaer Gebirges während des Pleistozän, haben die Kleine und die Große Ungarische Tiefebene periodische sinkende Bewegungen erlitten. Nach Untersuchungen von A. RÓNAI (1972) dort, wo die pleistozäne Schichtfolge vollständig ist, so bei Jászladány und Mindszent, konnte er 10 Abschnitte unterscheiden. Er setzt diese sinkende Abschnitte mit den in Ungarn festgesetzten 10 Stromterrassen gleich, und bringt den Zusammenhang von entgegengesetzten Bewegungen zwischen Beckenrahmen und Becken zum Ausdruck. Heute können wir die Korrelation — besonders infolge der Lückenhaftigkeit unserer Kenntnisse des Pleistozäns der Berggegende — noch nicht lösen, in der Zukunft wird aber eine wichtige Aufgabe der Pleistozänforschung die einzelnen Phasen der zwei isostatischen Bewegungen von entgegengesetzter Richtung, miteinander zu synchronisieren.

Die aufs Pleistozän bezeichnende Bewegungen setzen sich während des Holozän bis heute fort. Die rezenten Bewegungen haben nicht nur vertikale, sondern nach L. BENDEFY (1958) auch horizontale Bewegungen verursacht. L. BENDEFY zergliedert das Budaer Gebirge in drei Einheiten, die auf Grund sekulärer Bewegungen zu unterscheiden sind. Die südliche Tafel nennt er »Budaer Tafel«, die mittlere »Óbudaer«, die nördliche »Aquincumer«. Demgemäß weisen die waagrechten Komponenten der sekulären Bewegungen in der Budaer Tafel, welcher südlich vom Ördögárok sich erstreckt, nach SO. In den zwei nördlichen Einheiten merken sie die Bewegungen in S—SW Richtung. Der bis unseren Tagen dauernde Aufstieg des Budaer Gebirges ist bewiesen. Was dagegen die waagrechten rezenten Bewegungen anbelangt, so bedürften wir noch neuere Forschung, weitere Beweise, weil wir solche regionalartige Bewegungen weder im Neogen, noch im Pleistozän beobachteten, und eben deshalb deren Kontinuität wir nicht als bewiesen betrachten.

Das Budaer Gebirge umrandende Pester Ebene, bzw. das in jener herausgebildete Donautal und die Vortiefe des Bicskeer Beckens setzen die jetzige Lage des im Neogen begonnenen Gebirgsbildungs vorganges fest. Die tertiäre Insel, dann Halbinsel, welche zeitweise mehr oder weniger unter Wasser gerät, endlich hat während des 2 Mill. jährigen Pleistozäns seine jetzige Form erreicht und ist Budaer Gebirge geworden.

Wenn wir auf die Entwicklungsgeschichte des Gebirges zurückblicken, können wir diese Kontinuität in gewissen Beziehungen bis zu den altpalpinen Zyklus beendenden austrisch-mediterranen Phasen zurückführen. Die Hauptstrukturrichtungen der Ausbildung des »Budaer Streifbruches« sind im Tertiär mit Dilatationscharakter neu aufgelebt, und so können wir die Wirkung der geerbten Richtungen in den Tertiären Valer Graben, Ördögárok, Solymárer Tal, Piliser Brüche wiedererkennen. Die heutige Gestalt des Budaer Gebirges wurzelt tief in der geologischen Vergangenheit, durch welche wir auch die Richtung der Zukunft zu erkennen fähig sind. Die Entwicklung der Zukunft

ist im pleistozänen tektonischen Stil des Budaer Gebirges zu erwarten. Weitere selektive Erhebung des Gebirges, Entstehen von lockernden Brüchen dilatierenden Charakters, eine relative Vertiefung der Vortiefen geben die Umrisse des Entwicklungsstils der nächstfolgenden etlichen Mill. Jahren (Abb. 8).

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На основе структурно-геологической реамбуляции, автор выделяет 4 фазы в тектонической истории окрестностей Будапешта: I. Палеоальпийскую, II. Мезоальпийскую, III. Неоальпийскую, IV. Плейстоценовую.

В раннеальпийском тектоноцикле различаются геосинклинальная эпоха (с ладина по конец триаса) и орогеническая эпоха альпийско-карпатского типа. Движения австрийско-средиземноморской фазы создали сперва склады, потом поперечные разломы и чешуйчатую складчатость, и наконец горизонтальное смещение длиной 20—30 км. Так получилось, что Валатонская тектоническая линия продолжается в линии Дарно.

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CHANGES IN INTENSITY OF SALT AND COAL FORMATION IN THE NEOGENE

By

S. JASKÓ

BUDAPEST

In Europe and in the Near-East three climatic zones determine the spatial extension of the Neogene brown coal and salt beds (Fig. 1). Listed, from the north southwards, these climatic zones are as follows:

- a) Humid and semi-arid zone with coal formation but without evaporites.
- b) Semi-arid transitional zone, locally with coal formations, elsewhere with evaporites.
- c) Arid zone with evaporites and without coal formations.

This arrangement of the zones is based on the fact that when moving northwards the quantity of evaporites gradually decreases in general, and that of the coal formation gradually increases.

It is remarkable that in the zone between the longitudes of 28 to 30 degrees the location of the transitional zone has been displaced northwards by a distance of 5 to 6 latitude degrees. This displacement was probably caused by large-scale post-Miocene crustal movements.

The sequence of the consecutive periods of evaporite and coal formations can be exactly recognized in the central and western parts of the Paratethys as well as in the neighbouring areas (Fig. 2). In those places Neogene sedimentary sequences rich in fossils and classifiable in detail were formed. This provides a good opportunity for the exact determination of the geological age of each brown coal and salt bed. The graphs of Fig. 3 are deduced from the data of the areas marked in Roman numbers in Fig. 2. It can be stated from these graphs that in the central and western parts of the Paratethys (areas No. III-IX.) the time sequence of the formation of brown coal and salts is roughly the same everywhere. The intensity of evaporite formation assignable to the Tortonian (Badenian), further the transitional decrease of coal formation in the Sarmatian and Lower Pannonian are attributed to the general change of climate.

In the other areas the geological events followed each other with smaller or greater delays. It can be stated, in general, that in the second part of the Miocene the formation of evaporites was a uniform phenomenon. Large quantities of salt depositions and anhydrite beds were formed in Central Europe in the Badenian, in the Mediterranean in the Messinian and in the Middle East in the Gachsanan resp. in the Lower Fars and these form a uniform, elongated lithostratigraphic horizon.

Geophysical measurements and deep-bores put down in the basement of the Mediterranean Sea in the last decade revealed huge Miocene gypsum and rock-salt formations. Numerous new theories tried to explain the formation of Miocene evaporites* (Hsü, 1972; RYAN-CITA, 1972; NESTEROFF et al., 1972; CITA, 1972; KONRAD, 1973; SONNENFELD, 1974 and 1975; BENSON-RUGGIERI, 1974; DROOGER 1975; MALOVITSKY et al., 1975). These Paleogeographic hypotheses concern for the most part the basin of the Mediterranean

* Further, for sake of shortness, the so-called desiccated chemogenic sediments (anhydrite, gypsum, rock-salt, silvite) will be referred to as evaporite or salt formations according to the terminology proposed by E. SZÁDECZKY-KARDOSS (1930, p. 34).

sea itself. Studies as to how these new hypotheses correspond or contradict to the formerly accepted these regarding the Neogene of the European continent and containing rock-salt and brown coal beds, have been neglected.

The contrasts and relationships of the formation of evaporites and coals was first emphasized by E. SZÁDECZKY-KARDOSS (1927). The assignment of evaporite formation to certain geological periods was also demonstrated (E. SZÁDECZKY-KARDOSS, 1930; LOTZE, 1938 and 1965). The optimal conditions of coal formation in general were also discussed (STRAKHOV, 1967, p. 386; TIMOFEEV, 1968; LÜTTIG, 1971).

Detailed concrete investigation of such an aim of the epochs of the European Neogene has so far been lacking. Thus, by means of systematization and critical review of the data of enormous quantity in the paper below the author attempts portraying in a uniform way the relevant features of the Neogene as developed in Central and Southern Europe and the Middle East.

It is well known that saliferous sediments are characteristic formations of arid climatic facies. Salt formation is essentially caused by the evaporation of saturated solutions in rainless areas. The formation of coal beds, however, needs abundant precipitation and abundant vegetation. Consequently, coal formations cannot develop together with evaporites in space or in time (SZÁDECZKY-KARDOSS, 1927, p. 485).

Regarding the Miocene climatic zones as well as the spatial extension of coal and evaporite formations the following statements of universal validity can be said (Fig. 1). Three zones of nearly W—E strike have developed. Listed in a north to south direction these are as follows:

- a) humid zone and semi-arid zone with coal formations, without evaporites.
- b) Semi-arid zone locally with coal formations, elsewhere with evaporites.
- c) Arid zone with evaporites, without coal formations.

The arrangement of this type of the zones is based on the fact that, when moving northwards, the quantity of evaporites gradually decreases, and that of the coal formations increases, in general.

The fact can be also observed that when moving from eastwards the humid character also decreases within the zones *a* and *b*. The coal beds are frequent in the west, but can be found at the foot of the former mountains in the east. It is conspicuous that within the strip between the 28—30 degrees longitudes the position of the transitional zone was displaced in N—S direction by about 5 to 6 degrees latitude. In Europe the range of Miocene evaporites lies much farther north than in Asia Minor. This difference is also reflected in the extension of coal beds. South of 43 degrees latitude only a few exceptions represent the Miocene brown coal beds in the European continent. On the contrary, in Asia Minor the Miocene brown coal beds are frequently found also along the 37th degree of latitude. The assumption seems to be probable that originally the transitional zone extended from the west eastwards without

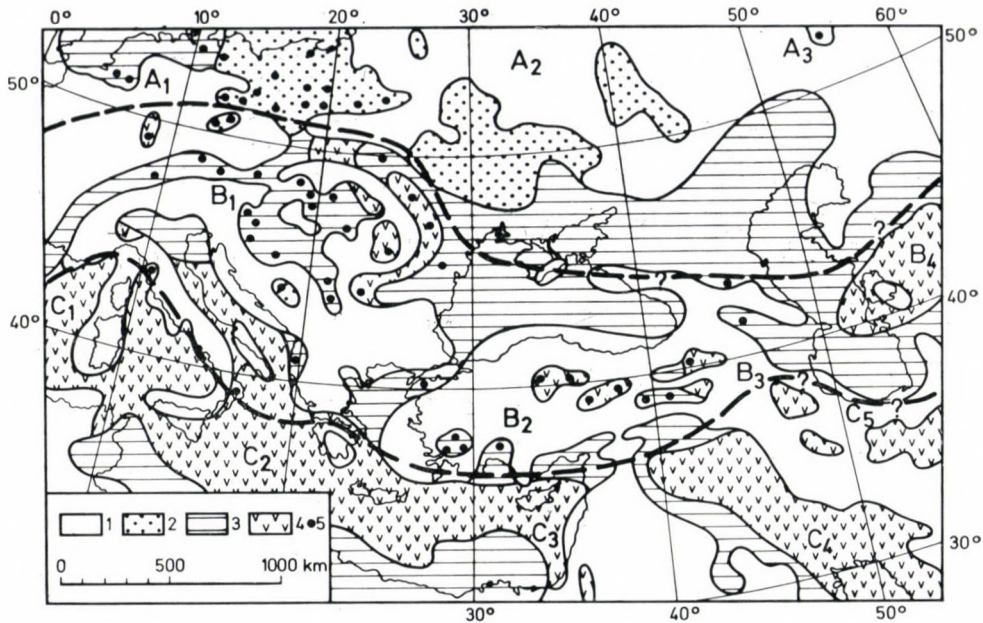


Fig. 1. Range of Miocene climatic zones. 1. Areas without Miocene sediments; 2. Miocene terrestrial, marshy, lacustrine and fluvatile deposits; 3. Miocene marine sediments (without evaporites); 4. Miocene marine sediments with gypsum, anhydrite and rock salt beds; 5. Brown coal occurrences; A. Humid zone; B. Semi-arid zone; C. Arid zone

any change, but the large-scale Post Miocene crustal movements displaced the European continent farther north as compared to Asia Minor. This assumption is supported by the fact that the Miocene sea-shore line turns to the south along the Vistula and Dnester, roughly in the region of Warsaw, Lvov (Lemberg) and Vinnitza.

Regarding the details the following can be stated.

East of the Netherlands numerous brown coal beds were explored on the German-Polish Plain (A_1)* as well as at the northern foot of the German Mittelgebirge, Erzgebirge and Sudeten. The eastern most coal traces were found on the eastern bank of the river Bug, near Brest (MANYKIN, et al., 1968, p. 63). On the German-Polish Plain the Miocene has a terrestrial facies poor in fossils. Marine sediments proved also by fossils are found in Lower Saxony and Schleswig-Holstein from the north (HINSCH-ORTLAM, 1974) resp. in Silesia from the south (LUCZKOWSKA-DYJOR, 1971). Thus, the coal-bearing strata of North Germany and Poland cannot be correlated exactly with the fossil-rich strata of marine and brackish facies of the Paratethys. By means of sediment-petrographic and pollen-spore analyses only the fact could be determined that the deposition of the Neogene terrestrial sequence of the area had been

* The types and numbers in brackets denote geological regions shown under the same marks in Fig. 1.

in progress for a longer geological time. Thus, the lowermost beds of the same coal occurrence are assigned to the Upper Oligocene, its middle and upper beds to the Lower, Middle, moreover, to the Upper Miocene by different experts (DYJÓR, 1970; CIUK, 1970; ZIEMBIŃSKA—NIKLEWSKI, 1966; AHRENS—LOTSCH—TZSCHOPPE, 1968; HUNGER, 1953; HOLUB—SKOCEK, 1968).

No brown coal and lignite beds are mentioned from the Neogene in the plain north of the Black and Caspian Seas (A_2). Only at the foot of the southern Ural Mountains does a minor coal basin occur (MOKRINSKI, 1962, pp. 439—443; MATVEEV, 1960, p. 421).

In the western part of the Paratethys, i.e. in the Neogene fore-deeps of the Alps and Carpathians (B_1) as well as in the Rhine graben and in North Italy the coal formations are intercalated in thick sedimentary sequences being distinguishable by marine fauna, thus their exact geological age can be determined, in general (JASKÓ, 1973 and 1975; MORETTI, 1962; NASCIMBEN, 1970; RAILEANU—GRIGORAS, 1963; SUF, 1952; ZAPFE, 1956; etc.).

Miocene brown coal occurrences are found in Europe usually north of 43° latitude. South of this line only exceptional brown coal beds which can be assigned to the Miocene can be found. E.g. in the southernmost part of the Appenine peninsula (CASTALDO—STAMPANONI, 1975) and in Albania (SHKUPI—BIBAJA, 1976) there are some smaller occurrences. When disregarding the brown coal occurrence in the southern Urals the eastern boundary of the Miocene brown coal beds can be drawn as far as 29° longitude.*

Concerning the extension of evaporites the following can be said. In Europe the northernmost occurrences of Miocene evaporites are found up to 50° latitude. The Aquitanian anhydrite and rock-salt beds of the Rhine-valley are found here (WAGNER, 1953, p. 709; SCHAD, 1964, p. 46).

No evaporites are known from the northern foreland of the Alps. Slight anhydrite traces are found only in the northern part of the Vienna Basin, at the Carpathian—Badenian boundary (JANÁČEK, 1974, p. 81). In the fore-deeps of the Northern Carpathians the first gypsum and rock-salt occurrences occur with 18° longitude and these can be traced as far as the foot of the Eastern and Southern Carpathians down to the Iron Gate in the south. The final part of this series extends well into the area of Bulgaria, where gypsum was explored by deep-bores. The evaporites of the fore-deeps of the Carpathians are found in two horizons in general, their minor part belongs to the Lower Miocene, while their major part can be assigned to the Middle Miocene (Badenian) (KWIATKOWSKY, 1974, p. 341; STOICA, 1961, p. 252; PARASCHIV, 1975, p. 41; JOVCHEV, 1961, p. 61; etc.).

The salt beds on the northeastern margin of the Pannonian Basin, in the Transylvanian Basin and in the Tuzla Basin of Yugoslavia were formed

* The geological age and extension of the brown coal beds of the Paratethys will be discussed in the second part of this paper.

also in the Middle Miocene (JANÁČEK, 1974, p. 79; VINOGRADOV, 1975, p. 103; CUIPAGEA—PAUCA, 1970, p. 246; SOKLIĆ, 1961, p. 73).

It is remarkable that the anhydrite traces of the middle part of the Pannonian Basin are of Sarmatian age (JÁMBOR, 1976, p. 304) i.e. these are younger than those mentioned above.

In Asia Minor (B_2) numerous Neogene basins are found separated from one another. (Roughly schematized, Fig. 1 shows only the greatest of them.) The Neogene strata of these basins consist mostly of marine, partly of terrestrial and volcanic formations. Due to correlation difficulties only the facts can be stated that the brown coal and lignite beds were formed in each basin independently of one another in different horizons of the Miocene and Pliocene. Numerous occurrences are found but most of them are insignificant. The only important occurrence is that of Elbistan, this is of Upper Pliocene age (BECKER—PLATEN, 1974, p. 683; STAESCHE, 1972, p. 332).

In the eastern part of Asia Minor gypsum strata are frequently found in marine sequences; locally rock-salt is also found. The exact geological age of the evaporites is unknown. According to NEBERT (1956, p. 84.) and IRRILITZ (1970, pp. 22. and 91.) these can be assigned to the Middle and Upper Miocene, resp. to the Lower Miocene.

In the environments of the Caucasus and Little Caucasus (B_3) only a poor coal formation took place in the Upper Oligocene, in the Upper Miocene and in the Pliocene (MOKRINSKI, 1962, p. 456.). Here a sedimentary sequence of marine facies and of great thickness was formed during the Middle and Upper Miocene in which significant gypsum and rock-salt beds occur (GABRIELIAN, 1962, pp. 344. and 359; ALIZADE, 1972, pp. 159. and 161.).

In the area between the Caspian Sea and Lake Aral (B_4) gypsum beds are frequent in a marine Miocene sediment and locally rock-salt is also found (SMOLKO—SMIRNOV—POPOV, 1972, pp. 337—365; SLIGIN—KOSTENKO—GALITZKI, 1971, pp. 257—261.). Here the boundaries between the three zones cannot be exactly traced thus it is possible that the whole of this area ought to be assigned to the arid zone (marked by C).

In the past decade several exploratory wells were carried out to explore the basement of the Mediterranean Sea. E.g. in the Gulf of Lions in the west (CRAVATTE et al., 1974), in the middle part of the basin (C_1) in the Tyrrhenean and Ionian Seas (C_2) (NESTEROFF et al., 1972) as well as near the shores of Israel in the east (C_3) (DERIN—REISS, 1973). Numerous hydrocarbon exploratory wells were also drilled in the Adriatic Sea (HARK—SCHÖNEICH, 1971). On the basis of these bores and geophysical measurements the evaporite-bearing Messinian strata could be determined in the basement also in the recently sea-covered areas. It has been proved that this formation is of much greater extension in the depth than assumed so far (MALOVITSKY, et al., 1975). In the terrestrial areas east of the Mediterranean Sea, i.e. in Lebanon, Israel and

Jordan, the Miocene sediments are for the most part absent. Only the tectonic depression of the Jordan trench is filled with a saliferous Miocene sequence (BENDER, 1968, pp. 89. and 162.).

From the eastern part of Syria far to the Persian Gulf enormous and continuous areas hide gypsum and saliferous deposits (C_4). These are found mostly in the "Lower Farsan" sequence which can be assigned to the Middle Miocene, and to a smaller extent in the Lower Miocene "Dibbane" strata (WOLFART, 1967; NAQUIB, 1967; METWALLI—PHILIP, 1974; WETZEL, 1974). The evaporite-rich strata of Persia which are probably contemporaneous with those of the "Lower Fars" are called "Gacharsan Formation" by several authors (GILL—ALA, 1972; JAMES—WYND, 1965; FALCON, 1974).

In the molasse zones of the mountain chains lying south of the Caspian Sea (C_5) the gypsum beds are frequent in the Upper Oligocene and Miocene sediments (Qom Formation), locally considerable halite and sylvine beds were also formed. The Neogene strata of the basin isolated from one another cannot be correlated with deposits occurring in either the Caucasus or in the environs of the Persian Gulf (ASOVSKI, 1949, p. 70; STÖCKLIN, 1974, pp. 225. and 230; ALAVI, 1972, p. 214.).

The alternations of the evaporite and coal formation periods during the Neogene can be well studied in the central and western part of the Paratethys as well as in the strata of neighbouring areas faunistically exactly classifiable. Here a good possibility is provided for an exact determination of the geological age of each bed.

In the fore-deeps of the young orogenic mountains the thickness of the marine, brackish and freshwater strata formed in the central and western part of the Paratethys and in the neighbouring areas amounts to 5—6000 metres. During the same period the adjacent massifs only slightly subsided and much thinner sequences were formed on their surface (JASKÓ, 1972, p. 168; and 1974, p. 91.). In the Middle Miocene marine strata locally gypsum and rock-salt deposit were accumulated. Limnic coal formations are frequently found as initial and final members of the Neogene sequences. Paralic beds are less frequent, these lie in the marginal areas.

In Fig. 2 the areas marked with dotted lines show only terrestrial (limnic, lacustrine and fluvatile) deposits. The parts shown illustrated by stripes are mostly marine but without an evaporite formation. In the crossed areas gypsum and anhydrite blocks are found in the marine sediments. Halite and sylvine deposits are shown by separate symbols.

On the basis of their reserves the Miocene brown coal and lignite occurrences are figured by three different signs (less than 50 million tons, between 50 and 500 million tons, and more than 500 million tons). In the map both operating and abandoned mines are shown. Notably, what the map shows is not an economic geographical portrayal of the contemporaneous

situation, but rather the spatial distribution of the rate of one-time coal genesis and the boundaries of generalized economic geological regions outlined. Concerning the reserves of coal numerous papers have been published. Some of the most important one are the following:

Austria: PETRASCHKE, 1921—1922; KIRNBAUER, 1967;

Hungary: VITÁLIS, 1939; BARTKÓ—HEGEDŰS—KÓKAY, 1963;

Yugoslavia: MILOJEVIĆ—ČIČIĆ, 1971; SIMIĆ, 1958;

Roumania: RAILEANU et al., 1963; PROTESCU, 1932;

Bulgaria: JOVCHEV, 1960;

Italy: PILOTTI—CASTRO, 1933; CASTALDO—STAMPANONI, 1975.

Reviews including the other countries of Europe: BISCHOFF—GOCHT, 1970; GORSKI et al., 1972; SCHWAHN, 1958; HAVLENA, 1964; MAKSAKOVSKI, 1968; FRIEDENSBURG, 1965.

Although the used data on mineral resources were compiled by different authors and according to different principles and in different times, their summarized results can be relied on as informative data. Map No. 2 as well as the graph of Fig. 3 were drawn on this basis. Out of the 97 brown coal deposits listed below 58 (60 per cent) possess coal resources less than 50 million tons. The coal resources of 26 occurrences fall between 50 and 500 million tons (27 per cent) and the number of deposits with coal resources amounting to more than 500 million tons is as low as 13 (13 per cent).

The Miocene brown coal and lignite deposits are as follows (in the same order as in Fig. 2).*

I. The Lower Rhine Valley and the coasts of the North Sea. 1. Limburg, 2. Lower Rhine Bay, 3. Hamburg, 4. Kiel, 5. Jüttland.

I/a. North-German and Polish Plain, German Central Mountains, Erzgebirge, northern foot of Sudeten. 6. Westerwald, 7. Nordhessen, 8. Wittenberg, 9. Bornhausen, 10. Rhoen Mountains, 11. Leipzig, 12. West Lausitz, 13. East Lausitz (Lusatia), 14. Legnica-Scinawa, 15. Gdańsk, 16. Trzcianka, 17. Zlochev, 18. Konin (Ochle), 19. Belchatov, 20. Rogozno, 21. Skierniewicz, 22. Morysin, 23. Adamov, 24. Brest, 25. Silesian Basin, 26. Turow, 27. Chomutov (Komotau), 28. Sokolov (Falkenau), 29. Cheb (Eger).

II. Mainz Basin and Rhine-trench. 30. Mainz Basin, 31. Rhine Trench,

III. Northern fore-deep of the Alps and the environment of the Vienna Basin. 32. Thurgau, 33. Wirtatobel, 34. Nördlingen, 35. Wackersdorf, 36. Budejovice-Trebon, 37. Trimmelkamm, 38. Amstetten, 39. Geras (Langau), 40. Wartberg, 41. Gloggnitz-Pitten,

IV. Fore-deep of the Northwestern Carpathians (No remarkable coal formations are known).

* The graphs No. I—X. in Fig. 3 were constructed from the data of groups No. I—X. The deposits of groups I/a, IX/a and X/a were neglected in these constructions because of correlation difficulties.

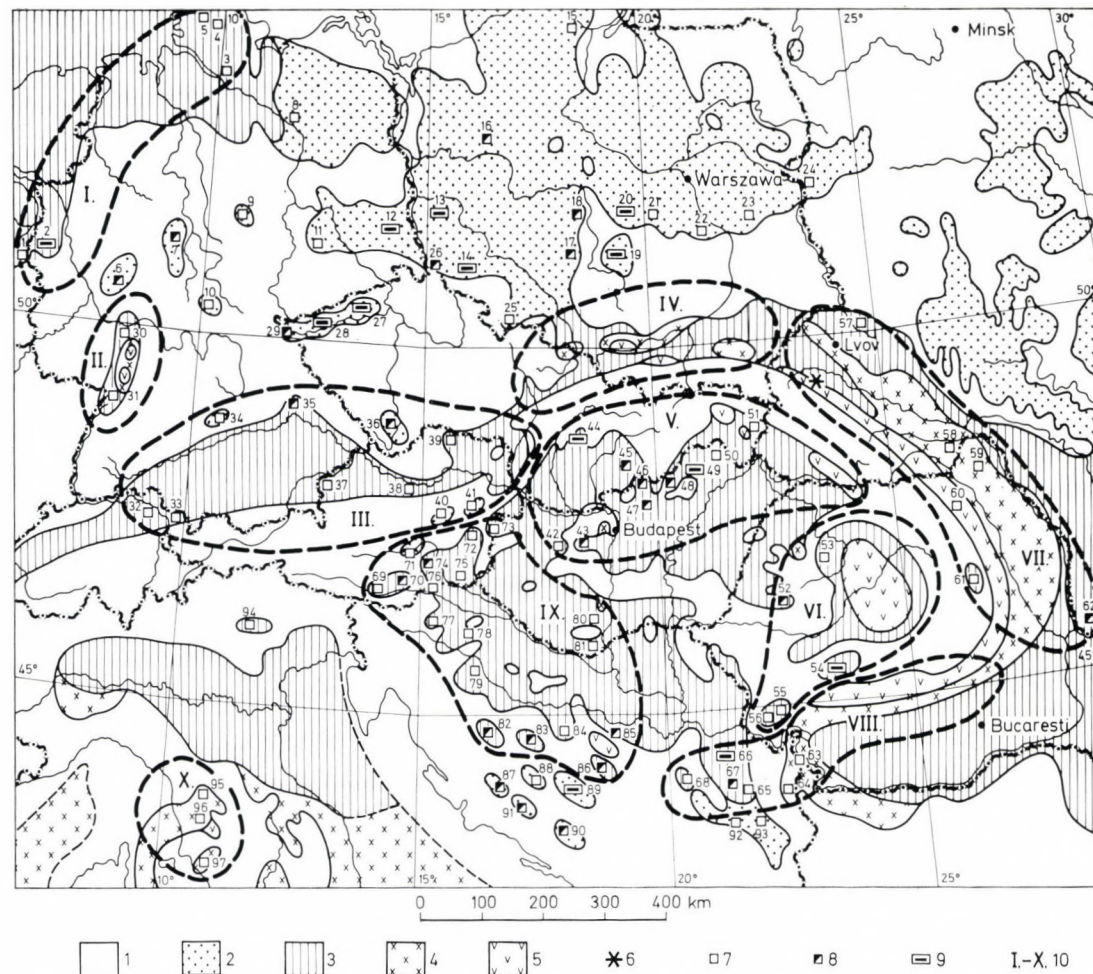


Fig. 2. Range of Miocene coal beds and evaporites in the central and southeastern part of Europe. 1. Areas without Miocene sediments; 2. Miocene terrestrial, marshy, lacustrine and fluvial deposits; 3. Miocene marine sediments (without evaporites); 4. Gypsum and anhydrite; 5. Rock-salt; 6. Potassium salt; 7. Minor brown coal deposit (less than 50 million tons); 8. Brown coal occurrence of medium size (between 50 and 500 million tons); 9. Major brown coal deposit (more than 500 million tons); 10. Serial numbers and boundaries of coal and salt regions

V. Northern margin of the Pannonian Basin. 42. Herend, 43. Várpalota, 44. Handlova-Nováki (Nyitrabánya), 45. Modry-Kámen (Kürtös), 46. Salgótarján, 47. Nagybátony, 48. Ózd, 49. Borsod, 50. Edelény, 51. Vihorlát.

VI. Transylvania. 52. Brad, 53. Almásvölgy, 54. Petroseni-Lupeni (Zsil valley), 55. Mehádia, 56. Bozovici.

VII. Fore-deep of the Eastern Carpathians. 57. Lvov, 58. Tshernovtzi (Tschernowitz), 59. Falciteni, 60. Baia, 61. Comanesti, 62. Reni-Vulkanești.

VIII. Fore-deep of the Southern Carpathians and the Morava Trench. 63. Mladenovac, 64. Zajecar, 65. Kucaj, 66. Petrovac, 67. Despotovac, 68. Arandjelovac.

IX. Southwestern part of the Pannonian Basin and the adjoining depressions. 69. Keutschach, 70. Lavanttal, 71. Knittelfeld, 72. Tauchen, 73. Brennborg, 74. Köflach-Voitsberg, 75. Feldbach, 76. Eibiswald, 77. Poljance (Pöltschach), 78. Radoboj, 79. Petrinja, 80. Hidas, 81. Pécsvárad, 82. Kamengrad, 83. Kotorvaros (Banja Luka), 84. Teslic, 85. Ugljevik, 86. Banovici.

IX/a. Smaller Neogene basins subsided into the Dinarides. 87. Livno, 88. Bugojno, 89. Middle Bosnian Basin, 90. Mostar, 91. Dubno, 92. Jankova-Klisura, 93. Aleksinac.

X/a. Southern margin of the Alps. 94. Valsugana.

X. Depressions accompanying the Northern Appennines. 95. Val d'Elsa, 96. Cecina, 97. Ombrione, Grosetto.

To make easier the correlation of different sequences any finer stratigraphic classification has been omitted, thus Fig. 3 shows several periods as one unit. Consequently, the Aquitanian and Burdigalian (Eggenburgian) were reduced to Lower Miocene, the Helvetian and Tortonian (Carpathian and Badenian) to Middle Miocene. The Sarmatian and Buglovian are in the Upper Miocene, the Meotian and Lower Pannonian (Pannonian s.str.) in the Lower Pliocene, the Upper Pannonian (Pontian s.str.) and Dacian in the Middle Pliocene, finally the Romanian and Levantinian in the Upper Pliocene.*

It can be seen in the graphs that the formation of brown coal and lignite could take place in all periods of the Neogene but with areally different intensity. It is conspicuous, however, that at the end of Miocene and in Early Pliocene time the lignite formation decreased, and locally completely ceased. In the Middle Pliocene, however, significant lignite formation took place again in the areas of the Vienna, Pannonian, Dacian and Thracian Basins. Apparently, except the Upper Miocene and Lower Pliocene periods, during the whole Neogene climatic conditions favourable for coal formation predominated in the central parts of the European continent. As to when and where more or less coal formations developed, this was dependent first of all on the temporally changing peculiarities of the local sedimentation processes.

* In the graphs the intensity of salt formation is interpreted quantitatively. Thus, the concept of SZÁDECZKY-KARDOSS diverging from that of LOTZE has been adopted.

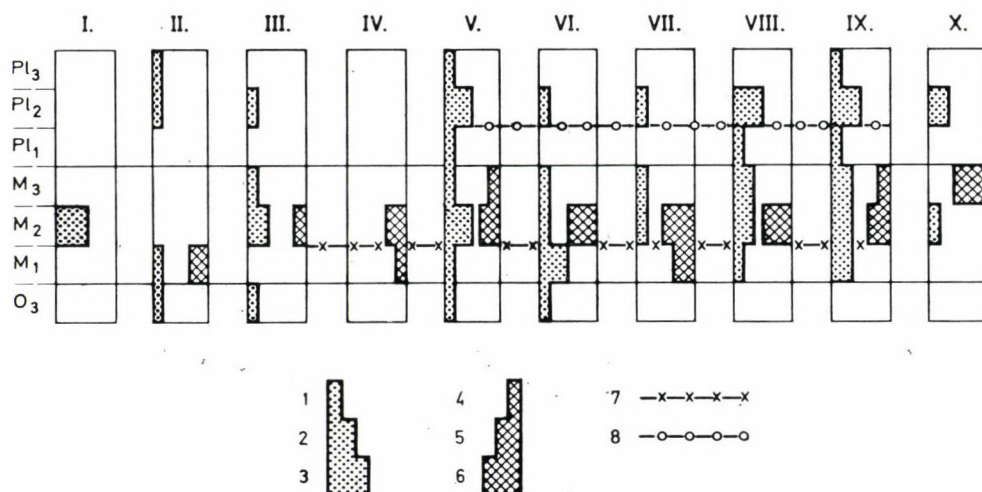


Fig. 3. Frequency of coal and evaporite formations according to geographic distribution and stratigraphic horizon (highly schematized graphs). I. Lower Rhine Valley and the shores of the North Sea; II. The Mainz Basin and the Rhine Trench; III. Northern Fore-deep of the Alps and the Vienna Basin; IV. Fore-deep of the Northwestern Carpathians; V. Northern part of the Pannonian Basin; VI. Transylvania; VII. Fore-deep of the Eastern Carpathians; VIII. Fore-deep of the Southern Carpathians and the Morava Trench; IX. Southwestern part of the Pannonian Basin and the adjoining depressions; X. Depressions accompanying the Northern Apennines; 1. Coal formation with small reserves; 2. Coal formations with reserves of medium quantity; 3. Coal formations with great reserves; 4. Evaporites with small reserves; 5. Evaporites with reserves of medium quantity; 6. Evaporites with great reserves; 7. Foot wall of the Badenian evaporite formations; 8. Footwall of the Upper Pannonian — Dacian lignite formations. O₃ — Upper Oligocene; M₁ — Lower Miocene; M₂ — Middle Miocene; M₃ — Upper Miocene; Pl₁ — Lower Pliocene; Pl₂ — Middle Pliocene; Pl₃ — Upper Pliocene

In the external fore-deep of the Carpathians, the Transylvanian Basin and the Pannonian Basin, evaporite formation took place mainly in the Middle Miocene, within the Tortonian. The quantity of evaporites formed in the Lower Miocene resp. in the Upper Miocene is smaller. Finally, the Pliocene of Europe does not show any trace of gypsum or halite.

The assumption seems to be probable that the high intensity of evaporite formation in the Tortonian as well as the transitional decrease of coal formation in the Sarmatian and Lower Pannonian can be traced back to the same reason, i.e. to the general change of climate in the European continent. Phytopaleontological research carried out in different countries showed, that in general the climate in the Early and Middle Miocene was humid, in the Late Miocene arid, in the Pliocene again humid (ANIĆ 1958, p. 191; BERGER, 1953, p. 232; NAGY, 1967, pp. 98, 101; SCHWARZBACH, 1966, p. 60. and 1974, p. 209. PLANDEROVA 1975, p. 386; BENDA & SICKENBERG 1975, p. 383.) This change was, of source, not of the same effect in different areas since the local influences might also effect microclimate.

When the decisive effect of different climatic changes upon the intensity of brown coal and evaporite formation be accepted, the question may arise that this had a contemporaneous effect not only on the European continent but in the basin of the Mediterranean Sea as well. In the latter region the formation of the Neogene evaporites is restricted only to the Messinian. According to the general view, however, the Messinian is younger than the Central European Tortonian (Badenian). As to several authors the Messinian is contemporaneous with the Sarmatian and Lower Pannonian (CICHA, 1970, p. 25; CICHÁ—SENEŠ, 1971, p. 211; GILLET—FAUGÈRES, 1974, p. 1178.). According to others the Messinian is contemporaneous with the Lower and Upper Pannonian (SELLI, 1970, p. 58; BARTHA, 1971, p. 21; BENDA—MEULENKAMP, 1972, p. 562; BERGGREN et al., 1974, p. 93; JIRICEK, 1974, p. 160). GRIVULESCU-ANDRESCU 1976, p. 1024.) In the Paratethys area evaporite formations are rare in the Sarmatian and absent in the Pannonian. In the Italian marine sediments probably contemporaneous with the Badenian, however, neither gypsum nor salt is found. These correlation difficulties are reflected by the difference of column X from the others (Fig. 3).

Two kinds of solution of this apparent contradiction may be offered:

a) The correlation concept ought to be changed, i.e. the Messinian would be contemporaneous with the Badenian and not with the Pannonian, thus the evaporite formation would be synchronous both in the Paratethys and in the Tethys. This, however, is in want of biostratigraphic evidence, thus only SONNENFELD proposed this possibility so far (1974, p. 1143).

b) The correlation classification is accepted without any change and the fact will be assumed that the effect of climatic changes is delayed in the region of the Mediterranean Sea as compared to Central Europe.

When accepting this latter variation the fact should be taken into account that, similarly to the formation of other sediments, that of evaporites is also affected by different local factors (e.g. geomorphological conditions, SZÁDECZKY-KARDOSS, 1927).

It is well known that the lot of inland seas is determined by the ratio of the evaporating water and of rain water, further by the discharge of the incoming rivers as well as by the size and direction of the joining strait communicating with the ocean. When evaporation is greater than the water supply, the basin will be dried. When the freshwater supply is greater than the evaporation (precipitation and water inflow), the water will be gradually limnified, the water surplus breaks leaks to the ocean. When, however, the freshwater supply is slight, but there is a possibility to water inflow from the ocean, evaporites will precipitate in the basin. In semi-arid zones the effects of these processes are manifested in different measures as a function of space and time. In this sense it can be understood that in the Middle Miocene anhydrite and rock-salt deposits were formed in the comparatively closed

bays and straits of the Paratethys, and in other basin parts this phenomenon did not occur. In the Middle Miocene no major river network existed. Along the line of the recent German Mittelgebirge, Erzgebirge, Sudeten and Podolia Ridge a flat watershed was extending and the rivers north of it did not supply the Paratethys. The ancestor of the recent river network started to form only at the end of the Miocene, resp. in the Pliocene, when the connection with the ocean was broken and the basin parts became successively limnified and filled up by alluvium. During the same period the sea predominated in the environs of the Mediterranean Sea and in the Middle East.

In the zone of transition between humid and arid climates the spread of vegetation producing the brown coal beds was influenced also by paleogeographical conditions. In the mountains more precipitation fell, thus at their foot forests and marshes could exist while in the semi-arid plains only steppe vegetation could be available. Under the transitional climate each subsequent climatic change had its own effect. Consequently, in the Miocene sequences of Europe the geohistorical changes of the factors influencing the sedimentation can be easily traced.

The enclosed maps show together the coal formations and evaporites formed during the whole Miocene. By drawing separately the sedimentary geological maps of the Lower, Middle and Upper Miocene, the differences between each stage could have been illustrated. The exact elaboration of such a study needs, however, further investigations.

Temporarily, the statement should be accepted that in the second part of Miocene a relatively short but rather wide-spread evaporite formation took place. The Badenian in Central Europe, the Messinian in the environs of the Mediterranean Sea, the Gachsaran, resp. the Lower Fars strata in the Middle East show huge rock-salt and anhydrite beds which can be traced over in great distances and form a continuous lithostratigraphic horizon. It is probable that these formations developed not at the same point of the absolute age chronological scale. Their formation, however, is the same, characteristic feature of the areas quoted above and can be satisfactorily separated from the preceding and subsequent geological events. Another peculiar event was the formation of lignite beds in the Vienna, Pannonian, Dacian and Thracian Basins at the end of the Pliocene sedimentation. This, however, spread over a much smaller region than the Miocene evaporite formation.

Preceding this evaporite formation the period between the Late Oligocene world-wide regression and Middle Miocene transgression had been suitable for coal formation. The beds formed during that time cannot be obviously observed on the sketches shown in Fig. 3. This is probably due to the fact that depending on local conditions these coal beds occur either as topmost members of the Late Oligocene regression or as initial members of the transgressive strata of different Miocene horizons.

Lotze has already created the rule of "contemporaneity", i.e. the intensity of salt formation reaches its maximum at the same time also in farther lying areas. According to him this is valid also of the coal formation (LOTZE, 1957, p. 188. and 1938., p. 152.).

All the afore-mentioned facts make this rule valid also of the Neogene of Europe and of the Middle East. Its hard-and fast application, however, is possible only within each isodiastrophic region at least, since among the areas of different evolution there are certain phase displacements.

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ИЗМЕНЕНИЯ ИНТЕНСИВНОСТИ ОБРАЗОВАНИЯ УГЛЯ И СОЛИ В НЕОГЕНОВОЕ ВРЕМЯ

Ш. ЯШКО

Резюме

Пространственное распространение неогеновых месторождений бурого угля и каменной соли в Европе и на Ближнем Востоке определяется тремя климатическими зонами (рис. 1). Перечисленные с севера на юг, эти климатические зоны следующие:

А) Влажная и полусухая климатическая зона с угольными месторождениями, но без эвапоритов.

Б) Полусухая (переходная) климатическая зона местами с угольными образованиями, а на других местах с эвапоритами.

В) Аридная климатическая зона с эвапоритами, без угольных пластов.

Такое распределение зон обусловлено тем, что в направлении с юга на север количество эвапоритов обычно постепенно уменьшается, в то время как количество угольных образований постепенно увеличивается.

Любопытно, что в полосе между $28-30^\circ$ долготы, местоположение переходной климатической зоны сдвинуто на примерно $5-6^\circ$ широты в направлении с севера на юг. Такие сдвиги вызваны, повидимому, интенсивными послемiocеновыми движениями земной коры.

Порядок периодов эвапорито- и углеобразования, следовавших друг за другом, хорошо опознается в центральной и западной частях Паратетиса, а также в сопредельных с ними районах (рис. 2). Дело в том, что здесь образовались неогеновые толщи, богатые ископаемыми организмами и поддающиеся более дробному расчленению. В связи с этим, вполне возможно точное определение геологического возраста отдельных пластов бурового угля и каменной соли. В результате обобщения частичных данных участков, указанных римскими цифрами на рис. 2, были составлены графики, приведенные на рис. 3. Из этих графиков вытекает, что в центральной и западной частях Паратетиса (участки III—IX) временная последовательность соле- и буроуглеобразования была более или менее одинаковой повсюду. Интенсификация в тортонском веке (бадениен) образования эвапоритов, а затем временное уменьшение углеобразования в сармате и раннем панноне были обусловлены всеобщим изменением климата.

В других участках геосторических событий последовали друг за другом с большей или меньшей задержкой. Все же можно сделать общий вывод, что эвапоритообразование было всеобщим процессом во второй половине миоценовой эпохи. Огромные по количеству запасов месторождения каменной соли и ангидритов проявляются в отложениях бадениена в Центральной Европе, в слоях мессиниано в районах Присредиземноморья и в отложениях гахсарана и нижнего фарса на Ближнем Востоке, причем они образуют выдержанный литостратиграфический горизонт, прослеживаемый на больших расстояниях.

RECENT LACUSTRINE DOLOMITE FORMATION IN THE GREAT HUNGARIAN PLAIN

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In the Danube—Tisza Interfluvium of the Great Hungarian Plain numerous saline lakes and sediments formed in former soda lakes are found. These lakes were formed in the depressions among the blown sand dunes of NW—SE direction produced by the prevailing wind. Late Pleistocene loess or Holocene blown sand lie at the bottom of the lacustrine sediments.

The climate of the Danube—Tisza Interfluvium is continental characterized by warm summer and frequently by dry periods of several weeks without precipitation. Under such conditions and due to strong evaporation the dissolved salts are concentrated in the waters of lakes of several decimetres depth and this is increased by the salt-rich ground water supply. The local term of this specific sedimentation environment is *semyek*.

Since the Early Holocene lime and dolomite have been precipitated from the basic water of high salt content just because of different factors. Major part of lime gas transformed into early diagenetic dolomite.

In the depressed areas surrounding the lakes the saline ground water evaporates and this results in the formation of limy-sodaic solonchak-solonetz type soils. Part of the soda precipitated on the surface is dissolved by rainwater and this also contributes to the increase of alkalinity in the lakes.

Introduction

Within the marine and terrestrial sedimentation areas numerous smaller but characteristic environments can be observed and each of them produces a specific sedimentary facies. Within the terrestrial sedimentation fluvial, lacustrine, eolian, glacial, etc. environments can be distinguished.

In the Danube—Tisza Interfluvium, one of the special areas of characteristic geological structure of the Great Hungarian Plain, special lacustrine sedimentation has been in progress (Fig. 1). The geological structure, the hydrogeological conditions of lakes, the morphology and climate of the area as well as the living world of the lakes have formed a peculiar sedimentation and which has been unique in the terrestrial sedimentation also on a world scale. In this paper this sedimentation process and the related sediments are introduced in order to enlarge the knowledge needed by the geological actualism.

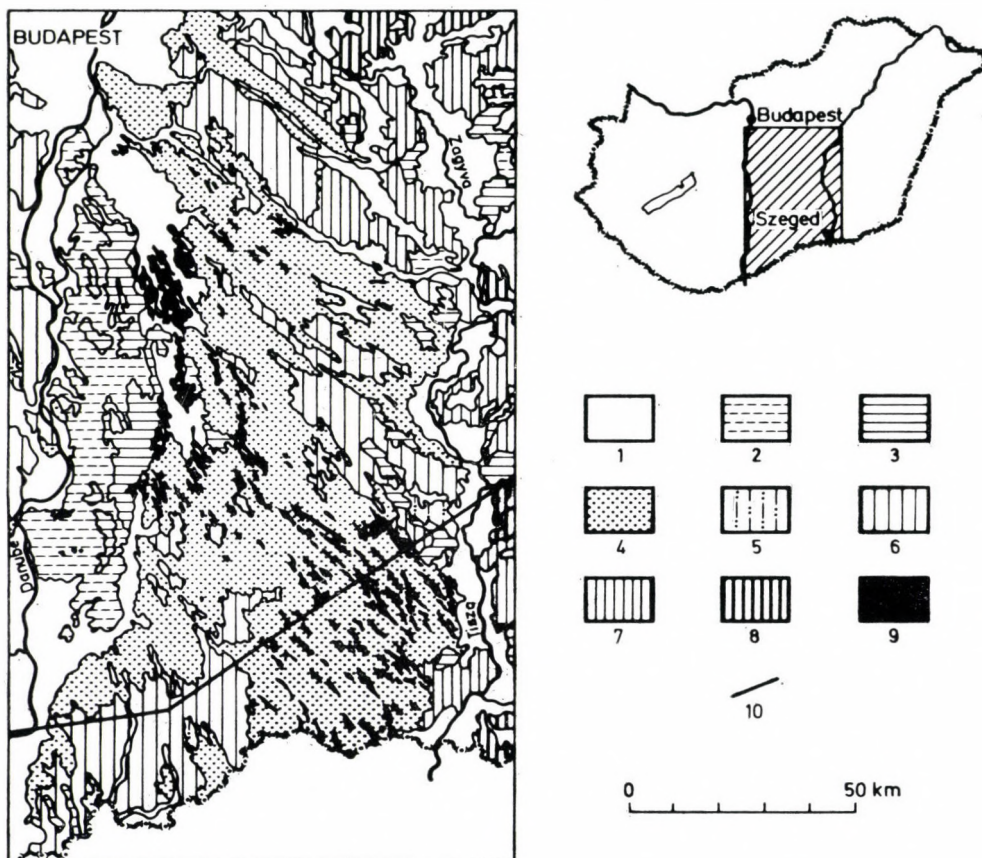


Fig. 1. Geological map of the Danube—Tisza Interfluvium and the extension of the *semyek* (recent and former lacustrine sedimentary environments). 1. Alluvium; 2. Redeposited loess; 3. Sodaic loess and clay, sand; 4. Blown sand; 5. Loessic sand; 6. Typical loess; 7. Alluvial loess; 8. Clayey loess; 9. *Semyek*; 10. Location of the profile of Fig. 2

Geological Setting and Quaternary Evolution of the Danube—Tisza Interfluvium

In the Early Pleistocene the direction of flow of the Danube was diagonal, i.e. to the southeast instead of the recent N—S direction. This direction was changed probably in the Günz—Mindel interglacial and the river started to form its recent valley subsequently. The winds of prevailing NW-direction indicating the Pleistocene glacial periods blew large quantities of sands from the Danube's flood plain and deposited onto the Danube—Tisza Interfluvium. The near-surface horizon of the Danube—Tisza Interfluvium is built up by this material and by loess deriving from the falling dust of the glacial periods (Figs 1 and 2). Consequently, at the end of the Pleistocene the surface of the Danube—Tisza Interfluvium was covered by loess and blown sand (I. MIHÁLTZ, 1953; B. MOLNÁR, 1973).

