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Metallogeny of the Baltic Shield

Proceedings of the symposium held in Helsinki, Finland, June 12—21, 1978

Edited by Jaakko Siivola

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FOREWORD — by **Aimo K. Mikkola**

The scope of the International Geological Correlation Program (IGCP) Project 91 »Metallogeny of the Precambrian» is wide. There are two principal paths to approach the matter: either according to the scientific content or the area concerned. On the first basis the major trends of the project have been determined as follows (Moscow meeting in 1974):

— Precambrian mineral deposits and geological environment of their formation,
— the role of sedimentation and organic matter in the Precambrian ore formation,
— the role of magmatism and metamorphism in the Precambrian ore formation, and
— ore bearing tectonic structures of the Precambrian.

On the second basis a Precambrian Shield forms a logical entireness for a metallogenic research. These both lines have been followed in the studies under the Project n:o 91 but basically, however, so that the broad trends mentioned above have been divided in smaller subgroups and dealt with rather independently in different Shield areas.

During the XXV International Geological Congress in Sydney 1976 Academician Sidorenko suggested to have the next meeting of the working group of the Project 91 on the Baltic Shield. This was based on the fact that the area of the Baltic Shield is rather limited and it is geologically well known. In addition, there was a possibility to bring together the results of studies in four countries on the Shield. Consequently the director of the Geological Survey of Finland Dr. H. Stigzelius invited the 1978 meeting of the working group to Finland with the intention that Finland should organize a symposium in this connection.

The Organizing Committee of the symposium (A. K. Mikkola chairman, H. Papunen, J. Siivola and M. Mäkelä, secretary) selected as main topics:

— Precambrian mineral deposits and geological environment of their formation, and
— the role of magmatism and metamorphism in the Precambrian ore formation.

A presymposium excursion was planned to five mines emphasizing ore deposits associated with basic magmatism and volcanic sedimentary environment. The excursion was held June 11—16 and the symposium June 19—21 at Helsinki University of Technology in Otaniemi.

Altogether 18 papers were invited. The authors presented 16 of them. The metallogeny of the Baltic Shield was object of eleven papers. The rest of them dealt with other areas or other IGCP-projects. It was agreed that a volume of the scientific proceedings of the symposium should be published with the assistance of the Academy of Finland and the Geological Survey of Finland. Director Dr. H. Stigzelius kindly promised to include the proceedings in the Survey's Bulletin. The Organizing Committee is indebted to him.

Representatives of eleven countries participated in the symposium.

Active support to organize the symposium given by the Ministry of Trade and Industry, Ministry of Education, Academy of Finland, Helsinki University of Technology, Geological Survey of Finland, and the Outokumpu and Rautaruukki companies deserves the great gratitude of the Organizing Committee.

The authors are responsible for their papers. The Organizing Committee has done only the editorial work.

METALLOGENY OF THE EASTERN PART OF THE BALTIC SHIELD

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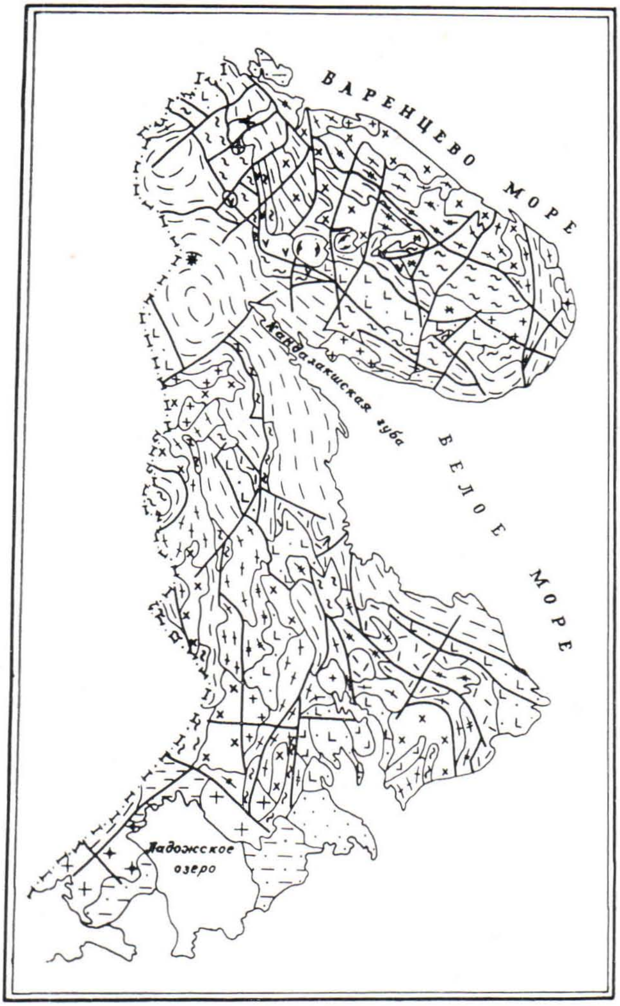
To study the metallogeny of the eastern part of the Baltic Shield we used the results of generalization of the latest geological, geophysical and geochemical data and took these as a basis for compiling maps on a scale of 1:1 500 000. The complete series includes metallogenic, geological formation, block structure, abyssal structure, metamorphic facies, geochemical zonation, petromagnetic and petrodensity maps. The editors of the maps are A. V. Sidorenko, T. V. Bilibina, A. O. Belyaev, K. D. Belyaev, P. V. Bylinskiy, L. V. Grigoe'eva, N. B. Dortman, M. A. Korsakova, V. L. Negrutsa, Yu. I. Rabinovich, D. V. Rundkvist, A. A. Smyslov. The main features of metallogeny of the eastern part of the Baltic Shield are considered in accordance with the complete set of maps.

The tectonic position of this region is determined by the block structure of the Earth's crust and the character of the interblock boundaries. On the whole the Baltic Shield is characterized by unconformities and boundaries of the Baltic Shield, the Karelia—Kola region is bordered by the Timan folded area in the north, the other boundaries are morphotectonic owing to the sedimentary cover of the Russian platform in the south and neotectonic, on account of the marine areas of the Barents and White seas. The inner structure of the eastern part of the Baltic Shield is determined by the predominant constituents of granite — metamorphic layer — structural-facies complexes.

According to A. Y. Semenov, a structural-formation complex consists of associations of geological formations formed in a definite type of geological regime and under specific physico-geographical conditions.

The geological structure of this region is represented by 5 structural formation complexes: I) Archean, older than 2 800 Ma; II) Early Protetozoic (Lopiy and Sumiy) older than 2 200 Ma; III) Middle Proterozoic, older than 1 800 Ma; IV) Late Proterozoic, older than 600 Ma; V) Paleozoic, younger than 550 Ma. The boundaries between structural-facies complexes are controlled by regional and angular unconformities. Each complex is characterized by definite associations of geological formations, metamorphism, dislocations caused by folding and those with a break in continuity. According to the terminology employed in the Soviet Union each complex is distinguished by the type of geotectonic regime as geosyncline (AR), protogeosyncline (PR_1), proto-orogenic (PR_2), protoactivations and platforms. In Paleozoic a regime of tectonic activation separated by a long period of tectonic quietness is superimposed on the craton structures of the Shield (Fig. 1).

Geological formations of structural complexes form the basis of geostructural areas extending to Finland, Sweden and Norway; these include the Archean folded structures of the Belomor area, the Svecofennian Early—Middle Protero-



I. Geological associations of the Palaeozoic structural-associational complex: 1 = agpaitic nepheline syenite associations and relicts of sandstone-alkaline basalt association; 2 = ultrabasite and alkaline rock plus carbonatite association.

II. Geological associations of the Middle Proterozoic structural-associational complex: 3 = sandstone-argillite and conglomerate-argillite-sandstone associations (undivided).

III. Geological associations of the Middle Proterozoic structural-associational complex: 4 = anorthosite gabbro—rapakivi granite association complex; 5 = alkaline granite association; 6 = granite association; 7 = aleurolite-sandstone association; 8 = fiyschoidic sandstone-argillite association; 9 = aleurolite-picrite-dolerite and dolerite gabbro association and dolomite-schist-dolerite association (undivided); 10 = conglomerate-sandstone-dolerite association; 11 = porphyry-dolerite association.

IV. Geological associations of the Early Proterozoic structural-associational complex; 12 = granite, migmatite-granite, charnockite-granite associations (undivided); 13 = anorthosite-gabbro association; 14 = keratophyre spilite, iron ore-quartzite and graywacke-porphyry-dolerite associations (unvidided).

V. Geological associations of the Archean structural-associational complex: 15 = migmatite-plagioclase granite-granite association; 16 = charnockite-diorite-granite association; 17 = migmatite-plagioclase granite association; 18 = gneiss, amphibolite-gneiss, crystalline schist etc. associations (undivided);

19 = fault dislocations.

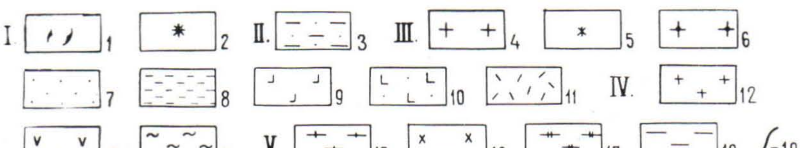


Fig. 1. Geological associations of the eastern part of the Baltic Shield.

zoic folded belts of Kola and Karelia and the Savoladoga Middle Proterozoic folded belt.

In the process of long cratonization, geostructural areas were transformed into megablocks composed of blocks of higher orders. The megablocks (Kola, Belomor, Bothnia, Karelia and Ladoga) which were established on account of their abyssal structure data are divided by sutural zones: Pechenga—Imandra—Varzuga and Kola—Vygozero—Kozhozero. These main

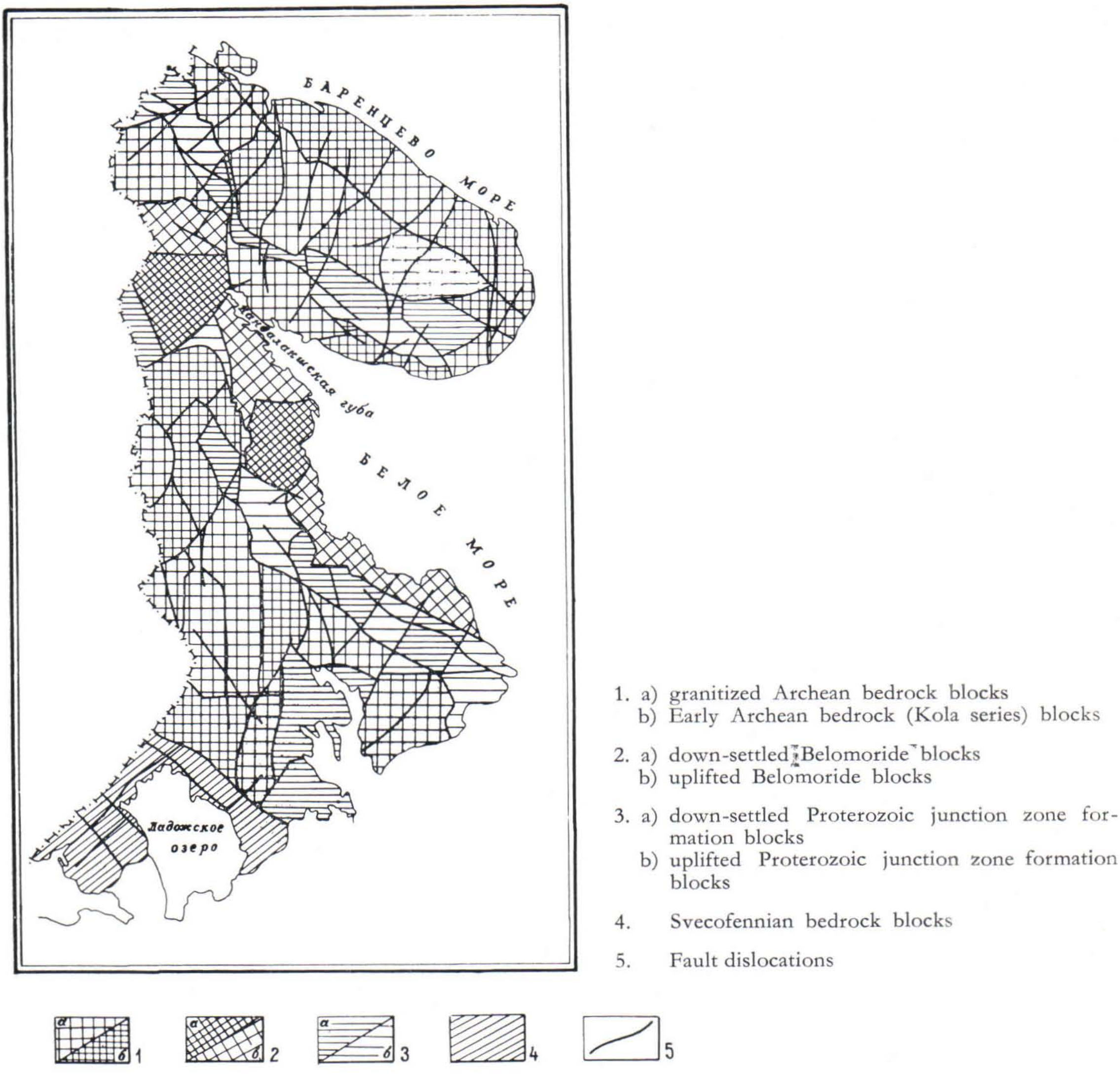


Fig. 2. Block structure of the eastern part of the Baltic Shield.

block structures are subdivided into blocks of the second and third orders. The Belomor block consists mainly of geological formations of the Archean progeosyncline complex; in the Kola and Karelia block Archean and Early Proterozoic pro- and protogeosyncline geological formations are united in sutural zones, and in the Ladoga synclinorium proto-orogenic and proto-geosyncline formations of the Middle Proterozoic are combined.

Derivations of processes of protoactivation and tectonomagmatic activation have been constructed within the Kola and Ladoga megablocks (Fig. 2).

The boundaries between the blocks are determined by three types of structures: 1) sutural zones located along crust-mantle fractures; 2) interblock crust-mantle fractures including the fractures that control the position of shore lines (Karpinsky fracture, Gorlo White Sea; 3) crust and crust-mantle fractures delimiting blocks of second and higher orders.

The rocks in the geological formations of the various structural complexes were regionally metamorphosed under conditions of low and high-pressure facies. Two types have been established: (A) andalusite-sillimanite; and (B) kyanite-sillimanite. The rocks were metarmorphosed under the conditions of the known facies although many of them also under those of various subfacies. In the evolution of the processes of metamorphism in the eastern part of the Baltic Shield there are three large cycles subordinated to the following epochs of diastrophism:

1) older than 3 500—3 300 Ma, PreReboly,
2) older than 2 800—2 700 Ma, Reboly
3) older than 1 900—1 700 Ma, Early Svecofennian.

The rocks of these cycles developed to various degree in different blocks of the shield.

In the eastern part of the Baltic Shield the useful minerals are mainly those of iron, nickel, copper, phosphate and aluminiferous ores in unique deposits of nepheline and apatite in agpaitic syenites and muscovite and ceramic pegmatites and phlogopites. Moreover, there are manifestations of tin, molybdenum and tungsten mineralization, pyrite and other useful minerals.

In the metallogenic analysis we classified the ore formations by means of the formation principle based on the following definition of an ore formation: »natural communities or groups of ore deposits with similar metal and mineral compositions in ores and modes of accociation with geological formations».

The position of an ore formation in the course of geological evolution in the region and the genesis of a mineralization is controlled in many respects by the position and geological type of related geological formations. Eighteen formation types of deposits were established in the eastern part of the Baltic Shield. Some of them occur repeatedly throughout the complex geological evolution of the shield, which brings the total number of ore formations up to 25. The formations are often complex and polychronic and include several mineral types. Typical examples of ore formations in this region are iron-ores in quartzites; sulphide copper-nickel ores in basic rocks or ultrabasites; iron-zinc-tin ores in skarns and eposkarn metasomatites; rare-metal-phosphate-iron ores in alkaline and ultra-basic rocks and carbonatites; rare-metal-alumina-phosphate ores in nepheline syenites; alumina (kyanite) in shales, muscovite and ceramic pegmatites, etc.

According to V. M. Goldschmidt's classification based on the geochemical features, the ore formations belong to siderophile, chalcophile and lithophile groups. The majority of formations in this particular regime belongs to the united siderophile-chalcophile group; the iron-ore formation in quartzites belongs to the siderophile group; and the phosphate, alumina, micaceous and rare-metal formations belong to the lithophile group. The siderophile-lithophile association is characteristic of complex deposits belonging, for example, to iron-ore-phosphate and phlogopite groups in alkaline-ultrabasic rocks and carbonatites.

The important feature of various types of ores is their complexity in formation characteristics. Thus uniting not only the associations of useful minerals in one formation, but the combinations of ore formations of various genetic types.

The variety of genetic types within deposits is characteristic of the metallogeny of the eastern part of the Baltic Shield.

Thus formations of the siderophile group belong to parametamorphic (iron) and magmatic types (iron, titanium); the group of the sidero-chalcophile formations mainly includes deposits of magmatic (nickel, copper) rheometamorphic type (cobalt, copper, lead, zinc) as well as hydrothermal-metasomatic and hydrothermal deposits (tin, tungsten, molybdenum, lead, zinc).

Volcanic prosesses played a key role in the formation of deposits of the siderophile-chalcophile group. The deposits of various genetic types from para- and orthometamorphic (kyanite schists, micaceous pegmatites) to magmatic, including pneumato- and hydrothermal-metasomatic deposits (apatite, nepheline, phlogopite etc.), are characteristic of the formations in the lithophile group.

The accumulation of local concentrations of ore metals in deposits is closely connected with the migration of ore matter in the Earth's crust, particularly in geological formations that show geochemical specialization.

The study of the migration of ore matter as shown on the geochemical map allowed the following main components of the dispersion and concentration of metals in the Earth's crust to be established:

1) regional geochemical background related to the content of ore elements in rocks close to clarke,
2) geochemically specialized geological formations with maximum content of syngenetic accumulation of metals in comparison with clarke,
3) superimposed geochemical anomaly zones connected with the redistribution of ore elements in the course of metamorphism and hydrothermal metasomatism,
4) local epigenetic concentrations of metals (ore manifestations, deposits) with maximum level of accumulation of elements in the Earth's crust.

The spatial distribution of various forms of concentration of minor elements determines the position of geochemically specialized geostructures—geochemical areas and belts differing in the degree of differentiation on the matter.

The evolution of processes of ore-formation in time is determined by the regular change of metallogenic and geochemical epoch reflecting in their turn in the type of geotectonic regime. In the geological evolution of the eastern part of the Baltic Shield six metallogenic epochs have been established: Archean, Early Proterozoic, Middle Proterozoic, Middle—Late Proterozoic, Paleozoic and Mesozoic—Cenozoic.

The Archean metallogenic epoch of siderophile specialization is characterized by restricted degrees of mineralization and small number of useful minerals.

The main metal is iron in quartz-hypersthene magnetite ores and iron quartzites. Geochemically specialized formations have increased clarkes of Fe, Cr, Ti, Sc, rarely Ni and Cu.

The Early Proterozoic epoch of the siderophile-chalcophile and lithophile profile includes three formation series:

1) leading ore formations associated with the process of sedimentary-volcanogenic lithogenesis, crust of weathering and metamorphism of amphibolite facies (iron quartzites, pyrite ores, kyanite schists). The age of the ores is 2 500—2 300 Ma and 2 000 Ma.
2) formations related to the processes of subsilicic-ultrabasite intrusive magmatism (copper-nickel deposits in basic rocks and ultrabasites, iron-titanium deposits in gabbroids). The age of the ores is 2 500 Ma and 1 870—1 820 Ma.
3) formations associated with the processes of granitization (molybdenum deposits in granites and feldspathic metasomatites). The age of the ores is 2 100 Ma.

Geochemical specialization of geological formations is a mixed sidero-chalco-lithophile one with a predominant sidero-chalcophile association.

The Middle Proterozoic epoch of chalcophile-lithophile profile also includes three formation series:

1) formations associated with the sedimentary-volcanogenic lithogenesis, weathering crust and metamorphism of epidote-amphibolite and greenschist facies (cobalt-copper ore deposits in sandstones and shales, copper-ore deposits in conglomerates and sandstones). The age of the ores is 2 200 Ma.
2) the leading formations related to the subsilicic ultrabasite intrusive magmatism (sulphide copper-nickel deposits in basic rocks and ultrabasites). The age of the ores is from 1 950—1 850 to 2 200 Ma.
3) formations connected with the granite, alkaline-granite and gabbro-granite intrusive magmatism (iron-zinc-tin deposits in skarns and aposkarn metasomatites; molybdenum deposits in apogranites and greisens, tungsten deposits in skarns and metamorphism under various facies conditions (muscovite and ceramic pegmatites). The age of the ores is 1 950—1 750 Ma.

Geochemical specialization of geological formations is camouflaged by associations of element admixtures.

The Middle—Late Proterozoic epoch of lithophile profile is weakly displayed. It includes the molybdenum occurrences in quartz veins and greisens and the lead-zinc-vein mineralizations not directly connected with magmatism. The age of the ores is 1 020—650 Ma.

The Paleozoic epoch of siderophile-lithophile profile is characterized by a sharp increase in the intensity of the mineralization process expressed in the formation of complex ores related to the alkaline-basic alkaline—intrusive magmatism (apatite, nepheline, phologopite, magnetite, etc). The age of the ores is 400—290 Ma.

The deposits associated with the weathering crust and alluvial sedimentogenesis (vermiculite, francolite, kaolinite clays, etc.) belong to the Mesozoic—Cenozoic epoch. The general scheme of evolution of ore formation related to the change in composition of sedimentary-volcanogenic and intrusive formations in time as well as the mineralizations accompanying them allow three types of blocks and zones to be established:

1) homodrom evolution series;
2) antidrom series; and
3) unstable or mixed character of evolution.

In this region the evolution trend is expressed in the general tendency of basicity—acidity of magmatic formations and associations of ore elements to change and also in the increase in alkalinity in the homodrom series.

The evolution of metallogenic epochs became somewhat shorter in duration from the Archean to younger epochs; at the same time orogenesis increased in intensity from the Archean to the Middle Proterozoic with a subsequent culmination in the Paleozoic and enhancement of the degree of differentiation and contrast of geological and ore forming processes.

The metallogenic zonation of the eastern part of the Baltic Shield obtained from the complex analysis of data of geological, geophysical, and geochemical maps allowed the metallogenic regions, structural-metallogenic zones, ore regions and ore knots to be established.

The metallogenic areas Kola, Belomor, Ladoga and Karelia, which correspond in broad outline to megablocks established by geophysical data and geochemical provinces differ in some important features. The Kola area is characterized by a crust of varying thickness. It underwent structural rebuilding during the Middle Proterozoic and Paleozoic under the conditions of Proto-orogenic regime and the regime of tectono-magmatic activation. Geochemically this area specialized in siderophile-chalcophile elements from the early to late metasedimentary-volcanogenic and intrusive formations; some of them even specialized in nickel.

In some parameters the Karelia metallogenic area is in an intermediate position between the northern and southern areas and is characterized largely by siderophile-chalcophile specialization (Fe, Cu, Ni, pyrite).

Geochemically the Ladoga metallogenic area or Ladoga arch is specialized in Sn, Mo, Zn, W, Ti. The rebuilding of the crust took place the Middle Proterozoic under the conditions of evolution of protogeosyncline regime and protoactivation.

In geochemical characteristics and structure of the crust the Belomor area is similar to the Ladoga area; it differs, however, in the type of rebuilding in the Middle Proterozoic. This rebuilding was expressed in the formation of tectono-metamorphic zones within which the processes of orthometamorphism (muscovite and ceramic pegmatites) were broadly developed. The formation of the Belomor arch was accompanied by the evolution of the rift zone in Svecofennian time.

Classification of the structural-metallogenic zones is given:

a) by metallogenic specialization,
b) by appurtenance to various metallogenic epochs,
c) by appurtenance to concrete tectonic structures.

The zones are subdivided into 5 types depending on their specialization in metals of siderophile, siderophile-chalcophile and lithophile groups, including non-metalliferous industrial minerals, and also on their complex character.

The direct connection with metallogenic epochs is provided by the age of the ore formations that determine the type of zone:

1) siderophile-iron ore and iron-titanium ($AR—PR_1$);
2) siderophile-chalcophile — copper-nickel, lead-molybdenum—copper-ore, cupreous ($PR_2—PR_3$);
3) lithophile and chalcophile-tungsten, molybdenum—tungsten, molybdenum ($PR_2—PR_3$);
4) complex siderophile-lithophile-iron-zinc-tin -ore, iron-apatite-phlogopite, nepheline-apatite; chalcopyrite PR_1, PR_2, PZ;
5) lithophile deposits of nonmetallic type muscovite-bearing and kyanite (aluminous) ($PR_1—PR_2$).

The zones are subdivided into intra- and interblock zones with respect to the elements of block and abyssal structure in the eastern part of the Baltic Shield. Within the limits of blocks and sutural mobile zones they are confined to systems of fold or block structures of higher orders bordered by fractures.

I. The intrablock zones include three groups:

1) iron, chalcopyrite and kyanite deposits corresponding to the Early Proterozoic synclinoria that divide the megablocks or are situated within the Archean blocks-anticlinoria;
2) muscovite-bearing and ceramic zones confined to the systems of blocks of the polymetamorphosed Archean rocks within the limits of tectono-metamorphic zones;
3) copper-nickel, molybdenum-copper and cupreous zones situated within the limits of sutural mobile zones (in structures of synclinorium type) and also confined to the chains of graben synclines within the megablocks.

II. The interblock zones are subdivided into:

1) copper-nickel and iron-titanium zones confined to the systems of fractures demarcating the sutural mobile zones and the adjoining Archean blocks;
2) tungsten, molybdenum, tin zones confined to the systems of open fractures of areas of protoactivation splitting blocks and megablocks of various orders;
3) complex iron-nepheline-apatite and apatite-phlogopite zones in more active parts of the craton in Paleozoic.

The position of the deposits is controlled by intrusive massifs; the boundaries of the zones, as a rule, are conditional.

In accordance with the establishment of three types of sections in the super-basaltic layer of the Earth's crust, a single granitic layer (sial), a single diorite layer (simatic) and a double intermediate layer including sial and simatic layers, in morphology corresponding to the Mohorovičič and Conrad surface, the following characteristics of structural-metallogenic zones were also obtained.

The intrablock zones correspond, according to their position, to intermediate and sial types of crust. They are sometimes confined to the slopes of dome-shaped uplift of Mohorovičič surface (pyrite zones) or to narrow troughs dividing them (micaceous zones). Copper-nickel interblock zones are confined to the blocks with intermediate or sial type of crust. The Pechenga zone correponds to a bend in the Mohorovičič surface where the crust is of maximum thickness (41—42 km), and the Onega zone corresponds to a dome-shaped uplift where the crust is from 32 to 37 km thick.

Interblock zones differ sharply from intrablock ones in the stratified structure of the Earth's crust. Copper-nickel zones are confined to the blocks with simatic and intermediate type of crust and associated with the boundary tectonic sutures.

Rare-metal zones connected with open fractures do not depend on the abyssal structure of the Earth's crust.

The complex zones and ore regions in areas of Paleozoic activation are distinctly associated with the blocks of simatic and intermediate type of crust and with boundary tectonic sutures. The Khibiny—Lovozero zone intersects the Central-Kola block of simatic type, where it is confined to the bend in the Mohorovičič surface diving dome-shaped uplifts at a depth of 39 km.

An additional basis to check the authenticity of establishment of the metallogenic areas, structural-metallogenic zones and ore regions is provided by the generalization of the information concentrated in set of maps.

The following direct and indirect features may be used to establish specific metallogenic zones;

1. Regional metamorphism of zones:

a) rocks of ore-bearing and enclosing geological formations, metamorphosed under the conditions of greenshist facies up to low-temperature subfacies of epidote-amphibolite facies;

b) rocks of ore-bearing formations of low-pressure Sill-Gr-Bi-Ort and Sill-Gr-Cord-Ort subfacies of amphibolite facies;

c) superposition of amphibolite facies on rocks of ore-bearing formations metamorphosed in granulite facies (polymetallic formations);

d) facies groups of moderate and high-pressure of disthene-muscovite and disthene-staurolite subfacies superimposed on gneissic strata, metamorphosed into rocks of amphibolite and granulite facies;

e) metasomatic formations, usually local, connected with regressive stages of metamorphism (magnesium, iron-magnesium, alkaline metasomatism).

2. Geochemical:

a) local anomalies of ore metals;

b) heterogeneity in distribution of large quantities of ore metals;

c) dispersion haloes in country rocks of geochemically specialized formations;

d) associations with geological formations specialized in the other ore-elements.

3. Petrophysical:

a) zones of increased magnetic susceptibility-magnetic and ferromagnetic;

b) zones of average magnetic susceptibility;

c) zones and fields of maximum density in rocks of ore-bearing geological formations;

d) average values of density in country rocks and local parts of contrasting area;
e) fields of minimum density in rocks of ore-bearing formations;
f) zones of heterogenetic physical properties associated with the alternation of sharply different petrophysical properties.

In conclusion it showed to be repeated that the present report is the first attempt to unify historical-geological analysis with the compilation of specialized maps synthesizing multiform information on geological structure and regional evolution. Besides its applied significance to the forecasting of deposits of useful minerals, this attempt is of methodical interest and may give additional information to solving the problem of the geological evolution of the Earth's crust.

THE SYDVARANGER TYPE OF QUARTZ-BANDED IRON ORE, WITH A SYNOPSIS OF PRECAMBRIAN GEOLOGY AND ORE DEPOSITS OF FINNMARK

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Introduction

The oldest rocks in Norway are encountered in East Finnmark near the northwestern boundary of the Baltic Shield (Fig. 1). Four ancient mobile belts of global extent intersect, underlain, by Archaean basement.

1. Early Archaean, 3 000 Ma or more.
2. Norwego-saamides, viz. quartz-banded iron formation, above 2 700 Ma.
3. Belomorides (Marealbides), viz. granulites and reworked Archaean.
4. Svecokarelides, 1 650—2 100 Ma (or Prekarelides?) viz. The Petsamo group (Holmvann, Caskias, Carravarre, Karasjok groups in West Finnmark).
5. Vendian and Late Riphean (Eocambrian and Late Precambrian).
6. Caledonides, metamorphosed rocks of the Kalak nappe.

The main tectonic units are separated by erosional and discordant structures and terminate with granite intrusions, as Sederholm observed about 50 years ago. The relative age relations are fairly well established, but data on the absolute ages are scarce and presented tentatively.

1. Early Archaean comprises granitic gneisses, migmatites and in the easternmost area of Jarfjord mica garnet schists and kinzigites, representing the oldest known sediments in Norway. The rocks are intruded by granites and scattered gabbro and peridotite bodies.
2. The quartz-banded sedimentary iron ore formation of Sydvaranger represent the oldest metallogenetic province of commercial importance in Norway and can be correlated with similar deposits of the same age on the Kola peninsula, in Canada and USA (e. g. Murmansk, Michipicoton, Vermilion). The iron formation is older than the Neiden granite complex (Wiik 1966), which has been dated to about 2 500 Ma (Meriläinen 1976).
3. The Belomorides (the White Sea block) forms a median massif between East and West Finnmark. It is composed of a variegated series of granulites with hypersthene, garnet, sillimanite and reworked Archaean gneisses. Only a small part extends into Norway where the eroded remnants dissappear underneath a cover of Late Precambrian/Eocambrian sediments. The complex has been thrusted above amphibolites of the Petsamo group in northeastern direction on one side, to the southwest on the opposite side (Bugge 1960).

Few metallic deposits are known from within the complex. Small cavity filling deposits of galena and barite of Early Paleozoic age intersect both granulites and the overlying Cambrian Dividal series. Similar deposits occur in USSR east of Grense Jakobselv.

In southern part of Pasvik valley gneisses and peridotite bodies are presently prospected for possible nickel-copper deposits of Allarachensk type. The age relations of the peridotites is not ascertained. The Allarachensk peridotites may to a certain degree be compa-

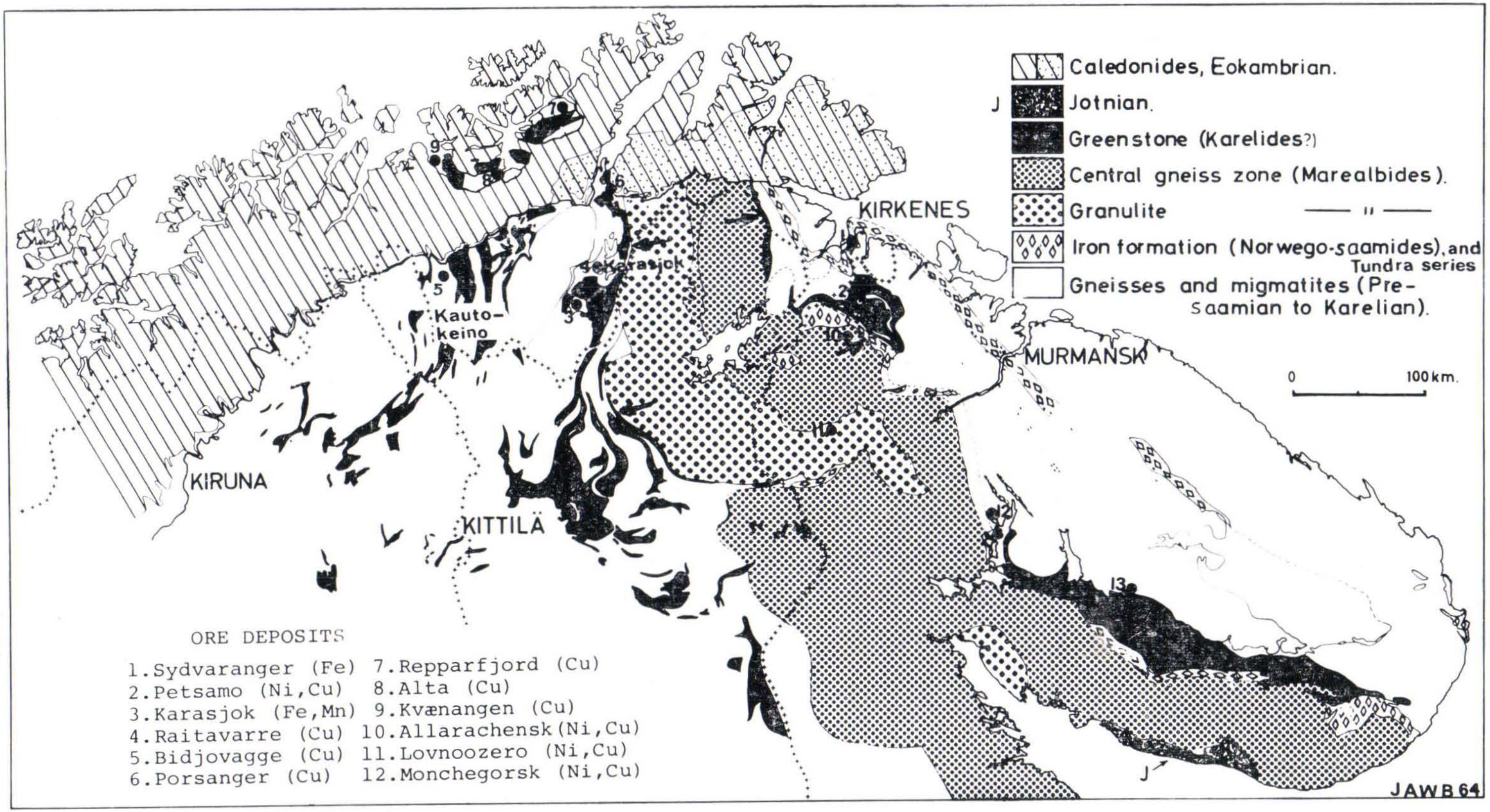

Fig 1. Northern Fennoscandia and Kola Peninsula.

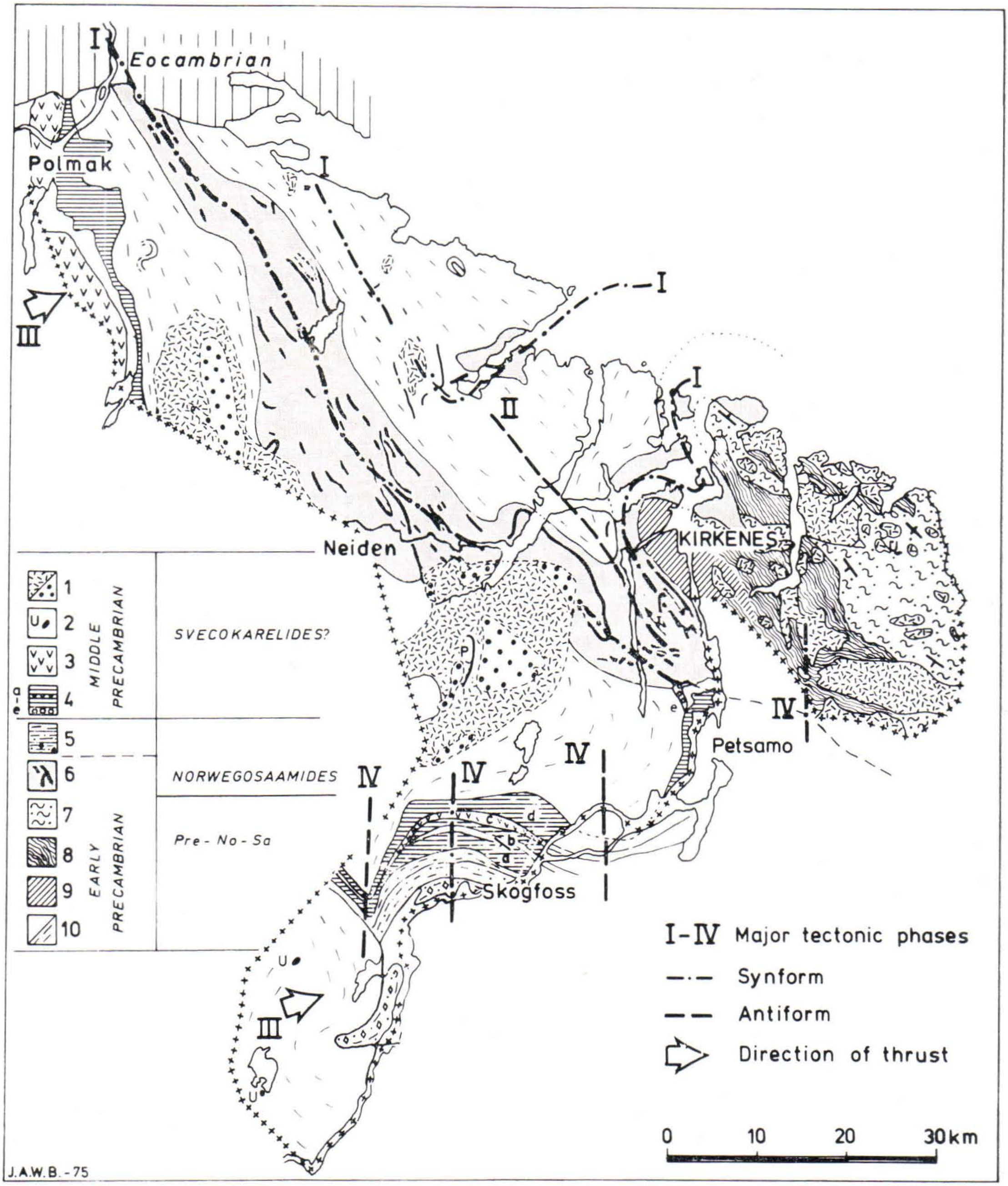

Fig. 2. Geology of Sydvaranger (after J. A. W. Bugge 1978) 1. Granite, 2. Ultramafics, 3. Mafic intrusives, 4. Petsamo—Skogfoss group, 5. Mica-, hornblende-, garnet-gneiss (Tundra series), 6. Banded iron formation, 7. Kinzigite, garnet mica gneiss, migmatite, 8. Mica gneiss, 9. Gneissgranite basement, 10. Archaean, incl. younger rheomorphic gneiss and migmatite.

red with the Pechenga type (Glazkovsky *et al.* 1977). The age has been determined provisionally as Proterozoic.

4. Svecokarelides constitute the Precambrian of the central and major part of Fennoscandia. Continental and shallow sea sediments dominate the northeastern, Karelian, section, while deepwater greywackes and turbidites are common in the Svecofennian section further southwest (Hietanen 1975). In Finnmark though, several rocks which have been accepted as Svecokarelian may be much older and correspond to Archaean greenstone belts, 2 800—3 000 Ma, in Finland (Gaál *et al.* 1978, Papunen *et al.* 1977).

Similar series of volcanogenic rocks and ultramafics occur on both sides of the Belomorian belt, in the Petsamo/Pasvik and the Karasjok area (Fig. 2). Both from geological and geochemical point of view they seem to be related. But radiometric dating of those on the east side points to Svecokarelian age.

The discovery in 1920 of large copper-nickel deposits in the Petsamo region of USSR encouraged the search for similar deposits on the Norwegian side of the border where serpentinized peridotite and metagabbro occur in the sedimentary-volcanogenic series in Pasvik and Karasjok.

In West Finnmark Precambrian occurs as windows within the Caledonides and also on Finnmarksvidda. Metalliferous deposits are copper sulphide deposits and iron-manganese oxide deposits.

5. Late Riphean and Vendian sediments, viz. epicontinental and shallow shelf sediments were deposited on the eroded remnants of Middle and Early Precambrian rocks.

A. East Finnmark

Quartz-banded iron ore

The iron formation can be traced 100 km westwards from Kirkenes (Fig. 2.) to the bottom of Varangerfjord where the folded and eroded remnants disappear below Late Precambrian and Eocambrian sandstones and tillites.

Two fold systems can be demonstrated. The old isoclinal F_1 phase is followed by a younger cross-folding F_2 phase.

Further west banded iron ores of the same type, but of a higher metamorphic grade (granulite facies) occur on the islands of Lofoten and Vesterålen on the opposite side of the Norwegian Caledonides. Exploration though, has only revealed a few million tons of ore with 25—30 % Fe and several additional small vein deposits of remobilized highgrade magnetite ore.

The supracrustals of the iron formation were deposited directly on the eroded, but not peneplained Archaean surface. The stratigraphy is best known in the mining area of Bjørnevann:

Biotite-hornblende gneiss (meta andesite)
Quartz-banded iron ore (iron sediments)
Bjørnevann gneiss (quartzite and arkosite with metarhyolitic interbeds)
Bjørnevann conglomerate

The ore beds occur in a banded and strongly folded series of metavolcanics and metasediments. They are interpreted as chemical sediments of exhalative-sedimentary genesis. The oldest sediments are quartzites and arkosites with a few metarhyolitic interbeds and some amphibolites, representing basic volcanics. The sediments are continental terrigenous and shallow water deposits. Polymict conglomerates occur in the mining area of Bjørnevann, but have not been recognized elsewhere.

The quartzites average 70—85 % SiO_2, and the minerals are quartz, feldspar, biotite, muscovite, epidote, garnet and kyanite occasionally in the more micaceous types.

Table 1.
Chemical composition and modes of rocks from Bjørnevann mining area.

	Bjørnevann gneiss		Meta-andesite			
	1	2	3	4	5	6
SiO_2	74.45	79.61	50.26	52.56	58.57	59.27
TiO_2	0.31	0.21	0.68	1.42	0.74	0.72
Al_2O_3	7.98	10.31	20.58	21.01	17.10	17.38
Fe_2O_3	5.13	0.91	1.88	1.07	0.75	1.58
FeO	2.26	1.60	4.84	4.75	4.73	4.76
MnO	0.09	0.10	0.39	0.29	0.25	0.15
MgO	2.08	0.10	6.00	3.82	5.78	4.24
CaO	0.92	0.71	7.05	8.32	4.74	4.96
Na_2O	1.57	3.29	3.97	3.95	2.99	3.67
K_2O	4.41	2.37	2.56	1.21	1.15	1.36
H_2O^+	0.59	0.59	1.50	0.91	2.69	1.60
H_2O^-	0.04	0.04	0.06	0.03	0.03	0.00
CO_2	0.06	0.40	tr	0.00	0.81	0.00
P_2O_5	0.02	0.02	0.02	0.08	0.05	0.09
S	0.02	0.02	—	—	—	—
	99.93	100.28	99.97	99.42	100.38	99.78

Anal. B. Bruun

Mode						
Quartz	47.6	47.2	4.9	11.1	23.7	21.2
Plagioclase	12.0	33.0	40.0	42.5	39.7	40.0
Microcline	10.0	11.0	—	—	—	—
Biotite	19.3	—	24.0	14.0	11.2	11.2
Muscovite	7.0	4.9	—	—	—	—
Amphibole	—	—	21.7	24.8	—	17.6
Epidote	4.0	3.3	7.7	6.4	2.9	1.6
Chlorite	—	—	—	—	17.7	7.8
Other	—	0.6	1.7	1.7	0.9	0.5
Calcite					2.2	

There is a marked difference in the chemical and mineralogical composition (Table 1) of the underlying rocks and those lying above and alternating with the iron ore sediments. The latter are quartz-biotite-hornblende gneisses with 50—60 % SiO_2, interpreted as meta-andesites and dacites. Rocks interpreted as tuffs and pyroclastics are found in a few localities.

The sudden change in geochemistry demonstrate a connection between ore deposition and volcanism and is in favor of an exhalative sedimentary mode of origin (Geijer 1911, Bugge 1960).

The ore is thin banded with magnetite-rich bands alternating with quartz-rich bands, each of them 2 to 10 mm thick. The same banding is well known from many other localities around the world and is beautifully developed. It is caused by rhytmic precipitation of iron oxide, ironsilicate, and silicagels, while the content of clastic material is low and carbonate and sulfide facies practically absent. Slumping structures formed in the unconsolidated sediment, sometimes have the appearance of disharmonic folds.

The ore is completely recrystallized and the principle minerals are magnetite, quartz, grunerite, common hornblende, epidote and biotite. The crude ore contains in average 30 percent magnetic iron and is low in phosphor and sulphur (0,04 % P and 0,05 % S).

Low-grade ore in Sydvaranger was first observed by the Norwegian geologist Tellef Dahl in 1866, but the importance and size not appreciated before the turn of the century, by which time improved methods of magnetic separation were available. In 1902 extensive prospecting resulted in the discovery of large deposits of quartz-banded magnetite ore (Fig. 3). The mining company A/S Sydvaranger was founded 1906 and has until now mined about 120 mill. ton of ore with a grade of 30 percent iron. The yearly production, all obtained from open pit mining, amounts to 2,4 mill. tons of pellets with a grade of 65 percent iron.

The iron ore beds were once nearly continuous througout the Pasvik Peninsula. The present irregular form was caused by repeated regional deformation and faulting. At least five tectonic stages can be distinguished:

I. Isoclinal and recumbent folds (F_1). The deposits of the central area south of Bjørnevann occur on both sides of a synform, Bjørnevann Mine lies at the northern crest of the synform with foldaxis plunging towards the south. In the mining area the plunge alternates between north and south. In the Ørnevann area the structure is more complex and recumbent folds and nappes are overturned to the west.

II. Crossfolds (F_2). The early phase of isoclinal folds was followed by shearfolds with

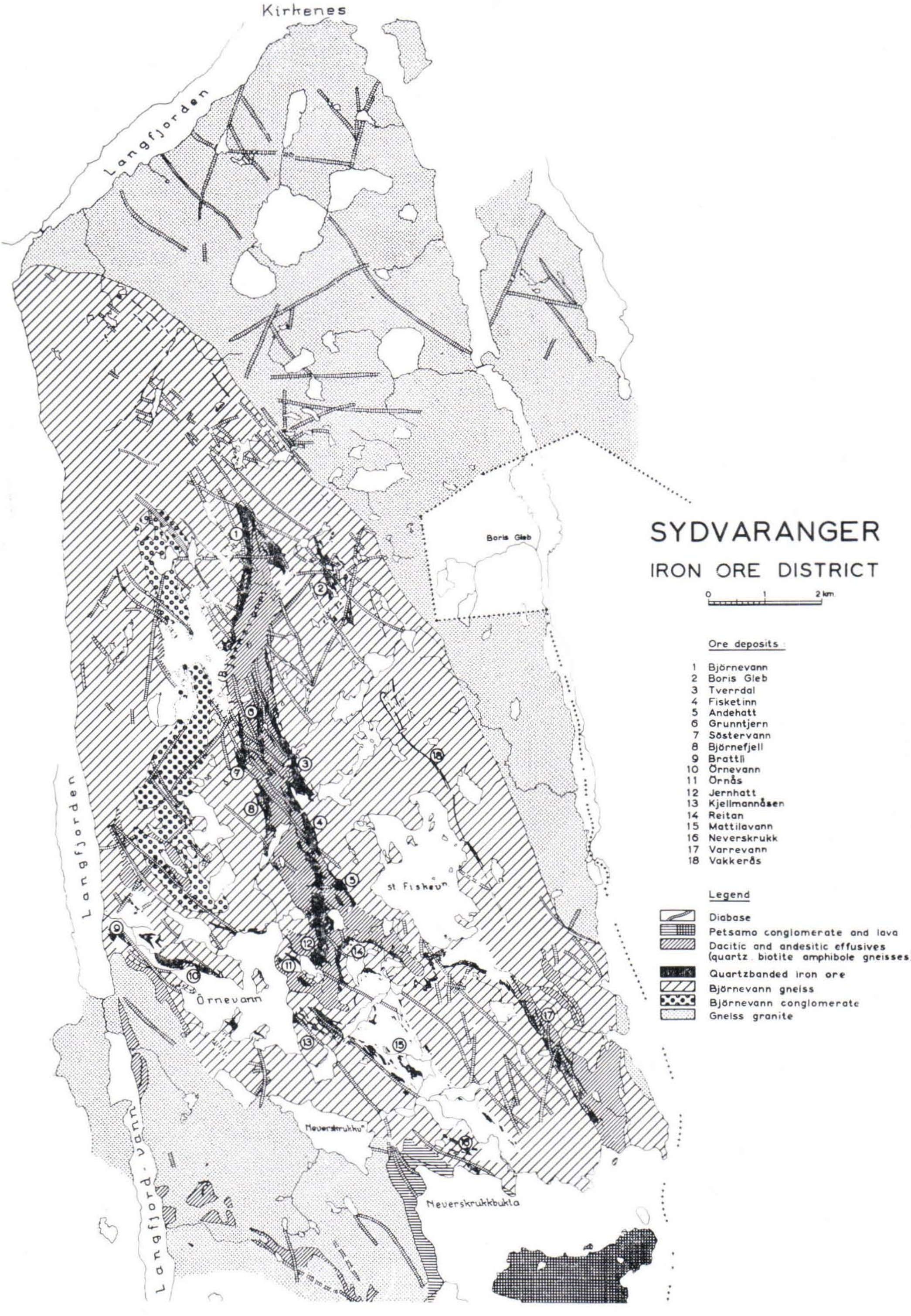

Fig. 3. Sydvaranger iron ore district (After J. A. W. Bugge 1960).

axes striking NW—SE. By this shear the western side of the deposits were displaced northwards relative to the eastern side. The N-form for several deposits was formed during this phase.

III. A deep rift, or subduction zone, formed in the crust between two Precambrian blocks, through which mafic and ultramafic magma in part extruded. These rocks are less metamorphosed and deformed than those on both sides. In a later stage the Belomorian rocks to the south-west were thrusted in a north-easterly direction above the mafic rocks of the Petsamo group. This deformation apparently also affected the banded iron ore formation, causing overthrusts towards north-east.

IV. Open folds with axes striking north-south has affected the Pasvik-Petsamo group of rocks. The influence on the iron formation is small.

V. Tensional stress causing fracturing and faulting is the last tectonic phase influencing the formation. Hundreds of diabase dykes were intruded along these fractures, and several of them are cut by even younger faults. Most of the dykes are probably of Svecokarelian age.

Table 2.
Chemical analyses from the Petsamo Group, Pasvik

	1	2	3	4	5	6	7
SiO_2	55.00	55.12	48.68	57.45	44.37	44.40	43.80
TiO_2	1.02	1.45	1.93	1.86	2.63	1.62	0.34
Al_2O_3	13.59	14.17	13.81	16.63	5.59	9.80	9.56
Fe_2O_3	1.87	2.87	1.23	7.59	0.28	0.88	1.27
FeO	9.78	8.95	13.17	4.03	16.35	11.10	10.90
MnO	—	—	—	—	—	—	0.14
MgO	6.19	4.12	6.13	0.67	18.95	14.98	18.31
CaO	7.69	6.93	8.62	3.80	6.01	9.56	8.32
Na_2O	2.51	1.84	1.88	4.32	0.11	0.84	0.92
K_2O	0.21	1.83	0.29	0.42	0.03	0.15	0.07
P_2O_5	0.12	0.18	0.10	0.19	0.18	0.10	0.11
Vol.	2.20	1.67	4.70	0.84	5.37	7.14	5.38
Other	—	—	—	—	—	—	0.51
	100.17	99.12	100.55	97.82	99.87	100.57	99.74
Katanorm							
Q	9.74	13.09	3.14	22.32	—	—	
C	—	—	—	3.63	—	—	
Or	1.29	11.42	1.83	2.64	0.19	0.94	
Ab	23.38	17.46	18.08	40.14	1.05	8.02	
An	26.15	26.43	30.14	18.35	15.55	23.95	
Σ sal.	60.56	68.41	53.46	87.08	16.73	32.91	
Hy	25.58	18.96	30.31	2.07	50.36	24.86	
Di	10.09	6.93	11.75	—	11.84	20.51	
Ol	—	—	—	—	16.45	18.02	
Mt	2.03	3.17	1.38	5.59	0.31	0.98	
Hem	—	—	—	2.06	—	—	
Il	1.47	2.14	2.88	2.77	3.88	2.40	
Ap	0.26	0.40	0.22	0.43	0.40	0.22	
Σ fem.	39.44	31.59	46.54	12.92	83.22	67.09	

1. Diabase-porphyrite, Neveskrukkbukt, Pasvik
2. Diabase-porphyrite, 4.5 km north of Svanvik
3. Metabasalt, borehole no 1, Skjellvann, Pasvik
4. Keratophyre, Skogfoss, Pasvik
5. Peridotite, borehole no 1, Skjellvann, Pasvik
6. Agglomerate, 4 km south of Skogfoss, Pasvik
7. Meta-picrite, Njuovcokka, Karasjok (Wennervirta 1969)

Petsamo-Pasvik group

The supracrustal rocks of the Petsamo group has been traced westwards from Pechenga through Norway and Finland, and back again into Norway further west until the eroded and metamorphosed rocks dissappear below Eo-cambrian cover of tillites and sandstones.

The oldest sediments were deposited on the eroded, but still not peneplained surface of the banded iron formation. Pebbles of quartz-banded ore are present in the Neverskrukk-conglomerate (Sederholm 1930). The first lava flows were diabase-porphyrites with amygdules and a few interbeds of felsic lavas and calcareous sediments, followed by keratophyres, pillowlavas and ore phyllite with peridotite and gabbro intrusions. Several of the rocks described from the Petsamo region (Väyrynen 1938) are encountered, except for the lava flows highest up in the series.

The Skogfoss area in Pasvik represent a deeper, more compressed and deformed section where the thermodynamic and chemical conditions apparently were less favorable for sulphide segregation. Ore deposits of commercial grade have not been discovered yet.

The Skogfoss petrographic province consists of a differentiated series of rocks (Table 2).

Tholeiitic metabasalts and andesites dominate, but acid varieties with dacitic and rhyodacitic affinities are common. The nature of the ultramafics and their genetic relationship to the metavolcanics is still obscure. They are probably cogenetic and form stratabound lens- to bed-like bodies in graphite schists and tuffaceous rocks. Ultramafic talc chlorite schists, talc conglomerates or agglomerates are also met with, suggesting the existence of true ultramafic lavas. Nearly identical pyroclastics and metapicritic lavas have been recognized in the Karasjok region (Wennervirta 1969).

