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CONTRIBUTION TO THE PROBLEM OF REGENERATION OF ORE DEPOSITS IN ALPINE OROGENIC BELTS (TUNISIA)

Post-variscan lead-zinc deposits of the Eastern Alps, the Carpathians, North Africa and similar mineralizations of Silesia (Mississipi Valey — type) form a worldwide distributed group, characterised by similar features. No genetic relations of the ore-forming processes to igneous rocks are usually observed — the deposits appear to be independent of any igneous activity. The host-rocks are mostly limestones, dolomites and calcareous shales. The ore minerals form disseminations, sometimes following certain sedimentary beds. Metasomatic bodies, fillings of zones of brecciation, of fault fissures a.s.o. are common too.

The mineralogy of these deposits is usually very simple. Galena-poor in silver and sphalerite without higher iron contents are the most important minerals. Colloform textures of the later are not uncommon in certain deposits of this type. Calcite, barite, pyrite, marcasite also occur, while other minerals are rather rare. The content of trace elements is usually low.

In spite of their simplicity the genesis of these ore deposits represents serious problems. They were vividly discussed at the International Geological Congress in Alger and during specialized sessions in München (1956), Klagenfurt (1955), Bleiberg (1959), Leoben (1955).

Regarding the origin the lead-zinc deposits of the Eastern Alps, the following wiewpoints have been expressed:

(a) The deposits are of syngenetic-sedimentary origin.

(b) Exhalative-sedimentary processes related to triassic volcanism are responsible for the formation of these mineralization.

(c) Supergene migration and lateral secretion were the main factors of the concentration of lead and zinc.

(d) The deposits are epigenetic, related to a tertiary magmatism.

(e) The deposits were regenerated during the Tertiary from older variscian deposits.

The theory of syngenetic-sedimentary origin of the east alpine Pb—Zn deposits is based mainly an the fact that the mineralizations are confined to certain stratigraphic horizons (Anisian, Ladinian, Carnian) and are characterized by concordant sedimentary textures.

An exhalative-sedimentary origin is suggested by the coincidence between the distribution of the ore deposits and the occurrence of volcanic rocks (porphyries) in the Anisian, Ladinian and Carnian.

Hegemann (1948, 1960a, 1960b), Schneider (1953), Taupitz (1954), Maucher (1956), Maucher—Schneider (1957), Schultz (1959) belong to those who advocate an syngenetic origin for the lead-zinc deposits of the Eastern Alps.

The epigenetic theory is based on tectonic analyses and evidently discordant mineralizations observed in alpine lead-zinc deposits. These are by syngenetists interpreted as: (a) hydrothermal epigenetic ore depositions in channel-ways of higher laying syngenetic mineralizations. Simultaneous deposition; (b) later mobilizations of primary syngenetic deposits "per ascensum or per descensum".

Petraschek (1926—45), Schneiderhöhn (1941), Colbertaldo (1948—1956) regarded the whole metallogenesis of the Eastern Alps as an epigenetic, unitarian act related to a plutonism of early to middle tertiary age.

Other followers of an epigenetic origin stress the importance of deep-seated metamorphic processes from which the metal contents and the hydrotherms were derived (Angel 1939; Clar 1945; 1955; Clar—Friedrich 1933; Friedrich 1956). These views regarding the role of hydrothermal solutions accompanying the alpine metamorphism are accepted by practically all modern investigators of alpine ore deposits.

An extreme view regarding the metallogenetic province of the Eastern and Western Alps was proposed in 1952 by Schneiderhöhn. The deposits of the Eastern Alps are in this hypothesis regarded as secondary hydrothermal mineralizations regenerated from older, variscian, deposits during the alpine orogenetic processes. The lead-zinc deposits of Bleiberg, Raibl, Auronzo are according to this theory typical examples of regenerated mineralizations (Schneiderhöhn 1953; Colbertaldo—Schneiderhöhn 1958).

Schneiderhöhn & Bolze (1951) studying the lead-zinc deposits of the Teboursouk Mts in Tunisia proposed a secondary hydrothermal origin for these mineralizations. Bolze (1953, 1954) extended the idea about the secondary hydrothermal origin over Pb—Zn deposits of Tunisia and Schneiderhöhn (1954) over those of North Africa. Two different groups of regenerated deposits are distinguished by Schneiderhöhn (1952—54):

1. Deposits regenerated by *epeirogenetic movements*. Deposited in faults

in rocks covering eroded basement (predominantly hercynian) series. The localization of the tectonic elements (faults, flexures a.so) in the cover is mostly identical with those of the socle which have been revived (so called "failles vivantes") during the younger tectonic processes. Deposits transported in this way are termed as *secondary hydrothermal*. The "search for paternity" of the deposits belonging to this group is characteristic.

2. Deposits *of alpine type* regenerated in orogens, recently regenerated. Typical for orogens not fully consolidated that pass again into a younger geosynclinal stadium. Two subtypes were discerned: (a) Deposits of alpine-type regenerated in an *orogene without synorogenic plutonism and subsequent volcanism.* Hydrothermal deposits similar to those of epirogenetically regenerated occur. By metamorphic mobilization of rocks in the regenerated orogene also new deposits may be formed and all transitions between those and the secondary ones may exist;
(b) Deposits of alpine type *regenerated in orogens with mighty synorogenic plutonism and subsequente volcanism.* The intensity of mobilizations and granitization is very strong leading to similar arrangement of deposits as around juvenile magmatic intrusions. The deposits are paragenetically impoverished, monotonous, or characterised by unusual mineral associations resulting from the mixing of components from different sources. In belts of alpine orogeny the problem of regenerated and secondary hydrothermal deposits is a serious one-as results from this short survey. Owing to the very complex history in parts where the synorogenic plutonism was intense, difficulties arise in the deciphering of their metallogenetical evolution. They gradually diminish towards zones where the processes of metamorphism, dissolution and transportation were less intensive.

In North Africa the moderately folded alpine zones with transitions into the outermost subhorizontally laying series of the cover seem favourable to the study of regeneration owing also to the fact that numerous lead-zinc deposits occur there. The north african, and especially tunisian deposits were regarded by Schneiderhöhn (l.c.) and Bolze (l.c.) as typical representants of secondary hydrothermal or regenerated mineralizations.

Detailed investigations of the lead-zinc deposits of the Teboursouk Mts. (Nefate, Fedj el Adoum and El Akhouat) formed one important basis for their hypothesis. An anticline of cretaceous sedimentary rocks is perforated by a triassic diapir. Along the contact the mineralization developed forming true veins metasomatic substitutions of the cretaceous limestones or cementing the breccia of the anormal contact.

The succesion is: galena, sphalerite, marcasite-pyrite, calcite. Sometimes galena forms after sphalerite, which displays colloidal textures. No quartz is present as gangue.

Schneiderhöhn & Bolze (l. c.) consider that older (variscian) mineralizations existed in the basement rocks before the deposition of mesozoic sediments. The brecciated zones along the contacts of the diapirs that formed at the beginning of the Tertiary allowed free passage to supergene waters during the Eocene and Miocene. They reached the deposits of the basement as hot, chloride-rich waters, dissolved their metal content and redeposited it in the secondary hydrothermal deposits in sediments of mesozoic to pliocene age. The age of the mineralization (secondary mobilization) is younger than the orogenic phase which took place after deposition of a continental series comprising pontian and doubtless also pliocene sediments.

The problem of secondary hydrothermal and regenerated mineralizations has been treated from many viewpoints. From isotopic investigations of ore lead follows that the unitarian theory of metallogenesis is for the Alps and Carpathians untenable. The lead of the triassic lead-zinc deposits (Tab. 1) displays a quite different isotopic contitution (older model ages) than the lead from mineralizations showing closer relations to the young tertiary igneous activity.

Two different sources of lead must be supposed for both types of deposits. The isotopic patterns of lead from triassic carbonate rocks do not exclude the

Tab. 1

	204	206	207	208
Poniky. Drienok.	1,360	24,98	21,29	52,37
	1,000	18,37	15,65	38,51
	5,444	100,00	85,20	209,62
	1,359	24,99	21,28	52,37
	1,000	18,39	15,65	38,54
	5,406	100,00	85,14	209,59
	1,356	24,97	21,27	52,40
	1,000	18,42	15,68	38,64
	5,429	100,00	85,15	209,81
Ardovo.	1,344	25,08	21,24	52,34
	1,000	18,79	15,81	38,95
	5,359	100,00	84,69	208,69
Píla. Ján Nepomucký.	1,370	24,82	21,42	52,39
	1,000	18,12	15,64	38,24
	5,520	100,00	86,30	211,08
Izdremec. Bulgaria.	1,361	25,01	21,31	52,32
	1,000	18,38	15,66	38,44
	5,442	100,00	85,20	209,21

possibility of a regenerated origin from variscian mineralizations though syngenetic deposits (exhalative sedimentary, for example) would be characterised by similar isotopic constitutions.

It seemed therefore very interesting to carry out isotopic investigations in areas, where according to Schneiderhöhn's hyphotesis the proper hydrothermal mobilization sensu stricto (secondary hydrothermal deposits) prevails over processes of metamorphic mobilization. In such belts the isotopic constitutions of lead from secondary hydrothermal deposits should be practically little affected by the addition of new lead during the transport. The ancient lead pattern of the primary deposits should be roughly preserved.

Several tunisian lead-zinc deposits have been studied with the aim to contribute at least from this point of view to the very complex and interesting problem of regeneration.

The metallogeny of Tunisia

The metallogeny of Tunis is characterized essentially by lead-zinc mineralizations, if we do not take into consideration sedimentary iron-, manganese and phosphate deposits. They are distributed mainly in the north-western part of the territory in the folded zones of the Atlas of Tunis Mts. The sedimentary deposits of phosphates, iron and manganese are confined mostly to the South, with a continually decreasing intensity of folding towards the platforme.

Only one unimportant lead-zinc deposit — that of Dj. Tebaga SW of Gabes — is known from the Sahara.

The territory of Tunisia is almost exclusively built by mighty sedimentary series of triassic to quaternary age.

Permian sediments (schists, limestones, dolomites) appear in the anticline of Dj. Tebaga, and are here mineralized. The Triassic consists mainly of gypso-argillaceous breccias with blocs of dolomites, dolomitic limestones, quartzites to sandstones and rarely of ophites. It is confined predominantly to the NW part of Tunisia. It is important from the tectonic point of view forming diapirs that penetrate through young, mainly cretaceous rocks and are considered as related in a certain sense to the mineralizing processes.

Jurassic rocks are known only from very limited outcrops in NE Tunisia (Djebel Ressas, Zaghouan). They are represented by massive limestones, marls, marly limestones and at places by slates. *Cretaceous marls, limestones, marly limestones and dolomites* belong to the most widely distributed rocks in Tunis.

The Paleogene is represented by marls, marly limestones, limestones and sandstones. Sandstones, conglomerats, marls (gypsous) and lacustrine limestones belong to *Pontian* and extend probably to Pliocene.

Outcrops of eruptive rocks of small dimensions are known only from the NW of Tunis (S of cap Négro, NE of Tabarka, sourroundings of the deposits

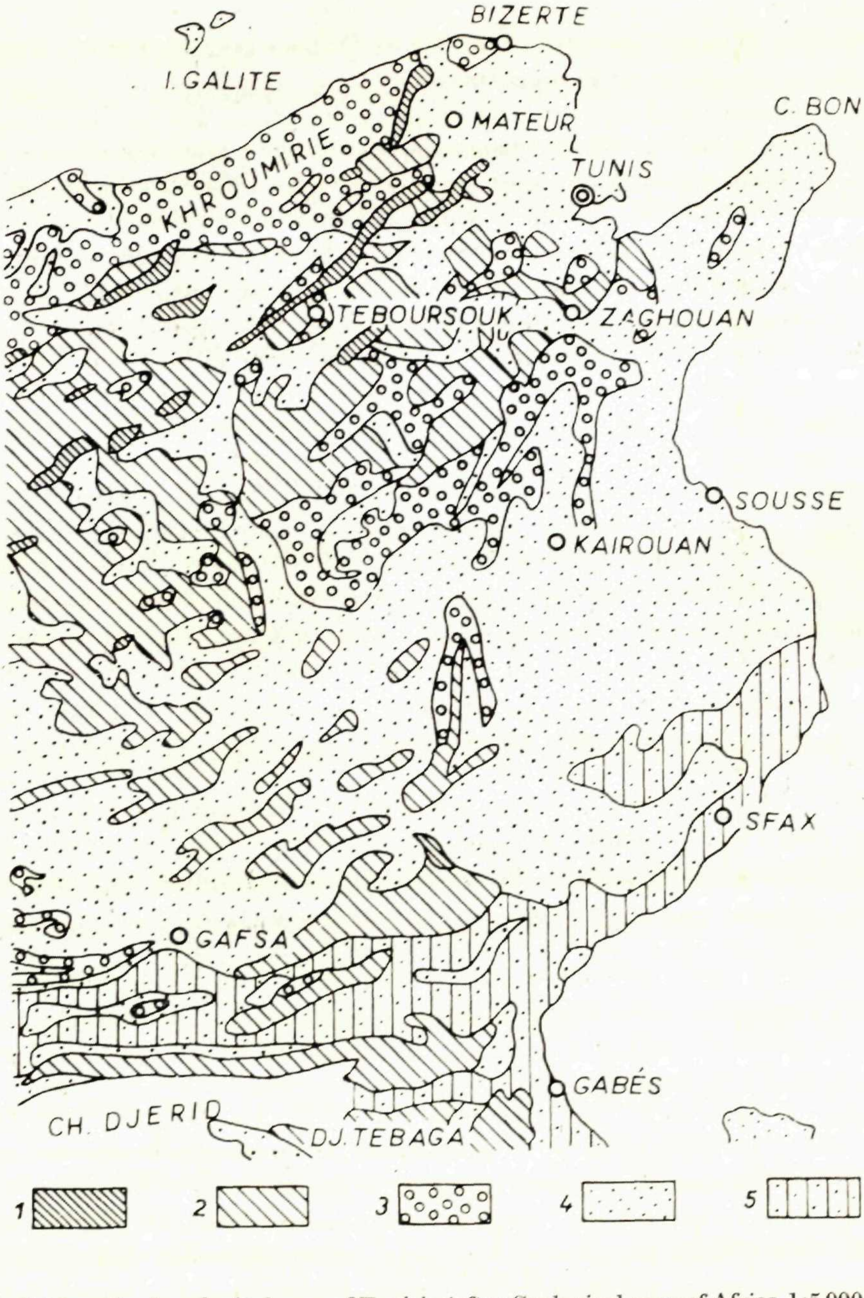

Fig. 1. Schematized geological map of Tunisia (after Geological map of Africa 1:5 000 000; ASG-UNESCO)

1 — Triassic, 2 — Cretaceous, 3 — Paleogene, 4 — Neogene-Quaternary, 5 — Nubic sandstones.

Sidi Driss). They are represented by dacitoide rocks intersecting oligocene sediments. Pontian basal conglomerats at Douaria contain already rollstones of these rocks, from which a middle miocene age of the volcanism was postulated by Solignac (in Sainfeld 1952).

Eruptive rocks (granodiorites and micrograniites) play a more important role in the archipelago of Galite some 50 kms N of the mediterranean coast of Tunis.

A survey of the lead-zinc mineralizations has been recently published by Sainfeld (1952), Gottis & Sainfeld (1952), Bolze (1952), Bolze—Schneiderhöhn (l. c.). Those of North Africa were treated by Bouladon (1949), of Alger by Glangeaud (1935), of Maroc by Bouladon (1952) a. o.

Isotopic investigations of „ore"-lead were carried out on galenas of the following Pb—Zn deposits of Tunisia: Ain Allega, Oued Maden, Djebel Arja, Bechater, Sidi bou Aouane, Sakiet Sidi Youssef, Fedj el Adoum, Slata, Bou Jaber, Djebel Ressas, Djebel Zaghouan, Djebilet el Kohol, Ain Nouba, Djebel Tebaga. A short characteristic of these deposits after Sainfeld (1952) is given below.

Ain Allega. One of the most important deposits of the Flysch-zone situated near the mediterranean coast some 15 kms E of Tabarka. An anticlinal bloc of cretaceous sediments sourrounded by flysch sediments of oligocene age is penetrated by triassic sedimentary rocks.

The main ore-body is localized along the contacts of the breccia of triassic and suessonian marls.

Galenite and sphalerite accompanied by some pyrite and marcasite are the principal primary ore minerals, while baryte with lesser amounts of quartz represent the gangue.

The deposits Sidi Khalifa in the vicinity of Ain Allega carries some mercury too. At Ras Rajel a breccia probably of Pontian age overlaying oligocene marls and sandstones contains dacites, that outcrop some 20 km NE in the Nefza as small bodies.

Oued Maden. As Ain Allega, the deposit of Oued Maden developed in the vicinity of the largest outcrop of triassic rocks in the SW part of the Flysch zone near the algerian frontier. The host rocks of the mineralization are mainly cretaceous (campanian) lime- and marlstones.

The largest ore-body is represented by disseminated lead-zinc minerals accompanied localy by stibnite, bournonite, cinnabarite and realgar.

The fault-vein of Koudiat el Oualahiss to Djebel Groura is regarded as the channel-way of the ore-bearing solutions. It carries lead, zinc, copper and antimony in a barite gangue. At Koudiat el Oualahiss cinnabarite and metacinnabarite impregnations are common in the vicinity of this vein.

Djebel Arja is characterised by a similar mineralization, carrying in places higher amounts of mercury.

Bechater. Ore-mineralizations of this district are situated some 15 km NWN of Bizerta in the immediale proximity of the Mediterranean Sea and of the borders between the Flysch-zone and that of the Atlas of Tunis.

The most important ore-bodies (Djebel el Rhezlane; Djebel el Graya) are localized in a triassic breccia. Zinc prevails over lead. The supergene alteration is very intensive.

Small, unimportant veins carrying Zn—Pb minerals in a calcitic gangue intersect at Sidi el Aoun campanian and at Djebel Soumeur, eocene limestones.

The deposits Sidi bou Aouane and Djebel Hallouf form an important district in the North of the Medjerda. Triassic rocks do not outcrop at this locality but are supposed in shallow depths under the axial parts of the cretaceous anticline of the Djebel in the western part of the district (Djebel Hallouf).

At Sidi bou Aouane the mineralization is developed in paleogene sediments and seems to reach even higher into pontian and pliocene rocks in the southern part of the deposit. Galena, sphalerite, pyrite and calcite are the most important primary minerals.

Sakiet Sidi Youssef. Near the algerian frontier, is essentially a vein deposit. The veins intersect marls and marly limestones (Emscher, Campanian) in the vicinity of a triassic diapir. The succession: pyrite and quartz, marcasite, galena, sphalerite (type "Shalenblende"), calcite.

Djebba. The deposit is situated on the SW border of the Medjerda. Veins and stockwerks are the main types of mineralization localized mostly in turonian and cenomanian limestones. The succession: galena, sphalerite, pyrite. Barite or calcite represent the gangue.

Fedj el Adoum. Vein deposits and stockwerks in cenomaniam limestones, bituminous marly limestones and marls. The mineralization is mostly confined to the vicinity of triassic rocks.

The usual mineral assemblage of: galena, sphalerite, marcasite, pyrite and calcite characterizes the deposit. Barite associated by calcite or celestite represents at places the gangue.

The deposit has been investigated by Schneiderhöhn—Bolze (l. c.) and regarded as a typical example of a secondary hydrothermal mineralization.

Djebel Slata. The deposit situated in SW Tunisia, near the algerian frontier is formed mainly by veins and closely associated replacement — bodies. They occur in aptian limestones, marls and sandstones, exceptionally in albian and turonian limestones and marls. Galena, sphalerite, bournonite, calcite, barite, dolomite and quartz are the primary minerals. The galena is argentiferous.

The deposit Djebel Bou Jaber is situated on the algero-tunisian frontier about 20 kms SW of Djebel Slata. It is prevailingly a replacement deposit in limestones of aptian age.

The following succession is reported: calcite, galena, sphalerite, barite, fluorite, calcite. The galena is rich in silver.

Djebel Ressas. About 30 kms SE of Tunis. One of the few localities where jurassic rocks autcrop and are mineralized (massiv tithonic limestones).

The mineralization consists predominantly of disseminations in brecciated zones. Galena, sphalerite (typ "Schalenblende"), small amounts of pyrite and a calcite gangue represent the primary minerals. The deposit his in the immediate vicinity of an important tectonic line.

Djebel Zaghouan. About 50 kms S of Tunis. The geological situation is similar to Djebel Ressas. One of the most important jurassic massivs of Tunis.

The mineralization is confined to jurassic limestones and is formed mainly by replacements. Galena, sphalerite are the usual ore minerals, while the gangue is represented by calcite, baritocelestite and fluorite.

The great NE-SW striking fault of Zaghouan is observable to a distance of about 100 kms.

Djebilet el Kohol. Deposit with similar features as Dj. Zaghouan, laying about 10 kms to SW.

Ain Nouba. Deposit localized far in the South-West. According to Sainfeld (l. c.) a typical metasomatic mineralization in aptian dolomites. Galena, ironbearing sphalerite and barite are the usual primary minerals.

Djebel Tebaga is the only lead-zinc deposit of Tunisia that occurs in the saharian platform and in pre-Triassic rocks. Mineralized are permian dolomites of a weak zone in the basement. It is essentially a vein deposit with local substitutions and fillings of caves.

Galena, sphalerite, smithsonite, calamine, calcite, barite and gypsum characterize this mineralization.

The isotopic constitution of lead from these deposits is given in table 2. Two distinct groups of deposits are distinguished according to the isotopic constitution of their lead.

To the first belong: Ain Allega, Oued Maden, Djebel Ressas, Djebel Zaghouan and Djebilet el Kohol. They are characterized by higher contents of radiogenic lead isotops — their model ages are younger. In this group the lead from Djebel Ressas displays an anomalous pattern of the Joplin-type.

Table 2.

Deposit	204	206	207	208
1. Ain Allega	1,331	25,38	21,09	52,19
	1,000	19,07	15,85	39,21
	5,244	100,00	83,11	205,61
2. Oued Maden	1,334	25,31	21,05	52,30
	1,000	18,97	15,78	39,21
	5,271	100,00	83,16	206,61
3. Djebel Ressas	1,330	25,46	21,03	52,17
	1,000	19,14	15,81	39,23
	5,224	100,00	82,56	204,87
4. Djebel Zaghouan	1,332	25,35	21,06	52,25
	1,000	19,03	15,81	39,23
	5,254	100,00	83,09	206,14
5. Djebilet el Kohol	1,335	25,34	21,07	52,26
	1,000	18,98	15,78	39,15
	5,268	100,00	83,15	205,77
6. Djebel Arja	1,341	25,23	21,08	52,35
	1,000	18,81	15,72	39,04
	5,315	100,00	83,54	207,49
7. Bechater	1,338	25,23	21,10	52,33
	1,000	18,86	15,77	39,11
	5,303	100,00	83,62	207,37
8. Sidi bou Aouane	1,340	25,25	21,09	52,31
	1,000	18,84	15,74	39,04
	5,307	100,00	83,51	207,14
9. Sakiet Sidi Youssef	1,341	25,22	21,11	52,33
	1,000	18,81	15,74	39,02
	5,317	100,00	83,72	207,51
10. Djebba	1,342	25,27	21,11	52,28
	1,000	18,83	15,73	38,96
	5,311	100,00	83,53	206,86
11. Fedj el Adoun	1,340	25,21	21,13	52,32
	1,000	18,81	15,77	39,04
	5,315	100,00	83,81	207,55
12. Slata	1,346	25,20	21,12	52,33
	1,000	18,72	15,69	38,88
	5,341	100,00	83,82	207,66
13. Bou Jaber	1,344	25,19	21,12	52,34
	1,000	18,74	15,71	38,94
	5,335	100,00	83,86	207,81
14. Ain Nouba	1,340	25,18	21,13	52,34
	1,000	18,79	15,77	39,06
	5,322	100,00	83,92	207,84
15. Djebel Tebaga	1,347	25,15	21,13	52,37
	1,000	18,67	15,69	38,88
	5,356	100,00	83,99	208,18
Deposit	204	206	207	208

It must be emphasized that Ain Allega and Oued Maden lie in the Flysch zone and represent the outermost lead-zinc deposits of NW Tunisia.

Djebel Ressas, Djebel Zaghouan and Djebilet el Kohol form on the other hand the eastern boundary of the mineralized zone of the tunisian Atlas and are arranged along the great tectonic line of Zaghouan.

Fig. 2. Lead — zinc mineralizations in Tunisia. 1. — veins; 2 — replacement deposits, 3. — impregnation dep., 4. — complex dep., 5. — smaller occurrences and indications.

Thus the eastern and western boundaries of the lead-zinc mineralization are marked by deposits containing lead with the youngest patterns. They indicate that tectonic elements governing the localization of these deposits are very young, too and the movements at least along the fault of Zaghouan probably still last.

Fluorite does not belong to the common minerals in Tunis. With the exception of Bou Jaber all other occurrences are confined to deposits characterized by the young lead pattern or to their sourroundings. In the area Djebel Ressas — Zaghouan fluorite occurs at: Djebel el Mecella, Djebel Makki, Zaghouan, Djebilet el Kohol, Kef el Azeiz, Hamam Zriba and Hamam Jedidi.

At the western boundary the deposits of Oued Maden and Ain Allega are not known to carry fluorite. Their mineralogy is very complex. Lead and zinc are accompanied by minerals of quicksilver, arsenic, antimony and copper. Fluorite occurs at Oued et Mtak, the NE continuation of the mineralized zone of Oued Maden — Ain Allega.

The average isotopic composition of the first group (with the exception of Dj. Ressas): $Pb^{200};^{204} = 19{,}01$; $Pb^{207};^{204} = 15{,}80$; $Pb^{208};^{204} = 39{,}20$.

The second group is formed by the deposits, Djebel Arja, Bechater, Sidi bou Aouane, Sakiet Sidi Youssef, Djebba, Fedj el Adoum, Slata, Bou Jaber, Ain Nouba, while the lead from Djebel Tebaga is characterized by a slightly different isotopic constitution. The rather uniform pattern of this group indicates an origin from one source. This would be in good accordance with opinions based on geological observations.

In all considerations regarding the genetic conditions of lead-zinc mineralizations in Tunisia the importance of the Triassic is more or less emphasized.

Termier (1895), De Launey (1913), and Berthier (1914) supposed a primary sedimentary mineralization in triassic rocks as the source from which lead and zinc were dissolved and during younger movements deposited in the vicinity of the Triassic.

Berthon (1922) noted the sterility of triassic rocks in lead and zinc and considered that the Triassic plays an important role in the tectonic sense only, allowing the circulation of ore-bearing solutions along its contacts with the surrounding rocks.

Glangeaud (1935) admitted the existence of two types of Pb—Zn deposits in Alger that were formed in a more or less close relation to eruptive rocks. Deposits of the first group were deposited in the Triassic or its vicinity from hydrothermal solutions accompanying the volcanic activity. The second group of "emigrated" deposits was primarily formed in the gypsum an salt bearing Triassic and later dissolved and redeposited near the anormal contact of the Triassic with surrounding rocks.

Schneiderhöhn (l. c.) and Bolze (l. c.) suppose a primary hydrothermal (veins) mineralization of variscian age in the basement series. In the territory of Tunisia no such deposits are exposed at the surface. These deposits were during the young Tertiary to Quaternary transported into the mesozoic and tertiary rocks of the cover. They form here secondary hydrothermal deposits that are according to the authors typical for the northafrican, especially tunisian metallogenic province. The paragenetic inversion in lead zinc deposits, older observations of Schneiderhöhn in Germany and the Alps as well as results of french investigators in North Africa formed the basis of this theory.

To the Triassic is ascribed a double role — a tectonic and a chemical one.

Sainfeld (1952) expressed in his monography about lead-zinc deposits of Tunisia doubts regarding the secondary hydrothermal origin of all deposits and admitted also certain relations to the young tertiary volcanism.

According to Schneiderhöhn's (l. c.) theory secondary hydro thermal deposits in orogens without synorogenic plutonism and subsequent volcanism — to which doubtlessly the territory of Tunisia belongs — were formed mainly by the upward transport of older mineralizations without considerable addition of new portions of metals derived from leaching of country rocks and from metamorphic processes.

The isotopic constitution of lead should be in such cases preserved and the secondary hydrothermal deposits characterized by the pattern of the old primary ones. Schneiderhöhn & Bolze (l. c.) suppose a variscian, Termier, de Launey, Berthier a triassic, and Glangeaud a triassic to liassic age of the primary mineralization. The average isotopic ratios are:

	$Pb^{206/204}$	$Pb^{207/204}$	$Pb^{208/204}$
Ist. group	19,01	15,80	39,20
IInd. group	18,79	15,73	39,00

In the Alps, Carpathians and Balcanides young tertiary deposits are characterized by analogous isotope ratios. The difference against typical variscian lead is remarkable.

If the tunisian deposits are regarded as secondary hydrothermal, derived from variscian deposits, then considerable addition of new portions of lead must have taken place during the transport.

It seems highly improbable that in each case just the amount of young lead has been added, which is necessary to form lead of a young tertiary pattern. The supposition that the northafrican metallogenetic province is characterized by different isotope ratios for the variscan and alpine orogens from those of Central and Western Europe is also untenable. There was no difference observed between variscian deposits of Europe and Marocco.

If the variscian patterns have been preserved in an orogen with intense metamorphism and synorogenic plutonism — as in Marocco, then there are no geological reasons to suppose greater admixtures of young lead in tunisian secondary hydrothermal deposits. On the contrary, their isotopic constitution should be nearly identical to that of primary variscian deposits as they were formed in an orogen representing the initial stages of regeneration.

The lead-zinc mineralizations were emplaced after the last intensive folding in the time span between young Tertiary and early Quaternary (Sainfeld l. c.). These observations are in harmony with model ages based on lead isotope ratios indicating a young tertiary age of the deposits.

As the time of emplacement corresponds with the isotopic constitution a homogenized source of lead derived from large, deep seated, regenerated blocs is probable. As a result the isotopic constitution of tunisian lead is similar to that of central europian young tertiary deposits.

No evidence of a direct transport of variscian deposits from the basement into mesozoic or tertiary rocks of the cover, without supposed addition of large quantities of young lead from other sources, was found. Schneiderhöhn's and Bolze's theory about the secondary hydrothermal origin and the derivation of the metals almost exclusively from variscian deposits seems therefore for the tunisian deposits unprobable and highly simplified. Similar views were in discussions at the Geological congress in Alger (1952) expressed by Bolfa, Agard, Rechenberg, Petraschek and Heseman.

The possibility of a secondary hydrothermal transport of old deposits must, of course be admitted. If it took place in the territory studied, then only in a limited extend and its recognition by isotopic or other geologic methods is veiled by other processes.

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KANTOR JÁN—ELIÁŠ KAROL

THERMO-VACUOMETRIC METHOD OF STUDY OF PRIMARY AND SECONDARY DISPERSION AUREOLES AS A GUIDE TO ORE

Abstract: Fluid inclusions in minerals from hydrothermal ore deposits, primary, secondary dispersion aureoles and barren country rocks have been studied by the thermo-vacuometric method. Examples are given illustrating the possibilities of the method in exploration and prospecting for hydrothermal deposits.

Introduction

Primary, and secondary aureoles of dispersion form the basis of classical methods of geochemical prospecting for ores. Their trace amounts of certain, with the ore-forming processes directly connected indicatory elements are usually determined by sensitive methods of analytical chemistry or in special cases (mechanical dispersion of heavy, resistent minerals in stream sediments, eluvial-, deluvial deposits, in altered wall-rocks a.s.o.) of mineralogy. The resulting geochemical anomalies indicate the probability of discovery of new hidden or buried deposits.

The principles of geochemical prospecting have been in an ilustrative manner outlined for example in the works of Hawkes H. E. and Webb J. S. (1962) and Ginzburg I. I. (1957).

In this paper the possibility of utilizing hydrothermal and fluid inclusions as a guide to ore is discussed. A special thermo-vacuumetric apparatus constructed at the Geological Survey D. Štúr in Bratislava was used in our investigations.

For paleothermometric studies samples in evacuated reactors (10^{-5} mm Hg) can be at a constant rate heated by an electronically controled device. The vacuum and its dropping caused by decrepitation of liquid and gas inclusions is continually recorded by a recording millivoltmeter. Ten samples can be measured in each of the 3 reactors without interruption of the vacuum.
The temperature can be also maintained at any value up to 900 °C if the total volume of gas and liquids released at a given temperature is of interest, as

for example in the study of dispersion halos. The sample is in this case inserted into the heated furnace, and the gases evolved are before the introduction of a new one pumped out. The rate of analyses depends mainly on the time necessary for a good evacuation of the system after each run.
Similar investigations based on the decrepitation method were newly carried out by Jermakov N. P. (1966).

Primary dispersion aureoles

The very complex nature of the ore forming processes is manifested at one side by the often multistadial development of the hydrothermal ore deposit itself, and by the variegated wall-rock alteration on the other side.

From the viewpoint of geochemical prospecting, the interaction of hydrothermal solutions with the adjacent rocks is of prime importance: broad zones of newly formed or altered minerals or the so called mineralogical aureole forms.

During the ore-forming stages of the hydrothermal activity elements, characteristic of the ore are introduced into the altered zone to a distance depending on the character of the hydrotherms, of the wall-rocks, their fracturing, and the mobility of the elements.

Different factors govern the deposition of the introduced metals either close to the ore-body or their distribution far from the ore contact, beyond the limits of the visibles mineralogical aureoles. Distribution patterns of this type are the so called geochemical aureoles, usually utilized in geochemical prospecting.

After having precipitated in geochemical dispersion halos their load of metals, the nearly spent hydrothermal solutions may through micro-fissures and intergranular openings penetrate to a considerable distance beyond the outer limits of geochemical aureoles. They do no more carry indicatory elements, and are also too weak for a chemical interaction that might result in new formed minerals or a visible alteration of the wall-rock. It is impossible to discover halos of this type neither by mineralogical, nor chemical methods. They were designated by Jermakov (l. c.) as „vapourized" aureoles, and the only relic accessible to investigations is represented by liquid, liquid-gas, or gas inclusions.

The outward movement of the mineralizing solutions reaches in this case its maximum. They are therefore of prime interest as a possible useful tool in attempts of a search for hidden, and buried postmagmatic deposits.

A schematic distribution of the different aureoles around an epigenetic hydrothermal ore-deposit according to Jermakov is reproduced in Fig. 1.

The termovacuumetric evaluation of dispersion aureoles is based on temperature differencies in degasification of their minerals (decrepitation of inclusions) in comparision to those of the unaffected country rocks.

The temperature dependent disclosing of liquid-gas inclusions from quartz of an epithermal — (curve 1) and mesothermal vein deposit (curve 2) is shown in Fig. 2. Quartz separated from fresh, unaltered granite does not release any gas or vapour up to 500 °C. Insignificant amounts escape between 500 and 700 °C, while the main degasing starts at about 700 °C (Fig. 2, curve 3).

Quartz from hydrothermal dispersion aureoles occuppies a position between the quartz of the ore-body and the fresh wall-rock. As a rule parts near the contact with the deposit are richer in inclusions than the distant ones. This

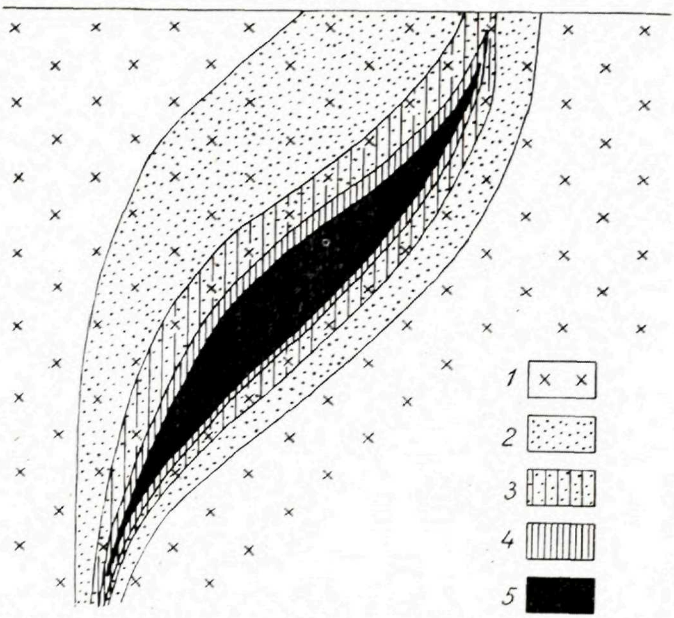


Fig. 1. 1. Granitoids; 2. "Vapourized" aureole with liquid—gas inclusions; 3. Geochemical aureole; 4. Aureole with visible alteration of country rock; 5. Hydrothermale deposit;

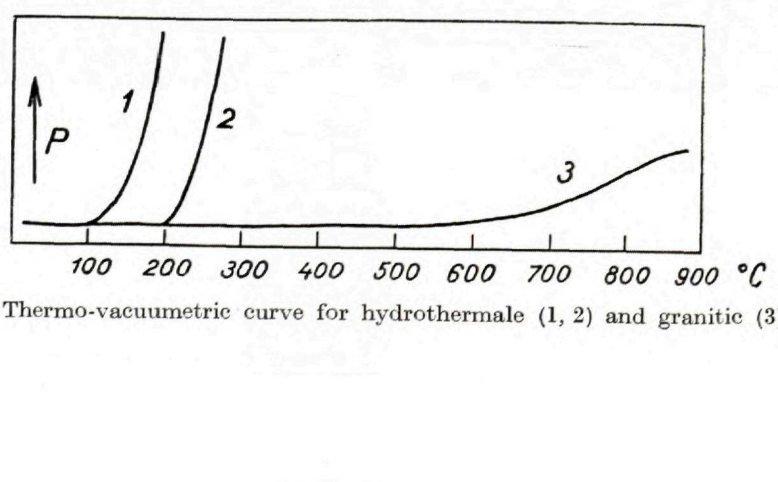


Fig. 2. Thermo-vacuumetric curve for hydrothermale (1, 2) and granitic (3) quartz.

regular pattern being sometimes disturbed by zones of intense fracturing leading from the channel-ways of the hydrothermal solutions to greater distances.

The thermo-vacuumetric curve of a mineral, for example quartz from a hydrothermal vein deposit is mainly a function of the formation temperature (filling ratio of the inclusions), the dimensions of the inclusions, their number and the grain size used in the experiment. Similar principles underlay the disclosing of granitic quartz. In dispersion aureoles the factors mentioned above, the temperature of the solutions as well as the ratio between the two main genetic types of inclusions determine the shape of the thermo-vacuumetric curve.

If there is a distinct lag between the formation of the epigenetic deposit, and the lowest temperature under which the country-rock, for example a granite, consolidated, an optimal heating temperature can be chosen. It allows the depression of the granitic inclusions and a volumetric determination of the hydrothermal ones. The volume of fluids, that ecape at a given temperature, from a given quantity of sample, depends on the character of the inclusions, and the grain size. A certain amount of deep-seated inclusions does not decrepitate even after an intense overheating above the filling (homogenization) temperature. The measured contents are therefore allways lower than the actual ones corresponding to a given temperature.

Minerals evolving upon heating gaseous substances, such as hygroscopic or chemically bonded water, carbon dioxyde a.s.o., are not suitable for thermo-vacuumetric investigations.

Quartz, the most important mineral of both the hydrothermal ore deposits and rocks, if free from admixtures, yields excellent results owing to its stability upon heating even to the highest temperatures of 1000 and more °C.

Primary dispersion halos were studied by the thermovacuumetric method around stibnite veins of the Nízke Tatry Mts.

Lens-shaped stibnite-quartz ore-bodies are located in shear zones intersecting granitoids. The vein filling consists of mylonitized, bleached granitoids, quartz and stibnite. Pyrite is usually present in small amounts, while carbonates, barite, other sulphides and Sb-sulphominerals occur only sporadically. Quartz and stibnite were not deposited uninterruptedly in the shear zones, their lenses being separated by barren parts.

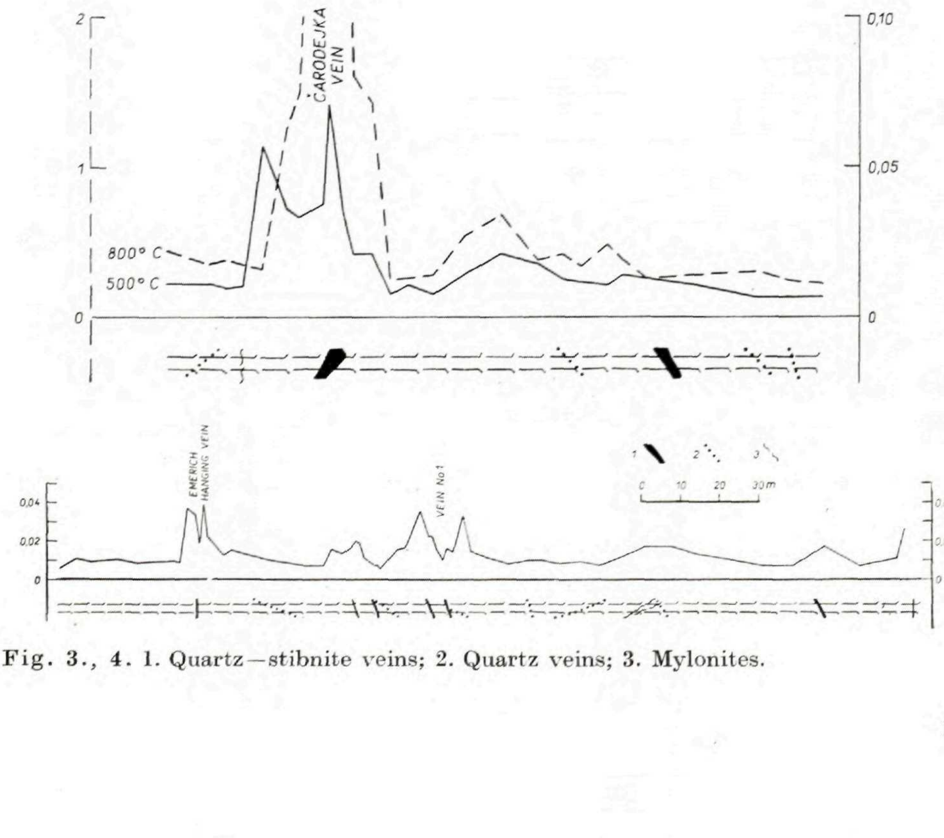
In mines, continuous series of chip samples oriented perpendicularly to the veins were taken for laboratory investigations. The results are shown in Fig. 3, and 4. The ordinate indicates the volumes of gaseous substances released during heating at a given temperature.

Samples of the section illustrated in Fig. 3 were run twice: once at a temperature of 500 °C and the second time at 800 °C. The scale at the left side belonging to the 800 °C, the right side to the 500 °C runs. A similar distribution pattern of liquid-gas contents near the veins is evident for both temperatures, though the volumes are considerably higher in the 800 °C series. This increase is due mainly to the overheating of minute hydrothermal inclusions to such an extent that their decrepitation is made possible. Primary inclusions of granitic quartz contribute only to a limited extent to the measured volumes.

In Fig. 4 a mineralized zone crossed in the mines of Liptovská Dúbrava is shown. The stibnite veins are located in granitoids as in the former case. The samples were heated at 500 °C. The section approaches near its right end a new vein-system, visible in the diagram by an increase of the measured volumes.

The vicinity of veins is in this section allways marked by anomalous concentrations of hydrothermal inclusions. Their maxima, governed by local fracturing systems do not necessarily coincide with the immediate foot- and hanging walls.

The frequency of decrepitations in the vicinity of veins was also measured on quartz by a method proposed by J e r m a k o v (l.c.). The results are reproduced in Fig. 8. A marked increase in frequency is observed with the approaching to the veins. Contrary to the investigations of Jermakov we could not find „non-sounding" quartz in granitoids. Even in barren rocks



F i g . 3., 4. 1. Quartz—stibnite veins; 2. Quartz veins; 3. Mylonites.

the decrepitations started already bellow 500 °C, what may be due to a superimposed, regional, hydrothermal activity in our case as well as to a greater sensitivity of the apparatus used.

Secondary dispersion aureoles

Owing to the necessity of preservation of the liquidgas inclusions during the secondary dispersion, the application of the thermo-vacuumetric method is restricted to mechanical or clastic patterns only. There is no possibility to apply it to the rich variety of patterns formed by chemical processes.

Active stream sediments, including flood-plain sediments and terraces represent them main domain where the survey of the distribution of fluid inclusions may yield positive results in the detection of anomalous patterns related to primary hydrothermal and pneumatolytic deposits.

Residual soils, glacial till and colluvial deposits may also be objects of interest in special cases.

Chemically, and mechanically resistant metals and minerals such as gold, platinum, cassiterite, chromite, columbite-tantalite, wolframite, scheelite a.s.o. in stream sediments form the base of the oldest, still successfully applied prospecting method. The disadvantage of the heavy mineral survey is, that only resistant minerals are preserved, and for mineralogical identification the grain size must attain certain minimal dimensions.

Many important constituents of ore-deposits, for example sulphides, and carbonates, are readily soluble under weathering conditions. Most of the metal. content of primary hydrothermal ore-deposits is therefore during the secondary dispersions transported in natural aqueous solutions.

This load is precipitated and adsorbed by clay minerals of the stream sediments.

Combined investigations, both of the heavy minerals and of the chemism (trace element contents) of the finest fractions in stream sediments are very advantageous. By the first, the pattern of mechanical dispersion, by the second the distribution of the readily (cold-) extractable part of the metal content and the so called hydromorphic pattern may be revealed. Their evaluation considerably rises the probability of detection of new hidden deposits.

But the adsorption of metals in sediments may be very irregular, depending on local conditions. Precipitation barriers may also hinder their downstream movement. The content of heavy, resistant minerals in the eroded parts of the deposits, and their primary aureoles may be quite low and their distribution in stream sediments very restricted.

Quartz, one of the most common constituents of ore-deposits and primary dispersion halos, may prove to be in all these cases an additional object worth to be investigated. It is not destroyed during the transport and its inclusions are well preserved.

By the method used in our investigations is the detection of hydrothermal quartz practically limited by its dilution ratio with granitic quartz. It is favourable in mineralized areas and in the vicinity of epigenetic hydrothermal deposits as demonstrated by examples.

Equal weights (0.1 gr) of samples were used in each run. They were sieve-sized, as the number of decrepitated inclusions depends - under same conditions - on the grain size.

Malé Karpaty Mts. Gold-quartz vein near Pezinok (Hliník).

The sampled area is built mainly by crystalline schists and variscian granitoids (Fig. 5; after B. Cambel 1959). The part of a larger granitic intrusion (left side of Fig. 5) is sterile in hydrothermal mineralizations. A gold-quartz vein outcrops in the apical parts of a granitic cupola surrounded by mica-schists. The mines are abandoned and data regarding extension, thikness, gold contents of the vein very scarce. A mean width of 0,3 to 0.6 m is very probable.

Another complex is represented by metamorphosed basic rocks: epigabbrodiorites, epidiabases and their pyroclastics of Lower Paleozoic age. Exhalative-sedimentary pyrite-pyrrhotite deposits are genetically related to this ophiolitic, submarine volcanism. They were metamorphosed by the granitic intrusion, and in places mobilized. A later, postgranitic hydrothermal activity is manifested by the superimposed stibnite mineralization (right side of Fig. 8).

Results of thermo-vacuumetric investigations of quartz from stream deposits are given in Fig. 8. Frames indicate location of samples, and the numbers in frames the amounts of fluids released at 500 °C. They are gradually increasing stream upwards and reach near the gold-quartz vein their maxima.

By a relative high volume (23) is distinguished the sample from a short tributary. The important mineralized zone of the Kolársky vrch (pyrite-pyrrhotite and stibnite) outcrops near the ridge over it, but almost entirely on the eastern slopes. The quartz-stibnite mineralization attaining some 400 m in striking and about 10—40 m in width is defined to a layer of graphitic schists. From the analysis follows, that at least a part of this ore-bearing zone or its aureole is drained by the small tributary stream.

Nízke Tatry Mts. Stibnite veins S of Liptovská Dúbrava

The area where samples were taken for thermo-vacuumetric examinations is located on the northern slopes of the Nízke Tatry Mts. It is built almost exclusively by granitoids (Fig. 6).

In the northern part the "Prašivá" type prevails. It is an autometamorphic

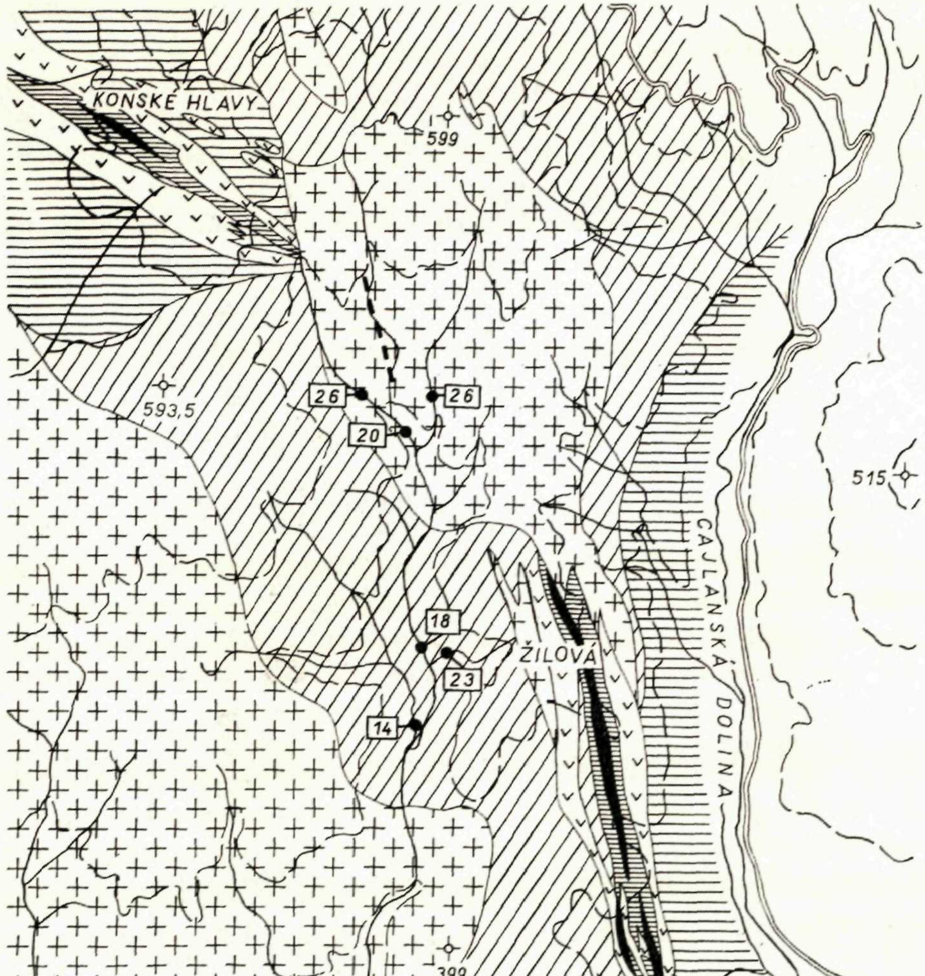


Fig. 5. 1. Granitoids; 2. Phyllites; 3. Graphitic schists; 4. Basic rock and their pyroclastics; 5. Biotite paragneisses, migmatites and micaschists. (Pegmatite dykes); 6. Exhalative—sedimentary pyrite—pyrrhotite mineralization; 7. Au—vein; 8. Samples with results of analyses.

biotite- to two mica granite with porphyritic orthose and microcline. The central part of the mountains is built by the „Ďumbier" granite — a more basic biotite-quartz diorite. In places a biotite granite to granodiorite with por-

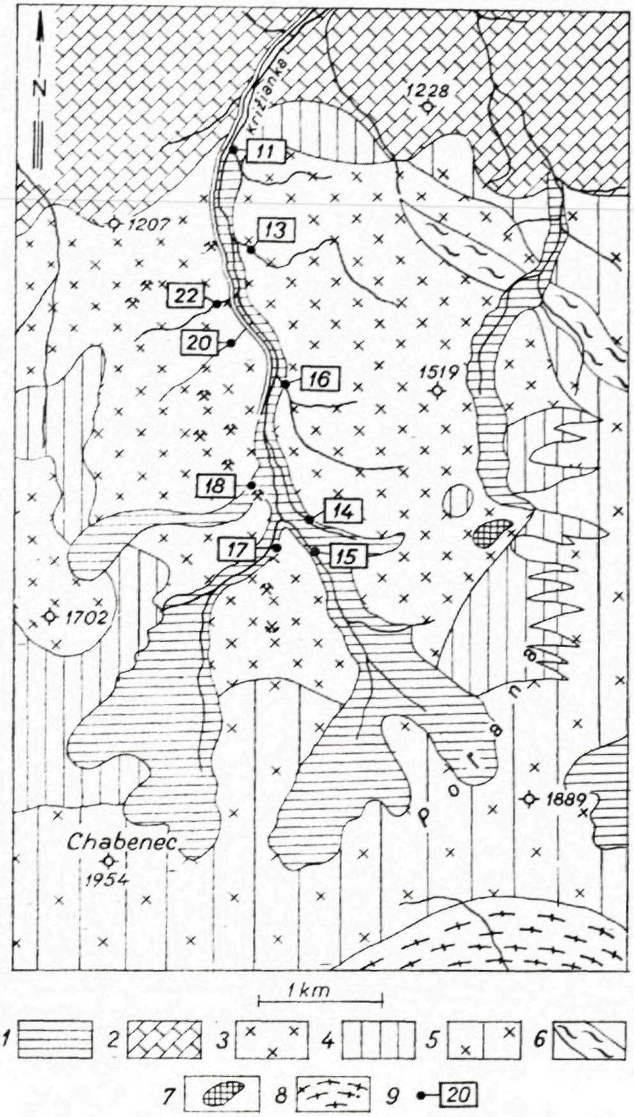


Fig. 6. 1. Fluvioglacial deposits; 2. Mesozoic (sediments); 3. Autometamorphic granites (Prašivá-type); 4. Biotite granites and granodiorites with porphyroblastic K-felspars; 5. Biotite quartz diorite to granodiorite (Ďumbier-type); 6. Syntectic granodiorites to granites; 7. Amphibol— and amphibole—biotite diorites; 8. Migmatites; 9. Samples with results of analyses.

phyroblastic K-felspars developed at the contact of both types by metasomatism of the Ďumbier type. The mesozoic cover appears in the northernmost parts of the territory represented in Fig. 6.

Stibnite veins represent the most important nowadays mined deposit of the Nízke Tatry Mts. The veins are confined to an about 4 km long and 700 m broad nearly N-S striking zone mainly in the leucocratic "Prašivá" granites.

The width of the individual veins varies usually between 0.3—0.5 m. The filling consists mainly of mylonitized granitoids with quartz and stibnite stringers and seams of only few centimeters thikness. In rare cases stibnite lenses attain a thikness of 0.5 m. The distribution of quartz and stibnite is very irregular, lens-shaped. Other minerals, with the exception of small amounts of pyrite, are insignificant, and occur sporadically.

Stream sediments were sampled. The results are given in Fig. 6. The productive zone is well marked by elevated release of fluids from inclusions in quartz.

Vysoké Tatry Mts. Gold and gold-stibnite veins

The SW part of the Vysoké Tatry is almost exclusively built by granitoids (Fig. 7). A type similar to the "Ďumbier type" of The Nízke Tatry prevails. This biotite-quartz diorite is with transitions connected with biotite- to biotite-muscovite granodiorites. Neither pegmatites nor aplites are frequent in the territory. Outside (N and NW) a leucocratic zone of autometamorphosed aplitic and pegmatitic granites forms the uppermost parts of the intrusive body.

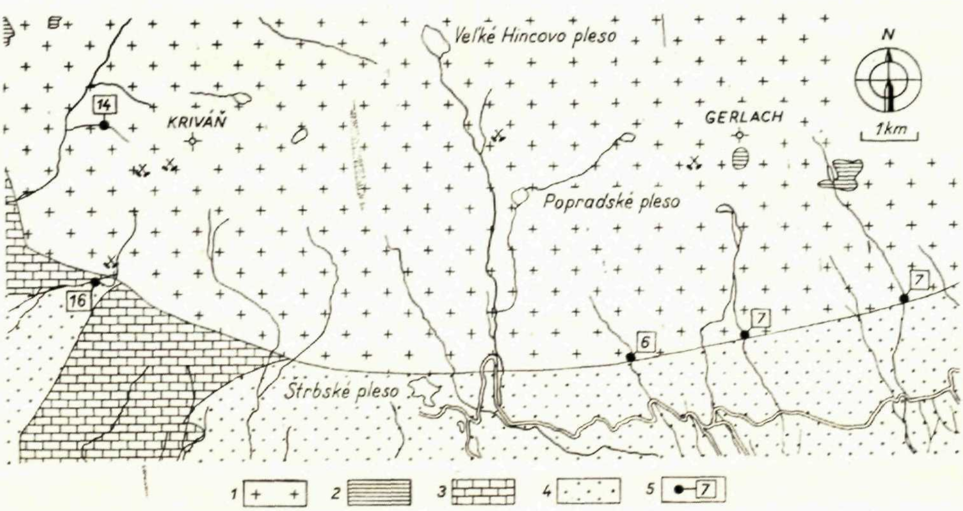


Fig. 7. 1. Granitoids; 2. Crystalline schists; 3. Mesozoic sediments; 4. Fluvioglacial deposits; 5. Sites of samples with results of analyses.

From metallogenetical viewpoint the Vysoké Tatry are to be regarded as almost sterile. The most important mines, since 1773 exhausted, are located on the western slopes of peak Kriváň. Quartz veins with native gold and little sulphides (stibnite, pyrite, galena a.s.o.) are usually thin. In places transitions to pegmatitic veins exist. Veins of this ore-formation were explored under the ridge of Úplaz and the locality Priehyba.

Near Tri Studničky a quartz, stibnite, chalcopyrite, gold mineralization was followed by now abandoned underground works of small extension. Nothing is known about the dimensions, and the metal contents of the vein.

Ore indications of pure scientific interest are known from some other places too: a copper-gold mineralization in the Mengušovská dolina valley and quicksilver-copper indications from the western slopes of peak Gerlach.

Thermovacuumetric analyses of quartz from some stream sediments are given in Fig. 8. In the barren zones of the granitoid body is quartz characterized by low contents of liquid-gas inclusions. Higher amounts were measured in quartz from streams draining the most important mineralizations W and S of peak Kriváň.

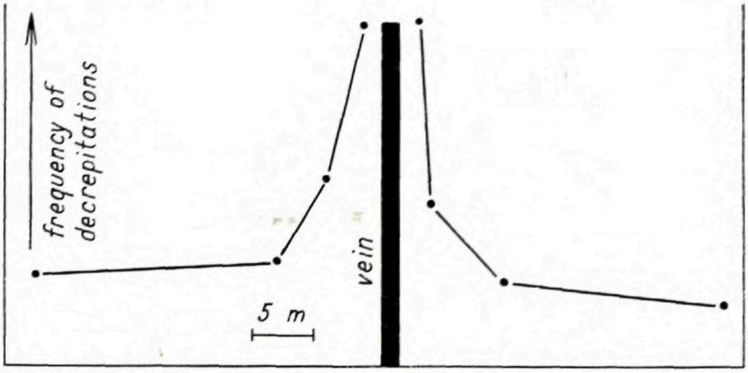


Fig. 8. Frequency distribution (ordinate) of decrepitations in the vicinity of a hydrothermal vein in granite.

Conclusion

The thermo-vacuumetric is a simple, rapid method for the study of liquid-gas inclusions in minerals.

Hydrothermal solutions usually penetrate beyond the limits of visibles mineralogical alterations of the country rock as well as beyond the deposition sites (geochemical aureole) of trace elements genetically related to the ore-forming processes, thus forming the largest aureoles of dispersion. They are there-fore easily detectable by the thermo-vacuumetric survey of rock splits from the surface, underground workings, bore-holes and of residual products of weathering. Quartz is the most suitable mineral. The method yields best results especially in terrains built by granitoids, though it is also applicable in quartzites, sandstones and some other rocks.

Secondary dispersion aureoles are easily detected by the thermo-vacuumetric survey of drainage sediments. Obviously only zones of hydrothermal to pneumatolytic activity-regardless of whether barren or ore-bearing- can be recognized.

A combination of the method wit the geochemical prospecting (trace elements) and heavy mineral survey can therefore secure most reliable results in the prospecting for buried, and blind ore-bodies as suggested by Jermakov (l.c.) for the decrepitometric method.

Examples from the West-Carpathians illustrate the applicability of the thermo-vacuumetric method.

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Г. М. ЗАРИДЗЕ

О ВРЕМЕНИ РЕГИОНАЛЬНОГО МЕТАМОРФИЗМА В РАЗВИТИИ ГЕОЛОГИЧЕСКОГО ЦИКЛА

Рассматриваемый вопрос не нов. Уже в первой половине прошлого века Лайалль (Lyeel, 1833) автор термина „метаморфическая порода“ считал, что метаморфизованные (превращенные) породы образованы за счет древнейших осадочных пород под воздействием внутренней теплоты Земли вследствие их погружения. Вопросы о метаморфизме получили надежную основу после введения в геологию учения о геосинклиналях. Но перелом в разработке теории метаморфизма произошел в начале текущего столетия благодаря работам Седерхольма, Барроу, Бенке, Гольдшмидта, Грубенманна, Эскола, Ниггли, Тилли, Харкера и др.

В геологии укоренилось мнение о связи процессов регионального метаморфизма и гранитизации (ультраметаморфизма) с погружением геосинклиналей. Оно различными авторами модернизируется и преподносится в различном виде, однако суть остается прежней (Заридзе — Татришвили, 1964).

Исследования древних метаморфических образований на территории Грузии дали нам основание заключить, что „*сколько-нибудь существенной роли метаморфизма в период погружения геосинклинального бассейна и седиментации материнских пород . . . не наблюдается*“. „Преобразование пород зависит не от степени их погружения, глубины залегания, а от температуры внедрившегося гранитоидного материала.“ Это преобразование осуществляется „*в стадию образования и воздымания складчатой системы*“ (Заридзе — Татришвили, 1953, стр. 130). Мы полагали, что метаморфизующий и гранитизирующий материал содержал в себе кремний, натрий и калий (Заридзе, 1952).

Следовательно, нами совершенно определенно указывалось, что региональный метаморфизм совершается в основном в стадию складчатости и что в период погружения геосинклинали роль процессов метаморфизма незначительна.

О происхождении метаморфизующих и гранитизирующих растворов мы полагали, что они зарождаются в базальтовой или перидотитовой оболочке Земли не в результате диференциации магмы, привлекаемой обычно для объяснения, а каких-то других более сложных процессов (Заридзе, 1955, стр. 395), которые для нас в то время не были ясны.

Исследования А. П. Виноградова (1959, 1961, 1967), по-видимому, приблизили геологов к решению этой весьма сложной проблемы. Он отмечает, что „вещество горных пород земной коры, вода и некоторые газы на земной поверхности появились одновременно, в результате единого процесса выплавления и дегазации вещества мантии Земли по механизму зонного плавления. При этом произошло расщепление силикатного вещества на две фазы — тугоплавких (дуниты, перидотиты, остаток первичной мантии) и легкоплавких пород (базальты — породы земной коры и легколетучие вещества, вода и кислые дымы)... В ходе этого процесса поднимаются вверх не более легкие, а более легкоплавкие и легколетучие компоненты (Виноградов, 1967, стр. 8). К последним автор относит щелочи, многие литофильные элементы, например U, Ca, Th и многие другие (там же, стр. 4).

Мнение о проявлении двух типов регионального метаморфизма в вулканогенно-осадочных геосинклиналях было высказано Г. Г. Ридом (Read, 1957). Первый тип (орогенический), проявляющийся в раннюю стадию складчатости геосинэлинальной толщи, по степени метаморфизма не превышает зеленосланцевой фации, второй тип регионального метаморфизма (плутонический) осуществляется позднее первого; он связан с процессами гранитизации и обусловлен подъемом высокотемпературных растворов, зарождающихся в симатической оболочке Земли.

Из сказанного следует, что Г. Г. Рид хотя и констатирует наличие двух типов регионального метаморфизма, но отрицает связь метаморфических процессов с погружением геосинклинальных толщ. Это мнение согласуется с нашим, несмотря на то, что мы все же допускали незначительные преобразования в стадию погружения вулканогенно-осадочных геосинклинальных толщ.

Представление о связи регионального метаморфизма с двумя основными этапами развития геосинклиналей, т. е. с ее погружением и воздыманием было высказано Б. Я. Хоревой (1966).

Из отечественных петрологов Н. А. Елисеев (1959, стр. 80; 1963, стр. 79) отмечал, что „нередко региональный метаморфизм связан с горообразовательными процессами...“ и что „нередко с региональным метаморфизмом в тесной связи находится интрузивная деятельность“.

Взгляд о региональном метаморфизме, совершающемся в связи со складчатостью и подъемом магматических масс защищает Н. П. Семененко (1963), отмечая, что „геотермический градиент недостаточен для объяснения метаморфизма“ (стр. 13).

Последующие наши исследования метаморфических образований Кавказа подтвердили в целом наши ранние взгляды.

Нами были описаны многочисленные случаи становления кристаллических сланцев и далее гранитоидов в результате проникновения в исходные породы лейкократового материала, привнесенного восходящими растворами в стадию складчатости. В частности, на Северном Кавказе было описано превращение древнейщих, по-видимому, синийских амфиболитов (З а р и д з е — Т а т р и ш в и л и, 1959, стр. 116). Эти породы инъецированы лейкократовым материалом с различной интенсивностью; в результате возникают полосчатые образования с послойным чередованием черного амфиболита и светлого привнесенного вещества. Дальнейшее превращение приводит к образованию инъецированных кристаллических сланцев, близких распространенным на Большом Кавказе. С увеличением количественной роли привнесенного материала, обусловливающего метасоматические процессы, кристаллические сланцы (биотит-роговообманковые, биотитовые, двуслюдяные, ставролит-андалузитовые и др.) постепенно преобразуются и мигматиты, гнейсы, гнейсовидные гранитоиды и гранитоиды, не обнаруживающие гнейсовидности или обнаруживающие ее в весьма слабой степени, либо участками. Среди гранитоидов, особенно гнейсов, встречаются в различной мере преобразованные уцелевшие останцы кристаллических сланцев. В данном случае становлению метасоматических гранитоидов предшествует образование кристаллических сланцев. Последние формируются при прогрессивном метаморфизме, а образование гранитоидов происходит в относительно более низкой температуре.

Вряд ли можно сомневаться в том, что метаморфические процессы особенно ярко выражены в вулканогенных геосинклиналях (интрагеосинклиналях) и совершаются в стадию их складчатости, т. е. во время формирования вулканогенно-осадочных интрагеоантиклиналей. Таких интрагеоантиклиналей, образованных за счет существующих частных вулканогенных геосинклиналей на Малом Кавказе в альпийском цикле последовательно возникало несколько, с каждой из которых связаны гранитоидные интрузивы: верхнебайосские (Атабек—Славянский и Гиланбирский), кимеридж-нижнемеловые, приуроченные к Шамхорскому и другим антиклинориям (Сомхито-Карабахская зона), эоценовые (Севано-Акеринская и др. зоны) и самый крупный на Малом Кавказе послеверхнеолигоценово-домиоценовый (мегри-ордубадский батолит), приуроченный к Зангезурскому антиклинорию. По нашему мнению, в альпийском цикле развитиа Большого Кавказа, Малого Кавказа и Горного Крыма орогенные гранитоиды не образовались. Региональные метаморфические процессы протекали в связи с становлением частных геоантиклиналей с образованием пород в пределах зеленосланцевой и эпидот-амфиболитовой минеральных фаций.

Мы неоднократно отмечали, что если геосинклинали непроницаемы для магматических масс, то они окажутся непроницаемыми, либо слабо проницаемыми для восходящих растворов, проявляющих особую активность после прекращения магматической деятельности и вызывающих повышение температуры в толще пород.

Исходя из той же позиции о проявлении метаморфических процессов в период формирования интрагеоантиклинальной структуры, мы отмечали, что возникшие при этом натяжения не могут рассматриваться как давление нагрузки. Интрагеоантиклиналь представляет собой сложную сводовую конструкцию — складчатую структуру с серией антиклиналей и синклиналей с развитием в отдельных местах разрывных нарушений взбросонадвигового характера. В этой структуре распределение сил не будет иметь ничего общего с давлением нагрузки. Давление в различных частях складчатой структуры будет различным, вследствие чего на одной и той же глубине в изотермических условиях могут образоваться различные метаморфиты (Заридзе, 1963, 1967; Заридзе — Татришвили, 1964).

Данные о геотермической температуре и давлении нагрузки, приведенные Н. П. Семененко (1963) показывают, что они недостаточны для образования глубокометаморфизованных пород. Так, под сиалической корой Земли мощностью в 25—30 км (граница „М") в пределах платформы, где геотермический градиент определяется 10 град/км температура достигнет примерно 250 °С. Даже в пределах складчатых вулканических областей на примере Карпат, в области вулканического хребта Закарпатской впадины, где геотермический градиент достигает 35,7 град/км, температура на глубине 18—27 км должна повыситься до 640—960 °С, а на глубине 30 км она достигнет 1070 °С. Эти глубины находятся ниже раздела Мохоровичича, ограничивающего область метаморфизма.

Относительно давления нагрузки автор правильно отмечает, что толщи, подвергшиеся метаморфизму обычно имеют мощность 5—10, реже 15 км, которые, по расчетам, должны находиться под статической нагрузкой соответственно 1300, 2700, 4000 атм. На глубине 25—40 км на границе сиалической коры (раздел „М") давление нагрузки повысится 6500—11 000 атм, в то время как по экспериментальным данным для образования дистена необходимо давление в 20 000 атм и температура в 600—900 °С. Для кристаллизации пиропа в кимберлитах требуется давление 20 000—30 000 атм. Альбит переходит в жадеит при давлении 20 000 атм и температуре 900 °С, а для образования алмаза в кимберлитах давление должно возрасти до 50 000 атм.

Приведенные выше примеры метаморфизма для альпийских частных вулканогенных геосинклиналей иллюстрируют преобразовательные процессы, протекающие в одном определенном цикле геологического развития, однако метаморфизм высокой ступени обычно имеет наложенный характер, т. е. совершается в результате повторных преобразований (полиметаморфизм), связанных с более поздними геологическими циклами.

В цепь полиметаморфических процессов следует поставить также и диафторез, фиксирующийся нередко в областях развития метасоматического гранитообразования.

Очевидно, что для толщи пород каждой из упомянутых выше частных вулканогенных геосинклиналей, метаморфизм, которому они подверглись, останется в основном неизменным до конца существования горно-складчатых структур Большого Кавказа, Малого Кавказа и Горного Крыма. То же можно сказать относительно среднепалеозойских зеленокаменных толщ Северного Кавказа, Урала (здесь метаморфизм протекал в каледонском и герцинском циклах — ордовик-конец палеозоя) и др. подобных эагеосинклинальных образований.

Другое положение в древних геосинклинальных толщах, сложенных ныне интенсивно метаморфизованными образованиями — кристаллическими сланцами, мигматитами, гнейсами и гранитоидами, образующими субстрат для более поздних толщ.

Наши наблюдения на Кавказе и анализ фактического материала по другим регионам показали, что метаморфизм вулканогенного и осадочного материала древних геосиклиналей произошел в течение, по-видимому, трех циклов геологического развития (Заридзе, 1966). Например, в феноскандии фиксируется три цикла. Если в качестве примера возьмем кристаллические сланцы Кавказа, то можно утверждать, что исходные их породы представляли собой вулканогенные геосинклинальные образования верхнего докембрия, которые первое метаморфическое крещение получили в стадии возникновения интрагеоантиклинали в верхнем докембрии. Повтороно высокотемпературному метаморфизму эти сланцы подверглись в каледонском цикле. Окончательное их формирование произошло уже в герцинском цикле. В это время метаморфизм имел регрессивный характер. Одновременно формировались микроклиновые, часто порфиробластические (порфировидные) граниты.

Сказанное доказывается сопоставлением докембрийских метаморфических образований с каледонскими и более поздними — среднепалеозойскими зеленокаменными геосинклинальными образованиями, наличием реликтовых участков и минералов, а также определениями аргонового возраста, цифры которого в отдельных случаях дают большой разброс (Заридзе, 1966).

Эти древние кристаллические породы в дальнейшем уже не подвергались заметным региональным преобразовательным процессам, которые могли быть установлены петрографическими методами, если не считать локальных изменений в контакте с секущими их интрузивами и аргонового омоложения некоторых их участков.

Выше рассматривались метаморфические процессы, совершающиеся в вулканогенных геосинклиналях (эвгеосинклиналях) или иначе в геосинклиналях фемического профиля. Вызывают большой интерес процессы регионального метаморфизма и магматизма, наблюдающиеся в терригенных геосинклиналях или в геосинклиналях сиалического профиля.

В настоящее время накоплены, тщательно проонализированные фактические материалы по сиалическим геосинклинальным областям. В них не наблюдается, либо слабо проявлен в собственно геосинклинальной стадии развития подводный вулканизм основного состава, однако в стадии складчастости этих терригенных геосинклиналей формируются вытянутые цепочкообразно вдоль глубинных разломов на сотни километров крупные тела гранитоидов, которым предшествуют массивы основных, иногда ультраосновных пород различного размера. В связи с становлением гранитоидов протекают процессы регионального метаморфизма.

С. С. Смирнов (1946) в составе Тихоокеанского пояса выделял две зоны с различным типом магматизма и металлогении — внутреннюю, расположенную непосредственно вблизи океана, и внешнюю окаймляющую первую со стороны континента. Внутренняя зона характеризуется вулканизмом основного состава и гипербазитовыми интрузиями, а внешняя — преимущественно кислым вулканизмом и редкостью гипербазитовых интрузий. „Переход от более основных к более кислым комплексам совершается в направлении от океана в глубь континента" (стр. 27). Автор отмечает, что в составе мощной (до 10—11 км) Верхоянской геосинэлинальной толщи (пермь-средняя юра) основные эффузивы очень скудны или отсутствуют. То же наблюдается в соответственных толщах Забайкалья. Редки также здесь интрузии базитов и гипербазитов. Синорогенные интрузии обычно представлены лейкократовыми гранитоидами, реже встречаются основные их разновидности (стр. 19). Связанное с гранитоидами оруденение олово-вольфрамовое.

Развивая мысль С. С. Смирнова, Е. А. Радкевич (1959) различает два рудных района в соответствии с геосинклиналями двух типов -- фемическими и сиалическими. Последние развиты в восточной части СССР (Приморье, Забайкалье, Верхоянье).

Прогибы сиалического профиля характеризуются мощными, главным образом терригенными, порой флишоидными толщами. Основные эффузии ранних стадий, если и проявлены, то имеют они резко подчиненное значение. В тех зонах, где основные эффузивы получили значительное развитие, соскладчатые гранитоиды проявляют повышенную основность, с ними связана не оловянная, а медно-молибденовая минерализация и золото, как это имеет место, например, в Гродековском прогибе Приморья (стр. 49).

Мезозойский (древнекиммерийский, киммерийский и новокиммерийский) магматизм Забайкалья и связанная с ним минерализация подробно рассмотрены Ив. Ф. Григорьевым и Е. И. Доломановой (1955). Среди древнекиммерийских и киммерийских интрузивов они выделяют: (1) биотитовые граниты; (2) эндоконтактовые гибридные образования, развитые в контакте с глинистыми сланцами с высоким содержанием меди, магния и железа, это роговообманково-биотитовые граниты, гранодиориты, кварцевые и бескварцевые диориты, граносиениты, кварцевые сиениты или лейкократовые граниты, мигматиты; (3) метасоматические измененные граниты (биотитовые и биотит-роговообманковые) с гигантскими кристаллами микроклина, розовые биотитовые граниты с дымчатым кварцем, двуслюдяные граниты, мусковитовые граниты и грейзены.

В эндоконтактовых гибридных образованиях появляется роговая обманка и, реже, моноклинный пироксен; образуется также относительно основной плагиоклаз (олигоклаз-андезин, андезин). При более интенсивном метасоматозе биотитовые граниты переходят сначала в двухслюдяные, а затем в мусковитовые граниты и, наконец, в грейзены. Минералы биотитовых и роговообманково-биотитовых гранитов замещаются мусковитом, кварцем, топазом, турмалином, гранатом и др. В связи с грейзенами образуются высокотемпературные касситериотово-вольфрамовые месторождения.

Новокиммерийские интрузии встречаются в разных районах редкометального пояса Забайкалья. Они представлены гранит-порфирами, залегающими в виде штоков и даек, и амазонитовыми гранитами, образующими пластовые залежи. Они более кислые и более щелочные, чем ранние гранитоиды и беднее железом, магнием и кальцием. Возраст этих интрузий устанавливается: в Сохондинском районе по пересечению гранит-порфирами верхнеюрских вулканогенных пород и в районе р. Сенькиной — Чикоконской, где древнекимерийские гранитоиды секутся гранит-порфирами.

С новокиммерийскими интрузиями связаны касситеритовые кварцевотопазовые и кварцево-амазонитовые, а также вольфрамито-кварцевые и касситерито-вольфрамито-кварцевые жилы.

Как видно из описания, окончательное формирование древнекиммерийских и киммерийских гранитоидов происходило метасоматическим путем. Образование гранитоидов более основного состава объясняется гибридизмом с глинистыми сланцами с высоким содержанием меди, магния и железа. Из статьи не совсем ясно, почему эти сланцы так богаты названными элементами. Не туфогенные ли они. Не ясно возникновение основного плагиоклаза (андезина) в эндоконтактовых гранитоидах. Если допустить, что глинистые сланцы богаты также известью, то тогда какие же это глинистые сланцы. Не следует ли полагать, что предшествующие гранитоидам интрузии были относительно основными, которые поглощены (ассимилированы и метаморфически преобразованы) гранитоидами. На это указывает также наличие диоритов. Непонятно и высокое содержание меди в глинистых сланцах. Не внесена ли она гидротермальными растворами.

По мезозойским терригенным отложениям Вылюйской впадины и Западного Верхоянья имеется обстоятельная работа А. Г. Коссовской (1962). Здесь выделяется четыре зоны преобразования пород после литогенеза, т. е. осадконакопления и уплотнения, причем характер изменения для платформенной и геосинклинальной областей отличные.

В первой зоне (начальный эпигенез по автору) преобразованные породы по составу и структуре близки к первично-седиментационным породам. Изменения здесь выражены в постепенном исчезновении в разрезе ряда наиболее нестойких обломочных минералов — пироксенов, амфиболов, биотитов (наблюдается каолинизация последнего) и др. и в быстром увеличении в направлении сверху вниз гравитационного уплотнения пород — возрастание их объемных весов и уменьшение пористости. Переработка углей не выходит за буроугольную стадию.

Во второй зоне (глубинный эпигенез) происходит сильная переработка глинистого цементирующего материала (гидрослюдизация с образованием мусковита и биотита) и исчезновение нестойких обломочных минералов: кальциево-натриевых плагиоклазов с возникновением кальциевого цеолита-ломонтита, пироксенов, амфиболов и др. Появляются значительные массы ломонтита. Однако существенного нарушения структур исходных пород не происходит, хотя и возникают мозаично-регенерационные структуры в песчаниках и явления микростилолитизации. В этой зоне гравитационное уплотнение пород как бы заканчивается.

В треьей зоне, которая характерна для геосинклинальной области (стр. 180), породы почти утрачивают седиментогенные структуры, приобретая черты метаморфических пород. Они интенсивно дислоцированы, степень их измененности обусловливается не только глубиной погружения, но и стрессом. За счет песчаников появляются структуры, свойственные кварцитовидным породам с зубчатыми или сложными лапчато-извилистыми поверхностями сочленения зерен. Обломочный биотит интенсивно изменен: пластинки его состоят из сложно-перемежающихся волокон светло-зеленого изотропного хлорита, бесцветного сильно двупреломляющего мусковитоподобного минерала и редких реликтов обеспеченно слабо плеохроирующего и двупреломляющего минерала; в нижней половине зоны гидробиотит полностью исчезает. Постепенное становление мусковита связано с высвобождением калия из обломочного калиевого полевого шпата при его разрушении.

Четвертая зона распространена только лишь в пределах наиболее дислоцированных участков Верхоянского прогиба и охватывает самые нижние горизонты разреза, относящиеся к нижней перми. Для зоны характерно наличие аспидных и филлитоподобных сланцев и кварцитоподобных песчаников со сложными „шиповидными" структурами, выразившимися в перпендикулярных ориентировках пластинок хлорита и мусковита, проникающих в сложно ориентированные зерна кварца и полевого шпата.

В этой зоне породы полностью утрачивают исходные обломочные структуры; начинается формирование сланцеватых текстур, связанных с явлениями сегрегации хлоритово-слюдяных минералов, обособляющихся в отдельные прожилки и крупные лепидобласты величиной до нескольких миллиметров, выделяющихся на фоне основной ткани породы, сложенной агрегатами хлорита, мусковита и кварца.

Таким образом, происходит последовательное изменение глинистого вещества с тенденцией к упрощению минерального состава и сведению его, в основном, к *четырем фазам: мусковиту, хлориту* (афросидерит или рипидолит), *альбиту* и *кварцу*.

Первые две зоны А. Г. Коссовская относит к стадии эпигенеза (диагенеза), при которой завершается гравитационное уплотнение пород и минеральная переработка нестойких фаз с сохранением структры осадочных пород, вторые две зоны — к стадии начального метаморфизма (метагенеза), при которой стираются признаки структур осадочных пород и оформляется минеральная ассоциация, соответствующая мусковитовой субфации зеленосланцевой фации. Стадию метагенеза автор считает переходной к региональному метаморфизму, когда начинается массовое появление биотита (биотитовая зона).

В связи с охарактеризованными зонами следует заметить, что нам кажется недоказанным образование метаморфитов третьей и четвертой зон в период погружения осадочных толщ и мнение об их развитии как в геосинклинальных, так и в негеосинклинальных областях. Это вытекает из характеристики названных зон, которые связаны с геосинклинальными отложениями, секущими стратиграфические границы и с крупными элементами Верхоянского мегантиклинория. Кроме того, эти зоны, особенно четвертая, обнаруживают пятнистое (участковое) развитие. Следовательно, можно полагать, что метаморфизм двух последних зон протекал в стадию складчатости. Поэтому и региональный метаморфизм в Верхоянье лучше начинать с третьей зоны, т. е. когда происходит перелом в геолого-структурной обстановке развития области, где в качестве нового важного фактора метаморфизма выступает стресс.

В связи с рассматриваемым нами вопросом вызывает интерес исследование Б. Я. Хоревой (1963, 1966) по магматизму и метаморфизму Иртышской зоны смятия и по общим вопросам метаморфизма. Классическим примером метаморфического пояса сиалического профила она считает изученный ею палеозойский Иртышско-Маркакульский метаморфический пояс, приуроченный к Иртышской зоне смятия Алтая (Хорева, 1966, стр. 80).

Геосинклинальные образования этого пояса сложены в основном терригенными породами, однако во время осадконакопления порой происходила вулканическая активность (среднедевонские кварцевые порфиры и дациты, порфироиды и порфиритоиды, пластообразные тела спилитоподобных вулканогенных пород). Гораздо более интенсивные магматические проявления в конце среднего девона имели интрузивный характер; они в Иртышской зоне смятия начались формированием габбро-плагиогранитов, приуроченных к Иртышско-Маркакульскому глубинному разлому, подвергшиеся впоследствии преобразовательным процессам. Надо полагать, что упомянутые выше тела спилитоподобных пород являются близкими по возрасту габбро-плагиогранитам.

В верхнем девоне происходит образование прииртышского офиолито-подобного пояса, приуроченного к Кальба-Нарымскому глубинному разлому. Следует отметить, что широко проявленный в сопредельной территории юго-западного Алтая базальтовый вулканизм, по-видимому, имеет общий с прииртышским поясом магматический очаг.

В дальнейшем в зоне Иртышско-Маркакульского глубинного разлома в связи с интенсивными складчато-глыбовыми поднятиями, создавшими шовную горст-антиклинорную структуру Иртышской зоны смятия в верхнем девоне — начале нижнего турне происходит формирование гранитоидов. Последние характеризуются линейно вытянутыми в северо-западном направлении полусогласными интрузиями трещинного типа, изменчивостью петрографического состава, бластезом, метасоматозом, гнейсовидностью, мигматизацией и глубоким метаморфизмом вмещающих пород с образованием микрокристаллических сланцев, контактовых метаморфических сланцев, переходящих в кристаллические сланцы, контактовые кристаллические сланцы и гнейсы.

На этом основании автор полагает, что гранитоиды представляли собой высоконагретые мигматит-плутоны, выжатые по глубинному разлому из областей ультраметаморфизма в тот период, когда еще окончательно не был оформирован очаг гранитной магмы. Его формирование совпадает с верхнепалеозойским этапом тектогенеза; в это время произошло внедрение посторогенных (орогенных, — Г. З.) умеренно-кислых калбских гранитоидов верхнего карбона, приуроченных к Калба-Нарымскому глубинному разлому (стр. 109).

Вряд ли можно считать убедительным столь высокую нагретость мигматит-плутона, выжатого из еще не сформировавшегося гранитового магматического очага и вызвавшего высокотемпературный метаморфизм вмещающей толщи. Более правдоподобным кажется нам, образование метаморфитов и по крайней мере части пород, входящих в состав гранитоидного комплекса, например, мигматитов, гнейсов, некоторых гранитов и др., в результате воздействия на габбро-плагиограниты и вмещающие терригенные отложения высоконагретых восходящих растворов в позднекаледонском, раннегерцинском и позднегерцинском этапах развития области. Следовательно, есть основания полагать, что метаморфические и метасоматические процессы продолжались до конца карбона, а может быть еще позднее; особенно интенсивное их проявление происходило в зоне Калба-Нарымского глубинного разлома, в которой сформировались калбские гранитоиды. Об этом говорят наблюдаемые автором интенсивный метаморфизм среднедевонских габбро-плагиогранитов, наличие в пределах развития массивов плагиоклазовых гнейсо-гранитов полос, линз и ксенолитоподобных тел метаморфизованных пород среднего и основного состава, представленных габбро-амфиболитами, гнейсо-диоритами, гнейсо-тоналитами и гнейсовидными роговообманковыми плагиогранитами, которые рассматриваются как породы более ранних фаз среднедевонского комплекса (стр. 63). Метаморфиты тяготеют к глубинным разломам и расположенным в них гранитоидам; на значительном удалении от них метаморфиты переходят в нормальные осадочные толщи содержащие фауну. Заметим, что метаморфизм вулканогенных геосинклинальных толщ происходит более или менее равномерно. Аналогичное в целом явление формирования гранитоидов и метаморфитов, возникших за счет частично вулканогенных, преимущественно терригенных отложений описано И. Я. Некрасовым (1962) в Верхно-Чукотской складчатой области.

Хорошим примером связи метаморфизма с постмагматическими гидротермальными процессами в период складчатости, является аспидная диабаз-порфиритовая формация, в которой ведущую роль играет фация аспидных сланцев, выделенная нами в качестве переходной метаморфической фации между нормально-осадочными диагенезированными и слабометаморфизированными породами, отвечающими зеленосланцевой фации П. Эскола (Заридзе, 1962, 1966).

Минеральная ассоциация для аспидных сланцев представлена: кварц-серицит-хлорит-углисто-глинистое вещество, а для измененных основных магматических пород и их туфов — кварц-хлорит кальцит (-эпидот)-рудный минерал (-реликты исходных минералов: основной плагиоклаз, моноклинный пироксен, амфибол). Необходимость выделения такой переходной метаморфической фации видна из того, что позднее Б. Я. Хорева (стр. 70) выделила филлитовую метаморфическую фацию для пелитовых пород, не зная, по-видимому, что подобная фация ранее уже была предложена нами с более удачным, как нам кажется, наименованием, так как филлиты скорее относятся к зеленосланцевой фации; в их состав, наряду с серицитом и хлоритом, входяат слюды и альбит.

Б. М. Келлер (1949), детально изучивший терригенные формации (аспидную, флишевую, молассовую) в Залаирском синклинории (общая мощность свыше 8000 м) и сопоставивший отложения последней с другими терригенными геосинклинальными отложениями (Памир, Средняя Азия, Рейские сланцевые горы и хорошо изученный внешный прогиб Уачиты в США), пришел к заключению, что „возникновение аспидной формации относится к тем ранним стадиям развития, когда геосинклиналь интенсивно прогибается, а расчленение ее на частные антиклинальные поднятия и синклинальные прогибы второго порядка (интрагеосинклинали и интрагеоантиклинали) еще не резко выражено. Геосинклинали, в которой происходит накопление песчано-глинистых пород аспидной формации, соответствует широкая уплощенная впадина рельефа; ограничивающие ее поднятия не имеют резкого геоморфологического выражения" (стр. 128).

Согласно названному автору, для Зилаирской аспидной формации (верхний девон-низы карбона) характерна большая мощность осадков, сложенных из чередующихся между собой песчаников (порой граувакковых) и аргиллитов с признаками ритмичности флишевого типа. Спутниками терригенных осадков является прослои вулканических туфов, пачки силицитов; наблюдаются также пластовые залежи диабаз-порфиритовых пород и, как следствие позднего динамического и гидротермального метаморфизма, образование за счет аргиллитов аспидных сланцев, ороговикованных в виду пропитывания их кварцем. Последний образует многочисленные жилы, секущие складчатые структуры.

Для песчаников аспидной формации Зилаирского синклинория характерно присутствие цемента, состоящего из новообразованного кварца и регенерированных полевых шпатов с чешуйками хлорита и серицита; в них содержится также раскристаллизованный пепловый материал, изредка встречаются линзы туфов. Обломки состоят из различных пород и минералов.

Глинистые сланцы сходны с цементирующей массой песчаников; они представляют собой мелкокристаллический агрегат, состоящий из новообразованного кварца, регенерированного полевого шпата, серицита, хлорита, соссюрита, лейкоксена. Интересно отметить, что интенсивность метаморфизма и количество кварцевых жил резко возрастают в восточной полосе распространения аспидной формации, примыкающей к антиклинорию Урал-Тау.

С формированием интрагеоантиклинали, сложенной из аспидной формации, в рядом расположенном прогибе второго порядка (интрагеосинклинали) продолжается накопление терригенного материала с образованием флишевой формации. В это время вулканическая деятельность, как правило, прекращается или имеет спорадический характер. Таким исключением, в частности, является Западный Кавказ, где в сеномане на короткий промежуток времени среди отложений флиша появляется туфогенная толща авгитовых песчаников. В это время в толще исчезает флишевая ритмичность и вновь появляется в туроне в связи с прекращением в прогибе вулканической деятельности.

По свидетельству Б. М. Келлера (1949), мощные толщи Зилаирской флишевой формации (возраст: средний-верхний карбон) обычно сложены уплотненными (агриллиты, песчаники), но не метаморфизованными породами.

В отношении метаморфизма осадков несколько отличное положение наблюдается в верхнеюрско-нижнемеловом флишевом синклинории Южного склона Большого Кавказа в пределах Рачи и Сванетии, осадки которого петрографически изучены И. Д. Чечелашвили, А. Д. Копалейшвили и Э. В. Варсисашвили.

Отложения рассматриваемого синклинория интенсивно дислоцированы. Они представлены псефитолитами, псаммитолитами, алевролитами и пелитолитами (кластогенные породы); известняками, мергелями и терригенно-карбонатными образованиями (карбонатные породы). В них нередко наблюдаются жилки кальцита, реже кварца, секущие складчатось, иногда их друзы и жеоды. Местами фиксируется окварцованность осадков. В Лухумском разрезе, где наблюдается чередование известняков с пачками темных сильно окварцованных сланцеватых аргиллитов, развитых в сводовой части антиклинальных складок, имеется реальгар-аурипигментовое месторождение.

В толще зафиксированы следующие флишевые фигуры: иероглифы проблематического происхождения, механоглифы, образованные в процесе передвижения осадка или потоков воды и биоглифы, возникшие в результате жизнедеятельности организмов. Косая слоистость наблюдается преимущественно в терригенном, меньше в карбонатном флише.

Осадки флишевой формации макроскопически в целом неметаморфизованы и в этом отношении резко отличаются от нижне-среднеюрской аспидной формации, однако микроскопическое изучение пород показало, что в них происходили некоторые преобразовательные процессы исходных осадочных минералов. Авторы разделяют представление Н. М. Страхова относительно того, что новообразованные минералы, в том числе альбит, возникают в стадию позднего диагенеза осадка, когда вследствие обменных реакций и движения растворов происходит перераспределение осадочного материала.

Далее отмечается, что альбиты, содержащиеся в различных разрезах карбонатного флиша Верхней Рачи фиксируются примерно на одном стратиграфическом уровне — в нижней части верхней юры и поэтому, разделяя взгляд И. А. Преображенского (1940), согласующегося с взглядом Тестера и Гордона (Tester — Gordon, 1934), авторы считают, что альбиты имеют коррелятивное значение. Гордону, как известно, удалось разграничить по вторичным полевым шпатам киватинскую формацию северовосточной Миннесоты от дресбахской формации.

И. А. Преображенский (1940), описавший регенерированные полевые шпаты верхнедевонских песчаников Тимана отмечает, что „образование автигенных полевых шпатов в межбазальтовых песчаниках могло быть связано с гидротермальными растворами из базальтовых эффузивов; ... но автигенные полевые шпаты находятся и в верхней свите верхнедевонских песчаников, лежащих выше базальтовых потоков, причех и морфологически, и оптически эти полевые шпаты не отличаются от находящихся в межбазальтовых песчаниках. Поэтому можно предположить, что и те и другие образовались в условиях обычных для нормальных осадочных пород. В нижней свите, лежащей под базальтовыми покровами, имеются горизонты конгломератов с баритовым цементом, относительно которого является вопрос, в каких условиях он образовался. Но автигенные полевые шпаты находятся как раз в тех породах, в которых нет барита, и которые лежат, по-видимому, выше баритовых конгломератов" (стр. 33). По приведенной объективной характеристике вряд ли можно сомневаться о постмагматическом гидротермальном происхождении как барита, так и полевых шпатов. Здесь невольно возникает у нас аналогия со спилитами, которые обычно приурочены к определенному стратиграфическому уровню эвгеосинклинальных толщ. Ими часто начинаются вулканогенные толщи и образованы в результате раннего натриевого метасоматоза под действием натрийсодержащих восходящих растворов в период складчатости и возникновения интрагеоантиклинальной структуры.

Другие новообразованные минералы, которые встречаются уже в вышележащих осадках флиша, представлены кварцем, кальцитом, слюдой (хлорит, серицит, мусковит); редко встречаются ломонтит, сидерит, пирит и титановые минералы (брукит, рутил, анатаз). В Верхней Сванетии и в ущелье р. Лухумис-скали зафиксированы аргиллиты с своеобразными новообразованиями „зонального" кварца и полисинтетически сдвойникованного альбита, имеющих местами идиоморфные очертания, придающих породе порфировоподобную структуру. Между этими минералами расположены параллельно нитевидные образования кварц-серицита.

Поскольку преобразовательные (метаморфические) процессы, вызванные активностью восходящих (гидротермальных) растворов в период складчатости рассматриваемой флишевой толщи не вызывают сомнений, то возникает предположение о закономерном характере этого процесса. Особую интенсивность восходящие растворы проявляют в том случае, когда им предшествовал эффузивный магматический процесс, т. е. если они являются постмагматическими. Чем интенсивнее предшествующий, особенно базальтовый вулканизм, типично представленный в эвгеосинклиналях, тем интенсивнее проявляются в период складчатости восходящие метаморфизующие, гранитообразующие и рудноминерализационные растворы. Это в той или иной мере относится и к внутриконтинентальным вулканогенно-осадочным толщам. В случае терригенных геосинклиналей, где основной и частью ультраосновной магматизм проявляется главным образом в интрузивной форме, преобразовательные процессы, тяготеющие к глубинным разломам, также зависят от масштаба предшествующего базитового магматизма.

Для полноты картины о региональном метаморфизме следует рассмотреть этот процесс на примере угольных бассейнов.

Некоторыми исследователями выделяются три вида метаморфизма углей: контактово-термальный, динамический и региональный. В. И. Яворский (1967) отмечает, что процесс контактово-термального метаморфизма сводится к внедрению в угленосную формацию магмы, эффект влияния которой зависит от ее состава (кислая или основная), массы, температуры и положения относительно пласта угля (стр. 119). Приводятся примеры из Сучанского месторождения и Тунгусской угленосной площади по В. С. Шехунову (1938).

В Сучане отмечаются многочисленные дайковые тела кварцевых порфиров и порфиритов, пересекающих как угольные пласты, так и вмещающие их осадки, которые вызывают их скоксованность в незначительной по ширине зоне, не превышающей нескольких метров. Наряду с эффузивами и субвулканическими телами отмечается наличие гранитоидов, в связи с чем изменение углей достигает графитизации. Исходя из этих данных, В. С. Шехунов высказывает мнение о метаморфизме всех углей Донбасса в связи с воздействием на них магматических тел. Это мнение не разделяет В. И. Яворский (1967), хотя и приводит пример из Кузнецкого бассейна, где в балахонской свите Томь-Усинского района широко развиты пластовые тела и дайки диабазов, вблизи которых угли подвергаются антрацитизации.