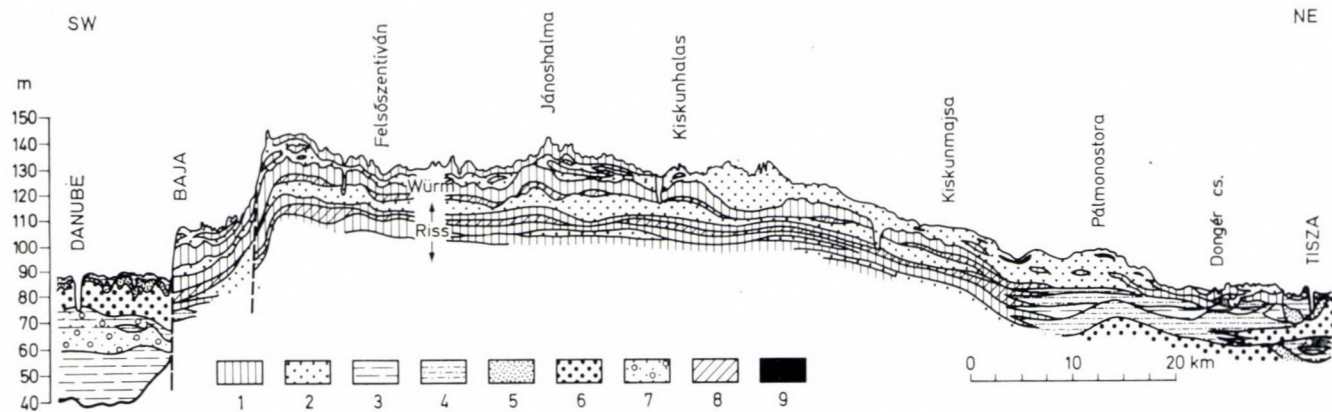


Fig. 2. Geological profile in the Danube-Tisza Interfluve according to MIHÁLTZ, I. (1953). 1. Loess; 2. Blown sand; 3. Clay; 4. Silt; 5. Fine-grained sand; 6. Medium-grained sand; 7. Gravelly sand; 8. Humic strata; 9. Carbonate rock

In the Holocene the climate has considerably changed also in the Danube—Tisza Interfluve and has become warmer as compared to the former periods. The climate of Holocene, however, has been not uniform since cool, warm dry and warm humid periods have alternated each other. Sedimentation followed these alternations. The blown sand deposited during the Pleistocene moved less in the humid and to greater extent in the dry periods. The blown sand has been arranged by the wind of prevailing NW—SE direction into dunes of NW—SE direction and among the dunes elongated depressions developed. When the climate became warmer and the ground water rose, part of these depressions was filled water, thus in these places constant or periodic lakes were formed.

Recently, the ridge of the Danube—Tisza Interfluve rises morphologically over the Danube valley by 30 metres and over the alluvium of the river Tisza by about 40 metres (Fig. 2). The lakes developed in this region have been studied by a complex research group. The research itself has extended over the formation, sedimentation and living world of the lakes.

The investigations carried out till now stated that the lakes of the Danube—Tisza Interfluve do not belong to any of the characteristic lake types and are most kindred to the ephemeric lakes but differ from them in numerous features. These can be characterized by the water depth of several decimetres and this makes possible the water movements caused by wind and this produces again flow phenomena in the sediments of the lake. Morphologically the lakes are frequently changing just due to the movement of blown sand. This resulted in the alternation of blown sand, soil and lacustrine sediments in the lakes' margins. The longitudinal axis and the shape of the lakes elongate in NE—SW direction (Fig. 3). The northwestern parts become narrow, the southeastern parts are generally widened. In certain cases several lakes lying parallel with each other is transformed into one lake system and this is caused by the burst of sand ridges (Fig. 3, Lake Kondor). The extension of the lakes varies but none of them exceeds 3 to 4 km length. Their widths correspond to the longitudinal measures, i.e. not wider than 1 to 2 km (B. MOLNÁR, 1971, 1973; B. MOLNÁR—I. M. MURVAI, 1975, 1976; B. MOLNÁR—M. SZÓNOKY, 1976).

The mean temperature in July is more than 22 °C in the Danube—Tisza Interfluve, the maximal average temperature of 50 years exceeds also 25 °C. The annual precipitation varies between 500 and 600 mm, the temporal distribution is not uniform. Droughts of high temperature and 4—5 weeks duration may also occur. The shallow lakes are of large surface as compared to their water quantity thus considerable evaporation is in progress. The lakes are supplied by rain water and by ground water infiltrating towards the local depressions. The rain water and surficial water supply of the lakes of the Danube—Tisza Interfluve is less than the annual evaporation (M. ANDÓ, 1964). Consequently, the water surplus may derive only from ground water.

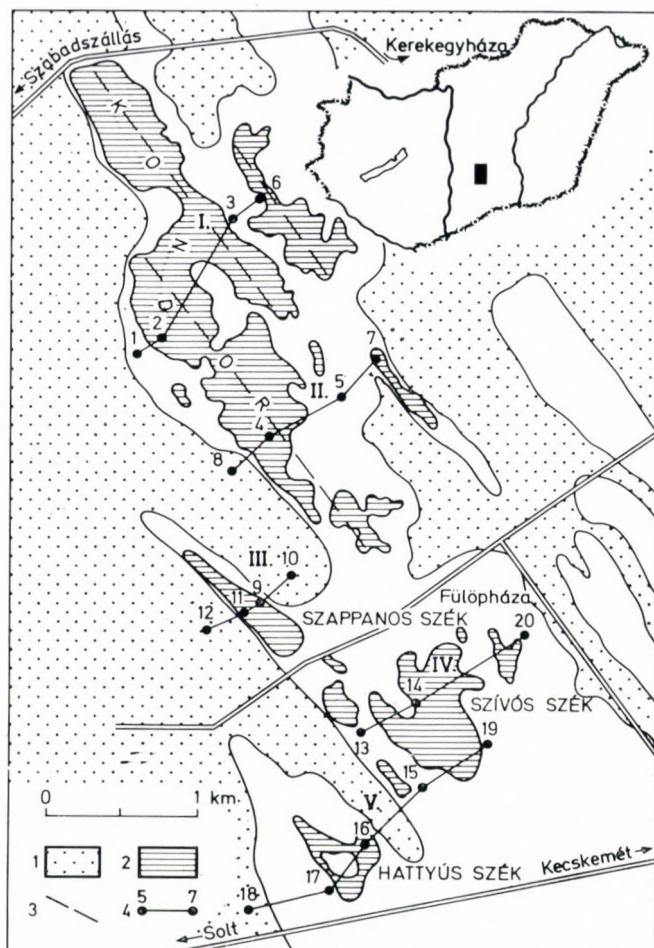


Fig. 3. Plot of the natron lakes of Fülöpháza indicating the locations of bores and geological profiles. 1. Blown sand; 2. Lacustrine (semyek) sediment; 3. Axial lines of the former depressions of Lake Kondor; 4. Location of the geological profile

The ground water table, however, depends on the weather conditions. All these factors are responsible for the periodically changing water surface and for the occasional drying of the lakes.

The peculiar sedimentation environment introduced above is called by the Hungarian folk-speech as "semyek" in order to make distinction from the lakes (the word "semyek" denotes the phonetic transcription of the Hungarian "semlyék" into English). In the Danube—Tisza Interfluvium there are a lot of semlyek (Fig. 1). Their detailed geological discovery has been carried out by choosing the typical areas, drilling geological exploratory bores, designing geological profiles resp. preparing trial trenches. All these explored in detail

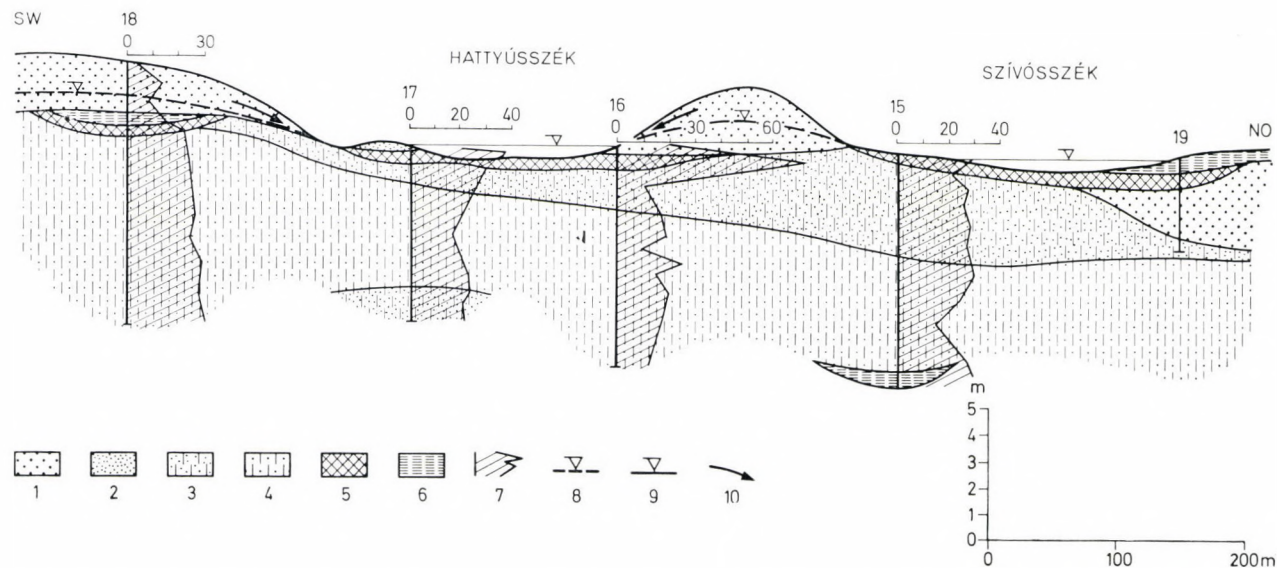


Fig. 4. Geological profile of the lakes Hattyússzék and Szívósszék. 1. Fine-grained blown sand; 2. Fine sand; 3. Loessic fine sand; 4. Fine sandy loess; 5. Carbonate mud; 6. Strongly humic unconsolidated silt; 7. Carbonate per cent; 8. The ground water table; 9. Water table of the lakes in July 1972; 10. Direction of filtration of ground water

the semyek, their environments and the sediments of them. The exploratory bores indicated among others that former lacustrine sediments are found in numerous sites where recently emergence is found, e.g. bore No. 18 in Fig. 4, or Fig. 2. Formerly shallow lakes were found in these places but the lakes were filled up by blown sand. It is obvious from the profiles that the extension of the lacustrine sediments is always greater than the recent surface of the lakes. Consequently, the extension of the lakes had been different as compared to the recent one. The greatest extension of lakes, however, reflect not the largest area of the lakes, but only the changes of it.

Formation and Study of the Semyek Sediments of the Danube—Tisza Interfluve

In the semyek sediments of the Danube—Tisza Interfluve the loess or blown sand is covered by lacustrine sediments of 30 to 110 cm thickness. Two types of lacustrine sediments can be distinguished. Mainly in the south-eastern part of the Danube—Tisza Interfluve the blown sand is overlain by hard carbonate rock of 20 to 60 cm thickness and this is covered by loose carbonate mud of 20 to 40 cm thickness (Fig. 5). In other regions of the Danube—Tisza Interfluve the lower hard carbonate rock is absent, only the loose carbonate mud of 20 to 80 cm thickness has developed. First the formation of the loose carbonate mud will be discussed.

1. Where the hard carbonate rock is lacking, the carbonate mud can be divided into three horizons according to its formation:

a) The bottom of the carbonate mud derives from the water flowing and infiltrating through its cover of higher carbonate content. This amounts to about 30 per cent of the carbonate mud or more (0.3 to 0.4 m) but it varies also within the same lense (B. MOLNÁR—I. M. MURVAI, 1975, 1976; B. MOLNÁR—M. SZÓNOKY, 1976).

The downward filtrating carbonate solution saturated the fine- or small-grained sand of the lake bottom, or occasionally the loess. The precipitated carbonate matter filled up first of all the pores of sand. Upward this shows transition without marked boundary. The carbonate content of this lower part is rather changing as reflected by laboratory analyses but varies between 25 and 50 per cent in general and depends considerably on the quantity of the infiltrating and side-ward moving water affected by the morphology. It differs from the middle horizon by the higher residue insoluble in hydrochloric acid. This residue consists of fine and small-grained sand, and of loess fraction when deposited onto loess.

b) The middle horizon of the carbonate mud is about 50 to 60 per cent of the total thickness of carbonate mud, i.e. 0.6 to 0.8 m. This carbonate precipitation has been of such an extent that it completely changed the original rock structure thus producing an individual geological formation



Fig. 5. The carbonate and carbonate mud excavation of Csolyospálos. *a*) Red-spotted sandy carbonate rock of looser structure; *b*) Hard light-grey carbonate rock; *c*) Dark-grey hard carbonate rock; *d*) Light grey carbonate mud with meadow soil in its upper part

often with 70 to 80 per cent carbonate content. This high value is rarely found but the carbonate content is always over 50 per cent. The grain size distribution of its insoluble residue is similar to that of the lower horizon, but regarding its quantity it is less. This strongly carbonaceous formation is loose, white or greyish-white under dry conditions. Its composition and features as well as its extension in the Danube—Tisza Interfluvium have been investigated by numerous researches (I. MIHÁLTZ, M. M. FARAGÓ, 1946, P. KRIVÁN, 1953, M. MUCSI, 1963, B. MOLNÁR—I. M. MURVAI, 1975, 1976, B. MOLNÁR—M. SZÓNOKY, 1976). The carbonate contains molluscs only in negligible quantity, and only a few species.

c) The upper horizon of the carbonate mud is only 10 to 15 per cent of the total thickness (0.1 to 0.3 m). Its carbonate content is less than that

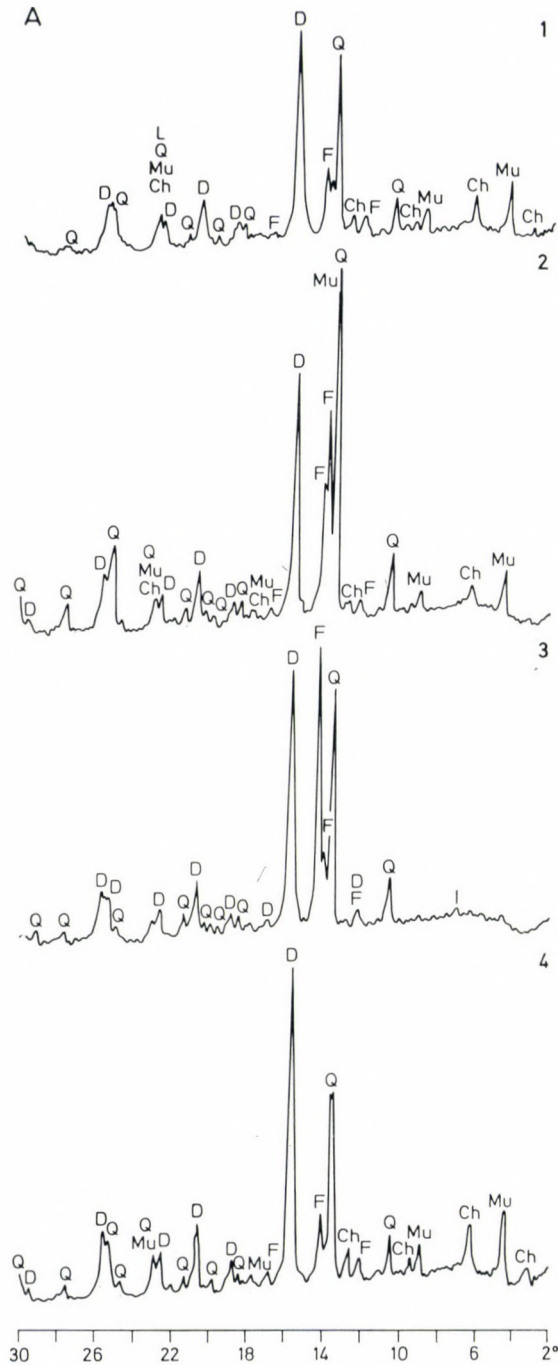


Fig. 6. X-ray diffractometric records of dolomite muds from the Danube-Tisza Interfluvium (Recording conditions: $\text{CuK}\alpha$, Ni-filter, 32 kV, 24 mA, 2 min). 1. Lake Kondor, Borehole No. 3. 0.5 — 0.6 m, typical carbonate mud; 2. Lake Hattyússzék, Borehole No. 17. 0.2—0.3 m fine and small sandy carbonate mud; 3. Lake Lodriszék, Borehole No. 14. 0.5—0.6 m fine and small-grained sandy carbonate mud; 4. Lake Kerek, Borehole No. 5. 1.6—1.8 m carbonate mud. Abbreviations in the X-ray diffractograms: D = dolomite; Ca = calcite; Q = quartz; F = feldspar; Ch = chlorite; Mu = muscovite; I = illite

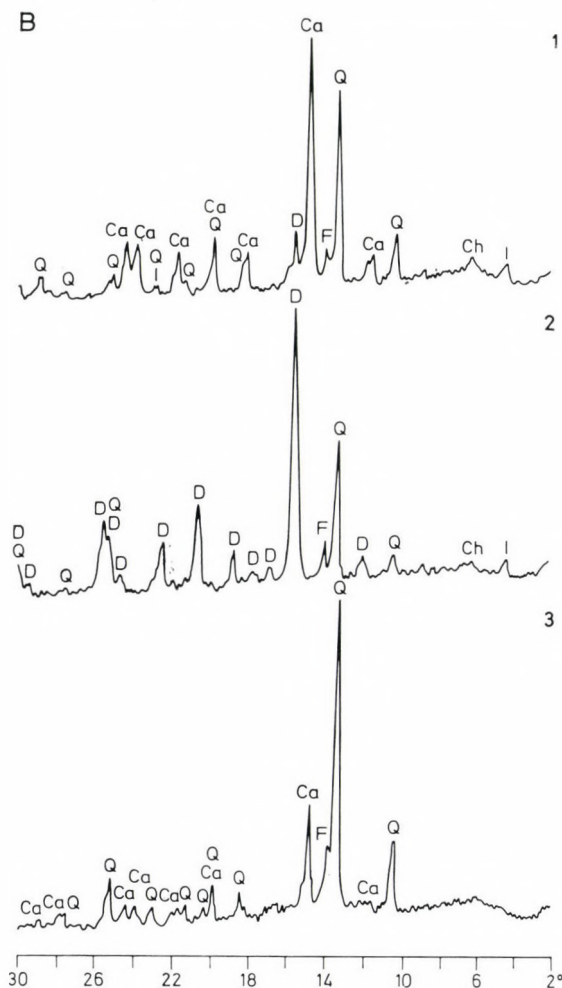


Fig. 7. X-ray diffractometric records of dolomite and limestone from the Danube—Tisza Interfluve (recording conditions see in Fig. 6.). 1. Nagybugac, loose dolomite; 2. Domaszék, loose limestone; 3. Dorozsma, hard dolomite (Legend of the records see in Fig. 6.)

of the middle horizon, i.e. is similar to that of the lower one (25 to 50 per cent). This was deposited in the lakes inundated for longest times. This upper horizon differs from the underlying ones first of all in its higher clay content in the insoluble residue, thus after drying it crackles down to 20 cm depth. In the geological profile shown in Fig. 4 the horizon of carbonate mud was not specified because technical reasons.

2. Mainly in the southeastern part of the Danube—Tisza Interfluve and overlying first of all blown sand harder carbonate rock of 20 to 70 cm thickness and loose carbonate mud of 20 to 50 cm thickness were deposited. This formation can be divided as follows.

a) At the bottom red spotty, veined sandy carbonate rock of looser structure developed everywhere. Its maximal thickness is 20 to 30 cm. In the rock the ratio of the carbonate cementing material and of the sand grains are nearly 1 : 1. Its formation is probably bound to the water flowing and infiltrating through its cover. The upward transition is without sharp boundary but can be exactly followed (Fig. 5).

b) The subsequent part is of 30 to 50 cm thickness and consists of hard, compact light-grey carbonate rock with plant marks. It contains first of all plant roots. Along the roots the rock is often porous. No or rare stratification is found. The sand grains are indistinguishable in the rock, in general. Its carbonate content varies at about 70 to 80 per cent. It is separated by a sharp boundary from its cover. In the Danube—Tisza Interfluvium this material was used to the foundations of houses.

c) The subsequent part is only of 5 to 8 cm thickness, very hard and often of more than 80 per cent carbonate content. The carbonate rock is of darker-grey colour and both up- and downwards it is bordered by sharp boundary and bedding plane. In the latter ones deep fissure traces resp. fissure fillings can be observed.

d) White-grey carbonate mud follows without sharp boundary and in a thickness of 30 to 60 cm and it is similar to the middle member of the strata described under 1. Its carbonate content varies between 50 and 60 per cent. As it is demonstrated in Fig. 5 its upper part often transformed into soil.

Among the semyek sediments especially the hard carbonate rock described in 2. was chemically investigated (I. MIHÁLTZ—M. M. FARAGÓ, 1946, P. KRIVÁN, 1953, M. MUCSI, 1963). X-ray diffractometric measurements were carried out by Nemeicz, E., and he stated that the carbonate muds are rather of dolomite, while the harder carbonate rocks of limestone composition (in P. KRIVÁN, 1953).

In our work the X-ray diffractometric investigations of several typical samples were also carried out. Results are shown in Figs 6 and 7, as well as in Table 1. In Fig. 6 the X-ray diffractometric records of the carbonate muds of the Danube—Tisza Interfluvium are shown. The dolomite peak occurs sharply in all records. In addition to dolomite quartz, feldspar, chlorite, muscovite and occasionally illite were determined.

Fig. 7 shows the X-ray diffractometric records of the harder carbonate rocks. Out of the samples analyzed those of No. 1 of Nagybugac and No. 3 of Dorozsma proved to be limestone, while that of No. 2 of Domaszék has been dolomite. The limestone and dolomite are somewhat poorer in minerals than the dolomite mud. In addition to calcite and dolomite quartz, feldspar and less illite were determined. Chlorite occurs, muscovite however, is lacking in this sequence.

Table I
Peculiarities of the carbonate rocks of the Danube—Tisza Interfluve

Mark of sample	Location	Macroscopic description	Mineral composition	
			by means of derivatograph	by means of X-ray diffractometer
A	1 Lake Kondor, borehole No. 3 0.5—0.6 m	typical carbonate mud	dolomite, quartz, chlorite	dolomite, quartz, feldspar, chlorite, muscovite, illite
	2 Lake Hattyússzék, borehole No. 17 0.2—0.3 m	small and fine grained sandy carbonate mud	quartz, dolomite, chlorite	quartz, dolomite, feldspar, chlorite, muscovite
	3 Lake Lódriszék, borehole No. 14 0.5—0.6 m	small and fine grained sandy carbonate mud	dolomite, (calcite), quartz, chlorite	dolomite, feldspar, quartz, (illite?)
	4 Lake Kerek, borehole No. 5 1.6—1.8 m	carbonate mud	dolomite, quartz, chlorite	dolomite, quartz, feldspar, chlorite, muscovite
B	1 Nagybugac, Lake Vész	loose carbonate rock	calcite, quartz, (dolomite?)	calcite, quartz, feldspar, dolomite, chlorite, illite
	2 Domaszék	loose carbonate rock	dolomite, quartz	dolomite, quartz, feldspar, chlorite, illite
	3 Dorozsma	hard carbonate rock	calcite, quartz	quartz, calcite, feldspar
	4 Zajná Hill	Triassic hard carbonate rock	dolomite	—

The derivatograph records of these samples were also carried out. Derivatograph records four kinds of changes when heating the sample to be analyzed. The instrument measures the temperature, the differential thermal (DTA) and thermogravimetric (TG) changes. The latter one indicates all the reactions which are accompanied by change in weight caused by heating. Finally, it records the derivated thermogravimetric (DTG) change which throws light upon the rate of the change in weight.

The carbonate muds which proved to be dolomite by X-ray diffractometric analyses, were determined also as dolomites by the derivatograph (Fig. 8, Table 1). The double peak of dolomite occurring at about 800 and 910 °C was of different intensity. The records showed not the usual dolomite curves, and the basic medium acting in the formation of dolomite has been responsible for this. In this way alkalis of carbonate and sulphate bonds of different quantities also developed and these disturb the regular shape of the curve.

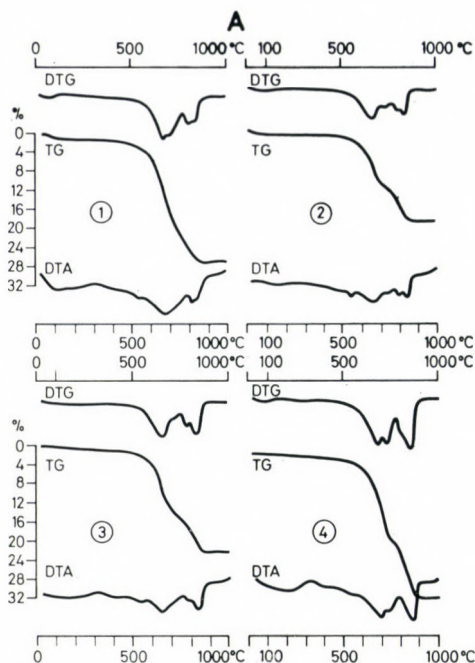


Fig. 8. Derivatographic records of dolomite muds from the Danube—Tisza Interfluvium. DTA = differential thermal analysis; TG = thermogravimetric; DTG = derivated thermogravimetric change. Sensibility: DTG = 1/10, DTA = 1/20, TG = 500 mg, disk-speed 100 min., heating rate 10°/min. Samples of this figure are the same as in Fig. 6.

The derivatographic results of the harder carbonate rocks are demonstrated in Fig. 9. The samples 1. and 3. show rather characteristic calcite peaks, thus these are limestones. The sample of No. 2 of Domaszék is of dolomite composition. The derivatographic curve of the latter one is somewhat elongated as compared to the dolomite muds.

The record No. 4 of Fig. 9 is the derivatographic curve of a typical Triassic dolomite recorded under the same conditions. When comparing this record with those of the semyek sediments the difference between the formation conditions becomes obvious. The Triassic dolomite is chemically much purer, therefore it shows the regular curve of dolomite.

The X-ray diffractometric and derivatographic investigations carried out in our work supported the results obtained by E. NEMECZ for carbonate muds. These called, however, the attention to the fact that the harder carbonate rock is not always of limestone composition, but dolomite is also found in them. Thus, the detailed investigation of several typical sections is needed since the composition within the carbonate profile can be determined only in this manner. These investigations are in progress. It is, however, obvious now that major part of the semyek sediments is dolomite.

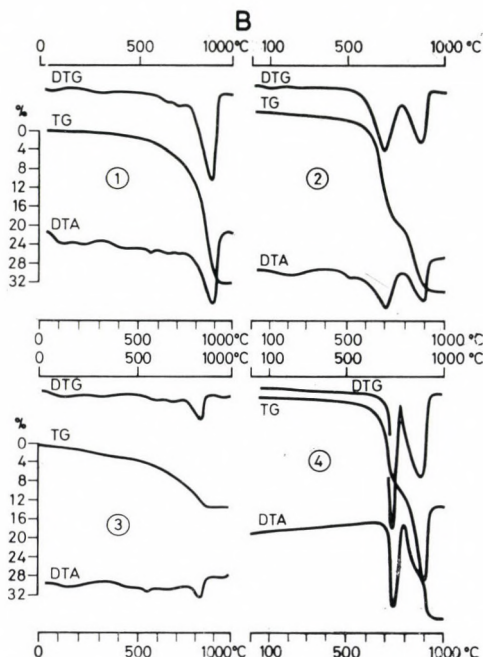


Fig. 9. Derivatographic records of dolomite (sample No. 1 and 3) and limestone (Sample No. 2) from the Danube—Tisza Interfluvium and of comparative Triassic dolomite (Sample No. 4). Recording conditions: see Fig. 8. Samples No. 1—3 are the same as in Fig. 7

Formation of the Semyek Sediment of the Danube—Tisza Interfluvium

To understand the deposition and precipitation of the semyek of the Danube—Tisza Interfluvium is highly promoted by the previous investigations of F. SMAROGLAY (1939), I. MIHÁLTZ—M. M. FARAGÓ (1946), P. KRIVÁN (1953), Zs. DVIHALLY (1970), J. SZÉPFALUSI (1970, 1976) and B. MOLNÁR—I. M. MURVAI (1975, 1976). Taking into consideration the geological processes as well as the former results, the formation of carbonates can be explained as follows.

The lakes are supplied by rain water and by the waters flowing towards the local depressions. The latter is especially important. The quantities of Ca and Mg are high in the sands forming the bases of lakes and blown out from the Danube valley, and in the loess and these are bound partly in form of CaCO_3 , or $\text{CaMg}(\text{CO}_3)_2$ or in form of other compounds. Dolomite occurs first of all in loess.

During soil formation the water filtrating towards the lakes dissolves the $\text{Ca}(\text{II})$ and $\text{Mg}(\text{II})$ and transports into the lakes. The chemical composition of some ground water samples found in the bores around some of the lakes has been analyzed (Table 2). The ground water contained dissolved salts in considerable quantities, i.e. 700 to 4000 mg/l. The quantities of $\text{Ca}(\text{II})$ and $\text{Mg}(\text{II})$

Table II

Chemical composition of the Lake Szappanosszék and that of the ground-waters from the bores in the neighbourhood

Sampling point	The water of the Szappanosszék		Lake Kondor, borehole No. 1		Lake Kondor, borehole No.2.		Lake Kondor, borehole No. 6		Lake Kondor, borehole No. 8		Lake Kondor, borehole No. 4		Lake Kondor, borehole No. 7		Lake Szappanosszék, borehole No. 12		Lake Szappanosszék, borehole No. 11		Lake Szívósszék, borehole No. 13		Lake Hattyússzék, borehole No. 17.		Lake Hattyússzék, borehole No. 16	
Chemical characteristics	mg/l	THAN's ionic equivalent per cent	mg/l	THAN's ionic equivalent per cent	mg/l	THAN's ionic equivalent per cent	mg/l	THAN's ionic equivalent per cent	mg/l	THAN's ionic equivalent per cent	mg/l	THAN's ionic equivalent per cent	mg/l	THAN's ionic equivalent per cent	mg/l	THAN's ionic equivalent per cent	mg/l	THAN's ionic equivalent per cent	mg/l	THAN's ionic equivalent per cent	mg/l	THAN's ionic equivalent per cent	mg/l	THAN's ionic equivalent per cent
Sodium Na ⁺	5689.2	97.19	140.2	59.92	108.9	27.32	22.3	10.58	48.1	17.67	264.6	53.13	239.4	64.91	400.2	55.63	3421.3	97.56	252.7	47.76	1080.2	88.58	256.5	68.75
Potassium K ⁺	184.0	1.85	9.0	2.26	8.3	1.22	0.8	0.22	2.0	0.43	9.6	1.13	5.5	0.87	16.5	1.35	120.0	2.01	1.6	0.18	31.3	1.51	4.5	0.71
Calcium Ca ⁺⁺	traces	—	53.1	26.03	160.7	46.26	98.9	53.79	95.0	40.6	54.0	12.44	49.2	15.30	6.4	1.02	traces	—	124.8	27.07	16.0	1.50	54.0	16.60
Magnesium Mg ⁺⁺	29.6	0.96	14.6	11.79	53.1	25.19	39.5	35.40	60.2	41.83	87.7	33.29	36.9	18.91	159.8	41.99	7.8	0.42	69.9	24.98	54.2	8.40	27.5	13.93
Iron Fe ⁺⁺	traces	—	traces	—	—	—	—	—	—	—	traces	—	—	—	—	—	—	—	—	—	—	—	traces	—
Ammonium NH ⁻	traces	—	—	—	—	—	—	—	traces	—	traces	—	traces	—	—	—	traces	—	—	—	traces	—	—	—
Chloride Cl ⁻	1631.7	18.08	74.2	20.55	67.2	10.93	3.2	0.98	97.7	23.28	39.4	5.13	46.8	8.22	66.8	6.02	975.2	18.03	150.6	18.46	368.2	19.58	92.8	16.13
Hydrogencarbonate HCO ₃ ⁻	3282.7	21.13	427.1	68.73	499.1	47.17	528.4	94.34	463.7	64.19	1199.6	90.73	771.2	78.74	1300.9	68.11	2092.9	22.48	646.8	46.06	2262.5	69.89	818.8	82.66
Carbonate CO ₃ ⁻⁻	4059.6	53.15	—	—	—	—	—	—	—	—	—	—	—	—	166.8	17.76	2400.4	52.46	—	—	—	—	—	—
Sulphate SO ₄ ⁻⁻	934.0	7.65	52.4	10.71	448.8	41.89	20.6	4.68	71.2	12.52	43.0	4.13	100.4	13.03	121.8	8.10	514.0	7.02	392.0	35.47	268.0	10.52	9.4	1.20
Meta-silica H ₂ SiO ₃	3.1	—	17.9	—	19.7	—	22.9	—	19.7	—	18.7	—	14.8	—	8.0	—	—	—	17.9	—	22.9	—	14.5	—
Total dissolved solids	15813.9	100.00	788.5	99.99	1265.8	99.99	736.6	100.00	857.6	99.99	1716.6	99.99	1264.2	99.99	2247.2	99.99	9531.6	99.99	1656.3	99.99	4103.3	99.99	1278.0	99.99
Alkalinity	189.10		7.00		8.18		8.66		7.60		19.66		12.46		26.88		114.30		10.60		37.08		13.42	
Total hardness in N°	6.80		10.78		34.71		22.92		27.14		27.73		15.37		37.65		1.79		33.54		14.70		13.88	
Carbonate hardness in N°	—		—		22.90		—		21.28		—		—		—		—		29.68		—		—	
Chemical reaction to phenolphthaleine	Alkaline		Acid		Acid		Acid		Acid		Acid		Acid		Alkaline		Alkaline		Acid		Acid		Acid	

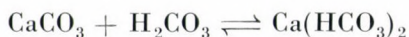
proved to be 16–160 resp. 8–150 mg/l. The Na(II), HCO^- , H_2SiO_3 and in some cases the CO_3^{--} contents have been also important.

Ground waters in which the total dissolved salt quantity exceeded 5000 mg/l were found by L. KUTI (1976) during the geological mapping of the Kecskemét environs. Ca(II) and Mg(II) amounted to 600–600 mg/l. In other regions of the Danube–Tisza Interfluvium ground waters of similar chemical character and rapidly changing chemical composition are found (A. RÓNAI, 1956). Consequently, the ground water carries salts of varied composition and of considerable quantity towards the lakes.

The shallow lakes of the Danube–Tisza Interfluvium are of large surface as compared to their water volume, thus in droughts there is considerable evaporation. The temperature fluctuation in the lake's water is high within a short section and within one day. Consequently, in the water there is a considerable chemical change both seasonally and during one day. This is reflected both by the quantity of the dissolved salts and by the ion/cation ratio. In general, the lakes are characterized by high total dissolved salt content and by strong alkalinity, i.e. their pH value is more than 9 in summer, moreover, it may exceed the pH value of 10 or 11.

From the Lake Szapannosszék near Fülöpháza (see also Table 2) water samples were taken in July 1972, parallel with those from bores. The total dissolved salt content of the lake's water exceeded at that time 15.000 mg/l. According to Zs. DVIHALLY (1970) and J. SZÉPFALUSI (1970, 1976) in extreme cases this value may amount also to 25.000 mg/l.

In winter most of the lakes in the Danube–Tisza Interfluvium lose their natron character, their waters contain then carbonate, their pH -value and alkalinity highly decrease. In winter, due to the lack of plant assimilation and to the predominance of dissimilation as well as to the autumn and winter precipitation, the carbon dioxide accumulates in the waters which gradually decreases the quantity of the precipitated CaCO_3 . The latter is transformed into calcium-hydrogencarbonate according to the following reaction:



In this case due to the accumulation of the free carbon dioxide the Ca(II) getting the lake and deriving from the ground water remains in solution, further, certain part of carbonate mud deposited in the lake's bottom will be dissolved, thus the quantity of Ca(II) may be ten times higher than in the summer period. In summer the carbonate and hydrogencarbonate content of almost all waters is nearly the same than that of the Na. In winter the major part of hydrocarbonate is bound to Ca(II).

In spring the assimilation increases, since the lakes contain algae and other aquatic plants in the near-shore strip. Their quantity is worthy of

mention. Parallel with increasing assimilation the free, then the equilibrated CO_2 will be used and this breaks the equilibrium developed during winter. The rise in temperature increases the evaporation which increases the salt concentration, i.e. also the Ca(II) and Mg(II) content but without the proportional increase of the solubility index of Ca(II) and Mg(II) . This process decreases at the same time the solubility of CO_2 . Parallel with the decrease of CO_2 content and with the increasing salt concentration the pH value will increase, thus CaCO_3 and to a smaller extent the $\text{CaMg(CO}_3)_2$ will gradually precipitate decreasing in this manner the Ca(II) and Mg(II) content of the water. After the disappearance of the free and equilibrated CO_2 the plants consume the half-bound CO_2 of NaHCO_3 . Consequently, in the water the quantity of precipitated carbonate will increase, the Ca(II) -ion and Mg(II) -ion contents decrease, while the pH and alkalinity values will increase again. In summer the illumination is very strong, thus dissimilation becomes predominant just due to the growth over the optimal value. In this case the whole process begins to turn back.

According to P. KRIVÁN (1953) the chemical processes of the lakes are affected by other physical factors (e.g. pressure decrease and waves), by the morphology of the lake's bed and shores, in addition to those mentioned above.

The precipitation of dolomite muds in alkaline medium is indicated also by their poor gastropod fauna. Only a few species less sensible to alkalinity and a few individuals are characteristic (A. HORVÁTH, 1950; A. RICHNOVSZKY, 1970). It is to be noted, however, that the gastropod fauna is much more abundant in the dolomite and limestone (M. MUCSI, 1963).

The geological consequence of the process outlined above is the repeated precipitation of Ca(II) - and Mg(II) -ion in form of CaCO_3 and $\text{CaMg(CO}_3)_2$ and the deposition of them in form of carbonate rock. The dolomite of these carbonates is formed partly by syndiagenetic precipitation, i.e. this is a syn-sedimentary deposit. Dolomite is formed in general in similar manner in the desert zones of the Earth in basic seas, though G. MÜLLER (1969) determined dolomite in a depth of one metre below the lake's bottom precipitated from the water of Lake Balaton of much less salt concentration.

This formation of dolomite has been dealt with in detail by H. E. USDOWSKI (1967, 1968). Accordingly, the transformation of calcite into dolomite may be promoted by the syngenetic replacement of Ca(II) - by Mg(II) -ion. This is the so-called early diagenetic dolomitization and this may be increased when the incompletely consolidated strata get the surface. Because of the frequent drying of the lakes of the Danube—Tisza Interfluvium the carbonate precipitations may get the surface, this phenomenon may also increase the dolomitization.

The evaporation of the saline ground water rising in a capillary way in the depressions of the lake's neighbourhood and the desiccation of the water

cause the concentration and precipitation of salts mainly in form of NaHCO_3 and Na_2CO_3 which are easily soluble in water. In these cases limy-sodaic solonchak—solonetz type soils will develop. Certain part of the soda precipitated in the surface will be washed into the lake, thus its alkalinity will be continuously increased.

It is questionable when the carbonate sediments were formed. It is a commonly accepted view that the loess discovered either by bores or found in the surface of the Danube—Tisza Interfluvium indicates the Würm-III glacial. Consequently, the overlying sedimentary sequence is of Holocene age. The Holocene sequence and its temporal evolution is well-known concerning the whole Danube—Tisza Interfluvium (I. MIHÁLTZ—M. FARAGÓ, 1946; B. ZÓLYOMI, 1953; A. HORVÁTH—S. ANTALFI, 1954; M. MUCSI, 1963, 1965, 1966; M. M. FARAGÓ, 1966, 1969; M. M. FARAGÓ—M. MUCSI, 1971; M. KOMLÓDI, 1966, 1969; M. ANDÓ—M. MUCSI, 1967; B. MOLNÁR, 1971; B. MOLNÁR—I. M. MURVAI, 1975, 1976; B. MOLNÁR—M. SZÓNOKY, 1976).

On the basis of the papers above the postglacial can be divided upwards by pollen, gastropod fauna and sedimentary investigations as follows: birch-pine, hazel, oak and beech phases. These phases can be difficultly parallelized with the Central European climate and vegetation phases of Firbas. By all means, however, these may correspond to the phases IV—IX of Firbas (F. FIRBAS, 1949).

According to M. M. FARAGÓ (1966, 1969) the main period of formation of the carbonates of the Danube—Tisza Interfluvium fell to the end of the dry hazel phase and to the first part of the oak phase of the Holocene. Absolute age determinations are in progress, it is, however, probable that the hard carbonate rock was formed in the main formation periods, i.e. in the hazel and oak phases. In other parts of the Danube—Tisza Interfluvium loose carbonate mud was simultaneously precipitated as heteropic facies. Finally, after the main formation period, everywhere the deposition of the latter one became predominant and this process is in progress recently, too.

No sufficient explanation exists regarding the facts that why only in certain sites the hard carbonate rock was developed and what is responsible for the sudden changes which occurs in the profiles by sharp boundaries and bedding planes between the hard carbonate rock and loose carbonate mud (Fig. 5). To solve this problem further investigations are required.

The natron lakes of the Danube—Tisza Interfluvium have come to an end in two manners. These are buried either by blown sand, or filled up by the carbonate deposited onto the lake's bottom as explained by F. SMAROGLAY (1939). The latter phenomenon causes the rise of lake bottom over the ground water table. The lake is first overgrown by plants and parallel with progressive senescence the aquatic vegetation is replaced by steppe vegetation (Fig. 5).

The present picture of the Danube—Tisza Interfluve is the result of the geological process described above. This shows the natron lakes covered by water as well as the filled lacustrine environments reflected by their sediments and their spatial alternation (Figs 1 and 3).

In older, mainly Triassic, but in younger continental formations geological phenomena are often found which can be hardly explained but as to our opinion are highly similar to the *semyek* formations of the Danube—Tisza Interfluve (J. MROCZKOWSKI, 1972; E. GRUMBT, 1974). The comparative investigation of these formations from this point of view seems to be worthy to carry out in the near future.

Summary

1. In the Danube—Tisza Interfluve of the Great Hungarian Plain numerous natron lakes and sediments formed in former natron lakes are found (Fig. 1). These lakes were formed in the depressions between the dunes formed by the prevailing wind of NW—SE direction. The basis of the lacustrine sediments consists of Late Pleistocene loess or Early Holocene blown sand.

2. The region in the Danube—Tisza Interfluve is continental climate characterized by warm summer temperature and droughts of several weeks. In this environment and due to strong evaporation the dissolved salts are extremely concentrated and this is increased by the salt-rich ground water supply (Table 2). The local term of this peculiar sedimentary environment is *semyek*, in order to distinguish it from lakes.

3. On the effect of different factors (vegetation, rapid change in temperature, waves, etc.) lime and dolomite have precipitated from the lakes' water of high salt concentration and high alkalinity since the beginning of the Holocene. The major part of the lime is transformed into early diagenetic dolomite (Figs 6—9; Table 1).

4. In the deep-seated areas and spots of the lakes' environment the rising saline ground water evaporates and limy, sodaic solonchak—solonetz type soils are formed. Part of the soda precipitated in the surface is washed into the lakes and this also contributes to the increase of their alkalinity.

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СОВРЕМЕННОЕ ОБРАЗОВАНИЕ ОЗЕРНЫХ ДОЛОМИТОВ НА ТЕРРИТОРИИ БОЛЬШОЙ ВЕНГЕРСКОЙ НИЗМЕННОСТИ

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Р е з ю м е

На Междуречье Дуная и Тиссы, являющемся частью Большой Венгерской низменности известны многочисленные солончаковые озера и осадки, накопившиеся в ранее существовавших озерах такого характера. Эти озера образовались в депрессиях, формировавшихся между сериями дюн эоловых посков СЗ—ЮВ простирания, обусловленных деятельностью преобладавших ветров. Базу озерных осадков образуют позднплейстоценовые лессы или раннеголоценовые эоловые пески.

Климат Междуречья Дуная и Тиссы является континентальным с относительно высокой летней температурой и частыми засухами, продолжающимися на протяжении нескольких недель. В таких условиях, из-за сильного испарения, сильно концентрируются соли, растворенные в воде озер глубиной, еле-еле достигающей нескольких дециметров, причем это дополнительно усиливается в результате восполнения грунтовыми водами с повышенной соленостью. Такая характерная среда осадконакопления получила местное название шемье.

Из щелочной озерной воды под влиянием различных факторов начиная с начала голоцена выделяются известь и доломит; большинство извести превращается в раннедиагенетический доломит.

На более глубоких участках, окружающих озера, соленая грунтовая вода испаряется и этот процесс приводит к образованию почв солночако-солонцевого типа. Атмосферная осадочная вода часть соды также унесет в озера, усиливая тем самым щелочность озерной воды.

MINÉRALOGIE, GÉOCHIMIE ET GÉNÉTIQUE DES BAUXITES DU VIETNAM DU NORD

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D'ALUMINIUM (ALUTERV-FKI) BUDAPEST

Entre 1972 et 1975 l'entreprise ALUTERV effectuait des recherches de bauxite sous la direction géologique de l'auteur, dans la province de Lang Son au Vietnam du Nord. Ce travail récapitule les abondantes observations recueillies au cours des recherches et les résultats d'examen des matériaux détaillés effectués dans les laboratoires et instituts de recherches scientifiques de Hongrie.

A cet espace de l'Asie du Sud-Est les gîtes de bauxite primaire à diaspore et chamosite paléozoïques diffèrent complètement tant du point de vue géologique, minéralogique que génétique des gîtes de la province de bauxite méditerranéenne et qui ont été altérés en argile et réaccumulés après oxydation, dans le Néogène; des gîtes secondaires précieux du point de vue industriel y ont été formés. Ces gîtes de bauxite présentent une affinité génétique avec les gîtes de bauxite de la Chine et de l'Oural.

Les gîtes de bauxite du Vietnam du Nord situent le long des principales lignes structurales, en direction du N-NW au S-SE, au bord de la plaque de la Chine méridionale, dans une zone de mio-géosynclinal.

Dès le Cambrien jusqu'au Trias supérieur, ce territoire présentait une dépression à interruptions plus au moins longues. Le Paléozoïque ancien et récent est représenté par calcaires biogéniques déposés aux milieux littoral et de mer ouverte. La seule période de genèse de la bauxite prouvable était dans le Permien supérieur, entre les étages kazanien et tatar. A cette époque, la surélévation épirogénique, accompagnée des mouvements oscillatoires était partielle. La bauxite a été formée en terre ferme et accumulée en partie en terre ferme, en zone littorale inondée en oscillation périodiquement et en partie en baies étant en voie d'isolation. En allant vers la mer plus ouverte, comme faciès hétéropiques on trouve de l'argilite, puis l'argilite divisé par bancs de calcaire. La bauxite déposée en zone littorale est divisée par formations houilleuses. En allant de la terre ferme vers la mer plus ouverte, des faciès de bauxite caractéristiques ont été formés caractérisés par la paragenèse minéralogique purement oxydante, oxydante-réductive et purement réductive. Les gîtes de bauxite ont été recouverts des sédiments transgressifs de l'étage tatar du Permien supérieur, composés de la marne houilleuse et du calcaire. Au toit plus récent, on trouve de l'aléurolite d'âge triasique et des volcanites acides surtout d'âge jurassique.

L'orogénèse est déroulée dans le Trias supérieur. Sous l'effet du tectonisme intense le complexe de calcaire ancien et plus récent a été fortement fracturé et écaillé (plus tard fortement karstifié), l'aléurolite triasique est plissée-fracturée.

Le calcaire du toit est dissous sans trace au-dessus des gîtes de bauxite restés en position de gisement relativement non perturbée. Le complexe affleuré est altéré en argile, en place. En fonction des conditions géomorphologiques, le complexe situé aux terrains topographiquement les plus élevés est continuellement érodé, réaccumulé aux pentes des collines abruptes et accumulé de nouveau dans les vallées plus profondes. De ce point de vue on peut distinguer trois types de gîte détritique: types de Leo Cao, Ma Meo et de Dong Dang. Des procédés oxydants intenses accompagnent l'effritement, la réaccumulation et l'accumulation qui exercent leur effet différemment dans les types de gîtes différents. Au cours de l'oxydation des changements considérables ont été déroulés et déroulent encore dans la composition minéralogique et chimique originale de la bauxite qui ont produit à la phase initiale l'amélioration qualitative des détritiques de bauxite, mais à la phase suivante c'était l'abaissement de la qualité.

Introduction

Sous la direction géologique de l'auteur l'Entreprise de Construction d'Industrie d'Aluminium (ALUTERV) a exécuté la recherche de bauxite, entre 1972 et 1975, dans la République Démocratique du Vietnam (RDV), dans la province de Lang Son. Au cours de celle-ci, il y avait de possibilité à la reconnaissance d'autres territoires bauxitifères de la RDV, parmi lesquels du point de vue industriel le plus important est le gisement de Tap Na, situé dans la province de Cao Bang. D'après leurs structure géologique et caractéristiques pétrographiques ces gîtes de bauxite diffèrent complètement des gisements connus de la province de bauxite méditerranéenne, et par suite notre expédition exigeait une conception de recherche et le développement d'une pratique complètement différentes de celle employée en Hongrie.

Dans le travail de l'Expédition hongroise et dans la composition du rapport géologique détaillé suivant ont participé L. HAVAS, E. HAVAS—SZILÁGYI, A. MINDSZENTY, A. VÉGH et I. VÖRÖS.

La recherche de bauxite touchait un territoire de 200 km² environ, outre les travaux géodésiques comprenant la cartographie géologique en échelle de 1 : 25.000 et la cartographie géologique de bauxite en échelles de 1 : 5.000 et 1 : 10.000, et aussi les recherches d'exploration: approfondissement des puits creusés et des sondages. A côté de celles-ci, la recherche du complexe à détritits de bauxite — à l'aide de la méthode géoélectrique — déroulait aussi avec succès (P. EGERSZEGI).

