

ORIGIN OF ORE MINERALIZATION IN MAFIC AND ULTRAMAFIC ROCKS OF THE SOWIE MOUNTAINS BLOCK SURROUNDING

Geneza mineralizacji kruszcowej mafitów i ultramafitów w otoczeniu bloku sowiogórskiego

Ryszard SALACIŃSKI

Institute of Geology, Faculty of Geology of the Warsaw University
al. Żwirki i Wigury 93, 02-089 Warszawa

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ABSTRACT. Signs of qualitatively and quantitatively variable ore mineralization can be observed in mafic and ultramafic rocks occurring in the Sowie Mountains block surrounding. The author has studied the ore associations in the Nowa Ruda and Ślęza gabbroic massifs and in the Gogolów—Jordanów and Przygórze (near Nowa Ruda) serpentinite massifs. The samples for studies have come from boreholes drilled by the State Geological Institute and the others have been collected in quarries and outcrops. The routine optic reflected light microscopy has been used by the author as the main investigating method. The studies have been completed by use of the electron microscopy and electron microprobe method (Salaciński, Zawidzki 1983). Experimental chemical investigations have been also performed (Białowolska, Salaciński 1984).

The Cr-Fe-Ti oxide and Fe-Cu-Ni-Zn-Pb sulfide mineralizations have been characterized in the present paper. The

analysis of the ore mineral alteration processes at the post-magmatic stage and recognition of the regularities in these processes, in formation of new mineral phases and crystal habits and ore deposit-type concentrations are the essential part of the report.

In the author's opinion the well-known chromite deposit at Taładla is of hydrothermal origin. Chromite occurring in the ore from this deposit is a secondary mineral with respect to the primary chromite dispersed in the rocks of the whole massif. The hydrothermal chromite was formed due to mobilization and redistribution of chromium from the primary chromite generation. The author indicates the possibility of the other chromite deposit occurrences in this massif, which should be connected with tectonic zones.

The existence of the ore concentrations of copper, nickel and iron at the deeper levels of the Nowa Ruda massif is also probable.

INTRODUCTION

Outcrops of plutonic mafic and ultramafic rocks occupy relatively small surface in the Sude-tes and in the Foresudetic area. Most commonly such rocks occur at the marginal zone of the Sowie Mountains block, where they form several massifs in the rocks of the northern, eastern and south-western sides of the block (Fig. 1). Their tectonic

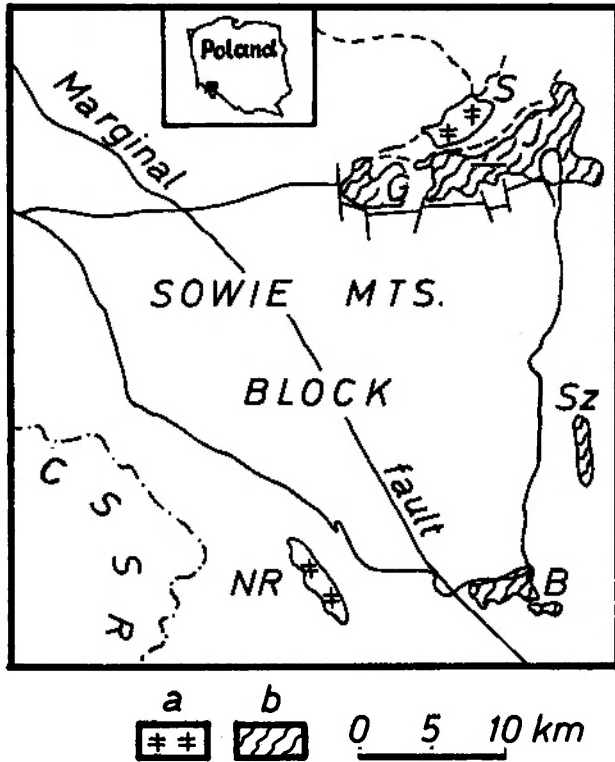


Fig. 1. Occurrence of mafic and ultramafic rocks in the surrounding of the Sowie Mountains block: *a* – gabbroic massifs; *b* – serpentinite massifs; *B* – Braszowice-Grochowa massif; *G–J* – Gogółów-Jordanów massif; *NR* – Nowa Ruda and Przygorze massif; *S* – Słęża massif; *Sz* – Szklary massif (after Oberc 1972)

Występowanie mafi-
tów i ultramafi-
tów w osłonie bloku s-
owio-
górskiego: *a* – masywy gabbrowe; *b* – masywy serpentynitowe; *B* – masyw Braszowic i Grochowej; *G–J* – masyw Gogółów-Jordanów; *NR* – masyw Nowej Rudy i Przygorza; *S* – masyw Słęzy; *Sz* – masyw Szklar (wg Oberca 1972)

positions, ages and origin have been the subjects of scientific discussions since years, and even presently, in spite of numerous observations and studies, the opinions on these problems are very variable. The fact, that “the Sowie Mountains block together with the mafic rocks, mainly serpentinites, which surround it and in significant part underlie this block, and in sequence underlaid by granitoids, are the deepest units of the tectonic structure of south-western Poland, and Lower

Silesia particularly ...” (Oberc 1972), is the reason of this situation. Mafic and ultramafic rocks in those units are the components of the oldest structural stages in this part of Poland. They have a complicated geological history, because they were submitted to multiple alterations and deformations during several orogenic cycles, from Precambrian till the youngest ones.

All this causes that the studies of the mafic and ultramafic massifs in the Sowie Mountains block surrounding are difficult. The geological investigations made hitherto have been performed in fragments and they have been pertinent to small-scale geological problems. Thus reports are lacking, which would present geological characteristics of all those massifs or each of them separately. Only recently an attempt of such complex elaboration of the Gogółów–Jordanów serpentinite massif in the north of Sowie Mountains block has been made (Ciemniewska *et al.* 1981). Problems connected with mineral raw materials have been the reason of these works. The investigation project realized by the then Geological Institute of the Central Geological Office (presently State Geological Institute) intended to elucidate the possibility of occurrence of chromium ores and other mineral raw materials connected genetically and spatially with the ultramafic rocks of the Gogółów–Jordanów massif. The works were performed in the area of Uliczno–Czarna Góra, in the direct neighbourhood of the chromium ore deposit at Tąpadla, which descriptions are known from literature (Birecki 1960b, 1962). This area has been selected for detailed studies on the basis of the earlier prospecting works, made by geophysical (Grzywacz 1959; Okulus, Cieśla 1971) and schlich survey methods (Fedak, Kowarz 1969; Fedak *et al.* 1971). As a part of the detailed works, twenty five drillholes have been made in the late 70s, of the depth of 34 to 980 m (the total length of the boreholes was ca. 2,900 m). These works have been accompanied by complex geophysical, geochemical, mineralogical, petrographic and metallogenic studies.

Similar prospecting works but of distinctly smaller extent have been carried recently by the State Geological Institute also at the areas of the other massifs. However, the works have been limited to the geochemical mapping and several drillholes (Sobótka 2a, Nowa Ruda IG-1, Przygorze 1, Przygorze 2) and to elaboration of the geological data obtained.

By the courtesy of the Director and the employees of the State Geological Institute in Warsaw, the author had an opportunity to participate in all these works, in the part connected with studies of the ore mineralization. The author would like to express his sincere thanks for this help. He feels especially indebted to Drs. Jerzy Fedak and Maria Lindner for an access to geological materials and long scientific co-operation during realization of the above studies.

In the present paper the author used his own results obtained both during the above studies and from the studies of the geological materials coming from the surficial sampling of the operating and abandoned quarries, and natural outcrops located in the areas of the mafic and ultramafic rocks occurrence around the Sowie Mountains block. This sampling has been performed by the author independently of the works carried jointly with the Geological Institute. The surficial sampling has not always given the valuable material because due to weathering of the rocks the ore associations present in them have been strongly altered or even removed in various degree. This caused that the author treats the results of these samples studies only as a completion of the results yielded from the borehole material studies.

In the laboratory studies the author had at his disposal the geological materials from the gabbroic massifs of Ślęża, Nowa Ruda, Braszowice and Brzeźnica and from the serpentinite massifs of Gogółów—Jordanów, Szklary, Grochowa and Przygórze.

For the present report the author used the

results, samples and geological data of his own studies of several years to present a systematic review and on this basis the synthetic picture of the mafic and ultramafic massifs metallogeny in this part of Lower Silesia. The microscope observations of ore minerals in reflected light was the main study method and it was completed, when necessary, by transmitted light microscopy.

Over 700 polished sections and several tens thin sections were the essential material for the ore minerals investigations. The microscope studies have been performed with use of an Amplival pol. u. ore microscope, an OKF-1 reflectivity measuring unit and a MPT-3 microhardness meter to measure the Vickers microhardness of minerals. The microphotographs have been made using a Zeiss MF-matic micrographic unit.

Studies of variations of chemical composition of certain ore minerals and tracing of chromium migration during chromite alterations have been performed with use of a scanning electron microscope and an electron microprobe (Salaciński, Zawadzki 1983) and by the experimental chemical methods (Białowska, Salaciński 1984).

Characteristics of the ore-bearing rocks is presented on the basis of the existing mineralogical-petrographical and geochemical descriptions prepared during the studies carried by Geological Institute (Maciejewski 1963, 1968; Niemczynow 1966; Ciemniewska *et al.* 1981) or published in other works (Szumilas 1963; Kubicz 1966; Harańczyk, Wala 1970; Białowska 1973; Białowska, Salaciński 1977, 1981).

ORE MINERALIZATION IN MAFIC AND ULTRAMAFIC ROCKS

Mafic and ultramafic rocks are important from the economic point of view as the places of concentrations of numerous metals, frequently of commercial importance. World's important ore deposits of chromium, platinum group elements, nickel, cobalt, titanium, vanadium and other metals are connected with ultramafic rocks or products of their alterations. Mafic rocks host first of all the deposits of ilmenite-magnetite ores and liquation-type copper and nickel sulfide ores. In connection with the development of the ore dressing methods one may expect that even dunites would become in future a complex ore of many metals, e.g. Cr, Ni, Co, V, Ti, Pt, if the total tenor of these elements in rocks equals about 1% (*vide* Ciemniewska *et al.* 1981).

Mafic and ultramafic rocks form massifs of

various size in the platform areas or in the margins of the old crystalline gneiss blocks adjacent to deep tectonic disjunctions within geosynclinal zones. The massifs sizes are very variable. After Konstantynowicz (1979), the dimensions of the ultramafic chromite-bearing massif outcrops in the geosynclinal areas range from 1 to 2,000 km², but in the platform areas they reach even several tens thousand km². The Bushveld magmatic complex in South Africa of the lopolite shape about 450 km in length and 250 km in width and consisting mostly of mafic and ultramafic rocks may be such an example. In this context, the Lower Silesian massifs of these rocks, cropping out on the surfaces from several tens to few km² can be regarded as smaller ones.

The enrichment of the discussed rocks in metals has two aspects. Increased clarks of certain metals are caused by their higher content in rock common forming minerals, other metals can form their own minerals, frequently concentrated in lithologically or tectonically isolated parts of massifs.

From the economic point of view, the second occurrence mode has the significant importance, whereas the first occurrence form is important only in this case, if the further rock or mineral alterations cause the release and mobilization of the dispersed metals; the metals may next secondarily concentrate as, for instance, in the nickel silicate deposits.

Concentrations of the above listed metals in mafic and ultramafic rocks are very variable (Tab. 1). Iron is the most abundant metal in the

both rock groups. During the magmatic processes iron enters first ore minerals formed by immiscibility and next at the stage of the early and main crystallization. It precipitates in form of magnetite, ilmenite, and subordinate chromite and sulfides. Its significant part is also bound in the rock forming silicates: olivines, pyroxenes and amphiboles. Local chromite or Fe-Ti ore accumulations of the economic deposit nature in ultramafic and mafic rocks (dunites, peridotites, gabbros) are geochemically important iron concentrations achieving 60 wt. % (Polanski 1974).

The other listed metals have various geochemical characteristics and they may be divided in two groups (Tab. 1). One group: Cr, Ni, Co and platinum group elements, concentrates mostly in ultramafic rocks, but the other group: Ti, V and Cu, are connected with mafic rocks.

Table 1. Content of certain metals in mafic and ultramafic rocks, in weight percent
Zawartość niektórych metali (w % wag.) w mafitach i ultramafitach

Clarks (klarki)	Elements (pierwiastki)							
	Fe	Ti	V	Cr	Ni	Cu	Co	Pt
For lithosphere* (dla litosfery)	5.10	0.57	0.0135	0.01	0.0075	0.0055	0.0025	5·10 ⁻⁷
For mafic rocks** (dla mafitów)	8.0	0.7	0.0160	0.017	0.016	0.090	0.005	2·10 ⁻⁶
For ultramafic rocks** (dla ultramafitów)	9.8	0.04	0.0020	0.3	0.14	0.00005	0.011	5·10 ⁻⁶
Concentration ranges in magmatic deposits* (zawartości brzeżne w złożach magmowych)	>40-45	>0.9-3	0.28	>27	>0.2-0.5	>0.3-0.7	>0.2-0.3	>2·10 ⁻⁴

*After (wg) Konstantynowicz (1979).

**After (wg) Polanski (1987).

Chromium, the most common metal of the first group, is bound essentially in chromite. Its local concentrations in ultramafic rocks, as well as in serpentinites being the product of these rock alteration, are the main ores of this metal. Ultramafic rocks (dunites) may bear 2000–2800 ppm Cr (Stueber, Goles 1967 – *vide* Ciemniowska *et al.* 1981), whereas mafic rocks contain 200 ppm Cr (Vinogradov 1962). A part of the chromium present enters rock forming silicates, especially pyroxenes and chlorites, and minerals with spinel structure.

Spinel is a mineral with variable chemical composition and they form several isomorphic series with general formula $R^{2+}R_2^{3+}O_4$; the bivalent cation sites (R^{2+}) may be occupied by Mg, Fe^{2+} , Zn, Mn, and Ni, and the trivalent cation sites (R^{3+}) – by Al, Fe^{3+} and Cr. Also maghemite ($\gamma\text{-Fe}_2\text{O}_3$) and ulvöspinel (Fe_2TiO_4) have the spinel structure. Maghemite is cation-deficient, and ulvöspinel has typically the replacement

$2Fe^{3+} \rightleftharpoons Fe^{2+} + Ti^{4+}$ (Deer *et al.* 1962). Many spinels containing also other cations have been synthesized, namely bivalent Co, Cd, Cu or Mn and trivalent V, Sn, Ga, In, Co, Mn or Ge (*op. cit.*).

The structure of the spinel group minerals has been first studied by Bragg (1915) who have proved that the elementary cell of these minerals contains 32 oxygen anions and 24 cations, eight of them with coordination number 4 (the position A) and the remaining sixteen – with coordination number 6 (the position B). Barth and Posnjak (1932) showed on the basis of the X-ray radiographic studies, that two types of the crystal structure can be detected in the spinel group. The structural types, differing in the cation distribution between the A and B sites, are called the normal spinel structure and the inverse spinel structure. The normal spinel structure has been detected e.g. in proper spinel ($MgAl_2O_4$), gahnite ($ZnAl_2O_4$), hercynite ($FeAl_2O_4$) and galaxite ($MnAl_2O_4$), the

inverse spinel structure — in magnesioferrite (MgFe_2O_4), ulvöspinel (Fe_2TiO_4) and magnetite (FeFe_2O_4).

The pure end members of the spinel group occur rarely as natural minerals. The varieties are distinguished and named on the basis of the

prevailing R^{2+} and/or R^{3+} . This situation caused the existence of numerous classifications of the spinel group minerals. According to a general scheme, three series in the spinel group have been distinguished as aluminum, iron or chromium prevails in the trivalent cation site (Tab. 2). In

Table 2. Systematics of the spinel group
Podział grupy spineli

		Series (seria)		
		spinel (spinelu właściwego) Al^{+3}	magnetite (magnetytu) Fe^{+3}	chromite (chromitu) Cr^{+3}
Trivalent cation (kation trojwartościowy)				
Bivalent cation (kation dwuwartościowy)				
	Mg^{+2}	spinel (pleonaste) (spinel właściwy)	magnesioferrite (magnezjoferryt)	magnesiochromite (magnezjochromit)
	Fe^{+2}	hercynite (hercynit)	magnetite (magnetyt)	chromite (chromit)
	Zn^{+2}	gahnite (gahnit)	franklinite (franklinit)	
	Mn^{+2}	galaxite (galaksyt)	jacobsite (jakobsyt)	
	Ni^{+2}		trevorite (treworyt)	

natural minerals within each series a more or less complete miscibility exists, however between the series the miscibility is distinctly limited and it depends strongly on temperature. Experimental studies of the system FeAl_2O_4 (hercynite) — FeFe_2O_4 (magnetite) performed by Turnock (1959, *vide* Deer *et al.* 1962) proved the continuous miscibility above 858°C , but at lower temperatures a miscibility gap separates two phases of these solid solutions.

Other classifications are more detailed and concern the distinguishing of members of an intermediate chemical composition within one or several solid solution series. Examples of such classifications of various mineral phases bearing chromium are presented in Figure 2a-d.

Nickel, due to its sulfophile properties, may occur in early-magmatic products as sulfide — pentlandite $(\text{Ni, Fe})_9\text{S}_8$, which forms inclusions in pyrrhotite or occurs separately. Simultaneously nickel displays a crystallochemical affinity to magnesium and iron, resulting in diadochy in early-magmatic magnesium-iron silicates, especially in olivines (Polański 1974). Minerals of the ultramafic rocks may contain nickel in the following concentrations (in ppm): olivine 4000, pyroxene 1500, amphibole 700, serpentine 2500, magnetite 3000 (*vide* Ciemnińska *et al.* 1981). Generally ultramafic rocks may contain nickel in concentration 1000–3000 ppm (Waleńczak 1962). Geochemically this metal is similar to chromium and there exists a strict correlation in occurrences of

these two elements (Turekian, Carr 1960 *vide* Ciemnińska *et al.* 1981).

Cobalt, similarly to nickel, has a distinct geochemical affinity to Fe^{2+} and Mg^{2+} and thus it may enter structures of minerals of these elements. Practically cobalt minerals are absent in ultramafic rocks and its total amount occurs dispersed in iron-magnesium rock-forming minerals. After Vinogradov (1962) ultramafic rocks may contain up to 200 ppm Co.

Platinum group elements concentrations in magmatic rocks are extremely low. Their relatively higher amounts like the above discussed metals can be found in the early magmatic crystallization products: dunites, peridotites and pyroxenites. They may form in these rocks their own minerals: alloys, intermetallic compounds or compounds with sulfur, arsenic, tellurium, bismuth or antimony. They may also be admixtures in some rock-forming minerals, and especially in chromite and chromium spinel. Platinum group elements, mainly platinum and palladium may as well occur in early-magmatic pyrrhotite as isomorphous admixtures. Such differentiated occurrence of forms of platinum group metals in ultramafic rocks cause large variations of their contents.

Titanium occurs in highest amounts in mafic rocks amongst magmatic rocks, mostly as ilmenite. Ilmenite forms monomineral aggregates and small exsolution lamellae in magnetite. The latter called titanomagnetite or ilmenomagnetite, comprises sometimes great deposits in mafic rocks,

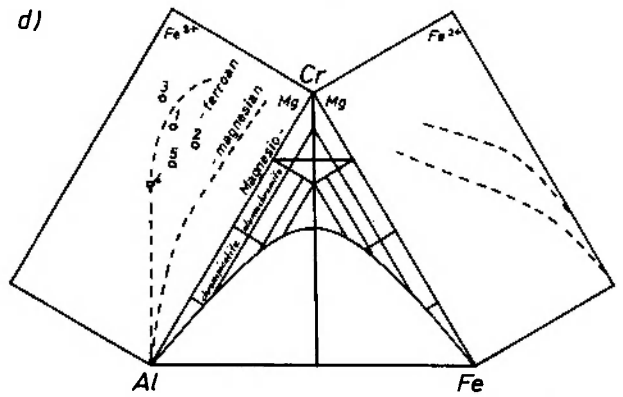
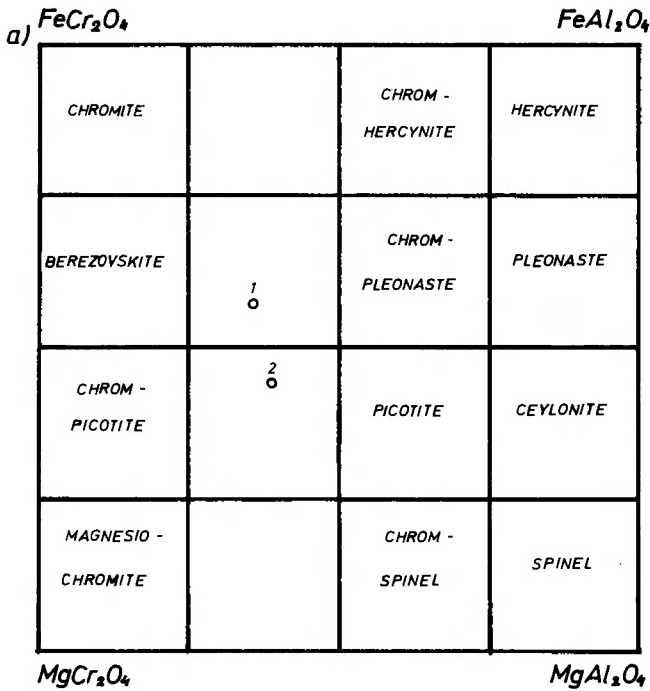
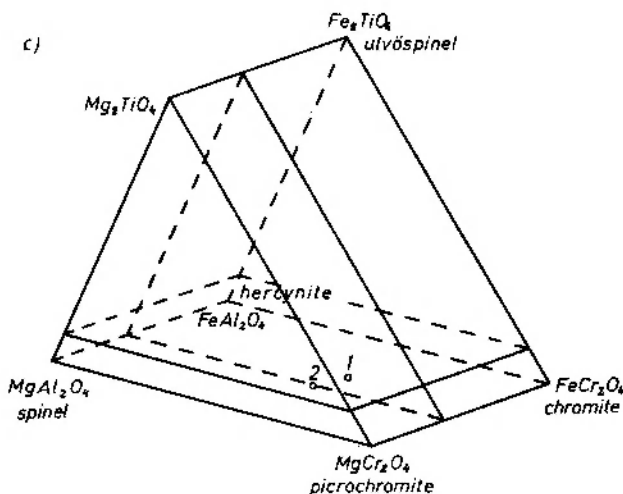
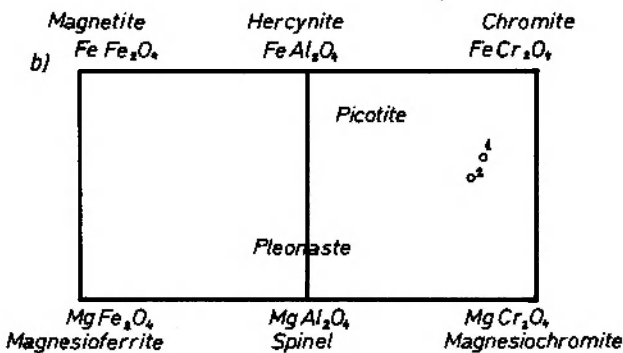


Fig. 2. Projection of the analyses of the chromite ore chemical composition from the Tąpadła deposit according to various classification schemes: a – after Winchell and Winchell (1951); b – after Deer *et al.* (1962); c – after Haggerty (*vide* Frondel 1975); d – after Pavlov (*vide* Smirnov 1986); sample numbers as in Table 3

Projekcja analiz składu chemicznego rudy chromitowej ze złoża w Tąpadłach na różne schematy klasyfikacyjne: a – wg Winchell i Winchella (1953); b – wg Deera *et al.* (1962); c – wg Haggerty'ego (*vide* Frondel 1975), d – wg Pawłowa (*vide* Smirnow 1986); numery próbek wg tabeli 3



being the largest geochemical titanium concentrations in the Earth's crust and simultaneously, a large iron concentrations (Polański 1974). A part of titanium enters dark rock-forming minerals. During pneumatolytic and hydrothermal alterations of the titanium-bearing minerals, this ele-

ment may be released and next it may form secondary rutile (Ramdohr 1975).

Vanadium in mafic rocks is a typical dispersed element, entering the main dark rock-forming minerals: pyroxenes, hornblende, biotite and accessory magnetite. High vanadium content is typical especially of the early-magmatic titanomagnetite deposits: 0.07 to 0.2 wt. %, sometimes even up to 1% (Polański 1974). Supposedly vanadium is present in the magmatic environment as the ion V^{3+} of the ionic radius close to Fe^{3+} , and thus it can substitute iron in various minerals. Significant dispersion of vanadium in the discussed rocks is a reason of absence of its greater concentrations of the economic importance.

Copper is a highly sulfophile element and thus practically it does not enter rock-forming silicates, but it precipitates in form of accessory sulfides. A limited enrichment in copper, up to several tens g per t, was found in plagioclases, hypersthene and olivine. Mafic rocks enrichment in copper is connected with immiscibility processes, resulting in early segregation of sulfide melt from silicate magma, but simultaneously in a paragenetic connection between this sulfide and crystallizing mafic rocks.

The presented geochemical tendencies are the reasons of so significant concentrations of some of the above named elements in mafic or ultramafic rocks, that it leads to origin of the deposits with considerable resources. The well known examples

are of iron deposits of the titanomagnetite formation, Taberg and Otänmeki type, of the copper-nickel liquation deposits in Sudbury, Norylsk, Pechenga and Bushveld, of the chromite deposits

in the Urals, Balkan Peninsula and Bushveld, of the platinum group elements deposits in the Urals, Bushveld and Sudbury.

PRESENT-DAY STATE OF THE RECOGNITION OF ORE MINERALIZATION IN THE SOWIE MOUNTAINS BLOCK SURROUNDING

The discussed above theoretical circumstances have been the basis for initiation of the mentioned geological-prospecting works at the mafic and ultramafic rock outcrops in Lower Silesia by the State Geological Institute. The recognition degree on the metallization of these massifs was very poor before these works started. Most data has been pertinent to the chromite deposit occurrences in the serpentinite massifs of Grochowa and Gogółów—Jordanów, and the weathering nickel silicate ore deposit in the Szklary serpentinite massif. The oldest data concerning the geological prospecting and exploitation of these deposits are from the 19th century (*vide* Birecki 1960a, b, 1962). The first exploitation attempts of chromite have been made in Lower Silesia already from 1824 till 1827 on the hill Twarda Góra near Grochowa. After a longer cessation the prospecting started again in 1866, and allegedly an ore vein 0.5–1.0 m thick and containing 19–40% Cr_2O_3 was found.