Названные авторы не принимают во внимание то обстоятельство, что на угли и вмещающие их осадки оказывают воздействие не только магматические массы, эффект которых проявляется на весьма ограниченном расстоянии, а главным образом постмагматические растворы, имеющие глубинное происхождение, лишь парагенетически связанные с данным магматизмом. Они перемещаются по тем системам трещин, часть из которых заполнена магматическими телами. В последнем случае они проникают вдоль контактов магматического тела с вмещающей породой и либо вызывают метаморфизацию последнего, либо завершают процесс, начатый в магматическую стадию. Последнее явление создает впечатление о воздействии на угли и вмещающие их осадки только лишь залегающего в их контакте магматического тела. В целом этот процесс охватывает большие площади, всю толщу, т. е. является именно тем процессом, который определяется под названием регионального метаморфизма.

Динамометаморфизм независимо от других факторов метаморфизма, вряд ли может вызвать изменение состава (качества) каменных углей. Он, как известно, обусловлен складкообразовательными движениями, во время которых, как отмечалось происходит активизация восходящих растворов. Именно так надо понимать мысль Ю. А. Жемчужникова (1948), отметившего, что „динамический метаморфизм углей в складчатых областях может сочетаться с региональным, но немыслим без него". Ряд авторов погружение угленосных толщ и региональный метаморфизм связывает со складкообразованием. В частности, И. И. Аммосов (1941) изменение степени метаморфизма углей Прокопьевско-Киселевского района Кузнецкого бассейна объясняет тем, что в момент метаморфизма, т. е. на максимальной глубине погружения угленосной толщи, пласты угля уже были сложены в пологоволнистые складки и, находясь вследствие этого на различной глубине, подверглись воздействию разных температур. В этом же районе И. И. Молчановым (1948) было установлено, что на современном денудационном срезе к замыканию брахисинклинальных складок всегда наблюдается повышение степени метаморфизма углей, максимум которого совпадает с антиклинальным перегибом длинных осей этих структур. Этот исследователь также указывает, что метаморфизм угля одних и тех же пластов увеличивается от замков антиклинальных складок к замкам синклиналей, погружающихся на большую глубину. В. И. Скок (1963) пишет, что „при более крутом залегании пластов угля метаморфизация их на глубину происходит интенсивнее, чем при пологом" (стр. 26). Далее он отмечает: „В Кузнецком бассейне основная метаморфизация угля происходила в период нахождения его на максимальной глубине погружения, а точнее — в момент второй инверсии, имевшей место в конце мела (первая была в средне- или верхнетриасовое время)" (стр. 32—33).

Вряд ли можно считать эту гипотезу приемлемой. Если уж связывать метаморфизм углей и угленосной толщи с погружением последней, то естественее связать его с предскладчатым периодом осадконакопления, когда глубина погружения осадков много больше, чем в синклинальных складках, возникших во время складчатости.

Заметим, что углеобразование на территории Грузии в батском веке связано с самой сильной для южного склона Большого Кавказа батской фазой складчатости, вызвавшей замыкание байосской эвгеосинклинали и образование гранитоидов. Батская регрессивная угленосная толща является конседиментационной.

Что касается мнения о проявлении относительно высокой степени метаморфизма углей и вмещающих их толщ на более низких горизонтах, совпадающих с замками синклиналей, то это можно объяснить воздействием на них восходящих растворов, которые при прохождении нижних уровней толщи были относительно более высокотемпературными, чем на верхних.

* * *

Наши представления о региональном метаморфизме различного типа толщ пород приведены в тексте настояйщей статьи. Они в целом не отличаются от высказанных нами ранее представлений (Заридзе — Татришвили, 1953). В настоящее время мы также считаем, что региональный метаморфизм совершается в период складчатости под воздействием восходящих высокотемпературных растворов, обусловливающих подъем геотермического градиента и течение минералопреобразовательных процессов. Давление, при котором совершается метаморфический процесс, возникает при складчатых натяжениях; образованная при этом структура (интрагеоантиклиналь) в отношении давления является анизотропной.

Региональный метаморфизм проявляется, как известно, в геосинклиналях, среди которых типичной для этого вида метаморфизма является эвгеосинклиналь. Первый этап регионального метаморфизма, совершающийся в стадию образования интрагеоантиклинали, обычно не превышает зеленосланцевую и эпидот-амфиболитовую фации для пелитовых пород. В это время формируются бедные калием гранитоиды (габбро — плагиогранитовая формация).

Образование метаморфитов с более высокотемпературными минеральными ассоциациями, происходит при повторных метаморфических процессах в течение нескольких циклов геологического развития. В конце второго или даже третьего геологического цикла формируются богатые калием и кремнием гранитоиды путем метасоматоза, за счет пород габбро-плагиогранитовой формации и их вмещающих метаморфитов.

В терригенных геосинклиналях, сопровождающихся иногда крупными интрузиями основных и частью ультраосновных пород, залегающих в глубинных разломах, интрузии гранитоидов накладываются на предшествующие им базиты, либо располагаются в тех же разломах, но лишенных базитов с образованием цепочкообразно расположенных крупных массивов. Становление гранитоидов в расматриваемых нами геосинклиналях происходит в течение не менее двух циклов геологического развития, как это имеет место, например, в Иртышско-Маркакульской и др. зонах.

В терригенных геосинклиналях флишевого типа, совершенно лишенных или почти лишенных всякого рода магматических образований как в эффузивной, так и в интрузивной формах, метаморфические процессы играют весьма незначительную роль. Примером могут служить палеозойские отложения флишевой формации Зилаирского синклинория на Южном Урале. В верхнеюрско-нижнемеловом флишевом синклинории южного склона Большого Кавказа в связи со складчатостью флишевой интрагеосинклинали имела место активность восходящих гидротермальных растворов, которым не предшествовало проявление магматизма; они произвели заметные преобразования в самой нижней части флишевой толщи, где наблюдается окварцеванность и новообразования альбита и др. вторичных минералов. В связи с активностью гидротермальных растворов в районе р. Лухумис-цкали образовалось, как отмечено, реальгар-аурипигментовое месторождение.

Сказанное дает нам основание полагать, что в период складчатости, повидимому, всегда происходит с той или иной интенсивностью, просачивание через осадочные толщи восходящих растворов различного состава, температуры и P_{H_2O}. Это наблюдается и в неогеосинклинальных внутриконтинентальных (эпиконтинентальных) бассейнах, в частности, в угольных бассейнах, в которых метаморфизм угля и угольной толщи, если он в какой-то мере обусловлен непосредственным воздействием магматического тела в случае его наличия, то главную роль в процессе метаморфизма играют восходящие растворы — постмагматические, либо не постмагматические, т. е. не имеющие связь с предшествующим магматизмом вследствие его отсутствия. О процессах регионального метаморфизма, совершающегося в стадию складчатости угленосных толщ у большинства исследователей сомнений не вызывает, однако о его причинах, как отмечалось мнения расходятся.

Одним из наиболее сложных является вопрос о том, какие преобразования происходят в стадию осадконакопления и погружения толщ пород. Ето все еще нельзя считать решенным полностью. Есть основание для утверждения, что эти изменения незначительны. Например, мы уверены в том, что метаморфизм аспидных сланцев (аспидносланцевая фация) Южного склона Большого Кавказа произошел в период складчатости, т. е. во время формирования аспидносланцевой интрагеоантиклинали (конец средней юры) и зарождения рядом с ней флишевой интрагеосинклинали, которая в свою очередь подверглась некоторым наложенным гидротермальным процессам, опять-таки в период складчатости. Однако некоторые исследователи, изучавшие терригенные геосинклинали считают, что метаморфизм, вызванный погружением толщи пород в период осадконакопления, является более интенсивным, достигая низкотемпературной субфации зеленосланцевой фации, а региональный метаморфизм более высокой степени совершается уже в период складчатости.

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G. M. ZARIDZE

DATING OF THE REGIONAL METAMORPHISM DURING THE DEVELOPMENT OF THE GEOLOGICAL CYCLE

The present article gives the author's conception of the regional metamorphism of different types of rock beds. This conception is generally not different from earlier conceptions (Zaridze—Tatrishvili 1953).

Now there is still a supposition about the regional metamorphism ending in the period of folding under the influence of then arising high-temperatured solutions conditioning the increase of geothermic gradient in the course of the processes of alteration of minerals. The pressure accompanying the final phase of metamorphism, arose due to the folding stress. Thus the structure (intrageoanticline) was formed, anisotropic in relation to the pressure.

Regional metamorphism may be observed in geosynclines, eugeosynclines being typical of this type of regional metamorphism.

The first stage of regional metamorphism completed in the stage of the forming of intrageoanticline, usually does not exceed the green-schistose and epidote-amphibolitic facies, as far as pellitic rocks are concerned. In this time, granitoides (gabbro-plagiogranitic formation) poor in K, arose.

The forming of metamorphites with high-temperatured mineral associations takes place in repeated metamorphic processes, in the course of several cycles of geological development.

At the end of the second and even third geologic cycles, due to metasomatism, there arose granitoides rich in K and quartz to the d./p. of the rocks of gabbro-plagiogranite formation, metamorphites included.

In terrigene geosynclines, accompanied sometimes by extensive intrusions of basic and partly ultra-basic rocks deposited in deep faults, intrusions of granitoides are deposited on the underlying basites or in the same faults without basites, with the forming of extensive chain-like massifs.

Granitoides in geosynclines under study were followed in the course of at least two cycles of geological development, as it was e. g. in the Irtyš-Markakul, and other zones.

Metamorphic processes have generally little importance in terrigene geosynclines of the Flysch type, completely or partly lacking any type of the magmatic formations as in effusive so in intrusive forms. As an example of this may be considered the Paleozoic deposits of the Flysch formation in the Zilair synclinore in Southern Ural. In the Upper-Jurassic — Lower-Cretaceous Flysch synclinore of the southern slope of the Great Caucassus in relation to the folding of the Flysch intrageosyncline, hydrothermal solutions were active. They were not preceded by magmatic activities, they evoked alteration in the lowermost part of the Flysch beds, where quartzification and new formations of albite and other secondary minerals, took place.

In relation to the activity of hydrothermal solutions in the territory of the Luchumns-ckali river, there arose an realgar-auripigment deposit.

From this it follows that in the time of folding, solutions of different composition, temeperature and P_{H_2O} penetrated the sedimentary beds. This may be also observed in neogeosynclinal innercontinental (epicontinental) basins, especially in coal basins. In these metamorphism of coal and coal beds — if conditioned by immediate influence of magmatic body — in case it occurs there — then postmagmatic or non-postmagmatic solutions have greatest importance in the process of metamorphism.

There are no doubts about the processes of regional metamorphism completed in the period of the folding of the coal-bearing beds, yet the opinions about the cause of these processes are contradictory. One of the most complicated questions is: which alterations took place in the stage of deposition and submerging of rock beds. This problem has not been completely solved yet. It may be supposed that these alterations were only slight. For instance, we may believe that metamorphism of the aspid shales (aspid-shale facies) of the southern slope of the Great Caucassus Mts. took place in the time of folding, i. e. in the period of forming of the aspid-shale intrageoanticline (the end of the Middle Jurassic) and of the rise of the Flysch intrageosyncline that was affected by some hydrothermal processes, again in the period of folding. Yet some investigators studying terrigene geosynclines supposed that metamorphism evoked by submerging of rock beds in the time of their deposition, was more intense, reaching the lower-temperatured subfacies of the green-shale facies, while regional metamorphism of higher degree ended as early as the time of folding.

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JÁN ILAVSKÝ

ZUR METALLOGENETISCHEN KARTE DER WESTKARPATEN 1:1,000 000

Auf Grund der Generalkarten der Mineralrohrstoffe 1:200 000, die für den Teil der Westkarpaten in den letzten zehn Jahren in der *Geologischen Anstalt Dionyz Štúr's, Bratislava* verfertigt wurden, haben wir eine Skizze der metallogenetischen Karte im Maßstab 1:1,000 000 zusammengestellt, die den Bestandteil der metallogenetischen Karte der Tschechoslovakei bildet, welche anläßlich des XXIII. geologischen Weltkongresses erscheinen wird. Bei dieser Gelegenheit möchte ich die wichtigsten metallogenetischen Probleme der Westkarpaten erörtern, u. zw. von den Standpunkten, die sich in den letzten Jahren in aller Welt geltend machen. Aus ihnen folgt, daß die Begriffe, mit welchen man in diesem Fach arbeitet, in folgende Kategorien gehören:
räumliche Metallotekte, die den Gehalt der metallogenetischen Rayonierung bilden;
zeitliche Metallotekte, die den Gehalt der Gliederung der metallogenetischen Entwicklung auf Epochen, Stadien, Phasen usw. bilden;
stoffliche Metallotekte, die geologische und metallochemische, sowie geochemische Erscheinungen und Daten über die studierten Lagerstätten einbegreifen.

Dadurch, daß die metallogenetischen Karten in vereinfachter Form solche Mengen von Daten zum Ausdruck bringen, erfüllen sie einerseits im gewissen Sinne die statistische Methode der bildlichen Darstellung der Ergebnisse in der metallogenetischen Erforschung und anderseits bieten sie die Unterlage für die wissenschaftliche Analyse über genetische Bedingungen der Akkumulation der Mineralrohstoffe und der Gesetzlichkeiten ihrer räumlichen und zeitlichen Verteilung (im Sinne P. Routhier's (1963) bieten sie die Unterlagen für die Strategie des Aufsuchens der Mineralrohstofflagerstätten).

Räumliche Metallotekte der Westkarpaten (metallogenetische Rayonierung)

Die Basis der metallogenetischen Rayonierung im Sinne J. A. Bilibin's (1953, 1955), W. E. Petrascheck's (1955, 1963), P. M. Tatarinow — V. G. Gruševoj — G. S. Labazin's (1957) und P. Routhier's (1963) ist die Teilung des bestimmten Raumes auf die Einheiten niedrigerer Ränge, die einen bestimmten typischen Bau, dabei jedoch einheitlichen stratigraphischen Gehalt, einheitliche Strukturen, gleichen Typus des Magmatismus, Vulkanismus und bestimmte genetische Typen der Erzlagerstätten mit ihren typischen Formen und mit der Intensität der Vererzung besitzen. Das ist im wesentlichen das *Prinzip der Vergleichungslehre über die Erzlagerstätten oder die Typologie der metallogenetischen Einheiten.*

Die Gliederung der geologischen und dadurch auch der metallogenetischen Systeme hängt also von zahlreichen spezifischen geologisch-tektonischen und metallogenetischen Merkmalen ab, die in einzelnen kontinentalen Blöcken nach dem Gehalt, Alter usw. unterschiedlich sind (Kordilleren-, Appalachien-, Alpino-Himalaja-, Ural-Systeme usw.). Im Sinne W. E. Petraschek's sind derartige Einheiten am zweckmäßigsten als *metallogenetische Megaprovinzen* zu bezeichnen, wobei durch die obenangeführte Beifügung ihre Konkretisierung ausgedrückt wird.

Die Megaprovinzen in den geosynklinalen, polyorogenen Systemen zerfallen im Sinne der bekannten Definitionen J. A. Bilibin's (l. c.) und W. E. Petrascheck's (1955) in *metallogenetische Provinzen.* Im wesentlichen handelt es sich um Abschnitte oder Zweige der Megaprovinzen, die sich durch bestimmten, typischen, dabei einheitlichen geologischen Bau, einheitliche Strukturen, einheitlichen Typ des Magmatismus, Vulkanismus und der Metamorphose auszeichnen. Die Provinz besitzt ihre typische Intensität der Vererzung, typische Formen und genetische Typen der Lagerstätten. Im Sinne Tvalčrelidze's (1957) besitzt sie auch ihre eigenen typomorphen Metalle, in bestimmten Entwicklungsstadien einzelner Epochen typomorphe genetische Typen der Lagerstätten und im Sinne P. Routhier's (1963) ist sie im bestimmten spezifischen Erosionsniveau abgedeckt. In solchen Provinzen sind oft auch mehrere metallogene Epochen vertreten und so sind sie im Sinne Turneaur's (1955), Tvalčrelidze's (1957) und Smirnow's (1960) *polyzyklisch.*

Dieser Definition nach sind also die Westkarpaten *eine selbständige westkarpatische metallogenetische Provinz* und man kann sie nicht mit den Ostalpen in eine Einheit zusammenfassen (was wir im Weiteren noch erörtern werden).

Mit Bezug auf die Gliederung der Westkarpaten nach D. Andrusov — A. Matějka (1931) und V. Zoubek (1937) oder H. Stille (1953), kann man die westkarpatische metallogenetische Provinz (nach W. E. Petrascheck's Kriterien) in 3—4 Subprovinzen zergliedern:

(a) *metallogenetische Subprovinz der Vortiefe der Westkarpaten*;

(b) *metallogenetische Subprovinz der äußeren Flysch-Karpaten*;

(c) *metallogenetische Subprovinz der Zentral-Westkarpaten*;

(d) *neovulkanische, oder pannonische metallogenetische Subprovinz.*

In jeder von ihnen ist bloß ein bestimmter Teil der stratigraphischen Kolonne der westkarpatischen Serien und Formationen mit dem zugehörigen Magmatismus, Vulkanismus und den Erzformationen vertreten. Die Subprovinzen besitzen ihre Eigentümlichkeiten des Baues; eine jede besitzt eigenen strukturellen Stil, andere genetische Typen und andere Erzarten. Auch die innere Gliederung einzelner Subprovinzen ist verschiedenartig.

1. *Metallogenetische Subprovinz der karpatischen Vortiefe* hat den Charakter einer strukturell-faziellen Zone, die durch Neogen (mit Erdöl und Gaslagerstätten) gebaut ist und die Grenze mit der Böhmischen Masse bildet.

2. *Metallogenetische Subprovinz der äußeren Karpaten.* Ihr Bau ist ausgeprägt linear und die Richtungen einzelner strukturell-faziellen Zonen gleich, d. h. bogenförmig: im Westen SW-NO, in der Mitte O-W und im Osten NW-SO. Abgesehen von der Detailgliederung der äußeren Flysch-Karpaten in zahlreiche längliche faziell-lithologische und tektonische Einheiten (Matějka—Roth, 1962, 1964) kann man diese Subprovinz in drei strukturell-metallogenetische Zonen zergliedern:

(a) *Strukturell-metallogenetische Zone des äußeren Flysches*, in welcher die Kreide von kleinerer Bedeutung ist und das Paleogen vorherrscht. Typische Vererzung bilden da pelosideritische Flöze in der Unterkreide (Valangien), im Urgon, Alb, Cenoman und Eozän der nordwestlichen Karpaten, d. h. in den Mährisch-schlesischen Beskiden. Vereinzelt kommen die Blei-Zink-Vererzungen in kleinen Körpern der Neovulkanite vor.

(b) *Strukturell-metallogenetische Zone der Klippen*, die hauptsächlich durch Jura und Kreide, weniger durch Paleogen gebaut ist. In dieser Zone sind kleine Flöze der oxydisch-karbonatischen Manganerze im Lias und Dogger bekannt (Lednické Rovné, Zázrivá, Šarišské Jastrabie). Auf dem polnischen Territorium gibt es in dieser Zone kleine Körper der mit dem Teschinit äquivalenten Eruptivgesteine mit unbedeutender Konzentration der Blei-Zink-Erze und des gediegenen Goldes.

(c) *Strukturell-metallogenetische Zone des Inneren Flysches*, die hauptsächlich paleogenen Alters ist, enthält nur sehr kleine Vererzungen der sedimentären, oxydisch-karbonatischen Manganerze und Pelosiderite, die in den Schiefern liegen (Orava).

Diese Subprovinz, mit drei strukturell-metallogenetischen Zonen, besitzt also Vererzungen des Eisens und Mangans, die sich in *zwei metallogenetischen*

Etappen wiederholt abwechseln: jurasische mit Mn und kretazisch-eozäne mit Eisenerzen. Diese Etappen fallen in das mittlere Entwicklungsstadium der alpidischen metallogenen Epoche der Westkarpaten. Beide Etappen sind räumlich differenziert und sind in verschiedenen strukturell-metallogenetischen Zonen eingelagert: *Manganerzetappe* jurassischen Alters in der inneren und *Eisenerzetappe* kretazisch-eozänen Alters in der äußeren Zone.

Solche räumliche und zeitliche Verteilung entspricht dem zeitlichen Verlauf der tektonischen Prozesse im tiefen Liegenden der Flyschzone.

3. *Polyzyklische, heterogene Subprovinz der Zentral-Westkarpaten* unterscheidet sich diametral von der vorherigen Subprovinz nicht nur durch abweichende Bauelemente, sondern auch durch ihren inneren Bau. Die Sedimentation, der Magmatismus, Vulkanismus, die Tektonik, Metamorphose und Metallogenese besitzen anderen Charakter, anderen zeitlichen Verlauf und ganz undere räumliche Verteilung (uzw. wiederum in einigen strukturellfaziellen Zonen). Vom metallogenetischen Standpunkte zeigt sich diese Subprovinz (was die Vertretung einzelner Erzformationen, die genetischen Typen, ihre zeitliche Einreihung, die Formen der Lagerstätten und ihre räumlichen und zeitlichen Beziehungen zum betreffenden Magmatismus oder Vulkanismus betrifft) als einheitlich.

Gegenwärtiges Niveau der metallogenetischen Terminologie, sowie das Niveau des Durchforschens des westkarpatischen Gebietes in letzten 15 Jahren, namentlich die neuesten isotopischen Pb-Untersuchungen der Galenite zahlreicher westkarpatischen Lagerstätten (J. Kantor 1960—1965) erlauben uns die metallogenetische Rayonierung dieses kompliezirten Gebietes auf moderne Art zu interpretieren, die nicht im Einklang mit bisherigen Ansichten von Máška (1957), Koutek (1958, 1964), Cambel (1958) usw. sind.

Im Einklang mit obenzitierten Definitionen von Bilibin, Petrascheck, Smirnov, Tvalčrelidze usw., ist es notwendig diese Subprovinz in zwei große strukturell-metallogenetische Zonen zu teilen: Tatroveporiden und Gemeriden.

Strukturell-metallogenetische Zone der Tatroveporiden besitzt im ganzen Gebiet einheitlichen geologischen Bau und fast gleiche Vertretung der geologischen Formationen: kristalline Kerngebirge mit variskischen Granitoiden inmitten. Um alle Kerngebirge herum treten die mesozoischen Serien (Untertrias-Unterkreide) u. zw. in der autochthonen oder tektonischen Position (subtatrische Decke). Nachorogene Formation ist Eozän. Die Metamorphose in allen Kerngebirgen ist mezo- und katazonal, wobei ihr Alter hauptsächlich variskisch, teilweise auch älter ist. In der alpidischen Ära wurden die kristallinen Kerne diaphtorisiert.

Die Erzlagerstätten der strukturell-metallogenetischen Zone der Tatroveporiden sind in allen Gebirgen, was das Material, Form, Genese und Alter betrifft, übereinstimmend und auch ihre ökonomische Bedeutung ist im ganzen einheitlich. Die Intensität und Größe der Lagerstätten in einzelnen Entwicklungsetappen sind im ganzen in den Kerngebirgen gleich.

Eingehendere Gliederung der strukturell-metallogenetischen Zonen der Tatroveporiden in lineare Teileinheiten ist auch bei ihrem Zonenbau unmöglich, da die Kerngebirge voneinander bedeutend entfernt und durch Areale des Mesozoikums und Paläogens isoliert sind.

Deshalb muß man als Einheit niedrigeren Grades den *Erzdistrikt* annehmen. Unter Erzdistrikt verstehen wir die Kerngebirge mit dem sie umgebenden Mesozoikum in der heutigen Deckenform.

Mit Bezug auf die ältere geologisch-strukturelle Gliederung dieses Gebietes (im Sinne Andrusov—Matějka—Zoubek's) kann man in der *Subzone der Tatriden* diese Erzdistrikte unterscheiden (von Westen nach Osten): Kleine Karpaten, Inovec, Kleine Fatra, Hohe Tatra, Branisko mit Čierna Hora und Zemplin-Insel und einen anderen Streifen von Tribeč, Malá Magura, Groß-Fatra und der Niederen Tatra.

Detaile Beschreibung der einzelnen Erzdistrikte ist auf dieser Stelle nicht zweckmäßig, da sie in der Vergangenheit von mehreren Verfassern gegeben wurde (zum Beispiel L. Maderspach 1879; J. Ahlburg 1913; M. Máška 1957; J. Koutek 1958, 1964 usw.).

In der *Subzone der Veporiden* kann man im wesentlichen nur einen: Vepor-Erzdistrikt ausgliedern, der mit vorhergehenden durch die Lagerstättenformen, Inhalt und Alter identisch ist.

Mit Bezug auf den Umstand, daß die metallogenetischen Prozesse in einzelnen Erzbereichen polyzyklisch waren, d. h. es gibt da mehrere metallogenetischen Epochen mit zugehörigen Entwicklungsstadien und metallogenen Phasen, werden wir jeden Erzdistrikt im Sinne Nekrasov—Šatalov's (1960) in *strukturellmetallogenetische Etagen* gliedern, die zeitlich mehrere metallogene Epochen enthalten (kaledonische, variskische und alpidische z. b.). In den Etagen kann man *strukturell-metallogenetische Subetagen* (räumlich) *mit zugehörigen Phasen* (zeitlich), als kleinere räumliche und zeitliche Einheiten, die durch das Entwicklungsstadium dem Orogen (Vorfaltungs-, Faltungs-, Nachfaltungsstadium), bzw. noch klieneren Zeitabschnitten — *den strukturell-metallogenen Etappen bis Stufen* mit den zugehörigen Erzformationen ausgliedern. (Diese Gliederung wird noch weiter näher beleuchtet, und zwar so, daß daraus weitere Erscheinungen wie Konzentrationsfaktor und Intensität der Vererzung hervorkommen).

In dieser Subzone sind hauptsächlich zahlreiche hydrothermale Gänge der siderit-sulphidischen Erzvormation vertreten, die das postvariskische Alter haben und in ihrer Zusammensetzung namentlich Fe, Cu, Sb, Au, Pb, Zn vorkommen. Eine kleinere Bedeutung gehört den sedimentär-vulkanogenen Pyrit-Pyrhotinerzen in metamorphierten Grünsteinen (Amphiboliten) des vorvariskischen Alters. In oberpermischen Schichten sind in manchen Gebirgen stratiforme Kupfererze mit Siderit vertreten.

Strukturell-metallogenetische Zone der Gemeriden hat einen von der Tatroveporiden-Zone vollkommen unterschiedlichen Bau, anderen tektonischen Stil, andere Magma- und Vulkanismus-Typen, die in dem stratigraphischen Profil auch quantitativ anders verteilt sind. Die Erzlagerstätten sind in der Gemeriden-Zone, was die Arten, Typen und Formen betrifft, reicher und ihre Dichte und Ausmaß sind viel größer als in den Tatroveporiden.

Die strukturell-metallogenetische Zone der Gemeriden ist die vollständigste, was den stratigraphischen Gehalt betrifft, da in ihr das Kambrium, Ordovizium, Silur, Devon, Oberkarbon, Perm, das Mesozoikum von der Untertrias bis Jura, wie auch die Gosaukreide, Paläogen und Miozän an den Rändern des Gebirges vertreten sind. Die Metamorphose in der ganzen Zone hat den epizonalen Charakter und hat sich vielleicht zweimal abgespielt (variskisch und alpidisch).

Über die metallogenetischen Probleme gibt es aus den letzten Jahren zahlreiche Literatur, die aber vom metallogenetischen Standpunkt nicht eingehend beurteilt werden.

Der geologische Bau der Gemeriden ist zum Unterschied von den Tatroveporiden ausgeprägt *linear-zonal* und dabei *symmetrisch.* In seiner Achse und Mitte ist die altpaläozoische Zone des Volovec, d. h. die kambrosilurische Gelnica-Serie. Diese Einheit entspricht *der kaledonischen strukturell-metallogenetischen Etage.* In ihr lagern die exhalativ-sedimentären Kies- und Magnetiterze *in einigen linearen Streifen*, die den drei zeitlichen strukturell-metallogenetischen Stufen (Kambrium, Ordovizium, Silur) entsprechen. Die selbständige Erzformation sind hier die sogennanten „metasomatischen“ Sideritlagerstätten von Nižná Slaná und Železník, die nach den geochronologischen Studien von J. Kantor (1965) kaledonisches Alter haben und wahrscheinlich sedimentär-vulkanogenen Ursprungs sind. Es handelt sich hier also um eine andere Fazies als bei Kies- und Magnetitlagerstätten.

In der kambrosilurischen Serie liegen weiter in zahlreichen parallelen *Erzzügen* hydrothermale *Ganglagerstätten* der sideritischen und sideritisch-sulfidischen Formation, die zeitlich aber jünger sind (variskisch). Von dem räumlichen Standpunkte zerfallen sich diese Erzzüge in zehlreiche *Erzfelder mit eigenen Gängen.*

Die variskische strukturell-metallogenetische Etage repräsentieren: die Phyllit-Diabas-Serie devonischen Alters und einige kleine Massivchen der variskischen Granite (mittelkarbonischen Alters), weiter Sedimente und Vulkanite des Oberkarbons und klastische Perm-Schichten. Das geosynklinale (Vorfaltungs-) Stadium repräsentieren Phyllite und Diabase (Rakovec-Serie) mit stratiformen exhalativ-sedimentären Hämatit-Magnetit-Erzen. Sie stellen die *vorfaltungsphasige variskische strukturell-metallogenetische Subetage* dar, die zonal und fast symmetrisch beiderseits der Gelnica-Serie in Form von zwei strukturell-metallogenetischen Streifen lokalisiert ist. Als jüngere Lagerung gibt es in ihnen wieder spätvariskische Erzzüge mit sideritischen und sideritisch-sulfidischen Ganglagerstätten.

Dem *Faltungsstadium* entsprechen kleine Massivchen der Granitoide, die längs des Kontaktes der Gemeriden und Veporiden-Zone vetreilt sind (in den Gemeriden ist keine metallogenische Phase vertreten).

Das spätvariskische Entwicklungsstadium vertritt das Oberkarbon, das *die spätvariskische strukturell-metallogenetische Stufe* mit den Magnesit-Lagerstätten am westlichen und östlichen Ende der Gemeriden repräsentieren, während in dem Zentralteil der Gemeriden (im länglichen Sinne) die Oberkarbon-Schichten von den sideritischen und sideritisch-sulphidischen Gängen durchbrochen sind, die in den karbonatischen Schichten metasomatische Vererzungen bilden. Das Alter der Sideritvererzung ist hauptsächlich jungvariskisch, die während der alpidischen Ära rekristallisiert wurde.

Das Oberkarbon bildet beiderseits der Volovec-Zone *zwei Streifen.* Vom metallogenetischen Standpunkt ist der nördliche Streifen von großer Bedeutung. Er bildet am südwestlichen Ende den metallogenetisch-faziellen Streifen *des Magnesit-Karbons* im Bereiche von Lučenec und Ochtina, und im Abschnitte Štítnik-Dobšiná-Mlynky den metallo-faziellen Streifen *mit „metasomatischen" Sideritlagerstätten,* während er weiter ostwärts bis nach Košice (von metallogenetischer Sicht aus) nicht für metasomatische Lagerstätten, sondern eher für hydrothermale Gänge vom Belange ist. Der Košicer Abschnitt ist wiederum Magnesit-führend.

Der südgemeride Streifen des Oberkarbons (ehemalige Rožňavsko železnícka-Serie) besitzt keine Karbonate, die für die Entstehung der metasomatischen Siderite geeignet wären; dagegen gibt es da zahlreiche sideritische und siderit-sulphidische Gänge permischen, oder auch alpidischen Alters.

Das finale Entwicklungsstadium der Varisziden repräsentieren *oberpermische strukturell-metallogenetische Schichten,* die analog wie das Karbon in zwei Streifen entwickelt sind. Es handelt sich um kontinentale Entwicklung des Verrucano (im Süden der Gemeriden auch mit Meeresfazies), die durch sauren porphyrischen Vulkanismus, mit eingelagerten stratiformen, alpidisch metamorphiesierten Uran—Molybdän—Kupfer—Erzen, sowie mit Flözen der Hämatiterze, begleitet ist. Sie sind im Perm in *mehreren metallogenetischen Horizonten* vertreten.

Des permischen Alters ist auch sehr ausgedehnte sideritische und sideritisch-sulfidische Gangformation hydrothermaler Herkunft, die auch in älteren Schichtenfolgen sehr verbreitet ist, u. zw. in Form von schon erwähnten *Gang-*, oder *Erzzügen.* Infolge der primären Diskontinuität dieser Gangzüge, die hie und da verschiedene Dichte haben, kann man *Erzfeld* als eine kleinere Einheit betrachten und nach den Ortsnamen bezeichnen. Diese Erzfelder bestehen aus kleinsten Grundeinheiten — *Lagerstätte* oder *Gang*, die ihre *Gangabschnitte* haben.

Alpidische Etage ist in den Gemeriden durch zwei Streifen vertreten: Galmus-Gebirge in Norden und Slovenský Kras in Süden, die zum paläozoischen Untergrund autochthone bzw. paraautochthone Beziehung haben.

Untertrias in beiden Streifen representiert das geosynklinale Entwicklungsstadium der Alpiniden. Sie stellt die *initiale strukturell-metallogenetische Stufe* dar, die einige *Erzhorizonte* mit stratiformen, größtenteils exhalativ-sedimentären Erzformationen besitzen, wie Hämatit, Hämatit-Baryt, Hämatit-Siderit, oder Hämatit-Kupfererz-Fazies.

Während des Faltungs-Stadiums der alpidischen Ära (Mittelkreide-Eozän) entstanden in den Gemeriden einige *hydrothermalmetasomatische* und teilweise auch *pseudohydrothermale, regenerativ-mobilisierte Erzformationen* (Hämatite in der Trias, Antimonit-Gold-Quarz-Gänge im Paleozoikum, Kupfererz-Gänge manchmal mit Siderit-Ankerit).

Die Gosauer Konglomerate, Eozän und Oligozän an den Seiten der Gemeriden-Zone, gehören schon den späteren Entwicklungsstadien der Alpiniden an. Einige metallogenetischen Stufen enthalten verschiedene Erzformationen (Manganerze im Eozän, Bauxit-Sedimente des kretazischen Alters usw.).

Zusammenfassend kann man also sagen, daß die metallogenetische Subprovinz der Zentralwestkarpaten *polyzyklisch* ist, da sich die Erzlagerstätten in ihr in drei metallogenetischen Epochen gebildet haben. Die Entwicklung der Erzlagerstätten war, was den Ursrpung der Erzstoffe und die Entstehungsarten betrifft, heterogen; deshalb ist für diese Subprovinz die Benennung polyzyklisch heterogene Subprovinz vollkommen begründet.

Neovulkanische-jungalpidische-pannonische Subprovinz der Westkarpaten.

Die räumliche und zeitliche Auffassung und den stofflichen Gehalt dieser Subprovinz haben schon früher mehrere Forscher vermittelt (Wendl 1939; Helke 1942; Kuthan 1960; Polák 1963; Koděra 1957, 1964; Böhmer 1958—64 usw.). Sie bildet einen gegen Süden gewölbten Bogen an der inneren Seite der Westkarpaten, mit dem Verlauf von dem Gebirge Kremnické pohorie über Štiavnické pohorie nach Nordungarn und kehrt wiederum zurück in das slovakische Gebiet ins Gebirge Tokajsko-Prešovské pohorie. Diese Gebirge sind überwiegend durch verschiedene Andesite, Rhyolite, Trachyte, Dazite bis Basalte tortonisch-pliozänen Alters ausgebildet. Die räumlichen metallogenetischen Einheiten in dieser Subprovinz sind bisher nicht zufriedenstellend gelöst, hauptsächlich deshalb, weil ihr Bau nicht ausgeprägt linear ist.

Vulkanite und Sedimente des Miozäns representieren *finale metallogenetische Phase der Alpiniden*. Sie enthalten in mehreren Stufen und Etappen, bis Horizonten verschiedene Erzformationen. Da diese nicht genügend geklärt sind

(Polák 1961), ist in den Arealen der neovulkanischen Gebieten deshalb die Gliederung mehr in *Erzrayone* üblich. Die Lokalisation der Erzrayone hat ihre Gründe in der Tektonik des tiefen Untergrundes der Neovulkanite.

Erzrayon Kremnica ist namentlich durch die Gold-Quarz-Gänge bekannt. An den Rändern sind die Limnoquarzite und Antimon, Quecksilbervererzungen entwickelt.

Erzrayon Banská Štiavnica ist durch die Gold-Silber-Blei-Zink-Erze bekannt. In die Tiefe kann man Übergänge in die Kupfer- und Wolfram-Erze bemerken. In der weiteren Umgebung von B. Štiavnica sind dann die Quecksilbererze, Manganerze, oder exhalative Körper vom Schwefel und kontakt-metasomatische Magnetitlagerstätten (Vyhne, Tisovec) in der Nähe der miozänen Diorite, bekannt.

Erzrayon Zlatá Baňa im Prešovsko-Tokajské pohorie besitzt auch solche subvulkanische hydrothermale Erzformation, die aus Blei-Zink-Silber-Erzen in der Mitte ausgebildet sind; an ihren Rändern kommen Antimonit und Quecksilber-Erze vor. In den synsedimentären miozänen Komplexen sind verschiedene Horizonte der Pelosiderite entstanden.

Auf dem Territorium Ungarns ist der *Erzrayon Mátra* mit Blei-, Zink-, Silber-Erzen und Cu-Erzen vom Belange.

Übergeordnete räumliche Einheit der Rayone ist der *Kranz der neovulkanischen Gebirge*, die im wesentlichen die Funktion der strukturell-metallogenetischen Zone besitzen.

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Die Flysch-Karpaten, östlich des Flusses Laborec, sowie neovulkanisches Gebirge Vihorlat-Gutin unterscheiden sich durch ihren Bau, Entwicklung und Metallogenese von den Westkarpaten. Mit Bezug auf ihren schwachen Erosionsgrad, bzw. Aufschließung und Anwesenheit von zwei Subprovinzen der Westkarpaten (Flysch-Zone und neovulkanisch-pannonische Subprovinz), kann man dieses Territorium, das in UdSSR liegt, als eine *selbständige, kleinere metallogenetische Provinz der Karpatoukraine* bezeichnen. Nach B. V. Merlič (1957) bildet diese Subprovinz drei lineare strukturell-metallogenetische Zonen, u. zw. von Norden gegen Süden: *Petroš-Zone, Perečin-Svaljava-Zone* und *Čop—Vyškov-Zone.*

Auf unseres Territorium reicht bloß die *Perečin-Svaljava-Zone,* u. zw. längs des nördlichen Randes des Gebirges Vihorlat, wo sich der *Vihorlat-Erzdistrikt* befindet. Er enthält kleine epithermale Gänge der Eisen-Opal, der Pyrite und Quecksilber-Mineralisation. Längs des südlichen Randes des Vihorlat, in den anliegenden sedimentären Becken kommen häufige Flöze der sedimentär-hydrothermalen Pelosiderit-Erze vor.

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Zulezt kann man einige allgemeine Probleme des räumlichen Charakters einwenig behandeln, u. zw. den Erosionsfaktor. Nach P. Routhier (1963) ist Erosionsfaktor eines der Grundkriterien der Vergleichungslehre über die Erzlagerstätten, Typologie der metallogenetischen Provinzen im Rahmen der Megaprovinzen und der geosynklinalen, polyorogenen Systeme. Seine Wichtigkeit wird hervortreten, wenn man die Westkarpaten mit Ostalpen im Westen und mit sowjetischen Karpaten im Osten vergleicht. Den *intensiv-*

sten Erosionsgrad besitzen die Ostalpen, *den mittleren* die Westkarpaten, während die sowjetischen Karpaten *den schwächsten Erosionsgrad* haben. Ostwärts von den Ostalpen wird die Intensität der Erosion des alpin-karpatischen Bogens immer schwächer und so ist ihre Folge nicht nur die unterschiedliche Abdeckung dieser Provinzen in verschiedenen Tiefenniveaus, sondern auch die übrigen metallogenetischen Merkmale (Dichte der Lagerstätten, Formen und genetische Type, Abdeckung der verschiedenen Etagen usw.). Unterschiedlichkeit der Geosynklinalen von geotektonischen und metallogenetischen Standpunkten aus im Sinne Tvalčrelidze's, Smirnow's, Routhier's usw., ist nicht nur die Folge ihrer primären Füllung, sondern auch der verschiedenen Intensität der Erosion. Infolge des ungleichen Erosionsgrades sind die strukturell-metallogenetischen Zonen der Ostalpen im bestimmten Sinne *symmetrisch*, in den Westkarpaten *asymmetrisch*; im Bezug auf die Lokalisierung der intensivsten sideritführenden Streifen, oder Magnesitlagerstätten, sind die Verhältnisse in beiden Provinzen umgekehrt (Ostalpen: nördlich von Tauern, Westkarpaten: südlich von den Kerngebirgen).

Zeitliche Metallotekte der Westkarpatischen metallogenetischen Provinz

Den Altersfragen der metallogenetischen Prozesse der Westkarpaten hatte man in letzten Jahren große Aufmerksamkeit gewidmet. Zu den geotektonisch-geologischen, mineralogisch-paragenetischen und geochemischen Studien über das Alter verschiedener Vererzungen (Ilavský, 1957, 1960; Cambel 1958; Varček 1963; Máška 1957; Zoubek 1953; Pouba 1957, 1958 usw.), hatte namentlich Kantor mit seinen geochronologischen Studien beigetragen (1957—1965). Mit diesen Arbeiten wurde in den Westkarpaten die Anwesenheit von *drei metallogenetischen Epochen* bestätigt. Die *variskische* und *alpidische Epoche* ist vollständig und man kann sie gut in Entwicklungsstadien, Phasen, Etappen, Stufen und Horizonte zergliedern, wie wir es schon früher (Ilavský 1957, 1960) und auch im Absatz über die Rayonierung angeführt haben. Unvollständig ist bloß die Kaledonische Epoche.

Kaledonische metallogenetische Epoche, die durch die Gelnica-Serie representiert ist, hat *drei Vererzungsphasen* entwickelt, u. zw.: *kambrische* mit Pb—Zn—Cu—FeS_2 Erzen; *ordovizische* mit Pb—Zn—FeS_2—Erzformation und mit den „metasomatischen" Sideriten; *silurische* metallogenetische Phase mit FeS_2—Cu Erzen.

Damit wir die ordovizische Phase besser charakterisieren können, zerteilen wir sie in *zwei Etappen*: eine mit Pb—Zn—Cu—FeS_2 Erzen und eine andere mit Siderit-Erzen. Jede von ihnen enthält einige Erzhorizonte.

Variskische metallogenetische Epoche ist durch *drei Entwicklungsstadien* vertreten, u. zw. das präorogene, orogene und postorogene, die sich vielleicht auch in fünf zerteilen lassen, d. h. geosynklinales, frühorogenes, orogenes, spätorogenes und finales.

Während *des geosynklinalen Stadiums* sind die sedimentärvulkanogenen Eisenjasspilite entstanden, die in Diabasserie devonischen Alters eingeschaltet sind (Zone der Gemeriden). Frühorogenes Stadium enthält keine Vererzungen.

Das orogene Stadium mit den Granitoidgesteinen ist hauptsächlich in den Kerngebirgen (Tatroveporiden) stark vertreten. An ihre Kontakthöfe sind manchmal kontaktmetamorphe Magnetit-Skarnlagerstätten gebunden. Die Pegmatitformationen in Granitoidarealen sind nur in sehr schwachem Maßstabe entwickelt, ebenso wie kleine Molybdänitvererzungen (Niedere Tatra, Čierna Hora).

Spätorogenes Stadium der variskischen Epoche ist vom metallogenetischen Standpunkt aus sehr wichtig. Es hat sich während des Oberkarbons abgespielt, der detritische Entwicklung mit basischem Vulkanismus aufweist. An dieses Stadium sind stratiforme Linsen der kristallinen Magnesite gebunden. Eine andere Erzformation sind „metasomatische" Sideriterze im Oberkarbon von Dobšiná.

Das finale Stadium ist durch die Permschichten in Verrucano-Fazies representiert. Diese metallogenetische Phase enthält *mehrere Mineralisationsetappen*, die in einigen Stufen und Horizonten liegen. Erzformation der Siderit und Siderit-sulphidischen Gänge in der zentralen Subprovinz kann man nicht genau datieren; sie ist aber sicher *postvariskisch* (siehe geochronologische Forschungen von Kantor 1960—65).

In den *Permschichten* sind in einigen Horizonten die Hematitlagerstätten entwickelt, die den Lahn-Dill-Typus representieren (zusammengebunden mit Vulkaniten), teils aber auch einer anderen Genese sein können (Ilavský 1958). Eine andere Erzformation ist durch Uran-Kupfer-, oder nur Kupfererze vertreten, die wiederum im Permkomplex liegen und mit dem sauren Vulkanismus zusammenhängen (Zone der Gemeriden). Dieser Etappe sind die stratiformen Siderit-Kupfererze im Perm der Tatroveporiden aequivalent (Špania Dolina, Ľubietová), was durch geochronologische Studien von Kantor (1965) bestätigt wurde (Pb-Isotope). Fast alle diese Erzformationen sind in mächtigen Evaporitserien (Anhydrite, Gypskörpern) eingeschaltet.

Die Evaporitserien des Oberperms, die Übergänge in die Untertrias aufweisen, decken sich teilweise mit dem geosynklinalen Stadium der alpidischen metallogenen Epoche.

Marine Sedimentation während *der Trias bis Kreide* in den Westkarpaten führte nicht zur Konzentration der Erze. Kleinere sedimentäre Manganhorizonte im Lias oder oolitische Eisenerze im Rhät kann man in Zusammenhang mit *frühorogenen Bewegungen* der alpidischen Ära geben.

Das orogene Stadium ist aber von metallogenetischer Sicht aus von großer

Bedeutung, weil es durch Rejuvenation der magmatischen und metamorphen Prozesse fast alle älteren Lagerstätten beeinflußt hatte (Rekristallisation, Regeneration, Mobiliesierung). Infolge dessen, sowie dank den erneuerten subkrustalen magmatischen Prozessen sind im Mesozoikum der Westkarpaten zahlreiche, aber sehr kleine Pb, Zn, Cu und Fe-Mineralisationen entstanden, teils im umhüllenden Mosozoikum, teils aber auch in den subtatrischen Decken.

Dem spätorogenen Stadium der alpidischen Epoche gehört Eozän-Oligozän. Während der Etappe, die von Oberkreide bis ins Paleozän eingreift, entstanden kleine Bauxit-Flöze, die aber keine praktische Bedeutung haben. Die Etappe des oberen Eozän ist namentlich durch die Sedimentation der karbonatisch-oxydischen Manganerze bekannt, die in mehreren Horizonten liegen.

Das finale Stadium der alpidischen Epoche ist auf die metallogenen Prozesse sehr reich; es dauerte das ganze Miozän-Pliozän durch mit bunt entwickeltem Vulkanismus und Magmatismus. Man kann es in mehrere Phasen und Etappen zerteilen.

Zum Burdigal gehören Rhyolitgesteine, die teilweise Quecksilber führend sind (Merník). Während der Phase *Helvet-Torton-Sarmat*, die sehr reich an verschiedene Typen der Andesite, Trachyte, Dazite sind, entstanden bekannte Erzformationen der subvulkanischen hydrothermalen Gänge mit Ag, Au, Pb, Zn und Hg, die den Inhalt der pannonischen jungvulkanischen Subprovinz bilden. Man kann in dieser Phase *mehrere Etappe* unterscheiden, u. zw. die mit *subvulkanischen Gängen*, andere mit der *stratiformen Schwefelmineralisation*, mit *Kontaktlagerstätten der Magnetitformation* usw. In den Sedimenten, die sich zusammen mit Vulkaniten im Meeresbecken abgelagert haben, sind mehrere Pelosiderit-Horizonte entstanden. Diese Sedimentation ist auch durch einige *Salzlagerstätten* ausgezeichnet (Torton, Helvet).

Im Pliozän ist noch eine metallogene Phase entwickelt, die hauptsächlich Quecksilber auf dem Territorium der Slowakei trägt.

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Im Zusammenhang mit zeitlichen Metallotekten in der westkarpatischen metallogenetischen Provinz muß man noch einige allgemeinen Probleme temporaren Charakters behandeln.

Die metallogenetische Kulmination im Sinne P. Routhier's (1963), die die Beziehungen zwischen der Kulmination der Orogene und Metallogenese im Rahmen einer Provinz abspiegelt, ist in den komplizierten, polyorogenen, polyzyklischen und polymetallogenen Einheiten sehr schwer zu unterscheiden.

Man kann im Prizip von verschiedenen Kulminationen sprechen. Im Laufe der kaledonischen Ära kann man als metallogenetische Kulmination die silurische Stufe (Etappe) betrachten, hauptsächlich in der Gemeriden-Zone.

In der variskischen Ära kann man von der orogenetischen Kulmination in der Zone der Tatroveporiden sprechen. Sie fällt in die bretonische Phase. Die metallogenetische Kulmination mit hydrothermalen Gängen wurde im Oberperm erreicht. *Der zeitliche Unterschied zwischen der orogenetischen und metallogenetischen Phase war etwa 80—100 Milionen Jahre.*

In der alpidischen Ära fällt die orogenetische Kulmination in den Zeitabschnitt der Mittel- und Oberkreide, die metallogenetische erst in das Endstadium (Miozän; Torton-Sarmat). Der Altersunterschied zwischen beiden Kulminationen ist wiederum etwa 100 Milionen Jahre.

Vom theoretischen Standpunkt aus ist also die zeitliche Übereinstimmung der Unterschiede zwischen der orogenen und metallogenen Kulmination bei der variskischen und alpidischen Ära der Westkarpaten beachtenswert.

So aufgefaßte metallogenetische Kulmination, in jeder Epoche behandelt, ist *monoorogenen Charakters*; dabei erfaßt sie nicht den richtigen Sinn der Kulmination in den polyzyklischen und polyorogenen strukturell-metallogenetischen Einheiten. Vom Standpunkte solcher Einheiten muß man nämlich von der metallogenen *Kulmination des polyorogenen, oder polyzyklischen Typus* sprechen.

Übereinstimmend mit Schlußfolgerungen über die Entwicklung der geologisch-tektonischen und metallogenetischen Prozesse aus breiten Gebieten einzelner Kontinente, die von Schneiderhöhn, Smirnov, Borchert, Turneaur, Raguin, Routhier etc., ausgesprochen wurden, kann man über die westkarpatische metallogenetische Provinz sagen, daß sie *polyorogenen Types ist*; *ihre metallogenetische Hauptkulmination fällt in die variskische Ära.* Diesen Zeitabschnitt kann man auch als orogenetischen und metallogenetischen *Kataklismus* bezeichnen, weil in den älteren Epochen haben sich die orogenen Zyklen nicht vollständig entwickelt; sie erreichten bloß geosynklinale Anfangsstadien und in ihnen sind nur stratiforme Lagerstätten der Kiesformation entstanden. Nach dem variskischen Kataklismus sind jedoch die orogenen Entwicklungen vollständig, mit allen Stadien, mit bunten genetischen Typen und Erzarten, die der Form und Mineralassoziation nach unterschiedlich sind.

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Nun möchte ich die Problematik der bezeichnendsten (Magnesit- und Siderit-) Erzformationen, besonders *vom Standpunkt ihres Alters*, erörtern.

Die sogennanten *„metasomatischen“ Sideritlagerstätten, die in der kambrosilurischen Gelnica-Serie* liegen (hauptsächlich Nižná Slaná und Železník), *gehören nicht der variskischen metallogenen Epoche.* Neuesten Forschungen J. Kantor's zufolge, die im geochronologischen Laboratorium unseres Institutes verfertigt wurden (mündliche Mitteilung), weisen die Pb-Isotopen in Geokronitten von Nižná Slaná, die bisdaher der nachsideritischen sulphidischen Erzformation eingereiht wurden, das altpaläozoische Alter auf. Diesen Resultaten nach muß man dann die Genese dieser Lagerstätte syngenetisch auffassen, d. h. im Zusammenhang mit sedimentären oder vulkanogenen Prozessen, wie darauf schon Kudenko—Stecenko (1963) hindeuteten.

Wenn man auch andere Erscheinungen an dieser Lagerstätte betrachtet (z. B. stratiforme Linsen, Feinstratifikation, Facieserscheinungen und Übergänge in die Ankerite und Kalksteine, stratigraphische und paläogeographische Kriterien bei den Aufsuchungsarbeiten), muß man auch die synsedimentäre Genese für möglich halten. Grobkristallinen Charakter und Endform haben diese Lagerstätten im Verlaufe der alpidischen Rekristallisationsprozesse angenommen.

Die westkarpatischen Magnesitlagerstätten liegen in dem Karbon, dessen Profil in fast allen Lokalitäten ständig ist, d. h. an der Basis Quartzsande oder Quartzite, die in die Sandschiefer übergehen; letztere weisen auch nach oben Übergänge in dunkle, organogene, graphitische Schiefer auf. In diesem Horizont pflegen stratiforme basische Eruptivgesteine der gabbroidischen Zusammensetzung (Gabbroamfibolite) und Lagen der Diabastuffe und — tuffite anwesend zu sein. Darüber kommen karbonatische Gesteine: Dolomite, Kalksteine mit Magnesitkörpern, in welchen hie und da auch noch Diabastuffe und -tuffite und fast immer dunkle graphitische, organogene Schiefer zu sein pflegen. Magnesitkörper sind größtenteils unregelmäßig, massiv (Magnesite), an den Rändern, wo feine Lagen von Magnesit, Dolomit, Schiefern, oder Diabasen vorkommen, sind sie aber klar geschichtet. Der Kristallinitätsgrad hängt von der Detritusbeimegung oder Bitumänen ab. Zahlreiche sulfidische Mineralien in den Magnesiten (Trdlička 1956, 1960) kommen in Form von sehr feinen Körnchen bis Äderchen in den Magnesitkörnchen und in ihren Intergranularen vor. Nur Galenit, Sphalerit, Chalkopyrit und Tetraedrit bilden auch größere Gänge in Magnesiten, oder in hangenden Dolomiten und Kalken.

Die Pb-Isotopen aus solchen Gängen und Äderchen des Galenites, weisen nach J. Kantor (mündliche Mitteilung) das variskische oder alpidische Alter auf. Magnesitkarbon ist sehr stark disloziert und alpidisch metamorphosiert.

Die Veporidenzone, die in der unmittelbaren Nähe des magnesitischen, auf sie aufgeschobenen gemeriden Karbons liegt, ist alpidisch metamorphosiert allerdings unter den mesozonalen, gegen Westen almählich bis katazonalen Bedigungen. Es handelt sich um die sog. Glimmerschieferzone der Veporiden.

In dieser Zone liegen zahlreiche *Magnesit-Talk-Lagerstätten.* Diese haben massiven, oder schichtigen Charakter und sind in den Dolomiten und Kalksteinen eingeschaltet, die in den Glimmerschiefern, Paragneisen, Metaquartziten, amphibolitischen Schiefern und Amphiboliten liegen. Talk ist an diesen Lagerstätten in zwei Formen vertreten: massig in scharf differenzierten Lagen an den Rändern der Magnesitkörper, oder selbständige Lagen in Magnesitkörpern bildend und schuppig, sehr intim mit Magnesitkörnchen durchgewachsen.

Die Gesteinskomplexe der Magnesit-Talk-Lagerstätten der Veporiden-Zone erinnern also sehr an die Verhältnisse in der nahen Gemeriden-Zone. Unterschiede bestehen bloß im Grad der Metamorphose: in den Veporiden die mesozonale Metamorphose, wo ihr Alter mit Hilfe der A-K-Methode Kantor (1960) als alpidisch neuer bestimmte; in den Gemeriden die Metamorphose epizonalen Charakters.

Nach den angeführten geologischen und isotopischen Angaben kann man sagen, daß die Magnesitlagerstätten des gemeriden Karbons, ebenso wie Magnesit-Talk-Lagerstätten der Veporiden-Zone syngenetisch, vulkano-sedimentär sein können, u. zw. abhängig vom basichen Vulkanismus im Oberkarbon. Ähnliche Ansicht haben schon längst für die ostalpinen Magnesitlagerstätten W. Siegl und Leitmeier, für die mantschourischen Nishihara, für die spanischen Llarena, für die uralischen Starostina, Ivanov u. a., ausgesprochen (bei uns Zorkovský 1955; Kudenko—Stecenko 1963).

In der alpidischen Ära, unter dem Einfluß der Metamorphose und Rekristallisation, erworben sie den grobkristallinen Charakter und die Folgen der „Metasomatose“ sind ausgeprägt geworden. Stärkere Metamorphose in den Veporiden hat die Entstehung bedeutender Talkmengen zu Lasten des Magnesites verursacht. Infolge der alpidischen Mobilisation und Metamorphose entstanden auch in Magnesiten und in ihrer nahen Umgebung Regenerationsäderchen und metasomatische Vererzung mit Pb—Zn—Cu—Erzen. Nirgends auf den westkarpatischen Magnesit-Lagerstätten treten Sideritvererzungen auf.

Bei der Beurteilung der postvariskischen metallogenetischen Phasen muß man noch eine Tatsache berücksichtigen, u. zw. die Existenz *der großen Siderit-Kupfer-Lagerstätten im Perm der Tatroveporiden bei Ľubietová und Špania Dolina* (Libethen, Herrengrund). Diese Lagerstätten sind alpidisch sehr stark disloziert und metamorphiert, in Permsandsteinen und Schiefern eingeschaltet, an manchen Stellen stratiform, anderswo gangförmig, zum Teil auch im liegenden Kristallin, oder im übergelagerten Mesozoikum.

Sehr zerstreute Galenitkristalle in der Gangausfüllung wurden einer isotopischen Pb-Analyse unterzogen und haben das permische Alter aufgewiesen (Kantor 1965). Es ist also möglich, daß diese Lagerstätten auch syngenetisch und stratiform sind und während des alpinen Orogens sehr umgearbeitet wurden. Analoge syngenetische und stratiforme Vererzung mit feinkörnigem, schichtigen Hematit und Siderit, wurde in den letzten Jahren in der Gemeriden-Zone studiert (siehe Ilavský 1957).

Stoffliche Metallotekte der westkarpatischen metallogenetischen Provinz

Wie es aus vorhergehenden Zeilen hervorgeht, gibt es in den Westkarpaten viele Metalle und Erze, die für einige räumliche und zeitliche Einheiten typisch sind. Mit Berücksichtigung der vorhererwähnten Metallotekte lassen sich diese Metalle und Erze in Gruppen einreihen.

Als *metallische, oder Erzkomplexe* bezeichnen wir die größten stofflichen Einheiten, die in einer metallogenetischen Epoche entstanden sind; sie gehören einem genetischen Typus und sind stofflich und geochemisch sehr nah oder identisch.

Erzkomplex der Kieserze, die der kaledonischen Epoche entsprechen, sind im allgemeinen durch stratiforme Erzlinsen in verschiedenen Eruptivgesteinen vertreten, an deren Zusammensetzung sich Pyrit-Kupfererzkies-Galenit-Sphalerit-Erze beteiligen.

Erzkomplex der stratiformen Lagerstätten des Postvaristikums ist durch Magnesiterze, Siderite und Kupfer-Uranium-Erze gebildet.

Ein anderer Erzkomplex ist durch *hydrothermale spätvariskische Siderit-Sulphid-Gänge* vertreten, die hauptsächlich in den Tatroveporiden und Gemeriden entwickelt sind. Die Hauptminerale sind durch Fe, Cu, Pb, Zn, Ni, Co, Hg, Sb, Au, Ag und Ba vertreten.

In der jungvulkanischen, pannonischen Subprovinz gibt es einen anderen Erzkomplex, der durch *subvulkanische Sulfidgänge* repräsentiert ist. An ihrer Zusammensetzung nehmen namentlich Blei, Zink, Silber und Gold, teilweise auch Kupfer, Antimon und Quecksilber teil.

Da die Erzkomplexe sehr breite räumliche oder zeitliche Einheiten vorstellen, müssen wir einen engeren Begriff, u. zw. *Erzformation* benützen, damit wir einen streng umrahmten Zeitabschnitt (metallogenetische Phase oder Etappe) mit gleichen genetischen Typen und gleichem Inhalt bezeichnen können.

Im kaledonischen Komplex der Gemeriden-Zone gibt es z. B. eine Formation der Pyriterze, eine andere mit Pyrit-Kupfererzen, Pyrit-Kupfer-Blei-Zinkerzen usw. Jede von ihnen befindet sich in einem anderen stratigraphischen Rahmen (Kambrium, Ordovizium, Silur) und ist durch abweichende *Mineralassoziation* vertreten. Das kaledonische Alter hat auch die Siderit-Erzformation (ehemalige metasomatische Siderite), die geologisch und stofflich ganz verschieden ist.

Im postvariskischen Erzkomplex der stratiformen Lagerstätten muß man vom stratigraphischen und metallogenetischen Standpunkt aus mehrere Erzformationen unterscheiden.

Magnesit-Erzformation im Oberkarbon der Gemeriden-Zone ist in einer Länge von 120 km bekannt, ohne daß sie ihre mineralogisch-geochemische Zusammensetzung, lagerstättenkundliche und geologische Verhältnisse ändert. Eine andere Erzformation bilden „metasomatische" Sideritlagerstätten im Oberkarbon, die sich geologisch von Magnesiten sehr unterscheiden, aber den altpaläozoischen stratiformen Sideritlagerstätten sehr nahe stehen.

Die Erzformation der hydrothermalen postvariskischen Siderit-sulphidischen Gänge ist in allen Kerngebirgen (insgesamt der Gemeriden-Zone) geologisch, stofflich und geochemisch einheitlich und selbständig.

Die Erzformation der stratiformen oberpermisch-untertriassischen Kupfer-Eisen-Uran-Vererzungen ist geologisch und stofflich wieder abweichend.

Eine andere, sehr ausgeprägt entwickelte Erzformation ist durch *subvulkanische hydrothermale Gänge* vom Banská Štiavnica und Kremnica-Typus vertreten, an denen hauptsächlich Silber, Gold, Blei und Zink vorkommen.

Man kann ganz gut sehen, daß eine Erzformation zeitlich mit einer metallogenetischen Phase zusammenhängend ist.

Es gibt aber auch Erzformationen, die oft *verschiedene Mineralparagenesen, oder mineral-paragenetische Assoziationen enthalten*, die etwa abweichend durch ihre Geologie, Zusammensetzung und Herkunft sind. Auf den stratiformen Kieslagerstätten im Altpaläozoikum handelt es sich z. B. um verschiedene metamorphe Fazies, auf sedimentären Lagerstätten um verschiedene Sedimentationsbedigungen, auf den Erzgängen um verschiedene Tiefenbereiche, oder Einflüsse der Nebengesteine, die die Zusammensetzung vom stofflichen Standpunkt aus beeinflußt hatten.

Der Ausdruck Mineralisationsperiode (Kutina 1956) soll die Folge der Ausscheidung einzelner mineral-paragenetischen Gruppen präzisieren, und zwar mit Rücksicht auf den ganzen Ausscheidungsprozeß der Lagerstätte. Auf den hydrothermalen Gängen postvariskischen Alters spricht man z. B. von der sideritischen Periode, oder Zufuhrperiode, von den sulphidischen Perioden usw. Die Namen sind dann nach den überwiegenden Mineralen benützt. Es gibt Fälle, wo sich die Zusammensetzung der Ausfüllung mehrmals repetiert; für solche Fälle hatte man den Ausdruck *Repetitionsperiode* benutzt (J. H. Bernard 1956).

Mit dem Begriff *Erzgeneration* will man denselben Mineral bezeichnen, der sich mehrmal in der Sukzessionsreihe auf derselben Lagerstätte repetiert.

Andere Begriffe, die sich sehr gut von der stofflichen Seite ableiten lassen, sind Intensität der Vererzung, oder *Akkumulationsfaktor* (Petrascheck 1955; Routhier 1963). Dadurch wird die Dichte der Lagerstätte auf gewissem Gebiet ausgedrückt. Auf der metallogenetischen Karte 1:1,000 000 haben wir insgesamt 260 Erzlagerstätten eingetragen, was ungefähr 10 % von der Gesamtzahl in Wirklichkeit ist. Daraus kommt heraus, daß man die Dichte der Vererzung in den Westkarpaten durch eine Lagerstätte pro 20 Quadratkilometer einführen kann.

Einzelne Subprovinzen der Westkarpaten haben *verschiedene Koeffiziente*, was für die Auswertung der Prognosen sehr wichtig ist. Von der angeführten Lagerstättenzahl befinden sich in der Subprovinz der äußeren Karpaten etwa 5 %, in der neovulkanischen, inneren Subprovinz etwa 15 % un der Hauptanteil mit 80 % befindet sich in der zentralwestkarpatischen Subprovinz.

Innerhalb der letzten Subprovinz sieht man sehr auffallende Unterschiede auch zwischen einzelnen strukturell-metallogenetischen Zonen, oder Etagen der Erzdistrikte. Zum Beispiel die Zone der Tatroveporiden ist fünfmal so groß wie die Gemeriden, die Intensität der Vererzung ist aber in der Gemeriden-Zone fünfmal größer.

Beachtenswert ist der *Vergleich einzelner metallogenetischer Etagen (Epochen)*. Der kaledonischen Epoche gehört von der Gesamtzahl der Erzlokalitäten etwa 5 %, der variskischen Epoche, deren Lagerstätten auch auf der kaledonischen Etage aufgelagert sind, 75 % der Lokalitäten und der alpidischen etwa 20 %.

Mit Bezung auf die polyzyklische Entwicklung dieser Subprovinz, muß man dann manche ältere metallogenetische Etage vom Standpunkte der Dichte der Lagerstätten als *„blutüberfüllt, d. h. nachträglich bereichert“* u. zw. auf *monophasische bis polyphasische Art,* betrachten.

Der *Akkumulations-, resp. Konzentrationsfaktor* im Sinne Routhier's (1963) und Brock's (1964) drückt die Beziehungen zwischen der Tonnage der Lagerstätten und ihrer Menge in bestimmten räumlichen metallogenetischen Einheiten, oder zwischen der Zahl der größten und kleinen Lagerstätten, aus.

Mit Bezug auf die Intensität der Vererzung im Verlaufe einzelner metallogenetischen Epochen, Stadien und Etappen, sowie auf die mono- bis polyphasische Bereicherung, oder „Blutüberfüllung“ einiger älterer metallogenetischen Etagen, muß man den *primären Konzentrationsfaktor*, d. h. im Sinne der Entwicklung, und den *resultierenden Konzentrationsfaktor* unterscheiden.

Der primäre Konzentrationsfaktor der kaledonischen Epoche in der zentralkarpatischen Subprovinz beträgt ungefähr 20 %, in der variskischen 10 % und in der alpidischen 5 %. Wenn wir einzelne Subprovinzen zwischeneinander vergleichen, wird im ganzen für die zentralkarpatische Subprovinz der Konzentrationsfaktor 10 %, für die neovulkanische 5 % und bei den äußeren Karpaten (Flyschzone) fast kein (es handelt sich hier um resultierenden Konzentrationsfaktor).

Weitere Beispiele können wir aus der Gemeriden-Zone nehmen. Im Magnesit-Karbon der westlichen Gemeriden ist der primäre Konzentrationsfaktor ungefähr 50 %, wenn es von Gesamtzahl ca 30 Magnesitlagerstätten 15 große mit Vorräten von über 10 Milionen Tonnen gibt. In der Gelnica-Serie ist der primäre Konzentrationsfaktor cca 10 %; wenn wir jedoch die jüngeren, aufgelagerten Lagerstätten der variskischen Ära in Betracht nehmen, ist ihr resultierender Akkumulationsfaktor ungefähr 50 %.

Man sieht also, daß der Begriff des primären, oder resultierenden Akkumulationsfaktors von beträchtlicher theoretischen und praktischen Bedeutung ist.

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Im weiteren wollen wir uns mit Begriffen *typomorphe Metalle* und *typomorphe Erzformationen*, sowie *typomorphe genetische Typen* befassen (Tvalčrelidze 1957; Petrascheck 1955, 1963; Smirnov 1960; Routhier 1963).

Mehrere Autoren erklären die Entstehung der wiederholenden Metalle und Erzarten mit dem Prinzip des *„Erbens“* (Raguin, Routhier, Smirnov). Andere brachten (im Sinne der metamorphen Regeneration und Mobilisation durch die überhitzenden Wässer) in die Typologie ein neues Element; die Konsanguinität in der Zusammensetzung läßt sich auch aus identischen Prozessen der Geotektonik und Metallogenese abführen (Konvergenzerscheinungen; Schneiderhöhn, Smirnov). Zur Klarstellung dieser Typologie trägt auch die Konzeption des *Transformismus* (Eskola, Sederholm, Marmö) u. a. bei.

Von den typomorphen Metallen hat das *Eisen* die größte Bedeutung in der westkarpatischen metallogenetischen Provinz. Es ist häufig in allen Entwicklungszyklen, u. zw. immer in derselben Etappe. *Blei* und *Zink* sind weitere typomorphe, in allen Epochen durchlaufende Metalle. *Kupfer* weist analogen Charakter der Typomorphie wie *Pb—Zn*. Es kommt mit ihnen hauptsächlich bei den syngenetischen Vererzungstypen zusammen vor, während sie sich bei den hydrothermalen voneinander räumlich differenzieren.

Weitere typomorphe Metalle sind *Gold, Silber und Quecksilber*. Ihre Mengen sind nicht groß, jedoch koinzidieren sie zeitlich und räumlich mit Pb—Zn und Cu—Erzen.

Quantitative Kulmination der einzelnen typomorphen Metalle, die wir angeführt hatten, ist für die ganze westkarpatische Provinz sehr typisch.

Eisen ist namentlich für die kaledonische Ära typisch, jedoch von der Gesamtmenge der Fe-Erze gibt es da nur gegen 10 %. Quantitative Kulmination erreichen Fe-Erze im Endstadium der variskischen Ära, wo sie bis zu 80 % konzentrieren. Der Rest, etwa 10 % ist unregelmäßig in der alpidischen Ära verteilt mit Maximum in der Untertrias; in den jüngeren Etappen (Rhät, Lias, Kreide, Miozän) sind sie verhältnismäßig selten. Umgefähr 90 % Eisen ist in der zentralkarpatischen Subprovinz konzentriert, u. zw. 80 % in der Gemeriden-Zone und Rest in den Tatroveporiden. Aber auch in der Gemeriden-Zone sind sie unregelmäßig verteilt, wenn ihre Anhäufung längs der Kontaktlinie der Gemeriden-Zone mit Veporiden liegt.

Die Entwicklungskurve der *Blei-Zink-Erze* unterscheidet sich diametral vom Eisen; sie ist umgekehrt. Ungefähr 10 % des Pb—Zn gehören den Kieserzen der kaledonischen Ära; analoge Mengen sind auch in der variskischen Ära auf den metasomatischen Lagerstättentypen. Maximale Konzentration erreichen sie in der alpidischen Ära, u. zw. in der finalen Etappe (Miozän) lokalisiert in der pannonischen Subprovinz. In dem Querprofil der Westkarpaten wächst die Zahl der Pb—Zn-Lagerstätten und ihre Größe von der äußeren NW Seite gegen die inneren Zonen im SO (Petrascheck 1955).

Einen unabhängigen Verlauf von den vorerwähnten weist die Kurve des *Kupfers* aus, dessen Bildung schon in der kaledonischen Ära hoch war. In der variskischen Ära ist sie auch hoch u. zw. mit maximaler Konzentration im Perm, während in der alpidischen Ära ihre Mengen niedrig sind. Räumliche Verbreitung des Kupfers kopiert im großen und ganzen das Eisen, in dem seine maximale Konzentration in der Gemeriden-Zone liegt.

Die Kulminationskurve des *Goldes* und *Silbers* koinzidiert mit jener der Pb—Zn-Erze; auch ihre räumliche Distribution ist mit ihnen gleich. Dagegen zeigt *Antimon* (was die quantitative Kulmination, als auch die räumliche Konzentration betrifft) nähere Affinität zum Kupfer.

Als Schlußfolgerung der erwähnten geochemischen Faktoren man kann sagen, daß das wichtigste und vorherrschende Metall in der westkarpatischen Metallogenen Provinz *Eisen* ist. Was die Tonnage betrifft, wird es bloß durch *Magnesium* übertroffen. Das weitere wichtige Metall ist *Kupfer*; die übrigen, wenn sie auch ubiquitären Charakter haben *(Blei, Zink, Antimon, Gold, Silber, Quecksilber)*, sind mengenmäßig unbedeutend.

Was die Typologie der genetischen Typen betrifft, man kann als ein instruktives Beispiel die *stratiforme Eisenerzmineralisation* anführen, die in allen Epochen, hauptsächlich mit dem basischen Vulkanismus verbunden sind, u. zw. in den initialen Entwicklungsstadien (Gelnica-Serie; Magnetit-Hämatite, Phylit-Diabas-Serie: Hämatit-Magnetite, Untertrias: Hämatite, Rhät-Lias: oolitische Hämatit-Vererzungen). Stratiforme Lagerstätten sind aber auch mit Spät- und Final-Entwicklungsstadien verbunden, u. zw. wieder mit dem basischen Vulkanismus (variskische Ära: Magnesite, schichtige Cu-Erze; alpidische Ära: Schwefelflöze, Pelosiderite usw.).

Die Gruppe *der hydrothermalen Gänge* ist umgekehrt nur für finale Entwicklungsstadien der Orogene typisch (Siderit und Siderit-sulphidische Formation des Varistikums, subvulkanische hydrothermale Gänge mit Sulphiden der alpidischen Ära).

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Zusammenfassend kann man zur Metallogenese der Westkarpaten sagen:

(a) der mittlere Erosionsgrad hat fast alle metallogenen Etagen, mit kaledonischer angefangen bis zur alpidischen, ideal abgedeckt, so daß man fast alle räumlichen Einheiten, wie Subprovinzen, Etagen, strukturell-metallogenetischen Zonen, Streifen, Distrikte, Erzfelder und Lagerstätten instruktiv sehen kann;

(b) die metallogenen Prozesse haben sich in diesem Raum in Abhängigkeit vom Vulkanismus in Anfangsstadien aller Epochen, weiter von dem sauren und intermediären Plutonismus in den mittleren Stadien und zum Schluß in den späten und finalen Stadien wiederum in der Abhängigkeit von basischen oder sauren Typen des Vulkanismus, abgespielt.

(c) Die polyzyklische Wiederholung der metallogenen Prozesse hatte sich oft auf dasselbe Gebiet im Rahmen der Zone, des Streifens, Erzzuges, Erzfeldes, bis der Lagerstätte einigemal aufgelagert. Bei mehrfacher Wiederholung der metamorphen Prozesse hat sich der heterogene Charakter des Ursprungs der Erzstoffe soweit kompliziert, daß man oft nur sehr schwer ihren magmatischen, resp. apomagmatischen, oder vulkanischen, sowie regenerativ-mobilisierenden, bzw. lithogenen Ursprung unterscheiden kann.

Demzufolge stellt das westkarpatische Gebiet einen Prototyp der *polyzyklischen, heterogenen Provinz* im Rahmen der alpin-karpatischen mediterranen Megaprovinz dar, die den Bestandteil des Alpen-Himalaya-Systems Eurasiens bildet.

Zum Schluß möchte ich betonen, daß der Polyzyklismus der sedimentären, paläogeographischen, tektonischen, plutonischen, vulkanogenen, metallogenen und hauptsächlich metamorphen Prozesse in den gefalteten Systemen, Provinzen bis Lagerstätten als die wichtigste Frage das Problem der Metamorphose, resp. Polymetamorphose der Erzlagerstätten und ihrer Wirkung auf sie (was die Textur, Struktur, Paragenese und Geochemie anbelangt) in Vordergrund stellt, worauf man sich in der nahen Zukunft konzentrieren muß.

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RELATION OF DEEP-SEATED STRUCTURE TO THE DEVELOPMENT OF SUBSEQUENT VOLCANISM IN CENTRAL SLOVAKIA

Abstract. The substratum of Central Slovakian neovolcanics has been studied for many years. The research was based on the results of general geological maps and also on geophysical (in particular gravimetric and of late also seismic) researches, on the results of structural borings and on special volcanologic studies. In 1965 the first results outlining the rough scheme of the relief of the substratum and the principal features of the tectonic plan of the substratum underlying the neovolcanics were published. Our further researches determined more accurately the areal distribution of the principal units, the course of tectonic lines, and the main structural elements of the deeper parts of the earth's crust. In this paper the results of our investigations are summarized.

After concluding our synoptic geological and geophysical researches (scale 1:200,000) we have started detailed geological-volcanological and geophysical studies (scale 1:25,000), which were supplemented by structural borings. The geophysical results along with the character of the rocks and the proper structure of the subsequent volcanism were taken as the base for the construction of the map of the pre-Neogene relief.

The considerable, in the peripheral parts of the neovolcanic complex absolute prevalence of pyroclasts over lava flows produces in the maximum part of the area in question a sufficiently distinct physical boundary between the Neogene complex and its substratum, thus enabling the successful application of geophysical methods.

From the geophysical point of view, the results of gravity measurements on scale 1:25,000 supplemented by geoelectrical and geomagnetic methods are of fundamental importance.

Maps of Bouguer anomalies, constructed with a uniform density $S = 2.2\,kg/dm^3$ used for the reduction procedure, have been analysed both on the regional and residual components (Griffin 1949), and also maps of second vertical derivatives of gravity according to Elkins' (1951) and Rosenbach's (1953) formulas have been deduced. In areas, where detailed gravity measurements have hitherto not been carried out, maps of residual gravity anomalies and maps of higher derivatives inferred from maps of regional gravity mapping (scale 1:200,000) have been used. For this purpose the interval of centering $S = 4$ km is the most favourable one as it best reflects the morphological structure of the hidden pre-Neogene substratum. A smaller radius includes to a considerable degree the density inhomogeneities of the complex of volcanic and volcanic-sedimentary rocks (fig. 1).

For the *construction of the relief of pre-Neogene formations* the residual gravity field map is relatively advantageous because the regional effect of the deeper zones of the upper part of the earth's crust is strongly suppressed in it (apart from the marked density boundaries at the contact of the granite-gabbroic and peridotite strata, i. e. Conrad's and possibly also Moho's plane of discontinuity, also the gravity effect determined by the boundary between the higher zone of metamorphic rocks and the deeper granite zone is obliterated).

In addition to the inferred gravimetric maps, the anomalous gravity field has been interpreted along profiles connected with structural borings and outcrops of the substratum. The interpretation of these profiles has been done by means of Gamburcev's transparent sheet. The modelling has been carried out both on analogic and digital computers. The effects of disturbing bodies producing areal anomalies of the gravity field (Both—Smith 1958; Nettleton 1954; Saxov 1954, and others) have been also interpreted. The so-called fault zones (steep Bouguer gradients) of the gravity field have often been evaluated as effect of the vertical step of the in density heavier substratum (Malovičko 1960), whether of tectonic or erosion origin.

As the resultant picture of the map of complete Bouguer anomalies and also of the inferred gravity maps reflects apart from the morphology of the substratum also various other disturbing effects which are for the construction of the scheme of the substratum undesirable, it has been necessary to consider these effects and to eliminate them to the maximum possible degree. In the main we have been concerned with the manifestations of more significant inhomogenities not only in the substratum, but also in the neovolcanic complex and in its cover (pliocene sediments at the north-western periphery of the Panon basin). In addition to the most advantageous methods of both qualitative and quantitative interpretations of the gravity field anomalies, results obtained by other geophysical methods have also been used. The presence of hidden basic bodies, determining relatively positive gravity anomalies, which could erroneously be interpreted as a phenomenon of morphological elevations of the substratum, has been in some places detected magnetometrically. Under favourable physical and geomorphological conditions, for the determination of the absolute depths of the substratum and of spatially extensive objects in the volcanic-sedimentary complex respectively, geoelectrical, and in the peripheral parts of the region also seismic methods have been employed.

The construction of a *tectonic map of the pre-Neogene substratum* requires a greater amount of data than the scheme of its relief. This is due, above all, to the necessity of determining the boundary of the individual geological units which can differ considerably in their stratigraphical position or lithology, but not in physical parameters. A conspicuous morphology of an otherwise in density homogeneous substratum can also be determined by the destruction of the relief in various geological periods, and not by tectonic processes which determine the origin of faults.

Under favourable conditions, the first case, that is the tectonic contact without a conspicuous density boundary (e. g. contact of metamorphosed schists with an in density homogeneous layer of limestone, on a hidden platform-like relief of the substratum) can be detected by some geoelectrical methods, exceptionally also magnetometrically. The second case, that is the determination whether the morphological step of the substratum was modelled by tectonic effects can be deduced only from the correlation of the linear course of the horizontal Δg gradient with known or assumed directions of the fault lines. It must be stated that the effects of tectonically determined density boundaries can be considerably deformed by bodies with a different density. These bodies occur both in the Neogene complex and in the substratum. The disturbed effects of these bodies obliterate to a certain degree the influence of their own tectonically conditioned step or the boundary of two mediums differing in density.

For the *detection of hidden morphological structures* as well as for the construction of a tectonic map of the pre-Neogene volcanic basement (using partly the results of gravity mapping), the knowledge of the density parameters of the rocks of the Neogene complex and of its substratum is unavoidable. Our main task has been to extrapolate the depth of the hidden substratum between the structural borings and its outcrops. It must be emphasized that manifestations (gravity effects) of hidden morphological structures might be erroneously ascribed to not identified density inhomogeneities, and the tectonic contact along the course of the assumed fault line to some horizontal gravity gradients.

For the *evaluation of the possibility of following the hidden morphological structure* — the relief of the pre-Neogene formations — with respect to its emplacement at depth it is also necessary to study the question of the vertical change of densities of otherwise both in petrography and lithologically analogous rock complexes. It is known that sediments at greater depth generally acquire an equivalent density or such approaching the densities of the substratum. In some specific cases the Tertiary sediments infilling the deep sedimentary area can thus acquire greater densities at their base than their substratum.

This occurs especially when the substratum is composed of acid magmatites relatively light in density. Consequently, the gravimetric research of considerably deep sedimentary areas is thus faced with the unfavourable fact of a limited or even totally impossible mapping of the relief underlying the Neogene cover. An inversion in the values takes place in case that the deep lying Neogene sediments acquire a greater density at their base than the crystalline basement. On a gravity map, the morphological elevations of the substratum occur as gravimetric depressions and, vice versa, the depressions in the relief of the substratum appear as gravitational elevations. These questions have been already studied by many authors (Uhmann 1959; Vyskočil 1959; Ibrmajer—Motlová 1961; Šutor 1964).

As already stated above, in the maximum part of the area studied there is a relatively considerable density boundary between the volcanic and/or volcanic-sedimentary complex and the pre-Neogene (pre-Tertiary) formations. In the central part, considerable density differences between the individual petrographical types of the volcanic masses can also be observed (fig. 1). The densities of neovolcanic rocks are determined both by porosity (tuff, agglomerate, lava, extrusive bodies) and by their basicity (rhyolites, andesites, basalts). In addition to pyroclastic rocks, an increased density variability can also be observed in the most frequent rock — the andesites. They vary from acid and porous (vitrophyric andesites) up to basic and massive (compact) types (pyroxene and basalt andesites). Of the rocks of the Neogene substratum, acid intrusive rocks and the flysch complex indicate lesser densities; medium densities are in metamorphosed schists and carbonates (limestone, dolomite); highest densities are in basic and particularly in ultrabasic igneous rocks. At the southern peripheral part of the volcanics and in some intravolcanic depressions, also Miocene sediments without or with a subordinate volcanic fraction are present. Generally, these sediments are relatively homogeneous in density with the exception of the less extensive conglomerate and breccia layers. The density of the Podunajská panva (basin) are much more variable (Orlický 1967). With respect to the nature of sedimentation and to the lithofacial and tectonical development of the various sedimentary areas, the complex of Paleogene rocks in the region of the neovolcanics are characterized by rather varying density parameters. This diversity is especially apparent in the densities of the Inner Carpathian Paleogene in relation to the densities of Paleogene rocks in the south Slovakian-north Hungarian sedimentary area.

As ensues from the discussed analysis of the density parameters of the various rock types or complexes, different densities occur not only within the volcanic complex and its substratum, but also in the proper stratigraphically equivalent rock formations. It is, however, generally known that rocks in which the densities attain the outer limit are relatively scarcely distributed both in volcanic formations and in the substratum.

Magnetic, resistivity, and elastic parameters also vary widely. Because the methods utilizing these parameters played in the research of the region of the Central Slovakian neovolcanics only a subsidiary part, their analysis is not given in this paper.

With respect to the construction of the scheme of the relief it must be stated that the course and the aproximate depth of the morphological forms of the pre-Neogene relief are determined with a greater degree of probability in those regions, from which detailed gravity measurements and also information obtained by other geophysical methods have been available and, in particular, where the physical, especially density boundaries between the Neogene complex and pre-Neogene substratum are more pronounced.

The interpretation of the course of tectonic lines is based on the correlation of the geological situation at the margin of the volcanics with the linear course of intensive horizontal gravity gradients, further on the results of structural borings, and on the geological structure of the inliers representing the exposed substratum of the volcanics.

The localization of tectonic lines on the surface in the neovolcanic complex is based on the mutual geological positions in the heterogenic formations; on the analyses of jointing and its effects on the extent and development of morphological steps in the monogenetic formations, and on the development and/or prevailing orientation of the river system in the final result.

Relief of the substratum

Until the beginning of the main volcanic activity, the relief of the substratum developed in agreement with the development in the other parts of the Inner Carpathians. Its modelling began after the main Cretaceous orogenic phase prior to the Upper Cretaceous (i. e. Gosau). It can be assumed that in the Upper Cretaceous deposition took place locally in some parts of the depression. Between the Upper Cretaceous and the Eocene, the relief was further shaped by erosion and already during this period a considerable part of the substratum was denuded down to the crystalline basement, especially in the southern parts of our territory. In the Eocene or Oligocene respectively the greater part of the territory was covered by deposits of the Inner Carpathian Flysch.

Pronounced morphological forming was effected in the period between the Paleogene and the main volcanic activity. The products of volcanism buried the old morphological forms fixing thus the final geological situation formed by erosion of that time. In the initial stages of the Miocene, marine deposition extended into the southern and south-western regions of the neovolcanics. The pre-volcanic relief was considerably even, rising gradually from the south towards the north. During the volcanic activity and after its termination, the substratum was affected by tectonic uplift and subsidence which gave the relief its final form (fig. 3).

In general, the present-day relief of the substratum is therefore characterized by a great morphological dissection (in the order of amplitude above 1500 metres). According to the morphological character, the substratum can be divided into two parts, the south-western and the north-eastern, ones, which are separated from each other by a conspicuous NW-SE striking ridge between Čaradice and Šahy (the so-called Santovce-Túrovce ridge and Čaradice elevation), which between Šahy and Levice locally occurs at the surface.

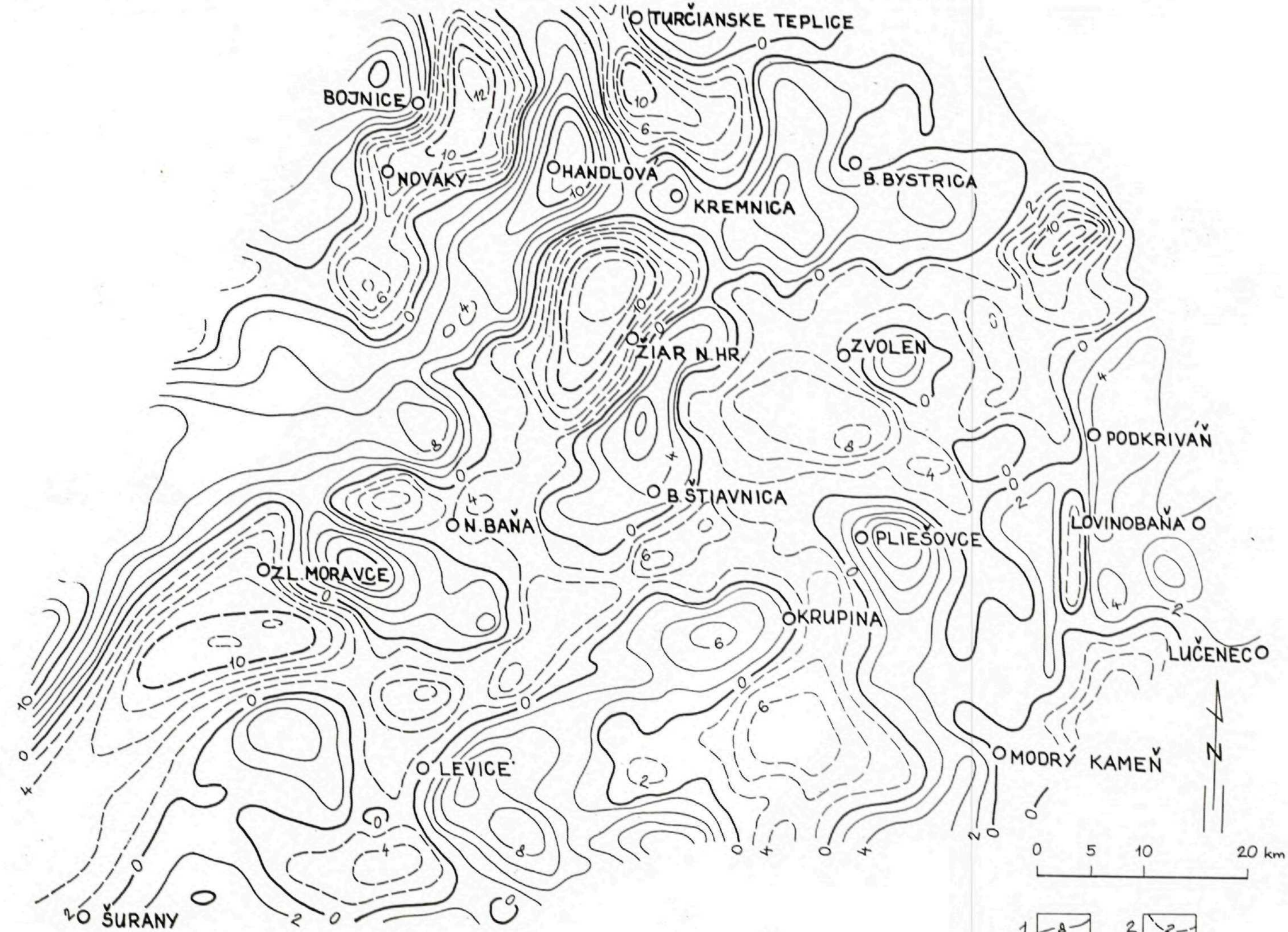
The south-western part represents essentially the margin of the Podunajská panva (basin) and it is morphologically relatively little dissected. Nevertheless, also in this part occur more pronounced basins attaining the depth of 2000 to 4000 metres, which are separated from each other by higher plateaus.

The north-eastern part comprises the substratum proper of the Central Slovakian neovolcanics. Today it is characterized by a great morphological dissection. In the north-east it is limited by the conspicuous morphological NW-SE striking Malachovo-Lieskovec ridge.

In the north-eastern part we have determined greater morphological elevation and depression forms. Of the elevation bodies it is especially the NE-SW striking Rudno ridge outcropping in the Hodruša-Štiavnica inlier, further the Handlová ridge of the same strike, which represents a narrow connection between Mt. Tríbeč and the Malá Fatra Mts. Concerning to the depression-forms it is in particular the Žarnovica depression with the conspicuous Žiarská kotlina (basin) exceeding in depth 2000 metres, the Krupina depression and Bzovik basin approximately 2000 metres deep, and the Bacurov basin attaining the depth of 1500 to 2000 meters. Moreover, there is the Modrý Kameň depression with the conspicuous Trenčská kotlina (basin) roughly 2000 metres deep. The plains and plateaus between the elevations and depressions constitute the connecting links. One of the largest is the Lučenec plateau and in the south-eastern part of the Central Slovakian neovolcanics, the Sebechleby plain.

Fig. 1. (p. 79) The map of residual anomalies of gravity force acc. to Griffin, $S = 4$ km, $\Delta g_{res} = \Delta g_o - \frac{1}{8} \sum_{1}^{8} \Delta g$; $R = s\sqrt{5}$; 1 — positive isolines; $\triangle g'$ 2 — negative isolines $\triangle g$.

Fig. 2. (p. 80) The scheme of relief of pre-tertiary neovolcanic basement. 1 — outcrop of basement, 2 — isolines of basement.