Au Vietnam, nous avons élaboré la méthode d'enrichissement du complexe composé en détritits de bauxite durs et en argile meuble. A côté de la nécessité industrielle de celle-ci, on a obtenu des informations précieuses sur l'effritement de la bauxite primaire in situ et sur la transformation minéralogique et géochimique profonde au cours du remaniement.

On a effectué les examens des matériaux aux instituts de recherche scientifiques, laboratoires et entreprises industrielles de Hongrie.

A côté des analyses routines, c'est l'Entreprise de Recherche de Bauxite à Balatonalmádi qui a effectué les examens des éléments en traces. (A. SELÉNYI, M. SIKLÓSI-JENEL.)

Pour reconnaître la texture et la composition minéralogique qualitative de la bauxite, on a préparé 150 lames minces à l'ALUTERV (A. MINDSZENTY).

C'est le Laboratoire de Recherches Géochimiques de l'Académie Hongroise des Sciences qui a effectué les analyses de diffraction à rayons X (GY. BÁRDOSSY et M. TÓTH).

Pour éclaircir les problèmes stratigraphiques, à l'Institut Géologique de Hongrie M. SIDÓ a effectué les études micropaléontologiques sur le calcaire paléozoïque, renfermant la bauxite, complétées d'études macropaléontologiques (A. VÖRÖS, Section Paléontologique du Musée National Hongrois).

La Chaire de Traitement de Minéral de l'Université Technique d'Industrie lourde à Miskolc s'occupait fort détaillément et efficacement du problème de la possibilité d'enrichissement industrielle du complexe de bauxite au niveau semi-industrielle (SZ. PETHŐ et K. SZIKLAVÁRY).

Ici aussi, je dois exprimer mes remerciements à mes collaborateurs qui, par leurs travaux dévoués, ont résolu avec succès le devoir chargeant l'expédition, aussi au Vietnam qu'en Hongrie.

Histoire courte de la recherche

J'ai trouvé les premières informations sur la bauxite du territoire de Lang Son, dans les ouvrages des géologues français R. BOURRET (1922), puis E. PATT (1927). On a préparé déjà dans les années de 1930 les premières estimations sur les réserves.

Au Vietnam du Nord, libéré en 1954, les géologues soviétiques ont comblé le manque en spécialistes dans les recherches géologiques commencées en grand élan. En 1959, on a commencé l'exploration systématique aux territoires bauxitifères de Lang Son et Cao Bang sur les résultats desquelles les rapports géologiques finaux, préparés en 1961 et 1969, ont rendu compte. On peut considérer de l'importance fondamentale, à la connaissance de la constitution géologique du Vietnam du Nord, la carte géologique en échelle de 1 : 500.000 et le texte explicatif y joint préparés par A. DOBCHIKOFF et al. (1965).

1. Aperçu général géologique et structural

Du point de vue de la tectonique générale, le Vietnam se trouve à l'extrémité est du système alpo-himalayen. Le système orogénique entier est situé au bord nord de la Téthys.

Parmi les lignes structurales les plus importantes de l'Asie du Sud-Est ce sont celles à direction du NW au SE qui dominent au Vietnam, l'arc légèrement courbé des Montagnes Fan Si Pan et Vietnam allongent à cette direction au Vietnam et la chaîne de l'Arakan Yoma en Birmanie, etc. Vers le NW, on peut suivre cette direction jusqu'au Tibet. Les grandes rivières suivent aussi cette direction structurale au Vietnam du Nord: Song Da (Rivière Noire), Song Hong (Rivière Rouge) et Song Chay.

On peut diviser la partie du NE de la République Démocratique du Vietnam — présentant l'entourage plus large de notre territoire de recherche — en deux unités structurales:

- a. Le territoire situé à W de la vallée de la Song Chay est du caractère d'éugéosynclinal, où on trouve des sédiments fortement plissés, intrusions et roches volcaniques.

- b. La zone de miogéosynclinal située à l'E de la ligne structurale précédente et formant le passage à la plaque du micro-continent de la Chine méridionale.

Les territoires bauxitifères du Vietnam du Nord se trouvent dans cette dernière zone. Ce territoire se divise encore en deux parties: celle de Song Hiem du N et la zone d'An Chau du Sud.

Dans la zone de Song Hiem, la taille profonde de Cao Bang à Lang Son est la plus importante dont la direction est du N-NW au S-SE. Le long de celle-ci jallonnent les gisements bauxitifères de Tap Na, Ha Quang et Quang Hoa dans la province de Cao Bang et de Ma Meo, Dong Dang, Tam Lung et Khon Pich dans la province de Lang Son (Fig. 1). Sur ce territoire-ci, dès le Cambrien jusqu'au Trias supérieur se trouvait essentiellement une dépression. Le faciès du Paléozoïque est carbonaté et caractérisable par la formation de la bauxite. Les roches du Mésozoïque sont terrigéniques, la plupart à faciès de détritits fins déposés en mer. Du point de vue structural il présente une synéclyse très allongée — en prédominance à direction du NW au SE — et composée des anticlinaux et synclinaux. En dépendant de l'échelle de l'érosion, dans les anticlinaux les couches paléozoïque inférieure et supérieure d'âges différents affleurent. Les assises triasiques remplissent les synclinaux. La phase orogénique principale était entre les étages carnien et norien avec plissements et émergence [13, 33]. Au cours de celle-ci, le complexe composé des roches argileuses ou à détritits fins est plus fortement plissé, le complexe de calcaire paléozoïque plus rigide est chevauché et faillé au territoire de Lang Son et plissé-faillé au territoire de Cao Bang. Les produits du volcanisme préorogénique triasique et postorogénique jurassique et crétacé se trouvent le long des lignes structurales principales (Figs 2 et 3).

La zone structurale d'An Chau diffère de celle de Song Hiem en ce que les anticlinaux se trouvent, ici, dans un arc structural courbé du W-NW à l'E-NE. Ici, la ligne structurale de Thay Nguyen, Day Ty et Dong Dang est la plus importante dont le long (resp. le long des lignes structurales parallèles à celle-ci) se trouvent les gîtes de bauxite de Bac Son, Bang Mac, Ba Xa, Dong Dang (Figs 1 et 2).

Le rôle de ces principales lignes structurales — dans la genèse et l'accumulation de la bauxite — est analogue à celui de la ligne structurale à direction de l'E-NE au W-SW, en Transdanubie en Hongrie.

2. Géologie des territoires bauxitifères

2.1. Formations paléozoïques

Les territoires bauxitifères vietnamiens se lient aux formations de calcaire paléozoïques, plus ou moins divisibles, appartenant au Dévonien, Carbonifère et au Permien (Fig. 2).

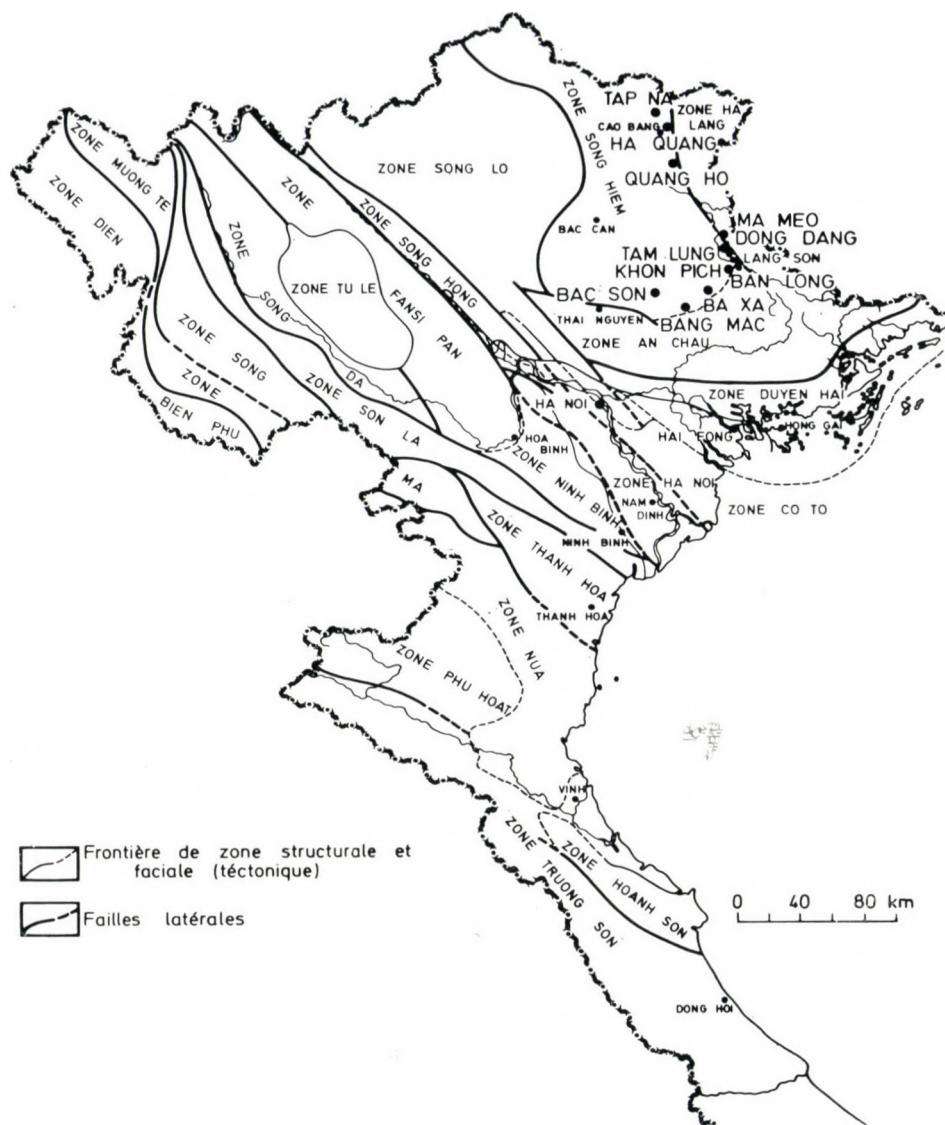


Fig. 1. Les unités mégatectoniques de Vietnam du Nord (DOBCHIKOFF et al. 1965) et les occurrences de bauxite

Les affleurements du complexe de calcaire — présentant les reliefs les plus caractéristiques de l'Asie du Sud-Est — se lient aux principales lignes structurales, décrites dans le chapitre précédent. Ceux-ci forment un désert rocheux composé des tours karstiques bizarres, dentelées, fantastiquement taillées, couvertes de végétation riche, élevées solitairement ou en masses

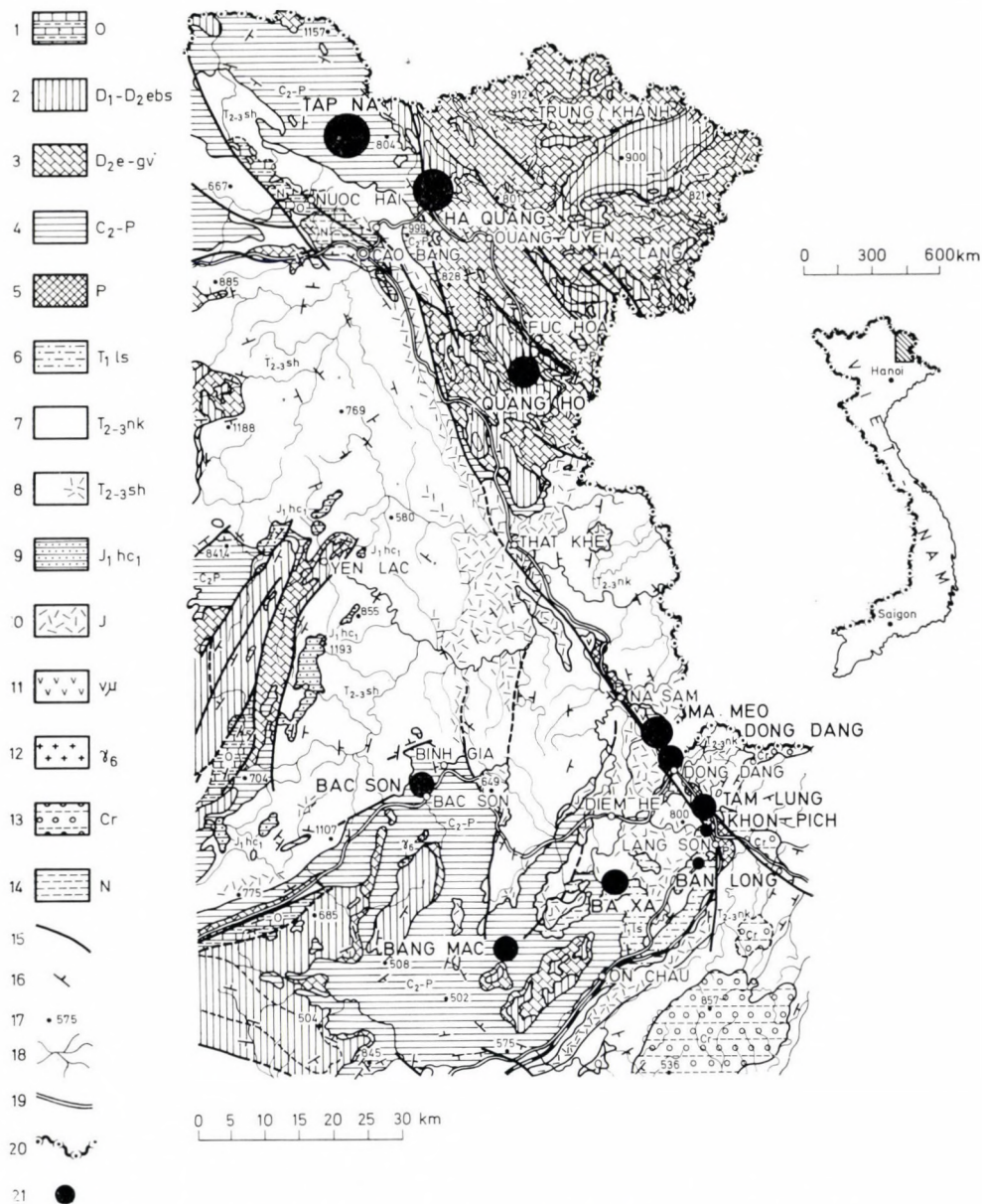


Fig. 2. Esquisse géologique des gisements de bauxite dans le provinces de Cao Bang et Lang Son (Vietnam du Nord) (DOBCHIKOFF et al. 1965). 1. Ordovicien: aléurolite, grès, gravier, calcaire; 2. Dévonien inférieur et moyen: calcaire; 3. Dévonien moyen: calcaire; 4. Permocarboneux: calcaire; 5. Permien: calcaire; 6. Trias inférieur: grès, aléurolite («Couches de Lang Son»); 7. Trias moyen-supérieur: grès, aléurolite; 8. Trias moyen-supérieur: grès tufacé; 9. Jurassique: aléurolite et conglomérat tufacé; 10. Jurassique: rhyolite, rhyodacite; 11. ? diabase; 12. Crétacé: «granit de Song Chay» 13. Crétacé: conglomérat, gravier; 14. Néogène: argile, sédiments continentaux en général; 15. lignes structurales principales; 16. pendage, direction; 17. point d'altitude; 18. rivière; 19. route 20. frontière du pays; 21. gisements de bauxite

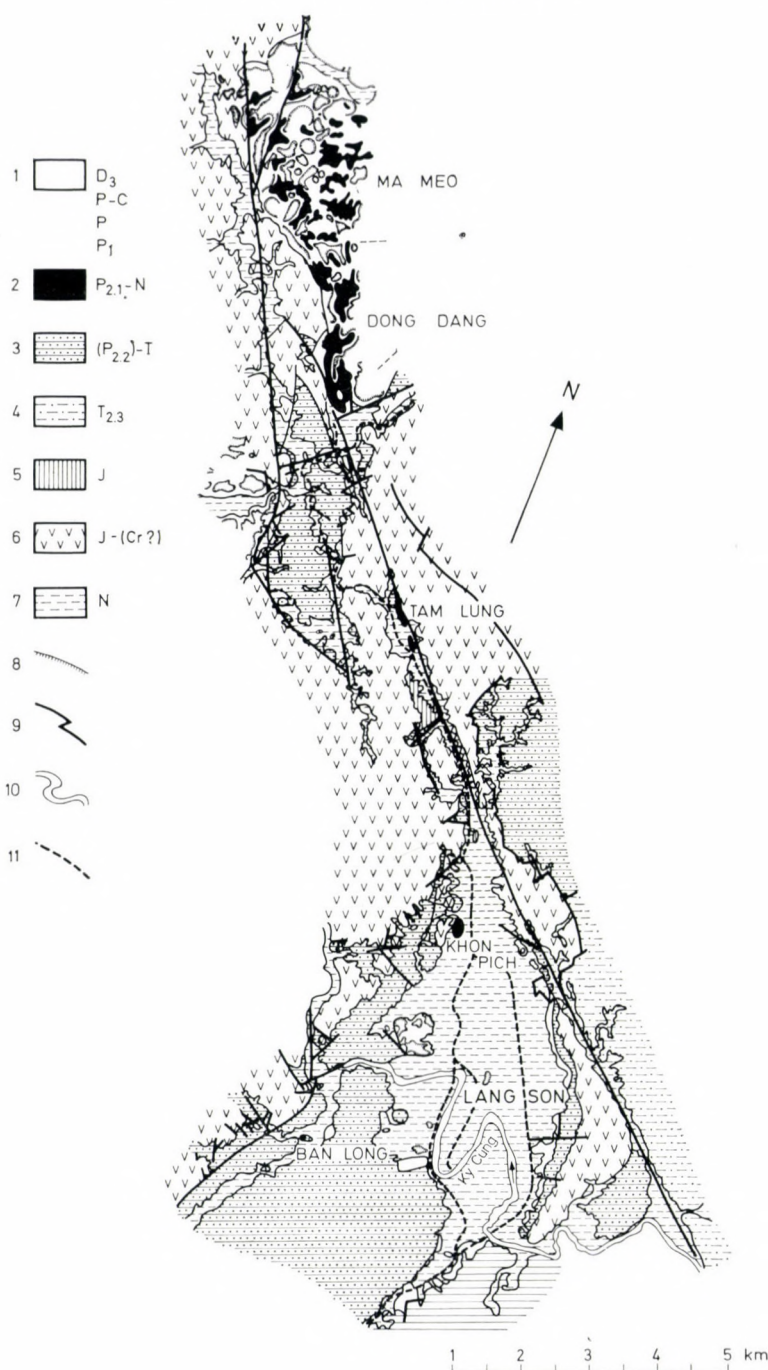


Fig. 3. Carte géologique de la région bauxitifère des environs de Lang Son (L. HAVAS, 1975). 1. Dévonien supérieur (D_3): calcaire; Permo-carbonifère (P-C): calcaire; Permien indifférencié (P): calcaire; Permien inférieur (P_1): calcaire; 2. Permien supérieur-Néogène: $P_{2,1}$ -N: bauxite; 3. Permien supérieur-Trias ($P_{2,2}$ -T) aléurolite: «couches de Lang Son» 4. Trias moyen-supérieur: ($T_{2,3}$): grès tufitique, conglomérat; 5. Jurassique (J): laves et tufs de rhyolite et de rhyodacite; 6. Jurassique-Crétacé (J-Cr): laves et tufs rhyolitiques et rhyodaitiques; 7. Néogène (N): argile, conglomérat, sédiments continentaux récents indifférenciés; 8. crêtes rocheuses; 9. lignes structurales principales; 10. rivière; 11. chemin de fer

plus grandes (Photo 1). Elles présentent les lambeaux des plateaux karstiques (13), jadis continus, auparavant tectoniquement fort fracturés, chevauchés en écailles, puis karstifiés sous le climat tropical-subtropical pluvieux.

Le terme le plus ancien du complexe de calcaire du territoire de Lang Son représente l'étage givétien du Dévonien trouvable à la partie sud (Ban Long) du territoire de recherche (Fig. 3). A côté des *Crinoïdes*, *Brachiopodes* et *Polypiers*, M. SIDÓ a déterminé *Thamnopora boloniensis* (GOSSELET), espèce caractéristique du Dévonien eurasién. Dans le district plus lointain, on peut retrouver les termes plus anciens du calcaire dévonien aussi qui présente le mur des deux gisements bauxitifères (Ha Quang et Quang Ho) à grandes extensions, mais à qualité inférieure de la province de Cao Bang (Fig. 2).

Dans le massif de Bac Son, la littérature spéciale mentionne une discordance nette entre le Dévonien et le Carbonifère [13, 34, 45].

Au territoire de Lang Son, l'étude de la microfaune a vérifié l'étage moscovien du Carbonifère, aux parties sud et nord du territoire de recherche. Au S le contact avec le Dévonien moyen est présumablement tectonique (Fig. 4).

M. SIDÓ a distingué trois niveaux à l'intérieur de l'étage moscovien:

- a. à *Pseudostaffella*, *Profusulinella* et *Dvinella connota*,
- b. à *Pseudostaffella*, *Eostaffella* et *Profusulinella*,
- c. à *Wedekindella* et *Pseudostaffella*.

Au territoire de recherche de Lang Son, l'horizon à *Schwagerines* du Permien inférieur (étage sakmarien), caractéristique à la région de la Thétys, manque. Aux territoires de Ma Meo et Dong Dang, le complexe Darvasites et Pseudofusulines, indiquant la partie supérieure du Permien inférieur, développe de l'étage moscovien lithologiquement en sédimentation continue: seul le diasthème paléontologique signale la lacune de sédimentation. Dans ce complexe de couches les fragments de *Beresella* sp. et les débris d'autres fossiles carbonifères remaniés sont aussi retrouvables. Chacune des espèces de *Marginifera* sp., *Athyris* cf. *nanthensis* (GRABAU), *A.* cf. *acutirostris* (GRABAU), *Marthinia semiglobosa* (TCHER.), *Phricodothyris nucleola* (GRABAU) est aussi retrouvable dans le calcaire "mapping" répandu en Chine méridionale [16]. Plus loin du territoire de recherche, au massif de Bac Son, DOBCHIKOFF et NGUYEN DINH KAT indiquent de la discordance tant dans le Carbonifère moyen qu'à la limite entre le Carbonifère et le Permien [13, 33].

Aux territoires de Ma Meo et Dong Dang, nous avons interprété le complexe lithologiquement homogène du Carbonifère moyen et du Permien inférieur (étages artinskien et d'ufa) comme une formation permo-carbonifère. Au S de ces territoires, ce complexe passe en calcaire ne contenant que d'espèces permien inférieur (Dong Dang S). En allant encore plus loin vers le S, les

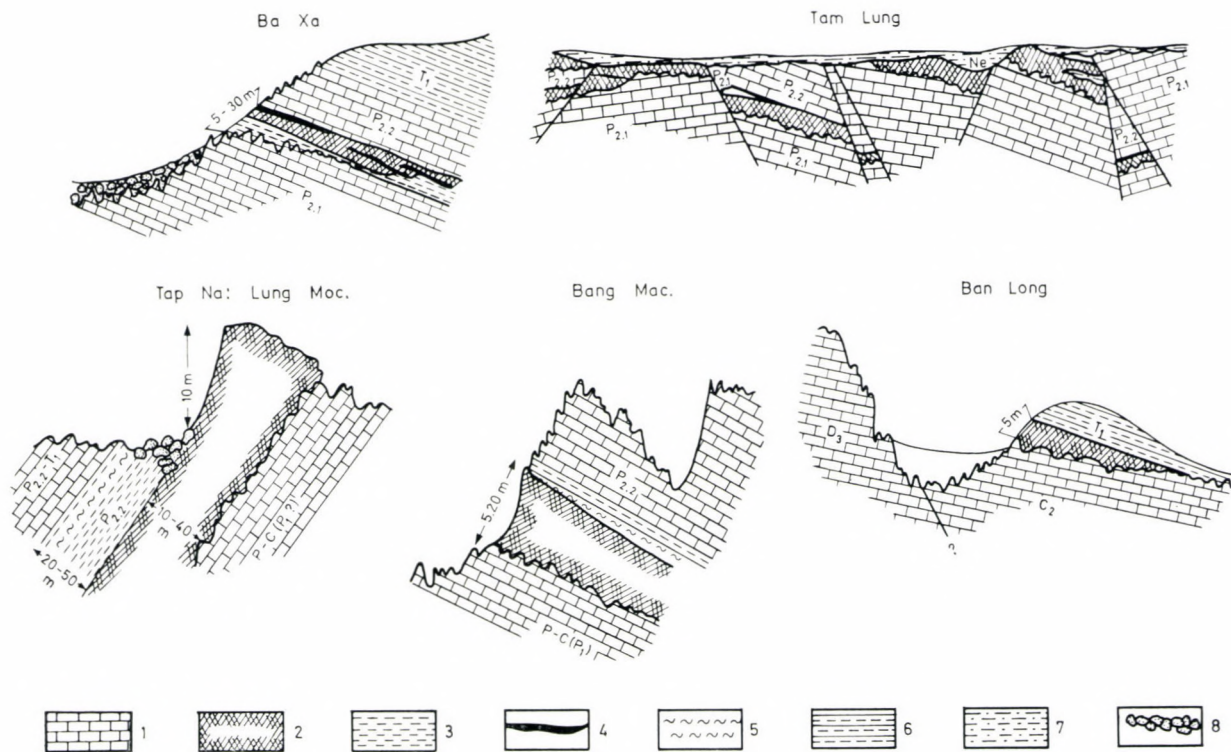


Fig. 4. Situation géologique des gisements couverts d'âge Permien supérieur. 1. Dévonien supérieur (D_3): calcaire; Permo-carbonifère (Permien inférieur?) ($P-C$, P_1): calcaire; Carbonifère (C): calcaire; Permien supérieur (étage kazanien) $P_{2.1}$: calcaire; Permien supérieur (étage tatar), $P_{2.2}$: calcaire; 2. Permien supérieur: bauxite; 3. Permien supérieur: argilite; 4. Permien supérieur: houille; 5. Permien supérieur: marne houilleuse, argile; 6. Trias inférieur (T_1): aléurolite; 7. Néogène (Ne): argile, conglomérat; 8. Néogène-Holocène: détritiques de bauxite

forages ont exploré des complexes paléozoïques de plus en plus récents (Tam Lung; Fig. 3).

M. SIDÓ a présenté l'image stratigraphiquement la plus complète sur le complexe permien supérieur. Elle a distingué les horizons suivants:

- a. à *Néoschwagerines*, *Cancellines*, *Missellines* et *Néofusulines*,
- b. à *Paléofusulines*, *Reichelines*, *Colaniella* et *Pachyphloia*,
- c. à *Nankinelles*, *Glomospires* et *Algues* calcaires,
- d. horizon à oolithes et brèche authigénique (Permo-Trias?).

A l'intérieur de la région de la Téthys, c'est l'horizon à *Paléofusulines* qui est le plus répandu. Les Foraminifères présents en grand nombre d'espèces et individus, *Colaniella* sp., *Leiodiscus* sp., *Robuloides* sp., convenables à la corrélation de longue distance, prouvent la communication directe entre les provinces est et ouest [42]. A l'intérieur de cet horizon, on peut distinguer deux lithofaciès:

- a. à dominance des *Algues* calcaires — littoral,
- b. à dominance des Foraminifères — mer ouverte, peu profonde.

On retrouve les espèces conductrices ci-mentionnées dans tous les deux faciès.

Au territoire de Lang Son, les formations paléozoïques de mer ouverte, plus anciennes que le complexe à *Glomospires*, se composent des calcaires lithologiquement uniformes dès le Dévonien jusqu'au Permien supérieur, en prédominance à texture grossièrement recristallisée, en gros bancs ou grossièrement stratifiée. Sur ce territoire, le pendage des couches ne dépasse pas 25°. Au territoire de Cao Bang, même le pendage à 60° n'est pas rare. Dans la suite, nous allons aborder de nouveau l'importance géologique de bauxite de la différence entre ces deux-ci.

La moyenne de la composition chimique des dix échantillons, provenant de ces formations, est la suivante:

Tableau I

SiO ₂	TiO ₂	Al ₂ O ₃	Fe ₂ O ₃	FeO	CaO	MgO	K ₂ O	Na ₂ O	+H ₂ O	CO ₂
4,2	0,04	1,80	1,30	0,31	46,5	0,52	0,59	0,43	3,65	35,59

La teneur en silice relativement plus élevée du calcaire est frappante, 60 à 80% de celle-ci est dû à la teneur de sables quartzifères fins.

Le niveau à *Glomospires* lithologiquement est plus variable dans son faciès littoral. L'argile houilleuse, bancs de houille argileuse et de houille, ainsi que des laies de marne houilleuse et le calcaire noir bitumineux y sont fréquents. Par endroits, on peut observer — dans le calcaire intercalé dans le complexe houillifère — des précipitations de silice en forme et grandeur de lentille et haricot, orientées parallèlement à la stratification.

2.2. Position stratigraphique de la bauxite primaire

On peut déterminer la position stratigraphique de la bauxite primaire le plus précisément aux gisements de Tam Lung et Ba Xa étudiés le plus en détail, où la bauxite se trouve dans le calcaire permien supérieur.

Au territoire de Ba Xa, la bauxite est située entre les horizons à *Paléofusulines* et à *Glomospires*, *grosso modo* à la limite entre les étages kazanien et tartar.

Au territoire de Tam Lung, la formation de bauxite était présente même dans l'horizon à *Glomospires*.

La littérature soviétique et vietnamienne y concernant indique le mur de la bauxite de Bang Mac comme permo-carbonifère [13, 33]. Il n'est pas exclu que le niveau inférieur du Permien supérieur, l'horizon à *Paléofusulines* soit présent aussi dans ce complexe. Le toit est, ici aussi, le calcaire noir bitumineux du Permien supérieur correspondant à l'horizon à *Glomospires*.

A certains gîtes du gisement de Tap Na, nous retrouvons la bauxite en position stratigraphique identique à la précédente, mais à cette différence qu'ici le mur est probablement plus ancien, un Permo-Carbonifère comprenant le Carbonifère moyen et le Permien inférieur.

C'est le territoire de l'indication de Ban Long, où nous trouvons la plus grande lacune de sédimentation, où le mur est le Carbonifère moyen et le toit est le Trias inférieur. Ici, à cause de l'émersion localement arrivée le calcaire du Permien supérieur du toit a été érodé, et la bauxite présente un lambeau insignifiant épargné de l'érosion. Elle a pu subir une réaccumulation plus ou moins grande dans le Trias inférieur, et en ce sens on peut la considérer, en partie ou entièrement, comme triasique inférieur.

Plus loin de ces territoires, on trouve de tels gisements bauxitifères à la surface nue (par ex. Ma Meo) que nous considérons en partie aussi à position stratigraphique permien supérieur. Ce sont des lentilles, resp. parties de lentilles situées topographiquement en altitude la plus élevée par rapport à l'environnement. Ce complexe de bauxite a été altéré *in situ*, lithologiquement indistinguible de la bauxite réaccumulée, mais par causes géomorphologiques l'apport récent par ici est impossible. Le toit pouvait être, ici aussi, le niveau à *Glomospires* du Permien supérieur qui a été «dissous» comme le sucre sans laissant des détritits (sauf les lentilles de silice décrites dans les précédents). A ce problème nous sommes d'accord avec GY. BÁRDOSY — visitant le terrain — et B. OTDOKIY, géologue soviétique.

2.3. Formations mésozoïques

On connaît la géologie des formations mésozoïques des environs de Lang Son en détail d'après le travail de cartographie géologique en échelle de 1 : 25.000, fait par L. HAVAS.

2.3.1. Formations sédimentaires

Aux environs de Lang Son, le complexe finement détritique à l'épaisseur totale dépassant 1000 m, et couvrant un grand territoire, représente le Trias entier. La sédimentation oscillatoire, les bancs à cause de la granulométrie variable caractérisent les formations tectoniquement fort affectées: les éléments morphotectoniques plissés et faillés-plissés y sont fréquents. Le produit d'altération latéritique actuel de sa surface est l'argile rouge à kaolinite.

Ces couches sont très pauvres en fossiles et stratigraphiquement ne sont corrélables plus précisément. Lithologiquement on peut distinguer deux niveaux:

a. Les soi-disantes «*couches de Lang Son*» composées du grès et de l'aléurolite qui développent du calcaire permien supérieur à l'intermédiaire de la marne argileuse divisée par bancs de calcaire. Ce complexe représente, *grosso modo*, le Trias inférieur. Aux environs de Ba Xa, il surmonte le calcaire ou la bauxite du Paléozoïque ancien à lacune d'érosion plus réduite mais plus grande à Ban Long. Dans le banc de grès fin de ce complexe *Gervileia costata* (SCHLOTH.) et *Myophoria costata* (ZEN.) sont provenues.

b. Le grès tuffitique et le conglomérat bariolés comprennent env. le Trias moyen et supérieur (et éventuellement le Jurassique inférieur). La matière du sédiment hétérogénique comprend des détritiques volcaniques non classés et des grains de quartz guère ou pas de tout usés, à côté de ceux-ci dans la fraction d'argile on peut démontrer l'illite et un peu de montmorillonite [25]. La formation est par endroits fortement silicifiée par affection post-volcanique plus récente. Ces parties siliceuses tombant de la roche plus meuble et facilement altérée accumulent en détritiques piémontais, en grandeur de cm à m. Au territoire de Ba Xa, par endroits on peut la retrouver ensemble avec la bauxite en voie d'érosion et résistante à manière pareille.

Au territoire de Tap Na, c'est le calcaire à bancs minces et gros, développé du Permien supérieur en continuité qui compose le Trias inférieur. Il est caractérisé par sa structure plus fortement plissée que le calcaire paléozoïque, présentant des voutes à diamètre pas plus de trois mètres.

2.3.2. Volcanites

Au Vietnam du Nord, dans le Trias supérieur, mais surtout à la limite entre le Trias et le Jurassique une activité de volcanisme acide intense, syn- et postorogénique a pris son commencement qui durait à interruptions plus ou moins longues encore dans le Crétacé inférieur, aussi [13]. A Lang Son et ses environs plus lointains les volcanites présentent les formes topographiquement les plus élevées.

Selon L. HAVAS à la phase initiale du volcanisme des laves globuleuses et cordées ainsi que des brèches de lave ont été produites. Sur la base des déterminations minéralogiques en lames minces et les analyses chimiques ce sont des andésito-dacites à amphibole verte passant en rhyolite avec l'intermédiaire rhyo-dacitique. Un type de rhyolite à texture granophyrique, et à composition chimique identique à la précédente, est généralement répandu à la surface, la relation entre celui-ci et les formations précédentes n'est pas encore éclaircie.

Au cours de l'altération, les feldspaths de la rhyo-dacite se transformaient en illite ou en kaolinite. On a trouvé tel gravier latéritiquement altéré dont 21% était de la gibbsite. Dans la lentille Ban Thau I. du gisement de Ma Meo, des graviers de bauxite à qualité excellente et montrant la texture de cette roche se présentaient aussi, mélangés avec de la bauxite karstique réaccumulée dans le Néogène.

Le Khao Ma (+800 m), pic le plus élevé de la région pouvait conserver la phase terminale de l'activité volcanique. Au flanc abrupt et au pic de celui-ci, la rhyolite non altérée à colonnes peut être du remplissage de cheminée.

Dans le complexe volcanique, des conglomérats tuffeux s'intercalent à plusieurs niveaux. La matière des graviers est en prédominance du grès accompagné du quartz et d'une volcanite aphanitique, tous les deux rares. La formation, n'indiquant pas un niveau constant, pourrait signaler le commencement des différentes périodes d'explosion.

Le complexe volcanique contient des imprégnations sulfureuses. Dans le gisement de Ba Xa, à côté des veinules de silice capillaires on peut trouver dans la bauxite des remplissages sulfureux aussi, comme produit téléthermal.

2.4. Formations néogènes, position stratigraphique des accumulations de bauxite secondaires

Avant la fin du volcanisme, cet espace de l'Asie du Sud-Est est devenu terre ferme. Dès le Crétacé jusqu'à nos jours — ne considérant pas le remplissage des bassins des grandes rivières et quelques sédimentations lacustres insignifiantes — ce sont l'érosion intense et la sédimentation terrestre qui régnaient.

Nous avons rangé dans le Néogène les formations plus récentes que les volcanites qui étaient produites par l'altération argileuse des roches de la surface et par leurs détritiques effrités et déposées dans des bassins sédimentaires périodiques. Aux environs de Lang Son, après une grande lacune d'érosion elles surmontent la surface karstique du calcaire paléozoïque et les lambeaux d'aléurolite triasiques épargnés par l'érosion et à Tam Lung la bauxite *in situ*.

L'argile encaissant les détritiques est le produit d'altération fin des volcanites, aléurolite et du grès ou calcaire, mais on peut y retrouver aussi la matière de la bauxite altérée en argile.

Du point de vue de la stratigraphie, il faut ranger ici les accumulations secondaires de la bauxite aussi qui se trouvent, comme complexe de détritiques monomictes, sur la surface des roches carbonatées paléozoïques d'âge différent. Elles se composent des détritiques de bauxite durs à granulométrie variée et variablement usés, lithologiquement résistants à l'altération et de l'argile encaissante meuble provenant de la bauxite et d'autres roches. En dépendant des conditions géomorphologiques, elles présentent des lentilles et gîtes, plus ou moins riches en détritiques de bauxite, accumulés aux flancs des collines, au fond des vallées d'érosion et en ponors karstiques profonds, non rarement sous une épaisseur de 50 m (Tap Na, Bang Mac, Ba Xa, Ma Meo, etc.).

Les ouvrages précédents ont rangé les gîtes de bauxite secondairement accumulés uniformément dans le Néogène [13]. Selon notre avis, ce processus pouvait commencer déjà avant le Néogène et dure à nos jours, aussi. D'après leurs caractères sédimentologiques et leurs positions géomorphologiques, on peut distinguer les accumulations plus anciennes et plus récentes sans pouvoir préciser leurs positions stratigraphiques fines. Au territoire de Lang Son, parmi les gîtes secondaires c'est la bauxite du gisement de Khon Pich qui est la plus ancienne. Sa source *in situ* n'est pas connue, nous probabilisons seulement que la bauxite primaire à position stratigraphique permienne se trouve recouverte des couches triasiques et permienes au W de la ligne de fracture longeant le côté ouest de la lentille. Après l'accumulation, le complexe de bauxite secondaire est à la plupart érodé et altéré. La lentille présente un lambeau à l'entourage d'un ponor karstique extrêmement profond épargné de l'érosion à "l'ombre" d'une colline d'aléurolite (Fig. 5).

Le complexe de bauxite secondaire de Tam Lung situé à la surface peut être un peu plus récent que le précédent, il se trouve encaissé en argile au-dessus du calcaire permien supérieur ou de la bauxite primaire permien supérieur.

Le complexe trouvable au fond des vallées de Dong Dang présente une accumulation encore plus récente, et les détritiques de bauxite argileux, étant en mouvement actuellement aussi aux flancs des collines abrupts de Ma Meo ainsi que les accumulations situées à proximité des gîtes primaires (Ba Xa, Bang Mac, Tap Na, etc.) sont les plus récents.

3. *Gîtologie générale, classement génétique des gîtes de bauxite*

3.1. *Situation structurale*

La genèse et l'accumulation de la bauxite karstique sont liées, à chaque âge et partout, aux mouvements épirogéniques et tectoniques, au changement des conditions paléogéographiques. Ce qui manifeste dans la répartition des

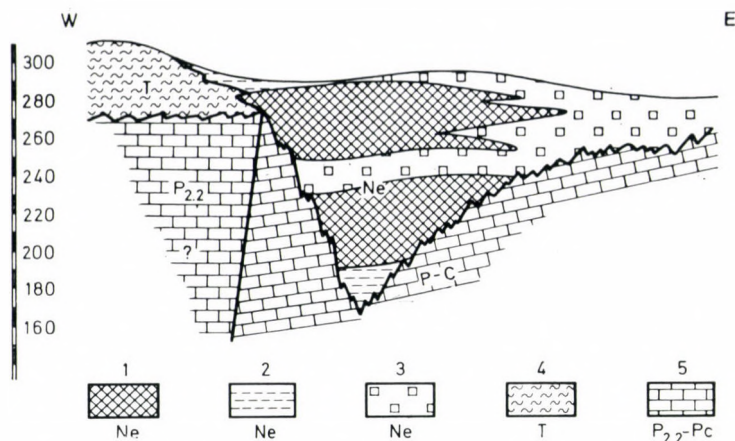


Fig. 5. Coupe géologique de gîte de bauxite à Khon Pich. 1. Néogène (Ne): complexe de détritits de bauxite; 2. Néogène (Ne): argile; 3. Néogène (Ne): argile à détritits d'aléurolite; 4. Trias (T): aléurolite; 5. Permo-carbonifère (P-C): calcaire. P_{2,2} Permien supérieur: calcaire

différents gisements liés aux directions structurales déterminées à l'intérieur d'une province bauxitifère, resp. des unités mineures: gîtes et groupes de lentilles [14].

Parmi les gisements bauxitifères de Lang Son, Dong Dang, Tam Lung et Khon Pich présentent des configurations allongées en direction de la principale ligne structurale du NW au SE. La forme des lentilles de Ma Meo, le relief du mur à l'intérieur des lentilles et la position des alignements rocheux abrupts entre les lentilles sont tous en harmonie avec la tectonique locale du NE au SW, perpendiculaire à la ligne principale (Fig. 3).

Les lignes structurales à direction du NW au SE caractérisent le territoire de Tap Na. Les affleurements de bauxite jalonnent le long de cette ligne, et les accumulations secondaires présentent des configurations allongées en même direction.

3.2. Classement des gîtes de bauxite

Nous allons classer les gîtes de bauxite vietnamiens d'après les gisements les mieux explorés et stratigraphiquement les mieux connus; nous avons distingué les types principaux suivants. Naturellement, à l'intérieur d'un gisement on peut retrouver plusieurs types de gîtes, les différents types passant sans limite nette, les uns dans les autres (Fig. 6).

3.2.1. Type de Ba Xa

Le complexe de bauxite *in situ* affleure en plusieurs kilomètres — à interruptions d'érosion plus ou moins grandes — le long de la direction du N-NE au S-SW. Son pendage est de 15 à 25°. On peut suivre les affleurements, en largeur de 10 à 60 m correspondant à l'épaisseur réelle de 5 à 45 m.

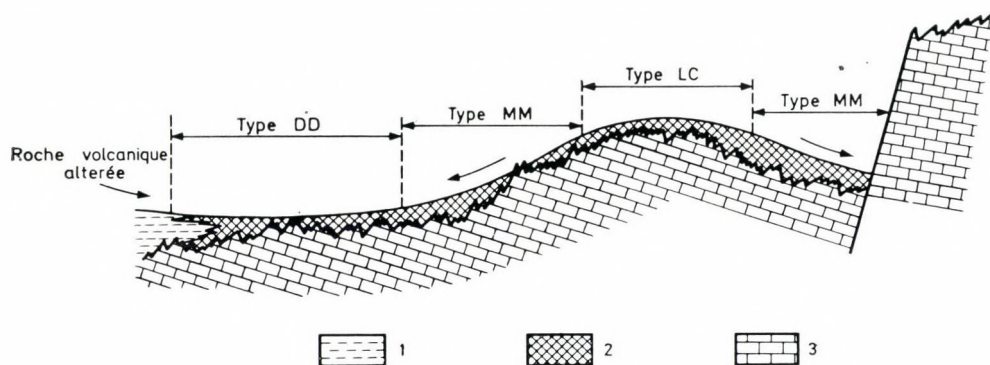


Fig. 6. Coupe géologique idéale des types de gîte détritique, de Leo-Cao, Ma Meo et Dong Dang. 1. argile; 2. complexe de détritits de bauxite; 3. calcaire

Lithologiquement le complexe de bauxite *in situ* est composé de la bauxite dure, rigide en bancs gros et de l'argilite plus finement stratifiée. L'argilite présente, à la plupart, l'un ou deux tiers inférieurs du complexe. Le passage est graduel entre la bauxite et l'argilite. A la zone de passage dans l'argilite en strie de 2 à 3 mm ou encore plus large on trouve des laies et cordons de bauxite à texture pisolithique, distinguables aussi à l'œil nu, qui deviennent de plus en plus larges en allant vers le haut.

Le complexe de bauxite est divisé par bancs houilleux, surtout dans son niveau supérieur. Les sondages implantés en direction du pendage hors le bord des affleurements, au niveau de la bauxite, ont traversé un complexe d'argilite divisé par bancs calcaires, où la bauxite manquait complètement. D'après les carottes de sondages notre impression était qu'ici par passage marneux, l'argilite est développée du mur en continuité, c'est-à-dire la sédimentation marine était continue.

Le complexe de bauxite est tectoniquement fort affecté, particulièrement c'est la bauxite relativement plus rigide qui est fortement fracturée et fissurée. A la surface ou à proximité de la surface, le long des fissures et surfaces de couches le complexe altère en argile. C'est l'argilite qui est la première victime de l'altération en tant que la roche dure, souvent stratifiée en lamelles, s'altère en argile meuble et plastique. En allant du haut vers le bas, le complexe de bauxite primaire *in situ* devient de plus en plus argileux. La bauxite en gros bancs, divisée par remplissages de fissures et de bancs d'argile, passe en continuité dans la bauxite pseudo-détritique *in situ*, altérée en place et encaissée dans l'argile (Photos 2 et 3).

Le mur en position plus élevée étant érodé, le complexe de bauxite pseudo-détritique arrivé à la surface peut être réaccumulé en dépendant des conditions géomorphologiques, car le long des parties cimentées par argile le complexe désagrège vite et passe de la surface du terrain plus élevée vers celle plus basse (voir: profil de Ba Xa dans la Fig. 4).

On peut ranger dans le type gîtologique de Ba Xa aussi les bauxites *in situ* du gisement Tap Na de Cao Bang à cette différence considérable que c'est la bauxite dure qui constitue le complexe de bauxite tout entier, alors ici l'effritement est plus modéré malgré l'affection tectonique plus forte. Ici, on ne connaît pas d'autres faciès du complexe de bauxite, étant donné que la recherche par forages ne les a pas encore explorées.

3.2.2. Type de Leo Cao

Comme on l'a mentionné dans le chapitre 2.2. là, où le complexe de bauxite se trouve à la surface nue, il y a de telles parties de gîtes, lentilles ou de parties de lentilles qui sont apparemment pseudo-détritiques, pareillement à la bauxite réaccumulée. D'après les analogies observées à Ba Xa nous considérons les lentilles, situées géomorphologiquement aux sommets des collines, comme complexe pseudo-détritique altéré en place (à position stratigraphique permien supérieur). Nous en trouvons les plus beaux exemples dans les lentilles Leo Cao du gisement de Ma Meo. Des valeurs extrêmes caractérisent l'épaisseur du complexe de bauxite. A l'intérieur d'une distance de quelques mètres des différences de 0,5 à 50 m peuvent arriver par conséquent du relief extrêmement irrégulier du calcaire du mur.

La proportion de poids entre la bauxite et l'argile est la plus élevée dans le type de gîte de Leo Cao. La quantité de la bauxite est en moyenne de 1,04 t/m³, mais il y a aussi de telles zones, où les valeurs de 2,4 à 2,6 t/m³ ne sont pas rares. Parce que le poids volumétrique de la bauxite compacte est 3, alors il est évident que l'on trouve ici telles «couches» qui sont composées par «détritus de bauxite» presque en leur entité complète. Si ces zones riches en bauxite se trouvent à la surface, l'enrichissement de la bauxite peut être aussi secondaire: les eaux de surface ont déjà lavé l'argile du ciment. Si les zones plus riches en bauxite se présentent en profondeur plus grande au-dessous de la surface, nous pouvons penser qu'il s'agit des restes de la couches, primordialement plus riche en bauxite, du complexe originalement divisé par l'argilite.

La dimension maximale des «détritus» reste en générale au-dessous de 1 m, en prédominance de 20 à 40 cm. La proportion entre les fractions plus grossières, resp. plus fines de 2 cm: 4 : 1. Les formes anguleuses ou seulement légèrement usées y sont fréquentes. Même au cours de l'altération en place — sans effet mécanique — peuvent se produire des formes bien arrondies, car les facteurs chimiques, aussi comme ceux mécaniques, attaquent premièrement les tranchants et les angles les plus exposés. La résistance de la matière homogène contre l'altération produit la surface sphérique.

Malgré son extension réduite, le type de Leo Cao présente la partie la plus précieuse du gisement de Ma Meo, car à côté du rendement relativement

riche la qualité des «détritus» est également la meilleure, ici. Voir en détails: chapitre 6.

Certaines parties d'autres gisements appartiennent aussi à cette unité géologique, comme par ex.: Ba Xa ou la partie du gîte de Tam Lung non recouverte du calcaire.

3.2.3. *Type de Ma Meo*

Au territoire des gîtes à type de Leo Cao l'argile à détritits de bauxite effrités est encore actuellement aussi en mouvement aux flancs des collines abrupts. On le voit bien à la photo 4, où on peut observer comment découle le complexe à détritits de bauxite entre les tours rocheuses à parois abruptes composées du calcaire du mur. En dépendant des conditions géomorphologiques la matière de roche désagrège en plus. A proximité des gîtes à type de Leo Cao, les blocs de 50 à 60 cm sont encore fréquents. En allant vers le bas sur les pentes, le diamètre maximal et prédominant diminue.

A ce type le rendement moyen n'est plus que 0,7 t/m³ seulement, et la proportion entre les fractions plus grossière, resp. plus fine de 2 cm est de 3 : 1. En allant vers les fonds des vallées, il arrive de plus en plus fréquemment qu'à certains niveaux, — mais plus particulièrement à la proximité du mur, — la fraction fine atteint, même dépasse la quantité de la fraction plus grossière.