B. West Finnmark

Copper deposits

In West Finnmark (Fig. 4; Table 3) conglomerates, arkosites and quartzites of the Karasjok group were deposited on the eroded remnants of a granitoid Archaean basement about 2 700 Ma old (Meriläinen 1976).

The argillitic gneisses above the quartzites consist of mica gneisses and mica schists in which lowgrade disseminated copper mineralization of Aitik-type has been found at Raitavarre SW of Karasjok.

Greenstone lavas of the Caskias group are widespread in the Precambrian of West Finnmark and contain several stratabound and vein type copper deposits (e. g. Bidjovagge, Porsanger, Porsa, Kvaenangen). Chalcopyrite is the dominating ore mineral, and gold is a small, but characteristic constituent in some deposits.

Chalcocite and bornite are sometimes important minerals and suggest a Late Precambrian weathering period with the formation of secondary supergene sulphide deposits (e. g. Souvrarappat, Porsanger).

Repparfjord mine is a disseminated lowgrade copper deposit with chalcopyrite, chalcocite and bornite in Late Precambrian sandstone from the Raipas windows within the Caledonides. The mode of occurrence is consistent with the idea of Late Precambrian weathering under warm arid conditions.

The two thousand km long lead belt at the Cambrian/Precambrian boundary on the east side of the Caledonides was probably formed during the same period of arid weathering (Bugge 1978). They represent the latest metallogenic province to be mentioned in this account.

Table 3.
Lithostratigraphy of Finnmarksvidda

LATE PRE-CAMBRIAN (l. Riphean)	**Gaissa nappe** (Porsanger sdst)	
LOWER CAMBR. Late Prec.	**Dividal group** (Hyolithus zone)	1. Stratabound Pb-Zn-baryte 2. Fissure vein Pb-Zn-baryte
SVECO-	**Granulites** (in part reworked Archaean)	
KARELIDES	**Carravarre group** (meta-arenite, albite-carbonate)	
PRE-	**Caskias group** (greenstone, mafics, ultramafics)	3. Stratabound and fissure vein copper deposits 4. Pyrite-pyrrhotite sediments
KARELIDES (?)	**Karasjok (Masi) group** (metapicrite?, metapelite and arenite, conglomerate)	5. Iron-manganese sediments. Iron quartzites 6. Disseminated copper
ARCHAEAN	**Granitic basement**	

In the Karasjok area there are large intrusive bodies of gabbro and associated peridotites

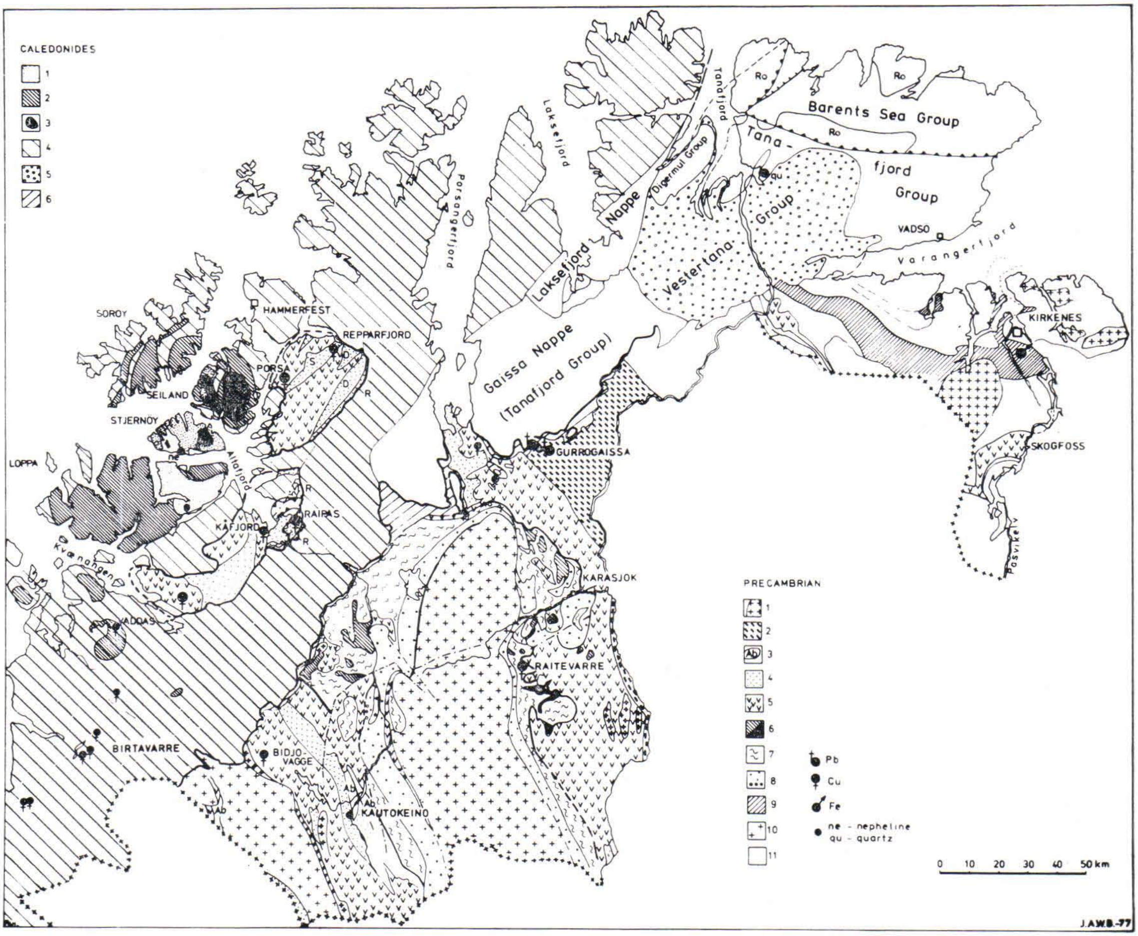

Fig. 4. Geology and mineral deposits of Finnmark.

Precambrian: 1. Granite, 2. Granulite, 3. Albite-carbonate rock, 4. Metasandstone [Carravarre (C), Rafsbotn (R), Bossekop (B), Doggelv (D), Saltvann (S)], 5. Greenstone group, 6. Ultramafics, 7. Argillites, 8. Arenites, 9. Quartz-banded iron formation, 10. Archaean basement, 11. Precambrian, undivided.

Caledonides: 1. Alkaline rocks, carbonatites, 2. Mafic intrusives, 3. Ultramafics, 4. Caledonides (Cambro-Silurian and Eocambrian metasediments), 5. Vestertana group, 6. Dividal group.

with small amounts of nickel and copper bearing sulphides. Commercial deposits have not been demonstrated, but the area is presently prospected for base metal deposits.

Iron-manganese deposits

South of Karasjok exhalative-sedimentary iron quartzites and iron-manganese deposits occur above ultramafic lavas and pyroclastics (Wennervirta 1969). Greenstone lavas seem to have been subordinate at the time of extrusion of the ultramafics. Such conditions are well known in Archaean time, but not in Svecokarelian, and is consistent with the idea of a higher, Prekarelian, age of the rocks than is commonly assumed.

In chemical respect though, the Karasjok and Skogfoss lavas differ from the Archaean high Mg-basalts and komatiites.

The Sydvaranger and the Karasjok iron ore deposits differ in geology, geochemistry as well as in age and represent two different types, calc-alkaline andesitic volcanism, low phosphor and manganese content in Sydvaranger, and ultramafic lavas, relatively high phosphor and manganese in the Karasjok type. But both types can be correlated with several other important iron ore deposits around the world. They are both situated within the Goodwin iron ore belt of Archaean and Proterozoic deposits, demonstrating that they probably originated in shallow basins at the margins of Early Precambrian continents.

References

Bugge, J. A. W., 1960. Precambrian of eastern Finnmark. *In* Geology of Norway, ed. by O. Holtedahl. Norges Geol. Unders. 208, 78—92.

— **1978.** Mineral deposits of Norway. *In* Mineral deposits of Europe, ed. by H. U. Bowie, A. Kvalheim & H. W. Haslam, IMM, vol. 1.

Gaál, G., Mikkola, A., & Söderblom, B., 1978. Evolution of the Archaean crust in Finland. Precambrian Res. 6, 199—215.

Geijer, P., 1911. Contribution to the geology of Sydvaranger. Geol. För. i Stockholm, Förh. 33, 312 p.

Glazkovsky, A. A., Gorbunov, G. J. & Sysoev, F. A., 1977. Deposits of nickel. *In* Ore deposits of the USSR, ed. by V. J. Smirnov. English ed. vol. 11, 3—29.

Hietanen, A., 1975. Generation of potassium-poor magmas in the northern Sierra Nevada and the Svecofennian of Finland. J. Res. U.S. Geol. Surv. 3, 631—645.

Meriläinen, K., 1976. The granulite complex and adjacent rocks in Lapland, northern Finland. Geol. Surv. Finland, Bull. 281. 129 p.

Papunen, H., Idman, H., Ilvonen, E., Neuvonen, K. J. Pihlaja, P., & Talvitie, J., 1977. Lapin ultramafiiteista. Summary: The ultramafics of Finland. Geol. Surv. Finland, Rep. Invest. 23. 87 p.

Sederholm, J. J., 1930. Några ord om berggrunden i Sydvaranger och nærliggande delar av Finland. Geol. Fören. i Stockholm, Förh. 52 (4), 435—454.

Väyrynen, H., 1938. Petrologie des Nickelerzfeldes Kaulatunturi—Kammikivitunturi in Petsamo. Bull. Comm. Geol. Finlande. 116. 198 p.

Wennervirta, H., 1969. Karasjokområdets geologi. Norges geol. Unders. 258, 131—184.

Wiik, V. H., 1966. Petrological studies of the Neiden granite complex. Norges geol. Unders. 237. 99 p.

VOLCANISM AND IRON ORES IN THE PRECAMBRIAN OF SWEDEN

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Introduction

The purpose with the present paper is to bring together some current concepts relating to Swedish Precambrian iron ore deposits, particularly with regard to the petrological and chemical features of the volcanic host rocks and the ores. The paper is a short presentation of work still in preparation. Detailed information about rocks, ores, analyses ect., and a more extensive discussion of the theoretical results is postponed to a forthcoming publication.

Sweden contains a relatively large part of the iron ore reserves of the Baltic Shield: almost 100 % of the apatite iron ores, about 89 % of the Ca-Mg silicate-rich skarn iron ores and about 10 % of the quartz-banded iron ores. The ores which occur in the Middle Precambrian formations of 2 200 to 1 600 Ma age, are encountered in two separate geographical regions, namely central and northern Sweden.

The ores of central Sweden, which comprise skarn ores, quartz-banded ores and some apatite-bearing ores, occur in a broad, semicircular zone west of Stockholm. They belong to a metamorphic volcanic-sedimentary complex which in this part of Sweden is the oldest unit of the Precambrian. This complex occurs in a large number of relatively deep basins and is dominated by acid volcanics (leptites) ranging in composition from sodic to potassic types. Intercalations of limestone-dolomite are common. In many parts there are detrital sediments, usually associated with basic volcanics. The internal stratigraphy of the volcanic-sedimentary complex is still undecided.

The supracrustal rocks were folded and metamorphosed in connection with the intrusion of the oldest group of Svecokarelian granitoids ranging from gabbro to granite. Their age is about 1 900 Ma which thus is the upper age limit for the volcanic-sedimentary complex. About 1 800 Ma ago, migmatization of the supracrustal rocks occurred in connection with the intrusion of the late Svecokarelian granites.

The iron ores in northern Sweden lie in a wide zone extending on both sides of Kiruna. The ores occur in supracrustal rocks. The oldest rocks are the basic volcanics of the Greenstone group. In the higher stratigraphic parts, there are intercalations of tuff, tuffite, graphite-bearing schist, limestone-dolomite, marl and chert. In association with these formations which were deposited in a closed basin environment on the border of the Archean craton during the evolutionary phase of the Svecokarelian orogeny, skarn iron ores and quartz-banded iron ores were formed.

The basic volcanics are underlain by a basement of gneiss and gneissose granite. N of Kiruna, the granite has an age of about 2 750—2 800 Ma. Similar basic volcanics in northern Finland have yielded radiometric ages in the 2 000—2 200 Ma range. The Greenstone group is intruded by syn-kinematic granitoids, mostly granodiorite and gabbro, with an age of about 1 880 Ma.

The Greenstone group is overlain by the Porphyry group which consists of acid to intermediate volcanics. These rocks occur in relatively shallow basins and are mainly found in the western part of the iron ore-bearing area. The age of the volcanics in Kiruna and SW of Kiruna is 1 605—1 635 Ma. The Porphyry group contains apatite-bearing iron ores and some deposits of quartz-banded iron ores.

The late-kinematic granites and monzonites are younger than the Porphyry group, being about 1 535—1 565 Ma old. Among them is a Perthite series which shows a close chemical and genetical relationship to the rocks of the Porphyry goup.

The main mineralogical and chemical features of the iron ores both in central and northern Sweden are summarized in Table 1. The skarn ores of central Sweden are divided into non-manganiferous and manganiferous types, the limit being placed at 1 % Mn.

The apatite-bearing iron ores are considered to be late-magmatic formations closely associated with the volcanics in which they occur (Geijer 1931 b, Magnusson 1970, Frietsch 1978). Oelsner (1961) and Parák (1975) have proposed an exhalative-sedimentary origin for these ores. The skarn ores and the quartz-banded ore types are considered to be syngenetic of volcanic-sedimentary origin (Frietsch 1977). The skarn silicates were formed by internal reactions caused by regional metamorphism. For the skarn ores both in central and northern Sweden, a pyrometasomatic origin has been proposed by Geijer (1931 a) and Geijer and Magnusson (1952).

Table 1

Some mineralogical and chemical features of the Precambrian iron ores of Sweden. Extreme average contents of phosphorus and manganese in certain deposits are given within brackets.

Ore type	Host rocks	Age Ma	Ore mineral	Gangue	% S	% P	% Mn
CENTRAL SWEDEN							
Apatite-bearing	Acid volcanics	> 1 900	Magnetite, hematite	Apatite	< 0.001	0.5—1.3	0.1
Quartz-banded	Acid volcanics and limestone-dolomite	> 1 900	Hematite, magnetite	Quartz, (Ca-Fe-Mg silicate)	0.001—0.1	0.007—0.03	< 0.2
Skarn	Acid volcanics and limestone-dolomite	> 1 900	Magnetite	Ca-Mg and Mg silicate Mn silicate	0.001—0.1 < 0.2	0.004—0.08 < 0.1	< 1 1—8
NORTHERN SWEDEN							
Porphyry group							
Apatite-bearing	Interm.-acid volcanics	1 605—1 635	Magnetite, hematite	Apatite	< 0.1	0.01—2 (5)	< 0.1
Quartz-banded	Interm.-acid volcanics and mica-schist	1 605—1 635?	Magnetite, (hematite)	Quartz, Ca-Mg silicate	< 0.1	0.03—0.06 (1)	0.03—0.4 (7)
Greenstone group							
Skarn	Basic volcanics and sediments	1 880—2 600	Magnetite	Ca-Mg silicate, (iron sulphide)	1—5	0.08—0.2 (3)	0.06—0.04
Quartz-banded	Basic volcanics and sediments	1 880—2 600	Magnetite	Quartz, Fe-Mg-Mn silicate, (iron sulphide)	1—5	< 0.1	0.003—0.8 (2)

Petrological-chemical features of the volcanics

Central Sweden. The iron-ore bearing volcanics in central Sweden are, in the main rhyolites and rhyo-dacites, if the content of silica is considered. Dacites occur to a limited extent, and andesites are quite subordinate. The volcanics are sub-alkaline rocks, belonging to the calc-alkaline series as shown by alkali/silica and AFM ($Na_2O + K_2O/FeO + Fe_2O_3/MgO$) diagrams. With increasing silica content there is a decrease in iron, alumina and calcium+magnesium, whereas the content of alkalies increases. The drop in the alumina content is more pronounced than is usual in a differentiation suite. In addition, the variation of the $Mg^{2+}/(Mg^{2+}+Fe^{2+}+Fe^{3+})$ atomic ratio is independent of the silica content (Fig. 1). Generally, in a calc-alkali (and alkali) series with progressive differentiation there is a steady decrease of this ratio with increasing silica content. Probably the »anomalous» behaviour of these values is not a volcanic feature, but instead related to sedimentary processes. The volcanics are intercalated with clastic and chemical sediments and in part the volcanics have undergone sedimentary reworking. This means that the rocks analyzed are unsuitable for further petrochemical considerations in this connection.

Northern Sweden. In the iron ore-bearing rocks in northern Sweden, some chemical features are shared by the Greenstone group and Porphyry group, but there are several chemical parameters which distinguish the two groups.

The alkali/silica diagram (Fig. 2) shows that the rocks of the Greenstone group are mainly basalts. In addition, a few andesites and peridotites are present. The Porphyry group covers a differentiation from basalt to rhyolite, however, there is a drop in the number of analyses in the dacite-rhyo-dacite field. In this diagram there is no discrimination between the Greenstone and Porphyry groups. Both fall almost equally within the alkaline and sub-alkaline fields. The alkaline tendency for both groups is thus much stronger than for the volcanics in central Sweden. However, all the acid volcanic of the Porphyry group are sub-alkaline according to the partition line drawn by Irvine and Baragar (1971). In the AFM diagram (Fig. 3), both groups fall equally within the tholeiitic and calc-alkaline+alkaline fields.

For both groups the Niggli values al, fm, c and alk, if plotted against the si value, cover similar fields and show typical trends for a

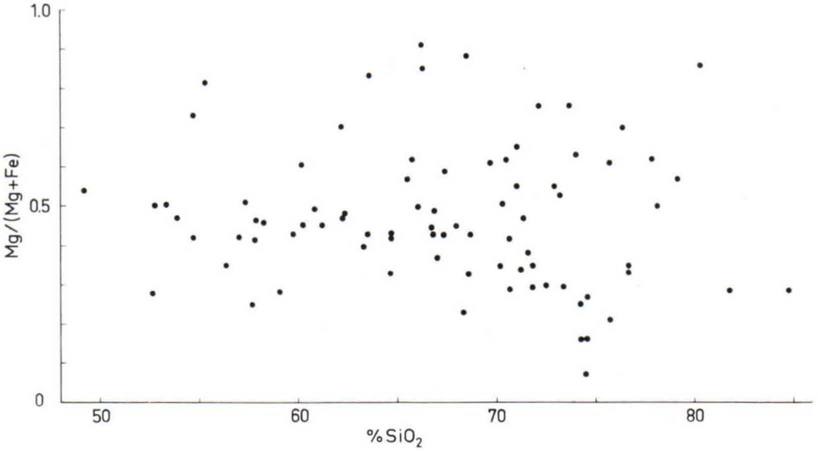

Fig. 1. The relationship between the atomic ratio Mg/(Mg+Fe) and weight % SiO_2 in the volcanics of central Sweden.

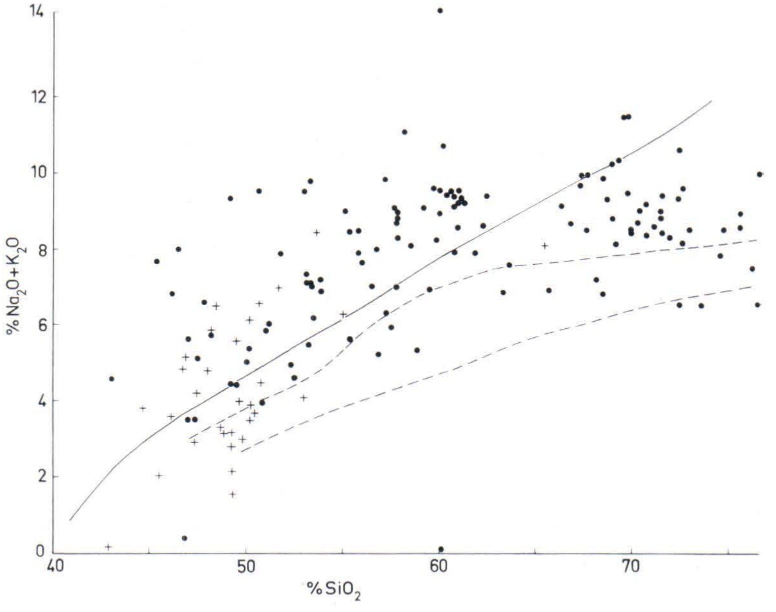

Fig. 2. Alkali-silica diagram (weight %) for the volcanics of the Greenstone group (crosses) and the Porphyry group (dots) in northern Sweden. The upper continuous line after Irvine and Baragar (1971) determines the (upper) alkaline and the (lower) sub-alkaline fields, the broken lines after Kuno (1968) determines (from above) the alkali basalt, the high-alumina basalt and the tholeiite fields.

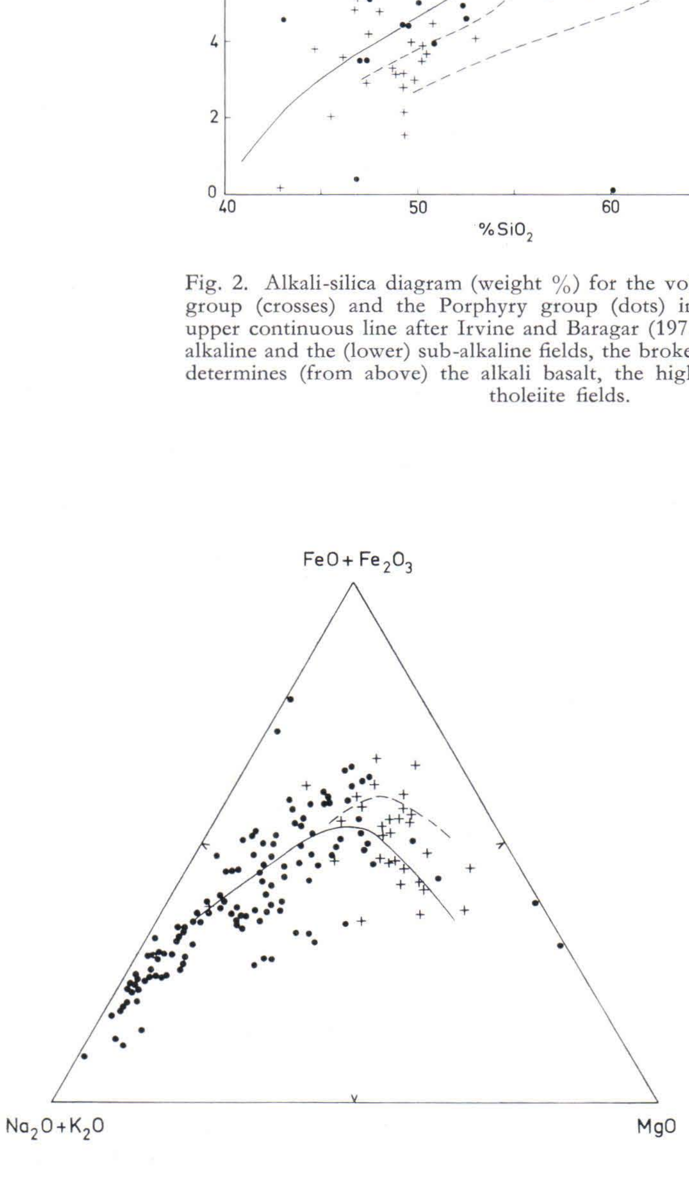

Fig. 3. AFM diagram (weight %) for volcanics of the Greenstone group (crosses) and the Porphyry group (dots) in northern Sweden. The continuous line after Irvine and Baragar (1971) determines the (upper) tholeiitic and the (lower) calc-alkaline + alkaline fields, the broken line after Kuno (1968) determines the fields of the (upper) pigeonitic and the (lower) hypersthenic rocks series.

uniform differentiation process: with an increasing si value the al and alk values increase and the fm and c values decrease.

The significant chemical differences between the two groups are as follows. In the Porphyry group, there is a rather abrupt decrease in the iron content with an increasing silica content, whereas the iron content in the Greenstone group remains relatively unaffected by the silica content (Fig. 4). The same feature is found when plotting the iron content versus the »solidification index» by Kuno (1968), calculated as 100 $MgO/(MgO+FeO+Fe_2O_3+Na_2O+K_2O)$. Most elements in the Greenstone and Porphyry groups vary according to a normal differentiation trend, e. g., the contents of CaO and MgO decrease and in the alkalies there is a slight increase (Fig. 5). With increasing differentiation in the Porphyry group there is a drop in the iron content, whereas in the Greenstone group it remains constant. In addition, the behaviour of silica is different. In the Greenstone group, the content is constant (45—55 % SiO_2) during differentiation, whereas in the Porphyry group it increases (from 45 to 80 % SiO_2). The two groups differ also in phosphorus content. The Greenstone group contains a constant, relatively low content (less than 0.2 % P_2O_5) whereas in the Porphyry group the content is higher (less than 0.8 % P_2O_5) and decreases with solidification (Fig. 5) and increasing silica content (Fig. 6).

In the Greenstone group, there is therefore during differentiation, a constant composition as regards iron, silica and phosphorus. In the Porphyry group, the iron and phosphorus contents decrease and silica content increases during solidification.

Another significant chemical difference between the Greenstone and Porphyry groups is that the degree of oxidation ($Fe^{3+}/(Fe^{3+}+Fe^{2+})$ atomic ratio) is above 0.4 in the Porphyry group

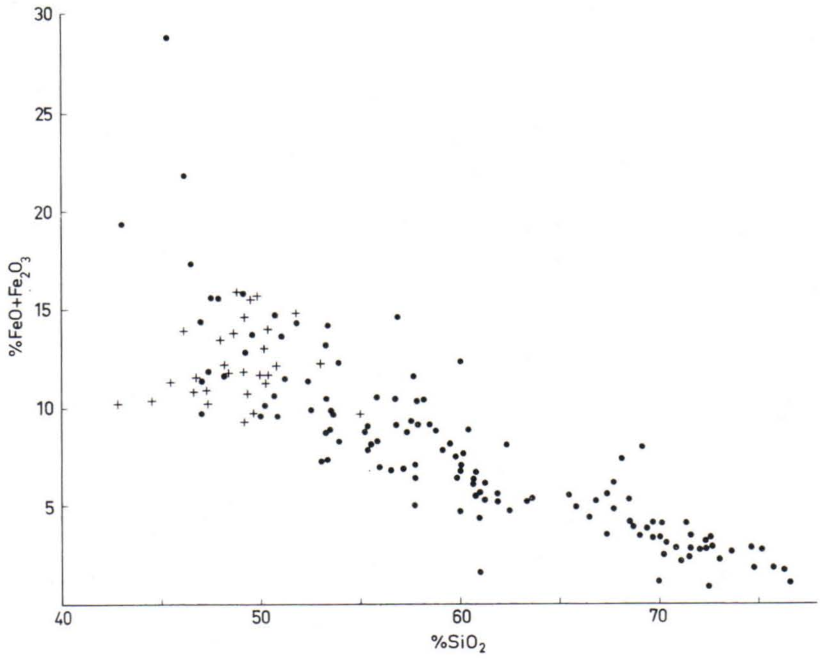

Fig. 4. Iron-silica diagram (weight %) for the volcanics of the Greenstone group (crosses) and the Porphyry group (dots) in northern Sweden.

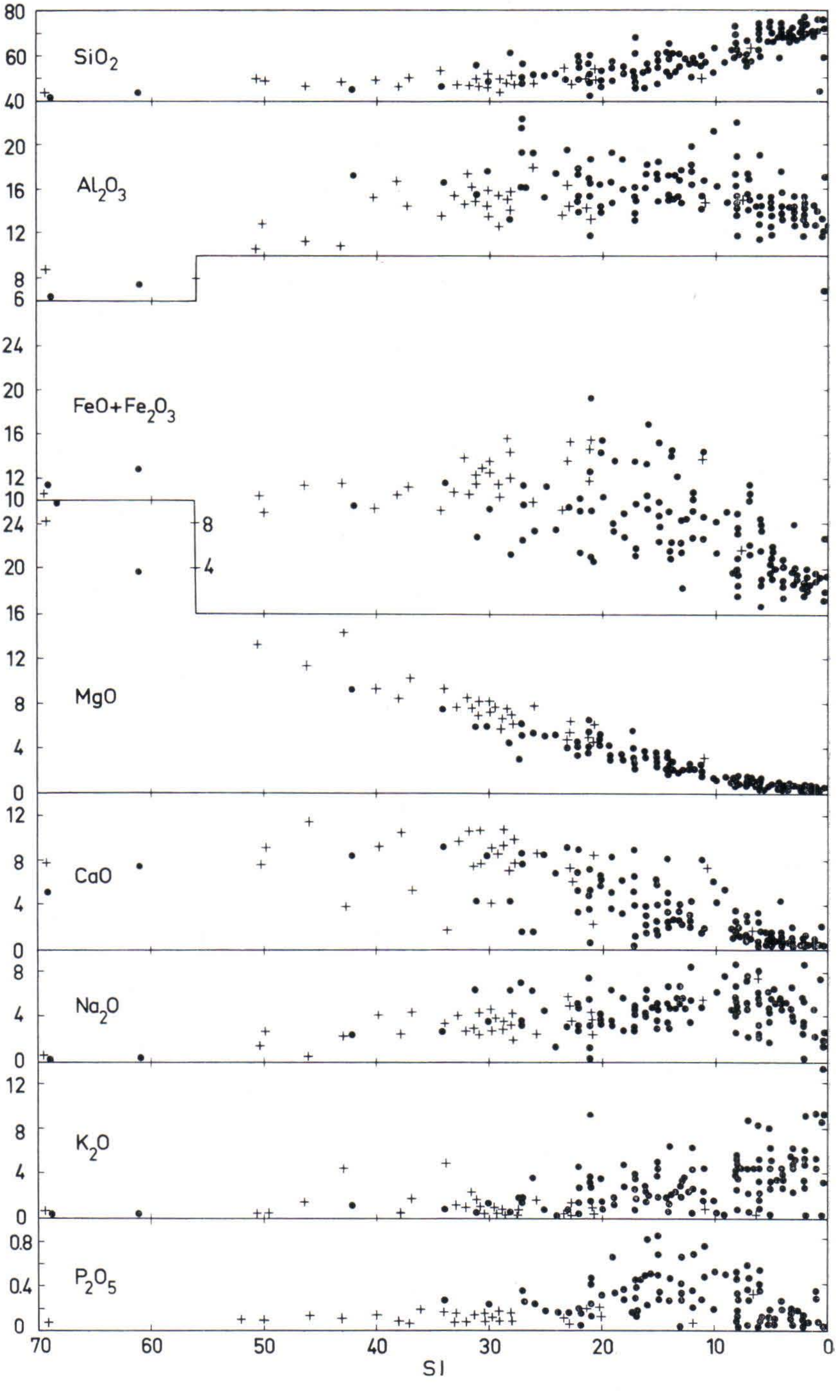

Fig. 5. The relationship between the contents of different elements (weight %) and »solidification index» (SI = 100 $MgO/(MgO+FeO+Fe_2O_3+Na_2O+K_2O)$ according to Kuno (1968) in the volcanics of the Greenstone group (crosses) and the Porphyry group (dots) in northern Sweden.

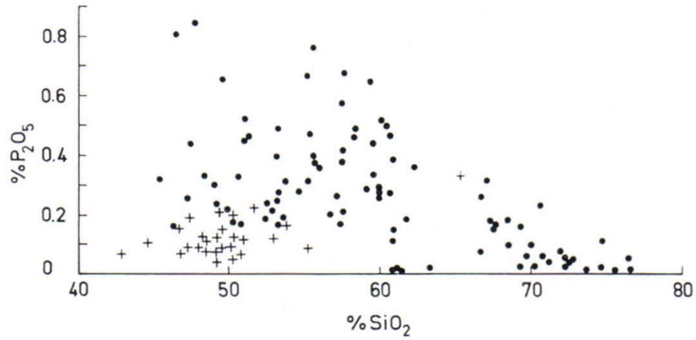

Fig. 6. Phosphorus-silica diagram (weight %) for the volcanics of the Greenstone group (crosses) and the Porphyry group (dots) in northern Sweden.

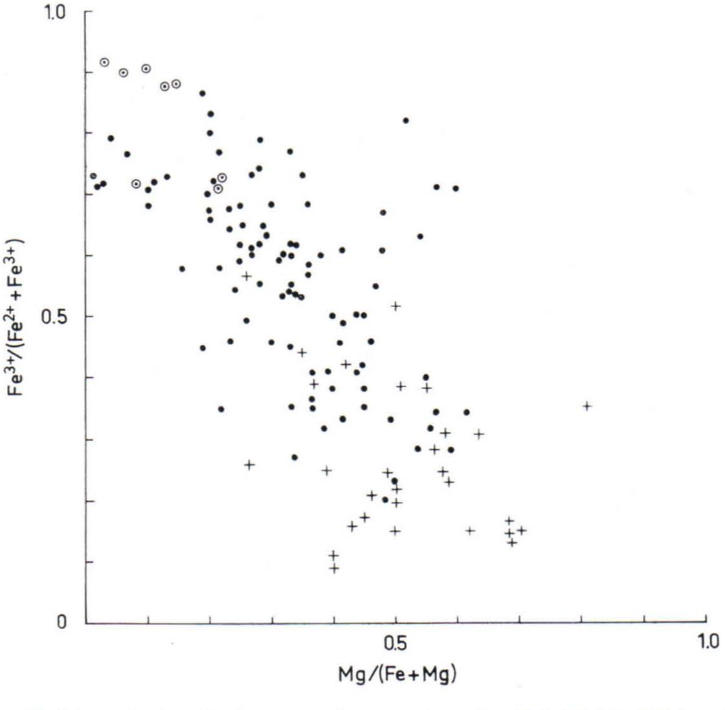

Fig. 7. The relationship between the atomic ratios $Fe^{3+}/(Fe^{3+}+Fe^{2+})$ and Mg/(Mg+Fe) in the volcanics of the Greenstone group (crosses) and the Porphyry group (dots; if sericite-altered: point within circle) in northern Sweden.

and below 0.4 in the Greenstone group (Fig. 7). In the Porphyry group the $Mg^{2+}/(Mg^{2+}+Fe^{2+}+Fe^{3+})$ atomic ratio is lower than in the Greenstone group. This ratio shows a clear decrease with increasing acidity in the Porphyry group (Fig. 8).

Dykes of metabasites and intermediate-acid porphyries are common in the Porphyry group.

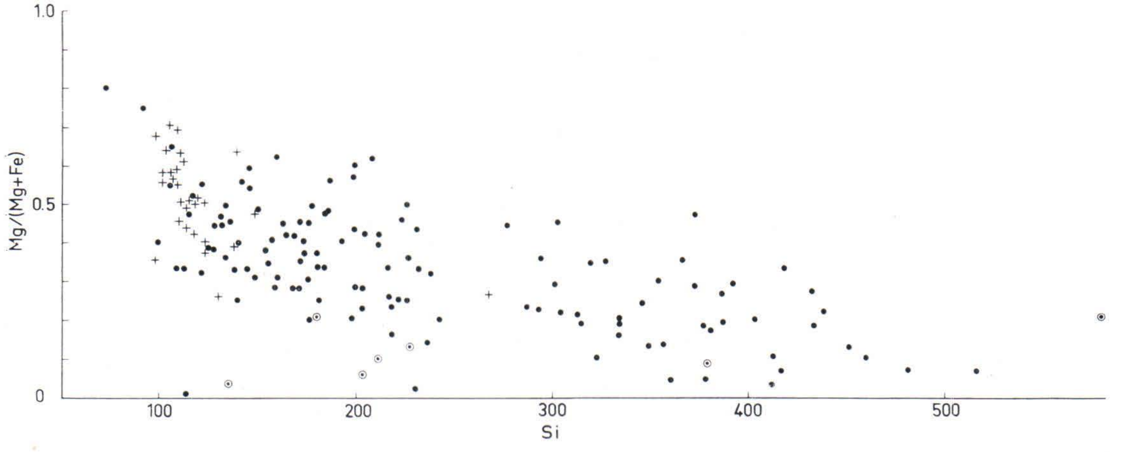

Fig. 8. The relationship between the atomic ratio Mg/(Mg+Fe) and the si Niggli value of the volcanics of the Greenstone group (crosses) and the Porphyry group (dots; if sericite-altered: point within circle) in northern Sweden.

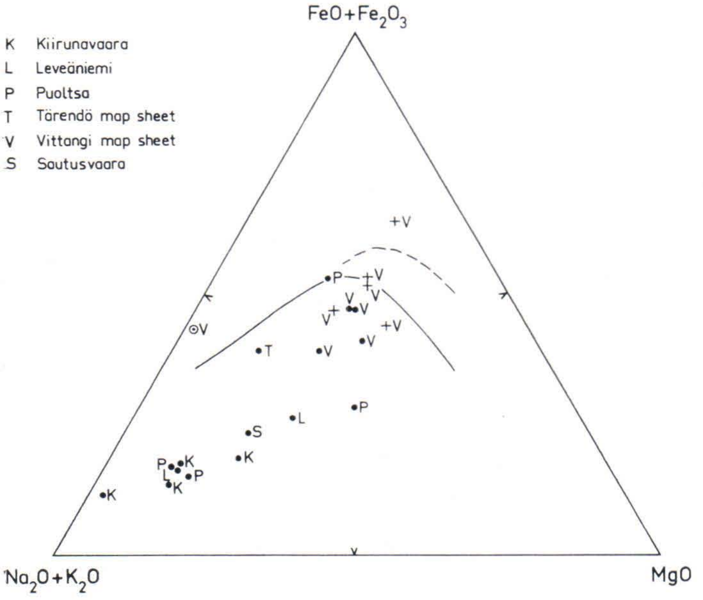

Fig. 9. AFM diagram (weight %) for dyke rocks (metabasites = crosses and acidic-intermediate porphyries = dots; if sericite-altered: point within circle) in northern Sweden. The continuous line after Irvine and Baragar (1971), the broken line after Kuno (1968).

To a small extent the dykes are found in the Greenstone group. The dykes are mostly younger than the apatite-bearing iron ores of the Porphyry group, but older than the younger granitoids. At Kiirunavaara, some of the dykes are older than the ore and some are younger. The porphyry dykes are in many respects, similar in composition, to the volcanics of the Porphyry group. One apparent difference is that some dykes (such as those at Kiirunavaara) have a low content of iron (less than 4 % $FeO+Fe_2O_3$). In the AFM diagram (Fig. 9), the dykes differ markedly from the rocks of the Porphyry group as they are shifted from the $FeO+Fe_2O_3$ apex. The content of phosphorus in the dykes is lower than in the major part of the Porphyry group.

Enrichment of iron in the volcanics

In both the iron ore-bearing magma types in northern Sweden there are therefore regular compositional changes indicating a clear differentiation trend. The evolutionary path has, however, been different for both groups. The different behaviour of iron is characteristic. With increasing silica content, the iron content decreases in the rocks of the Porphyry group wheras it remains constant in the rocks of the Greenstone group.

During fractional crystallization, the path of the development of the magma is largely determined by the composition of the liquid, temperature and oxygen pressure. Two contrasting ways for the magma development can be discerned depending on the oxygen pressure. These are largely determined by the stage at which magnetite starts to crystallize. If the magma has a low oxygen pressure, the liquid becomes enriched in iron and depleted in silica. At a high, constant oxygen pressure during fractionation the liquid is enriched in silica and alkalies without any iron enrichment.

The above diversity of differentiation trends can well be applied to the iron ore-bearing volcanics in northern Sweden. The Greenstone and Porphyry groups show different fractionation paths. The Greenstone group, with a relatively low degree of oxidation, has followed the trend of constant compositional fractional crystallization with an enrichment of iron, whereas the Porphyry group, with a relatively high degree of oxidation, followed the trend of constant oxygen pressure fractionation with a decrease in iron in the magma. The latest phase in the magmatic activity comprising the dyke rocks is depleted in iron. Also important in this connection is the decrease in the phosphorus content in the Porphyry group during fractionation. The fact that iron and phosphorus were withdrawn from the magma during differentiation supports the magmatic hypothesis of formation of the apatite-bearing iron ores. According to this interpretation these elements accumulated in the magma chamber and an iron-(apatite)-rich residual was injected into the volcanics at a late stage in the magmatism.

The iron-bearing volcanics of central Sweden show, with increasing silica content, a pronounced decrease in iron and an increase in alkalies. This would, according to the statements made above, indicate a high, constant pressure of oxygen during the crystallization. The degree of oxidation is, however, relatively low; the $Fe^{3+}/(Fe^{3+}+Fe^{2+})$ atomic ratio is mostly less than 0.3—0.4, but in part, up to 0.8. In addition the magnesium/iron ratio is the same for all rocks independent of the degree of solidification. The behaviour of these ratios would be in accordance with an iron concentration during fractionation. This inconsistency is possibly due to the statement made earlier that the volcanics in central Sweden are intercalated with sediments and affected by sedimentary processes.

Mineralogical features of the iron ores

The mineralogical composition of the different Precambrian iron ore types reflects the different trends of crystallization in the ore-bearing magmas. Besides iron oxides, apatite (in the apatite-bearing ores) and iron sulphides (in the skarn ores and the quartz-banded ores of the Greenstone group in northern Sweden), the ores contain silicates with calcium, magnesium, iron and to some extent manganese. The ferromagnesian minerals are particularly useful indicators of the degree of oxidation in the magma. An increased oxygen pressure results mostly in a shift towards more $Mg/(Mg+Fe)$-rich minerals. The higher the degree of oxidation, the more iron poor are the silicates. The same relationship can, however, be attained by a high sulphur fugacity. Both oxygen and sulphur are »iron-consumers» in this connection and consequently magnesia-rich silicates form.

In the apatite-bearing iron ores, Ca-amphibole and biotite are the only Mg-Fe-rich minerals of importance. Fig. 10 shows the composition of these minerals in the magnetite-hematite ore of Grängesberg in central Sweden and in the magnetite ore of Malmberget in northern Sweden. The degree of oxidation was higher in the Grängesberg deposits than in the Malmberget deposits as indicated by the higher Mg/(Mg+Fe) ratio in the silicates at Grängesberg.

In the non-manganiferous skarn ores in central Sweden, the main silicates are diopside, andradite, actinolite and hornblende (Fig. 11). The same silicates are also found in the manganiferous skarn ores; in addition they contain spessartite-almandite and other manganese-bearing minerals such as grunerite and knebelite (Fig. 12). In the manganiferous ores, there is a change towards silicates richer in ferrous iron than in the non-manganiferous ores. At the same time there is an increase in the manganese content: the highest contents are found in knebelite (27.87 % MnO) and in spessartite-almandite (9.58—23.51 % MnO), whereas relatively low contents are found in diopside (0.06—6.7 % MnO), tremolite (0.77 % MnO) and hornblende (2.45 % MnO). The presence of manganese in these ores seems to be caused mainly by compositional changes in the ore-bearing exhalations and not by changes in the environment of deposition. Possibly the ore-forming processes were of a reducing nature; this is supported by the fact that the manganiferous skarn ores contain some sulphur (up to 0.2 % S) and graphite is sometimes present in small amounts (0.5—1.5 % C).

In northern Sweden, in the skarn iron ores, the silicates are either Ca-Mg-rich (actinolite, diopside and hornblende) or Mg-rich (phlogopite-biotite, olivine and serpentine). The internal relationship between the Ca-Mg-rich silicates and the Mg-rich silicates is imperfectly known. In some deposits there seems to be a

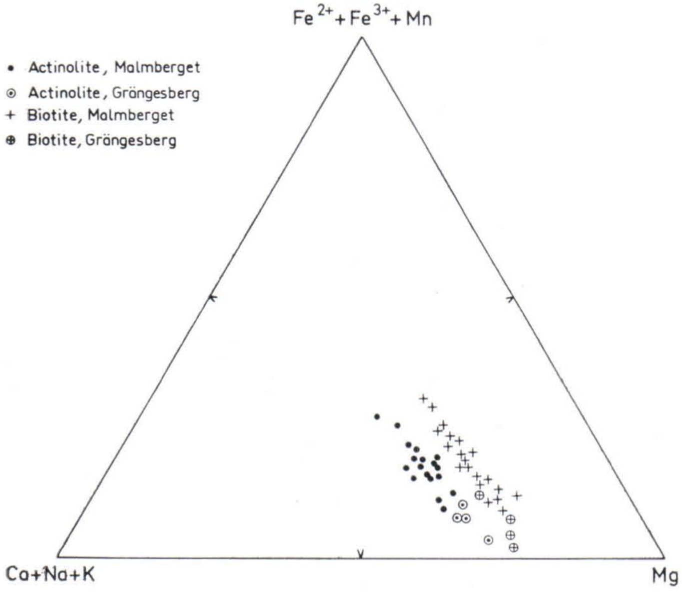

Fig. 10. The relationship between atomic % Ca+Na+K, $Fe^{2+}+Fe^{3+}+Mn$ and Mg in actinolite and biotite from the apatite-bearing iron ores Malmberget, northern Sweden and Grängesberg, central Sweden. Analyses from Annersten (1968) and Annerstren and Ekström (1971).

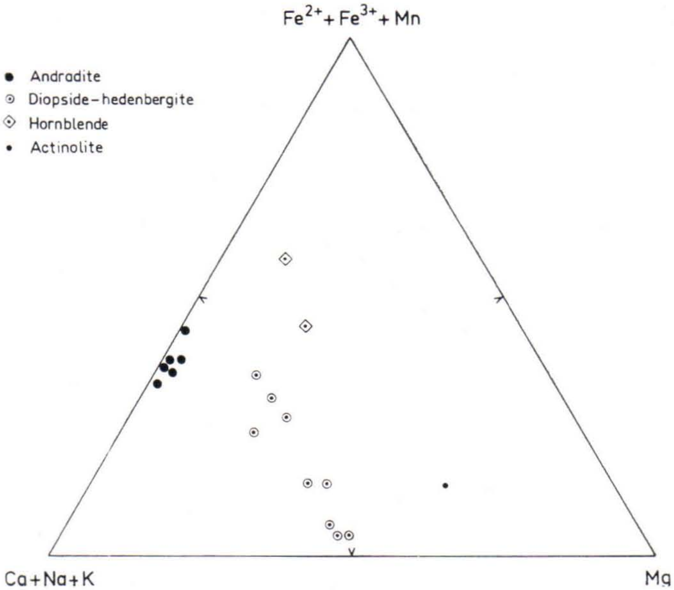

Fig. 11. The relationship between atomic % Ca+Na+K, Fe^{2+}+Fe^{3+}+Mn and Mg in silicates in the non-manganiferous skarn iron ores of central Sweden.

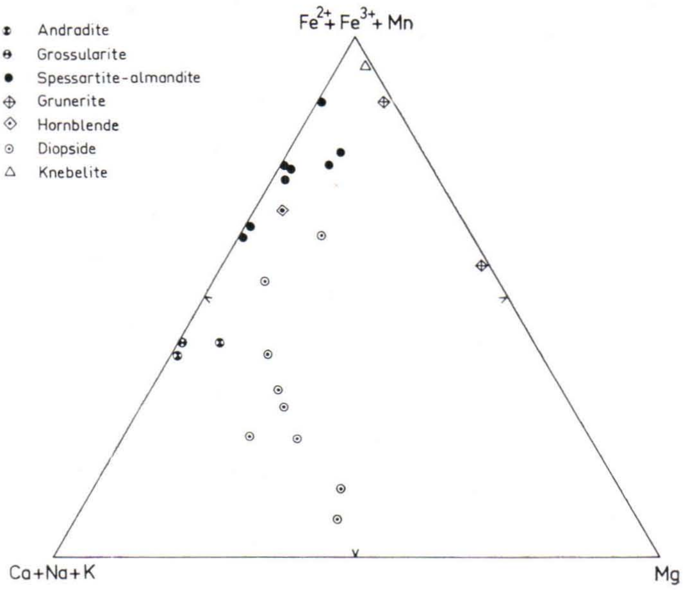

Fig. 12. The relationship between atomic % Ca+Na+K, Fe^{2+}+Fe^{3+}+Mn and Mg in silicates in the manganiferous skarn iron ores of central Sweden.

tendency for the Ca-Mg-rich silicates to form independent masses outside the ore and the Mg-rich silicates to be distributed in the ore itself.

The Mg-rich minerals are all very poor in iron (Fig. 13). Olivine and serpentine are almost iron free, and the same applies to the mica. The actinolite and diopside are also low in iron, much lower in iron than the same minerals in the skarn iron ores of central Sweden. The high Mg/(Mg+Fe) ratio in the silicates in the skarn ores of northern Sweden is not a result of oxidation — the degree of oxidation in the Green-

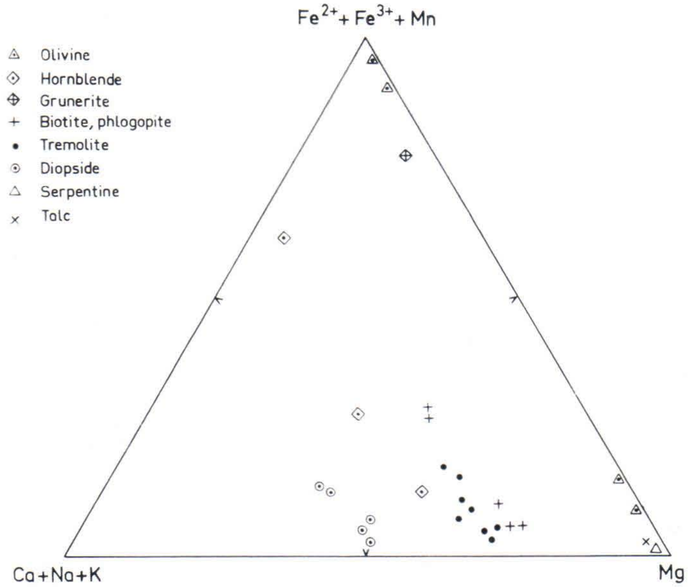

Fig. 13. The relationship between atomic % $Ca+Na+K$, $Fe^{2+}+Fe^{3+}+Mn$ and Mg in silicates in the skarn iron ores and the quartz-banded iron ores (4 upper marks) in northern Sweden.

stone group is rather low — but probably a reflection of a high fugacity of sulphur. The skarn ores and associated sediments (e. g. graphite-bearing schists and marls) are mostly relatively rich in sulphides (pyrite and pyrrhotite). In these sulphur-bearing sediments, sulphide-silicate equilibria interactions occurred during regional metamorphism. The primary silicates rich in iron or iron-magnesium reacted with sulphur to give Mg-rich silicates plus iron oxides and iron sulphides. The magnetite associated with the serpentine is mostly relatively rich in magnesia, the content reaching 5 % MgO.

The skarn minerals in the quartz-banded iron ores in the Greenstone group are ferromagnesian silicates such as hornblende, grunerite, cummingtonite, clinoenstatite, hedenbergite and almandite. Occasionally they are somewhat manganiferous. Analyses of the Marjajärvi quartz-banded ore show that the skarn minerals (4 uppermost points in Fig. 13) differ from those in the skarn ores by having a high content of iron and almost negligble content of magnesium. The manganese content varies here between 1.4 % MnO (in hornblende) and 12.4 % MnO (in olivine). In many respects this mineral assemblage is similar to that of the manganiferous iron ores in central Sweden. The presence of the manganese in the Marjajärvi deposit is also attributed to compositional changes in the ore-forming exhalations and not to changes in the environment of deposition. In the iron ores in both central and northern Sweden the appearance of manganese in the silicates in coupled with a decrease in the Mg/(Mg+Fe) ratio, giving iron-rich minerals.

References

Annersten, H., 1968. A mineral chemical study of a metamorphosed iron formation in northern Sweden. Lithos 1, 374—397.

Annersten, H. & Ekström, T., 1971. Distribution of major and minor elements in coexisting minerals from a metamorphosed iron formation. Lithos 4, 185—204.

Frietsch, R., 1977. The iron ore deposits in Sweden. *In* The iron ore deposits of Europe and adjacent areas, ed. by A. Zitzman. Explanatory notes to the international map of the iron ore deposits of Europe. Hannover, Bundesanstalt für Geowiss. u. Rohstoffe, v. 1, 279—293.

— **1978.** On the magmatic origin of iron ores of the Kiruna type. Econ. Geol. 73, 478—485.

Geijer, P., 1931 a. Berggrunden inom malmtrakten Kiruna—Gällivare—Pajala. Sveriges geol. Unders. C 366. 225 p.

— **1931 b.** The iron ores of the Kiruna type. Sveriges geol. Undersökning C 367. 39 p.

Geijer, P. & Magnusson, N. H., 1952. The iron ores of Sweden. XIX Congr. géol. int. Alger 1952, Symp. Gisem. Fer Monde, v. 2, 477—499.

Irvine, T. N. & Baragar, W. R. A., 1971. A guide to the chemical classification of the common volcanic rocks. Can. J. Earth Sci. 8, 523—548.

Kuno, H., 1968. Differentiation of basalt magmas. *In* Basalts, ed. by Hess, H. H. and Poldervaart, A. Interscience Publishers, New York, 623—688.

Magnusson, N. H., 1970. The origin of the iron ores in central Sweden. Sveriges geol. Unders. C 643. 364 p.

Oelsner, O., 1961. Zur Genese der nord- und mittelschwedischen Eisenerzlagerstätten. Geologie 6, 601—622.

Parák, T., 1975. The origin of the Kiruna iron ores. Sveriges geol. Unders. C 709. 209 p.

ASSOCIATION BETWEEN PRECAMBRIAN ORE DEPOSITS AND BASITE-ULTRABASITE INTRUSIVES ON THE KOLA PENINSULA

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The Kola Peninsula represents an area of extensive development of basic-ultrabasic magmatism and sulphide copper-nickel mineralization genetically related to it. Analysis of geological, tectonic, magmatic and metamorphic phenomena showed that the process of formation of the geological structure in the peninsula covered a large period of time: from the Lower Archean to the beginning of the Paleozoic. The process is divided into 4 tectonic- magmatic cycles:

1) the Belomorian (Saamian);
2) the Karelian;
3) the Hyperborean; and
4) the Caledonian-Hercynian (Geology of the USSR, 1958).

Two large stages are, in turn, distinguished within each tectonic-magmatic cycle; the first half of each stage is characterised by sedimentation and volcanic activity in the geosyncline zones, and the second half by intense folding and intrusive activity.

The Archean crystalline basement, which is composed of various gneisses, schists, amphibolites, and granulites, with beds of ferruginous quartzites, was formed during the Belomorian (Saamian) tectonic-magmatic cycle, the upper boundary of which has been dated at 2 600 $\pm$ 100 Ma. The rocks are everywhere compressed into steep isocline folds trending northwest and penetrated by intrusions of the two granite groups. They form elongated massifs and migmatite areas with micaceous, pegmatitic and lithium mineralizations.

Basic-ultrabasic magmatism in the Lower Archean is represented by volcanites of andesite-basaltic composition and comagmatic gabbroid sills, peridotite and lherzolite intrusions, forming in the gneissic areas clusters of bodies or in some places individual interstratal bodies along the boundaries of different rocks, not infrequently along the linearly elongated structural-metamorphic zones. No regularities have been observed in their distribution. All these intrusive bodies and their country rocks were subjected to regional metamorphism under PT conditions of granulite and hyperamphibolite facies. In a number of cases this has given rise to the »drusite structure», which distinguishes them from later basites and ultrabasites. This earliest stage of basic-ultrabasic magmatism produced no commercial copper-nickel deposits and was characterized only by an extremely weak manifestation of ore generating processes in the form of scattered and irregular dissemination of titanomagnetite, pyrrhotite, pyrite and more rarely chalcopyrite.