TURČIANSKE TEPLICE
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B. ŠTIAVNICA
N. BAŇA
PLIEŠOVCE
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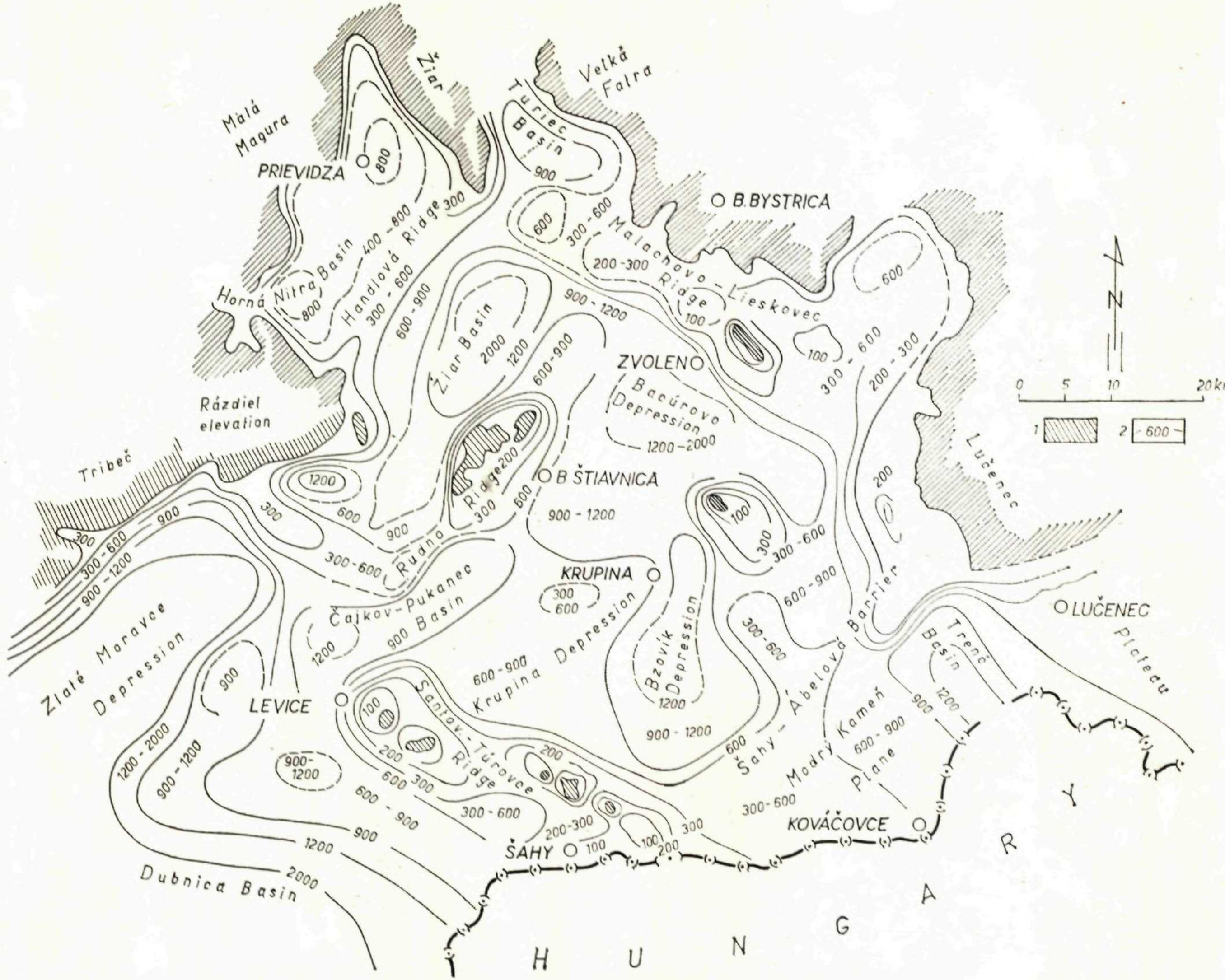
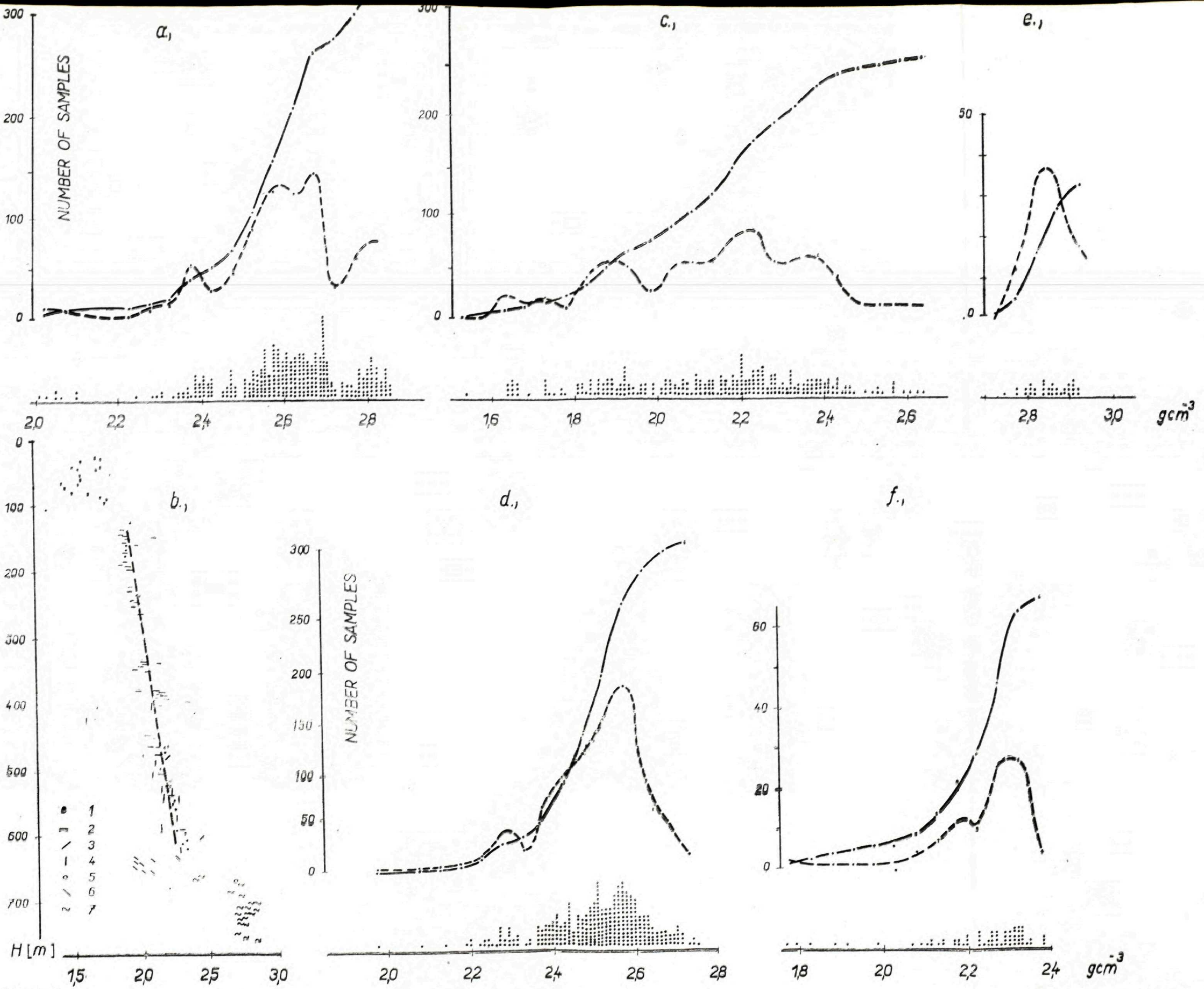


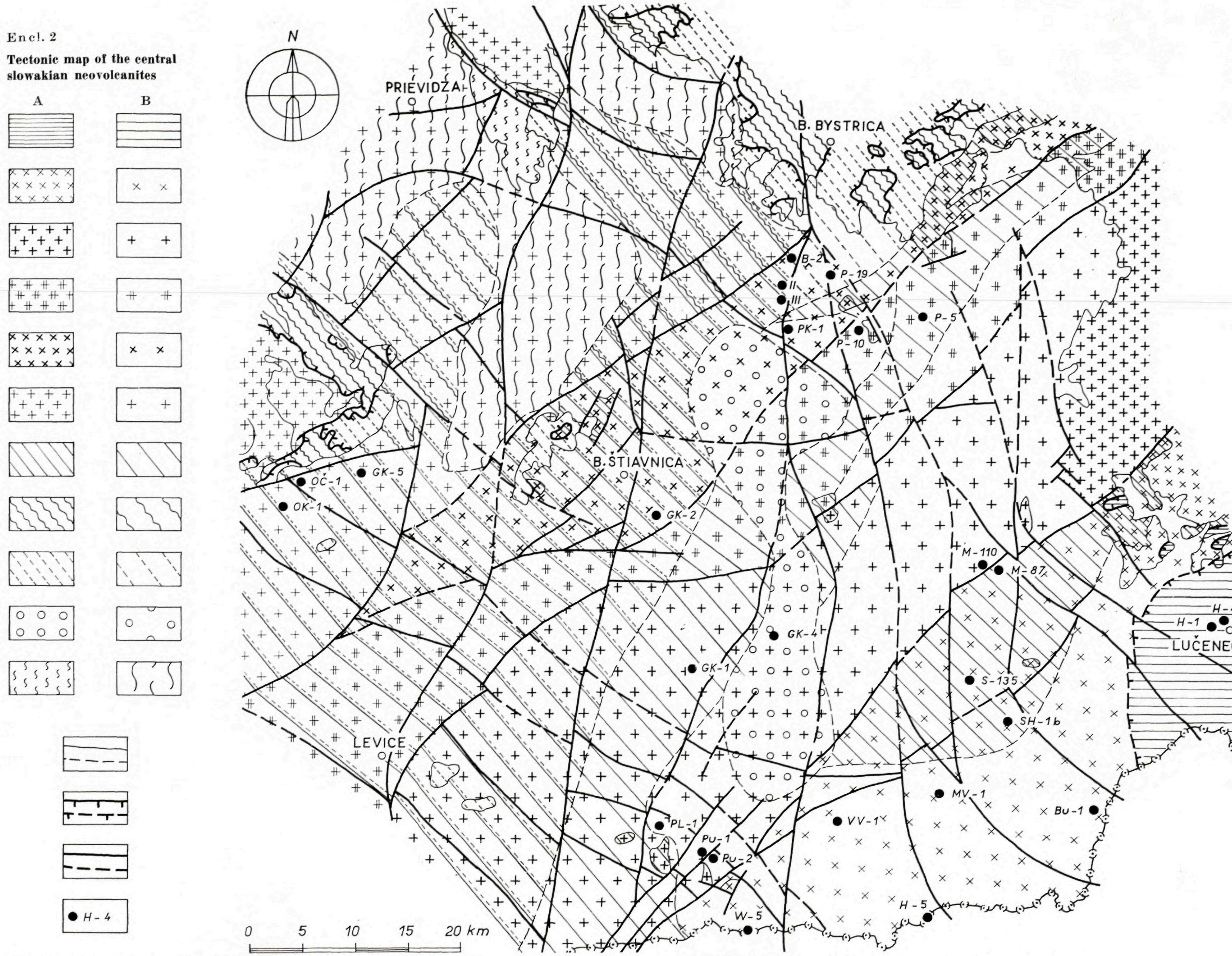
Malá Magura
Žiar
Velká Fatra
Turiec Basin
PRIEVIDZA
B.BYSTRICA
Handlová Ridge
Horná Nitra Basin
Malachovo - Lieskovec Ridge
Žiar Basin
ZVOLEN
Bacúrovo Depression
Rázdiel elevation
Tribeč
Ridge
B ŠTIAVNICA
Lučenec
Rudno
KRUPINA
Čajkov - Pukanec Basin
Krupina Depression
Bzovík Depression
Šahy - Ábelová Barrier
Trenč Basin
LUČENEC
Plateau
Zlaté Moravce Depression
LEVICE
Šantov - Túrovce Ridge
Modrý Kameň Plane
KOVÁČOVCE
ŠAHY
Dubnica Basin
HUNGARY
0 5 10 20 km
1
2 600



Encl. 1.
Volume densities of rocks (according to Orlický) — the integral and veration curves. *a* — rocks of basement, *b* — densities of rock samples from the boring SH_1b (Ipeľská kotlina basin): 1 — clay, 2 — marl, 3 — sandstones, 4 — sands, 5 — quarzite, 6 — agglomerate, 7 — paraschists; *c* — pyroclasts of andesite, *d* — andesites, *e* — basalts, *f* — rhyolites.

Encl. 2

Tectonic map of the central slowakian neovolcanites



Explanations: Tectonic units (A) on the surfaces and in the boreholes (B) in the substratum. 1 — Gemerides, 2 — Kohút Zone, 3 — Kráľova hoľa Zone, 4 — Krakľov Zone, 5 — Ľubietová Zone, 6 — Tatrides, 7 — Envelope series, 8 — Krížna nappe, 9 — Choč nappe, 10 — Cretaceous)?) — Paleogene, 11 — central Carpathian Paleogene; 12 — border-lines founds and supposed, 13 — tectonic lines founds and supposed, 14 — folds founds and supposed, 15 — borcholes.

The structure of the substratum*

All principal both geological and tectonic units of the Inner Carpathians participate to a various extent in the structure of the substratum. In the south-east the Gemerides represented mainly by the Carboniferous extend under the neovolcanics. They are insignificantly distributed reaching only as far as the eastern margin of the Trenčská kotlina (basin). In addition to this continuous occurrence of the Gemerides in the south-western margin of the neovolcanics, the Gemerides may also occur in the more northerly parts in form of small nappe relics. The greater part of the substratum is composed of the Veporides, in which all four tectonic belts can be distinguished. In the south it is the Kohút belt, extending from the Slovakian—Hungarian frontier as far as the Muráň line, which continues from the Tuhár region under the volcanics in the south-western direction towards Šahy. Its course interpreted according with intensive horizontal gradient of gravity source has been demonstrated by borings (M-110 — granitoids and M-87 — Lower Triassic quartzites). An anticlinal structure which is the continuation of the Podrečany anticline can be distinguished in the Kohút belt proper. It consists of crystalline schists (borings: Vu-1, MV-1, VV-1, H-5, W-5). In the northern part of the Kohút belt, the Tuhár syncline composed predominantly of Lower Triassic quartzites (established by borings M-87, S-135, SH-1b and in natural outcrops near Brusník) continues in the substratum. The carbonate members of this syncline occur only at the eastern margin of the neovolcanics (Encl. 2).

The Kráľova hoľa belt extends north of the Muráň line continuing under the volcanics in the same width known from the outcrops at the margin of the volcanics. In the north this belt is limited by the Pohorelá line which has been ascertained in the Pliešovce inlier. It continues south-west into the northern environs of Krupina. Here it turns directly towards the west and after this slight deviation takes up again the south-western course continuing into the environs of Levice. The eastern part of the Kraklová belt as far as the Pliešovce region consists prevalently of crystalline rocks (granitoids and crystalline schists), the western part towards Levice consists of crystalline schists and the Late Palaeozoic-Mesozoic envelope (borings: Pl-1, Pu-1, Pu-2 — crystalline schists, GK-1 — metamorphosed Mesozoic, inliers north of Šahy — Late Palaeozoic envelope). From the spatial extent of these formations it ensues that the anticlinal structure composing the eastern part of the belt wedges out in the environs of Šahy, whereas the synclinal structure constituting the western part, wedges out in the region of the Pliešovce inlier.

* Under the term ,,substratum" we understand the pre-Neogene formations with respect to the close succession between deposition and volcanic activity and/or the deposition of its products.

The Krakľová belt under the neovolcanics forms only a narrow stripe; in some sections its northern boundary is not certain and is only assumed. This belt is composed of crystalline rocks and the Late Palaeozoic and Mesozoic envelope. The crystalline rocks of this belt crop out in the environs of Lieskovec and in the Pliešovce inlier. They have been ascertained by boring (P-10) east of Zvolen. The sedimentary cover occurs at the surface in the Pliešovce inlier and has been established in the P-5 boring.

Likewise the Ľubietová belt forms in the substratum only a narrow belt which probably wedges out north of Levice. Complexes occurring north of Lieskovec and crystalline rocks ascertained by the Pk-1 boring are the established constituents of this belt. The crystalline complex and its metamorphosed Mesozoic envelope of the Hodruša-Vyhne inlier are assigned to this belt. Further data on this belt are so far not available. The northern limitation of this belt is a tectonic line which borders roughly the Rudno ridge.

North of this line the substratum is made up of the tatride elements which, however, do not rise from underneath the neovolcanics and probably mainly form the deeper part of the substratum, because the Krížna and Choč nappes rest on them. The complexes of the tatrides occur at the periphery of the volcanics in the Tríbeč and Žiar mountain ranges.

The Choč nappe crops out in the inliers south-east of Levice and has been established in borings (OK-1, OČ-1, GK-5, and more northerly of Zvolen in qorings B-2, P-19, and II., III.). The occurrence of this nappe in the substratum is, naturally, not continuous. According to our present-day knowledge, its southern limitation can be determined by the Zvolen-Semerovce connecting line.

On the basis of its relation to the Veporides (A. Biely — O. Fusán 1965), the occurrence of the Krížna nappe under the neovolcanics may be assumed north of the Ľubietová belt.

Apart from the principal tectonic units of the Inner Carpathians, in the structure of the substratum participated on a smaller scale also post-tectonic units, namely, the Inner Carpathian Palaeogene, possibly also the Upper Cretaceous in the Gosau development. However, the distribution of these units is small and they are preserved in larger depressions only. At the periphery of the volcanics Eocene is represented on the south-western slopes of Žiar as well as on the eastern slopes of the Kremnica Mts. near Kordíky. Moreover, this formation is also known from the northern margin of the Hodruša-Štiavnica inlier. On the basis of the morphological character of the substratum and of the established occurrences of the Eocene it can be assumed that it also extends under the neovolcanics between Žiarská kotlina (basin) over the Handlová ridge into the Hornonitrianska kotlina (basin). Moreover, it is assumed that this formation occurs in the depression part north of the

Malachovo-Lieskovec ridge, that is, it continues from the Kordíky region under the volcanics of the Kremnica Mts. into the Turčianska kotlina (basin).

The so far not closer stratigraphically determined sequence of varied polymictic conglomerate, breccia, aleurite, and breccia limestone and limestone, which has been ascertained by the GK-4 boring (Gosau? -Eocene) is assigned to the post-tectonic formations. This sequence infilling the Bzovík basin and, according to morphology it is assumed also the Bacúrov basin, has been established in a thickness of 2 000 metres.

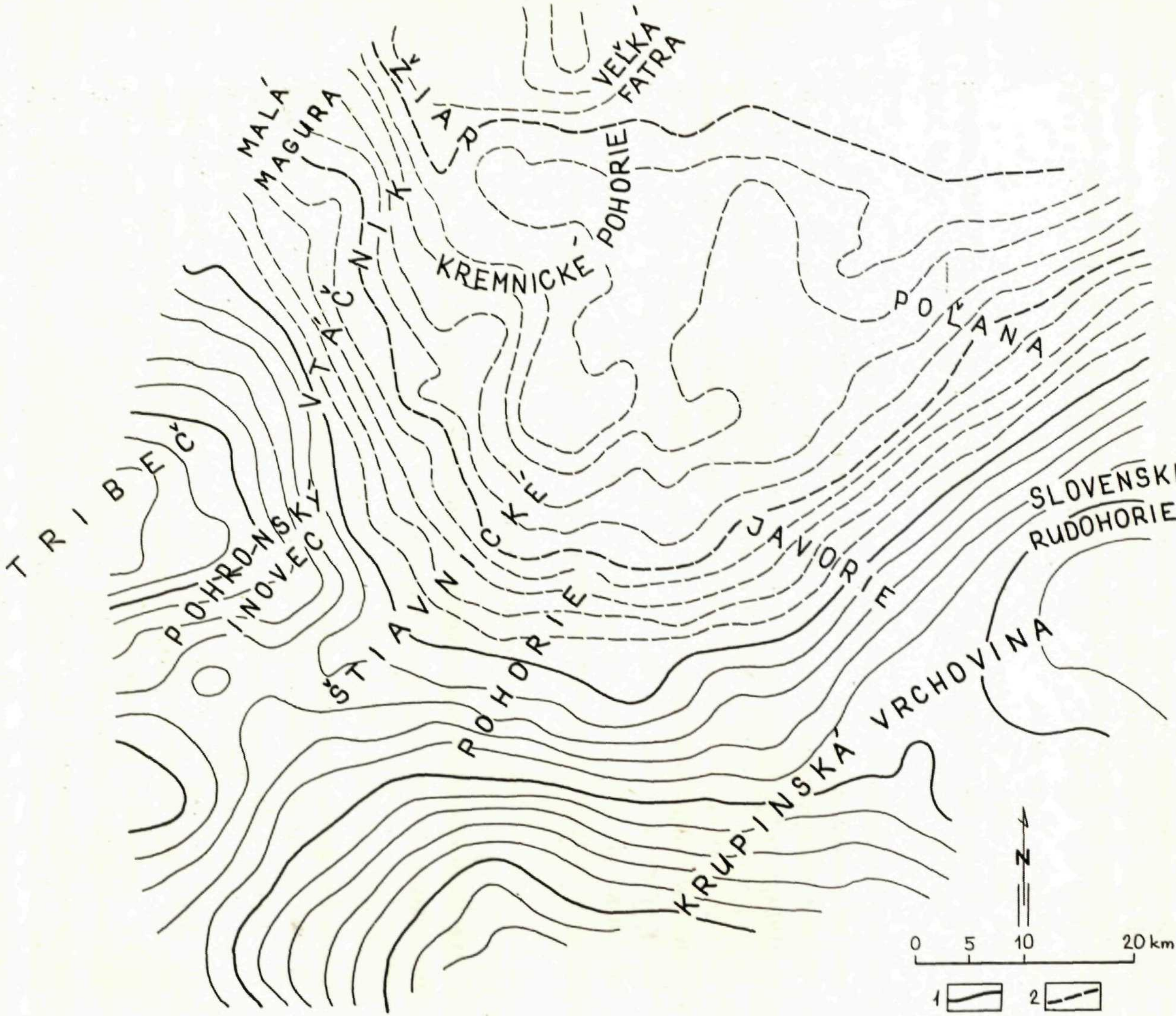
The determined tectonic units are disturbed by a system of conspicuous faults. The principal ones are those which separate the following tectonic units: Lubeník-Margecany, Muráň, Pohorelá, and Čertovica units. Their general strike is NE-SW.

Another system of an obviously older origin, and whose function was several times repeated, is a set of NW-SE striking faults. They are especially apparent in the marginal parts, where they affected the subsequent shape of the substratum of the neovolcanics. The course of the morphologically most conspicuous system of faults is N-S; it determined the origin of basins in the substratum of the neovolcanics and the subsidence of some of its parts. The fault system striking W-E played a lesser part in the subsequent forming of the relief of the substratum. The function of these faults was in the regularity of spatial distribution of the central and linear eruption types and areas of volcanic activity of volcanic clusters.

The deep-seated structure of the substratum

We have described the structure of the substratum as it appears more or less directly under the base of the neovolcanic complex. On the basis of the regional gravimetric map for $S = 2$ km and $S = 4$ km it is possible to give a rough outline also of the structure of the deeper parts of the substratum.

The interpretation is based on a greater spatial extent of the Inner Carpathians. According to the map of the regional gravity field for $S = 2$ km and $S = 4$ km in the Inner Carpathians and according to the intensity of the gravity field, two principal geophysically different belts can be defined which are separated from each other by a narrow transitional zone. The first belt occupying the southern part, is characterized by relatively positive values of Δg isanomalies, the value of which gradually increases towards the south. The second belt occurring in the northern part of the Inner Carpathians is characterized by relatively negative values, the gradient of which falls rapidly towards the north. In the southern belt, in the substratum proper of the neovolcanics, the Levice-Modrý Kameň structure can be delimited, which is from the north confined by the Žemberovce-Horné Strháre line and in the


VEĽKÁ FATRA
MALÁ MAGURA
ŽIAR
VTÁČNIK
KREMNICKÉ POHORIE
POĽANA
TRIBEČ
POHRONSKÝ INOVEC
ŠTIAVNICKÉ POHORIE
JAVORIE
SLOVENSKÉ RUDOHORIE
KRUPINSKÁ VRCHOVINA
N
0 5 10 20 km
1
2

south continues on Hungarian territory. Its northern boundary on the map S = 2 km lies between Horné Strháre and Hontianske Nemce. In the environs of Krupina it conspicuously bends southward. This deformation is called forth by the depression underneath the neovolcanics — the Bzovik basin. Its filling (sediments and volcanic rocks) attains a thickness of 1000 metres. On the map S = 4 km this deformation is no more apparent. On the basis of the gravity intensity in the region of this structure, and of the character of the gravity field it is generally assumed that Conrad's and/or Moho's plane of discontinuity rises gradually, and has farther to the south a more or less horizontal course. On ground of the above-outlined picture as well as from the character of the substratum and bore-hole data we assume that in its upper parts this structure is composed mainly of meso- up to katazonally metamorphosed crystalline rocks.

The gravity anomaly displays characteristic features on the S = 2 km, and S = 4 km maps between the southern and northern belts under the neovolcanics. Spatially it is limited by the Handlová—Banská Štiavnica—Zvolenská Slatina—Nemecká—Handlová connecting line. As compared with the wider environment on the two maps, the gravitational field in this region is relatively in equilibrium. On the S = 2 km map a conspicuous plateau belonging to the Žiar depression is apparent. It no more occurs on the S = 4 km map. At the south-western and south-eastern limitation of this space a great gradient of isanomalies can be observed. According to the character of the gradient and from the interpretation of the geological structure, this limitation is explained by both NE-SW and NW-SE striking tectonic lines initiated at depth. Apart from these principal tectonic lines, on the S = 2 km and S = = 4 km maps faults striking N-E are reflected in a conspicuous bend of isanomalies, which lie in the axis of the north-southern fault zone of the West Carpathians.

If we project the tectonic plan of the region of Central Slovakia and the plan of spatial distribution of the channels of supply of linear eruptions, of the areas of volcanic activity of volcanic clusters and of the central eruption types into transformed gravimetric maps, we can see that the maximum density of volcanic clusters and of central volcanism occurs at the south-eastern limitation of this field and in particular in its southern corner; moreover, at the south-western limitation these principal types are joined to a lesser extent by the linear type, whose dominant is however confined to the meridian fault system, which transgresses this anomalous field in the north-southern direction. The

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Fig. 3. The schematic map of regional component of gravity field acc. to Griffin. S = 4 km. $\Delta g_{reg} = \sum_{1}^{8} \Delta g$; $R = S\sqrt{5}$; 1 — positive isolines $\triangle g$; 2 — negative isolines $\triangle g$.

fact that the channels of supply of the subsequent volcanism are confined both to the narrow margin and to the area proper of this anomalous field, as well as its relatively very even density and its position (it lies between the northern and southern belts of intensive horizontal gradients) led us to interpret this geophysically conspicuous phenomenon as a concealed source area of the subsequent volcanism in this part of the Carpathians. The homogeneity of this field allows to assume the presence of material uniform in chemism, most probably of granitic composition.

The base of the relation of plutonism, from which the orogenic volcanism of the rhyolite — andesite row is inferred, must be sought already in the syntectonic phase of the orogenic stage, that is in the period of the main Alpine folding when the maximum concentration and directioning of kinetic energy took place. Its effects called forth the mobilization of the granitic mass. Extensive migmatization and migmatitic fronts with a granitic nucleus generated in situ. The mass thus mobilized and enriched in light volatile materials was passively squeezed in form of diapirs, and/or intruded, into the mantle. The passivity or activity of the granitic mass were determined by the energy gradient. The geotectonic position of the granitic mass and the direction of the migmatization front were determined by the direction of kinetic energy. This genetic origin explains the homogeneity of the granitic material within a single tectonic unit as well as within the entire orogenic system.

The Gemeride granites are the representatives of these young granites. In other belts of the West Carpathians equivalents of these granites have not been ascertained. On the other hand, phenomena of granitization observed in the Kohút belt are explained by Klinec (1961) in terms of a far-reaching extent of the migmatitic front which was called forth during the main Alpine folding.

From the point of view of further development, the granitic mass which had been mobilized during the syntectonic phase, and which consolidated in situ or as an intrusion in the mantle, finished its role. This was the reason why the synorogenic plutonism separated very strictly from the volcanism of the subsequent, namely the late orogenic phase accompanied by the subsequent volcanism.

The mobilized granitic mass which during the syntectonic phase did not consolidate and was not fixed in the mantle, has been subject to further development, though naturally under different conditions. The frontal migration of magma which advanced in the direction of the exerted stress slows down and the granitization processes are impaired. During the late orogenic stage, the stress of directional orientation ceases and is gradually replaced by hydrostatic pressure. Under these conditions the granitization processes are replaced by differentiation processes, modified locally by assimilation. Provided that as a consolidated element they did not participate in the structure of the orogen, the palingenetic granitic magmas which originated during the syntectonic phase of the orogenic phase, developed still further in the late orogenic stage by differentiation in the granite — granodiorite — quartz diorite — grano-gabbro — quartz-gabbro line.

The geotectonic position of these differentiates is given by genetic conditions which fulfil the conditions of the hinterland of the orogen. In this space the effects of tangential forces ceased first and the palingenetic granitic magmas developed quietly by differentiation under conditions of simple hydrostatic pressure. In some cases the acid differentiates and the pronounced intrusive forms still display slight indications of the granitic-metasomatic effects, whereas the basic differentiates occupy a distinct and exclusively intrusive position.

The trend of the differentiation of deep-seated rocks determined also the principal succession scheme of the effusives of the subsequent volcanism, namely: *rhyolite → andesite → basalt*.

The close petrochemical relationship of the deep-seated rocks to the effusives is also displayed in the same relationship of the acid rocks to the basic ones. The differentiation of palingenetic granitic magmas is characterized in that every following member in the differentiation is, as far as quantity is concerned, half of the preceding one. The basic end-products of differentiation represent only a small fraction of a percent from the total quantity of the parental magma. The same relationship holds good also for the effusives. It frequently happens that the basic effusives are even not developed but are represented by rocks, the chemical composition of which corresponds to andesites, whose intratelluric components however, e. g. plagioclases, are even extremely basic.

The close petrochemical relation between the trend of differentiation and the percentage proportion of the differentiates of deep-seated rocks to the succession of effusives was the reason why the effusives have been considered as their direct surficial equivalent (e. g. Judd 1876, Szabó 1879). The recent detailed research in the region of Hodruša—Vyhne (L. Rozložník 1961) indicates however that even though the effusives of the subsequent volcanism represent the petrochemical equivalent of deep-seated rocks (andesites — diorites) they are the product of a younger geotectonic-magmatic development phase. The conditions under which the geological-petrochemical or tectonic-petrochemical relations of the effusives to deep-seated rocks come to the fore are not sufficiently known. Actually, it has been determined that the texturally-structural development and the bond of mineral components characteristic of the effusive rocks were safely established also in subvolcanic forms (L. Atanasiu, R. Dimitrescu, A. Semeka 1953; A. Mihaliková 1958).

The spatial distribution of the superficial volcanism reflects the geotectonic position of deep-seated differentiates of palingenetic granitic magmas. While their development has optimum conditions in the hinterland of the orogen, the hinterland of the Carpathians was the scene of a widely developed subsequent volcanism. The asymmetrical structure of the Carpathians, determined by the effect of tangential forces which acted in one direction, is distinctly reflected also in the asymmetrical development of the subsequent volcanism, which at the northern (outer) margin occurred only rudimentarily, while in the southern (inner) margin in a great variety and extent.

The function of deep tectonics was not limited to the arrangement and direction of the channels of supply and thus to the forming of the respective type of volcanic apparatus, but played a fundamental role in the type of volcanism. The termination of the differentiation of the palingenetic granitic magmas, which on the surface is apparent by the ascent of basic effusives, led to stabilization — consolidation of the mobile zones; the inherited gravity disequilibrium was relaxated by the epeirogenic movements and in the inversion section by faults which extend deep into the earth's crust. The subsequent volcanism characteristic of the late orogenic stage is succeeded by the final volcanism. Its product is the association of plateau basalts. They produced extensive plateaus expanding from the southern slopes of the Carpathians over the Lučenec—Fiľakovo region as far as northern Hungary.

The character of eruptions of the final volcanism differs entirely from that of the subsequent volcanism. The share of pyroclasts constitutes only a small percentage of the total quantity of the eruptive material. The eruptions were of the quiet effusion type and only sporadically a small cinder cone originated around the crater (relics on Ragač). Moreover, the acid differentiates are characterized by a narrow spectrum of differentiation. In regions built up of basalts of the final volcanism, the proportion of acid differentiates does not even attain one percent of the total mass of these eruptives. The low explosivity as well as the petrochemical character of the subsequent volcanism is explained by V. Bemmelen (1950) by the low content of resurgent gases. Bemmelen maintains that the plateau basalts derived from juvenile alkaline-basalt magma contaminated in the upper parts by the assimilation of dry sialic rock substances from the base of the earth's crust.

The change of the tectonic style, from the alpine-type into the germanotype was accompanied by transition from the synorogenic plutonism through the subsequent volcanism, and terminated in the final volcanism. Apart from this fundamental change in the style of volcanism, the geological development of the neovolcanics of Central Slovakia indicates that also synvo lcanic tectonics played a certain role in the succession and nature of the eruptives.

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NAPPES - SYNCLINORES - FAULTS

In the last years a new setting in of nappe conceptions is observed in the tectonics of ridge mountains. It is evident that their importance in the structure of Alpides is really extraordinary. It would not be surprising if their presence were unequivocally proved in such regions as the South Carpathians of Yugoslavia and the Krajstides, where also synclinores, and anticlinoria have a very important share in the structure (Krajstides, southern part of the Meridional Carpathians). In this connection the question arises if previous period — characterized by a critical attitude of some geologists to a series of nappes, and partly also to the nappe conception — was something other than ,,a sad episode" hindering the development of opinions. Who more thoroughly follows literature can easily ascertain that some active geologists with abundant new substantial information that closer elucidates structural relations, frequently forces a new subdivision of tectonic units, and shows their new relations opposed the nappes (such important information is not easy to obtain in present advanced state of our information about the Alpides). Not lack of survey, even not blindness or overloading by facts were the reason that ,,because of trees they did not see the forest". The reason of the critical attitude is to seek elsewhere. New information enabled to judge from opposite attitude, in nappe conception so far not employed, not evaluated.

It is not an isolated case in science that facts are judged from two opposite poles. This is particularly valid for tectonics as a synthetic branch. Since the thirties already geologists have considered the structure of ridge mountains only through shifting of nappes; the other structure elements (synclinores, faults) were after their opinion of subordinate importance or connected with another, later development stage: the majority of such structure elements however was unknown.

The fundamental position, from which the followers of the anti-nappe conception started after much new information, can be in essentials summarized in two groups:

[1] The presence of *anticlinoria and synclinores*, genetically connected with the distribution and formation of tectonic units; several backward vergencies, refolding of nappes with the autochthon.

[2] *A block structure*, conditioned by the dissection of the geosyncline by many old faults, which played an important rôle in the development of geosynclinal complexes and their transformation into tectonic units.

The synclines or synclinores are supposed by some geologists (Grubič, Bončev) as the basic structural forms in the Yugoslav Carpathians, the Krajstides, in my opinion also in the West Carpathians (Mahel' 1961). I start however from the last named region. There are not only some local depressions, the shifting nappe masses accumulated in, e. g. the Goryczkova Depression in the High Tatra, known since long (Rabowski 1927) or depressions with higher nappes accumulated (Matějka—Andrusov 1931), but structures of regional importance that were manifesting in the time of formation of the structural plan, not only as passive forms-depressions but in long periods also as active structural zones. As such forms in the West Carpathians are particularly distinct the North Gemeride synclinore (Mahel' 1954, 1957) and the Hron synclinore (Máška—Zoubek 1961; Mahel' 1962). Their vergency and truncation by post-Paleogene structures but also a different character, not mediotype as in post-Paleogene structures, tectonically very complicated, point as I think quite unequivocally to the fact that they are Cretaceous structures. The presence of such structures is also confirmed by sedimentological study of paleocurrents (Marschalko). Each of these synclinores possesses its characteristic filling of Mesozoic units, its tectonic style, structural position. An admission of nappe masses in the synclinores means that they represented longitudinal depressions — *synclines* which entrapped the shifting nappe masses. The distribution of volcanic rocks, the manifestation of metamorphosis points to weakened zones of the earth's crust which manifested in the time of folding. They played an important rôle in the folding period not only by „entrapping" of the shifting nappe masses but also as active spaces, in which the nappe masses were passing through structural completing in later phases of the Cretaceous folding period. Structures (partial units, slices, folds) not only of outward vergency but to a considerable extent also of inward vergency formed in them. At many places nappe masses were refolded together with the autochthon and suffered from metamorphosis; they got characters proper to autochthonous series only. And just these characters, revelation (knowing) of which is an important step to forward, generally considered as characters of allochthonous series as well as hiding of the nappe character of masses by later manifestations of folding frequently are the starting attitude of the anti-nappe conception. It is necessary to stress that their margins are accompanied by old faults of upthrust character. In last time some geologists relate the root zones of nappes with these margins (Biely — Fusán 1965).

Emphasizing of the function of nappes in the structure of the Alpides resulted in neglecting the rôle of faults in the course of development of the geosyncline, and also in the folding period. The opinion prevailed that faults were the consequence of a late period (Neogene) of development, substantially younger than nappes. Detailed investigation showed however the important rôle of faults in formation of Mesozoic series and developments, and also of tectonic units. The accustomed idea of incompatibility of genetic relation between faults and nappes, prevailing in thought of geologists, led the defenders of the existence of old faults to an anti-nappe attitude. It was also the case in the West Carpathians which represent a strongly tectonically disrupted segment with many faults, fault-bordered basins, horsts, many mineral springs. That all according to the up to present conceptions is interpreted as a consequence of Neogene fault tectonics. Thorougher study of these faults in relation to the type and distribution of Mesozoic complexes as well as to the course of axes of Cretaceous structures shows that all four basic strikes of faults essentially manifested already in the Mesozoic (Mahel' 1959) (NW, NE, N-S, W-E = planetary systems). Many important young faults are actually rejuvenized only. Among the faults four systems play the main rôle.

The faults striking SE in the western part and NE in the eastern part, E-W in the central part are upthrusts related to Cretaceous structures. Their formation in the time of the Cretaceous folding is indubitable, they frequently diverge with the Neogene faults. The last mentioned show more frequently the tendency of deviation to N-S or to E-W. The fact that these upthrusts separate units of different development points to the presence of the faults already in the time of dissection of the geosyncline into longitudinal zones. There are many other proofs of their presence too-volcanism, bioherms.

More distinct however is the function in the time of development of the geosyncline of the faults transverse to the Cretaceous structures, NW in the western part, NE in the eastern part and N-S in the central part of the Carpathians.

In nature transversal faults of earlier origin are recognizable by the alternation of facies (particularly in the Jurassic) or of thickness of autochthonous series, by the change of the structure plan of autochthonous units, and also very clearly by other structure plan in nappe units. The faults are accompanied in their course by slice structure, sigmoids, and narrow transveral folded structures running parallelly.

The faults have separated the Carpathian geosyncline into several blocks.

These blocks played a rôle in the time of sedimentation and also of formation of the Cretaceous structures. Several faults played a rôle already before the Mesozoic, e. g. the lineament between the Alps and the Carpathians, the Hrádok—Skýcov dislocation, the Štítnik—Hornád dislocation, etc.

The enclosed tectonic outline of the West Carpathians shows the abundance not only of young Neogene faults but also of old faults, and their relation to Cretaceous and post-Paleogene folded structures.

Many young faults are rejuvenized Mesozoic and older faults, the substantially greater rôle of faults striking N-S, and the formation of new fault systems approaching this strike (NNE, NNW) in later periods is however conspicuous.

Conclusion

The presented article is neither meant to be an excuse for the anti-nappe attitude of the author nor a criticism of himself. I am too much aware of the fact that every movement is evaluated by history according to the results brought with it on the basis of a quantity and importance of new facts and new approaches, however not according to the solidity of the belief in ,,old truths". It is nothing peculiar in science that important new information leads its authors to other conceptional, frequently also extreme positions and causes a short-time desorientation.

The evidence of older faults and synclinores which originated in the same folding period with the units, and thus also with nappe units, is, as I think a contribution, and not only a contribution but also a starting-point for the next stage of research. It will be necessary, however, to bring together the dynamics of two categories apparently excluding in hitherto existing ideas-nappes on the one hand, synclinores and folds of old origin on the other hand. I think this is the way to the clarification of processes reaching greater depth. Of course we consider the folding period as a process consisting of several phases with change of intesity and quality in time and space.

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IDENTIFIZIERUNG DER BRUCHTEKTONIK IM ÖSTLICHEN ZENTRALKARPATISCHEN PALEOGEN (das Gebirge Levočské pohorie u. Šarišské hory)

Das Studium der Bruchtektonik im zentralkarpatischen Flysch hat neben der teoretischen auch große praktische Bedeutung. In den meisten Fällen representieren nämlich die Brüche Aufstiegswege der Mineralwässer, die aus den mesozoischen durch eine Flyschdecke bedeckten Gesteinen stammen. Die Feststellung der Vertikalamplitude der gesunkenen Strukturen ist wiederum bei der Aufsuchung der Mn-Erzgesteine von Wichtigkeit. Die Bestimmung, bzw. Begrenzung einzelner tektonischer Strukturen wird auch bei der Situierung der für die Erdölprospektion erforderlichen Bohrungsarbeiten eine entscheidende Rolle spielen.

Das zentralkarpatische Paleogen im studierten Raume zwischen den Städten Poprad und Prešov in der Ostslowakei ist überwiegend durch Flyschablagerungen verschiedener granulometrischer und fazieller Zusammensetzung des Mitteleozän-Unteroligozän-Alters gebaut (Marschalko 1966; Chmelík 1967; Marchalko—Gross—Kalaš 1966). Sie haben sich in einem Becken abgesetzt, dessen Untergrund durch den Kreide-Orogen formiert wurde und sich die Grundmerkmale der alpinen tektonischen Strukturen erhalten hatte. Bei der näheren Untersuchung des tektonischen Bildes des studierten Raumes ist die Tatsache, daß sich die Flyschfazies an einer *bedeutenden Biegung der karpatischen Strukturen* — sozusagen auf einem gebrochenen Bogen — gebildet haben, von entscheidender Bedeutung. Infolge dessen wird die Klärung der Genese der Brüche und vor allem die gegenseitigen zeitlichen (altersmäßigen) Beziehungen der Brüche der sog. karpatischen Richtung sehr erschwert. Deshalb zeigte es sich bei der Untersuchung dieser Fragen zweckmässig, zuerst die Lage der bathymetrischen und Strukturachse des Flyschbassins zu klären und die Beziehungen der Flyschfazies zu alten tektonischen Strukturen des Fundaments zu prüfen. Bei diesem Studium ist es unserer Meinung nach — besonders in den Ablagerungsbecken mit klastischer Füllung — notwendig, die *Transportrichtung* zu analysieren, die uns eine Grundlage für die Rekonstruierung des Beckenbaues geben kann.

Die nähere Bestimmung der Brüche in den Flyschgebilden der zentralen Westkarpaten ist vor allem durch folgende Tatsachen erschwert:

(1) *Mangel an faunistisch belegten Korrelationshorizonten* und infolge dessen nicht immer präzise stratigraphische Beurteilung der Lage;

(2) *eintöniges Aussehen der Flyschfazies*, besonders in größerer Entfernung von Dispersionszentren und Quellengebieten des klastischen Materials;

(3) *nicht befriediegende Kenntnisse über den Untergrund* des zentralkarpatischen Flysches;

(4) *nicht immer bekannte Gesamtmächtigkeit der Flyschablagerungen* und

(5) ziemlich *schlechter Erhaltungszustand der tektonischen Bruchflächen* im Terrain und oft verwischte Beziehung zu den morphologischen und topographischen Formen.

Als positive Elemente beim Studium der Bruchtektonik kann man folgende Tatsachen angeben:

(1) *Scharfe lithofazielle Grenzen* als Folge der verschieden situierten, aber zeitlich beständigen dispersen Zentren des klastischen Materials;

(2) ziemlich *leichte Messung und Auswertung der Strukturelemente* in den Flyschablagerungen;

(3) *linearer Aufstieg der Mineralwässer* als Indikatoren der tektonischen Linien;

(4) *lineare Verteilung der Flyschablagerungen* in Beziehung zum System der Paläoströme und zu alten tektonischen Strukturen.

(5) die Möglichkeit die *synsedimentäre Tektonik auf grund der gravitativen Bewegungen der Ablagerungen zu studieren.*

Einige Grundkriterien für die präzise Feststellung der Brüche in den Flyschgebilden der zentralen Westkarpaten

(1) *Verschiebung der Korrelationsflächen* (z. B. die scharfen Faziesgrenzen der stratigraphischen Leithorizonte) dient als einer der Ausgangspunkte bei der Identifizierung der Bruchflächen. Deshalb kann man die bei der Kartierung festgestellten faziellen Grenzen, Vertikaländerungen der Lithofazies, Rhytmonogramme und Histogramme der Mächtigkeit der Schichten, Berechnung und Bestimmung des klastischen Verhältnisses *Sandstein-Tonstein, Sandstein-Konglomerat* als Grundlage bei der Suche, Indizierung und Konstruierung der Bruchlinien in den Flyschgebilden betrachten. Die Diagonalbruchstörungen der SE-NW und SEE-NWW Richtung im Gebirge Šarišská hornatina wurden eben mit Hilfe solcher Korrelation festgestellt und in einer Länge von 18 Km verfolgt. Mit Hilfe der stratigraphischen Leithorizonte

konnte man die Vertikalsinkung auf 20—150 m, vereinzelt bis auf 150—220 m schätzen. Von den festgestellten und lokalisierten Brüchen haben wir auf Grund der Verschiebung der Korrelationsflächen und stratigraphischen Leithorizonte bis 60 %, und den Rest mit Hilfe der anderen erwähnten Methoden festgestellt.

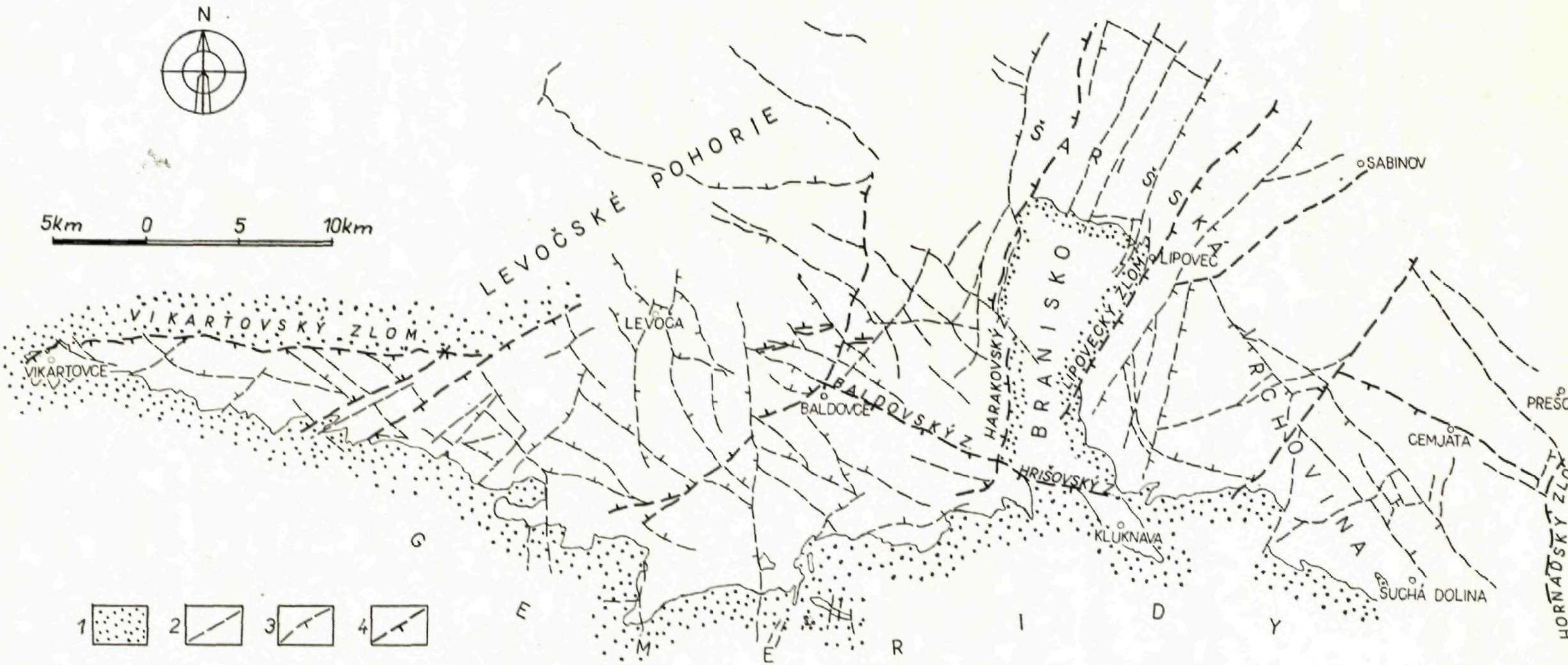
(2) Ziemlich präzis konnte man die Bruchlinien, bzw. Bruchzonen *am Kontakt der Flyschablagerungen mit älteren*, event. *jüngeren Gebilden (als Paleogen)* feststellen. Es handelt sich um die S-N und SSW-NNE gerichteten Brüche, die den Kristallinkern von Branisko vom Westen und vom Osten her begrenzen. Entlang dieser Brüche berührt sich das Kristallin (ohne mesozoische Hülle) mit höheren Flyschgliedern. Falls die Gesamtmächtigkeit des Paleogen nahe Branisko mehr als 700 m beträgt — wie man vermutet — dürfte die Sinkung etwa 1500 m hoch sein, was etwa jener der subtatrischen Brüche tenspricht. Ähnlichen Charakter hat auch der N-S verlaufende Hornád-Bruch, der die Flyschfazies in östlicher Richtung durchschneidet. Morphologisch besonders ausdrucksvoll ist dieser Bruch im Abschnitte Drienovská Nová Ves—Kendice, wo er seinen Verlauf nach Nordwesten ändert und dichotomiert. Zu den ausgeprägtesten Brüchen diesen Charakters zählt der O-W gerichtete Vikartovce-Bruch, bei welchem die Sinkung der Paleogenschichten gegenüber dem Perm-Werfen in östlicher Richtung von 100—220 m (in Westen) bis auf 450—600 m (in Osten) steigt.

(3) Ein anderes, indirektes Kriterium bei der Identifizierung der Brüche ist *die Verfolgung des Verlaufes und des Einflusses der vorpaleogenen Strukturlinien des Fundaments* (in unserem Falle der großen Aufschiebungsflächen und tektonischen Tiefenstrukturen) auf die Entwicklung der Brüche in hangenden paleogenen Schichtfolgen. Prägnante, SW-NO gerichtete Brüche solcher Art wurden bei der Talmündung von Veľká Biela Voda in Richtung Spišský Štvrtok festgestellt, welche die Fortsetzung der Muráň-Linie vorstellen. Das Bruchsystem im Bachtal von Ždiar und Tepličnô entwickelte sich parallel mit dem Verlauf der alten Linien am Kontakt des mesozoischen Vernár-Streifens mit der Melaphyrenserie der Choč-Decke. Ähnliche Beispiele finden wir am Kontakt der Gemeriden mit Tatroveporiden der Čierna hora. Den sog. Hrišovce-Bruch kann man in SO-NW Richtung bis zu Baldovce verfolgen, wo er verschwindet.

(4) Im Raume des zentralkarpatischen Flysches ist *der Aufstieg der Mineralwasserquellen* aus dem liegenden Mesozoikum bekannt. Solche „*Linienwasserquellen*", besonders des Calcium-Bikarbonat-Types, indizieren oft Brüche und umfangreiche Störungszonen, welche tief in den liegenden mesozoischen und paläozoischen Untergrund reichen. Zu solchen Störungen zählt vor allem das SO-NW gerichtete Bruchsystem von Cemjata und Žipov im Gebirge Šarišská hornatina, weiter die Brüche von Baldovce und die erwähnten, N-S verlaufenden Brüche von Branisko.

Abb. 1.

Tektonische Karte des zentralkarpatischen Flysches (das Gebirge Levočské pohorie, Šarišské hory, Hornád-Kessel).



1 — praepaleogene Gebilde im Ganzen; 2 — festgestellte, aber nicht präzis lokalisierte, bzw. bedeckte Bruchflächen (von geringerer Bedeutung); 3 — festgestellte (bedeckte) Brüche mit bekanntem Fallen; 4 — Brüche mit bekanntem Fallen, meistens von grösserer Bedeutung.

(5) Die Bruchflächen werden oft auch durch *plötzliche Änderungen im Fallen der Schichten*, oder durch *plötzliche Unterbrechung der subhorizontalen Lagerung* (verursacht durch die plötzliche steile Aufstellung der Gesteinsblöcke) angedeutet. Solchen Fall stellt z. B. eine cca 20 km lange O-W gerichtete Dislokation, entlang welcher der mittlere Teil des Gebirges Levočské pohorie in SSO Richtung abgesunken war. Die Dislokation bildet eine 50—300 m mächtige Zone, in welcher die Schichten unter einem Winkel von 80—90° steil aufgestellt sind. Sie begrenzt die sog. Klčov-Elevation entlang ihres Nordrandes. Der Rand der Elevation wurde in der Bohrung Klčov 1 nur etwa 137 m tief unter dem Flysch erreicht (Gross 1967). Diese Elevation, die auch durch gravimetrische Messungen bestätigt wurde, setzt sich in östlicher Richtung fort und ist durch die erwähnte Dislokation bis zu Spišské Podhradie begrenzt.

Die Indizierung der Brüche und Bruchstrukturen durch Studium der Veränderungen in der subhorizontalen Lagerung der Schichten wäre nicht ganz präzis, wenn man dabei noch *die ergänzende Korrelationsmethode* nicht benützen würde. So z. B. bei den regionalen Kartierungsarbeiten (im Maßstab 1:200 000) hat man in der Umgebung von Branisko eine Sigmoidalbiegung der Flysch-Schichten vermutet, was zur Vorstellung verleiten könnte, daß der Kristallinkern von Branisko eigentlich eine Megaantiklinale sei. Durch eingehende geologisch-stratigraphische Untersuchungen wurde keine solche Sigmoidalbiegung der Schichten festgestellt; die NNE-S gerichteten Bruchstörungen hatten da nämlich den Zerfall des Flysches in eine Reihe von nicht gleichmäßig gesunkenen Schollen zur Folge. Die Emporhebung des Kernes fand zur Zeit der nachpaleogenen Bewegungen stat. Aus diesem Grunde wäre es falsch anzunehmen, daß da während der Ausbildung des Flyschbeckens eine Elevation des Untergrundes als Keim der vermuteten Megaantiklinale existierte. Die geologischen Kartierungsarbeiten auch in anderen Regionen des zentralkarpatischen Flysches haben gezeigt, daß die Kerngebirge des karpatischen Verlaufs (die sog. Megaantiklinalen) — obwohl sie am Rande durch Überreste der ältesten Basalschichten gesäumt sind und so eine Vorstellung der ursprünglichen, bereits vor oder während der Sedimentation existierenden Strukturen erwecken — eigentlich junge tektonische Gebilde sind, die sich erst nach der Ausfüllung der ursprünglich einheitlichen Sedimentationsbecken gebildet haben. Die Identifizierung der Tiefenbrüche, welche den abgesunkenen Flysch (in den Becken) von den Kernen der sog. Megaantiklinalen begrenzen, bleibt somit die Aufgabe der künftigen eingehenden Kartierungsarbeiten.

Bruchflächen, Ausfüllung und Charakter der Brüche

Direkt im Terrain finden wir nur selten Beweise über das Fallen der Bruchflächen. Eine vor längerer Zeit durchgeführte Untersuchung in den Bergwerken von Švábovce hat gezeigt, daß da das Fallen der Bruchflächen zwischen 48 und 78° variiert, wobei die steiler fallenden Querbrüche nördlich der Vikartovce-Elevation beobachtet wurden. Die Bruchflächen waren nicht gekrümmt. Die Ausfüllung der Bruchstörungen bilden meistens Blöcke und Bruchstücke der zwischenschichtigen Flyschgesteine. Die nicht ausgeheilten Brüche hat man im Konglomerat- und Mikrokonglomeratflysch im Gebirge Šarišská hornatina beobachtet. Offene, bis zu 40 cm breite Risse sind in den Bergstollen von Švábovce zu beobachten; sie dienten als freie Aufstiegswege der Mineralwässer. Es wird angenommen, daß die ,,Linienwasserquellen" eben durch solche, nicht vollkommen ausgeheilte Brüche zirkulierten. Wie die abgestorbenen Travertinkörper und die Migrierung der Wasserquellen zeigen, wurden solche Wasserkanäle infolge der wiederholten Bewegungen auf solchen Flächen geschlossen und das Wasser suchte sich neue Wege durch die Flyschablagerungen.

Entlang der meisten identifizierten Bruchflächen kam es zur Schollensinkung, was auch durch stratigraphische Korrelation bestätigt werden konnte. Aus der Untersuchung *der Neigung der Schollen*, die durch Diagonalbrüche abgeschnitten sind, geht es hervor, dass es sich um *stufen- und kaskadenartige Sinkungen* handelte, welche den asymmetrischen Bau zur Folge hatten. Eine Serie solcher stufenartiger Sinkungen ist im Raume des Hornád-Kessels zwischen Spišská Nová Ves und Levoča, wie auch nördlich Branisko bekannt. Die Dislokation, welche in Norden die Klčov-Elevation begrenzt, weist ein sehr steiles Fallen (80—90°) auf; möglicherweise handelt es sich da um eine Schollenüberkippung. Die Bildung einer typischen horstförmigen Struktur konnte da nicht bestätigt werden.

Die zeitlichen Beziehungen zwischen den Brüchen der sog. karpatischen Richtung

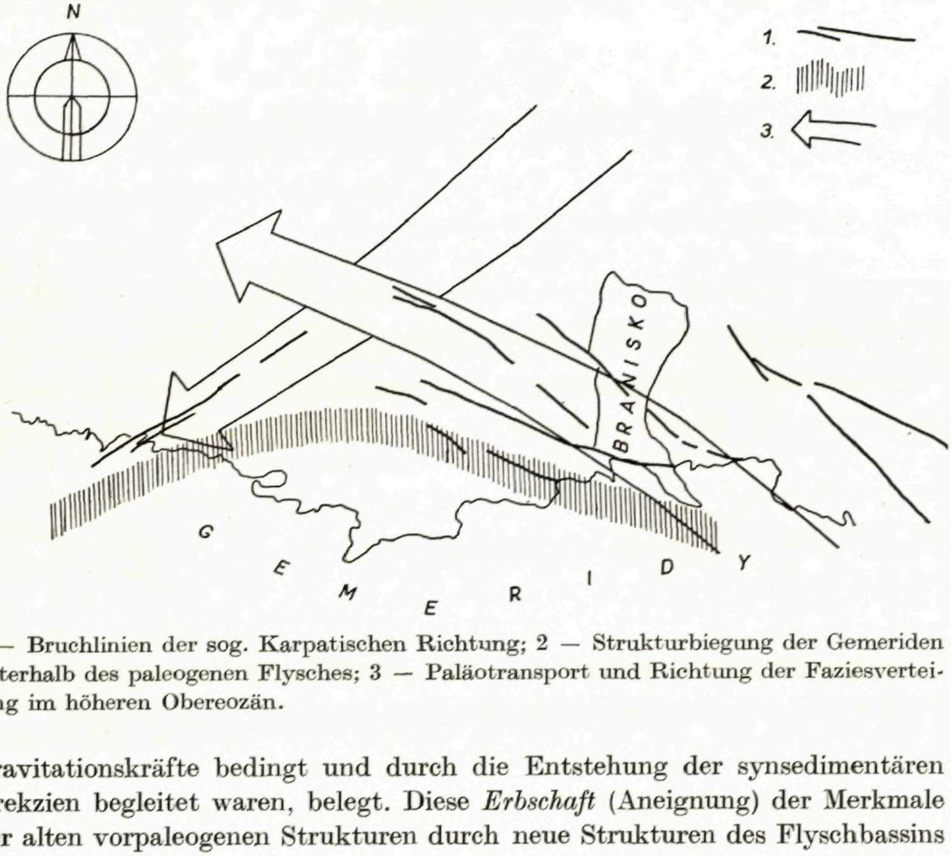
Sowohl die Genese, wie auch die zeitlichen Beziehungen zwischen einzelnen Brüchen kann man ohne Kenntnis der ursprünglichen Begrenzung und Form des Sedimentationsbeckens, seiner Struktur und der bathymetrischen Achse, bzw. der synsedimentären Bewegungen nicht zufriedend klären. Das Studium der Paläoströme in den Flyschfazien hat bereits große Fortschritte gemacht und ermöglicht uns die Richtung der Beckenfüllung zu erkennen. Die Systeme der Paläoströme weisen oft eine so große Stabilität in Zeit und Raum auf, daß sie eigentlich (zusammen mit dem entsprechendem Fallen des Beckenrandes) auch als beständige Faktoren der tektonischen Elemente betrachtet werden können, durch welche die Erosion, Transport und Ablagerung im Laufe der langen Zeitabschnitte geregelt wurde. Aus unserem Studium der Paläoströme folgt, daß die heutige Form des Flyschbeckens eigentlich nur einen Teil des abgestorbenen Körpers des Bassins vorstellt, der durch das klastische Material aus zwei verschiedenen Quellengebieten ausgefüllt wurde (Marschalko—Radomski 1960). Diese Quellengebiete waren auf beiden Seiten des Bassins voneinander unabhängig tätig. Ein solches Quellengebiet befand sich in der Zeit des Mittel- und Obereozäns im Raume des Kontaktes der Klippenzone mit dem zentralkarpatischen Flysch. Das klastische Material bewegte sich von NO in SW Richtung in den Raum des Gebirges Levočské pohorie und weiter bis ins Gebirge Stratenská hornatina. Falls die Paläoströme, durch welche sich an der Basis der gradiert geschichteten Fazies die Paläostromspuren, in höheren Lagen horizontale Lamination und Schrägschichtung gebildet haben, eine schlammige Masse bildeten, wurden sie auch durch die Gravitationskräfte kontrolliert und flossen als schwere dichte Suspension entlang der submarinen Abhänge in tiefer gelegene Räume hinab. Dieser Prozeß einverleibt in sich auch die Existenz einer bathymetrischen Beckenachse, die sich in dieser Richtung der Massenbewegung ausgebildet hatte.

Im mittleren Obereozän bis Unteroligozän entstand eine neue Quellenzone des Materials und zwar im SO Teil des Flyschbassins, welche das Material von SO her (in NW Richtung) für das Gebirge Šarišská hornatina und Levočské pohorie geliefert hatte. Aus dem Studium der Paläoströme (Marchalko 1961) folgt, daß auch in diesem Falle die Ströme durch Gravitationskräfte kontrolliert wurden und flossen entlang des submarinen Troges (der sich in NW Richtung immer mehr vertiefte) in den Raum des heutigen Gebirges Levočské pohorie. Infolge dessen entwickelte sich der hiesige Flysch an der gekreuzten bathymetrischen Bassinsachse und auch der Strukturachse (d. h. Achse der Strukturdepression mit maximaler Mächtigkeit des akkumulierten Materials).

Die gegenseitige Abhängigkeit zwischen der Distribution der Flyschfazies, der linearen Verlängerung der Schichtkörper und dem regionalen Paläostromsystem wurde durch die Untersuchung der orientierten Sedimenttexturelemente bestätigt. Die Verteilung einzelner Fazies im Bassin verlief in zwei Hauptrichtungen, und zwar von NO in SW und SWW Richtung, und von SO nach NW (bzw. NWW) und ist im guten Einklang mit dem Verlauf der vorpaleogenen Strukturelemente (siehe Abb. 2.). Daraus folgt, daß die Distribution der Flyschfazies zwangsläufig unter der aktiven Mitwirkung dieser älteren Strukturen verlief. Deshalb kann man annehmen, daß sich die sog. karpatisch (NO—SW) und SO—NW gerichteten Brüche wenigstens teilweise während der Subsidenz der Flyschfazies gebildet haben. Diese synsedimentären Bewegungen sind durch strahlenförmig geregelte klastische Gänge (die in Richtung alter tektonischer Strukturen im Untergrund des Paleogens orientiert waren; Marschalko 1966), oder auch durch Bewegungen der Ablagerungen, die durch

Abb. 2.

Das Paläostromsystem und Verteilung der Fazies im höheren Obereozän verfolgt die praepaleogenen Strukturelemente der Gemeriden. In diese Richtungen verläuft die bathymetrische und Strukturachse des Flyschbassins. Den Verlauf dieser sog. karpatischen Linien verfolgen auch grössere Störungen, die manchen Anzeichen zufolge während der Formierung des Flyschbassins aktiv waren und auch synsedimentären Ursprungs sein können.



1 — Bruchlinien der sog. Karpatischen Richtung; 2 — Strukturbiegung der Gemeriden unterhalb des paleogenen Flysches; 3 — Paläotransport und Richtung der Faziesverteilung im höheren Obereozän.

Gravitationskräfte bedingt und durch die Entstehung der synsedimentären Brekzien begleitet waren, belegt. Diese *Erbschaft* (Aneignung) der Merkmale der alten vorpaleogenen Strukturen durch neue Strukturen des Flyschbassins ist ein wichtiges Element für die Beurteilung der Genese der Brüche und ihrer zeitlichen (altersmäßigen) Beziehungen. Aus diesen Erwägungen folgt weiter, daß die N-S, NNO-SSW und NNW-SSO (also senkrecht auf die karpatische Richtung) orientierten Brüche (d. h. unsere Paläostromrichtung) jünger sind und *entstanden erst nach der Ausfüllung des Flyschbeckens*; sie verwischten daher das ursprüngliche Bild und Form dieses Beckens.

Obwohl das Paläostromsystem und Verteilung der Flyschfazies bis in den Raum des Gebirges Stratenská hornatina vorgegriffen haben, wurden die postpaleogenen Bewegungen längs dieses Systems (SO—WNW) so stark, daß in dieser Richtung keine Flyschablagerungen erhalten blieben. Eine ganz andere

Situation war östlich des Hornád-Bruches. Die Zufuhr des terrigenen Materials entlang der Beckenachse verlief von SO in NW Richtung von einem entfernten Quellengebiet. Falls der Bruch mittelmiozänen Alters ist, dann sollte man die Fortsetzung des Flysches in südöstlicher Richtung in unveränderter Mächtigkeit unter der vulkanischen terrigenen neogenen Formation erwarten.

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LITHOLOGICAL INVESTIGATION OF CONGLOMERATES OF MAGURA FLYSCH IN EAST SLOVAKIA

Abstract. In the present work the results of the lithologic and sedimentologic investigation of conglomerates in the southernmost (Čerhov) unit of the Magura Flysch in East Slovakia are given, as well as geological and lithological position of conglomerates within the unit under study, sedimentologic evaluation of structural and textural properties of conglomerates, and the evaluation of the conditions of the genesis of the former within turbidite formations.

Stratigraphic and Lithologic Characteristics of Partial Units of Magura Flysch

The Magura Flysch consists of the Paleogene and Upper-Cretaceous beds. Basing upon the lithofacial and facial-tectonic differences several partial units might have been distinguished in Magura Flysch. Matějka–Roth (1949), Roth (1960) divided the Magura Flysch in West Slovakia and in Moravia into the Rača, Bystrica and southernmost Belokarpatská partial units. Andrusov (1926) determined the southernmost partial Magura Unit in the Paleogene of the Oravská Magura Mts., called it the Oravsko-Magurská Unit. This partial unit is supposed to continue in the NE direction to Poland (Gorce Mts.).

In East Slovakia, division of the Magura Flysch Zone is complicated by the presence of Menilite and Malcov beds, most frequently occuring in morphologic depressions. The Malcov beds remind of the claystone strata of the Central Carpathian Paleogene or the Krosno beds of the Dukla Unit, by their lithologic composition. The basic division of the Magura Flysch Zone from the Western Flysch Carpathians from Slovakia and Moravia has been roughly applied to East Slovakia too. Two northern partial Magura Units (Encl. 1.) in East Slovakia are indicated as the Rača–Bystrica Units. In East Slovakia Kochanovce Unit (Leško 1960) or the Čerhov Zone (Matějka 1961) corresponds to the Belokarpatská Unit. With respect to the fact that the southernmost partial unit is most frequent in morphologically important Čerhov Mts., in agreement with the Andrusov's (1965) suggestion, it will be further indicated as the Čerhov Unit.

The Rača and Bystrica units in East Slovakia are lithologically and stratigraphic-tectonically so well differenciated, that there are only few unsolved problems. The substratum of both units is formed of the *Beloveža beds* passing up to the Upper Cretaceous in some places (the southern margin of the Smilno tectonic window; Nemčok—Koráb 1963), south to the Cigla village in the Výrava r. valley, to SW of the Nižná Jablonka village. The Beloveža beds are characterized by rhythmical alternation of clastics with varicoloured claystones, the latter predominating over sandstones (2:1 to 8:1). In external structures on the lower side of bed surfaces bioglyphs predominate over mechanoglyphs. On sandstone beds there are frequent traces of worm crawling and traces of the Bullia type. Among clastics fine-grained sandstones (64 %) prevail over siltstones (36 %). Concerning petrography, subgraywackes and quartzose sandstones occur in Beloveža beds in 1:1 ratio (Ďurkovič 1966). Claystones of the Beloveža beds are often varicoloured, alternating in stripes and streakes, while of the clastic sediments of the Beloveža beds horizontal lamination with frequent transition into convolute lamination is typical. Claystones are mostly gray, calcareous and noncalcareous, blue-gray, brown-gray, greenish, blue-green, red and violet-red. According to microfauna (Samuel 1960) the Beloveža beds in East Slovakia belong to the Paleocene-Lower Eocene.

Above the Beloveža beds about 1500 m thick series of the *Zlín beds* is resting. In differentiation of the Zlín beds of the Bystrica and Rača partial units, the presence of glauconitic micaceous sandstones, claystones, and marls of the Lacko type as well as the mutual ratio of claystones to sandstones are decisive. While glauconitic sandstones are abundant in Zlín beds of the Rača Unit, in the Bystrica Unit they are less frequent, indicated as the *Lacko beds* by Leško—Samuel (1968). Predominating clastic sediments are represented here by micaceous sandstones. An important differentiating characteristics of the overlying Beloveža beds is the presence of claystones and marls of Lacko type (*Lacko mergel,* Uhlig 1888) and gray-green claystones. The latter remind of the *Krosno beds,* or claystones of the Central Carpathian Paleogene. While in the Bystrica Unit there are mostly hard claystones and the marls of the Lacko type, in the Rača partial unit the intercalations between sandstones are formed also by soft claystones predominating in some East Slovakian areas.

The Zlín beds of the Bystrica unit are very poor in micro and macrofauna. According to Samuel (1960), the lower part of the Zlín beds of the Bystrica unit passes into the Lower Eocene. This opinion is also supported by the fauna of larger foraminifers (Vaňová in Nemčok 1961) *Nummulites burdigalensis* De la Harpe, *N. partschi tauricus* (De la Harpe), *Assilina aff. douvillei* Abrard & Favre. Hanzlíková (1960) quotes the Middle Eocene and Upper Eocene microfauna from the Zlín beds. The Lower Eocene age is also testified to by the presence of the Upper Eocene to Oligocene development of *Menilite and Maľcov beds* to the SE of Bardejov. This Paleogene series beginning with variegated claystones and ending with the Maľcov beds represents the normal overlier of the lover constituents of the southern partial Magura Units and of the Paleogene of the Klippen mantle. Leško (1959) gave the stratigraphic-lithological characteristics of the Čerhov Unit in the area of Kochanovce village, where over the Beloveža beds there is predominantly sandstone sequence equivalent to the Zlín beds of the Bystrica Unit. Over this extensive sandstone — conglomeratic complex near Kružlov, Maľcov and Raslavice villages there is a thick flysch sequence of Menilite and Maľcov beds with underlying variegated claystones with Cyclammina amplectens and fine-grained nummulitic conglomerates.

Čerhov Unit is formed of Paleocene to Upper Eocene beds. The lower part of the Paleogene of the Čerhov unit is represented by the Beloveža beds (Paleocene-Lower Eocene; Samuel 1959). It is actually fine-rhythmical flysch (Fig. 1) with green and red claystones alternating with thinbedded sandstones. In the Čerhov Mts. there is predominantly psammitic-pelitic Paleocene to Lower Eocene flysch development of the Beloveža beds with numerous bioglyphes. Claystone strata are represented by green to green-gray claystones, red claystones are only sporadically present. In the Beloveža beds red claystones are more abundant especially near Strihovce village in East Slovakia. The flysch development of the Beloveža beds in the Čerhov unit gradually passes into a flysch sequence with predominance of sandstones, in the overlier. In the Čerhov Mts. and to the east of the river Topľa up to the Czechoslovak-Soviet frontier this complex of calcareous graywackes and arkosic sandstone microconglomeratic strata shows the thickness of even 2500 m (Stráník 1965). Leško (1961) distinguished sandstone conglomeratic strata in the overlier of the Beloveža beds and called them Strihovce beds. In many places the sandstone-conglomeratic complex of the Čerhov unit remind of the wild flysch of the Central Carpathian Paleogene with monolitic sandstone beds and numerous slump bodies. These are most frequent in the area of Majdan, Proč, Matiaška, Domaša and Údavské villages (Encl. 1). They do not occur in a certain stable horizon in the overlier of the Beloveža beds of the Čerhov unit, position of their outcropping being irregular in the whole sandstone-conglomeratic complex (Fig. 1). The pebbles of carbonates in sandy matrix of the slump bodies are often of dm size, 40—50 cm pieces are less frequent. Sandstone beds forming the overlier and substratum of the slump bodies are

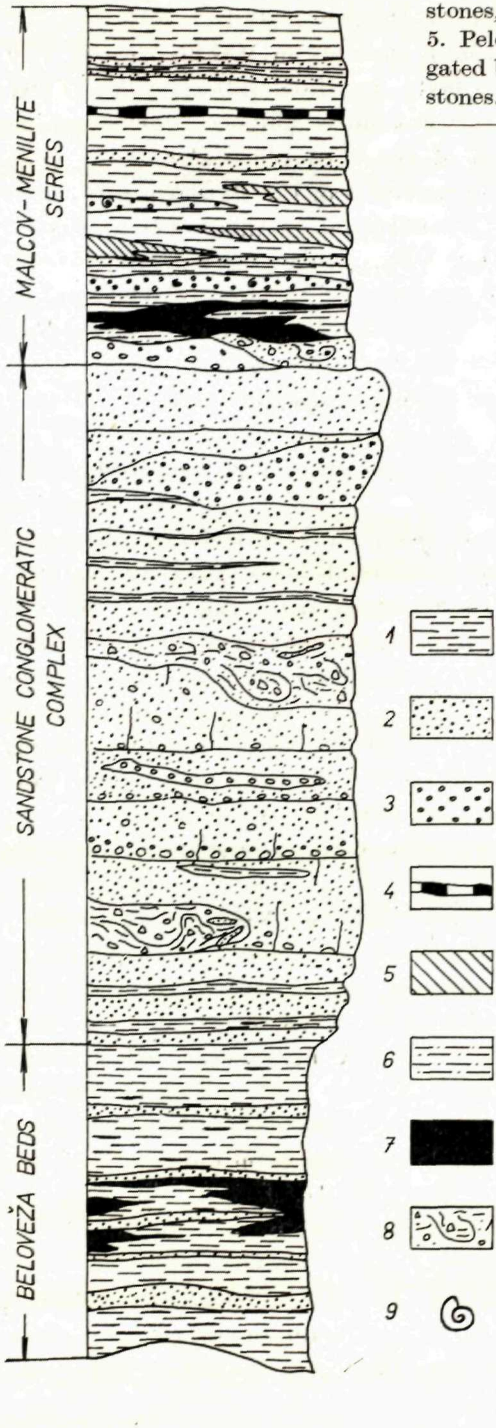


Fig. 1. Lithological scheme of Čerhov unit: 1. Claystones, 2. Sandstones, 3. Conglomerates, 4. Cherts, 5. Pelocarbonates, 6. Sandy claystones, 7. Variegated beds, 8. Slump bodies, 9. Nummulitic limestones.

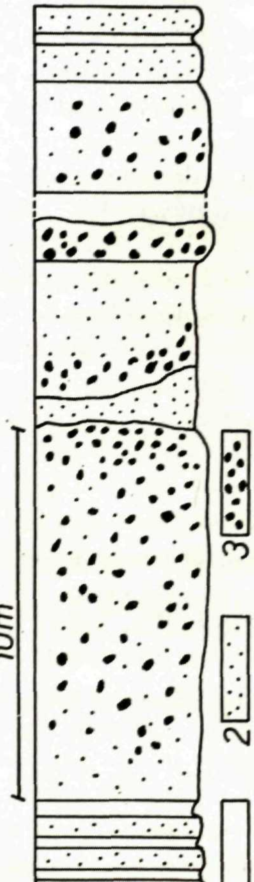


Fig. 2. Detail of the conglomerate-sandstone complex, locality Matiaška, 1. Claystones, 2. Sandstones, 3. Conglomerates.

0,5—5 m thick. Sandstones are gray-green, gray, fine to coarse-grained, mostly calcareous with muscovite. Usually on the lower part of beds small pebbles (2—5 mm) of quartz, cherts, fragments of phyllites are dispersed. Massive beds of coarse-grained sandstones frequently contain fragments of red, gray, green-gray to blue-gray claystones. In some of these Upper Cretaceous microfauna was found in the area of Kochanovce village. The claystone fragments are mostly of the Paleogene age. Sandstones of this type in upper parts of beds show platy parting. Plant debris and mica are found on the lower and upper surfaces of the beds. The thick sandstone beds are separated by soft sandy claystone.

In the overlier of the thick sandstone-conglomeratic complex (Fig. 1) there are Middle Eocene to Upper Eocene variegated beds characterized by alternation of red and green claystones. Less frequent are blue-gray, green-gray and gray calcareous and noncalcareous claystones. In this complex there are frequently concretions of Mn oxides, sometimes even 5—15 cm in size. Sometimes immediately in the overlier of Middle Eocene claystones with *Cyclamina amplectens* there is sequence of grass-green to gray-green strongly calcareous claystones with abundant Upper Eocene globigerina microfauna. Green clays only rarely exceed 50 cm in thickness. They are most thick near the NW ending of the Čerhov Mts. in the overlier of the Paleogene Klippen mantle in the Údol (Újak) village, — reaching about 5 m thickness. The Upper Eocene soft clays contain rich microfauna of quadrilobate globigerine (Samuel 1961): *Globigerina conglomerata* Schwager, *G. venezuelana* Hedberg, *Catapsydrax cf. dissimilis* (Cushman & Bermudez). This forms predominate over other species in green claystones.

The overlier of variegated claystones and clays is formed by an immense complex of the Maľcov-Menilite Series (Fig. 1) with thin intercalations of fine-grained nummulitic conglomerates and the Jaslo shales (Nemčok — Koráb — Ďurkovič 1961). These youngest flysch beds of the Čerhov unit outcrop in the synclinal zones in wider vicinity of Maľcov, Stebník, Rychvald, Raslavice, Kračúnovce and Veľká Domaša villages, in the Ondava river valley, and near Slovenské Volové village. In the Menilite-Maľcov beds in the Čerhov unit there are frequent strata of conglomeratic breccia with minor nummulites (Fig. 1). Carbonatic material of fine-grained conglomerates is represented by 3—10 mm limestone and dolomitic fragments. Basing upon the determination of nummulites by Bieda (1957, 1960), Vaňová & Köhler (in Nemčok 1961), the conglomeratic beds in the Menilite-Maľcov series in the Čerhov unit may be ordered to the lower part of the Upper Eocene.

In addition to the conglomeratic strata in the lower horizons of the Malcov beds or even on the base of the latter there are thin intercalations (max. 30 m thick) of Menilite shales. In Menilite beds there are frequent stripes and lenticles of brown to black cherts. The maximum thickness of cherts is approximately 10 m.

The Upper Menilite beds are laterally passing into an extensive flysch sequence of the Maľcov beds, gray, blue-gray, green-gray to brown-gray micaceous claystones predominating. Usually they are fine sandy with macroscopically observable mica, alternating with fine-grained gray and blue-gray calcareous sandstones. In the Maľcov beds claystones prevail over sandstones (3:1). Lithologically the Maľcov beds represent a flysch sequence with the thickness of 400—800 m. Their strike and dip indicate the synclinal character in the whole Magura Zone.

In the Maľcov beds of the partial Čerhov unit the Jaslo shales were found in three horizons above each other. The separate lithological horizons are max. 10 cm thick, resting above the claystones of Menilite type.

The Maľcov beds with the intercalations of claystones of Menilite type, cherts and Jaslo shales stratigraphically correspond to the Upper Eocene. Samuel (1960), Hanzlíková (1960) determined there microfauna with predominance of small globigerina: *Globigerina apertura* Cush., *Globorotalia centralis* Cush., *Globigerina rotundimarginata* Subb., *Gl. parva* Bolli etc. O. Samuel described small globigerina from the upper part of the Maľcov beds: *Globigerina postcretacea* Mjatliuk, *Cibicides cf. lopjanicus* Mjatliuk, and others occuring rather in the Lower Oligocene in the Lopjanice beds of outer Carpathians.

Form of Occurrence of Conglomeratic Bodies

According to the form of occurrence the studied conglomerates of the Čerhov unit belong to interformational conglomerates. They form 2—10 m thick intercalations in the surrounding flysch sequences. Usually they have sharp contact with the underlying and overlying beds. There are, however, also gradual transitions between sandstone beds and overlying conglomerate beds. Gradual transition is indicated by the increasing amount of coarser clastic components.

From the stand point of morphology the conglomerate bodies are irregularly shaped. Vertical and lateral fading-out of the separate granulometric classes may be observed.

The most frequent type of bedding of conglomerates is graded bedding, frequently repeated. This is especially characteristic of fine-grained fractions (10—20 mm). Coarse-grained varieties are irregularly bedded, characterized by the presence of fragments of claystones, sandstones, chaotically arranged in the matrix. There are sporadical occurrences of whole parts of bed sequences (Fig. 2, 3, 4) in the form of fragments in the surrounding conglomerate matter. These structural characteristics indicate the origin of the conglomerates under study in submarine slumps.

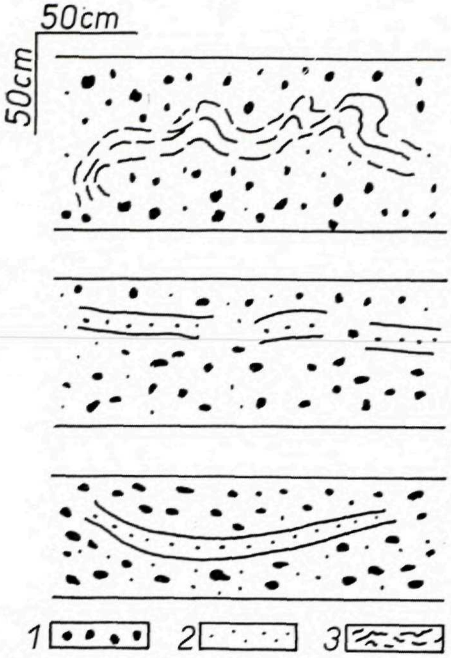


Fig. 3. Detail structure of the slump body, locality Údavské, 1. Conglomerate, 2. Sandstone, 3. Claystone.

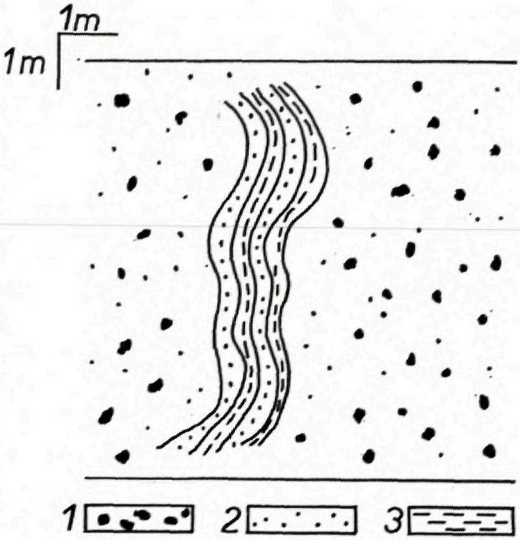


Fig. 4. Fragment of the flysch sequence in conglomerate, locality Údavské, 1. Conglomerate, 2. Sandy matrix and sandstones, 3. Claystones.

Granulometric investigation of conglomerates was carried on in the field by measuring of three axes (a, b, c) of pebbles according to Krumbein—Pettijohn (1939). 100 pebbles were measured from each exposure. The samples were chosen so that all granulometric varieties beginning with 1 cm. were included. From the results obtained, arithmetic mean and median of the "b" axis of pebbles (tab. 1) were calculated for the separate localities. Histograms of frequency are presented in Encl. 1. Calculations in tab. 1 were carried on without respect to petrographic types of rocks. Recalculation for the separate types of rocks occuring most frequently in conglomerates under study summarized for all the localities is presented in tab. 2.

Shape and roundness of pebbles. The shape of pebbles was studied on the ground of Zingg's (1935) classification based upon the relation of the axes of pebbles. The author divided pebbles into four groups according to the relations among the parametres of the axes of pebbles b/a, c/b (a-the shortest axis, c-the longest axis). In the Zingg's sense 1. spheroid, 2. disc, 3. blade, 4. roller pebbles were distinguished. Percentual representation of these morphological types on localities under study is presented in tab. 3. In similar manner also the main petrographic types of pebbles occuring in conglomerates under

Tab. 1. Textural data from conglomerates

locality	Md (b) mm	$\bar{x}$ (b) mm	Md s	Md r
10	29	33	0,65	0,32
3	37	36	0,61	0,18
5 – 6	33	34	0,63	0,17
82	25	25	0,63	0,31
38	28	27	0,62	0,37
49	31	28	0,60	0,37
P	19	20	0,60	
71	26	26	0,61	0,33
74	31	32	0,65	0,53
21	44	45	0,59	0,38
88	37	62	0,62	0,58

Md (b) – median of the b axis, $\bar{x}$ (b) – arithmetic mean of the b axis, Md s – sphericity median, Md r – median of roundness.

Tab. 2 Statistical data of pebble axes in studied conglomerates

pebble axis		a		b		c	
rock	n	$\bar{x}$ (mm)	s	$\bar{x}$ (mm)	s	$\bar{x}$ (mm)	s
limestone	340	32,64	21,14	29,84	18,52	21,87	16,10
sandstone	106	9,80	1,03	8,14	9,60	5,48	6,17
dolomite	110	5,14	5,82	3,77	4,18	3,77	4,18
quartzite	210	17,63	13,33	13,04	10,71	9,58	7,75
vein quartz	336	12,85	8,62	9,88	7,29	6,97	9,78

n – number of pebbles, $\bar{x}$ – arithmetic mean, s – standard deviation

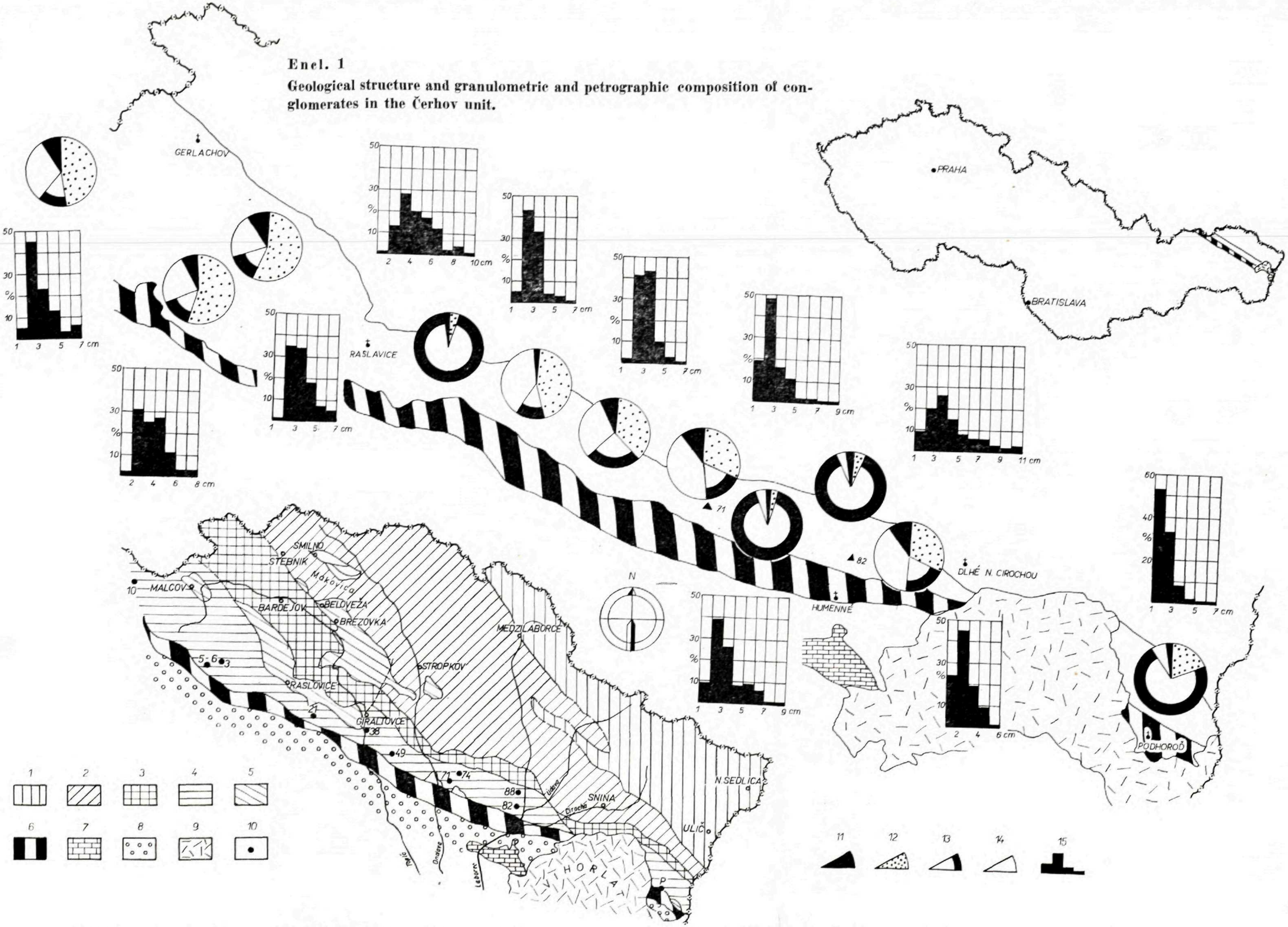
study (sandstones, quartzites, limestones + dolomites, vein quartz) were estimated. The results are in tab. 4, showing that the majority of sandstone pebbles is of the disc shape (50 %). Less abundant are spheroid shapes (23 %). Roller and blade pebbles are only sporadically present. In quartzites the spheroid shape (43 %) predominates over disc and roller shapes. The pebbles of limestones and dolomites show approximately equal representation of spheroid and disc shapes predominating over the roller and blade forms. In the vein quartz there is roughly the same occurrence of morphologic shape as in quartzites.

Sphericity of pebbles was studied on the ground of Krumbein – Sloss' (1953) classification. It is actually the Zingg's scheme modified by isolines of sphericity according to Krumbein. From the results obtained, the median of sphericity (tab. 1) was calculated. Median of sphericity was also calculated for the separate petrographic types of pebbles, with the following results: quartzites 0,64, limestones + dolomites 0,65, sandstones 0,58, vein quartz 0,65. Only the median of sphericity in sandstone pebbles is distinctly different. In the rest of pebbles median of sphericity is roughly the same.

Roundness of pebbles was visually studied on the ground of 5 degree scale by Pettijohn (1957). Results recalculated to the median of roundness for

Encl. 1

Geological structure and granulometric and petrographic composition of conglomerates in the Čerhov unit.



1. Dukla unit; 2. Rača unit of the Magura Flysch; 3. Bystrica unit of the Magura Flysch; 4. Čerhov unit; 5. Malcov and Menilite beds; 6. Klippen Belt; 7. Mesozoic of the Humenské pohorie Mts.; 8. Central Carpathian Paleogene; 9. Neovolcanites; 10. Localities of conglomerate occurrences (3 – Majdan, 10 – Legnava, 5 – 6 to the S. of Majdan, 82 – Údavské, 38 – Mičakovce, 49 – Matiaška, P – Podhoroď, 71 – Girovce, 74 – Červené, 21 – Proč, 88 – Nižné Ladičkovce); 11. Igneous and metamorphic rocks; 12. Sandstones and quartzites; 13. Limestones and dolomites; 14. Vein quartz (11 – 14 *petrographic composition of pebbles in the recalculation on 100 %*); 15. Granulometric composition of conglomerates.

the separate localities are given in tab. 1. Visual judgement may show that the best rounded pebbles are on the locality 74 (Encl. 1.) the least rounded — on the locality 3 and 5—6 (tab. 1.).

Orientation of pebbles. On four localities orientation of pebbles was studied. On fig. 5 there is graphical presentation of the relation between orientation of the longest axis (a) of pebbles and the predominating current system determined according to the orientation of flute casts on the lower part of sandstone beds in the separate profiles. Generally certain differences between the

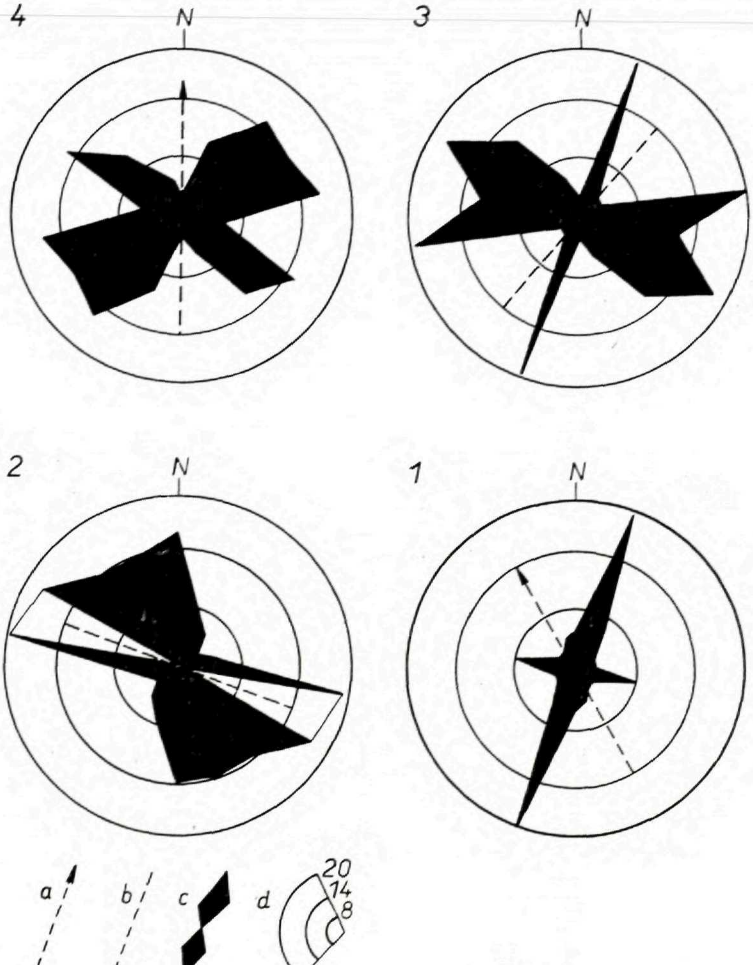


Fig. 5. relation between planar projection of orientation of "a" pebble axes and orientation of sedimentary structures on the undersurfaces of sandstones; *a* — flute casts, *b* — drag marks, *c* — orientation of pebbles, *d* — number of measurements. 1 — Matiaška, 2. Mičakovce, 3. Gírovce, 4. Údavské.

orientation of long axes of pebbles and the current systems may be observed. Fig. 6 shows spatial orientation of long axes of pebbles on the above localities. On petrographic diagrams no preferred orientation of long axes may be observed by which the opinion about slump origin of the conglomerates under study is supported.

Petrographic composition of conglomerates. Interformational conglomerates of the Čerhov unit are composed of two basic components: pebbles and matrix. Matrix is formed by sandy or clayey material. The ratio of matrix to pebbles is 1:1 or 1:2 on behalf of matrix. This ratio is changing on separate localities. Petrographic composition of matrix is represented by fine-, medium- to coarse-grained subgraywackes, their petrographic composition being generally identic with mineralogic composition of the predominating amount of sandstones in the Magura Flysch (Ďurkovič 1966).

Petrographic composition of pebbles for the separate localities is presented in tab. 5 and graphically illustrated on Encl. 1. Tab. 5 shows that the characteristic feature of the conglomerates under study is predominance of pebbles originating in sedimentary series (limestones, dolomites, sandstones, quartzites) over the igneous and metamorphic rocks. The triangular diagram on Fig. 7 shows graphical illustration of the relation between predominating pebbles. The composition of sedimentary constituents in the slump bodies (Encl. 1) is laterally changing, especially as far as the representation of limestones + dolomites, quartzites and sandstones is concerned. In direction from the W to the E distinct increase of the amount of carbonate pebbles may be observed. This may be explained perhaps by different composition of the rocks of the source area (Klippen Belt) in its W-E course.

Tab. 3. Pebble shapes in studied localities

locality / pebble shape	10	88	77	P	38	49	21	82	5–6	3
spheroid	51 %	37 %	45 %	36 %	41 %	41 %	29 %	32 %	54 %	29 %
disk	25	41	25	30	33	31	47	40	37	44
blade	6	2	3	6	5	6	10	3	5	7
roller	18	20	27	28	21	22	14	25	4	20

Tab. 4. Pebble shapes according to petrographic composition

pebble shape	sandstone	quartzite	limestone + dolomite	vein quartz
spheroid	23 %	43 %	38 %	42 %
disk	50	30	33	26
blade	10	5	5	4
roller	17	22	24	28

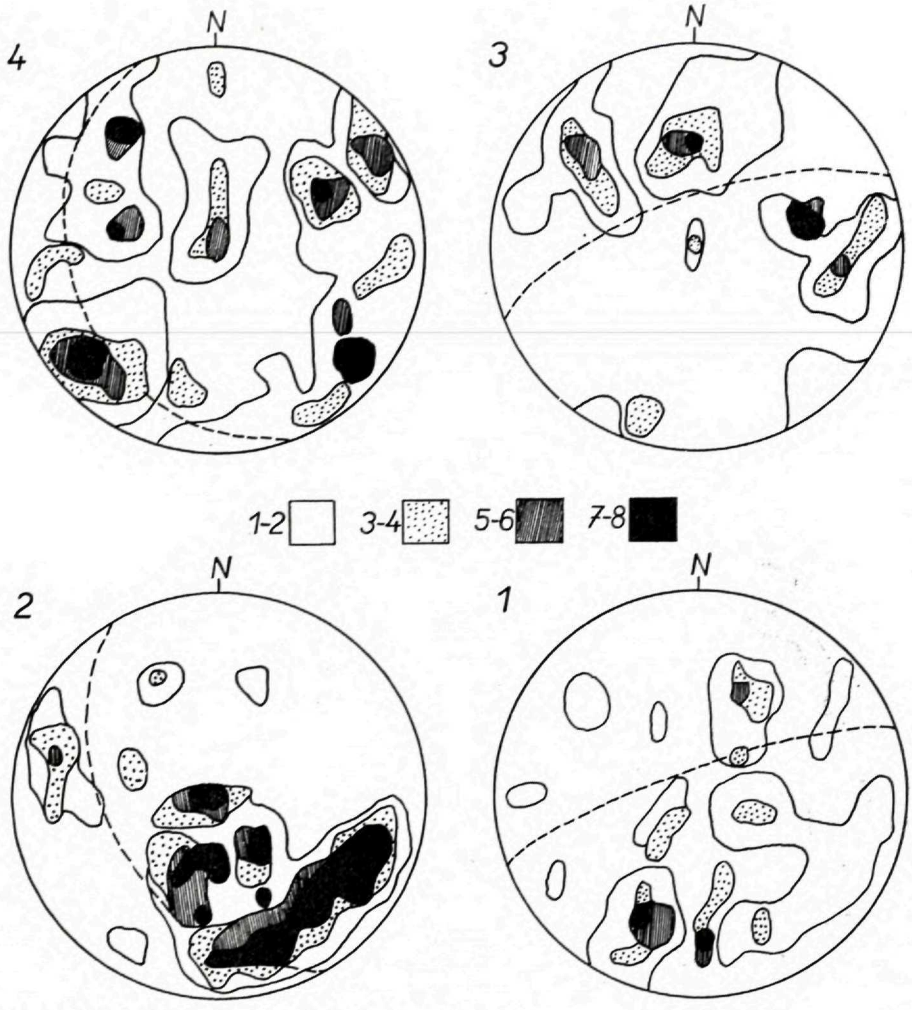


Fig. 6. Spatial orientation of the long axes of pebbles in conglomerates.