In 1877 chromite has been detected on the northern slope of Czarna Góra and since 1890 the exploitation of the nest A, occurring closely to the surface, has been started (Fig. 3). Exploitation has been going on till 1893. During the World War I a shaft has been deepened at Tapadla to the depth of 22 m and it has achieved the nest B, but galleries and an adit did not meet any ore. In 1937 the works, rendering the old shaft accessible, have discovered the nests C and D. Three years later a new shaft of the depth of 26 m has been deepened near the old one, achieving the nests D and E, and galleries leading from the new shaft have got the agglomerations F and G. In 1941 three further small nests H-1, H-2 and H-3, and the nest J have been detected by galleries and the nest K — by drilling holes. The works carried till 1945 did not reveal new nests. Birecki (1962) writes, that about 4000 t of chromite ore has been exploited from this deposit.

After the war, the geological, prospecting and documenting works, carried already since 1946, by the industrial surveys and the State Geological

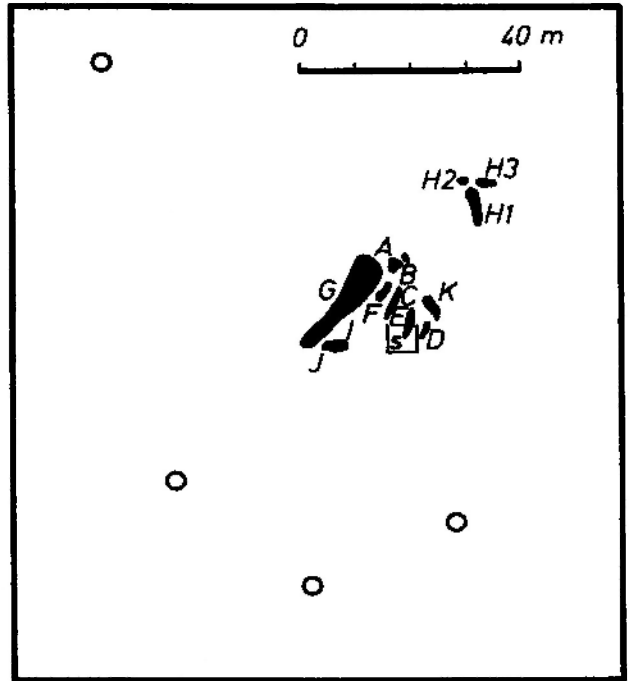


Fig. 3. Ore nests distribution (A–K) in the Tapadla chromite deposit; circles indicate boreholes made in the deposit vicinity (after Birecki 1960a, with the author's completions)

Rozmieszczenie gniazd rudnych (A–K) w złożu chromitu w Tapadłach; kółkami oznaczono wiercenia wykonane w pobliżu złoża (wg Bireckiego 1960a z uzupełnieniami autora)

Institute, have been pertinent to the nickel silicate ores and magnesite deposits. They caused the discovery and documentation of a number of ore zones in the western part of the Gogółów—Jordanów serpentinite massif at the region of Wiry, Gogółów and Shypica (Cierniewska *et al.* 1981). Beside the prospecting works, geochemical, mineralogical and petrographic studies have been performed at various massifs (Waleńczak 1962; Szumlak 1963; Pawlikowski, Salamon 1972; Białowolska 1973). In the early 60s in the then Section of Non-Iron Metal Ore Deposits of the Geological Institute, studies of primary ore mineralizations in mafic and ultramafic rocks of Lower Silesia have been started. These works have been continued over twenty years.

GENERAL GEOLOGICAL CHARACTERISTICS OF THE MAFIC AND ULTRAMAFIC MASSIFS IN THE SOWIE MOUNTAINS BLOCK SURROUNDING

Mafic and ultramafic rocks occur in the Sowie Mountains block surrounding as distinctly separated massifs. Some of them were altered into serpentinites. Such rocks comprise the main or even only component of the largest serpentinite massif in the Fore-Sudetic block called the Gogolów-Jordanów massif, and they form small massifs of Sobótka, Szklary and Grochowa near Żąbkowice Śląskie. In the Sudetic area small outcrops of these rocks are known from the Przygórze region near Nowa Ruda. Gabbros build up two large rock bodies: Ślęza massif in the Fore-Sudetic block and Nowa Ruda massif in the Sudetic area, smaller massifs in the Braszowice and Brzeźnica region and several small ones, from which there was not possible to obtain the fresh rock samples, and for this reason the latter have been omitted in further considerations. The author also did not studied diabases occurring with gabbros, because

he was interested only in plutonic rocks and their ore mineralization. Such mineralization in the rocks occurring with gabbros or serpentinites has been in the past and it is presently the object of interest of other scientists (Pawlikowski, Salamon 1972; Niskiewicz 1974).

Oberc (1960, 1972) considers the above named rocks to be of the Late Precambrian age; serpentinites are the oldest of them and gabbros and diabases are younger.

The most interesting materials have been obtained by the author from the gabbroic massifs of Ślęza and Nowa Ruda and from the serpentinite massifs of Jordanów–Gogolów and Przygórze, and thus the most attention has been paid to these massifs.

The gabbroic Ślęza massif forms a distinct culmination of 718 m a.s.l. in the landscape. Its outcrop has an oval shape and the surface of ca.

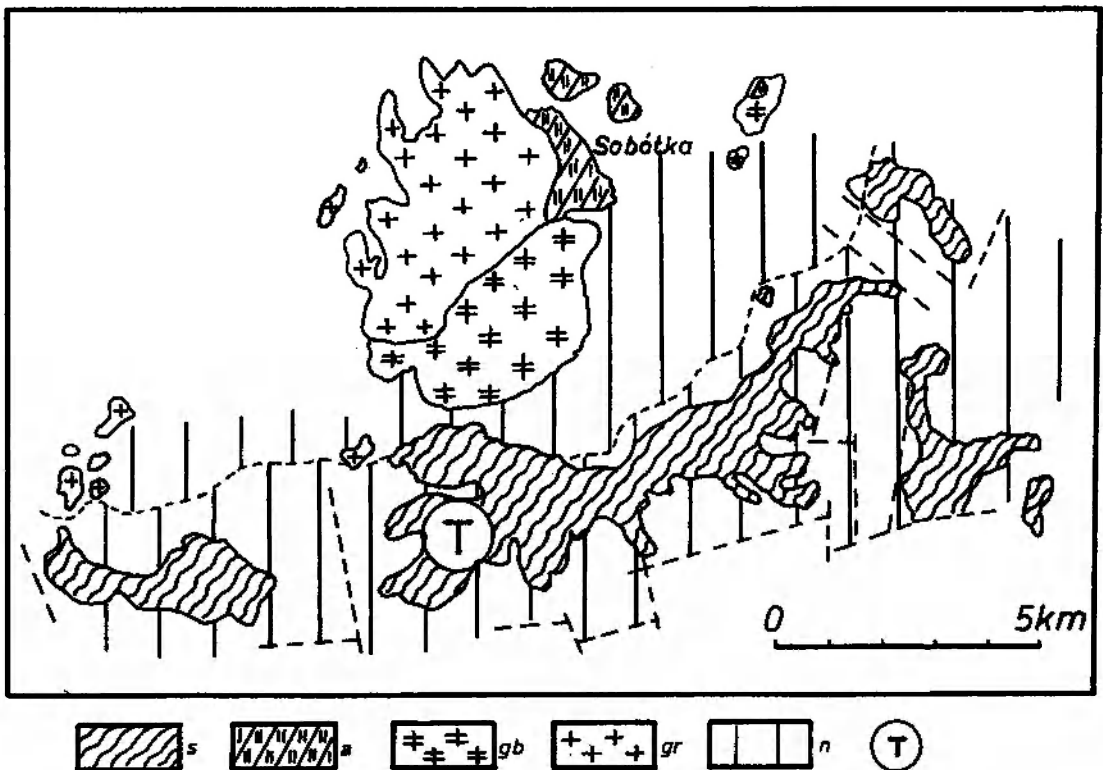


Fig. 4. Geological sketch map of the Gogolów–Jordanów and Ślęza massifs (simplified – after the *Detailed geological map of the Sudetes 1:25,000, sheet Sobótka* – Gaździk 1969 and *sheet Jordanów Śląski* – Trepka 1965): *s* – serpentinites; *a* – amphibolites; *gb* – gabbro; *gr* – granite; *n* – intrusive rocks covered with Tertiary and Quaternary deposits; *T* – investigation area in the Tąpadła deposit region

Szkic geologiczny masywów Gogolów–Jordanów i Ślęzy (uproszczony – wg *Szczegółowej mapy geologicznej Sudetów 1:25 000 ark. Sobótka* – Gaździk 1969 i *ark. Jordanów Śl.* – Trepka 1965): *s* – serpentynity; *a* – amfibolity; *gb* – gabro; *gr* – granit; *n* – skały intruzywne pod nakładem utworów trzeciorzędowych i czwartorzędowych; *T* – obszar badań w rejonie złoża w Tąpadłach

15 km². Serpentinities of the Gogolów – Jordanów massif and the small Sobótka massif, amphibolites and Strzegom granites occur in its vicinity (Fig. 4). The Słęża massif is partly covered by Tertiary deposits, which conceal its true shape and extension at a greater depth. From the west the gabbros are underlaid in a certain area by Strzegom granite. It has been evidenced by the above mentioned borehole Sobótka 2a near the village Sady, in the SW part of the massif (Fig. 4). More detailed characteristics of the massif rocks was given in the publications by Białowolska (1973) and Majerowicz (1979).

The Nowa Ruda massif is characterized by large and distinct differentiation, resulting in origin of such rocks as peridotites, pyroxenites, troctolites, olivine gabbros, anorthite gabbros, diallage gabbros, gabbro-anorthosites and anorthosites. Ultramafic and anorthosite rocks occur as schlieren within other rock varieties. There is no any regularity found in spatial distribution and in transitions from the olivine-bearing varieties to the pyroxene-feldspar ones. On the surface only a distinct prevalence of troctolites and anorthite gabbros in the NE part of the massif and diallage and olivine gabbros in the SW and S parts have been observed. The studies made by Borkowska (1985) have shown that gabbroic rocks of the Nowa Ruda massif crystallized at relatively low depth and their crystallization temperature was in the ranges of 860–1260°C. To the south from the line Slupiec–Dzikowiec the gabbros alter in diabases with decreasing crystal size, what indicates that to the southwest the intrusion achieved shallower zones (Fig. 5). As it was found in the mine pits of the Slupiec mining field of the Nowa Ruda mine in the SE massif, those diabases pass gradually with the increasing depth into typical diallage gabbros (Kowarz 1965; Bzowski *et al.* 1983). This indicates the genetic connections and similar age of those rocks (Wiewióra 1967).

The petrographic variability of the massif rocks in the vertical profile has been shown also by the borehole Nowa Ruda IG-1 made in 1965–1966 in the northern part of the gabbro massif near the village Wolibórz. This borehole has achieved the depth of 500 m. On the basis of the lithological profile described by Fedak and Kowarz (1967), the percentage of the main rock varieties in the drilling core has been as follows:

anorthite gabbro	– 43%,
olivine gabbro	– 37%,
troctolite	– 12%,
anorthosite	– 3%.

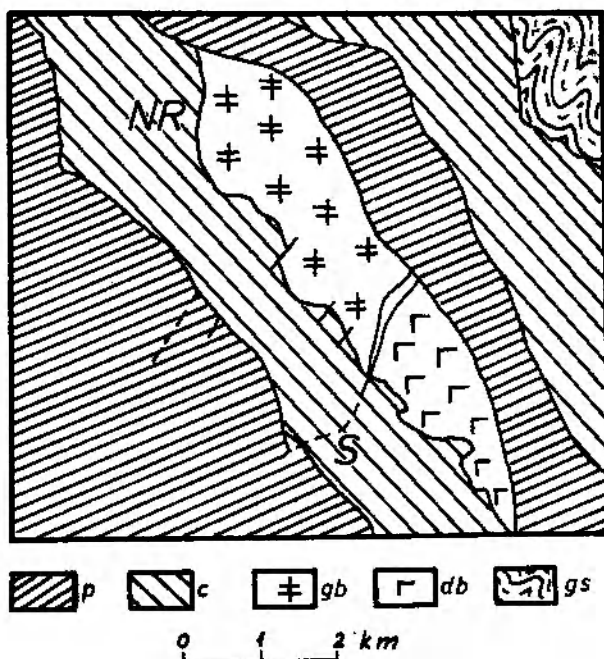


Fig. 5. Geological sketch map of the Nowa Ruda massif (simplified — after Bzowski *et al.* 1983 and the *Detailed geological map of Sudetes 1:25,000, sheet Nowa Ruda – Oberc, Wójcik 1965*); NR — the town of Nowa Ruda; S — the town of Slupiec; p — the Rotliegende deposits; c — the Carboniferous deposits; gb — gabbros; db — diabases; gs — gneisses

Szkiec geologiczny masywu Nowej Rudy (uproszczony — wg Bzowskiego *et al.* 1983 i *Szczegółowej mapy geologicznej Sudetów 1:25 000, ark. Nowa Ruda – Oberc, Wójcik 1965*); NR — miasto Nowa Ruda; S — miejscowość Slupiec; p — utwory czerwonego spągowca; c — utwory karbonu; gb — gabra; db — diabazy; gs — gnejsy

Moreover the quoted authors have stated the occurrence of small amounts of olivinites, dunites, pyroxenites and serpentinites. The main petrographic varieties occur in the drilling core profile zonally. Anorthite gabbros with anorthosites have been found at the depth ranges of 0–40 m, 248–418 m and 453–500 m, and in the remaining intervals there occur olivine gabbros and troctolites. Segregation of these two rock groups is relatively complete and only at the boundaries of the above defined zones their alternation has been noted.

An interesting petrographic variety, gabbro pegmatites, has been found in the borehole and on the surface (Białowolska, Salaciński 1981). Such rocks consist mainly of pyroxenes and they form in the borehole profile small nests, and their varieties built up of plagioclase, serpentine and subordinate pyroxene, occur as small veins and lenses. On the massif surface pegmatites can be observed that comprise plagioclase pierced by individual acicular crystals of green amphibole.

Between the Nowa Ruda gabbro massif and

Sowie Mountains cataclasites, a small serpentinite massif occurs in the Przygórze region, with the NW–SE elongation. It outcrops on the surface of the length of 700 m. Two drilling holes made in 1964 showed that the central massif part is built of serpentinites with numerous fault polishes and calcite and chrysotile veinlets (borehole P-2), but in the marginal massif zone (borehole P-1) slightly altered mafic and ultramafic rocks occur together with serpentinites. Among the non-serpentinite rocks olivine, anorthite and diallage gabbros have been found (Fedak, Kowarz 1967).

The massif Gogolów–Jordanów is the largest serpentinite massif in Lower Silesia (Fig. 4). Its elongated outcrop extends along the northern margin of the Sowie Mountains block, running ca. 20 km approximately evenly with a parallel of latitude from Gogolów in the west to the Jordanów Śląski vicinity in the east. Its part recognized geologically occupies the area about 100 km² with croppings of ca. 30 km². Along the whole southern border the serpentinites contact tectonically with the Sowie Mountains gneisses. Numerous faults, transverse to the contact and with the horizontal shift achieving 3 km, cause that the southern massif border has a zig-zag shape. From the west and northwest the serpentinites contact with the Strzegom granite, from the north – with the Ślęza massif gabbros and from the east they are surrounded by gneisses and fyllite schists.

The Gogolów–Jordanów massif consist mostly of the serpentinitized ultramafic rocks. Microscope studies revealed various degrees of the ultramafic rocks alteration, because together with serpentinites there occur the primary rocks poorly and moderately altered (Niemczynow 1966; Maciejewski 1968). On the basis of the detailed petrographic studies of the borehole material there have been distinguished: dunites, wehrlites, serpentinites with olivine relics, serpentinites, serpentinites with carbonates, gabbros, kersantites, rodingites and talc-carbonate, talc-serpentine, talc-chlorite, chlorite-carbonate, quartz-feldspar and biotite-carbonate rocks (Ciemniewska *et al.* 1981).

The rocks are strongly fractured. Oberc (1972) after Chmura and Sulkowski (1965), and Gajewski (1970), is of the opinion that there occur two essential shearing fracture directions: NW–SE and SW–NE, whose junctions are the directions running evenly with a parallel of latitude. These fractures are older than those in the neighbouring

Strzegom granite. They have been filled by various vein associations. According to Szatkowski (1982), four main directions of the tectonic disjunctions may be distinguished: N–S, W–E, NW–SE and NE–SW; the lineament directions determined by the method of the contour lines compaction and on the basis of the aerial photographs interpretation approximately coincide. It concerns especially the lineaments of the NE–SW and W–E directions. Distinct coincidence has been also found between the rock fracturing and lineament directions. All the above-listed tectonic directions, marked as lineaments, cross in the vicinity of the chromite deposit at Tąpadła (Fig. 6). Thus

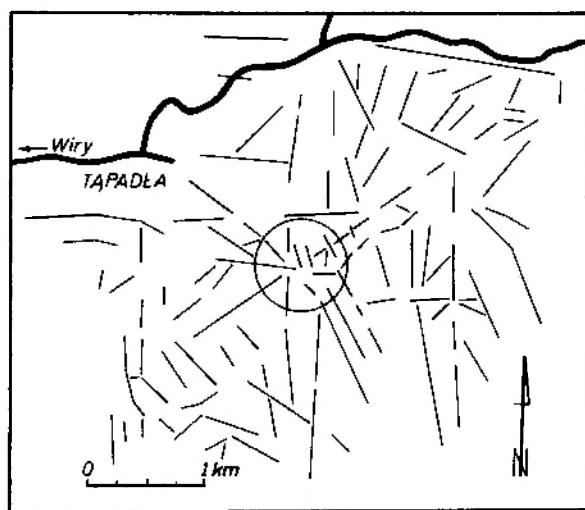


Fig. 6. Sketch of the lineaments in the Tąpadła vicinity, performed on the basis of the aerial photographs interpretation (after Szatkowski 1982); the circle marks the deposit region at Tąpadła

Szkic lineamentów okolic Tąpadł wykonany na podstawie interpretacji zdjęć lotniczych (wg Szatkowskiego 1982); kołem oznaczono rejon złoża w Tąpadlach

it is possible that the chromite accumulations occurrences are connected with zones of a thickening of various fracture systems.

The serpentinite Szklary massif from the eastern Sowie Mountains block surrounding, occurring within the Niemcza lineament, has not been included in the present paper, because the author was not able to obtain from its area representative samples of fresh rocks. This massif forms a distinct height about 5 km long and elongated meridionally. The massif rocks have been submitted to intensive weathering processes in Tertiary, what resulted in formation of thick weathering waste layer bearing nickel silicate ore deposits.

ORE MINERALIZATION

NOWA RUDA GABBRO MASSIF

From among the gabbro massifs the most abundant and quantitatively most variable ore parageneses occur in the Nowa Ruda gabbros. The author has analysed the ore mineralization found there both in the surface and drilling hole Nowa Ruda IG-1 specimens. Fedak and Kowarz (1967) have ascertained almost in the whole profile of the drilling core the presence of the macroscopic dispersed sulfide mineralization. They have observed the increase of the ore minerals amount in troctolites and olivine gabbros and their decrease in anorthite gabbros. They have determined: ilmenite, titanomagnetite, magnetite, pyrite, pyrrhotite, chalcopyrite, pentlandite, marcasite and spinel group minerals.

The present author, when sampling in 1974 the drilling core for his ore studies, has got an access to the incomplete core and most probably for this reason he observed the macroscopic ores only occasionally.

By the microscope studies of the drilling core profile the author has found the following minerals: chromite, ilmenite, rutile, leucosene, pyrrhotite, pentlandite, magnetite, chalcopyrite, sphalerite, bornite, pyrite, marcasite and the limonite group minerals. Chromite, ilmenite, magnetite and pyrite occur most frequently.

Chromite appears mainly in troctolite and olivine gabbro, but in anorthite gabbro and in anorthosite it is found rather rarely. It forms isometric grains of 0.0n to 2.0 mm in size, usually strongly cracked (Pl. I, 1). Chemical composition of chromite from the borehole Nowa Ruda IG-1 is variable. That occurring in troctolites and olivine-pyroxene gabbros bears exsolutions (Pl. I, 2) and isometric and acicular ilmenite inclusions (Pl. I, 3), what indicates a primary titanium admixture and the ulvöspinel molecule presence in chromite. The identification of the discussed mineral as chromite is based on its optical features determined in the ore microscope and very high microhardness ranging from 1023 to 1434 kG/mm², what agrees with the published data (Uytenbogaardt, Burke 1971; Isaenko *et al.* 1972). In other rock varieties there occurs chromite distinctly enriched in iron, what results in the magnetite rims origin (*vide* Ramdohr 1975), which may gradually thicken along grain contacts and fractures, replacing chromite (Pl. I, 4). Neither exsolutions nor ilmenite inclusions and its decomposition products

have been found in this chromite. High iron content in chromite is also indicated by the absence of the internal reflections (Uytenbogaardt, Burke 1971).

The mode of occurrence and the optical features of chromite are identical with chrome-magnetite described by Kowarz (1965) from this region. The present author however considers the name chromite as more adequate to the above listed features of this mineral, and the reference data indicate that a significant enrichment in iron is possible within the frames of the chemical formula of chromite (Fe, Mg) (Cr, Al, Fe)₂O₄ (Deer *et al.* 1962).

Chromite grains are subhedral, oval, roundish and only sometimes anhedral (Pl. II, 1), what proves their coeval crystallization from magma with other mineral components of the host rock and sometimes directly after them. Later chromite has been strongly cataclased, and the fractures formed have been filled by serpentine (Pl. I, 1) or rarer by sulfides (Pl. II, 2) like the fractures observed in olivine.

Ilmenite occurs almost in the whole borehole profile, but its amount and habit are very variable. Ilmenite I is of magmatic origin and it forms oval (Pl. II, 3) pale gray grains with brownish tint of the size of 0.2–1.2 mm, with strong anisotropy and with hardness ranging from 734–824 kG/mm². It is relatively rare, occurring mainly in troctolites and olivine gabbros. It crystallized probably contemporaneously with chromite. Ilmenite II forms finely dispersed exsolutions (Pl. I, 2) and lathy, isometric-oval or square segregations in chromite (Pl. I, 3, Pl. II, 1, 2). It bears titanium originally linked in the chromite structure. Ilmenite II crystallized after the chromite and ilmenite I crystallization had finished. Ilmenite III occurs as acicular microlites of the length of several hundredths of millimeter and of the thickness of few μm, and arranged according to the pyroxene cleavage (Pl. II, 4). It is of metasomatic origin and bears titanium coming from the primary diadochic admixture in pyroxenes.

All the three ilmenite generations are frequently replaced by rutile (Pl. II, 3), which forms pseudomorphs after ilmenite with strong, colourless or yellow internal reflections. It is light-gray, somewhat lighter than ilmenite. Moreover, ilmenite I is sometimes replaced by fine-grained leucosene aggregate with a distinct birefractivity and anisotropy, and very low reflectivity. Larger grains

of ilmenite I and II are sometimes cracked and the fissures, similarly as in chromite, are filled by serpentine (Pl. II, 3) and sometimes by later post-pyrrhotite pyrite.

Pyrrhotite is a distinctly later mineral than chromite and ilmenite. It forms grainy aggregates 0.5 to 1.0 mm in size, in a large part replaced by pyrite and marcasite (Pl. III, 1) as well as by magnetite. The mode of replacement may be various. Pyrite and marcasite replace pyrrhotite corroding it from the grain margins and forming rims of variable thickness either with the pyrrhotite relics preservation inside the aggregate or resulting in the complete replacement (Pl. III, 3). In such pseudomorphs pyrite prevails, forming cavernous aggregates with the preserved post-pyrrhotite cleavage, but marcasite comprises only fragments of these pseudomorphs (Pl. III, 1). Replacement of pyrrhotite by magnetite develops also zonally or it appears as crossing laths of Fe_3O_4 , arranged according to the pyrrhotite cleavage directions (Pl. III, 3,4). Pyrrhotite may be replaced by all three above mentioned minerals, or only by pyrite and marcasite. Post-pyrrhotite pseudomorphs consisting exclusively of magnetite have not been found.

Fresh pyrrhotite aggregates have yellow colour with brownish tint and they display strong anisotropy. Using crossed polaroids one may frequently observe intergrowths of the monoclinic and hexagonal pyrrhotite. Sometimes also in plain polarized light these both modifications differ in colour (*vide* Ramdohr 1975).

Pentlandite, frequently described from the Nowa Ruda gabbros in the past (Kowarz 1965), occurred occasionally in the specimens investigated by the present author. It has been found sometimes in the pyrrhotite aggregates as small (0.5 mm), cracked, isotropic grains of pale-yellow colour (Pl. III, 4). Its identification is not, however, completely sure due to its minute size and rare occurrence.

Magnetite occurs in three generations. Magnetite I forms rims around chromite (Pl. I, 4). Magnetite II has been found as the above described rims and laths in pyrrhotite and in pyrite-marcasite post-pyrrhotite pseudomorphs (Pl. III, 2, 3, 4). Magnetite III comprises tiny grains (0.01 mm in size) in a rock, arranged in chains few millimeters long, in pseudomorphs after olivine (Pl. IV, 1). All the three magnetite generations formed due to alterations of earlier existing minerals. It is possible however that certain magnetite clusters, included to the above named generations,

may be in fact directly of magmatic origin, but in course of later alterations the magnetite became similar to the described forms.

Chalcopyrite is relatively rare in the studied samples. It occurs as a subordinate component in the pyrrhotite aggregates and in the pyrite-marcasite post-pyrrhotite pseudomorphs, not altering in the disulfidisation process. It forms also small grains (0.5 mm) and microlites dispersed in the rocks, more common in the anorthite gabbros than in other varieties. Occasionally it intergrows pyrrhotite as thin parallel laths (Pl. IV, 2). It has yellow colour, and when compared with pyrrhotite – with greenish tint, displaying anisotropy. Its microhardness is about 245 kG/mm². Its origin is not clear, although the anisotropy indicates a high crystallization temperature. It could thus crystallize either together with pyrrhotite and pentlandite due to sulfide immiscibility or it might be of hydrothermal origin.

Sphalerite has been observed only in one preparation, in which it has formed small isotropic grain of gray colour at the edge of an aggregate of post-pyrrhotite pyrite and chalcopyrite (Pl. IV, 3). The absence of the internal reflections in sphalerite may indicate an appreciable iron admixture resulting from its high-temperature origin.

Bornite occurs also very rarely and it replaces chalcopyrite in its certain clusters (Pl. IV, 4). It has pink colour and displays no anisotropy. Its presence has been found only in specimens from the deepest parts of the borehole (400–500 m). Bornite crystallized probably during the stage of the hydrothermal alterations of chalcopyrite (*vide* Białowska, Salaciński 1977).