L'épaisseur du complexe de type de Ma Meo varie aussi entre valeurs extrêmes. L'épaisseur du complexe est la fonction de l'énergie du relief. On trouve la plus grande épaisseur là, où la matière mouvant vers le bas s'entasse en heurtant une paroi rocheuse (Photo 5), et là, où la barricade est interrompue, des vallées étroites courent vers le bas, et ici l'argile à détritits de bauxite «découle».

Ce type de gîte représente des réserves à quantité considérable aussi au territoire de Tap Na. Le processus de l'accumulation de la matière détritique dure ici moins longtemps, les détritits sont moins usés et moins classés.

3.2.4. *Type de Dong Dang*

D'après l'exemple du gisement de Dong Dang de Lang Son, nous avons inséré ici ces accumulations qui se trouvent géomorphologiquement au relief le plus bas. Au bord des vallées à fond large, le complexe arrivé dès les pentes de collines abruptes s'étale en fanglomérat (Photo 5). Si la construction de la vallée est géologiquement asymétrique, alors l'argile à détritits de bauxite, arrivée de l'entourage calcaire, est mélangée avec les détritits d'autres roches, aussi. En allant vers l'axe de la vallée, les détritits de bauxite diminuent de plus en plus dans le complexe d'argile, et la granulométrie abaisse brusquement. Ici, l'argile est déjà, en majorité écrasante, le produit d'altération des volcanites ou de l'aléurolite.

La bauxite peut être aussi mélangée avec les détritiques de ces roches. Dans la matière de la lentille de Khon Pich on peut reconnaître l'aléurolite triasique et dans les lentilles nommées Ban Thau du gisement de Ma Meo la matière des volcanites, aussi.

Nous avons rangé à ce type une partie des accumulations restées en entourage des carbonates, aussi: les sédiments déposés dans des vallées tectoniquement préformées larges de 70 à 200 m, limitée des parois calcaires abruptes, à déclinaison complètement légère et le complexe en voie d'accumulation dans les environs des ponors à rayon de 50 à 200 m, en situation topographique élevée.

Sauf Khon Pich qui est le plus ancien, le trait commun de tous les trois territoires est ce que la matière détritique accumule continuellement, il n'y a pas d'érosion ou elle est de l'échelle très limitée. L'altération mécanique a cessé, et c'est la nappe phréatique, resp. l'eau karstique qui effectue son effet d'altération ou oxydant.

Bien que l'on trouve dans les vallées des blocs énormes aussi, la granulométrie dominante est de 2 à 4 cm. La proportion entre les fractions plus grande, resp. plus petite de 2 cm est diminuée à 2 : 1. Le rendement moyenne reste aussi au-dessous de ceux décrits précédemment: 0,5 t/m³. La variation du rendement est extrêmement capricieuse tant horizontalement que verticalement. Les zones improductives — sans bauxite — y sont fréquentes, et les niveaux à détritiques de bauxite sont séparés du mur par argile improductive, par endroits épaisse de 10 m. En comparant aux types précédents, l'épaisseur du complexe de bauxite est la plus régulière dans les vallées tectoniques bordées par le calcaire du mur. Les plus fréquentes valeurs d'épaisseur se trouvent entre 10 et 15 m. Le relief du mur est moins variable au-dessous du niveau de l'eau karstique, resp. de la nappe phréatique.

L'épaisseur du complexe de bauxite, remplissant les ponors karstiques profondes, dépasse même 100 m, dans la lentille de Khon Pich explorée par sondages. Au territoire de Ma Meo, où nous n'avons exécuté que des recherches à puits creusés, nous les avons arrêtées à la profondeur de 50 m, dans le complexe à détritiques de bauxite.

3.2.5. Type de Tam Lung

On peut ranger à ce type la partie de gîte du niveau profond, recouverte du calcaire permien supérieur, du gisement de Tam Lung qui diffère pétrographiquement et génétiquement des gîtes recouverts des autres gisements.

Le complexe de bauxite est varié, sa caractéristique: les détritiques de bauxite durs en dominance de 2 à 6 cm, sont encaissés dans la bauxite meuble ou dans l'argile bauxitique. La différence essentielle par rapport aux types de gîtes détritiques précédemment décrits est ce que la matière encaissante

de la bauxite n'est pas le produit d'altération de la bauxite ou de l'argilite, mais elle est aussi de la bauxite primaire, même si elle est qualitativement plus inférieure que les détritits (voir en détails chapitre n° 7).

Aussi, comme à Ba Xa, le complexe de bauxite est divisé, ici aussi, par bancs d'argile houilleuse et de houille argileuse. L'épaisseur du complexe est assez variable due en partie à l'inégalité du mur, en partie à ce que le complexe de bauxite passe horizontalement, même en petite distance, dans des formations houilleuses et argileuses. Comme changement de faciès hétéroptique suivant, nous retrouvons aussi les sédiments purement marins consistant en schistes argileux à sidérose.

L'accumulation du complexe de bauxite était de temps en temps interrompue: le complexe renferme des bancs de calcaire, par endroits.

Comme effet épigénique, la pyritisation, causée par le milieu du marais, est fréquente et s'étend souvent sur l'épaisseur du profil entier, mais n'avance que dans la matière encaissante plus meuble.

Le territoire est fort tailladé tectoniquement, les affaissements ont été suivis de chevauchements. Il y a aussi de redoublements de couche à cause des chevauchements (Fig. 1).

4. Propriétés pétrographiques de la bauxite et de l'argilite

D'après les propriétés pétrographiques observables au terrain, on distingue les types suivants:

- A. Bauxite, noir à gris foncé, primaire, dure, compacte et ses variétés oxydées (Ma Meo, Tam Lung, Ba Xa).
- B. Bauxite, brun rougeâtre foncé et rouge brunâtre, primaire, compacte, dure et ses variétés oxydées en rouge (Ma Meo, Tap Na).
- C. Brèche de bauxite: en bauxite, gris ou rouge, meuble sont encaissés des détritits de bauxite des types A et B (niveau profond de Tam Lung).
- D. Argilite, gris clair ou plus foncé, à dureté moyenne et voisine de la bauxite et ses variétés oxydées (Ban Long, Ba Xa).
- E. Types de latérite.

La bauxite appartenant au type A est réduite en son état primaire: noir, gris foncé, gris verdâtre ou vert (Photo 6). Extrêmement dure ($D \geq 7$) à cassure conchoïdale ou esquilleuse. Texture pour la plupart oolithique — pisolitique, détritique ou pseudobréchique, plus rarement homogène. La structure des types à qualité meilleure est recristallisée, souvent à texture arénitique. La bauxite de Cao Bang est toujours plus grossièrement recristallisée que celle de Lang Son.

Ses affleurements, trouvables *in situ* à Ba Xa, Bang Mac et Tap Na, sont stratifiés en bancs, plus richement oolithiques et pisolitiques vers le toit. Sa

micro-stratification, souvent visible à l'œil nu, signifie l'alternance des stries à texture plus grossièrement ou plus finement recristallisée ou collomorphes.

L'oxydation secondaire de la bauxite, gris ou vert, présente un phénomène généralement observable. En zone épaisse de 1 à 3 mm, le processus d'oxydation commence déjà le long des fissures et surfaces de couches de la bauxite *in situ*, recouverte de couches permianes. L'épaisseur de la zone d'oxydation de la bauxite, noir, de Ba Xa et Tam Lung, reste pour la plupart au-dessous de 1 cm, sa couleur: jaune ou jaune rougeâtre.

Les graviers et blocs des types, gris clair ou plus foncé, sont à l'extérieur du rouge vif, à l'intérieur commence un anneau, rouge plus foncé, jusqu'à quelques cm. Il peut être suivi d'une surface limite, jaune, à 1 millimètre qui est en zig-zag le long des éléments de texture plus compacts et plus résistants au processus d'oxydation. Le noyau intérieur est du gris verdâtre, qui peut présenter tant l'état primaire que secondaire: il est la partie légèrement oxydée de la grise (Photo 6).

Tant à l'œil nu qu'à la loupe, on peut constater également que la croûte extérieure, secondairement oxydée à une structure plus lâche, moins compact que le noyau intérieur réductif.

Au cours de l'altération et réaccumulation à durée plus longue, les détritiques de bauxite, oxydés dans leur entité, se transforment uniformément en bauxite rouge qui conserve la texture originale et ne diffère de la bauxite primaire, gris et gris verdâtre, qu'en sa couleur. Par l'avancement des processus d'oxydation, la porosité de la matière détritique augmente aussi et dans des cas extrêmes de certains blocs s'altère à structure scoriacée.

On peut distinguer la bauxite, brun rougeâtre, en état primaire aussi oxydée appartenant au *type B* et les morceaux de bauxite du *type A*, secondairement oxydés en brun rougeâtre. La compacité du *type B* est identique à celle de la bauxite primaire, gris ou noir, resp. au noyau interne, primaire et réductif des blocs oxydés (Photo 7).

Par rapport au précédent, le *type B* est toujours plus homogène, la micro-stratification ne se présente que dans le gisement de Tap Na. En sens de la structure de cristal intérieure sa texture est bien cristalline, et il est caractéristique que les cristaux de diaspre brillent souvent et visiblement à l'œil nu.

Dans le gisement de Ma Meo aux parties de gîtes de type de Leo Cao ce type forme en prédominance le complexe (Photo 8).

Nous avons rangé dans le *type C* cette partie des brèches de bauxite qui est composée des roches à consistences très différentes; notamment elle consiste en ciment de bauxite meuble, friable sous la main à aspect d'argile et en détritiques de bauxite durs, en prédominance appartenant au *type A*.

La texture du ciment est pisolitique-oolitique, seulement rarement homogène, gris ou brun rougeâtre et rouge brunâtre. Le ciment est secondaire-

ment réduit, pyritisé. Dimension de la matière détritique anguleuse ou légèrement usée: 2 à 6 cm. Sa couleur, gris, gris brunâtre ou brun rougeâtre est plus foncée par rapport au ciment. Il y a aussi de tels blocs, gris, plus grands dans le complexe qui renferment des détritiques de bauxite dure, brun rougeâtre, de 1 à 5 mm (brèche dans la brèche) (Photo 9).

Le type *C* n'est connu qu'au niveau profond de Tam Lung. Il est bien possible que telles brèches de bauxite ont été formées ailleurs aussi, mais à la surface mélangées avec la matière détritique des gîtes de bauxite développés en gros bancs, son aspect bréchique «primaire» n'est plus reconnaissable.

Type D: d'argilite est connu dans les gisements de Ba Xa et Ban Long. D'après la description par A. MINDSZENTY cette roche est plus meuble que la bauxite précédente, dure, et non rarement taillable à couteau. Sa texture est à la plupart oolithique ou détritique. Elle est souvent stratifiée en bancs minces à structure feuilletée. C'est l'alternance des stries détritiques et homogènes ou compactes et moins compactes qui cause la stratification (Photo 10). Sa texture n'est ordinairement recristallisée seulement, polluée de l'aléurite de quartz. En état oxydé jaune clair, jaune rougeâtre, le long des fissures à stries et taches, lilas rougeâtre [32]. Stratigraphiquement elle situe au niveau de bauxite. En sens vertical un passage net se manifeste vers la bauxite, visible à l'œil nu. Horizontalement, elle se trouve en contact de digitation par rapport à la bauxite.

Nous avons rangé dans la bauxite de type *E* quelques échantillons, provenant des lentilles Ban Thau du gisement de Ma Meo, qui sont d'après leurs aspect, composition minéralogique, resp. étude microscopique les produits d'altération latéritique des volcanites. Ce type se trouve toujours mélangé avec des détritiques de bauxite de karst, au maximum en quantité de 5%. Sa structure est pisolitique-oolithique, sa texture est poreuse, pour la plupart meuble ou moyennement dure, friable sous la main, sa couleur: rouge jaunâtre, jaune brunâtre, jaune lilacé, rouge lilacé à taches blanches (Photo 11). Elle présente du minéral à qualité très bonne à teneur en Al_2O_3 de 70 à 80%.

Au territoire de Ma-Meo, dans le complexe de bauxite de karst, il y a quelques fragments de cuirasse rappelant la latérite dont la teneur en oxyde de fer est proche de 60%. Leur couleur est du gris, leur reflet brille comme le métal, leur structure est scoriacée.

5. Composition minéralogique et propriétés structurales de la bauxite

En ce qui concerne la composition minéralogique de la bauxite vietnamienne, nous n'en avons pas d'informations détaillées que sur le territoire de Lang Son. La composition minéralogique est connue d'après 230 essais

à diffraction à rayons X (GY. BÁRDOSSY et M. TÓTH) et des évaluations qualitatives sur lames minces (A. MINDSZENTY) [32].

Diaspore: sauf la petite indication de Ban Long, il était démontrable dans tous les échantillons en quantité au moins de 1%, le plus fréquemment de 40 à 45%, au maximum de 82%.

Sa quantité diminue, surtout relativement, au cours des processus d'oxydation, resp. de l'effritement (Tableau 3).

Selon les études microscopiques on distingue deux générations:

- a. Sous forme microcristalline adhérente avec les minéraux ferreux de la matrice ou — surtout dans les types oxydés — sous forme grossièrement cristalline à aspect de mosaïque.
- b. Comme phase pure bien séparée des minéraux ferreux, dans la matrice ou en ségrégation à contour irrégulier se réglant sur les éléments structuraux concentriques à l'intérieur des différents éléments texturaux.

Boehmite: sa quantité reste en moyenne au-dessous de 5% dans les gîtes secondaires oxydés, bien que le maximum a été trouvé justement dans la croûte ferreuse, rouge d'une bauxite remaniée de Ba Xa: 56%. Dans quelques échantillons de Ban Long, seule la boehmite de 25 à 30% présente l'unique oxy-hydroxyde d'aluminium. Sa variation quantitative, à l'intérieur d'un profil se fait toujours en défaveur du diaspore (Fig. 7).

Gibbsite: on la retrouve, en quantité plus ou moins élevée, presque en toutes les bauxites de surface. Sa quantité est le plus souvent de 4 à 7%, au cours de l'effritement sa moyenne monte de 2 à 4%. C'est un échantillon de Tam Lung qui a atteint le maximum: 22,4%.

D'après les études microscopiques son enrichissement est dû aux processus secondaires.

Kaolinite — méta-halloysite: d'après les études de diffraction à rayons X, on ne peut distinguer quantitativement ces deux minéraux. Leur quantité totale peut varier entre quelques dixièmes et 80%, mais à cause de l'inhomogénéité de la roche elle peut varier entre des limites larges, même à l'intérieur d'un seul échantillon. La strie grise d'un échantillon provenant de Bang Long comprend 33% de kaolinite à côté de la boehmite de 37%, et la partie gris clair contient 79% de kaolinite + méta-halloysite à côté de 1% de boehmite.

Chamosite: c'est le minéral le plus caractéristique de la bauxite vietnamienne. Malheureusement, lors les analyses de diffraction à rayons X nous n'avions pas la possibilité de distinguer ses variétés ferrique et ferreuse. Il est probable que dans les variétés de bauxite, noir, gris et vert grisâtre, c'est la *ferro-chamosite*, plus riche en aluminium, qui domine, et dans les types, brun foncé et rouge brunâtre c'est la *ferri-chamosite*. Dans les variétés de bauxite oxydées sa quantité est ordinairement moins élevée et au lieu d'elle ce sont premièrement la kaolinite, gibbsite et la boehmite qui s'enrichissent. Dans la moyenne de la bauxite de Lang Son elle est présente en 5%. Dans la bauxite à détrit, parallèlement à la diminution de la granulométrie sa quantité

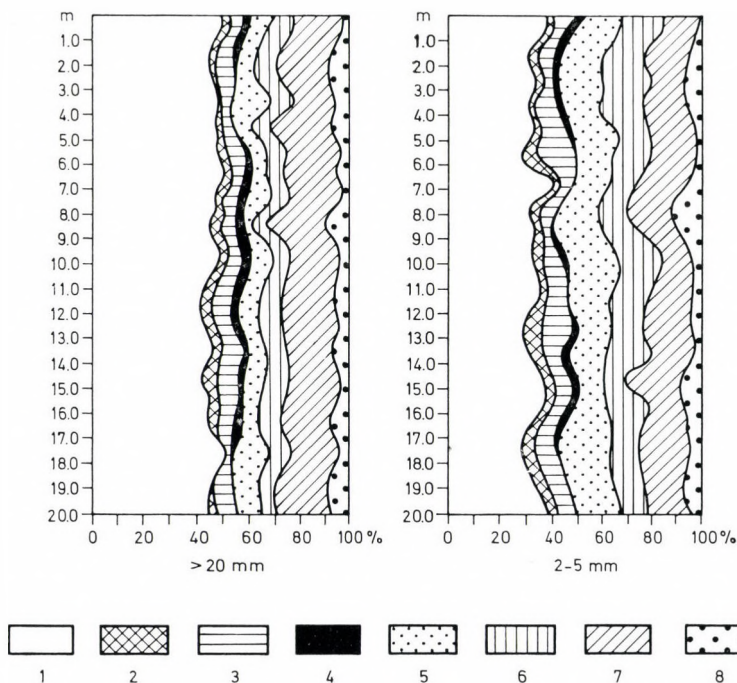


Fig. 7. Composition minéralogique des fractions granulométriques >20 mm et de 2–5 mm du complexe de détritits de bauxite dans un puits creusé: 1. diaspor; 2. boehmite; 3. gibbsite; 4. chamosite; 5. kaolinite; 6. hémate; 7. goethite; 7. autres minéraux

abbaisse aussi de 5,5 à 1%. Mais, tels échantillons ne sont pas rares non plus, où il y a 30 à 40% de chamosite. Dans la bauxite rouge foncé — primordialement oxydative — nous trouvons aussi de la chamosite en 34%, mais les valeurs maximales se rapportent aux variétés réductives.

Son aspect microscopique: aiguilles et écailles fines à dimensions pour la plupart de quelques microns, dans la matrice de la bauxite. Sporadiquement, elle présente des écailles idiomorphes distinctes de 100 microns. La variété finement écaillée compose rarement des amas radio-fibreux, montrant parfois une orientation nette (Photo 12a, b).

Goethite: elle est particulièrement répandue dans les types oxydés. Dans la croûte oxydative, jaune, extrêmement mince de la bauxite, noir, du territoire de Ba Xa, elle se présente entre 8 et 25%. Dans la bauxite à détritits, sa quantité augmente de 9 à 16% parallèlement à la diminution de la granulométrie. Sa quantité moyenne ne dépasse celle de l'hématite que dans les gîtes les plus oxydatifs de type de Dong Dang.

Hématite: En considérant la moyenne du territoire entier, c'est la dominance de l'hématite qui caractérise la bauxite par rapport à la goethite, dans certains cas elle dépasse trois fois la goethite.

Sous microscope: elle remplit les pores dans la matrice où présente des formations en stries, ailleurs forme des écailles pseudo-hexagonales de 1 à 2 microns. Composant constant des zones des pisolithes et celithes. Elle forme pseudomorphoses d'après la pyrite. Imprègne presque entièrement les types de bauxite fortement oxydés.

Parmi les autres minéraux il faut mentionner la *lépidokrokit*, supposablement d'origine secondaire et atteignant quelques pourcents dans certains échantillons. La *pyrite* est sédimentaire dans le gîte de Tam Lung, mais dans le gisement de Ba Xa elle est en prédominance d'origine téléthermale. Sa quantité peut atteindre 10 à 15%, dans certains échantillons manuels. La *sidérose* s'enrichit surtout dans l'argilite, et ici elle atteint 30% au maximum. Dans les bauxites primaires, en général elle reste au-dessous de 1%. Les minéraux du titane atteignent, en total, même 4 à 5%. C'est l'*anathase* qui domine, l'*ilménite* et le *rutile* ne présentent que quelques dixièmes de pourcent. Dans les tableaux illustrant la composition minéralogique, on n'a présenté que les quantités des sept minéraux principaux (diaspore, boehmite, gibbsite, kaolinite, chamosite, hématite, goethite) en pourcentages, la différence par rapport au 100% est égale au total des autres minéraux ci-énumérés.

La composition minéralogique ne présente pas de variations régulières ni verticalement, ni horizontalement, dans les gîtes de bauxite *in situ* ou à détritits étant place en ou en voie de remaniement. Comme on va le voir dans le chapitre 6.4. aussi, du point de vue minéralogique et géochimique les gîtes de bauxite paléozoïques sont considérablement plus homogènes que soit ceux de la province de bauxite méditerranéenne, soit ceux des territoires latéritiques.

Pour l'illustration de l'homogénéité verticale, dans la Fig. 7 est présentée la composition minéralogique d'un puits creusé de Ma Meo, concernant les fractions plus grossières de 20 mm, resp. entre 2 et 5 mm. Comme on peut le lire dans la figure aussi, la composition minéralogique de l'échantillon d'un intervalle présente une différence plus grande selon les différentes fractions que les intervalles de 1 m successifs du profil.

Étudions les changements de la composition minéralogique dans la brèche de bauxite du gîte primaire de Tam Lung, en comparant les détritits de bauxite durs et la matière bauxitique du ciment (cim) (Tableau II).

Il est bien visible que dans la matière du gîte primaire la différence entre les détritits durs et le ciment meuble est restreinte, pour la plupart, à la différence plus considérable manifestée dans la teneur en diaspore. Il n'y a qu'un seul cas, où une différence plus considérable apparaît dans la teneur en chamosite. Dans le ciment, l'hématite s'enrichit légèrement par rapport aux détritits.

Nous avons étudié dans le détail les variations de la composition minéralogique en fonction de la granulométrie au territoire de Ma Meo, d'où on a effectué en total 107 analyses par diffraction à rayons X sur les fractions plus grossières de 20 mm, entre 10 et 20 mm, entre 5 et 10 et entre 2 et 5 mm. D'après

Tableau II

Sondage LK-185.	78,0 m		80,8 m		81,1 m	
	détritus	cim.	détritus	cim.	détritus	cim.
diaspore	40,3	25,8	42,6	17,5	47,4	27,4
boehmite	2,3	10,0	—	8,0	—	11,0
gibbsite	—	—	—	—	—	—
kaolinite	—	2,9	—	—	—	—
chamosite	42,5	43,9	35,8	17,2	30,5	30,4
hématite	4,3	8,6	12,4	14,9	5,3	7,8
goethite	1,0	4,2	—	3,2	—	—
autres	0,6	13,8	9,2	39,2	16,8	23,4

celles-ci

- le diaspore diminue de 50% à 35% parallèlement à l'abaissement de la granulométrie,
- la boehmite varie entre 2 et 3% dans les différentes fractions,
- la gibbsite s'enrichit de 3,5% à 7% parallèlement à l'abaissement de la granulométrie,
- la kaolinite s'enrichit de 10% à 16%,
- la chamosite diminue de 3% à 0,7%,
- l'hématite varie, sans tendance, entre 16,5% et 18%,
- la goethite s'enrichit de 9% à 14%.

Dans le Tableau III. est récapitulée la composition minéralogique des parties de quelques échantillons de bauxite étant à de différents degrés d'oxydation.

En négligeant le fait que même un échantillon de bauxite manuel pourrait être plus ou moins inhomogène, on ne se trompe pas trop, en expliquant par l'oxydation les différences minéralogiques apparues à l'intérieur des échantillons. On peut constater que c'est la *chamosite* qui est la plus sensible à l'oxydation. Dans la croûte rouge sa quantité diminue en saut par rapport au noyau, gris ou vert, resp. à l'anneau. Dans ces cas-ci, d'après les analyses chimiques la présence de *ferro-chamosite* est probable. Dans l'échantillon provenant de 22,0 m du sondage LK-156, la chamosite ne diminue que de 32% à 19%, et dans ce cas-ci nous pouvons probabiliser la présence de la *ferro-ferri-*, resp. de la *ferri-chamosite* plus résistantes à l'oxydation. Parallèlement à la diminution de la chamosite, la kaolinite et l'hématite s'enrichissent davantage et la gibbsite moins (voir plus détaillément: chapitre 7.).

En récapitulation, nous présentons la composition minéralogique moyenne des différents gisements bauxitifères ensemble avec les valeurs minimales et maximales y observées (Tableau IV).

Tableau III

Tableau comparatif des parties réductives et oxydatives des échantillons de bauxite

localité	n° d'échantillon	bauxite	diaspore	boehmite	gibbsite	kaolin.	chamos.	hémat.	goethite
Tam Lung	Tmk-4	croûte rouge	63,3	—	—	1,0	1,8	19,0	4,0
		noyau gris	54,5	—	—	5,0	21,4	6,0	4,2
Khon Pich Lk-156	22,0 m	croûte rouge	57,7	—	—	3,5	19,0	3,4	9,7
		noyau gris	56,2	—	1,3	2,6	32,6	1,6	1,2
	102,0 m	croûte brun rougeâtre	51,9	—	1,5	3,7	4,5	15,8	13,5
		anneau jaune	50,5	—	1,0	1,3	30,0	4,0	7,5
Ba Xa	Bxk-7	noyau gris verdâtre	47,9	—	1,0	3,2	32,2	3,2	1,5
		croûte rouge	29,9	2,6	5,0	23,0	—	8,6	25,6
	Bxk-13	noyau vert	19,1	—	—	23,8	34,7	—	18,0
		croûte rouge	3,4	56,0	1,2	12,9	—	6,2	15,5
		anneau vert grisâtre	1,7	51,0	—	8,8	21,7	—	11,3
		noyau gris	1,4	53,5	—	6,0	32,3	—	2,0
	Bxk-23	anneau vert grisâtre	52,9	—	—	10,5	26,8	—	10,0
		noyau gris foncé	38,8	15,4	—	3,5	27,5	—	—
Ma Meo	Bt-5/9	croûte rouge	30,0	—	—	8,6	—	54,5	3,5
		noyau gris lilas	18,0	—	7,6	21,5	18,2	1,3	20,0

Tableau IV

Tableau récapitulatif de la composition minéralogique

	Ma Meo		Dong Dang				Tam Lung		
	max.	moy.	min.	max.	moy.	min.	max.	moy.	min.
diaspore	91,5	46,7	0,2	55,2	47,4	37,0	82,1	39,4	0,7
boehmite	44,1	2,2	—	2,0	0,5	—	44,9	8,3	—
gibbsite	13,0	4,2	—	2,0	0,7	—	22,4	1,4	—
kaolinite	40,0	11,4	—	23,6	16,8	11,0	64,8	11,6	—
chamosite	34,1	2,6	—	2,0	0,9	—	49,4	12,4	—
hématite	54,5	16,6	—	18,6	12,1	9,1	27,1	10,5	—
goethite	28,0	10,1	—	22,3	14,2	8,0	67,4	7,9	—

Le nombre des analyses à diffraction à rayons X:

Remarque: Ma Meo 110, Dong Dang 17, Tam Lung 64, Khon Pich 9, Ba Xa 20 et Ban

6. Composition chimique de la bauxite: conclusions géochimiques

6.1. Enrichissement de la bauxite, variation de la composition chimique en fonction de la granulométrie

Le fait que le complexe de bauxite à détritits est composé de bauxite dure et d'argile meuble encaissante a offert l'élaboration de la méthode d'enrichissement de la bauxite par voie physique, en place. Nous avons divisé sous jet d'eau puissant et à travers une série de tamis la matière écartelée en 300 kg environ de chaque mètre du profil du puits creusé à profil $1,5 \times 1$ m, en fractions >20 mm, de 10 à 20 mm, de 5 à 10 mm, de 2 à 5 mm et <2 mm. En connaissance du volume du puits creusé, en pesant le poids des différentes fractions par mètres, nous avons présenté la valeur de rendement exprimée en t/m^3 qui caractérise très bien les différents types de gîtes (comme nous l'avons décrit dans le chapitre 3.2).

C'est la matière des fractions plus grossières de 2 mm qu'on a mis sous l'analyse en série pour cinq composants, car nous avons supposé préalablement qu'elle soit économiquement traitable du point de vue industriel. Nous avons calculé les réserves de Ma Meo de Dong Dang et de Khon Pich et la qualité de ces réserves concernant la matière des fractions plus grossières de 2 mm [25]. A l'aide de la méthode d'enrichissement à lavage employée, nous avons amélioré la qualité des réserves de minerais, préalablement présentées et calculées, à Ma Meo moins, mais aux territoires de Dong Dang et de Kho-Pich considérablement. En contrôle, de temps en temps nous avons échantillonné et analysé aussi la matière de la fraction plus fine de 2 mm.

La Chaire de Traitement de Minéral de l'Université Technique d'Industrie Lourde à Miskolc a effectué des essais d'enrichissement sur quelques échantil-

des gîtes de bauxite de Lang Son

Khon Pich			Ba Xa			Ban Long		
max.	moy.	min.	max.	moy.	min.	max.	moy.	min.
57,7	45,4	0,5	52,9	16,3	—	2,5	1,7	—
36,6	4,6	—	56,0	21,4	—	37,9	16,0	—
3,1	1,0	—	5,0	0,6	—	6,3	1,5	—
20,2	6,6	1,3	61,0	23,5	—	78,9	34,9	—
32,3	19,9	—	37,5	16,7	—	83,0	16,9	1,0
15,8	7,4	1,6	8,6	1,2	—	36,1	3,8	—
13,5	8,5	1,2	25,6	13,4	—	14,5	4,4	—

Long 21. Alors, la fiabilité des valeurs présentées n'est pas identique pour les différents gisements.

lons caractéristiques, pesant plusieurs centaines de kgs, aux types de gîtes de Leo Cao, Ma Meo et de Dong Dang [51]. Ces échantillons présentaient la quantité de matière totale prise d'un mètre des différents puits creusés. Les essais du point de vue industriel importants ont livré des informations précises à la connaissance des processus géochimiques déroulés au cours de l'effritement de la bauxite en tant qu'à côté de la répartition des 12 différentes fractions granulométriques ils ont aussi livré les données des analyses à cinq composant des différentes fractions. Pour illustrer les données, nous avons construit une succession de figures (Fig. 8). Parallèlement à la diminution de la granulométrie, la quantité de l'aluminium, de fer et du titane diminue aussi, mais celle du silice, resp. de la perte au feu augmente, ce qui est caractéristique à tous les trois types de gîtes. La variation du titane et de la perte au feu est régulière, par contre la répartition d' Al_2O_3 , SiO_2 et de Fe_2O_3 présente un changement-extrêmement brusque dans le profil du type de Le Cao, entre les fractions de 2 à 0,8 mm, et de 0,8 à 0,07 mm. Dans le type de Ma Meo ce changement n'est plus aussi brusque et se présente au domaine de 1 à 0,07 mm. Dans le type de Dong Dang, la variation des trois éléments principaux est étonnamment plus régulière. La diminution, resp. l'enrichissement des trois éléments principaux est toujours parallèle à la diminution brusque de la quantité des fractions granulométriques. Dans les types de Leo Cao et de Ma Meo, dans les fractions plus fines de 0,8 mm, resp. de 0,07 mm la quantité de la matière est répartie plus régulièrement entre les différentes fractions granulométriques, et ici la quantité d' Al_2O_3 et de SiO_2 est presque constante, aussi.

Par comparaison des trois types il est bien visible que c'est la matière des gîtes de type de Leo Cao qu'on peut enrichir le mieux, non seulement,

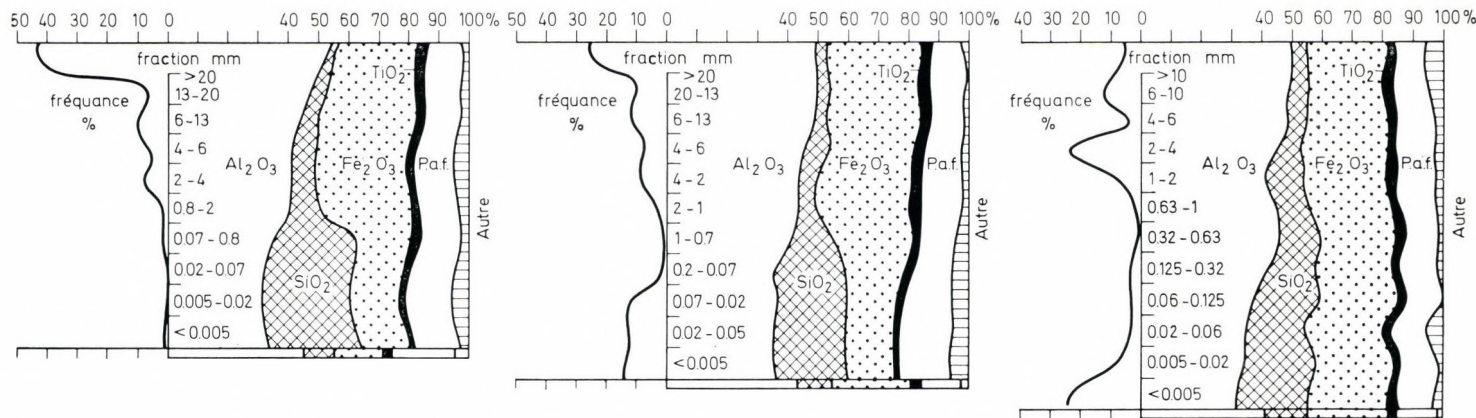


Fig. 8. La composition chimique des trois types détritiques de gîte en fonction de la granulométrie

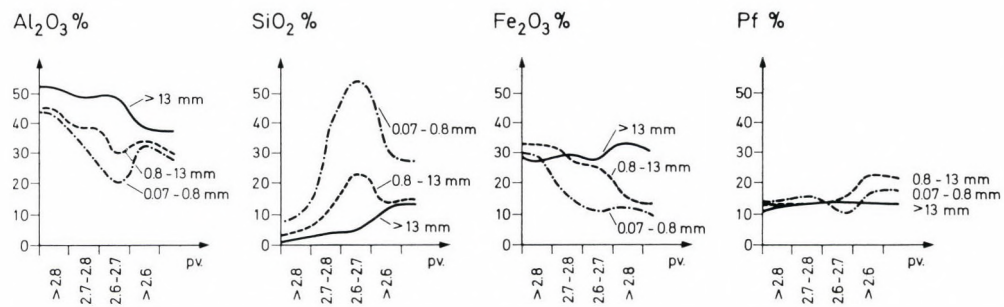


Fig. 9. La composition chimique en fonction de la granulométrie et du poids volumétrique

car ici la quantité des fractions plus fines de 2 mm est insignifiante par rapport aux autres, mais aussi à cause de cela, car à la granulométrie de 2 mm on peut exécuter l'enrichissement vite et à peu de consommation en énergie. D'après la qualité de la bauxite, il faut construire l'enrichissement industriel au type de Ma Meo à la fraction plus grossière de 1 mm et à celui de Dong Dang à celle plus grossière de 0,32 mm.

La proportion $\text{Al}_2\text{O}_3/\text{SiO}_2$ est la plus proche de l'argile pure dans le type Leo Cao, tandis que dans les types de Ma Meo et de Dong Dang la proportion entre les minéraux de bauxite et les minéraux d'argile est 1,5 à 2 : 1.

6.2. Variation de la composition chimique en fonction du poids spécifique

A la Chaire de Traitement de Minéral, on a étudié dans les domaines granulométriques >13 mm, 0,8 à 13 mm et 0,07 à 0,8 mm, comment change la composition des éléments dans les catégories du poids spécifique plus grand de 2,8, 2,7 à 2,8, 2,6 à 2,7 et plus petit de 2,6. Les résultats sont illustrés en diagramme (Fig. 9).

D'après la figure, dans la fraction >13 mm, parallèlement à la diminution du poids spécifique, la quantité d' Al_2O_3 monte légèrement au début, puis abaisse, la quantité de la perte au feu monte plus légèrement, mais celle de SiO_2 et de Fe_2O_3 plus abruptement.

Si nous négligeons provisoirement l'anomalie entre les valeurs de poids spécifique 2,7 et 2,6, parallèlement au poids spécifique diminuant, la quantité d' Al_2O_3 et de Fe_2O_3 diminue de plus en plus abruptement, dans les fractions granulométriques de plus en plus fines, la perte au feu un peu moins et SiO_2 un peu plus.

L'anomalie apparue entre les valeurs du poids spécifique 2,7 et 2,6 est le résultat de l'enrichissement en silice considérable en tant que dans les fractions granulométriques 0,07 à 0,8 mm à côté d' Al_2O_3 de 10% et de Fe_2O_3 de 11% il y a 54% de SiO_2 . Selon notre avis, il peut s'agir des détritits de silice à grains fins qui ont été épargnés après la dissolution du calcaire du toit et mélangés avec la matière de la bauxite. Bien que l'on peut observer des précipitations de silice à grandeur de lentille ou de haricot aussi dans le calcaire du toit, supposablement cette anomalie n'apparaît pas dans la matière des fractions plus grossières, car la masse de bauxite de la fraction «supprime» l'effet de cette addition en silice.

6.3. Géochimie de l'oxydation de la bauxite

La différence apparue dans la composition minéralogique entre le noyau intérieur réductif et la partie extérieure oxydatée de certains échantillons mérite l'attention particulière. Pareillement, d'après les informations des

analyses chimiques à 10 composants des six échantillons choisis, on peut suivre le processus géochimique déroulé dans les détritits de bauxite. D'après le Tableau V, la plus grande différence entre les parties oxydées et réduites apparaît dans les quantités de Fe_2O_3 , resp. de FeO en tant qu'au cours de l'oxydation la quantité de FeO peut diminuer même à deux ordres de grandeur. La valeur de FeO de la partie oxydée est autour de 10 et celle de la partie réduite est autour de 1. Le degré d'oxydation de la bauxite verte est seulement un peu plus élevé que celui de la bauxite grise (échantillon n° Bx-23).

Du point de vue industriel il est important que les parties oxydées contiennent successivement de moins en moins de SiO_2 que leurs paires réduites; au cours de l'oxydation, la teneur en SiO_2 des détritits de bauxite a diminué avec la quantité dépassant 60% (voir: chapitre 7.3.). La concentration en titane de la bauxite oxydée augmente et la bauxite devient plus hydratée qui s'exprime aussi dans la composition minéralogique. Parallèlement à l'oxydation, la quantité des alcalins et du soufre présente une tendance à diminuer, par contre la teneur en matière organique reste inchangée.

6.4. Caractérisation géochimique des gisements bauxitifères d'après la proportion entre Al_2O_3 , SiO_2 et Fe_2O_3

Nous avons construit les histogrammes des trois principaux composants d'après l'analyse chimique de l'ordre de grandeur de dizaine de milles. Au territoire de Tam Lung nous avons calculé la fréquence d'après l'analyse chimique du complexe de bauxite entier, c'est-à-dire des détritits de bauxite durs et de l'argile meuble encaissante, tandis qu'aux gisements de Ma Meo, Dong Dang et de Khon Pich nous n'avons tenu compte que des fractions plus grossières de 2 mm du complexe à détritits de bauxite. A Ba Xa les données concernent la partie enrichie de la bauxite remaniée et de la matière d'échantillons entière de la bauxite *in situ* (Fig. 10).

Dans les histogrammes du gisement de Ba Xa, nous avons séparé la répartition d'éléments de la bauxite primaire pure qui présente des courbes en cloche (de Gauss) symétriques à un maximum tant à l' Al_2O_3 qu'au SiO_2 ce qui indique l'homogénéité de la bauxite primaire. Les minces zones allongées, hachurée de l' Al_2O_3 au-dessus de 45% et du SiO_2 au-dessous de 10% indiquent la phase plus oxydée et à qualité meilleure de la bauxite réaccumulée. Les aires hachurées de l' Al_2O_3 au-dessous de 40% et du SiO_2 au-dessus de 17% indiquent la présence de l'argilite intercalée dans la bauxite *in situ*, resp. des détritits siliceux mélangés avec le complexe de bauxite réaccumulé.

Dans les diagrammes du gisement de Tam Lung, nous avons séparé, par hachures, la matière meuble encaissante et les détritits de bauxite durs. Bien que ce territoire est aussi caractérisé par les courbes symétriques à un

Tableau V

Composition chimique des parties oxydatives et réductives de quelques échantillons de bauxite

			Al ₂ O ₃	SiO ₂	Fe ₂ O ₃	FeO	TiO ₂	K ₂ O	Na ₂ O	+H ₂ O	CO ₂	S
Tam Lung	Tmk-4	croûte rouge	56,1	2,9	22,3	1,2	3,2	0,50	0,06	12,7	—	0,06
		noyau gris	54,6	8,7	6,3	13,5	2,6	0,90	0,10	11,4	0,03	0,11
Khon Pich	102 m	anneau jaune	50,0	3,0	27,3	1,3	4,0	0,09	0,20	13,5	0,09	0,0
		noyau gris	52,5	8,4	13,7	8,3	3,7	0,05	0,11	12,9	0,07	0,08
Ba Xa	K-276	croûte rouge	59,4	0,7	18,2	0,7	3,2	0,01	0,01	14,9	0,84	0,04
		noyau gris	58,2	8,6	3,9	12,0	2,9	0,04	0,04	12,9	0,63	0,09
	K-278	croûte jaune	54,0	4,6	22,6	0,8	3,8	0,05	0,04	12,6	0,14	0,04
		noyau gris	54,2	10,9	17,0	2,1	3,1	0,04	0,03	11,4	0,22	0,02
	K-284	croûte jaune	56,2	2,8	21,2	1,5	3,2	0,02	0,01	14,1	0,14	0,04
		noyau gris	54,1	9,6	9,9	10,0	2,9	0,03	0,01	12,4	0,14	0,05
	Bxk-23	anneau vert	55,5	8,0	9,6	10,2	3,6	0,03	0,09	12,8	0,07	—
		noyau gris	56,2	11,4	6,2	11,2	3,5	0,11	0,10	13,1	traces	—

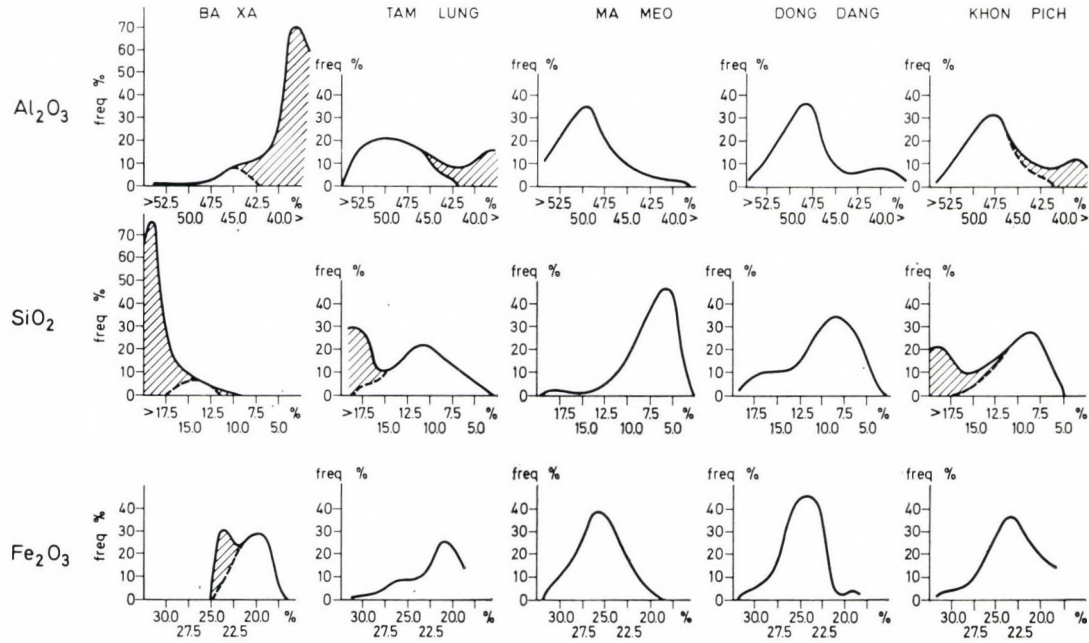


Fig. 10. Histogrammes des composants Al_2O_3 , SiO_2 et Fe_2O_3

maximum, ici la qualité des détritits de bauxites durs est déjà meilleure que la bauxite primaire de Ba Xa.

Les courbes asymétriques de répartition d'éléments du gisement de Ma Meo ont un maximum et leur descente est semblable à celle des gîtes à bonne qualité de la province de bauxite méditerranéenne.

Dans le gisement de Don Dang, la quantité de la matière à teneur en Al_2O_3 moindre que 45% et en SiO_2 plus que 12,5%, est plus fréquente qu'au territoire de Ma Meo ce qui est d'accord avec l'effritement plus poussé (chapitre 7.3.).

La bauxite remaniée du gisement de Khon Pich est polluée de détritits d'aléurolite. Cette contribution de matière étrangère s'exprime dans les histogrammes de l' Al_2O_3 et du SiO_2 à deux maximums. En détachant les distorsions des courbes causées par l'aléurolite, on obtient des courbes légèrement asymétriques à un maximum devenant semblables à celles du gisement de Dong Dang.

6.5. Géochimie des éléments en traces

Dans les échantillons des gisements bauxitifères de Lang Son nous avons fait effectuer 31 examens d'éléments en trace demi-quantitatifs pour pouvoir préciser lesquels sont les composants dont l'analyse quantitative mérite l'étude. D'après celle-ci on a fait effectuer les analyses spectrales quantitatives au Ga, Be, Zr et au V, à l'Entreprise de Recherche de Bauxite à Balatonalmádi. Le tableau VI récapitul les résultats, y sont aussi présentées les valeurs moyennes calculées par Schroll et Sauer [41] pour les bauxites karstiques méditerranéennes.

Tableau VI

gisements éléments ppm	Ma Meo	Dong Dang	Tam Lung	Ba Xa	bauxites médit.
Ga max.	36	39	43	45	
min.	18	20	8	6	
moy.	29	30	24	23	52
Be max.	11	10	10	4	
min.	3	4	1	1	
moy.	4	6	4	2	10
Zr max.	1288	1036	873	992	
min.	592	614	451	333	
moy.	816	872	672	784	574
V max.	346	216	193	242	
min.	186	138	82	108	
moy.	242	183	113	161	222

Nous avons composé le tableau d'après l'étude de 100 échantillons. En comparant la qualité des quatre éléments en traces aux concentrations d'éléments conformes des gîtes de bauxites karstiques méditerranéens, on n'observe pas de telle différence à partir de laquelle on pourrait déduire une conclusion génétique plus poussée, et d'autant moins, car les nombres des analyses diffèrent au moins par un ordre de grandeur, et ainsi la comparaison est plutôt informative.

En comparant les valeurs moyennes des différents gîtes du territoire de Lang Son, on voit que les éléments de traces s'enrichissent plutôt dans les gîtes réaccumulés et plus oxydés de Ma Meo et de Dong Dang, tandis que la bauxite primaire et celle secondairement réaccumulée à proximité de la précédente sont plus pauvres en éléments de traces.

7. Problèmes sur la genèse de la bauxite

D'après les observations de terrain au Vietnam du N et des examens des matériaux de Hongrie ainsi que sur la base de la littérature y relative, nous allons esquisser les conditions de la genèse, resp. de l'accumulation de la bauxite paléozoïque et de la réaccumulation plus récente.

Tant dans la recherche qu'au cours des examens des matériaux, nous étions obligés de suivre les intérêts industriels propres, alors nos études ne pouvaient aborder tous les détails, il est bien probable que les recherches ultérieures, les explorations minières et les nouveaux examens des matériaux plus détaillés vont préciser encore notre conception évoluée jusqu'à présent.

Nous pensons que cette conception est surtout valable aux gisements des environs de Lang Son, mais dans les autres gisements du Vietnam du N elle n'est utilisable que par endroits ou seulement partiellement. D'après la littérature nous supposons ainsi que les gîtes paléozoïques à chamosite et à diaspore des autres territoires de l'Asie de Sud-Est et de l'Oural et les gîtes primaires du Vietnam du N soient en parenté génétique.

7.1. Problèmes généraux sur la genèse de la bauxite et des minéraux

Au Vietnam du N, dès le Dévonien jusqu'au Trias, le Paléozoïque est caractérisé par la sédimentation marine à carbonates, produisant des formations de mer peu profonde, mer ouverte et littorales. Dans le Carbonifère et le Permien, la sédimentation a été plusieurs fois et localement interrompue, par endroits les discordances et les diasthèmes, plus ou moins marqués, indiquent une mer archipélagique, dans cet espace.

C'est au Permien supérieur que la régression générale est arrivée, comme résultat de l'épirogenèse accompagnée de lents mouvements oscillatoires.

Les émersiones n'étaient que partielles, car parallèlement à la sédimentation continentale, et à sa proximité immédiate, on peut démontrer par endroits la sédimentation marine continue. Dans la région littorale, à côté des bancs houilleux le changement des conditions paléogéographiques est aussi indiqué par ce que la sédimentation du calcaire a été interrompue et au lieu de celui c'est l'argilite qui a été déposée.

La cessation de la précipitation des carbonates peut être expliquée par ce que le long du littoral, divisé par lagunes en voie d'isolation, la mer a été adoucie, et par suite les conditions devenaient défavorables au biocénose marin précédent, c'est-à-dire le rôle des facteurs biologiques a cessé dans la sédimentation des carbonates [50].

La formation de bauxite a commencé dans la zone supralittorale, mais en terre ferme à la surface du calcaire. D'après la situation paléogéographique, c'est le calcaire l'unique roche-mère possible. Dans un ouvrage précédent, l'auteur [23] a prouvé, par analyses chimiques, la possibilité de la formation de la bauxite à partir du calcaire carbonaté, en soulignant qu'à cause de leur basse teneur en R_2O_3 on ne peut exclure les roches carbonatées de la gamme des roches-mères, car au cours de l'altération l'enrichissement des différents éléments ne dépend pas de la composition chimique de la roche-mère, mais de leur comportement géochimique. En ce sens les roches carbonatées présentent des bons producteurs de bauxite même, car elles sont composées de tels éléments qui sont extrêmement mobiles (métaux terreux alcalins) ou extrêmement immobiles (Al_2O_3 et Fe_2O_3) sous le climat tropical et subtropical. Au milieu alcalin SiO_2 se dissout en grande partie et en partant c'est de la bauxite qui se forme, tandis qu'au milieu moins alcalin ou neutre il reste en place et c'est de la *terre rouge* qui se produit.