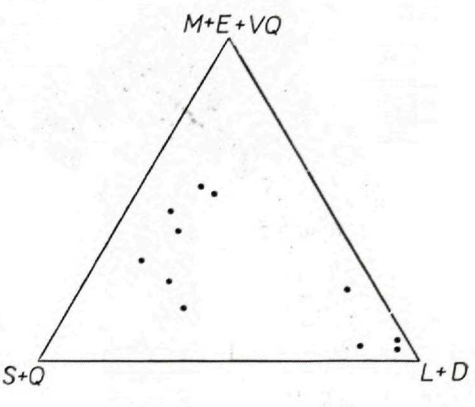
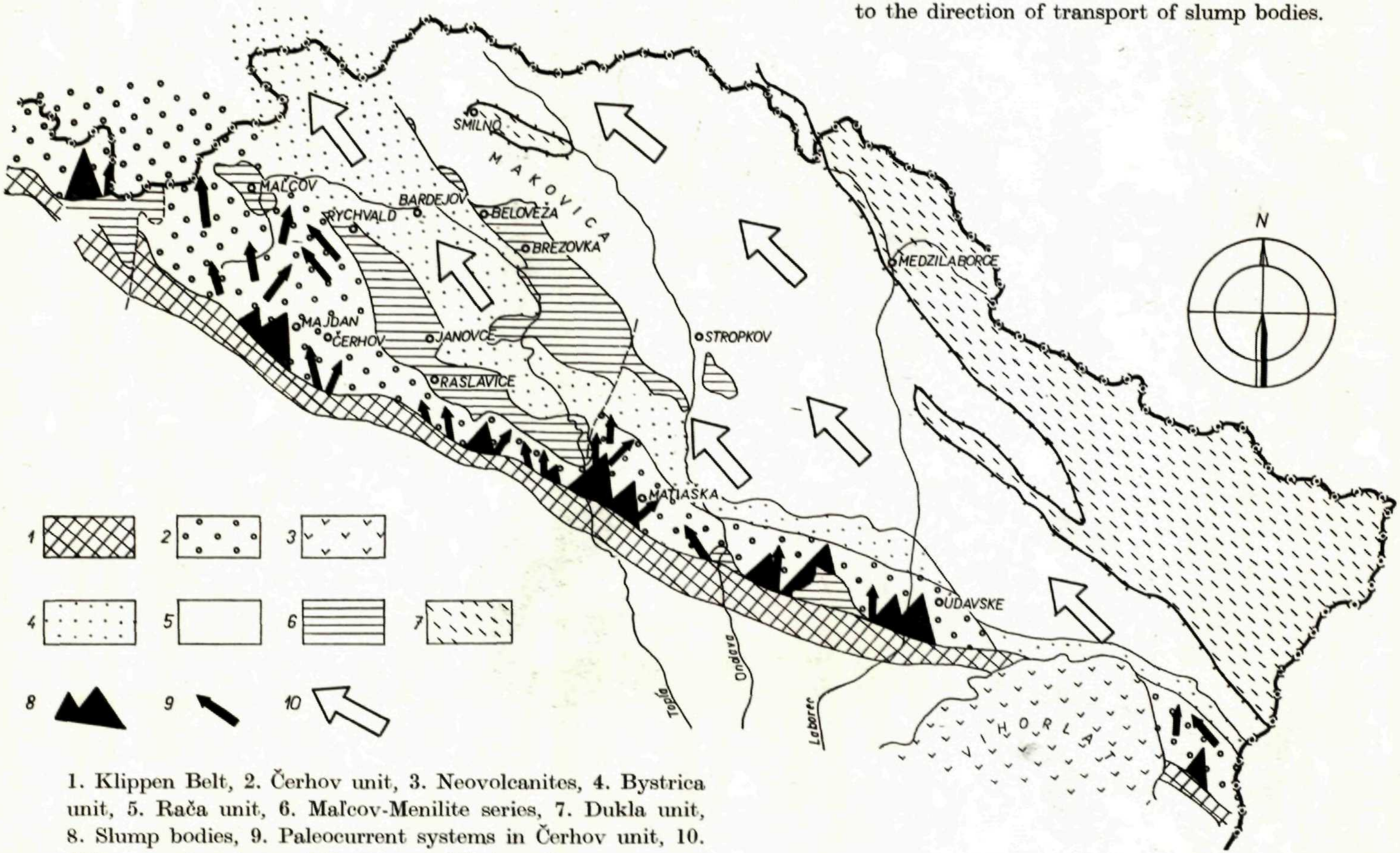


Fig. 7. Petrographic composition of pebbles in conglomerates. *S* — sandstones, *Q* — quartzites, *L* — limestones, *D* — dolomites, *M* — metamorphic rocks, *E* — eruptive rocks, *VQ* — vein quartz.

Fig. 8. Map of paleocurrent systems in relation to the direction of transport of slump bodies.



1. Klippen Belt, 2. Čerhov unit, 3. Neovolcanites, 4. Bystrica unit, 5. Rača unit, 6. Maľcov-Menilite series, 7. Dukla unit, 8. Slump bodies, 9. Paleocurrent systems in Čerhov unit, 10. Paleocurrent systems of Bystrica and Rača units.

Tab. 5. Petrographical composition of conglomerates

locality	10	3	82	74	71	P	5 – 6	49	21	88	38
limestone	19 %	19 %	17 %	48 %	12 %	29 %	12 %	24 %	70 %	67 %	13 %
marl						7					
dolomite		4	4	25	3	32	1	2	18	20	4
chert	1	1							3		1
sandstone	24	25	11	1	13	4	32	8	3	1	6
quartzite	42	31	20	2	19	17	27	24	3	3	40
vein quartz	7	18	46	3	48	7	21	34		3	34
granite granodiorite	7	1			3	2	5	2	3		2
gneiss mica schist		1	2		2		2	6			
melaphyre				1		2				4	

Manner and Conditions of Genesis of Conglomerates

Zonal distribution of different facies is a characteristic feature of the Paleogene sediments in Magura Unit (Encl. 1.). The distribution of facies is parallel with the course of Klippen Belt in East Slovakia. Beloveža beds in all the three partial units keep their stratigraphic-lithologic homogeneity, while their overlier is facially changing. Different lithofacial development of sediments in the overlier of the Beloveža beds offered the possibility to distinguish three partial lithostratigraphic complexes: the Rača, Bystrica and Čerhov units, in the Magura Flysch Zone. The common lithologic development of the Beloveža beds in all the three partial units is followed by the change of facial conditions within the Lower Eocene — upper part of Middle Eocene. These changes may be observed mainly in the supply of the clastic material. In the southernmost partial Magura Unit (Čerhov unit) situated nearest to the source area (Klippen Belt), lithofacial changes may be observed in the supply of coarse clastic material (coarse-grained sandstones, conglomerates). This coarse-grained sequence has its stratigraphic equivalent in the Zlín beds of the northern partial Magura Units.

Coarse clastic material is mostly distributed in the southernmost Čerhov unit. The supply of coarse clastic material into northern partial units was sporadical, indicated by rare occurrences of the slump bodies composed of coarse-grained material with max. thickness 0,5 m. Direction of the transport of clastic material to the Magura sedimentation area was studied on the ground of oriented sedimentary structures (Koráb and cons. 1962). In all the three partial units (Fig. 8) the preferred current direction from SE to NW was determined, found also in the Polish part of the Flysch Carpathians. It may be supposed that it is a lateral filling gradually turned in the axis of the basin into the direction roughly parallel with the Klippen Belt (Fig. 8), and the turbidity current, directions of which were determined according to flute casts, transported the greatest amounts of the clastic material into the Magura

Flysch geosyncline. The distribution of coarse clastic sediments (conglomerates, coarse-grained sandstones) in the southernmost Čerhov partial unit, and their gradual fading-out in northern direction, indicate another way of transport conditioned by nearness of the source area. Chaotic arrangement of pebbles in conglomerate bodies, the general interformational character of conglomerates, predominance of matrix in many conglomerates indicate gravitational transport in the form of submarine slumps.

The Klippen Belt (in wider sense) is considered the source area of conglomerates. This opinion is supported by determination of the orientation of flute casts on the lower sides of sandstone beds from the S to N (Fig. 8). Generally sediments of the Čerhov unit have the nature of turbidity formations with graded bedding and other accompanying structural characteristics, in which there was episodical gravitation sedimentation in the form of submarine slumps within — the Lover Eocene — upper part of Middle Eocene.

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JOINT ZONES IN THE FLYSCH OF THE CZECHOSLOVAK CARPATHIANS AND IN THE PALEOZOIC OF THE SE BORDER OF THE CZECH MASSIF

Abstract. This paper presents a summary of the work done over a period of more than five years. The object of this work was the investigation of rock jointing and other microtectonic phenomena in the Carpathian Flysch and in the SE border of the Czech Massif. Two types of macro-joints, i. e. (1) *bedding joints* and (2) *joint zones* have been established. Attention has been paid primarily to joint zones that represent an important element in the regional structural pattern. Bedding joints and joint zones distributed in the studied area exhibit variable density. The identification of joint zones grouped in sets can contribute to the knowledge of the tectonics in the entire area studied and on the basis of the results achieved there the tectonics in other regions can be solved. Joint zones indicate tectonic stresses the rocks were subjected to. The study has revealed close tectonic relations between the Carpathian system and the Czech Massif.

In 1960—1964 jointing of rocks and other microtectonic phenomena were investigated throughout the entire area of the Czechoslovak Carpathian Flysch, in adjacent Polish territories as far as Zywiec, and in the area of the SE border of the Czech Massif. The research work carried out during last few years (Plička 1960, 1962, 1963, 1964a, 1964b, 1964c, 1966a, 1966b, 1966c) has contributed to the knowledge of individual tectonic complexes in the studied area. The classification of joint systems and the evaluation of their regional significance has revealed two following joint types — (1) *bedding joints* (i. e. joints of the second order according to the conception of Soviet authors) and (2) *joint zones* (this term corresponds partly with the joints of the first order described by Soviet authors, but it refers predominantly to the term "systematic joints" applied by the American geologist R. A. Hodgson). This new term "*joint zones*" has been established for joints forming many systems persisting over long distances in the investigated area. Similar conclusions have been simultaneously inferred by R. A. Hodgson (1961, 1965) in the U. S. A. and by D. Spencer—Jones (1963) in Australia. The area subjected to investigation represents about a fifth of the entire Czechoslovak territory.

Joints extend throughout the entire studied area, but they show a variable areal and vertical distribution. Variable intensity of jointing in sedimentary rocks depends as much on the lithological development of formations as on the number and intensity of tectonic processes the rocks were subjected to. The distribution of *bedding joints* is controlled by local geological pattern. They are genetically related to the formation of individual folds. The distribution of *bedding joints* is confined to individual beds and/or to their overlying or underlying rocks. *Joint zones* pass through more beds or through whole complexes of sedimentary rocks and occur also in the crystalline rocks. They accomplish the pattern of great tectonic complexes, representing a characteristic structural unit within them.

Bedding joints are for the most part perpendicular to the bedding plane. Two systems of joints nearby perpendicular one to another predominate here. The density of *bedding joints* depends on the thickness of beds, their petrographic nature and the resistance of rocks. All evidence points to a tectonic origin of *bedding joints*. They are older than *joint zones*. In general, *bedding joints* are better developed in the Carpathian Flysch (except for the Ždánice-Subsilesian unit) than in the sedimentary rocks of the SE border of the Czech Massif, where they occur in the Culm graywackes and in thin beds of the Devonian and Carboniferous limestones. The relations established for *bedding joints*, (e. g. the ascertainment that the strikes and dips of strata affect the strikes and dips of *bedding joints*) cannot be applied to the second type of joints — to *joint zones*.

Joint zones are represented by more or less parallel running nearby vertical joints; these joints grouped in zones extend across the sedimentary rocks disregarding their strikes or dips and intersect even the *bedding joints*. They are developed in the Paleozoic and crystalline rocks of the SE border of the Czech Massif. In the Flysch of the western Carpathians *joint zones* show a variable distribution. The width of individual *joint zones* usually ranges from 0.5 to 6 m. *Joint zones* of the same strike and dip represent a *set of joint zones*. The spacings between individual *joint zones* belonging to one set exceed 3 m.

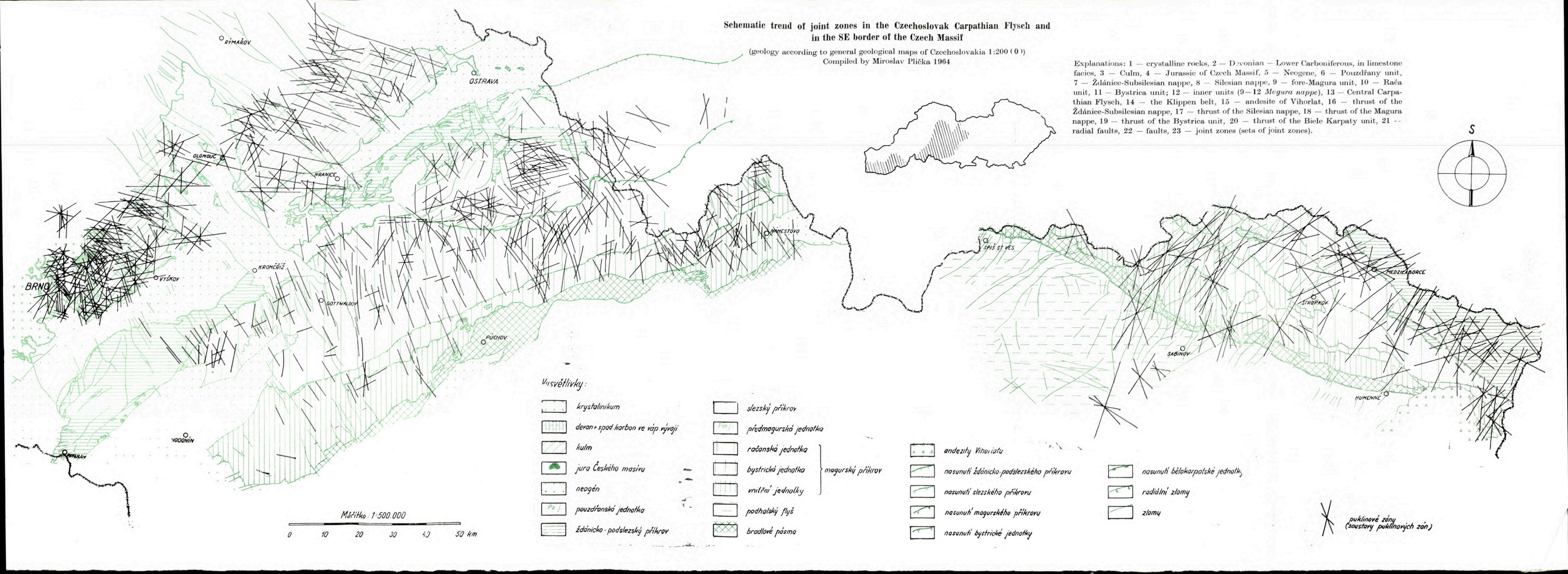
Joint zones (sets of joint zones) constitute systems of different age, strike and dip, and are closely related to tectonic processes that took place in this area. The number of *joint zones* forming a set increases towards the *faults* of regional characteristic exhibiting the same direction as the *joint zones*.

Consequently, it is evident that *joint zones* are closely related to *faults*. Field observations permit to make statements that *faults* found in the mapped area usually run parallel to two systems of *joint zones* oblique to each other (e. g. in the Culm of the Drahany Upland and in the Nízký Jeseník Mts.). It may be suggested that *joint zones* — as the result of tectonic stresses — may have formed earlier than these *faults*. The *faults* formed where the rocks were mostly weakened by *joint zones*. Judging from these observations, *joint*

Schematic trend of joint zones in the Czechoslovak Carpathian Flysch and in the SE border of the Czech Massif

(geology according to general geological maps of Czechoslovakia 1:200 (0))
Compiled by Miroslav Plička 1964

Explanations: 1 — crystalline rocks, 2 — Devonian — Lower Carboniferous, in limestone facies, 3 — Culm, 4 — Jurassic of Czech Massif, 5 — Neogene, 6 — Pouzdřany unit, 7 — Ždánice-Subsilesian nappe, 8 — Silesian nappe, 9 — fore-Magura unit, 10 — Rača unit, 11 — Bystrica unit; 12 — inner units (9—12 *Magura nappe*), 13 — Central Carpathian Flysch, 14 — the Klippen belt, 15 — andesite of Vihorlat, 16 — thrust of the Ždánice-Subsilesian nappe, 17 — thrust of the Silesian nappe, 18 — thrust of the Magura nappe, 19 — thrust of the Bystrica unit, 20 — thrust of the Biele Karpaty unit, 21 — radial faults, 22 — faults, 23 — joint zones (sets of joint zones).



zones indicate the *faults* even there, where they can be hardly ascertained by geological mapping.

Joints grouped in *joint zones* are almost vertical. *Joint zones* distributed near the thrust planes of the nappes in the Carpathians and parallelling the thrusts show dips ranging from 40° to 70°. *Joint zones* are nearby perpendicular to the thrust plane of the nappe.

The density of joints in a *joint zone* is variable. It can be defined by a fraction, where the numerator represents the number of joints per the zone width, and the denominator defines the zone width in metres. In the studied area, we have observed the following relations between joint density and the width of joint zone in different types of rocks of different age:

1. limestone, conglomerate (Devonian) — 3/1, 5/1, 10/1, 5/1, 10/2 ...
2. shale, graywacke, conglomerate (Culm) — 5/3, 3/0,5, 5/2, 4/1, 17/4 ...
3. limestone (Jurassic) — 7/3, 5/2, 5/1, 3/1, 6/3 ...
4. limestone (Lower Cretaceous of the Silesian nappe) — 5/1,5, 5/2, 30/6, 5/1, 10/2 ...
5. sandstone (Middle Cretaceous of the Silesian nappe, Godula beds) — 4/1,5, 6/0,5, 9/1 ...
6. sandstone (Upper Cretaceous of the Silesian nappe, Istebné beds) — 10/2, 10/1,5, 5/1, 10/4 ...
7. sandstone (Paleocene, Eocene of the Magura nappe) — 6/1, 4/1,5, 10/4, 15/2 ...
8. marlstone (Eocene of the Magura nappe) — 6/1,5, 4/1, 5/0,5, 5/1 ...
9. granite of the Brno Massif — 5/1, 4/2, 6/1,5, 5/2, 10/6, 3/1 ...

It is apparent that no relations exist between the number of joints and the zone width, nor the minimum or maximum width of a certain zone can be controlled by rock type, stratigraphy and/or tectonics in the investigated territory. Individual joints within a *joint zone* can usually reach a width of some mm to several cm. Joints are commonly filled by calcite or quartz.

Joint zones (sets of zones) are of tectonic origin and can reach a length of several tens of kilometres. In pelitic, plastic or less compact rocks they disappear locally towards the overlying rocks or towards the basement. They are best developed in graywackes and conglomerates of the Culm of the SE border of the Czech Massif, in the crystalline rocks of the Brno Massif, and in sandstone beds of the Carpathian Flysch (see the Magura and Godula nappe, the Dukla—Užok folds). An average width of a sett of *joint zones* exceeds locally 20 km (see the NW system in the SE border of the Czech Massif between Brno and Vyškov, the ESE system in the Culm between Opava and Olomouc, and the ENE system of the Godula nappe).

A great areal extent of *joint zones* and their persistance over long distances may be proved by the following data:

1. analogy to extensive *joint zones* in other areas over the world, examined at surface outcrops and by means of the aerial photographs (R. A. Hodgson 1961, 1965; D. Spencer—Jones 1963);

2. persistence of a specific system of *joint zones over* different localities in the area;

3. *fault directions* conforming to that of *joint zones* within an extensive area;

4. conformity between the water flow direction and the direction of *joint zones* in an extensive area (e. g. in the Godula nappe; M. Plička 1963).

Joints arranged in *joint zones* show discontinuities both in their vertical and areal trend, where by their ends overlap. In the mapped area *joint zones* are grouped in *sets* belonging to different systems. Locally one system can be dominant, (e. g. in the western portion of the western Magura Flysch) and sometimes many intersecting systems may be developed (in the crystalline and Paleozoic rocks of the SE border of Czech Massif, in the Godula nappe, etc.).

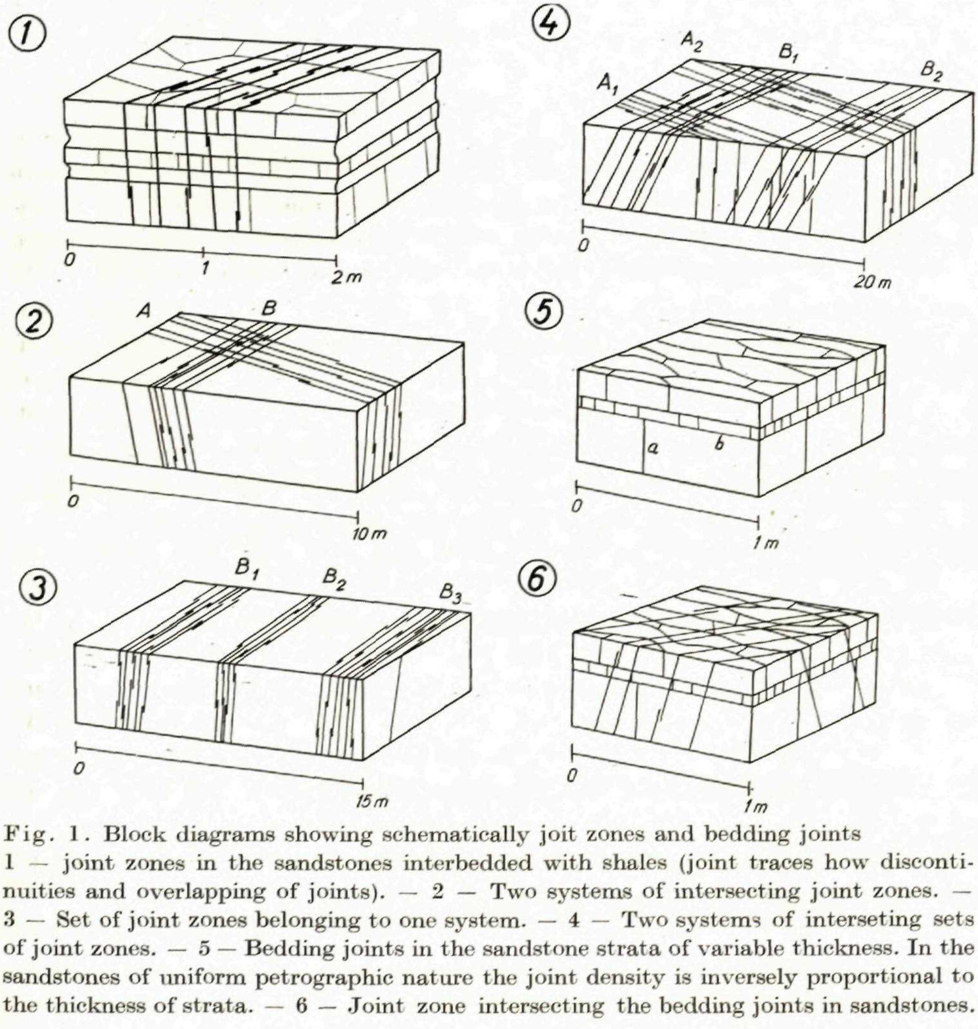


Fig. 1. Block diagrams showing schematically joit zones and bedding joints
1 — joint zones in the sandstones interbedded with shales (joint traces how discontinuities and overlapping of joints). — 2 — Two systems of intersecting joint zones. — 3 — Set of joint zones belonging to one system. — 4 — Two systems of interseting sets of joint zones. — 5 — Bedding joints in the sandstone strata of variable thickness. In the sandstones of uniform petrographic nature the joint density is inversely proportional to the thickness of strata. — 6 — Joint zone intersecting the bedding joints in sandstones.

In general, a difference has been found between the development of *joint zones* belonging to the Carpathian system and that of the SE border of the Czech Massif (see the map). Contrary to the Czech Massif, in the Carpathian Flysch *joint zones* are not so well developed and locally they are missing (the Ždánice-Subsilesian unit; the Hluk development of the Upper division of Paleogene of the Bílé Karpaty unit; the Zborov—Smilno anticlinorium in the eastern Slovakia). In the Carpathian Flysch, *joint zones* are almost perpendicular to the direction of the Carpathians and show a fan-like arrangement conforming with the Carpathian arc. In the Carpathians, a greater concentration of systems of *joint zones* has been recorded in the Magura Flysch east of Velké Karlovice, and in the Godula nappe.

A greater number of *joint zones* in the rocks of the Czech Massif (when compared with their density in the Carpathian system, e. g. in the Ždánice-Subsilesian unit) can be explained by different lithological and tectonic development of these two systems. In the Czech Massif an important factor is represented by the crystalline basement, the rocks of which were subjected to tectonic stresses; from there *joint zones* pass into the overlying Paleozoic rocks. Consequently, fault zones in the deeper basement rocks are indicated by *joint zones* (see the system of *joint zones* running from Vítkov towards the SE into the Carpathian system as far as the inner Klippen belt), and *joint zones* point to tectonic relations between the Czech Massif and the Carpathian system. Analogous tectonic relations between the Carpathain system and the Czech Massif can be observed where these two systems are in closer contact, i. e. in the area between Brno and Vyškov, and where *joint zones* run continuously from the Czech Massif into the Carpathian nappes.

Western part of the Carpathian Flysch

The Magura Flysch: *joint zones (sets of joint zones)* can be traced from the inner Klippen belt throughout the Magura nappe. They penetrate partial tectonic units of the Magura nappe and in the case of a favourable lithological development of formations they also pass across the Ždánice-Subsilesian unit. Apart from the aforesaid transversal *joint zones*, joint zones of a NNE to NE direction appear in the eastern part of the studied area from Makov towards the east. A multiplication of *joint zones* is evident where the Flysch formations are affected by the fault line of Raková—Semeteš trending towards the NE. This multiplication of *joint zones* has been recorded in the space of the maximum bend of the Carpathian arc.

In front of the Magura nappe, between Valašské Meziříčí and Jablunkov, we may observe that *joint zones* disjoin, some of them verging their directions towards the NW, the other towards the NE. *Joint zones* trending from NE to

SW follow the *fault line* Raková—Semeteš, the other *joint zones* that run in a north-western direction towards Rožnov p. Radhoštěm and Valašské Meziříčí are bounded in the west by a *fault line* directed from the western part of the Bystřička dam towards the SE. This disjoining of *joint zones* has been observed in the western part of the Magura nappe, in the space of its maximum bend, where the inner Klippen belt is strikingly displaced towards the north, and where, in the front of the Magura nappe, the outer Flysch zone is built by the Godula and Istebné beds containing thick sandstone strata. It is suggested, that two-directional tectonic stresses that operated in that portion of the Magura nappe, where the Godula nappe resisted to the front of the Magura nappe moving towards the north, are responsible for this divergence of *joint zones*. The movements of the Magura nappe also diverged towards the NW and NE, which has been proved not only by *joint zones*, but also by small faults and groove casts, shown on the joint surfaces. The eastern blocks were thrust towards the NE along smaller transversal faults showing negligible horizontal to oblique displacements. A fanlike divergence of *joint zones* directed from SSW to NNE and from SSE to NNW can be traced from the Makov—Mariková transversal elevation in the Bytča region. This arrangement, obviously influenced by this prominent transversal elevation, points to a contemporaneous genesis of the two tectonic phenomena.

Longitudinal *joint zones* predominate especially in fronts of the nappes and result from tectonic stresses the nappes were subjected to. Joint in zones are nearby perpendicular to thrust planes.

The outer Flysch zone: distribution of joint zones belonging to the outer Flysch zone is variable. In individual formations of the Ždánice-Subsilesian unit (see Ždánice—Hustopeče beds consisting from less compact rocks) and in the Silesian unit (see the lower Těšín shales) *joint zones* are only weakly developed; locally *joint zones* are missing. On the other hand, in massive sandstone strata of the Silesian nappe (the Godula and Istebna beds), *joint zones* are very common, especially in the sedimentary rocks which were exposed to intensive tectonic stresses. In the Godula nappe, (dominantly in tectonic scales of Javorník and Lysá Hora Mt.), *joint zones* are grouped in sets extending over several kilometres.

From observations on the relationship between *joint zones* in the Magura nappe and in the outer Flysch zone it is obvious that *joint zones* pass through both tectonic complexes. It is well apparent in an about 10 km wide belt in the Jablunkov depression and further in another belt trending to Staré Hamry and Frýdlant n. Ostr., the width of which is about 7 km; in the northern trend of this belt a 7 km wide zone of transversal faults deforming the front of the outer Flysch nappe has been observed.

Less pronounced are transversal joints passing from the Magura Flysch through the Ždánice unit; this is due to different lithological development of the two units.

In consideration of general distribution of *joint zones* in the western part of the Carpathian Flysch, prominent are longitudinal *joint zones* in the Godula nappe, most probably due to its slab-like character and to the presence of thick layers of massive sandstones. Two dominant directions of the sets of *joint zones* in the Godula nappe are SE—NW, and WSW—ENE (SW—NE). The first group of *sets of joint zones* (showing a SE—NW-direction) is prominent in the Jablunkov depression and east of Jablunkov. It is genetically related to the tectonics of the Jablunkov depression. The second group conforms with the longitudinal and diagonal tectonics of the Zubří—Pindula, Staré Hamry, and Predmier zone. In the tectonic scale of Lysá Hora Mt. *sets of joint zones* run roughly parallel to the bed strike and parallel the longitudinal normal fault at Nýdek. These parallel directions suggest a relationship between longitudinal *sets of zones* in the northern portion of the scale of the Lysá Hora Mt. and longitudinal tectonic element represented by the Nýdek fault.

Eastern part of the Carpathian Flysch

The eastern part of the Carpathian Flysch consists from the Magura nappe, the Dukla—Užok folds and a portion of the Central Carpathian Flysch E and SE of Vysoké Tatry Mts. In this area *joint zones* striking from NNE to SSW predominate; subordinately *joint zones* oriented from NNW to SSE have been encountered. *Sets of joint zones*, similar to those in the western part of the mapped area, pass through all tectonic units without changing their directions and extend even across the area south of the Klippen belt as far as the southern margin of the Central Carpathian Flysch at Spišské Vlachy. In this area we can also find close relations between *smaller faults*, *faults of regional characteristic* and *joint zones*, as stated for the western part of the territory. It are dominantly NNE—SSW and ENE—WSW-trending *faults* that have been found by geological mapping both in the Central Carpathian Flysch and in the Flysch nappes NE of the Klippen belt.

Joint zones are closely related to the macrotectonics and their continuous trend through individual tectonic units indicates their relationship to the regional tectonics of the basement, observed in the western part of the studied area too.

South-eastern border of the Czech Massif

Contrary to the Ždánice-Subsilesian unit of the outer Flysch zone, *joint zones* in the Paleozoic, Mesozoic and crystalline rocks of the Czech Massif exhibit a specific development. They constitute sets belonging to different systems. Certain analogy to joints observed by R. A. Hodgson (1961, 1965) in the Paleozoic and Mesozoic rocks of Utah and Arizona in the U. S. A. may be stated in this area. The origin of many systems of *joint zones* was influenced by the presence of crystalline basement that forms one tectonic unit with the observed sediments, which is not the case in the Carpathians. It has been found that some sets of *joint zones* persist over a distance of 60 km and their length reaches 100 km (see some *sets of zones* trending from the Czech Massif to the Carpathian arc, e. g. in the area of Horní Benešov—Nový Jičín—Rožnov p. Radh. and farther to the SE). The *sets of joint zones* in the south-eastern border of the Czech Massif are variable in width. Some *sets of joint zones* show a width of some kilometres only, other sets, such as a WNW—ESE-trending set, extend throughout the area in a 100 km wide belt without changing directions. In the NW set of *joint zones*, for instance, we may observe that *joint zones* are locally missing in belts up to several kilometres wide.

Close relations between *joint zones* and *faults* are reflected in detail as well as in the regional joint pattern of this area. On the basis of a detailed investigation carried out in the Drahany Upland, the *faults* running parallel to *joint zones* are suggested to change often their directions and follow other different systems of *joint zones* oblique to each other (e. g. NW-directed faults at Vyškov and Luleč.)

The *set of joint zones* trending from WNW to ESE is well developed in the entire area of the SE border of the Czech Massif between Opava and Brno. This set has been observed both in the Paleozoic and crystalline rocks. It may probably belong to older systems of *joint zones* reactivated during younger tectonic movements.

The *set of joint zones* directed from NW to SE is weakly developed in the area between Opava and Brno. Locally it fails to continue, but here and there it is very distinct (e. g. in the area of the Maleník block — see Z. Roth, 1962). In general, this *set of joint zones* may have resulted from younger reactivated tectonic processes.

The set of joint zones striking from NNW to SSE and extending through the sediments belonging to the Culm and Devonian between Brno and Vyškov may owe its genesis to tectonic processes active in the Carpathians.

The N—S-trending *set of joint zones* is distributed in the crystalline and Paleozoic rocks near Brno. It is obviously related to *faults* striking from N to S associated with diabase effusions.

The *set* oriented from NE to SW originates in the crystalline rocks of the Brno Massif and continues across the Devonian sediments of the Moravian Karst towards the NE — into the Culm sediments near Opava.

Joint zones composing the E—W-directed set are confined only to the area extending between Hranice, Prostějov and Brno.

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FAULT TECTONICS OF THE ŠTIAVNICA HORST

Abstract. Within the area of the so-called Štiavnica Horst, between Banská Štiavnica—Banská Hodruša—Vyhne—Sklené Teplice the pre-Neogene basement of young volcanic rocks crop out. The structure development of the Štiavnica Horst was controlled particularly by fault tectonics. Its importance is enhanced by the fact that part of these faults became ore-bearing elements of the well-known Banská Štiavnica and Hodruša ore districts. Up to the present, the fault systems were interpreted simply as the systems of normal faults within the framework of the horst-graben structure. Our investigation has shown that the fault systems are of a far more complicated genesis.

Outline of the structure of the Štiavnica Horst

The structure of the Štiavnica Horst consists of two conspicuous structural units which differ in age, tectonic style and lithology. One of them is the pre-Neogene substratum made up of Palaeozoic, Mesozoic and also Eocene complexes. The other, Neogene unit occurs in the form of a cover formation built up mostly of the supercrustal accumulations of the products of the West Carpathian volcanism.

The substratum of young volcanics is both stratigraphically and petrographically unusually varied. Among its components, the granite (called *Vyhne granite*) probably of Palaeozoic age, constitutes the crystalline core. Additionally, Carboniferous conglomerates, sandstones and micaceous shales, belonging to the Choč nappe (in part probably to its envelope) are present. Permian arkoses, sandstones and variegated beds developed in patches also make up slices of the Choč nappe. Mesozoic rocks occur in two incomplete units — the Krížna unit and the Choč unit. The former represents the Krížna nappe forming the rolled-out „polsters" in the substratum of the Choč nappe. It is composed of the Anis-Ladinian carbonates, the Keuper variegated sequence, Jurassic (Lias?) fragments and the Cretaceous (Neocomian-Albian) complex. The Choč unit built up of the typical Werfenian and Middle Triassic carbonates tectonically overlies, together with the Carboniferous and partly also Permian, the Krížna nappe. The Eocene nummulites-bearing conglomerates rest with a pronounced unconformity on the above-mentioned structures.

With regard to its character, the large Neoidic intrusion of diorite — quartz-diorite, granodiorite and some of its porphyry and aplite dyke forms — should also be ranged to the substratum. In the mode of occurrence, extent, composition and intensive contact effects, these rocks are reminiscent of the banatites of Rumania (Rozložník — Šalát 1963). Broadly speaking, they penetrated between the crystalline core and its sedimentary mantle formed of the above-mentioned nappes. The basic tectonic style of the substratum of young volcanics developed during the Cretaceous folding processes, which stage is called the Early Carpathian or Early Alpine. Distinctive of the Early Alpine stage is the nappe tectonics. The study of small and major tectonic structures has revealed two main trends of the Early Alpine structure: the main, approx. NE-SW and the subsidiary SE-NW strikes. They are reflected in the intricate brachyanticlinal structure, emphasized by the diorite-granodiorite intrusion (forming the huge core of the Štiavnica Horst) which gave rise to the Hodruša upfold with a number of partial elevations (e. g. Štiavnica elevation, elevation of Rosalia vein) and depressions (see the section).

The Neogene volcanic complex of the Štiavnica Horst is made up of an abundance of volcanic products corresponding in time with the 2nd and 3rd andesite phases and the 3rd rhyolite phase in the sense of Kuthan's classification (1963). Their position within the framework of the volcanics of the Slovenské stredohorie Highland is explained in another paper (Rozložník — Slavkovský, 1967). The rocks build up mostly surface forms; the only exception is the hornblende-biotite dacite which develops subvolcanic forms, mainly in the substratum, tending to die out towards the overlying young volcanic complexes. The rudimentary supercrustal accumulation of volcanics in the centre of the Štiavnica Horst, showing an incomplete development compared with the peripheral parts of the Horst, suggests that during the Neogene the Hodruša upfold experienced positive vertical movements. The terraces of streams at altitudes higher than 600 m are a good evidence that the uplifts have persisted up to present time.

The tectonicity of the Neogene period widely differed from that of the Early Alpine stage, being distinguished by fracture-fault style. During this Late Carpathian or Late Alpine stage, the Early Alpine and Late Alpine tectonic plans influenced each other. In the pre-Neogene substratum, the elements of the Late Alpine stage represent the ,,superimposed'' elements (dykes, veins, faults, a. o.). On the other hand, the anisotropy of the Early Alpine structural elements controlled to a certain extent the development of the Late Carpathian structural plan of the overlying volcanic cover.

Dating of fault systems

From this very brief outline of the evolution of the structure it follows that the faults are confined broadly to the Late Alpine stage. As it was clear already at first sight that the faults did not originate simultaneously, the restoration of the fault-tectonic history required first to date the individual fault systems. The dating of some fault systems could be based on several points of support.

The first demonstrable fault system closely preceded the formation of sills and dykes of dacite, the age of which is relatively precisely established within the succession of young volcanics (between the hornblende-biotite andesite and the pyroxene andesite of the "peak" phase).

The second, well demonstrable, so-called post-dacite group is far more varied and its generation was obviously attended by several partial stages. The date of its formation is defined by the metallization of the Banská Štiavnica—Hodruša ore district and by the rhyolite eruptions.

Understandably, the mineralized faults are of greatest economic interest. Their position within the framework of the structural plan cannot be studied irrespective of the evolution of the Late Alpine structures and, as will be shown below, of the Early Alpine structures either.

Therefore, we shall pay particular attention to the structural plan of the faults associated with the dacite stage.

The dacite stage of fault formation is manifested mainly by the dykes of hornblende-biotite dacite which are mostly well traceable in the field. They occur in the granodiorite-diorite complex, in the sedimentary sequence, as well as in the volcanic andesite (incl. hornblende-biotite andesite) complex. Their thickness ranges from the order of decimetres to several tens of metres. About 160 dykes of this type have been registered but, owing to a small thickness, some of them could not be plotted in the map. Whereas the determination of strikes is fairly reliable, the values of dips are less precise. The strike pattern is shown by the rose diagram. The maxima are concentrated to the NE- and approximately NW-trends. The dykes of the first group most frequently dip to the SE and those of the second group are steeply inclined to the NE and SW.

Apart from dykes, there are also dacite sills and bodies resembling laccoliths. The importance of sills lies in their indicating the dynamic-kinetic conditions under which the dykes were formed. The sills are concentrated in the sedimentary zone intervening between the granodiorite massif and the volcanic complex. They used the zones of weakness in the sedimentary complex developed during the translation of nappes in the Early Alpine stage and, invading them, they frequently enclose the whole blocks of sedimentary rocks.

The dykes show a comparatively complicated course. Branching and bifurcation are often observable. In most cases, evidence is available that their origin was accompanied by block-fault movements, i. e. that they are due to normal faults of a determinable amount of displacement. In some cases, however, the movement has not been proved.

The first question to be solved in analysing the structural plan of faults of the dacite stage relates to their possible connection with the structural plan of the Early Alpine stage. The feasibility of the influence of the earlier structural plan is strengthened by the fact that the overwhelming part of the dacite dykes is demonstrably founded in the lower structural layer, dying out upwards into the overlying volcanic complex. The relationship of the dacite sills to the Early Alpine structure is very expressive, but a considerable part of dykes also follows its main (NE) and subordinately even its transverse (NW) strike.

From this it is inferable that the structural anisotropy of the Early Alpine stage exerted influence on the origin of the dacite-stage faults. However, some deviations have also been observed.

The second question refers to the dynamic regimen which had caused the rejuvenation of the older structural strikes, the zones of weakness and the initiation of the fault system. In this regard the following three factors should be taken into consideration: (1) The fault system is connected in space and symmetric disposition with the domal structure in the core of which the granodiorite is emplaced; (2) during the Late Alpine stage this domal structure had a tendency to rise; (3) the common uniform filling of the faults generated attests that the systems of different strikes (NE and NW) are necessarily of contemporaneous origin.

As a result, the origin of faults of the dacite stage may readily be interpreted in the following terms:

The core of the domal structure — the granodiorite body — functioned as a bearer of the vertical uplift movements, as due to its relative homogeneity and isotropy in relation to the diverse geological environment, it displayed an "en bloc" behaviour. Tension developing in the apex of the rising core resulted in the formation of concentric and radial faults filled with dacite and of sills located in the most mobile zone (during the uplift) of sediments intervening between the granodiorite and young volcanics.

With respect to the three above-mentioned basic facts all other theoretically feasible interpretations are improbable and unprovable.

The post-dacite stage of fault formation comprises three principal representatives of fault systems:

(1) *The mineralized faults representing the Au—Ag-polymetallic formation* of the Banská Štiavnica ore district. Their plan is known with a fairly high accuracy.

(2) *The type represented by a single major fault* within the area studied,

called Považanská fault. This does not show direct signs of mineralization but its course is documented by rhyolite eruptions. The enclosed sketch map indicates that the Považanská fault bounds the Štiavnica Horst in the W and NW.

In the south-western part it truncates the Early Alpine structural strikes and in the northern tract it swings into the Early Alpine — NE strike.

(3) *Not very numerous faults which probably originated only after mineralization.*

As is seen from the rose diagram, the first group of postdacite mineralized faults show a rough symmetry to the faults of the dacite stage, except for the minimal distribution of the NW strike. In the W part, along the Považanská fault, there are westerly dips, whereas in the remaining part of the area the faults, with a few exceptions dip, to the E at an increasing angle towards the east. Similarly as the dacite dykes, the ore veins also branch, bifurcate, etc.For completness'sake it should be noted that ore veins mostly represent dislocations of normal fault character. They are often governed by the course of dacite dykes or are more or less symmetrical to the structural pattern of them. The uniform mineralization suggests that the pattern did not change in essentials during the development of the mineralized faults. As far as some differences do exist, we think them to be due to monoascendent zoning.

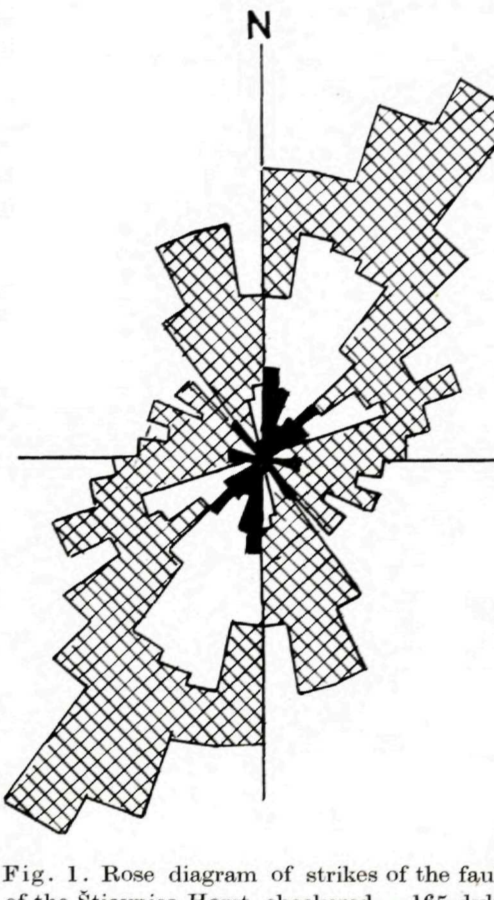

Fig. 1. Rose diagram of strikes of the faults of the Štiavnica Horst. checkered — 165 dykes of hornblendebiotite dacite; plain — 57 ore veins; solid — 44 faults (mostly without mineralization)

From what has been said about the mineralized faults it could be concluded that they formed under nearly the same dynamic conditions as the faults of the dacite stage. Even the uplift movement of the Hodruša elevation persisted during the post-dacite stage and, as mentioned above, it has been virtually active up to date. However, several facts do not accord with the existence of an identical dynamic regimen of the dacite and post-dacite stages.

On the Všesvätých stockwork (at the eastern outskirts of Banská Hodruša) two outstanding systems of mineralized faults are developed: one system, undoubtedly inherited, strikes approximately north-east and is transected by faults of the N-S strike. An analogous situation exists on the Anton Paduanský stockwork (between Vyhne and Banky), where the two strikes have an arrangement reminiscent of horse-tail structure.

The meridian trend is characteristic for the prevalent part of the Váh fault and is also outstanding with the postmineralization faults. The latter subordinately display the north-western and quite scarcely the E-W directions.

The differences between the structural plans of the dacite and the post-dacite stages cannot be ascribed to another anisotropy. The reason must be sought for in diverse dynamic conditions, which probably arose in consequence of tectonic regional events manifested in the Slovenské stredohorie Highland during the Neogene. According to Kuthan (1962, 1963, 1965), the margins of the Štiavnica Horst are followed by two major tectonic lines approaching in strike the meridian direction and extending as far as the Kremnické hory area. These played an important role in the course of the 2nd and 3rd andesite phases and the third rhyolite phase, including the mineralization processes. In our opinion, they also induced the above-mentioned changes in the development of faults of the post-dacite stage. In some aspects, the final phases of the fault formation in the Štiavnica Horst require still a more detailed research.

Conclusion

The existing knowledge of the development of faults in the Štiavnica Horst may be summed up as follows:

The formation of faults during the Late Tertiary took place in two partial stages — the dacite and post-dacite stage. The known mineralized faults are placed in the latter stage.

The formation of faults of the post-dacite stage was strikingly affected by the structural anisotropy of the substratum of the volcanic rocks that was inherited from the Early Alpine stage. The initiation of faults was associated with the upheaval of the Hodruša dome: concentric and radial faults filled with hornblende-biotite dacite were generated.

In the post-dacite stage, the location of faults was influenced by similar factors as that of the dacite-stage faults and, moreover, by new, different elements caused by regional tectonic effects. They gave rise to faults oriented in the N-S direction.

In the light of new information, the Štiavnica horst-graben structure appears to have undergone a far more complicated evolution than was presumed so far, i. e. a polyphasal development in which a number of diverse factors played role.

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DEPENDENCE OF THE DEVELOPMENT OF TERTIARY BASINS IN THE INNER SIDE OF THE WEST-CARPATHIAN ARCH UPON THE STRUCTURE OF THE SUBSTRATUM

At the end of the Paleogene, and especially in the Neogene, in the inner side of the West-Carpathian Arch there were superimposed basins of "the intermontane superimposed depression" type (T. Buday in Buday—Cicha—Seneš 1965), extending over the tectonic units of the inner West Carpathians and partly over the intermontane part. Such are the Ipeľ-Rimava basin (Map I) and the Danubian basin (Map II). In their basic regional features the structure of the substratum is not reflected, since they are generally oriented diagonally to the course of the Alpine tectonic units of the substratum. In the separate parts of the basin, however, the influence of the structure of substratum upon the filling is evident, observable mainly : 1. in the migration of subsidence, 2. in different facial development of sequences of the same age on different tectonic units of the substratum, as well as in different facial development of the separate stratigraphic stages. Stratigraphic division and the environment of sedimentation have been proved by rich fauna lastly dealt with by the following authors: E. Brestenská, I. Cicha, V. Kantorová, R. Lehotayová, V. Molčíková, A. Ondrejíčková, K. Slavíková, J. Seneš, M. Vaňová, printed in *Explanation to the general geological map 1:200 000 (Rimavská Sobota, Zvolen, Nitra, Nové Zámky, Bratislava).*

At the end of Paleogene, in the Rupelian, the marine transgression passed from the intermontane area in the S frontally into the Ipeľ—Rimava basin. The Rupelian flood penetrated only to the southern parts of Gemerides and Veporides, as indicated by the present-day distribution of sediments over southern Slovakia (borings in the southern parts of the Rimava and Ipeľ depressions). In borings near Štúrovo (the south-eastern part of the Danube lowlands, where the Tertiary substratum was formed by a mountain-range), the Rupelian developed from the Lower-Oligocene Cyrena beds. The "lower sandstone horizon" — a 100—150 m thick basal part of the Rupelian — consists according to Dobra (1960) of quartzose material of 0.5—1 mm granularity, with intercalations of pelites, gravels, conglomerates and allochthonous coal. In the Ipeľ depression, the Rupelian rests on the Veporide crystalline shales and on the metamorphosed mantle.

According to Krystek (1958) and Marková (1966), in the basal part there are arkosic sandstones (Md 0.3—0.6 mm, So < 2), including the strata of gravels and conglomerates composed of angular quartz and quartzite pebbles with sporadical fragments of phyllites and amphibolites. They pass into badly sorted subgraywackes (Md 0.06—0.15, So > 2) containing rounded pebbles of quartz, quartzites and cherts. Analogous sediments are there on the base of the Rupelian in the Rimava depression. In conglomerates carbonatic material (Homola 1954) of the Gemeride Mesozoic predominates.

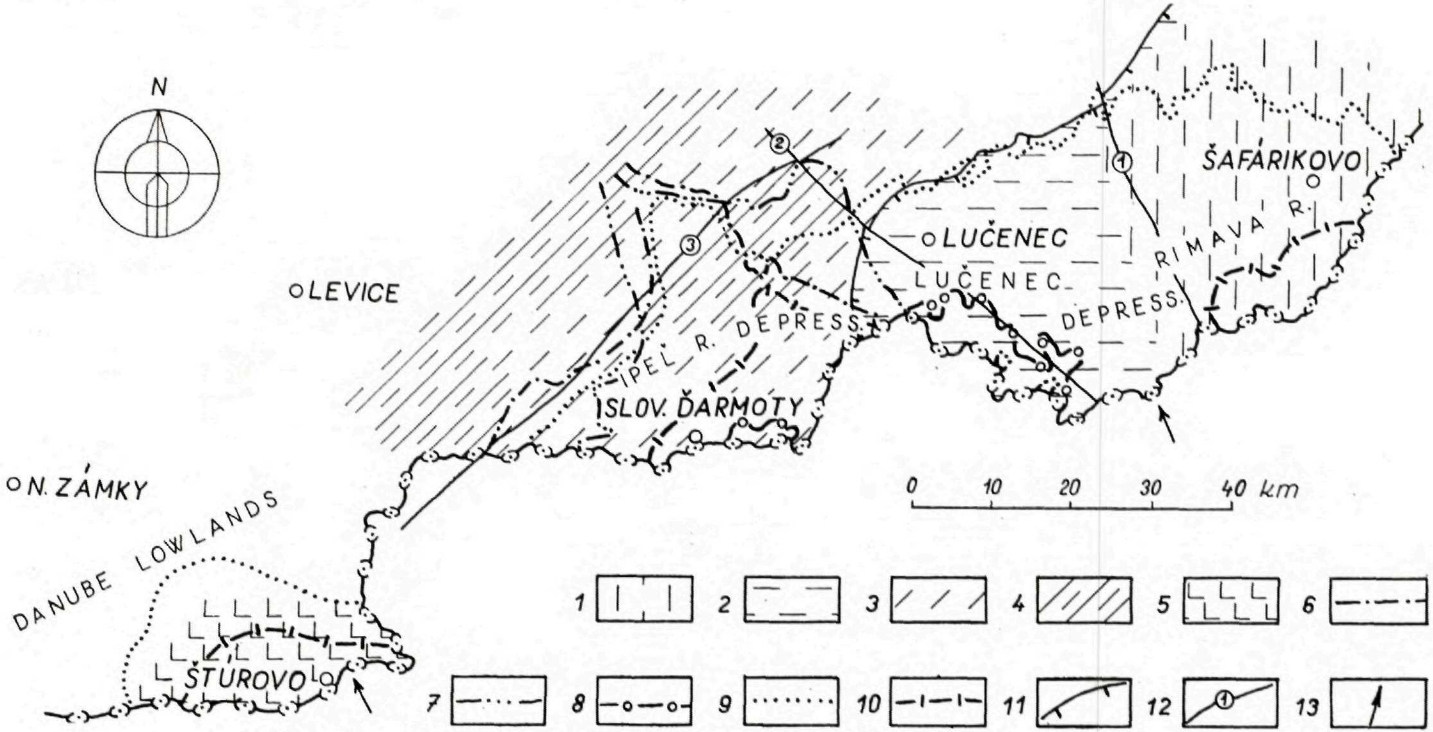
In the SE part of the Danube lowlands, in the Ipeľ and Rimava depressions, the substratum of the sandstone horizon consists of marly (predominantly 18 % $CaCO_3$) silty clays with Md=0.005 mm. They are characterized by micro-laminated parallel structure, stressed by parallel arrangement of micas. The clay ratio belongs to montmorillonite. Locally increased ratios of pyrite appears, indicating stagnant waters in local depressions.

In the Aquitanian—Chattian — in Čechovič's (1952) sense — the superimposed character of the basin was stressed by extension of transgression farther to the North. In that period the general uniform character of the subsidence was preserved in the basin, the influence of substratum being reflected mainly in various lithofacial development in the separate parts of the basin.

On the NE margin of the basin — in the Rimava depression — on the limestone substratum of Gemerides, there arose clastic limestones with the typical characteristics of littoral rocks of predominantly abrasive origin. There are present mostly the main types of littoral rocks, commencing with fragmentary to organodetritic types, with local occurrences of biotites.

The marginal development of the Aquitanian on the northern margin of the basin — in the Lučenec depression — is represented by the polymict conglomerate consisting of pebbles of metamorphic and granitoid rocks of Veporides, and of their metamorphosed Mesozoic mantle as well as of the Paleozoic sediments of Gemerides. As to its structure and degree of sorting — it belongs to the littoral deposits arising near the mountainous coast on the margin of the subsiding basin. In the deeper parts of the basin — in the Ipeľ depression — basal beds are present in the form of conglomerates with graywacke matrix — paraconglomerates — characterized by a very low degree of maturity of sediments, and by a high content of the unstable component. Predominantly they are composed of the Veporide crystalline.

Map I. Scheme of Extension of the Oligocene and the Neogene transgressions in the Ipeľ and Rimava Basins. Substratum of deposits of Ipeľ and Rimava Basins.



1. Mesozoic of Gemerides; 2. Paleozoic of Gemerides; 3. Crystalline rocks of Veporides; 4. Mantle Mesozoic of Veporides and the Choč nappe; 5. Rocks of Hungarian Central Mts. *Limitation of extension of Oligocene and Neogene sediments:* 6. Upper Helvetian (Carpathian); 7. Helvetian s. s.; 8. Lower Burdigalian; 9. Aquitanian (Chattian-Aquitanian); 10. Rupelian; 11. *Tectonic limitation between Gemerides and Veporides* (Lubeník and Margecany tectonic line); 12. *Dislocation belts:* a) Rimava dislocation belt, b) Lučenec dislocation belt, c) Muráň tectonic line; 13. General direction of transgressions.

Similarly differentiated is the composition of the sedimentary filling of the basin in the Aquitanian. In the Rimava depression, it is represented by the clayey silts (Fig. 1), in the Lučenec depression — by sandy silts and silty sands (Fig. 2) genetically corresponding to graywackes; — in the Ipeľ depression — by mixed clayey-sandy sequence (Fig. 3) with local occurrences of gravels, many-coloured clays, relicts of sedimentation near the river estuary.

The mineralogical composition of the heavy fraction of sediments is roughly equal. Only the ratio of the representation of garnets and micas is changed in dependence upon the granulometry (Figs. 4a—e).

The above lithofacial changes in the eastern, central and western parts may indicate commencement of differentiation of the basin, conditioned by the revival of movements on the Lučenec and Rimava zone of dislocations of the NW-SE directions. In consequence to the movements, the central part of the basin — its substratum being predominantly the Gemeride Paleozoic — has a shallower-water development in comparison with the eastern and western parts.

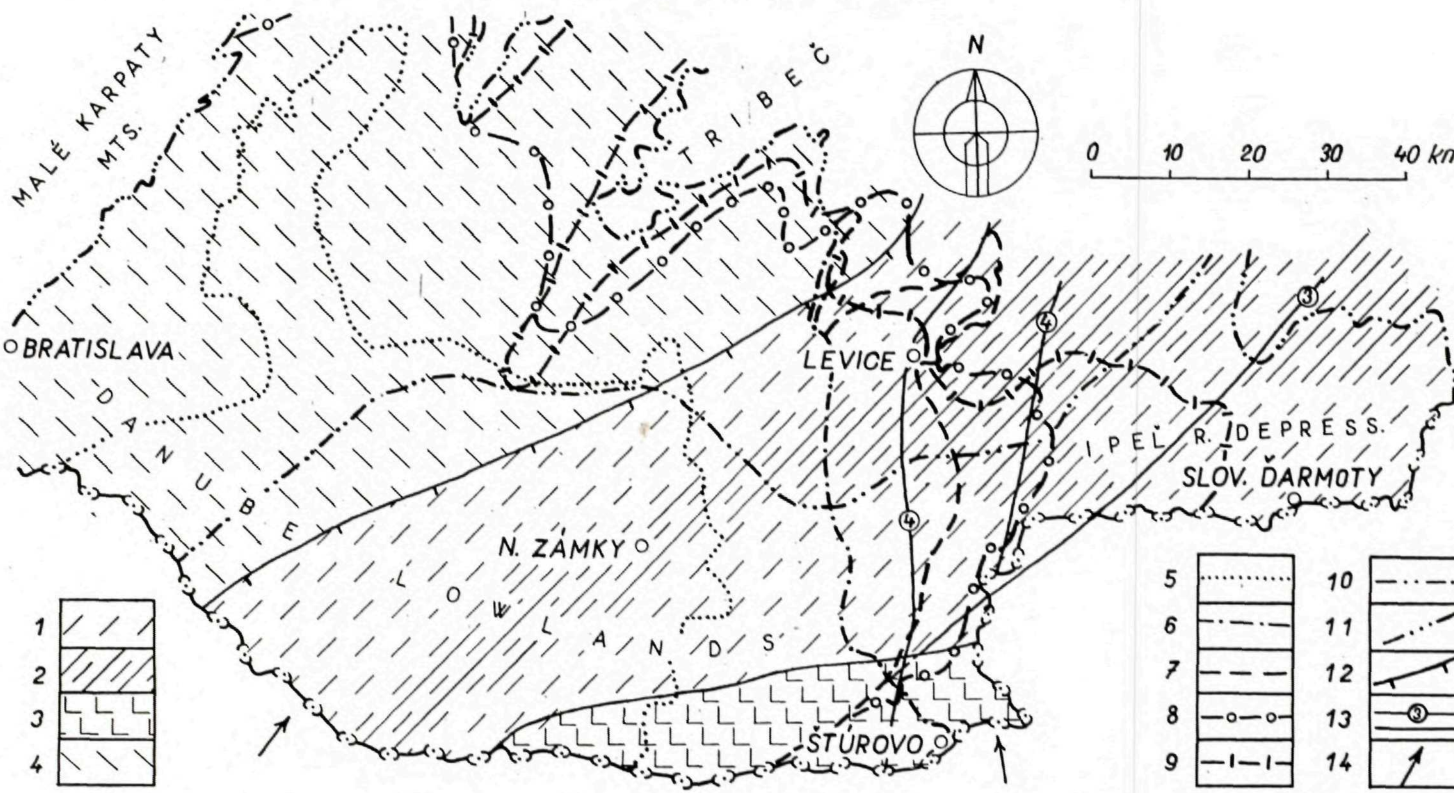
In the Burdigalian, the subsidence in the Ipeľ—Rimava basin was considerably lower — as to its extent and intensity— than in the Aquitanian. The subsidence affected only the central part of the basin between Fiľakovo and Slovenské Ďarmoty, and was evoked by the movements in the Lučenec dislocation zone. Larger subsidence took place in the eastern part of the area with the Gemeride Paleozoic predominating in the substratum. (The thickness of sediments is about 100 m here, while on the margin of the basin only 30 m).

Sediments of the Aquitanian in the Lučenec depression and in the southern part of the Ipeľ depression are covered by an accumulation of the marine Burdigalian, not too important as to its extent and thickness. In the Lučenec depression there are sandstone facies, facies of clayey-silty and silty-clayey sediments (Fig. 5). The green, coarse- to medium-grained sandstones with glauconite predominating, and with local occurrences of microconglomerate strata with the intercalations of tuffs, tuffites and silty-clayey sediments, are most frequent here. In all of these there is the volcanogene component in the form of feldspars, volcanic glass, pseudomorphoses α after β-quartz, biotites and zircons. Glauconite is richly represented; changes in accessory heavy minerals took place (Figs. 6a—c), andalusite, sillimanite and tremolite occur, and the granulometric coefficients of the sandstone facies were changed: Burdigalian: medium Md 0,18; medium So 1,9. — Aquitanian: Md 0,11; So 2,6.

The place of the facies in the genetic classification corresponds to the arkosic sandstones — opposite to the Aquitanian graywackes. This is for the first time that the beds of the wind-born vitric tuffs appear among the Burdigalian sediments in stagnant water zones.

The development of the Burdigalian in the Ipeľ depression has an in-

Map II. Scheme of Extension of the Neogene Transgressions in Danube Basin.



Substratum of the Neogene: 1. Crystalline rocks of Veporides; 2. The Mesozoic mantle of Veporides and the Choč nappe; 3. Rocks of Hungarian Central Mts; 4. Tatrides, in some places the Choč and Križná nappes. *Limitation of extension of the Neogene sediments:* 5. Levantian; 6. Pontian; 7. Panonian; 8. Sarmatian; 9. Upper Tortonian (Tortonian s. s.); 10. Lower Tortonian (Lanzendorf series); 11. Margin of basin; 12. *Tectonic limitation between Tatrides and Veporides.* 13. *Dislocation belts:* 3 — Muráň Tectonic line, 4 — Žiar and Levice dislocation belt. 14. General direction of transgressions.

shore character, being represented by the sandstone and microconglomeratic volcanogene-sedimentary facies.

At the end of the Burdigalian, on the inner side of the West Carpathians a regional upheaval took place (Seneš 1960, 1961) thus evoking temporary disappearance of the Ipeľ—Rimava basin. In Helvetian, the subsidence in the Ipeľ—Rimava basin was intense only in its western part, i. e. in the present-day Ipeľ depression, partly in the western part of the Lučenec depression. The eastern part of the Ipeľ — Rimava basin lost the character of a basin and became dry land. The substratum of the subsiding area is represented exclusively by the Veporide crystalline, in the Helvetian and Karpatian. With respect to subsidences of the Aquitanian and Burdigalian, distinct shifting of the subsidence to the West may be observed in the Helvetian and Karpatian. The Lučenec dislocation zone was important for the new configuration of the basin.

At the beginning of the Helvetian, the subsiding basin was filled with freshwater sediments, later on — in the Upper Helvetian s. s. the sea entered the basin and remained there until the Carpathian. The freshwater sediments commence with a productive sequence composed of the coal seams, of the carbonaceous montmorillonite — kaolinitic clays, quartz sands, and end with a 100—250 m thick stratum of overlying clays. The latter represents a monotonous facies composed predominantly of montmorillonite and illite, with the heavy fraction represented by authigene siderite in the lower part, and by pyrite in the upper part.

The freshwater sediments pass from the area with the substratum formed by Veporide crystalline massif to the south-western margin of the Gemeride Paleozoic. The marine sediments are concentrate on the Veporide substratum. They have been developed in three basic facies: a) fine-grained marly silts; b) non-marly „manganese" sands and marly sandstones; c) clayey marly silts. The fine-sandy silts (called Oncophora beds) were deposited in the quiet marine environment below the zone of wave action. They are characterized by a high content of mica (Fig. 7) originating in the epi- and mesozonally metamorphosed rocks of Veporides.

The overlying sandy sediments with a loose and non-marly lower part, and with solidified marly upper part, have the nature of well-sorted littoral sediments. Generally, their thickness is smaller than 100 m. Rare occurrences of conglomerate strata contain resistent Veporide material, with quartz, quartzite, chert and crystalline shales. Genetically, the sandstones belong to the arkosic sandstones. Marly sandstones contain 10—16 % of glauconite. Associations of heavy minerals are analogous in both facies (Fig. 8,9). In the source area, apoŗt from older sedimentary rocks and acid eruptives, metamorphosed rocks of the meso- and katazones were present also, and that might indicate the presence of tremolite (in diagram included with amphiboles), andalusite and sillimanite. The products of denudation of young volcanogene rocks represent the stable component.

In the Lower Tortonian — *the Lanzendorf series* — the maximum subsidence was in the E part of the Danube lowlands — the thickness of sediments reach up to 500 m here or even 1000 m, while in the Ipeľ depression only 100 m. Substratum of the subsiding part of the basin is formed mainly by the northern zones of Veporides, partly perhaps of Tatrides. In the area between Štúrovo and Komárno, the substratum is formed by Mesozoic of the intermontane region, and in some places of the central part of the lowlands — also by the Mesozoic of the Choč and Krížna nappes.

The basal beds of Lanzendorf series are exposed only in the Ipeľ depression, where the Veporide crystalline substratum is covered by sediments of earlier Miocene, and on the borderline of the Ipeľ depression and Danube lowlands, where the Tortonian is transgressing over the lower constituents of the mantle series of Veporides. The data about the composition of the basal beds in the Danube lowland are known from borings.

The predominating type of basal beds is represented by sands forming a continuous horizon, 30—40 m thick in the Ipeľ depression. They are predominantly horizontally bedded. There are local intercalations of organogene, algal limestones with the admixture of the volcanogene component passing into tuffites with calcareous cement, and the strata of light hyaline tuffs including amphibole and biotite. The presence of amphibole and rare pyroxene of neovolcanites is the main characteristic differentiating the association of heavy minerals of basal sands from the older sediments (Fig. 11). Feldspar content increases in the light fraction because of the presence of volcanogene component up to 30 %, the volcanogene β-quartz content being 5 %.

The conglomeratic facies of the basal Tortonian may be observed on the eastern margin of the Ipeľ depresssion and on the north-eastern margin of the Danube lowlands. Slavík (in Slavíková 1958) found quartzites, granitoides and crystalline shales in conglomerates of the eastern margin of the Ipeľ depression. On the NE margin of the Danube lowlands there are monomict quartzose conglomerates, gravels and sands, composed of the material of Veporides mantle series. These pass over into the marine sediments through a variegated intermediary continental-marine sequence. Farther southwards on the borderline between the Ipeľ depresssion and the lowlands, oligomict conglomerates of the littoral zone have been developed, including material of the Central-Hungarian Mts. i. e. limestones and cherts; pebbles of neovolcanites, andesites and their tuffs, in addition to the Veporide constituents (quartz, quartzite, quartzous sandstone and phyllite). The basal constituents of Lanzendorf series on the eastern margin of the Danube lowlands have been developed predominantly in the form of sands with close relation to the sedimentary rocks of their substratum. In some places, below the sands there are gravels composed of silicites, quartz, granites, quartzites, garnet mica-schists, arkoses, and sandstones transported from the West.

Between the basal beds and the proper sedimentary filling on the NE margin of the Danube lowlands, and in the Ipeľ depression there are large accumulations of andesitic pyroclastics and laminae of wirid transported volcanic glass. After the fading-out of volcanism, pelitic sediments of the Lanzendorfer series covered the Danube lowlands; they are, however, gradually wedging-in in the NE direction, and in the Ipeľ depression only littoral deposits of the resedimented, and eolian transported volcanic products may be found.

The medium granularity of pelitic facies is 0,004 mm in the basin area, increasing to 0,03 mm in the Ipeľ depression. Association of heavy minerals in the Ipeľ depression is a typical volcanogene one including hypersthene, augite and amphibole (Figs. 13,14). In the Danube lowlands the occurrence of these minerals is accessory, authigenic pyrite in the form of the cores of microorganisms (Fig. 12) showing the highest content. In both associations pelitic fractions are represented by montmorillonite with the lower content of illite. The source area in the period of the Lanzendorfer series in south Slovakia is represented by neovolcanite massifs.

In Upper Tortonian, the transgression passed into the western and north-western parts of the Danube basin becoming there a homogenous unit. Two tectogenetically different parts of the Danube basin were connected: the western part with its tectogenetic development common with the Vienna basin up to the Upper Tortonian, and the eastern part with independent development. The subsidence was affected here by movements along the faults of NE-SW or N-S directions, arising on the margins of the core mountains Tribeč, Inovec and Small Carpathians. The maximum subsidence passed farther west- and north-westwards of the Danube basin, where the thickness of sediments reached up to 3000 m. The substratum of this part of basin is formed predominantly by the Tatride elements or by the Krížna and Choč nappes. In the time of overthrust of subsidence to the West, the sea retreated from the Ipeľ depression to its westernmost part including the Upper Tortonian of about 1000 m thickness.

The lower part of the Upper Tortonian, the zone of agglutinations (Lehotajová in Seneš 1962) is not lithologically different from the pelites of the Lower Tortonian. At the end of the zone of agglutinations, due to the movements on the Žiar—Levice fault zone woking the uplift of the Ipeľ depression, extensive denudation of the southern margin of the central-Slovakian neovolcanites and deposition of sandy sediments on a large surface took place in the subsiding Danubian basin. The sequence is of the Flyschoid character and

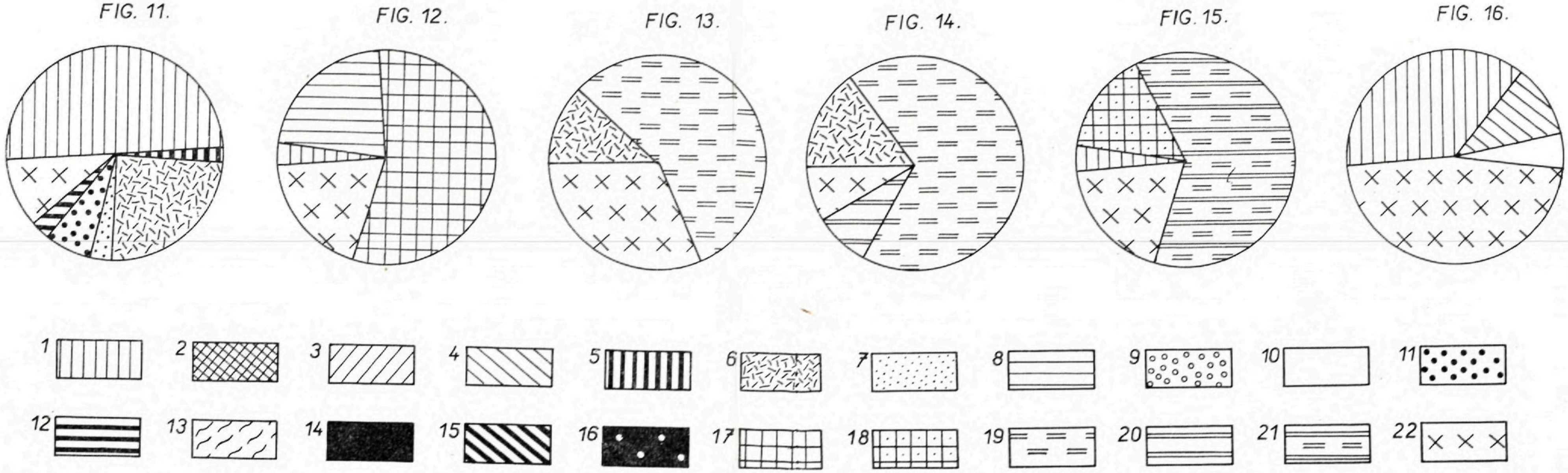


Fig. 1. The field (striped) of granularity of Aquitanian sediments in the Rimava depression.

Fig. 2. In the Lučenec depression.

Fig. 3. In the Ipeľ depression.

Fig. 4a. Heavy minerals (HM) of Aquitanian sediments in the Rimava depression. Accessories: Apatite, zircon, amphibole, biotite, staurolite, sphene, epidote-zoisite. The amount of HM (0,06—0,1 mm) 3,2 %. The ratio of opaque minerals (OM) 27 %, Md 0,1 mm, $CaCO_3$ 18 %.

Fig. 4b. HM of Aquitanian silts in the Lučenec depression. Accessories: Apatite, kyanite, tourmaline, rutile, sphene staurolite, carbonates. The amount of HM (0,1—0,25 mm) 1,9 %. The ratio of OM 34,3 %; Md 0,04 mm; So 2,6; $CaCO_3$ 0—20 %.

Fig. 4c. HM of Aquitanian sands in the Lučenec depression. Accessories: Apatite, cyanite, rutile, sphene epidote-zoisite. The amount of HM (0,1—0,25 mm) 4,2 %. The ratio of OM 15,5 %; Md 0,11; So 2,6; $CaCO_3$ 0—25 %.

Fig. 4d. HM of Aquitanian silts in the Ipeľ depression. The amount of HM 4,12; Md 0,01—0,04 mm; So 3; $CaCO_3$ 21,8.

Fig. 4e. HM of Aquitanian sands in the Ipeľ depression. Accessories: Rutile. The amount of HM 5,33; Md 0,07—0,2 mm; So 1,5—4; $CaCO_3$ 21,2 %.

Fig. 5. The fields — (striped) of granularity of Burdigalian sediments.

Fig. 6a. HM of Burdigalian sands. Accessories: Apatite, cyanite, zircon, tourmali, rutile, chlorites, phosphate, andalusite, carbonates. The amount of HM 2,1 %. The ratio of OM 23 %; Md 0,18 mm; So 1,9; $CaCO_3$ 0—15 %.

Fig. 6b. HM of Burdigalian clayey silts. Accessories: Apatite, cyanite, tourmaline, biotite, epidote-zoisite, phosphate. The amount of HM 19,2 %. The ratio of OM 19,2 %; Md 0,02 mm; So 2,1; $CaCO_3$ 20,7 %.

Fig. 6c. HM of Burdigalian sandy silts. Accessories; Zircon, amphibole, biotite, staurolite, epidote-zoizite. The amount of HM 1,1 %. The ratio of OM 13,5 %; Md 0,05 mm; So 3,9; $CaCO_3$ 13,4 %.

Fig. 7. HM of Helvetian s. s. sandy silts. Accessories: cyanite, zircon, amphibole, staurolite, anatase. The amount of HM 0,4 %. The ratio of OM 13,0 %; Md 0,06 mm; So 1,3; $CaCO_3$ 14 %.

Fig. 8. HM of Carpathian sands. Accessories: Apatite, cyanite, zircon, chlorites, sphene, phosphate, sillimanite. The amount of HM 1,9 %. The ratio of OM 50,6 %; Md 0,32 mm; So 1,6.

Fig. 9. HM of Carpathian marly sands. Accessories: Sphene, sillimanite. The amount of HM 1,4 %. The ratio of OM 24 %; Md 0,25 mm; So 1,99; $CaCO_3$ 7—24 %.

Fig. 10. HM of Carpathian clayey silts. Accessories: Apatite, cyanite, rutile, sphene, phosphate, sillimanite, carbonates. The amount of HM 1,5 %. The ratio of OM 24,8 %; Md 0,01 mm.

Fig. 11. HM of sands of Lanzendorf series in the Ipeľ depression. Accessories: cyanite, zircon, rutile, biotite, chlorite. The amount of HM 2,5 %. The ratio of OM 22 %; Md 0,28 mm; So 1,5; $CaCO_3$ 0,14 %.

Fig. 12. HM of clay of Lanzendorfer series in the Danube lowlands. Accessories: Apatite, zircon, tourmaline, amphibole, biotite, pyroxene. The amount of HM 6,75 %; Md 0,004 mm; So 2,8—4; $CaCO_3$ 5—35 %.

Fig. 13. HM of sands of Lanzendorf series in the Ipeľ depression. Accessories: Apatite, cyanite, zircon, tourmaline, chlorites, epidote-zoisite, phosphate, augite, andalusite, garnets, carbonates. The amount of HM 11,56 %. The ratio of OM 14,5 %; Md 0,24 mm; So 0,23.

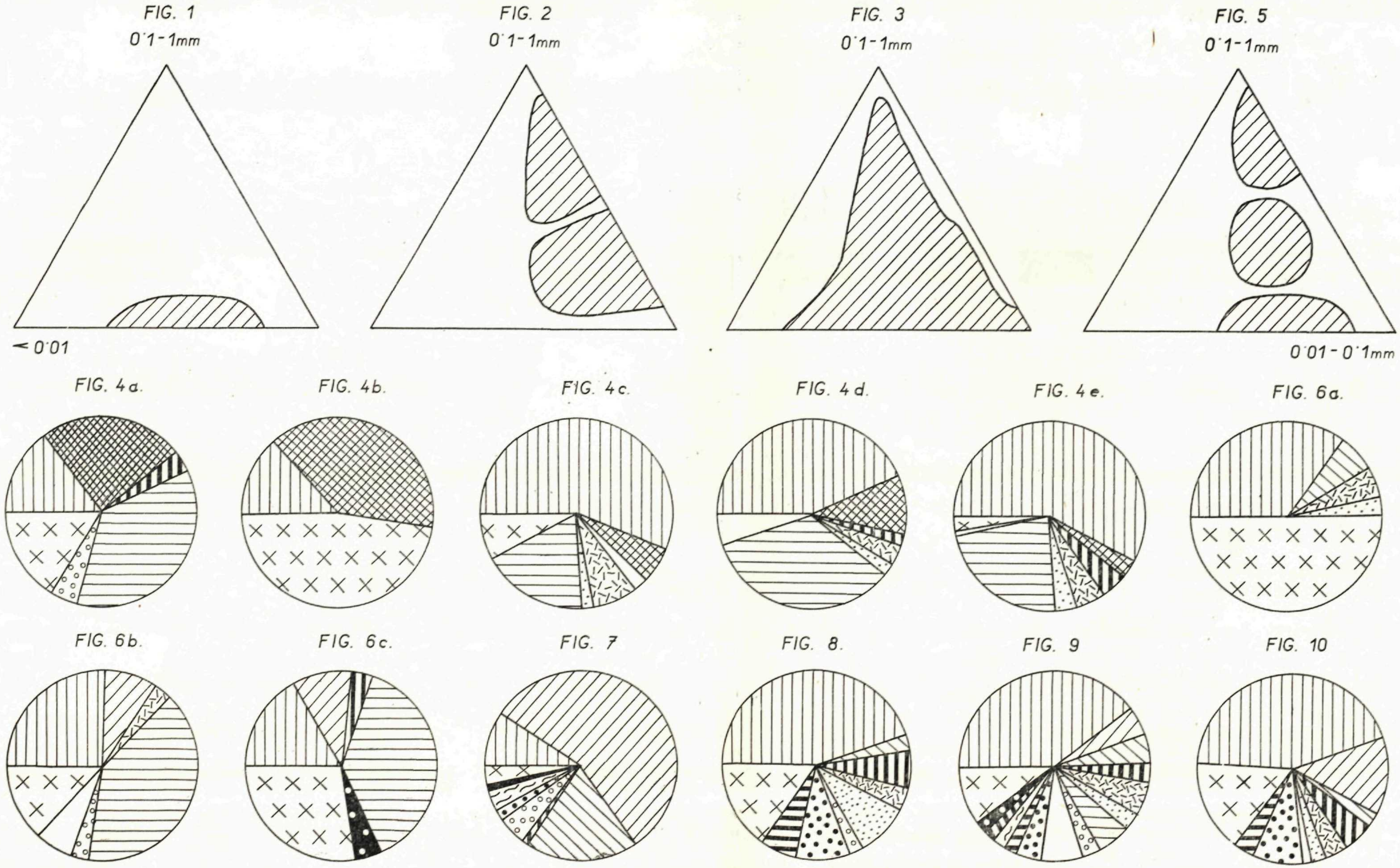
Fig. 14. HM of silts of Lanzenforfer series in Ipeľ depression. Accessories: Apatite, cyanite, zircon, tourmaline, staurolite, chlorites, phosphate, andalusite, garnet. The amount of HM 7,8 %. The ratio of OM 12,6 %; Md 0,024 mm; $CaCO_3$ 17,5 %.

Fig. 15. HM of sands of the Upper Tortonian zone of agglutinations. Accessories: Amphibole, biotite, ilmenite, magnetite. The amount of HM 13,75 %; Md 0,3 mm; So 1,4.

Fig. 16. HM of silts of Upper Tortonian the Bulimina-Bolivina zone. Accessories: tourmalin, amphibole, rutile, chlorites, staurolite, epidote-zoisite, pyroxene, carbonates. The amount of HM 1,2 %. The ratio of OM 48,6 %; Md 0,02 mm; So 3,1; $CaCO_3$ 5 %.

1. garnet; 2. micas; 3. chlorite; 4. biotite; 5. tourmaline; 6. amphibole; 7. staurolite; 8. carbonates; 9. rutile; 10. zircon; 11. epidote-zoisite; 12. andalusite; 13. cyanite; 14. sphene; 15. apatite; 16. phosphate; 17. pyrite; 18. ilmenite; 19. hypersthene; 20. augite; 21. pyroxene; 22. accessories.

Encl. 1. **Granulometric and haevy minerals (HM) diagram of the Neogene deposits**



consists of sandstone and clayey-silty strata, their mutual ratio increasing in behalf of sands, in eastern direction. There are numerous conglomerate strata with the washout contact with substratum. The material is almost exclusively volcanogene (Fig. 15), mineralogical composition is vertically and horizontally comparatively stable, graded-bedding being present there, too.

In the overlying sequence there are present clayey silts, stratigraphically belonging to the Bolivina-Bulimina zone (Lehotayová in Seneš 1962). In basal part, on the eastern margin of the basin, there are conglomerate strata containing as much as 20 % of the pebbles of quartz, quartzites, mica-schists phyllites and granitoides in the predominating mass of andesite pebbles. Lithological parameters of the Upper-Tortonian silts facies are conformable with those of the recent sediments of the inland seas. The facies is characterized by an abundant terriginous component (variegated association of heavy minerals Fig. 15), by the equilibrated ratio of the clastic and volcanogene components in the light fractions, abundant plant detritus). The regression of the Upper-Tortonian sea indicates definite retreat of marine sedimentation from the Ipeľ depression (from the area that was originally a part of the Ipeľ—Rimava basin).

The reduction of the subsiding area on the NE margin of the Danube basin is due to movements on the Žiar—Levice dislocation zone of the N-S direction. To a certain extent the faults of the NW-SE direction were revived.

In the Sarmatian no substantial replacement of subsidence in comparison with the Upper Tortonian took place, only its intensity lessened. In the eastern part of the basin movements continued along the Žiar—Levice fault zone owing to which the Sarmatian sediments were limited practically to the Danube lowlands. The Sarmatian sediments are predominantly of the brackish-marine origin. Their maximum thickness is about 100 m in the Danube lowlands. The marginal facies outcrop on the foothill of the Small Carpathians and on the eastern margin of the Danube lowlands.

In the foothill of the Malé Karpaty Mts. (Small Carpathians) there are mainly medium- to coarse-grained massive calcareous sandstones with isolated strata of polymict gravels, quartz predominating.

On the eastern part, gravels, conglomerates, sands, sandstones and oolitic limestones occur. The pebble material consists of quartz, quartzites, lydites, cherts, Mesozoic limestones and andesites. In the NE part of the Danube lowlands the Sarmatian consists mainly of the rhyolite and andesite pyroclastics. In the basin part slightly silty marly to non-marly clays, in some places variegated, have been developed. They alternate with the strata of fine-grained clayey-silty micaceous sands. The regression Sarmatian sediments include variegated clays, sands, with isolated occurrences of carbonaceous clays and lignite.

In comparison to the Sarmatian, in the Pliocene brackish-marine and lacustrine facies indicates that the subsidence of the Danube basin was intensified; the maxima of subsidence were, however, replaced farther southwards. In the eastern part of the basin the margin of subsidence retreated westwards, owing to — similar to the end of the Upper Tortonian and Sarmatian — the movements along the Žiar—Levice fault zone completed with the movements along the faults of NW direction.

The thickest Pannonian s. s. was found in the SW part of the Danube lowlands. The oldest Pannonian sediments in the area under study have been developed only in the maximally subsiding part of the basin, in the facies of marly clays, sands and gravels, in the north — only gravels. Above these there rest sands and sandstones covered by the sequence of silty-sandy marly clays in the whole basin, and lignites — mainly on the margins of the basin.

In the Pontian of the Danube lowlands the subsident area was not different from that of the Pannonian, only its intensity increased. The thickest Pontian in the SW part of the Danube lowlands is about 1500 m (T. Buday 1962). It is a sequence of fine-grained sands alternating with clays and locally with gravels, especially on the NW margin passing to gravels and freshwater limestones. In the upper part of the Pontian there are coal seams and marlaceous clays.

In the Uppermost Pliocene (perhaps in the earlier Quaternary, too) subsidence was fading-out. It is indicated by the lessening of the surface extension of the subsiding area and by considerable decrease of thickness of sediments (max. 100—130 m). It is a prevalently gravel and sandy sequence with clay strata. The thickness of clays increas eastwards.

In the eastern part of the Ipeľ—Rimava basin, local depressions with moderate subsidence accompanied with limnic-lacustrine sedimentation of clays, sands and gravels, arose at the end of the Pliocene. The subsidence was essentially limited to the Gemeride substratum. As to the surface, the subsidence was quite extensive, but — as indicated by thickness of the Pliocene sediments — general value of subsidence was low and lagged behind subsidence in the Danube lowlands.

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On the ground of this it may be stated that the dependence of development of the Tertiary basins on the inner side of the West-Carpathian arc upon the structure of substratum has been reflected:

[1] in migration of subsidence in the course of development of basins from one unit of substratum to the other, regularly from the East to the West;

[2] in the fact, that migration was not caused by old Alpine tectonic lines separating tectonic units of the substratum, but by the revival of older dislo-

cation zones of the NW-SE direction, or by revived or newly arisen dislocation zones of the N-S to NE-SW directions, in some places stressed by marginal faults of some core mountains;

[3] owing to the migration of subsidence, of various depths of sedimentary environment and changes of source area (degree of denudation, volcanism, lithological character of the filling of basins changed during the development of the Neogene)

[4] The superimposed basins were characterized by greatest planar extension in the initial stages of their development: the Ipeľ—Rimava basin — in the Aquitanian, the Danube basin — in the Upper Tortonian. In further development their planar extension decreased.

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Geologické práce, Zprávy 44—45. Bratislava 1968

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DER PERM-MESOZOISCHE VULKANISMUS IN DEN WESTKARPATEN UND OSTALPEN

Abstract. In diesem Artikel werden die neuesten Forschungsergebnisse der permischen und mesozoischen Eruptivgesteine der Westkarpaten und deren Beziehung zu ähnlichen vulkanischen Produkten in den Ostalpen behandelt. In den westkarpatischen Raum sind die permomesozoischen Vulkanite der äusseren und inneren (zentralen) Westkarpaten, incl. des anliegenden ungarischen Mesozoikums (die sog. Rudabánya-Entwicklung und Bükk-Gebirge) einbezogen. Der Beitrag stützt sich auf eigene Ergebnisse des Autors aus den zentralen Westkarpaten, wie auch auf Arbeiten anderer Forscher über die Westkarpaten und Ostalpen, vor allem: G. C. Amstutz (1954), F. Angel (1929), K. Balogh & G. Pantó (1953) E. Bederke (1959), F. Becke (1912), A. Biely (1960—66), C. Burri & P. Niggli (1945), H.P. Cornelius (1929—1949), V. Čechovič (1948), O. Fusán (1962, 1963), L. Hauser (1940, 1941, 1942) F. Heritsch (1932), M. Ivanov (1956—1962), J. Kamenický (1950—1958), J. Kamenický — L. Kamenický (1955), J. Kantor (1955, 1956), F. Kossmat (1937), A. Köhler & A. Marchet (1939), M. Kuthan (1959), G. Pantó (1951, 1952, 1953), A. Pilger (1940), A. Pilger — R. Schönenberg (1957, 1958), W.P. de Roevere (1941—1959), J. Salaj (1962), M. Slavkay (1966), G. Steimann (1926), Sz. Szentpétery (1918—1953), V. Šťastný (1927) K. Urban (1934), L. Waldmann (1926), V. Zorkovský (1949—1959).

Die permische Etappe

Die neoide — alpidische Geosynklinale hat sich in den Westkarpaten aus der variszischen Molasse entwickelt, die durch den saueren subsequenten Vulkanismus im Perm im nördlichen Teil des Zips-Gömörer Erzgebirges (Ivanov 1953) und im westlichen Teil der Veporiden (Zoubek 1956) charakterisiert ist. Beim heutigen Stand unserer Kenntnisse ist die präzise stratigraphische (zeitliche) Spannweite dieser Permentwicklung sehr schwer festzulegen. Aus dem erworbenen Material aus beiden Gebieten lässt sich schliessen, dass der saure Vulkanismus vor allem an tiefere Lagen der Permschichtfolge gebunden ist. Im nördlichen Teil des Zips-Gömörer Erzgebirges ist er älter als die Evaporitfazies, die etwa dem (?) Oberperm- Werfen (Schukow 1961; Rojkovič 1964, 1967; Mahel' & Vozár 1966) angehört. In den Veporiden ist der subsequente Vulkanismus an die Zeit der Sedimentation der basalen arkosenartigen Schichtfolge gebunden (in ihrem Untergrund lagert das Veporiden-Kristallin, im Hangenden die höheren Permschichten der bunten Verrucanofazies, auf welche die untertriassischen Quarzite transgredierten). Im südlichen Teil des Zips-Gömörer Erzgebirges bleibt die Lage der Diabasgesteine der Meliata-Serie, die zu jüngerem Paläozoikum (Kuthan 1947; Čekalová 1954) gestellt wird, ungeklärt. Diese vulkanischen Gesteine besitzen nicht überall — im Verhältnis zu den anliegenden Ablagerungen — den syngenetischen Charakter; aber wenigstens einen Teil davon kann man mit dem triassischen (J. Kamenický 1957), event. jüngeren Vulkanismus (beschrieben im weiteren Text) dieses Gebirges in Zusammenhang bringen.

Das Perm-Alter kann mit Sicherheit im basalen Abschnitt der Choč-Decke den Produkten des basischen bis mittelbasischen Vulkanismus im jungpaläozoischen sedimentär-vulkanischen Komplex — der oft auch als Melaphyrenserie, oder vorher Werfen mit Melaphyren (vergleiche auch Andrusov 1959) bezeichnet wird — zugesprochen werden. Dieser Vulkanismus ist mit der Permschichtfolge dieses Komplexes synsedimentär und im Sinne der neuesten stratigraphischen Gliederung der Choč-Decke (Biely 1961—1966) ist er eindeutig jünger als Karbon und älter als die Untertrias. Räumlich ist er in mehreren Gebirgen der zentralen Westkarpaten als ein Bestandteil der überschobenen Massen der Choč-Decke (deren ursprünglicher Sedimentationsraum, dessen Position und Ausmass trotz neuen mühsamen Detailforschungen noch immer diskutabel ist) ausgedehnt. Der Vulkanismus verlief in diesem Sedimentationsraum im subaqualen Milieu und hatte einen linienartigen Charakter. Der Aufstieg der Magmen verlief entlang einiger langen (stellenweise unterbrochenen), tief eingelegten Bruchlinien, deren Bildung etwa in die Perm-Zeit fällt. Die Entstehung dieser Linien hängt offensichtlich mit der Ausbildung des Sedimentationsraumes, der im Sinne de Roevere's (1959) das *Embryonalstadium* der alpinen Geosynklinalentwicklung erreicht hatte. In diesem Sinne kann man auch den vorausgesetzten Bruchlinien und dem mit ihnen verbundenen Magmatizmus den embryonalen Charakter zuerkennen.

Die vulkanische Tätigkeit im Choč-Sedimentationsraum weist einen pulsativen Charakter auf und konzentrierte sich in die bisher bekannten, bzw. erwiesenen zwei Eruptionsphasen; jede von ihnen ist durch eigenartige hydrothermale Phase und sie begleitende pyroklastische Ablagerungen charakterisiert und aus wenigstens zwei Ausgüssen bestehend. Die Produkte des permischen Vulkanismus gehören der Gruppe der normal gabbroiden, Gabbrodiorit-, weniger den Dioritmagmen (im Sinne der P. Niggli' schen Klassifikation 1936), mit relativ höherem Alkaliengehalt bei den Produkten der ersteren Eruption. Die Ergussaequivalente sind durch verschiedene Abarten der Melaphyre, Melaphyrporphyrite und Porphyrite vertreten. Der Charakter der pyroklastischen Ablagerungen (eine bunte Skala der Tuffe, Tuffite, tuffitischer Ablagerungen und Sedimente mit pyroklastischer Beimischung) weist auf eine verhältnismässig schwache Explosivität der Eruptionen hin. In den tieferen Basalabschnitten der Permschichtfolge und im liegenden Karbonkomplex hat man gangartige oberflächennahe Aequivalente dieser Effusivgesteine festgestellt (Šťastný 1927), die durch Diorit- und Gabbrodioritporphyrite vertreten sind. Die Überreste der Karbonschichtfolge zusammen mit Gängen der erwähnten vulkanischen Gesteine wurden auch isoliert von den hangenden perm-mesozoischen Massen der Choč-Decke, vor allem unterhalb des Muráň-Mesozoikums im Slowakischen Erzgebirge, unterhalb des karbonatischen Komplexes des Vernár-Streifens im östlichen Teil der Niederen Tatra und im Gebirge Čierna Hora im tektonischen Hangenden der älteren Gebilde festgestellt (auf diese Funde hat mich A. Biely und O. Fusán aufmerksam gemacht). Die Beziehung dieser Vorkommen zur Choč-Decke wurde in der Vergangenheit nicht untersucht. Die petrographische Verwandschaft der vulkanischen Gesteine und Art und Weise ihres Auftretens rechtfertigen ihre Eingliederung zur Choč-Decke, wodurch auch die Angaben über die Ausdehnung dieser Decke weiter präzisiert werden.

Wie bereits erwähnt, verlief diese vulkanische Tätigkeit während der Ablagerung der ganzen Permschichtfolge der Choč-Decke. Die ältesten Effusivkörper wurden unmittelbar oberhalb der Grenze mit Karbon, die jüngsten unterhalb der untertriassischen Quarzite festgestellt; dieser Vulkanismus endete vor Ablagerung der untertriassischen Quarzite (Biely 1965).

Die Untersuchung der Beziehung des permischen Vulkanismus im Choč-Sedimentationsraum und der Vulkanite des nordgemeriden und veporiden Perms zeigt gewisse Unterschiede in der stratigraphischen Spannweite, im Charakter des Sedimentationsbeckens und in der petrographischen und chemischen Zusammensetzung der vulkanischen Gesteine dieser Gebiete. Die saueren Gesteine des subsequenten permischen Vulkanismus aus dem Zips-Gömörer Erzgebirge ähneln durch ihren chemischen und petrographischen Charakter den aequivalenten vulkanischen Produkten im westlichen Teil der Veporiden (die permische Hülle der Ľubietová-Zone und Gangkörper der Quarzporphyre im Kristallin der Ľubietová-Zone), unterscheiden sich aber von permischen Vulkaniten der Choč-Decke. Die Quarzporphyre des nordgemeriden Perm gehören in die Gruppe der Granit-, weniger der Granodioritmagmen (im Sinne Niggli's 1936). Ähnliche vulkanischen Gesteine im westlichen Teil der Veporiden gehören überwiegend zur Granodioritmagma. Eine selbständige Gruppe vertreten die wahrscheinlich basischeren Abarten der Quarzporphyre im nordgemeriden Perm — die Porphyrite und Quarzporphyrite (Ivanov 1953; Rojkovič 1964, 1967), die ihrem chemischen Charakter gemäss der Gruppe der Dioritmagmen gehören. Als gemeinsames Glied des Gemeriden- und Choč-Perm könnte man eben die erwähnten Porphyrite — vor allem von chemischer Sicht aus bezeichnen; aber ihr petrographischer Charakter und Art und Weise des Auftretens ist verschieden.

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Auf den Charakter der Ausgangsmagmen der permischen vulkanischen Gesteine der Choč-Decke habe ich bereits oben hingewiesen. Auf Grund der bisherigen Studien bin ich der Meinung, dass der permische Vulkanismus der Choč-Decke einen anderen Charakter hat als jener in den Gemeriden; seine ebenbürtigen Aequivalente kann man in den Ostalpen nicht suchen.

In den südlichen Kalkalpen, im Raume der Lienzer Dolomite, in den Karnischen Alpen, bei Plos, Brixen, St. Peter sind im Unterperm Quarzporphyre und Augitporphyre bekannt. Auch der sog. Trotsburger Melaphyr mit Pyroklastiken und wenigstens ein Teil der Vulkanite der Bozener Porphyrmassen gehören dem Perm an. Aber alle diesen vulkanischen Gesteine werden meistens zum subsequenten, vereinzelt auch finalen variszischen Vulkanismus gestellt (Schaffer 1951).

In den Helvetiden, namentlich in der permischen Verrucano-Fazies der Mürtschener Decke im Raume Glarner-Freiberg hat Amstutz (1954) Quarzporphyre, Keratophyre und Spilite untersucht. Aus den uns zugänglichen Unterlagen lässt sich sagen, dass lediglich die Spilite durch ihren Charakter mit einigen feinkörnigen, stärker spilitisierten Melaphyren bis Spilit-Melaphyren des Perm in der Choč-Decke der Westkarpaten verglichen werden können.

Nach Bederke (1959) könnte man wenigstens einen Teil dieser permischen Vulkanite in den Alpen in die Etappe des sog. embryonalen Magmatismus (im Sinne de Roevere's 1959) eingliedern, ähnlich wie den permischen Vulkanismus der Choč-Decke in den Westkarpaten.

Über die Stellung des permischen Vulkanismus in den Ostalpen und Westkarpaten herrscht derzeit keine eindeutige Ansicht. Dieses Problem wird noch durch die verschiedene stratigraphische Beurteilung der Schichtfolgen mit vulkanischen Gesteinen kompliziert. Aber in beiden Gebirgskomplexen hat man Produkte der aciden (Granit-, Granodiorit-), intermediaren und basischeren (Diorit-, Gabbrodiorit-, Gabbro-) Magmen festgestellt. Es ist daher anzunehmen, dass eben die acideren vulkanischen Gesteine in den Westkarpaten zum subsequenten variszischen Vulkanismus gestellt werden können. Da hängt es vom Charakter des Sedimentationsmilieu, in dem die vulkanische Tätigkeit vor sich ging und von deren Beziehung zum variszischen Orogen und zu den Initialäusserungen der neoiden Geosynklinale ab, zu welchem Orogen dieser oder jener vulkanische Komplex gestellt werden kann, bzw. welches seiner Stadien er vertritt. Im Sedi-

Gesteinsvorkommen des ophiolitischen Vulkanismus in den Ostalpen (nach C. Burri & P. Niggli 1945)

mentationsraum der Choč-Decke in den Westkarpaten konnte es zur solchen Situation kommen, dass die magmatische Entwicklung erst nach der tektonischen folgte; auf Grund dessen ist der permische Vulkanismus (der bereits vor der Transgression der Untertrias statfand) zum alpidischen orogenen Zyklus zu stellen, wie darauf in den Alpen Bederke (1959) hinweist.