In the Early Archean, during the Lapponian phase of folding, extensive plagiomicrocline granite intrusions were emplaced succeeding the lens- and cigar-shaped bodies of norites, gabbro-norites, and gabbro-anorthosites that were also metamorphosed under PT conditions of granulite and amphibolite facies. The copper-nickel mineralization in the vicinity of Lake Lovno is genetically related to this group of basites. Ore-bearing norite and gabbro-norite intrusives occur in chains trending northeast in conformity with schistosity of the rocks; they are elongated lenses or flattened cigars in shape, from 100 to 800 m long. Disseminated ores prevail in the orebodies of the Lovnoozero deposit; massive and brecciated ores are of minor extension. The ore bodies are sometimes intersected by veins of granites and hypersthene diorites (Kozlov 1973).

The Karelian tectono-magmatic cycle, which, as is generally accepted, corresponds to the formation of the Proterozoic folded belts, appears to have been most productive in developing both basic-ultrabasic magmatism and copper-nickel mineralization. The Proterozoic belts overlie the Archean crystalline basement as narrow bands filling deep depressions, troughs and intercupola spaces in the formerly much more extensive geosynclinal area. Confined to the axial part of the Kola Peninsula two parallel Karelide belts are distinguished:

1) the Kola—Keiv, and
2) the Pechenga—Varzuga structural facies zones.

The Kola—Keiv zone, which is composed of biotite-garnet-staurolite gneisses, quartzites, metavolcanites, kyanite and other schists, displays numerous interformational bed-like bodies of gabbro-labradorites. They occur in chains along the linearly trending boundaries of the zone, not infrequently penetrating both the underlying Archean gneisses and the Proterozoic schists. The length of the massifs reaches 10—20 km and their thickness 1 500—2 500 m. The Tsaginsk and Achinsk gabbro-labradorite massifs with commercial titano-magnetic mineralizations are related to this group of intrusives. The sulphide copper-nickel mineralizations occur only in some layered bodies of orthoamphibolites and gabbro-labradorites; they are, however, subeconomic. All the known sulphide copper-nickel deposits on the Kola Peninsula, except the forementioned Lovnoozero deposit, are confined to the Pechenga-Varzuga structural-facies and the metallogenic Karelide zone striking from northwest to southeast across the whole Peninsula. The development of the zone was typical of a geosyncline (Gorbunov, 1968).

The lower structural stage, comprising tundra and other series of volcanogenic-sedimentary rocks with sills of metabasites, was formed during the Lower Proterozoic. Tectonic movements of that period caused steep isoclinal folding and displacements in sedimentary-volcanogenic rocks, as well as to numerous faults in the adjacent Archean gneisses along which basic-ultrabasic magma was injected.

Small intrusives of pyroxenites, dunites, and harzburgites (serpentinites), located along abyssal faults in the Archean basement and forming belts of up to 150—200 km in length oriented in the direction of the general strike, were initially connected with this stage of magmatic activity. Numerous massifs of ultrabasites of the »Serpentinite Belt» (the Pados, Chapes and other tundras) may also be related to it. They delineate the deeply eroded branch of the Pechenga—Varzuga folded belt, in which, near the Finnish border, only remnants of volcanogenic-sedimentary rocks are preserved. The ultrabasites are characterized by a high iron content (f = 10—12), chromite and magnetite mineralization and a very poor sulphide copper-nickel mineralization in serpentinites. Later large interformational intrusions of basite-ultrabasites were emplaced along large faults at the contacts with the Archean gneisses. In composition and metallogeny they can be divided into three groups:

a) gabbro-labradorites with titano-magnetite mineralization (the Main Ridge massif of the Monche—Chuna—Volchji and other tundras);
b) slightly differentiated massifs of norites and gabbro-norites with poorly disseminated sulphide copper-nickel ores (the massifs of Mount Generalskaya in the Pechenga region, and Fedorovy and Panskie tundras in the Imandra—Varzuga region);
c) nickel-bearing differentiated massifs of ultrabasic and basic rocks: peridotites (lherzolites), pyroxenites, and gabbro-norites. Related to this group are the Nittis—Kumushjya—Travyanaya massif and the Sopcha in the Monchegorsk ore region with its known deposits of massive vein-type sulphide ores and deposits of disseminated ores, as well as the differentiated massif of olivinites, peridotites and pyroxenites of the mountain Zasteid-II, which is located in the northeastern contact of the granulitic formation with biotite-amphibolite gneisses and contains, in the endocontact, a low-grade disseminated copper-nickel mineralization.

The upper structural stage of the Pechenga—Varzuga zone was formed during the Middle Proterozoic. The thick (over 10 000 m) Pechenga series of sedimentary-volcanogenic rocks: diabases (spilites), tuffs, tuffites, sandstones and phyllites with sills of gabbro-diabases developed at its northwestern wing; the Imandra—Varzuga series of almost analogous composition developed at the southeastern wing. Basic effusives predominate in both series. Owing to the culmination of folds and deep erosional truncation no sedimentary-volcanogenic rocks have survided in the area between the wings.

During the Late-Karelian phase of folding these rocks were compressed into minor folds and penetrated by numerous intrusions of nickel-bearing basic-ultrabasic magma that ascended along the system of deep faults whose roots penetrated deep into the mantle. At the same time, at the northwestern wing, the main mass of differentaited nickel-bearing intrusives of gabbro-pyroxenite-peridotite (wehrlite) composition were emplaced mainly as rootless interstratal bodies, phacoliths, within the dislocated tuffogenic-sedimentary rocks, where they formed a belt of ultrabasic intrusions that include the commercial copper-nickel deposits of the Pechenga ore field. At the south-eastern wing, within the Imandra—Varzuga series, the intrusive activity manifested itself in the formation of interstratal bodies of gabbro-diabases, gabbro-norites, and very rarely, gabbro-peridotites with low-grade sulphide dissemination. All the rocks and ores were subjected to regional metamorphism under PT conditions of greenschist facies.

During the development of the Pechenga—Varzuga structural-metallogenic zone, a large number of small intrusive bodies of ultrabasites were emplaced from both sides of the consolidated blocks of archeaides adjacent to the zone along the abyssal faults of the northwestern and northeastern trends. The intrusions often occur along their feathery branches in the contacts of amphibolites and gneisses. Two age groups of ultrabasites are distinguished:

a) the older and more widely spread group of massifs of peridotites (serpentinites), pyroxenites, and feldspathic pyroxenites, which have no noticable accumulations of copper and nickel whose formation can be allied with the Early Karelian folding phase;
b) the group of small massifs of serpentinized peridotites (harzburgites) with rich sulphide copper-nickel mineralization and which resemble the Pechenga nickel-bearing intrusions in age, rock and ore compositions and degree of metamorphism. The copper-nickel deposits of the Allarechka ore field, which are situated in the gneiss area south of the Pechenga

Table 1

The scheme of the development of basic-ultrabasic

Tectono-magmatic cycle	Stage of develop-ment	Age: Group	Age: Subgroup	Age: Ma	SEDIMENTARY-VOLCANOGENIC COMPLEXES
1	2	3	4	5	6
Caledonian-Hercynian	Platform	Paleozoic	Lower	380 ± 400	Conglomerates, quartzites, Turi Cape sandstones etc.
Hyperborean (Baikal)			Upper	600 900 ± 100	Clay schists with dolomitic intercalations, sandstones, conglomerates from Rybachi Peninsula
Karelian	Geosynclinal	Proterozoic	Middle	1 650 ± 100	Pechenga and Imandra—Varzuga series of thick augitic diabase sheets (spilites) alternating with phyllites, sandstones, tuffites, tuffs, quartzites, dolomites, basal conglomerates
			Lower	1 900 ± 100	Quartzites, mica schists, schistose amphibolites with greenstone layers (Tundra and other series). Kyanite, staurolite-micaceous schists, quartzites, biotite-garnet-staurolite and amphibole gneisses (Keiv and other series)
Belomorian (Saamian)	Progeo-synclinal	Archean	Upper	2 600 ± 100	Granulite complex, gneisses and schists of iron-ore series
			Lower	~3 500	Garnet-biotite, micaceous and amphibolite gneisses, feldspathic amphibolites and migmatites

magmatism and its metallogeny on the Kola peninsula

BASITE-ULTRABASITE MAGMATIC COMPLEXES			METALLOGENY
Effusions	Intrusions		
	Composition	Form and conditions of occurrence	
7	8	9	10
	Complex of ultrabasic-alkaline rocks: olivinites, pyroxenites, ijolites, melteigites, urtites, nepheline syenites	Massifs of central type, stocks	Commercial deposits of apatite-magnetite (Kovdor) titanium-magnetite (Africanda) and other ores
Picrite-porphyrites, augite porphyrites, alkaline basalts			
	Complex of diabases, olivine gabbro-dolerites, gabbro-dolerites	Steeply dipping dykes in Archean gneisses and granites, sandstones and schists	Barite-polymetallic veins of Murmansk coast
	Nickel-bearing olivinites, peridotites (wehrlites), pyroxenites and gabbros	Interstratal steeply inclined bodies, phacoliths	Commercial copper-nickel deposits of Pechenga and Allarechka regions
Augite diabases, spilites	Gabbro-diabases	Sills	
	Gabbro-norites, pyroxenites, peridotites (lherzolites), olivinites	Interformational differentitated nickel-bearing massifs	Commercial copper-nickel deposits of Monchegorsk region
	Norites and gabbro-norites	Interformational steeply and gently dipping large massifs	Low-grade disseminated copper-nickel ores (massifs of Fedorovy and Panskie Tundras and Generalskaya Mt.)
	Gabbros, gabbro-labradorites	Large interformational massifs	Titanomagnetitic deposits (Tsaginsk, Achinsk)
	Pyroxenites, dunites, harzburgites	Lens-shaped bodies confined to deep faults in Archean gneisses	Low-grade chromite and sulphide dissemination (massifs of Serpentinite Belt and others)
Spilites, diabases, mandelstones	Gabbro-diabases, pyroxenites	Sills	Pyrite-pyrrhotite dissemination
	Gabbro-norites, norites, gabbro-anorthosites, pyroxenites	Lens-shaped mostly interstratal bodies	Low-grade disseminated copper-nickel ores (Lovnoozero deposit)
Basic volcanites			Ferruginous quartzite deposits
	Gabbro-anorthosites, gabbro-labradorites, gabbros, norites, peridotites	Lens-shaped interstratal bodies	Disseminated pyrite-pyrrhotite and titanium-magnetite mineralization
Volcanites of andesite-basaltic composition			

deposits, are related to this group of ultrabasite. Intrusions of nickel-bearing serpentinized peridotites are also encountered among the Archean gneisses to the north of Pechenga deposits and in the area between the Monche tundra and the Pechenga (Priozernoe ore showing).

All the copper-nickel deposits of this age group are characterized by bed- and lens-like ore bodies of rich disseminated ores, which repeat the curves of the surface of the lower contacts of the parent intrusions. These simple forms are often accompanied by joint veins of massive sulphide and breccia ores, which are confined to the dislocations with a break in continuity along the contacts of ore-bearing intrusions and to the fractures of exfoliation in the country rocks.

Hence, the formation of copper-nickel deposits on the Kola Peninsula took place in the closing period of Karelian folding during the early stage of transformation of the geosyncline, (although at the early stages of its development) into a stable fold belt, the latest stages of which seem to be completely lacking. The absence on the Kola Peninsula of granites younger than the nickel-bearing intrusions allows the ore metamorphism to be connected with the postmagmatic stage of formation of ore-bearing intrusions and with regional metamorphism (Gorbunov 1968).

During the Hyperborean tectono-magmatic cycle magmatic activity manifested itself mainly in the peripheral parts of the Kola Peninsula as dyke series of diabases, gabbro-dolerites and olivine gabbro-dolerites injected along a system of vertical fractures trending northeast. The barite- polymetallic mineralization on the coast of the Barents and White Seas is associated with them.

During Postproterozoic epoch the Kola Peninsula was subjected to repeated tectono-magmatic activation, that expressed itself in renewed volcanic activity and the injection of subvolcanic intrusions of ultrabasic alkaline rocks; olivinites, pyroxenites, ijolites, melteigites, urtites and nepheline syenites. They are characterized by magmatic differentiation that gave rise to ultrabasites, alkaline rocks and carbonatites. The magma was undersaturated with silicium and aluminium and sodium had a distinct prevalence over potassium (Geology of the USSR 1958).

The formation of the Kovdor ultrabasic alkaline massif and its apatite-magnetite and phlogopite-vermiculite deposits, the Africanda, Salma-mountain, Habozero, Vuorijärvi and other massifs, is connected with these processes. The Gremyakha-Vyrmes massif should probably be related to this group, too.

The occurrence of alkaline rocks among Lower Paleozoic basic-ultrabasic massifs seems to have been a precursor of the intense development of alkaline intrusions in the Kola Peninsula, such as Khibiny and Lovnoozero with their known apatite and other deposits, which took place in the following Hercynian Epoch.

The outlined sequence of basic-ultrabasic magmatic evolution and nickel metallogeny against the background of the development of the geological structure of the Kola Peninsula is somewhat schematic; it is shown in reduced form in Table 1.

From the information available it follows that the intensity of the basic-ultrabasic magmatism, beginning with the Archean, increased continuously and reached its zenith in the Middle Proterozoic, in the period of completion of Karelian folding and transformation of the geosyncline into a fold belt. In the Upper Proterozoic and Lower Paleozoic the formation of complexes of diabase dykes followed by separate massifs of ultrabasic-alkaline rocks was connected with tectono-magmatic activation of the platform, synchronous with the Hyperborean and Caledonian-Hercynian epochs of tectogenesis in adjacent areas.

On the basis of chemistry, mineralogical composition and metallogenic specialization, the ultrabasites of the progeosynclinal stage of development were derived from under-saturated olivine-tholeiites. The basite massifs are probably derivatives of the same magma. During the stage of Proterozoic geosyncline development, massifs of ultrabasites with a higher iron content and chromite mineralization were initially formed in connection with the increase in fracture depth. Later, as the depth of the magmatic chamber in different structure-facies zones changed, large and mostly interformational intrusives were formed as derivates of the picritic and gabbroid magmas. They proved to be the most productive of copper-nickel and titanium ores. During the Paleozoic ultrabasic-alkaline magmatism with iron titanium rare metal metallogeny manifested itself widely in different parts of the Kola Peninsula. This magmatism was not, however, connected with the basite-ultrabasites of the calc-alkalic or ferromagnesian series and which, in its turn, was a precursor of the unique Khibiny and Lovnoozero alkaline massifs.

References

Geology of the USSR, 1958. Vol. 27, Murmansk Region, part 1. Moskva, »Gosgeoltekhizdat». 710 p. (In Russian)

Gorbunov, G. I., 1968. Geology and genesis of sulphide copper-nickel deposits of Pechenga. Moskva, »Nedra». 352 p. (In Russian)

Kozlov, Ye. K., 1973. Natural series of rocks of nickel-bearing intrusions and their metallogeny. Leningrad, »Nauka». 286 p. (In Russian)

DISLOCATION METAMORPHISM AND ENDOGENEOUS ORE FORMATION IN THE PRECAMBRIAN

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In this paper the term »dislocation metamorphism» is used, not as a synonym of cataclastic metamorphism, but as a more general conception embracing the whole complex of metamorphic phenomena in narrow zones of intense deformation (Turner & Verhoogen 1960). The processes of dislocation metamorphism play a key role in the formation of various Precambrian endogenous mineral deposits: they transform the more ancient ore bodies, cause remobilization of ore material and create tectonic structures favourable for the localization of metamorphic, postmetamorphic and postmagmatic mineral deposits.

Many Precambrian sedimentary-volcanogenic deposits of massive sulphides underwent metamorphism accompanied by folding, block movements and transformations of ore bodies. In mildly metamorphosed rocks they retain their original appearance, in strongly altered rocks show indications of redistribution of sulphide ores.

A good example is the Kholodninskoe lead-zinc deposit recently discovered near Lake Baikal in metasedimentary and metavolcanic rocks of Proterozoic age (Fig. 1). They fill a narrow through striking north-east and framing the Siberian platform. The degree of metamorphism usually corresponds to greenschist and epidote-amphibolite facies, except in some narrow tectonic zones where it attains amphibolite facies.

The fault zones enclose porphyroblastic high-temperature metasomatites, quartz-muscovite schists and regenerated lead-zinc mineralization that differs from stratiform ores and is interesting more from a genetical than a practical point of view. The stratiform orebodies consist mainly of schistose and recrystallized banded sulphide ores; the regenerated mineralization is represented by veinlets and impregnations of galena, sphalerite, pyrite, and pyrrhotite (Kuznetsov *et al.* 1977, Ruchkin *et al.* 1975).

Progressive dislocation metamorphism was among the main genetic factors in the sulphide nickel deposits in the Norseman-Viluna belt in Western Australia. According to R. Gee, D. Groves and C. Fletcher (1976), the presence of magmatic ores in these deposits is indicated by the basal location of disseminated sulphide minerals in differentiated lava flows and in the central parts of dunitic intrusions. However, the common occurrence of ores in dynamic-style belts, the structural control, strong metamorphic layering, and tectonic fabrics of ores all argue for a close relation between dynamo-metamorphism and the present location of rich nickel ores. The metamorphic grade of Archean sedimentary and volcanic rocks varies from prehnite-pumpellyite to high amphibolite facies (Fig. 2). The dominant style of low- to medium-grade metamorphic domains is that of static recrystallization resulting in preservation of primary textures and structures. In contrast,

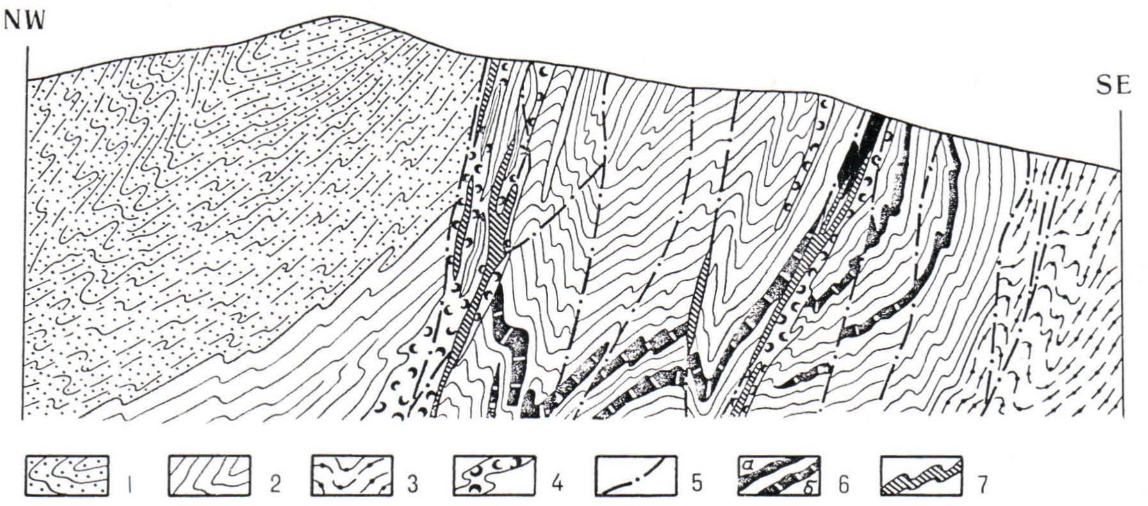

Fig. 1. Cross-section of the Kholodninskoe lead-zinc deposit (Ruchkin *et al.* 1975): 1 = quartzites, quartz-mica-garnet schists; 2 = quartz-carbonate, graphite schists; 3 = quartz-garnet-mica schists, quartzites, marbles; 4 = porphyro-blastic high-temperature metasomatites; 5 = faults; 6 = stratiform sulphide orebodies (a — determined, b — supposed); 7 = regenerated lead-zinc mineralization.

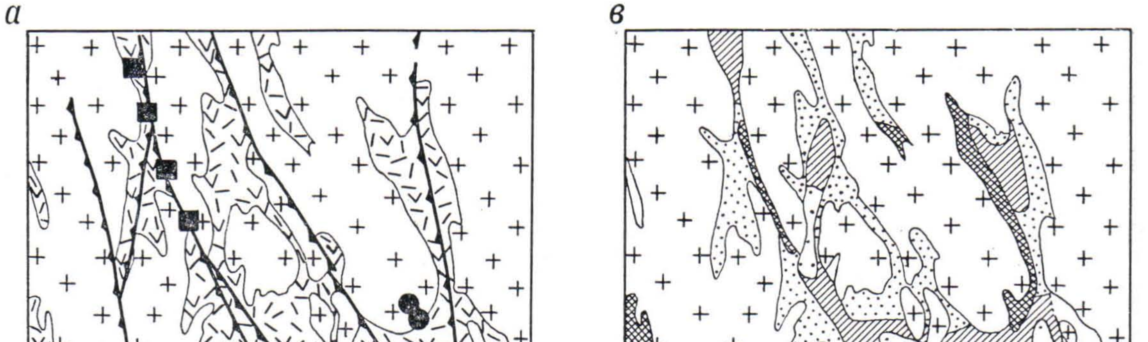

Fig. 2. Geological position of nickel ore deposits (a) and metamorphic domains (b) in the northern part of the Norseman-Wiluna belt (Gee *et al.* 1976): 1 = granitoids; 2 = metavolcanics and metasedimentary rocks; 3 = faults; 4 = volcanic type of deposits; 5 = dunitic intrusive type of deposits; 6—10 = metamorphic facies: 6 = prehnite-pumpellyite to low greenschist; 7 = low to upper greenschist; 8 = low to middle amphibolite; 9 = high amphibolite; 10 = no information available.

in high-grade areas of dynamic-style metamorphism, rocks are sheared. Most economic deposits are confined to middle-upper amphibolite facies domains of dynamic-style. The sulphides are intergrown with metamorphic silicate minerals and have experienced the same metamorphic processes as the enclosing rocks. The ores typically have tectonic fabrics, and pyrrhotite exhibits intense physical and optical orientation together with kink bands.

Regional zones of blastomylonites determine the localization of ore-bearing alkaline metasomatites, which represent a special type of postmetamorphic hydrothermal deposits. The formation of alkaline metasomatites is related to protoactivization of early Precambrian tectonic structures in the Earth's crust. The same geological position is characteristic for the Proterozoic hydrothermal uranium deposits in the Lake Athabaska region, where Archean gneisses and granites of the Tazin group are intersected by large faults that control uranium deposits (Beck 1970). Originally it was considered that the uranium ores at the Fay mine were located in argillite beds. But J. Krupička and G. Sassano (1972) discovered that the so-called argillites were formed as a result of multistage deformation and regressive metamorphism. The relatively older tectonites are represented by quartz-biotite-feldspar blastomylonites, the younger ones by cataclasites and breccias; both of them originated before uranium mineralization.

In regions of tecto-magmatic activization large rejuvenated faults of the crystalline basement control the distribution of postmagmatic hydrothermal ore mineralization. For example, in the central part of the Aldan shield many faults were formed in the Early Proterozoic, when dykes of metadiorites injected into the narrow tectonic zones (Fig. 3). Later on, the dykes and enclosing Archean gneisses, the crystalline schists, were transformed into biotite-amphibole blastomylonites and blastocataclasites accompanied by high-temperature quartz-feldspar metasomatic rocks. Mesozoic activization resulted in intensive magmatic and hydrothermal activity. The bulk of the Mesozoic intrusives and dykes occur in the crystalline basement independently of the Early Proterozoic faults, whereas low-temperature gold-bearing quartz-orthoclase metasomatites usually inherit the zones of blastomylonites. In contrast to blastomylonites formed through plastic flow, the mineralized disjunctives of Mesozoic age are characterized by brittle deformation and the development of brecciated, cataclastic and stockwork textures.

Investigations of the internal structure of ore-bearing faults in the crystalline basement permitted the author (1972) to develop the ideas of V. Kreyter (1956) concerning the vertical zonality of tectonic fractures and to distinguish five depth levels of dislocation metamorphism. These levels are characterized by different mineral associations and textures of deformed rocks and are represented by different structural element: I) zones of brecciation and fracturing; II) disjunctive displacements with gouges; III) epidote-chlorite mylonites and cataclasites; IV) biotite-amphibole blastomylonites and blastocataclasites; V) injection migmatites.

The structural elements of levels IV and V, formed under the conditions of amphibolite facies, as a result of recrystallization, partial melting of enclosing rocks, and protoclasis are of special interest for understanding the relationship between dislocation metamorphism and endogenous ore mineralization. The tectonites of amphibolite facies are similar to regionally metamorphosed rocks in mineral association, but differ from them in structure, texture and orientation of metamorphic minerals.

Geological information about these structures is, however, very limited, which is why special investigations have been carried out, e. g. on regional faults associated with the Kirovograd-Novoukrainsk pluton (Fig. 4). The pluton is located in the central part of the Ukrainian Shield among Early Proterozoic gneisses

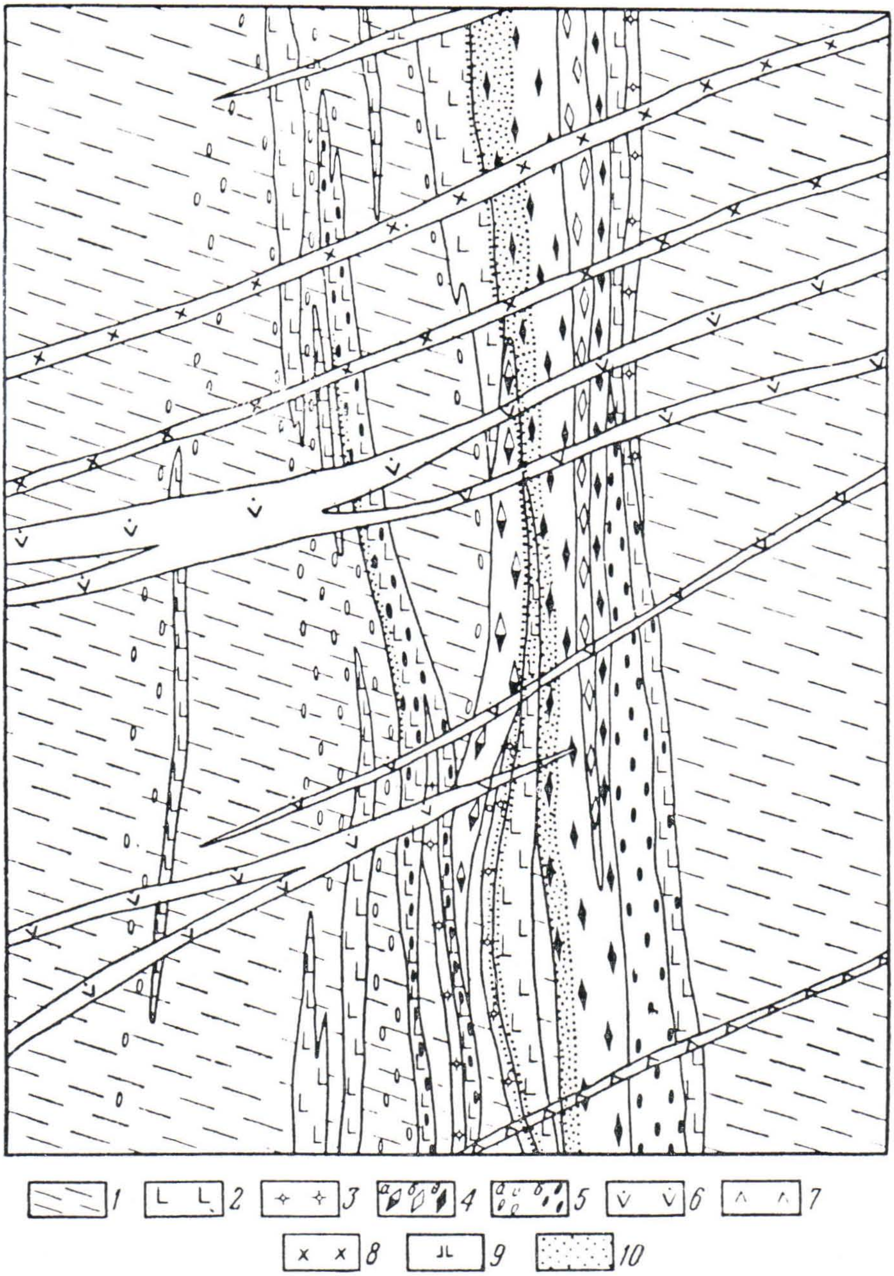

Fig. 3. Relationship between Mesozoic dykes and rejuvenated Early Proterozoic fault in cross-section (Kazansky *et al.* 1971): 1 = gneisses and schists; 2 = metadiorites; 3 = quartz-feldspar rocks; 4 = blastomylonites; 5 = blastocataclasites; 6 = aegirine-augite syenite porphyries; 7 = hornblende syenite diorites; 8 = aegirine syenite porphyries; 9 = orthophyres; 10 = low-temperature quartz-orthoclase metasomatites.

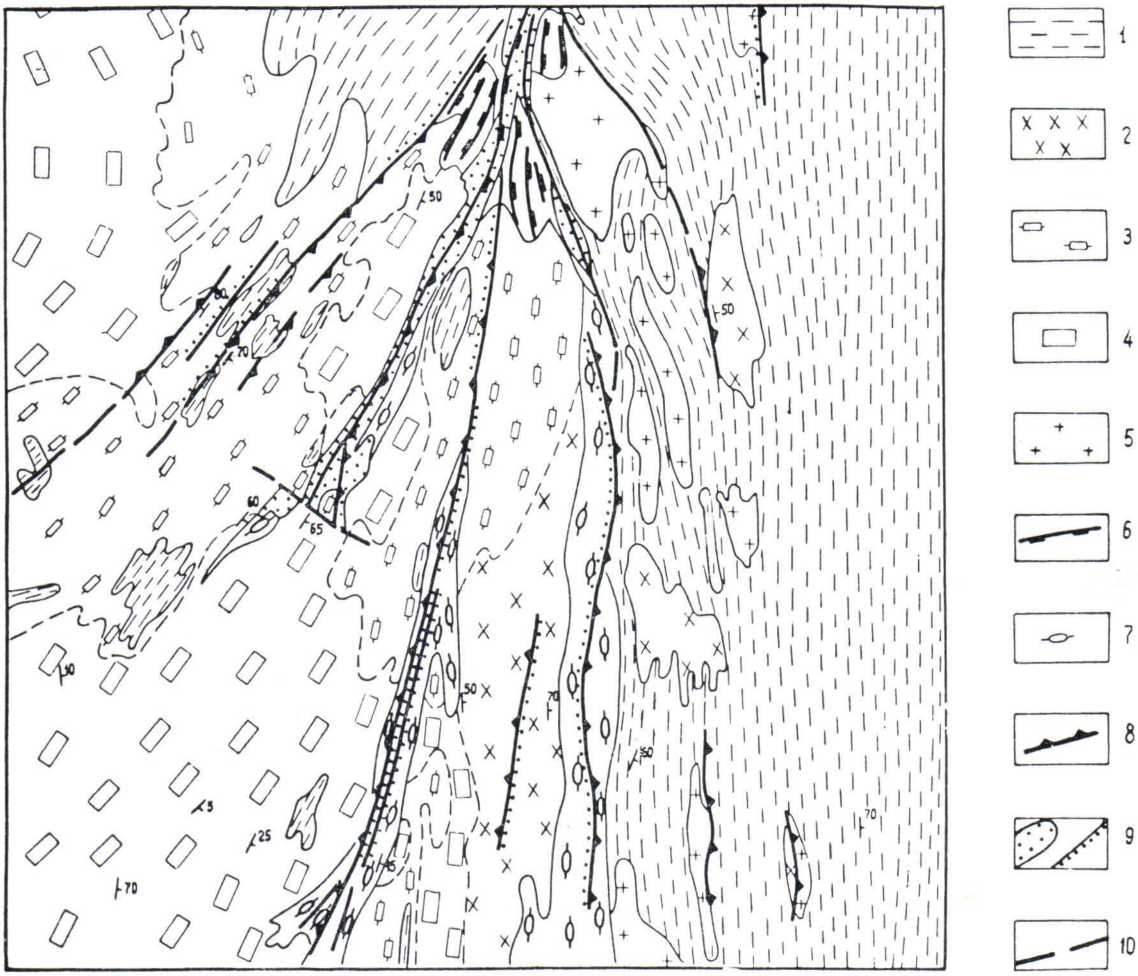

Fig. 4. Zones of dislocation metamorphism as related to the Kirovograd—Novoukrainsk pluton (plan): 1 = gneisses; 2—4 = granitoids of the first phase (2 = marginal, 3 = apical, 4 = central facies); 5 = granitoids of the second phase; 6 = injection migmatites; 7 = zones of protoclasis; 8 = biotite-amphibole blastomylonites; 9 = epidote-chlorite mylonites and cataclasites; 10 = faults.

and was formed in two phases 2 000—1 800 Ma ago (Kazansky 1978). Its eastern border coincides with regional tectonic zones marked by injection migmatites. Structural and petrological investigations show that after crystallization granitoids of the first phase were subjected to intense protoclasis. The granites of the second phase intruded into the zones of protoclasis. Later, granitoids of both phases were converted into biotite-amphibole blastomylonites and blastocataclasites and then underwent diaphtroresis of greenschist facies.

Regular variations of strain during formation of the Kirovograd-Novoukrainsk pluton and associated faults are clearly reflected in petrostructural diagrams of the granitoids and tectonites (Fig. 5). In the massive granitoids of the first and second phases and in the surrounding gneisses the optical axes of quartz grains show no preferred orientation, whereas in the zones

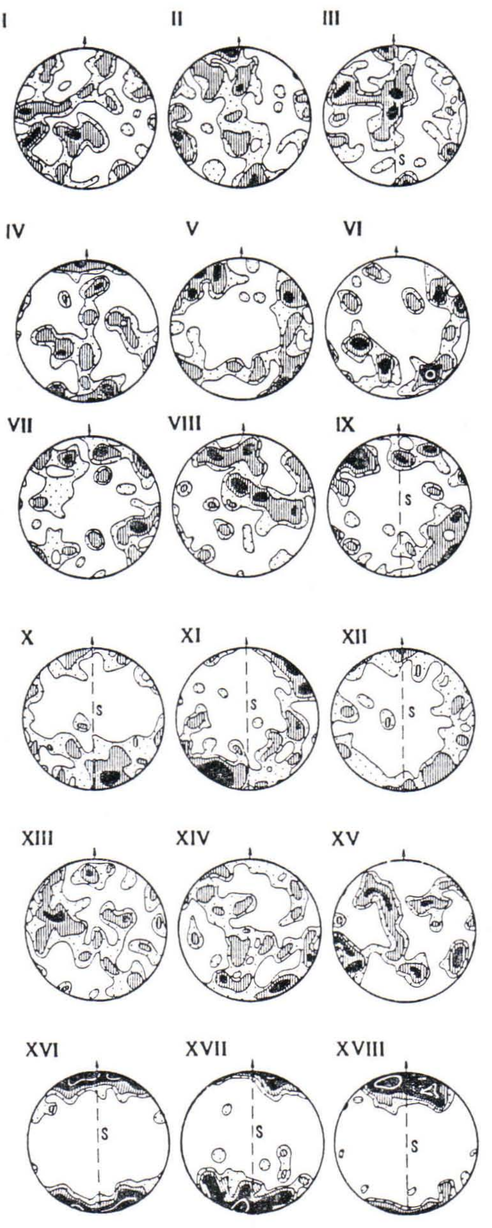

Fig. 5. Orientation of optical axes of quartz grains in vertical cross section: I—III = gneisses; IV—IX = granitoids of the first phase; X—XII = granitoids in the zones of protoclasis; XIII—XV = granitoids of the second phase; XVI—XVIII = blastomylonites. Density 1—2—3—8 %.

of protoclasis they are arranged according to the R-type tectonite, and in the blastomylonites they correspond to the S-tectonite with a single maximum whose centre coincides with the dip. This orientation testifies to vertical movements along the faults.

It should be emphasized that gneisses, granitoids, blastomylonites and blastocataclasites are all characterized by the same mineral assemblages and by similar composition of rock-forming minerals such as feldspars, amphiboles and biotites.

Development of structural elements of the Levels IV and V at high temperatures and pressures is reflected in their petrophysical properties. In contrast to the brecciated and fractured rocks of the upper level the tectonites of amphibolite facies hardly differ at all from enclosing ultrametamorphic rocks in density, porosity and elastic properties (Table 1). Hence these structural elements are poorly identified by geophysical methods.

In rejuvenated faults the upper structural elements are usually superposed on the deeper ones, a feature that is considered as a manifestation of regressive dislocation metamorphism. Some tectonic structures correspond to the progressive branch of dislocation metamorphism, e. g. the shear zones in the Petchenga complex on the Kola Peninsula. The Petchenga complex consists of metavolcanic and metasedimentary rocks of Middle Proterozoic age and contains magmatic sulphide copper-nickel deposits subjected to metamorphism together with ore-bearing ultrabasic intrusions (Gorbunov 1968). The temperature of metamorphism increased from the top to the bottom of the Petchenga complex (Duk 1977, Zagorodny *et al.* 1964). The increase in metamorphic grade from prehnite-pumpellyite to amphibolite facies was accompanied by deep changes in the textures and structures of the deformed rocks (Fig. 6 a). These changes are most distinct in the transitional zone between greenschist and epidote-amphibolite facies where massive relict textures of

Table 1.

Petrophysical properties of gneisses, granitoids and tectonites after L. Zwjagintsev (Kazansky 1978).

Groups of rocks	Density g/cm³	Effective porosity %	V_p km/sec	V_s km/sec	Modulus of elasticity $E \cdot 10^5$ kg/cm²	Poisson coefficient
1. Gneisses	2.66	1.08	5.20	3.03	6.41	0.23
2. Granitoids of phase I	2.63	0.70	4.74	2.65	4.92	0.25
3. Granitoids in zones of protoclasis	2.67	0.73	5.05	2.72	5.19	0.28
4. Granitoids of phase II	2.63	0.74	5.08	2.91	5.98	0.25
5. Blastomylonites	2.61	1.45	4.75	2.84	5.59	0.18

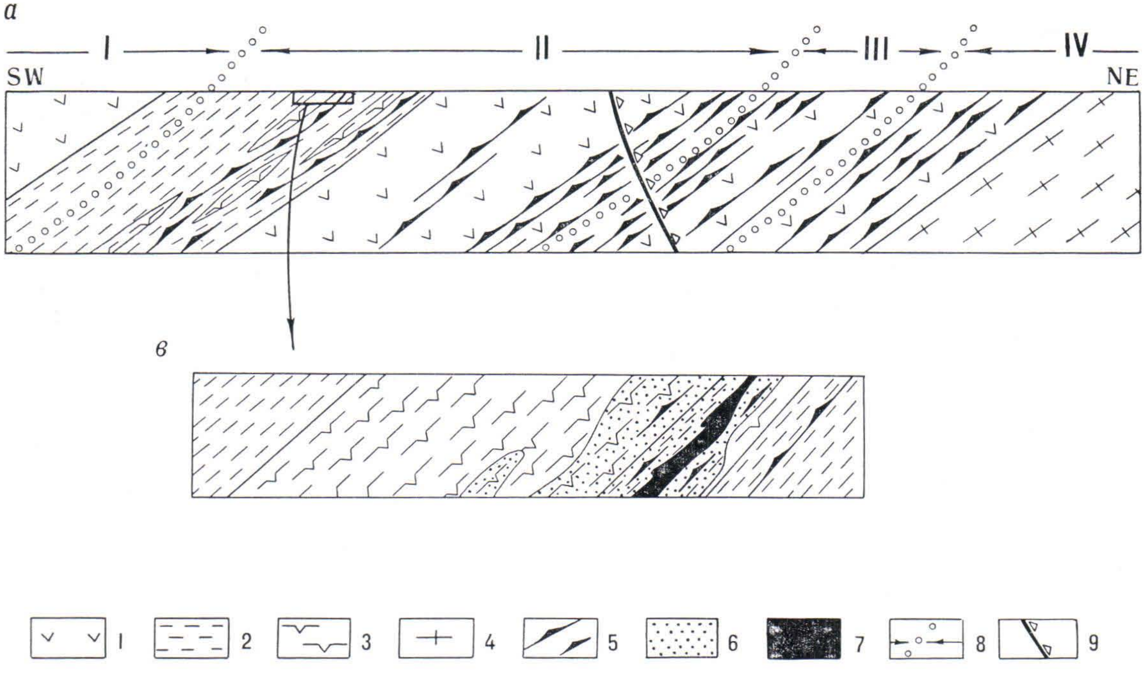

Fig. 6. Relationships between dislocation metamorphism (a) and sulphide copper-nickel mineralization (b) in the Petchenga complex (in section): 1 = metavolcanic; 2 = metasedimentary rocks; 3 = metaperidotites; 4 = gneisses; 5 = schistosity, 6 = disseminated ore, 7 = rich ore; 8 = facies of progressive metamorphism (I = prehnite-pumpellyite, II = greenschist, III = epidote-amphibolite, IV = amphibolite); 9 = regressive greenschist facies.

metavolcanic rocks grade into crystalline-schistose texture with progressive replacement of actinolite by hornblende containing 10—15 % Al_2O_3.

Schistose metavolcanic and metasedimentary rocks reveal blastomylonites and blastocataclastic textures and well-defined orientation of metamorphic minerals (Fig. 7). On the vertical cross-sections the optical axes of quartz and carbonate grains show preferred orientation of the R-tectonite type: chlorite, biotite and muscovite lamellae are arranged along schistosity planes while prismatic grains of amphiboles are elongated parallel to »b»-axis of R-tectonites. These data are compatible with the interpretation of the tectonic structures of the Pet-

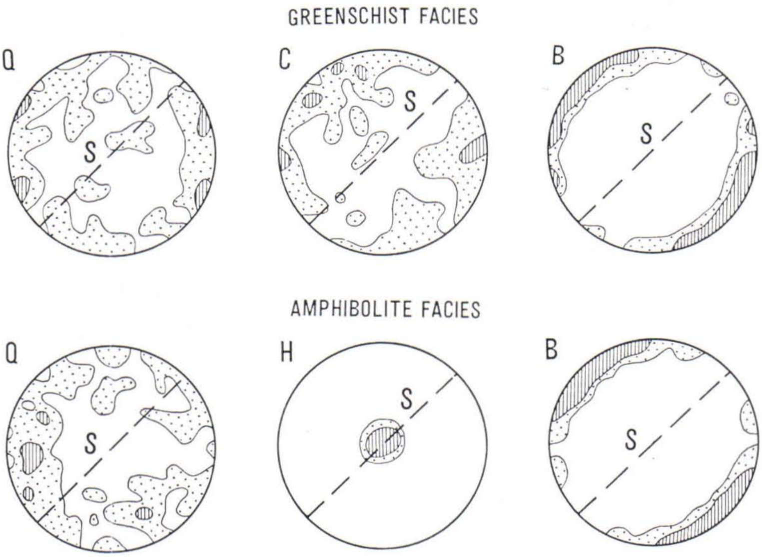

Fig. 7. Orientation of capital axes of quartz (Q), calcite (C), biotite (B) grains and axes »C» of hornblende (H) grains in schistose metavolcanic rocks in vertical cross section. Density 1—2—4—8 %.

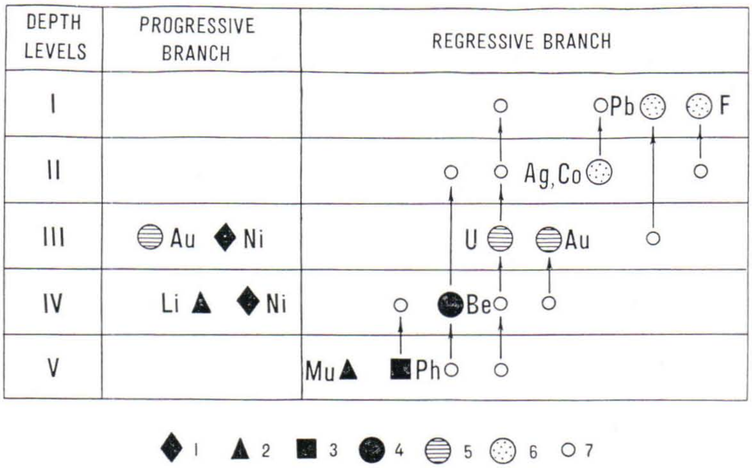

Fig. 8. Distribution of endogenous mineral deposits according to the depth levels of dislocation metamorphism: 1 = remobilized magmatic deposits; 2 = pegmatites; 3 = skarns; 4 = high-; 5 = middle-; 6 = low-temperature hydrothermal deposits; 7 = pre- and postmineralization structural elements (arrows indicate stages of ore-bearing fault developments), Mu = muscovite, Ph = phlogopite.

chenga complex as a series of thrust fault zones and thus support the idea suggested by H. Väyrynen (1938) many years ago.

During progressive dislocation metamorphism sulphide copper-nickel ores underwent cataclasis, recrystallization, redistribution and acquired schistose, corrugated and blastomylonitic textures. Rich epigenetic ores are confined to the basal parts of metaperidotite bodies and at the same time are controlled by shear zones after ultrabasic and sedimentary rocks (Fig. 6 b). In accordance with the general zonality in progressive metamorphism, mineral associations of ore-bearing shear zones correspond to greenschist facies.

As demonstrated by the author (1972), in rejuvenated faults of a crystalline basement different types of ore-bearing alkaline metasomatites are formed at different depth levels of dislocation metamorphism. High-temperature

Table 2.
Distribution of endogenous mineral deposits by the depth levels of dislocation metamorphism.

Depth levels	Character of deformations	Mineral assemblages of tectonites	Ore-bearing tectonic structures	Typical mineral deposits
I	Brecciation, fracturing	Crushed material of enclosing rocks	Zones of breccias, kakirites, tectonic fractures	Low-temperature hydrothermal deposits of lead, zinc, fluorite, gold, antimony etc.
II	Fracturing, cataclasis, grinding along slide planes	Hydromicas, clay minerals, carbonates, quartz, anthraxolite	Disjunctive displacements with gouges	Low- to middle-temperature hydrothermal deposits of silver, bismuth, cobalt, uranium etc.
III	Cataclasis, mylonitization, partial recrystallization	Epidote, chlorite, actinolite, albite, sericite, quartz, carbonates	Zones of cataclasites and mylonites	Middle-temperature hydrothermal deposits of gold, uraniferous sodic metasomatites, remobilized magmatic copper-nickel deposits
IV	Foliation and recrystallization	Biotite, hornblende, microcline, oligoclase, quartz	Zones of blastomylonites and cataclasites	High-temperature hydrothermal deposits of beryllium, tantalum, niobium in alkaline metasomatites, rare-element pegmatites, re-mobilized copper-nickel deposits
V	Foliation, recrystallization, partial melting	Biotite, hornblende, pyroxenes, microcline, plagioclase, quartz	Zones of injection migmatites	Phlogopite and magnetite deposits in magnesium skarns, muscovite, pegmatites

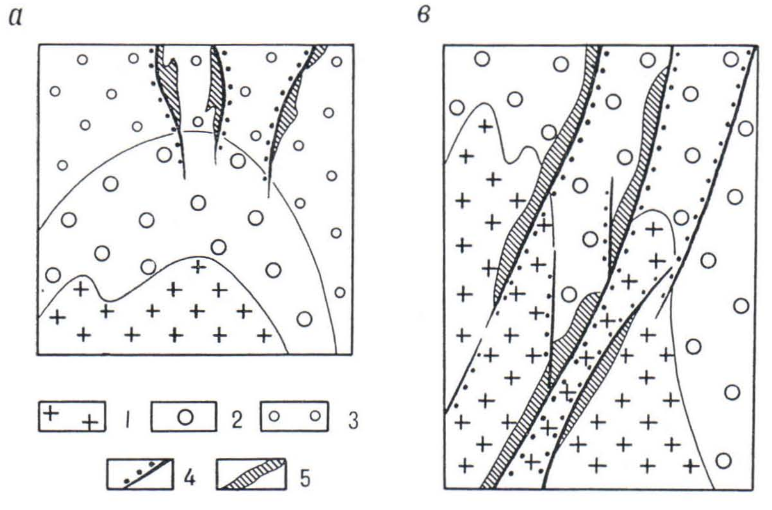

Fig. 9. Relationships between granitic intrusives and hydrothermal ore mineralization (in section): a = gold-bearing quartz veins; b = uraniferous sodic metasomatites; 1 = granites; 2 = metamorphic rocks of amphibolite; 3 = greenschist facies; 4 = epidote-chlorite mylonites and cataclasites; 5 = ore-bodies and hydrothermally altered rocks.

potassium metasomatites with rare element mineralization are associated in time and space with biotite-amphibole blastomylonites, middle-temperature uraniferous metasomatites are associated with epidote-chlorite cataclasites, and low-temperature auriferous quartz-orthoclase metasomatites are controlled by zones of brecciation and fracturing. Recent investigations testify to the facies correspondence between endogenous ore formation and dislocation metamorphism and show that both branches of dislocation metamorphism — regressive and progressive — are accompanied by definite types of mineral deposits (Fig. 8, Table 2).

In general, transitions from deeper to higher levels of dislocation metamorphism are accompanied by a decrease in the temperatures of ore deposition. The vertical range of endogenous ore formation in the Earth's crust depends on many factors and varies in different geological situations, including the vertical dimensions of depth levels. The most important of these factors are thermal aureoles, caused by active intrusive bodies and metamorphic domains. Vertical zonality is more contrasted if the geothermal gradient was high during ore formation but more continuous if it was low.

Comparison between some Precambrian hydrothermal deposits confirm this proposition (Fig. 9). According to R. Boyle (1961), the gold-bearing quartz veins of Yellowknife, which originated in the process of progressive metamorphism in an Archean volcano-sedimentary complex, are restricted to the narrow outer greenschist zone of the contact aureole of a granitoid pluton and reveal rapid changes in morphology and ore grade along strike and dip. In contrast, postmetamorphic sodic metasomatites crosscut and penetrate granitoid plutons along extensive zones of epidote-chlorite cataclasites; they are characterized by considerable vertical range and very clear vertical zonality in wall-rock alterations and ore deposition due to a slow decrease in temperature in an ascending column of hydrothermal solutions (Kazansky 1978).

References

Beck, L. S., 1970. Genesis of uranium in the Athabasca region and its significance in exploration. Canadian Min. Metall. Bull. 63 (695).

Boyle, R. W., 1961. The geology, geochemistry and origin of the gold deposits of the Yellowknife district. Geol. Surv. Canada, Memoir 310. 193 p.

Duk, G. G., 1977. Structural-metamorphic evolution of Petchenga complex. Leningrad, »Nauka». (In Russian).

Gee, R. D., Groves, D. I., Fletcher, C. I., 1976. Archean geology and mineral deposits of the Eastern Goldfields. Excursion guide No. 42A. 25th Intern. Geol. Congress, Canberra, 1976.

Gorbunov. G. I., 1968. Geology and genesis of sulphide copper-nickel deposits of Petchenga. Moskva, »Nedra» (In Russian) 352 p.

Kazansky, V. I., 1972. Dislocation metamorphism and endogenous ore formation in fault zones in crystalline basement. 24th Intern. Geol. Congress, Montreal, 1972. Section 4, Mineral deposits.

—, **(Editor), 1978.** Endogenous ore mineralization of ancient shields (evolution, structural and petrological conditions of ore formation). Moskva, »Nauka». (In Russian).

Kazansky, V. I., Krupennikov, V. A., Rozanov, Yu. A., 1971. Conditions of localization of Mesozoic auriferous metasomatites in crystalline base of Central Aldan region. Intern. Geol. Rev. 13 (4), 491—499.

Knight, C. L. (Editor), 1975. Economic geology of Australia and Papua New Guinea, 1975. I. Metals. The Australian Institute of mining and metallurgy, Monograph series, No. 5, 1126 p.

Kreyter, V. M., 1956. Structures of ore fields and mineral deposits. Moskva, Gosgeotekhizdat. (In Russian).

Krupička, I. & Sassano, G. P., 1972. Multiple deformation of crystalline rocks in the Tazin Group, Eldorado Fay Mine, NW Saskatchewan. Can. J. Earth Sci. 9 (4), 422—433.

Kuznetsov, V. A. & Distanov, E. G. (Editors), 1977. Problems of genesis of Sibirian stratiform lead-zinc deposits. Novosibirsk, »Nauka». (In Russian).

Ruchkin, G. V., Bushuev, V. P., Varlamov, V. A., Konkin. V. D., Kuznetsova, T. P., & Pirizhnyak, N. A., 1975. Kholodninskoe deposits as representative of Precambrian massive sulphide lead-zinc deposits. Geol. rudn. mestor. 17 (5), 3—17. (In Russian).

Turner, F. I. & Verhoogen, I., 1960. Igneous and metamorphic petrology. Second edition. New York, McGraw-Hill. 694 p.

Väyrynen, H., 1938. Petrologie des Nickelerzfeldes Kaulatunturi—Kammikivitunturi in Petsamo. Bull. Comm. géol. Finlande, 116. 198 p.

Zagorodny, V. G., Mirskaya, D. D. & Suslova, S. I., 1964. Geology of Petchenga volcano-sedimentary series. Moskva—Leningrad, »Nauka». (In Russian).

Fe-Ti-MINERALIZATIONS IN THE ÅNA-SIRA ANORTHOSITE, SOUTHERN NORWAY

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Introduction

This review repeats part of the findings obtained during a research program between 1970 and 1977. It was the purpose of this program to examine a representative example of the immediate and world-wide existing association of ilmenite deposits with plutonic complexes of the anorthosite-mangerite suite. The anorthosite massif of Åna-Sira in the south of Norway seemed to be particularly suited for this investigation because, not only are the largest deposits of titanium in Scandinavia situated here, but also a large number of minor deposits are to be found, which exhibit marked differences in composition, fabric, and spatial relation to their surrounding host rock.

Mining Fe-Ti ores in the southern Norwegian anorhosite province began in the 19th century, when about 90 000 t of ore were worked, mainly in the mines around Blåfjell, and shipped to England for iron production. From 1918 until 1964 the ilmenite ore body of the Storgangen (»Great Dike») was worked. From this ore deposit, which at times yielded 11—13 % of the world production of TiO_2, more than 10 mio t of crude ore with 17—19 % TiO_2 were won by underground mining. Diamond drilling up to 1960 proved evidence of a further 60 mio t of ore (Dybdahl 1960).

Since 1960 the mining changed gradually to open-pit mining of the ilmenite-norite body of Tellnes. Since then, besides sulfide and magnetite concentrates, between 600 000 and 1 000 000 t of ilmenite concentrate have been produced annually. The ilmenite concentrate is, for the most part, processed outside of Norway to TiO_2 and is used as a base for white pigment. According to exploration up to the present point the supplies of the Tellnes deposit hold about 250 mio t of ore with an average of 28—29 % ilmenite (Gierth and Krause 1973).

Regional geology

The andesine anorthosite body of the Åna-Sira massif, which has an area of about 200 km², plus a number of other anorthosite massifs, noritic intrusive bodies and extensive units of the mangerite suite, form the intrusive complex of Rogaland in southern Norway which covers about 1 200 km² (Fig. 1).