Pyrite is beside chromite the most frequently found ore mineral and it forms very variable occurrence habits. Pyrite I is a component of the post-pyrrhotite pseudomorphs (Pl. III, 1, 3). It has a cavernous texture. It is strongly fractured, and its shape and size depend on the original pyrrhotite clusters forms. It is almost always associated with magnetite and sometimes with marcasite. It achieves size of tenths of a mm to several mm. Pyrite II forms euhedral and anhedral metacrysts of various dimensions. It is of metasomatic origin and bears iron released from decomposed mafic minerals. This process starts from development of thin intergranular veins and dendrites filling cracks in rock-forming minerals (Pl. V, 1, 2), which are next corroded and replaced along contacts, especially at the veinlet crossings. This way isometric pyrite aggregates form first, and next during replacing of rock-forming minerals it

shows a tendency to develop euhedral habit of the pentagonal dodecahedron (Pl. V, 3). Later this crystallographic habit transforms into cube (Pl. VII, 2). The euhedral feature is more distinct in smaller crystals, however the larger ones commonly bear inclusions of rock-forming minerals and they have undulatory, corrosive contacts with non-ore minerals. Microlites and pyrite pigment, abundant in certain samples, also should be included to the pyrite II. Pyrite III is rare and occurs as veinlets 0.1 mm thick and several mm long (Pl. V, 4), probably of hydrothermal origin. Sometimes pyrite III is associated with fine-grained marcasite.

Marcasite does not occur independently, but always it is a minor associate of pyrite I in post-pyrrhotite pseudomorphs (Pl. III, 1) and of pyrite III in veinlets. It forms fine-crystalline, strongly anisotropic aggregates of light-yellow colour, a little lighter than pyrite.

Limonite group minerals replace marcasite and pyrite in post-pyrrhotite pseudomorphs, corroding these sulfides along contacts and fractures. Observed zonal change of colour and reflectivity in limonite clusters proves that different mineralogical varieties of iron hydrous oxides and hydroxides are here present, however, their exact identification is not possible due to their too small sizes.

Similarly interesting ore mineralization signs have been found on the massif surface. Hydrothermal ore mineralization has been described by Pawlikowski and Salamon (1972) from the diabase quarry at Slupiec. These authors have given a mineralogical characteristics of two ore veins. One vein consisted of chalcopryrite, chalcocite, digenite, covellite, limonite, goethite, malachite and azurite. These minerals occurred with calcite. Another vein was filled mainly with calcite with pyrite and chalcocite.

In the southern part of the Nowa Ruda gabbro massif, the Nowy Dzikowiec region, the present author in co-operation with Białowolska (1977) stated the occurrence of ore minerals in a quartz vein, cropping out at the northeastern slope of Banach Hill. This vein cuts the diagenetic gabbro, which is strongly altered due to the hydrothermal solutions activity. Hydrothermal processes caused the enrichment of the rock in magnesium, nickel and chromium and decrease of contents of calcium, strontium and aluminum. These processes led also to the filling of fissures in the gabbro with vein minerals. The essential part of the vein filling comprises quartz in three generations and an ore

mineral assemblage, spatially and genetically connected with the second quartz generation. Chalcopryrite, pyrite and probably digenite are the primary minerals formed by crystallization from hydrothermal solutions. These minerals can be observed only as relics. The remaining ore minerals: bornite, chalcocite, covellite, malachite and hydrous iron oxides, appeared due to the primary mineral replacement under hydrothermal and hypogene conditions.

Concentrations of the iron-titanium oxides were found in gabbros occurring in the underground pits of the Nowa Ruda mine, western field Slupiec (Bzowski *et al.* 1983). Among the ore minerals, titanomagnetite with the exsolution structures of the solid solutions magnetite-ilmenite and magnetite-ulvöspinel, several generations of ilmenite, hematite and titanite were stated. Moreover, rare sulfides of the paragenesis pyrrhotite-chalcopryrite-sphalerite were observed in pyroxenes and ilmenite.

ŚLĘŻA GABBRO MASSIF

Signs of the ore mineralization in gabbros of the Ślęża massif were analysed on the basis of the specimens from the core of the borehole Sobótka 2a by the present author. This borehole of the depth of 230 m was located at the southwestern slope of the Ślęża Mountain, about 1 km to the east from the village of Sady. In the borehole cross section down to the depth of 198.1 m gabbro occurs, sometimes with macroscopically visible sulfide clusters and quartz veinlets. Below this depth a fine- or middle-grained granite was found.

In the borehole profile ilmenite, rutile, leucocoxene, hematite(?), pyrrhotite, chalcopryrite, pyrite, marcasite and limonite were recognized in gabbro, and in quartz veinlet cutting gabbro at the depth of 169.0 m — molybdenite occurring in appreciable amount was observed. Pyrrhotite, pyrite and marcasite occur in granite underlying gabbro.

Beside pyrite, ilmenite is the most common ore mineral. It forms acicular inclusions arranged parallel to the cleavage in dark minerals and in pseudomorphs formed after these minerals (Pl. VI, 1). The cleavage planes are not so distinct, however, as in pyroxenes of the Nowa Ruda gabbro (Pl. II, 4). Similarly frequently ilmenite occurs as oval, isometric or irregular grains 0.1 to 0.4 mm in size. It has been observed that sometimes such grains are corroded and gradually replaced by rock-forming minerals, with ilmenite relics preserved between these minerals as elongated inclusions

arranged systematically parallel to the ilmenite main crystallographic factors. The both ilmenite varieties are frequently replaced by rutile, and larger grains — also by an aggregate of leucoxene and probable rutile. Ilmenite forming large grains is probably of magmatic origin, whereas that of acicular habit formed, resulting from the titanium and iron mobilization, is contained originally in dark minerals.

Pyrrhotite occurs relatively frequently, especially in the upper part of the borehole profile. It forms oval or irregular grains 0.0n to 1 mm in size, anisotropic, sometimes with intergrowths of the monoclinic and hexagonal modifications visible between crossed polaroids (Pl. VI, 2). It is replaced by a fine-grained aggregate of pyrite and marcasite or by marcasite solely. The latter corrodes pyrrhotite starting from its grain edges and forming rims with very jagged outlines (Pl. VI, 3). Small amounts of chalcopyrite frequently associate with pyrrhotite. Both these minerals are probably of the liquation origin.

Beside the above-mentioned clusters of 0.1–0.2 mm in size, associated with pyrrhotite, chalcopyrite frequently occurs as microlites below 0.05 mm in length and as a pigment dispersed in rock-forming minerals. In this case it was identified due to its anisotropy. Small aggregates of chalcopyrite were also observed in a pyrite veinlet, probably of hydrothermal origin.

Pyrite is the main ore mineral and it displays very variable forms of occurrence. It forms a pigment dispersed in the rock, microlites, grain aggregates and metacrysts of various automorphism degree. Pyrite grain size varies within the limits of 0.00n to 3.0 mm. Smaller grains are usually euhedral, their contacts with rock-forming minerals are straight and sharp, and they do not contain inclusions (Pl. VI, 4). Larger crystals have undulose contacts and contain numerous inclusions and intergrowths of barren minerals (Pl. VII, 1). It has been observed that the metacrysts had initially the habit of the pentagonal dodecahedron (hexagon in the cross-section), and next, developing angles, they altered in the cube (square or triangular cross-sections, Pl. VII, 2).

The above described forms of pyrite are of metasomatic origin. They developed at the places of removed silicates. Moreover, pyrite has been observed in veinlets up to 5 mm long and 0.n mm wide, probably of hydrothermal origin. These veinlets consisted of cube crystal aggregates, associated with rare, irregular chalcopyrite clusters.

As it has been above mentioned, pyrite and

marcasite jointly replace frequently pyrrhotite, forming pseudomorphs and rims around the pyrrhotite relics.

Pyrite is sometimes replaced by limonite developing along cracks and forming colophorm aggregates of variable mineral composition (Pl. VII, 3).

The occurrence of molybdenite in the form of pale-gray, strongly crumpled ribbons (Pl. VII, 4) of very high reflectivity, strong bireflectivity and very strong anisotropy has been observed in a quartz veinlet at the depth of 169.0 m. Molybdenite is the only ore mineral in this veinlet. The veinlet is of the epigenetic origin and it should be connected with a post-magmatic activity developed after the Strzegom granite intrusion. The gabbro surrounding the veinlet contains increased rutile and leucoxene concentrations.

GOGOLÓW-JORDANÓW SERPENTINITE MASSIF

Rocks occurring in the Gogolów-Jordanów ultramafic massif contain numerous signs of ore mineralization, although the number of mineral species is limited. The largest accumulations of ore minerals occur in the chromite deposit at Tapadla, in the western part of the massif. According to Birecki (1960 a, b), the deposit occurs in the summit parts of Czarna Góra hill near Tapadla and it is connected with a dunite intercalation 30–150 m wide and 700 m long, of the NE–SW strike. This zone has been investigated by the mine pits down the depth of 56 m from the Earth's surface. The surrounding rocks are diallage peridotites containing 15–25% of diallage (Birecki *op. cit.*). Accumulations of the chromite ore have most frequently elongated shapes. Immediate surrounding of the ore nests is “the carbonate serpentine, chlorites, carbonates, rarer antigorite, phlogopite, vermiculite and talc...” (Birecki *op. cit.*). Chlorites and carbonates comprise frequently a barren admixture within the ore nests themselves. The mine works have met thirteen ore bodies labelled with the letters A–K. The nest sizes range essentially from a few to about twenty metres. The largest ones had their lengths of 22 metres (the nest A) and 24 metres (the nest G). External ore contacts are sharp (Spangenberg 1943).

Within the ore nests three ore varieties have been distinguished: massive, dotty and racemose. The massive variety, consisting of the coarse-crystalline chromite, is the richest one. Grains of this mineral have diameter of 1 to 3 mm and

they occur in a magnesite-chlorite matrix, which displays between the grains a network of veinlets 0.1 to 1 mm thick. Such veinlets cut also frequently the grains. In thin sections the inner cores of the chromite grains are brown-translucent, but their outer rims are completely opaque. According to Horninger (1941), the opaque chromite would be a transitional phase to magnetite. The author's studies of the ore samples performed with use of the ore microscope did not confirm such a change of the outer rim composition in the chromite grains, which would cause an optical features change. Only inclusions of barren minerals are numerous in these grain parts (Pl. VIII, 1). The dotted ore consists of chromite grains up to 1 mm in size, randomly scattered in chlorite-carbonate matrix. In the racemose ore the chromite occurs as large (up to 5 mm) separate, oval grain aggregates inherent in the chlorite-carbonate matrix. The aggregate arrangement is frequently distinctly ordered, streaked, parallel to the nest elongation. Gradual and sometimes rapid transitions have been observed between the above named ore varieties.

Beside chromite and chlorite-carbonate (magnesite, ankerite, dolomite) matrix, also minor antigorite and thin veinlets of hydrous nickel silicates occur in the ore nests. In any place within the ore nests neither peridotite nor dunite have been found (Birecki 1960 a, b). The latter begins to appear at the rims of the nests and it bears then up to about 5% chromium. The complete ore analyses are shown in the Table 3, and projections of the chemical composition of ores are presented in various classification systems in Figs. 2a–d, assuming the total of cations or oxides comprising chromite as equal 100%.

The position of the chromite from Tapadla has been determined on the basis of the calculations of the quantitative ratios between cations or bivalent and trivalent elements oxides (Fe:Mg, FeO:MgO and Cr:Al, Cr₂O₃:Al₂O₃), respectively from individual analyses (Tab. 4). The calculated ratios have an approximate nature, because the chemical analyses give the ore compositions, with a part of barren minerals inclusively.

From the presented analyses and projections the conclusion appears that the chromite from the Tapadla deposit may be determined as ferrous alumochromite (analyses 1, 3, 5), magnesian alumochromite (analysis 2) and ferrous chrompicotite (analysis 4) according to the Pavlov's classification (Smirnow 1986) or magnesian-aluminous chromite according to the Haggerty's classification (Frondel

Table 3. Chemical composition of the chromite ores from Tapadla, weight percent (after Birecki 1960)

Skład chemiczny rud chromitu w Tapadlach, w % wagowych (wg Bireckiego 1960)

Compo- nents (składniki)	Sample no. (nr próbki)				
	1	2	3	4	5
	Massive ore (ruda zbita)			Dotted ore Racemose (ruda os- ore (ruda powata) groniasta)	
Cr ₂ O ₃	41.23	39.50	38.50	20.11	22.03
Al ₂ O ₃	24.58	25.80	16.21	24.70	20.28
FeO	19.04	13.35	-	-	-
Fe	-	-	16.53	9.44	10.61
MgO	14.77	18.00	2.62	6.80	9.59
CaO	-	-	1.10	0.55	0.90
MnO	0.58	0.42	-	0.02	0.03
SiO ₂	-	1.62	14.70	20.31	19.06
H ₂ O	-	0.75	-	-	-
NiO	-	0.32	-	-	-
Ni	-	-	0.12	0.16	0.09
Pb	-	-	0.00	-	-
Mn	-	-	0.02	-	-
Cu	-	-	0.00	0.02	0.00
Pt	-	-	-	0.00	0.00
Total (razem)	100.20	100.09	89.80	82.11	82.59

Note: Sample no. 1 was analyzed by Traube (1894), no. 2 by Horninger (1940) and 3–5 by Cracow Geological Undertaking of the Smelting Raw Materials (1957).

Uwaga: Próbkę nr 1 była analizowana przez Traubego (1894), nr 2 przez Horningera (1940), a 3–5 przez Krakowskie Przedsiębiorstwo Geologiczne Surowców Hutniczych (1957).

Table 4. Quantitative relations between the main cations and their oxides in chromite from the ore deposit at Tapadla (calculated by the author on the basis of the analyses from the Table 3)

Stosunki ilościowe między głównymi kationami i ich tlenkami w chromicie ze złoża w Tapadlach (wyliczone przez autora na podstawie analiz z tabeli 3)

	Massive ore (ruda zbita)		Dotted ore Racemose (ruda os- ore (ruda powata) groniasta)		
	1*	2	3	4	5
$\frac{Cr_2O_3}{Cr_2O_3 + Al_2O_3}$	0.63	0.61	0.70	0.49	0.52
$\frac{Cr}{Cr + Al}$	0.68	0.66	0.74	0.51	0.58
$\frac{FeO}{FeO + MgO}$	0.56	0.43	-	-	-
$\frac{Fe}{Fe + Mg}$	0.62	0.48	0.86	0.58	0.53

*Sample numbers and analyses as in Table 3 (nr prób i analizy jak w tabeli 3).

1975). On the classification schemes of A. N. Winchell and H. Winchell (1951) as well as on that of Deer *et al.* (1962) the projections fall in the fields without names.

In the late 70s in the neighbourhood of the deposit, twenty five prospecting boreholes have been made, one of them has achieved the depth of 980 m, and the others have had the depth ranging from 34 to 163 m. The boreholes did not revealed new ore zones, but in the samples from the drilling cores a continuous presence of the ore mineralization signs has been stated. The present author has investigated about 400 polished sections prepared from this material. On the basis of the performed studies one can state that the prevailing majority of ores either occurs in serpentinites as scattered mineralization or concentrates along the rock contacts with veinlets filled with serpentine or carbonates, cutting the serpentinites (Salaciński 1979; Ciemniwska *et al.* 1981). Dispersed ore minerals form large accumulations 1–4 mm in size, consisting of chromite, magnetite and an intermediate mineral phase (Pl. VIII, 2) or they occur as sparse magnetite impregnations in ore-forming minerals comprising pseudomorphs after olivine (Pl. VIII, 3) and after pyroxene. Ore accumulations along the veinlet contacts are formed from the magnetite clusters from 0.1 to one or two mm in size (Pl. VIII, 4). In serpentine and carbonate veinlets ore minerals occur extremely rarely. They are single grains of magnetite and native copper clusters ranging from 0.1 to 5 mm.

Sulfide aggregates having up to several mm in diameter and with variable mineral composition are a different form of occurrence connected with granitoid veins. These ores are not genetically connected with serpentinites.

Sparse sulfides also occur very rarely in serpentinites. It is mainly pyrite. On the other hand, in talc rocks chalcopyrite and pyrrhotite have been observed as acicular microlites with the dimensions of 0.001 per 0.1 mm arranged parallel one to another.

Thus, the minerals form the spinel group, belonging to the isomorphous series chromite-magnetite, with apparent admixture of other cations. Mg, Al, Zn, Ni has been found in variable amounts (Salaciński, Zawadzki 1983). In this group only chromite is a primary mineral crystallized directly from magmatic melt. According to the reference data (*vide* Bolewski, Parachoniak 1974), chromites in dunites can be one of the main mineral components together with olivines and pyroxenes. However, its concentration in the stu-

died samples does not exceed originally several volume percent. This can be concluded from the quantitative relations between chromites plus products of their alterations and the remaining minerals observed in the polished sections. In microscope the chromite is visible in form of relics in the inner cores of the mineral aggregates, consisting also of the transitional phase and magnetite (Pl. VIII, 2). The chromite relics are usually several hundredths to several tenths of a millimeter in size, and the diameter of the chromite-bearing aggregates ranges from 0.1 to five millimeters. The latter dimensions are typical of the primary chromite clusters.

Chromite was intensely corroded and replaced by a mineral being a transitional phase to magnetite. This process was connected with the chromium removal, as it appears from the low chromium content in the rock shown in chemical analyses (Ciemniwska *et al.* 1981) and from the fact, that the transitional phase was replaced in turn by magnetite; this transitional phase should be an intermediate mineral in the isostructural series chromite-magnetite.

The described chemical alterations have been studied with the use of an electron microprobe (Salaciński, Zawadzki 1983). The results obtained during these studies show certain regularities of the main and trace element changes in the studied minerals. Chromite relics in the aggregates, beside the main elements: iron and chromium, bear also magnesium, aluminum and zinc admixtures. Iron content increases in the intermediate phase, whereas magnesium and aluminium are absent, zinc content decreases and nickel first appears here. Magnetite, the final product of this alteration, consists of iron plus admixtures of chromium, magnesium and silicon; the nickel content increases and zinc is absent. The intermediate mineral phase can be determined as an iron-chromium spinel with the composition transitional between chromite and magnetite. This determination is not exact, because the proposed name can be also related to chromite, but more strict determination of the classification position of the iron-chromium spinel requires a precise establishing of the quantitative relations between the main elements in this mineral, what has been difficult due to small amount of the material studied and the form of its occurrence. It should be mentioned that for indication of the minerals, which can be included to the chromite species, thirty six names have been used in literature and fourteen other spinel-group minerals bears chromium in their composition (Fronde

1975). The present author uses in his whole publication the name "iron-chromium spinel" exclusively for the mineral phase formed during the chromite alteration to magnetite.

The chromite-replacement process starts from cracks and the grain margins, and in the cases of most advanced process, the chromite may be completely removed. In the most aggregates observed, the iron-chromium spinel and chromite are in equal amounts or spinel slightly prevails. Using large magnifications, it may be ascertained that chromite is corroded along the cleavage planes. In chromite along its contact with the iron-chromium spinel small oval spots of the latter mineral appear, next increasing and joining one another (Pl. IX, 1). Minute caverns have been observed in centres of these spots, what suggests that a volume decrease of the newly forming spinel is connected with the discussed process.

The iron-chromium spinel is in turn replaced by magnetite, which forms rims around the aggregates and enters them along cracks (Pl. VIII, 2, Pl. IX, 1). In the aggregates themselves magnetite occurs in subordinate amounts. During further alterations of serpentinites, in a significant part of aggregates the chromite and the iron-chromium spinel have been replaced by silicates and only magnetite rims have preserved as ribbon-arranged aggregates of complicated shape (Pl. VIII, 2). Similar phenomena have been observed also in serpentinites from the Przygórze region (Pl. X, 3).

Variations of the mineral compositions of the above described aggregates have been confirmed by geophysical studies. Magnetic thermoanalysis of samples from several drilling cores has shown the Curie points at 170, 320, 400 and 520–550°C, what indicates the presence of three different mineral phases (Mizeracka 1978). During these studies it has been also found that the samples containing the above described minerals have displayed after heating runs the magnetic effects typical of maghemite.

Chromite occurring in the ore from the Tapadła deposit looks distinctly different. In the microscope image it forms large, strongly fractured grains, bearing in marginal parts inclusions of silicate minerals (Pl. VIII, 1). Traces of alterations typical of the chromite described earlier, have not been here observed. Thus a conclusion is apparent that there are two chromite varieties of different origin, and that the second chromite variety had to crystallize after all the alterations of the first chromite variety had finished.

Various forms of the magnetite occurrence are

associated only with the first chromite variety. Besides the above described magnetite being the final alteration product of chromite, also magnetite has been found in pseudomorphs after olivine and pyroxene, as finely dispersed clusters showing linear arrangement and finally as large magnetite aggregates with lathy ilmenite inclusions.

Magnetite in post-olivine mainly serpentine pseudomorphs forms streaked accumulations along different directions of the former cracks in olivine. These accumulations consist of the magnetite grains 0.00n–0.0n mm in size, sometimes with single pyrite grains of similar size (Pl. VIII, 3). Post-pyroxene pseudomorphs bear oval magnetite grains a few hundredths of millimeter in size, distributed frequently along the original cleavage planes. The magnetite amount in the post-olivine pseudomorphs is low, equal to a few percent, however the post-pyroxene pseudomorphs may bear even about twenty percent of magnetite and its grains are larger.

The largest amounts of magnetite occur along the margins of the chrysotile and carbonate veinlets. It differs distinctly from the above described magnetite forms. This magnetite has coarser grains of tenths of millimeter in size, accumulating in aggregates occupying from about ten to thirty percent of the polished section. The magnetite grains intergrow at their edges with silicate minerals, resulting in complicated, tabular and radial structures (Pl. VIII, 4). Sometimes in the marginal parts of magnetite grains there are visible relatively rare, lathy ilmenite (?) inclusions, and in the central parts – individual pyrite microlites ca. 0.005 mm in size. This magnetite has lower microhardness than the one described before (Sałaciński 1979) and it does not occur in the rock parts without veinlets.

A different magnetite variety has been found in another borehole at the depth of 8.5 m. It forms large (up to 10 mm), oval aggregates consisting of fine (0.00n to 0.05 mm) lacinated grains. Structural and textural features of these aggregates remind certain varieties of the chromite ore from the nearby Tapadła deposit, what suggests, that this magnetite may be a pseudomorph after the ore aggregate chromite (Sałaciński 1979).

Sulfides and other ore minerals occur only in strongly altered serpentinite zones. Pyrite is, however, an exception, because it associates with certain magnetite aggregates. Chalcopyrite and pyrrhotite occur in talc rocks, where these minerals form tabular grains up to 0.1 mm in length and

0.01 mm in width, arranged parallel, according to the laminar talc rock texture. The sulfide inclusions together with talc flakes bend over and envelop only the relic aggregates of chromite, iron-chromium spinels and magnetite. Chalcopyrite and pyrrhotite inclusions seem to be later (epigenetic) both in serpentinites and in talc rocks.

Native copper has been found in carbonate veinlets occurring at greater depths (455, 911 and 947 m). It forms irregular accumulations from 0.1 to three or four millimeters in size and elongated according to the veinlet strike. This mineral crystallized most probably from hydrothermal solutions; these solutions were also the parent ones for the carbonate vein filling. The copper source, like that of pyrrhotite and chalcopyrite, was not genetically connected with serpentinites.

Rutile has been found in quartz-feldspar veinlets and in the strongly altered, beige-gray serpentinite contacting with these veinlets. It forms grainy aggregates and single oval grains 0.01 to 0.1 mm in size, being a result of the hydrothermal activity with a titanium source located probably in the neighbouring Ślęza gabbroic intrusion.

The serpentinites from the Tapadla region are also cut by vein granitoid rocks. Ore paragenesis met in these rocks is relatively rich in mineral species and their composition and form of occurrence are similar to those of the ore assemblages known from the Strzegom granitoids (*vide* Salaciński 1976, 1978), but they differ from the ore minerals occurring in serpentinites. Ores intensely impregnate the vein rocks forming irregular accumulations from 0.1 to several mm in size. The following minerals have been found: chalcopyrite, galena, sphalerite, bornite and native bismuth. Chalcopyrite is the main mineral, galena intergrowing with chalcopyrite occurs in somewhat lower amounts, and the remaining species are the minor constituents. Sphalerite and bornite have been found in certain chalcopyrite accumulations, and native bismuth forms tiny oval inclusions in galena.

Szatkowski (1982), when analysing the tectonic network connections with ore mineralization signs in the central part of the Gogolów—Jordanów serpentinite massif, has also described a number of ore minerals from this region. Beside chromite from the ore zone of Tapadla, he has found in serpentinites dispersed chromite and magnetite in forms analogous to these described above. Subordinate, in certain samples there occurred rutile, chalcopyrite, marcasite, pyrite, pyrrhotite, bornite, millerite, cobaltite and probable linneite.

In the eastern part of the Jordanów—Gogolów massif the ore accumulations of the economic importance have not been found till now, however, the present author's studies proved the common presence of dispersed ore minerals in serpentinites. Chromite, iron-chromium spinel and magnetite have been found in rock specimens collected in the quarries at Naslawice, Trzebnik, Przemilów and Mikolajów (Pl. IX, 2–4). The mineral composition of the ore paragenesis and relations between the minerals are analogous to the above described ones from the Tapadla region.

PRZYGÓRZE ULTRAMAFIC MASSIF

Ore mineral assemblage in serpentinites and associated mafic rocks of the Przygórze region is poor and it comprises chromite, magnetite, pyrite and presumably ilmenite. Replacement processes are very strongly developed in the ore minerals.

Chromite, like in the borehole Nowa Ruda IG-1, forms isometric or irregular grains up to 1 mm in size, sometimes strongly cracked with fissures filled by serpentinite (Pl. X, 1). Its microhardness equals ca. 1500 kG/mm². Sometimes occur tiny (0.01 mm) oval inclusions of pale-gray mineral of the higher reflectivity and an anisotropy difficult to determine. Probably it is ilmenite.

Chromite is an iron-high variety, especially in the outer parts of the grains, which were frequently replaced by magnetite in the form of rims (Pl. X, 2). Sometimes the inner, iron-poorer chromite grain parts were dissolved and there remain only magnetite rims, forming aggregates of long, ribbon-shaped aggregates (Pl. X, 3). Iron excess in the chromite caused that it was an unstable mineral and very frequently whole grains were replaced by magnetite. The replacement process began from cracks and fissures, and next it gradually affected larger chromite grain parts, up to complete replacement (Pl. X, 4). Certain pseudomorphs display internal structure of magnetite, underlined by the crumbling away triangles and the cleavage either according to (111), or irregular, like outlining the intergrowth borders.