Die mesozoische Etappe

Die mesozoische Etappe des alpidischen Sedimentationsraumes in den Westkarpaten beginnt mit der Transgression der untertriassischen Quarzite (wahrscheinlich des Seis-Alters), die in den Hüllenserien und im Subtatrikum den Ausgleich der Fazien gebracht hatte. In der letzten Zeit hat dieses Problem Biely (1960—1966) studiert. In den bisher bekannten Trias- und Juravorkommen (mit Ausnahmen des Tithon) ist das totale Fehlen der Eruptivgesteine in den Hüllenserien und subtatrischen Decken (Krížna- und Choč-Decke), wie auch in der Klippenzone festgestellt worden. Die Ergebnisse aus den letzten Jahren bestätigten die Existenz der Diskordanz zwischen dem permischen vulkanisch-sedimentären Komplex einerseits und den Trias-Quarziten in der Choč-Decke andererseits.

Von Sicht der vulkanischen Entwicklung aus hat das Mesozoikum in der internen westkarpatischen Einheit — in den Gemeriden und zu ihnen gestellten überschobenen mesozoischen Komplexen (Muráň-Mesozoikum, Drienok-Mesozoikum; Bystrický 1964), wie auch die Rudabánya-Entwicklung und Bükk-Gebirge in Ungarn einen ganz verschiedenen Charakter als die erwähnten Hüllen und subtatrischen Serien.

Das Gemeriden-Mesozoikum (wie es oben definiert wurde), das Mesozoikum der Rudabánya-Entwicklung und Bükk-Gebirge ist vor allem durch intensive vulkanische Tätigkeit in der Unter- und Mitteltrias gekennzeichnet. Problematisch erscheint die stratigraphische Stellung einiger Vorkommen der Quarzporphyre im Liegenden des Muráň-Mesozoikums. Sie können untertriassisch aber auch älter sein (V. Zorkovský 1959). Diese Quarzporphyre, begleitet vom pyroklastischen Material, gehören in die Gruppe der Granitmagmen und durch ihre petrographische Charakteristik erinnern sie an einige aciden Effusivgesteine der mesozoischen Drienok-Einheit, deren Obercampil-Alter faunistisch belegt ist (Slavkay 1965). Aus dem mir zugänglichen Material bin ich der Meinung, dass als Aequivalente dieser aciden, seltener auch mittelaciden

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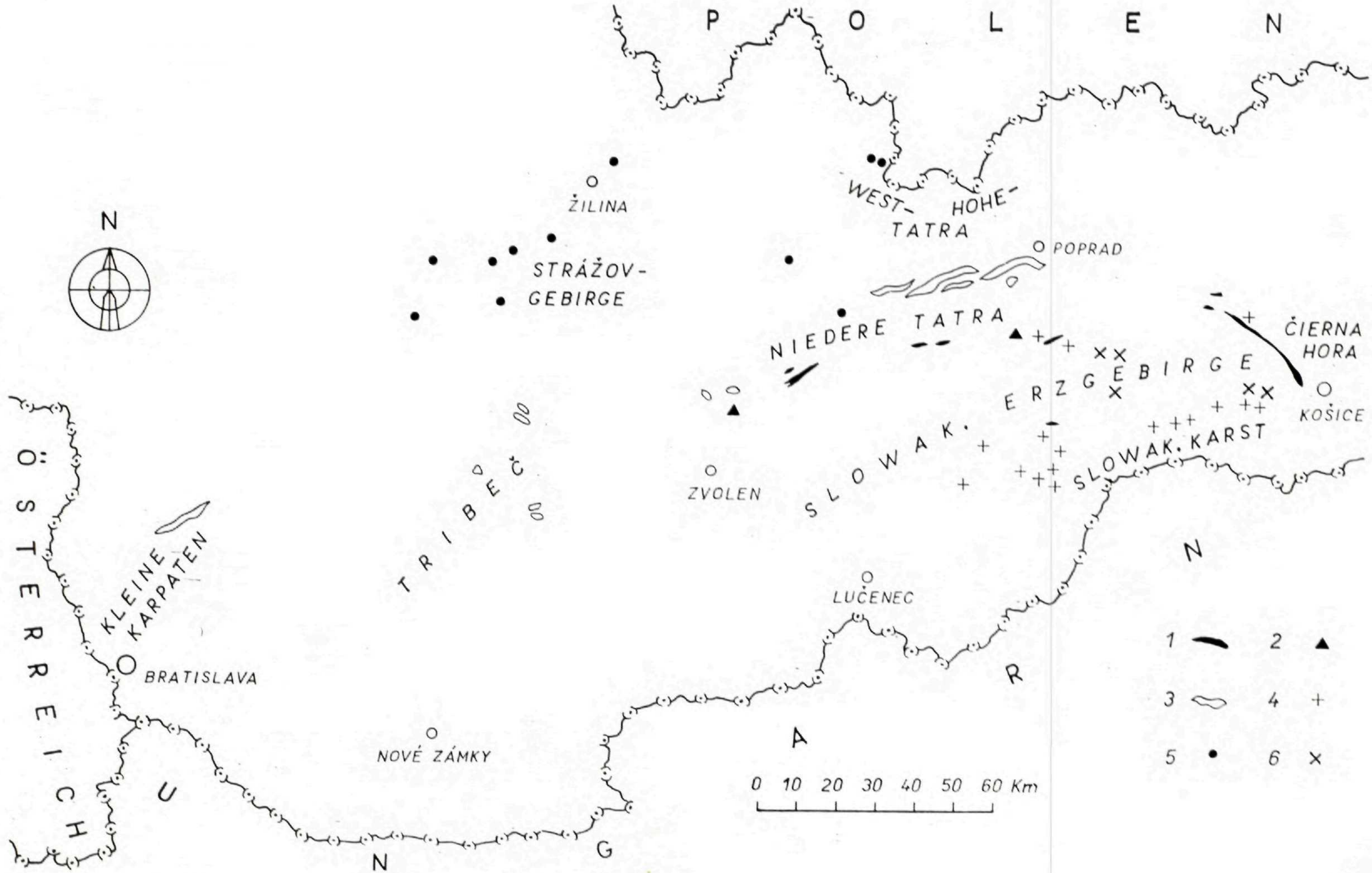
Gesteinsvorkommen des perm- mesozoischen Vulkanismus in den Westkarpaten (zusammengestellt nach der geologischen Übersichtskarte der ČSSR 1:200 000)

Erklärungen: der variszische orogene Zyklus: 1 — Produkte des subsequenten Vulkanismus — Perm; 2 — Produkte des saueren Vulkanismus (problematische Stellung — Perm - Untertrias). Der alpidische orogene Zyklus: 3 — Gesteine des embryonalen Magmatismus — Perm; Gesteine des Initialvulkanismus — Trias-Unterkreide; 5 — Gesteine des Initialvulkanismus — Tithon-Neokom; 6 — synorogene Magmagesteine — Oberkreide.



P O L E N
ŽILINA
WEST-
HOHE-
TATRA
POPRAD
STRÁŽOV-
GEBIRGE
NIEDERE TATRA
ČIERNA
HORA
SLOWAK. ERZGEBIRGE
KOŠICE
SLOWAK. KARST
ZVOLEN
TRIBEČ
KLEINE KARPATEN
ÖSTERREICH
BRATISLAVA
LUČENEC
NOVÉ ZÁMKY
U N G A R N
0 10 20 30 40 50 60 Km
1
2
3
4
5
6

Eruptivgesteine wenigstens teilweise (stratigraphisch und petrographisch) einige Vorkommen ähnlicher Gesteine in den südlichen Kalkalpen und der mesozoischen Decken der Centralalpen (in beiden Gebieten der sog. Bozener Porphyrkomplex — Perm-Untertrias) gehalten werden können. In den centralen Westkarpaten erinnern sie (von chemischer Sicht aus) an permische Quarzporphyre des Zips-Gömörer Erzgebirges, die wir bereits erwähnt haben

Unklar bleibt die Beziehung dieser permischen Quarzporphyre und der permuntertriassischen aciden Eruptivgesteine im Muráň-Mesozoikum bzw. in der Drienok-Einheit. Sie gewinnen an Bedeutung, falls wir ihre genetische Ver wandschaft mit dem zips-gömörer Mesozoikum erwägen; auf Grund dessen drängt sich die mögliche magmatische Verwandschaft der permischen und permuntertriassischen Eruptivgesteine in den Fordergrund. Leider, beim heutigen Stand unserer Kenntnisse kann man in dieser Frage keine eindeutige Stellung nehmen. Es wird aber die Möglichkeit nicht ausgeschlossen, dass die permuntertriassischen Eruptivgesteine der Drienok-Einheit und des mesozoischen Muráň-Plateau in die letzte, vielleicht verspätete (gewissermassen auch verlängerte) Phase des permischen Vulkanismus im Zips-Gömörer Erzgebirge fallen. Bei solcher Annahme dürfte die stratigraphische Spannweite dieses Vulkanismus etwa jener des Bozener Porphyrkomplexes entsprechen.

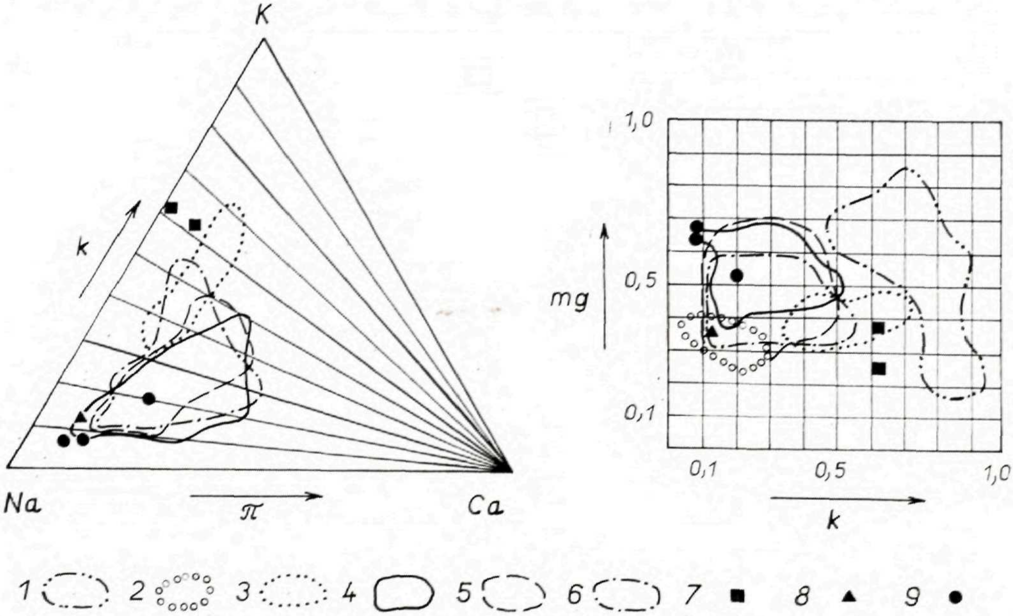
Den basischen Vulkanismus im zips-gömörer Mesozoikum hält man derzeit (J. Kamenický 1950—1957; J. Kantor 1955—1956) für Unter- bis Mitteltrias, mit initialem Charakter und in Beziehung zu den Alpen entspricht er dem ofiolitischen Vulkanismus. Wegen mangelnde Vertretung der jüngeren mesozoischen Glieder in den Gemeriden kann seine obere Zeitgrenze nicht festgesetzt werden. Es ist höchstwahrscheinlich, dass er mit den vulkanischen Äusserungen im anliegenden ungarischen Mesozoikum äquivalent ist (Rudabánya-Entwicklung, Bükk-Gebirge; Zs. Szentpétery 1923—1953; K. Balogh — G. Pantó 1953), wo der Vulkanismus nachweisbar zur Untertrias bis Unterkreide — ähnlich wie die bereits erwähnten Ophiolite der Alpen, event. auch der Dinariden (C. Burri — P. Niggli 1945) und die Mezzoeruptiva in Siebenbürgen (Zs. Szentpétery 1918) — gestellt wird. Allen erwähnten vulkanischen Äusserungen wird der initiale Charakter zugesprochen. Sie sind meistens basisch bis ultrabasisch, aber auch saurere Abarten werden nicht ausgeschlossen. Es handelt sich um Effusivgesteine mit begleitendem pyroklastischen Material und um Intrusiva, echte Gänge, Steilgänge, Daiken und Necken.

Bei den Intrusivformen ist das Alter sehr schwer feststellbar, genau so wie im Zipsgömörer Mesozoikum. Während das pyroklastische Material, das die vulkanische Tätigkeit in der Unter- und Mitteltrias bezeugt, nur an wenigen Fundorten bekannt ist, finden wir oft Beweise der Kontaktmetamorphose der erhaltenen Triasablagerungen durch die Intrusivformen. Aus dem Studium der Beziehungen zwischen dem Gemeriden-Mesozoikum und jenem des anliegenden ungarischen Raumes lässt sich schliessen, dass wenigstens ein Teil der Intrusiva das Obertrias- bis Unterkreide-Alter haben kann. Die stratigraphische Spannweite dieser vulkanischen Gesteine ist vorläufig noch nicht eindeutig geklärt; als einziges brauchbares Kriterium dabei ist die Parallelisierung mit den stratigraphisch identifizierten Gebieten.

Von der petrochemischen Sicht aus haben J. Kamenický (1950—1957), und

J. Kantor (1955, 1956) neben vereinzelten Vorkommen der pyroklastischen Gesteine auch verschiedene Abarten von Diabasen, Glaukophaniten und Serpentiniten festgestellt. Sie vermuten, dass es sich um äquivalente Produkte des ophiolitischen Vulkanismus in Alpen handelt. Die Entstehung der Glaukophanite verbinden sie mit der Umwandlung der Diabasen und Serpentinite mit den Peridot-Ausgangsmagmen. Diese vulkanischen Gesteine im gemeriden Mesozoikum und im anliegenden ungarischen mesozoischen Raume sind – mit Ausnahme der tithon-neokomischen Magmagesteine in den subtatrischen und Hüllenserien, bzw. in der Klippenzone – einzige ausgeprägtere Äusserungen des initialen Vulkanismus in der neoiden westkarpatischen Geosynklinale.

Im Vergleich mit basischen und ultrabasischen mesozoischen Eruptivgesteinen der Gemeriden sind ähnliche Gesteine in der Rudabánya-Entwicklung und im Bükk-Gebirge petrographisch viel bunter. Neben den basichen Abarten gabbroiden bis Gabbrodiorit-Charakters (Diabase, Diabasporphyrite, Porphyrite, Gabbro, Gabbroporphyrit), oft mit erhöhtem Alkaliengehalt (vor allem in den unterkretazischen Eruptivgesteinen – Natriumgabbro, Essexite, Spilite) erscheinen da auch Gänge der Eruptivgesteine aciden Charakters (Granitmagmen), die man im Sinne der ungarischen Autoren K. Balogh – G. Pantó 1953) mit dem synorogenen kretazischen Magmatismus verbinden kann.



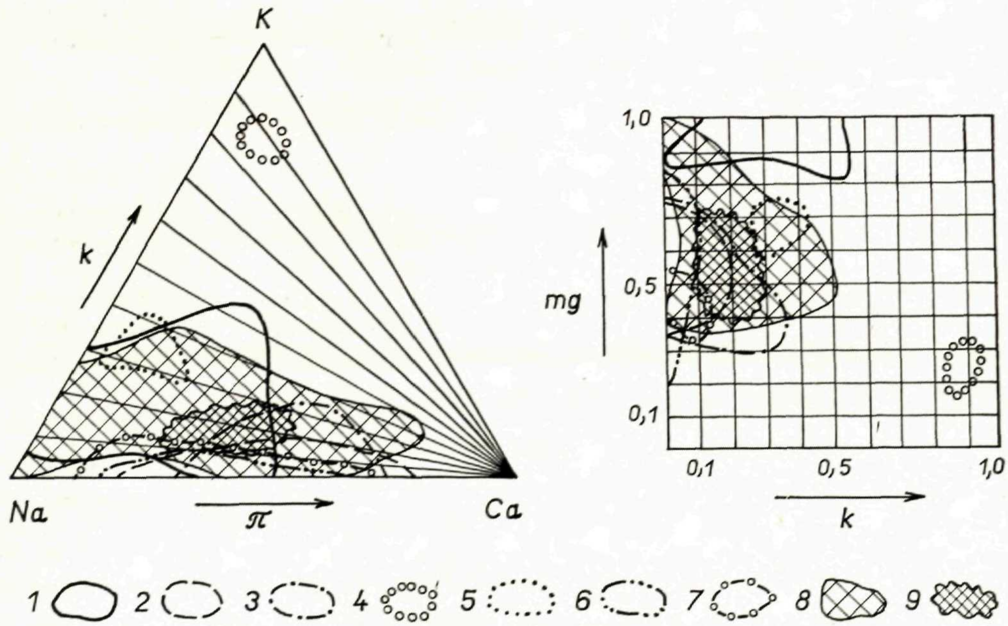
Der chemische Charakter der jungpaläozoischen Eruptivgesteine

Der subsequente variszische Vulkanismus im Perm des Zips-Gömörer Erzgebirges: 1 – Quarzporphyre (Ivanov 1953: 11 Analysen; Rojkovič 1965: 11 Analysen); 2 – Porphyrite, Quarzporphyrite (Ivanov 1953: 1 Analyse; Rojkovič 1965: 3 Analysen). *Der subsequente variszische Vulkanismus im Perm der Ľubietová-Zone (Veporiden):* 3 – Quarzporphyr (Vozár 1962: 3 Analysen); *Der Embryonalmagmatismus im Perm der Choč-Decke:* 4 – Melaphyre und Porphyrite in den Kleinen Karpaten (Zorkovský 1949: 3 Analysen; Vozár 1965: 20 Analysen); 5 – Melaphyre, Porphyrite und Dioritporphyrite an den Nordhängen der Niederen Tatra (Zorkovský 1949: 5 Analysen; Vozár 1964–67: 60 Analysen); 6 – Melaphyre, Porphyrite und Dioritporphyrite der Südhänge der Niederen Tatra, des Tribeč-Gebirges, des Oberneutra-Tales und der Insel von Sklené Teplice (Zorkovský 1949: 10 Analysen; Vozár 1965–67: 10 Analysen, Slavkay 1965: 1 Analyse). *Vulkanische Gesteine im Perm der Mürtschauer Decke im Raume Glarner-Freiberg:* 7 – Quarzporphyre, 8 – Keratophyre, 9 – Spilite (alles nach G. C. Amstutz 1954).

Von der ziemlich intensiven, wenn auch weit entfernten vulkanischen Tätigkeit im Raume des Bükk-Gebirges zeugen auch einige Funde der pyroklastischen Ablagerungen im Ladin der Südgemeriden (Kuthan 1959) und in den anis-ladinischen Karbonaten im Nordgemeriden Raume, im Galmus-Gebirge (Biely 1966).

*

Nun möchte ich noch einmal zur Frage der Eruptivgesteine in der Meliata-Serie im südlichen Zips-Gömörer Erzgebirge zurückkommen. Diese Eruptiva bilden oft Gangkörper und kontaktmetamorphieren die anliegenden Ablagerungen. Mit Rücksicht auf ihre auffallende Verwandschaft mit mesozoischen Diabasen des Bükk- und Zips-Gömörer Gebirges (Südslowakischer Karst) kann bei ihnen das Trias-, event. auch jüngeres Alter nicht ausgeschlossen werden. Wie erwähnt, auf die Verwandschaft des mesozoischen Vulkanismus der Gemeriden mit dem ophiolitischen Vulkanismus anderer Gebirge im mediterranen Raume haben bereits die älteren Arbeiten hingewiesen (J. Kamenický 1950—1957; J. Kantor 1954, 1955).



Der chemische Charakter der mesozoischen Eruptivgesteine

Der alpidische Initialvulkanismus in der Trias des Zips-Gömörer Erzgebirges: 1 — Serpentinit (J. Kamenický 1957: 6 Analysen; Kantor 1955: 15 Analysen); 2 — Diabas (Kantor 1955: 2 Analysen; J. Kamenický 1957: 8 Analysen; Zorkovský 1960: 2 Analysen); 3 — Glaukophanit (J. Kamenický: 1957: 10 Analysen). *Die Eruptivgesteine der Perm (?) — Untertrias-Zeit in der Drienok-Einheit und im Muráň-Mesozoikum:* 4 — Quarzporphyr (Zorkovský 1959: 2 Analysen; Slavkay 1965: 12 Analysen; Vozár 1964—67: 6 Analysen). *Der alpidische Initialvulkanismus im Mesozoikum des Bükk-Gebirges (Ungarn):* 5 — Gabbro, Diabias mit erhöhtem Alkaliengehalt (Zs. Szentpétery 1951: 6 Analysen); 6 — Diabas (Szentpétery 1953: 16 Analysen). *Eruptiva in der Meliata-Serie:* 7 — Diabas (Kantor 1955: 6 Analysen). *Der alpidische Initialvulkanismus:* 8 — Ophiolite der Westalpen (C. Burri & P. Niggli 1945: 55 Analysen); 9 — Ophiolite der Ostalpen (C. Burri & P. Niggli 1945: 10 Analysen).

Die vulkanische Tätigkeit im Mesozoikum der Ostalpen* kam in den südlichen Kalkalpen am reichsten zur Geltung. Wenn wir aus der mesozoischen Etappe den Bozener Porphyrkomplex ausschliessen, dann bleibt die stratigraphische Spannweite des Initialvulkanismus Obercampil, event. Anis bis Lias. Der bei einigen Vorkommen der Eruptivgesteine angenommene spätere Ursprung konnte paläontologisch eindeutig nicht bestätigt werden. Die vulkanische Tätigkeit der südlichen Kalkalpen kulminiert in der Ladinzeit und knüpft an jene Tätigkeit aus dem Anis, event. auch Campil an, bzw. setzt sich bis in die Obertrias (Karn-Nor) fort. In dieser Phase des mesozoischen Vulkanismus dominieren basische Eruptivgesteine — Aequivalente der Gabbro-, Gabbrodiorit-, bzw. auch (in geringerem Masse) Dioritmagmen. Oftmals konnte man auch Körper der ultrabasischen Gesteine finden. Neben den eindeutigen Effusivgesteinen treten auch Intrusivkörper mit der Kontaktmetamorphose der anliegenden Ablagerungen auf.

Aus diesem Raum werden von verschiedenen Forschern: Gabbro, Pyroxenit, Peridotit, Gabbroporphyrit, Diabas, Melaphyr, Diabasporphyrit, Dioritporphyrit, Porphyrit, Quarzporphyrit, Kersantit, Monchiquit, Essexit, nephelinischer Syenit, Syenit, Bostonit, Camptonit u. ä. erwähnt. Aus dieser Aufzählung ist der mannigfaltige petrographische Charakter hiesiger Vulkanite ganz klar. Manche, vor allem Intrusivgesteine sind einigen Autoren zufolge jünger (Jura- Kreide). Von petrographischer Sicht aus sind diese vulkanischen Gesteine den entsprechenden mesozoischen Eruptivgesteinen der Westalpen ähnlich; C. Burri & P. Niggli (1945) betrachten sie als Produkte des ophiolitischen Initialvulkanismus.

Die jüngere vulkanische Phase im Mesozoikum der südlichen Kalkalpen ist durch die porphyritähnlichen rhaet-liassischen Eruptiva vertreten, die nur vereinzelt zu finden sind und den Dioritmagmen angehören. Die ihnen entsprechenden Gesteine wurden auch in den Dinariden festgestellt.

Im Vergleich mit den südlichen Kalkalpen sind die mesozoischen Decken der Zentralalpen an vulkanische Gesteine viel ärmer. Die paläontologisch belegte Vertikalverbreitung der vulkanischen Gesteine ist Untertrias bis Oberkreide, mit Optimum in der Unter- und Mitteltrias. Zahlreiche Vorkommen der Werfener Diabas- und Gabbrogesteine sind in jenem Teil der mesozoischen Decken in den Zentralalpen erhalten geblieben, der sich an den Nordrand der südlichen Kalkalpen anlehnt (Schaffer 1951). Die Anis- und Ladinschichten dieser Decken sind reich an pyroklastisches Material, das von Cornelius (1941) mit der intensiven vulkanischen Tätigkeit in den südlichen Kalkalpen in Zusammenhang gebracht wird. In den Kreideschichten der mesozoischen Decken der Zentralalpen wurden Ehrwaldit- (austrische Phase) und Porphyritfunde (laramische Phase) gemacht.

Zeitmässig konzentrierte sich die vulkanische Tätigkeit der Nordalpen in die Unter- und Mitteltrias-Zeit; spätere Aktivität konnte nicht erwiesen werden. Neben den Effusivkörpern, die durch pyroklastisches Material begleitet sind, hat man auch Gangintrusiva gefunden. Petrographisch wurden identifiziert: Melaphyr, Diabas, spilitisierter Diabas, Spilit, Gabbroporphyrit und Gabbro, alles den Gabbromagmen gehörend. Stellenweise fand man auch alkalische Eruptivgesteine. Im Vergleich mit den Westalpen ist in den Ostalpen der jura-kretazische Vulkanismus viel seltener vertreten, bzw. bei einigen

* Bearbeitet nach Cornelius (1929—1949), Burri-Niggli (1945) und Schaffer (1951).

Vorkommen (vor allem bei Intrusivkörpern) konnte das nachtriassische Alter nicht bewiesen werden.

Der Vulkanismus *im Tithon-Neokom der Westkarpaten* ist als Ergebnis der Labilität (Unbeständigkeit) der alpinen Geosynklinale noch vor ihrer Ausfaltung zu betrachten. Diese Tätigkeit kam in den tektonisch am stärksten exponierten Gebieten des Sedimentationsraumes zur Geltung; als Aufstiegswege der Eruptivgeistene dienten aktive Bruchlinien. An das Unterkreide-, bzw. auch Tithonalter aller dieser Eruptiva in den centralen Westkarpaten (Hüllenserien und subtatrische Serien) und in der Klippenzone kann man aus deren alpinen Richtung und syngenetischer Beziehung gegenüber den anliegenden Sedimenten (Effusivkörper, begleitet durch das pyroklastische und paläontologisch belegte Material) schliessen. Diese tithon-neokomische Phase trägt einen initialen Charakter und ist zusammen mit den älteren mesozoischen Eruptivgesteinen des ophiolitischen Vulkanismus den anderen Gebirgsmassiven des mediterranen Raumes ebenbürtig.

Aus den bisher bekannten Vorkommen der unterkretazischen vulkanischen Produkte in den Westkarpaten lässt sich ihre beschränkte territoriale Verbreitung vermuten. Dabei muss man berücksichtigen, dass in vielen Gebieten die Kreide-Ablagerungen nur arm erhalten sind; andererseits ist es möglich, dass auch ein Teil der Gangkörper inmitten der älteren Gebilde dem tithonneokomischen vulkanischen Zyklus gehören kann. Petrographisch handelt es sich um pyroxenische Gesteine vom Gang- und Effusivcharakter (Augitit, augitischer Porphyrit, augitische mandelförmige Körper; K. Urban 1934; V. Zorkovský 1949; J. Salaj 1962). Pyroklastisches Material findet sich nahe der Ergüsse, aber auch isoliert, nicht zusammenhängende Lagen in den meistens unterkretazischen Sedimenten bildend.

In der internsten westkarpatischen Einheit — in den Gemeriden ist der kretazische synorogene Magmatismus durch den Aufstieg der oberkretazischen Granite (J. Kamenický & L. Kamenický 1955; J. Kantor 1959) gekennzeichnet. Die Granitintrusionen können mit der Bildung der Quarzporphyre und Granitporphyre im Mesozoikum des Bükk-Gebirges (Ungarn) in Zusammenhang gebracht werden, die K. Balogh & G. Pantó (1953) für Vertreter des kretazischen synorogenen Magmatismus halten. Den bisherigen Kenntnissen zufolge kann man die Vorkommen der Gemeriden-Granite für einzige Vertreter des synorogenen Magmatizmus bei uns halten. Zwecks möglicher Parallelisierung möchte ich noch erwähnen, dass die synorogenen Granite in den ostalpinen mesozoischen Decken dem Paleogen-Miozän angehören (Schaffer 1951).

Zusammenfassung

Den perm-mesozoischen Vulkanismus darf man nicht isoliert, sondern im Zusammenhang mit dem älteren — subsequenten variszischen und dem jüngeren subsequenten — alpinen Vulkanismus beurteilen. Wie wir es in diesem Artikel zu zeigen versuchten, kann man den basischen permischen Vulkanismus in der Choč-Decke als Ergebnis des embryonalen Magmatismus betrachten, der der eigentlichen tektonischen Entwicklung der alpinen Geosyklinale — also auch den alpinen Initialvulkaniten, den Ophioliten, zu welchen man alle trias-unterkretazischen Eruptivgesteine eingliedern kann, vorausging. Problematisch ist die Stellung der untertriassischen Quarzporphyre, die den variszischen Gesteinen des subsequenten Vulkanismus chemisch nahe verwandt sind. Mit Hilfe der Parallelisierung mit den ostalpinen Eruptivgesteinen konnte die bisherige Ansicht über die Ähnlichkeit der triassischen und teilweise auch jüngeren vulkanischen Tätigkeit in beiden Regionen bestätigt werden. Über die mögliche Parallelisierung des permischen Vulkanismus in den Ostalpen und Westkarpaten bestehen derzeit verschiedene Ansichten.

Zum Schluss möchte ich auf diesem Wege meinem Lehrer Herrn Prof. Dr. M. Kuthan für die fachmännische Führung und Beurteilung der Handschrift den aufrichtigen Dank aussprechen.

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VLASTIMIL KONEČNÝ—MIROSLAV KUTHAN

VOLCANIC FORMS AND TYPES OF ERUPTIONS OF NEOGENE VOLCANISM IN CENTRAL SLOVAKIA

INTRODUCTION

The development of the subsequent and final volcanism in the Tertiary and Quaternary periods in Slovakia was connected with the later evolutionary stages of the Carpathian orogene. Volcanic activity took place on the outer side of the orogene, still the full development of volcanism was concentrated first of all in the inner parts of the orogenetic arch.

The origin of magmatic mass of the subsequent volcanism characterized by the rhyolite andesite-basalt association is connected with the palingene character of magma, derived from the later differentiation stages of granitoid mass mobilized in the syntectonic phases of the orogene process (Kuthan 1967). The most distinct characteristics of the subsequent volcanism is the high explosiveness of eruptions. This specific feature of the subsequent volcanism is connected with the palingene origin of magma (Bemmelen 1949). The character of high explosiveness follows the absolute predominance of pyroclastic mass over the effusive. In regional aspect the ratio is about 9 : 1.

The development of volcanic forms, and types of eruptions are conditioned by widely differentiated spectra of material composition, and by its dependance upon spatial and chronological migration of the eruption centres; this being a reflexion of the complicated and variable tectonomagmatic position. In the manifold row of volcanic forms the most important are the forms of stratovolcanic type, characteristic of the most frequent type of volcanism of the andesite intermediary nature, especially its complex forms, i. e. the group volcanoes (multiple volcanoes in the ssnse of Cotton 1959).

The development of these forms is considerably modified by the environment of its origin, changed — in consequence of tectonic mobility of extensive regions — in dependance upon the oscillation of the shore-line of the Tortonian-Sarmatian sea, from the typically terrestrial to subaqueous environment with the development of brackish to lacustrine basins.

In the periods of the retardation of volcanic activity, characterized by predominance of destructive processes, extensive denudation of volcanic forms and redeposition of material into adjacent areas or into the place of the near continental intravolcanic basins, took place.

Products of the younger stages of activity were deposited on volcanic terrains distinctly modelled by denudation.

The subsequent volcanism was followed by the volcanism of the final stage in the sense of H. Stille (1953), in the time of consolidation of regions, when the gravity unequilibrium inherited from the preceding stages of mobility, was balanced by epeirogenetic movements accompanied with the rise of extensive deep faults. Along these faults slightly differentiated and contaminated magma was ascending from the deep-seated sources. In the sense of Van Bemmelen (1949, 1950), the origin of the magma is derived from the alkalinebasaltoid magma, partly contaminated by the assimilation of dry sialic mass, in the upper parts of the crust.

In difference from the subsequent volcanism, the products of the final volcanism are characterized by comparatively narrow differentiation spectrum, by lower explosiveness of eruptions with dominant activity of effusive processes. The forming of extensive basalt sheets is distinctly analogous to the plateau basalts (trappes). The rise of the scoria cones of predominantly monogene type, and of the scoria cones with lava effusions formed by the activity of the Stromboli type, indicate sporadic explosive activity.

CLASSIFICATION OF VOLCANIC FORMS AND TYPES OF ERUPTIONS

The development of the palingene magma and volcanic phenomena connected with the former, are evidently conditioned by its spatial tectonic position in relation to the orogene. The optimal conditions for the differentiation development of the palingene magma are offered by the environment of the simple hydrostatic pressure. This condition is fully provided in the inner part of the orogene, i. e. in the area of the substitution of tangential forces by epeirogenetic movements with the fault tectonic of the germanotype character. Due to this, volcanic activity in the area of the outer arch was less distinct, represented mainly by the dike and neck bodies accumulated in relatively narrow zones closely connected with tectonic systems (NW and N margins of the Carpathian arc — the Javorina, Lopeník and Polish Pieniny Mts.) These bodies do not show any distinct morphologic shape, neither do they form independent geological-orogenetic complexes.

The inner zone maximally corresponding to the conditions of the orogenetic back-land, is an area of widely developed subsequent and final volcanism with morphologic-orogenetic relicts of developed volcanic forms. The general direction of the development of magmatic mass is reflected in the basic succession scheme of the eruptives of the Neogene volcanism: rhyolite → andesite

→ basalt. In spite of the distinctness of this tendency, the development of the subsequent volcanism is far from being continuous: the interruption of the evolutionary cycle, and repeating of some segments are quite frequent. Owing to this, in the resulting scheme the repeating of the manifestations of rhyolite, intermediary-andesite and acid andesitic volcanism may be observed (Cf. the scheme of the succession of eruptives; Tab. 1).

The general analysis of forms and types of the separate stages of eruptions (phases) with approximately corresponding basic parameters of material composition (Ist, IInd, IIIrd rhyolite phases, Ist, IIIrd andesite phases, IInd, IVth andesite phases) shows the analogy of the main volcanic phenomena.

Basing upon this, the following types of volcanic forms and eruptions have been determined:

(a) Volcanism of the subsequent stage:

1. *Volcanism of the intermediary andesite to basalt-andesite* type; 2. *Volcanism of the more acid andesite, andesite-dacite and dacite type*; 3. *Volcanism of rhyodacite to rhyolite type*;

(b) Volcanism of the final stage: *basalt volcanism.*

Within the separate types of volcanism, types of eruptions and volcanic forms have been specified (Cf. Tab. 2—5).

Tab. 1. **Succession of Eruptives of the Neogene Volcanism in Central Slovakia**

	Petrographic type of volcanic products	Stratigraphic position	Volcanic phases	Index
Final volcanism	final basalts	Quaternary Upper Pliocene	Main phase of final volcanism	B_2
	basaltoid andesites, andesitoid basalts	Pliocene	Preparatory phase of final volcanism	B_1
Subsequent volcanism	rhyolites, rhyodacites	Sarmatian	IIIrd rhyolite phase	R_3
	pyroxene andesites augite, hypersthene	Sarmatian	IVth andesite phase	A_4
	amphibole			
	biotite			
	coarse porphyric andesites, biotite-amphibole andesites, dacitoid andesites, biotite-amphibole	Sarmatian	IIIrd andesite phase	A_3
	pyroxenic andesites ± amphibole, ± biotite, ± olivine)	Sarmatian Upper Tortonian	IInd andesite phase	A_2
	rhyolites, rhyodacites, dacites	Middle Tortonian	IInd rhyolite phase	R_2
	amphibole-pyroxene andesites ± garnet, amphibole-biotite andesites ± garnet	Lower Tortonian	Ist andesite phase	A_1
	rhyolites, rhyodacites	Burdigalian	Ist rhyolite phase	R_1

According to M. Kuthan 1967; J. Forgáč, K. Karolus, E. Karolusová, V. Konečný, M. Kuthan (in press).

The above classification schemes have not the character of the resultative schemes and definitions; they only represent systematic arrangement and evaluation of the existing determinations and definitions of volcanic phenomena. This standpoint admits further modification and complementing of these aspects.

VOLCANISM OF SUBSEQUENT STAGE

1. Volcanism of the intermediary andesite to basalto-andesite type

In the process of the development of subsequent volcanism, the volcanism of the intermediate to basalt-andesite volcanism was active in a wide regional area. In relation to other types of volcanism determined, it was a highly dominant phenomenon as far as the amount of erupted mass is concerned. Within the development of the subsequent volcanism in central Slovakia, two stages of this type of volcanism have been determined:

The first stage follows the closed acid andesite volcanism (Ist phase — Lower Tortonian). The commencement of the intermediate volcanism of andesite type (IInd andesite phase, A_2) are distinctly determined in the SW, and mainly in the eastern parts of the area, where the volcanic products are resting upon the peneplanized relief of the volcanic terrains of the Ist andesite phase, after the intra-Tortonian denudation stage (Konečný 1962, 1965; Vass 1964). The development of the basal complexes of the andesite volcanism of the IInd phase took place in the marginal sublittoral to littoral zones of the Tortonian sea, and in the adjacent continental areas in brackish to lacustrine basins. The commencement of this activity was in the bulimine-bolivine zone — the Upper Tortonian (J. Seneš 1962).

Previously, the succesion of proofs about the existence of the second stage of the andesite volcanism (pyroxenic andesites), following the closed andesite to andesite-dacitic and dacitic volcanism of the IIIrd andesite phase, in the succession scale, were collected. The second stage of the pyroxenic andesite volcanism has got a preliminary name, i. e. that of the andesite volcanism A_4 (Forgáč — Karolusová — Karolus — Konečný — Kuthan, in press).

Quite a row of authors brought the proofs of the activity of the andesite volcanism spatially and chronologically situated into the overlier of the IIIrd andesite phase:

F. Fiala (1933) was the first to quote pyroxenic andesite from the area of Kozelník and Hronská Breznica.

J. Šalát (1953) described pyroxenic andesite (andesine, diopside) from the western part of the central-Slovakian region, from the vicinity of Nová Baňa.

K. Karolus — E. Karolusová (1964) also presented leucocratic andesite. Some authors found pyroxenic andesite in the Štiavnické pohorie Mts., in the overlier of the IIIrd andesite phase: M. Čajková (1956) — diopside andesite and hypersthene, diopside andesite with olivine; M. Harman (1956) — hyaline pyroxene andesite. Burian (1965) described pyroxene andesite $\pm$ biotite (the Sitno type). L. Rozložník,

Tab. 2. **Volcanism of subsequent stage**

1. Volcanism of intermediate andesite to basalt-andesite type

			Volcanic forms	Types of eruptions	Products of volcanism	V. P.	S.
I.	Surface forms	in terrestrial environment	1. Stratovolcanic forms (polygene volcanoes) 2. multiple volcanoes	volcanoe type of eruptions Pelée type (nuées ardentes) Vesuv type (effusive-explosive)	pumicaous ash tuffs, chaotic agglomerates pumicaous ash tuffs, lavas	II, IV	A_2, A_4
		in subaqueous environment	3. Planar accumulations of lava flows and hyaloclastite breccias	non-explosive subaquaeous volcanism in Rittman's sense (1960)	lava flows of small thickness (2 – 8 m), hyaloclastite breccias	II	A_2
II.	Subvolcanic forms		4. Neck bodies	intrusion along tectonic directions (crossed and overthrust zones) penetrations explosive in the last stage	andesite breccia	II	A_2
			5. Dike, vein bodies	intrusion along tectonic lines	andesite	II, IV	A_2, A_4

V. P. = volcanic phase; S = symbol

J. Šalát (1966) quoted several types of pyroxene andesites from the area of Hodruša, J. Forgáč (1966) – pyroxene andesite from the northern part of the Štiavnické pohorie Mts. A. Brlay—V. Konečný (1966—67) described hypersthene andesite with sporadic augite and with hyaline development of the matrix from the southern part of the Štiavnické pohorie Mts., then hypersthene augite, augite-hypersthene ± brown amphibole, and hypersthenic amphibole, andesite.

The above quoted data concern the areas of the distinct superposition relation with respect to the products of the IIIrd andesite phase, developed mainly in the form of tuffs and agglomerates; further the areas offering no possibility of determining this relation, cannot be sufficiently characterized in this respect, since there are still lacking some data about the spatial distribution of volcanism of the A_4 stage. However, territorial distribution of relicts indicates that andesite volcanism A_4 has regional character, being represented by a richly differentiated spectrum.

In the Banská Štiavnica region there are especially suitable conditions for the determination of the superpositional relations to the products of the IIIrd andesite phase. The extensive spatial distribution of the relicts of andesite lava flows (volcanism A_4) situated on the summits of morphological elevations, the thicknesses of these flows (Sitno about more than 100 m), and structural features of subhorizontal to horizontal fluidality corresponding to the course of the base of the deposited flows, indicate the manner of distribution of flows over considerably flat and peneplanized relief.

Andesite volcanism A_4 is characterized by spatial situation, by materia composition in relation to the volcanism of the IIIrd andesite phase, separated from the latter by the stage of volcanic quiescence with a period of destruction of bodies of the IIIrd andesite phase. The age of the andesite volcanism A_4 may be determined indirectly. Stratigraphic position of sediments of the intravolcanic basin in the area of Banská Štiavnica in the substratum of products of the IIIrd andesite phase (Burian—Štohl—Kováčik 1964) indicates that the commencement of the activity of the IIIrd andesite phase could have been in the Lower Sarmatian to Lower Pannonian (E. Planderová 1965). If we take in to consideration the following development of volcanism of the IIIrd andesite phase closed by the stage of destruction and denudation then the determination of the commencement of the activity of the volcanism A_4 may be as late as in the higher parts of the Sarmatian.

Predominance of the explosive products over the effusive (their ratio being approximately 9:1) is the common characteristics of the IInd phase volcanism and the volcanism A_4. The analogy of material composition and petrographical character considerably impedes the mutual differentiation in the areas with the absence of products of the third andesite phase. The analogy of the basic structural elements of the volcanism A_2 (IInd andesite phase) and A_4 is also indicated by the analysis of relicts. This is why the forms and types of volcanic eruptions of both stages are compared simultaneously.

Surface Volcanic Forms

Volcanism of the andesite intermediate to basalt-andesitic type is characterized by the maximum regional distribution over almost the whole area of the central-Slovakian region. The maximum concentration, with greatest variability of petrographic composition is in the central part (the area of the Kremnicko-Štiavnické pohorie Mts., Inovec and Vtáčnik Mts.), while the southern, south-eastern and eastern parts of the region represent an area of sporadic activity with greater dispersion of active centers.

(a) **Development in terrestrial environment**

Volcanic products of the IInd andesite phase in the central and northern parts of the volcanic region in basal strata are characterized by the indications of deposition of pyroclastic products in aqueous sedimentation basins, with the transition into typically terrestrial environment in higher zones. Volcaniclastic material, too, is accumulated in the adjacent intravolcanic basins or in isolated brackish bays, in marginal zones of the Tortonian-Sarmatian sea (to the S of the line Hontianske Nemce, and in the western parts of the line

Kozárovce). In the southeastern and eastern parts of the region, volcanism has the character of volcanodetritic material deposited in subaqueous environment of brackish, desalinated basins to sublittoral zones of the sea.

Stratovolcanic forms (polygene volcanoes)

The most frequent type is represented by apparatuses with stratovolcanic type of structure, composed by the explosive stages (with production of pyroclastic material), and by effusive stages (with effusions of lavas of intermediate andesite composition) of constructional type. The character of the constructional type is typical of apparatuses of greater size (Cotton 1959). Owing to the considerably advanced denudation and almost complete destruction of the primary forms, it is difficult to identify the separate apparates. The destruction stages are connected with extensive destruction with immense explosions accompanied by the collapse of the upper parts of cones, and with destruction evoked by the ascent of domes and tholoides, and by the activity of exogene factors.

In the present reconstruction study, it is inevitable to count the fact, that the extensive denudation of volcanic terrains has reached the stage of inversion of the relief. The top parts representing distinct morphological elevations, are covered with flat, subhorizontally to horizontally situated lava flows (the top parts of Javorie Mts., the summits of the Štiavnické and Kremnické pohorie Mts.). It is evident, that such thick andesite flows (relict on the peak Sitno 1004 m above sea level, thick more than 100 m) could have been accumulated only in the areas of depressions or near the foothills of volcanic cones. The higher cone forms built of heterogene incoherent pyroclastic material, and of periclinal lava flows, representing the primary morphological elevations, were affected by total destruction and denudation. If we consider the sites of deposition of extensive regional relicts of the lava flows, then in case of their supposed connection the height values of the primary forms may be expected to have reached 2000 to 3000 m.

That means that the cone forms cannot be situated in the areas of the present-day morphological elevations. They may be present in the areas with the absence of horizontally to subhorizontally deposited lava accumulations, Thus it may be supposed that the situation of the original cone forms corresponds rather with the present-day morphological depressions than with elevations.

Mechanism of eruption may be characterized on the ground of the analysis of erupted material.

In the profiles with well-preserved elements of the stratovolcanic structure, the bodies of chaotic, coarse-fragmentary to coarse-block pyroclastics, separated by the strata of tuff or tuffitic sediments, were identified. The thickness of the bodies fluctuates between 5—20 m. Characteristic is the spatial distribution in the forms of flat bodies orientated in one direction. The fragmentary material of variable size beginning with 3 cm up to 50—60 cm is cemented by the tuff mass with indications of strong solidification to agglutination. Sporadically blocks of the size of several m^3 are present. Deposition of material is chaotic, indications of sorting and stratification within the separate bodies are absent. Some blocks with torn and strongly vesiculated central part and with splittering of peripheral parts indicate explosion in the hot state in the process of the movement of the whole mass.

These bodies represent the depositions "nuées ardentes" (Lacroix 1903, 1904), "glowing avalanches" (Williams, Meyer Abich 1955), lately defined as "pyroclastic flows" (Aramaki, Yamasaki 1963), produced by the eruptions of the Pelée type.

In the field study of these bodies considerably extensive planar distribution with respect to comparatively smaller thickness, may be observed. Perret (1937, 1935), Gilbert (1938) and others explain this fact by movements of the hot mass from the ash fragments, of fragments and blocks in highly comprimated and quickly expanding suspension exceptionally mobile due to the reduction of friction among fragments. This mobile mass is following the topographical dip of relief, in its movement.

The structure of stratovolcanic forms is shared — in addition to the agglomeratic coarse-fragmentary pyroclastic deposits — also by fine-detritic tuff, tuff-pumiceous and tuff-sandy material produced in the stages of high explosive activity of volcanic type, with eruptions of ash-pumiceous clouds (pumice clouds, ash clouds). The essential part of the ash material is transported into the upper parts of the atmosphere in immense explosions, and eolically transported behind the borderline of the proper region of volcanic activity. In the course of the aeolian transportation, the material is sorted according to the granulometric composition and specific gravity; in areas near the volcano the coarser particles are separated, the finer fractions fall out with growing distance from the centre of eruption, as it was demonstrated by Hay (1959) on the example of the eruption Soufrière (1952).

The granulometric selection in the course of the aeolian transport is shown by some segments of the borehole GK 3 to the S of the Štiavnické pohorie Mts., near the village H. Rykynčice represented by the strata of vitrocrystalic tuffs deposited in the area of subaqueous marine basin. In the course of aeolian transport mineralogic and chemical composition of material varies. The heavier components, such as pyroxenes, amphiboles and ore minerals fall down in the places nearer to the volcano, while the lighter particles, such as volcanic glass, are transported on greater distance. The change of mineralogical composition is inevitably accompanied with the change of the chemistry of the resulting fraction. In intravolcanic basins and in adjacent parts of the Tortonian and Sarmatian sea, the indicated strata are formed almost purely by accumulation of pumice, which is a result of their separation from the coarser clastic components and from the fraction of minerals in the course of the aeolian transport, and probably also by the selective processes in marine environment. It is evident, that the resulting chemistry of the pumiceous fraction does not correspond to the average chemical composition of the whole erupted fraction. The effect of separation in the aeolian transport in the strata of the Burdigalian rhyodacite tuffs was pointedout by M. Marková (1966).

The strata of *vitrocrystallic and pumiceous tuffs* indicating the stages of explosive activity, were preserved either in isolated continental basins or in deeper parts of the marine sedimentary area. In terrestrial zones they were evidently denuded and washed down into the adjacent aquaeous basins. In littoral and sublittoral zones, too, due to repeated redeposition and resedimentation of material, these tuffs were desintegrated and mixed with other volcanodetritic material, which was accompanied by the rise of beach sands (Hrušov, Rykynčice).

Redistribution of volcanodetritic material on the slopes of volcanic apparatuses of stratovolcanic type, built of incoherent, nonconsolidated fragmentary material of predominantly pyroclastic type, was shared by the flows of the lahar type ("volcanic mud-flows"). Saturation of the non-consolidated mass by water in rainfalls (cf. Williams 1941; Schmidt 1934 — flows of the type "rain lahars", also "cold mudflows") or volcanic earthquake (Benkulen 1933) represent the stimuli of the movement of this mass.

The rise of the lahar flows in the near-shore zones of the Tortonian sea were pointed out by M. Kuthan (1963); some structural characteristics from the area of the Krupinská vrchovina highland were presented by V. Čechovič (1960, 1962).

In the southern marginal zones of the Štiavnica region, a row of bodies of the lahar type was determined. Lahar flows mobilized in the most northern zones (to the N of Hontianske Nemce) on the slopes of the apparatuses of stratovolcanic type passed over the littoral zone of the Tortonian sea when moving to the S, and continued along the sea floor in the form of immense mud flows with high specific gravity. Thickness of the separate bodies varies within 8—25 m, the maximum length followed was that of 6 km. Deposits of the flows are formed of polymict fragmentary volcanic material with the essential admixture of non-volcanic rocks (pebbles of Mesozoic and Palaeozoic rocks). In addition to subangular and strongly rounded fragments, also angular fragments and blocks are present. Their arrangement in tuffitic to clayey-tuff cementing matter being chaotic. Low roundness to angularity of the lahar flows transported to greater distances is — according to Curtis (1954) — explained by the function of the muddy-tuff material, that is supressing the mutual friction among fragments, reducing their abrasion in the course of transport, helping the mobilization of the whole mass, and the results of the uniform movement of the whole mass (Anderson 1933; Blackwelder 1928).

With respect to the specificity of the lahar flows indicated in the near-shore zones of the Tortonian sea with the movement and deposition in submarine environment, the term "subaqueous mudflows" in the sense of Carrozzi (1960) was accepted.

The polymict nature of the material, the absence of indications of the temperature effects, presence of the wood fragments point out that the movement of non-consolidated material on emerged slopes of volcanic apparatuses covered by vegetation, took place there. Flows of this type, activated without any evident effect of volcanic activity in the sense of Bemmelen (1949), may be parallelized with the type "cold mudflows" (Konečný 1966).

The lava flows represent further important element of the forms of stratovolcanic type. The planar distribution of relicts shows considerable fluidality of the flows. With respect to their higher resistance against erosion in relation to slightly consolidated noncoherent heterogene pyroclastic masses, in the present day period they are covering a row of morphological elevations (Sitno, the elevation point 1004, slopes of the Vtáčnik Mts., the ridges to the south of Banská Štiavnica). According to their subhorizontal to horizontal deposition and the conformable course of fluidal textures, and their thickness (exceeding 100 m), their character may be determined as that of the accumulations in depressional and flat areas, or near the foothills of the cone forms. Relicts with more steeply arranged lava flows, representing original slope areas of stratovolcanic forms (e. g. the cut of the railroad Turčok—Horná Štubňa, F. Fiala 1931), may be rarely distinguished. In this case, the character of the original slope is documented by the folding of the lava flow at the movement of lava along the steeper slope (Kuthan 1958). The absence of the uniform periclinal dip within larger volcanic massifs (Poľana) is probably also due to the activity of greater number of eruptive centers, that have conditioned the rise of complicated systems with a row of partial parasitic forms.

Thicknesses of the lava flows represent a considerably variable element dependent upon the primary physical-chemical factors of lava and upon the topography of relief. The maximum thicknesses were observed in flows with horizontal to subhorizontal arrangement. There are thicknesses about 100 m (the summits of the Javorie Mts., the peaks in the area of the Kremnicko-Štiavnické pohorie Mts.). The fluidal lavas of more basic andesites to basaltoandesites are characterized by smaller thicknesses (2—10 m — the slopes of the Javorie Mts.).

In the flows with completely preserved vertical profile, in their upper part, the zone of brecciation (locality Bzenica — Kuthan 1958) or brecciation of the substantial part of the flow accompanied with the rise of the block lava, may be observed (cut of the road Beňadik—Tlmače; Karolus — oral communication). The types "pahoe hoe" were not found within the intermediate andesite volcanism.

Multiple volcanoes

In addition to monogene forms determined by spatial and most probably also by chronological localization, there is a number of proofs of the development of complex polygene volcanic forms with several feeding systems.

In the area of Fabova hoľa and Muránska plošina plateaus, a number of circular and elliptical neck bodies of small size (20 × 20 cm to 1 × 2 m) concentrated to a small planar extension (4—6 on 1 km²), were exposed there. The function of these centers with migrating activity in time and space evidently gave no rise to the development of a simple stratovolcanic form of the central type, but to the complicated system with a number of partial parasitic apparatuses, i. e. group polygene volcanoes (the multiple volcanoes in the sense of Cotton 1952). In this case, the eruptive center cannot be localized, and it is better to determine the field of (of multiple volcanic activity).

The analysis of larger orogenetic volcanic complexes shows that this polygene type is not exceptional, but quite the usual type of forms.

Monogene forms represented by tuff cones were not identified becuase of their complete destruction, still their primary development is highly probable.

(b) **Development in subaqueous environment**

The eastern and southeastern parts of the central-Slovakian neovolcanic region are — in relation to the central part — representing an area with greater spatial dispersion of active centers, offering thus better possibilities for detection and analysis of the separate forms. The rise of specific forms and substantial modification of the character of eruptions were conditioned by the regional distribution of subaqueous environment.

In the area of the Javorie Mts. considered the relict of an extensive stratovolcano (Kettner 1939), development of volcanism is characterized by a widely differentiated spectrum; in addition to basalt-andesite and andesite rocks also the rocks of the dacite-andesite and rhyodacite types have been determined (Valach 1966). In the stage of the development of this apparatus specific processes of subaqueous volcanism were active.

In effusions of *lavas of basaltoid andesite* in subaqueous environment, the explosive phenomena being supressed, non-explosive desintegration of the surface of lava flows accompanied with the rise of hyaloclastite breccias in the sense of Rittman (1960) and Cucuzzo Silvestri (1963), took place. This process of disintegration was indicated by G. Meska and F. Fiala (1948) as the subaqueous granulation of lava. The hyaloclastite breccia on the surface of flows represents the preserving crust preventing further disintegration of lava flows present in its substratum in the form of a continuity flow. The hyaloclastite crust is splitting in the course of movement, and due to the strain tension there arise fractures penetrated by the lava dykes (the cut of the road Stará Huta—Oremov Laz). Hydrothermal solutions (warmed water and volcanic gases) cause alterations of fragments of volcanic glass and of marginal parts of vesicles, connected with hydratation and oxidation, giving rise to palagonite fringe (V. Konečný – A. Mihalíková 1966).

Planar accumulations of lava flows and hyoclastite breccias

The development of the hyaloclastite complex in the area of the Javorie Mts. took place by the rapid sequence of effusions of highly fluidal lavas of basalto-andesitic composition (SiO_2 – 47,89 %) in aqueous environment. Thickness of flows fluctuates within 1–8 m.

According to Rittman (1960), the rise of hyaloclastite formations is connected with the effusive activity along extensive fractures on the floor of the aqueous environment. The majority of submarine eruptions of basalt lavas have the character of fissure eruptions similar to subareal linear eruptions. The fact, that they take place on the sea floor, does not differenciate the fundamental characteristics of eruptions (Cucuzzo Silvestri 1963), yet the surface phenomena are completely different. The weight of a water column at respective depth of a basin causes reduction or absence of explosive processes; the outcropping lava coming in contact with water is rapidly desintegrated, giving rise to hyaloclastite breccias.

Although mechanism of fissure eruptions has not been determined for the whole hyaloclastite complex of the Javorie Mts., still there may be observed a process of brecciation in the area of ascending of larger dyke bodies. In upper parts of the separate strata of hyaloclastite breccias, in the processes of redeposition and resedimentation by water currents, there were formed strata of finedetritic beds of pelitealeuritic character with the uniform subparallel course. The processes of resedimentation and redeposition pass in stages between the separate effusions of lavas. In case of more rapid sequence of effusions no redeposited hyaloclastite strata arise, and the complex acquires the character of alternation of lava flows and chaotic hyaloclastite breccias.

The repeated accumulation of not too thick lava flows alternating with the strata of hyaloclastite breccias and with the strata of resedimented finedetritic hyaloclastites, evoked the rise of planarly extensive form with slight slope on its periphery, and with gradual transition into the adjacent complexes of the Krupinská vrchovina highland, built predominantly of redeposited volcanodetritic material, deposited in subaqueous environment. The thickness of the hyaloclastite complex is preliminarily determined at 230 m, with the extent over about 30 km^2.

The flat relief of hyaloclastite formation is distinctly pointed out by the distribution of thick relicts of lava flows of acid andesites (amphibole, hypersthene) deposited in the overlier of the apical parts of the Javorie Mts.

The transport of hyaloclastite material into an extensive sedimentary area of the Krupinská vrchovina highland took place mainly owing to the transporting activity of water currents. In addition to this kind of transport, in the marginal zones of the hyaloclastite complex also transport of material by hyaloclastite flows may be observed. The rise of currents and character of movements of hyaloclastite material is explained as follows: at retardation or completion of movement, or at total brecciation of the frontal part of lava flow, the more mobile hyaloclastite originating by disintegration of the upper part of flow, is shifted over the front of the flow, and moves independently in direction of the original movement of the flow behind the termination of the proper lava flow. The movement of hyaloclastite mass has the character of mass transport in the form of turbidity current of high specific gravity. Besides gravity force the mobilizing factor may be represented also by the volume expansion of the hot hyaloclastite material in an aqueous environment (Konečný 1967).

The lava flows of pyroxene andesite (augite, hypersthene) *with glassy to microlitic development of matrix*, characterized usually as the Bohunice type (Szabó 1886), represent a specific type. Their fluidal texture is distinct due to the alternation of lighter and darker strips; the surface of flows is often strongly vesiculated. Characteristic is the increased content of SiO_2 (up to 62 %). Considerable planar extention of effusive masses is still unsolved.

The study of the borehole profile Ku-1 (near the village Počúvadlo) penetrating through the flow of glassy pyroxenic andesite, brought some new aspects. Vertical cross-section from the basis to the overlier determines the following characteristic variations of the development of matrix:

0,00 – 10 cm: In the basal part, andesite is brecciated, strongly vesiculated with typical desintegration into small angular fragments, the fissures being penetrated by opal veinlets. The glassy lightbrown matter and phenocrysts (hypersthene, andesine, and sporadic augite) show a network of minor fissures.

1,5 m: In the above-basal part in the glassy matrix the amount of microlites is increasing, splitting is less intense.

6,5 m: In the lower part of the flow, the andesite is distinctly laminated (the length of the lamination surfaces is up to 3 cm), grey-black, matrix acquiring hyalopillitic to pylotaxitic development.

54,5 m: Andesite is strongly vesiculated, vesicles being flattened, subparallelly placed in the direction of flow, glassy matrix and phenocrysts being split into minor fragments.

74,4 m: The surface part of the flow has the character of highly vesiculated cinder mass of grey-black colouring. Glassy matrix is strongly vesiculated, the vesicles being irregularly torn, or even flatly deformed. A dense network of fissures is splitting phenocrysts and matrix into fragments. The fissures pass over the phenocrysts from the glassy matrix.

This indicates the development of the glassy matrix to be characteristic of the basal, and especially of the upper part of the flow, i. e. of the zones of quicker solidification, while the central, comparatively more fluid and more crystallized parts of the flow are characterized by the development of the lamination textures. The upper part of the flow acquires highly vesiculated to cinder-like character due to the spontaneous fading of the gaseous components. The intense splitting of the glassy matrix of the upper and basal parts of the flow is due to the inner strain at the spontaneous cooling by contact with aqaeous environment (desintegration into minor angular fragments); brecciation of especially the basal part is due to the mechanical fragmentation of viscous, rapidly cooling marginal parts unable to balance with the general mobilization moment of the flow. The spontaneous cooling on contact with aquaeous environment is accompanied with oxidation of dispersed ore pigment, and evokes the brown colouring of the glassy matrix.

These facts illustrate the process of *the rapid cooling of the lava flow in subaqaeous environment* (the subaquaeous environment is also pointed out by textures of deposited pumaceous ash tuffs in the overlier and substratum of the flow). On the ground of these facts, the role of the aquaeous environment may be considered as one of the important factors in the rise of vitrofiric andesites of the Bohunice type. The presence of lava flows in these areas in pyroxenic andesites with normal development of matrix (less acid — 58 % SiO_2) points out to the necessity of consideration of the specificity of petrochemical character and physical state of lava in the time of its cooling, as further inevitable disposition for the rise of the glassy flows of the Bohunice type.

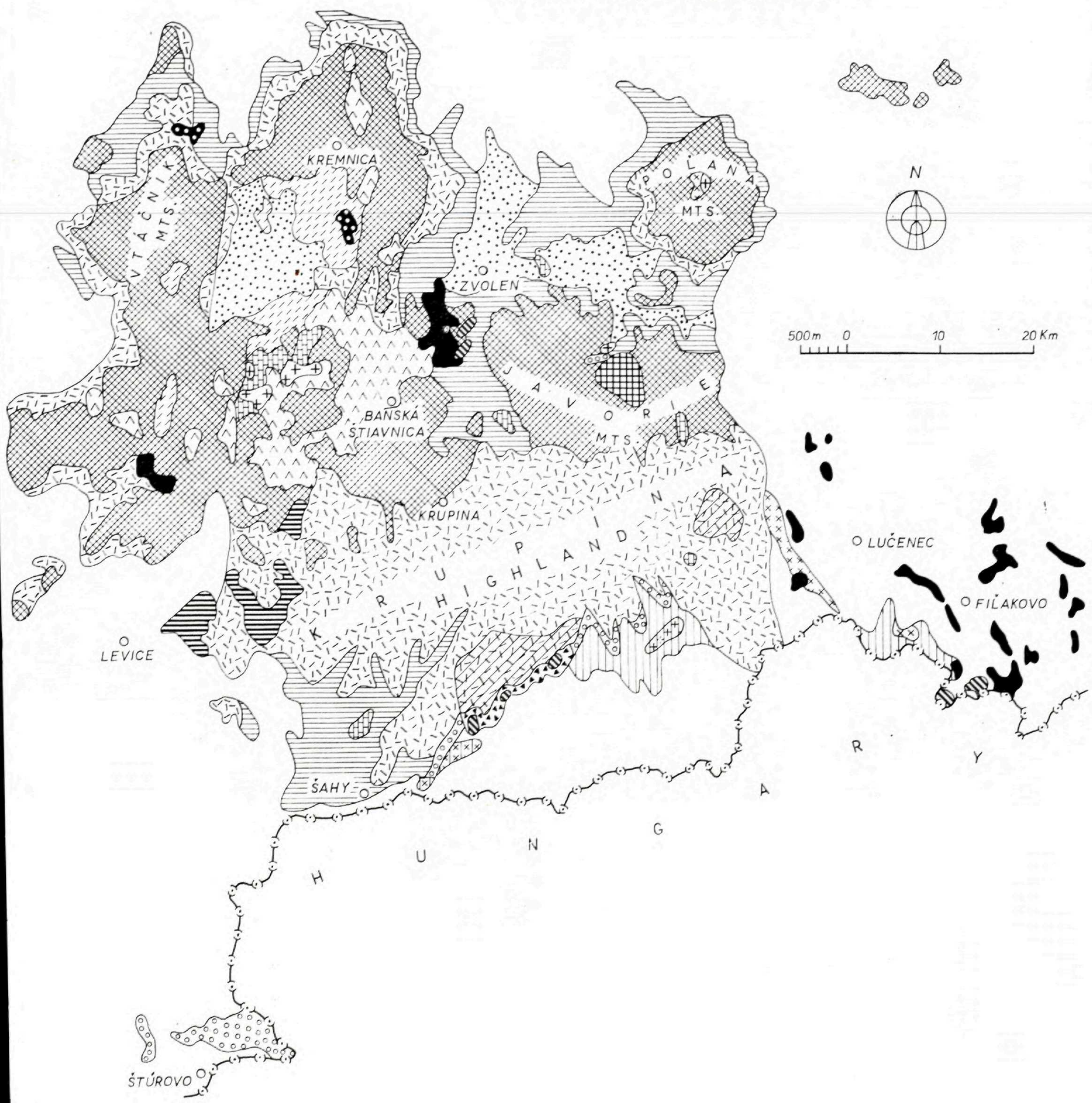
The conception of *the glassy pyroxenic andesite* indicated as the "Bohunice type" as of a uniform horizon or as an equivalent of one eruption stage, spread over an extensive territory, meets several obstacles. K. Karolus (1965) quotes some cases from the western parts of the distribution of glassy pyroxenic andesite of the Bohunice type indicating the pre-Lower Sarmatian age of the latter, and orders the Bohunice andesite into the second andesite phase. F. Fiala (1961) describes glassy andesite (the Bohunice type) in the overlier of the IIIrd andesite phase from the area of the Kremnické hory Mts. From the southern area of the Banská Štiavnica region come the proofs about the presence of the glassy andesite (borehole Ku-1) in the overlier of the products of the IIIrd andesite phase (A. Brlay—V. Konečný 1967).

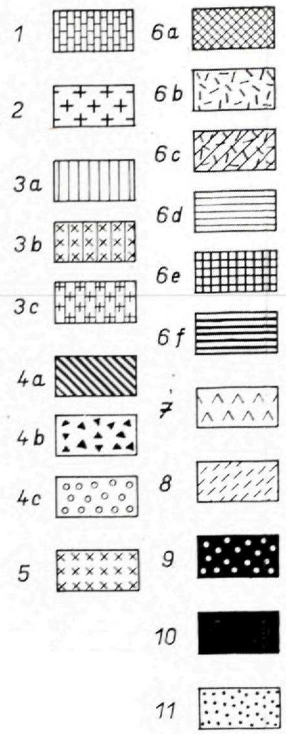
Consequently, it is impossible to use the term "Bohunice andesite" in the stratigraphical sense as an equivalent of one eruption stage or one lava horizon.

The suitable conditions for the rise of glassy types of pyroxenic andesites could occur also in several stages of activity in case of the external situation being convenient (subaqaeous environment), and of the optimal disposition of the physical-chemical state of lavas.

Great planar extension (to the S of the village Bohunice), limitation over the areas with the characteristics of subaquaeous environment, the absence of the elements of the structure of stratovolcanic type, close spatial relations with the strata of breccias showing the indication of the process of disintegration of lavas in subaquaeous environment (this process being analogous to the rise of hyaloclastites in more basic types) are the main characteristics of volcanism of pyroxene andesite with the glassy development of matrix (the Bohunice type). By the considerable planar extension the high fluidity

Encl. 1. Scheme of relicts of subsequent and final volcanism in the Central Slovakia
(by V. Konečný — M. Kuthan 1967)





1. Outcrops of the Substratum (Mesozoic and Paleozoic Rocks). — 2. The Granodiorite and Diorite bodies (pre — Tertiary ?). — *Products of the rhyolite phase* (R_1) *— area covered with sedimented pumiceous rhyolite tuffs and tuffites:* 3/a — in Upper Burdigalian; 3/b — in Helvetian (s. s.); 3/c — in Karpatian (Helvetian s. l.). — *Volcanism of acid andesite type* (I st. andesite phase — A_1): 4/a — extrusive bodies; 4/b — accumulations of coarse extrusive breccias; 4/c — area of redeposited volcanoclastic rocks in subaqueous marine environment; — *Volcanism of rhyolite type* (II nd. rhyolite phase — R_2): 5 — pumiceous rhyolite tuffs. — *Volcanism of intermediate andesite type* (IInd IVth andesite phase—A_2, A_4): 6/a — relics of the Stratovolcanic forms (explosive and effusive products); 6/b — coarse volcanoclastic rocks of allochtonous facies (redeposited material in subaqueous environment); 6/c — coarse volcanoclastic rocks of autochtonous facies (pyroclastic rocks); 6/d — volcanoclastic sediments of allochtonous facies (tuffites); 6/e — hyaloclastite formation of basaltoid andesite; 6/f — Lava effusions in subaqueous environment of pyroxene andesite (Bohunice type) probably fissure eruptions. — *Volcanism of acid andesite, andesite — dacite type* (III rd. andesite phase — A_3): 7 — extrusive forms (cumulodome, tholoids), lava flows and pyroclastic products are less frequent. — *Volcanism of rhyolite and rhyodacite type* (III rd. rhyolite phase — R_3): 8 — extrusive bodies predominate over explosive products. — **Final volcanism:** 9. Basaltoid andesites (B_1) — lava flows and dikes. — 10. Final basalts (B_2) — lava plateau, stratovolcanoes and scoria cones. — 11. The fillings of intravolcanic depressions.

of lavas is indicated. These facts point out rather the activity along the fissures with effusions into subaqueous environment than the effusions from the central feeding channels.

Subvolcanic Forms

The deep destruction of volcanic terrains in some areas caused denudation into subvolcanic levels and offered thus possibilities for determination of the feeding volcanic systems. The most frequent forms of subvolcanic bodies are necks and dykes.

Neck bodies

The optimal conditions for the determination of the neck bodies are offered by terrains where superficial volcanic forms were remowed. In the area of Fabova hoľa and Muránska plošina plateaus there was denuded a row of circular and elliptic necks of small size (20×20 to 1×2 m), irregularly accumulated in small areas (4—6 bodies on 1 km^2). Their positional situation shows a narrow connection with the areas of large overthrust dislocations. The accumulation of the neck forms (to diorite bodies) in the area Tisovec—Fabova hoľa—Muránska plošina plateau is in the area of the overthrust zone of Kráľova hoľa and the Krakľov zone not far from the transverse younger tectonic zone with the NW-SE course. Its form is most frequently circular to elliptic, with breccia filling, the edges straight and smooth. The bodies have the character of explosive penetration (explosive pipes) in the last stage of intrusion. The enclosed allogene material does not show any indications of intense thermal activity (Kuthan 1967).

The bodies of greater size are known in the area of the western part of the neovolcanic region near the Vtáčnik Mts. (Ostrovica). These bodies represent preparated forms — circular to elliptic in outline, elongated in one direction, formed by pyroxenic andesite (augite, hypersthene). The complicated columnar jointing is more or less perpendicularly oriented to the contact plane. On the periphery of the bodies there are no marginal breccia zones preserved, in marginal zones the surrounding lateral rocks are dragged out of the walls and transported by the moving lava (Karolus 1957).

In the area of the southern periphery of neovolcanites, there is a body of roughly eliptic form, exposed by denudation (Čelovce). The central zone formde by a lava mass, is surrounded on the periphery by a brecciated zone (surrounding the central core in about two thirds). In the lower part there is a gradual transition from coarse fractured andesite into higher, intensely splittered parts, in the upper part there is a transition into the zone of brecciation. The contact effects in relation to the surrounding sediments are not distinct, indicated only by firmness of the contact sediment or by alteration of its colour (zones thick several cm). The ascending movements are pointed by flow textures as well as by the transport of enclosed fragments with relicts of marine fauna. This body is connected with a dyke swarm.

Some intrusive bodies are terminated by ramification into several dykes and by gradual transition into the breccia parts, which points out to the subsurface mechanism of their origin (Kapolna near Vígľaš, Kuthan 1957).

Dyke bodies

The dike bodies may be divided into several types. *The dyke bodies with distinct contact on tectonic line* are of greater size, characterized by great directional length. As an example of this may serve the dyke from the area of the sout-Slovakian region, followed from northern Hungary (Benczurfalva) to our territory along 14,5 km, with thickness in some places smaller than 10 m. Magmatic masses ascending along such regional fractures, conditionned activity of linear type.

Another type of the outcropping of dike bodies is *roughly radial arrangement around intrusive to neck bodies* (e. g. around the elevation point 649 and the hill Rohy near Vígľaš). Further type is represented by *irregular accumulations of dyke bodies* with narrower spatial closing with the main neck body (Čelovce).

The central parts, i. e. the area with the maximum activity and accumulation of products of the eruptive phases, the dyke bodies are orientated frequently into distinct tectonic directions (the Carpathian, Sudetian direction).

2. Volcanism of Acid Andesite, Andesite-dacite to Dacite Type

Volcanism producing material of acid andesite types of andesite-dacite to dacite types is distinctly different from the volcanism of intermediary to basalt-andesite type.

While the intermediate andesite volcanism is a typical manifestation of the structure of forms of stratovolcanic type with the distinct predominance of explosive phenomena, and with subsidiary effusive activity, the acid andesite to dacite volcanism is characterized by predominatly extrusive processes with transitions into short, strongly viscous flows, and processes of intrusions with the forming of dyke bodies and larger intrusive forms. Although explosivity of eruptions is rather high (production of the pumaceous-ash clouds), the explosive phenomena are subsidiary to the amount of the masses produced.

Two stages of the volcanism of the above type may be distinguished within the subsequent volcanism:

1. *Volcanism in the Lower Tortonian*, with submarine extrusions of brecciated material, and intrusive-extrusive ascending of larger masses (the Ist andesite phase A_1); 2. *Volcanism in the Lower Sarmatian* with extrusions of

Tab. 3. Volcanism of acid andesite, andesite-dacite and dacite types

			Volcanic forms	Types of eruptions	Products of volcanism	V. P.	S
I	Surface to subsurface forms	Continental environment	1. Extrusive bodies (type cumulodome)	Extrusions with forming of cumulodomes	andesite, extrusive breccia		
			Extrusive bodies with transitions into short flows (type dome flows)	Extrusions passing in short flows	andesite, extrusive breccia, lava flow;		
			2. Protrusions (type tholoid or plug) accompanied with nuées ardentes	Protrusions in craters connected with emissions of nuée ardentes	andesite, agglomeratic pyroclastics,	III	A_3
			3. Lava flows	Effusions and flows connected with extrusions	andesite		
			4. Accumulations of pyroclastic and volcanodetritic material in adjacent sedimentary basins	Highly explosive eruptions process of progressive desintegration in the course of ascending of extrusions	pumaceous and ash tuffs, fragmentary agglomeratic material and volcanoclastic material,		
		Submarine environment	5. Irregular accumulation bodies of brecciated extrusive material (breccia piles)	submarine extrusions of brecciated material	chaotic agglomerates, extrusive breccias	I	A_1
						I, III	A_1, A_3
II	Subvolcanic forms		6. Dykes	intrusion along tectonic lines	acid andesites, dacite-diorite-porphyry	III	A_3
			7. Isometric to unidirectionally oriented intrusions	intrusion in tectonically pre-disposed directions	quartz-diorite-porphyry		

the cumulodome type with transitions into the lava flows, protrusions of tholoides, lava flows, intrusions of dike and larger bodies (the IIIrd andesite phase A_3).

Volcanism of the Ist andesite phase is a wide regional phenomenon, according to the existing data. In addition to volcanic products in the eastern and southern parts of the region, they may be found also in the central part (Banská Štiavnica; Burian 1965, E. Karolusová 1966), in the southern peripheral Štiavnica area (E. Karolusová 1965, V. Konečný 1965), and in the area of the Kováčovské kopce hills. In the last time they were also determined by the boring works from the area of the Zvolenská kotlina depression (M. Pulec 1966).

On the other hand, volcanism of the IIIrd andesite phase is localized in comparatively narrower territory in the central part, in the area of the rift fault, rimming the Kremnické pohorie Mts. in the west, near the S border of the Hodruša—Vyhne island on the S, SE and NW, and in the area of the Hronský Inovec Mts.

Volcanism of the Ist andesite phase has the character of submarine development in sublittoral to littoral zone; continental condition being typical of the forming of the bodies in the IIIrd andesite phase.

Their common characteristics, however, is the analogy of the main features, i. e. the predominating nature of extrusions and subsidiary explosive processes.

Surface to Subsurface Volcanic Forms

Process of extrusions with the development of dome forms and forms with transitions into short thick viscous flows, the rise of protrusions (of tholoid type — F. Fiala 1965), and processes of intrusions of dyke bodies and larger isometric bodies, are characteristic of the eruptions of highly viscous acid andesite and andesite-dacitic to dacite masses of the IIIrd andesite phase. Explosive activity, with the forming of nuées ardentes and pumaceous-ash masses are only an accompanying phenomenon. Although submarine extrusions of the Ist andesite phase show analogy of the basic mechanism, i. e. the process of extrusions, the bodies — in difference from the IIIrd andesite phase — are completely disintegrated and brecciated in the course of their ascension, and in the last stage they are extruded in the form of extensive mass of brecciated material. The latter is accumulated in the nearest areas of the extrusions, the accumulation forms being irregular. The difference of the surface forms is due to the specificity of the submarine environment of the rise of these extrusions, to a considerable extent.

(a) Development in continental environment

The ascension of the bodies of the IIIrd andesite phase took place in terrains built of the relicts of earlier volcanic stages (mainly products of the IInd andesite phase), in a distinctly dissected relief. The modelling of relief was due to the continental denudation stage connected with extensive devastation of the surface forms. The dissection of relief in the stage of the ascending of the bodies of the IIIrd andesite phase is pointed out by the development of intravolcanic basins with finedetritic and tuffito-lignite sedimentation (intravolcanic basin in the area of Banská Štiavnica — Burian — Kováčik — Štohl 1964). The ascension of the bodies of the IIIrd andesite phase conditioned further deep destruction of the original relicts of volcanic forms.

Extrusive bodies (cumulo-dome type); extrusive bodies with the transition into thick short flows (dome-flows type)

Accumulation on smaller terrains with the total predominance of solid matter over fragmentary detrital volcanic and pyroclastic material, is distinctly characteristic of the bodies of the IIIrd andesite phase. This is reflection of substantially different processes of genesis and development of forms with respect to the most frequent type of stratovolcanic apparatuses. On the ground of macrotextural elements, such as textures of vertical or near-vertical fluidality, zones of brecciation near the periphery of bodies, formation of some bodies by the extrusive movements, and cumulation of highly viscous magma in the areas of the extrusion — as the forms of the dome type, may be supposed. In the marginal parts of some bodies, characteristic zones of brecciation were developed. Their width is exceeding 50 m. Clastic angular to subangular material produced by mechanic desintegration is cemented by lava vesiculated matter with fluidal texture of the flowing-round of fragments. Orientation of fragments is chaotic, accidental, even roughly subparallel with the direction of fractures. Transitions of the zones of brecciation into massive andesite bodies may be observed.

The forming of brecciated zones in the peripheral zones of bodies is connected with the process of extrusions in the course of the extrusive processes (perhaps by mechanical fragmentation of more quickly solidified peripheral zones, and due to the abrupt expansion, accompanied with the extension of the space). The study and the proper interpretation of the zones of brecciation may serve — in addition to other macrotextural features — as a criterion for identification of spatial parameters of bodies, and for the determination of some aspects of their genesis.

In peripheral zones of the extruded masses there is a frequent transition into thick and short highly viscous, often brecciated flows. (The borehole Ku-1 — Počúvadlo show the thickness of 58 m brecciated flow, situated on a propyllitized complex of the IInd andesite phase). Due to high viscosity, the planar extension of flows is limited and localized more or less only to the nearest areas of the extruded masses.

In peripheral zones of the main accumulations in a narrow spatial association, in the area of depressions there are accumulated pyroclastic products (pumiceous ash tuffs, agglomerates with the nature of deposition in subaqueous basins). These facts show that the above bodies are formed predominantly by accumulation of highly viscous matters in the asceeding process, that are communicated with the surface.

In the sense of Williams (1932), the development of the above bodies is partly endo- and partly exogeneous. Jaggar (1920) reccomended the term "cumulo-domes" for this type, since the term "dome" was used in too general sense, and also for the "domes of the Hawaian type" (shield volcanoes) formed by the repeated effusions of basalt lavas i. e. by quite a different process.

Cotton (1952) suggested to call the transitional forms between domes and short, highly viscous flows, formed on the margin of the caldera Crater Lake (Oregon) in the extrusion of dacite lavas, passing from the dome form into a short flow on the slope — the "dome flows". It may be supposed, that the terms "cumulo-dome" and "dome-flows" are fully applicable for some forms of the IIIrd andesite phase.

Fiala (1961) pointed out to the extrusive process of some bodies of the IIIrd andesite phase in the Kremnické hory Mts. Brlay (1965—66), Konečný (1965—66) described the extrusive and intrusive character of the bodies of the IIIrd andesite phase.

There is a close relation between the spatial situation of bodies and zones of intense tectonics. This is first of all pointed out by the distribution of the bodies of the IIIrd andesite phase in the southern part of their arch-like zone along the line E-W to NE-SW, that is — according to geophysical research — a tectonic zone of the lst order, representing the southern tectonic border of the Hodruša—Vyhne tectonic group (Bodnár, Ďuratný, Zbořil, Orlický, Filo 1966). Geologically, the zone represents the contact of the propyllitized complex of the IInd andesite phase of more northern central area with the outer arch-like zone of the bodies of the IIIrd andesite phase. Analogous ~~co~~nnection may be observed in the spatial distribution of bodies of the IIIrd andesite phase near the eastern border of the central Štiavnica area, built of the propyllitized complex of the IInd andesite phase (Böhmer—Štohl, in press).

In the forming of bodies of the IIIrd andesite phase, their connection with the weakened zones, i. e. the zones of the distinct tectonic mobility, may be observed. This mobility in the course of ascending of bodies of the IIIrd andesite phase is proved by the forming and further development of the sedimentary basins on their periphery (borehole Ku-1 near the village Počúvadlo), filled in the course of gradual subsidence by detritic volcanic material and by pyroclastic products.

The ascending of bodies of the IIIrd andesite phase took place in the area of denuded relief of the IInd andesite phase, morphologically dissected, with the development of small intravolcanic basins (intravolcanic basin in the area Banská Štiavnica, with limnic sediments; Burian—Kováčik—Štohl, 1964).

Protrusions (of tholoid or plug type) accompanied by emissions of "nuées ardentes"

As an example of this form may be considered the protrusion of glassy amphibole-biotite andesite $\pm$ hypersthene (the IIIrd eruption group, according to Fiala 1956, 1961; the IIIrd phase, according to Kuthan 1963), penetrating through the older products of pyroxenic andesites with the stratovolcanic type of structure, near Horný Chom (the Kremnické pohorie Mts.)

In the first explosive stage conditioning the rise of „nuées ardentes" in lateral explosion (also glowing cloud), moving down the moderate slope in direction to the area of the Jastrabá sedimentary basin, a protrusion of highly viscous andesite accompanied with the rise of specific form of the "tholoid" or „plug" type, was formed in the top part of the volcano Horný Chom (the Kremnické pohorie Mts.). According to the nature of explosion, "the nuées ardentes" of the first explosion stage are considered as the type between the "directed blast" and "explosive incandescent avalanches" (Fiala 1965).

Lava flows

They do not form planarly extensive forms, they represent comparatively short and thick viscous flows closely associated with the main extrusive masses. They are often characterized by brecciation and by vesiculated to pumiceous surface. Similarly Fiala (1961) suggests that the viscous masses of amphibole-biotitic andesites (the third eruption stage) form extrusions rather than lava flows.

Andesite of the Obyce type, distributed over large areas, is a specific type. The problem of genesis of this type of andesite is still being solved.

Accumulations of pyroclastic and volcanodetritic material (in adjacent sedimentary basins)

Fragmentary volcanic material coming from the progressive desintegration of extruded bodies or representing deposits of nuées ardentes connected with growth of tholoids, and finally the explosive material from the stages of rapid explosive volcanic activity (pumiceous and ash tuffs), were accumulated in the adjacent depression, usually isolated in sedimentary basins.

The deposits of nuées ardentes are characterized by unsorted, chaotically deposited material (with the size from minor fragments to large blocks), with substantial admixture of fraction of the size of ash. The rise of nuées ardentes is connected with the growth of tholoides (Veľký Chom; Fiala 1961—65).

In addition to the mass of chaotic material of the type of nuées ardentes, the bodies of the IIIrd andesite phase are associated with agglomeratic pyroclastics with the fraction of 1 to 10 cm, rarely 25 cm. Fragments are well-rounded and sorted. Volcanoclastic material shows the character of deposition in a subaqueous environment, forming the strata thick several ten meters, alternating with the strata of lithocrystallic tuffs. Some textures indicate the floating down into intravolcanic depressions (Burian 1964). Volcanoclastic material is probably coming out from processes of disintegration of bodies of the IIIrd andesite phase, passing in the course of their development, and also as a result of the later destruction processes.

Thick strata of the pumaceous-ash tuffs alternating with the strata of pellitic sediments are a product of the stages of strongly explosive activity accompanying the forming of bodies of the IIIrd andesite phase.

(b) **Development in submarine environment**

In the time of Lower-Tortonian a transgression accompanied by subsidence of the southern part of neovolcanic region, near the connection of the southern neovolcanic area and the eastern part of the Podunajská nížina lowlands, along the faults of Carpathian direction, intense volcanic activity (Seneš 1962) of the Ist andesite phase (Kuthan 1963), took place. In the region of

Hungary amphibol-biotitic andesites, biotitic andesites and hypersthenic andesites $\pm$ amphibole $\pm$ garnet are being intruded.

In the southeastern and eastern parts of the central-Slovakian neovolcanic region, extensive bodies of pyroxene, amphibole- andesite $\pm$ garnet were formed (Breziny, Šiatoroš—Karanč), in the southern part submarine extrusions of smaller bodies of hypersthenic andesite with amphibole, took place. In the central and western parts of the neovolcanic region, the strata of agglomeratic or widely redeposited volcanoclastic material were determined by the boring work in the base of the volcanogene complex. The genesis of andesite bodies in the Neresnícka dolina valley and Karanč — Šiatoroš, cannot be determined so far because of the insufficient exposures in the terrains.

Accumulation bodies of extrusive breccia piles

Some aspects of the processes of genesis and mechanism of eruptions are offered by the study of submarine extrusions in the southern periphery of neovolcanites. The area of the piercing out of the extrusive bodies is situated in the zone between comparatively stabilized eastern zone and the western zone of the intense subsidence, i. e. in the zone affected by the maximum tectonic strain in the segment of sublittoral to littoral zone of the Lower-Tortonian sea. In the segment of 15—12 km there were determined extrusive bodies, concentrated along the line NE-SW (Carpathian direction), transversely dislocated by a NW-SE fault system.

Bordering of bodies is irregular, isometric (Širákov) to distinctly oriented in one direction. The maximum width followed is about 120 m.

Within the separate bodies, in vertical section, beginning with the lowest parts, the following zones were determined:

1. in the lower part: andesite-massive or penetrated by a network of roughly vertically oriented fractures,

2. in the upper part there is a transition into the zone of more intense frakturating, mainly in those parts that are adjacent to the most important fractures, characteristic by the jointing of angular fragments. Deformation of some fragments indicates fracturing in a highly viscous state, in the course of the piercing movements. Angular fragments in this zone, are not completely replaced from their original position, they preserve an orientation approximately along the main fractures,

3. in the next uppermost zone there is a spontaneous brecciation and disintegration with simultaneous increase of vesiculation of fragments. Orientation is chaotic, distinct rotation of fragments due to the mass movement, may be observed. The cementing matrix has the nature of vesiculated and disintegrated lava, altered to a granular mass. Chaotic breccia mass acquires the character of a coarse fragmentary agglomerate.

The boundary between the brecciated extruded material and the surrounding accumulations of thick-blocky agglomeratic masses is indistinct, in some bodies having the character of gradual transition. Fragmentary material of brecciated bodies and surrounding mass is identical as for texture and petrography. Basing on this, following process of genesis has been reconstructed: in the time of extrusion along tectonically predisposed directions into water-bearing marine sediments, at the cooling and evaporation of gaseous components, viscosity of the extruded lava is increasing so quickly, that the upper and marginal solidified parts cannot be balanced with the general mobilization moment, and intense fracturing takes place. Along fractures, in the differentiation movements, intense splitting into angular fragments preserving — in the initial stage — approximately subparallel orientation with the course of main fractures, takes place. In the continuous movement connected with the extension of the space, and in time of contact with the sea water, spontaneous brecciation and extrusion of brecciated mass onto to the sea floor, took place.

The fact, that desintegration takes place in highly viscous state, is pointed out by distorzion and deformation of some fragments. In the upper part, the gaseous component is spontaneously escaping from the extruding masses, which is indicated by increased vesiculation of fragments and desintegration of the vesiculated lava matter into a granular cement. The rise of the granular cementing matter is also due to the friction in the course of movement.

The origin of volcanoclastic material from extrusions of brecciated bodies is indicated: 1. by accumulation of coarse-block agglomeratic masses in the immediate neighbourhood of extrusions, 2. by monomict character and identity of petrographic composition of agglomeratic masses and extrusions of brecciated material, 3. by distinct angularity of fragments, 4. by absence of fragmentary material of explosive origin (bombs, ect.).

On the periphery of the accumulation bodies of brecciated material, their destruction, and transport of volcanodetritic material in sublittoral to littoral zones of the Tortonian sedimentary area, took place.

Subvolcanic forms

Identification of bodies of this type is conditioned by the degree of denudation of higher volcanogene complexes. Valuable data are offered by mining work, crossing the dyke systems (Banská Štiavnica—Kremnica). Dyke bodies and intrusions of larger masses of the IIIrd andesite phase are spatially distinctly associated with the areas of the ascending of extrusions of main masses. Dyke bodies show positional connection with pre-disposed tectonic systems. In larger intrusions, this connection is indistinct.

Dyke bodies

They represent forms of variable thickness (from several cm to 50 m), following the distinct tectonic directions in volcanogene complexes. The maximum direction length is as much as 5 km. The NE-SW to N-S directions represent the predominating orientation of the dyke systems in the central Štiavnica area. Besides the actual dikes also the bodies with the character of sills were determined, following the planes of discontinuity between the Mesozoic substratum and volcanogene complexes. Combination of the actual dike bodies with sills gives rise to complicated forms. The dyke bodies are predominantly indicated as subvolcanic forms without sufficient proofs of their primary communication with the surface. The course of ore veins of the Banská Štiavnica ore region (Štohl 1962) is also bound to the structural-tectonic system used for the piercing out of the dyke bodies.

On the ground of microscopical determination of characteristic variations of the development of structures of matrix in dependance upon the depth and geological structural characteristics, a suggestion was made to leave the original term "dacite" ("dacite dykes") and to use the indication "quartz-diorite porphyry and dykes of quartz-diorite porphyry" instead (Mihalíková 1966).

Intrusive bodies with isometric to orientated direction

In addition to dike bodies there are also forms of larger size (1,5 km, and more) with irregular isometric to those orientated in one direction. Bodies of this type occur mainly in the central area of the Banská Štiavnica region, in propyllitized complex of the IInd andesite phase. Identification of the bodies is conditioned by considerable degree of denudation of the upper volcanogene complexes and of lower levels. As an example may serve the dacite body to the N of the village Dekýš (in the area of the elevation point Šemetlov). Its intrusive character is documented by penetration of propyllitized complex of the IInd andesite phase (Brlay 1965). The rock has holocrystallic to hemicrystallic matrix, and amphibole-biotite, plagioclase, quartz phenocrysts.

Piercing-out of bodies of this type is also known in Štiavnica — Hodruša tectonic group, their distribution shows the orientation on tectonic directions.

3. Volcanism of rhyodacite and rhyolite type

In the course of the activity of subsequent volcanism, three stages with rhyolite and rhyodacite volcanism are known. The first stage passed in the Middle Burdigalian, sporadically reaching up to the Helvetian-Carpathian (Vass, Gabčo 1964). This stage is represented by deposition of pyroclastic material of rhyolite-dacite type, its centres might have been in the adjacent Hungarian areas (Kuthan 1963).

In the time of the second stage (the IInd rhyolite phase) in the Middle Tortonian, pyroclastic products with thickness up to 40 m (Ľuboriečka) were deposited. Some of eruptive centres are represented by the rhyolites in the area of Poľana (Húsenica 1958) and Vtáčnik Mts. (Slávik 1956, 1962).

Tab. 4. Volcanism of rhyodacite and rhyolite types

			Volcanic forms	Types of eruptions	Products of volcanism	Volc. phase	Symbol
I.	Surface to subsurface forms	Continental environment	1. Extrusive masses with distinct connection with tectonic directions	Extrusions	Rhyolite, marginal zones with perlitic and brecciated development	III	R_3
			2. Deposits of pyroclastic material transported				
			(a) in atmosphere and in aqueous environment	Eruptions of ash-cloud and pumaceous clouds	pumice and ash	I, II, III	R_1 R_2 R_3
			(b) by ash flows in time of ignimbrite eruptions	Eruptions of ash flows (Katmai type)	welded pumaceous tuffs (ignimbrites)	III	R_3
II.	Subvolcanic forms		3. Dike bodies	Ascend along tectonic lines	rhyolite	II, III	R_2 R_3

The maximum rhyolite volcanism was that in the last stage (the IIIrd rhyolite phase). There was most distinct concentration over smaller area, with respect to the preceeding stages of volcanic activity. Highly viscous lavas are limited to the rift fault area in the region of the Kremnické hory Mts., and of the Tekov—Žiar depression. In other places only sporadic penetrations of rhyolite dyke forms occurred (the area of Banská Štiavnica).

It is difficult to reconstruct the area occupied by the original extent of explosive material on the ground of the relict pyroclastic deposits. According to indications offered by acid rhyolite material in intravolcanic basins (the Zvolen—Badín depression, the Turiec depression, etc.), and in brackish sediments of the Podunajská nížina lowlands, extensive aeolian transport of exploded material in the form of pumaceous-ash clouds may be supposed.

In addition to that, also eruptions with the rise of ash flows following topographic relief in their movements, occurred. After deposition, there pass processes of welding, accompanied with the rise of the ignimbrite type.

Surface Volcanic Forms

Basing on forms, and characteristic manifestations, the following elements may be determined: (1) *Extrusions of extensive rhyolite masses*; (2) *Depositions of pyroclastic material*: (a) transport in atmosphere and in aqueous environment — pumaceous and ash tuffs, tuffites; (b) transport by ash flows — ignimbrites.

Extrusions of extensive rhyolite masses

Due to the extreme viscosity of the rhyolite mass of the IIIrd rhyolite phase there are no normal lava flows; the rhyolite masses are extruded along tectonically predisposed zones, which gave rise to masses, characteristically orientated in one direction. In relicts, conveniently denuded by erosion or artificially (the quarry near Szabova Skala), the ascension is pointed out by macrofluidal structures, orientated under the angle of 70°—90°. In the marginal parts of extruded masses, on contact with subaqueous environment in quick process of cooling, perlitic glassy masses arise (Szabova Skala; Forgáč—Karolus—Konečný, in press). In the course of the piercing out, brecciation of the marginal, more rapidly solidified parts takes place. These parts are unable to balance with the general movement, and due to this they are affected by strain and splitting, the fragmentary material being again penetrated by fluidal rhyolite mass.

The problem of inner structures is not completely solved, it may be supposed, however, that the primary forms were represented by dome type forms.

Pumiceous and ash tuffs, tuffites.

Rhyodacite volcanism (the Ist rhyolite phase) is represented by the strata of pumaceous tuffs in the territory of the central-Slovakian neovolcanic region. These tuffs are deposited in lacustrine to brackish Burdigalian basins of southern Slovakia. Manifold alternation of the minor strata of pumiceous-ash material and clayey-sandy sediments, indicates periodical activity of explosive type (pumiceous-ash clouds) of the volcanic types of eruptions. Volcanic centres are not known on our territory. They are, however supposed to be in the adjacent Hungarian areas of volcanic activity (Kuthan 1963). Volcanic material shows some characteristics of granulometric separation owing to the aeolian transport (Marková 1966). The activity of volcanism producing rhyodacite pyroclastic material with sporadic manifestations lasted up to the Helvetian-Carpathian (Vass, Gabčo 1964).

In the Middle Tortonian there are relict strata of rhyolite pyroclastic material (the IInd rhyolite phase), in some places represented by the strata of 20—40 m thicknesses (Trenč, Ľuboriečka, eastern slopes of Poľana). Eruptive centres are supposed relicts of rhyolite bodies from the Poľana massif (Húsenica 1958) and in the eastern part of the Vtáčnik Mts. (Slávik 1956, 1963). Products of the explosive activity of the IIIrd rhyolite phase are deposited in the adjacent intravolcanic basins (the Zvolen—Badín depression, Žiar depressions); fine volcanic material was transported by aeolian processes into the Sarmatian brackish sedimentary area of Podunajská nížina lowlands. Deposition of pyroclastic material in intravolcanic depressions was interrupted by sedimentation of clayeysandy and pellitic sediments with lignite. Separation on the ground of granulometry points out to the aeolian transport, and perhaps, also the transport in aquaeous environment. (Borehole P 2 — the Žiar depression — Pulec 1966).