Studies on the units of the intrusive complex have been carried out by Barth (1933, 1945), Duchesne (e. g. 1972), Kolderup (e. g. 1896,

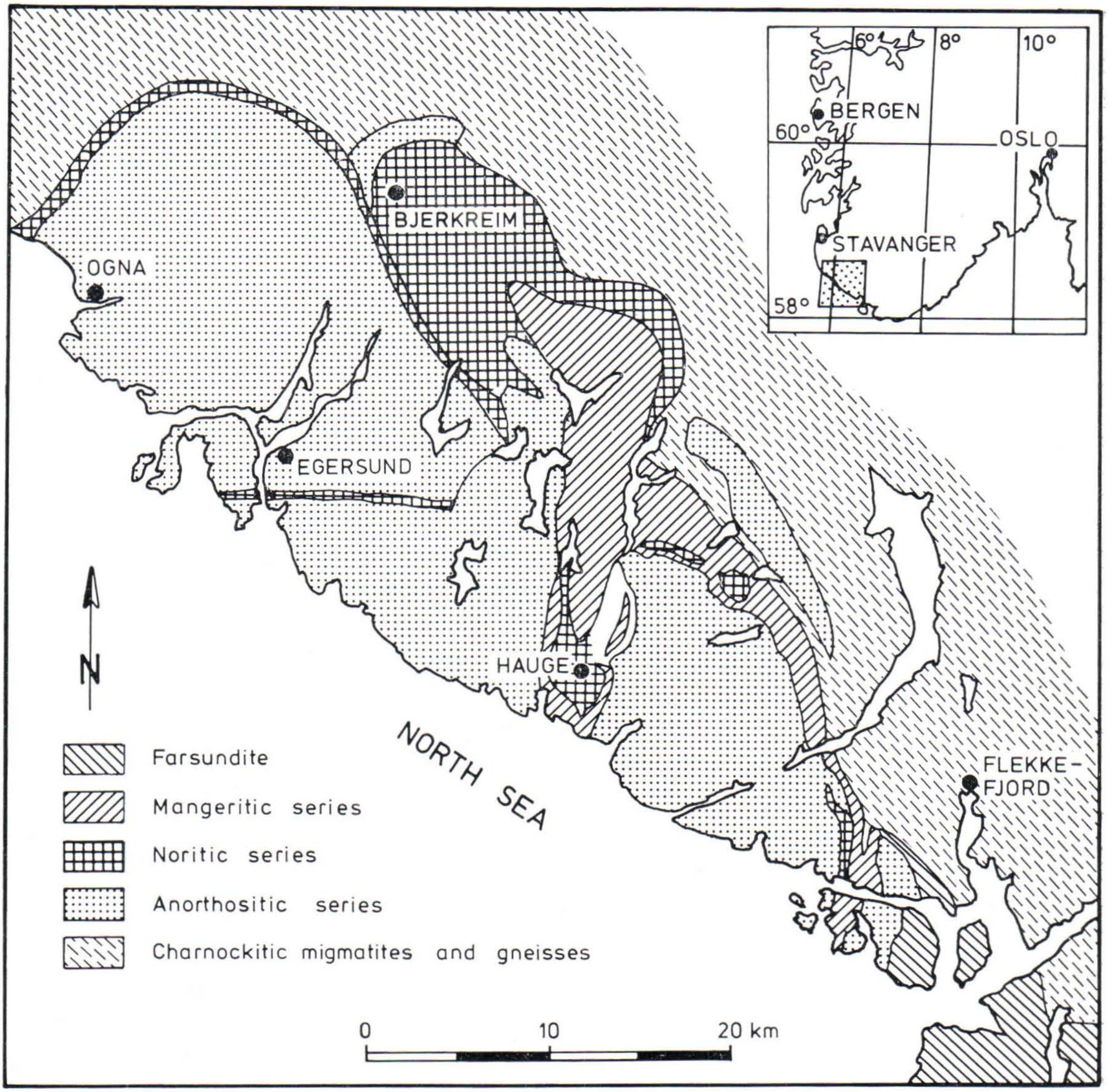

Fig. 1. Regional geological sketch map of the South Rogaland igneous complex (modified after J. and P. Michot 1969).

1914), Vogt (1910) and especially by P. Michot (1934—1970) and J. Michot (1952—1972). The latter summarize the total geological-petrological situation and develop models for the origin of the intrusive complex (e. g. P. Michot, 1960, 1969; J. and P. Michot, 1969; J. Michot, 1972). The units of the complex are assumed to be of partly magmatic, partly anatectic, and of partly metasomatic origin.

The intrusive complex of Rogaland is discordantly intercalated into charnockitic migmatites and gneisses of the Telemark-Rogaland region (Hermans *et al.* 1975). The isotope ages of the rocks occuring in this area are between 1 500 and 850 Ma. The second date marks the final consolidation of the latest units of the intrusive of Rogaland (Versteeve, 1974). Mangerites and the rocks of the Farsundite complex

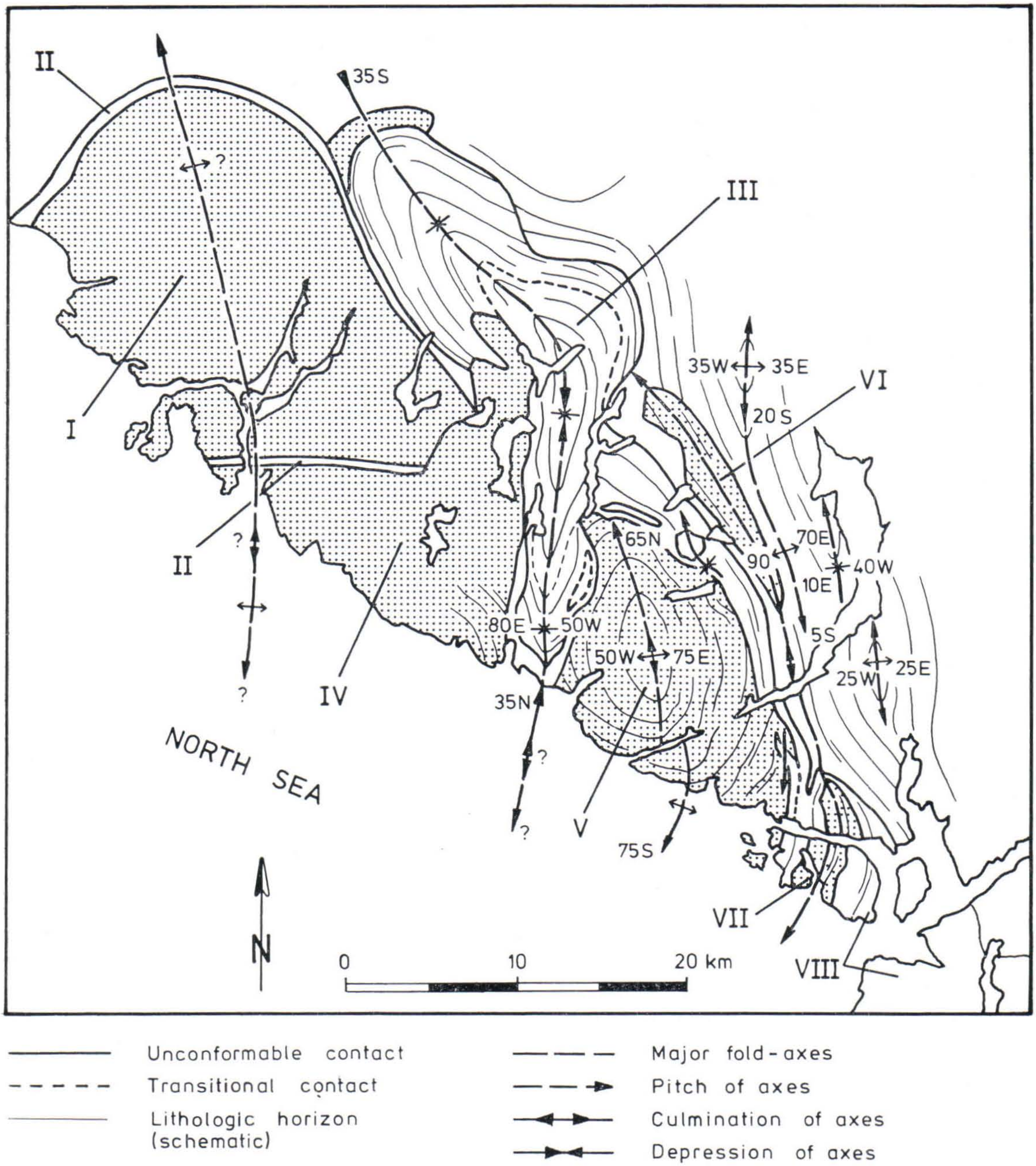

Fig. 2. Regional structural sketch map of the South Rogaland igneous complex and neighboring areas (modified after J. and P. Michot, 1969; Hermans *et al.* 1975).

I. Egersund-Ogna anorthosite; II. Noritic intrusions of Lakssevelefjeld-Koldal and Puntevold-Lien; III. Lopolith of Bjerkreim-Sokndal; IV. Håland-Hellaren anorthosite; V. Åna-Sira anorthosite; VI. Gasaknuten anorthositic outlier; VII. Hidra anorthositic outlier; VIII. Farsundite complex.

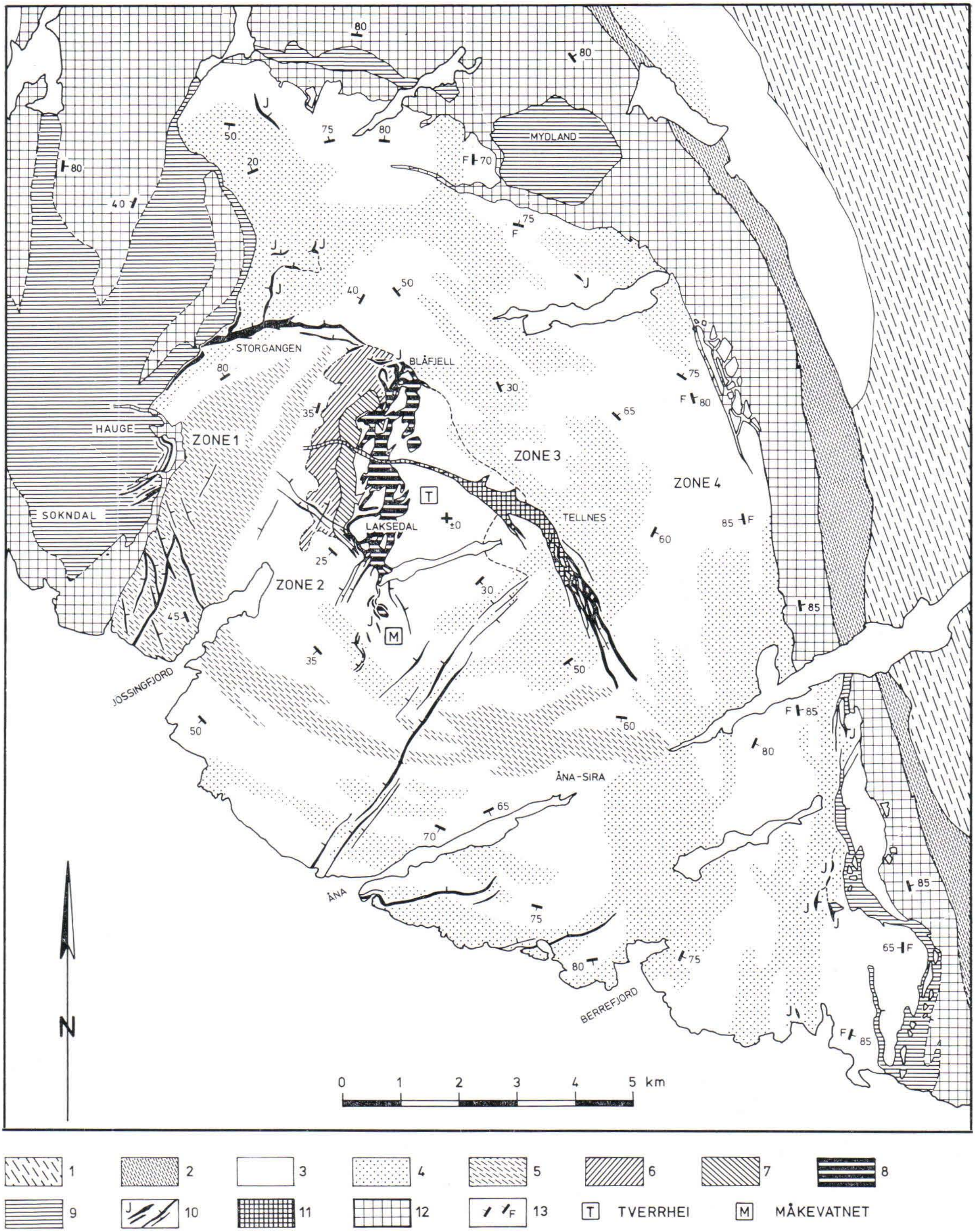

Fig. 3. Geological sketch map of the Åna-Sira anorthosite and related rocks (noritic, mangeritic and basalt dike systems partly omitted). 1 = Charnockitic migmatites and gneisses; 2 = Banded noritic-charnockitic gneisses; 3 = Anorthosite (0—10 vol % mafites); 4 = Leuconorite (10—25 % mafites); 5 = Rhythmic bedding of anorthosite and leuconorite; 6—8: Noritic internides; 6 = Layered intrusion of Böstölen, poorly stratified upper part; 7 = Layered intrusion of Böstölen, stratified lower part; 8 = Norite-pegmatite body of Blåfjell-Måkevatnet; 9 = Noritic externides (Lopolith of Bjerkreim-Sokndal); 10 = Norite, ilmenite-norite (J) and ilmenite (J) dikes; 11 = Ilmenite-norite body of Tellnes; 12 = Mangeritic externides (Lopolith of Bjerkreim-Sokndal); 13 = Igneous layering and secondary foliation (F).

form the end of the magmatic development within a Sveco-Norwegian orogenesis, which has to be considered a contemporary of the Grenville in North America (Broch 1964; Gerling *et al.* 1968, J. Michot and Pasteels 1968).

The most important tectonic element on a regional scale is a system of folding with non-linear N—S axes (Fig. 2). P. Michot (1960), within the intrusive complex, distinguishes two more phases of folding; in the gneiss area around Flekkefjord (Falkum 1966) up to five fold systems are differentiated. In a later article, Falkum (1972) points out that form and structure of the entire Norwegian southern tip might be determined by a complex megatectonic synclinorium with N—S axes plunging northward.

The massif of Åna-Sira is located in the southeast of the intrusive complex of Rogaland. It is surrounded by the intrusive units of the lopolith of Bjerkreim-Sokndal or its eastern equivalents. In the west an immediate contact to the noritic series exists (»Phase III a», P. Michot) as well as to the mangeritic rocks of the Eia-Rekefjord intrusion (»Phase III b», P. Michot). In the north and the east, the anorthosite massif is separated by intrusive mangeronoritic to quartz-mangeritic units from the Gasaknuten-outlier, the Hidra-outlier and the gneiss cover. To the south, the border is formed by the North Sea coast (Fig. 3).

All the rocks and ore bodies of the South Rogaland complex, in the area under investigation, can be classified as belonging to one magmato-tectonic development which, on the basis of geological and petrographical criteria, is to be subdivided into three sections. The sequence anorthositic, noritic, and mangeritic series marks the chronological order of a differentiation process. The noritic and mangeritic parts of the intrusion are further subdivided according to their spatial-structural relationships to the anorthositic series. Those parts of the noritic and mangeritic series, which are found outside of the Åna-Sira anorthosite and belong to the lopolith of Bjerkreim—Sokndal or its eastern equivalent are termed external units. The internal units are situated in a central position in the anorthositic series. Numerous dike systems of varying types form connecting links between noritic and mangeritic externides or internides respectively.

In all rock series concentrations of Fe-Ti-oxides can be found locally. However, concentrations of economical value are exclusively connected with the noritic and mangeritic series.

Anorthositic series

Anorthositic series is the term of the series of rock, poor in or free of mafitic components, which covers the major part of the area under survey and is only very little structured.

Mineralogy, modal analyses

The rocks of the anorthositic series consist predominantly or even entirely of plagioclase with 40—50 % An, and are frequently antiperthitic. The plagioclase holds about 3 % of nonexsolved Or-component (Zeino-Mahmalat and Krause 1976). Minor constituents are orthopyroxene (hypersthene, subordinated bronzite) and ilmenite. Ilmenite, as a rule, shows several generations of hematite exolution lamelae as well as spindle-shaped inclusions of spinel (»spinel-D», Gierth and Krause 1973). Accessories are apatite, biotite, K-feldspar, as well as sulfides (pyrrhotite, pyrite and secondary magnetite, chalcopyrite, linneite, covellite). Titanomagnetite within the anorthositic series is exclusively found in sporadically occuring mafitic intercalations. It exsolves ilmenite and spinel, but relics of ulvite are not present at all.

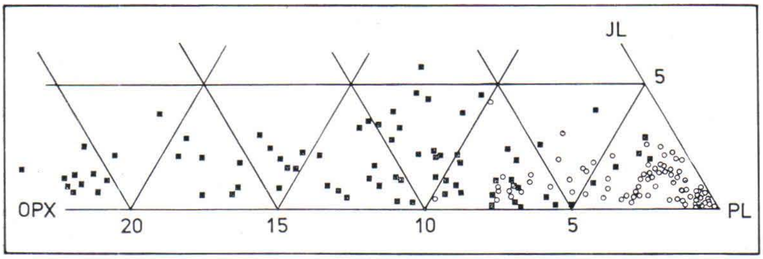

Fig. 4. Modal composition of 160 rock samples from the anorthositic series (in vol %). Circles: rocks, classified as anorthosite; squares: rocks, classified as leuconorites. OPX = Orthopyroxene; PL = Plagioclase; JL = Ilmenite.

Baddeleyite (ZrO_2) has not been observed. Modal analyses on about 160 rock samples have shown that in geological mapping rocks with less than 10 vol % mafites (range 0—14 vol %) were termed anorthosites while rocks with a mafite proportion up to 25 vol % (range 6—25 vol %) were classified as leuconorites (Fig. 4).

Noritic intercalations as well as pyroxenitic and ilmenitic layers (Ti-magnetite proportion of opaques 0—ca. 10 vol %) are quantitatively negligible. In the units of the anorthositic series, except for the minor noritic-mafitic intercalations, no rocks were observed where the ore content, consisting entirely of hemoilmenite, had surmounted 6 vol %. This fact is an important distinctive feature in comparison to corresponding rocks of the noritic series.

Numerous and various forms of inclusions are embedded in the rocks of the anorthositic series. Their spatial arrangement and distribution provide information for investigating orthomagmatic structures of the anorthositic series.

In mafite-poor rock types euhedral plagioclase crystals of labradoritic composition appear with lengths of up to 1,5 m. Sometimes these crystals display oscillating zonary banding at the edges and show, to some extent, distinctive labradorescence. Other crystals, aggregated in xenolith-like bodies show adcumulus growth (Wager *et al.* 1968). The appearance of these plagioclases indicates that they could represent earlier cumulates (Zeino-Mahmalat, 1972). Euhedral black orthopyroxene crystals (bronzite) are likewise considered as representing orthocumulus material. They can reach up to 2 m in length. A certain part of block-like inclusions within the anorthositic series is evidently of allochthone nature. Form, margins, and above all, marginal resorption of these xenoliths or autoliths by the wall rock demostrate this fact.

Zoning of the anorthositic series

Occuring within the anorthositic series are different zones of similar rock types, which have been mapped in detail. The demarkation of these zones was achieved by a comparatively time-consuming method of point and profile mapping. Between 500 and 4 000 rock classifications per km² were done. The resulting maps of rock varieties (scale 1:10000 and 1:25000) are unique for anorthosite massifs of the massive type (e. g. Pedall 1977, Krause *et al.* 1977).

Because of differences in the mafite proportion and mafite distribution in the anorthositic series and through the zonation of various inclusions and xenolithes, as well as through differences in textures and structures, the large dome-shaped anticlinorium of the Åna-Sira massif is represented as a concentric structure on the two-dimensional projection of the map (Fig. 3.)

A profile from west to east through the Åna-Sira anorthosite shows a distinct zonation and consists of the following units (cf. Fig. 3):

Zone 1: Rhytmic layering of anorthosite and leuconorite.

The western part of the profile is represented by a layered sequence of coarse-grained anorthosite and finegrained leuconorite. Here the orthomagmatic structures generally are well preserved. Between these anorthosite and leuconorite bodies, which in places are extensively interfingered as a result of plastotectonic deformation (»flow layering»), there are intercalated zones, being characterized by rhytmic igneous layering on a centimeter-meter scale. These zones of mesoscopic rhythmic layering show the characteristics of a key horizon. They can be followed along the entire west and south flank of the Åna-Sira dome. The entire sequence along the western part of the profile shows a roughly rhytmical megascopic layering of a type ABCD-ABCD. A corresponding type of rhythmical layering can also, in parts, be observed in the exposure. The layers are:

A: Almost entirely monomineralic anorthosite in the upper part of the sequence.
B: Rhythmic interlayering of anorthosite and leuconorite with irregular intensity of banding.
C: Leuconorite with faint parallel texture.
D: Leuconorite with subordinate conformable norite-, pyroxenite-, or ilmenitite layers close to the comparatively distinctly marked base of the sequence.

Zone 2: Anorthositic »core».

To the east, the zone of rhythmic bedding is followed by an anorthositic area, which is more or less monomineralic. On the map, this body appears as the »core» of the anorthositic series, concentrically enclosed by the other sequences (cf. Fig. 3). Between Blåfjell and Måkevatn this zone is inconformably intruded by the internides of the noritic series. On the other hand, numerous monomineralic anorthosite-xenoliths of all sizes are intercalated into these intrusive bodies.

Zone 3: Leuconoritic zone rich in inclusions.

Already with an eastern dip, a semi-circular zone, open to the west, of an overall leuconoritic appearance surrounds the central anorthosite body. It is characterized by a large number of various inclusions. Groups of angular leuconorite — and norite-xenoliths appear sporadically, often with well—developed igneous layering and gradual rhythmic transitions into plagioclase adcumulates, invariably directed to one side. Most common are swarms of black orthopyroxene crystals, pyroxenitic lenses and inclusions of coarse-grained plagioclase adcumulates. Most inclusions are arranged in trains, which may continue in the strike for up to 500 m. The resulting megascopic planar fabrics are comformable to the total arrangement of the entire leuconoritic zone and outline orthomagmatic structures.

Zone 4: Secondary foliated anorthositic area.

The eastern third of the Åna-Sira anorthosite is occupied by a petrographically extremely poor structured area. Leuconoritic lenses are of minor significance in the anorthositic zone.

Sequence of rock varieties, as well as arrangement of inclusions, forming primary fabrics, lie unconformably at an acute angle to the eastern contact of the Åna-Sira massif. The secondary foliation, however, runs approximately parallel to the contact. Extensive granulation of the rocks is obvious, especially with the mafites which show a pronounced parallel texture.

On the cryptic zoning of the anorthositic series

The detailed mapping of the anorthositic series was followed by systematic sampling. This was carried out along three profiles in the western, eastern, and central parts of the Åna-Sira anorthosite; the profiles being vertical to the respective primary structures. The An-contents of plagioclase were determined in about 150 samples in addition to the modal composition in order to be able to find a possible existing cryptic zoning or layering within the

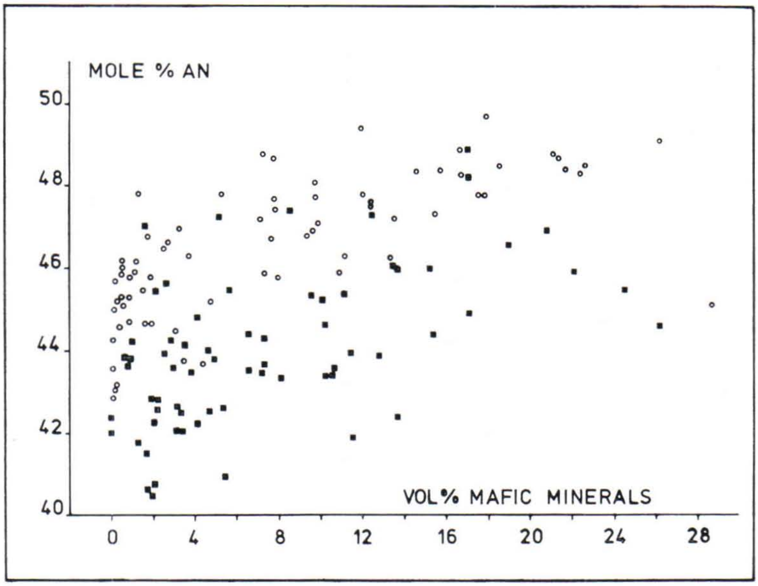

Fig. 5. Variation of mole % An in plagioclace with the amount of mafites in rocks from the anorthositic series. Circles: samples from central parts of the Åna-Sira anorthosite (between Hauge and Tellnes); squares: samples from marginal parts of the Åna-Sira anorthosite (west of Mydland and SE of Åna-Sira).

anorthosite massif. Per sample 6—12 single grains were measured by universal stage methods and the mean was calculated (Fig. 5). The investigations resulted in the following findings:

1. The An-contents are directly related to the mafite contents of the igneous rock varieties (Fig. 5). Since in these rocks opaques are only of inferior significance, a direct connection between the distribution of Fe-Mg-silicates and the An-contents is resulting.
2. Accordingly, the roughly isomodal, rhythmic sequence of the rock varieties within the anorthositic series is paralleled by a corresponding — in section of the massif — isochemic cryptic variation of the An-contents. The plagioclases of noritic-mafititic basis layers in the rhythmic bedded zone are labradorites (An 50—60 %), the transitions to the andesines of the overlying rock (An 40—50 %) are gradual; the change to the An-values in the rhythm directly underneath seems to be always abrupt. For the $FeSiO_3$-contents of the orthopyroxenes a corresponding development must be assumed; for the wall rock of the ore body of Storgangen this has been established by Krause and Pape (1975).
3. Plagioclases from the border area of the anorthosite massif generally display lower An-contents as such from the central areas (Fig. 5).

Primary structural elements

The concentric overall build of the Åna-Sira massif corresponds to a composition layering on a megascopic scale, as it is also described of several other andesine anorthosite massifs (e. g. Anderson, 1969). Like the rock zoning of the anorthositic series as a whole, also the layered sequence of anorthositic and leuconoritic rocks in the western part of the massif has to be considered as a product of igneous layering. The occurrence of anorthositic rock types in the upper parts of a rhythm, followed towards the basis of the rhythm by leuconorites and subordinate mafitites, gives evidence of this. The rhythmic layering on the whole, however, is developed isomodally. Neither towards the core of the anorthositic series, nor towards the margin, is there any dominance of leuconorites or anorthosites to be observed.

Besides the previously described megascopic-mesoscopic composition layering and the rhythmic, gradual layering to be observed in exposure, there appear a number of additional types of orthomagmatic layering:

Xenoliths arranged in series, often retaining their angular outlines with only minor sings of resorption, lie conformable to the surrounding rock varieties. The single blocks are commonly arranged parallel to the strike of the whole swarm; the surrounding rock shows no parallel texture. This is also the case in the plagioclase adcumulate trains and in the serially arranged orthopyroxene inclusions. Because of their chemism these cumulus-type insets can also be classified as orthomagmatic products (Zeino-Mahmalat and Krause 1976). Their spatial arrangement and distribution reprensent a primary planar fabric. Phase layering, that is, the sudden appearance or disappearance of a cumulus component (Hess 1960, Wager and Brown 1968) results from the varying proportion of orthopyroxene in the composition of the igneous rocks. Orthopyroxene here definitely belongs to the cumulus material. Often a parallel texture can be observed in the rocks, which is determined by the arrangement of the mafites, these running parallel to the rest of the surrounding planar fabrics. This structural element partly corresponds to igneous lamination, even though in places it can be superimposed upon by secondary fabric, like plastotectonic deformation or gneissification. Further primary structures are preserved in exposures, which correspond to magmatic equivalents of graded bedding, crossed bedding, angular unconformites, partition and thinning out of layers, as well as trough bedding and slumping. These structures often possess a striking similarity with those, which are described in typical layered intrusions, for instance, at Skaergaard (Wager and Brown 1968).

Structure

Location, form, and extension of the noritic and mangeritic intrusive bodies are, to a great degree, determined by the dome-shaped structure of the anorthositic series. On the other hand, the structures of these younger intrusions are superimposed upon the form elements of the anorthosite massif. Finally, the anorthositic, noritic, and mangeritic series are, according to their chronological position, deformed either plastically and /or by fracturing. Characteristic are convergence and superimposition of different fabric types, depending upon their spatial position within each series, and upon the composition of the rocks.

The orthomagmatic structures are to a certain extent obliterated by secondary deformation. With the exception of the eastern part of the Åna-Sira massif near the gneiss cover and a few smaller areas, conspicuous by their tectonic position, the secondary foliation, on the whole, superimposes itself conformable upon the primary structural elements.

The intensity of this gneissification varies. Wide areas, especially in the western and central parts of the Åna-Sira massif, seem to be secondarily foliated only to a small extent. In the mesoscopic range, the change from plastotectonic deformation to fracturing is mainly dependent on the rock type. For the most part, in anorthositic rocks, there can already be formed a wide set of joints, whereas leuconorites in interbedded layers are deformed plastically without fracturing. At the interface of the two rock varieties develops a small-scale crumpling, open at first, which with progressing deformation during the doming of the anorthosite body changes locally into shear folding. Last of all, the anorthosite layers in the course of this deformation process may be separated into single boudin bodies. This competent behaviour of anorthositic rocks in contrast to leuconoritic rocks during the syn-to postconsolidative deformation must be considered as the cause of the widespread migmatite-like structural phenomena on the map and in exposure.

For the purpose of obtaining a close view on details of the total structure, about 10 000 fabric data of primary and secondary planar fabrics were collected within the anorthositic series. The subsequent structural analysis yielded the following findings: The most important structural element of the Åna-Sira massif is a central N-S-running brachy-anticlinorium with a minor to obvious eastwardly inclined main axial plane. This antiform culminates in the area between Blåfjell and Måkevatn in a broad crested zone with open, and for the most part, symmetrical folding in the 10 to 100 m range and generally horizontal N—S axes. Progressing northwards as well as southwards, the main N—S axis shows an increasingly steeper pitch (northern part of the Åna-Sira massif: 170/65 NW; southern part: 170/75 SE).

The lopolith of Bjerkreim—Sokndal, adjacent to the Åna-Sira anorthosite in the west, represents a corresponding equally non-linear synform. In contrast to the complexly folded gneiss cover (Falkum 1966) it must be supposed that the intrusive complex of South Rogaland, after the regional N—S folding has gone through an autochthonous tectonic development. This is marked by vertical movements, leading to a central doming of the anorthositic series and accompanied by diapiric subintrusions of noritic and mangeritic material in the crest area of the thus reinforced brachy-anticlinorium. Specific elements of the noritic subintrusion, as in the example of a conjugated norite dike system on the southwestern flank of the Åna-Sira anorthosite, or part of the structures of the noritic internides, seem to have been additionally influenced by an E—W directed compression.

Initially the deformation of the anorthositic series was plastotectonic. Flow layering, as well as the occurrence of a regular minor folding (crumpling) on the western flank of the Åna-Sira massif, should be noted here. This folding is directed asymmetrically away from the crest of the Åna-Sira dome towards the core of the Bjerkreim-Sokndal lopolith and represents a kind of drag folding, resulting from the doming of the anorthosite massif and the accompanying depression of the lopolith. With progressing deformation of the anorthositic complex, fracture deformation gained increasingly in importance. The final stage of the upfolding, demonstrated by the opening of fissures and formation of the mangeritic internides, is entirely marked by fracturing. Resulting from the upfolding of the Åna-Sira massif, a second fold system was formed. The axes of these open and symmetrical folds are arranged radially around the crest area of the brachy-anticlinorium and the greater part of them plunge outward. On a regional scale, this folding $F_2 \wedge F_1$ is of subordinate importance. Connected with the doming of the Åna-Sira massif, the main tension concentrated in the crest area of the anticlinorium. Accordingly, the main fractures of the different internal subintrusions are grouped here. Corresponding to the temporal sequence of the various internides — layered intrusion of Böstölen (noritic series, inital stage) — norite-pegmatite body of Blåfjell-Måkevatn (noritic series, final stage) — ilmenite-norite body of Tellnes (mangeritic series, initial stage) there is a continuous development in degree of differentiation as well as a progradation from West to East in the position of these sub-intrusions, which might be due to the eastward inclination of the Åna-Sira antiform (cf. Fig. 3).

The units of the intrusive of South Rogaland, especially the anorthosite bodies reacted compe-

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tent upon the complex deformation within the surrounding gneisses. The intrusive complex shows distinct gneissification only in a marginal zone of differing width.

A regionally widespread basalt dike system from a structural point of view cannot be connected with the development of the South Rogaland igneous complex. The dikes follow a stringent non-conjugated system of WNW—ENE fissures and only show negligible dislocations.

Noritic Series

As described, the noritic series is subdivided into internides, »conjunctive links», and externides, according to their spatial relationship to the anorthosite massif of Åna-Sira. The noritic internides occupy a roughly oval-shaped area in the western and central crest area of the Åna-Sira anticlinorium (cf. Fig. 3). This area is about 2.3 km wide, and in N-S direction about 4 km long. The internides are distinguished by a great number of ore and rock types, which display well-developed magmatic features. Distinctive are numerous occurrences of oxidic Fe-Ti ores in layer-, streak-, or dike-shaped bodies.

Layered intrusion of Böstölen

The oldest subintrusion of the noritic internides is the layered intrusion of Böstölen. It also represents the most basic unit of the noritic series. Intrusion and differentiation of the probably, on the whole, noritic material, took place on the western rim of the crest of the Åna-Sira anti-form. The Böstölen intrusion consists of lower part with well developed igneous layering and a relatively poorly structured leuconoritic to anorthositic upper part (cf. Fig. 3). The lower part consists of a rhythmic sequence of intercalated ore and rock varieties (»Böstölen rhythmites»). Into leuconorite and norite, rich in opaques, are intercalated layers of titanomagnetite-ilmenite ores and pyroxenites. Additionally anorthosite and leuconorite layers as well as norite-pegmatitic lenses are found (Table 1).

The Böstölen rhythmites in a North-South extended semi-oval enclose the chief part of the remaining elements of the noritic internides. The dip of the layering is between 30° and 50° directed uniformly from the core outward. Upon the Tverrheia towards the East and against the dip, higher (marginal) parts of the layered intrusion are preserved together with their cover anorthositic rock. The dip here is shallow to horizontal and turns east because of slight folding. On the flat crest of the semi-oval, the thickness of Böstölen rhythmites reaches up to 180 m. After slow lateral decrease in thickness, this layered part of the intrusion wedges out to both sides. In the strike, the upper leuconoritic and pyroxenitic layered sequences within a rhythm always end in front of the basis layers richer in titanomagnetite. Numerous xenoliths, up to several hundred meters in size, are intercalated into the Böstölen intrusion. The purely anorthositic insets have irregular and angular forms and show always sharp contacts against the surrounding noritic rocks. At the contact, the Böstölen rhythmites break off or bend and follow the contours of the xenoliths. Because of composition, fabrics, An-values, and spatial relations, it is probable that the insets originate in the anorthositic »core» of the Åna-Sira massif, into which all noritic internides have been intruded inconformably.

Build and modal composition of the layered intrusion of Böstölen is shown in Table 1. Besides the main constituents orthopyroxene, plagioclase, titanomagnetite and ilmenite appear, like in all the other units of the noritic series, subordinate clinopyroxene and biotite as well as apatite, spinel, sulfide, and graphite as accessories.

The primary sulfide paragenesis contains pyrrhotite, pentlandite, and chalcopyrite in intergrowth typical of basic intrusives.

The secondary paragenesis consists of pyrite, linneite, millerite, and covellite. Accessory baddeleyite is present in almost every sample. In some ore-rich layers olivine, in distinct phase layering, occurs as a minor constituent (e. g. Table 1, samples 35—37).

The magnetite in all samples from the investigated profile is distinguished by fine tabular ilmenite lamellae and spinel exsolution in two generations. Ulvite exsolution in magnetite orthocumulus material can only be observed in the lowest part of the profile (up to sample 17). In the upper part, however, there are found tiny isolated magnetite insets within the silicate, which show the typical cloth fabric of ulvite exsolving magnetite. These grains appear in upward direction up to sample 52. Reaction rims between magnetite and ilmenite (Krause and Pape 1975) are found in almost every sample. With the serpentinization of olivine, secondary magnetite is formed. Ilmenite is always accompanied by spinel exsolution. In the lower part of the profile up to sample 52, ilmenite shows no sign of hematite exsolution. This appears increasingly from sample 52 on upwards. Graphite is totally absent above sample 52, whereas it always is present underneath as an accessory.

The lowest part of the Böstölen rhytmites is marked by series of ore layers rich in titanomagnetite, with grain sizes of about 1 mm in diameter. Invariably three to six individual layers with thickness of up to 60 cm are arranged to rhythms in centimeter-decimeter distance, the thickest ore layer almost in every case occupying the basis of rhythm. The total sum of this series of layers lies between 5 and 15; further upwards, the thickness of the single bed as well the thickness of the series decrease. Within the stratified part of the intrusion, the total ore contents also decreases from the basis upwards; in the middle third of the whole series a distinct dominance of orthopyroxene exists. In the uppermost part plagioclase to a growing extent takes part in the composition of the ore layers. A corresponding development can also be observed in the strike in direction of the lateral endings.

The transition from the Böstölen rhythmites proper to the leuconoritic-anorthositic part of the subintrusion is indistinct. The noritic-leuconoritic matrix of the layered part interfingers in the strike with the rock varieties poor in mafites.

As with the rocks of the anorthositic series, also with ore and rocks of the layered intrusion of Böstölen there exists an obvious dependency of the An-contents of the plagioclases on the mafite proportion. The An-values of plagioclase from anorthositic to noritic rocks (0—ca. 40 % mafite contents) lie between 44 and 55 %, the An-values of the rock types rich in mafite lie between 49 and 62 %. In Fig. 6 the relationship between the An-contents of the plagioclase and the mafite contents of the specific samples is illustrated. In connection with the composition layering of the Böstölen intrusion results a distinct rhythmic cryptic layering of the plagioclase. This is accompanied by a systematic variation of the magnetite proportion in the ore (Table 1). Like the An-contents, the titanomagnetite proportion of the ore gradually decreases from the basis towards the upper parts of the Böstölen rhythmites. Ulvite and graphite contents, frequency and size of the hematite exsolution lamellae in the ilmenite, as well as occurrence of antiperthitic exsolutions in the plagioclase, are also in conformity with this rhythmic development of differentiation, directed entirely upwards. Not yet finished geochemical investigations on opaques already allow the identification of relationships, which accompany this development described above and which are in line with findings from other layered intrusions (e. g. Bushveld, Skaergaard).

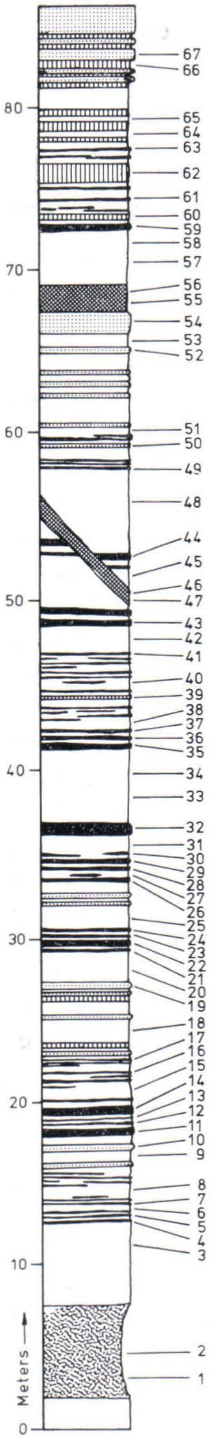

Table 1. Variations in the modal proportions of minerals and mole % An and Mt 100/(Mt+JL) in the lower part of the layered intrusion of Böstölen, west of Laksedalsvatnet.

Sample Nr.	Rock type	Jl	Mt	Sf	Sp	Opx	Cpx	Bi	Ol	Pl	%An	Mt·100 / Jl + Mt	Rem
UPPER END OF LAYERED SERIES, RHYTHM 3													
67	A	-	-	-	-	1.8	-	-	-	98.2	43.8	-	
66	LF	4.9	0.1	-	-	9.9	1.8	-	-	83.3	44.7	2	
65	L	5.1	0.1	-	-	15.6	0.7	-	-	78.5	45.0	2	
64	L	10.8	0.4	-	-	11.6	0.8	-	-	76.4	46.2	4	
63	E	20.6	0.8	-	-	30.2	2.1	-	-	46.3	48.7	4	
62	LF	5.1	0.2	-	-	13.6	1.2	-	-	79.9	44.1	4	
61	E	10.9	5.4	-	0.6	21.6	-	-	-	61.5	52.0	33	
60	A	-	-	-	-	-	-	-	-	100.0	47.8	-	
59	E	19.8	5.6	-	1.0	15.4	-	0.9	-	57.3	55.2	22	
58	A	3.1	0.7	-	-	0.8	-	-	-	95.4	46.8	18	
57	L	7.9	2.1	-	-	6.8	0.8	-	-	82.4	46.8	21	
56	NP	9.9	1.1	-	-	26.4	1.1	-	-	61.5	50.3	10	
55	NP	5.8	2.9	-	-	-	-	-	-	90.5	51.5	33	acc.
54	A	1.8	-	-	-	-	0.8	-	-	97.4	48.0	0	
53	L	15.4	1.8	-	-	1.0	1.0	-	-	80.8	48.0	11	
52	A	-	-	-	-	-	-	-	-	100.0	44.2	-	
51	L	8.4	1.1	-	-	14.8	0.8	-	-	74.9	47.8	12	
50	A	-	-	-	-	-	-	-	-	100.0	43.6	-	
49	E	18.0	0.4	-	-	36.7	0.9	-	-	44.0	49.1	2	
48	L	5.9	0.8	-	-	17.4	1.2	-	-	74.7	48.5	12	
NP - DIKE													
47	NP	18.2	1.1	-	-	4.6	-	-	-	70.4	37.1	6	1)
46	NP	3.0	0.5	0.9	-	7.4	6.1	-	-	80.5	39.5	14	2)
CONTINUATION OF LAYERED SERIES, RHYTHM 3													
45	L	9.2	3.9	-	-	19.6	1.2	-	-	66.1	49.7	30	
44	E	12.8	11.1	-	-	70.2	2.3	-	-	3.6	54.3	46	
43	E	15.8	28.8	-	1.0	46.4	4.9	-	-	3.1	55.3	65	
42	A	4.4	0.4	-	-	2.9	-	-	-	92.3	48.2	8	
41	L	5.1	3.1	-	1.1	35.8	2.0	-	-	52.9	50.2	38	
40	L	4.1	3.9	-	-	18.0	1.1	-	-	72.9	49.0	49	
39	A	-	-	-	-	-	-	-	-	100.0	49.0	-	
38	L	3.1	3.1	-	-	9.8	1.0	-	-	83.0	50.6	50	
37	E	12.2	24.1	-	0.9	45.8	2.1	-	14.9	-	-	66	
36	E	13.1	35.8	1.1	1.9	36.2	2.8	-	9.0	0.1	57.5	73	
35	E	18.2	28.9	1.0	1.0	37,0	3.0	-	0.9	10.1	55.7	61	
34	L	5.6	6.5	1.0	0.9	18,4	0.9	1.0	-	65.7	51.5	42	
33	L	7.7	4.4	1.1	-	13.0	0.9	1.0	-	71.9	50.9	36	
32	E	12.2	27.1	-	2.1	36.7	2.9	1.1	-	17.9	54.2	69	
31	A	n.d.	n.d.	n.d.	-	10.1	-	-	-	83.7	50.2	n.d.	3)
30	E	18.2	16.1	2.0	1.0	60.1	2.0	-	-	0.6	55.5	47	
29	E	18.1	18.9	1.1	2.1	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	51	4)
28	E	27.1	23.0	1.1	2.1	34.8	1.9	1.1	-	8.9	n.d.	46	
27	L	14.1	7.2	1.1	1.0	38.6	2.1	1.9	-	34.0	57.3	34	
26	E	24.1	31.8	1.0	2.1	26.8	1.1	-	-	13.1	55.3	57	
25	L	5.6	0.5	-	-	7.9	2.0	-	-	84.0	54.8	8	

Table 1, continuation

Sample Nr.	Rock type	Jl	Mt	Sf	Sp	Opx	Cpx	Bi	Ol	Pl	%An	$\frac{Mt \cdot 100}{Jl + Mt}$	Rem
24	E	18.8	31.1	-	1.9	45.6	2.6	-	-	-	-	62	
23	L	n.d.	n.d.	n.d.	1.1	23.9	0.9	-	-	59.1	58.3	n.d.	5)
22	E	18.2	24.1	-	2.0	52.3	1.0	0.8	-	1.6	n.d.	57	
21	E	19.9	26.8	0.9	3.8	38.7	1.8	1.2	-	6.0	61.5	57	6)
BASIS OF RHYTHM 3, LAYERED SERIES													
20	L	8.3	0.7	0.8	-	20.4	1.1	-	-	68.7	47.3	8	
19	A	1.2	-	-	-	2.8	3.9	-	-	92.1	45.3	0	
18	L	5.6	0.7	-	-	20.7	1.4	-	-	71.3	46.4	11	acc.
17	E	26.1	5.2	-	0.7	55.8	2.2	-	-	10.0	52.5	17	
16	E	11.8	6.9	-	1.0	37.1	1.2	-	-	42.0	50.8	37	
15	L	18.7	12.2	-	-	6.8	-	1.1	-	61.2	51.9	40	
14	E	29.6	25.8	-	1.9	34.7	0.9	0.8	-	6.3	52.0	47	
13	E	26.6	25.0	1.0	2.1	39.8	-	-	-	5.5	56.2	48	
12	L	7.5	5.0	0.8	-	15.2	-	1.0	-	70.9	53.4	41	
11	E	22.2	21.8	1.9	-	n.d.	n.d.	0.9	n.d.	n.d.	55.7	50	7)
BASIS OF RHYTHM 2, LAYERED SERIES													
10	A	-	-	-	-	-	-	-	-	100.0	44.1	-	
9	L	13.1	4.1	-	-	17.9	0.8	-	-	64.1	44.7	24	
8	A	3.1	-	-	-	0.9	-	1.1	-	94.9	44.9	0	
7	E	33.1	11.8	-	1.1	22.8	-	-	-	31.2	48.6	26	
6	L	n.d.	n.d.	n.d.	-	20.0	-	1.0	-	65.9	49.8	n.d.	8)
5	E	29.2	19.8	0.8	2.1	22.8	-	1.2	-	24.1	50.8	40	
4	E	12.4	5.9	-	-	24.1	0.7	0.8	-	56.1	50.6	32	
3	L	5.8	2.0	-	-	6.8	1.1	1.0	-	82.4	51.2	33	acc.
BASIS OF RHYTHM 1 AND BASIS OF LAYERED SERIES													
2	E	28.2	19.0	1.4	1.5	9.9	0.6	0.8	11.2	27.4	54.8	40	
1	E	35.4	29.8	-	1.8	27.0	0.6	-	5.4	-	-	46	
FOOTWALL OF LAKSEDALSGANGEN													

Rock types: 1 = Leuconorite, norite (L); 2 = Anorthosite (A); 3 = Ore (E); 4 = Fine grained leuconorite (LF); 5 = Norite-pegmatite (NP); 6 = Laksedalsgangen.
Modal analyses (in vol %): JL = Ilmenite; Mt = Magnetite; Sf = Sulfide; Sp = Spinel; Opx = Orthopyroxene; Cpx = Clinopyroxene; Bi = Biotite; Ol = Olivine; Pl = Plagioclase.
Remarks (Rem): 1. 5.7 % apatite; plagioclase anti- to mesoperthitic; 2. 1.5 % apatite; plagioclase anti- to mesoperthitic; 3. 6.2 % opaques; 4. 59.8 % gangue; 5. 15.0 % opaques; 6. 0.9 % apatite; 7. 53.2 % gangue; 8. 13.2 % opaques; acc. accessories

As an example, selected trace elements in magnetite and ilmenite from ore layers of the lowest part of the Böstölen rhythmites are represented in Table 2.

Like the noritic externides and the conjunctive link of the Storgangen system — described below — the subintrusion of Böstölen shows all the essential qualities of a layered intrusion. On the other hand the close spatial relationship, the high percentage of leuconoritic beds in the series, as well as the gradual transition of the stratified part of the Böstölen intrusion into a mafite-poor roof zone, establish the immediate relationship to massive andesine anorthosites. A distinction between massive-type labradorite and andesine anorthosites, as made by Anderson and Morin (1964; 1969) does not prove to be the fact, at least for the Åna-Sira massif.

Norite-pegmatite body of Blåfjell-Måkevatn

The second subintrusion of the noritic internides is the norite-pegmatite body of Blåfjell—Måkevatn. The intrusion of this body and the

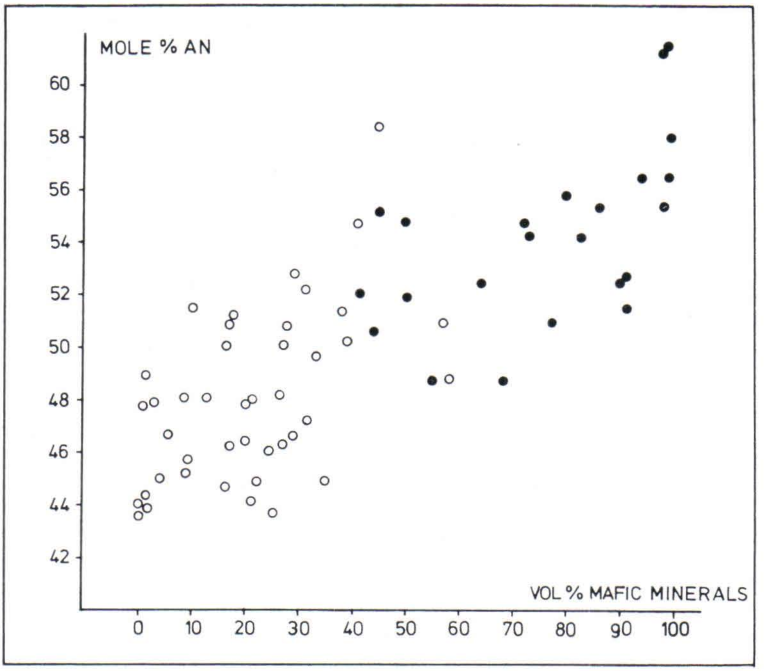

Fig. 6. Variation of mole % An in plagioclase with the amount of mafites in rocks and ores from the lower part of the layered intrusion of Böstölen, west of Laksedalsvatnet.
Open circles: anorthosite, leuconorite, norite, norite-pegmatite; dots: ore.

Table 2.
Geochemistry of magnetite and ilmenite on ore layers from the lower part of the layered intrusion of Böstölen, west of Laksedalsvatnet (location of samples cf. Table 1). Analyses done by AAS; Pedall and Swensen, 1974.

	MAGNETITE				ILMENITE		
Sample Nr.	Ti (%)	V (ppm)	Cr (ppm)	Mn (ppm)	V (ppm)	Cr (ppm)	Mn (ppm)
LAKSEDALSGANGEN							
1	5.86	5600	584	743	252	16	3180
2	4.87	5280	551	558	101	20	3460
LAYERED INTRUSION OF BÖSTÖLEN							
4	1.58	6130	1420	129	856	88	2690
7	1.04	6120	524	104	847	32	2600
11	1.47	5000	92	144	552	8	3370
21	5.06	4050	93	647	203	8	3370
24	3.15	4500	1290	315	971	152	2170
26	4.89	4900	61	680	303	0	3600
32	6.40	4190	91	847	151	0	3610
35	5.93	3970	152	818	150	8	3510
37	5.36	4010	46	834	203	8	3490
43	6.51	4090	106	958	202	12	3810

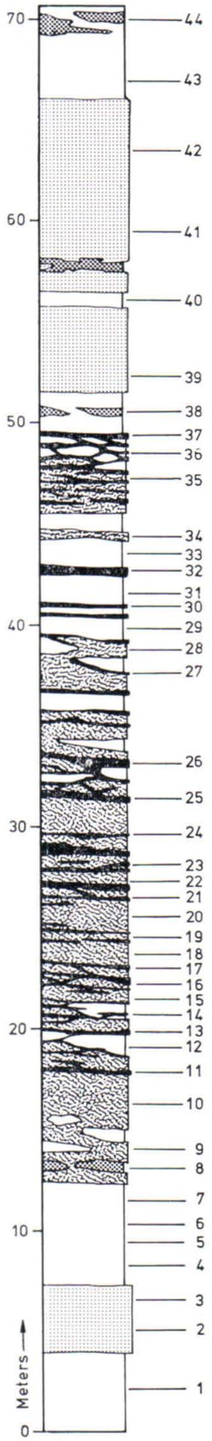

Table 3.
Variations in the modal proportions of minerals and mole % An and Mt 100/(Mt + Jl) in the Laksedalsgangen and neighboring rocks, gallery at Laksedalsvatnet.

Sample Nr.	Jl	Mt	Sf	Sp	Opx	Cpx	Bi	Pl	% An	$\frac{Mt \cdot 100}{Jl+Mt}$
44	5.1	0.3	0.3	-	18.0	-	-	76.3	46.7	6
43	9.0	0.1	-	-	25.9	0.8	-	64.2	51.4	1
42	5.9	-	-	-	0.9	-	-	93.2	49.8	0
41	2.0	-	-	-	7.7	1.0	-	89.3	49.0	0
40	5.9	0.1	-	-	14.4	0.9	-	78.8	48.5	2
39	0.9	-	-	-	10.0	1.1	-	88.0	48.8	0
38	28.6	-	0.4	-	20.8	-	-	50.2	50.8	0
HANGING WALL OF LAKSEDALSGANGEN										
37	51.0	7.4	0.9	2.1	10.8	0.4	1.1	26.3	55.3	13
36	59.2	9.7	0.7	1.7	10.6	-	-	18.1	53.5	14
35	5.0	0.6	-	-	8.4	0.7	-	85.3	50.7	11
34	6.8	1.7	0.8	-	-	-	0.6	90.1	50.3	20
33	2.3	0.4	-	-	6.8	2.1	-	88.4	52.3	15
32	39.8	9.9	0.7	0.6	43.2	-	-	5.8	n.d.	20
31	3.9	-	-	-	-	-	-	96.1	47.4	0
30	86.5	7.4	0.9	1.8	-	-	-	3.4	n.d.	8
29	9.0	0.1	-	-	0.8	-	-	90.1	50.0	1
28	27.7	7.1	2.1	4.6	-	-	0.8	57.7	53.7	20
27	54.7	15.6	1.0	3.9	16.0	0.8	-	8.0	54.8	22
26	51.4	5.9	-	1.7	11.8	-	1.1	28.1	n.d.	10
25	76.6	7.9	1.1	1.9	-	-	-	12.5	53.5	9
24	61.8	16.7	0.9	1.0	4.9	-	-	14.7	53.5	21
23	53.9	9.1	-	0.8	31.8	2.2	-	2.2	54.5	14
22	9.9	2.7	1.1	-	12.9	1.0	0.9	71.5	51.4	21
21	51.1	8.8	0.7	0.8	29.2	1.0	-	8.4	53.1	15
20	11.4	3.6	1.1	-	15.7	0.7	1.1	66.4	51.0	24
19	21.7	5.9	2.0	0.6	8.8	-	2.1	58.9	51.6	21
18	3.0	0.1	7.8	-	13.6	-	2.0	73.5	49.8	3
17	71.2	3.5	-	1.1	4.8	0.9	1.1	17.4	53.2	5
16	86.2	5.1	-	0.8	4.6	-	-	3.3	58.8	6
15	44.9	12.7	1.9	1.0	22.8	-	-	16.7	58.3	22
14	65.4	12.8	0.8	3.6	17.2	-	-	0.2	57.8	2
13	65.9	2.8	0.7	1.1	4.8	-	1.1	23.6	53.8	4
12	10.1	-	1.7	-	1.0	-	-	87.2	51.0	0
11	83.8	5.1	-	1.9	6.7	-	-	2.5	56.5	6
10	8.2	1.6	-	-	7.6	-	-	82.4	51.2	16
9	11.8	-	-	-	11.6	-	-	76.6	49.4	0
8	13.1	2.1	-	-	39.6	-	-	45.2	52.6	14
FOOTWALL OF LAKSEDALSGANGEN										
7	2.2	-	-	-	9.9	1.0	-	86.9	51.7	0
6	4.2	-	-	-	8.6	0.8	-	86.4	48.2	0
5	1.3	-	-	-	25.7	1.1	-	71.9	49.3	0
4	2.2	-	-	-	10.9	0.8	-	86.1	46.5	0
3	4.1	-	-	-	1.0	-	-	94.9	46.7	0
2	2.0	-	-	-	2.9	-	-	95.1	47.5	0
1	-	-	-	-	11.7	-	-	88.3	44.3	-

Rock types: 1 = Leuconorite, (norite); 2 = Anorthosite; 3 = Ilmenite norite, ilmenite; 4 = Norite-pegmatite; 5 = Ilmenitic norite, (norite).
Modal analyses: Jl = Ilmenite; Mt = Magnetite; Sf = Sulfide; Sp = Spinel; Opx = Orthopyroxene; Cpx = Clinopyroxene; Bi = Biotite; Pl = Plagioclase.

connected massive ore veins is situated in the immediate crest area of the brachy-anticlinorium east of the Böstölen rhythmites, which are unconformably cut by norite-pegmatite dikes and ore veins. The intrusive body consists of a number of massive norite-pegmatite stocks arranged in N—S direction like beads on a string. The stocks are connected by dike swarms and breccia zones with a norite-pegmatite matrix. The massive norite-pegmatites are, in each case, linked to inferiour culminations of the crest area proper. Preferably, along the contacts between norite-pegmatite and wall rock, massive accumulations of Fe-Ti-oxides appear, which often assume an independent vein-like character. The largest two Fe-Ti-deposits of this type are Laksedal to the south of the main body and Blåfjell to the north. Both mineralisations are connected to the lateral endings of the norite-pegmatite bodies. Within this second subintrusion no continuous differentiation development from the bottom to the top could be observed. Characteristic for wide areas seems to be an irregular schlieric to vein-like side-by side of norite-pegmatite, norite, ilmenite-norite, and ilmenite. Massive ore lenses reach up to approximately one meter in thickness and can be followed for several hundreds of meters in the strike. Laterally, they often change into ilmenite-noritic material.