Beside the above described structures, magnetite forms sometimes in the rock a finely dispersed pigment and laths within pyrite grains.

Pyrite is very rare and forms in the rock randomly scattered microlites and larger aggregate forms sometimes in the rock a finely dispersed pigment and laths within pyrite grains.

ORIGIN OF THE ORE MINERALIZATION

MAFIC MASSIFS

The above described ore mineral assemblage occurring in the Nowa Ruda massif has a complicated genesis. Origin of chromite and partly ilmenite was connected with the magma crystallization process. Differentiation of mineral composition of rocks and segregation of many lithological varieties was the reason of the variable ore concentrations and enrichment of the olivine-bearing rocks in chromite. Titanomagnetite found in the Slupiec region (Bzowski *et al.* 1983) may be included to the same mineralization stage. This mineralization assemblage indicates that in addition to chromium and iron, titanium also was present in appreciable amounts in the crystallizing magma. Titanium entered the structure of the above named ore minerals and the pyroxene (diallage) structure as a diadochic admixture replacing iron.

Between magnetite and titanium-bearing minerals: ilmenite and ulvöspinel, there exist solid solution series, which form at high temperature and are stable above 600°C. At lower temperature they exsolve giving characteristic structures. This process has been marked in the Fe-Ti oxide paragenesis from the Slupiec region gabbro (Bzowski *et al. op. cit.*) and it has been the reason of the ilmenite exsolution in pyroxenes.

Part of the ore minerals of the magmatic origin has segregated by liquation. To this group one may include the sulfide paragenesis: pyrrhotite, pentlandite, partly chalcopyrite and sphalerite. The two latter minerals have their liquation varieties occurring as disseminated aggregations, most frequently together with pyrrhotite. A pentlandite occurrence only as traces and absence of the magmatic magnetite, usually associated with sulfides in the liquation deposits, are apparent here.

After the magmatic stage, on the background of the temperature decrease, a stage of the primary mineral alterations has developed. This stage has been characterized by exsolutions in Fe-Ti oxides and mobilization and removal of chromium from chromite to chrome augites, leading to formation of the magnetite envelopes around chromite. Simultaneously silicates were undergoing to serpentinization and uralitization, and iron released during this process was fixed in magnetite of the next generation. The processes of the structural recombination started in sulfides, resulting in appear-

ance of two intergrowing pyrrhotite varieties and in twinning and characteristic anisotropy of chalcopyrite that proves its crystallization temperature exceeding 550°C. Due to metasomatic redistribution of the material, in the disulfidisation process the pyrite-marcasite-magnetite aggregates formed after pyrrhotite and pyrite metacrysts replaced silicates.

Hydrothermal processes have given quartz veins with sulfide paragenesis: chalcopyrite, pyrite and digenite. Penetration of hydrothermal solutions in rocks of the massif had the nature of pulsations, what is evidenced by the presence of three quartz generations in veins. The lowest temperatures of their crystallization were respectively equal 350, 310 and 250°C. The highest crystallization temperatures of ores (chalcopyrite) connected with the second quartz generation exceeded 550°C. Alterations of this paragenesis under hydrothermal conditions caused the bornite origin (Bialowolska, Salaciński 1977).

Hydrothermal processes, which caused the ore mineralization origin in diallage gabbro, resulted in quantitative and qualitative changes of mineral composition in the rock parts contacting quartz veins. The altered gabbro consists mostly of chloritized diallage and minor albite. Changes of the main and associated minor elements concentrations were significant. The gabbro has been enriched in magnesium, nickel and chromium, and the calcium, strontium and probably aluminum contents decreased (Bialowolska, Salaciński 1977). The increase of the nickel and chromium contents suggests that the gabbroic intrusion itself was a source of these metals.

Under the influence of the hypergene factors, the early ore minerals weathered and altered in chalcocite, covellite, limonite and malachite.

The ore mineral assemblage found in the drilling core profile of the Sobótka 2a borehole differs qualitatively from the ores stated in the Nowa Ruda gabbro and the ore content is lower. The chromite, magnetite and nickel minerals absence in the Ślęza gabbro is the most important difference. Magmatic immiscibility and crystallization produced here only ilmenite, pyrrhotite and chalcopyrite, but a part of the chalcopyrite aggregates formed during later metasomatism and hydrothermal activity, having their sources probably in the Strzegom granitoid massif. Also quartz veins in gabbro and molybdenite present in these veins are connected with this activity. Thus, the

original composition of ores connected genetically with gabbro was very poor. The only regularity in the primary ore variations is the pyrrhotite absence in the deeper part of the drilling core. Presently Jamrozik (1990) found on the massif surface an intensely titanomagnetite mineralized zone. This mineral completes thus the magmatic ore paragenesis.

After the magmatic stage, like in the Nowa Ruda massif, the essential alterations of the earlier ores are distinct. Disulfidisation process resulted in origin of the pyrite-marcasite pseudomorphs after pyrrhotite and the titanium excess was bound in the newly formed ilmenite inclusions. The later hydrothermal metasomatism alters ilmenite in rutile and hematite, and causes the saturation certain rock fragments with pyrite. The latter crystallizes initially in the habit of the pentagonal dodecahedron and next, at lower temperatures, in cubes. Especially rich pyrite mineralization occurs in quartz veins cutting gabbro at the western Śięża Mt. slopes near the village of Sady. This mineralization, like the molybdenite found in the drilling core, is genetically strange in gabbros and connected with the hydrothermal activity coming from the Strzegom granite intrusion. Niskiewicz (1974) in these veins stated also occurrence of sphalerite, chalcopyrite, chalcocite, covellite, limonite and probable digenite.

ULTRAMAFIC MASSIFS

The course of the mineral-forming processes, resulting in the present-day quantitative and qualitative states of the ore parageneses in serpentinites may be most completely deciphered by the example of the Gogółów – Jordanów massif, which has given the largest amount of the geological materials used in the studies. Ore mineral parageneses observed in the massif rocks, resulted both from the conditions and mode of the primary magmatic crystallization, and from the alteration connected with serpentinization and the later influences of the Śięża gabbroic and Strzegom granitoid intrusions.

Because most of the studied material has been collected from the drilling cores at the depths below the hypergene activity, the hypergene influence on the present-day ore preservation state may be omitted.

Dunites, zonally enriched in diallage, were the primary rocks of the studied massif. Moreover, wehrlites occurred there as well. The primary rocks relics have been found by means of the

petrographic studies of the rock samples from drilling cores (Ciemniewska *et al.* 1981). The chromite concentration in the original rocks did not exceed few percent. During the massif rock serpentinization the iron from olivines and pyroxenes was released and partly oxidized to the trivalent cation. This process resulted in origin of the finely dispersed magnetite, crystallizing in the cracks in olivine or along the cleavage planes of pyroxenes, and sometimes along the intergranular contacts. Trace sulfur concentration caused formation of single pyrite crystals. Chromite underwent also gradual alterations during serpentinization. In a moderately oxidizing environment chromium contained in chromite was partly removed and chromite was replaced by the iron-chromium spinel enriched in iron and containing trace nickel, the elements inherited from the altered olivines and pyroxenes. The remaining part of chromium was mobilized and removed during the next stage and from the left iron the magnetite rims crystallized around the chromite-spinel aggregates (Salaciński, Zawidzki 1983). These alterations were connected with the volume decrease, because the newly formed phases bear microcaverns.

Chromium in chromite occurs as trivalent cation and because of its chemical properties it is supposed to be inert contrary to the hexavalent cation. However, the observations made during the studies of the gabbroic pegmatites, the chromites occurring in serpentinites and the ultramafic enclaves in alkali-olivine basalts (Salaciński 1979; Maaløe, Printzlau 1979; Białowolska 1979, 1980; Białowolska, Salaciński 1981; Salaciński, Zawidzki 1983; Smulikowski, Kozłowska-Koch 1984) indicate the chromium mobility in the Earth's crust and in the upper mantle. Chromium mobilization from dispersed chromites and its transportation to the metasomatic chrome augites were observed in the Nowa Ruda gabbro massif. The chromium content exceeding 1 wt. % has been found in these augites (Białowolska 1979). In all these cases the chromium mobilization and redistribution have been connected with the solutions and volatiles activity. These factors supposedly displayed also an important role in the chromium mobilization and redistribution under the upper mantle conditions. After Spera (1981) the metasomatism is a common phenomenon in the Earth's mantle and the circulating water-rich solutions and carbon dioxide contain significant amounts of dissolved salts, especially under high pressure. The above listed observations indicated only the chrome redistribution results but not the mobilization

process itself. However, the possibility of the chromium leaching from chromites, and the same its migration in nature, has been evidenced experimentally (Białowolska, Sałaciński 1984). The experiments performed intended to elucidate the problem, whether the chromium mobilization from chromite by solutions is possible, and one of the experiment features was to maintain such conditions, which did not cause the chromium oxidation to the hexavalent cation, similarly to the natural environment.

As it has been mentioned above, the trivalent chromium ion bound in chromite has a low activity. Also the chromium sesquioxide Cr_2O_3 is the most stable chromium oxide, being insoluble in water, acids and bases. Its melting temperature equals 1990°C . For the mobilisation, the chromium bound in chromite should oxidize, what is rather not probable under the conditions of the discussed alterations. Thus, the possibility of the chromium mobilization by the iron chloride solutions due to the replacement of Cr^{3+} by Fe^{3+} from the solution, has been checked experimentally. A theoretical possibility of such exchange appears from the positions occupied by these two metals in the electrochemical series. As it can be found in the chemical properties tables, the normal chromium potential relative the hydrogen electrode equals -0.56 , and that of iron is -0.44 . This means that chromium should replace iron in its salts solutions. Thus experiments have been made to investigate the chromium behaviour when bound in chromite, under the action of the iron chloride solutions of various concentrations and under different vapor pressures acting on the solutions.

As it appears from the studies of gas-liquid inclusions occurring in hydrothermal minerals, the fluids in inclusions are concentrated chloride solutions, hence in the experiments the iron chloride solution has been used. The experiments included a 70-days-long period of heating at temperature 60°C of the air-tight sealed ampoules containing 2 g finely ground chromite and iron chloride solution of various concentrations and volumes. To exclude the oxidation possibility, the free volume of ampoules has been filled with carbon dioxide. After the runs, the chromium content in solution has been determined colorimetrically. The chromium content in the solution ranged from 0.20 to 1.01 wt. %.

The experiments performed indicate, that the transition of chromium present in chromite to solutions containing the Fe^{3+} ions is possible. The laboratory experiments obtained during short-time

runs (70 days) and at low temperature and low pressure indicate that the mobilization of chromium from chromite under natural processes in the Earth's crust and in the Earth's mantle might have developed during a geological time period to a significant extension. Chromium released from chromite might thus have been transported to the structures of other, newly formed minerals, mainly clinopyroxenes and chlorites, and secondary chromites crystallizing under hydrothermal conditions.

According to the experimental results (Białowolska, Sałaciński 1984), the chromium mobilization has been more intensive when iron chloride concentration has been lower and the vapor pressure in ampoules higher.

Thus, the processes of the replacement of chromite by the iron-chromium spinel and next by magnetite, connected with the mobilization and migration of chromium under natural conditions have been well evidenced. Such processes are a general regularity of the post-magmatic alterations of chromites, because the occurrence of the chromite-magnetite aggregates with distinct replacement features are shown in photographs published in numerous papers (Grafenauer 1977; Ziętek-Kruszewska 1978; Weineke, Wiesender 1982), but the interpretation of this fact is very vague or even not given.

The above mentioned reactions have been accelerated at a certain stage of the massif rock evolution, what has been marked by the removal of the chromite and spinel relics from the aggregates and the remaining of skeletal magnetite being the strongly crumpled former rims. Such acceleration might have been caused by the thermal influence of the gabbro intrusion. Chromium during these alterations has had probably a higher migration ability and it has been removed, but iron has remained at the place in solutions filling the intergranular pores. Tectonic stress caused by the gabbro intrusion might have caused the origin of the fracturing zones, next filled by hydrothermal solutions carrying Ca, Ti, CO_2 or CO and traces of Cu from the gabbro intrusion.

Reducing conditions existing in veinlets caused the native copper crystallization in these veinlets and origin of the reducing front in rocks adjacent to the veinlets. Along the veinlet/rock contacts a partial iron reduction resulted in origin of the next magnetite variety. Significant amount of this magnetite suggests that iron contained in this mineral was extracted from the rock, from the primary chromite and from the iron-chromium spinel that had replaced chromite. Probably by

a similar mode, large bodies of the secondary hydrothermal chromite, binding chromium from the primary chromites, have formed in certain massif parts.

Lower iron amount has been found in the chromite from the Tapadla deposit ore than in the primary chromite. The ore deposit chromite bears also admixtures of nickel as well as magnesium, aluminum and silicon; the three latter elements may be partly present in silicate inclusions found by microscope methods in the chromite (Pl. VIII, 1). Different composition of the trace elements in the ore deposit chromite, especially absence of zinc and presence of nickel suggest, that it bears chromium extracted from primary chromites and iron with nickel from the olivine and pyroxene decomposition. This indicates a later, hydrothermal genesis of the chromite from the Tapadla deposit.

Hydrothermal chromite deposits, whose existence was never mentioned in the literature, have been described only in 1949 by Uspenskiy (*vide* Moskaleva 1968) from the Tagilo-Nevianskiy region in the Urals, where chromite forms small nests and veinlets without commercial importance. After Uspenskiy, such chromite accumulations form due to hyperbasic rocks serpentinization, when accessory chromite of the altering rock is dissolved by hydrotherms, and next the ore substance is redeposited to form the ore bodies. Thus it is a confirmation of the regularities observed in the Tapadla region.

The reducing conditions in the above mentioned veinlets precluded the sphene formation despite the presence of titanium and calcium in solution, but titanium entered ilmenite occurring as lathy inclusions in the near-contact magnetite and calcium together with magnesium precipitated as carbonates filling presently the veinlets. Titanium origin from the gabbro intrusion is very probable, because appreciable amount of ilmenite has been found in the Słęża massif gabbros.

Origin of feldspar-quartz veins and several metres thick altered serpentinite zones adjacent to these veins are supposedly connected with a post-magmatic activity related to the gabbro intrusion. Both veins and serpentinite bear only rutile. Presence of this mineral proves a relatively extensive hydrothermal activity.

Presence of the vein granitoid rocks with appreciable amounts of ore minerals, mainly sulfides, is connected with the Strzegom granite intrusion. The paragenesis found there: chalcopyrite, galena, sphalerite, native bismuth, bornite, is

similar to parageneses observed in the Strzegom granite (*vide* Salaciński 1978) and it formed due to the hydrothermal solution activity at moderate temperatures. Activity of these solutions caused also origin of talc zones bearing scarce pyrrhotite and chalcopyrite.

The reasons of the replacement of chromite by magnetite found in the ore of the drilling cores at the depth of 8.5 m are not clear. Structural and textural features of this magnetite are different that those of other varieties, thus it cannot crystallize in any of the above described processes. Maybe, the replacement of chromite by magnetite was caused in this case by the mentioned hydrothermal solutions generated by the Strzegom granite, because this process should develop under oxidizing conditions, and solutions connected with the gabbro intrusion had reducing nature. If this hypothesis is correct and taking into account the observed in microscope intensity of the chromite replacement by magnetite, one may suppose that the activity of hydrothermal solutions generated by granite was able to cause a "destroying" of the chromite ore at the ore zones of the Tapadla type.

Supergene processes did not influence the chromite alterations, what is proved by the fact of the chromite aggregates occurrence in the weathered serpentinites; this chromite does not differ from the analogous aggregates at greater depths.

Ore mineral association occurring in the Przysgorze massif is very poor. However, taking into account the intensity of the primary rock alterations, one may suppose that the original ore mineral composition was more diversified, especially richer in pyrrhotite and other sulfides, which next could be dissolved and transported to the cover rocks.

Among the ore minerals found by the author in this massif, chromite and ilmenite are the primary ones. Initially pyrrhotite could associate them. Chromite occurring here bears less titanium and more iron than chromite from Nowa Ruda. This is evidenced by almost complete absence of ilmenite inclusions. Chromite alterations occurring at high temperature led to formation of frequently occurring rims or pseudomorphs of magnetite. The latter crystallized also from iron released during serpentinization from the mafic minerals structures, forming a finely dispersed pigment in the rock. Next magnetite variety formed during disulfidisation of pyrrhotite, resulting in the pyrite-magnetite pseudomorphs. Pyrite microlites scattered in the rock appeared due to metasomatism of silicate minerals.

EVALUATION OF THE OCCURRENCE PERSPECTIVES OF LIQUATION SULFIDE DEPOSITS IN MAFIC ROCKS

Prospecting works performed in the Lower Silesia mafic massifs areas tried to elucidate their occurrence perspectives for the liquation type sulfide mineralization. Segregation and concentration of the sulfide substance in the liquation-type deposits develops at the early magmatic stage. The same paragenesis of the main ore minerals including pyrrhotite, pentlandite, chalcopyrite and magnetite, with distinct prevailing of pyrrhotite, is the characteristic feature of these ores in all the deposits of the world. Moreover, such ores frequently bear cubanite, platinum group minerals (Pd-platinum, sperrylite, sulfides of platinum, palladium and others), sometimes nickel arsenides (niccolite), sulfides of lead, zinc and cobalt (linneite), in certain deposits also minerals of tin (stannopalladinite), molybdenum, tungsten, native gold and others. In subsurface deposits like Norylsk ores contain also minerals genetically later, belonging to a lower-temperature mineralization stage: Ni-pyrite, millerite, violarite, bornite, magnetite and others. Paragenesis of these minerals indicates more oxidizing conditions of their origin (Betekhtin, Genkin 1951).

Deposits of the copper-nickel sulfide ores are genetically strictly connected with mafic magmatic rocks distinctly enriched in magnesium: norites, gabbro-norites, olivine gabbro-diabases, rarer peridotites, comprising differentiated massifs (Betekhtin *et al.* 1964; Bolewski 1984). Studies of the relations between ore bodies and parent rocks and mineralogical studies of the ores prove that the sulfide crystallization took place after complete rock crystallization. Such geological facts like presence of the typical sulfide veins cutting parent rocks, sulfide cement of these rock fragments and even of the younger acid intrusive rocks indicate, that in the plutonic conditions the sulfide mass, segregated from silicate magma due to immiscibility at the early-magmatic stage, is able to exist for a very long time as a liquid phase.

These opinions are based first of all on the results of the geological structure studies of the copper-nickel deposits, their geological position in the mafic massifs and relations between ore bodies and wall rocks.

Experimental data have shown, that the copper sulfide solubility in silicate melts decreases during the temperature decrease, and basic melts enriched in iron and magnesium dissolve more sulfides than acid melts. Presence of sulfur in the iron-mag-

nesium silicate magmas undergoing liquation is a necessary condition of the sulfide deposit formation (Betekhtin *et al.* 1964).

The deposits connected with the following massifs of various size: Monchegorsk, Norylsk and Pechenga in Russia, Sudbury in Canada, Bushveld in RSA, are the most characteristic sulfide deposits of the liquation type. The Sudbury lopolite of the surface of over 1300 km² and the Bushveld magmatic complex of very complicate structure and surface of over 68,000 km² are the largest ones. It is difficult thus, when evaluating the ore perspectives to look for analogy, between these massifs and the Lower Silesia massifs having the surfaces of few to less than twenty square kilometres. The Monchegorsk deposit in the Kola Peninsula, the Norylsk deposit North of the Arctic Circle in the Krasnoyarsk country, Eastern Siberia and the Stare Ransko deposit connected with a small mafic massif near Jihlava in Czechoslovakia known to the author are more appropriate for comparisons.

A series of steep veins of massive sulfides has been found in the Monchegorsk deposit in a relatively small massif with the distinctly marked pseudostratification of the differentiated mafic and ultramafic rocks. Ores fill typical tension fissures with the detachment sines. The veins strike is parallel to the massif elongation and the veins occur above the level of the bottom sulfide impregnations, bearing locally also schlieren sulfide secretions consisting of small nests. The vein thickness decreases with the increasing depth, so that the veins do not reach the intrusion bottom. Large (up to several cm) almost not altered plagioclase and pyroxene crystals were found locally in the sulfides in the veins. Gradual transitions from sulfide ores to coarse-crystalline gabbro pegmatites were described from certain massif parts. Massive ores occurring in veins consist mainly of grainy aggregates of pyrrhotite with macroscopically visible inclusions of pentlandite, magnetite and chalcopyrite. Pyrrhotite and chalcopyrite contents vary significantly in the veins. There occur also veins consisting almost exclusively of chalcopyrite and cubanite.

Nickel and cobalt occur mainly in pentlandite. In nickel-rich ores in pyrrhotite matrix pentlandite forms large porphyrocrysts or anhedral grains up to 2 cm or more in size, with cleavage well-visible in microscope. Also pentlandite exsolutions in

pyrrhotite have been observed as a later generation.

Pyrite is usually absent in these ores. Only in places of post-ore deformations, where pyrrhotite mass changes in a fine-grained granoblastic aggregate, numerous small pyrite grains appear.

The Norylsk deposit ores are genetically connected with olivine gabbro-diabases (Betekhtin *et al.* 1964). Ores and their parent rocks solidified in the sub-surface conditions, what influenced distinctly their structural and textural features and composition. First of all it caused a wide development of the impregnation ores as fine sulfide nest aggregates of pyrrhotite, pentlandite and chalcopyrite. The aggregates occur mainly in lower, olivine-rich intrusion levels. Impregnation ores origin resulted from relatively quick cooling of the ore-bearing magma in the sub-surface conditions. Typical vein bodies of massive sulfides of several generations occur close to the gabbro-diabase footwall.

Overlapping by later, low-temperature mineral associations, usually not observed at the deposit formed under a deep conditions, is a typical feature of the Norylsk deposit ores. Decomposition of earlier minerals or their replacement by new ores: Ni-bearing pyrite, violarite, millerite and others, have been stated. The new ores formed under conditions of elevated sulfur concentration in the remnant solutions. With these alterations there connects also the origin of thin veinlets filled with typical hydrothermal minerals – carbonates, chlorite, sphalerite, galena and others.

The Stare Ransko massif occurs nearest to the mafic massifs of Lower Silesia. It is in the transitional zone between the Moldanubian of the Bohemian-Moravian Upland and Kutna Hora Krystallikum at the crossing of two tectonic zones striking NW–SE and N–S. The massif is very variable lithologically. In its volume different varieties of plagioclase-pyroxene gabbro to anorthosite randomly occur. Three isolated bodies of serpentized peridotites, plagioperidotites, troctolites and pyroxene gabbro were identified by Czechoslovakian geologists. They interpret them as a result of the deep-seated magma differentiation and subsequent intrusion of the gradually more acid differentiates. One of the intrusive stages gave magma rich in sulfides. Ores occur as impregnations, only locally passing in the nest aggregates. The ore deposit zones have irregular shape and cut discordantly different gabbro varieties. A part of the ore zones occurs along the contacts with ultramafic rocks.

Ore mineralization is typical of such type deposit. Pyrrhotite, chalcopyrite, pentlandite and magnetite are the main minerals. Moreover there appear cubanite, pyrite, marcasite and valeriite. Chromite and a green spinel, probably pleonaste, have been found in ultramafic rocks. Average total Ni and Cu content reaches only 0.4%, thus this deposit is not exploited despite relatively large resources.

From the above descriptions it appears that the Lower Silesian Nowa Ruda and Ślęza ultramafic massifs and the ores present in them differ distinctly from the above characterized deposits and massifs, although also certain analogies can be found. Gabbroic rocks of the Ślęza Mt. are very poor in ores, usually dispersed in the parent rock. The pyrrhotite content decreases with the depth, chalcopyrite occurs in trace amounts and the two remaining main minerals of the liquation paragenesis: pentlandite and magnetite, are absent. Also the other minerals associated with the liquation-type deposits have not been noted. The massif does not display any significant lithological differentiation. Thus this massif is not promising the presence of the liquation-type ores, but the titanomagnetite ore mineralization seems to be of certain importance.

The ore minerals assemblage in the Nowa Ruda massif is more interesting, and the geological structure and lithology of the massif are more diversified. Qualitatively and quantitatively richer there are the mineral assemblages in troctolites and olivine gabbros. Chromite, ilmenite, magnetite and pyrrhotite prevail, but the low amount of chalcopyrite, extremely rare pentlandite and absence of other nickel minerals are apparent here. Thus, in the investigated part of the massif, the paragenesis: pyrrhotite-chalcopyrite-pentlandite-magnetite, typical of the Ni-bearing mafic massifs, did not develop and nickel, found by the chemical analyses, is dispersed as diadochic admixture in pyrrhotite. Anorthite gabbros bear ore minerals in lower amounts and the ores consist mainly of pyrite and chalcopyrite.

The Nowa Ruda massif bears gabbro pegmatites, but it has not been ascertained, if the pegmatite composition changes with the depth like in the Monchegorsk deposit. Transition of gabbro into diabases suggests the comparison with the Norylsk deposit. If the mafic and ultramafic rocks are a part of the Nowa Ruda massif, an analogy with the Stare Ransko massif can be supposed.

These analogies may suggest the probable enrichment in ores in the deeper massif parts. Present-

ly the depth of the gabbro rocks occurrence is not known. The Nowa Ruda IG-1 borehole achieved the depth of 500 m and no lithological changes have been found that would indicate the close intrusion contact with the country rocks. Any distinct relation between the ores composition,

their amount and the borehole depth is not known here. Thus the possible economic importance of any ore concentration that would occur deeper is controversial. However, this massif potentially may contain economically valuable ores.

EVALUATION OF THE OCCURRENCE PERSPECTIVES OF CHROMITE DEPOSITS IN ULTRAMAFIC ROCKS

The chromite deposits are connected genetical-ly and spatially with ultramafic rocks, that may be the members of two primary magmatic formations: dunite-harzburgite and dunite-clinopyroxene-gabbroic.

Only the dunite-harzburgite formation is a parent one for the commercial ore mineralization. In all other rocks the chromite concentrations usually do not exceed the ranges of the mineralization signs and only very rarely can be called ore deposits.

Chromite deposits are not connected with all rocks of the dunite-harzburgite formation, but mainly with the dunites extremely enriched in olivine, bearing up to 5% enstatite and with dunite-harzburgites of similar composition but bearing 5–15% enstatite (Zabrodin, Prikhodko 1984). Rocks with such compositions are appropriate for origin of the massive ores and rich impregnations very high in Cr_2O_3 . Sizes of the ore bodies are proportional to the volume of the olivine rock, hence the larger dunite or dunite-harzburgite massif, the larger is the chromite body (Moskaleva, Morozova 1973).