Par rapport au Mésozoïque et Cénozoïque, dans le Carbonifère et le Permien l'air, resp. les précipitations étaient considérablement plus riches en CO_2 [45]. Ce qui influençait favorablement la dissolution du calcaire et supposablement la formation de bauxite a été aussi plus rapide que dans le Mésozoïque.

Une partie du résidu de dissolution a été déposée sur le continent — dans les dépressions karstiques remplies d'eau à la saison pluvieuse — et une autre partie transportée vers la mer est accumulée dans la zone littorale immédiate périodiquement envahie de la mer: dans les baies en voie d'isolation. Ainsi à partir du continent jusqu'à la mer plus ouverte des faciès de bauxite caractéristiques ont pris leur naissance et comme faciès hétéropiques suivants d'abord c'est l'argilite pure, puis l'argilite divisée par bancs de calcaire qui ont été déposées (Fig. 11).

Au continent, les minéraux de la bauxite précipitaient des gels amorphes. En dépendant du pH variable et de la puissance de l'érosion, le transport vers la mer déroulait sous formes des solutions (véritables et colloïdales),

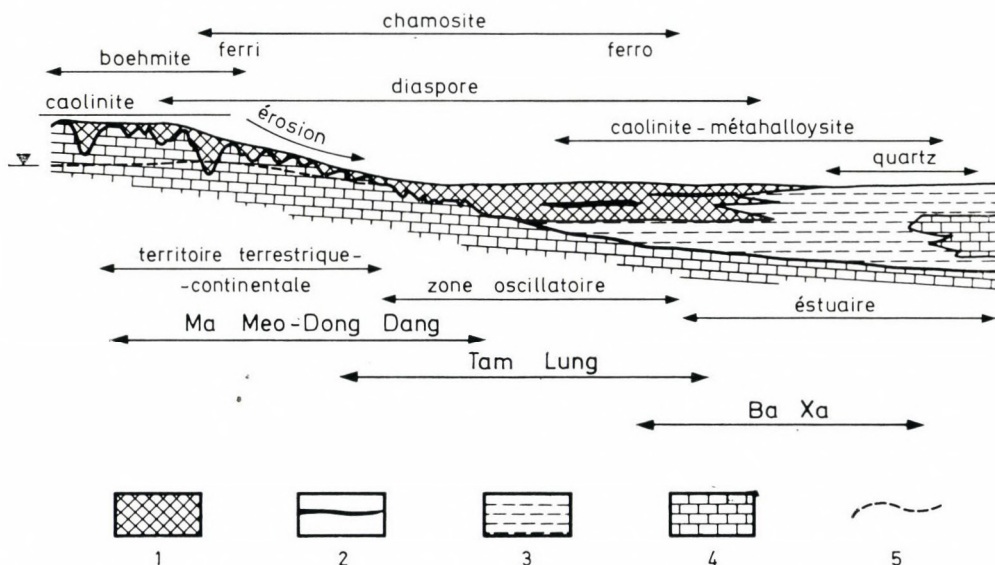


Fig. 11. Faciès de bauxite dans la zone côtière permienne supérieure 1. bauxite; 2. houille, argile à houille, houille argilleuse; 3. argilite; 4. calcaire; 5. niveau des eaux karstiques

et de suspensions à pélites et même des détritits fins (Fig. 2): détritits de bauxite, rouge, de 2 à 4 mm dans la bauxite, grise.

Le golfe marin en voie d'isolation devenait périodiquement marais. Ici, à cette période l'accumulation de la bauxite a cessé à cause du pH bas, et c'est le dépôt de l'argile meuble, pauvre en fer, qui accompagnât les bancs houilleux en voie de dépôt. Après l'extinction du marais, l'accumulation de la bauxite a recommencé. Il est bien frappant que même la bauxite noire ne contient plus que 2% de matières organiques.

La composition minéralogique de la bauxite et cette reconstruction paléogéographique sont en accord harmonique. Selon les opinions de I. VALETON, P. J. COMBES et V. T. D'ALLEN, le diaspore précipite de l'alumogel amorphe en milieu légèrement réductif [48, 10, 1]. En étudiant la genèse minéralogique des gîtes de bauxite de Hongrie, l'auteur est arrivé aux avis identiques et a montré que les conditions de Eh—pH sont les fonctions de l'échelle du drainage, c'est-à-dire de la profondeur de l'eau karstique à partir de la surface [24].

Nous avons construit le diagramme Eh—pH de la Fig. 12, valable aussi à la composition minéralogique de la bauxite vietnamienne, en utilisant les résultats d'essais de synthèse de H. HARDER, faites sur la nontronite.

D'après la comparaison entre la situation paléogéographique et du diagramme n° 12, tant au continent que dans les golfes marins c'est surtout le diaspore qui se précipite. Actuellement et pour la plupart, la bauxite contient de la boehmite et de la gibbsite, en quantité plus ou moins élevée. Les études de

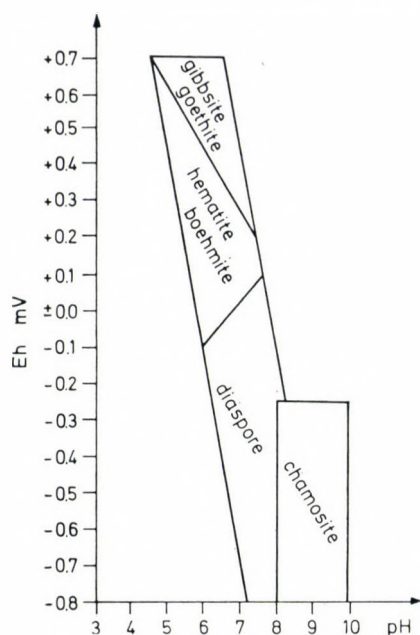


Fig. 12. Stabilité des principaux minéraux d'aluminium et de fer de la bauxite en fonction de et du pH (HARDER H.)

A. MINDSZENTY, faites sur lames minces ont décelé que la gibbsite est entièrement secondaire et la boémite en est aussi en partie: leur naissance est explicable par l'oxydation ultérieure [32].

Comme nous l'avons illustré dans la Fig. 11, la boémite a été aussi formée aux territoires plus élevés, relativement mieux drainés. L'érosion continuelle transportait périodiquement la bauxite, l'a déposée, vers les territoires situés plus bas, où elle est devenue instable et s'il y avait assez de temps, s'est transformée en diaspore. Si la réaccumulation dans le Permien a été déroulée immédiatement avant le dépôt du toit — par ex. au territoire de Tam Lung — alors des parties de gîtes, riches en boémite, se développaient.

Ce sont les minéraux du fer qui signalent le plus sensiblement les conditions d'oxydation. Dans cette paragenèse, la goëthite indique le milieu le plus oxydatif et la ferro-chamosite celui le plus réductif. Ne considérant ce que les processus d'oxydation secondaires ont modifié la composition minéralogique originale, l'image paléogéographique et les traits texturaux de la bauxite suggèrent que l'hématite (et un peu de goëthite aussi) pouvait être formée primitivement aussi aux territoires plus élevés, mieux drainés et, partant, plus oxydatifs. Ainsi, à côté de la boémite les minéraux de l'oxyde ferrique — présents dans la composition minéralogique actuellement observée — s'étaient formés en partie à manière primaire. De même, nous considérons la ferri-

chamosite aussi comme minéral primaire. Il n'est pas probable qu'elle soit formée secondairement à partir de la ferro-chamosite, car on peut difficilement imaginer l'oxydation de la ferro-chamosite sans désintégration du réseau moléculaire. La quantité de la ferro-chamosite a été influencée par celle de l'ion de Fe^{++} étant à disposition dans la solution qui a aussi fixé une partie de l'aluminium. Conformément aux conditions de $\text{Eh}-\text{pH}$, l'Al devenu en état instable dans la solution est précipité sous forme du diaspore. En allant vers la mer plus ouverte, la concentration de la solution en aluminium a diminué de plus en plus, et par suite ce sont la kaolinite, resp. la méta-halloysite qui ont été précipitées favorisées aussi par l'augmentation des conditions d'oxydo-réduction et la diminution de celles du pH. Ainsi a été développé le faciès d'argilite caractérisé par la quantité dominante des minéraux d'argile, où la chamosite existe encore et les modifications du SiO_2 sont déjà apparues, précipitées d'une solution pauvre en aluminium et sursaturée en silicium.

En récapitulant les précédents, en allant du continent vers la mer, la proportion Al/Si de la solution montre une tendance de décroissance.

Cette conception est en accord avec l'opinion de E. SZÁDECZKY-KARDOSS, qui a montré que parmi les minéraux d'argile c'est la kaolinite dont la formation indique les conditions les plus oxydatives [44].

Lors de la synthétisation de la nontronite, H. HARDER [19] a déterminé, par essais, les domaines de stabilité des silicates de fer ferreux et du quartz (Fig. 13). Le domaine de stabilité de la nontronite est identique à celui de la chamosite à cette différence que près la chamosite précipite des solutions plus pauvres en silice.

La kaolinite primaire a été aussi formée dans la zone continentale. Puisque que Si présente le minimum de mobilité dans le domaine de pH entre 5 et 8, nous pensons qu'à la saison chaude et pluvieuse les oxy-hydroxydes d'aluminium ont été formés, et à la période plus froide et moins pluvieuse c'est pour la plupart la kaolinite qui se précipitait ainsi dans les dolines karstiques plus profondes et de temps en temps des matières plus riches ou plus pauvres en kaolinite se déposaient d'après l'alternance des périodes saisonnières.

7.2. *Génétique de bauxite régionale*

Le territoire le plus élevé et le plus oxydatif était la région de Ma Meo et Dong Dang, où la bauxite primaire et oxydée a été aussi formée, en grande masse. L'émersion pouvait commencer présumablement avant l'étage kazanien, car ici le mur a été érodé jusqu'aux couches permo-carbonifères les plus anciennes. Après l'érosion du calcaire, la formation de bauxite a commencé à la période tranquille, syngénétiquement à la karstification, en partie au continent en milieu oxydatif, en partie à proximité immédiate du littoral sous conditions, en prédominance, réductives à la zone périodi-

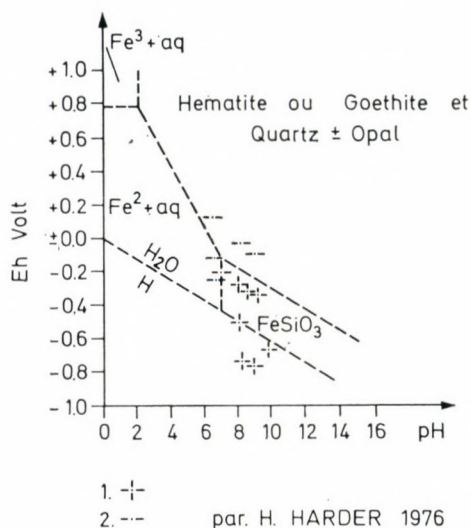


Fig. 13. Formation de nontronite en fonction de l'Eh et du pH. 1. essais à résultat positif, 2. essais l'Eh à résultat négatif

quement inondée par l'eau. Comme nous l'avons déjà mentionné préalablement, le toit pouvait être ici aussi le calcaire permien supérieur (étage tatar) qui plus tard a été complètement dissout sans traces, sauf les détritits siliceux menus. La bauxite déposée en différents faciès, en arrivant à la surface, a été désagrégée, plus ou moins oxydée et remaniée.

A la province Cao Bang, au gisement de Tap Na, à cause du tectonisme le pendage des couches de bauxite primaire est abrupt, la dissolution du calcaire du toit pouvait être partielle seulement. Ici les couches de bauxite primaire affleurent au dessous du toit.

Tam Lung présente un territoire intermédiaire entre le continent et les golfes marins couverts continuellement de l'eau, en voie d'isolation. Par rapport à Ma Meo, l'émersion a commencé plus tard (mur permien supérieur!) et a duré moins longtemps. A cause des changements géomorphologiques, sur ce territoire la matière produite sous faciès différents a été remaniée. C'est ainsi que la brèche de bauxite a été développée, dans laquelle les détritits de bauxite plus durs à diaspore et à chamosite ont été formés aux milieux plus réductifs, et le ciment plus meuble, plus riche en boehmite et kaolinite aux milieux plus oxydatifs. Par suite de la transgression la surface du complexe de bauxite, préalablement accumulée, est devenue marais qui a réduit secondairement surtout le ciment plus meuble.

La bauxite primaire de Ba Xa a été déposée dans les golfes marins en voie d'isolation, en milieu le plus réductif. L'accumulation de la bauxite se déroulait au cours de la régression caractérisable par les mouvements oscillatoires, où premièrement c'est l'argilite qui a été déposée et plus tard la

bauxite divisée par de laies de houille. Conformément à la succession génétique et sédimentaire esquissée dans le chapitre précédent, en allant vers la mer nous retrouvons les changements faciologiques aussi horizontalement en ce que la bauxite passe en argilite, puis celle-ci passe dans une formation de mer plus ouverte divisée par bancs de calcaire (Fig. 11).

7.3. *Processus d'oxydation*

Dans les précédents, on a déjà caractérisé en détail les changements qui déroulent tant dans la composition chimique que minéralogique au cours de l'oxydation et de l'effritement de la bauxite.

En conséquence le complexe secondaire est plus différencié par rapport à la bauxite primaire.

A la phase initiale, la teneur de la bauxite en ferro-chamosite devient instable, par suite la qualité de la bauxite s'améliore en dépendant de la teneur de la bauxite en ferro-chamosite. Une partie de la teneur en aluminium de la chamosite, à l'intérieur des détritits, précipite en boehmite secondaire ou en gibbsite, et l'autre partie, hormis les détritits, précipite en kaolinite, avec Si sortant en son entité, en solution. C'est le processus d'argilisation de la bauxite amenant le relâchement textural des détritits.

Par suite, les minéraux aluminifères s'enrichissent dans la bauxite, en sens tant relatif qu'absolu. Dans le type de gîte de Leo Cao — composé de détritits de bauxite grossiers à fraction d'argile relativement réduite — ce processus ne peut se réaliser en plus à cause de l'érosion continuelle: c'est pour cela que l'on trouve la bauxite à la meilleure qualité dans ce type gîtologique.

A la phase tardive, les détritits s'effritent de plus en plus et leur qualité abaisse progressivement. A côté de la kaolinite, l'argile encaissante contient toujours plus de minéraux aluminifères. Alors, ce processus produit l'homogénéisation. Dans le type de Dong Dang, où la matière chimique relâchée et physiquement effritée a été accumulée, c'est la nappe phréatique oscillatoire qui exprime son effet lessivant intense et dont le pH oscille aussi. L'acide huminique, produit par la végétation située au-dessus du complexe de bauxite, abaisse localement et successivement le pH au-dessous de 5, et par suite à cause de leur structure plus lâche les (oxy)-hydroxydes, d'aluminium secondairement et préalablement précipités seront dissouts. L'eau karstique mélangée avec de l'eau de la nappe phréatique neutralise vite la solution sortant des détritits et les aluminifères y formés améliorent, maintenant, la qualité des fractions d'argile. La teneur en fer des détritits sera aussi dissoute dont la partie plus considérable, ayant pour résultat une formation pareille à la cuirasse désertique, précipite à la surface des détritits. Ainsi, la qualité des détritits de bauxite abaisse, à la phase tardive, parallè-

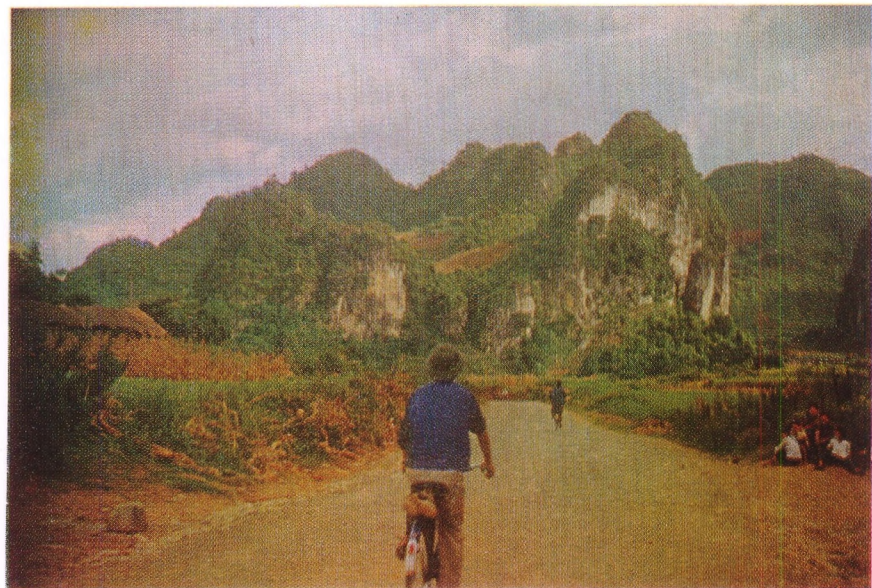


Photo 1. Les tours karstiques de Ma Meo

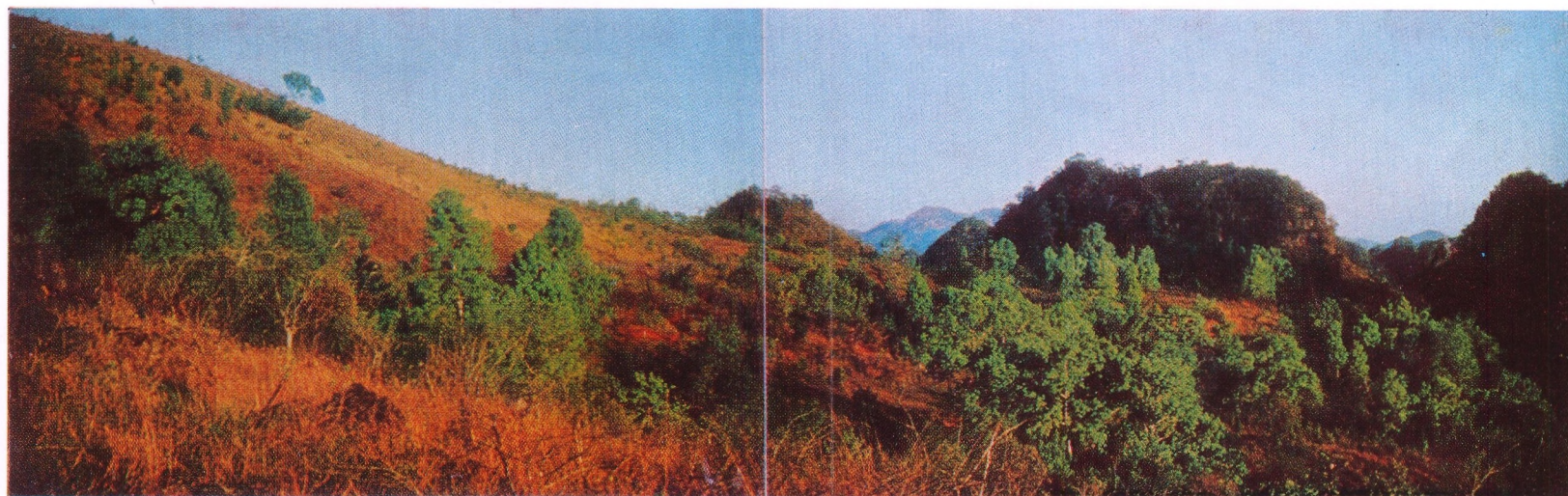


Photo 4. Type de Ma Meo: aux flancs des collines abrupts, le complexe de bauxite est en mouvement continu

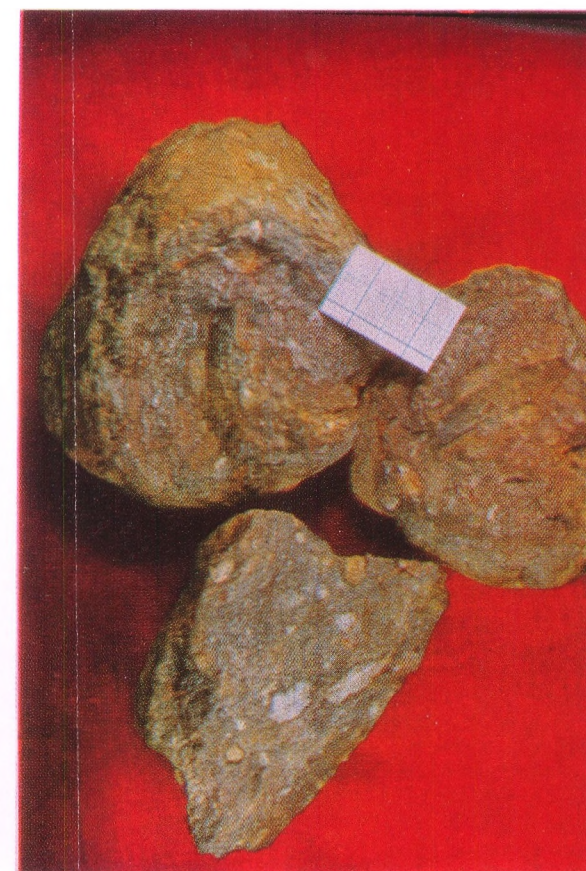


Photo 11. Gravier de rhyodacite latéritisés (Ma Meo: Ban Thau)



Photo 2. La bauxite en gros bancs, *in situ*, affleurant de dessous des couches du toit, altère en place et devient argileuse le long des fissures

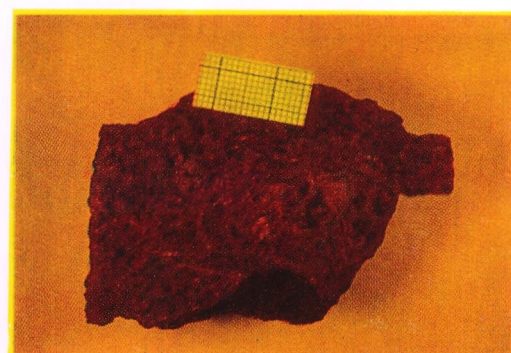


Photo 6. Bauxite primaire noire et bauxite verte à croûte oxydée rouge (Ba Xa)



Photo 7. Bloc de bauxite à texture scoriacée (Khon Pich)



Photo 8. Bauxite primaire oxydative (Ma Meo) à oxydation secondaire



Photo 9. Détritux oxydatifs primaires dans la bauxite primaire réductive (Tam Lung)

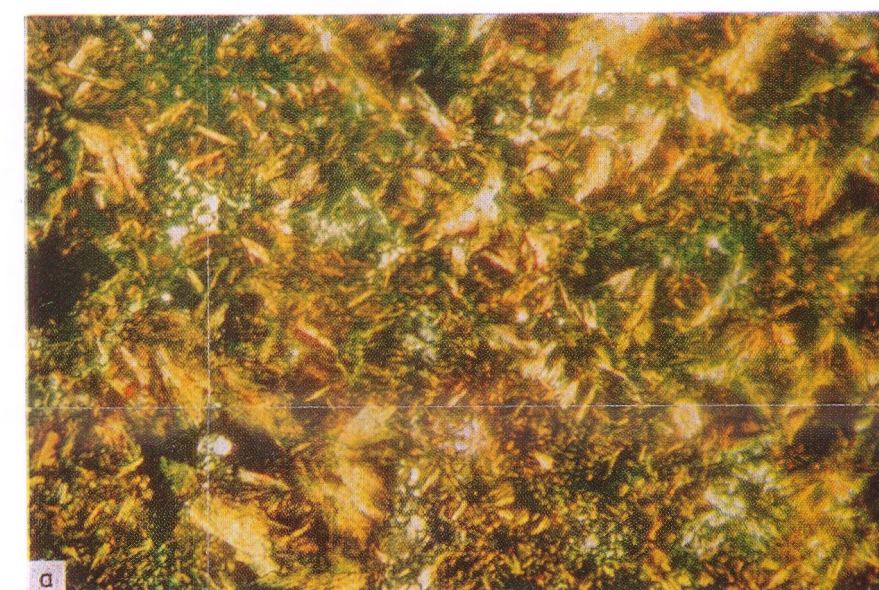


Photo 12. a-b. Aspect microscopique de la chamosite. Bauxite contenant de la chamosite (125 $\times$) et sa partie à grain fin (200 $\times$). Ba Xa

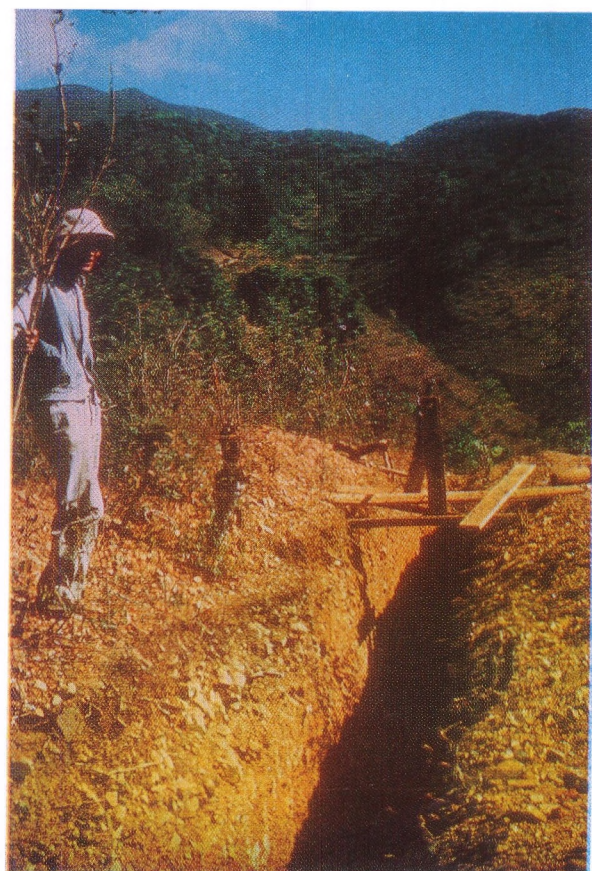


Photo 3. L'argilisation est plus intense vers le mur



Photo 5. La bauxite arrivée dès les flancs des collines s'étale en conglomérat aux fonds des vallées



Photo 10. Argilite (Ban Long), au milieu: strie de bauxite, au bord: oxydation secondaire



lement à la diminution du poids spécifique, et le silicium et le fer s'enrichissent relativement.

Il faut attirer l'attention à ce que lors les essais d'enrichissement, l'échelle de la détérioration qualitative, apparue avec la diminution de la granulométrie et du poids spécifique, n'est attribuable qu'en partie au processus ci-dessus. Selon notre avis, ici intervient aussi ce facteur de que lors l'effritement à cause de la surface spécifique augmentée, le nettoyage des grains à partir de l'argile devient de plus en plus difficile, c'est-à-dire malgré le criblage avec lavage à la surface des grains de bauxite reste de plus en plus de la «pollution» argileuse.

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МИНЕРАЛОГИЯ, ГЕОХИМИЯ И ГЕНЕТИКА БОКСИТОВ СЕВЕРНОГО ВЬЕТНАМА

ДЬ. КОМЛОШИ

Р е з ю м е

В 1972—75 гг. под геологическим руководством автора Проектное предприятие Аллюминиевой промышленности (АЛЮТЕРВ) проводил поисково-разведочные работы на бокситы в провинции Ланг-Сон Северного Вьетнама. В настоящей работе подводятся итоги разносторонних геологических наблюдений, а также детальных лабораторных анализов собранных в полевых условиях проб, исследованных в лабораториях и научно-исследовательских институтах ВНР.

В данном регионе Юго-Восточной Азии образовались палеозойские, диаспорово-шамозитовые, первичные бокситовые тела, резко отличающиеся от местнорождений Средиземноморской бокситоносной провинции как минералогически, так и по своим генетическим условиям. В неогеновое время эти рудные тела подверглись глинистому выветриванию, были окислены и переотложены. В результате этих процессов образовались промышленные вторичные бокситовые залежи, генетически близкие к бокситовым месторождениям, известным в Китае и на Урале.

Бокситовые месторождения Северного Вьетнама расположены вдоль главных структурных линий ССЗ-ЮЮВ простирания на краю южно-китайского микроконтинента, в условиях многоэпиклиналильной зоны.

Рассматриваемая территория с кембрийского периода по поздний триас представляла собою депрессию с более или менее длинными периодами поднятия. Нижний и верхний палеозой представлен прибрежными и пелагическими мелководными биогенными известняками. Единственный, доказываемый период бокситообразования имел место в верхней перми, между казанским и татарским веками. В этот период имело место частичное поднятие эпигенетического характера, сопровождавшееся колебательными движениями земной коры. Бокситы образовались частично в условиях суши, частично же в прибрежной зоне, периодически затоплявшейся морем частично же в морских заливах, изолированных от открытого моря. В сторону пелагического моря в более открытых морских условиях накапливались одновременно с бокситовыми телами аргиллиты, затем аргиллиты, прерываемые пакками известняков. В прибрежной зоне бокситовые скопления чередуются угленосными отложениями. Дальше от суши в направлении более открытого моря образовались своеобразные бокситовые фации, характеризующиеся чисто окислительными, окислительно-восстановительными и чисто восстановительными условиями, отражаемыми соответствующими минеральными парагенезами. Бокситовые залежи перекрыты трансгрессивными отложениями татарского яруса верхней перми, сложенными угленосными мергелями и известняками. В более молодой части кровли встречаются алевролиты и кислые вулканы в основном юрского возраста. Орогенез имел место в позднетриасовое время. Под влиянием интенсивного тектонизма более молодая и более древняя известняковые толщи были сильно разрушены, в них возникли чешуеобразные надвиги (впоследствии произошло закарстование), а триасовые алевролиты обнаруживают комбинированные складчато-разломные структуры.

С бокситовых тел, сохранившихся в относительно более спокойных, ненарушенных условиях, кровельные известняки были бесследно размыты. Вышедшая на дневную поверхность толща подвергалась местному выветриванию и аргиллизации. В зависимости от геоморфологических условий на участках с наибольшей отметкой земной поверхности толща подвергается постоянному размыву, на крутых склонах холмов она была переотложена, а в глубоких долинах произошло накопление переотложенных осадков. На таком основании можно выделить три обломочных типа залежей: типы Лео-Као, Ма-Мео и Донг-Данг. Раздроблению, переотложению и накоплению переотложенных осадков сопутствуют интенсивные процессы окисления, проявляющихся по-разному в отдельных типах залежей. В процессе окисления произошли глубокие изменения в минералогическом и химическом составе бокситов первичного склада, которые в начальном этапе привели к улучшению качества бокситового обломочного материала, а в позднем этапе к ухудшению качества руды.

GEOMECHANICAL INVESTIGATION OF THE SOUTHEASTERN MARGIN OF THE BAKONY MOUNTAINS AND THE AGE OF THE LITÉR FAULT LINE

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Relying on new geological data, the author refines his earlier observations concerning the fault system of the southeastern Bakony Mountains precisising the dates of the crustal movements and their mechanism as an extension of his former papers dealing with the geomechanics of this area.

The results of this paper can be summarized as follows:

- the main fault line of the southeastern margin of the Bakony Mountains unknown to detail till now is demonstrated by a characteristic geological profile,
- within each tectonic compression zones, in each longitudinal and transversal fault line zones faults of different ages join one another in different combination and variations,
- a reversed (downward turning) asymmetric wedge structure is also demonstrated,
- the rotation axis line of the basin platform was underthrust and tilted as a result of the horizontally acting orogenic forces on the southeastern margin of the basin: a process with which the author connects the upward migration of carbon dioxide between Székesfehérvár and the shore of Lake Balaton,
- the Moldavian orogenic phase consists of two phases,
- the age of formation of the Litér fault line is assigned to the Styrian movements,
- the asymmetric wedge structure is a common piled geomechanical form and it is in close relation with plate tectonics. Concerning the Bakony Mountains, this phenomenon denotes a secondary dispersion of the collision caused by the movements of large plates approaching to one another.

In 1968 the tectonic and geomechanical observations were published entitled "Tectonic theories in light of Bakony Mountains evidence" an extension of a publication of the author's similar observations concerning the Várpalotaregion (1956).

In both papers mentioned above the structure and the geomechanical explanation of the formation of the southeastern marginal main fault line of the Bakony Mountains was dealt with based on the information then available.

Two events known since 1968 have thrown a new light upon this interesting and important tectonic form of both national and probably even broader significance.

One of these events has been the publicity given to the most important results of recent development, i.e. to the theory on global tectonics or, in other words, plate tectonics. It seems to be evident to study the structure mentioned above also from the point of view of this new world-conquering theory.

Another significant event proved to be the drilling of numerous bores in the mobile strip of the structure in the Várpalota region which made essentially clear also in detail the character and nature of this structure developed during earth's history.

For a better understanding of the geological profile to be demonstrated below, the former statements will be summarized as follows:

The formation of the structure can be explained by the formation of an oblique fault plane caused by the horizontally acting crustal movements, along which the southern foreground was underthrust. At that time the background was tilted backwards to a certain extent (northwestwards), it was uplifted and in lack of suitable support it fell along the abrupt fault line. This fault line is a secondary one, thus younger than the underthrust side. The asymmetric wedge structure formed in this manner became to rise in order to ensure an isostatic equilibrium since underthrust causes a thickening of the crust (i.e. pile up). The uplift resulted in increasing erosion.

As a result of movements the crust was solidified and became resistant to subsequent movements. Thus, a new phase of movements occurred not along the existing displacement planes but it was displaced towards the less resistant foreground (foredeep). Consequently, the compression planes of each crustal movement phase move away from the mountains and become gradually younger. In case of movements of higher intensity new wedge structures develop during the outward movement. In case of movements of less intensity the existing wedge structure will be widened and this is caused by the occurrence of new and new underthrust planes in the subsequent phase moving towards the foredeep.

The Wedge Structure of Inota

In the underthrust strip of Inota deep bores were drilled in the past years under the guidance of the National Enterprise for Geological Exploration and Drilling (as key bores). The six bores were carried out within an area of 300 metres. Out of them that of Inota No. 138 proved to be most interesting and most profitable from the point of view of the profile in question. This was 800 m deep and was carried out by periodical core sampling accompanied by detailed geophysical investigations. The strata sequence of the bore shown in the profile has been compiled on the basis of core samples and well-logging results in collaboration with József Dorogi, senior geophysicist.

Having penetrated the Sarmation and thin Upper Badenian strata the bore traversed a transversal fault line of about 300 m height difference intersecting diagonally the geological profile. This was followed by the lowermost marine sediment sequence of the Upper Helvetian (Carpathian). This is underlain by the Lower Helvetian (Ottnangian) sequence with the fauna well-

known from the typical facies sections of the Ottnangian strata of Várpalota (J. KÓKAY, 1972, 1973). Downwards the bore traversed three underthrust zones in the manner demonstrated in the profile and at a depth of 741 m it reached the Middle Triassic (Ladinian) dolomite series.

The next key bore shown on the profile is the core drill of Inota No. 140 which was drilled at a distance of 180 m from the other and nearly north of it. This traversed the sequence demonstrated, too. It reached the Triassic dolomite already at 53 m depth, overlain by the Lower Badenian marine sequence.

Between the bores No. 138 and 140 there is the bore No. 136 being, however, not demonstrated and showed the same sequence, i.e. under the Lower Badenian the Upper Helvetian was found in a normal position.

On the basis of data the profile of Fig. 1. was constructed and from which the following statements can be concluded.

1. The outlined asymmetric wedge structure is crossed also by large-sized transversal faults. The fault of about NNW-SSE strike and of about 300 m shown here cannot be found in the Sárrét Basin and in the northern smaller basin, either. Thus, it is restricted only to the wedge structure. Its formation can be assigned to an intensive Moldavian phase, because this acted upon the Upper Badenian but left the Sarmatian, at least its lowermost part, untouched.

2. It can be stated that the underthrust planes become ever younger when moving away from the mountains.

3. In this structure between bores No. 138 and 140 the movement plane generated by the Main Styrian phase (Carpathian — Lower Badenian boundary) can be demonstrated. In accordance with the bore No. 136 lying between the two other bores the traversed Upper Helvetian sequence is not of littoral (schlier) facies, i.e. at that time the seashore was found about 200 to 300 metres north of here. The littoral sediments were eroded from the basement just after the main Styrian uplift and were replaced by the sediments of the Lower Badenian sea.

4. Movements relating to strong compression effects took place at the boundary of the Upper Badenian and Sarmatian, moreover these affected probably also the lower part of the Sarmatian sequence (Moldavian phase). All the three planes traversed by the bore No. 138 developed that time.

5. In this profile the Rhodanian (and subsequent Roumanian) movements started at the end of the Pannonian could not be exactly determined only another plane accompanied by smaller displacements is assumed in a southward direction. (On the wall of the former Upper Pannonian sand-pit small underthrusts could be observed, 1968, Fig. 3.).

6. The movement of the abrupt fault line lying on the rear of the wedge structure falls partly to the Sarmatian and mostly to the Pannonian and

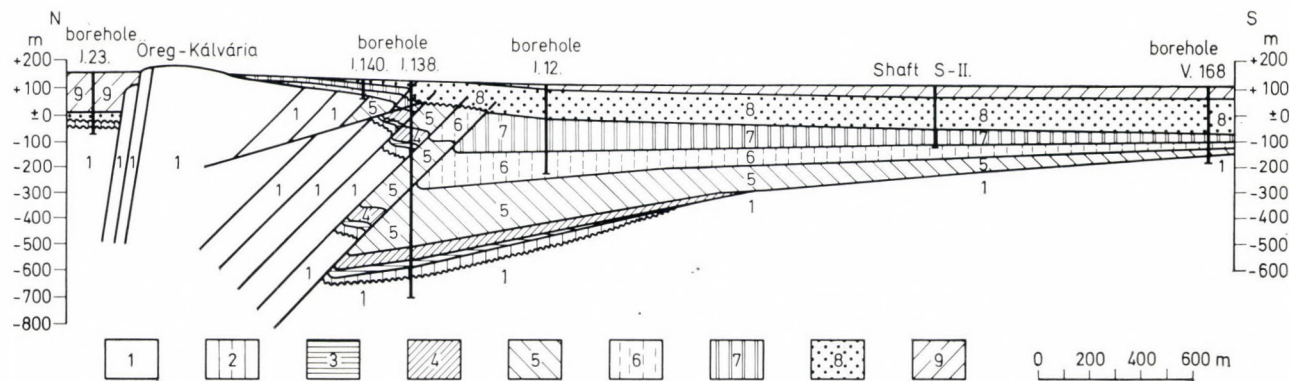


Fig. 1. Geological profile east of Várpalota. 1. Triassic; 2. Eocene; 3. Lower Miocene; 4. Lower Helvetian ("Ottományian"); 5. Upper Helvetian ("Carpathian"); 6. Lower Tortonian (Lower Badenian); 7. Upper Tortonian (Upper Badenian); 8. Sarmatian; 9. Pannonian

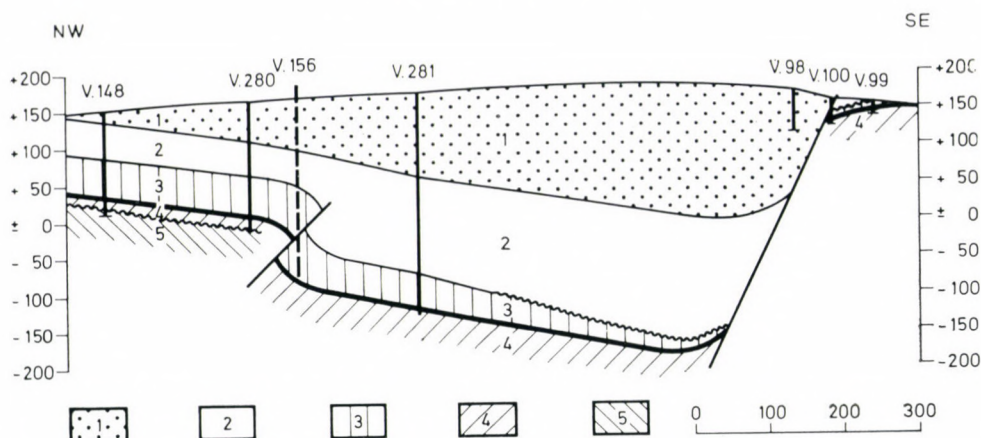


Fig. 2. Geological profile west of Várpalota. 1. Pannonian; 2. Sarmatian; 3. Upper Badenian; 4. Lower Badenian; 5. Carpathian

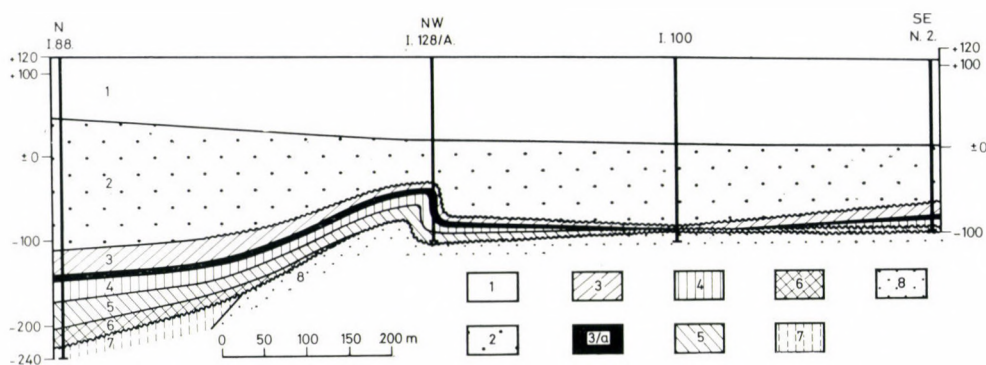


Fig. 3. Geological profile in the southeastern part of the brown coal basin of Várpalota. 1. Pannonian; 2. Sarmatian; 3. Upper Badenian; Brown coal bed; 4. Lower Badenian; 5. Carpathian; 6. Ottnangian; 7. Lower Triassic; 8. Permian

is characterized by synepiorogenic features other than by episodically occurring orogenic ones (1968).

7. An Early Styrian phase must also have occurred (the northernmost underthrust planes shown in the profile only as references), since after the deposition of the Lower Helvetian sediments a strong southward transgression can be observed in the Upper Helvetian throughout the Paratethys. This allows to conclude in favour of orogenic movements taken place at the Ottnangian — Carpathian boundary. Concerning these movements only a few data are available according to which in the area of Bántapuszta smaller erosion or abrasion unconformity is found at the Lower — Upper Helvetian boundary which indicates uplift (J. KÓKAY, 1967; Table 1., Fig. 1).

8. As a result of the series of Styrian and Rhodanian movements, the geosyncline of Sárrét of roughly NE-SW strike became narrower by about one km.

Fault Line Combinations

The most important faults of the Várpalota geosyncline were identified on the enclosed map. In general, it can be said that the oldest longitudinal faults lie in the north while the younger in the south making evident the theory of the movements of compression zones (Fig. 4).

Within each compression zone (J. KÓKAY, 1956) in each longitudinal fault line zone faults of different ages, i.e. different orogenic phases, join one another in different combinations and variations depending on the energy contents of each compression zone. Consequently, within the neighbouring compression zone the orogenic phases of the same age do not occur frontally, but show a mosaic-like displacement and are grouped in different combinations.

When analyzing this condition in detail, the following picture can be obtained:

The main compression zone of Várpalota is delimited by transversal faults between the horst-like Triassic dolomite range of Inota and the Triassic area of Pét. Its northern margin is bordered by the Higher Bakony, this being also the limit of the one-time range of the Eocene and Middle Miocene. On the basis of data and geological considerations the formation of the marginal fault line system lying at that place (underthrust zone) is due to the Pyrenean — Early Styrian — Main Styrian phases, but the latest Styrian phase (i.e. the Moldavian) did not occur here.

South of here, also in this compression zone, the main fault line of Várpalota (the so-called Lonsos Fault) is known since earlier times (K. TELEGI-ROTH, 1924; J. KÓKAY, 1956) and this of "normal fault" character. Its formation is connected to the Rhodanian phase started in the Late Pannonian and is accompanied by a throw of about 300 to 400 metres. Farther north, by about 300 to 500 metres and roughly parallel with it another fault line is found covered by Sarmatian and Pannonian strata as has been demonstrated in the geological profile west of Várpalota (Fig. 2). Southeast of this probable underthrust line the Sarmatian sequence suddenly thickens and in the borehole drilled in the assumed zone (V. No. 156) the Upper Badenian diatomaceous earth shows a steep dip (70 to 80°). Thus, the time of its formation can be assigned to the Badenian — Sarmatian boundary, i.e. it is connected with the Moldavian orogenic phase. It can be exactly seen on the profile that this is also an asymmetric wedge structure, but unlike that of Inota it is overturned downwards. Accordingly, in the asymmetric wedge structure of Inota the compressive underthrust limb lies in the south and the

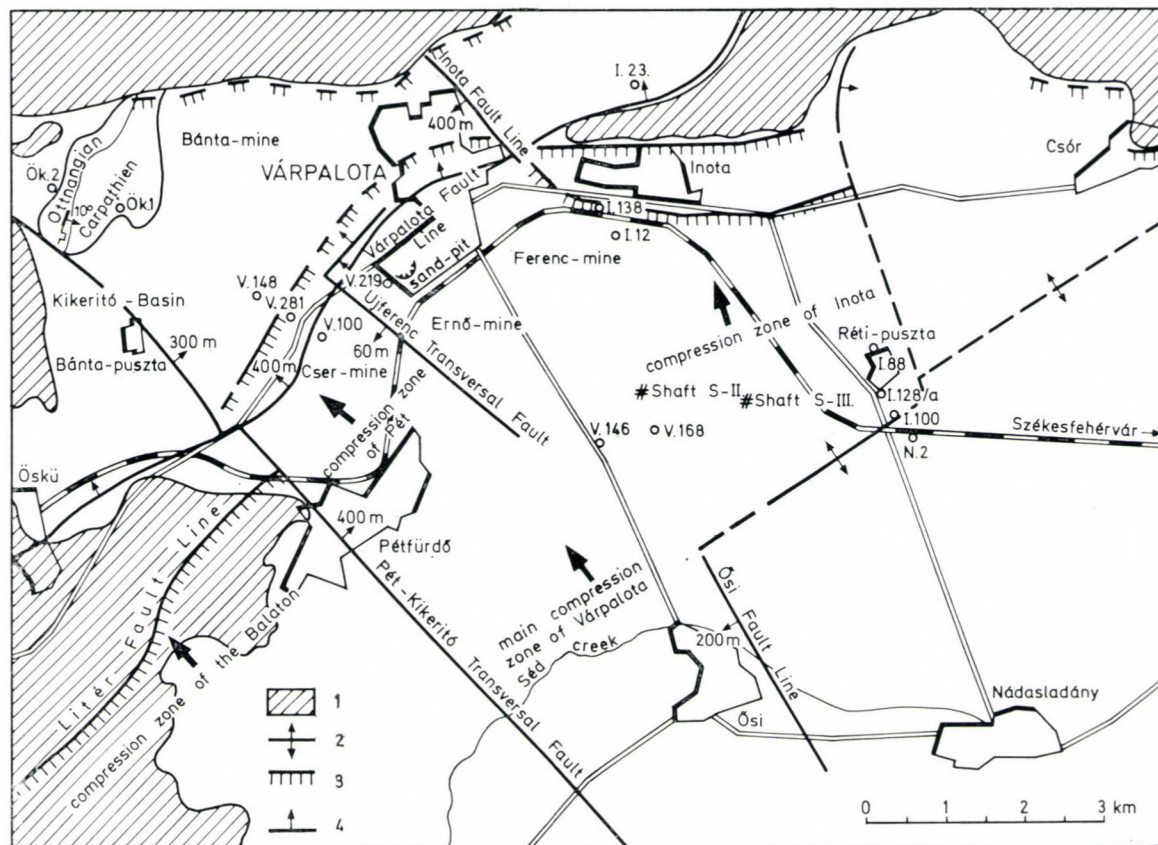


Fig. 4. 1. Triassic basement; 2. Rotation axis; 3. Underthrust; 4. Fault

younger "normal fault" in the north, while west of the town this is exactly reversed. The former statement, however, is true in any case (1968) that the "normal fault" limb is always younger than the underthrust one.

In the Inota strip lying east of the Várpalota main compression zone different orogenic phase combinations can be shown within each longitudinal fault line systems. The northern margin was produced by the Pyrenean movements (1956). Younger movements occurring in the western main compression zone cannot be identified here. On the basis of the geological section analyzed in detail, the northern margin of the south-lying horst-like dolomite range of Inota is of Rhodanian, while the southern part is of Early Styrian — Main Styrian — Moldavian — Rhodanian phase.

The previous statement is valid also to the transversal faults. In the individual fault line zones event of different ages join each other (fault steps) in rather different groups and variations. These transversal faults lie on the borders of different compression zones, i.e. these are the slip planes of orogenic movements acting horizontally from the southeast.

The western border of the main compression zone of Várpalota is the large transversal fault of Pét-Kikeritő which developed during the Early Styrian — Main Styrian — Moldavian — Rhodanian orogenic phases (the latter one only in the belt of Pét). The fault line near Inota and Ósi indicating the eastern border of the belt and characterized by a throw of 200 to 400 m forming the Early Styrian — Main Styrian and Moldavian a step-series of southeasterly direction at the surface of the basement.

Within the main compression zone of Várpalota (in the "Cseri Mine Field") transversal faults generated by the Moldavian movements can be identified (1956). The so-called "Újferenc Transversal Fault" known in detail from exploratory bores and mining operations is of Rhodanian origin. This fault forms the eastern border of the smaller compression zone of Pét. It is characteristic of this fault line of also Rhodanian origin, they mutually determine each other (J. KÓKAY, 1968).