The proportion of titanomagnetite in the ore is perceptibly lower than in the layered intrusion of Böstölen (Table 3) and varies between zero and maximal 25 %. In most samples, magnetite shows ilmenite exsolution, as well as spinel exsolution in two generations, but in other samples they can be lacking. The otherwise characteristic reaction rims between magnetite and ilmenite are only in place present. Ilmenite always shows hematite exsolution and spinel lamellae. Locally, ilmenite is changed into rutile. Especially characteristic for these mineralizations is a relatively high sulfide proportion, which may take up several percent of the modal composition. Once more pyrrhotite, pentlandite and chalcopyrite occur as primary sulfides, out of which in the course of deuteric readjustment are derived pyrite, bravoite, linneite, millerite and covellite. Baddeleyite is, intergrown with ilmenite, omnipresent as an accessory. In places, up to 5 % of the modal composition consists of green Cr-containing spinel.

The An-contents of the plagioclases with 45 to 55 % lie in the range of those from the older subintrusion of Böstölen, but perceptively above the values in the anorthositic series. Antiperthitic exsolutions in plagioclases are widespread in contrast to the Böstölen rhythmites. Characteristic of the ore is, above all, a subordinate orthopyroxene content. At lateral endings of norite-pegmatite dikes and in apophyses of the major bodies, differentiates with high proportions of quartz, mesoperthite, K-feldspar, muscovite, apatite, and magnetite occur. The younger subintrusion of the norite-pegmatite body of Blåfjell-Måkevatn seems to represent a more advanced stage of differentiation than the layered intrusion of Böstölen.

The system of Storgangen

One of the conjunctive links between noritic internides and externides is formed by the dike system of Storgangen, the other is represented by a conjugated norite dike system along the western and southern flank of the Åna-Sira antiform. The ore body of Storgangen, together with its lateral endings and numerous offshoots forms an unconformable intrusive dike system within the anorthositic series. It extends over approximately 4 km in length in a wide arch from the contact between Åna-Sira-massif and noritic externides of the lopolith of Bjerkreim-Sokndal in the west immediately up to the upper margin of the leuconoritic-anorthositic sequence, belonging to the subintrusion of Böstölen (Fig. 7). The dip of the ore body and its offshoots is directed from NW to NE at 35 to 55°. Its thickness varies between several meters near

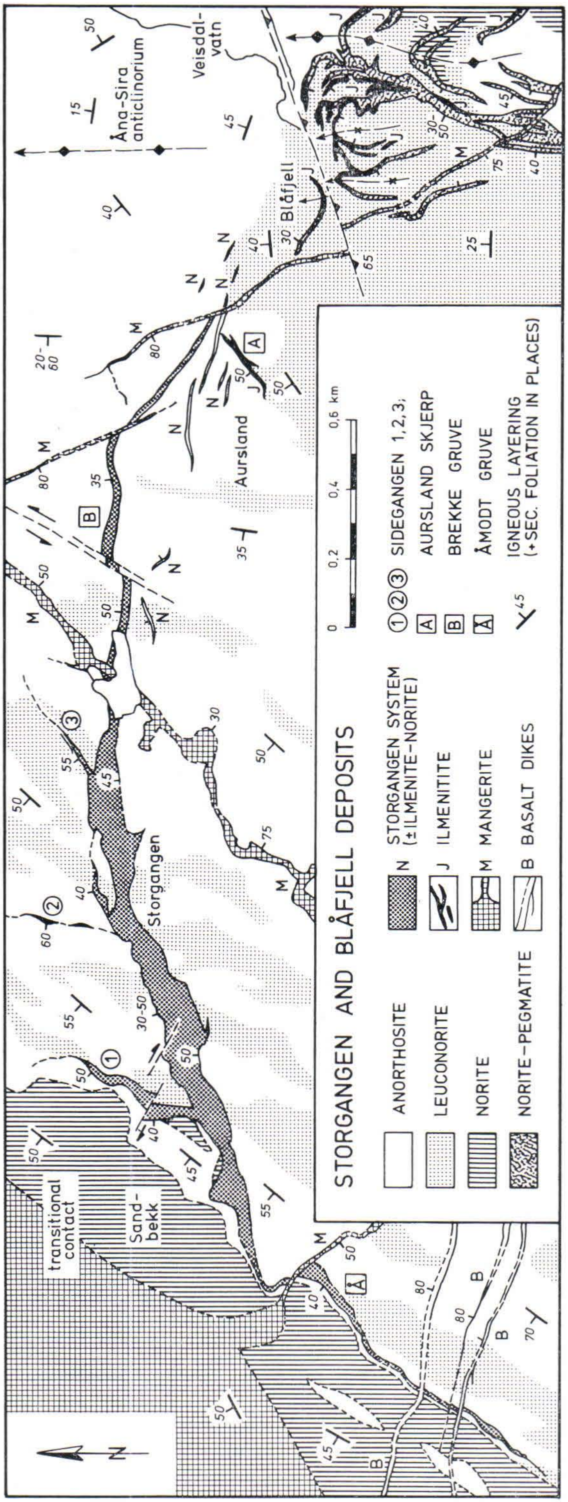

Fig. 7. Geological sketch map of Storgangen and Blåfjell deposits.

the lateral endings and, in places, 50 m in the main body. In exposure and hand specimen, a distinct layering on a centimeter—to meter-scale is typical, caused by a rhythmic sequence of a complex association of rock and ore. In a cross section, the structure of the Storgangen main body shows distinct relationships with the zoning of the layered intrusion of Böstölen, as well as with the arrangement of the rock and ore layers in the noritic externides. The ore body starts out at the basis with comparatively poorly structured massive ore and ilmenite-norite and shifts gradually to an upper part with conspicuous layering of material both rich and poor in ore. The uppermost part of Storgangen is mainly occupied by norite, leuconorite, and anorthosite. The main ore component is ilmenite; the titanomagnetite proportion in the ore alternates between 0 and 30 %. The gangue consists chiefly of plagioclase (43—55 % An) and orthopyroxene (25—30 % $FeSiO_3$). A cross section at the western end of the Storgangen main body shows that within the dike matter from the basis to the top a distinct differentiation development can be traced, which is marked by a continuous increase in the magnetite proportion of the ore, the $FeSiO_3$ contents of the orthopyroxenes, and a corresponding decrease in the An-values of the plagioclases (Krause and Pape 1975).

Norite dikes

The conjugated net of norite dikes possesses its greatest density in the west and to the south of the Åna-Sira massif. The thicknesses of the dikes amount to up to 10 meters. Northwards as well as eastwards, the number and thickness of the dikes decrease continuously. The layered intrusion of Böstölen is cut by those norite dikes, which end in the underlying norite-pegmatite body of Blåfjell-Måkevatn. The dike system can be followed westward up to the contact of the anorthositic series with the lopolith of Bjerkreim-Sokndal, where it is cut off by the externides of the younger mangeritic series. In the noritic externides no norite dikes have been observed. In the anorthosite massif of Håland-Hellaren, adjacent to the lopolith to the west, however, a corresponding norite dike system is once more to be seen. Occasionally, the dikes end in norite-pegmatite stocks, containing quartz and K-feldspar. Ilmenite tapestries, up to decimeter thickness, appear along the always sharp selvages. Some parts of the dikes, generally near the lateral ending, are completely filled with coarse grained ilmenite-norite to ilmenite. Titano-magnetite seems to have an absolutely subordinate share in the composition of the ore. The main gangue consists of plagioclase phenocrysts with An-contents between 30 and 50 %. The thicker ore dikes show a schlieric-rhythmic intercalation of material poor and rich in ore. Developments in differentiation within the dike matter are, if at all, only faintly developed and directed again to one side from the basis to the top. Almost all minor Fe-Ti-deposits outside of the noritic internides must be classified directly with this noritic-norite-pegmatitic dike system (Christiansen 1976).

Noritic externides

The anorthositic series are covered sheet-like by the noritic externides, as results from the distribution of noritic intrusives in the north and east of the Åna-Sira anorthosite. The thicknesses of this subintrusion are dependent upon the spatial-structural position in relation to the anorthosite massifs. The thickest of the norite bodies is situated in the non-linear Bjerkreim-Sokndal synform between the Åna-Sira massif respectively, in the West. In the northeast and the southeast, the Åna-Sira massif is girth-like surrounded by a number of further norite bodies.

The structural relations between anorthosite and norite are inconformably-intrusive. The contacts between norite and mangerite are unconformable as well as gradual. At Sandbekk in the northwest and also west and north of the Hidra anorthosite in the southeast, this contact possesses all the properties of a conformable transition caused by progressing magmatic differentiation. Because of the intensive gneissification in the eastern part of the Åna-Sira massif, the original inconformably-intrusive structural relationships of the noritic externides towards the anorthositic series are almost entirely obliterated; in the contact area anorthosite-norite, the impression of a straky heterogenous leuconoritic rock arises.

The most common rock of the externides is a homogenous leuconorite to norite usually displaying a pronounced secondary parallel texture. The most striking characteristic of the rock series in the noritic externides is, as in Böstölen rhythmites and in the Storgangen system, a rhythmic igneous layering, which is caused by differences in composition and/or grain size. Within a rhythm, upon a basis rich in ore and the thickest ore layer often at the bottom, follows a noritic zone with intercalations rich in pyroxene, changing upwards gradually into a rhythmic interbedding of norite, leuconorite, and subordinate anorthosite. This rhythmic composition layering again is accompanied by a regular variation of the titanomagnetite proportion in ore and rock, as well as a cryptic layering, which is demonstrated by systematic change of the An-contents of the plagioclases, as well as by the $FeSiO_3$-contents of the orthopyroxenes. As at the Storgangen, but in contrast to the layered intrusion of Böstölen, the titanomagnetite proportion of the ore and the $FeSiO_3$-contents of the orthopyroxenes increases in a upward direction of the single rhythm as well as of the total sequence, while the An-values decrease (Krause and Pape 1975; Christiansen, 1976). The titanomagnetite proportion of the ore lies between (0)—5—70 %, the An-values of the plagioclases vary between 32 and 50 %; the $FeSiO_3$-contents of the orthopyroxenes between 30 and 40 %. Characteristic is a comparatively high clinopyroxene-orthopyroxene ratio. Though subordinate the whole clinopyroxene in places makes up to 20 % of the modal composition of the noritic externides.

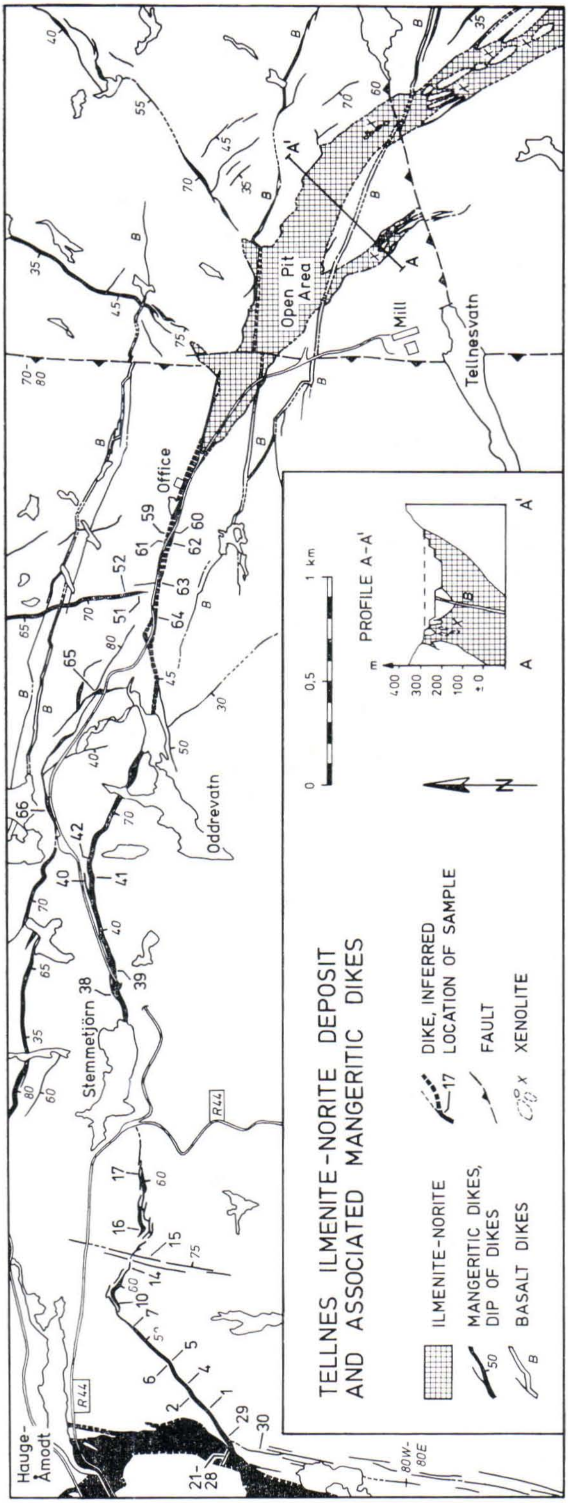

Fig. 8. Geological sketch map of Tellnes ilmenite-norite deposit and associated mangeritic dikes.

Mangeritic series

The units of the anorthositic as well as the noritic series are cut inconformably by dikes and stocks of a complex association of rock types which is termed mangeritic series, due to its general modal composition (Table 4). This subintrusion is considered to represent the final stage of the magmatotectonic evolution of the South Rogaland igneous complex. Conformable and gradual transitions to parts of the underlying noritic externides as well as approximation of basis rock types to noritic rocks indicate a comagmatic character of the mangeritic series. As in the noritic series, the mangeritic subintrusion may be subdivided into internides and externides, both being linked by a network of dikes. In the case of the mangeritic series, the internal unit (Tellnes ore body) also displays a more basic stage of differentiation than do the corresponding externides within the lopolith of Bjerkreim-Sokndal.

Structurally mangeritic internides, conjunctive links and most of the externides are positioned within a complex system of major or minor fissures. The mangeritic dike complex represents the final stage of the upfolding activity of the Åna-Sira anti-form, entirely characterized by fracturing without any plastotectonic deformation (Fig. 9).

The crest fracture, occupied by the mangeritic

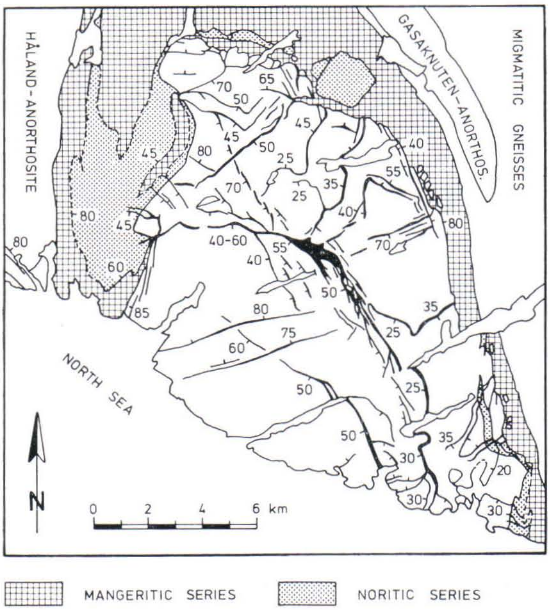

Fig. 9. Radial dike system of the mangeritic series within the Åna-Sira anorthosite.

internides and forming the ilmenite-norite deposit of Tellnes, is situated at the eastern margin of the culmination of the Åna-Sira dome. This ultimate crest fracture is connected immediately with a dike system, which in relation to the dome, is arranged radially as well as concentric. The dike system, in turn, is linked on all sides with the mangeritic externides.

The mineralogy of the mangeritic series is complex. In addition to the previous main constituents plagioclase (An-contents in the mangeritic series 50—20 %, average 27 %, cf. Table 4) and orthopyroxene (hypersthene) in an increasing proportion quartz, K-feldspar and mesoperthite are found. Even in acid members of the rock series there is always a pronounced dominance of plagioclase in relation to K-feldspar. Characteristic is a deficiency or even total lack of amphiboles, muscovite, or biotite; the latter forming only a minor constituent in material which, on the average, is rich in ilmenite. Other minor constituents are clinopyroxene, apatite, olivine, zircon, spinel, ilmenite and magnetite. In presence of biotite, ilmenite always shows coarse exsolution lamellae of hematite, as is the case in the Tellnes ore. Biotite and graphite, the latter otherwise forming a common accessory in the mangeritic series, are as well as never found together in one sample. Other accessories are baddeleyite and sulfides in the primary and secondary parageneses, described above.

Primary magnetite in mangeritic rocks of the profile M1—68 (Fig. 10 D, Table 4) is predominantly formed idiomorphic to hypidiomorphic. It consistently contains exsolution lamellae of ilmenite, which can penetrate the entire grain and are accompanied by fine spinel dust. Besides, oriented exsolution of spinel in one generation appears. Ulvite (cloth-fabric and exsoluted in (100)) is to be observed, when at the same time graphite was found in the section. In that case, ilmenite never contains hematite exsolution lamellae. Reaction rims between magnetite and ilmenite are very narrow, if present at all. Ilmenite is characterized by the absence of hematite exsolution. Hematite exsolution has been observed only in a few samples from the main mangeritic dike near Tellnes, as well as in the deposit itself. An absence of spinel-D is equally characteristic (Gierth and Krause 1973) in ilmenite of the mangerites. Among the oxides, martite belongs to the secondary paragenesis. It cannot be determined, which part of the martitisation belongs to the deuteric readjustment and which part is due to surface processes. Ilmenite is altered through all the transitional phases to rutile. The overall proportion of rutile, however, is always very small. The rock varieties of the mangeritic series form a monocyclic and seemingly continuous sequence of differentiation, which extends from noritic varieties (special type: ilmenite-norite) via mangeronorite, quartzmangerite, to quartzpegmatite and to pure quartz veins. All rock types may occur within one dike, grading into another gradually, but normally the sequence seems to be arranged concentric within the Åna-Sira massif.

Table 4.

Variations in the modal proportions of minerals and mole % An and Mt 100/(Jl+Mt) in dikes and intrusive bodies of the mangeritic series between Åmodt (Hauge) and Tellnes ilmenite-norite deposit. (Location of samples cf. Fig. 8).

Sample Nr.	Jl	Mt	Gph	Opx	Cpx	Ap	Zr	Bi	Q	KF	MP	PL	%An	Mt·100 / Jl+Mt	thickness of dike
WESTERN EXTERNIDES (MARGINAL MANGERITIC INTRUSION)															
21	5.8	4.7	-	19.5	3.9	3.8	0.1	+	2.2	12.6	0.3	47.1	28.5	45	90.0
22	2.8	3.7	-	14.0	1.5	3.2	0.2	0.2	1.1	15.9	-	57.6	29.6	57	90.0
23	3.6	3.2	-	19.1	3.3	4.2	0.3	-	2.2	7.3	11.1	45.8	28.7	47	90.0
24	1.0	3.8	+	5.7	1.2	0.9	0.4	-	5.4	10.9	41.2	29.5	27.6	80	90.0
25	5.0	3.4	-	23.6	4.1	4.7	0.2	+	1.3	5.3	3.7	48.8	33.6	41	90.0
26	5.7	3.2	-	16.8	12.2	4.6	-	+	0.1	6.2	5.4	45.8	34.8	36	90.0
27	7.8	1.5	-	20.1	6.3	5.4	-	0.3	0.2	2.0	0.5	55.9	32.3	16	90.0
28	5.4	3.1	-	24.8	7.1	6.3	-	0.1	0.4	0.8	2.8	49.4	36.5	36	90.0
MANGERITIC DIKE BETWEEN WESTERN EXTERNIDES AND TELLNES DEPOSIT															
29	1.5	2.6	+	2.3	0.4	0.8	0.4	4.8	17.8	15.3	36.0	18.2	28.1	64	3.0
1	1.6	5.1	+	2.2	1.2	1.4	0.2	0.2	5.0	39.2	0.4	43.5	25.8	76	1.0
2	1.0	2.6	+	1.8	0.8	0.6	0.5	-	12.3	34.4	11.3	34.7	28.2	72	8.0
4	0.7	2.7	-	3.0	0.2	0.7	0.1	-	8.0	25.0	9.8	49.7	27.1	79	6.0
5	1.1	6.4	-	6.9	2.6	2.7	0.5	-	0.4	9.0	1.0	69.4	31.3	85	2.0
6	2.0	4.3	-	10.2	0.3	2.6	0.1	-	0.9	3.8	2.5	73.4	31.3	68	2.0
7	1.0	3.1	-	3.2	-	0.1	0.3	1.6	19.4	33.0	15.8	22.5	27.3	75	4.0
10	4.6	7.0	-	7.4	0.1	2.1	0.7	-	5.7	39.6	0.6	32.3	27.9	60	1.5
14	7.3	14.2	-	9.5	7.0	4.7	0.8	-	0.5	7.1	12.1	36.8	28.4	66	0.3
16	1.0	3.3	-	1.8	0.7	0.6	-	0.2	10.9	15.1	38.5	27.9	27.3	76	2.0
17	1.5	4.0	-	2.6	0.8	1.1	0.1	-	3.8	32.4	6.1	47.5	27.9	73	0.5
38	2.8	4.4	+	7.0	4.8	1.9	0.1	-	5.4	28.5	14.2	31.1	25.8	61	10.0
39	1.9	4.9	+	8.1	2.8	2.4	0.4	-	0.4	27.7	28.9	22.8	27.4	72	10.0
41	2.6	2.9	+	9.6	3.5	1.4	0.1	-	1.5	21.7	17.0	39.8	26.7	53	10.0
42	7.3	1.9	-	19.4	1.6	4.9	0.1	-	1.5	12.7	13.6	37.0	26.5	21	8.0
66	2.2	2.4	+	6.5	0.1	1.2	-	1.8	8.8	31.2	7.2	38.6	25.3	52	7.0
65	1.5	4.0	-	4.5	0.3	1.2	0.5	-	4.1	12.4	37.1	34.5	27.0	72	6.0
51	3.1	5.7	+	6.0	1.5	1.9	0.3	0.2	14.2	13.1	33.6	20.5	28.4	65	1.0
52	2.1	4.8	+	6.5	2.3	1.7	0.2	0.1	12.8	15.4	34.3	20.2	27.7	70	1.0
64	3.1	3.4	+	11.3	1.2	2.2	0.1	-	8.4	24.0	9.8	36.4	27.3	52	8.0
63	2.8	4.8	+	20.5	4.5	5.9	0.2	0.4	2.5	9.4	2.4	46.9	30.2	62	10.0
62	5.2	2.5	-	21.3	2.3	6.7	-	0.2	1.3	14.7	5.7	40.3	31.7	33	15.0
61	5.3	2.2	-	17.4	3.4	5.4	+	0.3	0.7	22.1	1.4	41.9	26.1	29	15.0
60	6.0	2.2	-	20.5	2.9	6.2	-	0.2	0.1	18.3	3.8	39.8	27.0	27	15.0
59	6.8	2.3	-	18.3	5.7	5.6	-	0.2	-	16.0	3.6	41.6	27.7	25	15.0
TELLNES DEPOSIT															
TG 1)	28.6	0.7	-	10.2	0.8	0.5	-	3.9	-	-	-	53.2	35 - 55	2	30 - 300
TD 2)	39.0	2.0	-	15.0	n.d.	n.d.	-	3.5	-	-	-	36.0		5	
DIKELETS AND APOPHYSES															
30	7.1	3.5	-	9.1	14.9	5.4	0.1	-	0.6	0.8	4.1	54.5	35.8	33	0.4
15	7.7	5.4	-	17.9	6.1	5.1	-	1.0	2.2	3.0	5.8	45.8	31.3	41	0.1
40	1.2	3.9	-	1.0	1.3	1.2	0.2	5.5	42.8	33.1	1.8	8.6	30.5	76	0.1

Modal analyses: Jl = Ilmenite; Mt = Magnetite; Gph = Graphite; Opx = Orthopyroxene; Cpx = Clinopyroxene; Ap = Apatite; Zr = Zircon; Bi = Biotite; Q = Quartz; KF = Kali-feldspar; MP = Mesoperthite; PL = Plagioclase.

Thickness of dikes in metres.

Tellnes deposit: 1) (TG); average composition of Tellnes ore according to Gierth and Krause, 1973; 2) (TD): average composition of Tellnes ore according to Dybdahl, 1960.

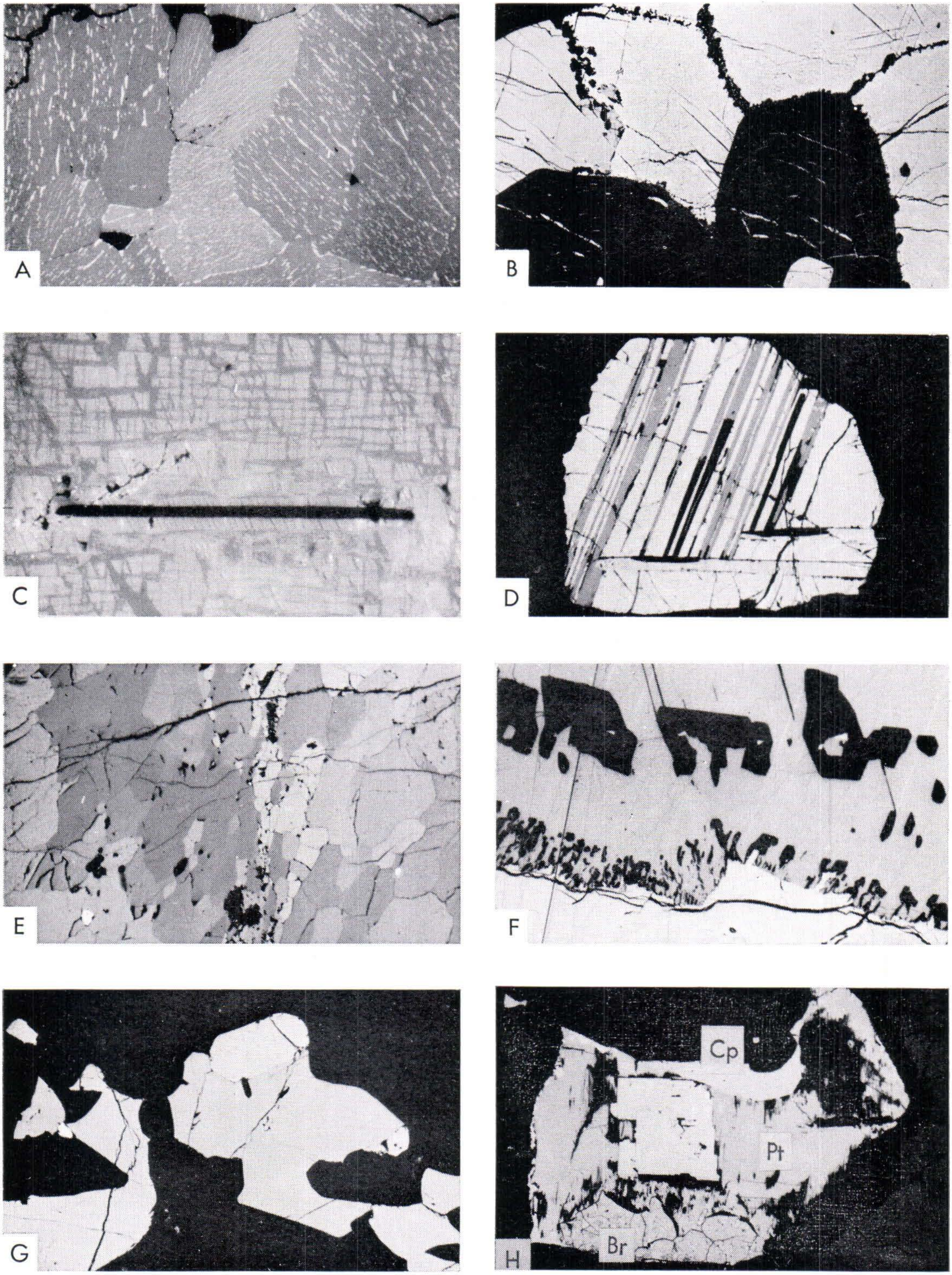
A
B
C
D
E
F
G
Cp
Pt
Br
H

Fig. 10. Photomicrographics of opaques from the Åna-Sira anorthosite and related rocks and ores.

A. Coarse-grained hemo-ilmenite. Lamellae of exsoluted hematite, often thickening towards the margin of the ilmenite grains. Further generations of finest hematite exsolution are present, but not visible at the scale of the photograph. Ilmenititic basis layer within the anorthositic series, NW of the junction between Tellnes road and main road. Width of picture about 0.35 mm.

B. Intergrowth of magnetite (light gray) and ilmenite (dark gray) with olivine-orthocumulate (black, hypidiomorphic). Spinel reaction rim between ilmenite and magnetite. The relief inside the magnetite grains is due to a minute exsolution of ulvite. Inside the olivine in the course of serpentinization formation of secondary magnetite along cracks. At the lower right a magnetite cumulus crystal is included in the olivine. Layered intrusion of Böstölen, west of Laksedalsvatnet, sample Nr. 37 (cf. Table 1). Width of picture about 1.1 mm.

C. Magnetite-ulvite intergrowth, the ulvite (dark gray) having exsolved in box-like shapes and forming a »cloth»-microtexture. Black is a large exsolution lamellae of spinel. Magnetite orthocumulus, included i northopyroxene. Layered intrusion of Böstölen, west of Laksedalsvatnet, sample Nr. 32. Width of picture about 0.06 mm.

D. Isolated magnetite grain in orthopyroxene, showing »sandwich»-microtexture. Two sets of differently orientated, broad ilmenite lamellae (dark gray) are intergrown with magnetite (light gray) and spinel lamellae (black). Layered intrusion of Böstölen, west of Laksedalsvatnet, sample Nr. 44. Width of picture about 0.9 mm.

E. Secondary foliated and recrystallized ilmenite-magnetite-ore from Laksedalsgangen, gallery at Laksedalsvatnet. Magnetite light gray, gangue black. Sample Nr. 16. Width of picture about 0.9 mm.

F. Spinel-bearing reaction rim between ilmenite (dark gray) and magnetite (light gray). Two series of spinel grains (black) lie within the ilmenite, the upper one marking the original contact between ilmenite and magnetite (cf. Krause and Pape 1975). Included in one spinel is an isolated remnant of magnetite. The magnetite shows fine exsolution lamellae of spinel (lower right), as well as the ilmenite. Laksedalsgangen, gallery at Laksedalsvatnet, sample No. 24. Width of picture about 0.4 mm.

G. Ilmenite and magnetite from mangeritic dike matter. Magnetite cumulus (upper middle) is intergrown with xenomorphic ilmenite aggregates. Magnetite displays fine exsolution of spinel, ilmenite shows no sign of hematite exsolution. No reaction rim appears along the interface of both minerals. Included in ilmenite is apatite orthocumulus (black, euhedral). Mangeritic main dike, west of Tellnes, sample Nr. 60. Width of picture about 0.9 mm.

H. Complex intergrowth of sulfides: The primary paragenesis comprises nickeliferous pyrrhotite (Pt, Pentlandite not visible in the picture) and Chalcopyrite (Cp). Secondary minerals are pyrite, which forms an idioblast inside the pyrrhotite and bravoite (Br). Laksedalsgangen, gallery at Laksedalsvatnet, sample Nr. 25. Width of picture about 0.5 mm.

The crest fracture (Tellnes) and broader dike sections are occupied by basic rocks: more acid differentiates are situated in narrower dike portions, quartz pegmatite in offshoots and lesser branches. Table 4 shows the great range of variations in the modal composition of the rocks within a single specific dike matter. In approaching the ore body of Tellnes, the transition quartz mangerite-mangerite-mangeronorite within the dike takes place gradually and without apparent discontinuity (Table 4; cf. also Mylius 1977). The average composition of the Tellnes ore is represented in Table 4. The deposit, its mineralogy, the occurring paragenesis, along with the chemical properties of the ore have been investigated and described by Gierth and Krause (1973). The length of the ore body in the strike extends to 2,7 km and the breadth in the central part amounts to up to 400 m. The overall dip is directed towards the SW at 40 to 45°. Downwards, the ore body probably narrows. None of the drilling has as yet reached beyond the sole of the body. Towards the SE, at the surface, the deposit deteriorates increasingly into dike zones and offshoots with numerous anorthositic xenoliths. Further down, however, the ore forms a compact body with a thickness of several hundred meters as has been established by drilling. Northwestward, the thickness of the ore body gradually diminishes to the mangerite dike extending WNW. Indications of distinct gravitative separation in the ore body are absent. However, a roughly trough-shaped zoning can be observed in the mineral distribution and composition. Toward the contact the contents of ilmenite, magnetite and sulfide diminish within the ilmenite-norite. Likewise, the Cr_2O_3-contents of the ore, which is dependent upon the hematite exsolutions of the ilmenite, as well as upon the spinels and the magnetite, diminish towards the contact. The highest apatite concentrations, however, appear within a 10 m wide zone along the ore body margin. The magnetite and sulfide contents decrease further down, whereas the apatite and Cr_2O_3-contents increase perceptively in this direction.

As a result of considering the ilmenite norite deposit of Tellnes as part of the differentiation sequence of the mangeritic series and especially from the evidence of the immediate association of the ore body with the characteristic radiating mangerite dike system, in the course of an exploration for Fe-Ti-mineralizations and connected components of economic interest, the area to be investigated in detail could be limited considerably. At the time being, however, such an exploration is unnecessary, owing to the large reserves at Tellnes and the present world market situation in titanium.

Conclusions and sequence of events

Anorthositic, noritic, and mangeritic series with all their subunits are of unmistakeably intrusive character. The intrusion of the original material took place in the deep katazone, within an environment which is now represented by migmatitic charnockitic gneisses (P. Michot 1960). It took place under synkinematic conditions during the formation of an eastward inclined fold system with probably non-linear N-S axes from the beginning. This system seems to correspond to the main folding of southern Norway (cf. Falkum 1972).

Considered as a main mechanism of the anorthosite formation is gravitative separation of primary plagioclase crystals (andesine) of melt and mafite orthocumulates in an early stage of crystallization of a magma under almost stable conditions. Caused by relative variations, in density and possibly supplemented by convection currents, the plagioclases rise and/or the mafites sink. The latter may again become unstable in the hotter core of the magma chamber. For the magma below the covering anorthosite layer, at first a tendency to become more noritic must be assumed. The differentiation processes within this region are illustrated in the finding of xenoliths of olivine-magnetite-noritic composition in the noritic internides at Måkevatn (Hense 1974). This finding permits drawing conclusions about non exposed basal units just as the occurrence of labradorite- and bronzite inclusions and adcumulates and more basic xenoliths in the anorthositic series does.

Through continuous EW-shortening as a result of the main folding, the remaining magma chamber is increasingly narrowed. In the overlying anorthositic series, for the most part consolidated, characteristic transversal fissures split open, one of them represented by the Storgangen system. A part of the material from the upper parts of the internal magma chamber is pressed along the contact between anorthositic series and the gneiss cover, the contact representing a pre-existing zone of weakness. The melt is concentrated mainly within the developing synform between Sokndal and Bjerkreim and also along mechanically favourable parts (from the point of E-W compression), on the northeast and southwest flank of the Åna-Sira massif. In these areas the noritic externides develop. By progressing crystallization differentiation, the more developed norite and norite-pegmatite dikes, as well as the conformable intercalated mangeritic externides, are formed. The intrusion of heavier noritic-mangeritic material into areas above the lighter anorthositic layer leads to isostatic equalizing movements; the folding with N-S axes is discontinued in favour of vertical movements, leading to a further lowering of the Bjerkreim-Sokndal synform and correspondingly to a continuous doming of the neighbouring, eastern inclined antiforms in the Åna-Sira massif and the Håland-Hellaren massif. Plastotectonism is superimposed upon the orthomagmatic structures in the anorthositic series and goes along with formation of a subordinate $F_2 \wedge F_1$ with the axes radially arranged around the crest of the upfolding. With continuous upfolding of the Åna-Sira massif, three major crest fractures split open in its culmination area. Caused by the eastward inclination of the anticlinorium and going along with the continuous vertical movements, an eastward directed progradation of the crest fractures takes place. Whereas with the oldest of the internal synkinematic intrusions, the layered intrusion of Böstölen, plastotectonic deformation predominates, obliterating its immediate contacts to the anorthositic wall rock, the breaking up of the younger norite-pegmatite body of Blåfjell-Måkevatn already occurs within a predominantly rigid frame. The structural relationships of the youngest and farthest east situated crest fracture, the ilmenite-norite body of Tellnes, together with its accompanying dike system are exclusively formed by fracturing of an entirely rigid frame.

The series: layered intrusion of Böstölen-norite-pegmatite body of Blåfjell-Måkevatn-ilmenite-norite body of Tellnes not only marks the age and spatial arrangement, but simultaneously characterizes a progressing differentiation development within the magma chamber below the anorthositic series. Predifferentiated melts might be separated from the remaining material underneath and pressed diapirically into the crest fractures of the overlying anorthositic series.

The layered intrusion of Böstölen represents a typical basic intrusion with a far-reaching and one-sided internal development, directed from the bottom to the top. Such a development is only indistinctly present in the more differentiated noritic-leuconoritic norite-pegmatite body and is almost completely lacking in the ilmenite-norite body of Tellnes. Whereas the ilmenite-norite of Tellnes still shows a relationship to the older noritic units, the other rock types of the mangeritic series-mangerite, quartz-mangerite, quartz-pegmatite in an increasing measure take on the character of rest differentiates. The mangeritic sequence is primarily zoned cupola-shaped within the Åna-Sira antiform, shows however, at the same time a certain dependency upon the width of the dikes: Basic ilmenite-noritic to mangeronoritic dike material occupies crest fracture and dike thickenings, acid differentiates are situated in narrow parts of the dikes, quartz-pegmatite in offshoots and accompanying branches.

The formation of the mangeritic externides forms the end of the magmato-tectonic development of the intrusive of South Rogalands. As a result of the vertical movements the antiform of the Åna-Sira massif has been torn from its bonds on all sides; into the opening, which surrounds the anorthositic series similar to a ring dike, mangeritic-quartzmangeritic material is intruded.

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References

Anderson, A. & Morin, M., 1964. Anorthosites in the Grenville of Quebec: a classification. Trans. Am. Geophys. Union 45, p. 125.

Anderson, A. T., 1969. Two types of massive anorthosites and their implications regarding the thermal history of the crust. *In* Origin of anorthosite and related rocks, ed. by Y. W. Isachsen. N. Y. State Museum and Science Service, Mem. 18, 57—69.

Barth, T. F. W., 1933. The large pre-cambrian intrusive bodies in the southern part of Norway. Internat. Geol. Congress, XVI Session, 1, 297—309.

— **1945.** Geological map of the western Sørland. Norsk Geol. Tidsskr. 25, 1—9.

Broch, O. A., 1964. Age determination of norwegian minerals up to March 1964. Norges Geol. Unders. 228, Årbok 1963, 84—113.

Christiansen, T. B., 1976. Die Vererzungen und ihre Verbandsverhältnisse im Norden des Åna-Sira-Massivs (Süd-Norwegen). Diss. TU Clausthal. 83 p.

Duchesne, J. C., 1972. Iron-titanium oxide minerals in the Bjerkreim-Sogndal-Massif, South-western Norway. J. Geol. 13, 57—81.

Dybdahl, I., 1960. Ilmenite deposits of the Egersund anorthosite complex. Mines in south and central Norway. Internat. Geol. Congress, XXI Session, Guide to excursion C 10, 48—53.

Falkum, T., 1966. Structural and petrological investigations of the Precambrian metamorphic and igneous charnockite and migmatite complex in the Flekkefjord area, Southern Norway. Norges Geol. Unders. 242, 19—25.

— **1972.** On large-scale tectonic structures in the Agder—Rogaland region, Southern Norway. Norsk Geol. Tidsskr. 52, 371—376.

Gerling, E., Kratz, K. & Lobach-Zushenko, S., 1968. Precambrian geochronology of the Baltic Shield. Internat. Geol. Congress, XXIII Session, 4, Geology of the Precambrian, 265—273.

Gierth, E. & Krause, H., 1973. Die Ilmenitlagerstätte Tellnes. Norsk Geol. Tidsskr. 53, 359—402.

Hense, J., 1974. Lagerstättenkundliche Untersuchungen an den Ilmenitvorkommen am Måkevatn. Dipl.Arb., TU Clausthal, (unpublished).

Hermans, G. A. E. M., Tobi, A. C., Poorter, R. P. E. & Maijer, C., 1975. The high-grade metamorphic Precambrian of the Sirdal—Ørsdal area, Rogaland/

Vest-Agder, SW-Norway. Norges Geol. Unders. 318, 51—74.

Hess, H. H., 1960. Stillwater igneous complex, Montana. A quantitative mineralogical study. Geol. Soc. Am., Mem. 80. 225 p.

Knorn, H. & Krause, H., 1977. Die Verbandsverhältnisse südlich von Tellnes im Zentralteil des Åna-Sira-Massivs (Süd-Norwegen). Norsk Geol. Tidsskr. 57, 85—95.

Kolderup, C. F., 1896. Die Labradorfelse des westlichen Norwegens, I. Das Labradorfelsgebiet bei Ekersund und Soggendal. Bergens Mus. Aarbog 5.

— **1914.** Fjeldbygningen inden rektangelkartet Egersunds omraade. Norges Geol. Unders. 71. 60 p.

Krause, H. & Pape, Hg., 1975. Mikroskopische Untersuchung der Mineralvergesellschaftung in Erz und Nebengestein der Ilmenitlagerstätte Storgangen (Süd-Norwegen). Norsk Geol. Tidsskr. 55, 387—422.

Krause, H., Pedall, G. & Christiansen, T. B., 1977. Abschlußbericht zum DFG — Forschungsvorhaben Kr 242/1—11, Untersuchungen von Ilmenitlagerstätten und Nebengesteinen im Åna-Sira-Massiv (Süd-Norwegen), (unpublished).

Michot, J., 1972. Anorthosite et recherche pluridisciplinaire. Ann. de la Soc. Géol. de Belgique 95, 5—43.

Michot, J. & Michot, P., 1969. Geological environments of the anorthosites of South Rogaland, Norway. *In* Origin of anorthosites and related rocks, ed. by Y. W. Isachsen. N. Y. State Museum and Science Service, Mem. 18, 411—423.

Michot, J. & Pasteels, P., 1968. Etude géochronologique du domaine métamorphique du sud-ouest de la Norvége. Ann. de la Soc. Géol. de Belgique 91, 93—110.

Michot, P., 1960. La géologie de la catazone: Le probléme des anorthosites, la palingenese basique et la tectonique catazonal dans le Rogaland méridional (Norvège méridional). Internat. Geol. Congress, XXI Session, Guide d'Excursion A 9. 54 p.

Mylius, H.-G., 1977. Lagerstättenkundliche Untersuchungen in der südlichen Fortsetzung der Ilmenitlagerstätte Tellnes (Süd-Norwegen). Dipl. Arb. TU Clausthal, (unpublished).

Pedall, G., 1977. Geologisch-lagerstättenkundliche Untersuchungen an Ilmenitvererzungen und ihren Verbandsverhältnissen im Nördlichen Zentralteil des Åna-Sira-Massivs (Süd-Norwegen). Diss. TU Clausthal. 173 p.

Versteeve, A. J., 1974. Isotope geochronology in the high-grade metamorphic Precambrian of Southwestern Norway. Norges Geol. Unders. 318.

Vogt, J. H. L., 1910. Forekomsterne av titanjernsten hovedsagelig med 38—40 % titansyre ved Ekersund—Soggendal in Norges jernmalmforekomster. Norges Geol. Unders. 51, 132—138.

Wager, L. R. & Brown, G. M., 1968. Layered igneous rocks. Oliver and Boyd Ltd., London. 588 p.

Zeino-Mahmalat, R., 1972. Untersuchungen an Plagioklasen aus dem Anorthosit-Komplex von Åna-Sira (Süd-Norwegen). Diss. TU Clausthal, 140 p.

Zeino-Mahmalat, R. & Krause, H., 1976. Plagioklase im Anorthosit-Komplex von Åna-Sira-, SW-Norwegen. Petrologische und chemische Untersuchungen. Norsk Geol. Tidsskr. 56, 51—94.

PETROLOGIC AND PHYSICO-CHEMICAL ASPECTS OF METAMORPHOGENIC ORE MINERALIZATION

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Introduction

Petrologic and physico-chemical aspects of metamorphogenic ore formation in Precambrian rock complex constitute a problem of scientific and practical importance. Until recently Precambrian rocks were thought to be relatively poor in mineral deposits. But now the concept has aquired a new aspect (Sidorenko *et al.* 1978). Abundant ore mineral deposits (such as alumina, iron and iron-manganese ores, graphite, etc.), attracting a close attention of numerous geologists, are discovered and explored in metamorphic Precambrian rocks.

All recent research on metamorphism, including the metallogeny of metamorphic complexes, is based on the facies classification of the conditions of endogenous mineral formation processes proposed by P. Eskola.

The latest facies schemes dealt have been concerned essentially with developing and elaborating the ideas of P. Eskola. In their attempt to radically revise and even refute these ideas, H. Winkler and W. Fyfe have in our opinion, been inconsistent.

Moreover, the facies characteristic alone is evidently insufficient for genetic metallogeneous reconstructions, which require that the various rock compositions be also taken into account. Here we must note the essential contribution of some of the Scandinavian geologists (e. g. Eskola, Saksela, and Marmo), who stressed the importance of specific rocks such as black shales, leptites and others for ore generation in metamorphic Precambrian rock series.

The contribution of Soviet authors has played an important role in studies of metamorphogenic deposits. Particular mention should be made of D. S. Korzhinskii and his group, who concentrated on paragenetic analysis and the study of metasomathism. The most important basis for further investigations and generalization was laid by the Metamorphic maps of the USSR (Dobretsov *et al.* 1966) and of continents (Metamorphic Map of Europe 1973; Metamorphic complexes of Asia 1977).

Nevertheless metamorphic controlling factors applied to the formation and distribution of mineral deposits in the Precambrian metamorphic complexes have not been sufficiently studied and the subject is still at the discussion stage.

Therefore, in the present paper we have endeavoured to shed some light on the results obtained in this field and to describe the numerous attempts made to solve these problems.

Petrologically, we shall briefly discuss the principles underlying the classification of metamorphogenic deposits, their formational control and their position in the tectonometamorphic cycle; physico-chemically, we shall suggest some possible mechanisms of metamorphogenic mineralization (oreforming process), mainly the problems of migration of matter during metamorphism, in fluid regime as well, and the relationship between the progressive and regressive stages.

However, owing to the specific nature of the subject discussed in this paper our main attention will be devoted to physico-chemical rather than general aspects of geology, even with regard to the first of one problems. We think that it is important to outline in general the problems with which we shall deal in future studies.

Classificational principles and metamorphic criteria

The concentration of ore minerals in metamorphic rock series may be either prometamorphic or generated during metamorphism by the process of accompanying metasomatism and palingenesis: through redeposition of prometamorphic concentrations (rheometamorphic) or the generating of new (syn- and orthometamorphogenic) deposits.

To establish the criteria by which metamorphogenic deposits can be distinguished in each particular case it is necessary to take into account all the combinations of the various controlling factors (Table 1):

1) metamorphic (belonging to specific metamorphic zones);
2) lithological (belonging to terrains of specific composition);
3) tectonic (connection with specific fold or fault structures in a metamorphic formation);
4) geochemical (areal connection with specific kinds of accompanying metasomatism, geochemical anomalies of the ore and disseminated elements);
5) magmatic (the existence or lack of areal and geochemical connection with a specific magmatic formation).

A more detailed subdivision of the metamorphogenic deposit classes can be made picking out the main groups of metallic (Fe, Mn, Au etc.) and nonmetallic (apatite, mica, piezoquartz, etc.) mineral products and calling them according to their formal features i.e. the key mineral product and major ore minerals.

The main unit of the classification may be a genetically related group of deposits of similar mineral composition, which occur in a metamorphic series of the same composition and metamorphic regime (metamorphic formation) and under similar geological conditions.

This unit is similar to the definition of an ore formation given by S.S. Smirnov and of a metamorphogenic-metallogenic formation given by IPGG investigators (Velikoslavinsky *et al.*, 1968; Sokolov, Glebovitsky and Turchenko 1976). Syn- and rheo-metamorphic classes are subdivided into metamorphic types and facies (nonzonal or monofacial complexes; andalusite-sillimanite, disthene-sillimanite and jadeite-glaucophane types of »zonal» metamorphism). Prometamorphic deposits are subdivided only into low-, middle-, and high-temperature groups.

Genetically this class can be divided into metamagmatic, metapostmagmatic, metavolca-

Table 1.
Criterions of different classes of metamorphogenetic deposits.

Genetic class	Controling factors				
	Stratigrafic and lithological	Metamorphic	Geochemical	Tectonic	Magmatic
Prometamorphic	Main	Subordinate	Only non-metamorphic	None	None
Rheometamorphic	Clear	Clear	Clear and	Clear	None
Synmetamorphic	Not clear	Main	metamorphic	Not important	Slightly

nic, metavolcanic-sedimentary, metasedimentary and metadestructional deposits. If necessary polymetamorphism is taken into consideration when distinguishing between subtypes (subformations) of the deposits.

Let us illustrate the rheo- and synmetamorphic deposits with the aid of a few examples. As usual the rheometamorphic deposits are located in the same series as the prometamorphic ones, but they are represented by rich orebodies in the form of ore-chutes, nests and lenses redeposited during the metamorphic process. According to the calculations by Y. N. Belevtsev (1968), more than 60 percent of the orebodies of industrial significance in the Krivoi Rog iron rocks are redeposited.

It is noteworthy that like some other ore deposits, the sulfide deposits of the Baltic Shield are also characterized by marked metamorphic and lithological control and specification (Turchenko 1978; Peltola 1976).

Hence, the metamorphic transformations of the andalusite-sillimanite facies of the Svecofennian belt, which were accompanied by migmatization and followed by late-alkaline magnesian metasomatism of the rocks of specific leptite formation, led to the formation of specific metasomatites, the so-called »skel» (cordierite-andalusite-almandine, cordierite-antophyllite-gedrite rocks) and the associated polymetallic ores of the Ommeberg and Falun group in Sweden, and Orijärvi and Aijala in Finland (Eskola 1914; Tuominen and Mikkola 1950).

Metasomatic changes in the White Sea—Lapland belt of quartz-kyanite facies are characterized by low iron mobility which, when passing into an inert state, gives rise to higher concentration zones in alumosilicates and is accompanied by the formation of disthene-garnet-hornblende metasomatites analogous to the »skel» in low pressure belts.

Further evolution, expressed in the appearance of quartz-muscovite facies and accompanied by a decrease in pressure and a change in the alkalinity of the environment, led to the deposition of mobilised components and a change in ore paragenesis. At the same time pyrrhotite mineralization formed first, and was followed by the redeposition of the Cu and polymetallic ores of North Karelia. Analogous processes of Cu-Ni mineralization at the acid dealkalization and late alkaline-magnesian metasomatism formed during the decreasing pressure have also been encountered in granulite high pressure complex e. g. in the Kandalaksha—Kolvits zone (Turchenko 1978).

Numbers of petrographers raise no objections to the metasomatic nature of the above-mentioned cordierite-antophyllite rocks and their genetic relationship with sulphidic deposits, even though mineralogical study of the latter has not been completed. Therefore the above scheme cannot be regarded as finally established; it reflects essentially only the concepts of the authors of this paper. However, detailed studies of the mineral paragenesis and their minerals by microprobe technique permit more rigorous generalizations to be made and the most important problem, that is their significance as a criterion for the deposits of this type, to be solved. Joint investigations are to be carried out in the next few years.

Deposits of the synmetamorphic class, whose ore concentrations were formed directly during regional metamorphism, include palingene and hydrothermal-metasomatic groups. Example of these under different PT conditions are: phlogopite-bearing magnesian skarns, metamorphogenic muscovite and rare-metal pegmatites, gold-bearing metamorphogenic-hydrothermal black schists and quartz-vein rocks, crustal and granular quartz. This series shows most distinctly the relationship between pegmatite formation and the thermodynamic regime of metamorphism.

Pressure corresponding to the lower part of the disthene stability field (P = 5.5—6.5 kbar), temperatures of garnet-disthene-biotite-muscovite subfacies (nearly 630°—680°C) and high fluidal acidity are optimal for the forming of industrial muscovite mineralization. At the same time, the temperatures of muscovite stability indicated suggest, not disthene-orthoclase paragenesis, but a high water partial pressure of about 4—4,5 kbar in fluid. Hence, the pH of solutions is low and the acid alteration processes wide-spread in both granitoids and gneisses (Glebovitsky and Bushmin 1978).

Rare-metal pegmatites (quartz-plagioclase-microrocline-spodumene-albite-lepidolite) in combination with genetically related granitoids shown by thermodynamic parameters are associated with metamorphic complexes of And-Sill type ($P \leq 4.5$ kbar and T = 520°—670°C). For this case, relatively low P_{H_2O} is typical which accounts for low acidity in solutions, that controls the reducing processes in the acid alteration stage and intensive development of the late-alkaline stage processes, with which the rare-metal mineralization is associated.

Metallogenic specialization of metamorphic formations

The empiric peculiarities of the spatial-time localization of metamorphogenic deposits are being investigated in the USSR in two parallel directions:

1) metamorphic formations;
2) metamorphic and mineragenic cycles.

The first direction, which is being developed mainly in Novosibirsk (N. L. Dobretsov, V. S. Sobolev, K. B. Kepezhinskas, V. V. Khlestov and others), proposes that metamorphogenic ore mineralization processes are determined by both the composition of the series and the features of metamorphism with which the other factors (accompanying metasomatism, local geochemical features, inter-formational tectonics, see Table 1) are correlated.