Between the two ultramafic formations mentioned at the beginning of this chapter, there exists an interformational rock association with relics of the dunite-harzburgite formation (dunite, harzburgite, chromite bodies) placed usually deeper, and with structure features, rocks and ores typical of the dunite-clinopyroxene-gabbro formation at the upper levels. This association includes lherzolites, diallagites, wehrlites, dunites, gabbros, anorthosites, norites, deposits of platinum group elements and titanomagnetite. Due to their position, in such massifs there may coexist the raw materials belonging to the both formations, e.g. chromite, platinum group elements, titanomagnetite.

Certain concentrations of the chrome spinels are connected with the peridotite-orthopyroxene-norite formation in platforms and with the gabbro-norite-harzburgite and peridotite formations in the folded areas. The peridotite formation

has the widest distribution in the geosynclinal areas of the whole world. In fact it includes the massifs of the harzburgite composition with subordinate dunites, lherzolites, wehrlites, pyroxenites and troctolites. Large concentrations of rich chromium ores are connected in this case with the most differentiated massifs belonging to the dunite-troctolite-harzburgite and dunite-harzburgite subformations.

In the platforms the harzburgite-orthopyroxene-norite formation, called also the stratified formation, has the economic importance. The name "stratified formation" has been given due to the distinctly occurring layering of the rocks of various composition in the intrusive bodies. In the bottom part of such intrusion there occur frequently ultramafic rocks, and in the middle and upper levels — gabbros and more acid rocks. The chromite deposits are here the layered accumulations of relatively small thickness ranging from 0.1 to 1.0 m, but of large extension, up to tens and sometimes to hundreds kilometres. Chromite ores and their wall rocks have high iron content, thus they are worse than ores from the geosynclinal areas, although their resources can reach hundreds thousand tons (Pavlov, Grigoreva 1974). This chromite deposit type is named also the segregational one (Moskaleva 1968; Bolewski 1981).

The deposits connected with the formerly listed formations belong to the hysteromagmatic deposits. Two hypotheses exist on their origin. The first one indicates, that the lenticular and vein-type massive and richly impregnated ores, comprising the essential part of the commercial deposits, have sharp boundaries with the country rocks. This proves that ore formation occurred after the silicate crystallization. An ore melt might intrude in the fissures formed due the thermal contraction of the wall rocks. The fact, that such ore bodies are locally densely cut with dunite veinlets and sometimes with veins of coarse-crystalline pyroxenites being a kind of pegmatites in the ultramafic rocks, evidences that the origin time of the chromi-

te deposits being epigenetic, with respect to the wall rocks, did not exceed the limits of the main magmatic process (Betekhtin *et al.* 1964). The chromite crystallization temperature had to be relatively high. The studies of the Vogelsberg (FRG) limburgites performed by Kozłowski and Metz (1986) have shown that the Cr-Ti-Fe spinels occurring in them and having zonal structure bear Cr₂O₃ from 33 wt.% in the grain centre to 3 wt.% in the outer rim. Crystallization temperature of the individual grain zones determined on the basis of the melt inclusion homogenization ranged from 1360 to 1280°C, respectively. Thus one may suppose that most of chromium present in magma enters the spinel (also chromite) structure at a temperature above 1300°C.

The second hypothesis indicates the facts proving the origin of the chromite parent rock, i.e. inclusively, is usually presented in the literature (Birecki 1960a, 1962; Bolewski 1981).

The second hypothesis indicates the facts proving the origin of the chromite parent rock, i.e. dunite, by the metasomatic mode due to olivinization of the enstatite-harzburgite substrate (*vide* Heflik 1972, 1976). Taking this into account, and regarding the nature of the chromite aggregates in dunite, some investigators conclude that ores of this type might have formed by metasomatism, i.e. as a result of the mobilization of the scattered ore substance from the primary enstatite-harzburgite rocks (Moskaleva 1968, 1974; Zabrodin, Prikhodko 1984).

The petrographical characteristics of the Gogolów–Jordanów massif rocks given in the previous chapters and descriptions of the ore zone at Tąpadla and the dispersed mineralization in samples from the drilling cores show distinct differences with respect to the features characterizing the economic chromite formations. The Gogolów–Jordanów massif before its serpentinization consisted mainly of clinopyroxene-bearing dunites and wehrlites (Ciemniewska *et al.* 1981), whereas in the descriptions of the above named chromite-bearing formations, their paragenetic connection with orthopyroxene-bearing dunites and harzburgites is expressed.

The ore zone of Tąpadla cannot be included to the interformation rock association, because the petrographic composition of this zone agrees with descriptions neither of the lower nor of the upper level of the above described association. Platinum group elements are absent (Popiel, Waleńczak 1978) and titanomagnetite as well; magnetite with

inclusions of the titanium minerals observed in microscope is a later, secondary mineral.

Thus it should be concluded that the Tąpadla deposit does not belong to any of the listed commercial formations and the Gogolów–Jordanów massif belongs to the dunite-clinopyroxene-gabbro formation.

On the basis of the performed studies one may say that two chromite generations occur in the massif. Chromite I as an accessory mineral formed poor dispersed mineralization in the whole rock. Presently it is preserved only in relics. It was a source of chromium for the secondary chromite II, occurring in the Tąpadla deposit, which does not display the alteration signs observed in chromite I. Clusters of the both chromite generations in the contact-adjacent ore body zones occur in a distance of few centimeters, thus they should be of various origin and they formed at various time. The dispersed chromite is a primary mineral connected genetically with the massif rock crystallization. Chromite from the Tąpadla deposit formed distinctly later than the primary rocks of the massif and it is of hydrothermal origin. It is evidenced by:

- 1) the paragenesis of the hydrothermal minerals (carbonates, chrome chlorites), which form frequently common aggregates with chromite,
- 2) the observed vein forms of the ore aggregates with symmetric structure,
- 3) the experimentally proved possibility of the chromium mobilization by hydrotherms,
- 4) very probable connection of the Tąpadla deposit with tectonic zones.

Till now this deposit has been described in literature as a magmatic ore and for this reason the prospecting works in the massif have intended to localize the dunite zones, because with these zones new deposits could be connected genetically and spatially. On the basis of the hydrothermal hypothesis such works should intend to find tectonic knots and main discontinuities, which could be the solution conduits and zones of the secondary chromite concentration.

Primary chromite as an accessory mineral occurred in dunites and wehrlites of this massif in concentration of few percent, but it was common in rocks of the whole massif. It is evidenced by the investigations of the rocks from the eastern massif part, where the results of the primary chromite alterations have been observed. It indicates the appreciable amounts of chromium, which has migrated within the massif ranges and probably

secondarily crystallized. This amount significantly exceeds the ore resources found at Tapadla. A potential possibility exists for the occurrence of the commercial chromium ore deposits in the Gogółów—Jordanów massif. However, one should suppose that a part of these deposits was eroded together with the massif rocks. These rocks were reached by erosion probably in Lower Permian,

because a significant amount of chromium has been found in the Zechstein beds in the Foresudetic monocline area. The eroded ultramafic rocks and chromite deposits could be the source of this element.

The Przygórze serpentinite massif probably bears no chromite deposits because of its very small size.

RESULTS OF THE INVESTIGATIONS AND CONCLUSIONS

The presented results makes reasonable the following statements:

1. Ore mineralization in mafic and ultramafic rocks of the Sowie Mountains block cover has polygenic nature. Endogeneous ore minerals occurring there formed due to immiscibility and crystallization at the magmatic stage, crystallization from the hydrothermal solutions and various alterations under endogeneous conditions. Primary ore parageneses undergo weathering at the outcrops and they are replaced by secondary minerals.

2. A distinct differentiation of the ore mineral composition has been found in individual massif. More diversified mineralogically parageneses occur in gabbro massif. The Nowa Ruda massif is especially interesting, as well as the Gogółów—Jordanów massif among the serpentinite units.

3. The present composition of the ore mineral assemblages, their distribution in the massif rocks and the structural and textural features are a result of the post-magmatic processes, mainly the hydrothermal ones.

4. The ore assemblage occurring in the Nowa Ruda gabbro massif has the most diversified origin. Chromite, ilmenite and titanomagnetite origins are connected with processes of the magmatic melt crystallization. The sulfide paragenesis: pyrrhotite, pentlandite, chalcopyrite and sphalerite, formed due to the sulfide and silicate melt immiscibility. During the temperature decrease a stage of the primary mineral alterations occurred, giving the exsolution structures in iron-titanium oxides, mobilization and removal of chromium from chromite and iron from mafic silicates, structural rearrangement of pyrrhotite and chalcopyrite, disulfidization of pyrrhotite and appearance of pyrite metacrysts. Pyrite initially forms crystals of the pentagonal dodecahedron habit, next changing to cube. Later hydrothermal solu-

tions activity yielded the quartz-sulfide-calcite parageneses filling veins.

5. The Słęża gabbro massif bears a poor ore assemblage. Magmatic minerals: ilmenite, pyrrhotite, chalcopyrite and titanomagnetite, occur there. These minerals underwent similar alterations as in the Nowa Ruda massif. The hydrothermal ore paragenesis includes chalcopyrite, molybdenite, pyrite and sphalerite. The neighbouring Strzegom granitoid massif was the hydrothermal solutions source.

6. The serpentinites of the Gogółów—Jordanów massif bear mainly the spinel group minerals: chromite, magnetite and a transitional phase between the two formed minerals, called in this paper the iron-chromium spinel. Two chromite generations have been found. Primary chromite was of magmatic origin and it occurred as an accessory in the rocks of the whole massif. At the post-magmatic stage a mass chromium leaching from the primary chromite occurred due to hydrothermal activity, resulting in the iron-chromium spinel and magnetite origin. The mobilized chromium migrated in solutions and it was reprecipitated as the second generation chromite. Accumulations of this chromite formed ore bodies in the Tapadla region. Thus this deposit is of hydrothermal origin. The possibility of the chromium migration under endogene conditions has been proved experimentally.

7. The studied massifs mostly can bear an economic ore mineralization. The massif Gogółów—Jordanów is the most interesting one, because it may contain chromite deposits of hydrothermal origin, located at the strongly tectonically engaged zones. A probability of the liquation sulfide accumulations occurs in the Nowa Ruda gabbro massif a greater depths, and in the Słęża massif the zones enriched in titanomagnetite may be of the economic importance.

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GENEZA MINERALIZACJI KRUSZCOWEJ MAFITÓW I ULTRAMAFITÓW W OTOCZENIU BLOKU SOWIOGÓRSKIEGO

Ryszard SAŁACIŃSKI

Instytut Geologii Podstawowej, Wydział Geologii Uniwersytetu Warszawskiego
al. Zwirki i Wigury 93, 02-089 Warszawa

Streszczenie

ABSTRAKT. W mafitach i ultramafitach występujących w osłonie bloku sowiogórskiego są obserwowane przejawy mineralizacji kruszcowej o zróżnicowanym składzie jakościowym i ilościowym. Autor przebadal asocjacje kruszcowe występujące w masywach gabrowych Nowej Rudy i Słęzy oraz w masywach serpentynitowych Gogolowa-Jordanowa i Przygorza koło Nowej Rudy. Materiał do badań pochodził z wiercen wykonanych przez Państwowy Instytut Geologiczny oraz z opróbowania kamieniołomów i odkrywek. W badaniach kruszców autor stosował przede wszystkim optyczne badania mikroskopowe w świetle odbitym uzupełnione wcześniejszymi badaniami przy użyciu mikroskopu elektronowego i mikroanalizatora rentgenowskiego. (Sałaciński, Zawadzki 1983). Wykonane zostały również eksperymentalne badania chemiczne (Białowska, Sałaciński 1984).

W pracy scharakteryzowano tlenkową mineralizację Cr, Fe, Ti oraz mineralizację siarczkową Fe, Cu, Ni, Zn, Pb. Istotną częścią pracy jest analiza procesów

przeobrażeń kruszców w etapie pomagmowym, a także stwierdzenie prawidłowości przebiegu tych procesów i powstawania nowych faz mineralnych, postaci krystalograficznych oraz koncentracji złozowych.

Zdaniem autora znane złozę chromitu w Tąpadlach w masywie Gogolów-Jordanów jest pochodzenia hydrotermalnego. Chromit występujący w rudzie z tego złozu jest minerałem wtórnym w stosunku do chromitu pierwotnego, rozproszonego w skałach całego masywu i powstał w wyniku uruchomienia i przemieszczenia chromu między dwiema generacjami tego minerału. Autor uważa, że istnieją uzasadnione przesłanki występowania innych złoz chromitowych w tym masywie, związanych ze strefami tektonicznymi. Prawdopodobne jest również istnienie koncentracji złozowych Cu, Ni i Fe w głębszych poziomach masywu noworudzkiego.

WSTĘP

Wychodnie plutonicznych skał mafitowych i ultramafitowych zajmują w Sudetach i na obszarze przedsudeckim stosunkowo niewielką powierzchnię. Najliczniej skały te występują w obrzeżeniu bloku sowiogórskiego, gdzie tworzą szereg masywów wśród skał jego osłony (fig. 1). Pozycja tektoniczna tych masywów, ich wiek oraz geneza są przedmiotem wieloletnich sporów naukowych. Przyczyną tego jest fakt, iż „blok sowiogórski wraz z otaczającymi go i na znacznych odcinkach podścielającymi skałami zasadowymi, głównie serpentynitami, pod którymi występują z kolei granitoidy, stanowią najgłębsze jednostki struktury tektonicznej południowo-zachodniej Polski, a Dolnego Śląska w szczególności...” (Oberc 1972). Mafity i ultramafity tych jednostek wchodziły w skład najstarszych pięter strukturalnych tej części Polski. Mają one skomplikowaną historię geologiczną, gdyż ulegały wielokrotnym przeobrażeniom i deformacjom w czasie kolejnych orogenez, począwszy od prekambryjskich cykli tektonicznych aż do najmlodszych (Oberc *op. cit.*).

Wszystko to spowodowało, że badanie masywów skał mafitowych i ultramafitowych w osłonie bloku sowiogórskiego jest trudne. Dotychczasowe badania geologiczne były wykonywane w sposób fragmentaryczny i dotyczyły wycinkowych problemów geologicznych. Dopiero w ostatnich latach podjęto próbę takiego kompleksowego opracowania dla masywu serpentynitowego Gogolów—Jordanów (Ciemniewska *et. al.* 1981). Temat badawczy zrealizowany przez Państwowy Instytut Geologiczny dotyczył wyjaśnienia możliwości występo-

wania rud chromu i innych surowców mineralnych w ultramafitach tego masywu. Pod koniec lat siedemdziesiątych wykonano na obszarze Uliczno—Czarna Góra 25 otworów wiertniczych o głębokości od 34 do 980 m, mających razem około 2900 mb. Pracom wiertniczym towarzyszyły kompleksowe badania geofizyczne, geochemiczne, mineralogiczne, petrograficzne i metalogiczne. Badania innych masywów ograniczyły się do wykonania zdjęcia geochemicznego i kilku wierceń (Sobótka 2a, Nowa Ruda IG-1, Przygorze 1, Przygorze 2).

W niniejszej pracy autor wykorzystał własne wyniki uzyskane zarówno z wyżej wymienionych badań, w których uczestniczył dzięki uprzejmości Instytutu Geologicznego, jak i z badań materiałów geologicznych pochodzących z opróbowania powierzchniowego.

W badaniach laboratoryjnych autor dysponował materiałami geologicznymi pochodzącymi z masywów gabrowych Słęzy, Nowej Rudy, Braszowic i Brzeźnicy oraz z masywów serpentynitowych Gogolowa—Jordanowa, Szklar, Grochowej i Przygorza. Podstawowy materiał do badań kruszców metodami optycznymi stanowiło ponad 700 preparatów połączonych i kilkadziesiąt płytek cienkich. W celu ustalenia zmian składu chemicznego niektórych minerałów kruszcowych oraz wyjaśnienia możliwości migracji chromu w czasie przeobrażeń chromitów wykonano badania w mikroobszarze oraz przy użyciu mikroskopu skaningowego (Sałaciński, Zawadzki 1983), a także eksperymentalne badania chemiczne (Białowska, Sałaciński 1984).

OKRUSZCOWANIE SKAŁ MAFITOWYCH I ULTRAMAFITOWYCH

Skały mafitowe i ultramafitowe są ważne z punktu widzenia surowcowego jako miejsca koncentracji wielu metali. Wzbogacenie omawianych skał w różne metale ma dwójaki charakter. Podwyższony klark jednych metali jest spowodowany zwiększonym ich udziałem w składzie minerałów skałotwórczych, inne zaś mogą tworzyć własne minerały, skupione często w wyodrębnionych litologicznie lub tektonicznie częściach masywów. Z punktu widzenia złożowego istotne znaczenie ma druga forma występowania, podczas gdy pierwsza ważna jest o tyle, o ile w wyniku późniejszych przeobrażeń skał lub minerałów nastąpi uwolnienie i uruchomienie rozproszonych metali, a następnie wtórna ich koncentracja.

Metalem występującym w największej ilości w obu grupach skał jest żelazo (tab. 1). Lokalne nagromadzenia rud żelazo-tytanowych lub chromitu w mafitach i ultramafitach przedstawiają ważne geochemicznie koncentracje złożowe żelaza sięgające 60% wag. (Połański 1974).

Chrom wchodzi przede wszystkim w skład chromitu, który jest głównym minerałem złóż występujących w skałach ultramafitowych, jak również w serpentynitach powstałych z ich przeobrażenia. Część chromu wchodzi w skład skałotwórczych krzemianów, zwłaszcza piroksenów i chlorytów oraz w skład minerałów zaliczanych do grupy spineli, do których pod względem strukturalnym należy również chromit. W grupie spineli czyste, skrajne człony rzadko występują w postaci naturalnych minerałów. Poszczególne minerały są wyodrębniane i zazwyczaj nazywane na podstawie dominujących kationów R^{2+} lub R^{3+} . Doprowadziło to do powstania wielu różnych klasyfikacji tej grupy minerałów. Niektóre z nich są przedstawione w tabeli 2 i na figurach 2a,b,c,d.

Nikiel występuje w produktach wczesnomagmowych w pentlandycie, który tworzy wrostki w pirotynie lub niezależne skupienia. Równocześnie nikiel wykazuje pokrewieństwo krystalochemiczne z magnezem i żelazem. czego wynikiem są

diadochowe podstawienia tych pierwiastków w krzemianach magnezowo-żelazowych, a zwłaszcza w oliwinach (Połański 1974).

Kobalt, podobnie jak nikiel, wykazuje wyraźne pokrewieństwo geochemiczne z Fe^{2+} oraz Mg^{2+} i może wchodzić do struktur minerałów tych pierwiastków. Dlatego też w skałach ultramafitowych nie ma w zasadzie minerałów kobaltu, a cała jego ilość występuje w stanie rozproszonym.

Częstość występowania platynowców w skałach magmowych jest zupełnie znikoma. W stosunkowo większej ilości występują one w dunitach, perydotytach i piroksenitach. Tworzą w tych skałach własne minerały lub wchodzą jako domieszki do niektórych minerałów, a zwłaszcza do chromitu, spineli chromowych i pirotynu.

Tytan występuje w największej ilości w skałach mafitowych, głównie w postaci ilmenitu lub tytanomagnetytu. Część tytanu wchodzi w skład minerałów krzemianowych.

Wanad w skałach mafitowych jest typowym pierwiastkiem rozproszonym, wchodzącym głównie w skład piroksenów, hornblendy, biotyty oraz magnetytu. Duże rozproszenie wanadu w omawianych skałach jest jednocześnie przyczyną braku w nich koncentracji tego pierwiastka o znaczeniu złożowym.

Miedź praktycznie nie wchodzi w skład skałotwórczych krzemianów, lecz wydziela się w postaci akcesorycznych siarczków. Wzbogacenie w miedź skał mafitowych jest wynikiem procesów likwacyjnych.

Przedstawione tendencje geochemiczne są niekiedy przyczyną powstawania dużych złóż wymienionych pierwiastków w skałach mafitowych i ultramafitowych. Należy tu wymienić znane złoża żelaza formacji tytanomagnetytowej typu Taberg i Otänmeki, likwacyjne złoża miedziowo-niklowe Sudbury, Norylsk, Pieczenga i Bushveld, złoża chromitów na Uralu, Półwyspie Bałkańskim i w Bushveldzie oraz złoża platynowców na Uralu, w Bushveldzie i Sudbury.

DOTYCHCZASOWY STAN ROZPOZNANIA MINERALIZACJI KRUSZCOWEJ W OSŁONIE BLOKU SOWIOGÓRSKIEGO

Większość danych dotycząca występowania kruszców odnosi się do niedawna do złóż chromitów w masywach serpentynitowych Grochowej i Gogołowa—Jordanowa oraz występowania złoża wietrzeniowego krzemianowych rud niklu w masywie serpentynitowym Szklar. Pierwsze próby eksploatacji chromitu podejmowano na Dolnym Śląsku już w latach 1824—1827 na wzgórzu Twarda Góra koło Grochowej (Birecki 1960a, b, 1962). W 1877 r. odkryto chromit w Tapadłach na północnym stoku Czarnej Góry i od 1890 r. rozpoczęto eksploatację, która trwała z długimi przerwami do 1945 r. Birecki (1962) podaje, że wyeksploatowano z tego złoża około 4000 t rudy chromitowej. Górnictwymi pracami rozpoznawczymi i eksploatacyjnymi stwierdzono szereg gniazd rudnych, które oznaczono symbolami literowymi od A do K, w tym trzy

małe gniazda o symbolach H-1, H-2 i H-3 (fig. 3).

Podjęte od 1946 r. prace poszukiwawcze doprowadziły do odkrycia i udokumentowania stref rudnych w zachodniej części masywu serpentynitowego Gogołów—Jordanów w rejonie Wir, Gogołowa i Słupicy (Ciemniewska *et al.* 1981). Oprócz prac prospekcyjnych były wykonywane w różnych masywach badania geochemiczno-mineralogiczno-petrograficzne (Waleńczak 1962; Szumlas 1963; Pawlikowski, Salamon 1972; Białowolska 1973). Na początku lat sześćdziesiątych w ówczesnym Zakładzie Złóż Rud Metali Nieżelaznych Instytutu Geologicznego podjęto badania pierwotnych mineralizacji kruszcowych w mafitach i ultramafitach Dolnego Śląska. Prace te były kontynuowane przez ponad 20 lat.

OGÓLNA CHARAKTERYSTYKA GEOLOGICZNA MASYWÓW MAFITOWYCH I ULTRAMAFITOWYCH W OSŁONIE BLOKU SOWIOGÓRSKIEGO

W osłonie bloku sowiogórskiego skały mafitowe i ultramafitowe występują w postaci wyraźnie wyodrębnionych masywów. Niektóre z tych skał zostały przeobrażone w serpentynity. Tworzą one obecnie duże masyw Gogołów—Jordanów

oraz małe masywy Sobótki, Szklar, Grochowej koło Ząbkowic Śląskich i Przygórze koło Nowej Rudy. Gabra występują w dwóch dużych masywach Ślęży i Nowej Rudy oraz w kilku niewielkich wystąpieniach w rejonie Braszowic i Brzeźnicy.

Wymienione skały Oberc (1960, 1972) uważa za późnoprekambryjskie.

Ponieważ najciekawsze materiały do badań autor uzyskał z masywów Słęzy, Nowej Rudy, Przygórza i Gogolowa—Jordanowa, dalsze rozważania dotyczą tylko tych masywów.

Masyw gabrowy Słęzy zajmuje powierzchnię kilkunastu km². W osłonie jego występują serpentynity masywów Gogolowa—Jordanowa i Sobótki, amfibolity oraz granit strzegomski (fig. 4), który również częściowo podściela gabra od strony zachodniej. Dokładniejsza charakterystyka skal masywu zawarta jest w pracach Białowskiej (1973) i Majerowicza (1979).

Masyw Nowej Rudy charakteryzuje się dużym i wyraźnym zróżnicowaniem. Występują w nim perydotyty, piroksenity, troktolity, gabra oliwinowe, gabra anortytowe, gabra diallagowe, gabroanortozyty i anortozyty. Na południe od linii Słupiec—Dzikowiec skały gabrowe przechodzą w diabazy (fig. 5). Zmienność petrograficzną skal masywu w profilu pionowym ilustruje wiercenie Nowa Ruda IG-1 opisane przez Fedaka i Kowarza (1967). Interesującą odmianą petrograficzną są pegmatyty gabrowe, które stwierdzono w wierceniu i na powierzchni masywu (Białowska, Sałaciński 1981). Badania wykonane przez Borkowską (1985) wykazały, że skały gabroidowe masywu noworudzkiego krystalizowały na niewielkiej głębokości, a ich temperatura krystalizacji mieściła się w przedziale 860–1260°C.

Między noworudzkiem masywem gabrowym a kataklazytami sowiogórskimi występuje w rejonie Przygórza niewielki

masyw serpentynitowy. Wykonane w 1964 r. dwa wiercenia wykazały, że centralna część masywu zbudowana jest z serpentynitów silnie zlustrowanych tektonicznie. w strefie brzeżnej natomiast oprócz nich występują również mafity i ultramafity zmienione w niewielkim stopniu. Wśród nich wyróżniono gabro oliwinowe, anortytowe i diallagowe (Fedak, Kowarz 1967).

Największym masywem serpentynitowym na Dolnym Śląsku jest masyw Gogolów—Jordanów (fig. 4). Dotychczas rozpoznana geologicznie jego część zajmuje około 100 km², przy czym wschodnie mają około 30 km². Od południa graniczy z gnejsami bloku sowiogórskiego, od zachodu i północnego zachodu z granitem strzegomskim, od północy z gabraami Słęzy, a od wschodu z gnejsami i lupkami fyllitowymi. Masyw zbudowany jest z zserpentynizowanych skal ultramafitowych o różnym stopniu przeobrażenia. Na podstawie badań petrograficznych materiału pochodzącego z wierceń wyróżniono: dunity, wehrlity, serpentynity z relikami oliwinu, serpentynity, serpentynity z węglanami, gabra, kersantyty, rodingity oraz skały talkowo-węglanowe, talkowo-serpentynowe, talkowo-chlorytowe, chlorytowo-węglanowe, kwarcowo-skaleniove i biotytowo-węglanowe (Ciemniewska *et al.* 1981). Skały są silnie spękane. Wyróżnia się kilka kierunków nieciągłości tektonicznych (Chmura, Sulkowski 1965; Gajewski 1970; Oberc 1972), przy czym w rejonie złoza w Tąpadłach (fig. 6) następuje zagęszczenie różnych systemów spękań (Szatkowski 1982).