Rotation Axis and Carbon Dioxide Line

On the basis of the data of the southeastern part of the brown coal basin of Várpalota a strong flexure is shown in the profile recorded near the railway station of Csőr-Nádasladány. This was traversed by the borehole Inota No. 128/a in the heavily thinning Miocene strata characteristic of the southern margin. It is also obvious that the surface of the Permian-Lower Triassic basement reached in several bores rather slight dips from the bore Inota No. 100 towards the bore Nádasladány No. 2 while it dips steeper towards the bore Inota No. 88. At the same time in the zone of the bore Inota

No. 100 the Miocene sequence underlying the Sarmatian is rather incomplete and thinning. [More concretely: directly under the Sarmatian sequence the lowermost part of the Upper Badenian diatomaceous clay marl was traversed but the coal bed was lacking and did not form, either; this directly overlies the thin remnants of the Upper Helvetian (Carpathian) marine sedimentary sequence. The Lower Badenian and the Lower Helvetian (Ottangian) are lacking.]

When evaluating the profile and the surrounding bores together, the following statements can be concluded:

The zone of the bores Inota No. 100 and 128/A proved to be a belt of a ridge uplifted like an anticline during the Miocene on the southeastern margin of the basin. Only a discontinuous sedimentation took place on this ridge, resp. denudation was frequent during the subsequent periods. The existence of the ridge can be unambiguously explained from the geomechanical point of view. *This belt is the rotation axis line of the underthrust and tilted basin plate on the southeastern margin of the basin and was generated by horizontally acting orogenic forces.*

In the marginal region of the southeastern Bakony Mountains the Moldavian phase proved to be the strongest orogenic period (1956, 1968). Thus, it is obvious that on the margin along the basin the flexure was generated by the strong horizontal orogenic impulse (demonstrated in the profile) and actually replaces an underthrust and is also accompanied by tensional deformations. This flexure developed really in the second part of the Moldavian phase. In the first part of this phase only denudation of limited extent took place as a result of the uplift. The second stronger period acted only on the denudated formations.

In the strip of the rotation axis line (the ridge) CO₂ gas explosion was generated by several exploratory bores. The geomechanical explanation of this is as follows: the zone of the rotation axis line is a zone heavily affected by tensile and flexing forces accompanied with far-reaching lithoclase system which made possible the upward migration of CO₂ generated probably by thermal dissociation.

This carbon dioxide migration is not a local phenomenon but it is a part of the carbon dioxide line lying between Székesfehérvár and the shore of Lake Balaton. The origin of carbon dioxide explained by this geomechanical theory is much more obvious than according to an eventual basaltic post-volcanic phenomenon.

In the walls of lithoclasts found in the brown coal bed of the Upper Badenian of Várpalota bluish white hydroquartzite (chalcedony) coat precipitating from the thermal waters can be observed (1956, p. 26). These phenomena relating to hot spring activity indicate a strong tectonic effect and the formation of faults reaching greater depths.

According to the decrepitation investigations of I. Csillag the precipitation of chalcedony took place at relatively high temperatures (probably over 200 °C). The fact whether the hydrothermal formation is assignable to the Moldavian or to the Rhodanian phase, is not decided yet.

Type and Age of the Litér Fault

When studying the geological maps of the Várpalota Basin and of the Balaton Highland it can be easily recognized that the tectonic structure of Inota-Várpalota can be traced westwards, as well.

The Várpalota main fault line of 200 to 400 metres height of throw connected with the Rhodanian phase and dividing into two parts the Várpalota Basin is the "normal fault" limb of the young asymmetric wedge structure crossing the great transversal fault of Pét-Kikerítő and extending farther along the southern margin of the Bántapuszta Basin and then continuing on the southern side of the villages Őskü and Hajmáskér up to the Kádárta-Veszprém region on the southern margin of the Hajmáskér-Gyulafirátót Basin. The Gyulafirátót Basin and the southern part of the Bántapuszta Basin contain a terrestrial sedimentary sequence with interbedded Sarmatian bentonite layers. As communicated orally by Á. JÁMBOR, this statement is supported also by the data of a new survey drill. The fault line joins the Sarmatian tectonically. This fact relates to the Rhodanian origin of the movement.

Tracing further the "underthrust" limb, one can see that west of the transversal fault of Pét the Litér Fault line reaches the area of the Triassic blocks of Pét. In the relatively well-exposed Triassic area this can be traced, even at the surface, south-westwards towards Litér village.

The compression line is nearly of southwest direction while the Hajmáskér limb ("normal fault") of the asymmetric side is of nearly westerly strike, thus the structure grows gradually wider. Between Hajmáskér and Litér the wedge structure is about 4 km wide, while in the region of Inota only about 1 km. In the wedge structure of Inota, drags taken place along compression planes are less significant, while along the Litér Fault Line regular thrust-sheets, drags, steep or overturned dips can be observed. Consequently, the asymmetric wedge structures of both Inota-Várpalota and Hajmáskér-Litér are geomechanically identical, the only difference being the fact that the tectonic compression zone of the Balaton Highland endured horizontally acting crustal movements of greater intensity than that of Inota and especially than that of Várpalota. The dividing line between the tectonic compression zones of Várpalota (southeastern Bakony Mountains) and of the Balaton Highland is the transversal fault of Pét-Kikerítő.

In accordance with the facts quoted above it is obvious that the Hajmáskér-Litér structure is the continuation of that of Inota-Várpalota, in fact,

the "normal fault" limb is also Rhodanian as testified to by direct evidence. Consequently, the time of formation of the underthrust line of Litér should be assigned also the Styrian phases including the "latest Styrian" (Moldavian) phase. It is not unimaginable that the fault started as early as the Pyrenean phase (when extrapolated from the Várpalota Basin), but a Cretaceous start looks impossible owing to the outward movement of the compression zones (1968). Thus, the older compression zones lie to the north.

On the southeastern margin of the Várpalota Basin when extrapolating both the rotation axis demonstrated in the foregoing part and the flexure replacing the underthrust, to the tectonic pressure zone of the Balaton Highland generated by stronger effects, it seems to be probable that the fault line of the Lake Balaton shore formed in the first phase also during the Moldavian. Here, however, a real underthrust developed instead of flexure of 50 m. This assumption is supported by the fact that in the Balaton region (Tihany, Udvari, Révfülöp) the Sarmatian sea transgressed onto older formations, as a result of the epeirogenic movements that followed the orogeny.

On the basis of oral communication by I. SZABÓ and of other data it is also probable that on the region Lake Balaton shore also Rhodanian compression movements took place, because locally (i.e. at Alsóörs) some folding of the Upper Pannonian strata can be observed.

Taking into consideration the facts mentioned above, the tectonic picture of the Balaton Highland can be solved in a much easier manner than by the sophisticated "air-saddled" tectonics frequently referred to by earlier workers.

General Frontal Piling Up

The term "underthrust" is used (1968) instead of the overthrust or reversed fault terms thus far in general use, since this is the exact definition of the horizontally acting movement from the point of view of its direction. The active orogenic force acted undoubtedly from SE since the southern side of the Bakony Mountains of counter-limb structure (Balaton Highland) is more uplifted, and endured stronger tectonics, while the reactive side (the Bakony of Pápa) is less disturbed. (It follows from the facts above that in the latter case really an "overthrust" has to be done with, in reality!)

The term underthrust is used here essentially in the sense of "subduction" used in plate tectonics. Thus, in connection with the geomechanical mechanism of the genesis of the Bakony Mountains an active subduction limb (e.g. the Várpalota-side) and reactive subduction (thus, as the margin at Pápa) can be spoken of.

Accordingly, the term "subduction" is used here to denote a geomechanical structure form. In the case of the formation of the Bakony Mountains

not a typical oceanic underthrust should be imagined along the southern margin, but this is the secondary dispersion of the collision generated by the approaching movement of the Eurasian and African plates of the lithosphere (STEGENA, 1972) which occurs along the farther-lying mobile strips. Independently whether this be a real "subduction" or not or a "disperse subduction" similar to that discussed above, these produce geomechanical structure forms of close affinity. Consequently, the resulting structures genetically similar to one another but different in their dynamism which can be traced back to horizontal effects producing the underthrust.

This third- or fourth-order subduction (restrained system) may be regarded to be one step of the Igal-Bükk subduction zone (E. SZÁDECZKY-KARDOSS, 1971).

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ГЕОМЕХАНИЧЕСКИЕ ИССЛЕДОВАНИЯ ЮГО-ВОСТОЧНОЙ ОКРАИНЫ ГОР
БАКОНЬ И ВОЗРАСТ ЛИТЕРСКОЙ ЛИНИИ РАЗЛОМОВ

Й. КОКАИ

Р е з ю м е

Продолжая свои, ранее опубликованные работы по геомеханике рассматриваемого района, автор статьи на основании новых геологических данных уточняет свои наблюдения за системой линий разломов на юго-восточной окраине гор Баконь, а также возраст и механизм движений земной коры.

Результаты исследований сводятся к следующему:

- На характерном геологическом разрезе показывается подробно еще не изученная, юго-восточная, главная краевая линия разломов гор Баконь.
- Внутри отдельных зон тектонических усилий сжатия, в отдельных зонах продольных и поперечных разломов сопряжены между собой разновозрастные разломы в виде различных сочетаний и вариаций.
- Кроме того, показывается также и обратная (перевернутая вниз), ассиметричная, клинообразная структура.
- Дается характеристика той осевой линии вращения бассейновой плиты, подвинутой под чехлом и опрокинутой в результате горизонтальных орогенических усилий на юго-восточной окраине бассейна, которой автор статьи приписывает и восходящий поток углекислоты на участке между г. Секешфехервар и побережьем оз. Балатон.
- Молдавская тектоническая фаза делится на две части.
- Время формирования известной Литерской линии разломов автор относит в основном к штирийским движениям складчатости.
- Описывается ассиметричная, клинообразная структура, являющаяся фронтальным геомеханическим сооружением, обусловленным усилиями типа сжатия и тесно связанным с тектоникой плит. В случае гор Баконь речь идет о вторичной дисперсии процессов столкновения, обусловленных сближением крупных плит.

OUTLINE OF THE GEOLOGY OF THE GRAND-DUCHÉ DE LUXEMBOURG

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Two distinct geologic-geomorphologic regions are found in Luxembourg. To the N is the Oesling, a region of highly disturbed Palaeozoics, extensive plateaux and deep, narrow, winding valleys, representing a post-Hercynian peneplain uplifted in the terminal Neogene. To the S lies the Gutland, where Mesozoics outcrop, topography is more varied, valleys more open and younger, faulting more significant. The Oesling depression, marking the boundary of these two areas, collects the drainage which exits eastward.

Eodevonian, Triassic and L.-M. Jurassic outcrop extensively, but the Tertiary has only small representation. The stratigraphic column incorporates some 4300 m of outcropping sediments.

Tectonic features all follow the Variscan (NE-SW) or Rhenic (NNE-SSW) trend, being influenced by Caledonian and Hercynian events. Large folds are but parts of structures which continue into Belgium and/or Germany. As the Moselle Trough is approached, faulting of the Mesozoics is marked, fracturing extending into Quaternary times.

The iron deposits in the SW gave rise to the iron-steel industry. Today only the Aalenian minette deposits are worked, though the industry began with exploitation of Tertiary deposits. At present, some 80% of iron used must be imported.

General

The small country of the Grand-Duché de Luxembourg occupies a geographical position of historical importance. Geologically no less, the location is significant, with the Ardenne massif to the N, the Eifel—Lorraine depression striking NNE-SSW through the country, the Luxembourg Gulf of Mesozoics trending NE across the southern part, highly disturbed Palaeozoics in the N unconformably overlain by more gently deformed Mesozoics in the S, thus giving rise to the two major geomorphologic-geologic divisions of the country. Oesling in the N and Gutland in the S. Luxembourg thus lies at the transition of the Mesozoic plains of Lorraine and the Palaeozoic plateaux of Western Germany, between the Ardenne Palaeozoic massif to the N and the Saar Basin Palaeozoics and Lorraine Mesozoics to the S (Fig. 1).

The first important contribution to the geology of Luxembourg is that of STEININGER (1828). The bibliography shows the earlier studies, both classical and important, as a matter of general interest. However it is only relatively recently that systematic, detailed mapping was undertaken, resulting in eight

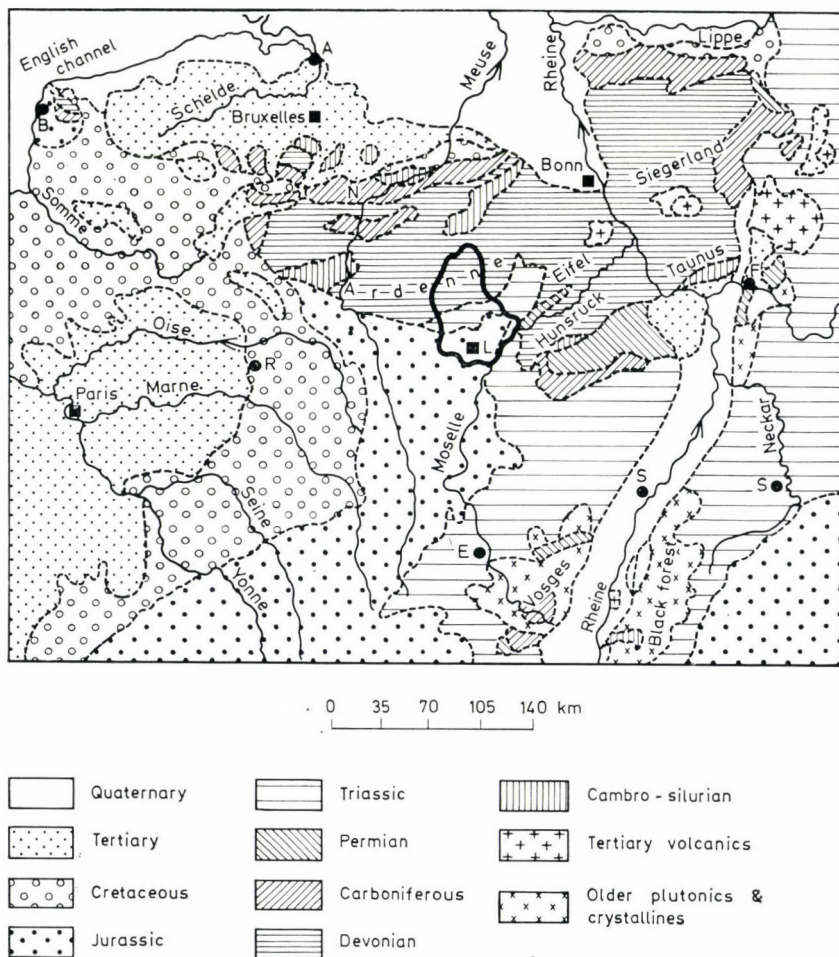


Fig. 1. Stratigraphic Environment of the G-D. de Luxembourg. 1. Quaternary; 2. Tertiary; 3. Cretaceous; 4. Jurassic; 5. Triassic; 6. Permian; 7. Carboniferous; 8. Devonian; 9. Cambro-Silurian; 10. Tertiary volcanics; 11. Older plutonics and cristallines

geological sheets, scales 1 : 25.000 and 1 : 50.000, and six tomes, appearing during 1937–1950. This most worthy contribution was the work of the late MICHEL LUCIUS, who organized the Service Géologique de Luxembourg in 1936.

Upon the appearance of newer, more precise topographic sheets of the country in 1954, it was decided in 1964 to begin a revision of the LUCIUS maps. The many new building and constructional projects, engineering and water borings, quarry developments, etc. no less presented an opportunity for further subsurface investigations. It is intended to publish sixteen new geological sheets on a scale of 1 : 20.000, of which three are at present complete.

In 1966 the Service Géologique published a geological map of Luxembourg on the scale 1 : 100.000.

Luxembourg occupies 2587 km², 82 km in length, 54 km in breadth. The population is approx. 357.000 of which some 24% are foreigners.

Geomorphology

Elevations vary from 130 m at Wasserbillig, where the Sûre river joins the Moselle, to 559 m at Buurgplaatz in the N very close to the Belgian border.

The Gutland or southern part of the country, where agriculture, communications, habitations are more extensively developed, has an average elevation of 300 m. In the N lies the Oesling, part of the Ardenne, average elevation 450 m, with wide, open plateaux transected by deep, narrow, winding valleys.

The Oesling represents a peneplain which has been upraised and disturbed by epeirogenic movements in recent geological times. First Caledonian then Hercynian orogeny created the Ardenne, and as rivers set to work here, elevations were lowered, slopes reduced, deposits spread forth and filled-in surface irregularities, so creating the post-Hercynian or pre-Triassic peneplain, formed at the base of the Permian and Triassic. Beginning in the Triassic and extending throughout the Jurassic, transgressive seas invaded this Eifel—Lorraine depression, forming sediments on the earlier-formed peneplain surface. Regression, followed by continental conditions, partially exhumed the old peneplain, forming now the pre-Senonian peneplain. In the U. Cretaceous this peneplain too was immersed, again in the Eocene erosion set to work, forming the Eocene peneplain, followed by an Oligocene transgression, the last to affect the Ardenne. We thus have pre-Triassic, L. Cretaceous, Eocene and post-Oligocene peneplains produced, but the Tertiary continental erosion periods scarce effected the earliest pre-Triassic peneplain. However, since the terminal Tertiary this early peneplain underwent changes in altitude and deformation more extensive than any previously experienced. Differential movements exhumed the first peneplain, creating “waves” of great radius of curvature, clearly outlining the peneplain and its aureole of Mesozoic sediments, and causing profound hydrographic changes. Thus the essential geomorphological feature of the Oesling is the old, exposed pre-Triassic peneplain, which in more recent times was uplifted and deformed.

EoDevonian is exposed in the Oesling, occurs at depth in the Gutland. Mesozoics once covered the Oesling but were stripped away during the U. Pliocene—Pleistocene, and hence before this event there was little geological-geomorphological differentiation between the two regions. Towards the end of the Pliocene, differential block movements took place, in the N the Mesozoics were eroded away, exposing the pre-Triassic peneplain, for here uplift was

more pronounced than in the S. This peneplain surface plunges abruptly (average slopes of 5–9%) under the southern Mesozoics, the buried peneplain levelling off to slopes of 1–2% southwards. From the late Pliocene, erosion planed off a surface at ca. 500 m elevation in the Oesling, coinciding more or less with the pre-Triassic peneplain. In the Gutland, an erosion surface of some 400 m altitude can be established, formed during the same time interval.

Contrary to geographical opinions, in Luxembourg we do not have two planation levels; the higher and lower levels unite in the peripheral depression by means of a warp or flexure, of tectonic origin, caused through differential block movement at the southern edge of the Oesling. There is thus continuity in time and space of this erosion surface throughout the country, the northern block having undergone greater uplift than the southern one. This peneplain surface is excellently and extensively displayed in the Oesling, but the same surface at lower elevations in the Gutland is much more sporadic in occurrence, less clearly defined.

In pre-Neogene times, the general slope of the terrain was towards the N, hence major streams flowed in the same direction. Over a low plain, with lakes and stagnant stretches of water, Miocene rivers developed further the primitive antecedent drainage network over what is now Luxembourg. During a period little short of some 30 million years, the network evolved, rivers entrenching themselves, and in essence, this network is what we see today, originally established over the planated Ardenne massif and peripheral area to the S, where Mesozoics and Tertiaries cover a Devonian basement. But then at the end of the Pliocene there began en bloc uplift of both massif and foreland to different degree so that warping occurred between the higher Oesling and lower Gutland. Such tectonism had a profound effect upon the drainage network. Some streams, such as the Ourthe and its tributaries the Salm, Amblève and Warche in Belgium were “decapitated” and their upper courses forced to reverse direction. Greater uplift of the massif thus initiated reversed streams over the Oesling which now flowed southwards. Some streams, e.g. the Alzette, which initially continued northward across the Oesling, were now forced to turn eastwards at the peripheral depression, and so to the lower Sûre (Fig. 2). Fig. 3 clearly shows this gathering together of streams in the 10–12 km wide peripheral depression, and exiting eastwards. Larger, more powerful rivers such as the Meuse and Rhine managed to maintain their northward flow directions during uplift of the Hercynian massifs, hence are antecedent rivers, but even the Moselle was forced to change course and turn northeastwards.

In the Gutland, the post-Palaeozoics all dip to the SW, a consequence of late Pliocene epeirogenic movements. Here, however, streams were able to entrench themselves in this lesser uplifted region, so that no reversal of drainage occurred. Hence we have what LUCIUS termed “un réseau hydrographique

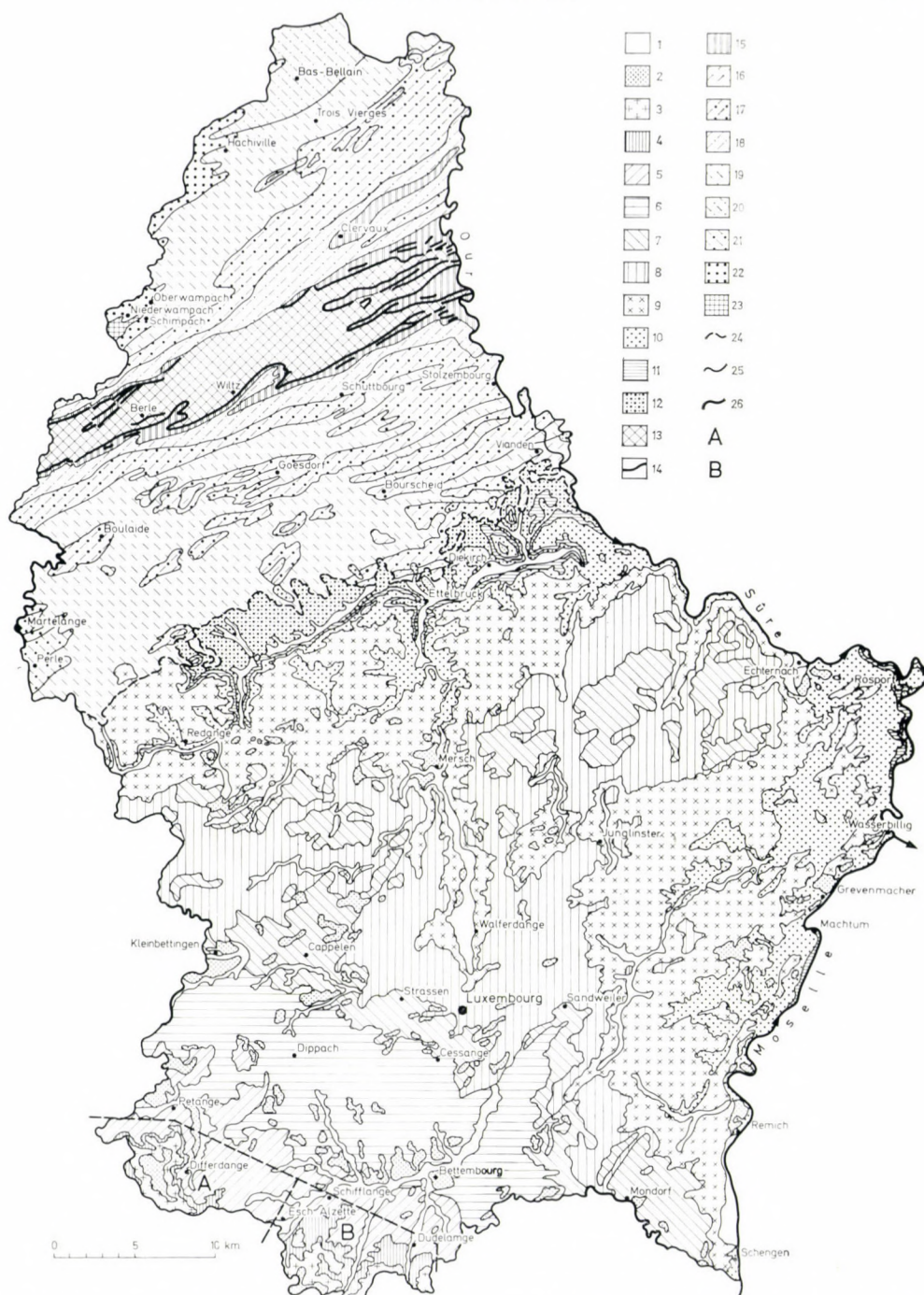


Fig. 2. Generalized Geologic Map of the G-D. de Luxembourg. (Modified after carte géol. gén. de G-D. de Lux. serv. Géol. de Lux. Échelle 1 : 100 000, 1966.) 1. Holocene (Alluvium); 2. Tertiary. *Jurassic*: 3. Bajocian; 4. Aalenian; 5. Toarcian; 6. Pliensbachian; 7. Sinemurian; 8. Hettangian. *Triassic*: 9. Keuper; 10. Muschelkalk; 11. L. Muschelkalk; 12. Buntsandstein. *Lower Devonian*: 13. Upper Emsian; 14. Berlé quartzite; 15. Middle Emsian; 16. Lower Emsian; 17. Stolzenbourg beds; 18. Schuttbourg beds; 19. Upper Siegenian; 20. Sandstone facies; 21. Slate facies; 22. Middle Siegenian; 23. Lower Siegenian. 24. Paleozoic mesozoic boundary; 25. Land frontier; 26. River frontier; A. Differdange iron basin; B. Esch iron basin

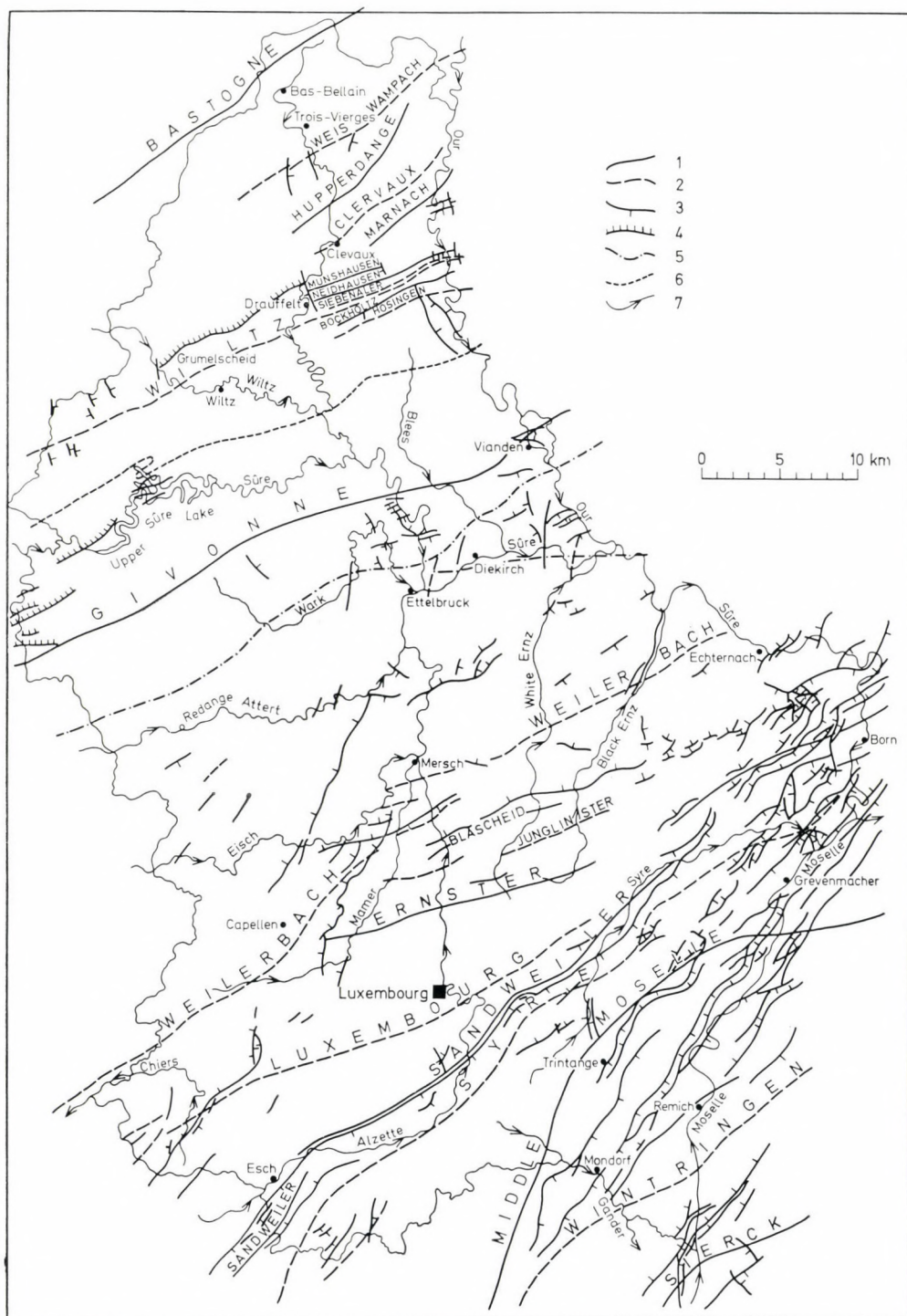
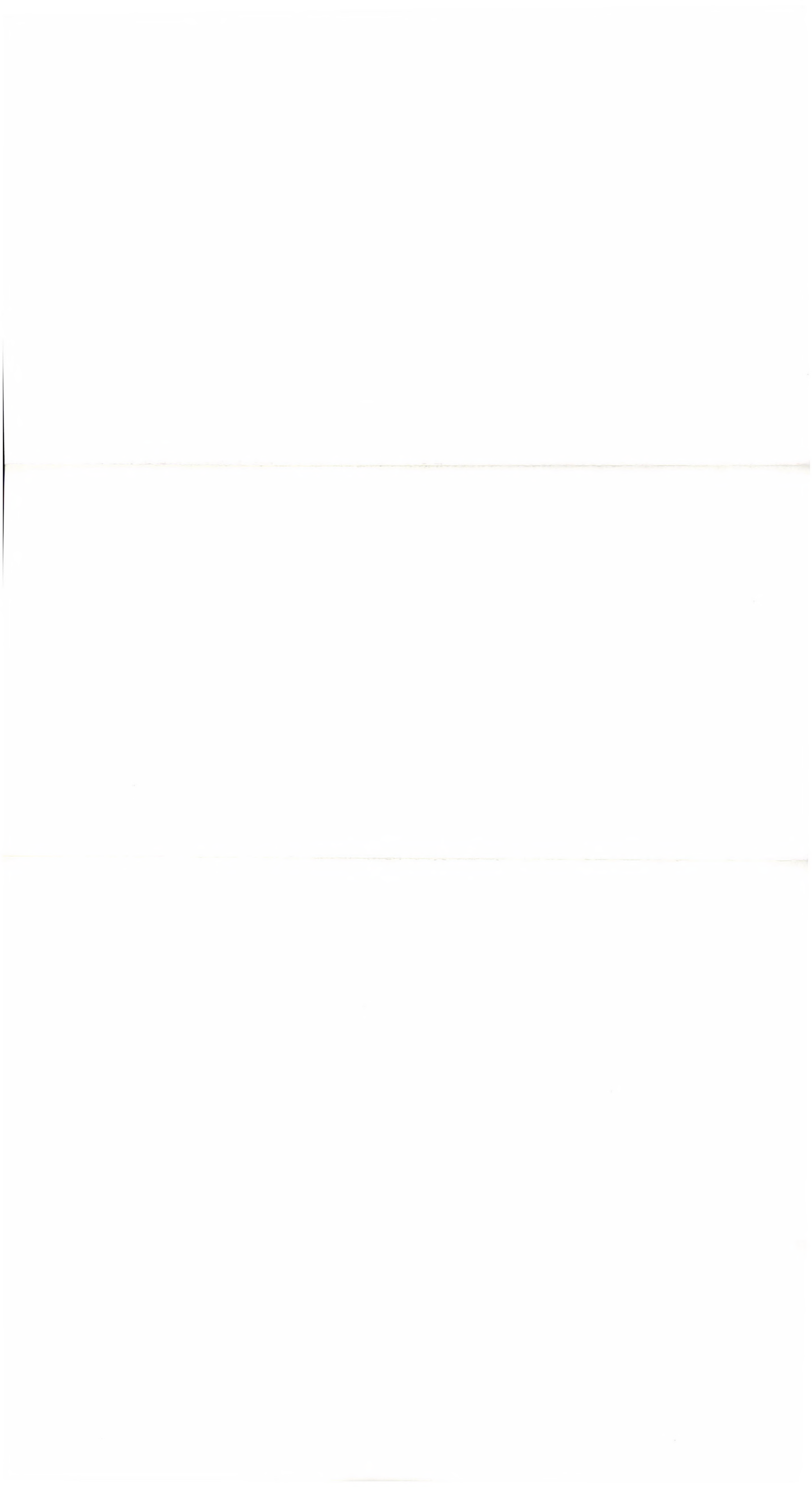


Fig. 3. Tectonic map of Luxembourg. (LUCIUS 1948, 1955, Ser. Géol. Luxembourg 1966).
 1. Anticlines; 2. Synclines; 3. Gravity faults-ticks on downthrow side; 4. Thrust faults-ticks on direction of thrusting; 5. Major flexure; 6. Limit of wiltz syncline and givonne anticline; 7. Rivers

Table I
Stratigraphic Column, Luxembourg

ERA	Period	Epoch	AGE		Formations			Thick. (m)	
Kainozoic	Quaternary	Holocene			Alluvium, calcareous sinter, rock falls, etc.			8	
		Pleistocene			Loam, sand, gravel, talus, terrace deposits, etc.			8	
	Tertiary	Pliocene — Mio- cene			Plateaux loams with pebbles, etc., Fe-ore concretions, silts with quartzitic concretions (Pierre de Stonne), clayey sands, clays, silts.			10	
		Eocene					Pisolitic Fe-ores		
Mesozoic	Jurassic	Dogger	Bajocian		Coral Limestone			30	
					Haut-Point Limestone			40	
					Sowerby Beds	Sonninia Beds	Ottange Lst.	4	
							Sonninia Lst.	14	
							Sonninia Cgl.	30	
					Micaceous Marls			12	
			Aalenian		Oolitic Iron Ores (Minette)	Concavum Beds		1	
						Murchisonae Beds		17	
						Opalinum Beds		20	
						Dumortieria Beds	Pseudoradiosa Beds	15	
							Levesquei Beds	18	
						Fallaciosum Beds			30
			Toarcian		Sriatulum Beds			20	
					Voltzia Marls			10	
					Crassum Marls			0.15	
					Posidonomya Beds	Bifrons Marls		20	
						Falciferum Bit. Marls		25	
		Pliensbachian		Domerian	Spinatus Beds (Dippach Sandstone)			30	
					Margaritatus Beds			80	
				Carixian	Davoei Limestones and Marls			5	
		Sinemurian		Upper of Lotharingian	Raricostatus Beds			3	
					Dudressieri Beds			6	
					Fossil-poor Marls (Fossilarme Tone)			25	
				Lower	Strassen Marls and Limestones			10	
		Hettangian		Luxembourg Sandstone			135		
				Psiloceras Beds			30		
	Triassic	Keuper	Rhaetic		Red Clays			8	
					Foliated Shales and Sandstones			15	
			Norian		Marly Keuper (Grès à marnolites, Steinmergelkeuper)			60	
					Red Gypsiferous Marls			10	
			Carnian		Reed Sandstones (Grès à roseaux)			40	
					Marls with salt pseudomorphs (Kauper salifère)			185	
					Muschelkalk		Upper	Myophoria Beds	
			Hauptmuschelkalk	Ceratites Beds				30	
				Crinoidal Dolomite				30	
			Middle	Anhydrite Group			Lingula Dolomite		8
							Mottled Gyp. Marls		60
			Lower	Muschelsandstein	Orbicularis Dol.		5		
		Shelly Limestone			45				
		Buntsandstein		Upper	Voltzia Sandstone			20	
					Transition Beds			85	
				Lower	Vosges Sandstone			152	
		Permian	Zechst-Rothlieg.				Shales, Sandstones, Dolomites		
Devonian	Lower		Emsian	Upper			Wiltz Shales/Schists		250
							Berlé Quartzite		15
				Middle			Clervaux Mottled Shales/Schists		300
							Lower	Schuttbourg Quartz-Phyllites	
				Stolzembourg Phyllites					
				Siegenian			Upper	Rambrouch, Ufflingen, Bas-Bellain, Kautenbach Shales/Schists, Martelange Slates	
			Middle				Martelange and Niederwampach-Hachiville Beds		600
			Lower				Phyllites, Quartz-Phyllites, Sandstones		150



qui est un paradoxe géographique", for Oesling and Gutland are drained by one and the same type of superimposed streams, one set flowing northwards, the other southwards, both unadapted to the present structure of either region. This hydrographic network is a heritage left by a rock covering which has disappeared as also a consequence of differential tectonic movements. In the Gutland, erosion has made compensation, whilst in the Oesling drainage reversal has resulted, and in both regions the network takes advantage of the peripheral depression to seek an exit.

Stream characteristics in the two regions differ. In the Oesling, streams have a younger, rejuvenated character, steep longitudinal and transverse profiles, flow is faster, they are deeply entrenched, meandering is very common, whilst above stretches the level plateau surface. In the Gutland, streams are more evolved, thalwegs are reduced, valleys more broad, rejuvenation not apparent, the flow is gentler. LUCIUS was inclined to believe that rock consistency had little if any control on stream development in Luxembourg, but we are inclined to agree with HEUERTZ (1949) that in the Oesling the rocks present a harder, more compact appearance than in the Gutland, that rock character and lithology is influential in valley development.

The studies of HEUERTZ (op. cit.) and STEFFEN (1951) regarding river profiles and terraces respectively, demonstrate the effects of differential uplift. The former study exemplifies the steep thalwegs of Oesling streams in contrast to much gentler ones of the Gutland. Excellent preserved river terraces occur in the Wark and Sûre valleys, less so in the Our valley. Three terraces can be defined, varying in elevations above mean stream level from 1 to 16 m, 16 to 62 m, 64 to 118 m, with uplifts of the order of 11–70 m, lowerings of ca. 10 m. In general the river terraces rise in elevation from SSE to NNW. As even the lower terraces show appreciable elevation, we can presume that slow uplift has continued into very recent times, possibly still operative.

Stratigraphy

The stratigraphic record includes Eodevonian, Triassic—Jurassic, small development of Palaeogene and Neogene (Fig. 2, Table I). In the Oesling, only Palaeozoics outcrop, in the Gutland, younger beds. The L. Devonian descends in depth southward, so that at Mondorf-les Bains on the French border, it lies 713 m below the surface, yet surprisingly enough, faulting brings a tiny Devonian outcrop to the surface in the Moselle valley at Schengen.

Palaeozoic

Only L. Devonian outcrops. In essence, this is a neritic facies, only exceptionally bathyal. Lithologically the rocks are either of arenaceous types — sandstones, quartzites, psammites — or then argillaceous types — shales, schists, phyllites, quartz-phyllites, slates.

Gedinian outcrops to the W in the Gedinne—St. Hubert—Libremont region of Belgium, forming part of the Eifel—Neufchâteau synclinorium which plunges eastwards so that in Luxembourg and Western Germany younger Devonian outcrops. Within Luxembourg only Siegenian and Emsian are exposed.

The L. Siegenian has small representation at Schimpach, in the upper Wiltz valley, and extending across into Belgium. Rocks in question comprise phyllites, quartz-phyllites and quartzose sandstones, forming a sharp anticlinal fold striking NE, some 150 m of rock being exposed. M. Siegenian occurs close to the Belgian border, occurring in two areas known as the Bastogne and Givonne “saddles” separated by an Emsian syncline. The southern exposures, in the vicinity of Martelange, show slightly foliated sandy shales-schists, sandy or clayey quartzphyllites, slates, psammites and quartzose sandstones. Northern occurrences show a greater preponderance of argillaceous rocks, usually tending to have a reddish or then dark blue tinge. No particular formational names are given the M. Siegenian, but we might refer to the southern occurrences as the Martelange beds, the northern ones, the Niederwampach—Hachiville beds. Thicknesses are not easy to judge but may be in the neighbourhood of 600 m.

The U. Siegenian has the greatest extent of outcrops of any Devonian strata, in areas to the N and S of the Wiltz Basin. In the northern area, sandstones are common, lesser so, sandy shales-schists and phyllites, forming a sandstone facies (Bas-Bellain shales-schists) and a more argillaceous facies (Ulflingen or Troisvierges shales). S of the Wiltz Basin there is a larger extent of U. Siegenian. Basal beds include fine-grained, compact phyllites of high fissility, which at Martelange and Perlé have been exploited for roofing slates. Above occur almost entirely dark-coloured, slightly fissile sandy shales-schists, known as the “Schiste grossier” or “Grobschiefer” facies. In the region of Martelange, the U. Siegenian is some 500 m thick, thickening northward to 1500 m at Bastogne, Belgium.

Before the canalization of the Moselle, small outcrops were to be seen in the bed of the river and also by the mineral well at Schengen, near the French—German borders, of fine-grained, reddish, micaceous, clayey sandstones with red clay inclusions, all having a siliceous cementation. These beds are correlated with the “Taunusstufe” of Sierck, across the Moselle in Germany, and are known in Luxembourg as the Sierck Quartzites. A thin basal conglomerate

rate marks the passage up into the Voltzia Sandstone. The top of the Eodevonian lies at some 700 m depth a Mondorf-les-Bains; ca. 10 km distant, faulting having brought these Palaeozoics to the surface. The Taunus Quartzite of Germany is considered of M.-U. Siegenian age.

In Luxembourg the Siegenian and Emsian strike NE across the country, exposures in dip directions covering some 35 km. Very often it is hazardous to estimate thicknesses of the Eodevonian, as the general lithology is monotonous, making the determination of limits uncertain, and then the strata are highly disturbed. FOURMARIER (1934) estimated that the Eodevonian in the Eifel basin was some 9000 m thick at the German border, whilst ASSELBERGHS (1946) gave a figure half this amount, and KNETSCH (1963) mentioned 5–8000 m in the Siegerland and Southern Ardenne. In Luxembourg Eodevonian has a thickness varying from ca. 2150 m to 2850 m.

Nowhere in the country are younger Palaeozoics outcropping, nor were they known at depth. Then during 1946–47 a boring was put down at Mondorf-les-Bains to a total depth of 730 m. Between depths of 692 m and 699.60 m occurred angular *débris* of shales and quartzose sandstones, with large quartzite pebbles at the base, dominance of sandstones higher up, cellular dolomite and red brecciated sandstone at the top. These poorly stratified sediments rest with marked angular unconformable relations on an Eodevonian abrasion surface. LUCIUS has classified the beds as Rothliegendes and Zechstein. (A 771 m deep boring at Longwy, some 40 km W of Mondorf-les-Bains, showed 231 m of Permian, beginning at a depth of 540 m.)

Mesozoic

S of the Oesling depression, Mesozoics extend throughout the Gutland—Triassic and L.-M. Jurassic.

The Buntsandstein or Grès bigarré is divided into lower and upper units. The former, Vosges Sandstone, in a strict sense the lower part of the basal Triassic sandstones, has no outcrops in the country, but at Mondorf-les-Bains the above-mentioned boring encountered the Sandstone at a depth of 589 m, where it measured 152 m thick. It is a brick-red rock, of angular quartz cemented by iron oxide, with small, flat pebbles of quartz and quartzite, but this basal conglomerate is only slightly developed. Above occur the Transition Beds (*Zwischenschichten*), of reddish and brown colour, coarse-grained sandstones with intercalations of quartzose gravels and dolomite nodules, interbedded with purplish dolomitic marls. At the top occurs the Voltzia Sandstone, of reddish hue, with thin interlayering of micaceous grey clays. The chief exposures are in the Attert-lower Sûre valleys.

The Buntsandstein measures some 260 m in thickness.

The Muschelkalk has four groupings. Lowest is the Muschelsandstein which comprises, at the base, shelly, clayey dolomitic sandstones and mottled marls. Above comes a 5 m thick bed of dolomite, containing *Myophoria orbicularis*, elsewhere rare in Luxembourg.

The second group forms the Anhydrite Group. Below are mottled marls with thin intercalations of dolomite and hard marls. Frequent lentils of exploitable gypsum occur in the mottled marls. To date no fossils have been found. Above are thin beds of dolomite containing fish scales and *Lingula tenuissima*.

There follows the Hauptmuschelkalk, 60 m of dolomites and limestones. The lower part, Crinoidal Dolomite, in places incorporates much crinoidal débris and has a crystalline or pseudo-oolitic texture. Higher up crinoids disappear, the rocks become more sandy and tabular, containing the very rare fossil (for Luxembourg), *Ceratites nodusus*. Glauconite grains are irregularly disseminated throughout the section, especially the upper part, whilst thin layers of clayey marls are common in the lower part.

Youngest Muschelkalk are the *Myophoria goldfussi* beds, with thick dolomites at the base overlain by an alternation of greyish marls and dolomites, then mottled marls with irregular intercalations of sandstone and dolomite, and at the top, yellowish dolomites of variable thickness containing the above Bivalvia species.

The Muschelkalk totals some 200 m in thickness.

The main unit of the Keuper is the Gypsiferous or "Hauptkeuper", comprising four divisions. At the base are mottled marls intercalated with thin dolomites and lamellae of quartziferous sandstone on whose surfaces are pseudomorphs of rock salt ("Keuper à pseudomorphoses"). Frequently thin intercalations of gypsum are present, some of which are exploitable. This littoral facies has yielded no fossils but has the most extensive development of all the Triassic and is sub-divided into five divisions, from base to top: reddish sandstones and mottled marls, 6 m thick; lower conglomerate of sandy dolomite with quartzose pebbles, 8 m thick; very hard quartzose sandstone alternating with mottled marls, 8 m thick; upper conglomerate of hard pure dolomite and quartzite pebbles, 6 m thick, and finally reddish marls with sandstones towards the top, 7 m thick. Above these beds come clayey sandstones of continental origin, known as the Grès roseaux, with frequent shalt-clay intercalations. Plant impressions, usually of *Equisetum arenaceum*, are quite common. Then come red gypsiferous marls, often showing intercalations of gypsum and cellular dolomite — the "Roten Gipsmergel" of German geologists. Above these beds are hard mottled marls — "Steinmergelkeuper" — with lentils of important gypsiferous deposits.

The stratigraphic position of the Rhaetic is open to debate, whether to be classed as topmost Triassic or basal Liassic. In the Lxxique stratigraphique international volume on the Lias, the Rhaetic of France is discussed,

whilst the Rhaetic of Luxembourg is treated in the Trias volume. In Luxembourg it is the custom to consider the Rhaetic as Triassic. The Rhaetic then begins with black foliated shales 5 m thick, with siliceous sandstones and conglomerates interbedded, the whole totalling some 15 m thick, but this can decrease to a few centimetres, represented only by pebbles. Above occur red clays, up to 8 m thick, but may thin away to zero. Good outcrops of the Rhaetic are rare in the country.

Placing the Rhaetic in the Triassic, the Keuper then totals some 320 m.

Jurassic in Luxembourg begins with the Hettangian. Plastic or sandy marls, becoming shaley towards the base, and alternating with limestones containing echinoid debris and typified by *Psiloceras planorbis* form the lowermost division. There follows the best-known and exploited stratigraphic unit in the country, the Luxembourg Sandstone. This sandstone facies is intercalated with marly and calcareous facies of the L. Liassic, so that its stratigraphic position rises in going from SE to NW across the Gutland, actually forming a large arenaceous lentil within the normal Liassic facies which trends obliquely across stratigraphic and palaeontologic units and zones. *Schlotheimia angulata* is the diagnostic fossil, but *Arietites bucklandi* and *Pentacrinus cf. angulatus* are characteristic in places. This compact, grey, buff or beige sandstone, extensively quarried for its sand and building stone, is the principal aquifer of the country, yielding some 27 million m³ water annually, 65 million m³ water percolating down to the water table annually over an area of exposure of 300 km², or a total area of 493 km² if we include areas covered by the impermeable marls lying above. The age of the subsurface waters varies between six and twelve months, hence constant renewal so that the groundwater is well oxygenated, the physical, chemical and organic nature is such as to render waters highly favourable for domestic consumption (BARTHEL, 1969; BINTZ, 1969).

The Sinemurian begins with the Strassen Marls and Limestones, dark, shaley marls with intercalations of bluish limestones, having a great abundance of *Gryphaea arcuata*. Two zones are recognized, those of *Belemnites brevis (acutus)* and *Arietites bucklandi*. In general these beds form a very thin covering over the Luxembourg Sandstone, are usually missing in quarries and road ditches. The U. Sinemurian (Lotharingian) is represented first by black, highly shaley marls, grading upwards into yellowish marls with calcareous concretions, poor throughout in fossils. There follows alternations of marls and limestones, with the zonal ammonites *Aegoceras dudressieri* below and *Ophioceras raricostatum (Arietites)* above. Sinemurian totals some 45 m in thickness.

The appearance of the Pliensbachian is marked by greenish sandy marls containing *Waldheimia numismalis*, followed by limestones with *Deroceras davoei* (Davoeikalke). W of Cessange there occur only two thin concretionary limestone zones — *Arietites raricostatum* and *Deroceras davoei* — whilst at

Kleinbettingen these zones are represented by sandy limestones and sandy-shaley marls, marking the beginning of the marly-sandy facies further W in Belgium. U. Pliensbachian (Domerian) is represented by foliated, concretionary and septarian marls, known as the *Amaltheus margaritatus* beds, up to 80 m thick, but only the septarian marls (calcaire lumachelle) show good outcrops. Above lie grey marls passing westwards into calcareous sandstones (Dippach Sandstone), with phosphatic nodules common towards the top. The septarian marls identify the lower limit in the area of the marly facies. *Amaltheus spinatus (costatus)* is the characteristic fossil. Pliensbachian totals some 115 m in thickness.