It is easy to take all these factors into account if one adheres to the lithological-metamorphic subdivision, namely, the metamorphic formation which is a lithological unit (suite, series of suites) of some definite composition that has undergone some definite type of metamorphism (Dobretsov *et al.* 1975). The types of metamorphism may be delineated on the basis of the scheme of metamorphic facies (Fig. 1); the spatial features are generalized in Table 26 (see Dobretsov *et al.* 1975). By such an approach the metamorphogenic deposits are included in metamorphic formations, and the metamorphic formations are often confined to some definite epochs (or structural stages).

Examples of metamorphic formations of shields and central massifs, folded zones and ophiolite belts are given in Table 2. Note that

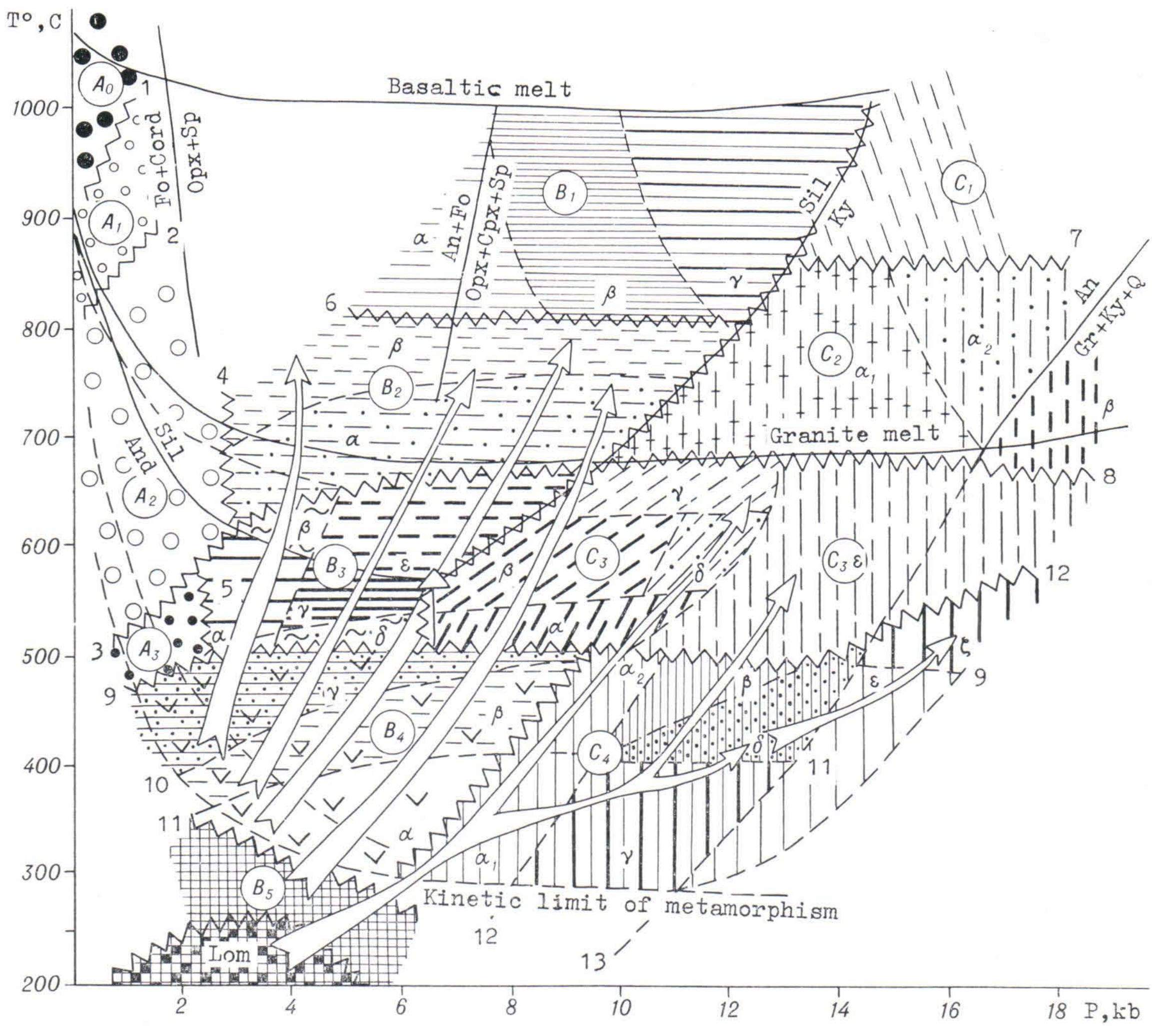

Fig. 1. The scheme of metamorphic facies. The letters α, β, γ, δ, ε and the dashed lines indicate subfacies. The figures stand for the curves confining the stability of the minerals and associations: 1 = spurrite, tillite, mervinite; 2 = dolomite, magnesite, calcite with quartz, almandine; 3 = muscovite with quartz (± biotite), grossular with quartz; 4 = iron cordierite (± garnet, Kspar); 5 = staurolite with quartz; 6 = ortho- and clinopyroxene with feldspars; 7 = common hornblende in metabasites; 8 = muscovite and staurolite with quartz; 9 = Mg-cordierite, almandine, andalusite (kyanite), hornblende in metabasites; 10 = biotite (in common metapelites); 11 = pumpellyite with chlorite; 12 = lawsonite; 13 = jadeite with quartz. The arrows indicate mineralogically diverse types of metamorphic zoning.

the shield formations corresponding to granulitic facies B_1, are poor in ore mineral deposits; carbonate-granulite formation, with phlogopite-bearing and iron ore skarns is of main importance here. The more prospective are middle and low temperature formations.

Examples of these are the zonal disthene-phyllite-gneissic and andalusite-phyllite-gneissic formations of folded zones. As was noted above, the former contains all industrial deposits of muscovite pegmatites. The muscovite pegmatites and the formations in which they occur

Table 2.
Examples of metamorphic formations (metamorphic terrane types) and related metamorphogenic ore deposits.

Tectonic position	Metamorphic formation	Type metamorphism from table 26	Facies	Lithology	Type of mineral deposits	Examples
1	2	3	4	5	6	7
Shields and Median Massifs	Charnockite-basite Eulysite-quartzite	1 1—2	B_1 $B_1 \pm B_2$	$B+Ch \pm P$ $FR+P \pm B$, C	— Fe, Mn	Antarctica Baltic and Aldan shields
	Khondalite	1	B_1	$P+Ch+B+G$	TR, Graphite and high - aluminous rocks	Ukraina, India
	Carbonate-granulite Carbonate-migmatite	1 2	$B_1 \pm B_2$ B_2	$C+P \pm B$, Ch, M $P+B+M+C$	Phlogopite, spinel, lazurite, sometimes Fe, B, Mg	Aldan, Madagascar
	Amphibolite-plagiogneiss	2	B_2	$B \pm P$, M	Ti	Urals
	Eclogite-gneiss	3	C_2	$P+B$ (E)$+M$	Ti, diamond? Abrasive	C. Kazachstan, Norwey
Folded Regions	Zoned andalusite-phyllite-gneiss	7	$B_2+B_3+B_4$: $_2 \pm B_3$ $B_3 \pm B_2$ $B_4 \pm B_3$	$T+P$	Al, Sn, Li, Mo, W, Cu, Pb, Zn, Cu	Baikal region, South Finland, Sweden, Rhodesia
	Zoned kyanite-phyllite-gneiss	B	$C_3 \pm B_4$		$T+P$	Al (kyanite), muscovite, abrasives	Baikal and Kola provinces, Ural
	Jaspilite	5, 7, 6	$B_4 \pm B_3$, C_3	$FR+B+T \pm SS$	Fe	Ukraina, U.S.A.
	Gondite	5, 8	$B_4 \pm B_3$, C_3	FR $(\pm M)+T \pm C$	Mn	India, Brazilia
	Quartzite-phyllite, metaconglomerate	5	B_4	$P+T+G$	Au, U	Witwatersrand, North Baikalian
The zones of the deep-seated faults and extensive nappes	Ophiolite	4	B_4+B_5	$B+U \pm T$, C	Talc, asbestos Cu—Au Pb, Zn, $BaSO_4$	Ural, Dinarides
	Zoned glaucophaneschist	9 δ	C_4+B_4	$T+B \pm U$, SS	Mn, Cu	Japan
	Eclogite-glaucophaneschist	10	C_4+C_3	$T+B(E) \pm U$, SS	Ti, Cu, Pb, Zn, Au	Italy, South Ural

Note: B) metabasites (metamorphosed volcanogenic sequens of basic composition); C) carbonate rocks; Ch) true charnocites (acid); E) eclogites; EL) eclogitic rocks; FR) ferruginous rocks; M) migmatites; P) metapelites; T)terrigenous complex of rocks; SS) siliceous shales and quartzites; U = ultramafic, G = graphite rocks.
Under the heading of metamorphic formation (or metamorphic terrane type) we understand a regular parageneses of metamorphic rocks of defined composition, which have undergone a definite type of metamorphism (see Table 3).

belong mainly to the Proterozoic group (Sokolov, 1970). The initial composition of all these complexes correponds to the terrigenous series of flysch type, but the muscovite pegmatites are formed only on the disthene-garnet zone. Many sulphide deposits associated with zonal andalusite-phyllite-gneissic formation in Precambrian folded zones are also characterized, as was mentioned earlier, by metamorphic and lithological control and by specific metasomatism (andalusite-staurolite or biotite zone, presence of black graphitic shales, cordierite-antophyllite metasomatites). At the same time, however, no ore mineralization occurs in series that are of analogous composition or analogous structure, but which differ in zonal subfacies $B_3\alpha$ (or $B_3\beta$), or in zonal metamorphic complexes. This points to the close genetical connection between ore mineralization and the processes of regional metamorphism of a specific type.

The metallogeny of metamorphic formations in ophiolitic zones, the most characteristic of which are the ophiolites themselves and glaucophane zonal formation, is specific enough for each zone (for example, ancient Benioff zones), owing to their »transparent» and long-lasting development. Many deposits were evidently imposed and sheared during metamorphism (for instance, asbestos, nephrite, jadeite, some sulphide ores). These are mainly Phanerozoic formations and belts, but at recently some examples of Precambrian formations have come to light that are interesting in terms of Precambrian metallogeny.
The Maksyutov complex in the South Urals (Dobretsov 1974) is one example of Precambrian eclogite-glaucophane-schist formation which contains deposits of rutile (in eclogite) and industrial white mica (in schist). Another example is the Precambrian ophiolitic formation of West Sajan with its large deposits of metamorphogenic asbestos (Dobretsov *et al.* 1977).

Tectono-metamorphic and mineragenic cycles

The main idea of this line developed in Leningrad (Yu. M. Sokolov, V. A. Glebovitsky, S. I. Turchenko and others) is that the origin and form of the concentration and dissipation of the components of a deposits are also controlled by stages of tectono-metamorphic cycles in rock complexes, whose thermo-dynamic parameters pertinent to the isofacial origin of metamorphic processes in mineragenic and/or metallogenic formations have optimum values.

Let us examine this statement taking metamorphogenic pegmatites as an example. The analysis of the data on the maps of the metamorphic belts of the USSR (Metamorphic belts of the USSR, 1975) and the pegmatite localities of the USSR enables us to designate empiric regularities in the distribution of the main industrial pegmatitic formations, which are connected with endogenous processes in the metamorphic belts. Rare-earths-ceramics formation is generated at early stages of the evolution of metamorphic belts. These stages are characterized by a lack of significant P and T gradients in horizontal and vertical directions; this explains the weak intensity of postmigmatitic phenomena. The specific regime of fluid, corresponding to P-T conditions of the granulite facies, and low P_{H_2O} in particular, does not stimulate the generation of ore mineralization in pegmatites of the given formation. The muscovite formation (T = 650°—680°C) takes place only in kyanite-sillimanite facies series and mainly in the zone of kyanite-almandine-biotite-muscovite subfacies of the almandine amphibolite facies. Clear differences between pegmatites formed during a single tectono-metamorphic cycle in a given place (monometamorphic belts — Northern Baikal mountains) and pegmatites formed during an overprinted tectono-metamorphic cycle (polymetamorphic belts — White Sea region, Stanovoi mountain range, Eastern Sajan) were established for the pegmatitic provinces and muscovite formation fields.

Quartz-feldspathic association of pegmatites in monometamorphic complexes is formed during the early and middle stages of the evolution of endogenous processes (»early inversional» and »late inversional» pegmatites), whereas the same paragenesis is formed in pegmatites of polymetamorphic complexes only during the middle stages (»late inversional» pegmatites). By all parameters the full cycle of metamorphogenic pegmatitization corresponds to the stages in the metamorphic cycle from the moment of generation of a pegmatite-forming substance of quartz-feldspatic composition to its turning into pegmatite with subsequent transformation. The first stage of pegmatitization, corresponding to the final stages of formation of vertical metamorphic zonality during the early inversion stage, is expressed in dia- and metablastesis, which occurs only in pegmatite fields in zonal monometamorphic complexes. The second stage, corresponding to the period of thermal dome forming, horizontal metamorphic zonality and formation of ultrametamorphic granitoid, is observed in all pegmatite provinces and fields. This is the magmatic stage of high-grade regional metamorphism (vein facies forming); it is simple in genetic plan (selective melting) and differs in different regions only in scale and intensity.

The generation of metamorphogenic deposits is a consequence of the definite combination in space and time of exogene and endogene differentiation in the earth's crust. The degree of differentiation and fractionation increases in the course of time during repeated processes of sedimentation, weathering, magmatism and ensuing metamorphism. In this respect the role of rheometamorphic and orthometamorphic deposits increases relative to that of prometamorphic ones.

The same tendency of gradual change and complication on the composition of metamorphogenic deposits together with the appearance of new metals in the ore-forming process characterises the passage from the Archean to the Phanerozoic.

Physico-chemical aspect

Physico-chemically metamorphogenic ore generation has been reduced almost entirely to two alternative versions: differentiation during the process of metamorphism and the supply of some components from external sources (fluids of mantle origin or related to magmatic intrusions). Trivial cases of ordinary metamorphism in primary ore concentrations deserve no special discussion here.

The essential differentiation of matter in metamorphism can be due to only two mechanisms:

1) redistribution by means of fluids and
2) partial melting and segregation of the anatectic melt.

For the first case we practically always have to suppose the transfer of ore components at the expence of fluid infiltration, while the pure diffusional model can be taken into consideration only in those rare examples when metasomatic ore generation takes place in direct contact with highly disequilibrated rocks (possibly some bimetasomatic phlogopite deposits or corundum plagioclasites).

The supply of matter from external sources owing to infiltration is even more possible. Hence, the genetic problems of metamorphogenic ore generation are essentially reduced to problems of the fluid sources, their chemistry, flow intensity, and shape and the causes of their localization. Theoretical works on metamorphism devoted to these problems (Dobretsov *et al.* 1970; Khlestov 1975; Fluidal regime of the Earth's crust and upper mantle 1977) and the animated discussion of this topic in recent years aim to help solving exploration tasks.

To date the following aspects of the problem have been investigated and discussed:

a) the composition of fluid in metamorphism and its influence on the P-T parameters of mineral equilibrium;
b) correlation between fluid pressure and the pressure in solid phases;
c) the sources of deep-seated fluid, the scales and mechanisms of its transfer.

After long debates on the deficiency or excess of H_2O during metamorphism the concept was introduced of the presence of fluids with complex composition where $P_{H_2O} < P_{fl}$. It was proposed that P_{H_2O} does not exceed a definite limit (about 1 500 atm.) (Sobolev 1961). This concept was widely adopted, however, it needed some elaboration, namely, that P_{H_2O} varies in a wide range; the limits of these variations are still a subject of discussion.

One of the main factors causing these variations is the effect of anatexis in high-temperature facies. The melt formed dissolves the major part of the water and dries the coexisting fluids. After numbers of discussions on the concept of nonmagmatic granitization, most petrographers have returned to treating migmatites as a result of partial melting. Owing to recognition and homogenization of recrystallized inclusions (Table 3) direct evidence has been obtained at the Institute of Geology and Geophysics for the molten nature of leucosomes in migmatites and for the specific temperatures of their formation.

The gratest disagreement in the evaluation of the role of water pressure lies in the low-temperature facies.

Unanimity has not been reached in this topic, not even by the authors of this paper. Some of the authors believe that water pressure increases monotonously with increasing total pressure under isothermal conditions. Other believe that there is a maximum of P_{H_2O} at P_{total} 4—5 kbars, which gives us lower estimates of temperatures at higher pressures.

Another important achievement in studies of metamorphic fluids is the discovery of high-temperature complex inclusions containing water and salt, liquid carbon dioxide and gas. These were firstly observed during studies of kyanites in muscovitic pegmatites from the North Baikal region (Sobolev and Bazarova 1963) and later from some other regions in the USSR (Sedova 1977). Here the estimates of temperatures, given by the dissolving of the solid phase, were to be at 600°—700°, while the study of the CO_2 inclusions suggested pressure limits from 3 to 8 kbars and the fluid composition ($X_{CO_2} = 0.3$ in fluid).

Some small carbon dioxide inclusions were found in rocks of granulite facies (Tomilenko *et al.* 1976; Cuney *et al.* 1976). Thus the significant role of carbon dioxide in fluid has been proved outside the decarbonizing rock series. Some peculiar inclusions of nitrogen and neutral gases have been established, but these have not been studied enough (Dolgov *et al.* 1970).

The following models exist for the sources and effect of fluids:

Table 3.

Temperatures of homogenization of melt inclusions in quartz (after Dolgov Yu. A., Bacumenko I. T. *et al.*)

Object of investigation	Locality	Temperature of homogenization (T °C)
Granite and leucosome of migmatite of granulites	Upper Aldan Region	900—840
Unclean graphic aggregates of granulites	—»—	850—820
Pegmatoid and graphic aggregates of granulites	—»—	840—790
Bi—Amph anatectite of amphibolite facies	Olecma river	790—760
Unclean graphic aggregates in migmatites of amphibolite facies	Western Tuva	≈800—790
—»—	Upper Aldan region	820—790
—»—	South-Chuisky Range	≤710

1) Fluids of deep-seated origin leading to granitization without melting. As described above, this widely held earlier hypothesis has now been rejected as a result of direct observations.
2) Deep-seated fluids, as heat transfer and source of some components, alkalies in the first place lead to the melting of rocks, carrying out of some components — the theory of magmatic substitution and »through magmatic» solutions by D. S. Korzhinskii.
3) The deep-seated fluids from the mantle or from crystallized deep magmatic masses supplying alkalies favour the formation of granitic eutectic in rocks rich in Al_2O_3 without significant loss of components.
4) The deep-seated fluids supply no essential amount of other components, but, owing to the presence of water, reduce the temperature of granitic eutectic by »drying» the circulating fluids.
5) Fluids formed mainly due to metamorphic dehydration and decarbonation reactions; their migration may affect the alkali transfer and ore components in particular.

Most authors believe that the main discussion should now turn to the relative importance of the processes formulated in items 3—5. However, so far, many of the objections to items 2) and in part to item 1) have not been eliminated. We are now obliged to continue a detailed study of granitic rocks in migmatite zones, which will be of great importance for solving the problems staded above.

The metamorphic rock series produce abundant fluids in the progressive stage owing to dehydration of the water-bearing minerals and decomposition of some carbonates. In view of the effect of the increase in temperature on solubility this apparently results in mobilization (transformation to fluid) of the majority of the ore components. The scheme of the processes of metamorphogenic ore formation may be constructed wihout the outer (mantle) fluid sources. Numerous data confirm this assumption. It has recently been established for high-temperature migmatite complexes as for most deep-seated members of metamorphic crustal sections (Shkodzinskii 1976; Kepezhinskas *et al.* 1977) that various chemical conditions exist for synmetamorphic fluids within limited and tectonically homogenous blocks. This proves the absence of sufficient intensive fluid flows from the mantle.

One of the most important aspects in the treatment of metamorphogenic ore generation is the scales of the transfer of matter. For middle- and low-temperature metamorphism, the isochemical nature of the progressive stage is shown for the main rock-forming components, if one means the bulk rock composition of layers of some km thick (Dobretsov 1975; Kepezhinskas *et al.* 1973, 1977). The variations in the main components in this case are limited (see Tables 4, 5) and local, caused entirely by local processes (the trend towards »homogenization» in rock series at metamorphism). This is additional proof of the lack of sufficient fluid flows from the mantle. Discussion of the possible mechanisms of migration of fluids within the mantle leads to the same conclusions (Khlestov 1975).

The isochemical nature of metamorphism in rock-forming components ensures at least that sedimentary concentrations of Fe, Mg, Al are conserved, resulting in the generation of parametamorphic deposits.

»The large-scale» isochemistry of metamorphism established for the major elements does not contradict the possibilities of the redistribution of ore components in metamorphic rock series owing to movement of synmetamorphic fluids. Some of the matter mobilized at lowered temperatures during the progressive stage inevitably re-entered into minerals, since, during the regressive stage, a considerable part of the fluid is bound as a result of hydration and carbonation reactions. Here we also add the effect of parallel change in solubility at cooling and drop in pressure The upward infiltration of

Table 4.

Average values ($\bar{x}$) and standard deviations (δ) of oxides for metapelites ($67 \geq SiO_2 \geq 57$, $CaO \leq 3.00$) from different metamorphic zones (after Kepezhinskas K. B.).

	Facies or zone	Sedimentary shales		Greenschists		Garnet + staurolite		Sillimanite-muscovite	
		I		II		III		IV	
	Components	$\bar{x}$	δ	$\bar{x}$	δ	$\bar{x}$	δ	$\bar{x}$	δ
Wt %, recalculated on dry material and made up to 100 %	SiO_2	63.39	2.42	62.62	2.63	62.27	2.47	62.57	2.38
	TiO_2	0.90	0.43	0.88	0.25	0.97	0.30	0.90	0.23
	Al_2O_3	19.15	1.84	18.89	2.00	18.84	2.34	18.57	2.39
	Fe_2O_3	3.93	2.48	2.99	2.08	2.03	1.64	1.81	1.58
	FeO	4.43	2.40	4.38	2.00	5.79	1.76	6.00	1.51
	Σ Fe .... as Fe_2O_3	7.58	1.90	7.83	2.00	8.45	1.76	8.46	1.29
	MgO	2.80	1.09	3.18	1.24	2.94	1.06	3.10	1.01
	CaO	0.75	0.60	1.08	0.83	1.49	0.74	1.39	0.70
	Na_2O	1.69	1.11	1.99	1.10	1.94	0.86	1.94	0.76
	K_2O	3.89	1.08	3.78	1.33	3.50	1.10	3.53	0.89
H_2O before recalculation		4.88	1.72	3.55	1.07	2.28	0.86	1.99	0.77
Number of analyses		83		97		136		69	

Table 5.

Average values ($\bar{x}$) and dispersions (s) of amounts of oxides for metabasites of Ilpenei suite, Kamchatka.

Components	Glaucophane schists		Glaucophane-bearing metabasalts		Greenschists		Green-stones	
	$\bar{x}$	s	$\bar{x}$	s	$\bar{x}$	s	$\bar{x}$	s
Al_2O_3 ...	14.14	2.26	13.24	2.47	13.51	2.31	13.54	2.46
CaO	9.30	2.65	10.29	5.50	9.83	4.35	11.02	3.28
Na_2O ...	2.67	0.81	3.19	1.08	2.38	1.35	2.31	1.28
K_2O	0.49	0.66	0.67	0.61	0.90	0.83	0.94	1.34
Na_2O/ Al_2O_3 ...	0.19	—	0.23	—	0.18	—	0.18	—
Number of analyses	20		21		16		10	

the fluid accompanied by progressive metamorphism seems to lead to the coincidence of zones of regressive deposition with zones of progressive dissolution. However, in homogenous upward infiltration a mechanism of this sort cannot lead to marked concentration of the ore components. Hence, the principal importance may be attached to two cases only:

1) for ascending flow of frontal fluid over the short section of its route the solubility of the ore components suddenly decreases;

2) the ascending fluid flows are distinctly localized in particularly permeable zones, where the main mass of the ore matter is deposited (metamorphogenic hydrothermal systems after A. A. Predovskii 1972).

A combination of both factors seems to be especially favourable.

A rapid lowering in solubility may allow a fluid to pass the boundary of chemically sharply contrasting rocks, in which case a metasomatic interaction takes place between the fluid and the new environment. The drop in solubility in fluid owing to the sudden reduction in pressure may be marked when the pore fluid flows to well drained fractures. The latter may exist for a long time only under rather low-temperature conditions; in amphibolite or the high-temperature part of epidote-amphibolite facies fractures of this sort have a short life because of rock-creeping deformation. This may serve as a partial explanation for the predominant confinement of the ore deposits to metamorphogenic hydrothermal genesis, to low-temperature metamorphic zones and to regressive stages of higher temperature zones. The case is essentially the same for the piezoquartz deposits of alpine vein type, the formation of which requires the necessary open cavites. The degree of jointing evolution with localization of the fluid flows apparently depends on intraformational tectonics.

The interval of the middle and low temperatures is specified by considerable changes in the activity coefficients of acids in solutions, the maximum of some of them being in this region. When such acids are present in the pore fluids their ability to dissolve may also pass through the maximum. This provides an additional explanation for more active hydrothermal mobilization and differentiation of the ore matter during the regressive stage of metamorphism.

Acid leaching and late alkaline metasomathism may be associated with the pH change in solutions caused by this mechanism.

Current models of metamorphic processes imply that the most intense fluid filtration generated by metamorphic rock series specifies the completion of the progressive stage and the beginning of the regressive stage. Empiric data (Sokolov 1970) indicate that the formation of synmetamorphic (ortho-metamorphic) deposits is synchronous with the beginning of regressive epoch. This is especially clear in instances involded with »anatectic differentiation». At partial melting the scattered ore components forming no independent minerals in the original rock series may mainly concentrate in anatectic melts (when the partition coefficient between the melt and mineral substrate is more than 1). At the crystallization of local aggregates of a anatectic masses in the regressive stage, rare elements »separated» from the substrate would tend to reach the residual melt, where at their final crystallization they would finally form concentrates of industrial importance with ore minerals present in the orebodies but lacking in the surrounding rock series. The forementioned »melt-substrate» partition coefficients may vary conspicuously as a function of pressure, largely because of the differences in water saturation of anatectic melts. Therefore, the degree of »separation» of the scattered ore elements at anatexis in various facies series is not identical, which, for instance may be explained by the confinement of rare-metal pegmatites to andalusite-sillimanite type of metamorphism.

The alternative confinement of muscovite pegmatites to kyanite-sillimanite facies series has a different aspect. At corresponding pressures, the crystallization of anatectites terminates in the T-range of muscovite stability, with high water saturation of the melts ensuring the possibility of considerable reproduction of pegmatite bodies by later metasomatism. Moreover, at corresponding values of PT parameters, it is possible to assume the crystallization of muscovite directly from anatectic melt (»Muscovite pegmatites of the USSR» 1975).

As stressed earlier by A. V. Nikolaev the extensive evolution of independent processes of diaphtoresis is possible only in zones in which abundant fluid flows rich in water circulate. However, it is still not clear whether these fluids derive directly from the mantle or from magmatic crystallization. In this case matter is replaced on a greater scale than in progressive metamorphism but the question of the essential supply of the ore components is still not answered. The problems of formation of the rich ore pools in the deposits studied and explored, for example, the Krivoy Rog iron ores are still being discussed. At this stage the authors of this paper think that the deposits are mainly formed by the replacement and concentration of the local components, as a result of an increase in their concentration.

Thus, even a short physico-chemical discussion of the genetic aspect of synmetamorphic (orthometamorphic) ore generation permits one to conclude that the main factors are: geochemical specialization of the original rock series, intensity and localization of the fluid flows depending considerably on intraformational tectonics during the epochs synchronous with metamorphism. Intensity of fluid generation and the chemistry of the fluids are due to the primary lithology and peculiarities in the metamorphic process (facies-formational control).

As we have already seen, new methods and direct observations have narrowed the framework of reasonable discussion. However, there is still considerable scope for hypothetical reconstructions, requiring further development of the methods of research combined with experimental work to specify the models and digital characteristics. This may be applied in greater degree to the problems of transfer of the elements that form metamorphogenic deposits.

References

Cuney, M., Pagel, M. & Touret, J., 1976. L'analyse des gaz des inclusions fluides par chromatographie en phase gazeuse. Bull. Soc. fr. Mineral. Cristallogr. 99, 169—177.

Dobretsov, N. L., 1974. Glaucophaneschist and eclogite-glaucophaneschist complexes of the USSR, Novosibirsk, »Nauka». 430 p. (In Russian).

— **1975.** Metamorphic formations and metamorphogene deposits. *In:* »Muscovitic pegmatites of the USSR», Leningrad, 36—50. (In Russian).

— **1975.** Petrochemical peculiarities in oceanic and early geosynclinal basalts. Geologia i geofiz. 1975 (2), 11—25 (In Russian).

Dobretsov, N. L., Reverdatto, V. V., Sobolev, V. S. et al., 1966. Metamorphic map of the USSR facies, scale 1 : 750 000. Novosibirsk (In Russian).

Dobretsov, N. L., Reverdatto, V. V., Sobolev, N. V. & Khlestov, V. V., 1973. Metamorphic facies. Moscow, »Nedra», 1973, 432 p. (In Russian).

Dobretsov, N. L., Sobolev, V. S., Sobolev, N. V. & Khlestov, V. V., 1975. Facies of regional metamorphism of high pressures. Canberra. A.C.T., 361 p.

Dobretsov, N. L., Moldavanov, Yu. B. & Kazak, A. P. et al., 1977. Petrology and metamorphism of ancient ophiolites (exemplified by Polar Urals and West Sayans). Novosibirsk 220 p. (In Russian).

Dolgov, Yu. A., Mel'gunov, S. V. & Shugurova, N. A., 1970. Thermodynamic conditions of formation of metamorphic rocks of South-Chuisk Range. Dokl. AN SSSR 192 892—895 (In Russian).

Fluidal regime in the Earth's Crust and Upper Mantle, 1977. Irkutsk, 1977, 113 p. (In Russian).

Glebovitsky, V. A. & Bushmin, S. A., 1978. Acid alkalization and Fe-Mg metasomatites in metamorphic complexes of the Baltic shield. *In:* Processes of deep petrogenesis and minerageny of the Precambrian of the USSR. Leningrad (In Russian).

Kepezhinskas, K. B., 1973. Behaviour of the rock-forming components in the process of regional metamorphism of metapelites. Geol. i geofiz. 1973 (10)' 49—56 (In Russian).

— **1977.** Paragenetic analysis and petrochemistry of the middle temperature metapelites. Novosibirsk, »Nauka», 200 p.

Kepezhinskas, K. B., Korolyuk, V. N. & Khlestov, V. V., 1977. Peculiarities in paragenetic thermobarometry of migmatites. *In* Multicomponent systems. IGIG SO AN SSSR, Novosibirsk, 57—93. (In Russian)

Khlestov, V. V., 1975. Fluid regime of the Earth's Crust and Mantle. *In* Geodynamic study. Moskva, »Nauka», N3, 145—180.

— **1972.** Ore formations. Geol. i geofiz., 1972 (6). (In Russian)

Metamorphic map of Europe, 1973. 1 : 500 000 (In Russian).

Metamorphic belts of the USSR, 1975. Expl. Note and Metamorphic map of USSR, scale 1 : 5 000 000. Leningrad. 47 p. (In Russian)

Metamorphic complexes of Asia, 1977. Novosibirsk, »Nauka». 350 p. (In Russian)

Moskovchenko, N. I. & Turchenko, S. I., 1975. Metamorphism of kyanite-sillimanite type and sulphidic mineralization. Leningrad, 143 p. (In Russian)

Peltola, E., 1976. Origin of Precambrian sulphide ores, Outokumpu district, Finland. 25th Intern. Geol. Congress, Sec. 1, 179—180.

Predovskii, A. A. 1972. Factors and peculiarities in metamorphogenetic hydrothermal ore formation process. *In* Metamorphogenetic ore formation. Kiev, »Naukova dumka».

Sedova, I. S., 1977. Determination of PT-parameters of metamorphism in the light of inclusion studies of mineral-forming media. *In* Thermo-barometry of metamorphic rocks. Leningrad, »Nauka», 138—174. (In Russian)

Shkodzinskii, V. S., 1976. Problems of physico-chemical petrology and genesis of migmatites. Novosibirsk, »Nauka», 224 p. (In Russian)

Sidorenko, A. V., 1978. General problems of metallogeny of the Precambrian in the light of exogenic and metamorphic processes. Abstract *In* Metallogeny of the Precambrian, Helsinki.

Sobolev, V. S., 1964. Physico-chemical peculiarities of mineral-formation in the Earth's Crust and Upper Mantle. Geology and geophysics, 1964 (1), 7—22.

Sobolev, V. S. & Bazarova, T. Yu., 1963. On the temperature of kyanite crystallization in pegmatites. Dokl. AN SSSR 153 (4), 920—922.

Tomilenko, A. A., Chupin, V. P. & Dolgov, Yu. A., 1976. Conditions of formation of metamorphic rocks in the light of inclusion study. *In* Genetic studies in mineralogy. Novosibirsk, IGIG, 138—142 (In Russian).

Tuominen, H. V. & Mikkola, T., 1950. Metamorphic Mg-Fe enrichment in the Orijärvi region as related to folding. Bull. Com. Geol. Finland, 150, 67—91.

Turchenko, S. I., 1978. Metallogeny of metamorphogenetic sulphide deposits of the Baltic Shield. Leningrad. »Nauka».

Velikoslavinsky, D. A., Sokolov, Yu. M. & Glebovitsky, V. A., 1968. Zoning of regional progressive metamorphism and metallogenetic specialization of metamorphic zones. *In* MGK XXII Sess. Dokl. Sov. geol., Problema 4, Precambrian geology. Leningrad, 196—210. (In Russian)

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PRECAMBRIAN BASIC MAGMATISM AND THE Ti-Fe ORE FORMATION IN CENTRAL AND NORTHERN FINLAND

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Introduction

During the present decade in Finland has started to pay more attention to fracture tectonics analysis, i.e. to the significance of fractures in the geological picture of the Precambrian and in ore formation. A useful starting point for this may be found in the Bothnian Bay—Lake Ladoga »main sulphide belt» or zone (Fig. 1), inside which three metallogenic sulphide provinces can be outlined (Kahma 1973): 1. the Outokumpu Cu and Co province, 2. the Vihanti Zn, Cu and Pb province, and 3. the Kotalahti Ni belt. The Outokumpu province is located in the sedimentary cover of Block B indicated in Fig. 1, the Vihanti Province in connection with the Svecofennian volcanics (cf. Simonen *et al.*, 1978), being distributed alongside a NNE fracture zone (Gaál *et al.* 1978), and the Ni belt extends through Finland in the southern parts of the Bothnian Bay — Ladoga zone. This belt is characterized by row-like arranged nickeliferous clusters of ultrabasic-basic intrusions showing a syntectonic age of about 1 900 Ma. Their main place of deposition is regarded to be the sphere of influence of the NW-striking deep fracture, wrench fault.

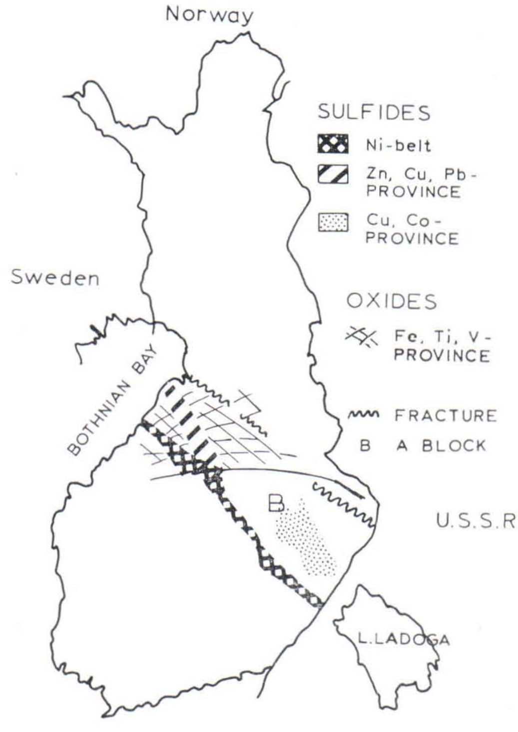

Fig. 1. Location of different metallogenic provinces within the Bothnian Bay-Lake Ladoga zone.

Since no opinions have yet been expressed on the association of oxidic ores and related basic magmatism within the Bothnian Bay—Lake Ladoga fracture zone, this will be discussed briefly, together with wider perspectives, in the following.

Deformation trend of the Baltic Shield

Fig. 2 includes a rough stratigraphic division of the Finnish bedrock. The oldest unit is the Archean basement which consists of a granitoid association of cratonized rocks and a greenstone belt association. The age of the former varies between 2 600—2 800 Ma, as determined from their zircon content by the U-Pb method, and 2 400—2 500 Ma in the domes of the intensively fractured Block B indicated in Fig. 1 (Simonen *et al.* 1978). Here, for instance, the Proterozoic sedimentary formations, the epicontinental and transgressive Jatulian group, associated with multiphase basic volcanism, have been deposited on the basement complex (starting more than 2 200 Ma ago, op. cit.). No Jatulian formations or Archean basement have been recognized in the area to the southwest of the »main sulphide belt» (Block B) where one finds Svecofennian pelitic and sandy geosynclinal sediments and volcanics (op. cit.). All the rocks mentioned above are penetrated by Middle Proterozoic ultra-basic and acid orogenic plutonic rocks, the youngest among which are the granitoid comp-lexes marked in this figure. The ages recorded for this area to the southwest of the main sulphide beltlie entirely in the Middle Proterozoic orogenic range (less than 1 900 Ma). Any possible older rocks presumably having been obliterated. The youngest, anorogenic, formations are the rapa-kivi granites in southern Finland, and the Jotnian red beds and dolerites.

Fig. 2, which also represents the fractures schematically is based mainly on morphology. The oldest fracture group is presumably the Early Proterozoic one to be seen on the Archean basement or »Jatulian continent». This includes: 1. strong multiphase and often curved fault traces, which partly coincide with the N-S trending Upper Archean greenstone belt (A-A on the map), 2. E-W fractures, which may continue from the basement complex westward into the Svecofennian area, where they would then appear as reactivated ring fractures around the granitoid complex of Central Finland, 3. longitudinal faults subparallel to 2, 4. wrench faults. The distribution of the products of the Jatulian magmatism and older greenstone belts suggests that the fractures mentioned here are of Jatulian and older origin. Their continuations in the Svecofennian formations point to the possibility that later reactivation was the rule there. It is also apparent that Block B (in Fig. 1), for instance, received its outlines during and before the Jatulian sedimentation (Laajoki 1977, Fig. 72; Talvitie 1978b).

The Svecofennian area in particular features of a further group, the Middle Proterozoic one (Fig. 2): 5. NNW transcurrent faults, 6. N-NNW fractures in Block B (Fig. 1), 7. NE fractures, 8. ring fractures around the granitoid complex of Central Finland, which are possibly reactivated and curved structures of earlier origin.

One general problem in the Svecofennian area is the estimation of the extent to which the pre-existing basement anisotropies have con-trolled the Middle Proterozoic fracture group described above. Recent seismicity indicates that they are still repeatedly being reactivated even today. NE fractures, especially, have been reactivated throughout the Late Proterozoic until the present time, the movements involving a rifting mechanism. This possibly denotes some kind of initial-state symmetry with the symmetry

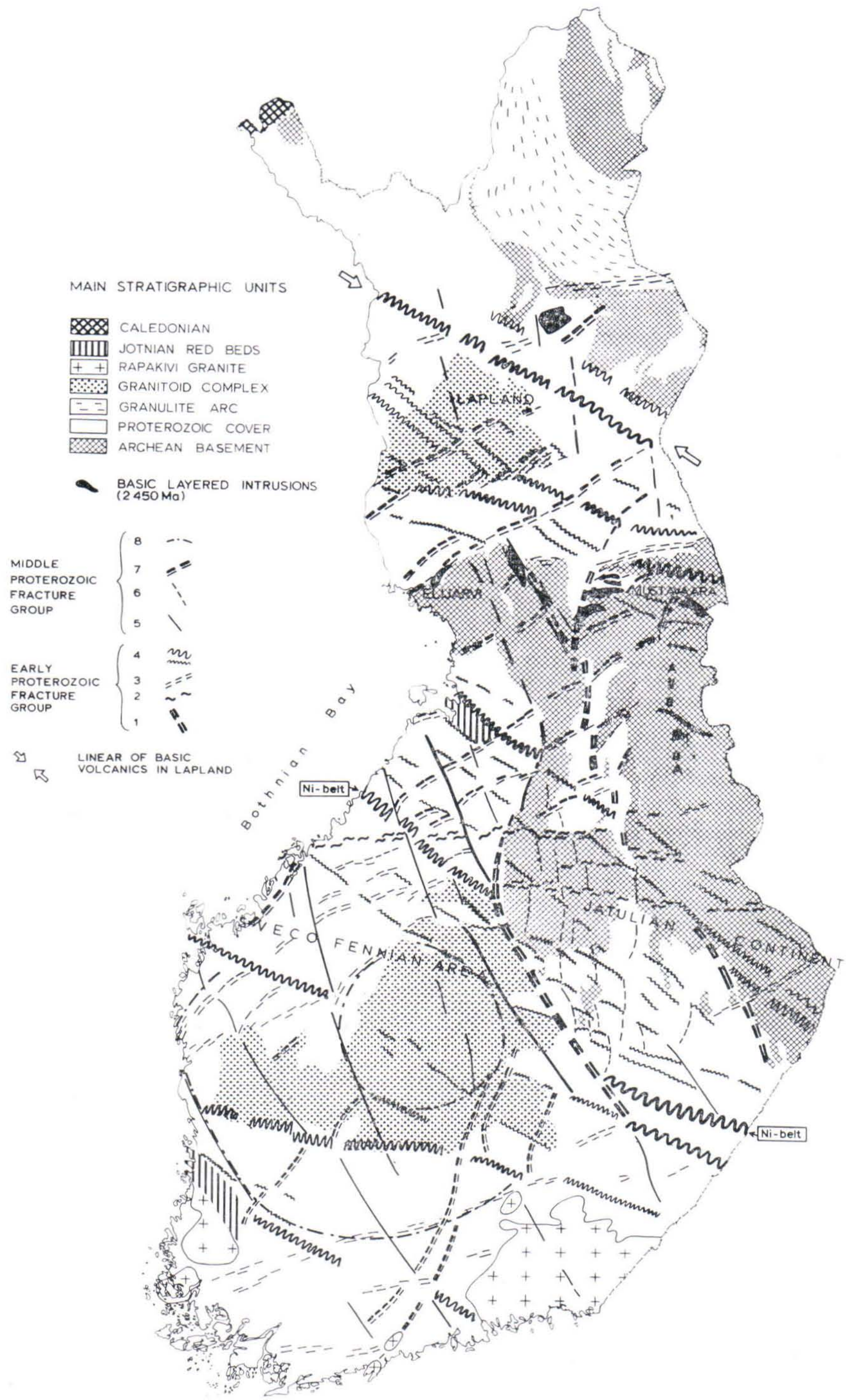

Fig. 2. Fracture tectonic outline.

of the Atlantic plate tectonics (cf. Paarma 1963, Valeyev *et al.* 1970, Surlyk 1977, Gorbunov *et al.* 1978, Talvitie 1978 a, Kröner 1977).

The faults of a transcurrent kind (NW, NNW; 3 and 5 in Fig. 2) are linear and show considerable long continuity, the NW linears appearing in their most closely spaced form, like a broad set or zone, in Lapland and the Bothnian Bay—Ladoga zone. One of these situated in Lapland controls in particular the distribution of the basic volcanics (arrow in Fig. 2) and another the above-mentioned Ni belt.

Fe-Ti ores, prospects and occurrences

Fig. 2 also depicts the known layered intrusions with an age of about 2 450 Ma (Kouvo 1976). Two economically important deposits, the Elijärvi Cr mine and the Mustavaara (Fe, Ti) V-mine, occur within these layered intrusions, both in Lapland. Taken individually, their locations are controlled by the above-mentioned Early Proterozoic fault group.

Fig. 3 shows the Otanmäki Mine and the gabbro and anorthosite-gabbro complexes with Fe, Ti and V indications as determined by Rautaruukki Oy. Often these rocks resemble layered intrusions, but closer examination (Fig. 4) has shown the primary magmatic differentiation conditions to be unstable, dynamic and autometasomatic, and not simply gravitational in situ. V-content in magnetic concentrate varies between 0.5—1.0 %, occasionally higher. Only the Otanmäki vanadium-magnetite-ilmenite ore is of economic value at present. The Otanmäki complex shows a Jatulian age (Table 1, Figs. 4, 5), which is clearly younger than that of the layered intrusions mentioned above. The age of the Otanmäki complex was determined from gabbro inside the complex (samples A 671-2) and from an alkalic granite cutting across the complex (A 100; cf. Marmo *et al.* 1966, Kouvo 1977).

The Kotalahti and Ylivieska basic intrusions inside the Ni belt give ages of 1 900 Ma (Kouvo 1976). In its NW-part, this belt also controls the distribution of Fe and Ti mineralization. The ages of the Ti and Fe intrusions are otherwise unknown, except, preliminary, for that of the Misi intrusion (Fig. 3) in Lapland (2 100 Ma, Kouvo, personal comm. 1978).

The prospects and occurrences in Lapland, (belt 1 in Fig. 3) and in southern Finland (belt 3, cf. Härme 1955) are located primarily in the sphere of influence of the NW wrench fault. The NW part of the broad Bothnian Bay—Lake Ladoga fracture zone forms an obvious tectonic unit, a Ti-Fe block, (controlled by the Early Proterozoic fault group). The southern border of this province is indicated as belt 2 in Fig. 3.

Fig. 6 shows prospect-type samples taken from glacial boulders by members of the public (amateurs) and submitted to Rautaruukki Oy. Their distribution fits well with that of the corresponding bedrock samples. This figure supports the view of the fracture and magma tectonics presented here, and complements it as far as the NW linears in Lapland and the Fe, Ti and V province in the Bothnian Bay—Lake Ladoga fracture zone are concerned, for instance. In other words, this provides new justifications for ore prospecting within these metallogenic units with the aid of fracture analysis. In general, the combination of crossing area of the NW and almost E-W fractures serves to control the Fe, Ti and V province (Fig. 1). When the Early

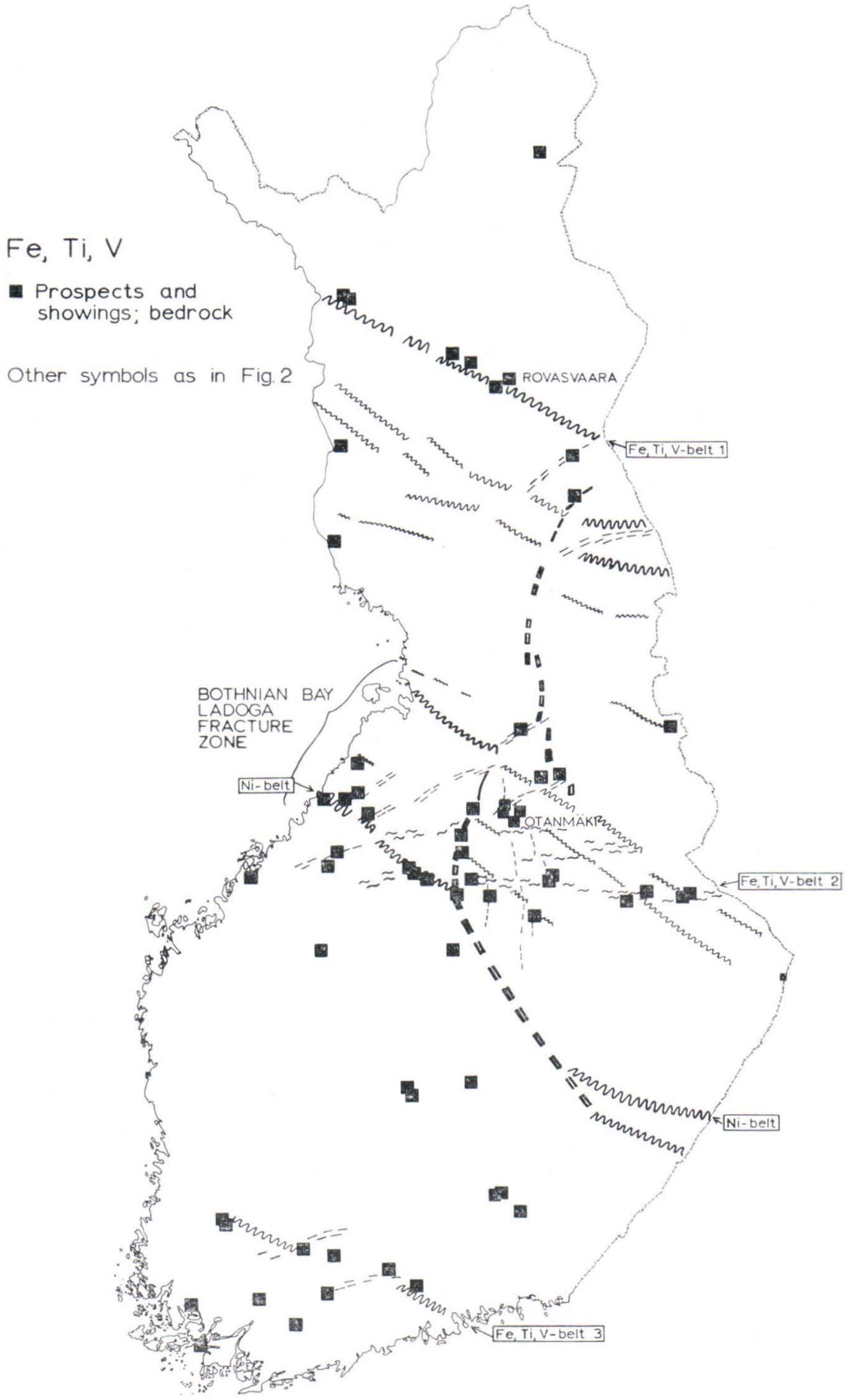

Fig. 3. Fe, Ti, V-mineralizations in bedrock in connection with basic intrusives.

Table 1.
U-Pb analytical data and radiometric ages for zircons from the Otanmäki ore field, according to O. Kouvo.

Sample No.	Zircon fraction ($g.cm^{-3}$/mesh size)	^{238}U ppm	Radiogenic ^{206}Pb, ppm	$^{206}Pb/^{204}Pb$ (measured)	Isotopic abundance relative to ^{206}Pb (= 100)			Radiometric ages. Ma		
					204	207	208	$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A671A	4.2—4.6/+100 light coloured ...	152.3	47.81	12 185	0.0030368	12.713	24.949	1 995±13	2 019±7	2 044±4
A671B	4.2—4.6/+100 dark coloured	312.1	96.37	10 093	0.006803	12.707	22.670	1 967±13	2 001±7	2 036±4
A671C	4.2—4.6/+100 from plagioclase	244.4	75.48	4 606	0.018569	12.854	26.611	1 967±13	2 000±7	2 034±3
A671D	4.2—4.6/—100 from plagioclase ..	239.2	73.56	2 733	0.032811	13.024	25.757	1 960±13	1 995±7	2 031±3
A671E	4.2—4.6/+125 from hornblende .	302.3	94.05	3 175	0.026863	13.001	23.453	1 980±13	2 009±7	2 039±3
A671F	+4.6/—	154.4	48.3	4 353	0.017469	12.913	24.756	1 990±13	2 017±7	2 045±4
A671G	4.0—4.2/—	915.3	275.43	4 054	0.0231753	12.827	20.710	1 923±10	1 971±6	2 022±3
A672A	+4.2/+100	416.4	129.3	12 131	0.0058735	12.690	30.070	1 976±13	2 005±7	2 035±3
A672B	4.0—4.2/+125	815.8	243.6	5 052	0.018366	12.786	23.356	1 911±13	1 966±7	2 025±2
A672C	4.0—4.2/+125	785.6	236.2	7 852	0.008401	12.611	21.858	1 922±10	1 969±5	2 020±4
A672D	+4.6/—	323.3	100.3	2 324	0.043896	13.225	31.889	1 975±11	2 006±6	2 039±6
A100A*	+4.2/+200	500.1	147.54	2 405	0.0273	12.698	15.256	1 891±13	1 941±9	1 995±14
A100B*	+4.2/—200	502.6	146.86	2 817	0.02123	12.587	15.214	1 875±13	1 931±13	1 991±20
A100C	3.8—4.2/—	1 040	259.8	723	0.127713	13.647	24.503	1 633±13	1 769±9	1 934±10
A100D	titanite	67.6	12.6	185	0.433185	17.093	81.098	1 261±12	1 482±32	1 814±53

*From Marmo *et al.*, 1966

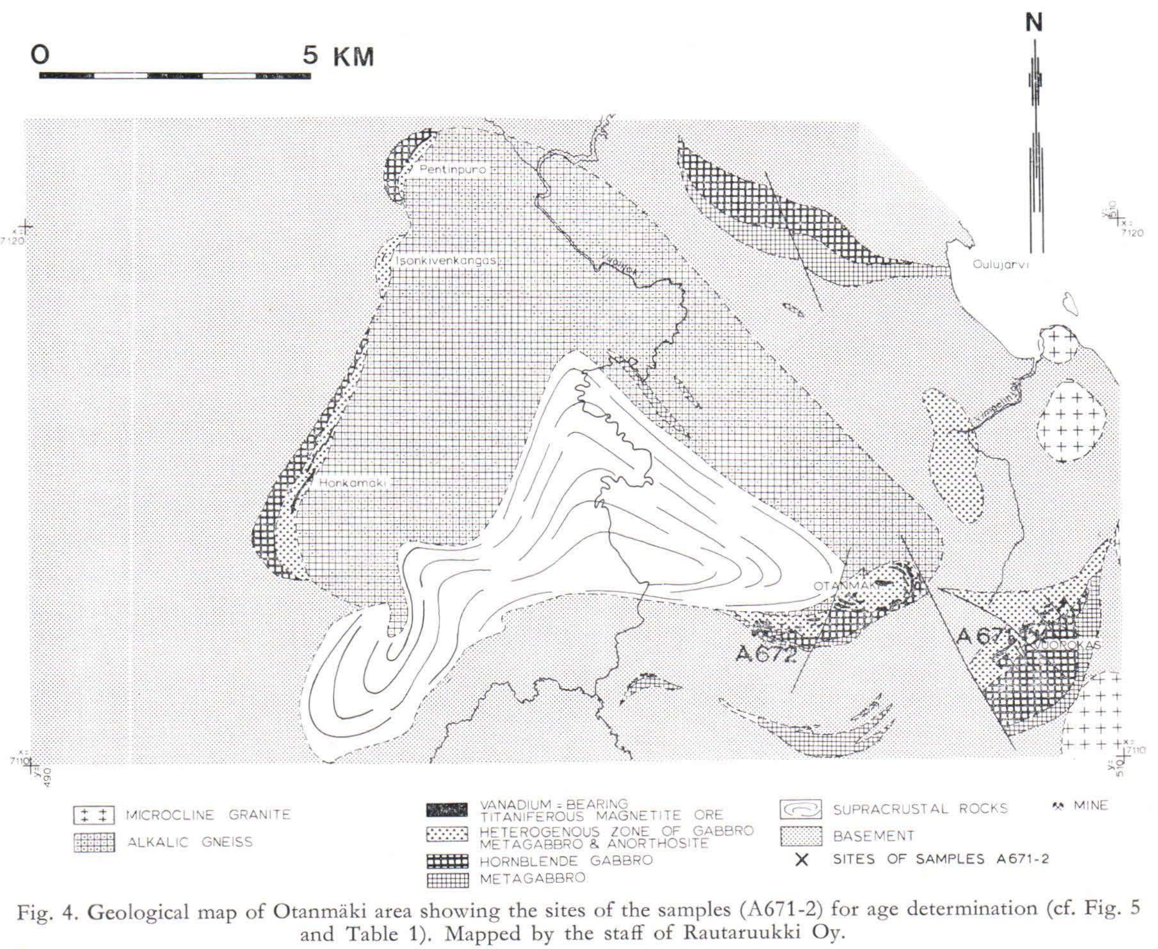

Fig. 4. Geological map of Otanmäki area showing the sites of the samples (A671-2) for age determination (cf. Fig. 5 and Table 1). Mapped by the staff of Rautaruukki Oy.