MINERALIZACJA KRUSZCOWA MAFITÓW I ULTRAMAFITÓW W OSŁONIE BLOKU SOWIOGÓRSKIEGO

Z wymienionych masywów gabrowych najliczniejsze i najbardziej zróżnicowane paragenezy kruszczowe występują w gabrach Nowej Rudy. W badaniach mikroskopowych autor stwierdził występowanie w profilu wiercenia: chromitu, ilmenitu, rutylu, leukokseniu, pirotynu, pentlandytu, magnetytu, chalkopiryty, sfalerytu, bornitu, piryty, markasytu i limonitu (pl. I—V). Najczęściej z tych mineralów występuje chromit, ilmenit, magnetyt i piryty. W południowej części gabrowego masywu Nowej Rudy autor wraz z Białowską (Białowska, Sałaciński 1977) stwierdził występowanie mineralów kruszczowych w żyłce kwarcowej. Była to grupa hydrotermalnych mineralów pierwotnych, które zachowały się tylko w reliktach (chalkopiryty, piryty i digenit), oraz grupa mineralów powstałych w wyniku zastępowania tych trzech mineralów w warunkach hydrotermalnych lub hipergenicznych (bornit, chalkozyn, kowelin, malachit i limonit).

W profilu wiercenia wykonanego na masywie Słęzy autor stwierdził w gabrze ilmenit, rutyl, leukoksen, hematyt, pirotyn, chalkopiryty, piryty, markasyt i limonit, a w żyłce kwarcowej przecinającej gabro występujący w znacznej ilości molibdenit. W granicy znajdującym się pod gabrem występuje pirotyn, piryty i markasyt (pl. VI, VII).

Skały występujące w ultramafitowym masywie Gogolów—Jordanów zawierają liczne, chociaż mało urozmaicone pod względem składu mineralnego, przejawy mineralizacji kruszczowej. Największe nagromadzenia mineralów kruszczowych występują w złożu chromitu w Tąpadłach (fig. 3), w zachodniej części masywu. Skład rudy ze złoza przedstawiają tabele 3, 4 i figura 2, a obraz mikroskopowy występującego w niej chromitu plansza VIII, fot. 1.

W wierceniach poszukiwawczych wykonanych w sąsiedztwie złoza nie stwierdzono nowych gniazd rudnych, autor natomiast zaobserwował powszechne występowanie mineralów kruszczowych w postaci rozproszonej lub też wzdłuż kontaktów z żyłkami węglanowymi lub serpentynowymi (Sałaciński 1978; Ciemniewska *et al.* 1981). W postaci rozproszonej minerały kruszczowe tworzą duże agregaty (1–4 mm), złożone z reliktyw chromitu otoczonego „spinelem żelazowo-chromowym” i magnetytem (pl. VIII, 2), oraz występują w postaci drobno-dispersyjnych impregnacji (głównie magnetytem) mineralów skalotwórczych (pl. VIII, 3). Wzdłuż kontaktów z żyłkami występują agregaty magnetytu złożone z kryształów wielkości do 2 mm (pl. VIII, 4). Osobną formę występowania stanowią skupienia siarczków w żyłkach kwaśnych skal magmowych. Stwierdzono w nich chalkopiryty, galenę, sfaleryt i bornit. Występuje też bizmut rodzimy. W żyłkach węglanowych przecinających serpentynity autor stwierdził skupienia miedzi rodzimej. We wschodniej części masywu Gogolów—Jordanów autor zaobserwował powszechne występowanie w serpentynitach, w postaci rozproszonej, agregatów kruszczów o składzie analogicznym, jak w rejonie Tąpadel (chromit, spinel żelazowo-chromowy, magnetyt — pl. IX).

Skład mineralny i ilościowy kruszczów w skalach masywu Przygórza jest bardzo ubogi. Występuje tu chromit, magnetyt, piryty i przypuszczalnie ilmenit (pl. X).

GENEZA MINERALIZACJI KRUSZCOWEJ

Zespół minerałów kruszcowych występujących w masywie Nowej Rudy ma złożoną genezę. Z procesami krystalizacji stopu magmowego związane jest powstanie chromitu i częściowo ilmenitu. Do tego samego etapu mineralizacji należy zaliczyć również tytanomagnetyt zarejestrowany w rejonie Shipca (Bzowski *et al.* 1983). Parageneza ta wskazuje, że w krystalizującej magmie oprócz chromu i żelaza w dużej ilości znajdował się tytan. Wszedł on również w skład piroksenów, z których następnie uległ odmieszanemu w postaci ilmenitu (pl. II, 4). Na drodze likwacji powstał pirotyn, pentlandyt oraz częściowo chalkopiryt i sfaleryt. Spadek temperatury po etapie magmowym spowodował wiele przeobrażeń minerałów pierwotnych. Nastąpiło odmieszanienie roztworów stałych w tlenkach żelazowo-tytanowych, uruchomienie i odprowadzenie chromu z chromitu do augitów chromowych, uwolnienie żelaza w czasie serpentynizacji i uralizacji oraz związane z tym powstanie kolejnej generacji magnetytu, a także powstanie przerostów dwóch modyfikacji pirotynu i zbliżniczeń chalkopiryty. W procesie dysulfuryzacji powstały agregaty pirotyto-marksytowo-magnetytowe zastępujące pirotyn, a w wyniku metasomatozy krzemianów metakryształy piryty. W procesach hydrotermalnych powstały żyły kwarcowe z siarczkami. Działalność hydrotermalna miała charakter pulsacyjny. Najwyższa temperatura krystalizacji chalkopiryty przewyższała 550°C, natomiast najniższe temperatury kolejnych generacji kwarcu wynosiły kolejno 350, 310 i 250°C. Gabro kontaktujące z żyłami zostało wzbogacone w Mg, Ni i Cr, a zubożone w Ca, Sr i Al.

Asocjacja kruszców w gabrze Ślęży różni się jakościowo od minerałów z gabry Nowej Rudy i jest uboższa ilościowo. Brak tutaj chromitu, magnetytu i minerałów niklu. W czasie kry-

stalizacji magmy i likwacji powstał tylko ilmenit, tytanomagnetyt, pirotyn i chalkopiryt. Część chalkopiryty związana jest również z późniejszą metasomatozą i działalnością hydrotermalną, mającą źródło w granitowym masywie strzegomskim. Z działalnością hydrotermalną związane jest również powstanie molibdenitu i żył kwarcowych z pirytem.

Serpentynitowy masyw Gogołów–Jordanów składał się pierwotnie z dunitów wzbogaconych w diallag. Udział chromitów w pierwotnych skałach nie przekraczał kilku procent. W czasie serpentynizacji doszło do uwolnienia żelaza z oliwinów i piroksenów oraz powstania magnetytu. Równocześnie chrom zawarty w chromicie ulegał częściowemu odprowadzeniu i na miejscu chromitu utworzył się spinel żelazowo-chromowy. Po kolejnym etapie odprowadzenia chromu powstał magnetyt (pl. VIII, 2, pl. IX). Uruchomiony chrom był przenoszony roztworami hydrotermalnymi i mógł wchodzić w skład wtórnego chromitu, krystalizującego w strefach tektonicznych. W takiej strefie zlokalizowane jest złożo rud chromu w Tąpadłach (fig. 6). Możliwość uruchomienia chromu w warunkach naturalnych została udowodniona eksperymentalnie przez autora i Białowską (Białowska, Sałaciński 1984). Roztwory hydrotermalne spowodowały również powstanie w żyłkach węglanowych skupień miedzi rodzimej, a w skale wzdłuż kontaktów z żyłkami dużych ilości magnetytu.

Asocjacja kruszców w masywie Przygórza jest bardzo uboga. Za minerały pochodzenia magmowego można uznać tutaj tylko chromit i ilmenit. Początkowo mógł im towarzyszyć pirotyn zastąpiony obecnie pirytem. Późniejsze przeobrażenia spowodowały usuwanie chromu i powstawanie otoczek i pseudomorfoz magnetytowych (pl. X, 2–4).

OCENA PERSPEKTYW ZŁOŻOWYCH

Porównując skład mineralny opisanych paragenez kruszcowych oraz skład i zmienność petrograficzną dolnośląskich masywów mafitów i ultramafitów z opisami budowy geologicznej i składu mineralnego złóż: monczegorskiego na półwyspie Kola, norylskiego we wschodniej Syberii, Starego Ranska w Czechosłowacji, i opisami chromitowych formacji złozowych

autor uważa, że istnieją perspektywy występowania przemysłowych koncentracji kruszcowych Cu, Ni i Fe typu likwacyjnego w głębszych częściach masywu Nowej Rudy oraz złóż chromitów pochodzenia hydrotermalnego w masywie serpentynitowym Gogołów–Jordanów, w strefach silnie spękanych tektonicznie.

ZESTAWIENIE WYNIKÓW I WNIOSKI

Przedstawiony materiał faktograficzny upoważnia do następujących stwierdzeń:

1. Mineralizacja kruszcowa mafitów i ultramafitów w osłonie bloku sowiogórskiego ma charakter poligeniczny. Występujące tu endogeniczne minerały kruszcowe powstały w wyniku likwacji i krystalizacji w etapie magmowym, krystalizacji z roztworów hydrotermalnych oraz różnorodnych przeobrażeń w warunkach wgłębnych. Pierwotne paragenezy kruszców ulegają wietrzeniu na wychodniach i są zastępowane minerałami wtórnymi.

2. Zaznacza się wyraźne zróżnicowanie składu mineralnego kruszców występujących w poszczególnych masywach. Bardziej urozmaicone mineralogicznie paragenezy występują w masywach gabrowych. Szczególnie interesujący jest masyw Nowej Rudy, a z serpentynitowych – masyw Gogołów–Jordanów.

3. Obecny skład zespołów minerałów kruszcowych, ich rozmieszczenie w skałach masywów oraz cechy struktural-

no-teksturalne są wynikiem procesów pomagmowych, głównie hydrotermalnych.

4. Zespół kruszców występujących w gabrowym masywie Nowej Rudy ma najbardziej urozmaiconą genezę. Z procesami krystalizacji stopu magmowego związane jest powstanie chromitu, ilmenitu i tytanomagnetytu. W wyniku likwacji wydzielila się parageneza siarczkowa: pirotyn, pentlandyt, chalkopiryt i sfaleryt. Wraz ze spadkiem temperatury nastąpił etap przeobrażeń minerałów pierwotnych, wyrażający się odmieszanieniem roztworów stałych w tlenkach żelazowo-tytanowych, uruchomieniem i odprowadzeniem chromu z chromitu, a żelaza z minerałów ciemnych, przebudową strukturalną pirotynu i chalkopiryty, dysulfuryzacją pirotynu oraz powstaniem metakryształów piryty. Zaobserwowano, że piryty początkowo tworzy kryształy o pokroju dwunastocianu pentagonalnego, które następnie ulegają przebudowie w szescian. Z późniejszych roztworów hydrotermalnych powstały paragenezy kwarcowo-siarczkowo-kałcytowe wypełniające żyły.

5. W masywie gabrowym Słęzy zespół kruszców jest ubogi. Występują tu minerały pochodzenia magmowego: ilmenit, pirotyn, chalkopiryt i tytanomagnetyt. Minerale te ulegają podobnym przemianom, jak w masywie Nowej Rudy. Parageneza hydrotermalna kruszców obejmuje chalkopiryt, molibdenit, piryty i sfaleryty. Źródłem roztworów hydrotermalnych był sąsiedni granitoidowy masyw strzegomski.

6. W serpentynitach masywu Gogołów—Jordanów występują przede wszystkim minerały z grupy spineli: chromit, magnetyt oraz minerał będący fazą przejściową między nimi, nazwany w tej pracy spinelem żelazowo-chromowym. Zarejestrowano dwie generacje chromitu. Chromit pierwotny jest pochodzenia magmowego i występował akcesorycznie w skałach całego masywu. W etapie pomagmowym nastąpiło masowe hutowanie chromu z pierwotnego chromitu przez hydrotermę, w wyniku czego powstał spinel żelazowo-chromowy i magnetyt. Uruchomiony chrom migrował w roztworach, a następ-

nie został redeponowany w chromicie drugiej generacji. Skupienia tego chromitu utworzyły ciała złozowe w rejonie Tąpadeł. Złozę to jest więc pochodzenia hydrotermalnego. Możliwość migracji chromu w warunkach wglębnych została udowodniona eksperymentalnie.

7. Badane masywy są w większości perspektywiczne pod względem złozowym. Najbardziej interesujący jest masyw Gogołów—Jordanów, w którym mogą istnieć złozy chromitu pochodzenia hydrotermalnego, zlokalizowane w miejscach o intensywnej tektonice. W masywie gabrowym Nowej Rudy istnieje prawdopodobieństwo występowania likwacyjnych skupień siarczków na większej głębokości, a w masywie Słęzy znaczenie złozowe mogą mieć strefy wzbogacone w tytanomagnetyt.

Warszawa, luty 1989 r.

PLATES AND EXPLANATIONS

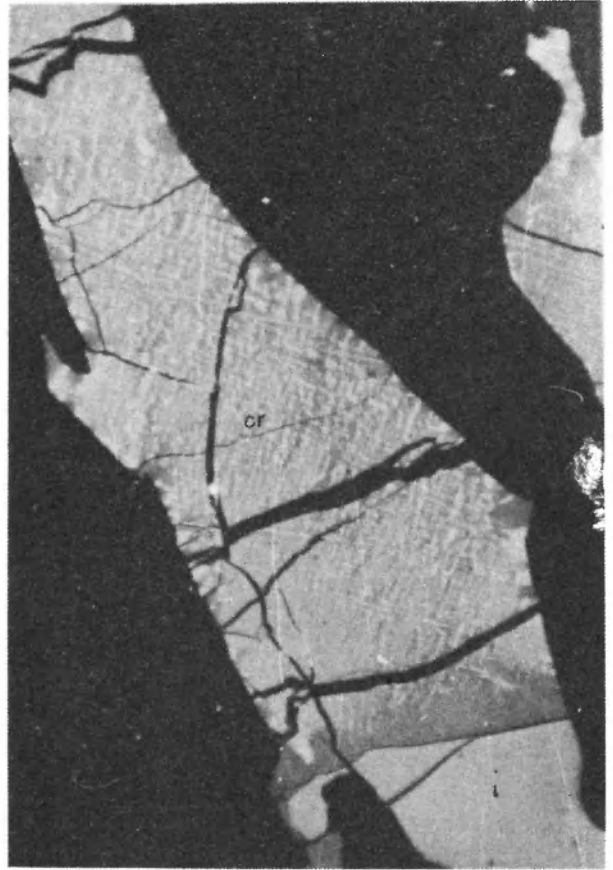
PLATE I – PLANSZA I

Ore minerals in gabbros of the Nowa Ruda massif (borehole NR-IG-1)
Minerały kruszczowe w gabrach masywu Nowej Rudy (wiercenie NR-IG-1)

1. Cracked chromite grain with fissures filled with serpentine, $\times 110$, one nicol, reflected light
Skatylazowane ziarno chromitu ze szczelinami wypełnionymi serpentynem, pow. $110\times$, 1 N, światło odbite
2. Ilmenite (i) and chromite (cr) with ilmenite exsolutions, $\times 420$, one nicol, reflected light
Ilmenit (i) i chromit (cr) z odmieszaniami ilmenitu, pow. $420\times$, 1 N, światło odbite
3. Chromite (cr) with large ilmenite (i) inclusions and magnetite (m) in pseudomorph after olivine, $\times 55$, one nicol, reflected light
Chromit (cr) z dużymi wrostkami ilmenitu (i) oraz magnetyt (m) w pseudomorfozie po oliwinie, pow. $55\times$, 1 N, światło odbite
4. Chromite (cr) replaced by magnetite (m), $\times 210$, one nicol, reflected light
Chromit (cr) zastępowany magnetytem (m), pow. $210\times$, 1 N, światło odbite



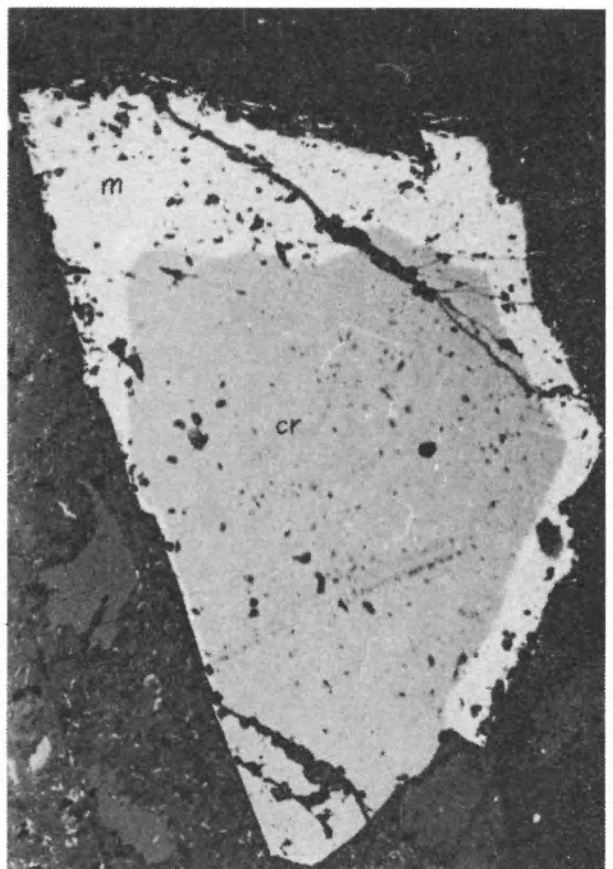
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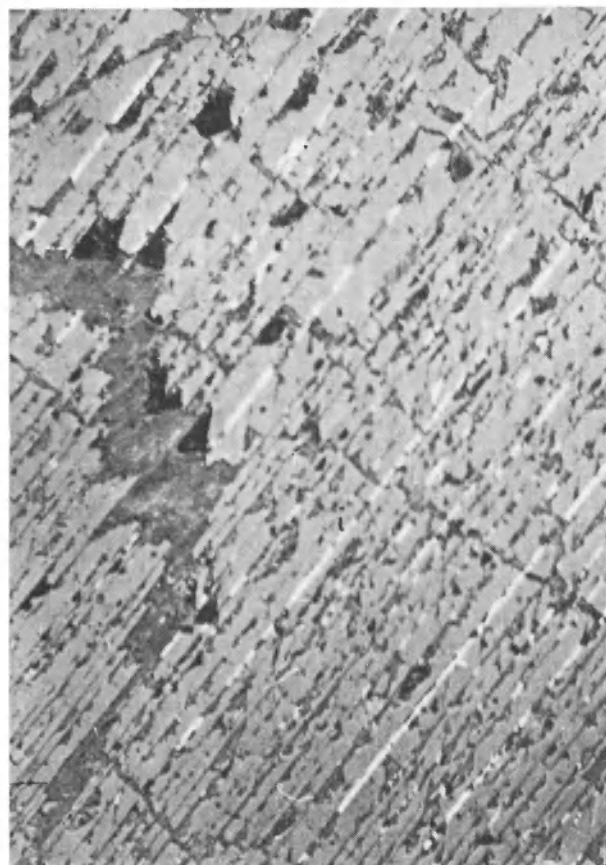
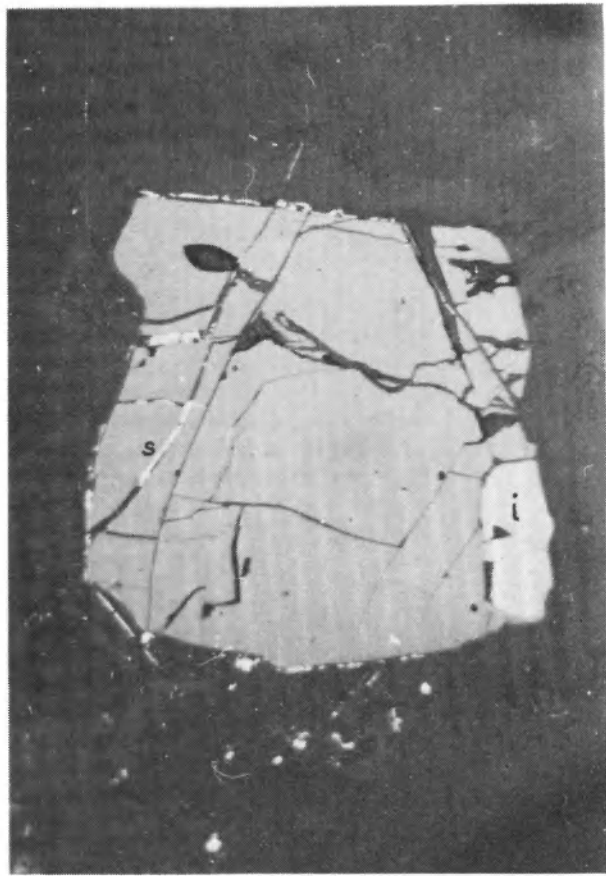
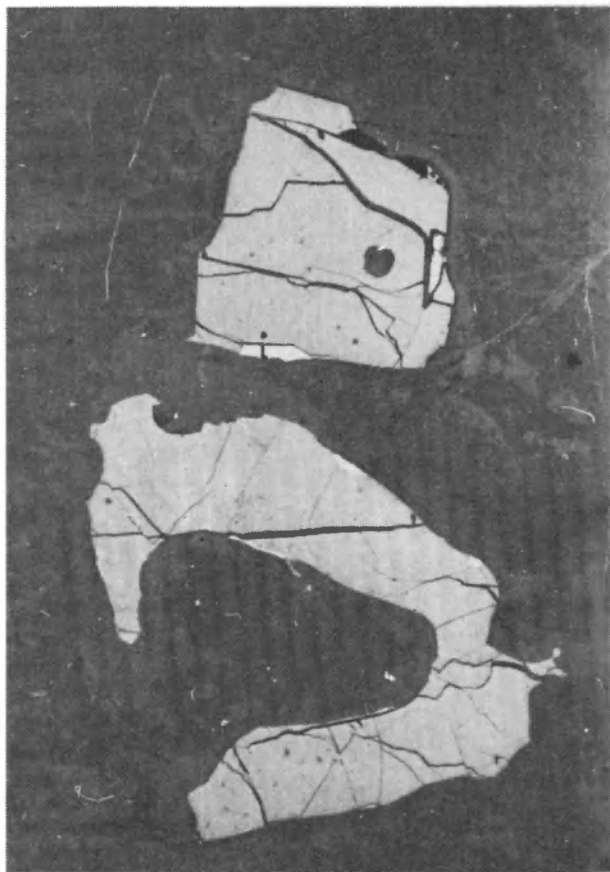
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Ryszard SAŁACIŃSKI - Origin of ore minerals in mafic and ultramafic rocks of the Sowie Mountains block surrounding
 Geneza mineralizacji kruszczowej mafitów i ultramafitów w otoczeniu bloku sowiogórskiego

PLATE II – PLANSZA II

**Ore minerals in gabbros of the Nowa Ruda massif (borehole NR-IG-1)
Minerały kruszczowe w gabrach masywu Nowej Rudy (wiercenie NR-IG-1)**

1. Chromite grain morphology variation; i – ilmenite, $\times 55$, one nicol, reflected light
Zróznicowanie morfologii ziarn chromitu; i – ilmenit, pow. $55\times$, 1 N, światło odbite
2. Sulfide veinlets (s) in chromite; i – ilmenite, $\times 110$, one nicol, reflected light
Żyłki siarczków (s) w chromicie; i – ilmenit, pow. $110\times$, 1 N, światło odbite
3. Ilmenite (i) replaced by rutile (r), $\times 210$, one nicol, reflected light
Ilmenit (i) zastępowany rutylem (r), pow. $210\times$, 1 N, światło odbite
4. Ilmenite laths (i) in pyroxene, $\times 240$, one nicol, reflected light
Listewki ilmenitu (i) w piroksenie, pow. $240\times$, 1 N, światło odbite

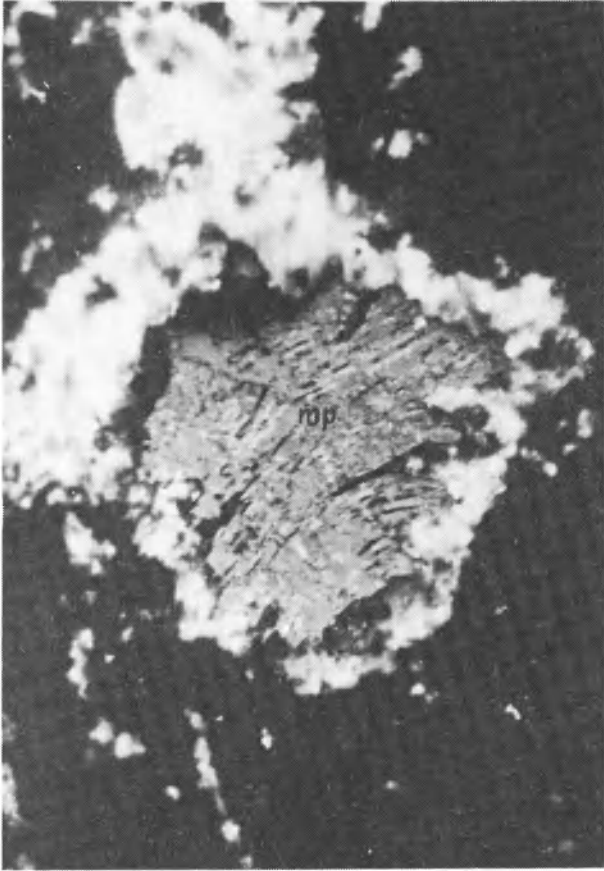


Ryszard SAŁACIŃSKI – Origin of ore minerals in mafic and ultramafic rocks of the Sowie Mountains block surrounding
Geneza mineralizacji kruszcowej mafitów i ultramafitów w otoczeniu bloku sowiogórskiego