The character of manifold alternation of pyroclastic material with clayey sediments (representing stages of rest in the course of explosive activity) indicates periodical eruptions of explosive type, producing pyroclastic material eolically transported in the form of pumiceous-ash clouds to considerably distant places.

Ignimbrites

Fiala (1961) quoted some relict strata of consolidated deposits with pyroclastic structures from the area of the Kremnické hory Mts. Due to welding pyroclastic structure is wiped out in some places, and the rock acquires the character of fine-grained glassy rhyolites. In some occurrences there are transitions between the tuff material into the strata of homogenized to rhyolite — like material. Basing on this fact, and on the flat deposition of relicts, the mechanism of genesis in the sense of Marshall (1935), Fenner (1923, 1937) of the products of ignimbrite eruptions, i. e. of eruptions of "ash flows" (Ross Smith 1961) and of "sand flow" (Fenner 1937) connected with the eruptions of the Katmai type was supposed.

The extent of the known relicts gave no possibilities to suppose the regional distribution and other aspects of the genesis till now.

Subvolcanic Forms

In rhyolite volcanism, subvolcanic forms are known only sporadically (predominantly the IIIrd rhyolite phase). No subvolcanic bodies of the Ist rhyolite phase are known in our territory. With the IInd rhyolite phase, relicts of complicated, strongly destructed rhyolite bodies of the central part of the Poľana massif (Húsenica 1958) and rhyolite bodies from the area of Vtáčnik Mts. (Slávik 1963, 1959) are connected.

Dyke bodies

The main accumulation of rhyolite extruded masses of the IIIrd rhyolite phase are in regional relation connected with dike bodies of small size (the average thickness is 10—20 m, sometimes more, the maximum directional length followed is 1,5 km).

In orientation of the dike bodies, mainly in the area of the Štiavnica region, the N-S direction is predominating (Burian 1965). In the last time there were determined the dyke bodies of rhyodacites in the area of the Javorie Mts., penetrating through the propyllitized pyroxene andesites (Valach 1966).

With respect to the large extrusive masses of the IIIrd rhyolite phase, the subvolcanic dyke bodies represent only a sporadic phenomenon.

VOLCANISM OF FINAL STAGE

Due to the close of the differential development of palingene magmatic masses of subsequent volcanism, consolidation and cratogenization of volcanic regions took place. In the course of processes leading to the balance of the inherited unequilibrium of the then mobile zones, deep faults arose mainly in the inverse segments; the volcanism of the subsequent stage is replaced by the final volcanism.

Basalt volcanism

The difference of volcanic forms and nature of eruptions of basalt volcanism of the final stage, with respect to subsequent volcanism is conditioned by its different tectonomagmatic position. Along deep balancing fault zones, magmatic matters of basic composition are mobilized from deep situated sources. The low gas content of slightly differentiated and contaminated magma is reflected in very low explosiveness of eruptions and almost in total predominance of the effusive activity. If the ratio of explosive material to the effusive is about 9:1, in final basalt volcanism it is contrary.

The effusions of basaltoid andesites is connected with the forming of fault systems of inverse zones from tectonically mobile zones, in the area between the Štiavnické and Kremnické pohorie Mts. and the NE part of the Vtáčnik Mts. The dike forms are orientated predominantly in the E-W ($\pm 20°$) direction, penetrating through the gravel-conglomeratic formation of the Žiar depression.

The commencement of the main stage of the final volcanism is in the Upper Pliocene, with effusions of large effusive masses. The maximum activity in the eastern part of the region caused the rise of an extensive basalt plateau.

The proper volcanism in the central-Slovakian neovolcanic area is younger, there are known sporadic occurrences of the dyke bodies, necks (Kalvária), of lava flows (Dubová—Bacúrov; Fiala 1933; Čajková 1957) and scoria cones with effusions of lavas (Pútikov vŕšok; Fiala 1952).

Surface Forms

The surface forms of the basalt volcanism, formed in the course of explosive and effusive activity, may be divided into several main types: (1) extensive

Tab. 5. **Volcanism of final stage**

			Volcanic forms	Types of eruptions	Products of volcanism
I.	Surface forms	Continental environment	1. Lava sheets of plateau type	Extensive effusions of basalt lavas	Lava
			2. Monogene scoria cones (Ragáč)	Activity of "fire fountains"	Scoria, ash, lapilli,
			3. Polygene forms of scoria material, bombs and lava flows	Activity of the Stromboli type	Scoria, lapilli, ash, boms, lava flows,
II	Subvolcanic forms		4. Dykes, groups of dikes, multiple dykes, necks?	intruding along tectonic directions, or along the crossing of tectonic directions	Basalt

lava sheets of the plateau type (eastern part of the region); (2) monogene scoria cones (Ragáč); (3) polygene forms (cones of mixed type formed by explosive products and effusions of lavas (Pútikov vŕšok).

Extensive lava sheets of the plateau type

Effussions of basic lavas evoked the formation of an extensive basalt plateau reaching deep to Hungary in the south (Jugovics 1940, 1944), in the eastern part of the neovolcanic region. The distribution of relicts indicate the extent of the basalt plateau from the southern foothill of the Carpathian Mts. (Podrečany) over the area of Lučenec—Fiľakovo to the area of Šalgotárján. On the basis of the plateau there is more or less continuous strata of tuff to tuffitic depositions of the limnic type. The maximum thickness of relicts of the original plateau is 20—30 m. The problem of the forming of the plateau has not yet been completely solved.

According to the generally accepted opinion, the forming of extensive basalt plateaus is connected with the effusive activity along the fissures of considerable directional lengths (fissure eruptions). There is, however, only a limited number of data proving the opinion.

Reck (1930) supposed effusions from the group of feeding channels concentrated on small areas, instead of effusions along fractures. Tyrrel (1937), too, supposed the origin of plateau basalts in greater number of feeding channels (multiple vent) of generally smaller size as the supposed fissures.

The basal complex of products of the final volcanism is penetrated by dyke bodies of length not exceeding 100—200 m or systems of dykes grouped in small areas. The dyke systems are considered as feeding channels of the main masses of plateau basalts (Karolus in Forgáč, Karolus, Konečný 1967).

The original continuous plateau was dissected by the later tectonic processes into a row of isolated relicts in the form of table mountain.

Monogene scoria cones

The scoria cones of small size formed by accumulation of incoherent to weakly consolidated material of the scoria nature with areas filled with minor lapilli material, are the product of explosive activity of final volcanism. Relicts of the scoria cones are preserved sporadically (Ragáč near Fiľakovo), in predominant part they were completely destructed.

Forms of this type are built by the activity of "fire fountains" (the term suggested by Dana 1888; or pyroexplosion — Cotton 1959). Explosions of hot fluid lava were conditioned by periodic expulsion of gas mounting in the lava column, expanding in the surface by explosion, throwing out pieces of hot lava. The scoria material is after falling down accumulated in the nearest area, forming irregular heaps (scoria piles) of roughly conic form.

The existing relicts indicate that the steep conic forms built predominantly of the activity of "fire fountains" had monogene character and smaller size.

Polygene forms (cones of mixed type)

As an example may serve the well-preserved relict of the youngest volcanic forms — Pútikov vŕšok (elevation point 405 near Tekovská Breznica; Fiala 1951).

The relict of the volcanic cone is situated on the crossing of tectonic lines, and represents a morphological elevation formed by the scoria, ash, lapilli material and by numerous bombs. In its summit part there are no distinct traces of crater. On the northern slope there are relicts of thick flows, at southern foothill there are denuded parts of feeding channels. On the NW foothill of the scoria cone, near the Hron river, there are three flows, the longest of them reaching and covering the gravel terraces of the Hron river. In convenient exposures there are some flows formed by other partial flows separated by the strata of scoria and bombs, indicating the characteristic structure of polygene character.

Volcanic bombs have isometric to flattened forms, unipolarly and bipolarly shaped in the course of their flight through the atmosphere. Half-plastic bombs, deformed when fallen, acquire characteristic forms (cow dung). In eruptions of whole chains of bombs at differentiation movements, complicated rolled forms arise.

Subvolcanic Forms

The dyke bodies do not reach greater size. The dyke forms with great directional length, indicating mechanism of fissure eruptions (i. e. along the fractures of regional extent) were not determined. The most frequent cases are short dyke of several ten m, or the dyke swarms and systems characterized by diverging of greater number of dykes (multiple dykes), and forms with elliptic, to roughly circular diameter, indicating forms of the neck type.

Dykes, groups of dykes, multiple dykes systems and necks

The dyke bodies have either the form of isolated bodies of small size or groups, concentrated over small areas. Isolated bodies represent sizes of several m, with the followable length to 100 m. As an example may serve basalt dyke (nepheline, basanite, near the village Kyshýbel), penetrating andesites of the IIIrd phase. Columnar jointing is oriented perpendicularly to the contact planes. A similar dyke body of basaltoid andesite in the northern part of the Žiar depression is penetrating through the gravel-conglomeratic formation. These bodies were probably connected with the activity of sporadic importance, with the forming of small monogene forms.

In the eastern part of the neovolcanic region, with the maximum activity of the final volcanism, there may be observed the grouping of isolated dike bodies or their complicated diverging on smaller areas. The result of this is a complicated system of multiple dykes. Near the village Šurice, there is a preparated body of basalt tuffites, penetrated by several dyke bodies of limburgite basanites (Mihalíková 1959, Karolusová 1959).

DESTRUCTION OF VOLCANIC FORMS

In the development of volcanic regions, processes of construction and destruction are active. Processes of construction pass in stages of the accumulation of volcanic masses in the time of volcanic activity, and represent positive process for the building of volcanic forms.

The stages of destructions are an accompanying phenomenon of volcanotectonic processes (destructions at enormous eruptions, collapsing of some parts of apparatuses, extrusions of cumulodomes, etc.), and a consequence of the surface exogeneous factors active in the course of volcanic activity, and especially after the ending of the activity.

The intensity of destruction of volcanic forms is directly proportional to the duration of the activity of exogeneous destruction factors, i. e. intensity is greater the longer the volcanic forms are affected by erosion and denudation.

When considering the duration of subsequent volcanism (Lower Tortonian — Pliocene) and the following final volcanism (Pliocene-Quaternary), we may state that destruction could have affected not only volcanic regions as a whole, but that it could have reached high stages also within the separate stages of the activity (volcanic phases), still before the deposition of products of the next volcanic stage.

Before starting the discussion of the destruction of volcanic forms, it is necessary to determine — in the sense of Cotton (1959), Rear (1957) — the main stages of destruction of volcanic forms of the stratovolcanic type, and to point out the manner of application in regions of subsequent and final volcanism.

The *primary stage* ("volcano stage") is represented by undestructed volcanic form (composed cone) immediately after the calming down of the volcanic activity.

In the *next stage* the slopes of the cone are grooved by erosion. The grooves are radially reaching farther from the central area. On volcanoes of the stratovolcanic types these grooves serve as the ways of movement of the lava flows. The parts, separating the individual lava flows formed by incoherent pyroclastic material with respect to smaller resistance against erosion, are removed by the continuing erosion, and the flows are preparated, and separated by deep-cut valleys in the next stage. This stage represents the proper commencement of the inversion of relief. Sectors of the original cone separated by the deep-cut valleys and grooves, are called "planezes", and this destruction stage — the "*Planeze stage*".

The higher stage of denudation "*residual mountain stage*" was defined by Rear (1957). Erosion gives rise to relict forms (residual hills) without indications of the original cone forms, except some preserved relicts of their basal parts.

In the *final stage* — "*Skeleton stage*", the upper parts of the volcanogene complexes are removed, and the more resistent subvolcanic bodies of the feeding system (necks, dykes, multiple veins) are separated.

When carrying on the reconstruction study of the volcanic forms of the subsequent and final volcanism, it is necessary to consider the fact, that destruction proceeded into higher stages, being accompanied by the inversion of relief in regional extent.

Volcanic forms of the Ist andesite phase were most strongly affected by destruction. In the eastern part of the region, extensive isolated relicts of andesite bodies (Karanč—Šiatoroš) were preparated.

Accumulations of brecciated material in the area of the piercing out of the extrusive brecciated bodies in submarine environment were extensively redeposited and resedimented on the periphery, in the sublittoral to littoral zones of the sea. Large tuffitic-sandy complexes with rounded clastic material were formed by the redeposited material, on the base of the volcanogene complexes of the central-Slovakian neovolcanic region.

After the ending of the stage of volcanism of the Ist andesite phase, in the period of intravolcanic denudation stage, the region was peneplanized in its SE part (in the eastern part the thickness of the original complexes was reduced to 150—50 m), and the extensive flat relief was formed there. That means that the stage of the high degree of destruction was reached already before deposition of the products of the next stage of volcanic activity.

The development and destruction of volcanic forms of the following volcanism of the IInd andesite phase took place partly in subaquaeous (sublittoral to littoral zone and in an environment of isolated lacustrine embayments), and partly in a terrestrial environment.

Volcanoclastic material transported from the area of emerged zones of the volcanic region is accumulated in the near-shore areas of basins and of the continental brackish basins. In the near-shore parts of the area thick tuffitic-sandy complexes (exceeding 250 m) of the beach type arose (Hrušov, Rykynčice). Farther from the shore, i. e. in the deeper parts of the basin, finer material with the strata of pumaceous-ash tuffs of the explosive stages of volcanic activity was deposited. Incoherent material forming stratovolcanic apparatus on the volcanic slopes is mobilized, being oversaturated by water, giving rise to the flows of the lahar type. Muddy flows replace volcano-clastic material into deeper parts of the sedimentary basins.

In some parts of the region, the destruction reached the final stage, bodies of subvolcanic levels are denuded and preparated (dike and neck bodies — Fabova hoľa, Muránska planina plateau, subvolcanic bodies near Vígľaš, dike body near Seleštány, neck and dike bodies near Čelovce. etc.). Destruction corresponds to the final stage — the Skeleton stage.

Morphologically differentiated relief in the central part is pointed out by the forming of intravolcanic basins with tuffitic to tuffitic-lignite sedimentation. Further extensive destruction of the forms of the IInd andesite phase in the central area was conditioned by the piercing out of the bodies of the IIIrd andesite phase and the IIIrd ryholite phase. Fiala (1961—1965) pointed out to the process of destruction of complexes of the second eruptive phase (IInd andesite phase) in connection with the piercing out of the tholoid body (Veľký Chom). Fragmentary material produced by the processes of progressive disintergation, in the course of the ascending of the bodies is accumulated into adjacent basins having often a subsiding tendency.

Although the basic feature of the development of forms of the IIIrd andesite phase (A_3), the accumulation of highly viscous lava flows in the area of feeding channels, accompanied by the rise of forms of the cumulodome type, the morphological aspect cannot be used for their determination with respect to considerable denudation. The denudation — still before the activity of andesite volcanism A_4 conditioned the rise of flat relief in some segments. Simultaneous morphological elevations are conditioned first of all by the selective character of erosion of heterogenous bodies of the IIIrd andesite phase (A_3), formed by masses of homogenous andesites with the zones of brecciation.

Distribution of the lava flows of the next andesite volcanism A_4 in some parts shows the character of deposition on comparatively flat relief, which is indicating the denudation stage following the formation of the bodies of the IIIrd andesite phase. The whole extent of andesite volcanism A_4 is not yet known. According to relicts forming the top parts of the existing morphological elevations it is evident, that it has a regional character. The lava flows, forming the top parts (the highest elevation points in the central part of the region, e. g. 1404 — Sitno) and making them look like flat summits or prolonged plateau, separated by deeply cut valleys of the U or V profile of the canyon- like type, represent quite a classical type of the inversion of relief. Morphological relief built of the products of andesite volcanism A_4 indicates the stages called Residual mountain stage and the Skeleton stage.

Basalt volcanism, present mainly in the eastern part of the region, took place in slightly morphologically dissected neovolcanic region. That evoked the planar extension of basalt masses which formed plateau; while sporadic activity in the central-Slovakian area passed in deeply destructed volcanic relief.

By tectonic destruction of the sheets of plateau basalts disposition for further division was offered. This was done by erosion which evoked the rise of table mountains near the margin rimmed by the rock blocks. The scoria cone forms were distinctly destructed, but for some sporadic relicts (Ragáč) preserving the character of primary form. Denudation of the cone forms on some places in subvolcanic levels exposed the dyke and multiple dyke systems, forming

— in some cases — an armature in the tuff complex, that became thus proof against erosion (Šurica; Karolusová 1959).

In the destructed volcanic relief, relicts of cones indicate — in addition to the destruction stage of the Residual mountain type and the Skeleton type — also the younger destruction stage with preserved parts of the primary cones, i. e. the Planese stage.

Morphologically best preserved relict is represented by the youngest volcanic form — the Pútikov vŕšok, with distinctly preserved relicts of the original conic form, in the lower part grooved by erosion.

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CHRONOLOGY AND TECTONIC BACKGROUND OF THE NEOGENE VOLCANISM IN EASTERN SLOVAKIA

Abstract. Systematic geological studies carried out in eastern Slovakia during these fifteen years have enabled us to develop a synthetizing conception of the trend, location and succession of volcanic events.

The distribution of the volcanic activity along the periphery of the Neogene of East Slovakia suggests that the volcanism was connected with those geological processes that modelled the Transcarpathian Inner Deep, a part of which is the East Slovakian Neogene. From this point of view, the Neogene subsequent volcanism of East Slovakia is only one of the forms of the tectogenetic development of the area. It may be regarded as a manifestation of the synchronous tectonics of a deeper layer of the Earth's crust, i. e. of the tectonics connected with the uplift of the depositional area of the Flysch geosyncline and the formation of the Inner Deep.

Succession of volcanic events

The Burdigalian volcanic rocks are represented by fine to medium-grained crystallo-vitroclastic rhyolite tuffs deposited in the subaqueous brackish environment of the Čelovce Formation sedimentary area. They have an apparent thickness of 3 m and occur in the basal part of the sequence bearing the following fossils: *Polymesoda brongniarti* (Bast), *P. b. costulata* (Coss. & Peyrot), *P. b. allongée* (Coss. & Peyrot), *Theodoxus (Th.) gregarius* (Thom.), *Pirenella hoernensis* (Schaff.) and abundant representatives of the genus *Chillostomella* and *Cibicides ungerianus* and relatively frequent *Lagenidae* (Slávik et al. 1960). These finds are evidence of the Burdigalian age. Similar layers of montmorillonitized and seladonitized fine-grained rhyolite pyroclastics have been found in the Burdigalian beds near Terňa by J. Březina (1960). According to Zipser (1817), fragments of obsidian have been collected in Prešov. Uniform thickness and good sorting of material point to the aeolian transport of the volcanic material. The presence of one to two millimetres large fragments of non-porous felso-spherulites indicates that the source of material was not very distant. The same may be inferred from the obsidian fragments (Zipser 1817) found on Táborisko (in the area of the town Prešov) which is of Burdigalian age.

Thin (maximum 5 m) layers of rhyolite pyroclastics nowadays bentonitized fine-grained rhyolite tuffs crop out in the vicinity of Fintice village. Their subaqueous deposition is evidenced by the macrofauna *(Cardium* sp., *Ostrea* sp., *Macularia* sp., *Planorbis* sp., *Pitaria* sp., *Venus* cfr. *haidingeri* Hoern.; Švagrovský 1962); microfauna is represented mainly by *Nonion commune* d'Orb. and *Rotalia beccarii* L., *Robulus* sp., *Spiroplectammina carinata* d'Orb. (Gašparíková 1962), which are indicative of the Lower Miocene age and the alternation of marine and brackish environment.

The Karpatian age of volcanic activity has been proved by several layers of rhyolite pyroclastics interbedded in the complex near Fintice. The thickest of the three layers established is several tens of metres thick. It consists of breccias, lapilli tuffs, xenotuffs and tuffaceous sandstones with seams and intercalations of sedimentary rocks. Kuthan (1948) also records a subordinate amount of effusives from this complex. The rock is thought to be the material of the near volcanic centre deposited beneath water. The volcanic activity was presumably of the Vulcanian type. The presence of pyroxene andesite fragments in the rock suggests that a complicated petrochemical regime already existed during the magmatic activity.

The intensity of volcanism began to increase abruptly in the Lower Tortonian and so did the share of volcanic rocks in the filling of the basin. Volcanic deposits make up five to six layers of rhyolite tuffs, related in petrography and petrochemistry. The thickness of individual beds ranges from five to hundred metres. Of the greatest thickness is the Hrabovec layer of rhyolite tuff, which is traceable continuously from Šarišská Poruba (Slávik 1967) through Nižný Hrabovec, Oreské as far as the foot of Boroľa hill near the frontier of the USSR. Over its whole length, the thickness varies from 50 to 150 metres. This layer of a fairly constant thickness has been established throughout the area of the East Slovakian Neogene.

Gašparíková (in Slávik 1964) records the following micropalaeontological finds from the beds close above the tuff: *Pullenia miocenica* Kleinpel, *Globorotalia scitula* Brady, *G. bulloides* Orb., *Bulimina elongata* Orb., *Dentalina acuminata* Rss., *Orbulina suturalis* Bronn., *Elphidium* cf. *flexuosum* Orb., *Globigerinoides triloba* Rss., *Nonion pompilioides* Ficht. & Moll., *Rotalia beccarii* L., *Cassidulina* sp., *Bolivina* cf. *dilatata* Rss., *Nonion commune* Orb., *Cibicides lobatulus* Walk. & Jac., *Lagena* cf. *hispida* Rss., *Uvigerina* cf. *acuminata* Hoss, *Bolivina hebes* Macf., *Dentalina* cf. *fenicostata* Cush. & Benn., *Robulus inornatus* Orb. The microfauna including *Globorotalia foshi barisanensis* LeRoy, *Globorotalia scitula* Brady, *Turborotalia mayeri* Cush. & Ell., *Globigerina bulloides* Orb., *Globigerina concinna* Reuss, *G. diplostoma* Reuss, *G. opinata* Pisch, *G. praebulloides* Blow, *G. tarchaensis* Subb. & Chutz, *G. bisphaericus* Todd, *G. immaturus* LeRoy, *G. obliquus* Boll., *G. rubrus* Orb., *Globigerinoides trilobus* Reuss, *Globoquadrina altispira* Cush. & Jac., *G. quadraria*

Cush. & Appl., *G. rotundata* (Orb.), *Orbulina biloba* Orb., *O. suturalis* Bronn., *O. transitoria* Blow, and *O. universa* Orb. has been collected immediately below the tuff layer.

Benthos is represented by *Robulus calcar* Linnaeus, *R. cultratus* Montf., *Bulimina* cf. *elongata* Orb., *B. pupoides* Orb., *Uvigerina macrocarinata* Papp & Turn., *U. pygmoides* Papp & Turn., *Bolivina hebes* Macfayden, etc. (Gašparíková—Slávik 1967).

The above-mentioned associations prove that the Hrabovec tuff layer lies at the boundary between the Tortonian s. s. and Lanzendorfer series and that it may be used as a key horizon. The rock is a light-grey and grenish pelitic to psammitic pumice-like tuff of vitroclastic texture with a negligible amount of clasts (andesine, quartz, biotite, ilmenite, zircon, apatite). The regional distribution of ash deposits and gradual changes of the thickness point to explosions of the Plinian type probably from subaqueous fissure volcanoes. The centres of this type of volcanism are unknown on the present-day surface. The analysis of granulometric composition leads to the conclusion that the centres are located inside the Tortonian sedimentary area and that they show affinity to the main longitudinal fault systems that existed at the time when the sedimentary area of the marine Lower Tortonian was developing. It is debatable whether these products may be connected with the subvolcanic bodies of rhyolite between Merník and Majerovce (Leško 1955; Buday in Matějka 1964), as the latter show a distinct affinity to the younger tectonic lines. The Lower Tortonian rhyolite pyroclastics are also known from the southern border of the Miocene sedimentary area, from the north-western margin of the Zemplin Palaeozoic; they are traceable along its periphery (Merlič et al. 1967) and occur even in facies indicating the proximity of volcanic centres.

In the marine Upper Tortonian, the rhyolite volcanism persisted dominantly in the southern part of the area; volcanic products in the central part of the basin are of substantially smaller thickness and occupy a position characterized by the rich microfaunal association with *Uvigerina hispido-costata*, *Bathysiphon filiformis*, *Bulimina intonsa* and *B. ovula* (Čopianová 1966). The lowest Upper Tortonian pyroclastics have been ascertained in the area along the frontier with the USSR and (by boring) in the *Spiroplectammina carinata* Zone (zone with agglutinants) around Stretava (Čverčko in Merlič et al. 1967). The rhyolite tuffs of the Upper Tortonian Bolivina-Bulimina Zone have already been mentioned by Seneš—Švagrovský (1957). Čechovič (1963) classed with them the rhyolite pyroclastics from the vicinity of Dvor Mária, for which J. Březina claims the direct deposition into aqueous environment. Volcanic material consists of pumiceous, fine to medium-grained vitroclastic tuffs with β-quartz and plagioclase. Of interest is the presence of pyroclastic rock with fragments of biotite rhyolite in the Bolivina-

Bulimina Zone near Trebišov, as encountered by deep borings. Dark coloured pyroxene andesites in the upper, palaeontologically evidenced part of the Bolivina-Bulimina Zone (Čverčko op. cit.) have been newly found in the areas of Žipov and Kráľovský Chlmec. The andesite complex is about 100 metres thick. A thin veneer of andesite tuff on the pyroxene andesite proves its surface-volcanic (probably also subaqueous) character. This finding is in good agreement with the data on the andesite boulders in the freshened Upper Tortonian deposits (Čechovič 1963). Surprisingly, no contemporaneous volcanicity is known from the northern periphery of the basin. The tectonic and volcanic rest is demonstrated there by the deposition of salt bodies.

In the freshened Upper Tortonian complex (Rotalia Zone) the volcanic products appear in the northern part of the Prešov—Tokajské pohorie Mts. and extend far to the south, into the Košice depression. They are dated on macrofaunal evidence. The fossils include *Nucula nucleus* (L.), *Arca (Anadara) diluvii* (Lam.), *Aloidis gibba* (Olivi), *Clithon pictus tuberculatus* (Schreter), *Turbonilla spiculum bergeria perpusilla* (Grat.), *Nassa adac* (Boettg.), *Nassa schöni* (Hoern. & Auing), *Nassa dujardini* (Pesh), *Clavus pustulatus* (Brocc.), *Mitrella scripta* (Linn.), *Clavatula contorta* (Švagr.) (Švagrovský 1964).

Rhyolite tuffs of this facies are pumiceous lapilli tuffites in the peripheral areas which pass into medium-grained facies. They crop out on the surface around Zamutov, where their maximum thickness has also been established. In this area they are distinguished by intercalations of sediments bearing the redeposited Tortonian fauna with abundant Silicispongiae. The Bolivina-Bulimina Zone has been safely found at some distance beneath the tuffs. The presence of rhyolite lavas and coarse pyroclastic rocks reveals that the volcanic centre was situated near Zamutov. Čverčko & Rudinec (op. cit.) found fine-grained tuffites at an equivalent position in the area near the USSR frontier, to the north of Čierna nad Tisou.

The boring Svinica 2 at the eastern margin of the Košice depression struck a layer of pyroxene andesite in the freshened complex (Čverčko in Merlič et al. 1967). This occurrence has so far been unknown and may be regarded as evidence of the Upper Tortonian age of the andesite volcanism.

Volcanic activity increased abruptly in the Sarmatian. The earliest Sarmatian volcanism probably gave rise to the sandy andesite tuffites in the Olšava Beds (after Švagrovský, 1964). Their rich macrofaunal content comprises *Abra reflexa* Eichw., *Congeria moesia* Jekel, *Ervilia dissita* Eichw., *Caliostroma planata sternia angulata* Eichw., *Bittium reticulatum* Costa, *Clavatula doderleini* M. Hoernes, *Clavatula striata* Fried, *Dorsanum duplicatum* Koles, *Mohrensternia banatica* Jekel, *Cerithium gracile* Sim. & Barbu, *Pirenella picta mitralis* Eichw., and *Acteocina lajonkaireana*

Bast. The hypothesis on the beginning of the andesite volcanism in the earliest Sarmatian has been confirmed by the find of 50 m thick beds of altered sandy pyroxene andesite tuffites in the sediments with the mass occurrence of *Cibicides badenensis* (Orb.) (Jiříček 1966).

The Sarmatian andesite volcanism was at its maximum in basal parts of the Zone of large Elphidia, when it produced a longitudinal belt of andesite strato-volcanoes which at present form a buried range in the south of the East Slovakian Neogene area (Merlič et al. 1967). The height of the buried volcanoes exceeds 1000 m; their age has been reliably determined on the Malčice body, where the volcanic complex is underlain by the Cibicides horizon and overlain (according to R. Jiříček and K. Čopianová) by the Zone of large Elphidia. The association is represented by *Elphidium reginum* Orb., *Elphidium josephinum* Orb., *Cythereidea hungarica* Zalanyi, *Nonion bogdanowiczi* Volosh., *Ammonia beccarii* (L.), *Hydroides pectinata* Phill. and *Cibicides badenensis* Orb.

The upper part of the Zone of large Elphidia is probably represented by Švagrovský's (1964) Myšľa Beds which yielded abundant macrofaunal species as follows: *Solen subfragilis* Eichw., *Ervillia dissita* Eichw., *Cardium vindobonense jekeliusi* Papp, *Cardium pseudoscoreni* Jekel, *Modiolus incrassatus* Orb., *Muscullus sarmaticus* Gatt., *Gibbula andrusovi* Švagr., *Clithon pictus* Fér., *Rissoa soceni* Jek., *Mohrensternia pseudoinflata* Hilb., *M. styriaca* Hilb., *Hydrobia elongata* Eichw., *Odostomia fuchsi* R. Hoern, *Valvata sarmatica* Papp, *Mohrensternia hydrobioides* Hilb., *M. pseudoangulata* Hilb., *Pirenella picta mitralis* Eichw., *P. nodosoplicata eichwaldi* Hilb., *Terebralia bidentata lignitarium* Eichw., *Polynices catena sarmatica* Papp, *Ocinebrina sublata* Bast., *Retusa truncatula sarmatica* Berber, *R. truncatula pappi Berger*, *Acteocina lajonkaireana* Bast., etc. These beds also contain numerous layers of pumiceous and lapilli rhyolite tuffs. To this zone may belong also rhyolite pyroclastics and ignimbrites found near Komarovce in the Košice depression by Vass (op. cit.).

Analogous rhyolitic rocks have been encountered in the borings in the vicinity of Ptrukša. An identical chronological interpretation may be inferred from the occurrence of the rhyolite tuff with andesite pebbles at the western periphery of the Prešovské pohorie Mts. Stratigraphical correlation is rather difficult owing to the progressive freshening of the area of deposition after the beds of the Zone of large Elphidia had been laid down.

The relatively well-traceble *Elphidium hauerinum* Orb. Zone (Upper Volhynian; Buday in Matějka 1964) itself bears beds of rhyolite tuffs and breccias, up to 30 m thick, in the area of Kráľovský Chlmec. The bentonitized rhyolite pyroclastics from the neighbourhood of Lastovce may also belong to it. The Hauerinum Zone is overlain by a complex of agglomerates, tuffo-breccias of pyroxene-hornblende andesites which alternate with lava flows in the Sub-Vihorlat basin.

Superjacent on this volcanogenic complex are the sediments with the following fauna: *Cardium gracile plicatofitoni* Sinz, *C. politioanei* Jek., *Musculus sarmaticus* Gat., *Irus naviculatus* Andr., *Cardium latisculus jammense* Hilb., *C. rineiseni* Jek., *C. vindobonense jekeliusi* Papp, *Replidacna cf. carasi* Jek., *Macha vitaliana vitaliana* Orb., *Cardium vindobonense breviformis* Papp, *C. gracile bessarabiensis* Gillat and *Donax lucidus* Eichw. (Seneš 1957). The microfaunal assemblage consists of *Nonion granosum* Orb., *Elphidium* aff. *crispsum* L. and Ostracea from the *Cythereis* group. This layer may be considered as the lower part of the Bessarabian or, in the sense of the present micropaleontological correlation for the East Slovakian region, as part of the *Porosononion subgranosum* Zone. On the basis of the petrographical and spatial affinity of the semi-intrusive and intrusive bodies of hornblende-pyroxene diorite-porphyrites, occasionally garnetiferous, with the above-mentioned volcanic rocks intervening between the Hauerinum and Porosononion Zones, we regard them as equivalent in age. In our opinion, the occurrences at Oblík, Kura Hora, Malá and Veľká Straž, and others are of the same date.

To establish a precise stratigraphical correlation of the Pliocene complexes is a complicated problem. Consistently with a number of authors (Čechovič 1964; Buday in Matějka 1964; Sheremeta 1966), we place the Sarmatian/Pliocene boundary at the end of the Bessarabian but, contrary to the former conception of Brodňan et al. (1964) and Buday (1964) we do not include the garnet bearing tuffite series and the upper coalbearing series into the Sarmatian. According to the recent boring results from the eastern part of the East Slovakian Neogene, the former series lies with a distinct unconformity on a number of stratigraphical horizons and begins with conglomerate beds, especially in the marginal tracts. Therefore, we range this garnet bearing tuffite series, represented chiefly by garnetiferous pumiceous tuffs, to the Pliocene in the area studied (Sasin, Slávik 1967). When correlated with the sequences of the Transcarpathian region (Sheremeta 1966), they would correspond to the Meotian, at the earliest. This fully agrees with Čverčko's opinion (1966) on the Pannonian age of the garnet bearing tuffite series. The 30—40 m thick tuffogenic complex increases in thickness (to 70—80 m) and grain-size in the proximity of the rhyolite bodies. Coarse-grained rhyolite tuffo-breccias fringe necks and thinly laminated cupola-shaped and subvolcanic bodies of rhyolites (plagio-rhyolites) at Beňatina, Michalovce, Lesné, in the Prešovské pohorie Mts. (Zlatá Baňa, a. o.)

The age of young pyroxene andesite bodies of the Vihorlat and Prešovské pohorie Mts. still lacks reliable faunal evidence. On the basis of lithological correlation with the West Carpathian Pliocene sections, J. Čverčko arrived at the assumption that the andesite volcanism is Pontian-Levantinian in age. The safely proved Levantinian age of andesite tuffs of Albinovská hôrka indicates that volcanism was active in the Upper Pliocene even in the Prešovské pohorie Mts. The inaccuracy of the stratigraphical term "Levantinian" led Sheremeta (1966) to dating the Vihorlat volcanogenic complex as Apsheron-Akchagylian. In the coal-bearing series with interbeds of andesite pyroclasts the Upper Pliocene fauna with *Candona albicans* Brady, *C. angulata* G. W. Müller, *C. candida* O. F. Müller, *Iliocypris* cf. *gibba* Raul, *Cyprideis* sp., a. o. has been collected (L. V. Buryndina 1966).

The comparison of the age and succession of volcanic rocks shows that the volcanic events of central and western Slovakia, i. e. the West Carpathians themselves, had a decidedly different pulse from those of eastern Slovakia. Therefore, Kuthan's attempt (1964) to develop a scheme of volcanic phases which would be valid for both areas raised considerable difficulties, as is apparent from a detailed study of his volcanic phases in the east of Slovakia (Kuthan in Matějka 1964). The term "*First rhyolite phase*" meant to designate the volcanicity between Burdigalian and the end of Tortonian calls for revision, as it comprises intensive andesite volcanism not only on Slovakian territory (Svinica, Žipov) but also in the Tokaj pohorie Mts. right beyond the frontier. In this sense, the range of the First rhyolite phase should be restricted excluding the rhyolite volcanics of Michalovce, Lesné and Vranov which are Lower Pliocene in age. Still more difficult is the correlation of andesite phases, as pointed out by Kuthan himself (in Čechovič 1964). However, it is hardly possible to accept his conclusions on the equivalent evolution of deep-seated magmatic masses in central and eastern Slovakia which imply the concurrence of the hornblende-pyroxene volcanic ($\pm$ garnet) phases. Even if this correlation is accepted, no equivalent to the buried andesite range stretching along the Zemplin-Beregovo elevation can be found in central Slovakia. Therefore, we think it more plausible to presume that a definite type of volcanic rock is the function of the tectogenic history of the area and, consequently, that the hornblende-pyroxene andesites and their subvolcanic intrusive equivalents are the products of a long-lasting development of the magmatic hearth and of the anatexis and differentiation of magma derived from it. From this consideration it follows that every large crustal block should be regarded as a separate unit showing a particular rhythm of geological and thus also volcanic activity.

Kuthan's conception of the Lower Sarmatian Second rhyolite phase should be completely altered, as the main representatives (Lesné, Michalovce, Zlatá Baňa) are decidedly of Pliocene age (Sasin, Slávik 1967). Consequently, this phase was not synchronous with the so-called First andesite phase of undation volcanism. Owing to great chronological and, to a certain extent, also terminological difficulties connected with the division of volcanism into phases (as documented by the above facts), we shall deal with this problem in an other paper.

The tectonic-volcanic relations

The close association of volcanic activity with the tectonic history of the area is, as has already been pointed out in the introductory paragraph, a very characteristic feature of the East Slovakian area.

During the post-Palaeogene period, the geological trend of East Slovakia differs in some respects from that of the Inner West-Carpathians. First, it is the rapprochement of the Miocene molasse with the geosynclinal sedimentary area of the Palaeogene Flysch. They are separated by a belt, only a few kilometres broad, that follows the longitudinal elevation of the Pieniny Klippen Belt.

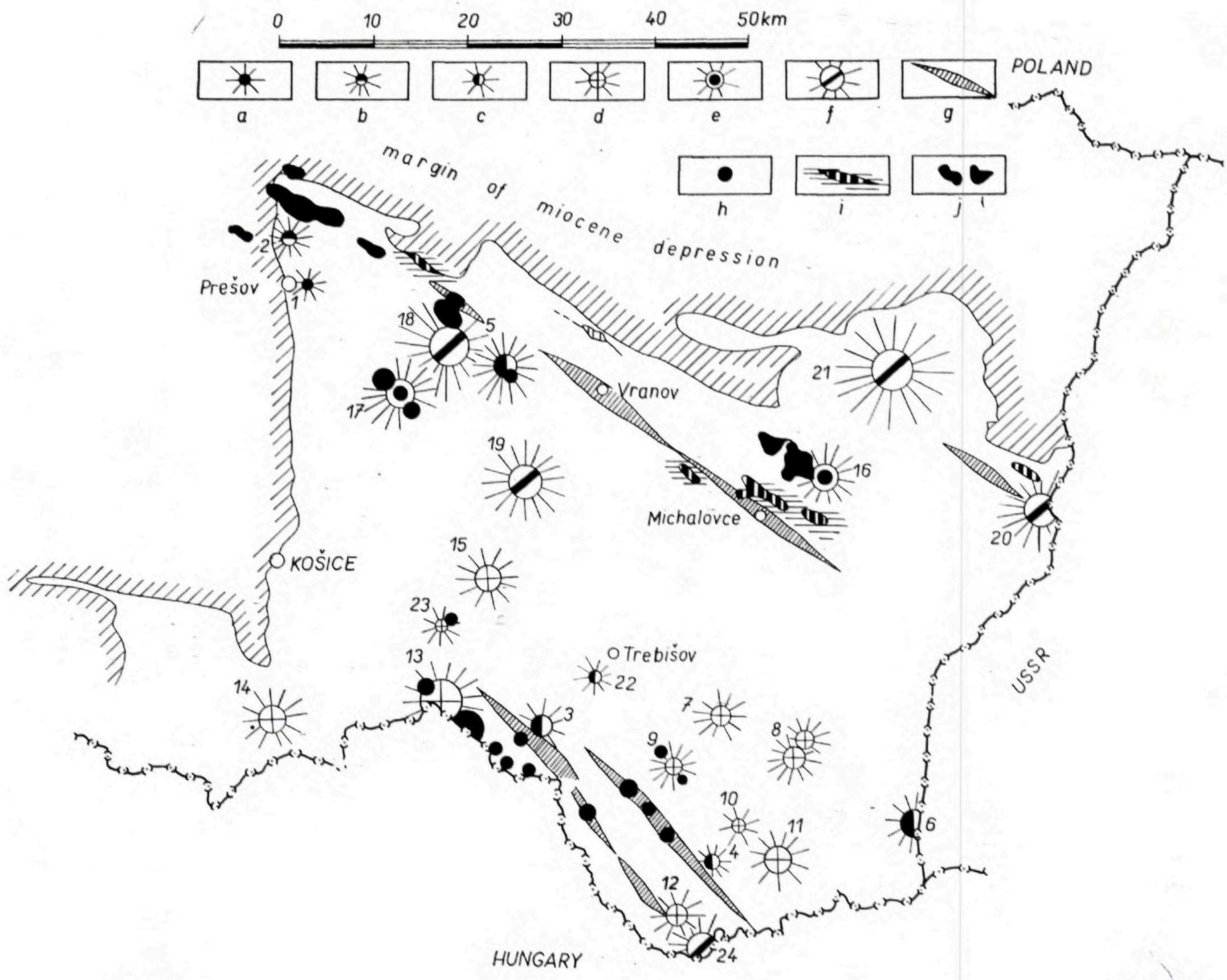
The development and the filling of the Trans-Carpathian Inner Deep depression, the part of which is also the East Slovakian Neogene, possess typical East Carpathian features as far as the Hornád fault. The most distinctive of these are as follows:

(1) The position of the "molasse" between the Pannonian block and the Flysch geosyncline;
(2) the predominantly marine to brackish character of the area persisting from the Burdigalian to the Pliocene;
(3) the presence of evaporites, mainly salt in the Miocene marine sediments;
(4) the position of volcanics and time succession of volcanic events, which link up the volcanism of East Slovakia with that of the East Carpathians.

Fig. 1. Diagram of Neogene volcanism in eastern Slovakia

a — Burdigalian volcanic centre (rhyolite), b — Karpatian volcanic centre (rhyolite), c — Tortonian volcanic centres (rhyolite, rare andesite), d — volcanic centres of the (marine) Sarmatian (andesite, subordinately rhyolite), e — volcanic centres of the freshened Sarmatian (andesite), f — Pliocene volcanic centres (andesite), g — Lower Tortonian fissure explosions (rhyolite), h — domes and subvolcanic bodies (andesite, rhyolite), i — fissure eruptions and effusions of Lower Pliocene age (garnetiferous rhyolites and their pyroclastics), j — semi-intrusive bodies and domes of Upper Sarmatian age (andesite in Upper Sarmatian).

1 — Prešov, 2 — Fintice, 3 — Žipov, 4 — Zemplín, 5 — Zamutov, 6 — Ptrukša, 7 — Malčice, 8 — Beša—Vojany, 9 — Sirník—Hraň, 10 — Bodrog, 11 — Kráľovský Chlmec, 12 — Streda nad Bodrogom, 13 — Veľký Milíč, 14 — Komárovce, 15 — Bogota, 16 — Kaluža, 17 — Kokošovce, 18 — Zlatá Baňa, 19 — Herľany, 20 — Popričný, 21 — Vihorlat, 22 — Trebišov, 23 — Ruskov, 24 — Kamenec.



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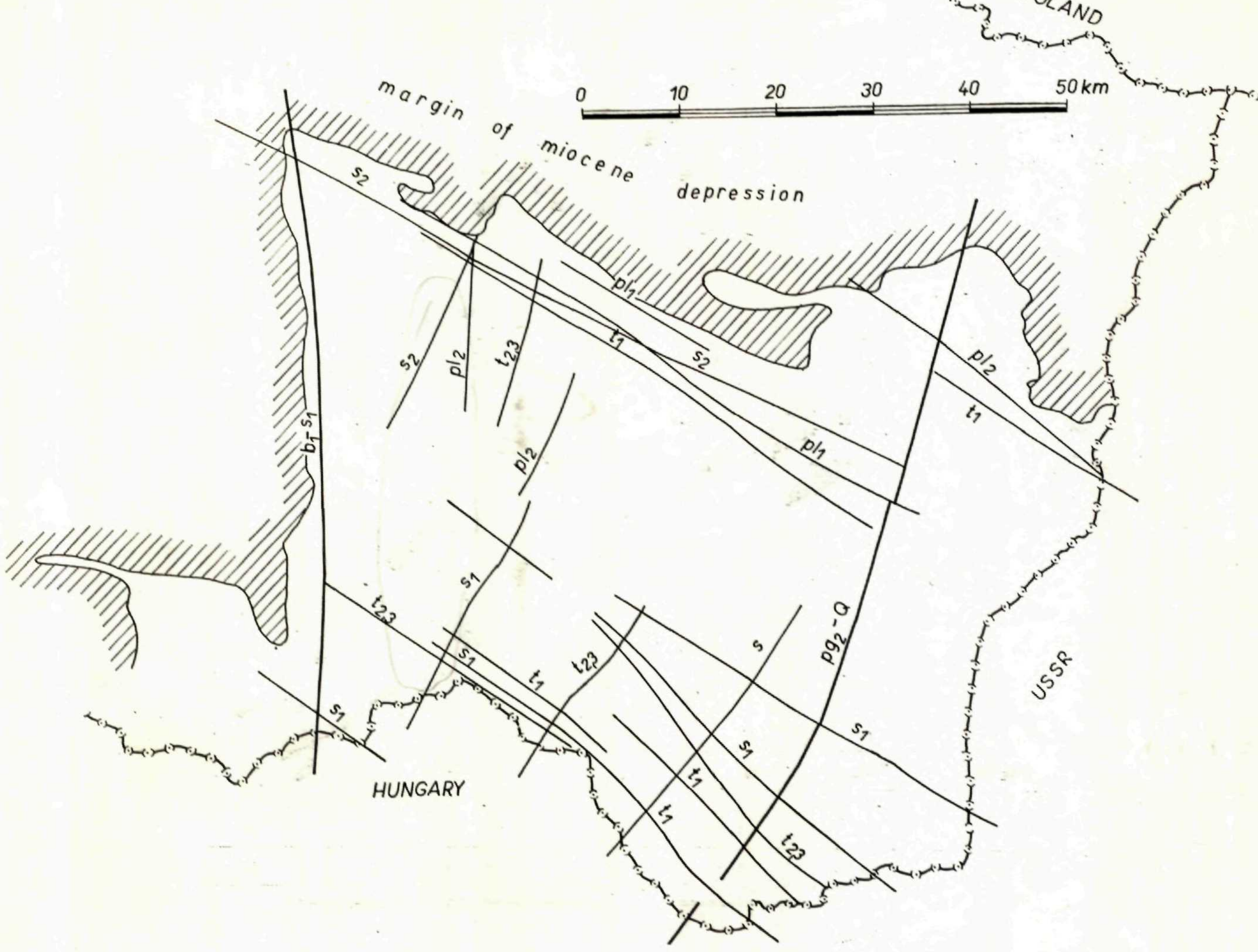
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The tectonic position of volcanic centres is inferred from their spatial relations to the major tectonic lines. The location of volcanic centres has been established by various observations, but mainly with the help of the analysis of the thickness and grain-size relations of pyroclastic materials.

The location of the centres of supply stilll remains to be solved for the Burdigalian volcanics. With a certain probability, the occurrence of obsidian on Táborisko Hill at the north-eastern outskirts of Prešov (Fig. 1) may be connected with them. The existence of a centre there is conditioned by the tectonic predisposition at the crossing of fault systems bounding the sedimentary area of the Hornád fault with some of the longitudinal faults bordering the Lower Miocene area of deposition. Indicative of this interpretation is also the centre of the Karpatian volcanism to the west of Fintice, the proximity of which is suggested by perivolcanic coarse rhyolite pyroclastics (breccias, tuffs up to xenotuffs) with sporadic subrounded fragments of pyroxene andesite. Whereas the Lower Miocene volcanicity was but rudimentary and of small extent, the volcanic activity of the Lower Tortonian shows a discrete affinity to the major boundary longitudinal fault structures and is connected with the downfaulting on the step-like fault system bounding the Transcarpathian depression in the north and on the peri-Pannonian fault system bordering it in the south. The uniform distribution of ash component, the decrease of thickness towards the centre of the molasse, petrographic homogeneity and the persistence of pyroclastics for more than 80 km on the Slovakian territory all point to fissure explosions of the linear type, presumably subaqueous for the most part. In the northern belt there are only fine-grained facies but in the south, along the present-day Zemplin-Beregovo elevation, the fissures brought to the surface in addition to coarse pyroclastics and agglutinates probably also dome-shaped and subvolcanic bodies of fluidal rhyolite whose linear arrangement agrees with the conception of linear eruptions (Cejkov, Hraň, Veľká Trňa). The symmetrical distribution of the Tortonian volcanic activity at the two margins of the Miocene sedimentary area was strikingly disturbed in the marine Upper Tortonian time, when a rhyolite volcano of central type (?) arose in the area of Michaľany, at the Hungarian-Slovak boundary of the Zemplin-Beregovo Horst. The proximity of a composite volcano is evidenced by almost 300 metres of rhyolite clastics and a lava flow of pyroxene andesite. The existence of a minor centre in the Trebišov district is suggested by biotite rhyolite encountered in borings (Čverčko —

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Fig. 2. Map of faults in the substratum of the East Slovakian Miocene

pg_2 — Palaeogene — Eocene, b — Burdigalian, t — Tortonian, s — Sarmatian, pl — Pliocene, Q — Quaternary.

Indices 1, 2, 3 are used for Lower, Middle and Upper respectively.

Rudinec op. cit.). Thick layers of pumiceous rhyolite tuffs occurring throughout this area afford additional evidence (Seneš—Švagrovský 1958). At that time, tectonic and volcanic activity in the northern part of the depression was at rest, which is demonstrated by the regional precipitation of evaporites, particularly of salt. In the brackish to freshened Upper Tortonian the tectonic rest was disturbed by the formation of a large monovolcano near Zamutov, to the east of Vranov in the northern part of the basin. The volcano produced mainly pyroclastics but also pumiceous agglutinates and lavas of porous biotite rhyolite that hints at a close nearness of the volcanic centre. Minor volcanic centres are postulated around Ptrukša near the ČSSR—Soviet frontier.

During the time of marine Lower Sarmatian, the conditions were analogous to those of the marine Tortonian. In the southern part of the East Slovakian Miocene region, extensive pyroxene-andesite volcanism was active. High strato-volcanoes were erected forming two longitudinal belts. The northern belt involves the buried volcanoes Beša—Vojany, Malčice and the lower part of the Bogota massif. The southern belt is located on the fault system that separates the Zemplin-Beregovo elevation from the central part of the Miocene depression. It consists of volcanoes whose apical parts rise to the surface. They are the following: Kráľovský Chlmec, Sírnik—Hraň and the surface edifice of the northern part of the Tokaj Mts. (the group Veľký Milič as far as the Ruskov Pass). It is notable that the volcanoes are aligned also in the transverse direction. This distribution pattern may readily be interpreted as having been controlled by the crossing of longitudinal and transverse tectonic lines. Of interest is the spatial separation of the Lower Sarmatian rhyolite volcanics from the andesite volcanoes. From the present state of knowledge it is inferrable that the Lower Sarmatian rhyolite volcanic centres follow the linear axis of the Zemplin-Beregovo elevation, right on the Hungarian—Slovak frontier. On this linear fissure, a series of pyroclastic cones (nowadays degraded) was piled up and their emplacement was followed by rhyolite effusions, dykes and necks which distinctly mark the course of the fissure. However, the rhyolite volcanism of significant extent was not simultaneous with the andesite volcanism. It was active subsequently to the latter and probably persisted up, to the time of brackish up to fresh-water Upper Sarmatian, when pumiceous tuffs and tuffites of moderate thickness were deposited. These became later the parent rock of bentonites. The Lower Sarmatian volcanism was not active in the northern tract of the basin. The surface volcanic activity did not begin before the brackish Sarmatian, from where it is known as intervening between the *Elphidium hauerinum* (Orb.) Zone and the Bessarabian Cardium horizon. Distinctive of this volcanic activity is its close affinity to the longitudinal tectonic structures rimming the Miocene area in the north as well as the abundance of intrusive (Malá Straž, Oblík, Kura Hora) and or cupola-like bodies (Lysá Stráž, Maglovec, Hrb, Lancoška) aligned along the margin of the basin. In the northern part of the Prešovské pohorie Mts. there are typical volcanic structures (Varhaňovce—Lesíček) with pyroclastic rocks, and similar but buried volcanogenic complexes belonging to the agglomerate-tuffite formation were drilled in the Sub-Vihorlat basin. Intrusive and subvolcanic bodies are distinctly connected with the longitudinal elevation zones. These are represented by the Kapušany Horst and its eastern prolongation up to Oblík in the Prešovské pohorie area and by the Humenné—Užgorod Horst in the Vihorlat area. Typical volcanic formations with pyroclastics are developed in the sections distant from the elevation structures.

The rejuvenated igneous activity at the base of the Pliocene gave rise to a string of rhyolite (rhyodacite) bodies of linear arrangement. The domes of garnetiferous rhyolite are impersistently distributed along the northern periphery of the basin (Zlatá Baňa, Merník—Čičava, Komariany, Lesné, Hrádok, Biela Hora, Zalužice, Beňatina). The linear course sufficiently proves the connection of volcanism with the fault system bounding the depression. The intrusion of magma through the plastic Miocene rocks affected the morphology of the bodies who assumed the form of necks and doms. It is indubitable that in the rigid rocks of the basement of the Miocene deposits the magmatic body would be of a linear dyke form, as indicated by structures of this type near Vranov. The time equivalent of this volcanism in the southern part of the area under consideration does not show a close connection with tectonic lines. The volcanic structure seems to be a group or, more precisely, a central volcano in the vicinity of Streda nad Bodrogom.

It is beyond dôubt that the youngest volcanism of eastern Slovakia is intimately related with the tectonic conditions; it is clearly associated with the movements of the northern margin of the Neogene depression that were active in the course of the Upper Pliocene. On this marginal weakened zone there appeared huge strato-volcanoes of the central type, accompanied by not very numerous subsidiary craters, so that the centres plotted on the map virtually represent the group volcanoes, the differentiation of which requires further detailed study. The group volcano Zlatá Baňa—Zamutov in the Prešovské hory Mts. and the Vihorlat and Popričný volcanoes in the Vihorlat area belong to them. The Vihorlat volcanism is tied up with the movements on the fault system at the southern border of the Klippen Belt; it broke up in places where this system cuts across the major transverse zone of discontinuity, i. e. the Vihorlat fault zone (Leško, Slávik 1967). The Pliocene volcanoes of the Prešovské hory Mts. are linked up with the movements along ancient fault systems which during the Pliocene led to the intensive uplift of the northerly blocks relatively to the generally shallow Pliocene residual lake situated to the south of Zamutov. A transverse structure also exists in this area, running roughly along the line of Šimonka—Dubník—Červenica. The crossing of these tectonic systems conditioned the rise of a strato-volcano, which at present is fully degraded, especially at the northern and north-eastern sides.

The submitted chronological and tectonic analysis of the volcanic activity in eastern Slovakia enables us to construct a map of disjunctive tectonic structures transecting the substratum of the Neogene basin. In compiling this map we have issued from the presumption that the volcanic bodies reflect the course of faults in the deep parts of the Earth's crust and, as a result, they are not necessarily consistent with the tectonics of the plastic cover. The comparison of the tectonic structure of the lower level with that of the upper level may also help to clear up how the disjunctive surface dissipates from the deep-seated rigid block into the plastic cover.

The above-mentioned succession of volcanic events and their connection with the tectonic history suggest a few conclusions on the dynamics of the volcano-tectonic process. Volcanism is the function of the tectonic activity of the area and may be regarded as the manifestation of the tectonics of the plastic parts of the Earth's crust. The penetration of volcanic eruptions up to the surface depends on the formation of faults which in this type of volcanicity constitute a system of ascent. As is evident from the above-said, the marginal faults of the depression or elevations play the most important role.

The migration of volcanic centres shows a striking feature. Within the crustal block including the East Slovak Neogene, volcanic activity was synchronous in the longitudinal direction and displayed some differences across the Transcarpathian interior depression. In broad outline, the migration of volcanic centres from the Pannonian massif towards the Flysch geosyncline is observable. The shift is most pronounced when the volume of piled-up volcanic masses, mainly since the beginning of the Upper Tortonian, is taken into consideration.

The appearance of volcanic masses at the Earth's surface is not a function of the increase of magma chambers but of tectonic activity. Having analyzed the relationship of the dynamics of fault movements to the volcanogenic formations, I came to the conclusions that volcanic bodies are not erected on the surface during the periods of orogenic activity; thus, for example, the periods in which the Magura Flysch sedimentary area underwent compression and which in the interior depression are characterized by the interruption of sedimentation and by angular unconformities, are free of volcanic events. In contrast, the peaks of volcanism coincide with the periods of the strongest subsidence of basins. Volcanic centres show a definite affinity to the synchronous marginal master faults. Another relevant problem to be solved bears on the reason for the changes in the chemical composition of volcanic rocks. Geological findings lead to the conception of one, more or less uniform magmatic hearth of a uniform petrochemical development, which had been reached by faults at different intervals of its duration. The petrogenetic evolution of volcanic rocks in the regional sense is conceived as follows: The magmatic hearth was generated concomitantly with the upheaval of Flysch geosyncline and with the evolution of the Transcarpathian Inner Deep during the Neogene. Owing to the processes of magmatic differentiation, until the end of the Lower Tortonian only the acid differentiates of the mantle of the magmatic hearth appeared on the Earth's surface. The production of andesites in the Lower Sarmatian indicates the supply of less differentiated pyroxene-andesite magma. As a result of the related tectonic unrest, the proceeding differentiation and assimilation of this magmatic portion gave rise to garnet-hornblende-pyroxene andesites at the northern boundary of the molasse towards the close of the Sarmatian. I presume that the Pliocene rhyolites with analogous garnet variety are the product of a subsequent differentiation of the same magma.

The termination of volcanism in the northern tract of the basin is distinguished by the ascent of low-differentiated pyroxene-andesite magmas, up to basalt-andesites in the final phase, which furnish evidence of the cratogenization of the area.

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Geologické práce, Zprávy 44–45. Bratislava 1968

JÁN SLÁVIK–JOZEF ČVERČKO–RUDOLF RUDINEC

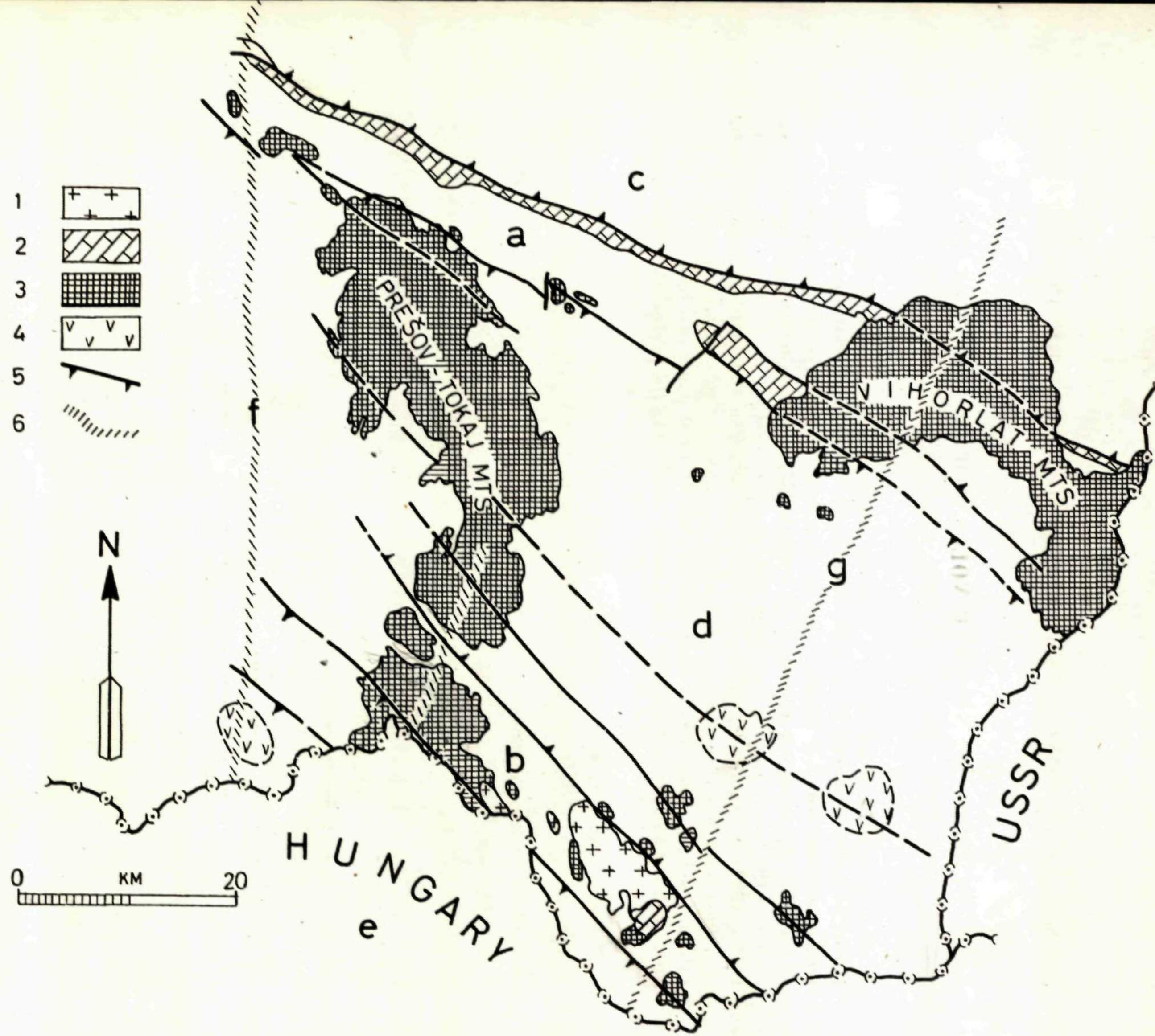
GEOLOGY OF NEOGENE VOLCANISM IN EAST SLOVAKIA

Abstract: Volcanic activity in the Neogene of East Slovakia has attracted special attention of geologists since the fifties of this century. The results of investigations by numerous authors are summed up in the papers by M. Kuthan (1948), J. Seneš & J. Švagrovský (1957), J. Janáček (1959), J. Slávik (1967) a. o. The most recent synthetizing review of the East Slovakian volcanism has been presented by M. Kuthan (in A. Matějka 1964; V. Čechovič 1963; B. Leško 1963). The systematic modern research, combined with numerous borings, furnished a number of new data which modify the existing opinions. These new findings are the object of our paper (for brevity, the earlier views will not be discussed, as their evolution is described in the papers cited above).

Geological-tectonic conditions of volcanic activity

East Slovakia represents one of the most suitable areas for the study of interrelationships between volcanism and tectonics. This is clearly shown by the connection of volcanic mountain ranges with significant tectonic zones which separate units of the first order of magnitude. The first of these is the elevation zone that separates the geosynclinal sedimentary area of the Magura Flysch from the Miocene molasse and the second — the elevation zone separating the Miocene molasse (Transcarpathian Inner Deep) from the Pannonian massif. These elevation zones are called the Peri-klippen elevation* and the Zemplín-Beregovo Horst, respectively. The elevation is bounded in the north by a steep thrust plane (Aldemirov fault system; Gofštejn et al. 1966) intervening between the zone of Pieniny Klippen and the Palaeogene sedimentary sequences of Magura. In the south, it is limited by the system of normal faults which form part of the mobile zone called the Peripieninian Lineament. The Zemplín-Beregovo elevation is a regional longitudinal horst structure constricted between the Myslava and the Chlmec faults (in the sense of

* The term "Peri-klippen elevation" is meant for the longitudinal elevation block situated between the geosynclinal area of the Magura Palaeogene and the Trans carpathian inner depression which also comprises the East Slovakian Neogene region.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T. Buday, 1965). The boring and gravimetric data, however, document its trend and extent as far as north-western Rumania (Zemplín—Cibleš elevation; G. Sasin 1965). In this sense, its origin may be regarded as the result of the activity of the Peri-Pannonian fault system. The horst structures represent the deep-founded fault systems on which various differentiated movements occurred during the individual intervals of the Tertiary. Thus, for example, the sedimentary area of the Magura Flysch sank along the northern margin of the Periklippen elevation, locally up to the Oligocene, and from the beginning of the Miocene the Transcarpathian Inner Deep subsided along its southern border. Along the faults forming the north-eastern boundary of the Zemplín-Beregovo elevation, the Transcarpathian inner depression experienced a strong subsidence during the Miocene, but the rhythm of this intermittent subsidence was different. The Pannonian (central?) massif sank mainly in the Pliocene along the southwestern margin of the horst. The principal volcanic events are connected with these longitudinal zones of the first order (Fig. 1). Transverse tectonic lines also played an important, although subordinate role in the localization of the volcanic centres. Of primary importance are the branch of the Hornád fault system (or the Balaton line), which largely controlled the distribution of volcanics in the southern part of the Prešovské pohorie Mts., and the Vihorlat fault system (Leško—Slávik 1967).

The direct connection of these main tectonic elements of the East Slovakian Neogene with the elements of the East Carpathians, the time equivalence of tectonic and volcanic events, the contemporaneity of palaeogegraphical changes from the Palaeogene to the end of Pliocene, the consistent lithogenetic development, particularly in the inner depression as well as the geotectonic features of the East Slovakian area, that all justifies us in ranging both the Neogene sedimentary formations and the young volcanic mountain ranges to the East Carpathians.

The spatial and facies analyses of Neogene volcanic rocks

The oldest Neogene volcanic rocks are represented by rhyolite pyroclastics occurring in the Upper Burdigalian sediments of the Čelovce Formation (Fig. 2; Slávik et al., 1960) and by fine-grained seladonitized acid tuffs near

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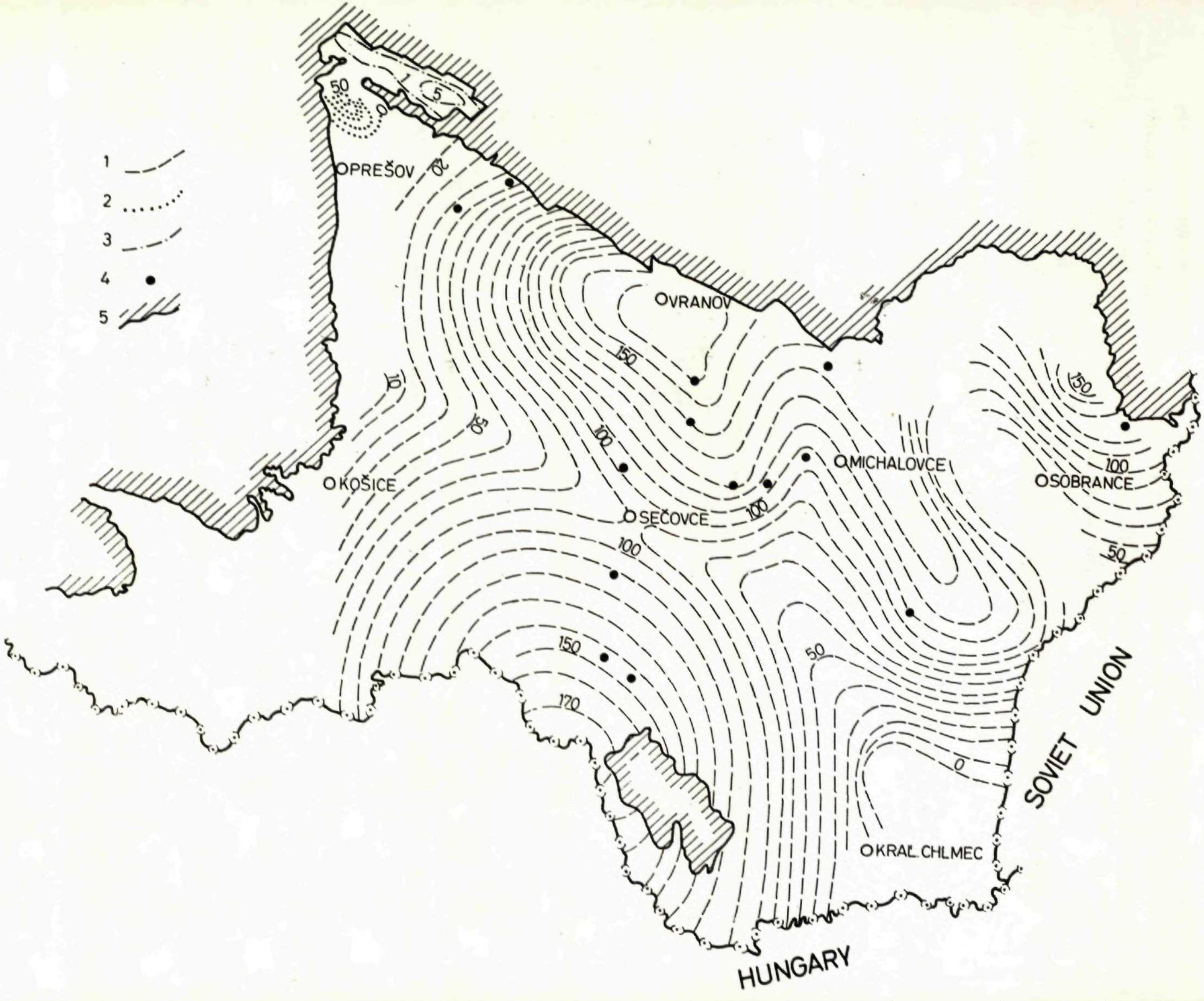
Fig. 1. Main tectonic and volcanic units of the East Slovakian Neogene area

a — Peri-klippen elevation, b — Zemplín—Beregovo elevation, c — geosynclinal area of the Magura Flysch, d — Transcarpathian inner depression, e — marginal zone of the Pannonian massif, f — Hornád fault system, g — Vihorlat fault system.

1 — Pre-Mesozoic formations of the Zemplín—Beregovo elevation, 2 — Mesozoic members of the Peri-klippen elevation, 3 — surface bodies of young volcanics, 4 — buried young volcanics, 5 — principal longitudinal fault systems, 6 — principal transverse fault systems



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Terňa (J. Březina 1960). These well-sorted crystal-vitroclastic pyroclastics, transported probably by wind, form a thin layer. They were deposited after a sudden explosion into a slightly freshened sedimentary basin (autochthonous tuff facies of a distant zone — in the sense of Malejev, 1963). The ash material covers an area of about 15 km^2 and the volume of preserved pyroclastics is 0.112 km^3. The timing of the bentonitized rhyolite pyroclastics found in three thin layers (maximum 5 m thick) in the area of Fintice remains still questionable. V. Gašparíková (in J. Harcek 1962) thinks them to be the product of the Burdigalian volcanic activity. They also are fine-grained ash tuffs deposited directly into the water basin, as documented by stratigraphically little distinctive brackish microfauna (J. Švagrovský in J. Slávik 1967). In case these bentonitized (subaqueous) tuffs are Burdigalian, the estimated volume of the preserved Burdigalian volcanics should be raised to 0.2 km^3.

The volcanic rocks of the Karpatian age crop out on the surface to the west of Fintice. These bedded, medium to coarse-fragmentary rhyolite pyroclastics are of the type of pumice tuffs, tuff breccias and xenolithic tuffs with the allothigenous admixture of quartz and older sedimentary rocks. The maximum thickness of the complex is 30 m and the distribution of volcanics is extrapolated to an area of about 28 km^2. Their volume is estimated at 0.420 km^3. From the point of view of facies, these rocks may be considered as the autochthonous pyroclastic facies of the transitional zone (zone between the near-crater and distant facies) of the tuff up to xenolithic-tuff type. The presence of rhyolite lavas referred to by Kuthan (1948) enables us to class these facies with the mixed effusive-extrusive group. The obsidian occurrence in the Karpatian complex in the Prešov suburb (Táborisko) recorded by Zipser (1817) may be regarded as a subvolcanic or vein equivalent of the same volcanism. However, it should be mentioned that some scientists, e. g. Šalát (op. cit.) believe this occurrence to be an archaeological finding place to which obsidian was transported. No significant types of alteration have been observed.

Volcanism synchronous with the deposition of the Lanzendorf Formation (Lower Tortonian) was of extraordinary importance. Products of this volcanism form several (four to six) layers; the most notable of them is the

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Fig. 2. Isopach map of Burdigalian, Karpatian and Lower Tortonian volcanic rocks in the Neogene of East Slovakia

1 — Isopachs of rhyolite tuffs of the Lanzendorf series (Lower Tortonian), 2 — isopachs of the Karpatian rhyolite pyroclastics, 3 — isopachs of the Burdigalian rhyolite pyroclastics, 4 — points of measurement of the volcanic rocks thickness, 5 — boundary of the sedimentary Miocene

Hrabovec rhyolite or rhyodacite tuff developed at the boundary between the Lanzendorf Formation and the Spiroplectammina Zone (Slávik 1967). Cumulative isopachs of the products of this volcanic phase show an interesting course. Their peaks discretely coincide with the longitudinal peripheral parts of the Miocene depression, and with substantially smaller thicknesses in the central part of the basin (Fig. 2). The composition of rocks representing this volcanism ranges from coarse-grained tuff breccias and xenolithic tuffs, chiefly at the southern periphery, to fine-grained vitroclastic ash tuffs at the northern periphery. The coarsest products have been found around Klečenov; they are breccia-like xenotuffs with fragments of limestone and dolomite and with fissures often healed by post-genetic anhydrite. The conditions in the area of Sečovce are analogous. Pelitomorphic to psammitomorphic ash tuff facies free from allothigenous admixtures occur along the whole northern margin. They have been found at numerous localities (Šarišská Poruba, Pavlovce, Hermanovce, Ruský Hrabovec, Oreské, Boroľa). The rocks in surface exposures are occasionally unaltered whereas in deep borings they are affected by diverse alterations, most frequently by pyritization (Trhovište), carbonatization (Stretava, Albinov, Sečovce, Iňačovce), chloritization (Albinov, Dlhé Klčovo) and even by weak kaolinization (Stretava). These volcanic products cover an area of almost 3000 km^2 and their volume attains 578 km^3. The thicknesses of the Lower Tortonian rocks do not suggest an aeolian transport from distant areas, as presumed by some geologists thus far. At present, we are inclined to regard them as fissure explosions (probably subaqueous) along the marginal faults of the Miocene sedimentary area. It is noteworthy that lavas have not yet been found in this group, which comprises only pyroclastic facies of the autochthonous type: lapilli, tuff to xenolithic tuff facies in the Zemplín-Beregovo elevations and psammitic to aleuritic-tuff facies of the transitional zone in the vicinity of the Peri-klippen elevation. The complex attains maximum thickness to the SE of Vranov (up to 170 m) and in the Žipov area (up to 250 m). Surprisingly, there is no distinct relationship between the thickness of pyroclastics and their granulometric composition. It is possible that the dependence has been distorted by constructing the cumulative isopachs, but if the isopachs were plotted for individual horizons, the map would loose its clarity. The map of secondary alterations could not be compiled for the insufficient number of observations.

During the marine Upper Tortonian (Tortonian s. s.) volcanic activity

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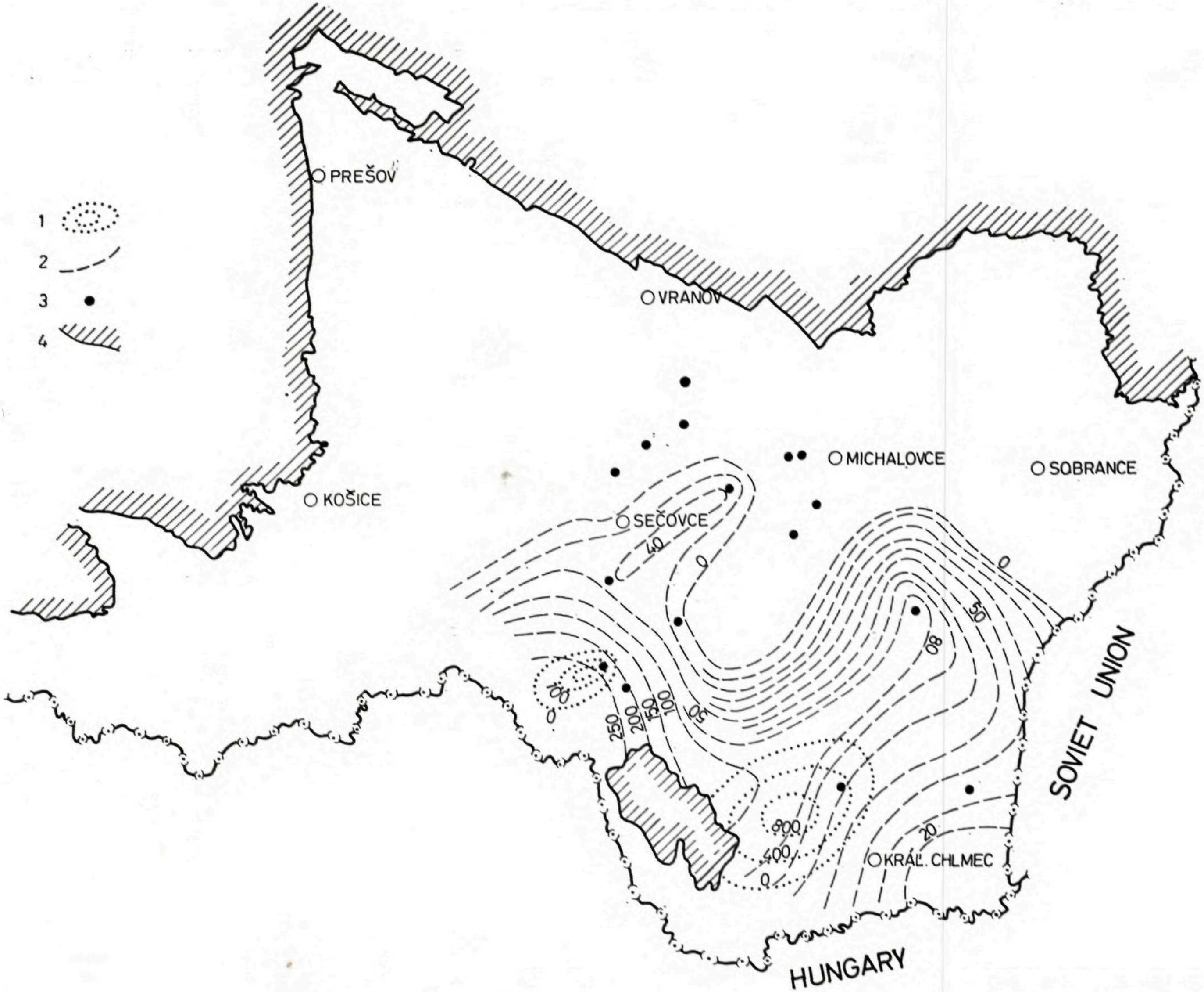
Fig. 3. Isopach map of the volcanics of the marine Tortonian s. s. (Spiroplectammina and Bolivino-Bulimina Zones)

1 — Isopachs of andesitic rocks, 2 — isopachs of rhyolite pyroclastics, 3 — points of measurement of the volcanic rccks thickness, 4 — boudary of the sedimentary area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on northern boundary faults died down and volcanicity was restricted to the area of the Zemplín-Beregovo elevation in the S part of the region. There, volcanic centres produced the bulk of rhyolite material right in the elevation, to the NW of the present-day Zemplín Horst and minor flat pyroclastic cones round Stretava and Trhovište, nearer to the centre of the basin (Fig. 3). In the environs of Klečenov and Žipov, the thickness of pyroclastics exceeds 250 m. They are coarse-grained rhyolite pyroclastics of the breccia type, obviously representing the near-crater facies. Secondary alterations include chloritization, silicification and carbonatization. In some parts anhydrite veins have been detected. This rhyolite volcanism makes up several layers in the Spiroplectammina Zone and especially in the lower part of the Bolivino-Bulimina Zone. In the upper part of the latter it is replaced by the first major effusions and probably also weak explosions of pyroxene andesites. These were struck by boring near the village Zatín between Kráľovský Chlmec and the Zemplín Horst. Andesites are more than 400 m thick and form four successive lava flows overlain by sedimentary beds. Analogously, in the neighbourhood of Žipov there is a 130 m thick layer of andesite overlain by a thin bed of fine-grained andesite tuffite with albitized plagioclase. In the Upper Tortonian the rocks of rhyolite composition have a volume of 135 km^3 and rhyolite ash material occupies an area of almost 500 km^2. The volume of andesites is 119 km^3 and their areal extent is 170 km^2. The predominating alteration type is carbonatization, observed mainly south of Sečovce, near Trebišov, Žipov, etc. In the environs of Ptrukša, the rocks are affected by silicification and round Stretava by carbonatization and pelitization, locally connected with the chloritization of glass. A characteristic feature of this volcanism is the lateral lobate alternation of salt and volcanogenic sedimentations. This phenomenon may be plausibly explained in that the salt deposition avoided the areas of volcanic unrest and sedimentation.

In the freshened Upper Tortonian, a huge volcano producing rhyolitic rocks appeared in the neighbourhood of Zamutov (Fig. 4). The maximum thickness of volcanics exceeds 350 m. Biotite rhyolite alternates with layers of coarse rhyolite pyroclastics which indubitably belong to the crater formation, i. e. to the facies of the volcanic centre. The apex of the Zamutov volcano rises to the surface but its predominant part is buried by the sediments of the Upper Tortonian. The volcanic products have a volume of 197 km^3 and cover an area of 545 km^2. Other Upper Tortonian pyroclastics appear near Ptrukša, to the north of Čierna nad Tisou; they are fine-grained rhyolite pyroclastics

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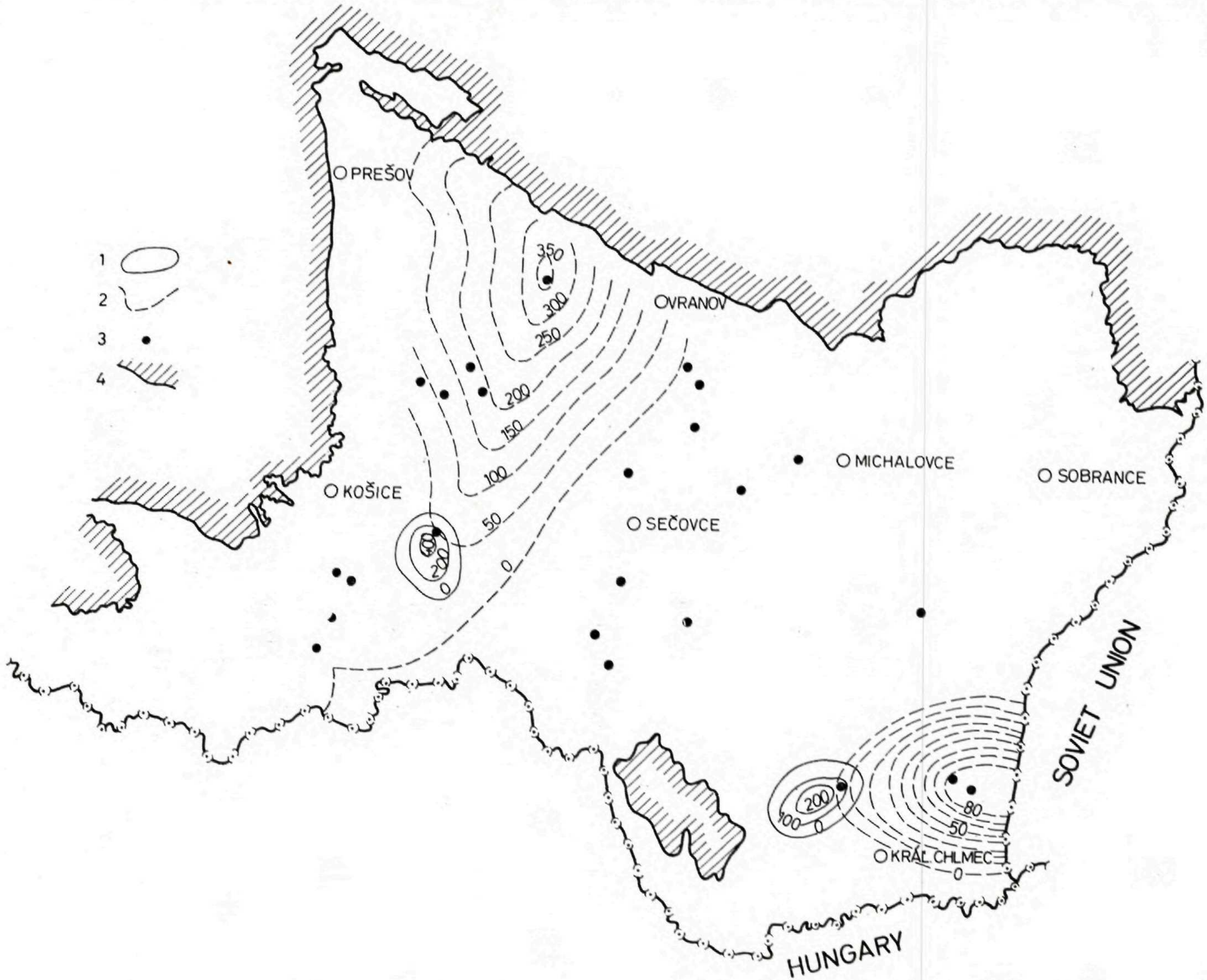
Fig. 4. Isopachs map of the Upper Tortonian volcanic rocks (Rotalia Zone)

1 — Isopach of andesite, 2 — isopach of rhyolitic rocks, 3 — points of the thickness measurement, 4 — boundary of the Miocene sedimentary area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forming flat bodies, maximum 80 m thick. Their areal extent amounts to 279 km^2 and the volume of ejected material is 16 km^3.

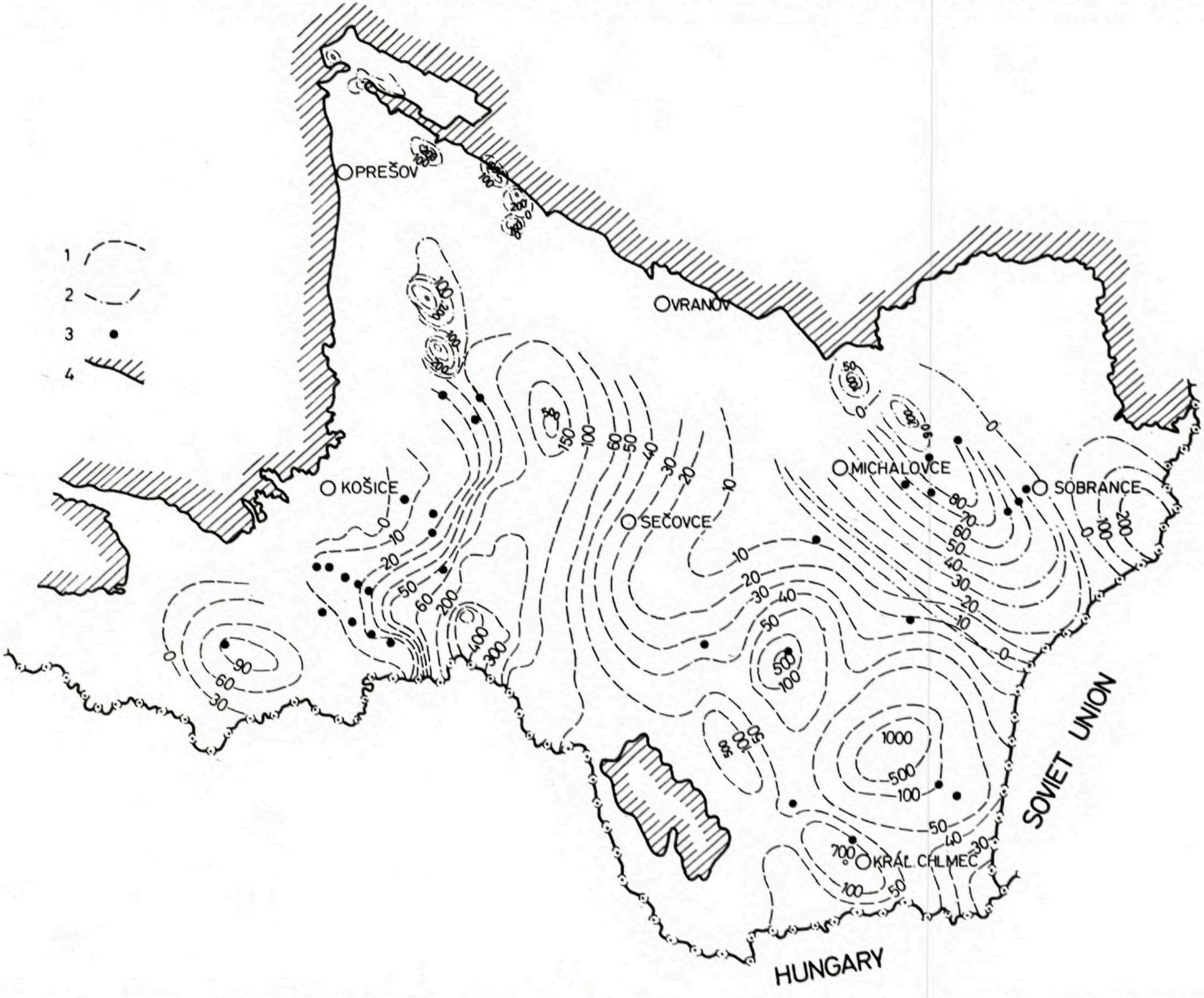
Pyroxene andesites and andesite tuffs round Svinica, on the western side of the Prešovské pohorie Mts., also occur in the Upper Tortonian freshened complex. The complex of silicified andesite tuffs (about 100 m thick) near Zatín seems to represent a complex of andesite rocks from the time of the Bolivino-Bulimina Zone that was buried during this period. In the area of Svinica, the andesites are probably the product of a new volcanic event, which reveals the shift of the andesite volcanism farther to the west. The volume of andesites belonging to the Upper Tortonian is 22.4 km^3 and the area covered by them is 90 km^2 large. There is no doubt that the lava effusions are directly connected with the near tectonic zone from which the lava flows poured out. Apart from carbonatization in the centre of rhyolite volcanism (in the Zamutov area) and pyritization of some andesite tuff layers at Zatín, no conspicuous types of alteration have so far been observed. The Zamutov volcano is a body of varied facies composition, ranging from effusive-extrusive crater facies through the transitional facies of outer slopes of the volcano to televolcanic facies made up of an alternation of auto- and allochthonous beds of pumice tuffs.

The Lower Sarmatian volcanism is documented south of the central part of the basin, between the Soviet frontier and Košice. Its products are andesite tuffites of Švagrovský's (1964) "Olšava Beds" and andesite tuffs in the Cibicides Zone encountered in the borings near Ptrukša. They form maximum 25 m thick beds of fine to medium-grained tuffs and tuffites, which at a greater depth are strongly decomposed (carbonatized, kaolinized, etc.), whereas the equivalent rocks outcropping at the surface are generally slightly weathered and mostly unaffected by other types of alteration. This pyroclastic onset of the andesite volcanic activity is restricted to the lowest Sarmatian, i. e. the Cibicides Zone. In the basal parts of the zone of large Elphidia the intensity and volume of andesite volcanism increases abruptly and two parallel lines of andesite volcanoes came into being. The bulk of these volcanoes is buried by the younger sediments. The northerly line extends from Beša through Vojany and Malčice to Bogota and the southerly line is traceable from Kráľovský Chlmec through Sírnik and Hraň to Hradisko.

The huge volcanic structures (Fig. 5) of these lines are indubitably of the strato-volcanic type and are distinguished by the multiple alternation of

Fig. 5. Isopachs map of the Sarmatian andesitic rocks

1 — Andesitic rocks of the marine (Lower) Sarmatian — Cibicides Zone and Zone of large Elphidia, 2 — andesitic rocks of the Upper Sarmatian (*Elphidium hauerinum* Zone; *Nonion granosum* Zone boundary), 3 — points of the thickness measurement, 4 — boundary of the Miocene sedimentary area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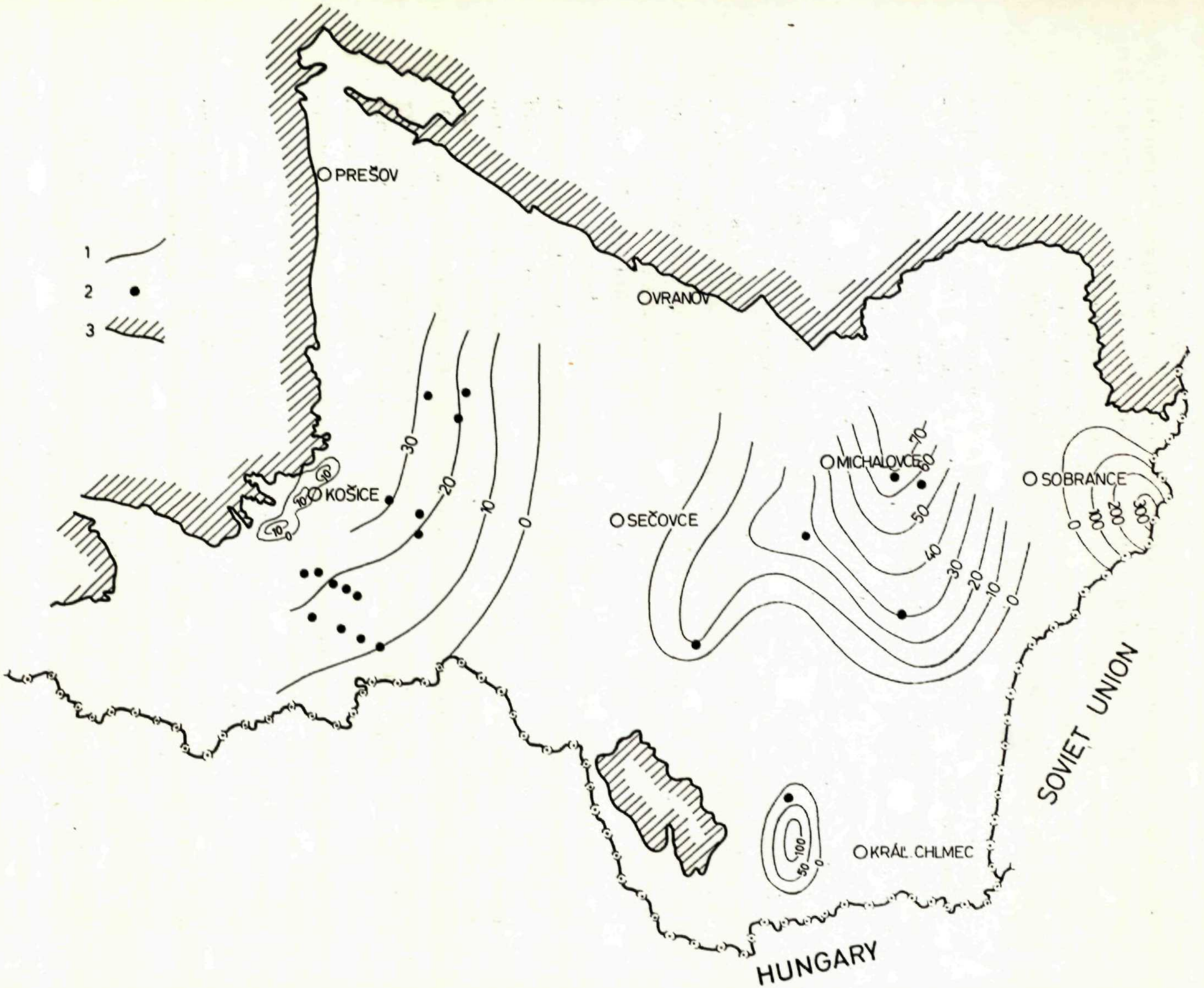
pyroclastics and lava flows; thus, for example, the Malčice boring established more than ten lava flows overlain by pyroclastic beds. The thickness of volcanics exceeds 1000 m but drops abruptly sidewards to die out at a distance of 15–20 km. The area covered by volcanic rocks of this phase surpasses 1000 km^2 and their volume attains 310 km^3. The andesite volcanism culminated by hydrothermal alteration which is regarded as responsible for the origin of hydrothermal quartzite in the top part of the buried Malčice volcano. The buried parts of volcanoes are represented by near-crater facies groups with intra-crater flows (probably Streda nad Bodrogom), domal structures (Sírnik), pyroclastics and, additionally, with facies of the transitional zone — of the outer hillsides — distinguished by the alternation of flows and coarse pyroclastics. Televolcanic facies are subordinate. The yellowish subaqueous andesite tuffs of the Olšava beds and the tuffite-tuff layers of andesite provenance on the SE slopes of Bankov Hill in Košice may be grouped with them.

In the north-eastern and western parts of the East Slovak Neogene, the products of andesite volcanism are overlain by numerous layers of rhyolite tuffs. They have been best recognized to the south-east of Košice and described as "Myšla Beds" by Švagrovský (1964). They are mainly developed in the form of pumice tuffs. In the northern part of the basin they converge into the area of Michalovce, where their thickness, as evidenced by Iňačovce boring, is up to 70 m. Their coarse-fragmentary tuff breccia material points to the proximity of the volcanic centre. In some sectors, chloritized and more frequently carbonatized portions have been ascertained. Their thickness exceeds 80 m and their volcanic centre is postulated in the proximity of the Zemplín-Beregovo Horst. These rhyolite pyroclastics have a volume of 53 km^3 and occupy an area of about 820 km^2. The mode of deposition and the analysis of grain-size distribution indicate the volcanic centres to be located near Michalovce and Komarovce, but occurrence of coarse pumice near Košice suggests that one of the explosion centres may also have been in that area. In the Hauerina Zone, in the stratigraphically higher Sarmatian, the medium-grained to pelitomorphic pumice tuffs and tuffites of the tuff-terrigenous facies form up to 50 m thick layers, mainly between Kráľovský Chlmec and the Prešov-Tokajské pohorie Mts., in the southern tract of the region. The volcanogenic material was partly deposited directly and partly redeposited by water in shallow lakes on the Zemplín-Beregovo Horst. Their volume is about 5 km^3. As they were uncovered up to the Upper Pliocene, they underwent weathering in the supergene zone and minerals of montmorillonite type were generated. As a result, deposits of bentonite originated at Michaľany, Lastovce, Kuzmice and probably also in other places. The weathering process was of a specific type. Rain water infiltrated into the ground was the principal factor: during its percolation it caused the decomposition of silicate glass (Slávik).