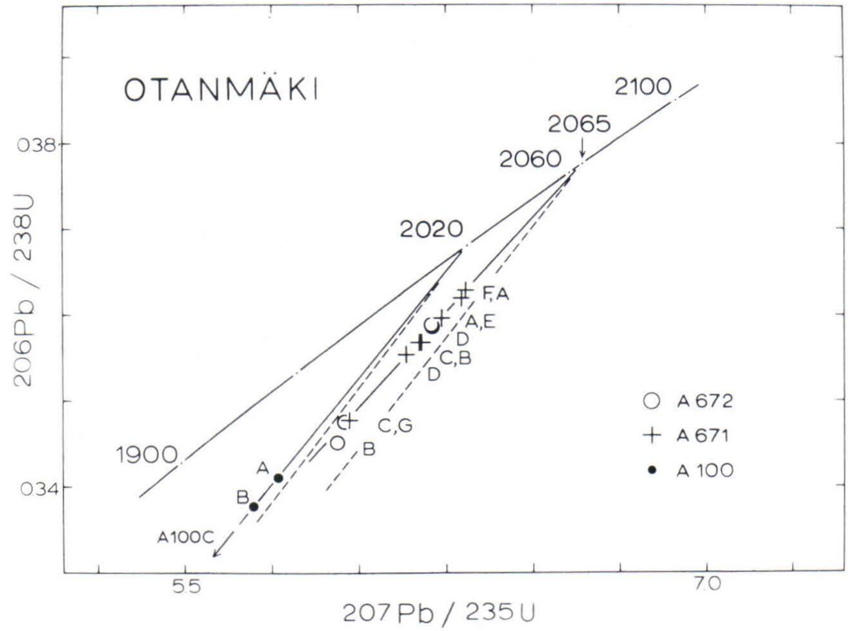

Fig. 5. Concordia diagram and U-Pb isotopic rations for zircon samples from two gabbro pegmatoids (A671-Vuorakas and A672-Otanmäki) and from the Honkamäki granite. The chords drawn are unbroken for the data defining an isochron and dashed line corresponding to the continuous diffusion trajectories of Wasserburg (1963). The parameters used for calculation are those given by Jaffey *et al.* (1971). The data are according to O. Kouvo.

Geochronometric results

»Data for three zircon fractions from the Honkamäki granite (Marmo *et al.*, 1966) and seven fractions from two mafic pegmatoid samples in Vuorokas and Otanmäki are listed in Table 1 and show distinct but different degrees of discordance on the concordia plot (Fig. 5). The Vuorokas—Otanmäki group (A671 and A672 respectively) form an excellent discordia array, indicative of a simple cogenetic suites of magmatically crystallized zircons, with an upper concordia intercept at 2065+—4 Ma and with the lower intercept of 883+—40 Ma according to York (1966). The most discordant fraction (A672B) was reanalyzed (A672C) and omitted. In case the first result is also included the values would be 2060+—4 Ma and 770+—78 Ma respectively. The three zircon fractions from the Honkamäki granite show an age 45 Ma younger than the mafic pegmatoids. The fitted discordia line (York 1966) has an upper intercept with concordia at 2020+—1.4 Ma and a lower intercept at 555+—3 Ma.

The significance of the U-Pb ages of zircons from the Honkamäki granite and the Vuorokas—Otanmäki mafic pegmatoids must be examined in the light of the slight disturbances of the U-Pb systems on both the zircon suites. This is indicated by the high lower intersection at concordia and the fact that the linear projections through the data-points do not fit the diffusion curve drawn according to Wasserburg (1963) but rather point to the episodic loss of lead shown by Wetherill (1965). This also is supported by the geologic evidence of later synorogenic metamorphism at 1800—1900 Ma. The both ages predate the Svecokarelian syntectonic events and the Honkamäki granite and the mafic pegmatoids are considered to have been emplaced close to the time of Jatulian volcanism 2 000—2 200 Ma ago.»

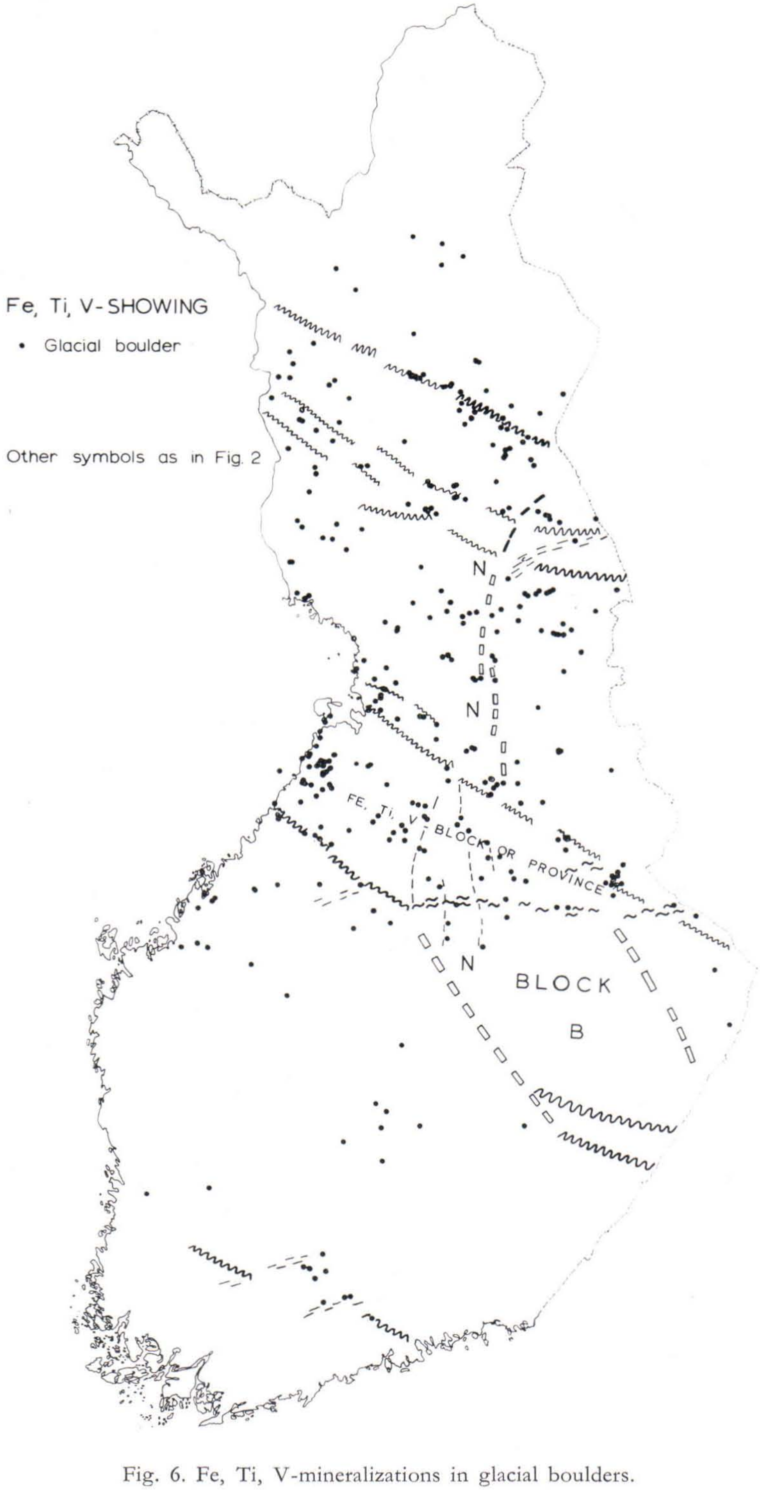

Fig. 6. Fe, Ti, V-mineralizations in glacial boulders.

Proterozoic layered intrusions are to be taken into account, attention must also be paid to the role of the fractures striking almost N-S (1 in Fig. 2) which run through the central part of Finland, for these include the ore-critical crossing area inside the Fe and Ti province (indicated by N-N in Fig. 6). In this case the problem lies in the extent to which the oldest traces of fractures have been obliterated by later deformations and may escape notice.

Discussion

When reviewing the Soviet geological maps for the area bordering on Finland one may conclude that the NW wrench lines in Lapland and the Bothnian Bay—Lake Ladoga fracture zone may have considerable SE-trending extensions (Archangelsky *et al.* 1937, Petrov 1970, Kratz 1974, Belyaev *et al.* 1977, Salop 1977, Kuosmanen *et al.* 1978). The estimated age for the primary development of fault movements along these fracture lines varies from Early to Middle Proterozoic on the Soviet side.

Thus, one could predict that under the upper Archean there was an already cratonized, rather stable, large continental platform which had begun to break into pieces during the Early Proterozoic according to a certain stress field and a certain symmetry, that is, according to the symmetry of the older fault group presented here. In this way blocks, and the fracture belts between them, would have been developing up to Middle Proterozoic times. The Early Proterozoic and later basic magmatism in Finland as described here would thus apparently follow an identical kind of tectonic-dynamic system. In other words, a certain deep fracture or block, once developed, may continue in the manner of an individual tectonic feature, and in this way the Bothnian Bay—Lake Ladoga fracture zone, for instance, can come to represent different metallogenic provinces and epochs. In respect to the basic magmatism and the Fe-Ti ores there seem to be three epochs in Finland: 2 450, 2 000—2 200 and 1 900 Ma (Early Proterozoic, Jatulian and Svecokarelidic basic magmatism). The problem of basic magmatism in southern Africa seems to be a parallel one, i. e., a multiphase event, according to recent radiometric datings of Great Dyke, etc. (Davies *et al.* 1970, Kröner 1977, Windley 1977). Since as this paper does not strictly speaking concern the situation in southern Finland, the Fe and Ti mineralizations and especially their possible association with the rapakivi granites are not dealt with.

A fracture tectonics analysis applied to the Finnish Fe-Ti ores is briefly described here, and we have seen that the system could be divided into different magma tectonic - metallogenic epochs. In the area of the Baltic shield there are known to be a number of basic-ultrabasic intrusions, containing Fe, Ti, V, Cr and Ni ores and indications of Pt. It would be of prime interest to collect corresponding material concerning the whole Baltic shield in order to obtain more detailed information on their metallogenic epochs and provinces, and on the related key questions concerning the magma tectonics of this area.

Acknowledgements

Thanks are due to Rautaruukki Oy for authorizing publication of the data on Fe-Ti mineralization. Construction of the tectonic analysis was assisted by financial support from the Finnish Academy. The authors also wish to thank Dr. Olavi Kouvo for highly fruitful discussions on the radioactive dating of the formations.

References

Archangelsky, A. D., Mikhailov, A. A., Fedynsky, V. V. & Lustich, E. N., 1937. Geological significance of gravitional anomalies in the SSSR. Bull. Akad. Nauk. SSSR, Ser. Geol. No. 4, 710—742.

Belyaev, K., Proskuryakov, V., Korsakova, M., Rabinovich, Y., Zagorodnyi, V., Bylinski, R. & Dolivo-Dobrovolski, A., 1977. The tectonic pattern of the eastern part of the Baltic Shield. Proc. Finnish-Soviet Symp. in Finland, 35—48.

Davies, R. D., Allsop, H. L., Erlank, A. J. & Manton, W. I., 1970. Sr-isotopic studies on various layered mafic intrusions, southern Africa. Geol. Soc., South Africa, Spec. publ. 1, 576—593.

Gaál, G., Parkkinen, J. & Talvitie, J., 1978. Tectonics of Ladoga—Bothnian Bay zone (in Finnish). *In* Laatokan—Perämeren -malmivyöhyke, symposium in Otaniemi 16. 2. 1978, Vuorimiesyhdistys ry., 20—35.

Gorbunov, G. I., Makijevskiy, S. I. & Nikolajeva, K. A., 1978. Metallogenitseskaja Zonalnost, svjazannaja s tektono- magmatitseskoi aktivizatsiei Baltijskogo stsita. Sov. Geol. No. 4, 15—26.

Härme, M., 1955. Kulonsuonmäen titaanirautamalmialueen geologiasta. Summary: On the geology of the titaniferous iron ore area of Kulonsuonmäki. Geol. Tutkimusl., Geotekn. julk. 59. 19 p.

Jaffey, A. H., Flynn, K. F., Glendenin, L. E., Bentley, W. C. & Essling, A. M., 1971. Precision measurement of half lives and specific activities of ^{235}U and ^{238}U. Phys. Rev. 4, 1889—1906.

Kahma, A., 1973. The main metallogenic features of Finland. Geol. Surv. Finland, Bull. 265. 29 p.

Kouvo, O., 1976. Kallioperämme kronostratigrafiasta. Suomen Geol. Seura ja Geologiliitto r.y., Koulutusmoniste 2.

— **, 1977.** The use of mafic pegmatoids in geochronometry. An abstract. 5th European Colloquium of Geochronology, Cosmochronology and Isotope geology, Pisa, Sept. 5—10, 1977. 1 p.

Kratz, K. O. (editor), 1974. Map of metamorphic zones in USSR. Geol. Ministry and Acad. Sci USSR.

Kröner, A., 1977. The Precambrian geotectonic evolution of Africa: Plate accretion versus plate destruction. Precambrian Res., 4, 163—213.

Kuosmanen, V., Paarma, H. & Tuominen, H., 1978. Laatokan—Perämeren vyöhyke osana suuremmasta systeemistä. *In* Laatokan—Perämeren malmivyöhyke, symposium in Otaniemi 16. 2. 1978. Vuorimiesyhdistys ry., 70—83.

Laajoki, K. & Saikkonen, R., 1977. On the geology and geochemistry of the Precambrian iron formations in Väyrylänkylä, South Puolanka area, Finland. Geol. Surv. Finland, Bull. 297. 187 p.

Marmo, V., Hoffren, V., Hytönen, K., Kallio, P., Lindholm, O. & Siivola, J., 1966. On the granites of Honkamäki and Otanmäki, Finland. Bull. Comm. géol. Finlande 221. 34 p.

Paarma, H., 1963. On the tectonic structure of the Finnish basement, especially in the light of geophysical maps. Fennia 89, 1, 33—36.

Petrov, A. I., 1970. Old faults in the eastern part of the Baltic Shield and movements along them. Doklady Akad. Nauk SSSR 191, 56—59.

Salop, L. J., 1977. Precambrian of the northern hemisphere. Elsevier Sci. Publ. Co. 176 p.

Simonen, A., Helovuori, O. & Kouvo, O., 1978. On the special features and age of the bedrock in Ladoga —Bothnian Bay zone (in Finnish). *In* Laatokan—Perämeren malmivyöhyke, symposium in Otaniemi 16. 2. 1978. Vuorimiesyhdistys ry., 10—19.

Surlyk, F., 1977. Mesozoic faulting in East Greenland. *In* »Fault tectonics in N.W. Europe», Geol. Mijnbouw, 56, 311—327.

Talvitie, J., 1978 a. Fracture tectonics and seismicity in Finland. Dept. Geoph. Univ. Oulu Contrib. 93. 14 p.

— **1978 b.** Deformation phases in the region of the Blue Road, Finland. Dept. Geoph. Univ. Oulu. Contrib. 95, 8 p.

Valeyev, R. N., Klubov, V. A. & Ostrovskiy, M. I., 1970. A comparative analysis of the conditions of formation and spatial distribution of the aulacogens on the Russian platform. Internat. Geol. Rev., 12 (4), 439—446.

Wasserburg, G. J., 1963. Diffusion processes in lead-uranium systems. J. Geophys. Res. 68, 4823—4846.

Wetherill, G. W., 1956. Discordant uranium-lead ages I. Am. Geophys. Union Trans. 37, 320—326.

Windley, B. F., 1977. The evolving continents. John Wiley and Sons Ltd. 385 p.

York, D., 1966. Least-squares fitting of a straight line. Can. J. Phys. 44, 1079—1086.

DEGASSING OF SEDIMENTARY ROCKS DURING METAMORPHISM AND A METAMORPHOGENIC MINERALIZATION

V. A. Tenjakov and Sv. A. Sidorenko

New data gathered during the past few years have enhanced our knowledge of Precambrian sedimentary geology and given rise to a fundamentally new concept of the dominant and decisive role played by exogenic, biogenic and metamorphic processes in the formation of the Earth's crust (Sidorenko 1975).

The development of this new approach to understanding Earth's history entails an essentially new consideration of the main processes producing most of the rock-types, at least in the upper sialic portion of the Earth's crust. Inevitably the cause-effect-time relationship between the processes responsible for the formation and modification of our planethas had to be reconsidered as have other important processes and phenomena in the Earth's crust, particularly the role of gases escaped from the Earth's interior.

It has been established that the carboniferous primary sedimentary rocks containing as a constant abundant gaseous hydrocarbons, including those of oil, were rather common in the Precambrian. In 1970 this evidence was used to formulate a new notion of »hydrocarbon breathing» in the oldest primary sedimentary metamorphic rocks (A. Sidorenko and Sv. Sidorenko 1970), and to make a conclusion that this hydrocarbon source affected the formation and balance of hydrocarbons accumulated in the sedimentary mantle of the continents.

Simultaneous studies of Precambrian metamorphic clayey-carbonate rocks supplied data that allowed us to conclude that hydrocarbon acid had been released from the rocks of this class undergoing metamorphism (A. Sidorenko *et al.* 1973). This conclusion sponsored the concept of hydrocarbon breathing of sedimentary metamorphic rocks and allowed us to suggest that this process influenced the evolution of CO_2 in the Precambrian atmosphere and hydrosphere as well as in some oregenerating processes associated with metamorphism of primary sedimentary rocks.

The synthesis of these new ideas, taking into account huge masses of water and gases of atmospheric (Sinitsyn 1972), biogenic (Volynets 1975) and other origin (Sokolov 1971) liberated by metamorphism, as well as ore elements mobilised by these agencies, enabled us to introduce the scientific notion of the process of »Liquid-gaseous-ore breathing» and its influence not only on the composition and development of the Earth's atmosphere and hydrosphere, but also on Precambrian metallogeny (Tenjakov 1975).

It became obvious how right and how farsighted Vernadsky had been when he considered sedimentary-metamorphic processes as a powerful and exceptionally extensive inversion mechanism governing and digesting many important components of the Earth's outer shell during the cycles of development in the Earth's crust.

Further analysis and the application of this concept to a few gases (A. Sidorenko, Sv. Sidorenko and V. Tenjakov 1978) allowed a relationship to be established for the first time on a global scale between sedimentary metamorphic processes and the liberation of certain elements such as gases and gaseous compounds from the Earth's interior. We have succeeded in showing that these processes act as a kind of breathing in the Earth's crust: first, the forces in the exogenic zones release many important components from the atmosphere, hydrosphere and biosphere: these combine and form new compounds, and then the metamorphism in the primary sedimentary rocks creats favourable conditions for their re-release and return into the surface zone. It is this liberation of H_2O, CO_2, N_2, hydrocarbons and some other gases (H_2) and fluids that probably leads to mobilisation of most ore and non-ore components and their redistribution and accumulations as a result of processes of metamorphogenic ore formation and mineralisation related to the period of ultrametamorphism and ultrametamorphic magma formation.

In the present paper we shall once again discuss the possible geochemical and metallogenic role of CO_2 released from metamor-

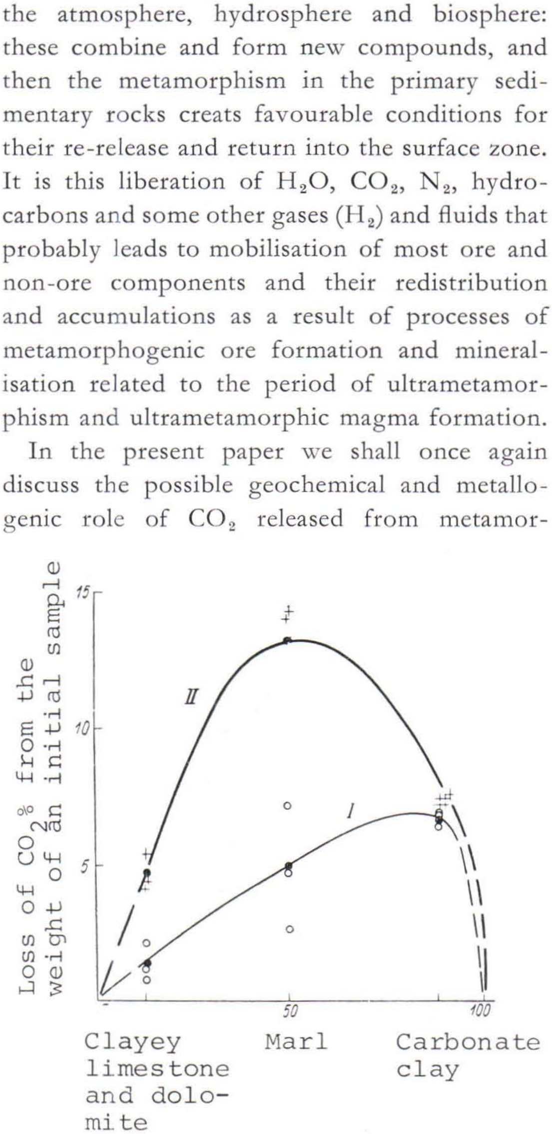

Fig. 1. Relationship between released hydrocarbon and the content of clayey component in a range from limestone (I) to dolomite (II).

phosed clayey-carbonate rocks in or near zones of metamorphic rocks and in the surficial exogenic zone which, as we assume, is likely to accumulate a significant amount of this »deconserved» carbon dioxide acid from sedimentary successions. Some time ago, in our paper on carbon dioxide (A. Sidorenko *et al.* 1973), we briefly described the exceptionally valuable data collected by Tugarinov and Naumov (1973) on gas fluid inclusions in metamorphic and hydrothermal minerals. Carbon dioxide appeared to be most common and typical component in hydrothermal solutions. Observations at low temperature often show that the inclusions contain CO_2 in the liquid from; a fact that points to CO_2 content exceeding 80—100 g/l.

Current data available from numerous observations confirm that CO_2 is commonly present in hydrothermal minerals (Tugarinov and Naumov 1973).

All the information so far gathered has allowed the authors to reach the extremely important conclusion that hydrothermal solutions frequently consist of CO_2-saturated water and that these solutions can transfer ore components in the form of carbonate complexes (Tugarinov and Naumov 1973).

Thus, it has been revealed that decarbonization of buried clayey-carbonate Precambrian sediments was of prime importance in the origin and maturation of metamorphogenic fluids and associated solutions as well as in the mobilization and accumulation of the elements. This idea of the connection between hydrothermal mineralization and regional metamorphism facilitated by a mechanism of CO_2 conservation is potentially attractive for further studies. If we take into consideration the fact that water carbon dioxide components in hydrothermal solutions, as jugded by inclusions, characterize the hydrothermal solutions formed at a distance from the metamorphic stratum, then this concept is basically important for understanding the origin and ore load source feeding nonmagmatogenic hydrotherms in general.

We previously described in some detail the influence of the carbon-dioxide breathing of metamorphic sedimentary rocks on the composition and evolution of the atmosphere around our planet (A. Sidorenko *et al.* 1973). Let me remind you that an approximate qualitative evaluation of this phenomenon shows that during individual periods of metamorphism (if we assume that they totalled at least ten) the atmosphere could absorb about 0.14×10^{17} tons Co_2. But even in this case the contentration of carbon dioxide in both the atmosphere and the oceans increased more than 100 times.

Naturally it was possible to assume that so many significant variations in the atmospheric composition affected the intensity of weathering, composition and other features of Precambrian sediment. It was also necessary to find out what kind of effect a periodically increasing partial pressure of CO_2 produced in the Precambrian atmosphere. The answer to this question is surely significant, and not only in terms of our concept of carbon dioxide breathing in the Earth's crust. The idea is commonly held that the Piecambrian atmosphere differed in some principal features from the later, post-Precambrian one. First of all, it was free or contained only small amounts of free oxygen but was rich in carbon dioxide. The qualitative analysis of these supposed features of the Precambrian atmosphere led a few scientist to fashion (hypothetical) conclusions on the specific nature of the weathering crust and of the sedimentary rocks that were formed during those epochs. Not infrequently it was the tentative high partial pressure of CO_2 in the Precambrian atmosphere that was supposed to be associated with acidic or even highly acidic rain water and low pH value (2.5—1.5 and less). Consequently some scientists assumed that the trend of weathering and sedimentation, and the development of global hypergene-chemogene mineralization process was different in the Precambrian from the Phanerozoic.

With the accumulation of our knowledge of the geology, lithology and geochemistry of the concrete products of Precambrian weathering (»Dokembrijskie kory vyvetrivanija» 1975) and of the oldest sedimentary rocks on Earth (»Problemy osadotshnoi geologii dokembrija» 1975; »Korreljatsija dokembrija» 1975), it has become clear that even the earliest Precambrian weathering did not produce rocks that were either exotic or basically different from later types of rocks, and that trends of exogenesis and lithogenesis were basically similar to those of the surface processes in later epochs (A. Sidorenko 1975; Tenjakov 1975).

Thus, although there is every reason to believe that the Precambrian atmosphere was really different from the Post-Precambrian one, at least in CO_2 content (varying during individual periods), we have no data on the regularity of specific features characterizing the products that originated in such an unusual environment and which would reveal the dissimilarity in the gas composition of the ancient atmosphere. In other words, we are unable to establish the »acidifying» effect of meteoric water or highly acidic surface water on the progress of hypergene processes and even more so on the trend of the main reactions of the surface processes.

In this connection we analysed the relationship between possible (and assumed) concentrations of CO_2 in the ancient atmosphere and the pH in corresponding atmospheric water (Tenjakov and Kopejkin 1977). The results show that the considerably acidic atmospheric water with pH about 3.4 can be attributed to carbon dioxide gas only if CO_2 pressure is very high, which is highly improbable, that is 1 000 times and more than the present level. Thus, the hypothesis of super acidic rains and super acidic surface water due to atmospheric CO_2 should be rejected (Table 1).

On the basis of data obtained from many experiments, including those of Pedro (1971), it may be considered that a periodically significantly enhanced partial pressure of CO_2 in the

Table 1.
pH values of atmospheric waters due to atmospheric carbon dioxide.

Atmosph.	T °C				
	5	25	50	75	90
$10^{-3.5}$	5.595	5.66	5.75	5.85	5.905
10^{-3}	5.345	5.41	5.5	5.6	5.655
10^{-2}	4.845	4.91	5.0	5.1	5.155
10^{-1}	4.345	4.41	4.5	4.6	4.655
1	3.845	3.91	4.0	4.1	4.155
10	3.345	3.41	3.5	3.6	3.655

Precambrian atmosphere only intensified the weathering processes speeding up the maturation of rocks and the growth of weathered rock successions (including lateritic products) (Tenjakov 1975). This CO_2 rise did not alter the general trends of known processes, such as hypergene alteration of alumo-silicate rocks. We should not disregard the fact that as a rule the alumo-silicate rocks act as an exceptionally resistant alkaline buffer that constantly and effectively withstands increasing concentration of hydrogen ions in surface water, and suppresses the activity of global high-acidic media coming in contact. Thus if we assume that the pH in the Precambrian atmospheric precipitation dropped as low as 4.5 to 4.0 on average (owing to the rise in CO_2 pressure), the surface water pH could have been maintained within this range; it would inevitably have been higher (on average) on account of constantly progressing mobilization of K^+, Na^+, Ca^{2+}, Mg^{2+} supplied from land rocks. This assumption is well supported by pH values in atmospheric water (average 5.7) (Bass Bekking *et al.* 1963) and in current values recorded in water from rivers, streams, soil and underground (average 6,8). It is known (Garrels and Mackenzie 1974) that, while weathering, a continent inhales rather much CO_2 from the atmosphere, thus receiving its acid stock.

The normal progress of weathering in Precambrian is disobeyed by the periodic process of laterite and bauxite formation throughout practically the whole period of the geologic history of the Earth from as far back as 3 200 Ma (Tenjakov 1975; Sidorenko and Tenjakov 1976). The geochemistry of bauxites leaves (Tenjakov 1972) no doubt that lateritization has been the main and only process producing a bauxite matter during the geological history of the Earth, and that if its pH was equal to 4.0 its physico-chemical conditions (Tenjakov 1973; Bronevoi and Tenjakov 1976) did not promote this development. Consequently we may assume that even of humid lithogenesis was most intense and contrasting in the Precambrian and followed the trends displayed in later epochs the conditions of its realization in the Precambrian, including alkaline-acid media, did not depend on the assumed high concentration of CO_2 in the ancient atmosphere.

The data available require that some theories and hypotheses of the nature of Precambrian hypergene processes be revised, especially those of global hypergene and chemogene ore formation epochs during the early stages of the geochemical history of the Earth and which were caused by high CO_2 in the atmosphere.

Thus, the widespread metamorphism that affected the primaiily sedimentary Precambrian rocks is responsible for mobilization and extraction of exceptionally large masses of not only water but also carbon dioxide, various hydrocarbons and other gases as well as many ore and non-ore minerals.

It is impossible to go on ignoring the significance of this breathing of the Precambrian metamorphic successions.

On the basis of the widespread distribution of Precambrian rocks, our data on possible amounts of carbon dioxide generated by metamorphism in clayey-carbonate rocks, and the migration of hydrocarbons from Precambrian strata into overlying loose rocks, we can conclude that the breathing of the metamorphic successions must influence both the evolution and composition of the Earth's atmosphere and hydrosphere during periods of geological time. It is evident that CO_2 masses periodically expelled into the atmosphere affected biological activity and hence the biological productivity of living matter, and that this later pronoted the accumulation of hydrocarbons in the primarily sedimentary Precambrian rocks and finally brought about their transformation into hydrocarbon rock-types penetrating the sedimentary rocks of the overlying loose complex.

References

Bass Bekking, L. G. M., Kaplan, I. P. & Mur, D., 1963. Predely kolebanij pH i okislitel'no-vosstanovitel' nyh potencialov prirodnyh sred. *In* Geohimija litogeneza. Moskva, II.

Bronevoi, V. A. & Tenjakov, V. A., 1976. Osnovnye fizikohimičeskie parametry lateritnogo processa (primenitel'no k gibbsitovym boksitam) Dokl. Akad. Nauk SSSR 228, 192—194.

Dokembrijskie kory vyvetrivanija, 1975. Moskva, VIMS.

Garrels, R. & Mackenzie, F., 1974. Evoljucija osadočnyh porod. Moskva, Mir.

Korreljacija dokembrija, 1975. Tezicy dokladov Mezdynarodnoga simpoziuma. Moskva.

Pedro, Ž., 1971. Eksperimental'nye issledovanija geohimiceskogo vyvetrivanija kristalliceskih porod. Moskva, Mir.

Problemy osadočnoi geologii dokembrija, 1975. Moskva, Nedra.

Sidorenko. A. V., 1975. Osadočnaja geologija dokembrija i ee značenie gnja poznanija dopaleozoijskoj istorii Zemli. Sov. Geol. 1975 (2).

Sidorenko, A. V. & Sidorenko, Sv. A., 1970. Ob »uglevodorodnom dyhanii» dokembrijskih grafitsodepžaščih tolšč. Dokl. Akad. Nauk SSSR 192, 184—187.

Sidorenko, A. V. & Tenjakov, V. A., 1976. Boksitoobräzovanie i geologičeskoj istorii zemlii i »princip shodstva» ekzogennyh processov v dokembrii i fanerotzoe. Dokl. Akad. Nauk SSSR 226, 1150—1153.

Sidorenko, A. V., Rozen, O. M., Tenjakov, V. A. & Gimmel'farb, G. B., 1973. Metamorfizm osadočnyh togšč i »uglekishoe dynanie» zemnoj kory. Sov. Geol. 1973 (5), 3—11.

Sidorenko, A. V., Tenjakov, V. A., Sidorenko, Sv. A., 1978. Osadočno-metamorfičeskie processy i »gazovoe dyhanie» zemnoj kory. Dokl. Akad. Nauk SSSR 238, 705—708.

Sinitsyn, V. M., 1972. Sial. Leningrad, Nedra.

Sokolov, V. A., 1971. Geohimija prirodnyh gazov. Moskva, Nedra.

Tenjakov, V. A., 1972. Problema istočnika i sposoba formirovanija večšestva boksitov. *In* Problemy genezisa boksitov. Moskva, Nauka.

—, **1973.** Geochemistry and mechanism of principal act laterite process. I. C. Sob. A, 3-a Congress Intern., Nice.

—, **1975 a.** Boksitoobrazovanie v geologičeskoj istorii zemli i problema boksitov dokembrija In: Dokembrijskie koru vyvetrivanija. Moskva, VIMS.

—, **1975 b.** O nekotoryh diskussionnyh problemah eksogenno-metamorfogennoj geologii dokembrija. In: Problemy osadočnoj geologii dokembrija. Moskva, Nedra.

Tenjakov, V. A. & Kopejkin, V. A., 1977. O probleme kislotnosti atmosfernyh i poverhnostnyh vod v dokembrii. Dokl. Akad. Nauk. SSSR 234, 429—342.

Tugarinov, A. I. & Naumov. V. B., 1973. Fizikohimičeskie parametry gibrotermal'nogo mineraloobrazovanija. I. Meždunar. geohim. kongress., Moskva, Nauka.

Volynec, V. F., 1975. Krugovorot azota i processy preobrazovanija organiceskogo vecsestva. In: Problemy osadočnoj geologii dokembrija, vyp. 4, kn. 2, 189—194. Moskva, Nedra.

A PROGRESS REVIEW OF THE HAMERSLEY BASIN OF WESTERN AUSTRALIA

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Introduction

The term 'Hamersley Basin' was first used by Trendall (1968) to refer to the depositional basin of the Hamersley Group, a 2.5 km thick Proterozoic sedimentary and volcanic sequence with abundant banded iron-formation (BIF), which crops out over an area of some 60 000 km^2 in the northwestern part of Western Australia. The scope of the term was later implicitly (Trendall and Blockley 1970), and then explicitly (Trendall 1973, 1975) extended to apply also to the depositional basin of the underlying Fortescue Group and the overlying Wyloo Group. The basin was thus regarded as a continuing tectonic entity which was infilled by these three successive groups, together included by Halligan and Daniels (1964) within the Mount Bruce Supergroup.

This concept of the Hamersley Basin, which is to some extent modified in this paper, grew out of a programme of regional mapping at a scale of 1:250 000 by the Geological Survey of Western Australia, to which a total of 14 geologists contributed. The programme began in 1956, and reached peak intensity in 1962—64. The results were summarised by MacLeod (1966), who systematically described the stratigraphy of the basin but did not formally name it.

Before the initiation of the programme the rocks of the Hamersley Basin were virtually unknown in the international geological literature. It is true that Maitland (1909) and Talbot (1920) had long before made known the existence of Proterozoic (»Nullagine») sedimentary rocks in the area, and also that Miles (1942) had given a petrographic description of the banded iron-formations near Wittenoom; but it was only in the late 1960s that the Hamersley Basin achieved the widespread familiarity that is now so evidently appropriate to the excellence of its exposure, extent, and standard of preservation.

At present the situation is almost reversed, with sufficient published accounts existing to make it difficult for those outside Western Australia to know what the best single source of information is, which aspects of earlier accounts may have been superseded, and whether some accounts merely summarise information published elsewhere. The purposes of this further brief description of the geology of the Hamersley Basin are:

1. to provide a brief guide to the existing literature, particularly for geologists working on the Baltic Shield who may not have ready access to Australian publications,
2. to provide a brief summary description of the basin which incorporates amendments of some concepts expressed in earlier descriptions, and
3. to refer to various current studies of the basin which may lead to resolution of various remaining problems.

Documentation of the basin

The best representation of the Hamersley Basin for those concerned with local detail is given by the 1:250 000 scale maps of the Geological Survey of Western Australia. Sixteen sheets, each 1° of latitude by 1°30' of longitude cover the outcrop area of the basin between latitudes 20° and 24° South and longitudes 115° 30' and 121°30' East. Sheet names and coverage are shown in publication catalogues and Annual Reports of the Geological Survey. MacLeod (1966) provided an excellent synthesis of the results of the 1:250 000 mapping and his Plate 2, a map at a scale 1:500 000, is still the best geological map covering most of the basin in a single sheet. The earlier paper by MacLeod and others (1963) was completely superseded by the 1966 bulletin, but retains documentary significance as being the first formal definition of the basin stratigraphy.

The later bulletin of Trendall and Blockley (1970) was restricted to a detailed account of the iron-formations of the Hamersley Group, in which respect it superseded MacLeod's (1966) earlier work. The brief account of the geology of the Hamersley Basin in this paper essentially summarises the information in these two basic references. Reference to later publications modifying and augmenting specialised aspects of the geology of the basin are mainly included in the text below, but specific reference may be useful here to three papers which are practical field guides to visitors wishing to visit the basin independently. These are a paper by Trendall (1966) which was intended as a field guide to the outstanding exposures of the Dales Gorge Member at Dales Gorge, another (Trendall 1968 b) similar guide to the superb exposures of the Joffre Member in the gorges south of Wittenoom, and an excursion guide prepared for the 1976 International Geological Congress, which with the aid of route notes available from the Geological Survey permit independent repetition of a 6-day excursion covering all aspects of the basin's development (Trendall 1976 c).

It may also be useful to note here that several reviews of Hamersley Basin geology (Trendall 1975 a, 1975 b) contain no primary information not included in other publications referred to in the following texts.

Summary description of the basin

Location, area, shape, and immediate geological environment

The location of the Hamersley Basin is shown in Figure 1. The present outcrop area of the three depositional groups initially regarded as representing the contents of the basin (see following heading) is about 125 000 km². Of this area the Fortescue Group crops out over about 40 000 km², the Hamersley Group some 60 000 km², and the Wyloo Group about 25 000 km². The depositional area of a basin limited to the Fortescue, Hamersley, and recently erected Turee Creek Group still seems likely to have been at least 150 000 km² and to have been ovoid in shape, with a long axis running roughly west-northwest. However, this estimate would need to be modified with some regional interpretation as discussed further below, under the heading 'Regional relationships'.

On the northern side of the basin, the basal deposits of the Fortescue Group were laid down with marked unconformity over the Archaean rocks of the Pilbara Block. However, as noted under the heading 'Age' below, it is not yet established what order of time interval is represented by this unconformity. Along the eastern and the eastern part of the southern sides of their outcrop area the rocks of the basin are themselves unconformably overlain by younger sedimentary rocks, either Phanerozoic or Middle Proterozoic, except where they unconformably overlie the Archean rocks of the Sylvania Dome.

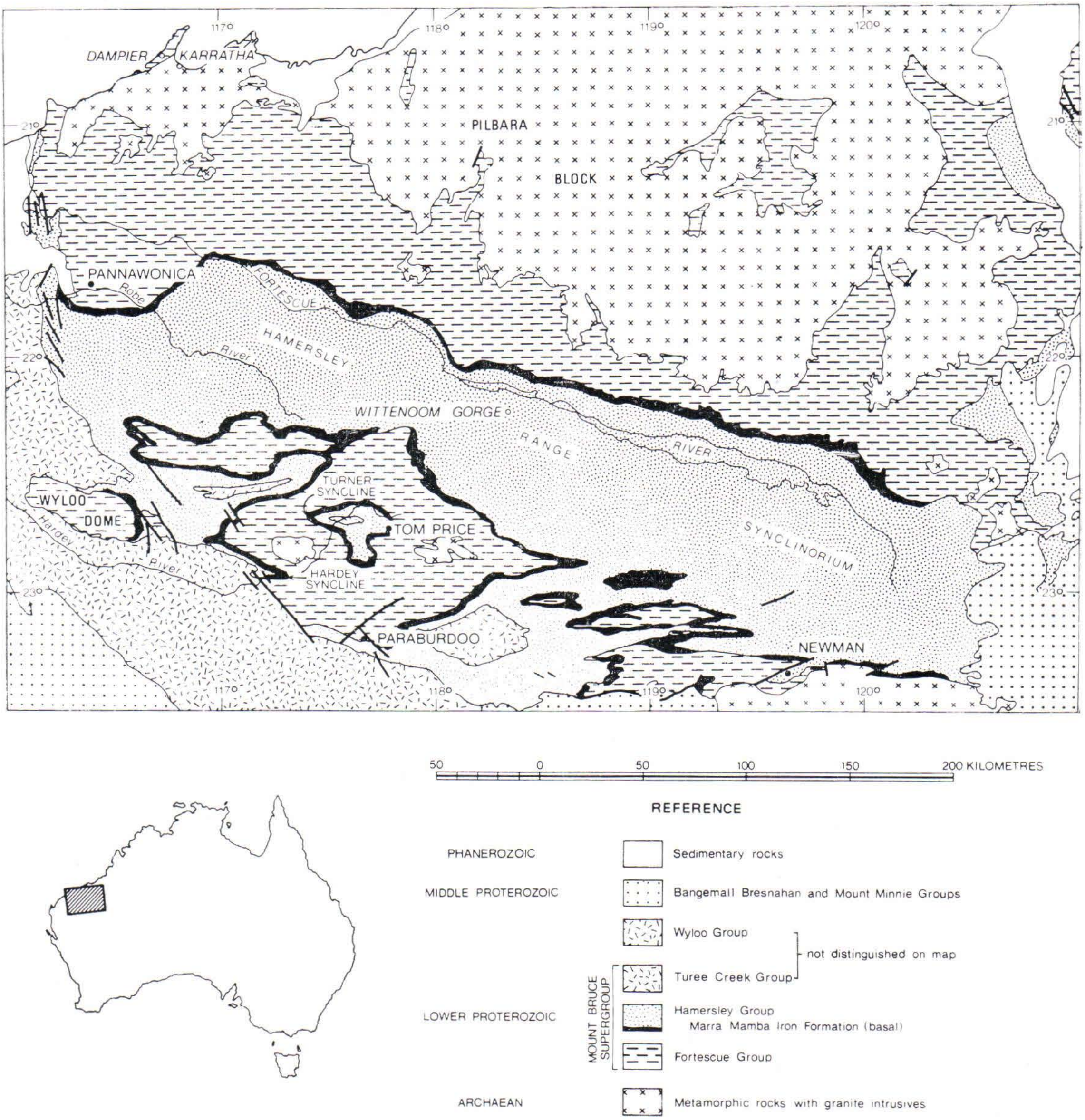

Fig. 1. Geological map of the Hamersley Basin area.

To the west and southwest the rocks of the Hamersley Basin are limited by the basal unconformity of the redefined Wyloo Group, which is now, as discussed under the following heading, excluded from the basin. This complex boundary is still incompletely studied.

Stratigraphy, and its tectonic implication

Fortescue Group

Stratigraphic nomenclature within the Fortescue Group is complex and in need of revision, and need not be detailed in this paper; readers concerned with details are referred to reviews by Trendall (1975 a, 1976 a).

The bulk of the material of the group is directly or indirectly volcanogenic, and dominantly mafic. Only in the extreme northeastern outcrop area of the Fortescue Group are acid lavas abundant (Hickman 1975), and these seem to be associated with an unusual local volcanic event. Extensive flows of dark fine-grained, massive, amygdaloidal or vesicular basalt, locally pillowed, are separated by equally extensive sheets of tuff of major derivation, much of which shows evidence of accumulation in shallow water. However, felsic and acid lavas also occur, many tuffs have an admixture of terrigenous (often granitic) debris, and shallow-water and stromatolitic carbonate intercalations are locally well developed, particularly in the northeast. The lateral continuity of the lowest stratigraphic units is generally restricted, and their thickness varies sharply along strike. As the stratigraphy is followed upwards, individual units tend to be more easily traceable over large strike distances, with less lateral variation in thickness.

In detail, the most complete stratigraphic sequences were deposited over greenstone belts in the underlying Archaean, with the lower stratigraphic units progressively pinching out over the granitic plutons. Hickman and Lipple (1975) interpret this as indicating both that the granitic terrain was higher than that of greenstone areas at the start of deposition, and that preferential downwarping of greenstone areas took place during deposition. On a broader scale the greatest thickness of the Fortescue Group developed in the central part of the basin, around the Turner Syncline (Fig. 1); here the total thickness is about 3 700 m. That this area was subject to greatest subsidence during Fortescue Group time is supported by a higher proportion of pillowed (subaqueous) to massive (subaerial) basalts.

A broad generalisation derived from Fortescue Group stratigraphy is that long periods of lava extrusion alternated with long periods of volcanoclastic accumulation, rather than there being a continual association of the two. The uppermost formation of the Fortescue Group, the Jeerinah Formation, is exceptional in that its mafic volcanic affinity is not demonstrable, and its lithology is closely similar to shale of the Hamersley Group. Like many of these it is, where fresh, black due to the presence of free carbon and commonly pyritic and cherty. It is also exceptional in that it is laterally continuous over the whole outcrop area of the Fortescue Group. Although thin lavas are intercalated locally even at this level the volcanicity which contributed directly to this first phase of basin development seems to have become stifled by the thick blanket of resultant material.

Thick sills of both basic and acid material were injected at various levels of the Fortescue Group. They are presumed to represent cogenetic volcanic material which failed completely to penetrate this blanket.

Hamersley Group

In contrast to the Fortescue Group the stratigraphic subdivisions and nomenclature of the Hamersley Group are clearly defined and demonstrably valid and continuous over the entire 60 000 km^2 of its outcrop area; the only exceptions to this generalisation are in the northeasternmost part of the basin and possibly in the equivocal Wyloo Dome area, referred to later in this paper.

The characteristic feature of the Hamersley Group is the presence of thick and abundant iron-formations. This can be seen from Figure 2, in which the stratigraphic subdivisions established by MacLeod and others (1963) and modified by Trendall and Blockley (1970) are displayed. The formation thicknesses represented in Figure 2 are those of the central part of the basin. Isopachs are available only for the Dales Gorge Member of the Brockman Iron Formation (Fig. 3) These show maximum thicknesses along a west-northwest trending trough in the

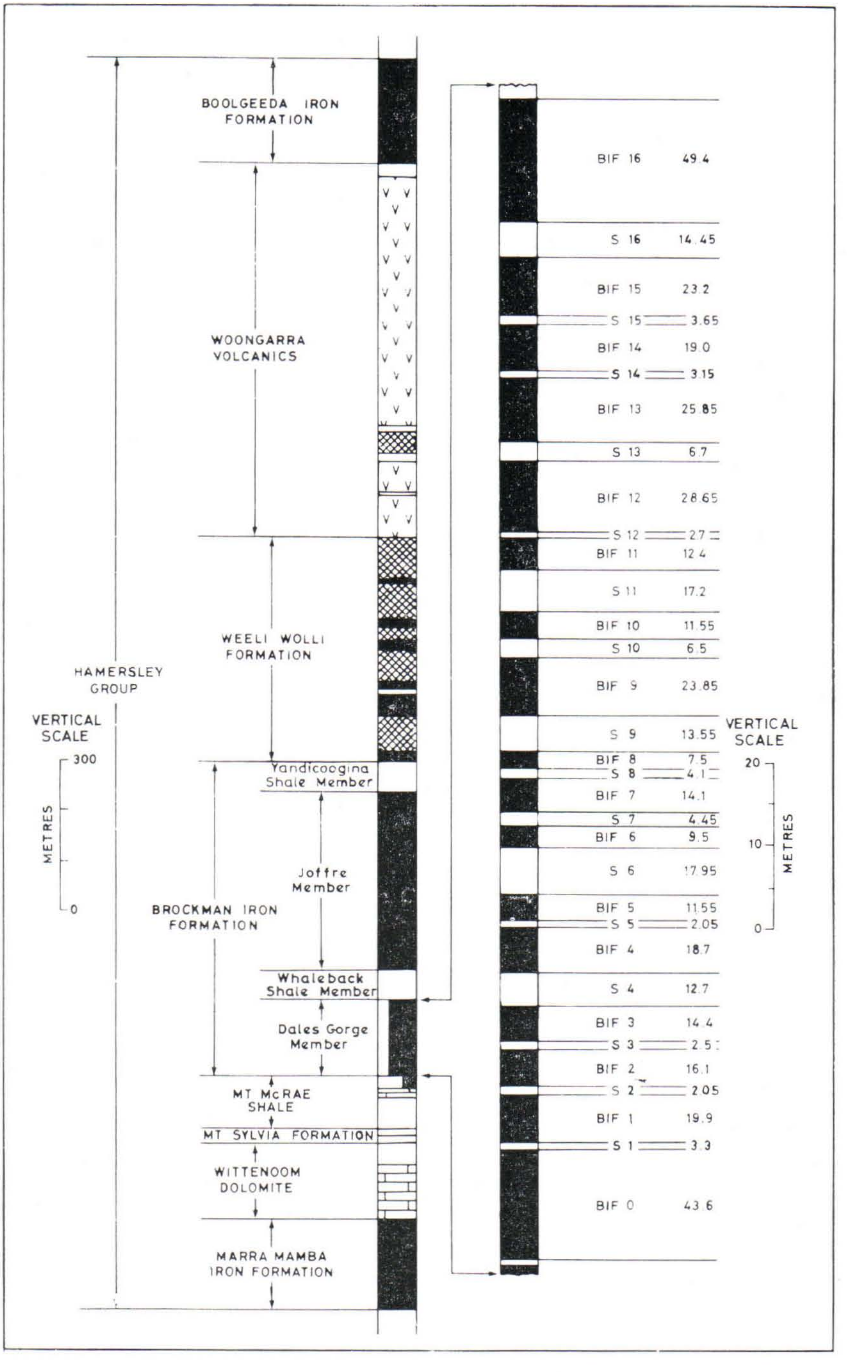

Fig. 2. Stratigraphic subdivision of the Hamersley Group.

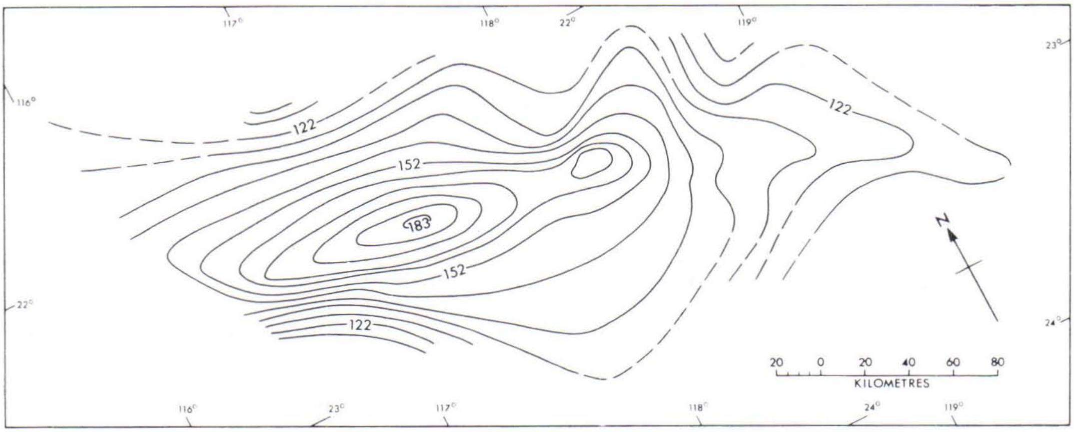

Fig. 3. Isopach pattern of the Dales Gorge Member. The thicknesses marked on the isopachs are in metres, and the isopach interval is approximately 6.1 metres. The locations of the measured sections, on which the isopachs are based, are given by Trendall and Blockley (1970).

central part of the basin, with a steady outward decrease in thickness in all directions, to give a generally elliptical pattern with some irregularities. The regional variation in thickness of the other formations, with the exception of the Woongarra Volcanics, appears to follow a similar pattern.

In the Table below one important feature of the vertical distribution of the sedimentary components of the Hamersley Group is emphasized; this is a broad alternation between shale (or shale and dolomite) and banded iron-formation (BIF), usually with subordinate shale. This alternation extends down into the Fortescue Group, and may be summarised as follows:

Unit	Approximate thickness (metres)
8. Weeli Wolli Formation and Boolgeeda Iron Formation	380
7. Yandicoogina Shale Member of Brockman Iron Formation	60
6. Joffre Member of Brockman Iron Formation	370
5. Whaleback Shale Member of Brockman Iron Formation	60
4. Dales Gorge Member of Brockman Iron Formation top of Mt McRae Shale	180
3. Wittenoom Dolomite+Mount Sylvia Formation+lower part of Mt McRae Shale	260
2. Marra Mamba Iron Formation	180
1. Jeerinah Formation of Fortescue Group	150—300

Each of the (even-numbered) BIF units in this sequence is lithologically different from the others, and in the case of the topmost unit the Weeli Wolli Formation and Boolgeeda Iron Formation also differ from each other to give a total of five major lithologically distinctive BIF units. These are described under a separate heading below.

Comparatively little work has been carried out on the sedimentary units other than the iron-formations (that is, the odd-numbered units in the Table above). The units which contain shale as the main, or an important, constituent include the Mount McRae Shale, the Wittenoom Dolomite, the Mount Sylvia Formation, and the Whaleback and Yandicoogina Shale Members of the Brockman Iron Formation. Although it is poorly exposed, much of the shale appears to share common characteristics: it is very fine-grained, dark grey or green to black when fresh, and white or buff when weathered, often finely laminated, and usually with a proportion of intermixed chert and carbonate, and sometimes a little iron-formation. The upper part of the Mount McRae Shale contains thin bands of shard-bearing volcanic ash, but there is no other direct evidence for a volcanic contribution. In spite of its apparent uniformity, there seem to be at least two varieties of shale: firstly, very fine-grained, soft, poorly laminated black shale rich in free carbon and pyrite, and secondly, dark green, finely laminated shale often with some microbanded chert and with evidence of volcanic activity. Intermediate types, may both lithologically and in depositional significance, represent mixtures of these two end-members, but insufficient work has yet been done to test this suggestion.

The lower part of the Wittenoom Dolomite is composed of massive, medium to thin-bedded dolomite with rare beds of black chert. Where fresh, the dolomite is finely crystalline and brown, pink, or grey with faint colour banding. It contains sedimentary structures such as cross-beds and slumps, and features such as stylolites and chert nodules. Stromatolites are present in the outlying stratigraphic equivalent of the Wittenoom Dolomite in the northeasternmost part of the basin, but are absent over the main part of the outcrop area. Davy (1975) has provided chemical and petrographic details of the lowermost part of the Wittenoom Dolomite, and Button (1976) has made comparative comments on the Wittenoom Dolomite and the extensive dolomites of the South African Transvaal Basin.

The igneous rocks which form over 40 per cent of the Hamersley Group comprise thick dolerite sills largely within the Weeli Wolli Formation, and acid volcanics of the overlying Woongarra Volcanics. Over most of the outcrop area over half the thickness of the Weeli Wolli Formation consists of dolerite in several separate sills. Although local variations have been recorded in the number and thickness of the sills, down to their complete absence in the Yarraloola Sheet area, there is insufficient information to relate these to other aspects of regional stratigraphy. A dolerite sill within the Woongarra Volcanics may be assumed to be the uppermost representative of this group of sills. The dolerite is uniformly massive and homogenous, dark green, and medium to coarse grained (de Laeter, Peers and Trendall 1974). It invariably shows strong deuteric modification, with uralite, albite, epidote and leucoxene replacing the presumably primary pyroxene, labradorite and ilmenite.