The Toarcian begins with the *Posidonomya bronni* beds. The lower part comprises extremely fissile bituminous shales, the *Harpoceras falciferum* beds, followed by less fissile marls with large calcareous concretions, constituting the *Hildoceras bifrons* beds, this fossil occurring only in the concretions. The next unit comprises a very thin (max. 15 cm) dark, sandy marl containing phosphate nodules, the *Coeloceras crassum* horizon, rich in general in ammonites. Above come dark marls, striped with thin sandy laminae with abundant *Harpoceras striatulum*. Though the ammonite *Astarte voltzi* is absent, these beds are named the *Astarte voltzi* marls as this species is abundant in neighbouring Lorraine. These marls pass upward through friable sandstones to sands with large calcareous concretions and a thin limestone layer at the summit. This is the *Harpoceras striatulum* Sandstone, which forms the boundary corresponding in part to the "Grès supraliassique". Then come shaley black marls, in part similar to the *Posidonomya* bituminous shales. In the iron basins of the SW, the marls become more sandy, difficult to distinguish from the underlying marls. These black shales are termed the *Harpoceras fallaciosum* beds.

As with the Rhaetic, so too with the Aalenian its stratigraphic position is open to debate. By some it is considered as topmost Liassic, by others, basal Dogger, i.e. the boundary between Lower and Middle Jurassic is speculative. In Luxembourg the *Dumortieria* beds were considered as Dogger Aalenian by LUCIUS (1948), as Liassic Aalenian by LUCIUS in the Lexique stratigraphique international volume (1956) and Liassic Toarcian by the Service Géologique de Luxembourg (1966) (Table II). The above beds comprise a lower unit of sandy marls and sandstones containing *Dumortieria levesquei* amongst other fossils, and an upper unit comprising sandstones with shale intercalations towards the top, containing, amongst other fossils, *Dumortieria pseudoradiosa*. Above there occur reddish-yellowish limestones with *Harpoceras plicatellum* and *Harpoceras opalinum*, followed by reddish limestones, conglomerates and sandstones containing *Harpoceras munchisonae*. Finally there occurs above a thin layer of marly sands with *Harpoceras concavum*. All the above Aalenian strata are represented in both the Esch and Differdange iron

ore basins, with the exception of the *Harpoceras murchisonae* beds which are lacking in the latter basin.

The Aalenian of Luxembourg has a thickness of some 40 m or 70 m, depending upon whether or not the *Dumortieria* beds are classed as Aalenian or Toarcian. All these beds form what are known as the Oolitic Iron Ores, Minette or Oolitiches Eisenerz (Table II).

Table II
Aalenian Stratigraphy of Luxembourg

LUCIUS 1948				Lex. stratig. intern. 1956			Serv. Géol. Lux. 1966		
Dogger	Aalenian	Concarum Beds		Lias	Aalenian	Ditto	Dogger	Aalenian	Ditto
		Murchisonae Beds				Ditto			Ditto
		Opalinum Beds				Ditto			Ditto
		Dumortieria Beds	Pseudoradosa Beds			Ditto	Lias	Toarcian	Ditto
			Levesquei Beds						

The Bajocian is represented by three principal units. Lowermost are the Sowerby beds, beginning with micaceous marls with numerous small phosphate concretions and a sandy limestone conglomerate with round and flat limestone pebbles. Rare specimens of *Sonninia sowerbyi* are present. Over this limestone occur similar rocks alternating with friable marls, the latter disappearing towards the summit. These constitute the *Sonninia* Limestone and the Ottange Limestone. The next unit is a 40 m thick bluish sandy limestone with a thin clay bed at the top, known as the Haut-Pont Limestone, representing the zone *Sphaeroceras sauzei* and *Sphaeroceras polyschides*. The youngest M. Bajocian strata are massive, white, detrital coralliferous limestones including the zones *Cadomites humphriesie* below and *Cadomites blagdeni* above.

Within Luxembourg the Mesozoics have a thickness of some 1400 m, outcropping over ca. 65% of the country.

Kainozoic

Of Palaeogene age are the pisolitic iron ores (Bohnerz) associated with the minette basins in the SW. These were formed under dry, warm conditions of Eocene times, giving rise to a lateritic-type formation up to 30 m thick.

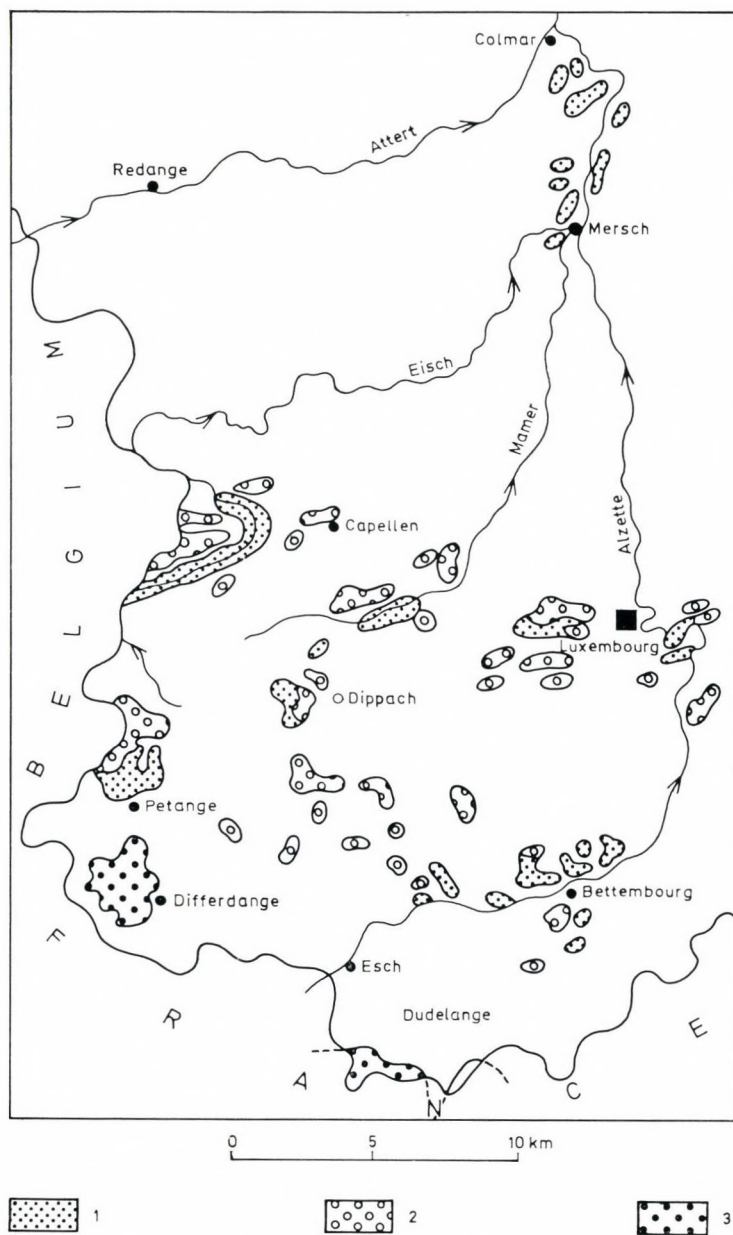


Fig. 4. Tertiary iron ores of Luxembourg. 1. Re-worked limonitic ores (lake-bog-iron ores, Post Miocene; 2. Primary limonitic ores, Miocene; 3. Pisolitic iron ores, M.-U. Eocene

Limonitic iron ores, quartzites, lignitic quartzites, of Neogene age, occur sporadically throughout the Gutland (Fig. 4), only a few metres or so thick. These are considered of Miocene age mostly.

Outwash products of the Pleistocene glaciation include loams, pebbly loams, sands, gravels, pebbly conglomerates, slump deposits, talus accumulations etc. Alluvium is restricted to valleys where it may attain thicknesses of 7–8 m.

The Kainozoic probably totals some 55–60 m in thickness.

Tectonics and Geological Evolution

Three major tectonic features occur in the Oesling, all striking NE-SW. Farthest N is the Bastogne anticline. Known as the Rocroi anticline further W in Belgium, at Serpont it has a Cambrian core but as the feature continues NE it plunges and successive Eodevonian come to form the core. Just W of Luxembourg, L. Siegenian forms the core, but this plunges rapidly near the frontier and only relatively small outcrops occur at Schimpach. No less the M. Siegenian plunges rapidly, hence also small outcrops in the country, whilst the U. Siegenian has the most extensive outcrops. In western Luxembourg, M. Siegenian digitations form symmetric folds, becoming monoclinal to the N. E of the Clerf valley the anticlinal structure dies out. The level or slightly rolling plateaux have few outcrops, slump deposits occur in valleys, so that detailed deciphering of structures is difficult.

To the S of the above is the Eifel syncline, known in Luxembourg as the Wiltz Basin, the Neufchateau Basin in Belgium. Beginning at Nouzon on the Meuse, the structure enters Luxembourg at Harlange, continues on through Dasbourg to Prum in Germany. In the western Belgian extremity, the structure is very pinched and highly dislocated, but broadens eastwards, folding, though tight is more regular. In Belgium and Luxembourg only Eodevonian outcrops, but as the syncline plunges E. M. and U. Devonian occur in the Eifel. In Luxembourg, Emsian forms the syncline, the Upper having widest extent. W of the Clerf valley the axis rises, with several secondary folds forming minor basins. This pronounced axial rise corresponds to an acceleration in plunge resulting in the secondary foldings. Highly irregular digitations of the borders of the main Basin are especially evident in the configurations of the thin Berlé Quartzite outcrops, prominent on both sides of the structure, and in the western sector in Luxembourg we also find long, thin, secondary, tight symmetrical folds plunging to SW and NE, disturbed by faulting. The variation in character, type and degree of foldings within the Wiltz Basin reflect variable reactions at depth of older strata forming the vicinal anticlinal structures.

S of the Eifel structure lies the Givonne anticline, whose SE flank is partially covered by Mesozoics. This major structure is marked by Cambrian outcropping at Givonne in France, gradually plunging eastward and in the neighbourhood of Prüm, in the Eifel, the anticline is completely covered by Mesozoics, but rises again eastwards, with Devonian outcropping in the "Eifel Hauptsattel". At the Belgian—Luxembourg border, the core is formed of M. Siegenian, plunging rapidly so that the extent of outcrops is small. In western Luxembourg the structure is strongly faulted. Proceeding NE, the U. Siegenian outcrops across the country. The anticlinal axis is believed to lie on the southern half, relatively close to outcropping Mesozoics, so that the northern flanks are more extensive. In the W, the boundary of the U. Siegenian—L. Emsian is taken to mark the limits of the Wiltz and Givonne structures, whilst in the E, between the Sûre and Blees valleys, the alternating outcrops of these stratigraphic units are considered as belonging to the Givonne anticline, and the more extensive development of L. Emsian in the northern part between the above two valleys belongs to the Basin. In the Our valley, the L. Emsian almost alone represents the Givonne structure. From W to E, exposures of L. Emsian increase in the Givonne anticline, forming relatively long, narrow outcrops lensing out westwards. Common throughout the structure are tight isoclinal foldings, symmetrical foldings. In the western sector S-dipping strata are more common; in the eastern sector, N-dipping beds at high angles are more typical, whilst in the Our valley dips are more gentle and symmetrically disposed. Faulting occurs in all three major structures. Commonest are smaller transverse gravity faults cutting across the "noses" of secondary folds, with general NW-SE strikes, downthrow to either the NE or SW. Such faulting accentuates the irregular, digitated outline of these folds. On the S flank of the Givonne anticline, in the eastern sector where Mesozoics are involved, gravity faulting, with strikes more parallel to the major structure are common, with downthrow to the S usually. In the western sector of this structure two important faults, Aiglemont and Herbeumont, enter from Belgium. The latter extends for some 75 km westwards into France, and in Luxembourg thrust faulting brings M. Siegenian in contact with U. Siegenian and the latter in contact with L. Emsian, faulting occurring on the N flank of the Givonne anticline, partly on the S flank of the Wiltz Basin. Between Perlé and Martelange, several splays of the thrust occur, resulting in highly intricate structures. Another significant thrust occurs in the Wiltz Basin. Between Grumelscheid and Drauffelt the N flank shows the Wiltz Schist thrust northwards against the Berlé Quartzite, so that the former is partially in abnormal contact with the L. Emsian.

The tectonic elements of the Hercynian foldings, forming "posthumous" features related to Caledonian deformations, have had considerable influence on sedimentation in the Oesling. The Eifel depression, running approx. N-S

through the region, between the Ardenne and Eifel of prime significance in the general sedimentation of Luxembourg, made itself manifest at the beginning of the Devonian, when and where maximum sedimentation took place on a sunken region already outlined in Cambro-Silurian times. The Givonne anticline in the L. Emsian period acted as a shallow separating two different facies. To the N, in the Oesling, the L. Emsian comprises quartz-phyllites and quartzose sandstones; to the S, in the Hunsruck, are shales, schists and slates, the last-mentioned of M.-U. Siegenian age. In turn, the U. Emsian shows two facies, to N and S of this shoal area, which also forms the southern limit of the red shale facies of the M. Emsian. The Bastogne anticline indicates the boundary of the U. Siegenian facies, for N of this structure the sandy facies of Bas-Bellain contrasts with the shaley facies of the rest of the Oesling. The Eodevonian of the Oesling is characterized throughout by extensive schistosity, here almost entirely of mechanical and not metamorphic origin. It represents an advanced consequence of the folding, the compression of rock material in the direction of axial planes due to tangential pressures, cleavage planes coinciding with axial fold trends. The inclination of foliation planes varies but is most common between 60° and 80° , in direction usually parallel to the bedding. Further, these L. Devonian Luxembourg strata have been subjected to extensive jointing, generally disposed in N-S and E-W directions. Inclinations vary from quasi-horizontal to vertical. In homogeneous rocks, joint planes are smooth but very rough in heterogeneous ones. A certain degree of movement along joints is indicated by striae in various directions. All lithologic types show jointing but the network is more dense as the rocks become more granular, hence more developed in arenaceous rocks. Jointing developed subsequent to folding.

Hercynian diastrophism resulted in only a moderate relief, intense metamorphism and crystalline massifs are lacking as exposures, except in the highest uplifted parts of the Ardenne where plutonic dykes and apophyses occur, though crystallines presumably occur at great depth. Continental conditions thereafter continued till the beginning of the Mesozoic. *Pari passu* with the lowering of the mountains, detritals formed, rivers brought débris to fill-up irregularities of the surface, carved away slopes and smoothed them, so producing the post-Hercynian or pre-Triassic peneplain previously mentioned, a disaggregation surface where only very rarely fresh basement rocks were exposed. Under arid, warm conditions, the red and brown beds of the Permian Rotliegendes were formed, to be succeeded by the Zechstein transgressive sea with its dolomitic limestones. As noted above, this Permian has a thin development at some 700 m depth in the far S of Luxembourg. The Zechstein marine invasion continued on through the Mesozoic, where in Luxembourg the Triassic and Jurassic sediments were formed in the Gulf of Luxembourg, penetrating NE across the country between the Ardenne and the

Hunsrück. On the southern edge of the Oesling L. Triassic deposition in this marine environment occurred, in angular discordant relations to the highly disturbed Eodevonian below. During Triassic sedimentation, transgressions as well as subsidence of the Gulf area — a southern extension of the N-S oriented Eifelian depression — were characterized by abrupt movements alternating with uprise of the base, so that the three Triassic series display cyclical sedimentation in direct relation to epeirogenic movements experienced. These series are to be interpreted as evidences of physico-palaeogeographic changes rather than biostratigraphic or biochronologic units. Liassic and Dogger seas later penetrated into the Gulf, a NW-SE section across the Gulf showing disposition of Mesozoics comparable to that in a SW plunging syncline, though in reality the Gulf deepened in that direction. Facies distinctions in the Jurassic are due to changes in the directions followed by transgressions, manifest by modifications in the trends of littoral currents, as well as by the degrees of epeirogenesis.

At the end of the Jurassic the seas regressed, and during the L. Cretaceous continental conditions prevailed. But in the Senonian a transgressive sea once again penetrated eastwards from the Paris Basin, to infringe upon the Ardenne and over part of Luxembourg, marking the last phase of marine sedimentation, since when continentality has prevailed.

Compared to the intense deformations associated with the Hercynian orogeny which affected the Oesling — and of course the Palaeozoic substratum at depth in the Gutland — the Mesozoics experienced only distant echoes of the Alpine orogeny, forming folds of large radius of curvature and faulting. Again structures adopt the Variscan trend, superimposed upon Hercynian foldings at depth, which in turn were influenced by Caledonian structures, with the result that the Palaeozoic and Mesozoic structural pictures are duplicated, i.e. the Mesozoic foldings reflect the Palaeozoic foldings. Duplication of course is not exact, for a pronounced unconformity separates the covering sediments from the basement, nor can we say that the directional trends of the sets of folds agree exactly, but there is overall similarity.

Extending SW-NE through the centre of the Gutland is the Weilerbach syncline. This structure is more clearly developed E of the Alzette valley, whilst W hereof its character is less well defined, but the occurrence of some 100 m of Luxembourg Sandstone in borings at Differdange and Longwy indicate its presence extending beyond the frontiers towards the SW. This structure is superposed on a Hercynian basin which outcrops in the Eifel, with Eodevonian between Daun and Gerolstein.

Likewise the Born anticline is superimposed upon a Hercynian anticline, where in the Eifel U. Siegenian forms the core. In Luxembourg the Born anticline has a core of Buntsandstein in the vicinity of Born on the lower Sûre. Towards the SW, it appears that this structure divides into several

anticlines, all of which extend W of the Alzette — the Blascheid anticline to the N, the Ernster anticline in the middle, the relatively long Sandweiler anticline in the S, with the Luxembourg and Junglinster synclines in between. Another major fold is the Middle Moselle anticline which curves through the SE part of the country, convex towards the NW. At Nachtum on the Moselle, L. Muschelkalk forms the core; SW of here the Hauptmuschelkalk crops out in the centre of the structure, whilst the high structural position of the Rhaetic and Hettangian above the Trintange valley marks the location of this anticline.

Faulting of significant dimensions is much more evident in the Gutland. The genesis of this fault network is inter-dependent with the Mesozoic foldings, so that faults and folds mutually replace one another. Mesozoic deformations, both folds and faults but especially the latter, increase in number and degree from NW to SE across the Gutland, the degree of fracturing increasing as the Moselle Trough is approached. Faulting represents adaptation to folding brought about by the compression of rocks of little elasticity and lacking thick sedimentary coverings. Initial tectonism created undulations of great radius of curvature, but repeated movements, especially if of shock type, distorted these rocks showing little elasticity or plasticity by means of fracturing.

Like the folds, the faults in the Mesozoics have a Variscan trend in relation to the major folds or then a Rhenic trend (NNE-SSW) in relation to the transverse Eifel depression. This latter trend characterizes the faulting in the Moselle region, changing to NE-SW further westwards. No thrusting occurs in the Mesozoics. The larger faults all show a tendency to be down-throw on the SE side, but in the Sûre — Moselle corner especially, many smaller faults create graben and horsts. This Moselle faulting marks the approximate boundary of the Hunsruck massif and the transverse Eifel depression, in genetic relation with the progressive sinking of this depression. Considerable faulting, affecting both the L. Emsian and the Buntsandstein is found N of Diekirch between the Sûre and Our valleys. It is to be noted that many faults in the Gutland also affect the thin Tertiary occurrences, even Pliocene, testifying to the relatively young age for the fault network.

As already commented, beginning in the late Pliocene, uplift of the entire Luxembourg region occurred, more pronounced in the Oesling than the Gutland. There thus developed a pronounced warp or flexure — the Oeslinger Randverbiegung — marked by the overlapping of the Buntsandstein on to the Eodevonian.

Economic Geology

Unquestionably the most renumerate natural mineral resources of Luxembourg have been the iron deposits. A considerable literature has developed on this subject, of which the more recent ones of LUCIUS (1945, 1952) are of prime importance.

The iron oxide ores occur as Aalenian oololiths and Tertiary concretions. The former are comprised essentially of more or less hydrated ferric oxide with which are associated chlorites, siderite, more rarely magnetite and hematite. Concretions comprise only hydrated ferric oxide. Goethite the principal ferric mineral of the Lorraine—Luxembourg iron deposits.

Tenors in Fe in the various stratigraphic units vary from 2.6–3.6% L. Devonian; 1.5–2.8%, Triassic; 1–25%, Jurassic. The U. Domerian (*Amalthæus spinatus* beds) have the highest tenors, the U. Sinemurian (Lotharingian) the lowest tenors, whilst the M. and U. Keuper have the widest extents — 624 km² — as also the greatest thicknesses — 150 m. The average Fe tenor for all rocks except the Aalenian is 2.8% throughout an area of 2484 km². Down to a depth of 20 m over this area there is a total of 2800 million metric tons iron. The iron basins of the SW had an initial extent of some 82 km², but this has been reduced to about half today.

The ores which have brought heavy industry to the country are located in the Esch and Differdange basins, with thicknesses of up to 50 m in the former, up to 30 m in the latter, the basins being separated by a fault. The ores occur in two stratigraphic horizons, the L. Aalenian siliceous horizon and the U. Aalenian calcareous horizon (Table III).

The iron ore deposits are classified as alluvial ores and minette, the latter word signifying “little mine” as in early days it was thought that the oolitic iron ores were inferior to the Tertiary ones. The former comprise Tertiary concretions and actually formed the basis of the iron industry of the country before 1860. The minette forms the basis of present industry though this is now largely supplied by imported ore. The minette is classified on the basis of its accessories into siliceous and calcareous, a distinction of metallurgical importance. Phosphorous is always typical in the minette, the limonitic meadow or lake-bog iron ores (“mineral de fer des prés”, “Raseneisenerz”) being a phosphatic deposit where P_2O_5 varies between 0.5 and 3.5%. On a practical basis, the deposits are classified into pisolitic deposits (“Bohnerz”; “mineral en forme de fèves”), high tenor ores occurring as concretions or crusts, formed of Fe in argillaceous residues of the Dogger-Lias limestones in a dry Eocene climatic environment. The other group comprises the “Wascherz”, “mineral lavé”, “mineral de fer des prés” or low tenor ores, concentrated as nodules and concretionary crusts of Fe contained in sandy clay deposits of the Miocene, precipitation occurring in a warm, humid climate. The third group is the Aalenian oolitic deposits or minette, comprising siliceous and calcareous varieties.

The oolitic minette results from the aggregation and cementation of oolites and clastic material. On a genetic basis, four types can be recognized, which result from: (a) chemical action whereby oolites and cement or then only cement is formed in situ whilst the oolites are formed elsewhere and have

Table III

Aalenian Stratigraphy in the Iron Basins of Luxembourg

			Esch Basin				Differdange Basin			
Aalenian			<i>Harpoceras concavum</i> Beds (Erosion surface with pebbles)							
	<i>Murchisonae</i> Beds	Calcareous Horizon	<i>Harpoceras murchisonae</i> Beds	Roof Sandstone Conglomerate		Dislocation Zone of Audun-le-Tiche Fault	Roof Limestone			
	<i>Opalinum</i> Beds		Red Sandy Bed	4	Upper Red Bed		3b	Gap in Sedimentation		
			Lower Red Bed	3a			Principal Red Bed		3	
	<i>Dumortieria</i> Beds	Siliceous Horizon	<i>Harpoceras opalinum</i> Beds	Second Yellow Bed	2a		Principal Yellow Bed	2	Upper Limestone	2
			<i>Harpoceras plicatellum</i> Beds	Grey Bed				1	Lower Limestone	1
<i>Dumortieria</i> Beds	Siliceous Horizon	<i>Dumortieria pseudoradiosa</i> Beds	Supraliasic Sandstone	Brown Bed	I		Red Bed	Grey Bed	I	
		Black Bed		II	Black Bed	III				
		<i>Dumortieria levesquei</i> Beds		Green Bed	III	Green Bed		IV		

Dislocation Zone of Audun-le-Tiche Fault

been transported thither; (b) mechanical action, whereby the oolites and clastic material have been transported and concentrated. Of the four types, those in which the oolites are alien to the milieu where now occurring are dominant. These oolites, along with clastic material, were transported via currents and consolidated by cementation, either of ferruginous or calcareous nature. In beds I, II and III, and as traces in beds 1 and 2 (vide Table) chlorites and siderite are present in the ferruginous cement. Oolites are more strongly hematized than the cement, having undergone oxidation during transportation. The calcareous types are lacking in chlorites and siderite, typical of beds 3. In rising stratigraphically, hematization increases, due to longer period of exposure to exterior agents favouring oxidation. The oolites form through chemical precipitation in a littoral milieu, being oxidized in the process of current transportation, the clastic débris of oolites and quartz also showing advanced oxidation, cementation taking place in situ.

The Esch-Differdange iron basins are separated by the Audun-le-Tiche fault, with a downthrow to the SE of 125 m. One or two smaller faults as well

as folds have the same NE-SW orientation (Figs 2, 3). The siliceous horizon is well developed throughout the Differdange basin and in the western part of the Esch basin, the calcareous horizon elsewhere in the latter.

During the L. Aalenian, tectonic conditions were the same in both basins, but in the Upper period, a distinct change occurred, with the formation of the calcareous horizon, the Audun fault outlining this distinction. Tectonism resulted in the marine currents not only changing directions but also energy changes took place so that the siliceous horizon formed in a uniform, quiet environment, whereas in the calcareous horizon, the oolites were entirely or then strongly hematized, are associated with much detritus and the influence of the submarine relief and current velocity variations became marked. Uplift of the two basins caused a change in configuration, depth, sedimentation and mineralization: in the Esch basin the calcareous horizons expanded whilst in the Differdange basin sedimentation ceased.

The minette represents a sedimentary cycle, weathering taking place under particular climatic, tectonic and topographic conditions on land adjacent to the Lorraine Aalenian sea. In a warm climate, the abundant rains accomplished much erosion, iron was dissolved by surface waters and arrived in the shallow vicinal sea where it was precipitated in littoral or sublittoral regions as oolites. In the deeper iron horizons, chlorites and siderite always are present along with hematite, the former precipitating simultaneously without filtration, the latter resulting from the oxidation of chlorites or siderite as well as by direct precipitation, for in the upper calcareous horizons, no trace of chlorites-siderite is seen in the oolites, and further, substitution of a ferruginous oolite by a calcareous one has never been observed. The oxidation of the iron salts results from exterior causes, the degree of oxidation depending upon the duration of such and its intensity. In the siliceous horizons, chlorites-siderite decrease upwards, with only traces in the calcareous horizons. Oxidation is directly related to tectonism and the energy of currents which increase the degree of aeration. The lower siliceous horizons were formed in a quieter environment than the upper calcareous horizons, these two horizon types representing extremes of environmental deposition, with cementation occurring after precipitation of the iron solutions in the basins.

The nuclei of the oolites in the calcareous horizons are formed of quartz, organic débris, fragments of envelopes of broken oolites and pieces of pre-existing minerals, whilst in the siliceous horizons, the constitution of the minette is more complex (Table IV, V).

The Lorraine—Luxembourg iron deposits are but a special case of the iron oolite facies so extensively developed in the L. Dogger (or U. Liassic) of Central and Western Europe, associated with particular tectonic conditions and palaeogeography. Thus the facies is found in the vicinity of emerged massifs in shallow seas, e.g. Swabia, Franconia, Bohemia, Massif Central,

Table IV

Chemical Analyses of Thirteen Samples of Pisolitic Iron Ores.
(Adapted from Lucius, 1948)

Total SiO ₂	14.00—63.12%
Al ₂ O ₃	0.00—22.00%
CaO	Tr— 2.04%
MgO	Tr— 1.17%
P ₂ O ₅	Tr— 0.57%
Fe	9.77—56.50%
Mn	Tr— 0.24%
S	Tr— 0.04%
Heat loss	4.58—12.25%
Insol. resid.	14.72—66.85%

Table V

Chemical Analyses of Thirty-two Samples of Limonitic Bog-iron Ores.
(Adapted from Lucius, 1948)

Total SiO ₂	3.20—43.87%
Al ₂ O ₃	0.70—13.96%
CaO	0.00— 4.90—
MgO	0.00— 0.45%
P ₂ O ₅	0.55— 2.60%
Fe	29.20—56.00%
Mn	0.00— 0.45%
S	0.00— 0.17%
Heat loss	7.70—12.12%
Insol. resid.	3.48—44.98%

Ardenne. Deposition of this Jurassic oolitic iron is related to the general geology and local tectonic states of the regions in question, so that necessary concentrations took place only under prescribed circumstances. Yet the vastness of such sedimentary ores even in the Lorraine region — estimated at over 5000 million metric tons — surely implies that purely local conditions were not the sole causal reasons. These iron deposits occur at the passage from Liassic to Dogger, in epicontinental seas close to emerged massifs which were undergoing weathering by copious rains in warm climatic environments. Concentration and distribution of the ferruginous oolites were controlled probably chiefly by local tectonic activity, topography of the shallow sea floor, direction and force of marine currents.

The distribution of the Tertiary iron ores, no longer of industrial import but geologically interesting, are shown in Fig. 4.

Extraction of iron deposits from the mining basins during the years shows the influence of wars, occupations, economic depressions, etc. From 1868 to 1975 this has varied from 720.000 metric tons in 1868 to 7.76 million tons in 1937 to 10.3 million tons in 1975. On the other hand the production of crude pig-cast iron and steel amounted to 4.8 mill. tons, 1974, and 3.5 mill. tons iron in 1975 and 4.2 mill. tons steel in the last year. The Luxembourg iron-steel industry depends to the extent of some 73% of its iron from Lorraine, ca. 4% from Sweden and Brazil. All but a microscopic amount of finished iron-steel is exported.

In the Oesling (LUCIUS, 1948, 1950) copper ore occurs at Stolzembourg in the shales and quartz-sandstones of the L. Emsian, containing up to 18% Cu. These deposits began to be exploited in 1755 and continued up until World War II when all was destroyed during the Ardenne offensive of 1944–45. Antimonite occurs in compact U. Siegenian shales at Goesdorf, worked from 1745 to 1938. Galena was exploited at Oberwampach between 1826 and 1939, from S. Siegenian sandstones and sandy shales. An aeromagnetic survey of the country recently showed strong anomalies in this region and across into Belgium where galena is presently exploited. So far field checks in the Luxembourg area have not been made.

Iron excepted, exploitation of the above metallics was never of great economic significance, but the geological interest of such remains. Of the non-metallics, gypsum has been of greatest economic importance.

Of various mineral-thermal springs, those of two localities only are worthy of mention. At Mondorf-les-Bains, during 1841–1846 a 730 m deep boring was put down to the Eodevonian, and for a quarter of a century, was the deepest bore-hole in the world. In the U. Buntsandstein at depths of 460 m and 502 m waters were encountered flowing at the rate of 872 m³ per 24 hours. In 1913 another boring (Adelheidquelle), some 100 m N of the first, was put down to 589 m into the Vosges Sandstone, mineral waters being encountered between 464 m and 580 m. These waters have temperatures varying between 19° and 25 °C, and contain approx. 15 gr salts per kgm water of which 9 gr per kgm water are sodium chloride (Table VI). In neighbouring Lorraine several thermal-mineral springs occur in the Buntsandstein, some with temperatures as high as 39 °C. The springs at Mondorf-les-Bains have been commercially developed as a health resort, parks, hotels, casino, etc. as well as the bottling and distributing of the mineral waters.

Then in 1959 the carbonated mineral water springs at Rosport were put into commercial production, the only gaseous spring in Luxembourg. A 67 m deep boring into the Voltzia Sandstone encountered carbon dioxide emanating at 20 m, with 1300 cm³ gas per litre water, 1520 cm³ gas at 36 m.

Table VI

Chemical Analysis (in gr.) of Mineral Waters, Mondorf-les-Bains

Sodium chloride	9.415
Potassium chloride	.202
Calcium chloride	3.426
Lithium chloride	.008
Sodium bromide	.167
Sodium iodide	.001
Calcium sulphate	1.501
Magnesium sulphate	.134
Calcium carbonate	.001
Strontium carbonate	.090
Magnesium carbonate	.454
Ferrous carbonate	.033
Manganous carbonate	Tr
Ferrous phosphate	Ti
Calcium arsenate	.001
Metasilicic acid	.015
Silicon dioxide	Tr
Total	15.448

The gas exerts a pressure of 1.3 atmospheres, 4.2 m³ water delivered per hour, with 2650 cm³ gas per litre water below 36 m depth. Samplings show 2.99 gr. CO₂ per kg water, and is to be classed as dull-alkaline carbonated water, like those of Gerolstein, Daun, Apolinarisbrunnen, etc. Since the above date, the "Source de Rosport" has been developed, a bottling plant built. There are numerous carbonated springs in the Eifel, the gas originating in deep basaltic magma chambers which gave rise to basaltic lavas, with extensive faulting in the Rosport region, so providing channelways from depth to surface.

Conclusion

With its small area, the geology of Luxembourg represents but a regional part of Ardenne, Lorraine and Eifel geology, the geological features of these three areas occurring within the country.

Fortunately in geology it is not size of a region or country that stimulates scientific interest. The somewhat jingoistic contention that on a per capita basis, Luxembourg is the greatest steel producer in the world might impress Chambers of Commerce, but makes little impact on the geological fraternity. The sober reporting of the essential geological features which can be integrated with neighbouring countries is the sole aim intended here.

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ОБЗОР ГЕОЛОГИИ ТЕРРИТОРИИ ВЕЛИКОГО ГЕРЦОГСТВА ЛЮКСЕМБУРГ

ПАУЛ С. МИТЧЕЛ-ТОМЭ

Резюме

Территория Люксембурга состоит из двух районов, отличающихся по геолого-геоморфологическому строению. На севере находится район Эслинг, сложенный сильно складчатыми палеозойскими отложениями с обширными плато и глубокими, узкими, извилистыми долинами; эти формы рельефа обусловлены формированием в послеварисское время пенеplena, приподнятого в позднеэоценовое время. На юге расположен район Гутланда, где широко распространенные мезойские образования повлияли на формирование разнообраз-

ной топографии — открытых и более молодых долин; здесь разрывные нарушения являются более выраженными на морфологии рельефа. Впадина Эслинг образует границу между обеими единицами, в неё направлены все водотоки обоих районов, причем сток осуществляется в восточном направлении.

Додевонские, триасовые, лейасовые и доггерские отложения широко развиты в коренном залегании; третичные отложения, однако, представлены в весьма ограниченной мере. Толща упомянутых отложений достигает мощности около 4300 м.

Тектонические структуры приурочены в общей сложности к варисским (СВ—ЮЗ) или рейнским направлениям простираения (ССВ—ЮЮЗ), обусловленным каледонскими и варисскими движениями, соответственно. Крупные складки являются лишь частями структур, прослеживающихся в направлении Бельгии или Германии, соответственно. Приближаясь к депрессии Мозел, мезозойские отложения в значительной мере нарушены, причем разрывные движения остаются активными даже в четвертичное время.

На базе месторождений железной руды на юго-западе развивалась черная металлургия. В настоящее время осуществляется разработка только лишь руд типа «минетт» (ааленский ярус), хотя разработка этих трутичных рудных тел скоро приведет к истощению промышленных запасов сырья. Примерно 80% потребностей страны в железе должны быть обеспечены импортом железной руды.

OUTLINES OF THE HYDRODYNAMICS AND HYDROCHEMISTRY OF THE PANNONIAN BASIN

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The regional flow-region of the basin is recharged by the groundwater of the mountains around and inside the basin (Figs 1 and 3). The intermediate flow-regions of the sub-basins are recharged from the mountains as well as from topographic highs inside the basin (Figs 1, 3, 6 and 7).

The potential distribution shows the hydrodynamic characteristics. Areas of strong positive vertical gradients coincide with the portions of cross-sections where the equipotential lines are parallel to major geologic features. This is an indication of stagnant deep groundwater. Other proofs of stagnant deep groundwater are the 85 m island-like contourline and very low horizontal gradients in the deep depressions (Fig. 2). The deep-lying undrained flats are the main natural discharges of the plains (Fig. 2).

Highly permeable formations serve as deep recharge belts indicated by bulging equipotential lines (Fig. 7) or by vertical equipotential lines showing the hydrostatic conditions (Fig. 5).

The hydrodynamics of flow-regions is in harmony with hydrochemistry and geothermics. A flow-region represents a thermal circulation system being in equilibrium under natural conditions. The equilibrium is produced by the descending cold bicarbonate water of the recharge area and by the ascending warmer salty water. In the same depth in the flow-region the descending limb shows lower water temperature, lower salinity and higher total hardness, whereas the ascending limb is of opposite characteristics and often contains gaseous water.

Existence of large and deep groundwater flow-regions requires considerable energy potential, precipitation, and extensive deep-reaching permeable formations outcropping over large areas.

The surface configuration of the Pannonian Basin provides for the required differences in elevation necessary to energy potentials to maintain deep flow-systems.

Precipitations over 500 millimetres/year and outcrops of thick and highly pervious formations are rendering possible the continuity of the flow-system.*

The deep groundwater flow-system of the Pannonian Basin is a consequence of elevation differences in the surface configuration as well as the lithology of the rocks there.

* The basic work on the theoretical principles of a groundwater flow-system was written by HUBBERT in 1940.

A major advance started when Tóth presented his computer-developed models.

Studies on groundwater flow carried out in the last fifteen years cleared the essentials of the hydrodynamics of regional flow-systems (BROWN, FARVOLDEN, FREEZE and WITHERSPOON, LE BRETON and JONES, MAXEY, MEYBOOM, WINOGRAD, etc.).

The character of a flow-region is shown by (a) the pressure conditions, (b) chemical composition of groundwater, (c) geothermic anomalies, and (d) natural isotopes.

In a multi-layer artesian flow-system there are two kinds of vertical pressure patterns:

1. Pressure due to differences in energy potential within the same formation shows a more or less uniform increase with depth.

A uniform increase is not always found when pressures are measured in different aquifers. Where the energy potential increases with depth there is an upward movement of water through poorly permeable confining strata.

2. There are extensive areas where pressure in the same borehole is less in a deeper than in a shallower aquifer. The downward increase of pressure is less than hydrostatic.

The decline in pressure with depth shows that an aquifer is recharged by a downward seepage of water through the overlying poorly permeable confining bed. Since energy is used in this movement, energy potential decreases with depth. The resistance to flow of the rocks should be duly considered with regard to depth and extension of the flow-systems.

A downward decline in pressure occurs when the deeper-situated aquifer is much more permeable than the higher-situated one.

A further explanation may be that a lower water-bearing stratum is sealed off by an unconformable formation. It is a possibility too, that the recharge area of the deeper-situated aquifer is less elevated than that of the shallower one.

Short Description of the Lithology and Geometry of the Pannonian Basin

The Pannonian Basin is a large and non-uniform flow-system and structurally a deep depression formed mainly during the late Tertiary and Quaternary (Fig. 1).

a) The Paleozoic basement consists of crystalline rocks.

b) Karstic Mesozoic limestones and dolomites occupy large areas around and inside the basin and constitute an important factor of the hydrogeology of the entire region.

c) Both basement complexes, the Paleozoic and the early Mesozoic (Triassic — Early Jurassic) are overlain by diverse *overburden formations*:

1. Deposits of excellent permeability: post-Liassic hard fractured Mesozoic limestones, Eocene nummulitic limestones and Miocene limestones.

2. Medium to poorly permeable deposits: thick sandy marine sequences of the Upper Oligocene, and Lower and Middle Miocene (Schlier); Lower Eocene

and Middle Miocene continental and littoral deposits, and coal measures. Sandstones and flysch deposits of great variety exposed only in the mountains outside the central basin; continental, deltaic and littoral beds of the Pliocene outside the central depression in the smaller intermontane basins.

3. The Tertiary volcanics — mostly Middle and Upper Miocene of age — are of different permeability owing to vertically alternating lava flows and tuffaceous beds as well as tectonic dislocations.

4. Impermeable formations: a thick sequence of post-Middle Liassic Mesozoic, marine formations Upper Cretaceous continental clays, thick Upper Eocene and Middle Oligocene marine clays.

The hydrogeologic boundary of the Pannonian Basin runs along the surface divide connecting river gaps cut into impermeable bedrock (Fig. 1) where there is no underflow, consequently the total amount of water entering and leaving the central basin through the gaps is surface water and can therefore be assessed there by river discharge measurements that are necessary for estimating the total water balance of the basin.

The boundary of the central basin runs along the surficial contact between the Pliocene deposits and older formations (Fig. 1). This is a natural boundary showing marked differences in hydraulic conditions, in topography and in paleogeography. The central depression was formed by rapid subsidences during the Pliocene and Quaternary. This boundary, almost everywhere, is also a zone of deep marginal faults. The basin fill thickens considerably, sometimes within very short distances, away from the bounding fault zones. The young geologic age is very well shown by the fact that the crystalline basement well inside the basin is covered over larger areas by deposits not older than the Upper Pannonian (Pliocene).

The central basin can be divided into two structural storeys. The upper one consists of surficial Pliocene deposits with or without a thin Quaternary cover. The lower storey is the area of Quaternary subsidence (Fig. 1).

Hydrodynamics of the Pannonian Basin

Research on the hydrodynamic conditions of the Hungarian Basin has been started comparatively early. Results of the investigations carried out in the 1950's were published in 1962 (Almássy in Schmidt: Hydrogeological Atlas of Hungary).

The piezometric contour map (Fig. 2) shows the original undisturbed pressures (pressures unaffected by withdrawal).

Plotting the piezometric contour map and the evaluation of head measurements for the calculation of vertical hydrodynamic gradients have been permitted by favourable circumstances:

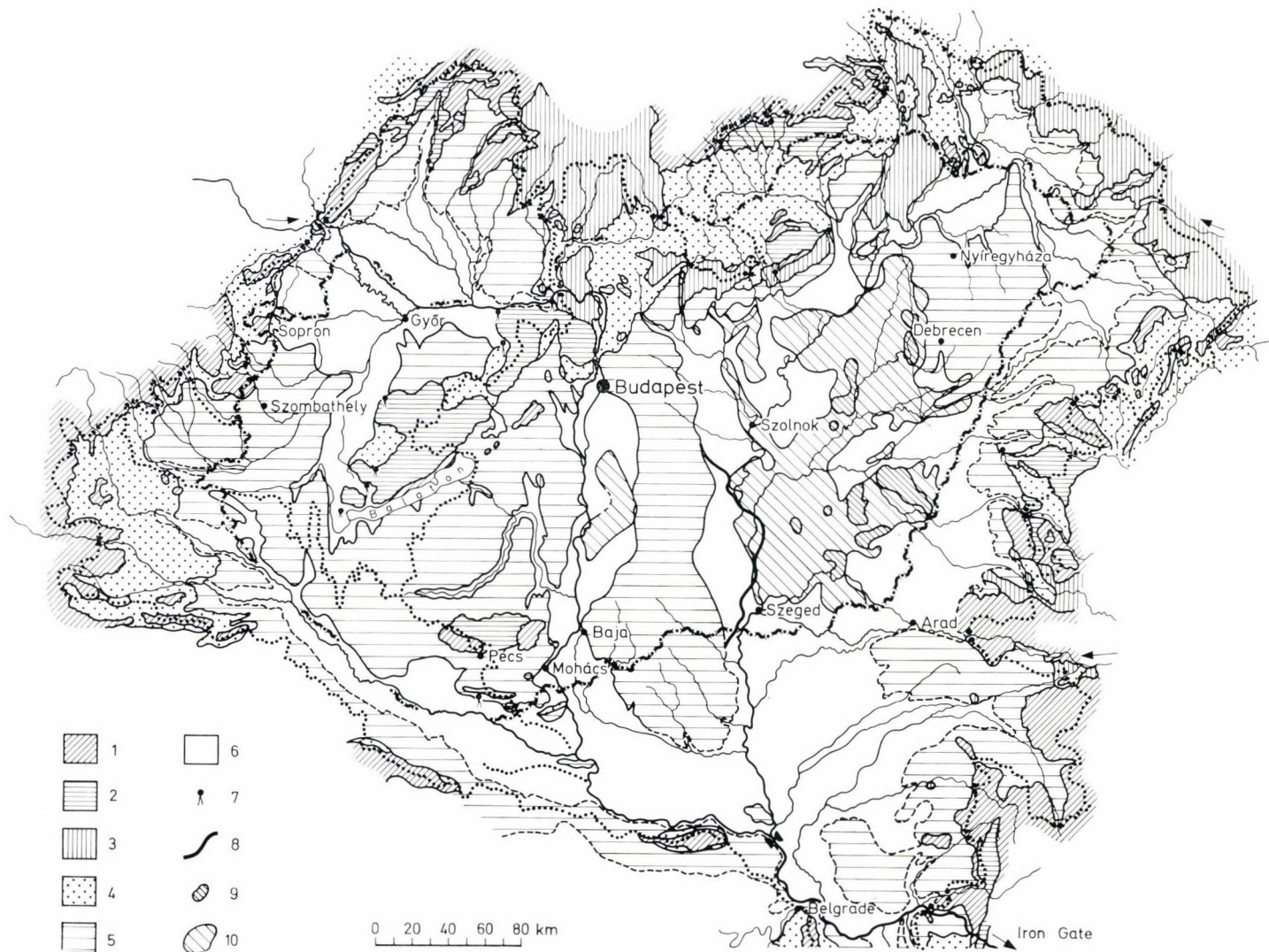


Fig. 1. Hydrodynamic and hydrogeological map of the Pannonian Basin (ERDÉLYI, 1973). Recharge areas: 1. Crystalline rocks; 2. Karstic rocks; 3. Volcanics (Neogene); 4. Recharge area. Sedimentary deposits: Eocene, Oligocene and Miocene; 5. Recharge area. Sedimentary deposits: Pliocene and Quaternary; 6. Discharge area; 7. Large karstic springs; 8. Zero hydraulic gradient at 400–600 m depth below ground; 9. Area of gaseous wells prior to hydrocarbon prospecting (depth of wells less than 100 m); 10. The same but depth between 100 and 400 metres;

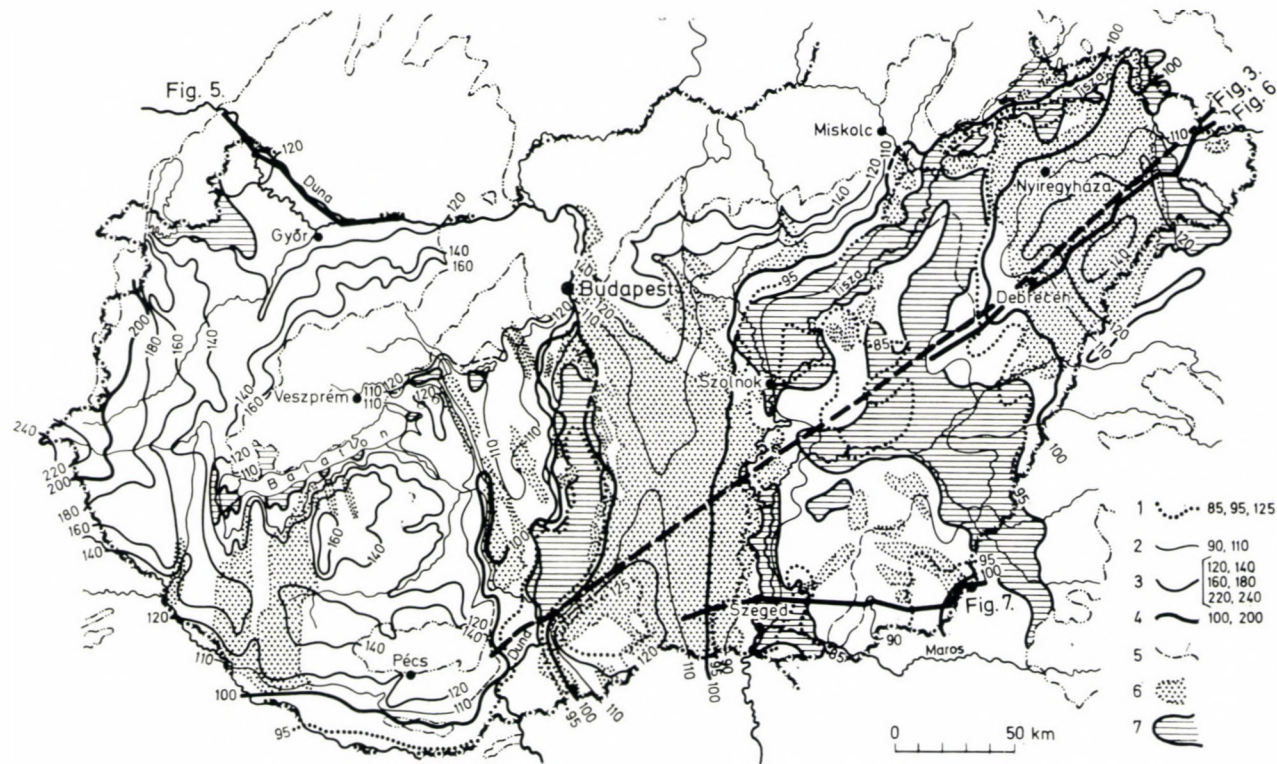


Fig. 2. Piezometric contours at depth of 150 to 400 metres below ground (ERDÉLYI, 1972). 1—4. Piezometric contours; 5. Boundaries of the hills; 6. Highly permeable soils (dune sand, loessy sand and sandy loess over 85 per cent of the area shown; 7. Undrained and poorly drained land

1. The number of drilled wells of known aquifer depth with elevations obtained by geodetic survey was about 8200 at the end of 1970.

2. Packer-tests were performed in hundreds of boreholes traversing two or more aquifers.

3. A great number of bored wells close to each other tap different aquifers.

The vertical hydrodynamic gradients were calculated from pressure heads of consecutive deeper aquifers not separated by impermeable or poorly permeable strata.