The Upper Sarmatian andesite volcanism shifted to the north of the Miocene depression, into the area of the Vihorlat and the Prešovské pohorie Mts. Its products were studied in detail chiefly in the Vihorlat area, where they constitute the so-called agglomerate-tuffite series (Brodňan et al. 1958). Rocks of this volcanic episode are dated as intervening between the zone with *Elphidium hauerinum* and that with *Nonion granosum*. This type of volcanics occurs at the surface mainly in the form of subvolcanic and intrusive cupola-like bodies and diorite-porphyrite intrusions, which are typically developed particularly in the northern part of the Prešovské pohorie Mts (Fintice, Kapušany, Maglovec, Oblík, Vinné). This hornblende-pyroxene-andesite volcanic and magmatic activity is intimately linked up with the northern margin of the basin. Moreover, the intrusive forms (diorite-porphyrite) are strictly confined to the tectonic elevation zone, whereas extrusive-effusive formations represented mainly by the agglomerate-tuffite series of the Sub-Vihorlat basin and some pyroclastics of the agglomerate type (between Zlatá Baňa, Varhaňovce and Ruská Nová Ves) developed farther to the south. As these subvolcanic and intrusive rocks frequently possess devitrified groundmass, they were called the "autometamorphosed hornblende-pyroxene andesites". They have a volume of 95 km^3 and extend over an area of 350 km^2.

The Pliocene volcanic activity begins with acid explosions. These outbursts produced flat layers of acid pyroclastic material. They are of regional distribution and occur at the base of the Pliocene sequence which was deposited after a hiatus that began in the latest Sarmatian. A particular facies of these pyroclastics has been found at the foot of the Vihorlat mountain group and in the vicinity of Lesné; they are distinguished by the presence of perfectly automorphic almandine crystals in medium-grained rhyodacite pumice tuffs — garnetiferous tuffs (Brodňan et al. 1958; Sasin — Slávik 1967). Garnetiferous and analogous tuffs are traceable along the northern border of the molasse from the Soviet frontiers as far as the northern environs of Zlatá Baňa. In the south they are well developed as coarse-grained breccia -like rhyolite pyroclastics between Kráľovský Chlmec and Streda nad Bodrogom (Fig. 6).* Their thickness which at the periphery of the basin is up to 100 m decreases inwards so that in the centre of the basin it is about 10 metres. This also suggests that the volcanic centres are confined to the faults bounding the Neogene sedimentary area. The effusive and subvolcanic equivalents of the pyroclastics show a linear arrangement along the northern margin of the Miocene molasse, extending from Zalužice (buried domes) across Biela Hora, Hrádok near Michalovce, Lesné, and Vranov to Zlatá Baňa. They form dome-like bodies of minor extent and steep sides so that they plunge abruptly into

* But a part of these pyroclastics can be perhaps of the Upper Sarmatian age.



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the Neogene sediments. Apart from occasional kaolinization, viz. chloritization of glass (pumice) no significant alterations have been observed in garnetiferous pyroclastics or their equivalents. Their areal extent exceeds 2000 km^2 and their volume is more than 180 km^3. The great number of observations that are available enables us to conclude that round the centres located invariably within the range of domal structures there are coarse-fragmentary pyroclastics of the breccia-type with fragments of garnetiferous rhyolite or rhyodacite. Farther away from the centre these pass into purely pumice tuffs with white and greyish clods of pumice and in the most remote parts into fine-grained sandy tuffs up to tuffites. Frequent traces of re-deposition by water, i. e. volcano-terrigenous facies generally occurring at the base of this complex led Čech (1959) to an incorrect assumption of the allochthonous origin of this pyroclastic complex.

Subsequently to the period of volcanic rest including the Pontian and part of the Meotian, the vigorous andesite volcanism broke out whose main stages took place in the Levantinian (Apsheronian, Akchagil; Sheremeta 1967). Large volcanoes formed in the Vihorlat area and to the north of the Dargov Pass in the Prešov—Tokajské pohorie Mts. (Fig. 7) and produced strato-volcanic complexes composed of the alternation of coarse agglomerates and tuff breccias with lava flows. In addition, dykes and sills (Čeremošná), dome-like bodies (Vechec, Veža) and sporadic conduits (in the foreland of volcanic massifs — Beňatina, Inovec) have been ascertained. The area covered by these youngest volcanic rocks (monotonous pyroxene andesites and scarce basaltoid andesites) equals 650 km^2 and the volume of ejected material attains 425 km^3. It seems that the lava flows of pyroxene andesite overlying the rhyolite pyroclastics at Streda nad Bodrogom may be the product of the same Upper Pliocene volcanism. If this assumption is confirmed, it will prove the closer time relations between the events on the two elevation zones. The andesite explosions terminate the volcanic activity in East Slovakia. The post-volcanic phenomena are documented by the conspicuous zones of silicification, carbonatization, chloritization and alterations of the argillitization type. These are developed mainly in the central part of the Vihorlat mountain group. Pyritization is known especially from the southern foot of the Vihorlat, from the area between Vinné and Remetské Hámre. In the Prešov—Tokajské pohorie Mts., the above-mentioned post-volcanic alterations are accompanied by opalization. Several mineralization phases gave rise to cinnabar, antimonite and probably also polymetallic veins.

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Fig. 6. Isopach map of the Sarmatian rhyolitic rocks

1 — Rhyolitic rocks of the Hauerina Zone, 2 — points of measurement of the volcanic rocks thickness, 3 — boundary of the sedimentary area

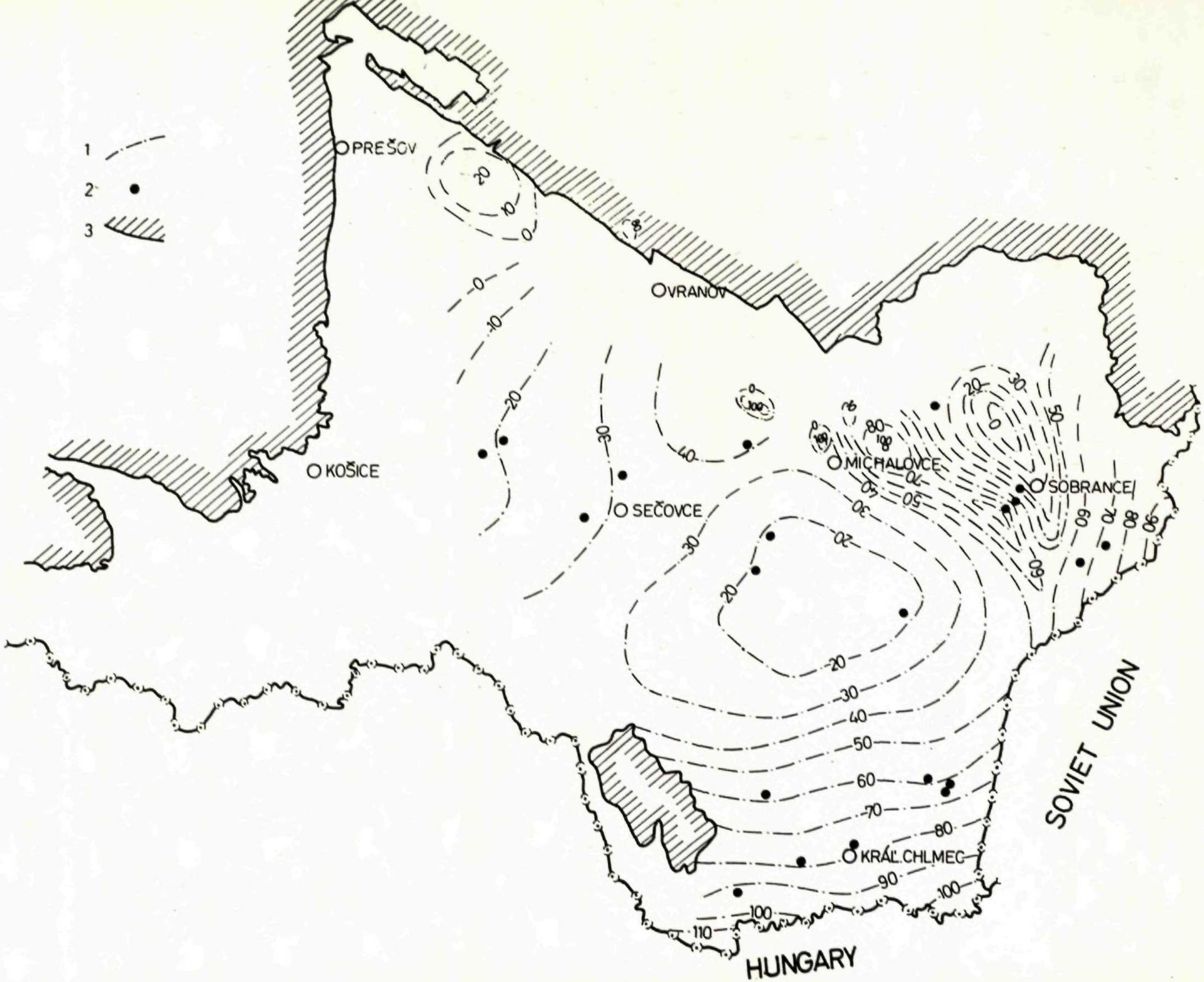

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Table 1 lists the volumes of volcanic products of the individual volcanic activites in chronological order.

Table 1. Volumes of volcanic rocks in the Neogene of East Slovakia

Age of volcanism	Volume in km³:		
	andesite	rhyolite	total
Burdigalian	—	0.2	0.2
Karpatian	—	0.42	0.42
Lanzendorf series	—	578	578
Tortonian, marine (s. s.)	119	135	254
Tortonian, fresh-water (s. s.)	22.4	213	235.4
Sarmatian, marine	310	53	363
Sarmatian, fresh-water	95	5	100
Pannonian	—	180	180
Levantinian	425	—	425
Total	971.4	1164.60	2136.0

The above Table clearly shows that from the Lower Miocene to Pliocene the volume of rhyolites generally decreases and, on the other hand, the volume of rocks of andesite composition increases. The explanation of this phenomenon is given in the text below. If the volume of the Neogene depression in East Slovakia is roughly estimated at 7000 km³, the considerable share of volcanic rocks is well apparent.

The relationship of volcanic activity to the basic tectonic events

The upheaval of the Magura part of the Flysch geosyncline to the north of the Klippen Belt, which began during the Oligocene, led since the Burdigalian to the synchronous formation of the Transcarpathian Inner Deep in the backland of the orogen area. In our opinion, it was caused by the origin of normal faults isostatically compensating for the mass deficiency in the deeper crustal parts, which accompanied the strong uplift of the Magura geosynclinal area. The inner depression developed at an accelerated rate during the Miocene and gradually declined in the course of Pliocene. However, this process was intermittent, showing periods of shoaling to transitory dwindling away of the area of deposition, as indicated by the breaks in sedimentation and the

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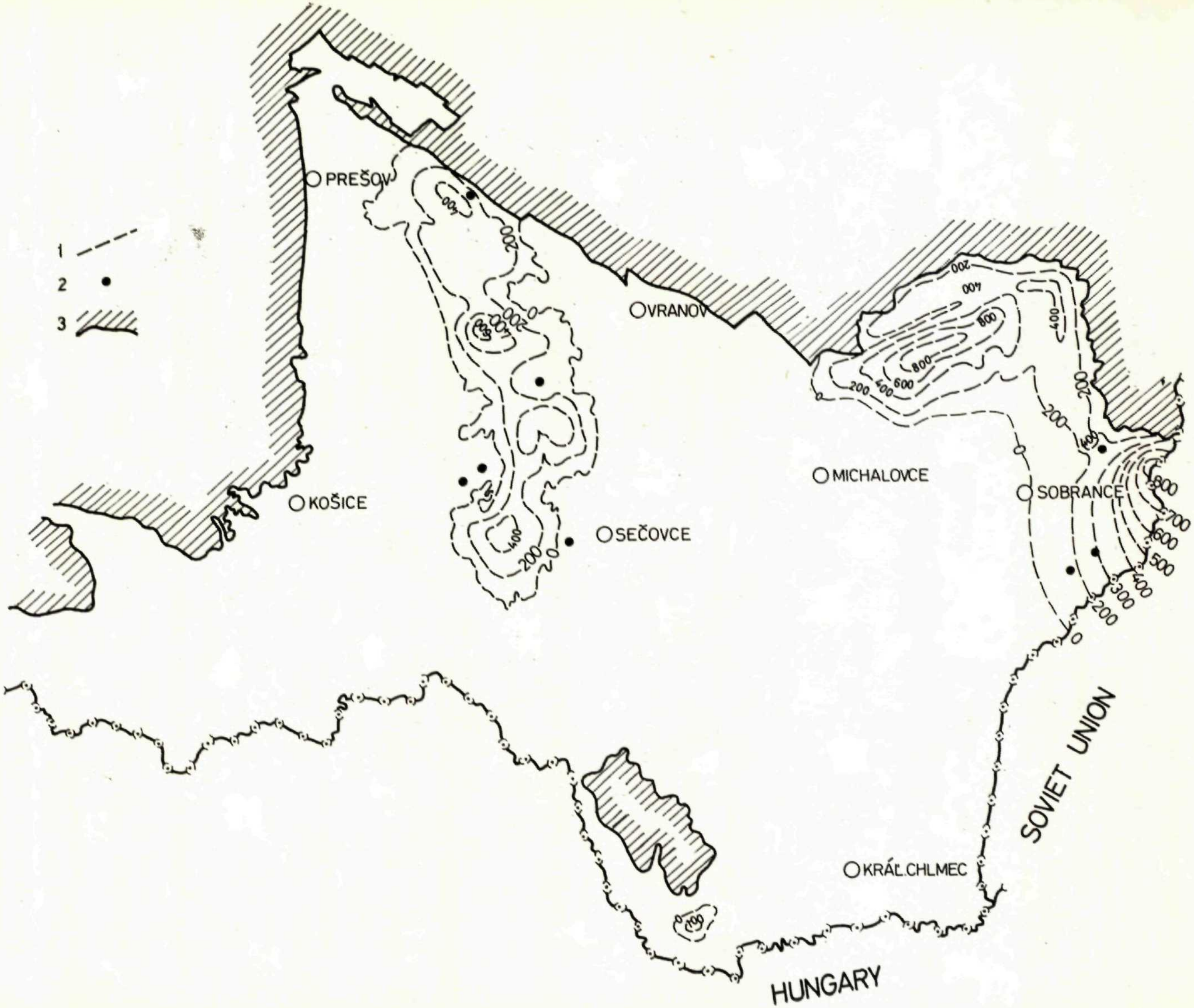
Fig. 7. Isopach map of the Pliocene rhyolite (rhyodacite) rocks

1 — Rhyolite tuffs, 2 — points of measurement of the volcanic rocks thickness, 3 — boundary of the sedimentary area (the southern part of fig. comprisses the uppermost Sarmatian pyroclastics too).



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onset of transgressive sedimentary facies. During the geological history of the interior molasse there were five main breaks in sedimentation which affected almost the entire area of deposition. According to Buday (1966) and the observation of the authors of this paper, the notable discordances characterized by angular unconformities, hiatuses and the onset of transgressive facies occurred at the following time-intervals: between Oligocene and Burdigalian; between Burdigalian and Karpatian (interruption comprising the whole Helvetian s. s.); at the boundary between the Karpatian and the Lanzendorf Formation; at the boundary between the marine and the brackish Tortonian, and at the Sarmatian/Pliocene boundary. Additionally, three transgressive movements of minor extent occurred: (1) at the base of the Zone with large Elphidia (Lower Sarmatian), (2) at the beginning of the Pontian, and (3) at the beginning of the Levantinian. From the correlation of the time sequence of these events with those in the folded Magura geosynclinal area it may be inferred that the above mentioned tectonic phenomena wer econtemporaneous with the folding phases of the Magura Flysch, i. e. that they represent the periods of strong tangential compression of the whole area. Each of these tectonic episodes was obviously followed by the release of tangential pressures which resulted in the abrupt sinking of individual crustal blocks in the Miocene molasse. Let us turn our attention the relationship of the volcanic events to this chronological pattern of the tectonic history of the area.

Table 2, which correlates the tectonic and volcanic manifestations, reveals that the volcanic activity avoids the periods of the main folding phases, i. e. that volcanic rocks are absent from the intervals of the interruption of sedimentation. In contrast, volcanic rocks occur discretely in those periods when tangential pressures fade away, i. e. in the periods of maximum subsidence movements in the sedimentary area. The analysis of volcanic activity, as shown in Fig. 2, discloses an asynchronous pulsation of rhyolite volcanic centres, in case the volume of ejected material was small, but a synchronous pulsation of the centres in both tectonic zones when the amount of ejected products was large. The andesite volcanism displays progressive migration from the southern Zemplín-Beregovo elevation northwards into the area of the Peri-klippen elevation between the Tortonian and Pliocene.

A certain analogy in the intensity of the two geological processes is also apparent from the relation of the volume of volcanic rocks to the thickness of sedimentary complexes. The maximum rates of subsidence coincide with the maximum volume of contemporaneous volcanic rocks (see Table 3).

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Fig. 8. Isopach map of the Pliocene andesitic rocks.

1 — Andesites of Upper Pliocene, dominantly Levantinian age, 2 — points of measurement of the volcanic rocks thickness, 3 — boundary of th e sedimentary area

Table 2.

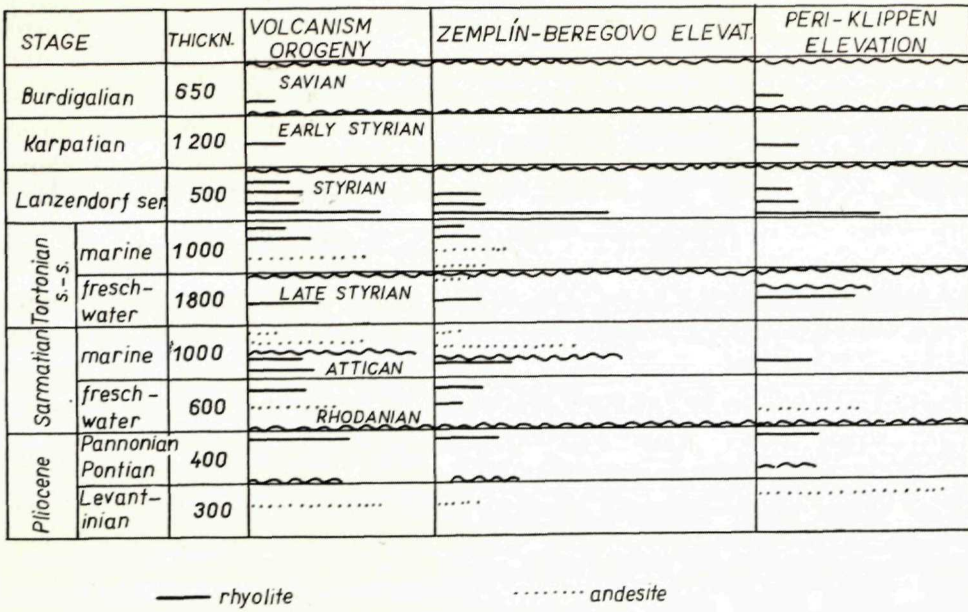

Table 3. Relationship between the thickness of sediments and the volume of synchronous volcanic rocks

Age	Maximum thickness of sediments in m	Volume of volcanics in km^3:
Burdigalian	500	0.2
Karpatian	1200	0.42
Tortonian s. l.	3300	1167
Sarmatian	1600	463
Pliocene	800	605

The facts mentioned above afford conclusive evidence that there does exist a causal connection between the volcanic and the tectogenetic history of the area. This relationship may be generally expressed as follows: Volcanic rocks ascending to the surface during the period of folding of the geosynclinal area and of the formation of interior molasses appear at the surface in the intervals between the main folding phases. The beginning of the subsequent volcanism is characterized by the predominance of acid magmas, whereas the end of it is distinguished by the prevalence of intermediate ones. The location of volcanic centres of both the acid and intermediate magmas into the marginal fault systems and their alternating activity prove that the magmas have common magma chambers, from which the material is brought to the surface during different stages of the differentiation process. However, it seems plausible that in the initial stages of the generation of magmatic foci acid magmas of the palingenic type originate and that the emptied magma chambers are afterwards invaded by portions of more basic magma rising from the deeper parts of the Earth's crust. In the magmatic focus it may undergo further differentiation and other processes. The comparison of volcanic and tectonic events during the Neogene development of the East Slovak region facilitate a new definiton of volcanic phases in eastern Slovakia. Table 2 clearly shows that the intervals between the main stages of tectonic activity were the periods of volcanicity that exhibits a general trend from acid to more basic differenciates. This observation is consistent with our geological conceptions, as after the termination of a given volcanic activity and a long-term interruption caused by the spatial compression, the process of differentiation, probably also of palingenesis and anatexis may occur in the magma chamber. When, afterwards, horizontal pressures are released and the magmatic focus is reached by deep faults, the palingenic or acid differentiated magmas are the first that come to the surface. The subsequent andesite magmas may represent undifferentiated components of the magmatic hearth or, more probably, new portions of magma derived from the deeper parts of the Earth's crust. This conception of the volcanism-tectonics relationship enables us to re-define the volcanic phases for the East Slovakian region. According to our definiton, *a volcanic phase covers a time interval during which a complex of volcanic events takes place within one discrete tectonic block, between two major tectonic processes.* In the area studied, magmatic masses, at first acid and basic towards the close, ascend discontinuously to the surface in the course of one volcanic phase. (It's not impossible that in other areas magmas of only one type are brought to the surface. Our proposal of the division of volcanic events in East Slovakia into phases is given in Table 4.

Conclusion

The volcanism of East Slovakia represents a geological process which differs from the young volcanism of the West Carpathians in its position, rhythm, succession in time and to a certain extent also in composition. Owing to its relationship to the sediments of the Miocene molasse, the tectonic position and chronology of events, it stands nearer to the volcanism of the western part of the East Carpathians, i. e. to that of the Transcarpathian Ukraine and north-western Rumania and partly also of north-eastern Hungary.

There is a discrete connection between the tectonic history of the area studied and the succession of the volcanic events. The volcanic activity is linked up with the fault systems that in the NE and SW constitute the longitudinal boundaries of the Transcarpathian Inner Deep, a component of which

Table 4. Scheme of Neogene vol

Compiled by

Age	Orogeny phase	Volcanic phase	Volume of volcanics (km^3)	Rock type	Thickness of volcanics in metres
Burdigalian	Savian	I	0.2	rhyolite	5
	Early Styrian				
Karpatian		II	0.420	rhyolite (andesite)	30
	Styrian				
Lanzendorf series (Lower Tortonian s. l.)				rhyolite	170
Tortonian marine, s. s. (Zone of aggl. foram. and Bulimina-Bolivina Zone)		III	832	rhyolite	250
				andesite	400
	Late Styrian				
Brackish to fresh-water Tortonian s. s. (Rotalia Zone)				rhyolite	350
		IV	546	andesite	150
Marine to brackish Lower Sarmatian (Cibicides Zone and Zone with large Elphidia)				andesite	1000
	Attican				
				rhyolite	30
Upper Sarmatian (Hauerina Zone)				rhyolite	50
		V	153		
Fresh-water Upper Sarmatian				andesite (diorite porphyrite)	100
	Rhodanian				
Lower Pliocene (Pannonian)				rhyolite (rhyodacite)	100
		VI	605		
Upper Pliocene (esp. Levantinian)				andesite basalt andesite	1000

canism in East Slovakia

J. Slávik, 1967

Facies types	Locality	Geological area
Pelitic to psammitic subaqueous tuffs and tuffites, occasionally seladonitized and bentonitized, free of lavas	Terňa, Čelovce, Fintice (?)	Prešovská kotlina (Prešov depression)
Parallel-bedded subaqueous layers free of lavas, pumiceous, xenolithic and lapilli tuffs with sporadic fragments of pyroxene andesite; tuffites: obsidian?	Fintice, Prešov – Táborisko (?)	Prešovská kotlina (Prešov depression)
Parallel-bedded layers of aleuritic vitroclastic tuff to pumiceous-lapilli and xenolithic tuffs, subaqueous, lava-free. Local traces of hillwash and bentonitization.	Šarišská Poruba, Petrovce, Nižný Hrabovec, Oreské, Boroľa, Klečenov – Žipov, Sečovce, Albinov, Stretava, Klčovo Dlhé,	Neogene area of East Slovakia
Rhyolite lapilli tuffs to tuff breccias, xenolithic tuffs, locally tuffites. Bedded structures and lenses.	Nižný Žipov, Klečenov, Stretava, Ptrukša	Southern margin of the East Slovakian Neogene (area of the Zemplín-Beregovo elevation), partly also the central tract
Lava flows interlayered with sedimentary rocks and thin beds of tuffites and sandy tuff.	Zatín, Svinica, part of the Veľký Milič massif (?)	
Lava flows, domes and pyroclastics of biotite ryolite, from breccia tuffs through lapilli, and pumiceous tuffs to tuffites	Zamutov, Kráľovce, Ptrukša	North-western and eastern parts of the East Slovakian Neogene
Lava flow	Svinica, Zatín	South-western part of the Neogene
Psammitic subaqueous crystalvitro clastic andesite tuffs. Stratovolcanic complex of coarse pyroclastics alternating with lava flows of pyroxene andesite	Ptrukša, Olšovany, Košice, Sirník, Hraň, Brehov, Malčice, Ruskov, Beša, Čičarovce, Kráľovský Chlmec, Hradisko	Southern border and central part of the East Slovakian Neogene
Pumiceous to lapilli tuffs, tuff breccias, perlite, ignimbrites and tuffo-lavas (domes?)	Nižná and Vyšná Myšľa, Skároš, Inačovce, Košice, Budimír, Kavečany (Viničky?), Byšta	Mainly the south-western and north-western margin of the East Slovak Neogene basin
Mainly pumiceous and lapilli tuffs alternating with redeposited tuffs and tuffites with strong bentonitization; domes and lava bodies	Lastovce, Michaľany, Kuzmice, Egreš	Southern border of the East Slovak Neogene, the zone of Zemplín-Beregovo elevation
Subvolcanic bodies, intrusive bodies, domes Coarse pyroclastics, agglomerates and tuff breccias	Veľká a Malá Stráž, Šariš-Castle Hill, Kapušany-Castle Hill, Maglovec, Hrb, Kura Hura, Vinné Lancoška, Medveďova, Zlatá Baňa, Varhaňovce, Lučky, Závadka, Hnojné, Sejkov, Vyšné Nemecké	Northern margin of the East Slovakian Neogene
Garnetiferous pumiceous to lapilli tuffs, tuffites to tuffite-conglomerates, dominantly in subaqueous and partly in terrigenous facies, domes, dykes	Hnojné, Závadka, Petrovce, Remetské Hámre, Sobrance, Sejkov, Vyšné Nemecké, Banské, Ptrukša, Stretava, Michalovce, Lesné Komarany, Čičava, Merník, ZlatáBaňa	Neogene area of East Slovakia Northern periphery of the East Slovakian Neogene
Andesite lapilli tuffs, tuff breccias, agglutinates, agglomerates, lava flows, lava breccias, domes, dykes. Hydrothermally altered zone.	Petrovce, Remetské Hámre, Boroľa, Jovsa, Beňatina, Herľany, Rankovce, Dargov, Sninský kameň, Klokočov, Husák, Šarišská Poruba, Ladomírov, Inovce, Vechec a o., depression of Veľké Okno, area of Zlatá Baňa	Northern part of the Prešov – Tokajské pohorie, the Vihorlat and the Popričný Mountains

is also the East Slovakian Neogene. The chief stages of volcanic activity were during the maximum subsidence of the interior depression and the volcanicity died down during the compression of the sedimentary area, i. e. in the folding periods of the Palaeogene geosynclinal basin. On the basis of the major folding periods, six phases of volcanism have been defined in East Slovakia. Each of them begins with manifestations of acid volcanism and ends with intermediate volcanism. The first three phases are characterized by a high coefficient of explosiveness and a complete absence of lavas, as well as by the exclusively rhyolitic composition of volcanic rocks. The successive phases are distinguished by both acid and basic (intermediate) lavas and pyroclastics. From this it may be inferred that the tectonic lines reached deeper parts of the Earth's crust or that the emptied magma chambers were invaded by new portions of magma. Regarding the migration of volcanic manifestations, an overall shift of volcanic centres is observable from the southern elevation zone (Zemplín-Beregovo elevation) towards the boundary faults of the northern elevation zone (Peri-klippen elevation). The volcanic rocks form a considerable part of the filling of the Neogene Inner Deep, constituting approximately one third of the total volume of the depression.

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GEOLOGICAL HISTORY OF THE NORTHERN PART OF THE EAST SLOVAKIAN LOWLAND IN THE QUATERNARY

Introduction

The East Slovakian Lowland is the extreme northern part of the vast Great Hungarian Lowland (Alföld). In contrast to the Danubian Lowland (Podunajská nížina) and the lowland called Záhorská nížina it forms a geomorphologically uniform whole of the character of a plain. The rivers (Ondava, Laborec, Uh, Latorica, Bodrog) developed broad flood plains which coalesce with one another. The monotonous ground surface is in the south diversified by aeolian formations, particularly by sand dunes. Between the border mountain range and the lowland there intervenes a *mountain-foot step of hilly relief*, which rises to 150—400 m above sea level. It is built up mostly of volcanic rocks, only at the foot of the Humenské pohorie Mts. the Palaeogene sandstone-claystone facies occurs and Neogene complexes are found in the western part of the Vihorlat mountain group. Various erosional forms, mainly dells, gullies and valleys of diverse types dissect the mountain-foot step, particularly in its western and eastern parts.

Several prominent topographical forms emerge from the lowland, such as the *flat-topped Trhovište—Hrabovec Upland* (Kvitkovič 1961) and the *Biela Hora—Zalužice—Fekišovce Upland.* The former has a moderately rolling surface and rises to 190—230 m a. s. l. It trends roughly from north to south and its relative height drops from 100—120 m to 10—12 m in the same direction. Simultaneously, the intensity of relief and of the erosional dissection into mounds and periglacial hollows decreases from the N to the S. The upland has a straight limitation both in the west against the Ondava plain and in the east from the adjacent flat country. The pre-Quaternary formations making up the upland are dominantly represented by Neogene sedimentary rocks and subordinately by tuffs, tuffites and rhyolites. The Pozdišovce Gravels occuring approximately to the south of Nižný Hrušov build up a thick complex of gravels with clay intercalations. Gravels crop out on the surface in places and as shown in Fig. 5 they are strongly disturbed by periglacial action. Of the Quaternary cover deposits, solifluction and hill-wash colluvial sediments, loess loams and loess are relatively most abundant.

The Biela hora—Zalužice—Fekišovce Upland also shows a moderately undulated smooth relief (Kvitkovič 1961) with the maximum altitude of 159.0 m above sea level (Biela hora). It is formed of clays and sands of the Upper Pliocene and, in the environs of Biela hora, of the Sarmatian and Tortonian sediments and the relics of the *Pozdišovce Gravels*. Quaternary deposits are represented by colluvial-eluvial and loess loams.

The lowland itself is interspersed by abandoned river meanders and south of Michalovce there are well-developed aggradation levees of the river Laborec and its tributaries Duša and Ondava. Between Biela hora and Tibava spreads the *Sub-Vihorlat Quaternary depression*; its eastern part is inclined moderately to the south and bordered by the mountain-foot step on the north.* Its western part between Biela hora and the Jovsa—Hnojné tie line is topographically more marked. This young tectonic depression is still affected by subsidence movements. Its lowest absolute altitude is 103.6 m. In the environs of Senné, in the drainage area of the Čierna voda, it represents the lowest lying tract in Czechoslovakia (98.4 m a. s. l.).

The relief is accentuated by small valley flats of streamlets flowing down the marginal ranges. Near Porostov and Svatuša, the Porostov erosional-tectonic relic rises 2—3 m above the lowland. A sheet of loess loams flanks the Čierna voda river from Vyšné Revíšte southwards to Blatná Polianka and from Hažín south- and south-westwards as far as the southern limit of the area studied. In the south and south-west they plunge under the young fluviatile sediments. South of Nižná Rybnica, through Blatná Polianka towards Senné the monotonous flat relief is diversified by sand dunes and relics of a continuous sand cover rising to a height of 1—4 metres. Isolated rhyolite domes are conspicuous landscape features; these weathered-out relics of rhyolite bodies reach a relative height of up to 110 metres. (e. g. Hrádok near Michalovce). In the area of Petrovce, Naciná Ves and on the Zbudza terrace, there are striking elevations of small size, formed of loam and finely sandy loam, which in the flood plain of the Laborec river stand on the youngest Holocene sediments. They are probably relics of barrows from historical time (?). In the Ondava area, the lowland is a flat river plain traversed by abandoned channels of the Ondava. Topographically marked aggradation levees and inter-aggradation depressions occur occasionally. The unevennesses in the lowland itself do not exceed 3 to 4 metres.

* It is completely filled with solifluction-proluvial sediments capped with loess and fluviatile loams.

The Quaternary development

The Quaternary sediments of the area studied are fairly widespread and show a considerable thickness, different modes of genesis and varied lithological properties. They underwent a complicated development which is reflected in the sediments of individual genetic types. Except for that part of the Trhovište—Hrabovec Upland and Biela hora where the Neogene Pozdišovce Gravels crop out on the surface, they lie on Neogene sediments of a great extent and thickness, on Neogene volcanics and on the Neogene complexes of the Inner Carpathians.

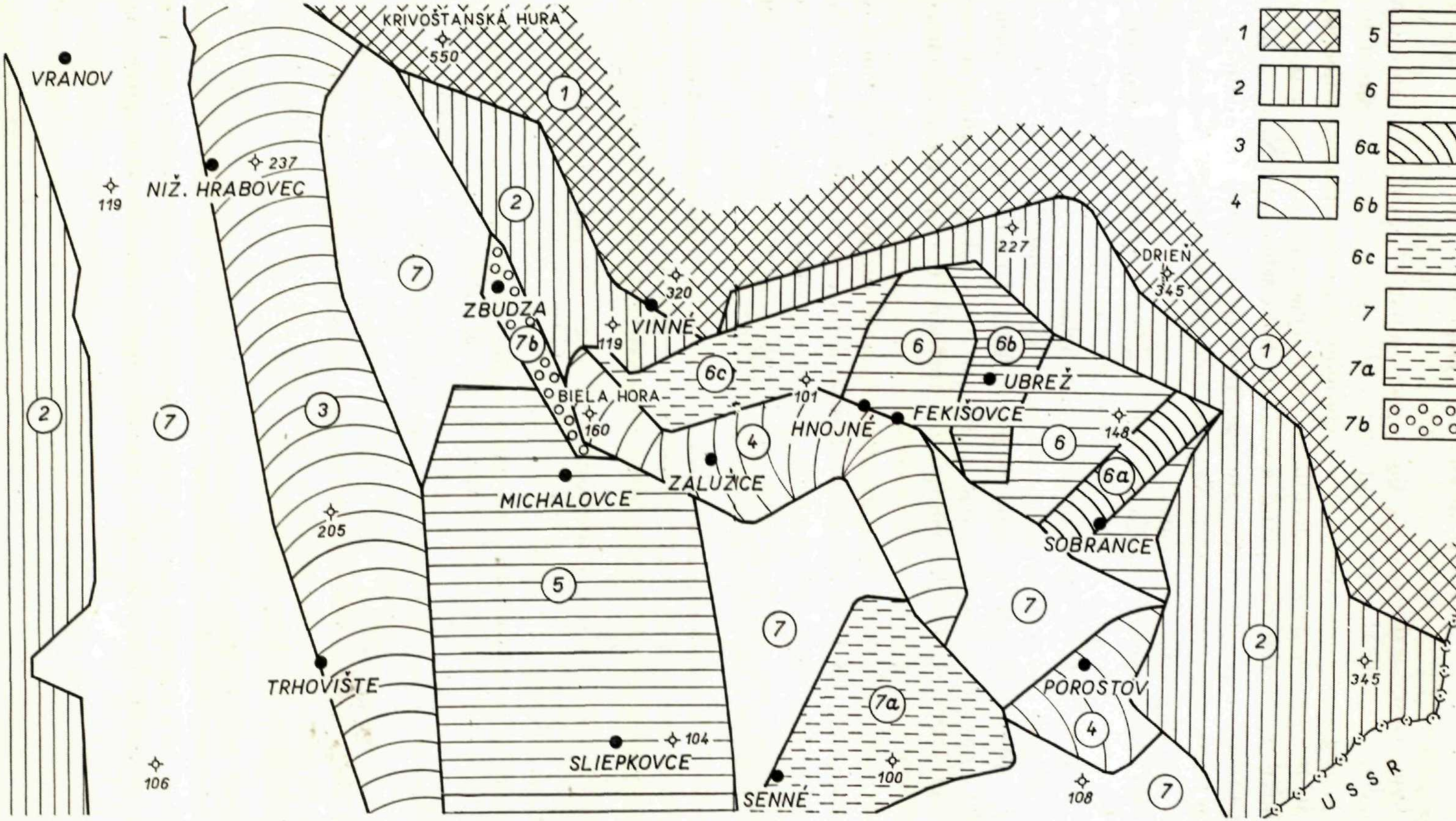
The area under consideration is the key area for the study of the Quaternary of the East Slovakian Lowland. The geological conditions are comparatively complicated. Since the Lower Tortonian the lowland was affected by intensive tectonic movements and volcanic activity which persisted up to the Pliocene, when the present-day surface relief began to form. The subsequent sculpturing of the landscape is already the result of Quaternary evolution which occurred under the influence of the continuing tectonic movements and climatic fluctuations.

The *fluviatile sediments* are of great importance for the study of the geological history of the East Slovakian Lowland during the Quaternary. Their distribution, thickness and evolution are mostly controlled by the tectonic mobility of the area. Apart from the Zbudza aggradational terrace, they usually stand in a normal superposition.

The greatest thickness of fluviatile sands and gravels has been established in the Michalovce—Sliepkovce depression, where the water courses — the most important agents in re-modelling the pre-Quaternary relief — reacted most sensitively to tectonic and climatic changes. The deposition of the Pozdišovce Gravels was followed by intense erosion which removed a considerable part of the Pozdišovce gravel and clay deposits, as evidenced by the relics of gravels on Biela hora. In the south, the Humenské pohorie Mts. was an obstacle to the fluvial activity of the Laborec river. In consequence of the gradual uplift of the Carpathians in the Pliocene, the Laborec developed an antecedent valley through the Humenské pohorie Mts. The relic of a Pliocene surface at an altitude of 230 m a. s. l. in the Brekov gap is a clear evidence. Other Pliocene surfaces are preserved in the Trhoviště—Hrabovec Upland and on the mountain-foot step which extends from Staré through Oreské to Vinné. During the Pleistocene, the antecedence of the Laborec river continued owing to further downmovements in the basin and a moderate uphea val of the border mountains. The river was progressively downcutting and carved a 500—600 m broad valley across the Humenské pohorie — the Brekov gap — through which it enters the East Slovakian Lowland proper.

The period of intensive erosion was followed, probably in the Middle Levantinian, by floods caused by rivers flowing from the north, as a result of tectonic movements accompanied by the last volcanic eruptions. In the area of the present Michalovce—Sliepkovce depression, minor lakes with through-drainage

Fig. 2. Distribution of geologico-tectonic units. Compiled by V. Bàňacký, 1966



1 — Marginal mountain ranges, 2 — mountain-foot step, 3 — Trhovište — Hrabovec elevation, Biela hora — Zalužice — Fekišovce horst structure, 5 — Michalovce — Sliepkovce Quaternary depression, 6 — Quaternary Sub-Vihorlat depression, 6a — buried Sobrance horst structure, 6b — Úbrež partial depression, 6c — young Blatné partial depression, 7 — plain, 7a — young Senné depression, 7b — Zbudza terrace

and isolated swamps came into being. In this environment, sands and gravels were cemented by Fe- and Mn hydroxides precipitated from subsurface waters to form beds, locally up to 2.5 m thick. Occasionally, the gravels are strongly loamified or cemented by clayey material. The absence of pyroxene-andesite material suggests that these gravels were deposited before the last eruptions and effusions of volcanics, being thus of pre-Quaternary, i. e. Pliocene age.

In our opinion, the alluvial cone, preserved as a relic between Vinné and Biela hora (Fig. 1), and basal sands and gravels in the Michalovce—Sliepkovce depression are the oldest Quaternary (Mindel) sediments. The cone is made up of fragments, pebbles and sand of brown-rusty colour. The material of andesites, sandstones, quartzites and limestones is considerably loamified.

Sands and gravels in the Michalovce—Sliepkovce depression are up to 20—40 m thick. Pebbles are little rounded, somewhat more weathered. Compared with the underlying Pliocene gravels, they contain pebbles of pyroxene-andesites, which suggests their deposition subsequently to the last effusions of andesite.

In the Michalovce—Sliepkovce depression these gravels are separated by a loamy-clayey layer from sands and gravels that are ranged to the Mindel/Riss interglacial. E. Krippel refers the following flora from this layer: *Pinus, Abies, Podocarpus, Carya, Picea, Betula, Salix, Alnus, Quercus, Carpinus, Ulmus, Tilia, Corylus*; NAP: *Poaceae, Asteraceae tub., Artemisia, Daucaceae, Silenaceae, Helianthemum, Cyperaceae, Sparganium-type, Typha, varia Polypodiaceae, Equisetum* and *spores.*

On the basis of weaker weathering, superposition, differing lithological properties, the content of pyroxene andesite and cool flora, these upper gravels are dated as the Riss. They are maximum 15 m thick. From this complex the following flora has been determined by E. Krippel: *AP — Picea, Betula, Pinus, Alnus, Carpinus*; *NAP — Daucaceae, Poaceae, Artemisia* and abundance of spores. At that time, the Zbudza terrace originated in the north-western part of the lowland. The material of this terrace consists of well rounded pebbles and sands which are interbedded with seams of medium to coarse-grained sands. The complex is rusty and rusty-brown in colour, with layers of gravel tinted black by Mn-coating. The analyses made by Minaříková has shown that petrographically the gravels are formed of 70—90 % of glauconitic sandstone, the remainder being black and grey chert, greyish-brown quartzites and more or less weathered andesites. In addition, pebbles of rhyolite and andesite tuffs are present in the northern part. The terrace was supplied with material from the mountain-foot step and the border mountains by the streams and solifluction.

Whereas in the Michalovce—Sliepkovce depression the subsidence went on during the Riss, the area of the Zbudza terrace was uplifted and, at the earliest, towards the end of the Riss, sank again. The terrace deposits are 5—12 m thick; at present it lies below the Laborec-river level and only at the north-western margin its surface is flush with it. The terrace was deposited on grey marly clays of the brackish Upper Tortonian. It is covered with 5—15 metres of loess loams and colluvial loams.

In the subsequent period, the climate became warmer and the area of the Sliepkovce—Michalovce depression was covered by warm forest vegetation with abundant swampy forests (with alder) which accompanied the streams. Taking into consideration also the absence of the Fagus pollen grains we may presume that the sedimentation took place in the first half of the last Riss-Würm interglacial. E. Krippel relates the following flora: AP — *Pinus, Abies, Picea, Betula, Alnus, Quercus, Tilia, Acer*; NAP — *Poaceae, Cyperaceae, Daucaceae, Menyanthes, Typha, Artemisia, Centaurea* and varia.

After a short break, the fluviatile sedimentation of sands and gravels, loamy-clayey sediments and locally also of organic sediments and aeolian loams was re-assumed. In the eastern part, sands and gravels are about 3 m thick and in the western part up to 14 metres. Pebbles are less weathered, fresher and intercalated with loamy-clayey sediments in the eastern part of the depression. During this period aeolian activity was fairly strong and weathered material was moved by hillwash from the slopes of the Biela hora—Zalužie—Fekišovce Upland into the depression. This complex is placed into the Würm glacial stage, when loamy organic layers were deposited during warmer interstadials. E. Krippel determined Carpinus, Quercus, Betula, Corylus, Picea and Poaceae and Myriophyllum as representatives of the warm-loving flora of the Würm interstadials.

At the time of the Würm glaciation the Tarnava, Vinné, Kaluža, Klokočov and Kusín alluvial cones (Fig. 1) were deposited on the mountain-foot step. Two of them, the Tarnava and Vinné cones coalesce morphologically. The former is up to 23 m thick and is composed of fragments, pebbles and boulders with sporadic blocks drawn-in by solifluction. The latter is only 4—15 m thick, and strongly loamified. The poor sorting of the material is due to a short transport. The rock debris was deposited in a separate, small, almost closed depression. The Klokočov, Kusín and Kaluža cones forming another group are fanwise spread and submerge into the young Blatná depression. The Kusín cone is of the greatest thickness — up to 28 metres.

In the wide area studied loess loams were deposited from the beginning of the Würm stage. They are widely distributed, occurring on the mountain-flood step, flattish upland and in the lowland itself. They are up to 20 m thick. In the Krčava brickyard and near Zbudza (in the Laborec valley) they are well exposed. On the Zbudza terrace, the loams are interlayered by three fossil soils which, according to Vaškovská's studies and by analogy with the Trnava and Hron plateaus, probably date from two interstadials and one interglacial. The uppermost soil, better to say, the soil relic, corresponds to the W 2/3 interstadial, the middle soil to W 1/2 and the lowest one to the Riss/Wurm interglacial (Fig. 3, Photograph pl. V).

Ložek (1963) refers a relatively great amount of *Pupilla sterri* (VTH), *Pupilla muscorum* (L), *Vallonia tenuilabris* (A. Br) from the loesses adjacent to the Humenské pohorie Mts., but does not mention their stratigraphical assignment. Towards the end of the Würm stage the deposition of aeolian loess took place; relics of the loess cover are preserved in the eastern part of the Michalovce—Sliepkovce depression, yielding remains of malacofauna. Schmidt (1966) determined *Trichia hispida* (L.), 7 specimens, and *Succinea* s. *Oxytoma*, 10 specimens. In this period, gravel aggradation in the western part of the depression died out (Table 2).

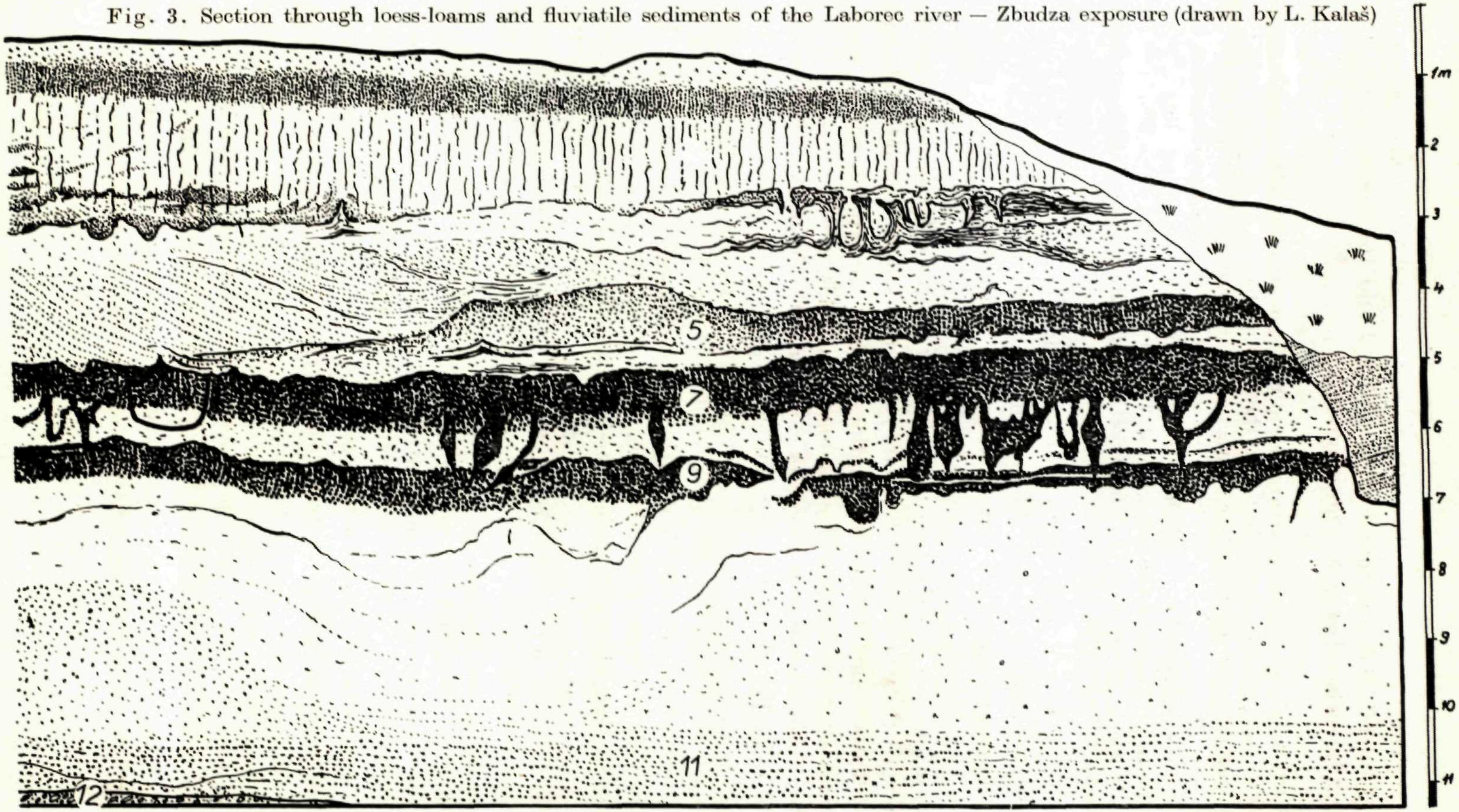
Subsequently to the accumulation of sands and gravels, fine wind-blown sands were deposited (e. g. near Žbince, in the environs of Blatná Polianka and Senné) presumably at the end of Würm and at the beginning of the Last Glacial.

The following period of the Last Glacial is distinguished by extensive floods which changed the whole area into swamps and fens silted by flood muds and partly by loess loams washed down the adjacent elevated grounds. Many swamps and stagnant waters of abandoned channels grew over by paludal vegetation. Various hydrogene processes gave rise to the deposition of a greyish-blue to blackish-blue loamy-clayey complex with occasional lenticles, seams and beds of peat, sapropel and sand. According to the results of pollen analyses, these deposits date from the cold interval of the Last Glacial.

In the north-western projection of the lowland, sands and gravels of the Laborec flood plain were covered by loamy and loamy-clayey sediments at that time. In fact, they are re-deposited loess loams and colluvial material which at the end of the Würm formed a continuous cover. The drop of the base-level of erosion in the Last Glacial resulted in intensive erosion. From the sedimentary cover only a relic was preserved below the eastern foot of the Trhovište—Hrabovec Upland.

The subsequent evolution up to the Atlantic occurred under the influence of increasing temperature and humidity and a uniform subsidence of the area. Not only in the depression but also in the adjacent areas clayey-loamy flood sediments were intensely aggraded. Towards the close of the Atlantic and at the onset of the Subatlantic, the streams were at spate and, as a result, swamps developed in the eastern part of the depression, where conditions farourable for the formation of organic loamy-clayey sediments with layers of peat and sapropel had existed. This local sedimentation was followed in the Subatlantic by the deposition of flood loams of loessic character which covered the whole valley flat of the Laborec river. The sedimentary period ends with the aggradation of levees along the river (Table 1).

Fig. 3. Section through loess-loams and fluviatile sediments of the Laborec river – Zbudza exposure (drawn by L. Kalaš)



1 – Dark-grey humic soil, 2 – loess loam, 3 – convolutions, 4 – loess loam, 5 – buried fossil soil of Würm 2/3, 6 – loess-loam, 7 – buried fossil soil of Würm 1/2, 8 – loess-loam, 9 – buried fossil soil of Riss/Würm, 10 – flood muds, gleyed sandy-loamy at the base, 11 – fine grained loamy flood sands, 12 – sands and gravels

It should be mentioned that in the concave bank of the Ondava river a complex is exposed, the upper part of which consists of 2 m-thick sub-Recent flood loams and sands with paludal soils. The latter form three to five layers in the south and one to two in the northern part. The swampy horizons cannot be connected with climatic fluctuations during the Holocene nor be regarded as stratigraphical key horizons. From the number of layers of flood loams and paludal soils, it is only inferable that the valley flat underwent a polyphasal development; the flood loams and sands are distinctive of the period when the aggradation of levees was intensified. The muds buried the Holocene humic paludal soils (1 m thick) developed on Pleistocene loess loams, 2—3 m in thickness. Below the gleyed loamy and loamy-clayey sediments (3—6 m) in the valley flat, which are locally rusty stained by Fe hydroxides (Plate VI), there is invariably a 2 m-thick layer of fluviatile sands. These in turn overlie a 2—3 m thick pedocomplex with clayey sedimentation, which is superjacent on sands and gravels emplaced in the sequence of beds building up the flood plain (Fig. 4). The gravels, which are an important component of the lowland structure do not crop out on the surface, being exposed only in seveal erosional valley sides of the Ondava river.

In the Sub-Vihorlat depression the Quaternary history was different from that of the Michalovce—Sliepkovce depression. In the Upper Pliocene (Levantinian) the Sub-Vihorlat area sank and an extensive lake came into being, in which clays with lignite seams were laid down. In the proximity of the Klippen Belt, strong eruptions of pyroxene andesite took place. After a partial regression of the lake, the material of the weathering marginal parts of the volcanic range was transported into the central part of the depression which still at the beginning of Pleistocene was flooded. Variegated clays (brown, dirty-brown, yellow-brown, bluish-brown and grey) of tuffaceous character and with pebbles and fragments of andesite and sapropel partings were deposited in this environment.

In the Sub-Vihorlat depression, the geological Quaternary history began with the deposition of the variegated complex which, as mentioned above, passed without break of sedimentation into the Early Pleistocene, representing the oldest Quaternary sediments of the East Slovakian Lowland and the contiguous area. In the depression, the clays of the variegated complex with sapropel and peat beds were encountered in borehole VN-18.

The pollen analysis of beds from a depth of 9.8—11.3 m (made by Planderová) revealed the following main species of the Upper Pliocene flora: *Alnus, Castanea, Pinus diploxylon, Polypodiaceae, Carya, Sphagnum, Stereosporites-type; Pinus haploxylon, Tilia, Cupuliferae, Quercus, Abies, Pterocarya, Carpinus, Salix* and *Ericaceae* are subordinate; NAP: *Asteraceae*. The composition of vegetation is similar to the pollen spectrum from the Pliocene (Variegated complex) of the south-western part of the West Carpathians. Some Miocene elements are absent, such as *Myricaceae* and *Taxodiaceae*, which in southern Slovakia penetrate sporadically into the Variegated complex. Thus, the floral assemblage represents presumably the uppermost Pliocene, in our opinion — the Levantinian. The Variegated complex persists upwards, bearing thin peat interbeds; at 6.9—7 m it yielded an abundance of pollen grains. Tertiary vegetation elements were totally absent from the sample taken at this depth (except for *Caryum* which occurred sporadically). On the

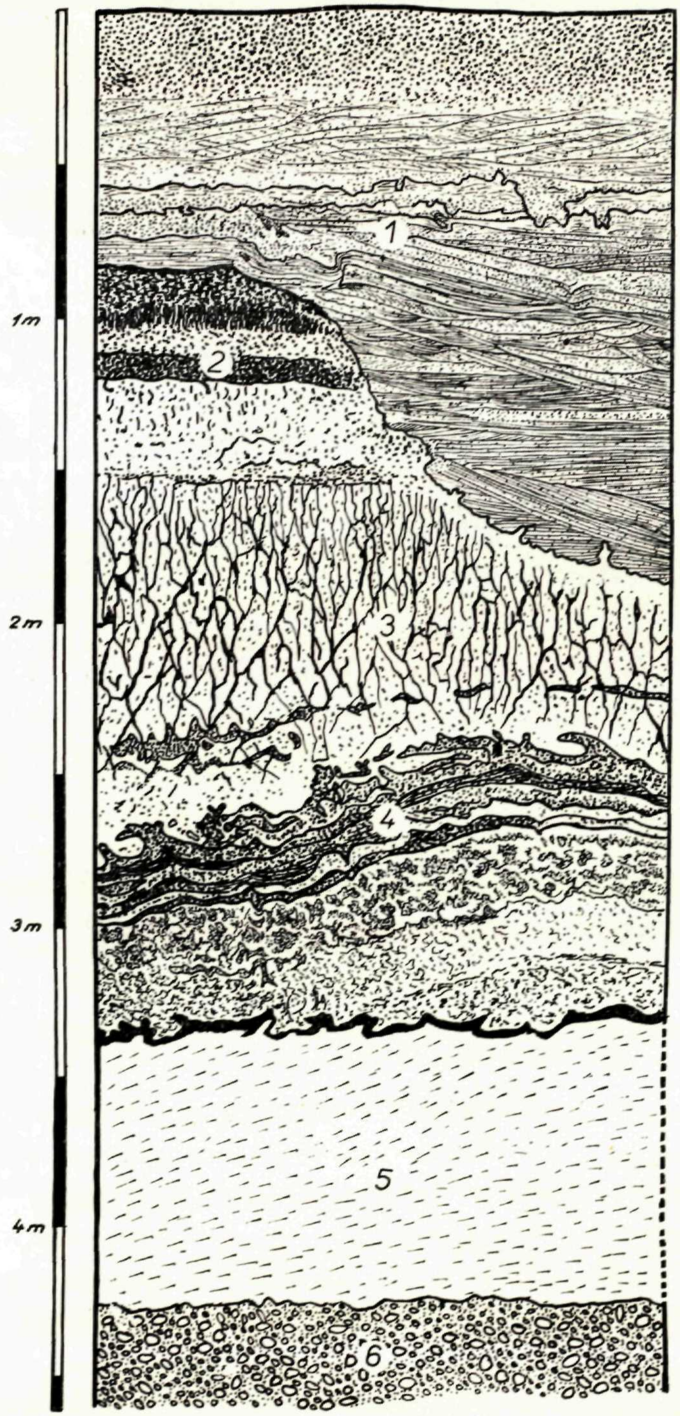
Tab. 1.

Stratigraphic table of the late and post-Glacial of the northern

	Division acc. Firbas	Division acc. Ložek (1953)	Absolute Chronology	Fluvial and organic sediments	Fluvial and organic sediments of the Blatná depression
post-Glacial – Holocene	X	sub-recent	+2 000 +1 200	loam, fine sandy loam	sandy-loamy, clayey-loamy and peat sediments, on the surface with viscous and peat soils
	IX	sub-Atlantic	500	loam, with predominating of inundation mass	
	VIII	sub-Boreal	−2 500		loamy sediments with sporadic pebbles
	VII	late Atlantic	4 000		erosion – interruption of the peat sedimentation
	VI	early Atlantic	5 500	clayey-loamy sediments with sapropel and peat layers	
	V	Boreal	6 800		peat
	IV	prae-Boreal	8 100		loamy sediments
Late Glacial	III	late Dryas	9 000	greyish blue to dark blue loamy-clayey beds with sandy, sapropel and peat layers	peat and mudflows
	II	Allerődian	10 000	loamy-clayey beds near the E hillfoot of the Trhovištsko-hrabovecká pahorkatina hills	
	I	early Dryas	15 000	termination of the gravel-sandy sedimentation	coarse grained clayey sands, loam and gravel layers

part of the East Slovakia lowland compiled by V. Baňacký 1967

Solifluction-fluvial sediments	Deluvio-fluvial sediments hillwash colluvium	Eolitic sediments
	deluvio-fluvial loamy cone on the E hill-foot of the Trhovišťsko-hrabovecká pahorkatina hills; stabilization of the slide activity, termination of deluvial stony sedimentation; loamy deluvial sedimentation (in the Vihorlat Mts.)	
	spatial rich representation of deluvial stony sediment and slides in the Vihorlat and Popričný Mts.	
loamy cones (on the W hillfoot of the Trhovišťsko-hrabovecká pahorkatina hills)		finest grained powdery sands (Žbince, Blatná Polianka)


1m
2m
3m
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1
2
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6

whole conifers were dominant, especially *Abies, Picea, Cedrus* and *Pinus diploxylon type*; they occur together with the sporiferous *Polypodiaceae* family. Other types were found sporadically (*Osmunda, Juglans, Tilia,* more abundant *Alnus, Betula, Typha,* less *Ericaceae, Lycopodium, Carya, Pterocarya* and NAP). The composition of the vegetation suggests the Early Quaternary age and no longer the Pliocene. This finding confirms the opinion on a gradual transition of the Variegated complex from the Tertiary into the Pleistocene.

After the sedimentation of the Variegated complex was brought to an end in the final phase of the Günz glaciation, an extremely strong erosion was active in the Sub-Vihorlat depression, being caused by the termination of synsedimentary tectonic movements. The bulk of the Variegated complex, for the most part of Early Pleistocene age succumbed to denudation. The oldest Quaternary sediments were preserved only in the western part of the Sub-Vihorlat depression, but in a far smaller thickness than they had originally. After the period of intense erosion towards the end of the Günz glaciation the clastic material of the volcanic mountains moved downslope and was transported into the depression. Thus, during the Mindel glaciation a voluminous mass of solifluction-proluvial material progessively filled the sinking depression.

In the Mindel-Riss interglacial strongly loamified deposits without clastic material were laid down. In the Riss the conditions were favourable to intensive weathering, the products of which are preserved as solifluction-colluvial deposits over a large area of the mountain-foot step. The loamified sheets of fragmentary-boulder-block material are most extensive. The mass of solifluction material was formed in periglacial periods when, under suitable petrographical conditions, products of strong frost-weathering was removed by solifluction form the mountain slopes and deposited on the mountain-foot step. The material of thick solifluction covers ranges in size from small fragments to 2 m large blooks (Plate VII). The covers reach the greatest thickness (14—15 m, max. 22—23 m) to the east of Jovsa. In places solifluction-colluvial deposits were completely removed by denudation.

In the R $^1/_2$ interstadial the deposition of proluvial material was interrupted and a loamy horizon developed. The solifluction-proluvial sedimentation was re-assumed in Riss 2 and in the Sub-Vihorlat depression a large amount of loamified proluvia was piled up (Plate VIII). Towards the end of the Riss glacial and at the onset of the Würm the watercourses transported proluvial material from the Vihorlat-Popričný Mountains into the lowland, where layers

◁

Fig. 4. Section through the flood-plain sediments of the Ondava river (drawn by L. Kalaš)

1 — Light-brown sandy-loamy flood muds, 2 — black humic subhydatic soil, 3 — light-brown-yellow loess loam, 4 — convolutions, 5 — dark-grey humic subhydatic soil, with convolutions on the surface, 6 — sands and gravels

Tab. 2.

Stratigraphic table of the Pleistocene of northern part

		Alpine glaciation	Fluvial sediments	Solifluction-fluvial sediments of proluvial cones	Loess sediments
Pleistocene	young	Würm 3 Stadial	sands-gravels with inundation loams		loess sediments in the Michalovce – Sliepkovce-depression, i. e. marshy loess (Iňačovce); loess loams near Krivošťanka, loess loams of the hill-foot etage
		Würm 2/3 Interstadial	clayey-loamy sediments		
		Würm 2 Stadial	sands-gravels, sands and loamy sediments	proluvial cones of Tarnava and Vinné with loamy-fragmentary sediments	loess loams of the hill-foot etage (Zbudza – Krčava), loess of Bánovce – Ložín – Malčice
		Würm 1/2 Interstadial	loamy-clayey sediments		
	middle	Würm 1 Stadial	sands-gravels, sands and loamy sediments		loess loams of the hill-foot etage (Zbudza – Krčava)
		Riss – Würm Interglacial	clayey-loamy sediments		
		Riss Glacial	sands-gravels in the depressions, sands-gravels and mudflows of the Zbudza terrace		
	old	Mindel – Riss Interglacial	clayey-loamy sediments sporadically with pebbles		
		Mindel Glacial	sands-gravels	sandy-loamy-fragmentary proluvial cone between Vinné and Biela Hora	
Neogene	young Pliocene	Levantinian	gravels to sandy gravels, clays		
		Pontian	gravels and variegated clays (Variegated series) of the Pozdišovce gravel formation		
		Panonian	erosion — denudation		

of the East Slovakia lowland compiled by V. Baňacký 1967

Petrographic characteristics	Fossile Flora	Fossile Malakofauna
paleogene sandstones predominate with admixture of pyroxene andesite; 80 % of the heavy min.: ore-minerals		near Krivošťanka: Pupilla sterii (VTH), P. muscorum (L.), Valloni tenuilabris (A. Br.) in Michalovce – Sliepkovce depr. Trichia hispida (L.), Succinea s., Oxytoma sp.
paleogene sandstones predominate with admixture of pyroxene andesite		
	AP: Carpinus, Quercus, Betula, Corylus, Picea NAP: Poaceae, Myriophyllum	
paleogene sandstones predominate with admixture of pyroxene andesite		
	AP: Pinus, Abies, Picea, Betula, Alnus, Quercus, Tilia, Acer NAP: Poaceae, Cyperaceae, Daucaceae, Menyanthes, Typha, Artemisia	
sandstones with enclosures of the pyroxene andesite; heavy min.: more hypersthene augite, 3 – 4 % hornblende	AP: Picea, Betula, Pinus, Alnus, Carpinus NAP: Daucaceae, Poaceae, Artemisia	
loamy-limonite konkretions predominate; heavy min.: ore-min.	AP: Pinus, Abies, Podocarpus, Carya, Picea, Betula, Salix, Alnus, Quercus, Carpinus, Ulmus, Tilia, Corylus NAP: Poaceae, Asteraceae, tub., Artemisia Daucaceae, Silenaceae, Helianthemum, Cyperaceae, Sparganium-typ, Typha,	
paleogene sandstones predominate, sporadically with pyroxene andesite; more tuffs, tuffits; heavy min.: hypersthene and sporadic augite		
glaukonite sandstones (50 – 60 %), rest hornblende; pyroxene andesite absent, in heavy minerals garnet		

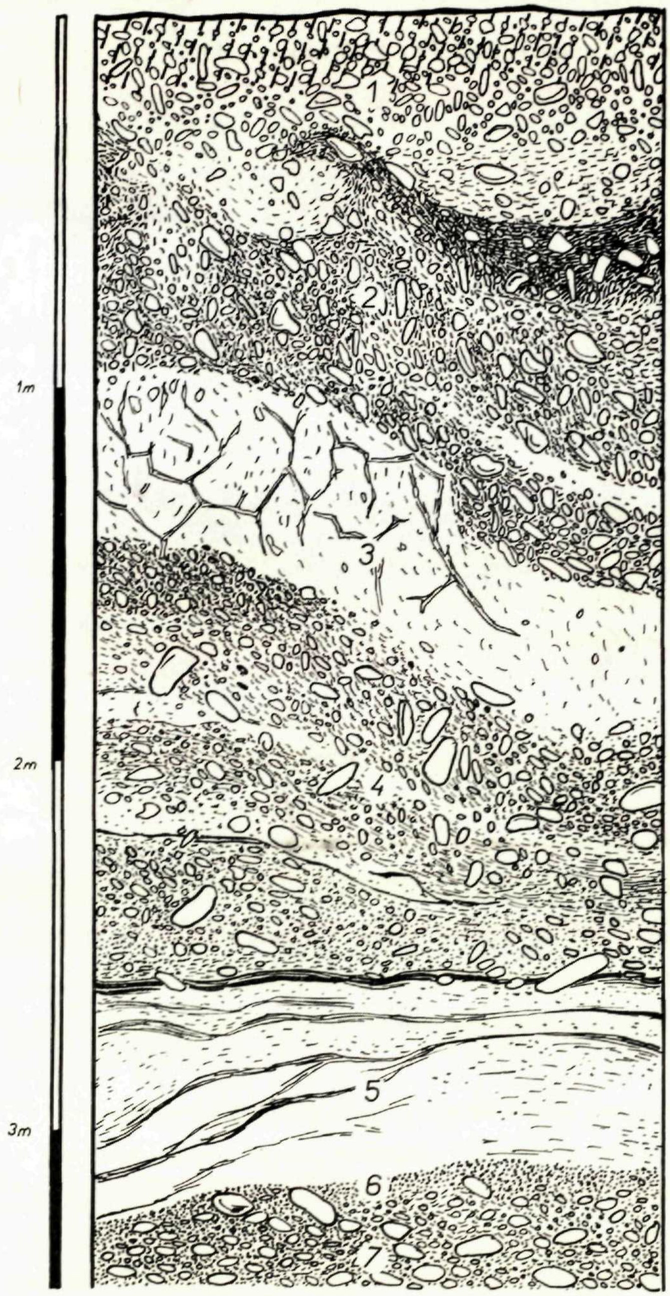
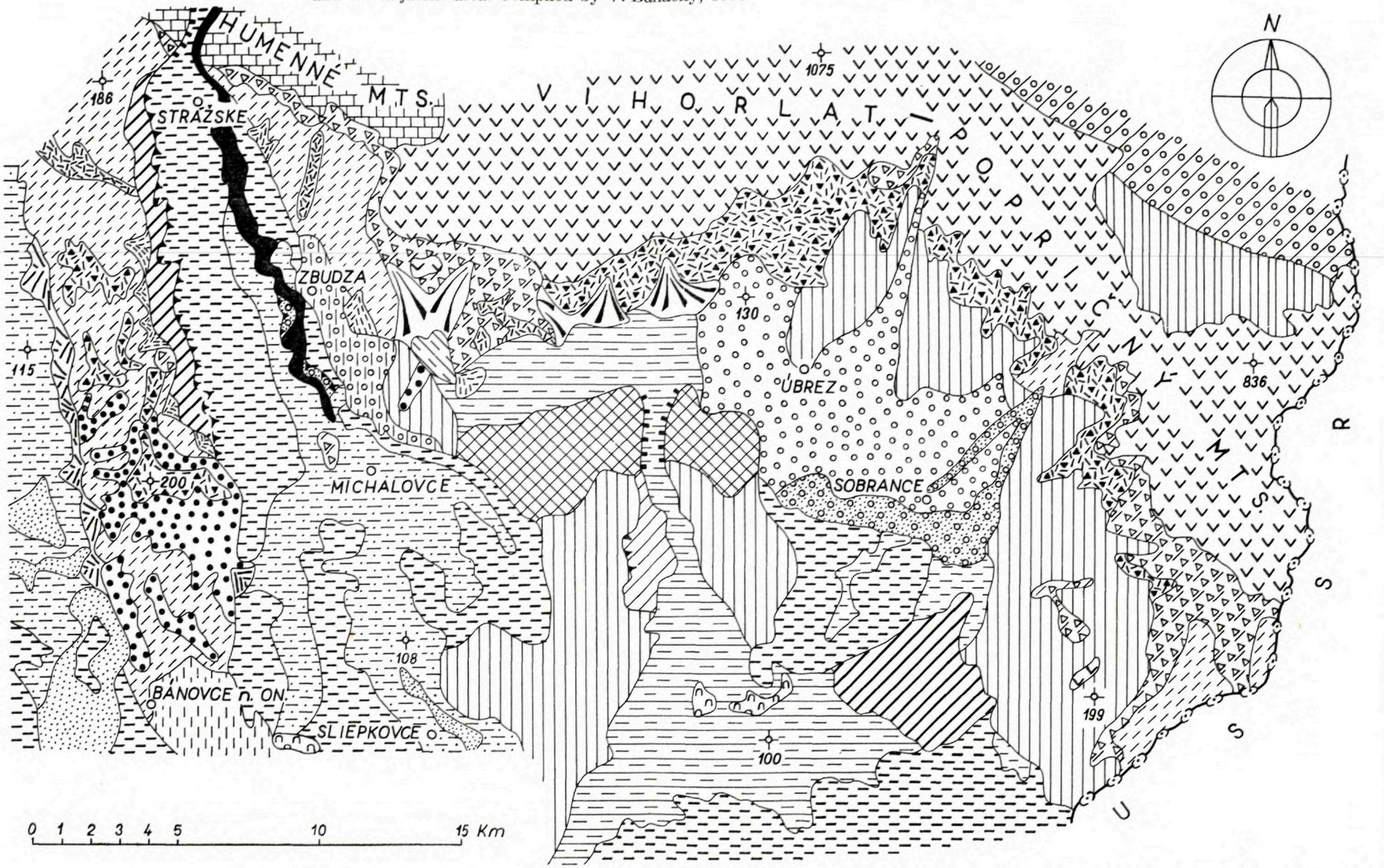
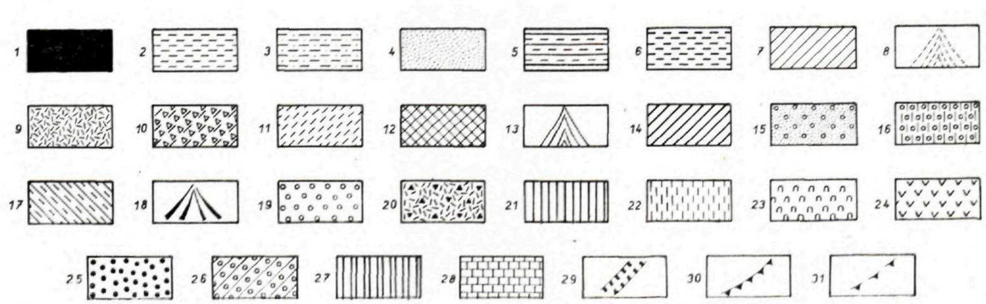


Fig. 5. Pozdišovce Gravels at the southern outskirts of Pozdišovce, convoluted and disturbed by cryoturbation (drawn by L. Kalaš).

1 – Strongly loamified light-brown grey gravels, 2 – brown loamified gravels with randomly scattered pebbles, 3 – rusty brown clay with cracks filled with grey and whitish-grey clay, 4 – dark-rusty-brown gravels with randomly scattered pebbles, 5 – whitish grey clay with rusty smudges, 6 – rusty-brown medium-grained sand, 7 – rusty-brown, slightly loamified sands and gravels.

Fig. 1. Quaternary geological map of the northern part of the East Slovakian Lowland and the adjacent area. Compiled by V. Baňacký, 1966





1 – 8 *Holocene sediments:* 1 – 5 younger fluviatile deposits: 1 — Subrecent sands and gravels, 2 — loam 3 — finely sandy loam, 4 — fine-grained sand, 5 — clayey-loamy and loamy-sandy sediments in the Blatné and Senné depressions; 6 – 7 early fluviatile deposits: 6 — loam, 7 — clayey-loamy material with sand seams; 8 — colluvio-fluvial, dominantly loamy cone; 9 – 12 *Holocene up to Pleistocene sediments:* 9 — stony and loamy-stony eluvium, 10 — loamy-stony hillwash colluvium, 11 — loamy hillwash colluvium, 12 — loamy eluvio-colluvial sediments; 13 – 23 Pleistocene sediments: 13 – 16 *fluviatile deposits:* 13 — Last Glacial, — predominantly loamy cone, 14 — Last Glacial — clayey-loamy and loamy sediment, 15 — Würm sands and gravels, 16 — Riss sands and gravels of the Zbudza terrace with a loess-loam cover (max. 15 – 18 m); 17 – 18 *solifluction-fluvial sediments:* 17 — Mindel — loam with weakly worn rock fragments and sand, relic of a cone, 18 — Würm cone of loamy-stony material; 19 — Riss solifluction-proluvial sediments, weakly rounded fragments, pebbles and boulders with loam admixture, 20 — colluvial-solifluction sediments — fragments, boulders and blocks with loam admixture; 21 – 23 *aeolian deposits:* 21 — Würm loess, 22 — Würm loess loams, 23 — fine-grained sands of Würm-Late Glacial stage; 24 – 28 *pre-Quaternary formations:* 24 — volcanics of the Vihorlat-Popričný mountain group, 25 — Pozdišovce Gravels of Pontian age, 26 — Intracarpathian Palaeogene, 27 — Klippen Belt, 28 — Mesozoic of the Humenské pohorie Mts., 29 — antecedent valley, 30 — prominent edges in the terrain, 31 — buried or denuded edges in the terrain.

of sandy gravels and sands were deposited in normal superposition. Gravels are well sorted, pure (only locally loamified) and well permeable. They settled down in the phases of intensified stream activity, whereas sand beds forming in places continuous sheets sedimented in tranquil phases.

The beginning of the Würm was distinguished by the removal of slope debris by streams onto the slope of the mountain-foot step and in to the western part of the Sub-Vihorlat depression to form there cones and fans. Soon afterwards, these were disturbed and their material was transported to the centre of the western part of the Sub-Vihorlat depression, where a sequence composed of clayey-loamy sands and gravels with fragments of volcanic rocks was piled up. These conditions persisted throughout the Würm up to the beginning of the Last Glacial (early Dryas). In the whole Sub-Vihorlat area intensive sedimentation of loess occurred in Würm time.

In the late Dryas, the western part of the Sub-Vihorlat area (Blatná depression) was the site of peat and sapropel development which took place on the wet grounds and swamps. In this depositional environment flood loams were deposited, the sedimentation of which ended by the beginning of the Preboreal. Favourable conditions for the formation of peat persisted in small depressions during the whole Preboreal to Boreal. The sedimentation of peat and sapropel was interrupted in the Atlantic stage by erosion. In the course of this stage until the beginning of the Subatlantic, poorly rounded pebbles of andesite with loam admixture (abrupt sedimentation) were laid down. They pass upwards into clayey and loamy sediments (tranquil sedimentation). This succession of strata repeats three times.

From the early Atlantic until the Subrecent the top part of the depression was covered by clayey-loamy and loamy-sandy sediments on which peat soils and gley soils developed.

In the eastern part of the Sub-Vihorlat depression the loess sedimentation was replaced by strong erosion which removed a great part of loess loams; they succumbed completely to erosion in the section from Úbrež to Tibava and solifluction proluvial deposits were laid bare there. In the Last Glacial until the early Holocene, on the major part of the area affected by erosion, fluviatile loams (1—3 m) derived mostly from the loess covers sedimented.

In the last, Late Holocene phase, the eastern part of the Sub-Vihorlat depression was covered by loamy and partly also gravel sediments which fill mostly narrow valley flats of the streams (Table 3).

The East Slovakian Lowland underwent a similar tectonic process in the Quaternary as the other lowlands of the West Carpathians: the predominating differential movements mostly acted on older fault lines, but the tectonic development in Neogene and probably also in the earlier phases of the Pleistocene was influenced by volcanic activity.

Tab. 3. **Stratigraphic table of Pleistocene of the Sub-Vihorlat depression compiled by V. Baňacký 1967**

Stratigraphic		Alpine glaciation	Fluvial sediments	Solifluction-proluvial dep.	Solifluction-proluvial cones	Loess loams
Pleistocene	young	Würm glacial	sands and gravels, loamy sediments		loamy-stony cones of Kaluža, Klokočov Kusín	loess loams of Úbrež – Jasenov
	middle	Riss 2 Stadial		fragments, pebbles and boulders with loamy admixture		
		Riss 1/2 Interstadial	clayey-loamy sediments			
		Riss 1 Stadial		fragments, pebbles and boulders and boulders with loamy admixture		
	old	Mindel-Riss Interglacial	clayey-loamy sediments			
		Mindel Glacial		fragments, pebbles and boulders with loamy admixture		
			variegated beds (only in the W part) erosion			
young Pliocene		Levantinian	variegated beds			

Towards the end of the Pliocene the relief of the East Slovakian Lowland was considerably levelled (Kvitkovič 1959, 1961). At that time, tectonic movements slackened and the ground surface was levelled down, as evidenced by the relics of the Upper Pliocene surface in the piedmont and the Trhovište—Hrabovec Upland lying 80—100 m above the lowland (Kvitkovič 1961). The investigation of the structure of Quaternary sediments based on extensive boring works, allround geological appraisal and geomorphological studies by Kvitkovič (1959, 1961) confirm that in the individual parts of the northern tract of the East Slovakian Lowland uneven movements of both subsidence and uplift character occurred, even if the lowland as a whole sank relatively to the adjacent mountain ranges. At present, the following geological-tectonic units may be differentiated in this area: Marginal mountain ranges, mountain-foot step, Trhovište—Hrabovec elevation, Biela hora—Zalužice—Fekišovce horst structure, Michalovce—Slipkovce depression, Quaternary Sub-Vihorlat depression, buried horst structure of Sobrance, Úbrež partial depression, young Blatná partial depression, plain, young Senné depression, Zbudza terrace.

The marginal mountain ranges (Vihorlat—Popričný and Humenské pohorie) were rising during the Quaternary, as evidenced by the antecedent valley of the Laborec in the Humenské pohorie with preserved relics of the successive surfaces. In the course of moderate uplift, the Laborec river deepened its valley to about 115—120 m. According to Kvitkovič (1960), the boundaries of the Humenské pohorie are rectilinear on either side. Facetted surfaces of lateral ridges and the topographical position of the mountains suggest that they are a horst (Kvitkovič 1960).

The mountain-foot step shows a character of a moderate uplift. Its contact with the mountain range is tectonic, initiated in the pre-Quaternary period. The boring works carried out for mapping purposes confirmed this assumption.

The Trhovište — Hrabovec elevation. The W and partly also E limitation of the upland by steep, short and even slopes of fault character indicates that the upland represents an elevation not only in geomorphological but also in tectonic sense. The elevation character of the upland is especially due to the intense downmovements of blocks to the east and west, subsequently to the formation of the riverine plain in the Upper Pliocene, which was attended by lateral erosion. From the disposition of slopes in the individual parts of the upland and from the relationship of the flat upland ridge to the long profile of the Ondava river, a differential character of the movements is inferable. The subsidence of the south-western part was more intensive than that of the area to the north-east of the upland; as a result of greater downmovements of the southern part (relatively to the northern part) the riverine plain tilted towards the centre of the basin.

From the data on the northern part of the lowland and adjacent areas it may be concluded that during the Pleistocene the pre-Quaternary relief underwent considerable changes.

In the area studied there are two Quaternary depressions with sedimentary fillings; they are called the *Michalovce—Sliepkovce* and the *Sub-Vihorlat depressions*.

The Michalovce—Sliepkovce depression of Quaternary age is bounded by rectilinear slopes of fault character in the west. The Biela hora—Zalužice—Fekišovce horst structure and its submerged southern part forms its eastern boundary. In the north the depression is bordered by the upheaved Upper Tortonian block (Janáček 1961) and in the south the limit is not quite clear. It is possible that the depression extends southwards and becomes shallower. This conception is to a certain degree supported by the results of Pospíšil's hydrogeological studies. This author records that the ground waters display a fan-shaped flow pattern and turn from the southern direction to the south-eastern and south-western as if they came across a barrier. The depression is filled with a Quaternary sedimentary complex up to 70 m thick. The upper part (4—10 m) consists of loamy aluvia with peat and sapropel and is underlain by fluviatile sands and gravels with two interglacial and two interstadial horizons. The results of investigations suggest that the depression started to subside in the younger phase of the Early Pleistocene (Mindel) and enlarged its extent with the increase in subsidence. The correlation of the thickness of sediments from the individual Quaternary phases with the length of these (Mindel beds — 30 m, Riss beds — 17 m, Würm beds — 15 m, Holocene — 8—10 m) shows that the intensity of subsidence has increased. Whereas in the Early and Pleistocene it amounts to about 50 m, it results in the deposition of a sequence up to 30 m thick during the Würm and Holocene which cover a much shorter time interval. The intensity and amount of movements are evidenced additionally by the conditions of the Zbudza terrace, which I establish as an, even tectonically, separate unit, because its height and space relations were considerably altered by the movements. This terrace developed in the Early Riss but soon afterwards it sank and was disrupted under the influence of movements in the Michalovce—Sliepkovce depression.

The Sub-Vihorlat depression is differentiated into the western and eastern parts separated by a fault line. As mentioned above, the top part of the Variegated complex of the western part is of Early Quaternary age. It presumably represents the deposits of the final phase of Upper Pliocene subsidence dying out in the earliest Pleistocene. At that time, the Biela hora—Zalužice—Fekišovce horst underwent a gentle uplift and prevented partly the removal of the rocks of the Variegated complex farther to the south. In consequence, they are preserved in the W part of the Sub-Vihorlat depression.

In the following period of relative tectonic rest a considerable part of the Variegated complex from the earliest Pleistocene was denuded. Intensive subsidence did not begin before the Mindel, when the Úbrež depression sank 18—22 m. In this time interval *the Sobrance horst structure* was upheaved and many springs issued on its slopes. In the Riss the area sank again; in Riss 1 the subsidence reached approximately 10 and in Riss 2 it amounted to 18—20 m. The depressions were filled with solifluction-proluvial sediments which are intercalated by interglacial and interstadial loamy-clayey horizons.

In the western part, tectonic and sedimentary rest lasted until the Würm. This part of the depression (Blatná) began to sink in the Würm and the down-movements have persited until the present time. Additonal evidence of its subsidence is provided by the alternation of the loamy, clayey-loamy and loamy-sandy sediments, with an admixture of pebbles and fragments of cone material, with horizons of sapropels and peat. These beds are also dated phytopaleontologically. The occurrence of peat and gley soils and annual flooding of the depression likewise indicate the sinking of this area.

The Senné depression developed in the structural plain during the Würm and Holocene. The depression is flooded annually, except for the summer and early autumn months. It is one of the lowest lying places in Czechoslovakia.

Intensive sinking movements in the latest Pleistocene and Recent are also inferable from the buried sand dunes by the youngest fluviatile sediments and from the swampy character of the depression.

The great thickness of Holocene alluvia not only in depressions but also farther to the south, virtually throughout the lowland, reveals that this subsided as a whole. This overall subsidence caused the burial of depressions which are not apparent topographically; they have been ascertained by boring.

In the marginal parts of the lowlands, in a number of places relatively more stable tracts exist, forming elevations, such as, for example, *the Biela hora—Zalužice—Fekišovce horst structure* (Kvitkovič 1961). As already mentioned, the structure started to rise in the Early Quaternary and prevented the removal of rocks from the western part of the Sub-Vihorlat depression towards the outh. Between Lúčky and Závadka the horst is cut through by the antecedent valley (Kvitkovič 1961). It is built up of Upper Pliocene sediments and its uplift took place over the great part of the Quaternary. Some other horsts (e. g. Sobrance horst) are hidden beneath the Quaternary sediments which also points to intense young subsidence. All these facts show that in the East Slovakian Lowland (compared with the Záhorská Lowland and the Danube Lowland) the conditions were specific to a great extent.

The complicated tectonic evolution of the area continued in the Quaternary; the intensity of block movements did not decrease, on the contrary, in some periods it increased and attained its maximum in the youngest Pleistocene phases and in the Holocene. Stratigraphical results allow to estimate the age and duration of the movements, etc. It is postulated that the Quaternary tectonic movements do not only represent the dying-out of the Upper Pliocene phase but that they represent a new, separate phase in both the character of movements and extent, which commenced in the Early Pleistocene (Mindel). This is suggested by the Early Quaternary beds in the western part of the Sub-Vihorlat depression, their abrupt termination and further sedimentation reassumed not before the Late Pleistocene. Such conditions indicate that the Upper Pliocene phase of subsidence could extend up to the earliest Pleistocene, when it receded to a period of tectonic rest (viz. relative rest with regard to synsedimentary subsidence), as proved by stratigraphical and lithofacies studies of Quaternary complexes in the depressions. These conclusions also indicate a plausible connection of the last andesite effusions in the Vihorlat and Popričný mountain groups with the beginning of the Quaternary tectonic phase in the lowland. According to this conception, the effusions of presumably Quaternary volcanics occurred in the later Pleistocene phases — in the Mindel.

The Quaternary area of the northern part of the East Slovakian Lowland is geologically quite particular and very complicated. The solution of the basic Quaternary-geological problems must be based on integrated investigation, detailed geomorphological analysis and palynological studies. Complicated spatial and superpositional relations of the individual genetic-tectonic-geomorphological units, of sediments and forms, or beds and formations call for both a detailed analysis and synthesis using the data of wide regional and thematical bearing. The findings and conclusions on the geological structure of the area studied are meant to be a contribution to the study of the Quaternary structure and evolution of the whole East Slovakian Lowland. The data related proved first a substantial influence of volcanic activity on the post-volcanic geological history of the region, as well as the great importance of synsedimentary and postsedimentary (adjusting) block movements differentiated in time and space within the framework of the general subsidence of the area. Of no less importance were antithetic movements, i. e. the uplifts of border mountain ranges and the subsidence of the lowland. They exerted direct influence on the intensity of erosion and aggradation processes, which enable us to trace and date the intensifying tectonic activity in the later phase of the Early Pleistocene, subsequent to a period of relatively tranquil period. The finding that geological activity has been fairly intensive even in the youngest Pleistocene phases and in the Holocene may also imply a substantial practical significance.

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ANTON PORUBSKÝ

HYDROGEOLOGY OF THE CZECHOSLOVAK SECTION OF THE DANUBE

The Czechoslovak section of the Danube forms together with the territory, its river—bed is situated in, one natural hydrogeological unit. Its borders in the Czechoslovak territory we indentify with the border of the partial river-basin of the Great Danube with regard to the construction of a system of hydroelectric stations. This border cannot be ascertained in detail in relation to the adjoining river-basins because on the basis of present state of knowledge of the ground water regimen of the whole Danube Lowland the boundary of the divide with adjoining river-basins, mainly of the Čierna voda, Dudváh the lower Váh and the area between two rivers (Small Danube—Váh—Nitra—Žitava) cannot be ascertained accurately. In the delimitation of the territory of the partial Danube water-basin we therefore start for the present from the water-basin nomenclature according to the state regional subdivision of water economic plan, after which the territory of the partial Danube water-basin includes: (a) *islands above Karlová Ves*, (b) *four communities on the right side of the Danube*, (c) *the territory around Bratislava*, (d) *the Žitný ostrov Island*, (e) *a narrow stripe of territories bordering the main stream of the Danube between the Váh—Hron and the Ipeľ*.

The mentioned delimitation from the standpoint of hydrogeology and ground water regime is roughly for orientation no accurate ascertainment of the boundary is available, mainly in the northeast and east for the Small Danube as far as the hydrological activity of the Danube reaches.

If we determined the hydrogeological unit in its full extent we should have to include in it the whole Danube Lowland with adjacent plateaus and uplands in Czechoslovak as well as Hungarian territory. For the clarification and knowledge of the problem under study I propose the division of hydrogeology of the Czechoslovak section of the Danube into two hydrogeological units as follows:

[1] hydrogeological unit of the Danube Lowland in a broader sense bound to the right and left side of the Danube;

[2] hydrogeological unit in a narrower sense bound to the partial Danube river-basin as it was delimited above.

The first as well as the second unit in the broadest sense are divided into regions and areas conditioned by the geological, geomorphological and hydrogeological structure, with regard to the occurrence of ground water and its regimen regularities.

In this work I am going to deal with only the hydrogeological unit of the Czechoslovak section of the Danube in a narrower sense, thus in an extent that coincides with its partial river-basin in our territory.

The river Danube, as it appears and as we know it at present, is only a consequence of a genesis throughout several thousands of years. During the time of its formation it was passing through a turbulent way of development with several stages, each of which has left distinct traces and indications of activity in all the territories, it flowed through in the past and flows through at present. Its way of development is much more complicated as we imagined till now. Because of small number of facts we frequently simplified the real relations and present state and we were not always based on a clear idea of the origin of the whole Danube Lowland and the development of all its stream system — not only of the Danube. This is confirmed by difficulties in the determination and seeking regularities of the individual sedimentation cycles in the whole lowland and the varied alternation of gravelous and sandy facies one upon another, one under another and also side by side — so called polycyclic sedimentation. We take as a basis of all the subdivision and study of individual facies only sediments of the Danube and so far we did not pay attention (in basic investigation) to accumulations of the Carpathian rivers, the Váh and of Alpine rivers: the Leitha, Raab and others.

The Danube did not always flow to our territory through the Devín defile. In ancient past it flowed round it — it did not reach this defile with its river-bed but flowed through the Leitha (Bruck) gate, flowed under the southern slopes of the Hundsheimer Hills. Its stream flowed through Petržalka under the southern slopes of the Malé Karpaty Mts. and changed its course to southern direction as far as to the environs of Bernolákovo and its high water surely inundated far to the NE from Bernolákovo to Pezinok. Similarly the Leitha and other rivers also flowed to the Danube Lowland quite independently and only the formation of their own aggradation ramparts made them form the present stream system.

The river Váh flowed in our territory also nearer to the southern slopes of the Malé Karpaty Mts. and flowed into lakes, swamps and arms of the Danube somewhere in the wider environs of Sládkovičovo.

With all the consideration of sedimentation cycles we cannot neglect even the fact that the Danube did not change its bed at once and did not carry over it from the Leitha (Bruck) gate to the Devín gate. Several sedimentation cycles must have passed till the aggradation ramparts of the Danube dammed up the way of its previous Leitha river bed and all the outflow with tributaries came to the present river bed. The importance of the knowledge of mentioned regularities we understand with detailed evaluation of the hydrogeology of the Žitný ostrov Island. If we speak now about the variegation and frequently about unlawful sedimentation in the centre of the Danube-Lowland we may affirm that we do not understand it till now until we synthetically join to the formation of individual facies developments not only the tectonics but also the syntehtics of accumulations and of the influence of the streams Leitha, Raab, Váh and other Alpine-

Carpathian rivers that played an important role in the cyclic filling of the Inner Carpathian Neogene basins. I suppose that only analytical information of individual factors having a share in the formation of the Inner Carpathian basins can make us understand the synthesis of their lawful development and the present variegation of the hydrogeological structure.

Hydrogeological subdivision of the Danube region and adjacent territory

Along the total length of its stream of 2857 km the Danube has not formed such complicated and interesting hydrogeological relations as just in our territory. The total area of its river-basin is 817.000 km², in our territory 2.791 km². The Danube connects in its course several Late Tertiary basins, the most known of which for us are the Inner Alpine Vienna Basin and the both Pannonian basins — the Komárno Basin (our Danube Lowland and the Small Hungarian Lowland) and the Alföld (the Great Hungarian Lowland). The Danube comes to our territory below the Devín ruin, where the Morava flows into it. It flows through the Devín defile to Bratislava and to the Danube Lowland and still keeps it high mountainous character. Its maximum volume of water is mostly in June (less in July) and reaches an amount of up to 14.000 m^3/sec. High water with a volume of water of 8—10.000 m^3/sec. uses to be quite regularly twice a year, at the end of winter, so called "ice water" and in June with intense melting of snow and glaciers in the Alps, so called "green water".

The average volume of water in Bratislava is 2.020 m^3/sec., the minimum one 570 m^3/sec. The river gradient in the river bed near Bratislava is 0,4 ‰, near Palkovičovo 0,15 ‰, and gradually diminishes to 0,06 ‰ towards Štúrovo.

The spring area of the Danube in the Schwarzwald- the spring streams Borg and Brigach- are in an altitude of 304 m above sea level, the harbour in Passau 171 m, in Vienna 130 m, in Bratislava 104 m, in Štúrovo 97 m, in Budapest 73 m and in Beograd 67 m. The basic source of completing the volume of water in the Danube river bed is the water from the Alps, its right side tributaries.

On the basis of present state of investigation and knowledge of geological, geomorphological and hydrogeological structure we can divide the Czechoslovak section of the Danube into five basic hydrogeological areas as follows: (1) *the Devín defile area*; (2) *the Subcarpathian-Petržalka area*; 3) *the area of the Gabčíkovo depression*; (4) *the Komárno area and the area between two rivers*; (5) *the Štúrovo—Chľaba area.*

The individual areas we divide after various characterizations still into partial areas, in which we include various geomorphological or geological units with equal ground water regimen.

The previous subdivision of the Czechoslovak section of the Danube into three areas: the Devín defile, the Veľký Žitný ostrov Island and the territory of the Danube from the Váh to Chľaba (Porubský 1961) does not fit to the present state of investigation any more. I stress that the hydrogeological subdivision of the region of the Czechoslovak section of the Danube does not concern the whole Danube Lowland, in which case we should have to distinguish also areas of individual river-basins, plateaus and uplands.

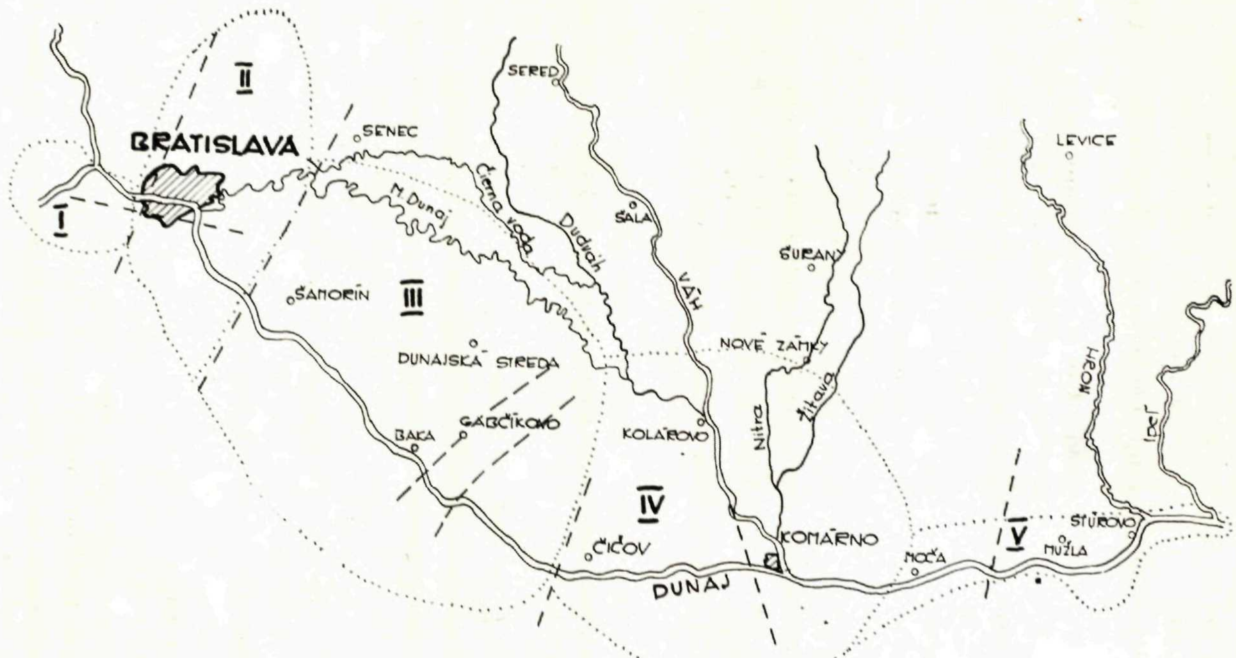

Fig. 1. Hydrogeological rayonization of the Czechoslovak part of Danube river

Explanation: - - - - faults, ... rayon borders. Rayons: I — Devín; II — sub-Carpathian-Petržalka; III — Gabčíkovo; IV — Komárno; V — Štúrovo—Chľaba.