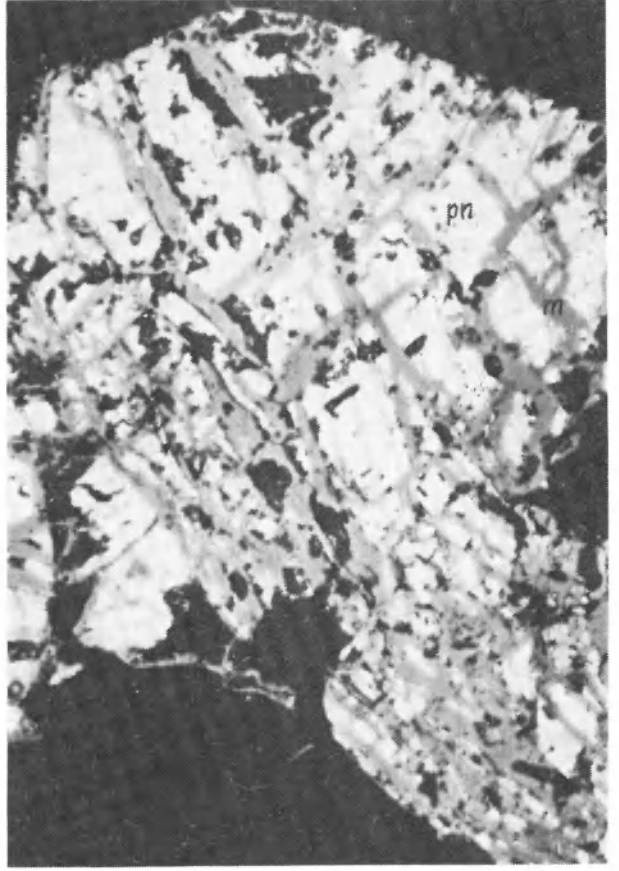
PLATE III – PLANSZA III

**Ore minerals in gabbros of the Nowa Ruda Massif (borehole NR-IG-1)
Minerały kruszczowe w gabrach masywu Nowej Rudy (wiercenie NR-IG-1)**

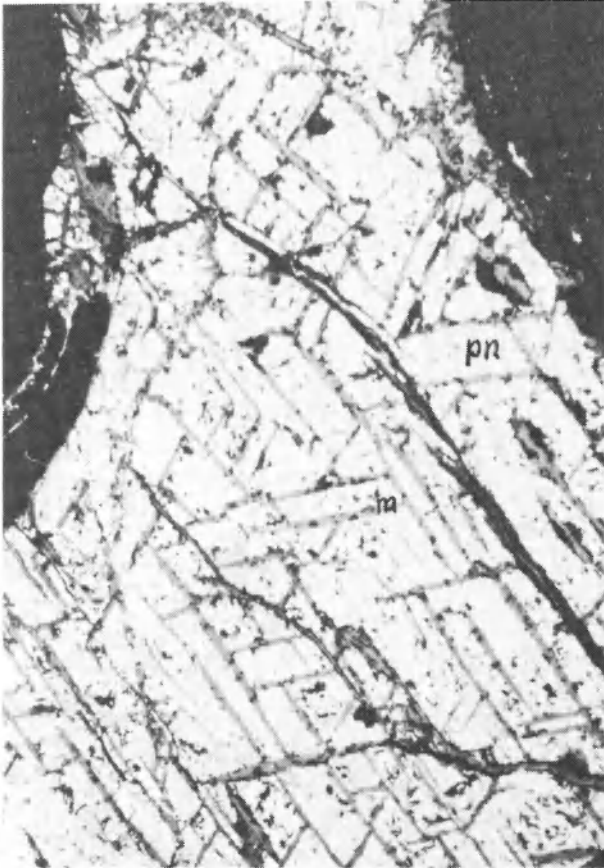
1. Pyrite-marcasite (mp) pseudomorph after pyrrhotite, $\times 210$, nicols crossed, reflected light
Pseudomorfoza pirytowo-markasytowa (mp) po pirotynie, pow. $210\times$, nikole skrzyżowane, światło odbite
2. Pyrrhotite (pn) replacement by magnetite (m), $\times 240$, one nicol, reflected light
Zastępowanie pirotytu (pn) przez magnetyt (m), pow. $240\times$, 1 N, światło odbite
3. Pyrite (pn) pseudomorph after pyrrhotite with magnetite (m) laths, $\times 120$, one nicol, reflected light
Pseudomorfoza pirytu (pn) po pirotynie z listewkami magnetytu (m), pow. $120\times$, 1 N, światło odbite
4. Pyrrhotite (pn) with magnetite (m) laths and pentlandite (pl), $\times 240$, one nicol, reflected light
Pirotyt (pn) z listewkami magnetytu (m) oraz pentlandyt (pl), pow. $240\times$, 1 N, światło odbite



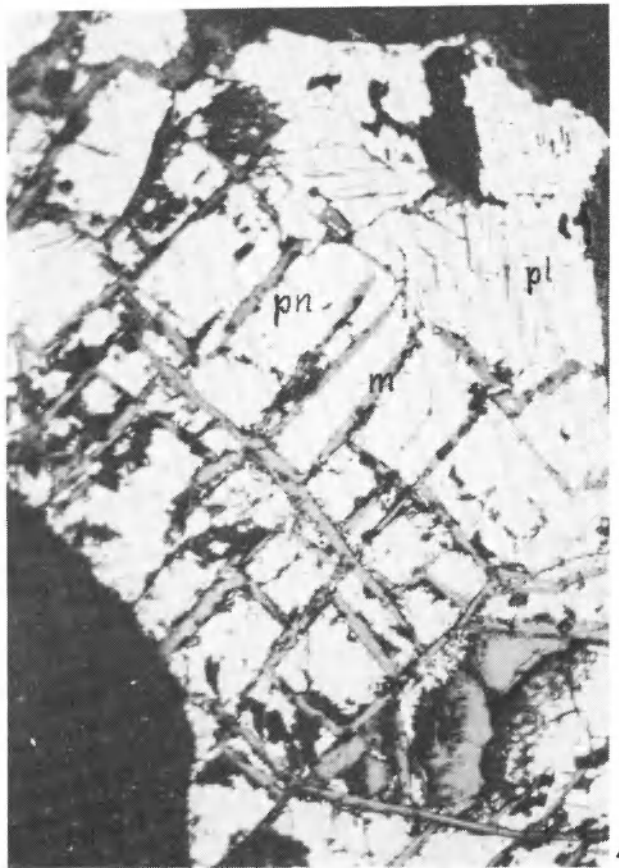
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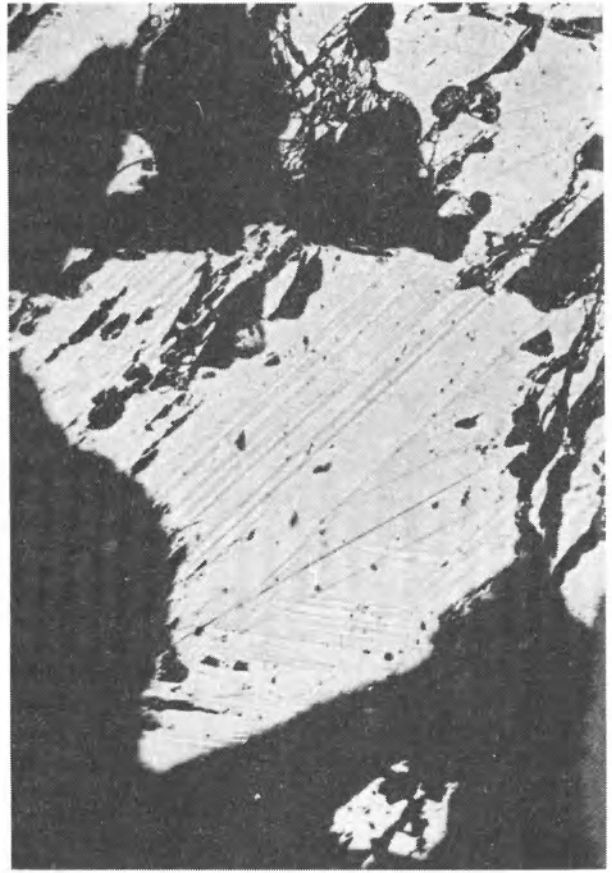
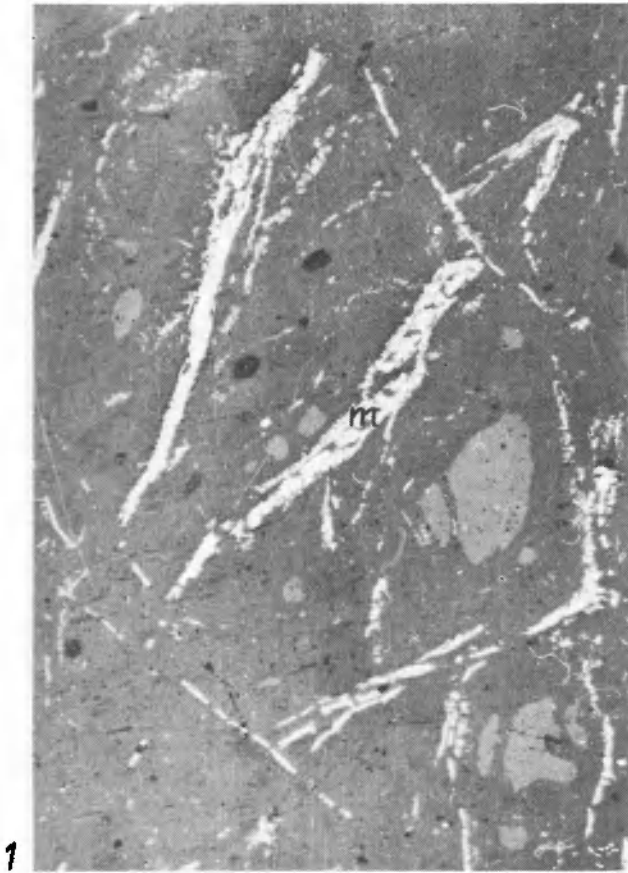
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Ryszard SAŁACINSKI — Origin of ore minerals in mafic and ultramafic rocks of the Sowie Mountains block surrounding
 Geneza mineralizacji kruszczowej mafitów i ultramafitów w otoczeniu bloku sowiogórskiego

PLATE IV – PLANSZA IV

Ore minerals in gabbros of the Nowa Ruda massif (borehole NR-IG-1)
Minerały kruszcowe w gabrach masywu Nowej Rudy (wiercenie NR-IG-1)

1. Magnetite (m) aggregates in a pseudomorph after olivine, $\times 100$, one nicol, reflected light
Skupienie magnetytu (m) w pseudomorfozie po oliwinie, pow. $100\times$, 1 N, światło odbite
2. Pyrrhotite (lighter) with chalcopyrite intergrowths (darker), $\times 240$, one nicol, reflected light
Pirotyt (jasniejszy) z przerostami chalkopirytu (ciemniejszy), pow. $240\times$, 1 N, światło odbite
3. Aggregate of sphalerite (s), chalcopyrite (ch) and pyrite (p) formed after pyrrhotite, $\times 240$, one nicol, reflected light
Agregat sfalerytu (s), chalkopirytu (ch) i pirytu (p) popirotynowego, pow. $240\times$, 1 N, światło odbite
4. Replacement of chalcopyrite (ch) by bornite (b), $\times 240$, one nicol, reflected light
Zastępowanie chalkopirytu (ch) przez bornit (b), pow. $240\times$, 1 N, światło odbite

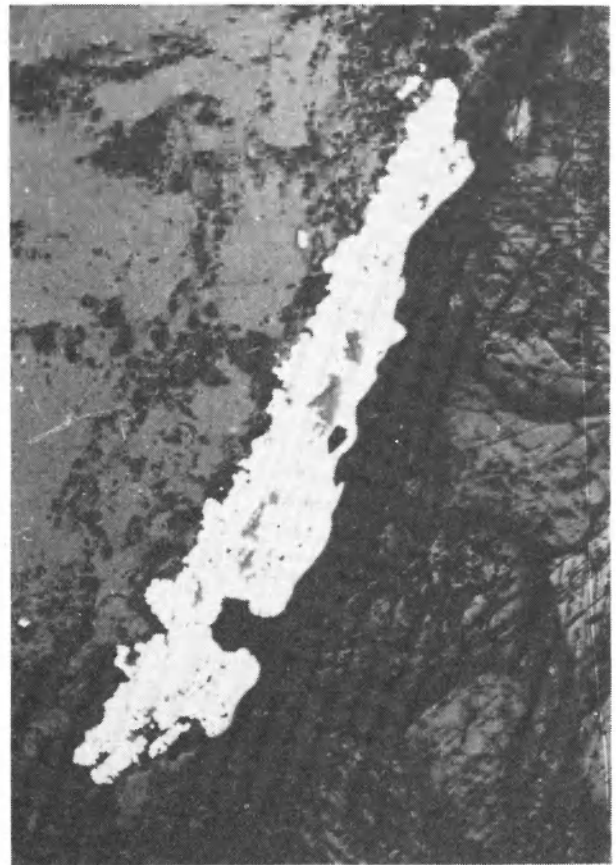
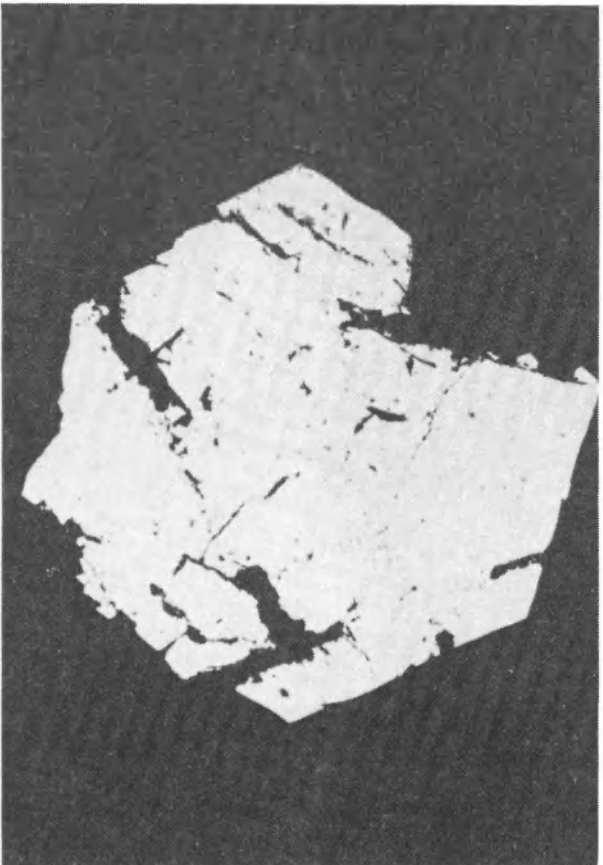
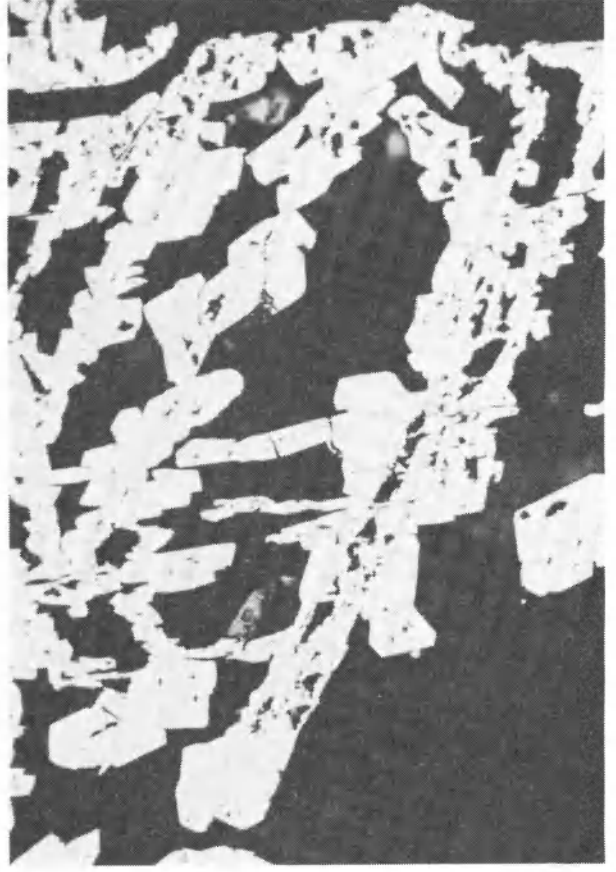
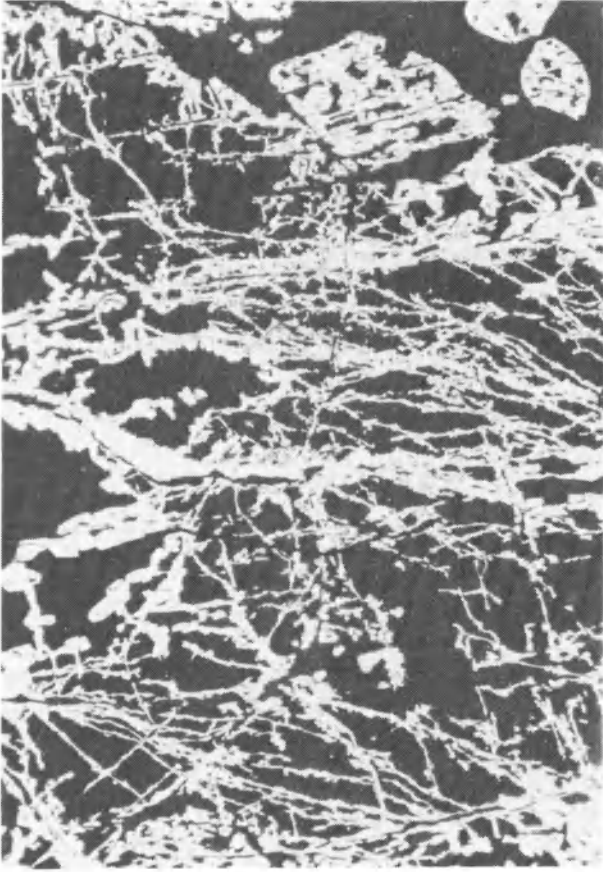


Ryszard SAŁACIŃSKI — Origin of ore minerals in mafic and ultramafic rocks of the Sowie Mountains block surrounding
Geneza mineralizacji kruszczowej mafitów i ultramafitów w otoczeniu bloku sowiogórskiego

PLATE V – PLANSZA V

Ore minerals in gabbros of the Nowa Ruda massif (borehole NR-IG-1)
Minerały kruszczowe w gabrach masywu Nowej Rudy (wiercenie NR-IG-1)

1. Pyrite in form of dendrites and intergranular fissures fillings, $\times 70$, 1 N, reflected light
Piryt w postaci dendrytów i wypełnień szczelin intergranularnych, pow. $70\times$, 1 N, światło odbite
2. A detail of the photograph 1, $\times 240$, one nicol, reflected light
Fragment fotografii 1, pow. $240\times$, 1 N, światło odbite
3. Pyrite metacryst having the pentagonal dodecahedron habit, $\times 70$, 1 N, reflected light
Metakryształ pirytu o pokroju dwunastokąta pentagonalnego, pow. $70\times$, 1 N, światło odbite
4. Pyrite in form of a veinlet, $\times 55$, 1 N, reflected light
Piryt w postaci żyłki, pow. $55\times$, 1 N, światło odbite

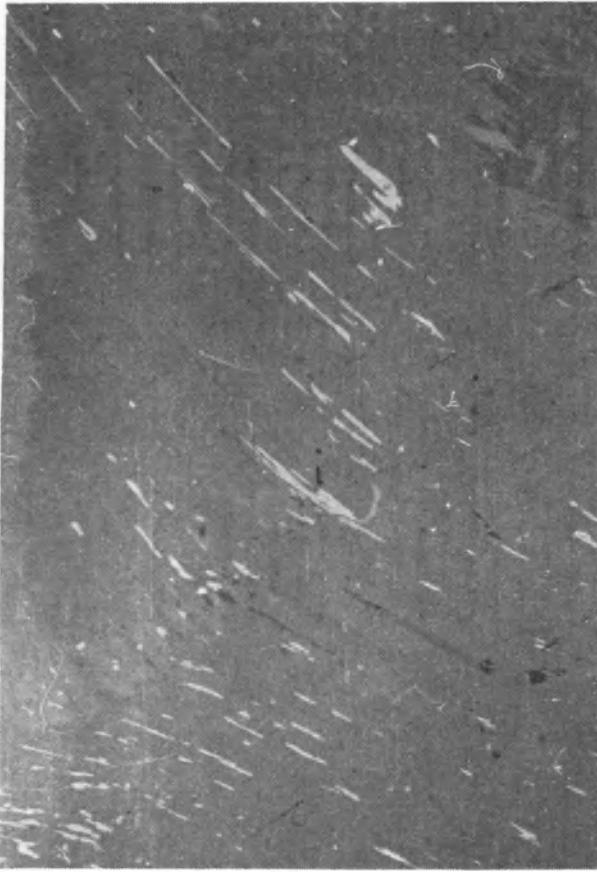


Ryszard SAŁACIŃSKI — Origin of ore minerals in mafic and ultramafic rocks of the Sowie Mountains block surrounding
Geneza mineralizacji kruszcowej mafitów i ultramafitów w otoczeniu bloku sowiogórskiego

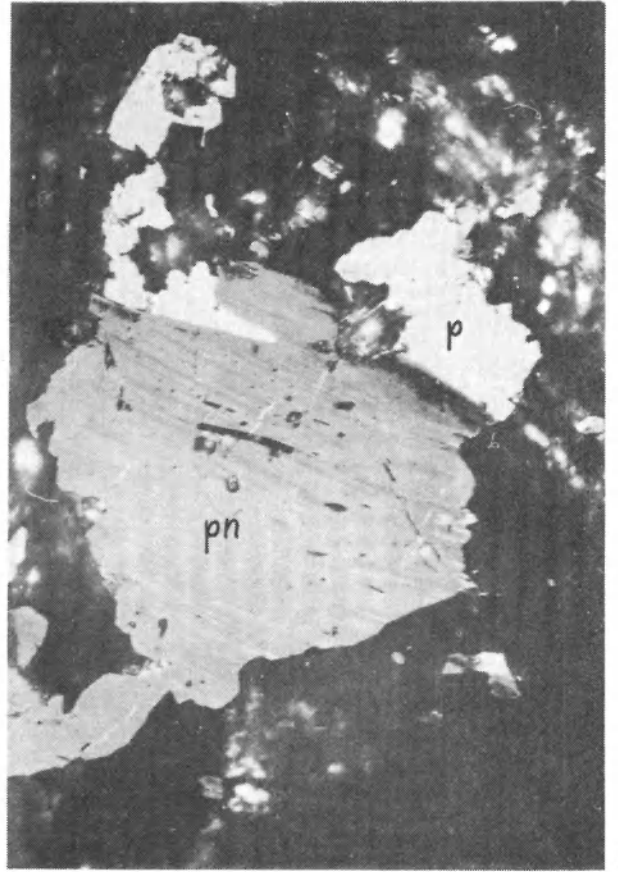
PLATE VI – PLANSZA VI

Ore minerals in gabbros of the Ślęża massif (borehole Żobótka 2a)
Minerały kruszcowe w gabrach masywu Ślęży (wiercenie Żobótka 2a)

1. Ilmenite (i) laths in pyroxene (?), $\times 110$, one nicol, reflected light
Listewki ilmenitu (i) w piroksenie (?), pow. $110\times$, 1 N, światło odbite
2. Pyrite (p) and intergrowths of the monoclinic and hexagonal modifications of pyrrhotite (pn), $\times 240$, one nicol, reflected light
Piryt (p) oraz przerosty modyfikacji jednoskośnej i heksagonalnej pirotytu (pn), pow. $240\times$, 1 N, światło odbite
3. Pyrrhotite (pn) replaced by marcasite (k), $\times 380$, one nicol, reflected light
Pirotyt (pn) zastępowany markazytem (k), pow. $380\times$, 1 N, światło odbite
4. Pyrite metacryst, $\times 240$, one nicol, reflected light
Metakryształ pirytu, pow. $240\times$, 1 N, światło odbite



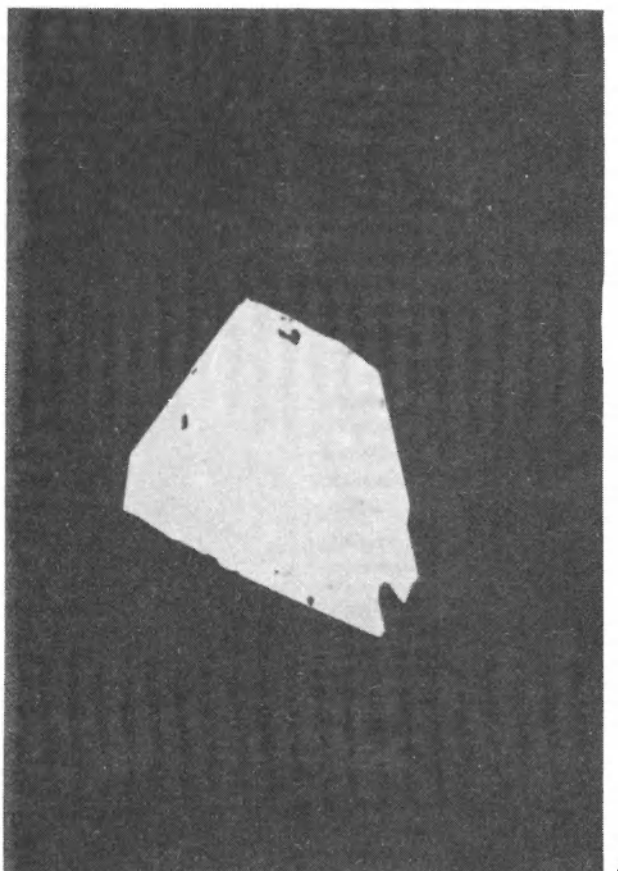
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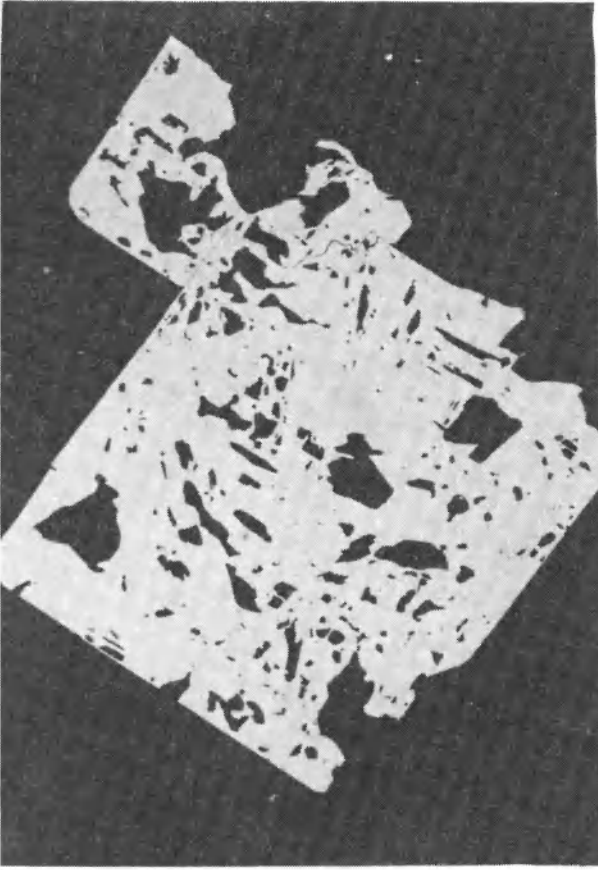
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Ryszard SAŁACIŃSKI – Origin of ore minerals in mafic and ultramafic rocks of the Sowie Mountains block surrounding Geneza mineralizacji kruszcowej mafitów i ultramafitów w otoczeniu bloku sowiogorskiego