Figs 1 and 6 show the intake and discharge areas of the intermediate and regional flow-systems. Fig. 3 shows the flow-system diagram of the Great Hungarian Plain.

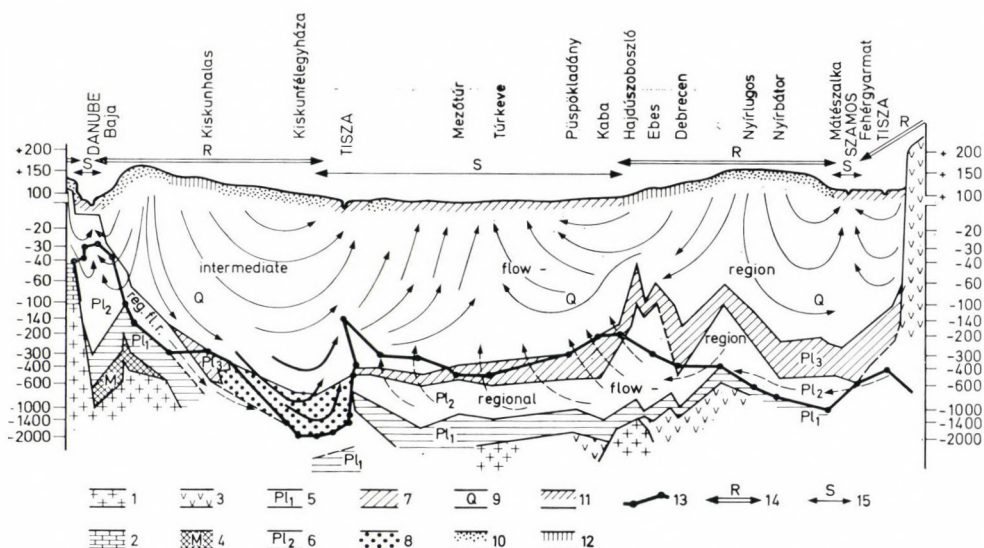


Fig. 3. Diagrammatic flow-pattern of the Great Hungarian Plain (ERDÉLYI, 1972). 1. Crystalline basement; 2. Mesozoic limestone; 3. Miocene volcanics; 4. Miocene deposits; 5. Lower Pannonian (Pliocene); impermeable, poorly permeable; 6. Upper Pannonian (Pliocene), permeable, poorly permeable; 7. Upper Pliocene, poorly permeable; 8. Upper Pliocene, highly permeable; 9. Quaternary in general; 10. Dune sand, loessy sand; 11. Silt, clay, soda soils; 12. Loess, sandy loess; 13. Fresh-water and brine interface; 14. Recharge; 15. Discharge area

1. Unconfined Water of Mountains and Hills

The mountains and hills around and inside the basin are characterized by an unconfined water table. The water table of the crystalline basement follows the topography. The karstic water table of the carbonate basement is a deep-situated regional water table which becomes confined only when carrying a considerable overburden.

The downward-decreasing pressure pattern in the mountains depends on the primary and secondary porosity of the rocks. Bordering lowlands and river valleys are the natural discharge areas of the mountains.

2. *Areas with a Decreasing Energy Potential Inside the Basin*

There are topographically distinct places of the central basin where the downward increase pressure is less than hydrostatic to great depths, mostly down to the neutral pressure zone separating the intermediate flow-regions from the regional one.

a) A comparatively narrow belt, more or less parallel to the foot of the mountains (Fig. 1). In this zone the shallower artesian aquifers are exposed to the surface and recharged by the unconfined groundwater of the outcrop area.

Deeper confined beds, however, are not recharged directly, but partly through the fractured karstic and volcanic rocks of the mountains bordering the basin (Figs 1 and 3). This is an example of a hydrodynamic short-circuit.

b) Low undulating terrains and higher hills made up of Pliocene deposits of continental, deltaic, littoral and lagoon origin. They are exposed over large areas and in small outcrops in Transdanubia and along the boundary of the Great Plain or capped only by thin Quaternary sands, sanddunes, loess and sandy loess of various thickness (5–55 m).

c) Areas of low undulating topography in the southern part of Transdanubia between Lake Balaton and the river Drave underlain by a Quaternary alluvial sequence of a thickness between 20 and 160 m.

d) In the Hungarian Basin the most important areas where the energy potential is declining with depth are the topographic highs inside the central basin. These areas are underlain by thick Pleistocene alluvial deposits and covered in more than 80 per cent by dune sand, thin loessy sand and sandy loess, all highly permeable.

Packer-tests carried out in pilot boreholes in areas with declining potential showed that a deeper-situated coarse-grained or cavernous carbonate aquifer has a much lower head than expected.

The unfavourable drainage conditions may partly be due to a tectonic process (subsidence) also affecting other areas with high sand percentage. Possibly the major factor accounting for the unfavourable drainage conditions has been the highly different rate of compaction between the areas of low and high sand percentage with the same amount of subsidence.

The sand percentage was calculated from the total thickness of the clean sand and gravel beds of an individual thickness over 1,5 m.

The actual sand percentage is higher than that shown on the profiles because not only the sand-layers less than 1,5 m of thickness but also silty and clayey sand ("dirty sands"), are not included.

3. Areas with an Energy Potential Increasing with Depth

The deepest parts of the Pannonian Basin are typical areas of artesian flow (Fig. 1), where the downward increase of pressure is more than hydrostatic.

Vertical distribution of the potential

The vertical gradients were calculated from heads measured before intensive withdrawal had begun.

Fig. 4 shows the approximate vertical change of pressure between 100 and 400 m depth below ground.

The 400 m depth is a practical limit, because:

1. Down to about 400 m depth the change in water density is negligible.
2. Depth of the intermediate flow-regions exceeds only exceptionally 400 m below ground.
3. Scarcity and areally uneven distribution of artesian head measurements in wells deeper than 400 metres.

The calculation of vertical gradients out of head measurements in wells over 400 m of aquifer depth would have distorted the picture owing to the following geophysical conditions of the basin:

1. The average geothermal gradient of the Pannonian Basin is 0.055° centigrade, two and a half times higher than that of areas around it.
2. The geothermal gradient varies greatly within short distances.
3. Occurrences of gaseous artesian water over large areas of the Great Plain (Fig. 1).
4. Abrupt changes in the gas content of the water within short distances.

Conclusions Drawn from Potential Distribution

1. The areal distribution of strong positive hydrodynamic gradients (Fig. 4) coincides broadly with those portions of the cross-sections where the equipotential lines are parallel to major geologic features. The most striking one of this kind is the base of the Quaternary formation. Where the groundwater is stagnant the upward-moving amount of water is equal to the amount of water pressed out due to the compaction. Water pressed out is moving upward either locally along tectonic disturbances or regionally by very low-velocity seepage through the confining strata. It may be objected here that the seepage velocity parallel to the bedding is many times higher than that perpendicular to the layer. It is true but large amounts of water must have moved upward considering the long time that has elapsed since surface and subsurface geometry and lithology of the Pannonian Basin have become very similar to the present-day pattern.

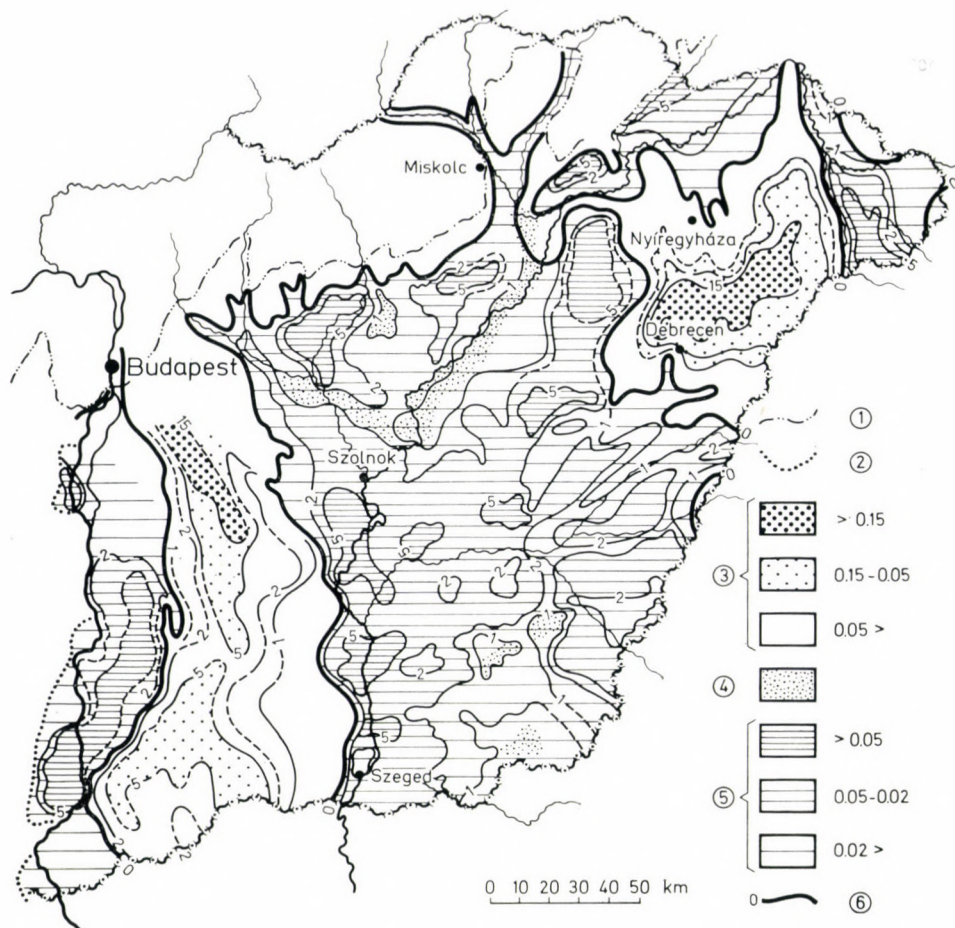


Fig. 4. Map of the vertical hydraulic gradients, Great Hungarian Plain, referring to aquifer depths between 100 and 400 metres below land-surface (ERDÉLYI, 1973). 1. Boundary of the hills; 2. Boundary of the Great Plain; 3. Negative vertical gradients; 4. Area of low near-surface vertical gradients; 5. Positive vertical gradients; 6. Neutral pressure gradient

The existence of stagnant deep groundwater is also indicated by portions of the piezometric contour map (Fig. 2). The "island" of the 85 m contour line east of the Tisza river is an evidence in itself. Another proof is the 10 m difference in potential between the contours of 95 and 95 m over a distance of 150 km along the middle stretch of the river Tisza. This drop in potential is equivalent to a 0.0007 horizontal hydrodynamic gradient.

2. Low vertical gradients indicate a high seepage velocity. An extreme case is the almost hydrostatic pressure distribution within the thick gravel of the Danube river in the Little Plain (Fig. 5).

3. Equipotential lines indicate the small role of the rivers in the natural discharge of the basin. The main natural discharge of the plains are not made

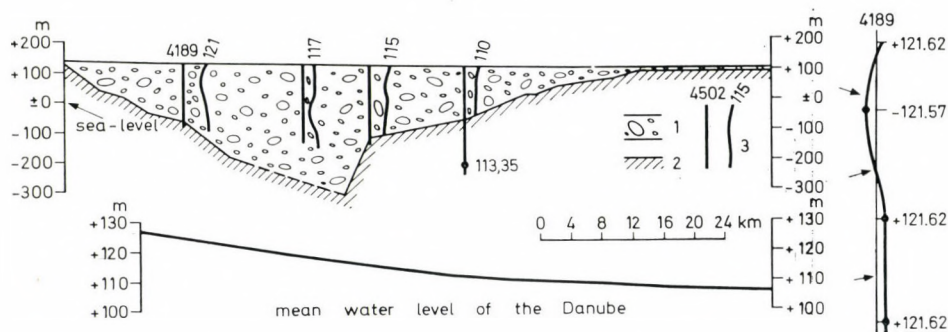


Fig. 5. Hydrodynamic cross-section along the Danube, Little Hungarian Plain. 1. Quaternary gravel; 2. Pliocene; 3. Exploratory borehole and equipotential contour

by the rivers, they are the large deep-lying undrained or poorly drained flats, (Fig. 2) covered by fine-grained deposits.

4. A cross-section shows how the recharge part of an intermediate flow-region is "floating" on the thermal water-bearing regional flow-region (Fig. 3 and 6).

5. A belt of near-surface negative vertical pressure gradients inside a region of positive gradients coincide always with an area where:

a) The shallow groundwater-bearing sand is cropping out in long strips and in patches from under the poorly permeable or impermeable soils. The phreatic water infiltrates through the sand outcrops into the aquifer and dilutes the upward-moving water. The consequence is a dynamic equilibrium between upward- and downward-moving parts of the flow-region causing a weakening of the artesian-bound trend, eventually changing the positive gradient into a weak negative one in the upper portion of the section (Fig. 7).

b) The river bed has cut through the impermeable surface layer and is draining and recharging the subsurface water-bearing beds underlying the impermeable cover. The river drain intercepts the upward-moving highly saline water.

Chemical Aspects of Groundwater Flow-regions and Tracing the Path of the Subsurface Water by Means of Chemistry

Szádeczky-Kardoss started investigations into the origin and nature of deep brines in the Pannonian Basin.

The formation water of most sedimentary areas increases in mineral concentration with depth excepting the shallow groundwater.

An important source of salinity may be deep brines ascending along faults or other "weak" zones or through an upward seepage of low velocity in the deep depressions.

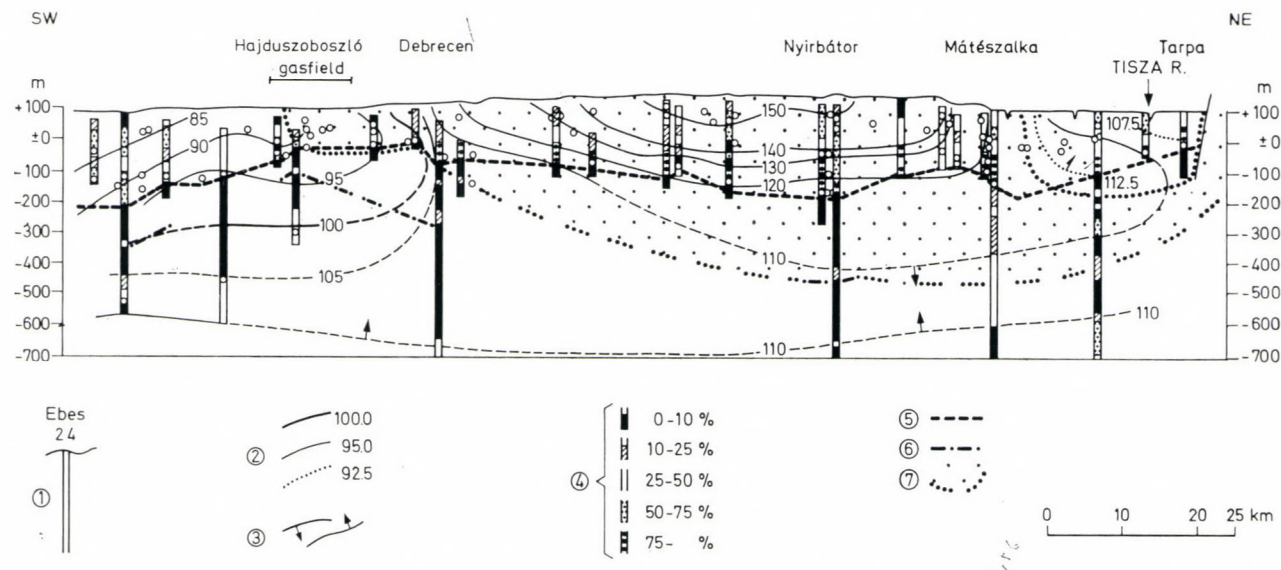


Fig. 6. Cross-sections in NE Hungary: equipotential (A) and hydrochemical (B). A: 1. Drilled well; 2. Equipotential lines; 3. Neutral zone between regional and intermediate flow-regions; 4. Sand percentage; 5. Quaternary and Upper Pliocene interface; 6. Top of the Upper Pannonian (Pliocene) formation; 7. Cross-section of the recharge zone. B: Total hardness: 1. more than 10; 2. between 10 and 5; 3. less than 5 German hardness grade; 4. Impermeable deposits; 5. fresh-water and brine interface; 6. 10 mg/l chloride content of the groundwater 7. chloride content (mg/l) in the indicated depth

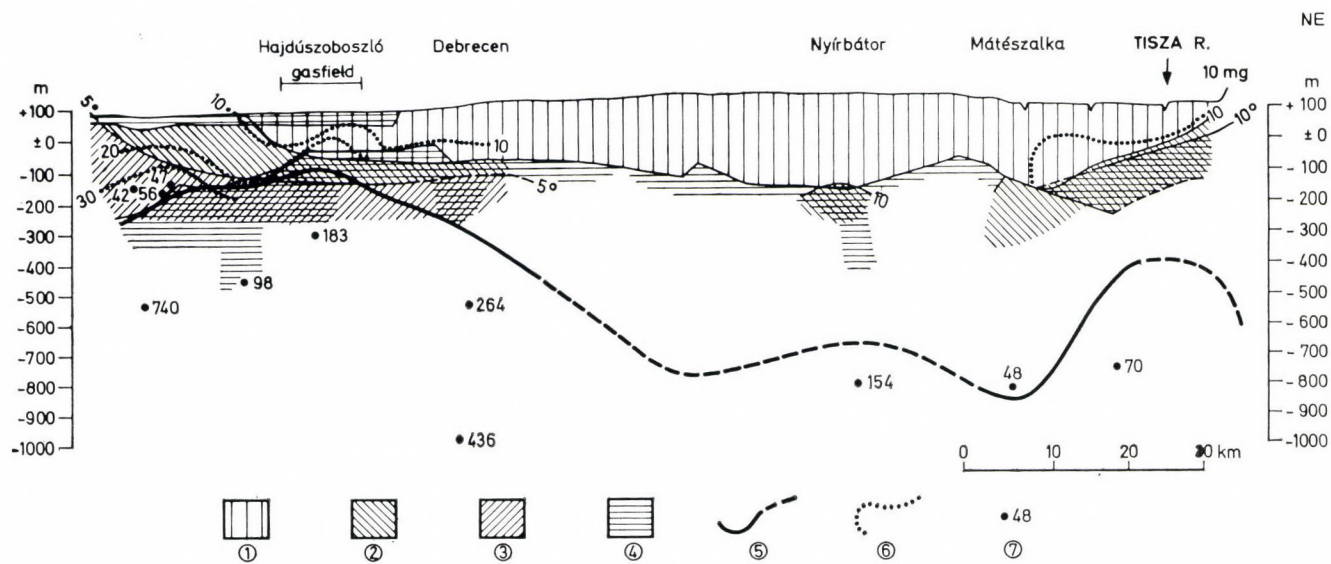


Fig. 6b.

The high percentage of fine-grained deposits, the general low-angle dip or horizontal bedding of the strata, consequently, low seepage velocity and time for a high rate of dissolution, furthermore the low topography, the comparatively large area covered by impermeable soils, and the continuous upward movement of deep groundwater, are factors that may account for the high salinity of the water of all horizons in most of the sub-basins.

Hydrodynamic conditions of the groundwater flow-regions are in complete accord with the hydrochemical concept. Precipitation water enters the soil of the topographic highs, recharge areas of the groundwater flow-regions. The bicarbonate type water moves down and away from the recharge areas (Figs 6 and 7). A zone of dynamic chemical equilibrium has been formed between the downward-moving colder bicarbonate water and the warmer salt water ascending from greater depths (Figs 6, 7 and 8). The result of the mixing of the two kinds of water is an intermediate sort of water in the transition zone with lower hardness and higher salinity (Fig. 7). The depth, shape and *thickness* of the transition zone is a result of many factors: geometry of the layers, lithology of deposits, topography and the infiltration capacity of soils, and climatic conditions.

Fresh-water is understood here to be a water containing a total of dissolved solids not exceeding 1000–1100 mg/l, chloride 100 mg/l and sulphate 100 mg/l. Figs 7 and 8 are also a proof of freshwater movement to great depths.

The central sub-basin is also characterized by a very thick and highly permeable Quaternary and Upper Pliocene (Figs 7 and 8) fluvial sequence deposited by a big river, the ancient Danube, in an area of rapid subsidence. The low salinity of the thermal water of the area testifies to the existence of a high-velocity deep groundwater flow (Figs 8 and 9).

The sub-basins of the Great Plain lie between the recharge areas outside and inside the main basin (Fig. 1). The hydrodynamic centre of each one is the area where all the adjacent flow-regions meet the upward-moving water of each circulation system (Fig. 7).

Characteristics of the centres are:

1. Lack of drainage shown by island-like or very wide-spaced piezometric contours.
2. Equipotential lines running parallel to the horizontal or low-angle dipping beds.
3. High values of vertical hydraulic gradients.
4. A thin freshwater-bearing formation (Fig. 9).
5. High salinity of the artesian water.

Thickness of the fresh-water-bearing formation is exclusively a consequence of the hydrodynamics of the basin. An area of poor porosity may coincide with a thick freshwater-bearing formation and conversely, a thin freshwater formation may be one of high porosity.

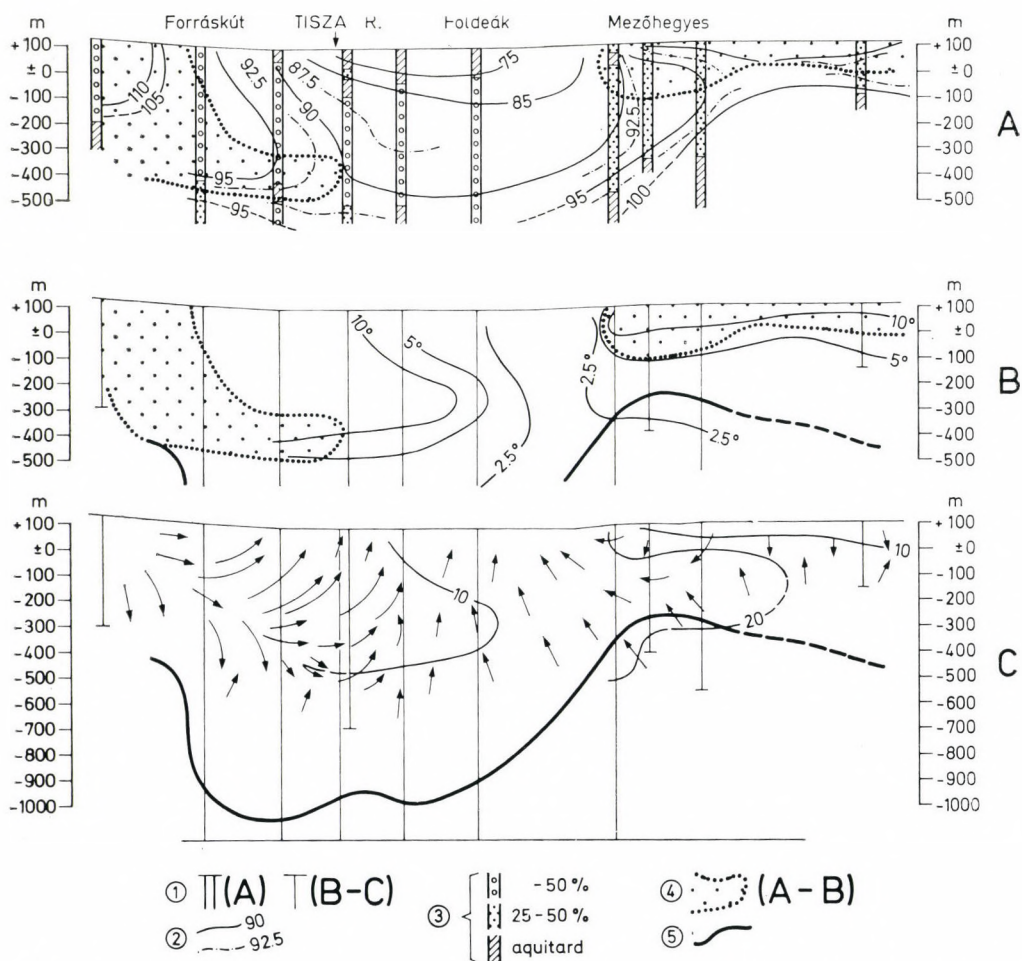


Fig. 7. Cross-sections in S Hungary: equipotential (A), total hardness of the water (B) in German hardness grade and chloride content of the water (C) in mg/l. 1. Drilled well; 2. Equipotential lines; 3. Sand percentage; 4. Recharge zone; 5. Fresh-water and brine interface

The area of the *gaseous wells* (depth of aquifers between 100 and 400 m below ground) roughly coincides, as expected, with the area of downward-increasing potential. There are no gaseous wells where the permeable surface deposits are underlain by high-porosity fluvatile deposits because here the gas is pushed aside by the downward moving water.

Lack of river drain or that of outcropping fluvatile sand may lead to salt accumulation in the topsoil (soda soils) of the sub-basins characterized by high vertical gradients. Whenever the pressure head of the groundwater lies in the capillary fringe of the poorly permeable surface deposit during the hot and dry months, i.e. the capillary fringe is thicker than the depth of the ground-water table, salts are accumulating in the topsoil and on the ground.

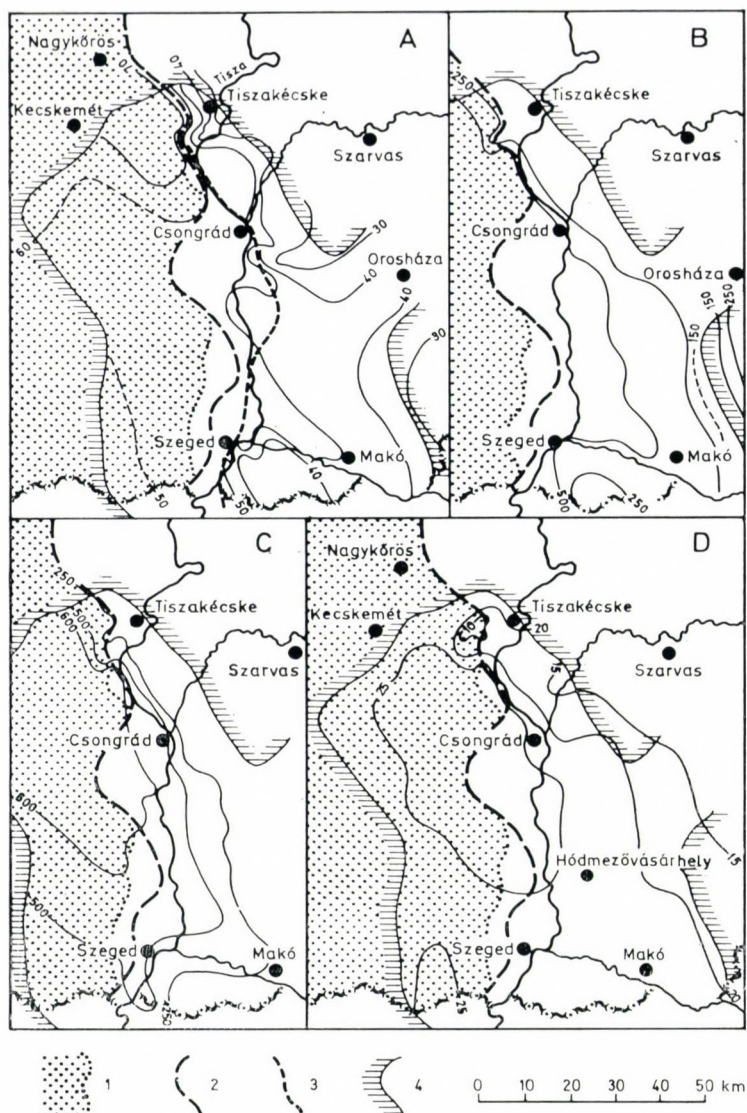


Fig. 8. Sketch maps of the Middle Tisza region: (A) sand percentage, (B) depth of groundwater over 10 German hardness grade, (C) depth of groundwater over 10 mg/l chloride content, (D) reciprocal geothermal gradients. 1. Recharge area; 2. Neutral pressure at 150–200 m depth below ground; 3. Same at 400–600 m depth; 4. Boundary of the uniform deep flow-region

The upward-moving water of the discharge areas is of dual origin, partly moving up in the flow-regions and partly being squeezed out of the compressed deposits. Both parts increase the salt content of the near-surface water in an unidirectional traffic of salt transportation.

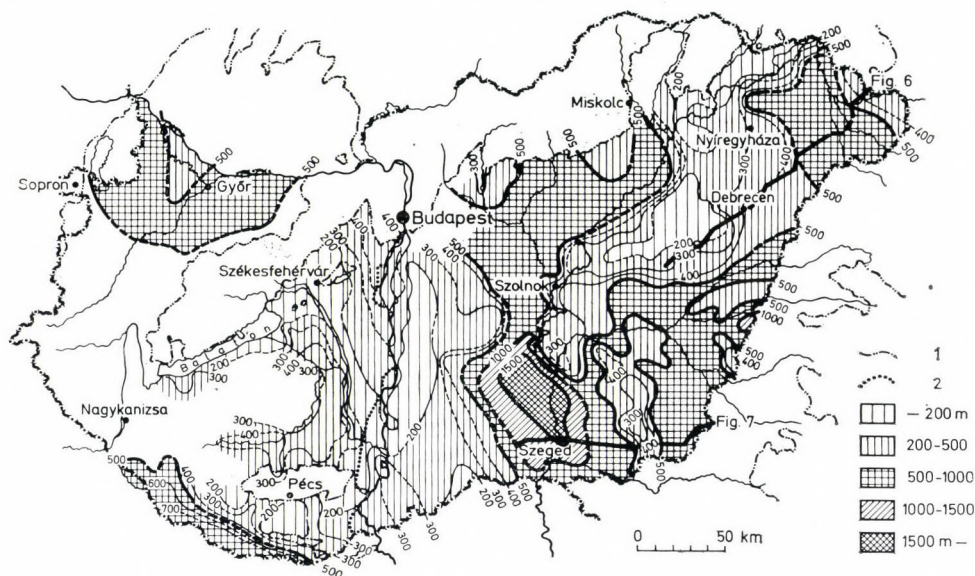


Fig. 9. Thickness of the fresh water-bearing deposits in Hungary (ERDÉLYI, 1972). 1. Boundary of the mountains; 2. Western border of the Great Plain

Deep Groundwater Flow and Geothermy

A flow-region represents a thermal circulation in equilibrium under natural conditions. The equilibrium has come into being between the descending cold bicarbonate water of the recharge area and the ascending salty water of the flow-region (Figs 7 and 8).

The descending limb of the thermal circulation system in the Middle-Tisza region lies on the Danube–Tisza watershed which is the outcrop area of a thick and highly permeable aquifer formation (Fig. 8). The large quantity of cold water seeps to great depths and pushes the warmer water toward the eastern half of the region.

The discharge belt of the upper few hundred metres of the flow-region is manifested by a zone of high vertical hydraulic gradients parallel to the Tisza river (Fig. 4). The water in the deep lower part of the flow-region is being forced to ascend owing to both the eastwardly decreasing grain-size shown in lower sand percentages (Fig. 8A) and the structural elevation of the deep underground which separates this sub-basin from the Eastern one (Fig. 11A). The eastern belt of the Middle Tisza flow-region is also the area of an ascending deep-flow coming from the Transylvanian margin of the Great Plain (Fig. 11D). The merging of both ascending limbs is manifested in high salinity (Figs 7, 8 B–C and 9), in natural gas occurrences (Figs 1 and 11) and higher temperatures of the deep groundwater (Fig. 8D). The positive thermal (Fig. 8D) and chemical anomaly (Figs 9 and 11) is here locally higher along

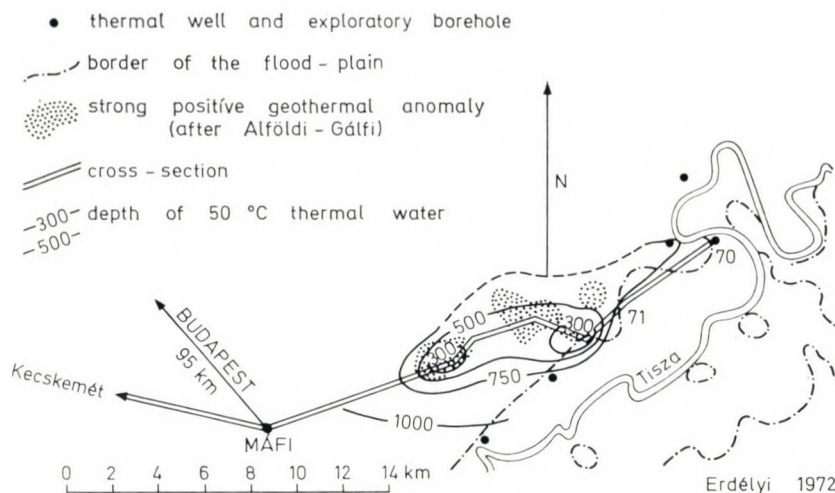


Fig. 10. Cross-sections of the strong positive geothermal anomaly in the Middle Tisza region (Tisza-kécske). A: Potential distribution; B: Total hardness in German hardness grade; C: Chloride content of the water (mg/l); D: Temperature distribution

the NNE margin of the flow-region where thermal water is forced up by the strong artesian pressure from under the confining regional Upper-Pliocene formation (Fig. 4). An exceptionally strong thermal circulation around the village of Tisza-kécske, well known since the 1920's (SÜMEGHY), is shown in Fig. 10.

Deep Groundwater Flow and Locating Hydrocarbon Occurrences

Hydrodynamics is becoming more and more an important tool for locating not only ground-water but also hydrocarbon resources.

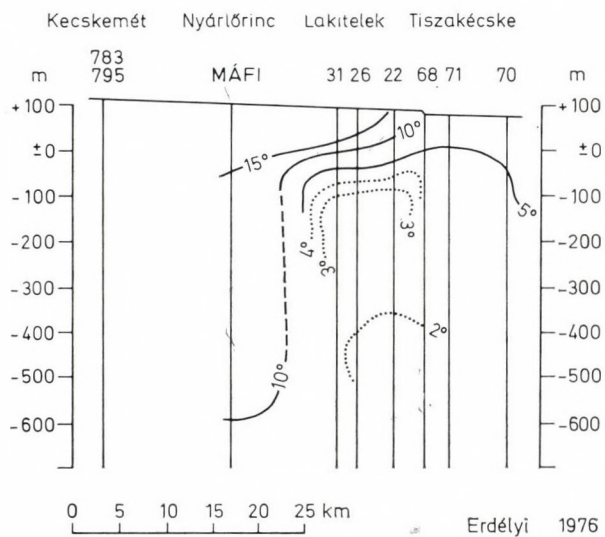
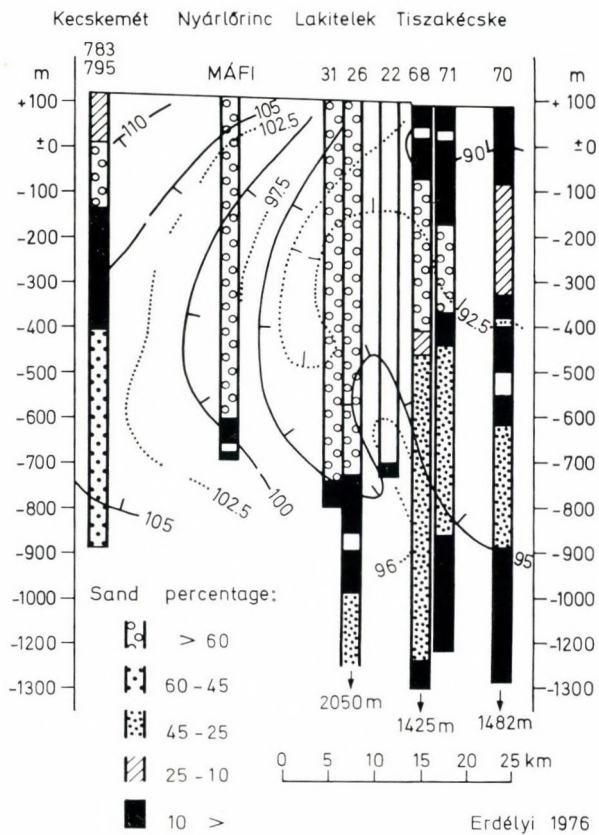
Study of regional ground-water hydrology prior to drilling may greatly reduce the cost of hydrocarbon explorations. Notably, the existing distribution of hydrocarbon accumulations is related to the present pattern of groundwater flow, which, in turn, can be realistically approximated by relying on the configuration of the water table and on the principle of regional hydraulic continuity.

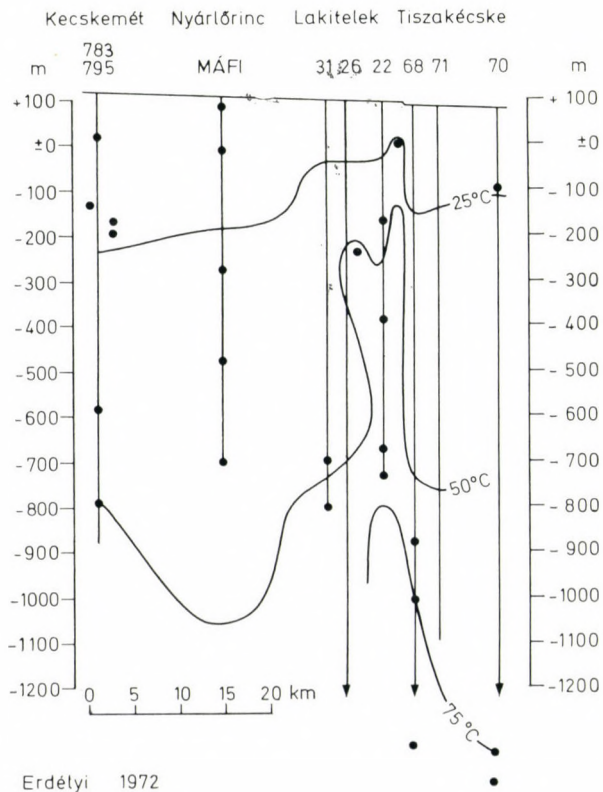
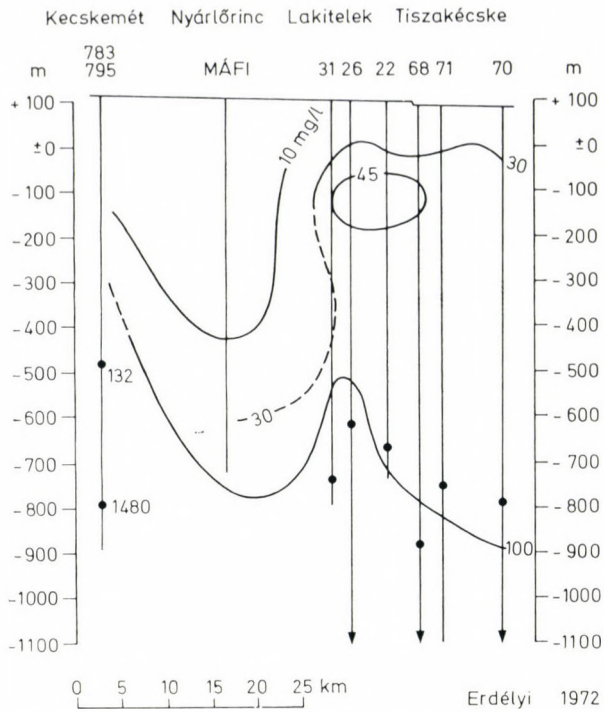
Results of studies indicate that the relative probabilities of hydrocarbons being associated with the most important identifiable flow-pattern features are as follows: ascending limbs and stagnant zones (TÓTH, 1970).

Hydrocarbon deposits in the Great Plain are associated with ascending limbs of the flow-systems (Figs 6, 8 and 11).

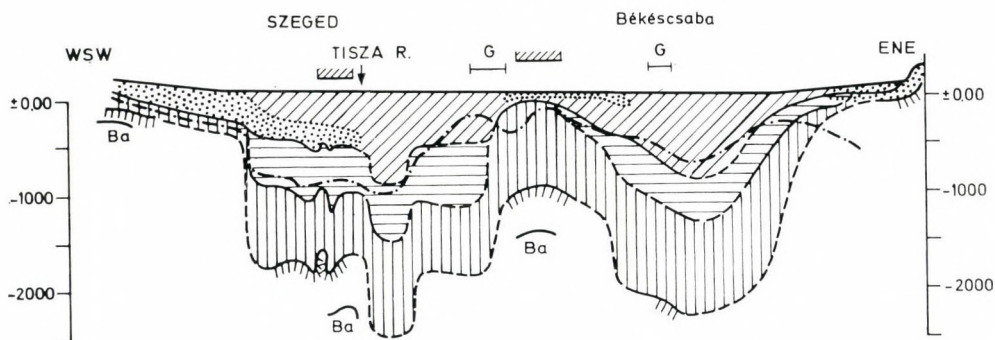
Relying on practical results of the hydrodynamic approach in hydrocarbon prospecting, we may formulate some conclusions as to the Pannonian Basin, too.

There are great vertical differences, even within short distances, in the configuration of the basement complex in the Pannonian Basin both in out-

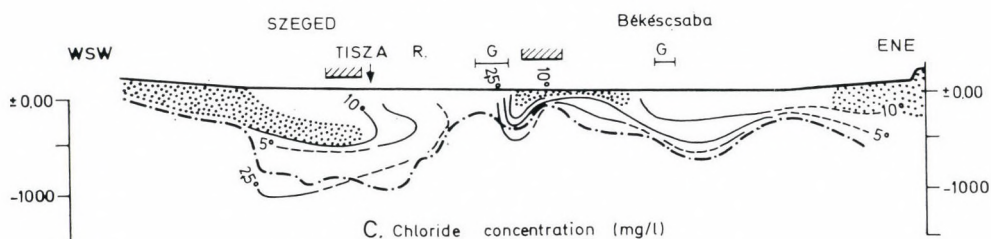




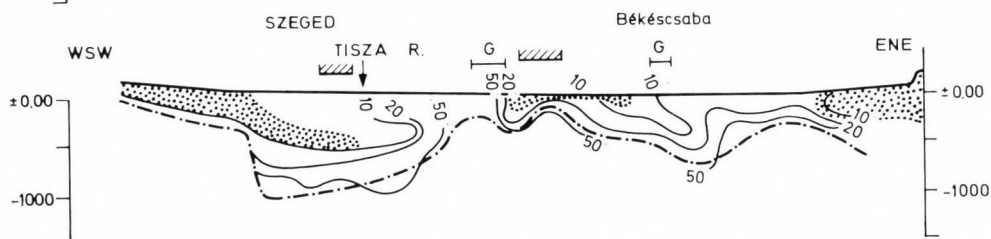
A. Geology



B. Total hardness (German hardness grade)



C. Chloride concentration (mg/l)



D. Schematic potential distribution and flow-directions

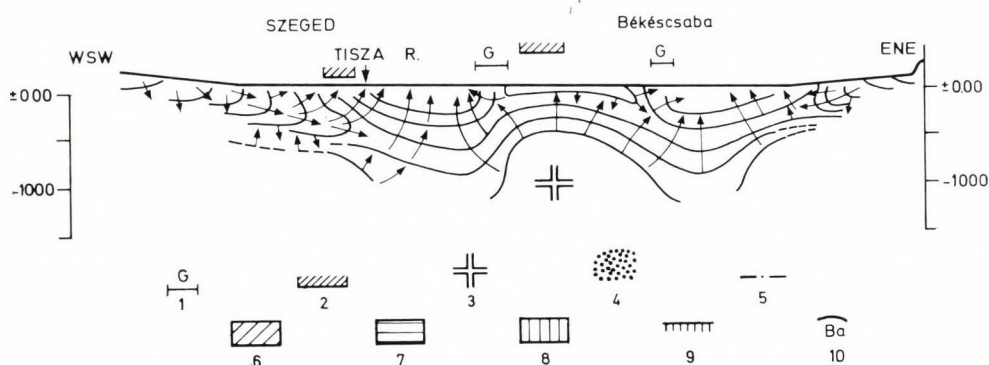


Fig. 11. Diagrammatic cross-sections summarizing structural features, water chemistry and deep groundwater flow of the Great Hungarian Plain. 1. Strong concentration of natural gas in drilled wells; 2. Known hydrocarbon occurrences; 3. Crossing of the hydrodynamic axis of the regional flow-system; 4. Recharge zone; 5. Fresh-water and brine interface; 6. Quaternary; 7. Upper Pliocene; 8. Upper Pannonian; 9. Lower Pliocene, practically impermeable; 10. Basement complex

crop areas and basins to depths about 1000 metres evidenced by coal, ore and hydrocarbon prospecting. It is highly probable that the same geometry of the basement prevails in the deepest parts of the sub-basins where the basin fill has not yet been penetrated by exploratory boreholes. We may suppose that the cover formations, first of all the thick Pliocene deposits, are more or less indistinctly parallel to the topography of the very deep-lying basement because of the differential compaction due to differences in thickness and lithology. Compaction anticlines and synclines are characteristic features in the Pannonian Basin (KÖRÖSSY). There are many known examples of this phenomenon from shallower depths of the Pannonian Basin. We may see this when we look at the great vertical differences in the top of the Lower Pannonian unit in places where many deep boreholes were put down for thermal water utilization, e.g. at Szeged (Fig. 11). There may be many deeper-situated structural features similar to those of Szeged in both sub-basins partly evidenced by a major hydrocarbon deposit at the western border of the deepest part of the Middle-Tisza region. Here the topography of each stratigraphic horizon is a subdued replica of the basement here known from drilling (Fig. 11). Both deep sub-basins are located in areas of the ascending limbs in the regional flow-system of the Great Plain (Fig. 11).

Stagnant zones ("dead cells" or "saddles" located between flow-regions) may exist in a flow-system complex where velocities approach to zero (Fig. 12).

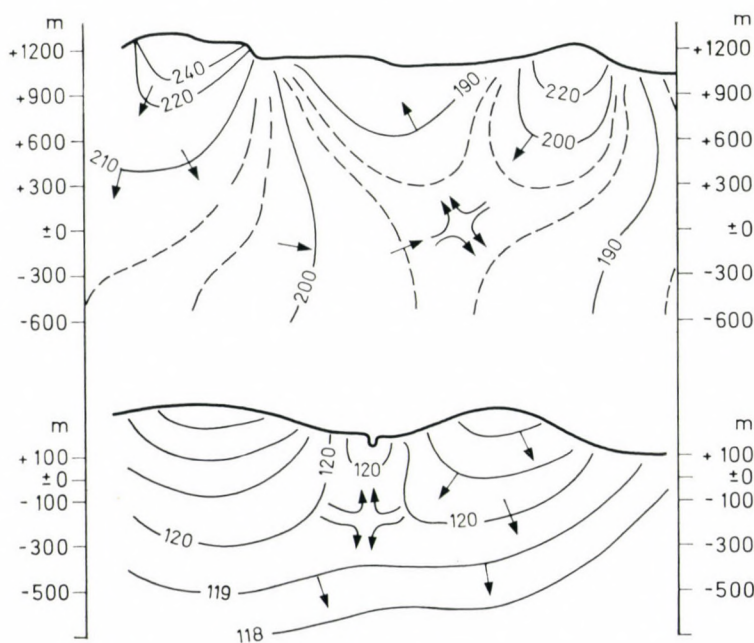


Fig. 12. Examples of stagnant zones in flow-systems

Stagnant zones for hydrocarbon accumulation may have lesser importance in the Pannonian Basin, however, the probability of their existence in the deep south-western Transdanubian basin deserves a new careful appraisal of the hydrodynamic data of the area concerned.

Emplacement in stagnant zones might well mean permanent disposal of some radioactive and toxic industrial wastes as well as saline water.

Interdisciplinary approaches in the evaluation of lithologic, structural, pressure, hydrochemical and geothermal data are indispensable for the understanding of ground-water flow under deep and large sedimentary basins. The author hopes to have given evidence of this concept in this paper abridged from a monograph made ready for publication in 1973.

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ОЧЕРК ГИДРОДИНАМИКИ И ГИДРОХИМИИ ПАННОНСКОГО БАССЕЙНА

М. ЭРДЕЛИ

Резюме

Региональная гидродинамическая система бассейна питается подземными водами окружающих и внутрибассейновых горных массивов. Восполнение водой промежуточных гидродинамических регионов частичных бассейнов (рис. 1 и 3) осуществляется за счет ресурсов прилегающих горных областей и приподнятых участков в пределах бассейна (рис. 1, 3, 7).

Гидродинамические особенности рассматриваемых площадей отражаются распределением их потенциалов. Сильной положительными вертикальными градиенты соответствуют тем участкам разрезов, где кривые эквипотенциалов параллельны главным геологическим пограничным горизонтам. Это один из признаков наличия застойных подземных вод. Другие доказательства: 1. островоподобное положение 85-метровой линии и 2. весьма заниженные горизонтальные градиенты глубоких депрессий (рис. 2). Эти глубоколежащие депрессии являются естественными главными факторами разряда гидродинамических систем низменностей (рис. 2).

Высокопроницаемые формации представляют собою зоны глубинного восполнения подземных вод. Об этом свидетельствуют выпуклости эквипотенциальных кривых (рис. 7) или вертикальные эквипотенциальные линии (рис. 7), отражающие гидростатическое состояние режима подземных вод.

Гидродинамика систем водотоков находится в полном единогласии с гидрохимией и геотермией. Гидродинамическая система это — циркуляция тепла, находящаяся, в природных условиях, в состоянии равновесия. Равновесие установилось между гидрокарбонатными прохладными водами, движущимися сверху вниз и более сильно минерализованными теплыми, восходящими водами. На той же глубине в нисходящей ветви систем стока вода сравнительно более холодная, твердая и содержит меньше солей, чем в восходящей ветви, где она содержит также и газы.

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