The Devín Defile Area

The Danube river bed in the Devín defile section is 11 km long, 290—300 m wide, the average depth is 2.80 m and the speed of stream 1,5—1,6 m/sec. The Devín defile (gate) is geologically formed by a graben of the core mountains of the Malé Karpaty, strongly disrupted tectonically by faults striking northwest — southeast, the dominant faults of which are the Devín Fault bordering, the Slovak part of the Malé Karpaty Mts., and the Hainburg Fault bordering as a fault line the Hundsheimer Hills on Austrian side (Myslil 1958).

From the hydrogeological aspect aquiferous Quaternary sediments of the Danube flood-plain are interesting. They predominantly are accumulations of the Danube and only to a smaller degree of the Morava. The whole Devín gate is besides morphological features mainly characterized by the presence of pre-Quaternary formations-granites, Sarmatian and Panonian - in small depth under the Quaternary gravel terrace. So far much attention has been paid to hydrogeological investigation of the Quaternary and of earlier formations, mainly from the aspect of their permeability and ground water regimen, in the first place with regard to the construction of the planed dam on the Sihoť (Käsmacher) Island. The Quaternary gravels and sands of the Danube alluvium are well permeable. Water forms in them a coherent subsurface stream, the magnitude of which would increase with the construction of the dam and endanger by soak adjacent territories of Petržalka and Bratislava. The territories will be protected, mainly on the Austrian side by drainage system. Quaternary sediments occur there in a layer 12—15 m thick. The ground water is in the depth of water-level in the Danube river bed. Important from the standpoint of hydrogeology is the ground water in the vicinity of Petržalka that can be mainly employed in industry.

Most important from the standpoint of hydrogeology is the Sihoť Island, the territory of the waterworks for the aqueduct of Bratislava. It is formed by a flowing round arm of the Danube between Devín and Karlová Ves in 1873—1876 km of the river. It is about 3 km long and 1 km wide. About 1000 l of water per sec. is taken from it into the aqueduct of Bratislava. At middle and lower water-level of the Danube the water is good for drinking also without preparation. At high water-level of the Danube it must be prepared by chlorinating. In the foregoing years the water specialists from Bratislava supposed the ground water of the Sihoť Island is derived from the Alps and comes to Sihoť by joints and tectonic lines. A proved matter of fact however is that it is water of the Danube, percolating from the river bed into alluvial deposits.

The Sihoť Island is for the low relative height in relation to the Danube river bed and state of discharge of water in the Danube inundated with high water, twice a year regularly. Once at the end of winter with ice shove and once at the beginning of summer, when snow melt sets in in the Alps, connected with efficient rainfall.

The permeability of gravelous sands of the Danube on the Sihoť Island is considerable, the highest of gravel and sand in our country. The value of the filtration coefficient "k" moves in them in the limit $6 . 10^{-2}$ m/sec. to $6 . 10^{-3}$ m/sec. The Sarmatian sands underlying the aquiferous Quaternary show the "k" value of $2{,}72 . 10^{-4} - 2.14 . 10^{-5}$ m/sec.

The stream directions and water table of the ground water are under permanent correlation with water level of the Danube and its arms.

From the aspect of quality the ground water of the Sihoť Island fits to the criteria of the Czechoslovak norm for drinking. Preparation by chlorinating is only necessary in the time of inundation of the island with flood. According to its chemism it belongs to the bicarbonate type of water.

The Subcarpathian — Petržalka Area

This area was previously ranged in the Bratislava-Petržalka area but the extension of the Subcarpathian-Petržalka area is wider. Geographically and geomorphologically it includes the territory of the southern slopes of the Malé Karpaty Mts. and of the lowland to as far as Jur near Bratislava, Bernolákovo, the Small Danube and all the right side of the Czechoslovak section of the Danube with Petržalka and adjacent communities.

Regarding to the occurrence, accumulation and ground water regimen we can divide the territory into two hydrogeological areas: the Bratislava and the Petržalka area and these again regarding to locally different hydrogeological characteristics into several subareas.

The whole Subcarpathian-Bratislava area belongs to the territory below the Devín defile on either side of the Danube. The territory is covered by a coherent aquiferous layer of Quatenary alluvia of gravel and sand, which is overlain by loam, sandy or humic loam, sometimes by loess or swampy sediments and sediments of old arms. The aquiferous Quaternary sediments rest upon Neogene sediments, little permeable to impermeable, only at the base of the Malé Karpaty Mts. on older rocks.

We can divide the Subcarpathian hydrogeological area regarding to the permeability and composition of the aquiferous material into three subareas:

(1) *subarea of weakly aquiferous loamy gravels and sands* of small thickness at the base of the south-eastern slopes of the Malé Karpaty Mts., extending from Bratislava in the direction to Rača—Šúr—Vajnory and Čierna Voda. The territory is covered by surficial, mostly sandy and muddy loam, underlain by loamy gravel and sand, cropping out at places. They reach to 2.60—6.20 m under the surface. The impermeable underground of the Quaternary sediments

is in the depth from 2 to 6 m. The value of the filtration coeficient "k" is in the limit $1.6-3.7 \cdot 10^{-4}$ m/sec. The strongly loamified gravels and sands pass to the south-east and south into purer ones and their thickness increases.

The water table of the ground water is in the depth of 0,0 m to 2,80 m under the surface with a gradient in southeastern direction. In the time of snow melt and precipitation the water was - and at places also is — in depressions and formed swamps, already drained now.

II. *subarea forms the territory north of the Small Danube* (north of the Biskupice Fault) in the direction to Vajnory as far as Bernolákovo. The territory is formed by surficial loam, at places by sandy loam and clayey sand with various facies transitions to the depth of 0,5—3,0 m. Under them is a coherent layer of gravel and sand, 8—9 m thick. Its thickness decreases towards the base of the Malé Karpaty Mts., increases to the south and south-east.

The permeability of aquiferous gravel and sand confirmed by pumping tests has the value of filtration coefficient of 0,0005—0,005 m/sec. The water table of the ground water is in the depth of 1,0—4,0 m, mostly in gravel.

In this territory the permeability of Neogene sand was also ascertained in Upper Panonian beds. Their value of the filtration coefficient moves from several dm to several m in a day.

The first subarea is the territory of weakly permeable Quaternary deposits west of Chorvátsky Grob. The Quaternary is formed of loam with rolled gravel and sandy loam to the depth of 0,5—1 m. The Quaternary sequence is without any aquiferous horizon and practically contains only soil humidity maintained by precipitated water.

The underlying rock of the Quaternary sediments is formed of Neogene (Upper Panonian?) beds — calcareous clays and sands as well as transitions between them in the depth of 0,5—2,5 m. The water table of the ground water is 0,8—2,7 m under the surface already in the Neogene. The Neogene aquiferous beds of this territory are in relation to other Neogene beds also in wider surroundings of Bratislava the most permeable. Their value of the filtration coefficient reaches to 10 m/day.

Precipitated water fallen down in this area soaks in rocks of the Subcarpathian territory and thus supplies the ground water of Neogene beds. The ground water flows off through the Neogene beds, infiltrates farther southerly and south-easterly and shares in the supply of water resource of Quaternary sediments lying southerly as well as of Neogene sands underlying the Quaternary of the described territory.

The water of Neogene beds is artesian water with positive negative level; according to the depth of aquiferous horizon and in the hydrogeological regional subdivision of Slovakia it is ranged in the artesian area of Bratislava (Porubský 1964).

In the territory around Bratislava two main sources share in the supply of ground water resource. They are water from the Danube river bed and precipitated water from the Malé Karpaty slopes. The Danube water starts to soak in gravelous-sandy deposits already in Bratislava and moves at the altitude 131—133.5 m. The ground water from the base of the Malé Karpaty Mts. apparently has a higher water table at the altitude of 131—136,0 m, thus proving that the Danube water in Bratislava and close surroundings affects the ground water in eastern and south-eastern direction, whereby we can give the hydroisohyps with the point 131,00 m as orientation boundary of meeting of the Danube water with the Carpathian water.

In the Petržalka hydrogeological unit the hydrogeological relations are different from those of the Subcarpathian unit. The right side of the Danube-Petržalka is a continuation of the valley and plain part of the Danube territory of the Devín defile. The territory is furrowed by old river beds and arms, covered by loam, loess-loam, sandy and muddy loam and the filling of old river arms. They are underlain by a layer of gravel and sand, which forms the so called Petržalka terrace (Myslil 1958), 12—18 m thick. The underlying rock in the surroundings of Petržalka is formed of Upper Neogene clay (Panonian ?), southerly of Upper Neogene sand (Pontian-Levantinian ?). In the direction along the Danube stream the thickness of the aquiferous Quaternary increases.

The direction of the ground water stream is under the influence on the water level of the Danube. The permeability of aquiferous gravel is 10^{-3} m/sec., of sand 10^{-4} m/sec.

The Danube has after flowing out from the Devín defile to the region of the Danube lowland - after passing over the Carpathian fault that borders in NE-SW direction the proper Neogene basin from the Malé Karpaty crystalline — its river bed high above the water table of ground water in Quaternary sediments of its accumulation "tub"; accumulations of its "dejection cone" except older higher Bratislava terraces. This peculiarity of the position of the Danube river bed enables the supply of ground water resources by the river on either side by soaking in under every water level also the lowest. In this way the Danube flows on the "edge" of its roof to nearly as far as the profile of Baka and Gabčíkovo.

The Gabčíkovo Depression Area

In the foregoing works of many hydrogeologists, hydrologists, geomorphologists we ranged the Žitný ostrov-Island as a unit in one hydrogeological unit-area, mainly on the basis of morphological features. It was simply because the territory was and is bordered by the Small Danube and the Danube from Vlčie hrdlo to Komárno, forming an area of 1,620 km².

At present however, if we have to make a more detail hydrogeological subdivision of the Danube territory, we must start from the geological structure, water-bearing and ground water regimen. All these factors mentioned we consider as basic and particularly under the conditions of the Small Danube Lowlands as a Neogene basin with thick filling of aquiferous gravel and sand and abundant static and dynamic ground water reserves must also become a criterion for the delimitation of an independent area.

The ground water of all the Žitný ostrov Island is under the influence of the Danube directly or indirectly by means of the Small Danube. The results of various kinds of investigation show that the main river bed (and in the past they were several) migrated throughout the Žitný ostrov Island a major part of the Small Danube Lowland- in Czechoslovak as well as Hungarian territory- as we have supposed till now. The importance of the river beds and arms of the Danube changed under the influence of the volume of water, formation of aggradation ramparts, accumulation of alluvium, tectonic-seismic processes, climatic changes and many other known and unknown natural factors. Thus once also the river bed of the present Small Danube was the main Danube bed and the Danube also flowed more to the north and north-east than its present boundary. On the other hand it is an evidence that also the stream system of its tributaries changed the place of its mouth, whether into the river bed or the arms of the Danube, into lakes and swamps, which covered all the territory of the Small Danube Lowland, also the Hungarian part. The rivers Leitha and Raab carried to the Small Danube Lowland material of sediments identical with that of the Danube and accumulated it also in the Neogene basin as the Danube.

The geological structure of the Žitný ostrov Island, in the central part of the Danube Lowland, is not uniform. Whereas the Quaternary sediments in the territory of Bratislava and Petržalka are 10—12 m thick, in marginal parts the thickness increases to 15—18 m. Towards the centre of the Žitný ostrov Island the thickness increases under the influence of tectonics and formation of the Neogene depression. The underground of the Quaternary in the Subcarpathian area is formed of Upper Panonian sediments, in the area of the Neogene depression predominantly by Levantinian and Pontian sediments. The Levantinian beds represent a facies of loose sediments and form together with Quaternary loose sediments one thick aquiferous complex with free groundwater surface. This aquiferous Quaternary-Levantinian complex (Janaček 1966) reaches the greatest thickness in the area of Gabčíkovo and Baka, to 400 m on the whole. In its geological structure share gravel and sand in various layers, strata and alternation, frequently in unlawful succession and layers different in thickness and granulometry. Just this so called polycyclic structure is not only a consequence of the activity of the Danube but also of its tributaries and of tectonic processes, which conditioned gradual sinking of the Komárno Basin and its filling up by gravel and sand in the succession as we see them now.

The main Gabčíkovo depression is tectonically bordered by fault lines, by the Sládkovičovo Line in the north and the Kližská Nemá Line in the south-east. According to recent investigation two tectonic lines were proved in the surroundings of Gabčíkovo: the western and the eastern line (Janáček 1956). The Upper Neogene formed of clay steeply plunges southerly and to the south-west near Podunajské Biskupice and appears again in 10—12 m under the surface in the elevation (horst) in Kližská Nemá. This territory we range for the specific geological structure, large amount of gathered water in the form of static reserves and specific ground water regimen in an independent hydrogeological area, where the territory on the Hungarian side also must belong because it forms with our territory one hydrogeological unit.

The territory of the Žitný ostrov Island to the south-east, east and south of Kližská Nemá we range on the basis of hydrogeology in the area of Komárno and between two rivers.

The eastern and south-eastern limitation of the Gabčíkovo depression we still do not know exactly but on the basis of up to present investigation we suppose its central part in the direction Gabčíkovo, Baka, Dunajská Streda with continuous decrease in thickness of the aquiferous Quaternary-Levantinian complex towards the Small Danube. To the east and north-east it joins to the Kolárovo Beds represented by Levantinian sand also in the underground of the Quaternary of the Gabčíkovo depression.

The influence of tectonic processes (as recently has proved Janáček 1966), mainly of young tectonics is also visible in the Quaternary-Levantinian complex of gravel and sand, in which the Kolárovo Fault south of Gabčíkovo was ascertained. The Gabčíkovo-Fault, which many specialists tried to prove regarding to the construction of Gabčíkovo dam, could not be proved as yet.

In the hydrogeological area of the Gabčíkovo depression the ground water regimen depends on the following factors: 1. the Danube and Small Danube river bed; 2. precipitation -snow; 3. other influence.

D. Duba in his works distinguished in the Žitný ostrov Island in accordance with real natural conditions three areas with different ground water regimen as follows: (a) area of the riverian zone, where the permanent influence of the rivers manifests; (b) mixed area; (c) area, where the ground water regimen forms under the influence of precipitation (in central part of the Žitný ostrov Island).

Recent investigation of the territory and the ground water regimen, mainly in the Gabčíkovo depression, however refines the ground water dynamics and regimen, which is essentially more complicated than we have assumed till now. Changes in pressure induced by large amount of water, the thickness of water column and uneven surficial and vertical distribution of variously aquiferous sediments (granulometric difference), form 2 dynamic regimens — the surficial and the deep-seated. The surficial regimen shows all features and regularities of common ground water regimen, Quaternary alluvial deposits, as previously treated by D. Duba. This regularity is however bound to zones and confined to the thickness of aquiferous horizons. In the conditions of the Gabčíkovo depression it is to the maximum depth of 30 m. Under this depth the influence of the deep-seated regimen with all its dynamic features manifests.

The transmission of hydrodynamic pressure to the depth as reflection on the existence of the ground water regimen could be proved already in 3 cases. In investigation for the construction of dams and hydroelectric stations of the Danube piezometric pipes were put in deeper borings in various depth at 3 chosen localities. The results of observations proved that in the time of flood and with higher water-level of the Danube ground water in deeper part or water circulating in layers of gravel among sand beds reacts first. In the surroundings of Šamorín in the time of flood in 1965 water flowed from the piezometer in the depth of 260 m 20 hours sooner than from the piezometer in the depth of about 16 m.

The whole Žitný ostrov Island represents with its aquiferous alluvial deposits a ground water reservoir with an accumulation of about 8 milliards m^3 of water of relatively good quality.

As we have mentioned already, the Danube flows with its river bed and many arms in the section of Bratislava—Gabčíkovo high above the ground water table and thus its water always supplies in the time of flood and also of low water-level the ground water reserves of the territory on either side. In the surroundings of Gabčíkovo as far as Palkovičovo both the water -levels are almost equally high so that the ground water table is hydroglogically dependent on the discharge of water in the Danube river bed and the inflow of ground water from higher part of the Žitný ostrov-Island. This regularity is then valid in all the farther territory in the Danube section as far as Chľaba. The mentioned "ridge" position of the Danube river bed above the surrounding ground water table is of extraordinary importance for the Žitný ostrov Island. Under its influence the ground water reserves of the island increase or decrease. Although these oscillations of the water table are small compared to the total thickness of aquiferous sediments, they equally signify a change in ground water reserves. The amplitude of these changes reaches the amount of several tens of m^3/sec., what markedly manifested e. g. in the time of the flood in 1954, when the maximum average inflow of Danube water into aquiferous beds of the Žitný ostrov Island was according to communication of water economists to 201 m^3/sec.

The influence of the Danube, although it is decisive, does not manifest equally in the whole territory. It is mostly evident of course along the river bed, the riverain zone, and then in almost all the upper part of the island. The ground water has there first a gradient towards the Small Danube and then gradually turns to the east and south-east.

The ground water "flows" under the influence of gravitation and hydrodynamic regularities in permeable beds and stops on the tectonic line of the elevated Neogene block in before Kližská Nemá or in the partial depression behind it. There it includes the strongest kinetic energy, its effects catastrophically manifested in all floods as yet.

A large amount of ground water flowing from the upper part of the island to the centre and farther to the east comes back to the Danube through the Small Danube and drainage channels by means of pumping stations.

The influence of precipitation on the ground water also cannot be neglected; it mainly manifests in the inner part of the island. In the upper part of the island the influence of inflow of ground water from the Malé Karpaty Mts., is proved. On the other hand high water from the Váh on the eastern side can from time to time interrupt the runoff of the ground water from the island and also swell up the water of the Danube. The meeting of two large flood waters on the Danube and on the Váh can have a catastrophic consequence, as we also saw in the flood in 1965.

The ground water table is 8 m in the upper part of the Gabčíkovo depression, about 4 m in the central part as far as Dunajská Streda, 0—2—4 m under the surface in the lower part and everywhere along the Danube.

The gradient of the ground water table is in the upper part of the island several times greater than in the lower part. If we take into consideration that the permeability of aquiferous material in the axis of the island gradually decreases to the east, we see clearly that the ground water inflow from the upper part is greater than the lower part of the territory is able to accept. This causes together with geological obstacles (Neogene block of Kližská Nemá) the swelling up of the ground water table and the break of gradient. Such an area is the territory near Čalovo and Okoč, where before the construction of drainage channels were large swamps.

The aquiferous material is regarding to the ground water regimen most markedly characterized by the filtration coefficient "*k*". In conditions of the Gabčíkovo depression it was determined many times with regard to the planed construction of the dams.

If we take into cosideration the complex of the aquiferous sequence we can regarding to the filtration coefficient delimit the following areas:

(a) area of Podunajské Biskupice—Štvrtok na Ostrove—Tonkovce with the filtration coefficient $2{,}32-3{,}47 \cdot 10^{-3}$ m/sec.

(b) area of the proper Gabčíkovo depression from the Čilistov trough as far as Topolovec, filtration coefficient $1{,}15 - 3{,}47 \cdot 10^{-3}$ m/sec.

(c) territory in the area Ostrova—Lehnice—Dunajská Streda—Veľké Dvorníky—Bohelov—Čalovo—Okoč, filtration coefficient 5,8 — 1,15 m/sec., $5{,}8 \,.\, 10^{-4} - 1{,}15 \,.\, 10^{-3}$ m/sec.

(d) marginal territory of the Small Danube with the filtration coefficient 50—80 m/day $5{,}8 \,.\, 10^{-4} - 9{,}25 \,.\, 10^{-4}$ m/sec.

In contrast to the mentioned values locally also more different "*k*" values can be expected.

In the evaluation of aquiferous sediments in the territory of the Gabčíkovo depression regarding to the construction of the planed dams we have determined the value of "*k*" separately for gravel, separately for gravel with sand and for sand and compiled a graph of "*k*" values on the basis of sand admixture with grain fraction to 2 mm according to granularity curves.

The value of the filtration coefficient for gravel and sand of the Danube we have ascertained on the basis of long standing investigation as follows:

gravel value $k = 3 - 5 \,.\, 10^{-3}$ m/sec.; gravel with sand $k = 6 - 8 \,.\, 10^{-4}$ m/sec.; sand $k = 5 \,.\, 10^{-5} - 6 \,.\, 10^{-4}$ m/sec.

Regarding to the complex of the sequence as to the vertical subdivision we have determined for deposits of Danube gravelous sands to the depth of 30 m the average value $k = 3 \,.\, 10^{-3}$ m/sec., under the depth of 30 m $k = 2 - 8 \,.\, 10^{-4}$ m/sec. In the ascertainment of anisotropy by way of experiment we have determined its value 2—4.

Komárno Area and the Area between Two Rivers

The delimitation of this hydrogeological area of the Danube territory in the Czechoslovak section is most difficult. Its geographical borders extend approximately between the Danube, Kližská Nemá, Čalovec, the Small Danube, Dedina Mládeže, Nesvády; the southernmost territory of the Nitra and Žitava river-basin the plain territory of the left side of the Danube as far as Kravany. The centre of the territory is Kameničná—Komárno with the confluence of the Váh and the Danube.

The territory is geologically and tectonically very complicated. Along the Danube between Kližská Nemá, Kravany as far as Patince it is frequently tectonically disrupted with larger and smaller Neogene elevations and depressions filled up by Quaternary or Levantinian sediments. The northernmore territory with the centre near Kolárovo and the confluence of the Váh and Small Danube forms the extensive Kolárovo depression filled up by Levantinian and Pontian sediments, the so called Kollárovo Beds (Buday 1962). These are represented in the upper part by pure and clay sand, at places also with fine gravel with abundant calcareous-sandstone concretions. The Quaternary aquiferous gravels and sands are deposited directly on the Kolárovo Beds and form with them together one hydrogeological horizon with free groundwater surface.

The underground of the whole complex of the Quaternary and of the Kolárovo-Beds is formed of Upper Panonian sequence. This crops out to 8—10 m in the surroundings of Komárno and extends approximately as far as the Kravany Fault in the east of the territory.

The Danube and its tributaries deposited in the Gabčíkovo depression still coarse gravel, frequently with larger rolled pebbles. In the territory of Komárno and between two rivers, thus as far as Kližská Nemá to the north-east and east the Danube deposited and also deposits with its tributaries finer gravel and sand. At the boundary of the hydrogeological areas of the Gabčíkovo-depression and of Komárno-area between two rivers to the east of Gabčíkovo the fact of change of the deposited aquiferous sediments is evident as to the granulometric size and thickness of beds. The character and cause of the occurrence of the Kolárovo-Beds (Levantinian sands) as far as the Gabčíkovo-depression area to the west have not been studied in detail and solved unequivocally as yet. The essential influence was surely that of tectonics, which resulted in gradual and long lasting subsidence of the central part of the Komárno Neogene basin and to not a small degree also in the paleogeographical conditions at the boundary between the Pliocene and Pleistocene. In this area of the Danube share in the geological structure mainly Later Tertiary and Quaternary rocks. The Neogene beds are exclusively composed of sedimentary rocks. The underlying rock of Neogene beds is a complex of Mesozoic carbonate rocks, which crop out only on the right side of the Danube in Hungary. In our territory we have found Triassic limestones with the investigation of the water construction Nagymaros in the depth of 130 m near Patince in the area of points of issue of thermal water. They build up the underground of the Upper Panonian clay-sand sequence.

The youngest beds are Quaternary formations as alluvial deposits, gravels, sands, flood loams, loess-loams, loess and dunes of drift sand, with many muddy and swampy depressions. The alluvial deposits belong partly to the Later Pleistocene and partly to Recent alluvial deposits of the Danube, Váh, Nitra and Žitava.

Regarding to the ground water regimen and local geological structure we can distinguish essentially 2 hydrogeological areas in the described area.

(a) *Area of the riverain zone of the Danube from Kližská Nemá as far as the Kravany Basin.* The territory is of plain character sloping to Komárno, Patince, Kravany, in eastern direction. Geologically it is formed of the Quaternary and Upper Panonian or Pontian. Several smaller tectonic depressions alternate there with low elevations. In the underground of the Quaternary 8—20 m thick are at places clays, situated near Komárno, Iža and Patince in the depth of 8—12 m. The territory in the east is bordered by the Kravany Fault, at which Upper Neogene sedimentary rocks end and towards Štúrovo start to appear Earlier Neogene, Paleogene or volcanic sedimentary rocks in the underground of the Quaternary.

The filtration coefficient of the aquiferous Quaternary moves in the limit of 10^{-4} m/sec. group and only seldom reaches also values of 10^{-3} m/sec. group.

The Neogene sands underlying the Quaternary or in Neogene beds show the value $k = 1.10^{-5} - 2.10^{-4}$ m/sec.

The ground water regimen is in direct hydraulic relation to the water in the river bed of the Danube and Váh with the Small Danube. The stream direction of the ground water changes under the influence of both rivers in NE to SE direction. The ground water table is 0,5—3 m under the surface.

Hydrogeologically important is also the Červená flotila Island in Komárno, the waterworks island for Komárno. Its hydrogeological structure is different from the structure of the territory on the left side of the Danube. The thickness of the aquiferous Quaternary is greater, 15—18 m. It is underlain by aquiferous sand of the uppermost Neogene. The ground water reserves are directly supplied by water percolated from the Danube. In their quality they fit to utilization in waterworks-similarly as on the Sihoť Island above Karlová Ves.

(b) *Area of the Kolárovo Beds forming a Pontian-Levantinian depression with the centre between Kolárovo and Čalovo.* Its finger- to fanlike projections extend to all sides of the right and left side of the Váh—Danube, Nitra, the left side of the Danube and the right as well as left side of the Small Danube. Their presence was confirmed by investigation also in the surroundings of Topolovec, Gabčíkovo even farther to the north-west.

In the area of the Kolárovo-Beds the aquiferous Quaternary sediments rest directly upon aquiferous Upper Neogene sands and form together with them one hydrological ground water horizon with free water surface. The stratigraphical boundary between the sequences of strata is very difficult to determine and near Kolárovo it is given in the depth of about 15—18 m. The aquiferous Kolárovo Beds are 100—120 m thick. They are formed of sands, fine gravels, sands with clay admixture and calcareous-sandstone concretions. At places also layers of light-grey clay and sandy clay occur in lenticular position. They form a large ground water reservoir, which is directly connected with the ground water reservoir of the Gabčíkovo depression and with the aquiferous Quaternary of the Váh and Small Danube.

The permeability of the aquiferous Quaternary overlying the Kolárovo Beds is represented by the value $k = 5.10^{-4} - 2.10^{-3}$ m/sec., the aquiferous Kolárovo Beds show the value $k = 10^{-5} - 10^{-4}$ m/sec.

The ground water regimen of this area is very varied and seldom stable. It is under permanent influence of ground water inflow from the adjacent areas and under regular and irregular influence of discharge of water in the river beds of individual rivers and of drainage channels. Once it is influenced by high water of the Danube and of the Small Danube with surplus of ground water from the Gabčíkovo depression, drainage system with pumping stations, another time by high or low water of the Váh, Váh—Danube and Nitra that can supply or drain the ground water.

The Štúrovo— Chľaba Area

This area represents the territory on the leftside of the Danube from Kravany as far as the mouth of the Ipeľ into the Danube. From the aspect of geology it is to stress that this area is separated from the foregoing by the important Kravany Fault, to the east of which under the Quaternary Panonian sedimentary rocks do not appear any more and mostly Paleogene and volcanic Neogene rocks are present. On the basis of hydrogeology it can be divided into 3 areas and that: the Kravany-Obid area, the Štúrovo area (confluence of the Hron and the Danube) and the Chľab area (confluence of the Ipeľ and the Danube).

(1) *The Kravany-Obid area* represents the flood-plain territory on the left side of the Danube. It is bordered in the east, north and north-west by a markedly appearing older Danube terrace extending on the line Moča—Buč—Mužla—Obid and on the Boží kopec Hill near Štúrovo it touches the Danube. Northerly above the Danube terrace is the upland of the Belanské kopce Hills, under the slopes of which the Danube flowed in the past.

All the territory in the elevation of 105—110 m above sea-level is a part of the not large Kravany—Esztergom Basin, in low morphological position, moderately undulated, and at places with swamps.

The surface of the older (probably Riss) Danube terrace lies 15—20 m above the flood-plain and is also of plain character, moderately undulated by the action of wind and smaller brooks.

The described area is characterized by 4 types of ground water: ground water of the older Danube terrace; ground water of the Holocene flood-plain; Neogene ground water; thermal water.

The regimen of artesian common water from the underground of the Quaternary is little known. The artesian water is not of greater importance for waterworks. Boring for its exploitation to greater depth is not recommended because in the underlying rock of the relatively shallow Tertiary is the Mesozoic consisting of carbonate rocks with a horizon of thermal mineral water. Its presence also manifests by the issue not far of the state farm of Obid. There are the same thermal mineral waters as in the surroundings of Štúrovo and Patince and also in Hungarian territory.

The deposits of the older Danube terrace under the southern slopes of the Belanské kopce Hills are 10—15 m thick. On the base the terrace consists of a layer of well permeable aquiferous gravel and sand with the water column about 4 m high. The underground of the terrace is built up of Tertiary impermeable rocks, in which a smaller depression is modelled - the old Danube river bed. Its filling is formed of well permeable gravel and sand, their permeability is expressed by the filtration coefficient $k = 3 \, . \, 10^{-3}$ m/sec. The ground water of the terrace is supplied only by precipitated water. The territory of the terrace fits well for artificial infiltration. On the edge of the terrace there is a spring line of debris terrace springs of common water with the yield of 0,2—5 l for one spring.

The Holocene flood-plain is formed by a plain along the Danube, in low position, with many swamps. The thickness is 5—12 m, the aquiferous bed is formed of gravelous sand and sand, 4—10 m thick. The filtration coefficient of aquiferous deposits is $k = 1 - 8{,}3 \, . \, 10^{-4}$ m/sec. The ground water is in the depth of 1,5 m and deeper. It is directly dependent on the water-level of the Danube. The stream directions of the ground water are in the time of low and medium water-level of the Danube to the SE, with high water to the NE.

(2) *The Štúrovo area- confluence of the Hron and Danube* is bordered by the Danube river bed in the south, by the Boží kopec Hill in the elevation 132 m above sea-level in the west, the older Danube terrace in the north and northeast, passes into the flood-plain of the Hron and the western slopes of the Kováčovské kopce Hills. The territory is generally characterized by a moderately undulated plain in low position in the elevation of 106—110 m above sea-level, with distinct depressions. The depressions are relics of old arms and meanders of the Danube and Hron before their inflow into the Chľaba defile. In this area 4 types of ground water are present: water of the Hron—Danube alluvium, water of the Danube terrace, water of the underlying Tertiary and thermal water derived from the underlying Mesozoic, which crops out on the Hungarian side.

The ground water in the flood-plain of the Hron river circulates in about 1—2 m under the surface and corresponds to the water in the Hron river bed. The permeability of the aquiferous bed is of the average value $k = 5 \, . \, 10^{-4}$ m/sec. The thickness of all the Quaternary is 6—8 m.

The Danube Holocene in the flood-plain is 9—10 m thick. The aquiferous gravelous-sandy sediment is predominately formed of quartz grains, at places with chert. The ground water circulates in 3 m under the surface and is in direct hydraulic relation with the water in the Danube river bed. The permeability of the aquiferous bed is characterized by $k = 3 \, . \, 10^{-4}$ m/sec.

The ground water related to older rocks underlying the Quarternary, the sedimentary and volcanic Tertiary, is of no essential practical hydrogeological importance.

The thermal waters of this area are analogous to those in Patince and Obid. The first warm water horizon (22 °C) in Štúrovo was met by boring in the depth of 102 m, the next in the depth of 116—117 m with water temperature of 37 °C and then in the depth of 120 m with water temperature 41 °C. They yield about 7 l/sec.

(3) *The Chľaba area — confluence of the Ipeľ and Danube r.* represents the

smallest geological, morphological and hydrogeological area of the whole Czechoslovak Danube section. The territory can be divimited by the confluence of the Ipeľ and the Danube, the Danube and Ipeľ river bed and the slopes of the Kováčovské kopce Hills. The area is manifold, there is in essentials the Danube—Ipeľ flood-plain, which is only a small projection of the Szob Basin in the elevation 107—112 m above sea-level.

The underground of the Quaternary is built up of the Upper Tortonian, predominately represented by volcanic- terrestrial and volcanic - freshwater sediments.

Hydrogeologically important are only the Quaternary sediments - alluvial deposits of the Danube and Ipeľ. The Quaternary is formed of flood loam, underlain by loamy and dusty sand. The base is formed of sandy gravel, mostly also representing the aquiferous bed. The Quaternary is 7—9 m thick, the water column 2—6 m high. The ground water circulates in the depth of 3,5—7,5 m. At places it manifests as confined water.

Aquiferous gravelous sands are predominantly formed of rolled pebbles of quartz, quartzites, granite, sometimes limestones and neovolcanics. Their permeability value changes within the limit $k = 3{,}7 \cdot 10^{-5} - 2{,}6 \cdot 10^{-4}$ m/sec.

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WATER BALANCE CALCULATIONS FOR THE KARST REGION IN THE SMALL CARPATHIANS

Abstract. Authors determine dependences of the regime of the runoff and of the regime of producing ground water storage by precipitation by simple water balance calculations in aquifers of carbonate rocks in the NW hill-side of the Small Carpathians.

The basic factor is the infiltration of the precipitation and its changes in the time which generally influence the regime of ground water that means also the regime of the karstic ground water. It is necessary to know this relation so that we could estimate the regime of the karstic water for the solution of theoretical problems but mainly for the solution of water development problems in connection with the utilizing of karstic water.

We are offering the explanation of relations between precipitation a part of which shares in the producing of undergroud runoff and in the origin of storage of karstic water for the region of carbonate strata of the Krížna and Choč tectonic units in the Small Carpathians.

The Small Carpathians are built of crystalline core and of Mesozoic cover. The crystalline core represents the central part of the mega-anticlinorium of the Small Carpathians. The Mesozoic cover lies on the crystalline core. This cover is found in three tectonic units: in the autochthonous cover unit and in the two shifted units — the Krížna and Choč units. The Krížna and Choč units are of basic importance from the point of view of the solved problem and have the following geologic building:

The Krížna unit creates the Mesozoic zone between Kuchyňa and Lošonec (20.6 km^2), which is limited on the SE by the younger members of the cover unit which are not interesting from the hydrogeological point of view and on the NW side by the melaphir serie of the Choč unit. The strata of the actual unit are laid monoclinalously with the general slope to the NW. They have been built by sediments since Lower Trias to Cenoman (Mahel' 1962).

The Choč unit occurs in the western part of the mountains. The investigated part of this unit (68.6 km^2) stretching in the SW-NE direction has as a whole a very simple building with the slope of strata to the NNW and to the NW. The lower part of this strata hydrogeologically unfavourable is built by a melaphir serie, the upper part of that strata is built by Middle and Upper Triassic limestones and dolomites (Buday—Cambel—Maheľ 1962). The Choč unit is separated from the attached Záhorská nížina (Záhorská plain) by the borderal Small Carpathians' break line. The Mesozoic strata fell deeply along the break line in that plain. The Mesozoic sediments of the mentioned Choč unit along the break line contact the Tertiary and Quaternary strata which fill up the Záhorská plain.

The detailed hydrogeological characterization in both units shows (Kullman 1964, 1965, 1965a) the Middle and Upper Triassic limestones and dolomites which are permeable through clefts and karstic hole as strata which produce the highest yields. The above mentioned units create two independent hydroglogical wholes (Fig. 1) with independent regimes of ground water. This paper analyses the regime of their karst water in relation to rainfall.

The runoff of karst water from the shown individual hydrological units is measured in springs. Their discharge from the Krížňan unit is determined

Fig. 1. The schematic map of investigated carbonate complexes in the Small Carpathians: (1) of the Choč unit, of the Krížna unit

in 18 remarkable springs (measured by the Hydrometeorological Institute in Bratislava) and from the Choč unit in 20 springs. The results of these systematic measurements of yields as a total of all springs separatlly for every water-bearing complex and for the period 1957—1963 are shown in Fig. 2 (Kullman 1965).

In Fig. 2 the regularity of occurrence of two main annual extremes of yield is evident. The minimum of yield is observed in November—December and the maximum in March—April. Between them we can see the regular decrease of yield from spring to the end of autumn and the increase of yield from autumn to spring. The increase of yield of springs during the winter period is caused by the seepage of rainfall. The sum of rainfall in this period is higher than the sum of total evaporation. The gradual decrease of yield in the summer half-year is caused by the discharge of ground water storage (Duba 1967). The yield in this period is increased only sporadically by seepage of rainfall, because the prevalent part of rainfall is consumed by total evaporation. The increase of the yield of the springs would answer the period of maximum rainfall, i. e. July and August. If these maximum rainfalls occur, they also contemporaneously mean an annual maximum of yield only in the case of extraordinary high sum of rainfall which must be essentially higher than the total evaporation. Such a sum of rainfall has been measured in July 1957, also in 1959 and mainly in the year 1966. Otherwise the above mentioned spring time maximum of yield prevails. These considerations are also proved comparison with the karst region of Strážovská hornatina (highlands), where the summer time rainfall maximums take part in the ground water storage recharge more expressively than in the above decsribed case, in consequence of smaller evaporation (Kullman 1965b).

The correctness of the conclusion about the prevailing of ground water recharge by seepage if precipitation in the winter half-year, can be seen in Fig. 3. It illustrates the relation between the half-year sum of rainfall from November till April (Z) and the sum of discharge of ground water which is measured in springs (O_v) for the same period and which is expressed in mm (Tab. 1).

Straight line relationships in Fig. 3 comply the best with years 1959, 1960, 1961, 1962 and 1965. The years 1957, 1958 and 1966 are placed to the right of these lines, it means, that the sum of rainfall of definite highness causes a higher runoff of ground water from springs in these years than would follow from the line of relationship. As seen in Fig. 2 and Tab. 2 it is a result of occurrence of high sums of rainfall in the summer half-years which preceded the above-mentioned years, i. e. in the period from May till October 1956, 1957 and 1965 (441 mm, 427 mm, 557 mm in Sološnica) and which had conditioned in comparison with other years, the highest minimum of ground water storage in the beginning of November. Afterwards also the relatively small

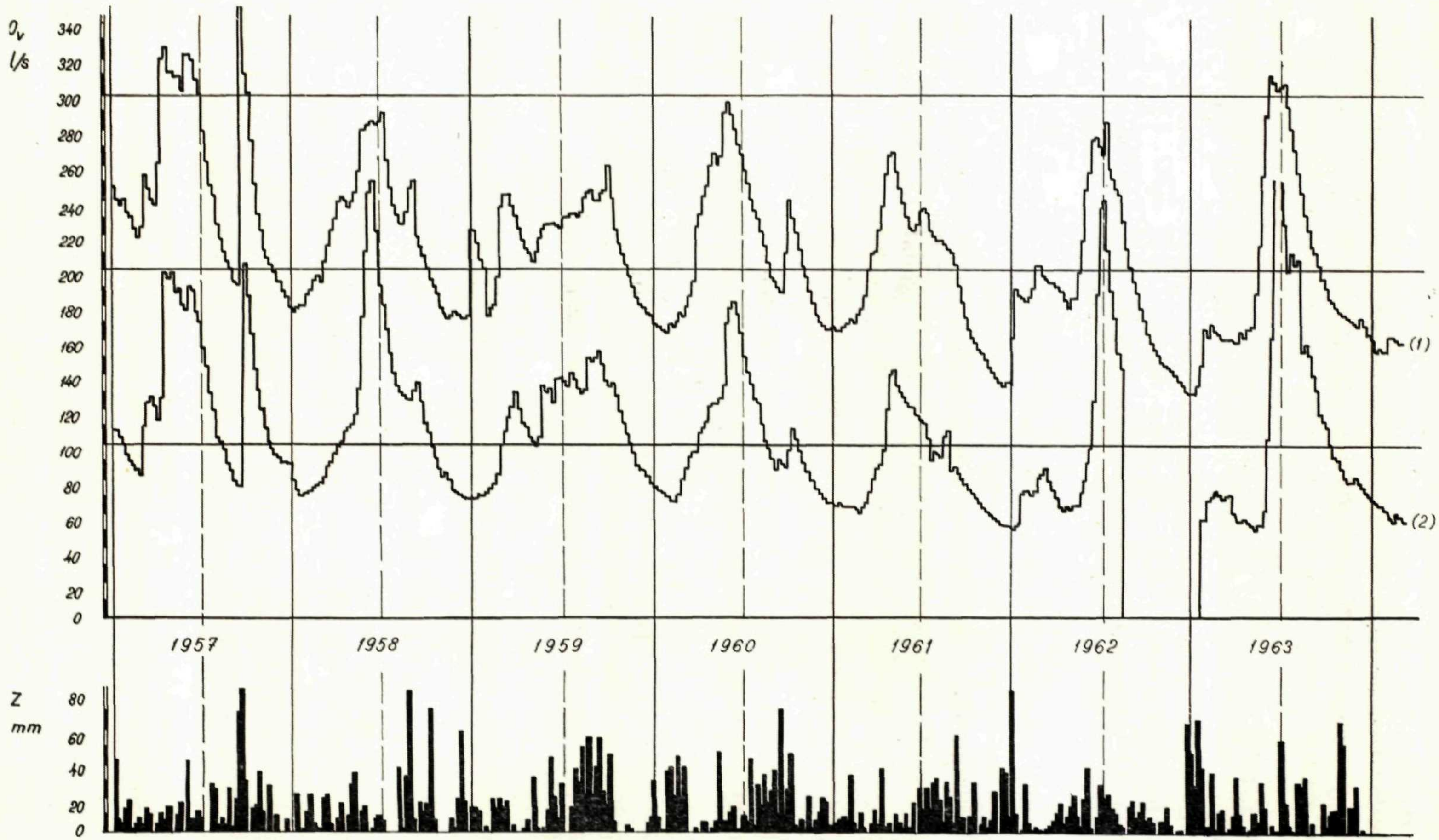

Fig. 2. The time course of the total yield of springs in carbonate complexes of the Choč (1) and Krížnan (2) units in the Small Carpathians (measured by the Hydrometeorological Institute in Bratislava)

Tab. 1.

The sum of rainfall of the winter half—year and the total discharge of springs from carbonate complexes in the Small Carpathians

Year	Rainfall November – April, in mm		Total discharge of springs November – April, in mm	
	Sološnica	Kuchyňa (Vývrat)	Krížna unit	Choč unit
1957	281	317	103,18	83,43
1958	241	252	93,30	52,33
1959	256	290	80,44	49,58
1960	326	337	85,46	51,01
1961	233	226	73,23	48,36
1962	309	308	74,15	46,65
1963	357	396	72,69	41,35
1964	187	207	57,55	37,20
1965	306	319	85,85	48,73
1966	270	282	83,50	55,76

sums of rainfall in the following winter half-years, which were added to the higher yield of ground water coming from the storage of the summer half-year, gave higher runoff of ground water (Duba 1967).

We observe the opposite appearence in the years 1962, 1963 and 1964 (Fig. 3). The relatively high precipitation of winter half-years in 1962 and 1963 causes, in relation to other years, only small yields of ground water discharge. It is caused (Fig. 2, Tab. 2) by small rainfall in the precedent summer half-years 1961 (395 mm), 1962 (292 mm) and 1963 (393 mm), which mostly evaporates and builds the surface runoff, which reflects in the lowest minimum value of yield in the beginning of November during the investigated period. Afterwards also the following high precipitation in the winter half-year firstly has supplied the ground water storage and secondly took part in the building of ground water discharge from the springs.

Tab. 2.

The sum of rainfall of the summer half—year and the total discharge of springs from carbonate complexes in the Small Carpathians

Year	Rainfall May – October, in mm		Total discharge of springs May – October, in mm	
	Sološnica	Kuchyňa (Vývrat)	Krížna unit	Choč unit
1957	441	531	91,2	58,6
1958	427	519	79,6	47,8
1959	407	520	90,8	50,4
1960	444	526	71,0	46,9
1961	395	365	56,8	41,0
1962	292	330	—	43,0
1963	393	412	89,7	47,9
1964	506	550	69,45	42,24
1965	462	581	124,55	59,77
1966	557	642	—	57,93

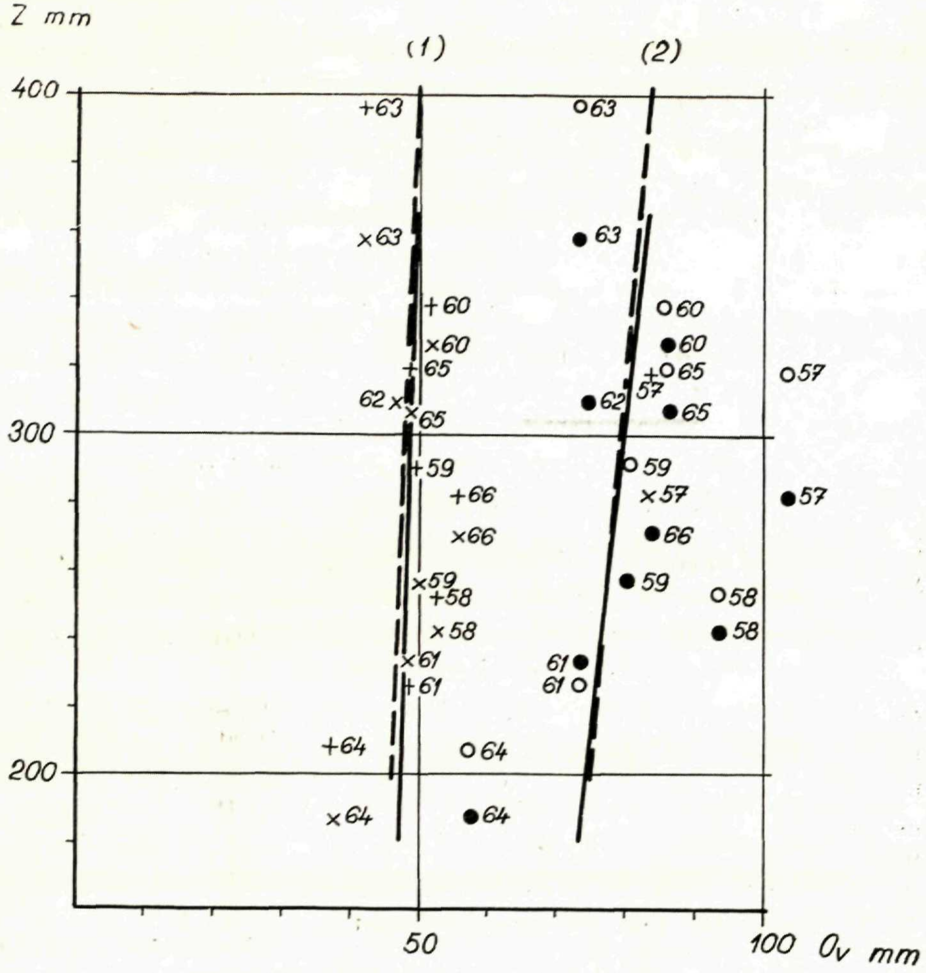

Fig. 3. Dependence of the total discharge of springs from carbonate complexes of the Choč (1) and Krížnan (2) units in Small Carpathians from November till April (O_v) on the sum of rainfall (Z) for the same period (1957—1966)

More unfavourable conditions have occurred in comparison with the year 1962 and 1963 in the winter half-year 1964. In that year two unfavourable factors have effected, namely the very small storage of karst water from the summer half-year 1963 (in consequence of low rainfall in this half-year — 393 mm) and the low rainfall sums in the winter half-year 1964 (187 mm). The prevalent part of seepage water from this rainfall has recharged the storage of ground water. Therefore the lowest discharge of karst water has been expressed in that year during the investigated period.

It is possible to work out a more exact expression of these orientate relations with the help of simple balance equation, which we arrange for the solved case in the following way (Duba 1959, 1961, 1964):

$$Z = O + O_{pr} + O_p + V \pm R_p \, , \tag{1}$$

where Z is the rainfall; V — the total evaporation; R_p — the changes of ground water storage;

O — the runoff in the water courses, which will have two component parts namely the total yield of springs (O_v) as sources of individual brooks and the surface runoff (O_d) which builds an inflow along them (the inflow in surface water courses will be zero in conditions of the peak basin);

O_{pr} — the underground runoff into brooks, which represents the increase of flow rate along the brook, in consequence of inflow or drainage of ground water;

O_p — the underground runoff along the perimeter of the partial basin or in other words, the hidden outflow of ground water into the neighbouring partial basin or other hydrogeological units (we eliminate the underground inflow into basin in consequence of its peak position near the watershed line).

We can determine the O and O_{pr} by direct measurements, where the O_v and O_{pr} represent that part of underground runoff, which mostly interests us from the point of view of utilizing the ground water.

We have used for the numerical composition of average water balance the average of rainfall data of the two stations for the period 1957—1965 (Tab. 1 and 2) and the average monthly data of the total evaporation from the surface of soil calculated by Tomlain. The difference between that data is given in Tab. 3 in the first group.

The increase of runoff has been determined by repeated single hydrometrical measurements (Kullman 1965) in surface courses in the investigated complexes of the Choč unit and by calculation using the equation (2) for the balance of Krížnan unit. These increases include the members of balance O_d and O_{pr} commonly and also include the surface runoff in the Eastern direction in the Krížnan unit (the discharge of springs occurs in the NW foot of the Small Carpathians). The direct measurements of yield of the mentioned discharge of springs in the NW boundary of both tectonic units include the further member of balance O_v.

We can compose afterwards the special balance equation for each of these seasons in accordance with the knowledge of different regime of changes of the ground water storage in the winter and summer half-years (Duba 1962). It is possible to calculate from two equations the remaining two unknown, namely the hidden outflow of ground water (O_p) and changes of the ground water storage ($\pm R_p$), whose tendency of changes we know. It is suitable to write this equation for such a solution in the following form:

$$(Z - V) - (O_d + O_{pr}) - O_v = O_p \pm R_p \,. \tag{2}$$

Results of these calculations are summed up in Tab. 3. From them it is obvious that the total runoff (is meant specifically) is nearly equal, also for the individual seasons. The difference will be in the participation of rainfall in the formation of runoff (in the case of considering single seasons). In the winter half-year the runoff is formed by 38—39 % of rainfall, 28—27 % of rainfall raises the storage of water in the basin (in the longterm average we can suppose, that it mainly raises the storage of the ground water) and 34 % of rainfall is consumed by total evaporation. Up till 93 % of rainfall is consumed by total evaporation in the average of summer half-years and so only 7 % can take part in the formation of runoff. This is only due to the discharge of ground water from sub-surface accumulations, if the underground runoff in the summer half-year is equal to that in the winter half-year. We can convince ourselves about the relatively perfect compensation of runoff in bot investigated tectonic units also in Fig. 3.

Tab. 3.

The average water balance of carbonate complexes in the Small Carpathians for the period 1957—1956

Members of balance and their groups	Krížna unit						Choč unit					
	Average of winter half-years		Average of summer half-years		Annual average		Average of winter half-years		Average of summer half-years		Annual average	
	mm	l/s	mm	l/s	mm	l/s	mm	l/s	mm	l/s	mm	l/s
Z	274,7		450,1		724,8		274,7		450,1		724,8	
V	94,0		420,0		514,0		94,0		420,0		514,0	
$Z - V$	180,7	236,2	30,1	39,3	210,8	137,8	180,7	786,2	30,1	132,0	210,8	459,0
$O_d + O_{pr}$	23,0	30,1	23,0	30,1	46,0	30,1	18,1	78,6	18,1	78,6	36,2	78,6
O_v	80,7	105,3	84,1	110,0	164,8	107,7	51,0	222,0	48,6	211,8	99,6	216,8
O_p	—	—	—	—	—	—	37,5	163,6	37,5	163,6	75,0	163,6
$(O_d + O_{pr}) + O_v + O_p$	103,7	135,4	107,1	140,1	210,8	137,8	106,6	464,2	104,2	454,0	210,8	459,0
R_p	+77,0	100,8	−77,0	100,8	0,0	0,0	+74,1	322,0	−74,1	322,0	0,0	0,0

It is also shown (Fig. 3, Tab. 3), that differences between values of underground runoff O_v which is formed by equal precipitation in the nearly analogical hydrogeological, hydrological, climatical conditions in the Krížna and Choč units indicate in the last unit the possibility of greater hidden runoff of ground water (Kullman 1964, 1965c) in comparison with the Krížna unit. This conclusion is proved by the results of hydrogeological investigation of Záhorská nížina (plain) (Kullman 1966), which show the remarkable outflow of karst water from the Choč unit into Quaternary sediments of Záhorská plain in this region (about 200 l/s).

The achieved calculations have attested the above-mentioned conclusions (Tab. 3). We determine the underground runoff (O_p) with the value rounded to 164 l/s, which is in agreement with the above given number, because our balance calculations do not consider the whole area of carbonate rocks of Small Carpathians, which can take part in the forming of the mentioned underground runoff (the part of carbonate rocks attaching the Záhorská plain between Sološnica and Plavecký Mikuláš). The Záhorská plain forms as if a sub-surface compensatory reservoir, what is evident from a nearly vertical course of straight line relationship in Fig. 3.

The submitted balance calculations have proved the high compensatory effect of the investigated carbonate rocks by its sub-surface accumulations, which condition the considerable leveled average underground runoff.

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Geologické práce, Zprávy 44—45. Bratislava 1968

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THE RELATIONSHIP BETWEEN THE RAINFALL AND THE REGIME OF COLD AND THERMAL KARST WATERS OF THE WEST CARPATHIANS

Introduction

The West Carpathians, in particular their central belt extending across the territory of Slovakia, are formed to a great extent of Mesozoic carbonate (limestone-dolomite) sediments. On the surface they cover approximately 3,280 km^2 (Kullman 1964). An additional part of carbonates of considerable extent is covered by Late Tertiary and Quaternary deposits and volcanic complexes in the lowlands. Considerable quantities of karst water circulate in the carbonate complexes occurring at the surface. They are estimated roughly at 26 to 33 m^3/sec. (Kullman 1964). The Mesozoic of the West Carpathians consists of a row of parallel anticlinoria and synclinoria which originated during Cretaceous folding (Mahel' 1965). The anticlinoria and synclinoria which follow roughly the direction of the Carpathian arc (SW—NE, W—E, NW—SE) are composed of several partial synclines and anticlines. The post-Palaeogene structural forms (morphotectonic units) — megaanticlines (mountains), and magasynclines (depressions and lowlands) are superimposed on the Cretaceous structural forms, diverging with them in direction. It is especially the synclinoria which are important from the hydrogeological point of view, because particularly in them occur the carbonate complexes. The partial synclines are of great importance for the circulation of karst waters (Mahel' 1966). The carbonate complexes occurring on the slopes of the "Core mountains" plunge under the adjoining depressions and lowlands frequently forming tectonic grabens.

Part of the karst waters confined to the carbonates of tectonic blocks which do not reach greater depth lacks the possibility of deep circulation. The karst waters producing in the rock complexes subsurface water flows mostly issue in contact springs on the contact with impermeable substratum. That part of karst waters which is confined to carbonate complexes reaching greater depth, participates in the deep circulation. Part of the water appears at the surface mainly as barrier springs, mostly in places affected by young tectonics (Kullman 1964). The remaining portion of water is involved in deep circulation and contributes to the replenishment of thermal water storage. On the basis of available data, about 700 lit./sec. of thermal waters tied up with carbonate complexes have been documented (Franko 1964, 1965).

The thermal waters constitute a sort of thermosiphon the descending branch of which is formed by plunging folds and the ascending one generally by young longitudinal and transversal faults. Chemically the thermal water is of a uniform calcium-sulphate-magnesium-hydrocarbonate type with several modifications (Hynie 1963; Franko 1964).

Approach to the problem

One of the problems implied in the research of thermal waters of the West Carpathian Mountains is the determination of the retardation of changes in thermal water discharge behind the changes in precipitation amount in the catchment areas (Franko 1964). This paper informs about the influence of infiltration on the cold and thermal karst water regime and analyses the time lag between infiltration and its manifestation in cold and thermal springs discharge. The evaluation of this relationship is based on the knowledge of these relations in shallow karst waters and of their diversity in individual areas of the West Carpathians. The studies made so far (Kullman 1965) have verified the basic relationship between precipitation and infiltration. It has been ascertained that the winter-half of a hydrological year has primary importance for re-charging the cold karst-water reserves (infiltration of snowmelt water), which is reflected in the main maximum of spring discharge (Fig. 1-I.). This relationship holds true although the precipitation maximum within the whole of the Mesozoic Carpathian ranges falls into the summer-half of the hydrological year. Analogous relations have also been found for thermal waters appraised up to date (Franko 1967). The reflection of precipitation conditions in the karst-water regime (the attainment of maximum discharge) depends on two basic factors: (1) Time factor of infiltration of snowmelt-water accumulations and (2) distribution of total evaporation in time, depending on temperature conditions. The interaction of these two factors affects quantitatively the infiltration and its maxima in time. Depending on the altitude above sea level, the maximum infiltration shifts from March in the lowest catchment areas (up to 400—600 m) to July in the highest lying catchment areas (1,000—1,300 m; Zaťko 1965). Temperature and, consequently, the rate of evaporation control the influence of summer precipitation on the replenishment of karst waters. This influence is negligible in the low-lying areas, although the summer precipitation represents a prevalent part of the year total. Thus, for example, in the carbonate rocks of the Krížna unit in the Malé Karpaty (Little Carpathians) Mts. (450—500 m above sea level), the annual mean precipitation in the period of 1957—1966 has been 724.8 mm (hydrological year considered); from this amount 274.7 mm falls to the winter half-year and 450.1 mm to the summer half-year. In the winter half-year, infiltration and surface run-off cover 66 per cent and evaporation 33 per cent; in the summer half-year evaporation reaches up to 93 per cent, infiltration and surface run-off being restricted to 7 per cent (Duba—Kullman 1967). With the increase in the absolute altitude of the area, evaporation decreases and the share of infiltration of summer precipitation substantially rises. This phenomenon is reflected in a second, subsidiary maximum of infiltration (e. g. the Strážovská hornatina Mts., 600—950 m above sea level, Fig. 2-III.). As a result of the delayed infiltration of snowmelt water in the regions of highest altitude, which roughly coincides with the time of maximum summer precipitation, one common infiltration maximum develops. In accordance with Zaťko's view, (1965), these changes in the regime of cold karst waters may be correlated with the altitude above sea level, which affects the time differences in their manifestations.

In comparing the regime of shallow karst waters with the regime of deep thermal karst waters in their relation to infiltration, the same method was used as in studying the ascending regimes of cold karst waters in the presumed catchment areas of thermal waters, or in the neighbouring areas of approximately the same altitude (e. g. the Žihľavník karst waters in relation to those of Trenčianske Teplice).

The relationship shown on several examples

The analysis of the relationship studied is shown on the Vajar Spring, as its temperature (15 °C) places it at the transition between cold waters (7—8 °C) and thermal waters (more than 20 °C). Of thermal springs those with long-term discharge records have been chosen for analysis. They are as follows: The springs in Bojnice (28 °C), in Trenčianske Teplice (40 °C) and in Sklené Teplice of a temperature of 53 °C.

The Vajar Spring issues on the southern margin of the limestone-dolomite complex of the Choč unit of the Malé Karpaty Mts. This rock complex extends between Rohožník and Smolenice, and to the north of the spring it is buried by the Intracarpathian Palaeogene, which in this part causes a deeper circulation of percolated water on its way to the issue on the ground surface. The spring has a constant temperature of 15 °C and the discharge varying between 36.3—96.1 lit./sec. (measured during the decade 1957—1966).

For the characterization of the shallow regime of this area, the regime of springs of the adjacent carbonate complex of the Krížna unit (average from 18 springs) has been selected. The area is distinguished by a non-uniform distribution of precipitation whose mean annual amount reaches 724 mm (determined from the period of 1957—1966). The winter half-year average is 274 mm and the summer half-year amount averages 450 mm. Maximum precipitation falls in the months of June to August.

The shallow karst-water regime shows one infiltration maximum in the months of March—April (in the year 1958 falling to the second week of April). The June—August precipitation maximum is virtually not reflected in the discharge curves of most years observed. The regime of the Vajar Spring with a deeper circulation also displays one clear-cut maximum reflecting the infiltration of snowmelt water. The retardation of this maximum relative to the maximum of the shallow karst-water regime, as appraised from the values of several year records, averages 3/4 to 1 month (Fig. 1-I.).

Thermal springs of Bojnice (5,6) appear on the western margin of the Upper Nitra valley (Hornonitrianska kotlina), at the southern foot of the Malá Magura Mts. at 297—332 m above sea level, i. e. at 50—70 m above the surface of the depression. The springs issue from the Palaeogene flyschoid sequence (underlain by Middle and Upper Triassic dolomites) at the crossing of longitudinal (NE—SW) and transversal (NW—SE) faults. They are confined to the Cretaceous synclinorium of the polystructural Fatro-Tatric zone, in which the Upper Nitra depression is also included, and ascend to the surface along the marginal (NE—SW) faults at the eastern boundary of the mountain range. The main part of their catchment area spreads at the north-western margin of the Žiar Mts., in the Mesozoic carbonates at an altitude between 400—800 m above sea level. The subordinate part of the catchment area extends to the W and NW of the issues of thermal waters in the basal Palaeogene dolomitic breccias, at 400—600 m above sea level. The springs constitute two spring lines differing in temperature. One of the spring lines has a temperature of 22 °C to 40 °C and the other of 40 °C to 47.6 °C. The discharge of the Bojnice thermal springs ranges from about 33 to 39 lit./sec. The springs of the line with a higher temperature are presumed to derive from the Mesozoic dolomites submerged deep beneath the Tertiary deposits, whereas the springs of a lower temperature are interpreted in terms of mixing of water from depth with water from the Palaeogene breccias occurring to the W and NW of the issues. For our observations the Strand Spring (of a temperature of 28 °C) has been selected because two-year discharge records are available. For the springs of higher temperature not even one-year discharge record that would be unaffected by human interference was available.

The characterization of the shallow karst-water regime in the zone cor-

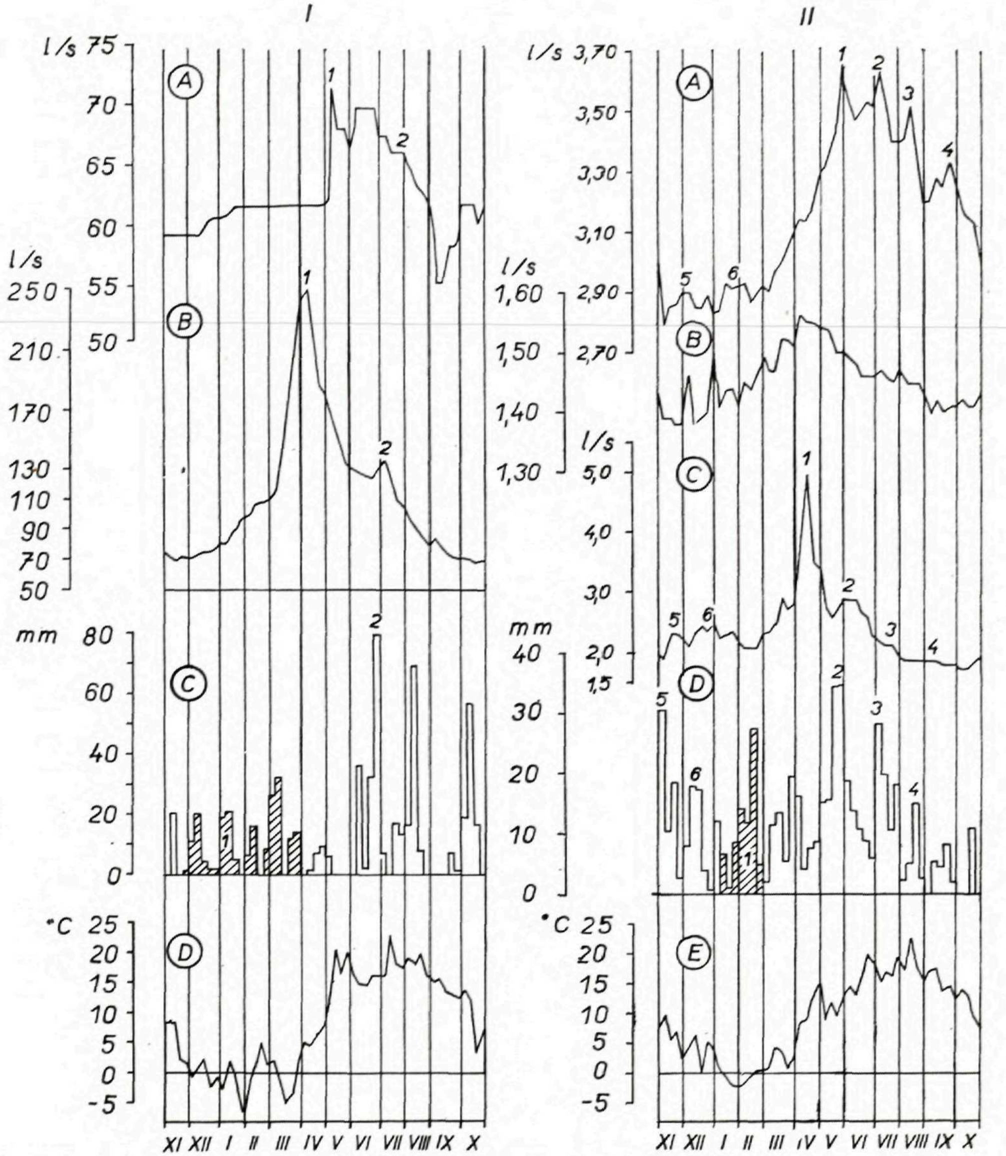

Fig. 1. Relation of rainfall to the yield of cold and thermal waters

I — Vajar spring (1958): A — Rohožník village (Vajar spring 15 °C); B — Malé Karpaty Mts. (the Krížna] tectonic unit); C — rainfall (Sološnica village); D — temperature of air (Kuchyňa village)

II. Bojnice spa (1961—62): A — Bojnice spa (the strand thermal spring 28 °C); B — Dubnica springs (1955—59); C — Mokrá dolina valley; D — rainfall (Prievidza); E — temperature of air (Prievidza) responding in altitude to the main catchment area has been inferred from the analysis of the spring called Mokrá dolina 3, and in the zone agreeing in altitude with the subordinate catchment area from the analysis of the Dubnica springs (in this case the analysis is based on the data for the mean hydrological year during the period 1955—1959). Data on precipitation and temperature were provided by the Observation Station Prievidza—Školka (Fig. 1-II). The region studied is distinguished by a relatively uniform distribution of precipitation in the course of the hydrological year. Maximum precipitation amount has been measured in the months of May—June.

The shallow spring regime shows one main maximum in April produced by snowmelt water. The Dubnica springs display the maximum in the first week of April, the Mokrá dolina 3 spring one week later, which delay is due to the higher altitude of the area concerned. The precipitation maximum of May—June does not assert itself substantially in infiltration or in the replenishment of the karst water storage, probably because of strong evaporation. The precipitation of December is expressed more pronouncedly, which corresponds to decreased evaporation.

The regime of thermal Strand Spring with a deeper circulation shows similarly as the springs of shallow circulation, one main maximum towards the end of May produced by snowmelt. As is seen from Fig. 1-II, the retardation of this maximum behind that of the springs with shallow circulation and thus behind the infiltration of snowmelt water amounts to 1 1/2 to 1 3/4 months. The discharge of the Strand Spring lags behind the summer and autumn precipitation (May to December) by the same time interval (1 1/4—1 3/4 months).

Thermal springs of Trenčianske Teplice occur in the valley of the Teplička brook, in the north-western part of the Strážovská hornatina Mts., at 270 m above sea level. The springs issued originally from the stream alluvia, along a transverse fault of NW—SE strike following the valley. At present the thermal water is being captured in the Triassic dolomites underlying the alluvial deposits. The thermal water is confined to a partial syncline of the polystructural Tatro-Fatrian zone of Cretaceous age. The catchment area of the springs is considered (Mahel' 1962) to occur in the carbonates of the Manín—Inovec Group in the Strážovská hornatina Mts., at an altitude of about 400—900 m. The springs are caught in three drilled wells (V-2, V-3 wells in one basin and P-1 well in the other). The thermal water has a temperature of 39—40 °C and a discharge of 15—23 lit./sec. For the springs four-year records are available. The P-1 spring, selected for observations on account of its being least affected by the fluctuation of water level in the stream and its alluvia, has a discharge of 4.8 to 7.0 lit./sec.

For the characterization of the shallow karst-water regime (Fig. 2-III) in the zone at the height of the presumed catchment area, the analysis of the karst-water system of the carbonate Žihľavník block (outside the catchment area) has been selected. Its preponderant part is situated at 600—950 m above sea level. Precipitation amount is rather unevenly distributed in this area. The winter months are poorer in precipitation than other seasons. The main precipitation maximum falls into June and a subordinate one into the month of November.

The shallow karst-water regime of Žihľavník displays two maxima: The first and principal maximum produced by spring thaw occurs in the fourth week of March and the second, lower maximum is observed in the fourth week of June. The regime of deep thermal waters shows only one maximum expressing the infiltration of snowmelt water. This maximum, however, is manifested as late as at the end of November or the beginning of December. As is seen from Fig. 2-III, the retardation of the discharge maximum of the thermal spring behind the maximum shown by cold waters and, thus, also behind the infiltration of melt water is about 8 3/4 months. Other peaks of this thermal spring compared with those of cold karst waters exhibit roughly the same delay.

Thermal springs of Sklené Teplice issue in the valley of the Teplá brook at 350 m above sea level. They belong to the group of thermal springs joined to the Palaeozoic-Mesozoic inliers emerging from beneath the Late Tertiary volcanics of central Slovakia. The springs issue partly from a travertine mound, partly from the stream alluvia or directly from the Triassic carbonates. The outlets are located on the a NNE—SSW trending fault that forms the western boundary of the inlier and runs through the stream valley, and on a NE—SW fault bounding the inlier in the north. At present the thermal springs are caught by drilled wells. The ST-1 well captures the spring (discharge 25 lit./sec., temperature — 51 °C) on the NNE—SSW striking fault and ST-2 well captures the spring (discharge — 4.5 li./sec., temperature — 50 °C) issuing on the NE—SW fault (V. Struňák in Myslil—Franko 1967) The thermal springs along with those of the Zvolen depression (Kováčová, Sliač) are tied up with the Hron synclinorium. Their catchment area is presumed to occur in the Mesozoic carbonates on the southern slopes of the Veľká Fatra and the Nízke Tatry Mts., at an altitude of about 400—1 000 m above sea level; a certain proportion of water derived from young volcanic complexes is anticipated.

The temperature of thermal springs ranges from 31 °C to 53 °C. Their discharge varies from 4.3 to about 9.4 lit./sec. Of greatest temperature (53 °C) and discharge (4.73 to 6.19 lit./sec.) is the Joseph Spring. A systematic three-year record of the spring yields is available. For our observations the Ľudovít Spring with the maximum temperature of 49 °C and the combined discharge of the Ľudový and Viera Springs (max. temperature of 50 °C) have been taken

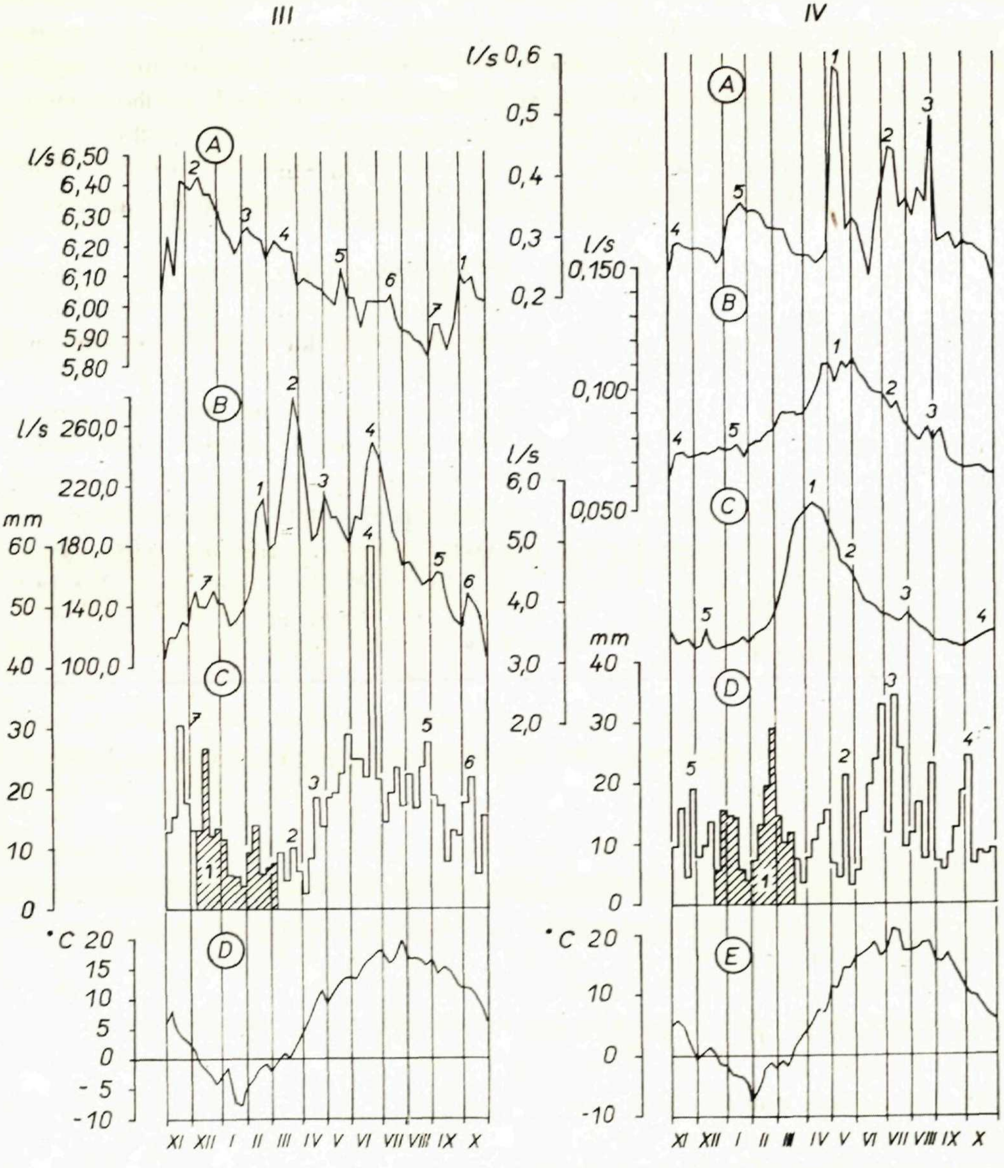

Fig. 2. Relation of the rainfall to the regime of cold and thermal water

I — Trenčianské Teplice spa (1963—66): A — Trenč. Teplice (P-1 spring 40 °C); B — Žihlavník hill; C — rainfall (Trenč. Teplice); D — temperature of air (Trenč. Teplice spa).

II — Sklené Teplice spa (1956—58): A — Sklené Teplice (Ľudový and Viera springs 53 °C); B — Sklené Teplice (Ľudovít spring 53 °C); Riečka vill. (Posviacené spring; 1957—63); D — rainfall (Banská Bystrica); E — temperature of air (Banská Bystrica town).

as a basis. The maximum temperature, however, was taken as 53 °C, i. e. that of the Joseph Spring. The other springs considered do not reach this temperature on account of their small discharge (they are considerably cooled during their ascent). The discharge of the Ľudovít Spring is 0.008—0.153 lit./sec., the aggregate discharge of the Ľudový and the Viera Springs is 0.114—0.746 lit./sec. The discharge measurements for the Joseph Spring are not accurate enough.

The characterization of the shallow karst-water regime in the zone corresponding in height to the presumed catchment area is based on the observations of the Posviacené Spring issuing to the NNW of Riečka at 600 m above sea level (Fig. 2-IV). In this area, the precipitation amount is relatively evenly distributed through the hydrological year. The mean precipitation maximum falls into the months of June—July, the subsidiary (autumn) maximum occurs in the period of October—December.

The regime of the karst spring with shallow circulation shows the main discharge maximum produced by snowmelt water in the mid-April. The precipitation maximum from June—July does not affect essentially the recharge of the karst-water storage. This is partly enhanced by the October—December maximum.

The deep thermal spring regime, similarly as the shallow spring regime, exhibits the main discharge maximum produced by snowmelt water in about mid-May. As is seen from Fig. 2-IV the retardation of this maximum behind that of springs with shallow circulation and, consequently, behind the infiltration of snowmelt water is, at the first sight, one month. Yet, with respect to a greater delays of principal maxima shown by shallower water regimes (Bojnice — 28 °C — 1 1/4 month, Trenčianske Teplice — 40 °C — 8 3/4 months), the retardation is not one month but roughly 13 months long. The other peaks of thermal springs related to the precipitation of summer and autumn seasons lag behind those of cold karst waters within approximately the same interval.

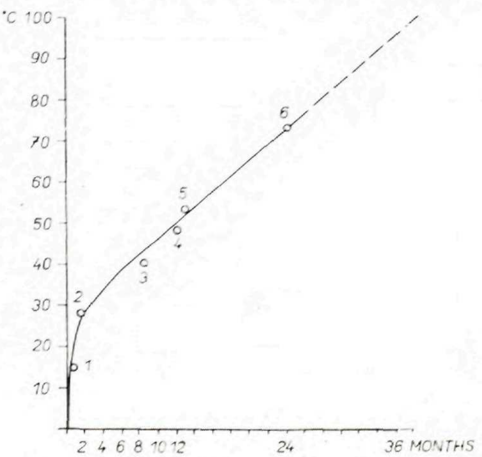

Fig. 3. Relation of the yield retardation of thermal waters to their temperature. 1. Vajar, 2. Bojnice, 3. Trenčianske Teplice, 4. Baden, 5. Sklené Teplice, 6. Karlovy Vary.

Conclusion

The submitted paper presents new information about the effects of infiltration on cold and thermal karst waters of the West Carpathians. We have assessed the time delays between infiltration and its manifestations in the cold and thermal springs. The following data are inferable from the analysis of these relationships:

(1) The storage of cold and thermal karst waters is replenished for the most part by the infiltration of snowmelt water, when evaporation is small owing to low temperature.

(2) Summer and autumn precipitation, by far exceeding the winter precipitation has a relatively small share in the replenishment of water storage.

(3) For the above-mentioned reasons, both the cold and thermal karst waters display one principal discharge maximum produced by spring thaw. In the course of the year, subordinate replenishment of water reserves takes place, depending on the altitude and thus also on the climatic conditions.

(4) The principal discharge maximum has been established in the cold karst springs immediately after infiltration, towards the end of March or in the first half of April.

(5) Compared with the maxima of cold springs, the main discharge maximum in the thermal waters examined is delayed. It is thought to be due to the depth of circulation, as is documented by the temperature of waters.

The values of retardation and temperature of individual thermal springs are listed in Table

Locality	Spring	Temperature of water in °C	Retardation
Rozhožník	Vajar	15	3 – 4 weeks
Bojnice	Strand	28	1 1/4 month
Trenč. Teplice	P – 1	40	8 3/4 months
Baden (Switzerland)	—	48	12 months
Sklené Teplice	Ľudovít Ľudový + Viera	53	13 months
Karlovy Vary (Bohemia)	—	73	2 years

From the determined values, the curve of retardation maxima of the discharge and temperature of thermal waters have been constructed (Fig. 3).

We have compared the obtained values for the waters of Slovakia with those of thermal waters in Karlovy Vary (Myslil – Václ 1966) and from Baden in Switzerland (Cadisch 1931). The values are given in the Table above. The data correlate well with the constructed curve. With respect to conformable geological conditions of the Slovak localities and the Baden springs, the coincidence is not surprising. But it is strange in the case of Karlovy Vary springs where the crystalline rocks form the environment of ground water circulation. This would suggest that the environment of circulation is not a controlling factor of retardation.

The curve, as given above, does not obviously represent the final solution of the problem. It must be expected that observations and results obtained from other localities will contribute to a greater precision of its shape.

(6) From the data recognized at the localities so far examined it may be tentatively inferred that the retardation depends mainly on the depth of circulation and not on its length. We think it to be the result of the regime. This assumption is suggested by the more sensitive reactions of thermal with confined level water regime to the changes of infiltration, compared with the responses of the cold water regime.

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GEOLOGICKÉ PRÁCE, ZPRÁVY 44–45

Vydal Geologický ústav Dionýza Štúra v Bratislave
vo Vydavateľstve Slovenskej akadémie vied
z príležitosti konania XXIII. medzinárodného geologického kongresu v ČSSR
Vytlačila PRAVDA, vydavateľstvo ÚV KSS v Bratislave zo sadzby garmod Extendet.
AH 27,56, VH 28,04

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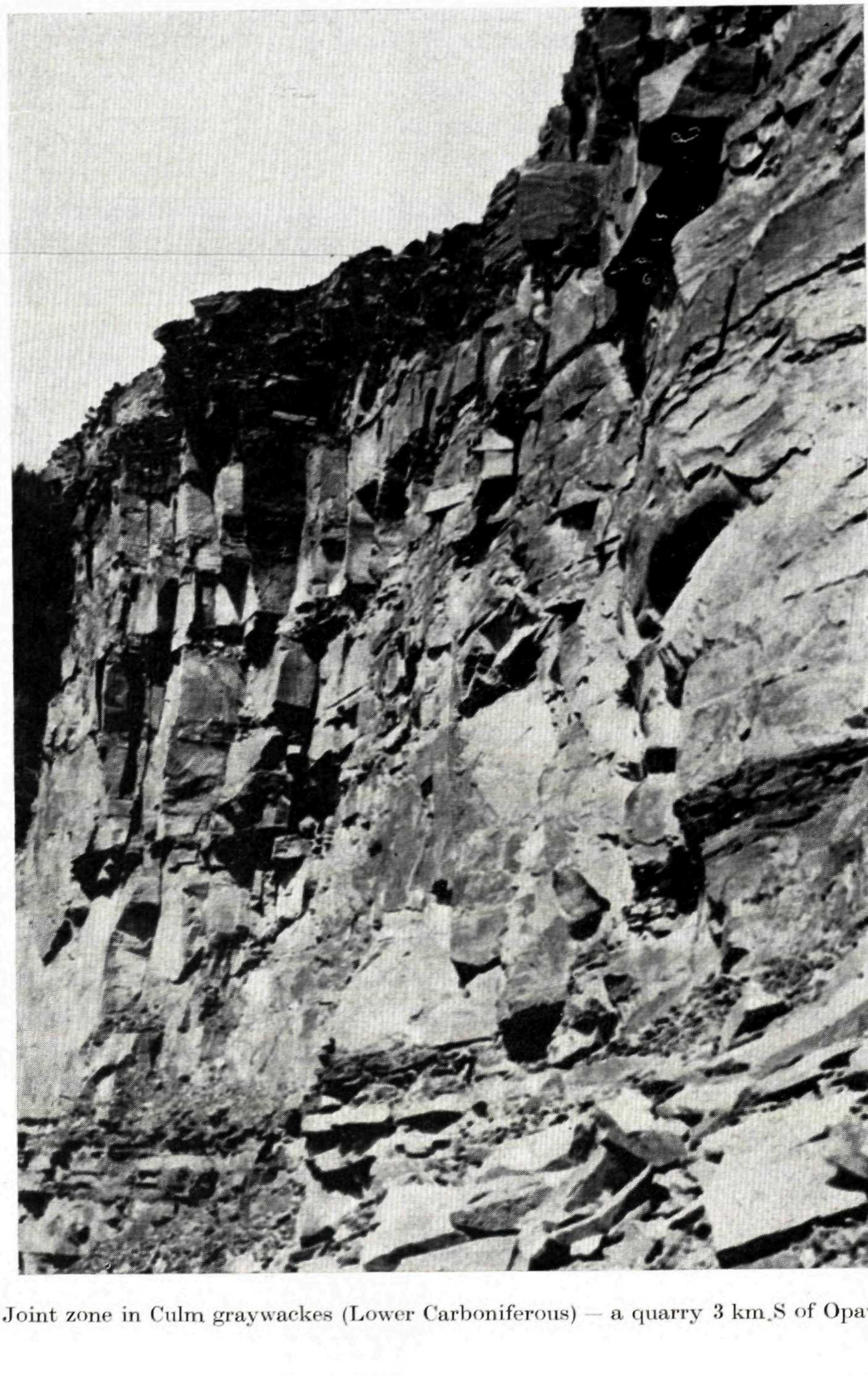
Fig. 1. Bedding joints in the Eocene beds of the Magura Flysch in a quarry at Sv. Štěpán.

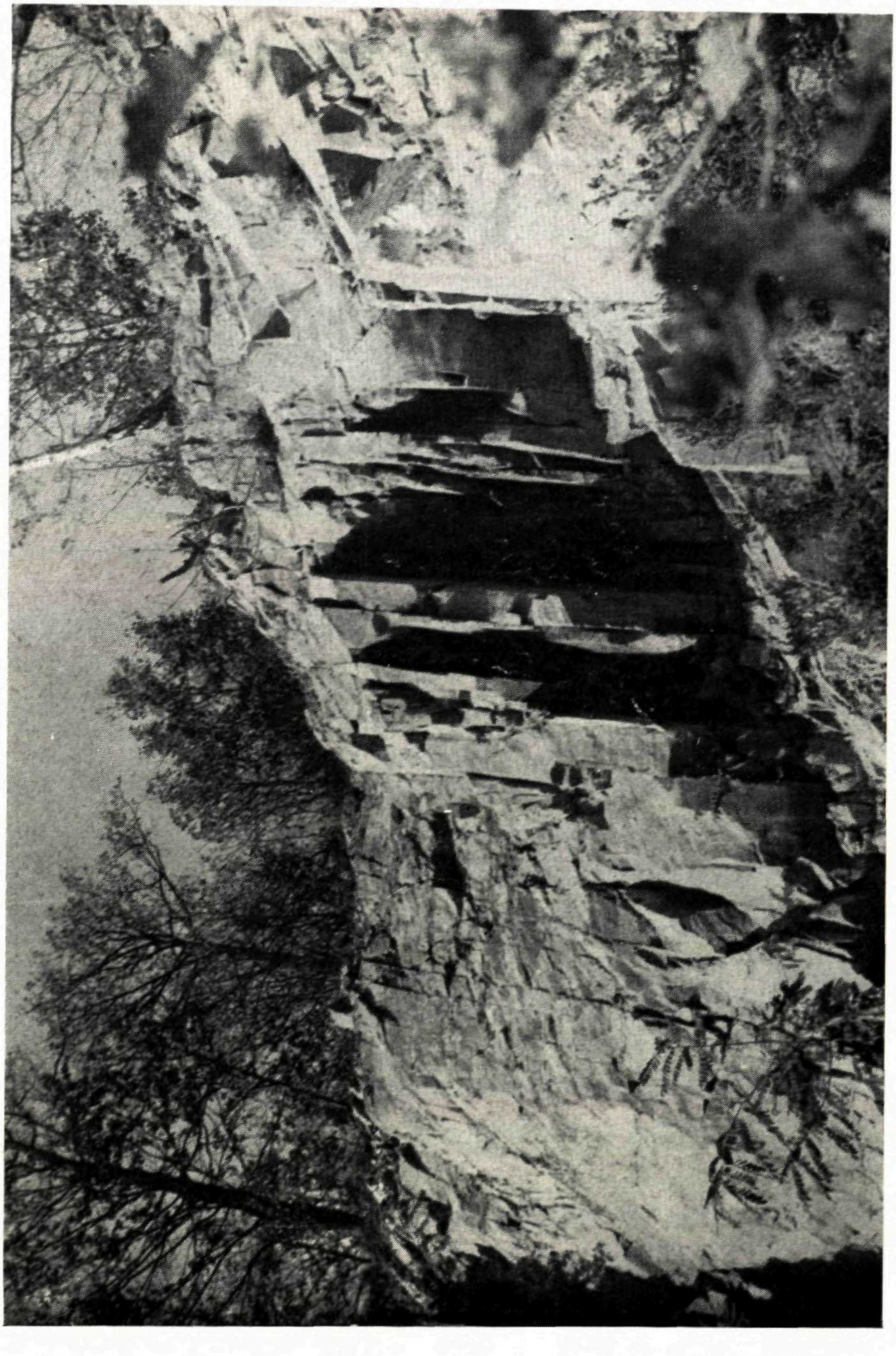
Fig. 2. Joint zone (see the left part of the Fig.) in Culm graywackes (Lower Carboniferous), in a quarry at Hrabůvka.

Fig. 1. Joint zone in the sandstone of the Doláň beds in the Magura Flysch (Paleocene). Dips of beds are about 15°; a quarry 1 km NE of Ježov.

Fig. 2. Joint zone in shales of the Zlín beds of the Magura Flysch (Eocene). Outcrops at Vyšné Vladiče in the East Slovakia.

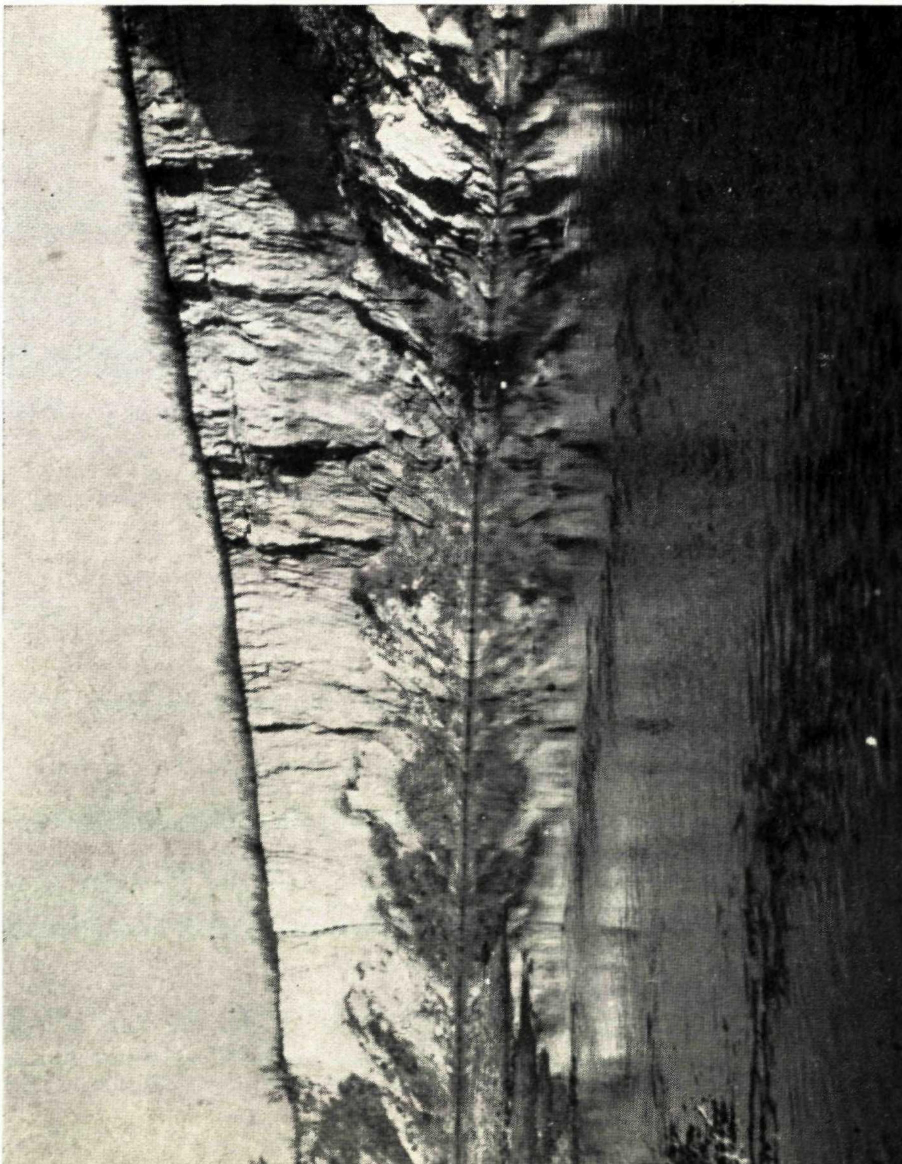


Joint zone in Culm graywackes (Lower Carboniferous) — a quarry 3 km S of Opava.



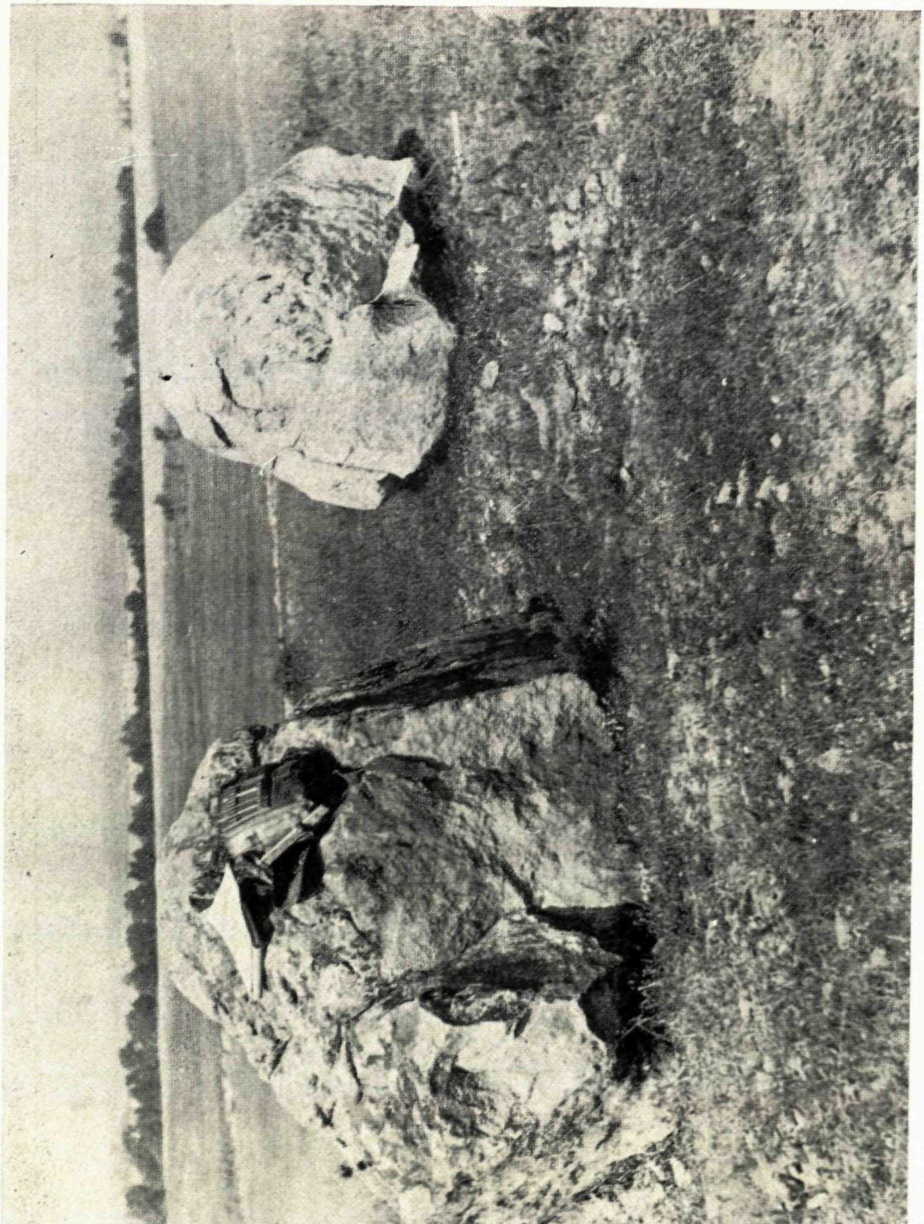
Joint zone deforming graywackes of the Lower Carboniferous at a dip of about 25° — a quarry NE of Olomouc.

Loess-loam deposited on fluviatile sediments of the Laborec river (Photograph by V. Baňacký)



Erosive bank of the Ondava river in loess and fluviatile sediments (Photograph by V. Baňacký)

Solifluction-proluvial sediments in the Sub-Vihorlat depression (Photograph by V. Baňacký).



Blocks of pyroxene andesite dragged onto the mountain-foot step by solifluction (Photograph by V. Baňacký)