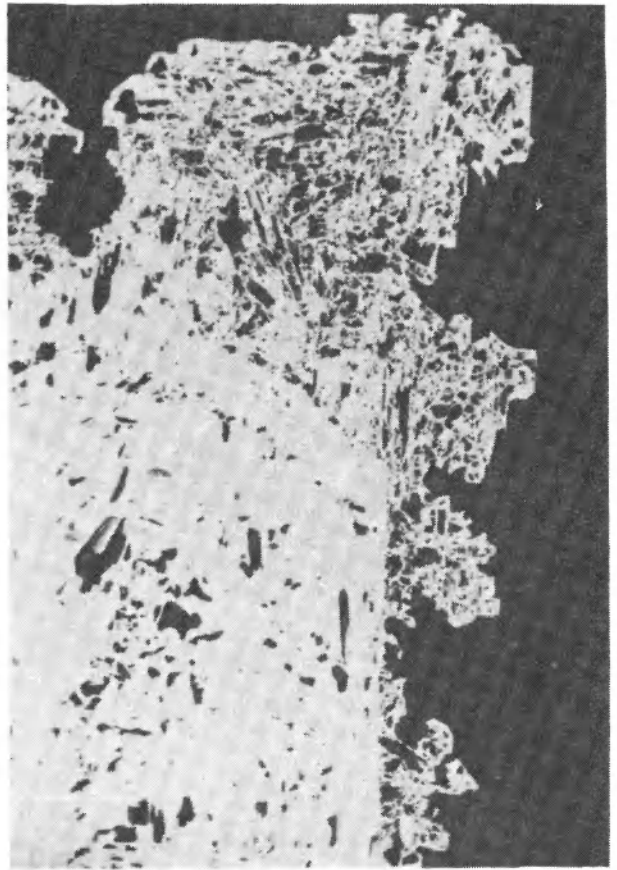
PLATE VII – PLANSZA VII

Ore minerals in gabbros of the Ślęża massif (borehole Sobótka 2a)
Minerały kruszcowe w gabrach masywu Ślęży (wiercenie Sobótka 2a)

1. Pyrite metacryst with silicate minerals intergrowths, $\times 240$, one nicol, reflected light
Metakryształ pirytu z przerosciami mineralów skalotwórczych, pow. $240\times$, 1 N, światło odbite
2. Change of the pyrite metacryst habit from pentagonal dodecahedron to cube, $\times 70$, one nicol, reflected light
Przebudowa metakryształu pirytu o pokroju dwunastokąta pentagonalnego w sześcian, pow. $70\times$, 1 N, światło odbite
3. Replacement of pyrite (p) by limonite (l), $\times 240$, one nicol, reflected light
Zastępowanie pirytu (p) przez limonit (l), pow. $240\times$, 1 N, światło odbite
4. Molybdenite clusters in a quartz vein, $\times 120$, one nicol, reflected light
Skupienia molibdenitu w żyłach kwarcowej, pow. $120\times$, 1 N, światło odbite



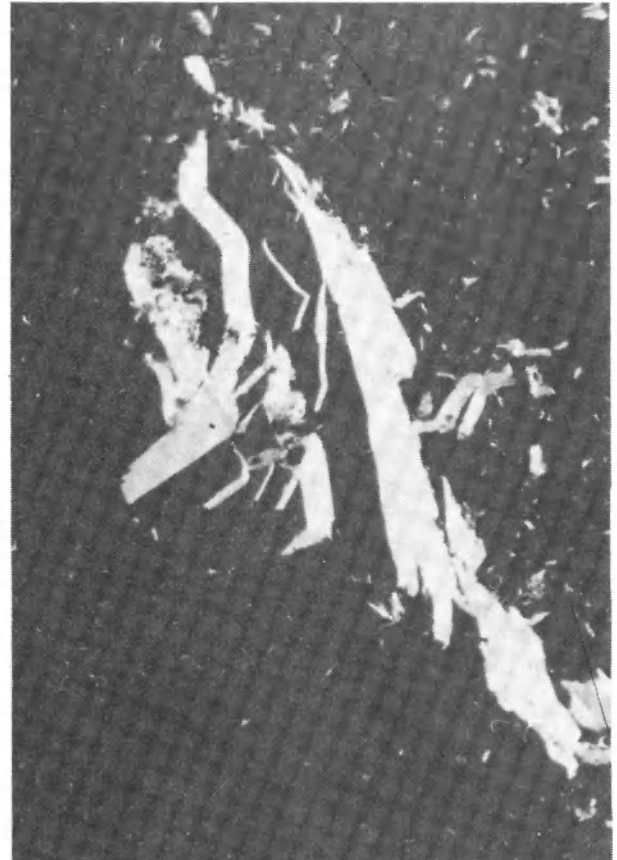
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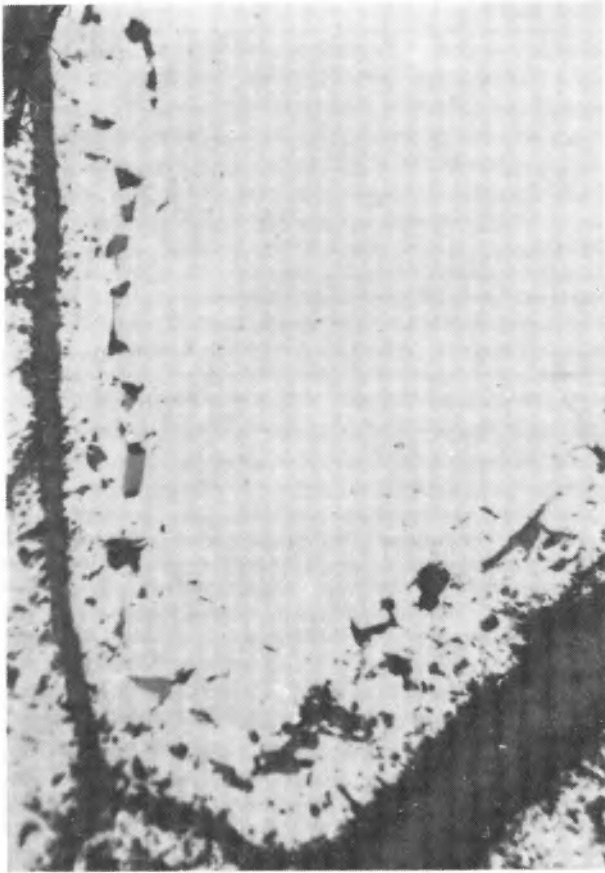
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Ryszard SAŁACIŃSKI – Origin of ore minerals in mafic and ultramafic rocks of the Sowie Mountains block surrounding
Geneza mineralizacji kruszcowej mafitów i ultramafitów w otoczeniu bloku sowiogórskiego

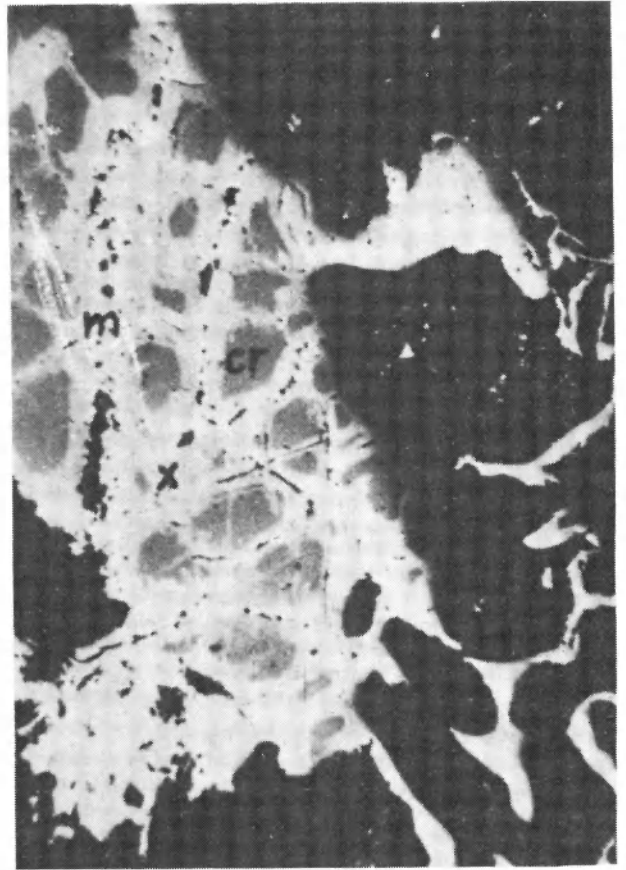
PLATE VIII – PLANSZA VIII

**Ore minerals in serpentinites of the Gogółów-Jordanów massif
Minerały kruszcowe w serpentynitach masywu Gogółów-Jordanów**

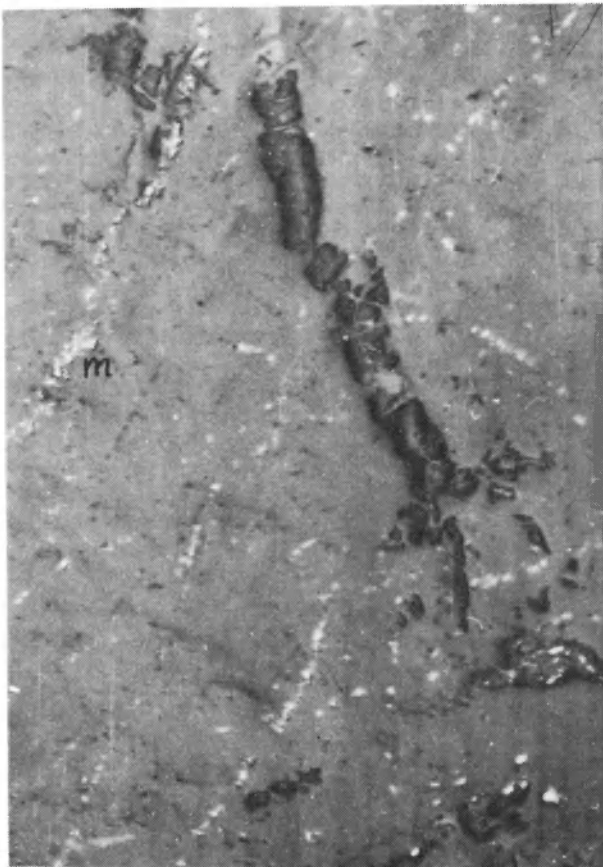
1. Chromite with silicate minerals intergrowths, $\times 120$, one nicol, reflected light, ore deposit at Tapadła
Chromit z przerostami minerałów krzemianowych, pow. $120 \times$, 1 N, światło odbite, złożo w Tapadłach
2. Aggregate of chromite (cr), iron-chromium spinel (x) and magnetite (m), $\times 45$, one nicol, reflected light, borehole at Tapadła
Agregat chromitu (cr), spinelu żelazowo-chromowego (x) i magnetytu (m), pow. $45 \times$, 1 N, światło odbite, Tapadła – wiercenie
3. Magnetite (m) in a serpentine pseudomorph after olivine, $\times 55$, one nicol, reflected light, borehole at Tapadła
Magnetyt (m) w pseudomorfozie serpentynowej po oliwinie, pow. $55 \times$, światło odbite, Tapadła – wiercenie
4. Magnetite (m) from the zone adjacent to veinlets, $\times 430$, one nicol, reflected light, borehole at Tapadła
Magnetyt (m) ze strefy przykontaktowej żyłek, pow. $430 \times$, 1 N, światło odbite, Tapadła – wiercenie



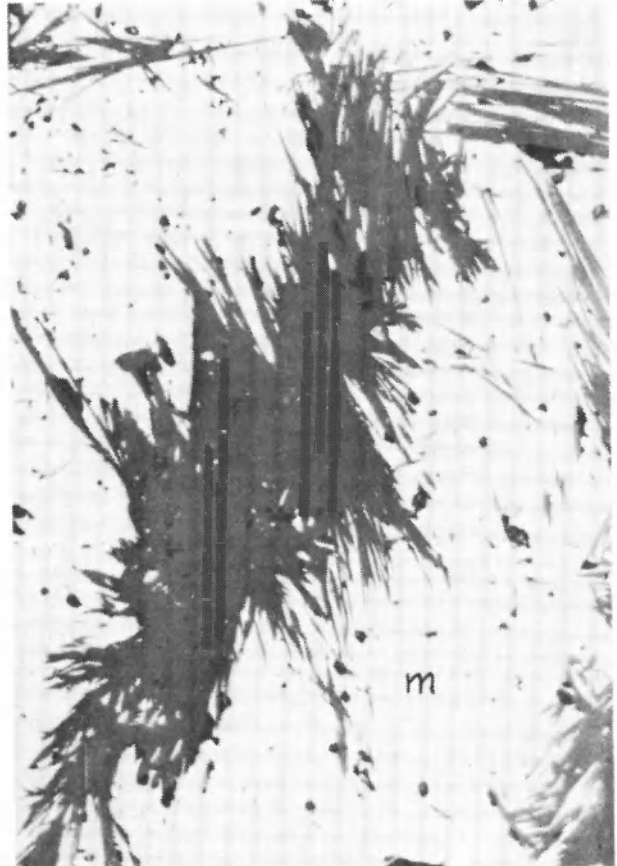
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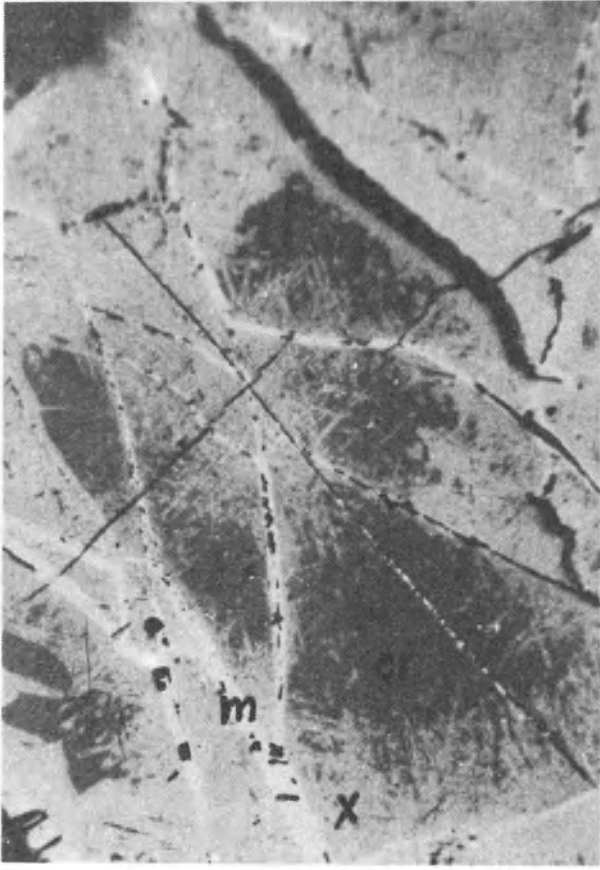
Ryszard SAŁACIŃSKI — Origin of ore minerals in mafic and ultramafic rocks of the Sowie Mountains block surrounding
 Geneza mineralizacji kruszczowej mafitów i ultramafitów w otoczeniu bloku sowiogórskiego

PLATE IX – PLANSZA IX

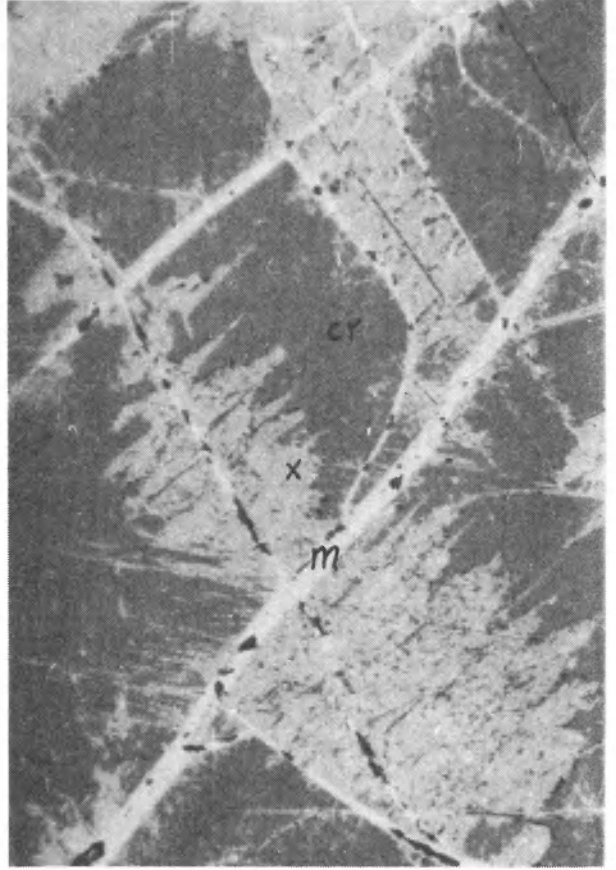
Ore minerals in serpentinites of the Gogołów-Jordanów massif. Replacement of chromite (cr) by iron-chromium spinel (x) and magnetite (m) in various parts of the massif

Minerały kruszczowe w serpentynitach masywu Gogołów-Jordanów. Zastępowanie chromitu (cr) przez spinel żelazowo-chromowy (x) i magnetyt (m) w różnych częściach masywu

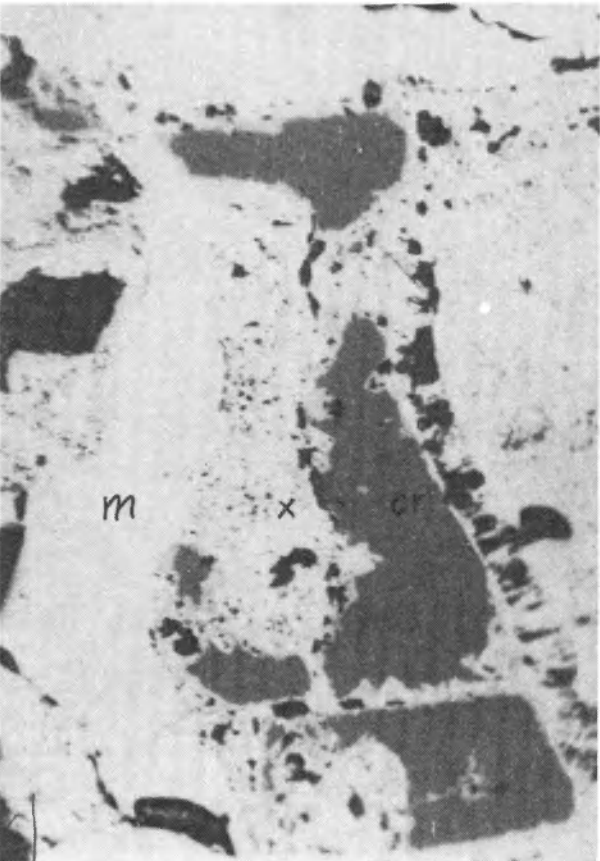
1. A borehole at Tapadła, the replacement most advanced, $\times 250$, one nicol, reflected light
Wiercenie w Tapadłach, proces zastępowania silnie zaawansowany, pow. $250 \times$, 1 N, światło odbite
2. A quarry at Mikołajów, less advanced replacement, $\times 60$, one nicol, reflected light
Mikołajów – kamieniołom, proces zastępowania mniej zaawansowany, pow. $60 \times$, 1 N, światło odbite
3. A quarry at Trzebnik, strong development of magnetite, $\times 60$, one nicol, reflected light
Trzebnik – kamieniołom, silny rozwój magnetytu, pow. $60 \times$, 1 N, światło odbite
4. A quarry at Przemków, laths of magnetite visible both in chromite and in iron-chromium spinel, $\times 60$, one nicol, reflected light
Przemków – kamieniołom, listewki magnetytu widoczne w chromicie i spinelu żelazowo-chromowym, pow. $60 \times$, 1 N, światło odbite



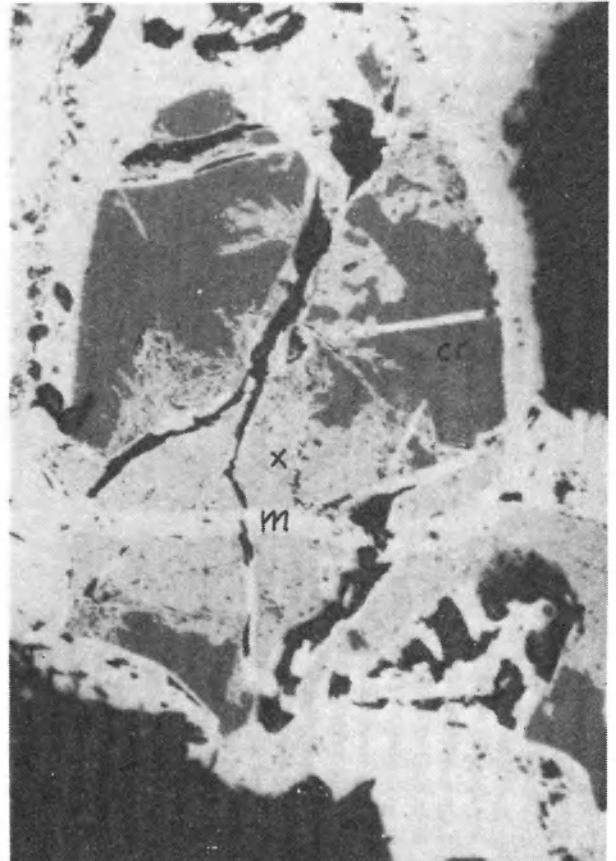
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Ryszard SAŁACIŃSKI – Origin of ore minerals in mafic and ultramafic rocks of the Sowie Mountains block surrounding Geneza mineralizacji kruszcowej mafitów i ultramafitów w otoczeniu bloku sowiogórskiego

PLATE X – PLANSZA X

Ore minerals in serpentinites of the Przygorze massif (borehole P-2)
Minerały kruszcowe w serpentynitach masywu Przygórze (wiercenie P-2)

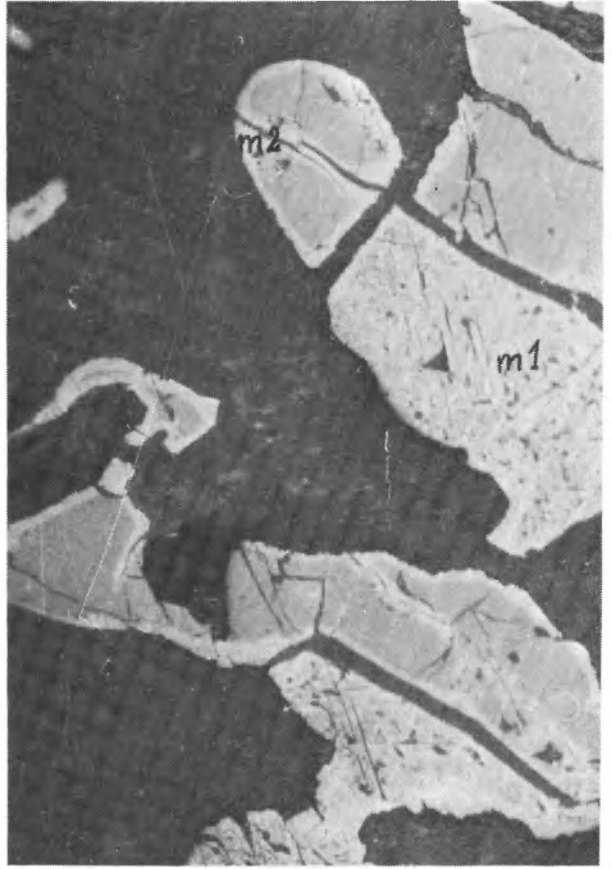
1. Cracked chromite with fissures filled with serpentine, $\times 60$, one nicol, reflected light
Spękany chromit ze szczelinami wypełnionymi serpentynem, pow. $60 \times$, 1 N, światło odbite
2. Chromite with magnetite replacements (m1) and rims (m2), $\times 240$, one nicol, reflected light
Chromit z zastąpieniami (m1) i otoczkami magnetytowymi (m2), pow. $240 \times$, 1 N, światło odbite
3. Ribbon-like magnetite aggregates, $\times 240$, one nicol, reflected light
Wstęgowe skupienia magnetytu, pow. $240 \times$, 1 N, światło odbite
4. Replacement of chromite (cr) by magnetite (m) following the cleavage planes, $\times 240$, one nicol, reflected light
Zastępowanie chromitu (cr) przez magnetyt (m) wzdłuż spekań, pow. $240 \times$, 1 N, światło odbite

All photos taken by the author

Wszystkie zdjęcia wykonane przez autora



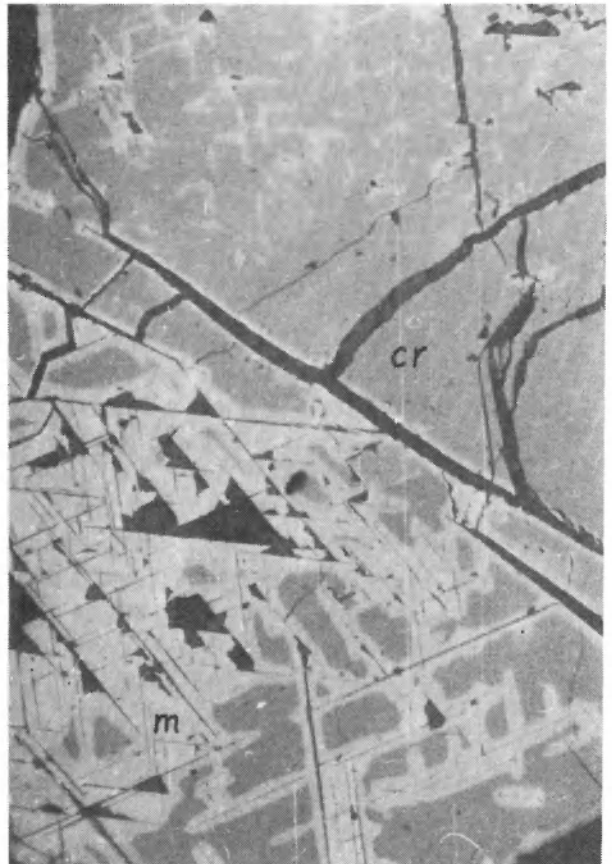
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Ryszard SAŁACIŃSKI – Origin of ore minerals in mafic and ultramafic rocks of the Sowie Mountains block surrounding
 Geneza mineralizacji kruszczowej mafitów i ultramafitów w otoczeniu bloku sowiogórskiego