The acid Woongarra Volcanics, the thickest formal stratigraphic unit of the Hamersley Group, includes discrete bands of both non-porphyritic and porphyritic rhyolite and dacite as well as thinner horizons of stratified tuff and agglomerate, at least one major intercalation of BIF and a dolerite sill. Although there is apparently a high order of lateral stratigraphic continuity and concordance with the overlying rocks, recent unpublished re-examination has raised some doubt whether the volcanic rocks may be in part, or locally, intrusive rather than lavas.

Turee Creek Group and Wyloo Group

In all accounts of the Hamersley Basin so far published an uppermost, mainly clastic, stratigraphic division — the Wyloo Group — has been represented as the final, terminating component of the content of the basin. However, several early reservations (Trendall and Blockley 1970, p. 295; Trendall 1975, p. 119) have been expressed concerning the validity of this concept, and recent stratigraphic revision (Trendall 1979) is linked to the tectonic concept put forward by Gee (1979) that the bulk of the Wyloo Group was laid down in a separate elongate depositional basin to the south of the depositional basin of the Fortescue Group and Hamersley Group. Gee referred to this later basin as the Ashburton Trough, and to its present folded content as the Ashburton Fold Belt.

The stratigraphic revision of Trendall (1979) consists basically of redefining the Turee Creek Formation, the basal formation of the Wyloo Group as the name has so far been used. Redefinition has involved, apart from fulfilling certain stratigraphic nomenclatural requirements which had been omitted previously, the raising of the formation to group status, and of defining within the Turee Creek Group, a basal Kungarra Formation. The Turee Creek Group now forms, with the Fortescue Group and Hamersley Group, the Mount Bruce Supergroup. The redefined Wyloo Group now has at its base the Beasley River Quartzite, the basal conglomerate of which — the Three Corner Conglomerate — overlies both the Hamersley Group and Turee Creek Group with marked angular unconformity. This unconformity defines the initiation of the new depositional regime of the Ashburton Trough, the stratigraphy of which is not further dealt with here. The unconformity has so far been mapped and defined only in a restricted area between the Wyloo Dome and the Hardey Syncline, and no attempt has therefore been made to represent its regional extent on Figure 1. Much of the western boundary of the Mount Bruce Supergroup, in the revised sense, is probably structural, with the Hamersley Group faulted directly against the Wyloo Group.

The Turee Creek Group (Trendall 1979) includes the basal Kungarra Formation, with a thickness in the type area of about 3 km, and at least a further kilometre thickness of quartzite, shale, and dolomite not so far formally named. The Kungarra Formation overlies the Hamersley Group with apparently perfect conformity wherever the transition is exposed. It consists of a uniform and monotonous sequence of greyish green siltstone, fine-grained greywacke, and fine sandstone. Thin carbonates are interbedded with siltstones in the higher parts of the formation, and dolerite sills are abundant.

Within the Kungarra Formation, slightly above the middle, the 300 m-thick Meteorite Bore Member forms a distinctive lithology extending over a discontinous outcrop length of some 30 km. It consists of an essentially unstratified siltstone matrix in which are sparsely and randomly distributed boulders, up to 50 cm across, of fine sandstone, acid volcanic and other exotic material. Most of the boulders are rounded, but a proportion of them are distinctly faceted, and these facets commonly bear parallel grooves and striations consistent with a glacial origin (Trendall 1976). The acid volcanic boulders are petrographically indistinguishable from material of the underlying Woongarra Volcanics of the Hamersley Group, but it is emphasized that no discordant contact of the Kungarra Formation with the Hamersley Group is known.

The iron-formations

Lithology and petrography

The stratigraphic position of the iron-formations has been set out above, and appears in Figure 2. Because the Dales Gorge Member is the most intensively studied, and therefore the best known, of the five iron-formation units, this description begins with it.

The type section of the Dales Geoge Member of the Brockman Iron Formation was defined by Trendall and Blockley (1968) in drill core from Wittenoom Gorge and Yampire Gorge. The division of the 142.2 m of type section core into two main lithologies is shown in the right-hand column of Figure 2. Each of the 33 numbered subdivisions of the member is designated a macroband. The 17 BIF macrobands are numbered upwards from 0 to 16 and each of these except the lowermost is underlain by a similarly numbered shale, or S, macroband. Trendall and Blockley (1968), in their published description of the type section, included continuous photographic coverage of the defining drill core at a scale of one fifth natural size.

The S macrobands consist mainly of dark green to black, iron-rich, stilpnomelane-bearing shale, often finely laminated, and of interbanded chert and green siderite, which may be very finely laminated, more or less structureless, or thinly bedded with a ghost clastic structure defined by slight colour variations within the fine-grained siderite. Limestone and breccia bands also occur within the S macrobands. Thin bands of stilpnomelane within the shale have textural variations which define the shapes of volcanic shards (La Berge 1966).

The BIF macrobands of the Dales Gorge Member conform in general lithology with »typical» Precambrian banded iron-formation of many continents, but resemble most closely those of the Cape Province and Transvaal of South Africa (Trendall 1968 a). Alternate thin bands (called mesobands) of chert and fine-grained iron-rich material were designated chert-matrix by Trendall and Blockley (1970). Chert mesobands are mostly 5—15 mm thick, and in common with other types have by definition a minimum thickness of 1 mm; the frequency of thicker cherts falls off rapidly up to the measured maximum of 87 mm. Mesobands of chert-matrix have similar thicknesses, and these two main mesoband types, chert and chert-matrix make up about 60 and 20 per cent respectively of the total BIF volume. The remainder are magnetite, carbonate, stilpnomelane, riebeckite, and minor miscellaneous types.

Within most chert mesobands, a small-scale regular lamination defined by layers of some iron-bearing mineral within the general fine mosaic of quartz is present, and is known as microbanding. Although microbands were initially defined (Trendall 1965), as consisting of a simple iron-rich/iron-poor couplet, subsequent detailed examination shows that a good deal of fine structure may be distinguished within this. Up to six internal subdivisions of single microband couplet may be present, some of which may themselves possess evident substructure. The defining mineral of the main iron-rich component of each microband is normally hematite, siderite, or ankerite, but may exceptionally be magnetite, stilpnomelane or riebeckite. The microband interval (between the centres of adjacent iron-rich laminae) in different cherts may be between 0.2 and 1.5 mm but there is by comparison, negligible variation between successive laminae in a single mesoband. The total iron content of chert mesobands varies inversely with microband interval, between about 3 and 30 per cent. Chert-matrix mesobands, by contrast with cherts, consist of a fine-grained mixture of quartz, magnetite, hematite, stilpnomelane, ankerite and siderite and have no regular microbanding, although there is often a vaguely defined and irregular streakiness. Chert-matrix has an average total iron content of about 40 per cent. Although the ideal geometric situation of perfectly planar mesobands is approached by much of the BIF there is also a great variety laterally discontinuous cherts (pod cherts or cross-pods), at whose lateral terminations the microbanding passes without disruption, but with an approximate 7:1 reduction in thickness, into the vague streaky lamination of the adjacent chert-matrix.

In many BIF macrobands of the Dales Gorge Member there exists a cyclic sequence of mesoband types, over a stratigraphic thickness of 10—15 cm which Trendall and Blockley (1970, p. 55—60) have called the Calamina cyclothem. In this sequence a single thick chert mesoband with fine microbanding, or a group of such cherts separated by thin magnetite mesobands, alternates with a mixed mesoband group of chert-matrix, magnetite, and comparatively thin and coarsely microbanded cherts; these two alternating components of the Calamina cyclothem are called the chert-magnetite group and the mixed group respectively. The thin microbanded cherts of the chert-magnetite group are normally pink or red, and have their microbanding defined by hematite, while the more coarsely microbanded cherts of the mixed group are white, with carbonate-defined microbanding.

Riebeckite occurs within restricted stratigraphic sections (riebeckite zones) of the Dales Gorge Member in some areas, mainly in the form of thick massive mesobands (actually consisting of randomly interlocked fibres) which are stratigraphically equivalent to, and result from the replacement of, the chert-magnetite group of particular Calamina cyclothems. Crocidolite, or blue asbestos, is among the several minor textural types of riebeckite occurrence.

The other four major stratigraphic units of BIF, and also the comparatively minor BIF development of the Mount Sylvia Formation (Fig. 2) are each characteristically different in some way, although all share the fundamental defining characteristics of high iron content and the presence of quartz (chert) and iron oxides as the dominant constituent minerals. They differ in such subtle characters as the nature of the sheet silicate component, the abundance and distribution of riebeckite, and the amount of shale intercalation, and the thickness and degree irregularity of the chert mesobands; one or more of these have resultant effects on the field expression. For detail reference should be made to Table 3 of Trendall (1976 a).

Lateral stratigraphic continuity

From the foregoing descriptions it is apparent that there exists, in the Hamersley Group, a wide range of regular depositional alternations differing in scale and in their exact degree of repetitive regularity. These are summarised in Figure 4.

A remarkable feature of all these scales of stratification is the very high degree of their lateral continuity throughout the area of the basin. No example of significant non-continuity of the major stratigraphic boundaries (F of Figure 4) has yet been well documented, although Trendall and Blockley (1970) and Daniels (1970) have noted anomalies in the Wyloo Dome area (Fig. 1) attributed by Horwitz (1978) to »removal by slumping, during sedimentation». On the next smaller scale the lateral continuity

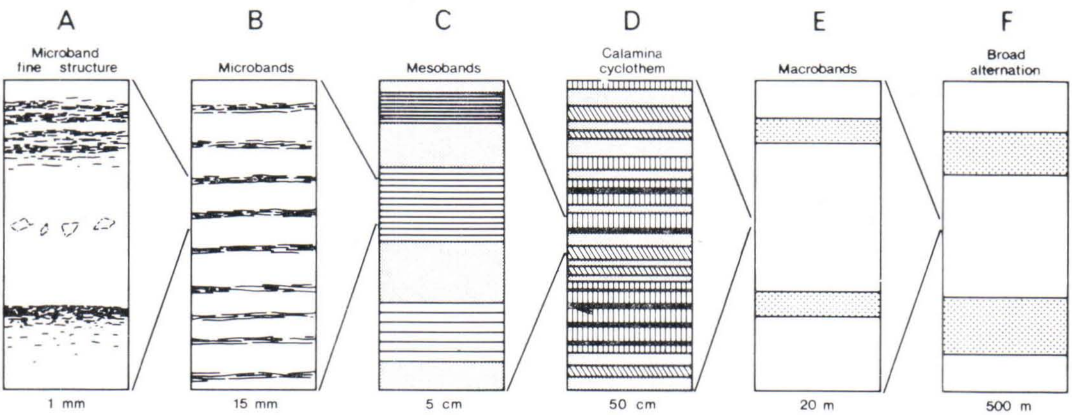

Fig. 4. Summary of stratification scales within BIFs of the Hamersley Group. In A are represented two iron-rich components, of successive microbands, each defined by streaks of 'dusty' hematite with chert. The upper hematite layer is subdivided into two parts (Trendall 1973 b), while the lower one is 'graded' (Trendall and Blockley 1970, Fig. 26 c). A few ankerite rhombs between the two hematite layers mark an event within the iron-poor layer. In C, microbanded chert mesobands of different microband interval are represented, separated by chert-matrix (light stipple). In D, the Calamina cyclothem is shown, defined by chert-magnetite groups (red chert with vertical lines, magnetite black) alternating with mixed groups (white chert with diagonal shading). The heavy stipple in E and F represents shale, while BIF is left blank.

of the Dales Gorge Member macrobands (E of Figure 4) is demonstrably unbroken over the whole 60 000 km^2 of the Hamersley Group outcrop area. The continuity of the smaller scale stratification (A-D of Fig. 4) is not so clearly demonstrable, but Trendall and Blockley (1970) and Trendall (1972) have shown examples of basin-wide microband continuity. At present it seems a valid generalization that all scales of stratigraphic banding of the Hamersley Basin iron-formations initially possessed basin-wide continuity and regularity, and that over virtually the whole basin their continuity has survived all subsequent processes of modification.

Chemical composition

Trendall and Blockley (1970, Chapter 5) provided 48 complete and 19 partial analyses of various samples from the Hamersley Group iron-formations, including BIF, shale, and individual mesobands. However, a more systematic assessment of the chemical composition of the Dales Gorge Member has been made more recently by Trendall and Pepper (1977). They report the following bulk composition (all figures weight per cent): SiO_2 44.34, Al_2O_3 0.89, Fe_2O_3 29.30, FeO 13.45, MgO 2.31, CaO 1.79, Na_2O 0.53, K_2O 1.26, H_2O^+ 0.98, CO_2 4.63, FeS_2 0.12, TiO_2 0.05, P_2O_5 0.18, MnO 0.17. Trace elements were generally at or below clarke values, but Ag, Bi, Cu, Ge, Pb and Sn appeared high.

Stable isotope studies

Becker and Clayton (1972, 1976) have carried out carbonate carbon and oxygen isotopic studies on both the iron-formations and some other stratigraphic units of the Hamersley Group, and Oehler and others (1972) have reported carbon isotope data for kerogen carbon. Becker and Clayton (1972) concluded from their carbon study that the iron-formation was precipitated in a basin isolated from the ocean, but probably in close proximity to it, and that organic activity may have played a significant role in the genesis of the iron-formation. From their (1976) oxygen isotope work the most significant finding was that the minerals of the Dales Gorge Member had undergone isotopic exchange at a temperature estimated to be above 270°C and probably less than 310°C, during burial metamorphism. Oehler and others (1972) found no anomaly in organic carbon isotopes in the Hamersley Group compared with other Precambrian sediments.

Vander Wood (1977) has reported initial $^{87}Sr/^{86}Sr$ values from the Dales Gorge Member, and has concluded that they are consistent with non-marine deposition.

Structure

It has already been noted that in the northern area of the basin the Fortescue Group was laid down with strong angular unconformity over the Archaean Pilbara Block (Figure 1), and from the evidence of scattered inliers the whole of the basin is assumed to have had such a stable cratonic floor of ancient rocks. However, whereas in the north neither the Fortescue Group nor the Hamersley Group has undergone significant subsequent deformation the intensity of folding increases steadily southwards, until, along their extreme southern outcrop area, the folding is relatively tight with the axial planes steeply south-dipping and south synclinal limbs overturned or vertical. In general the folding over much of the basin is open and simple, with average dips of 20°—40° on fold limbs, and a general axial trend from east-west in the eastern area, through west-northwest in the centre, to northwest in the westernmost area. In detail, however, the folding is highly irregular, with a frequent lack of axial continuity, and a multiplicity of scales of minor folds, generally parallel to the main regional trend. At the western and eastern ends of the basin faulting is prominent, in predominantly northwest and northeast directions respectively.

Although MacLeod (1966, p. 63—65, and Plate 2) has presented a good structural description and representation of the basin there has been relatively little analysis of basin structure from the viewpoint of overall tectonic evolution. The paper by Miyano (1966) represents one such attempt, which basically relates the present structure to the formation of a broad gentle dome centred in the north-central part of the present outcrop area. Other accounts of tectonic evolution are more concerned with broad regional syntheses of deposition and tectonics, and these are referred to farther below.

Metamorphism

Trendall and Blockley (1970, p. 294) asserted that the »Hamersley Group has nowhere undergone regional metamorphism», and supported this view with unpublished evidence of Hoering, work by Grubb (1967) on riebeckite synthesis, and preliminary oxygen isotope work by Becker and Clayton. A maximum temperature of 160°C was suggested, and it was considered that »the evolution of the Hamersley Basin was not accompanied by heating of the sediments to temperatures above those associated with ordinary geothermal gradients». Since that time a final synthesis of the oxygen isotope data (Becker and Clayton 1976) indicates temperatures »above 270°C and probable less than 310°C, during burial metamorphism». Ayres (1972) also suggested that there was significant metamorphism of the Dales Gorge Member from a consideration of the mineral paragenesis, and suggested »pressures of 4 to 6 kilobars produced by the load of overlying strata and a temperature of 300°C due to a geothermal gradient of 15°C/km». Current studies by Klein of the sheet silicates of the Hamersley Group iron-formations should provide valuable additional data for specifying conditions of metamorphism. Smith's (1975) study of the metamorphism of the mafic volcanic rocks of the Fortescue Group has shown that these have locally reached greenschist facies, although over the bulk of their outcrop area they have attained only the prehnite-pumpellyite facies.

Age

In spite of a massive amount of accumulated isotope analytical data, it is at present impossible to be certain of the time of the initiation, duration, or termination of the Hamersley Basin within several hundred million years. It is not feasible to present and discuss all the available data in this review. This has been done most recently by Trendall (1976 a) whose conclusions to that date are first summarised.

A *maximum* age for the initiation of the basin is given by the *youngest* granitic rock from the Pilbara Block; this may be as old as about 2 600 Ma or as young as 2 500 Ma. Fortescue Group rocks are not seen to directly overlie granite of this age; a variety of geological evidence argues against the likelihood, but does not exclude the possibility, of such granite being younger than the Fortescue Group, in which case the maximum possible initiation age would extend to about 3 000 Ma, based on granites actually overlain by the Fortescue Group. A *minimum* age for basin initiation is given as about 2 200 Ma by a granophyre intrusive along the basal unconformity at the northwestern edge of the outcrop area. The widest possible age range for basin initiation is thus 2 700 to 2 200 Ma.

From Trendall's (1976 a) data review he concluded that deposition in the Hamersley Basin could have started as early as 2 700 Ma and lasted until 1 850 Ma; or alternatively it could have started as late as 2 200 Ma and finished within an isotopically indistinguishable interval — less, say, than 100 Ma.

Since that review a substantial amount of additional data has been obtained, but little has been published, and it would not be appropriate to anticipate the full effect of recent work here.

However, Hickman and de Laeter (1977) have presented data which suggest that the Hardey Sandstone close to the base of the Fortescue Group, may be between 2 700 and 2 600 Ma old. U-Pb work by Compston and others (in prep.) is likely to support an age for the Hamersley Basin closer to the older than the younger of the limits already referred to. Speculative generalisations based on earlier published age estimates (e. g. Trendall and Blockley 1970, p. 31) should be avoided until the work of Compston and others (in prep.) has been published.

Mineral deposits

The rather sudden rise of the Hamersley Basin to a position of significance in the international literature on iron-formation is closely paralleled by the rapid development of its resources of iron ore. In response to the lifting of an Australian Government embargo on the export of iron ore in 1961, export of high-grade hematite-goethite ore began in 1966 after intensive exploration and development. In 1976, ten years later, over 85 million tonnes of iron ore were mined in Western Australia, of which over 77 million tonnes were exported, representing about 18 % of all iron ore transported across international boundaries throughout the world in that year. The bulk of this came from the Hamersley Basin.

Although it is hardly possible to summarise the geology of the Hamersley Basin without reference to this economic aspect it will be sufficient here to indicate where details of the occurence, distribution, and development of the iron ore associated directly or indirectly with the iron-formations of the Hamersley Group can be found. MacLeod's (1966) bulletin is an excellent summary of the geology of iron ore in the Hamersley Basin which has not been superseded by later work. However, Trendall (1975) has reviewed general aspects of ore genesis, and in the same volume, papers by Gilhome, Baldwin, Evans and Clint, Kneeshaw, Ward and others, Neale, and Adair (all 1975) give more detail of individual ore bodies. Lord and Trendall (1976) have provided a review of mining development while a review of the iron ore resources of the Hamersley Basin, totalling over 36×10^9 tonnes, is given by Morrison (1978). Ayres (1971) provides mineragraphic detail.

Iron ore is overwhelmingly the most important exploitable mineral of the Hamersley Basin. Blue (riebeckite) asbestos, or crocidolite, was mined up to 1966 (Trendall and Blockley 1970), while small deposits of lead, zinc, silver, copper, and gold have been worked (Blockley 1975), all of them within the Fortescue Group.

Depositional environment

Fortescue Group

The Hamersley Basin was initiated by the basic volcanic activity of the Fortescue Group, accompanied by subaerial accumulation of terrigenous debris, at several scattered centres to form the basin. These centres were preferentially located over sections of the narrow greenstone belts of the Archaean floor of the basin, but the actual sources of the lavas may have been concurrent fissures represented by dyke swarms now most conspicuous within the adjacent granitic domes (Lewis *et al.* 1975). In this interpretation the granite plutons may be envisaged as the site of fissure eruptions, as a result of which lava streams poured down, together with granitic debris washed down by accompanying torrential rains, to accumulate over the lower ground of the greenstone belts. It is uncertain to what extent, if this concept is valid, the difference in elevation was due to long standing denudational effects, or to concurrent crustal deformation (Hickman and Lipple 1975; Hickman 1975).

During the deposition of the Fortescue Group the extent of volcanicity gradually spread, and became of basinwide extent, so that the initially local depressions lost their identity within the broad regional sinking of the whole future area of the basin. This appears to have been greatest in the Mount Bruce Sheet area. Long periods of quiet lava effusion alternated with periods of explosive activity during which tuffs were the principal volcanic product. The volcanicity was a self-regulating process, which was finally reduced to a very low level through the stifling effect of the thick blanket of accumulated material; at this stage, some magma, unable to break through to the surface, spread locally to form extensive stratiform intrusions, often near the base of the volcanic succession.

Hamersley Group

By the close of Fortescue Group deposition the final form which the basin possessed during deposition of the Hamersley Group appears to have become established. The best evidence for the reconstruction of the basin environment at this stage comes from a consideration of the most characteristic lithology — banded iron-formation (BIF). The arguments used by Trendall and Blockley (1970) to establish a self-consistent hypothesis for BIF deposition were a refinement of Trendall's (1966) earlier suggestions; later revisions (Trendall 1972; 1973 a and b) have not altered its main features so far as direct basin interpretation are concerned, but viewpoints on more speculative problems, such as the source of iron, have since been modified. The following summary more or less reproduces that of Trendall (1976 a) in which most of the detail of the evidence was omitted for clarity.

Microbands are regarded as the only primary depositional features within the BIF, and the regularity of microband spacing within any one chert mesoband is taken to indicate that the iron-rich/iron-poor cyclicity by which they are defined is related to some equally regular natural environmental event controlling deposition. Only two natural events involving repeated systematic environmental changes have great regularity: the day and the year, controlled respectively by Earth's rotation and revolution. The general order of thickness of microbands precludes the day as a control, and it is concluded that microbands are seasonally controlled annual layers: they are varves. The origin of the fine structure within microbands is not understood.

The regional continuity of microbands then precludes any depositional mechanism other than chemical precipitation for microbands. It is inconceivable that an undisturbed layer of an even thickness no greater than a few millimetres could be spread over an area of at least 50 000 km^2 by any physical means, unless by the infall of dust.

The sheer bulk of BIF laid down in the basin makes it unlikely that gross systematic chemical alteration has taken place, and it is therefore accepted both that silica and iron were almost the only stable constituents of the deposited material. If this is so, then its origin as a precipitate from the basin water seems more acceptable than an origin as either volcanic or terrigenous dust (Carey 1976).

The lack of contemporaneous disturbance in the BIFs indicates an exceptionally still and quiet subaqueous environment. There is complete absence of evidence of any supply of material to the basin by incoming drainage. If the isopachs of the Dales Gorge Member (Fig. 3) reflect basin shape, evaporative concentration of iron and silica within a closed basin becomes an attractive hypothesis. A completely arid circum-basinal terrain is consistent with both this, the preceding point, and a close relationship between climate and deposition. Nevertheless, absence of any major stratigraphic disturbance within the Hamersley Group that could be attributed to dessication suggests a limited connection with the open ocean, and a barred basin acting as a circulating cell has potential for replenishment by incoming seawater. If the edges of the basin lay not far beyond the remaining outcrop area of the Hamersley Group a basin area of about 150 000 km^2 is indicated.

If microbands are chemical varves then an origin needs to be suggested for mesobands which is consistent with this hypothesis. Two questions are relevant: firstly, why does the microband interval vary between different chert mesobands although it remains constant within any one mesoband, and secondly, what does chert-matrix represent? The question of the status of chert-matrix seems to be resolved by the evidence provided by the lateral terminations of podded cherts; chert-matrix appears to represent grossly compacted material that originally was the chert. If compaction can have such a radical effect then it may possibly also explain the variation of microband interval between different chert mesobands on the following hypothesis.

Suppose that, each year, about 5 mm of extremely hydrous precipitate were laid down evenly over the entire basin. As successive layers accumulated, the initially gelatinous colloidal material became compressed and dehydrated, with some silica departing in solution in the water, but the iron remaining stable. An inverse relationship would be produced between microband thickness and Fe content of chert mesobands. This is the actual situation (Trendall and Blockley 1970), and the hypothesis is currently accepted as most consistent with the available evidence.

Within the Weeli Wolli Formation there is evidence for a systematic 23.3-year cyclicity of microbandings. An increase in the amplitude of variation of whatever depositional conditions were actually responsible for the expression of this rhythm in the Weeli Wolli Formation is believed to have been responsible for mesobanding in the Dales Gorge Member and other BIFs. A longer term cyclicity of depositional environment is believed to have caused the Calamina cyclothem, and related alternations of red and white chert mesobands, but the cause of this is still doubtful.

Volcanoclastic structures in the S macrobands of the Dales Gorge Member and in some of the major shale units indicate that these represent intervals when 'normal' quiet BIF accumulation was prevented by steady influx of volcanic debris. For the Dales Gorge Member, these intervals are probably related to spasmodic sinking of the basin.

The immediate cause, or causes, of precipitation in the basin are uncertain, but the presently preferred hypothesis is that the precipitation of iron may have been caused by the oxidation of ferrous iron in the basin water by oxygen evolved during photosynthesis by algae, floating either at or below the water surface. The presence of apparent microfossils in the Brockman Iron Formation (La Berge 1967; Karkhanis 1976) and of isotopically light carbon in carbonates (Becker and Clayton 1972) supports the presence of plant life in the basin. The precipitation of silica, on the other hand, may have been caused purely by evaporative concentration. In this hypothesis, both processes would be independently related to annual insolation, so that there would be scope for textural variation due to minor differences each year in such parameters as basin temperature or turbidity. Neither of the main constituents of the precipitate need have been present in large concentrations in the basin water.

Turee Creek Group

The simplest interpretation of the apparently concordant and quiet transition from the Hamersley Group to the Turee Creek Group is that subsidence along the southern margin of the basin effected a connection with the open ocean, with two main results. Firstly, the unique chemistry of the basin water was altered to prevent further BIF deposition, and secondly, an influx of fine terrigenous material led to the accumulation of siltstones and associated clastic rocks. There is no clear evidence that both changes could not have been due entirely to tectonics. The distribution of the lowest formation of the Turee Creek Group — the Kungarra Formation — suggests that it covered at least the southern quarter of the basin. However, the presence within it of possible glacial rocks, containing material possibly derived from the Hamersley Group, indicates the possibility that there was an associated climatic change and that in some areas of the basin — possibly the northern part — the Hamersley Group was exposed during deposition of the Turee Creek Group, which was never, in those areas, deposited.

Wyloo Group

The lowest formation of the Wyloo Group (Trendall 1979) has a thick, coarse, unstratified conglomerate composed entirely of Hamersley Group debris. The violent event so recorded is considered to mark the initiation of the Ashburton Trough of Gee (1979), which lies outside the scope of this paper.

Regional relationships

The preceding account of the depositional environment of the Hamersley Basin may be considered as an 'orthodox' interpretation following on from the study of Trendall and Blockley (1970) and elaborated in subsequent publications. In 1975 a further Proterozoic sedimentary basin containing iron-formation in the 1 000 m-thick Frere Formation in the lower part of the section was recognised to the south of the Hamersley Basin, and separated from it by the unconformably overlying and younger (c. 1 000 Ma) Bangemall Basin; Hall and Goode (1975) named it the Nabberu Basin, and more details of its stratigraphy and structure have been given by Bunting *et al.* (1977), Hall *et al.* (1977), and Hall and Goode (1978). Horwitz and Smith (1978) have associated the Nabberu Basin and the Hamersley Basin in a broad synthesis of Proterozoic depositional and tectonic history. They (p. 316) suggest that »the conditions of sedimentation in the Hamersley Group are not indicative of a barred basin, but existed over a ... shelf which... extended from the northeastern part of the Yilgarn Block and the eastern part of the Pilbara Block into a sea to the north-west». They point out that» the Hamersley Group and the Frere Formation are overlain» by rocks which» could equate in age to parts of the Wyloo Group» and the impression that they consider the iron-formations of the Hamersley and Nabberu to be stratigraphically and temporally equivalent shelf deposits is supported by Horwitz' (1976) view that »Both basins could well ... be parts of the same one basin». Button (1976) has also emphasized the stratigraphic similarities between the Nabberu and Hamersley Basins.

So far, no interpretation of the Hamersley Basin and Nabberu Basin as separate sections of a formerly continuous marine shelf has satisfactorily integrated the isopach evidence for the Hamersley Group (Fig. 3) or the conclusions from stable isotope studies that the Hamersley Basin was substantially isolated from the open ocean; such an interpretation is also basically at variance with an evaporative component in the genesis of the iron-formations, and is not supported by major lithological dissimilarities between the iron-formations of the two basins. More importantly than all these, however, is the increasing geochronological evidence that the Nabberu Basin is significantly younger than the Hamersley Basin, a relationship recognized by Gee (1979) in his alternative depositional and tectonic synthesis of the region. As in many other areas of Precambrian rocks throughout the world, there is at present an exciting increase in the scope and depth of evidence which can be applied to supply a degree of interpretative detail undreamed of only a couple of decades ago; this review of progress on the Hamersley Basin is likely to be superseded quite quickly as the results of current work appear.

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References

Adair, D. L., 1975. Middle Robe River iron ore deposits. *In* Economic geology of Australia and Papua-New Guinea. Australasian Inst. Mining Metall., Monograph 5, 943—945.

Ayres, D. E., 1971. The hematite enrichment ores of Mount Tom Price and Mount Whaleback, Hamersley Iron Province. Proc. Australasian Inst. Mining Metall. No. 238, 47—58.

— **1972.** Genesis of iron-bearing minerals in banded iron formation mesobands in the Dales Gorge Meber, Hamersley Group, Western Australia. Econ. Geol. 67, 1214—1233.

Becker, R. H. & Clayton, R. N., 1972. Carbon isotopic evidence for the origin of a banded iron-formation in Western Australia. Geochim. Cosmochim. Acta 36, 577—595.

— **1976.** Oxygen isotope study of a Precambrian banded iron-formation, Hamersley Range, Western Australia. Geochim. Cosmochim. Acta 40, 1153—1165.

Blockley, J. G., 1975. Hamersley Basin — mineralization: *In* Economic geology of Australia and Papua-New Guinea. Australasian Inst. Mining Metall., Monograph 5, 413—415.

Bunting, J. A., Commander, D. P. & Gee, R. D., 1977. Preliminary synthesis of Lower Proterozoic stratigraphy and structure adjacent to the northern margin of the Yilgarn Block. West. Australia Geol. Surv. Ann. Rept. 1976, 43—48.

Button, A., 1976. Transvaal and Hamersley Basins — review of basin development and mineral deposits. Mineral Sci. Eng. 8, 262—293.

Carey, S. W., 1976. The Expanding Earth. Developments in Geotectonics, 11, Elsevier, Amsterdam.

Daniels, J. L., 1970. Wyloo, W. A. West. Australia Geol. Surv., 1:250 000 Geol. Series, Explanatory Notes.

Davy, R., 1975. A geochemical study of a dolomite-BIF transition in the lower part of the Hamersley Group. West. Australia Geol. Surv. Ann. Rept 1974, 88—100.

de Laeter, J. R., Peers, R. & Trendall, A. F., 1974. The petrography, chemical composition and geochronology of two dolerite sills from the Precambrian Weeli Wolli Formation, Hamersley Group. West. Australia Geol. Surv. Ann. Rept 1973, 82—91.

Evans, W. J. & Clint, P., 1975. Iron deposits of the Brockman Syncline, Hamersley Iron Province. *In*

Economic geology of Australia and Papua-New Guinea. Australasian Inst. Mining Metall., Monograph 5, 906—910.

Gee, R. D., 1979. Structure and tectonic style of the Western Australian Shield. Tectonophysics 58, 329—369.

Gilhome, W. R., 1975. Mount Tom Price Iron Orebody. *In* Economic geology of Australia and Papua-New Guinea. Australasian Inst. Mining Metall. Monograph 5, 892—898.

Grubb, P. L. C., 1967. Asbestos. CSIRO Division of Applied Mineralogy, Ann. Rept 1966—67, p. 6.

Hall, W. D. M. & Goode, A. D. T., 1975. The Nabberu Basin — a newly discovered Lower Proterozoic basin in Western Australia. Geol. Soc. Australia, 1st Aust. Geol. Convention, »Proterozoic Geology» (Abstract), 88—89.

— **1978.** The Early Proterozoic Nabberu Basin and associated iron formations of Western Australia. Precambrian Res. 7, 129—184.

Hall, W. D. M., Goode, A. D. T., Bunting, J. A., & Commander, D. P., 1977. Stratigraphic terminology of the Earaheedy Group, Nabberu Basin. West. Australia Geol. Surv. Ann. Rept 1976, 40—43.

Halligan, R. & Daniels, J. L., 1964. Precambrian geology of the Ashburton Valley region, North-West Division. West. Australia Geol. Surv. Ann. Rept, 1963, 38—46.

Hickman, A. H., 1975. Explanatory Notes on the Nullagine 1 : 250 000 Geological Sheet, W. A., West. Australia Geol. Surv. Rec. 1975/5.

Hickman, A. H. & Lipple, S. L., 1975. Explanatory Notes on the Marble Bar 1 : 250 000 Geological Sheet W. A., West. Australia Geol. Surv. Rec. 1974/20.

Hickman, A. H. & de Laeter, J. R., 1977. The depositional environment and age of a shale within the Hardey Sandstone of the Fortescue Group. West. Australia Geol. Surv. Ann. Rept 1976, 62—68.

Horwitz, R. C., 1976. Two unrecorded basal sections in older Proterozoic rocks of Western Australia. CSIRO Division of Mineralogy Rept FP17.

— **1978.** The Lower Proterozoic of the Wyloo Anticline. CSIRO Division of Mineralogy Rept FP20.

Horwitz, R. C. & Smith, R. E., 1978. Bridging the Yilgarn and Pilbara Blocks, Western Australia. Precambrian Res. 6, 293—322.

Karkhanis, S. N., 1976. Fossil iron bacteria may be preserved in Precambrian ferroan carbonate. Nature 261, 406—407.

Kneeshaw, M., 1975. Mt Whaleback iron orebody, Hamersley Iron Province. *In* Economic geology of Australia and Papua-New Guinea. Australasian Inst. Mining Metall. Monograph 5, 910—916.

La Berge, G. L., 1966. Altered pyroclastic rocks in iron-formation in the Hamersley Range, Western Australia. Econ. Geol. 61, 147—161.

— **1967.** Microfossils and Precambrian iron-formations. Geol. Soc. America Bull. 78, 331—342.

Lewis, J. D., Rosman, K. R. J. & de Laeter, J. R., 1975. The age and metamorphic effects of the Black Range dolerite dyke. West. Australia Geol. Surv. Ann. Rept 1974, 80—88.

Lord, J. H. & Trendall, A. F., 1976. Iron ore deposits of Western Australia, geology and development. American Ass. Petroleum Geol., Memoir 25.

MacLeod, W. N., 1966. The geology and iron deposits of the Hamersley Range area. West. Australia Geol. Surv. Bull. 117.

MacLeod, W. N., de la Hunty, L. E., Jones, W. R. & Halligan, R., 1963. Preliminary report on the Hamersley Iron Province, North West Division. West. Australia Geol. Surv. Ann. Rept 1962, 44—54.

Maitland, A. G., 1909. Geological investigations in the country lying between 21°30' and 25°30'S lat. and 113° 30' and 118°30'E long., embracing parts of the Cascoyne, Ashburton and West Pilbara Goldfields. West. Australia Geol. Surv. Bull. 33.

Miles, K. R., 1942. The blue asbestos bearing banded iron formations of the Hamersley Ranges. West. Australia Geol. Surv. Bull. 100.

Miyano, T., 1976. Geotectonic history of the Proterozoic Erathem in the Hamersley area, Western Australia. Mining Geol. 26, 207—220 (Tokyo, in Japanese).

Morrison, R. J., 1978. Iron ore resources of Western Australia as at 31 December 1977. West. Australia Geol. Survey, Rec. No. 1978/11.

Neale, J., 1975. Iron ore deposits in the Marra Mamba Formation at Mining Area »C», Hamersley Iron Province. *In* Economic geology of Australia and Papua-New Guinea. Australasian Inst. Mining Metall. Monograph 5, 924—932.

Oehler, D. Z., Schopf, J. W. & Kvenvolden, K. A., 1972. Carbon isotopic studies of organic matter in Precambrian rocks. Science 175, 1246—1248.

Smith, R. E., 1975. Metamorphism of the Proterozoic Fortescue Group, Western Australia — a reconnaissance study. CSIRO Division of Mineralogy, Rept FP9.

Talbot, H. W. B., 1920. The geology and mineral resources of the North-West, Central, and Eastern Divisions, between Long. 119° and 122°E, and Lat. 22° and 28°S. West. Australia Geol. Surv. Bull. 83.

Trendall, A. F., 1965. Progress report on the Brockman Iron Formation in the Wittenoom—Yampire area. West. Australia Geol. Surv. Ann. Rept 1964, 55—65.

— **1966.** Second progress report on the Brockman Iron Formation in the Wittenoom—Yampire area. West. Australia Geol. Surv. Ann. Rept 1965, 75—87.

— **1968 a.** Three great basins of Precambrian banded iron formation deposition: a systematic comparison. Geol. Soc. America Bull. 79 (11), 1527—1544.

— **1968 b.** The Joffre Member in the gorges south of Wittenoom. West. Australia Geol. Surv. Ann. Rept 1968, 53—57.

— **1972.** Revolution in earth history. Geol. Soc. Australia J. 19, part 3, 287—311.

— **1973 a.** Iron-formations of the Hamersley Group of Western Australia: type examples of varved Precambrian evaporites. *In* Unesco, genesis of Precambrian iron and manganese deposits. Proc. Kiev Symp., 1970.

— **1973 b.** Varve cycles in the Weeli Wolli Formation of the Precambrian Hamersley Group, Western Australia. Econ. Geol. 68, (7) 1089—1097.

— **1975 a.** Hamersley Basin. *In* Geology of Western Australia. West. Australia Geol. Surv., Mem. 2, 118—141.

— **1975 b.** The Hamersley Basin — regional geology. *In* Economic Geology of Australia and Papua-New Guinea. Australasian Inst. Mining Metall., Monograph 5, 411—413.

— **1975 c.** Geology of Western Australian Iron Ore. *In* Economic geology of Australia and Papua-New Guinea. Australasian Inst. Mining Metall., Monograph 5, 883—892.

— **1976 a.** Geology of the Hamersley Basin. 25th Intern. Geol. Congress, Sydney, Australia, Excursion Guide No. 43A.

— **1976 b.** Striated and faceted boulders from the Turee Creek Formation — evidence for a possible Huronian glaciation on the Australian continent. West. Australia Geol. Surv. Ann. Rept 1975, 88—92.

— **1979.** A revision of the Mount Bruce Supergroup. West. Australia Geol. Surv. Ann. Rept 1978.

Trendall, A. F. & Blockley, J. G., 1968. Stratigraphy of the Dales Gorge Member of the Brockman Iron Formation, in the Precambrian Hamersley Group of Western Australia. West. Australia Geol. Surv. Ann. Rept 1967, 48—53.

— **1970.** The iron formations of the Precambrian Hamersley Group, Western Australia, with special reference to the associated crocidolite. West. Australia Geol. Surv. Bull. 119.

Trendall, A. F. & Pepper, R. S., 1977. Chemical composition of the Brockman Iron Formation. West. Australia Geol. Survey Rec. No. 1976/25.

Ward, D. F., Coles, I. G. & Carr, W. M. B., 1975. Jimblebar and Western Ridge iron ore deposits, Hamersley Iron Province. *In* Economic geology of Australia and Papua-New Guinea. Australasian Inst. Mining Metall. Monograph 5, 916—924.

Vander Wood, T. B., 1977. Strontium isotope systematics in the Hamersley Range. Florida State University, M. Sc. Thesis.

CALEDONIAN-APPALACHIAN STRATABOUND SULPHIDES

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Introduction

In character and scope, CCSS is much more restricted and specialised than is project 91, which has taken upon itself the whole time — and space — span of the Precambrian areas of the earth — a formidable task indeed!

CCSS is concentrating on a much narrower time interval — that of the late Precambrian to perhaps middle Palaeozoic — as exemplified in a single (though now fragmented) mobile belt; that of the Caledonian-Appalachian orogen in northwestern Europe and North America. This represents a considerable feature of the earth's crust, all the same, with a total present length of some 12 000 kilometres from Svalbard to Alabama — as illustrated in Fig. 1. It will also be apparent that the region involved encompasses a number of political units on both sides of the present Atlantic Ocean, so that the project is truly international in aspect. In fact, the shaded area in Fig. 1 includes parts of seven countries which constituted the nucleus of CCSS as it was originally conceived.

In addition two other countries on the southeastern margins of the orogen proper are members of the project and it is hoped others will join in the near future.

So much for the geographical and geopolitical aspects of CCSS; metallogenetically, too, the scope of the project has been deliberately restricted, in that it is dealing only with stratabound sulphide deposits within the Caledonian-Appalachian orogen — this in spite of the fact that there are other, important oretypes present, both non-sulphide and non-stratabound. This restriction has been built into the project with good reason, it is felt. For the first it makes for a much more »manageable» project. Secondly, the »Correlation» aspect of IGCP is more easily handled in the case of stratabound mineralization, and correlation in time and space, in environments of formation and in periods of deformation and metamorphism, is a strong aspect of the project.

To conclude this introduction to CCSS, its »official» objectives, as originally transmitted to IGCP, may be cited: »The correlation of data on the geological environments and characteristics of the Late Precambrian and Early Palaeozoic stratabound base-metal sulphide deposits in the Caledonian-Appalachian orogenic belt, (with special emphasis on their stratigraphical and lithological associations, structural settings and the subsequent tectonic/metamorphic modifications) with a view to elucidating their genesis, predicting further ore occurrences and facilitating the exploration and development of individual prospects».

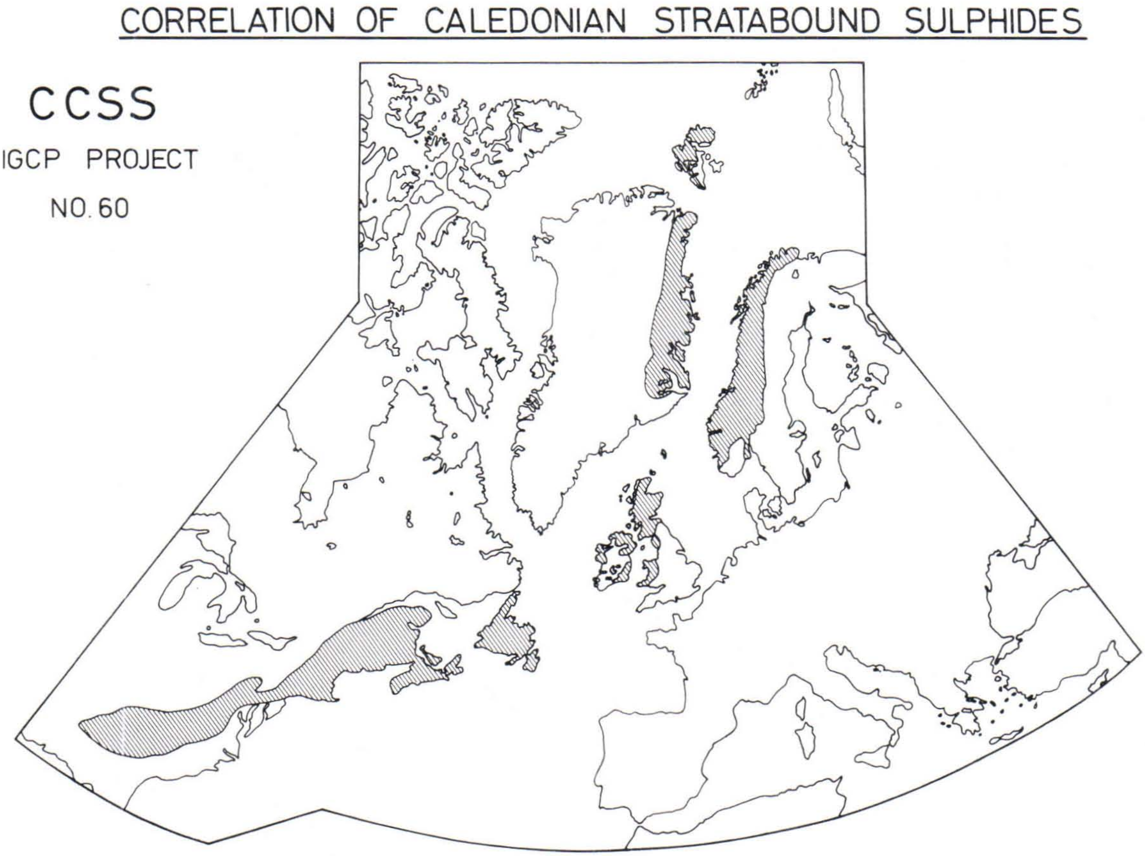

Fig. 1. Predrift reconstruction of the Caledonian—Appalachian orogen (shaded).

Progress to date

A wide range of scientific problems is involved in a project of this type, but, in keeping with the objectives just cited, the principal areas to which attention is being given at the present stage of the project, are;

1) The stratigraphical positions of the various categories of stratabound sulphides within the Eocambrian-Palaeozoic successions in different regions of the project area, with a view to correlating the ore-forming events in both time and space.
2) Lithological settings of the sulphides with a view to determining the palaeographical and structural environments of sulphide deposition and correlating these with the type and, if possible, the importance, of the resulting deposits.
3) The subsequent tectonic/metamorphic modifications to which the deposits have been subjected, with the object of coming to a greater understanding of the characteristics of the ore bodies, e. g., their morphology, mineralogy and texture.

Up to now, it has not been possible to give equal emphasis to all these objectives due to problems of financing and due to the varying interests of participants. On the other hand, several projects include elements of all three areas, a fact which makes it difficult to assess progress in any particular one of them.

The lines of research being pursued involve most geological, mineralogical and geochemical techniques; from regional and detailed mapping of mineralized areas or individual deposits, through petrological/geochemical investigations of the host lithologies, to detailed mineralogical, fluid inclusion and isotopic investigations of the ores.

Categories of research

Field and laboratory studies of problems within the scope of CCSS constitute the main effort of the project and are provding new, basic knowledge of the Caledonian stratabound sulphides and their geological environments. Some of these studies have been in progress for approximately three years and preliminary results are beginning to appear. A summary review of some of the results so far achieved is given below.

Geological and geochemical studies of host-rock lithologies

A) Sediment-hosted deposits. Sediment-hosted stratabound sulphides are, of course, abundant in many countries within the project region. They occur at most stratigraphical levels within the Late Precambrian and Palaeozoic successions involved, but are definitely of greater relative importance at some levels than at others. In Scandinavia and Greenland the earlier, Eocambrian to Cambrian, sedimentary sequences are of special significance and much effort has been directed towards these sequences in the northeastern part of the project area. Parallel projects have been in progress on these deposits both in Norway and Sweden and are providing us with new and more detailed knowledge regarding the lithological environments of important Pb-(Zn) ores in this area. Fig. 2 shows the general paleogeographical setting of deposits of the lead-in-arenite type in this region.

One sub-project, terminated at the end of 1977 after 3 years of funding, has involved field work ranging from western Norway at latitude 61°N, and has involved mapping of type-profiles and sampling for geochemical analysis. The work has shown that the sedimentary sequence hosting the Pb-(Zn) sulphides was deposited partly under tidal water conditions (evidenced by scour channels, desiccation cracks, etc) and partly as more homogeneous subtidal deposits. These facies types alternate and form several transgressive and regressive sequences, some of which are illustrated by the accompanying section from the Osen area (Fig. 3) (Bjørlykke 1977).

A parallell subproject has been concerned with the autochthonous sulphide-bearing lithostratigraphic unit, 10—100 m in thickness, which is traceable along the whole of the eastern margin of the Swedish Caledonides. Preliminary interpretation of the sedimentary environments of deposition of the unit suggests the interaction of fluvial, deltaic, tidal and shallow marine deposition. Regressive sedimentary cycles predominate, while the formation as a whole reflects an overall, easterly directed transgression.

The work in Greenland, carried out by the Danish CCSS group, has been aimed at detecting the presence of stratabound mineralization in a hitherto uninvestigated area of the Caledonides where large thicknesses of Late Precambrian

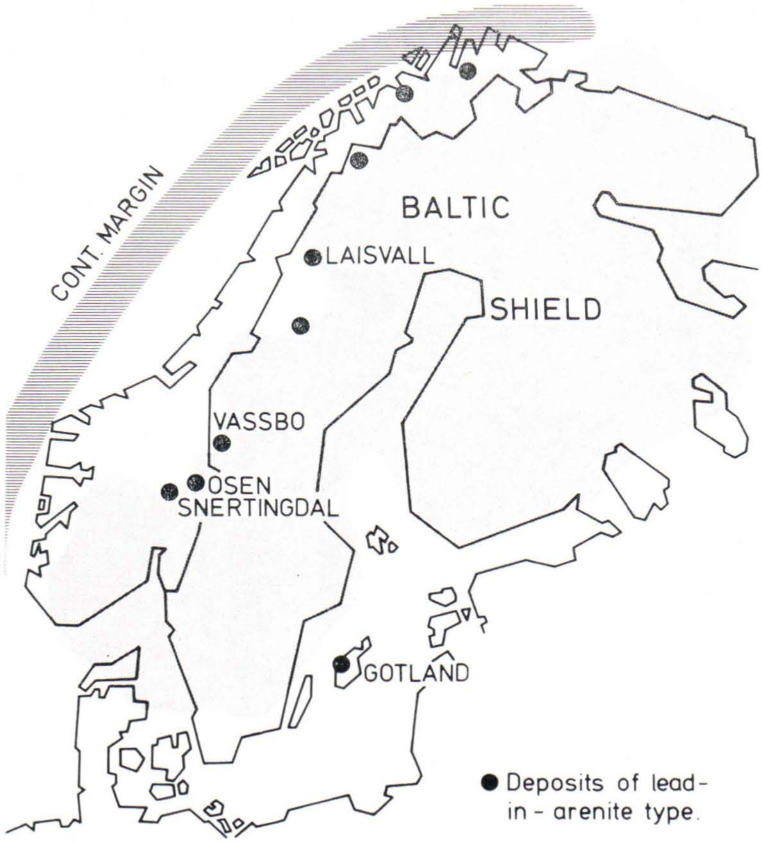

Fig. 2. Sketch map of the Scandinavian region showing a paleogeographic reconstruction for Lower Cambrian times and the location of deposits of the lead-in-arenite type (Bjørlykke 1977).

and Early Palaeozoic sedimentary rocks occur. This work will be referred to in more detail below.

Within the Lower Palaeozoic sedimentary sequences of the Caledonian-Appalachian orogen a very important type of stratabound sulphide deposit consists of zinc and lead sulphides in carbonate host rocks. (Often referred to as Mississippi Valley type deposits). Although these are of greater importance in North America than in Europe, they are known east of the Atlantic, and work has begun to investigate mineralization of this type in the central Caledonides in Northern Norway. Where later metamorphism has not obscured the relationships, it can be seen that the sulphides occur in primary and diagenetic structures within a dolomitic lithology. Solution collapse breccias have acted as host structures at at least one deposit. Fossil evidence at one locality indicates a Llandovery age for the dolomite, which agrees well with an earlier Pb/Pb isotopic age of 420±90 Ma on galena from the deposits (Moorbath and Vokes 1963).

At the younger end of the Caledonian successions, Canadian workers have included successor basins in their studies. Considerable progress has been made in relating mineral occurrences

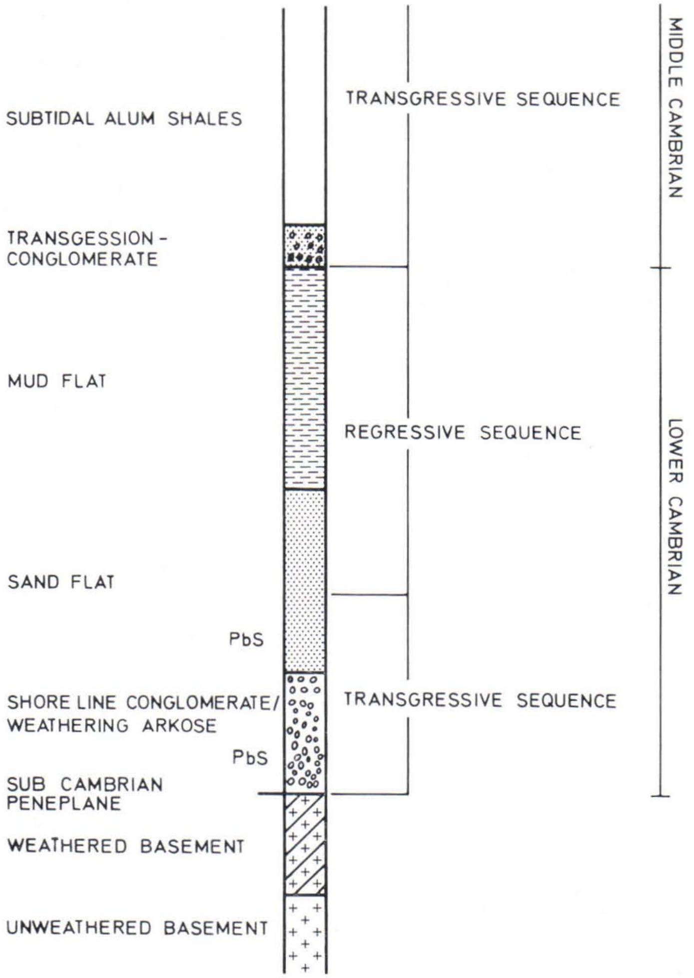

Fig. 3. Interpretation of drill core from the Osen lead-in-arenite occurrence showing the depositional sequences.

to restricted stratigraphical intervals related to a maiine transgression-regression cycle.

B) Volcanic and volcanic/sediment-hosted deposits. These constitute the most common and ubiquitous type of Caledonian-Appalachian stratabound sulphide deposits (»Massive» sulphides, Kieslagerstätten, etc). Thus it is only natural that deposits of this type are being paid considerable attention by CCSS workers in most of the countries participating in the project. Studies of the probable paleo-geographical and global tectonic environments of formation of the volcanic members hosting deposits of this type have made considerable progress in several countries, especially in Scandinavia.

Projects in both Norway and Sweden have been concerned with the geochemistry of greenstone metavolcanics hosting polymetallic, often massive, sulphide ores in the central Caledonian belt. The aims of the projects have been to find possible correlations between metavolcanic sequence characteristics and types and distribution of the contained sulphides. Preliminary results seem to indicate that such correlations do exist.

The massive sulphides occurring in the Cambrian—Early Ordovician metabasaltic series of the western Trondheim area seem to be related to a certain level within the volcanic sequence, being overlain by more differentiated metabasalts than the typical primitive ocean floor tholeiitic metabasalts which occur below the sulphide horizon. Massive sulphide deposits in the eastern Trondheim district are more numerous than in the west and show environments of island arc affinity. In addition, these eastern deposits are characterized by higher abundances of Zn and Pb and show different ore body morphologies. (Grenne 1977). Similar results are being obtained on the geochemistry of the metavolcanic and metaintrusive rocks associated with stratabound polymetallic sulphides in the low-grade Köli nappe sequence in northern Jämtland and Västerbotten, Sweden. Metamorphosed volcanic and subvolcanic rocks occur at four different structural levels in the sequence. Stratabound, partly massive, Cu- and Zn-bearing sulphides are most conspicuous in association with mixed acidic and basic volcanic rocks of the Ashgillian Tjopasi group and the calcareous metagreywackes and mixed Stekenjokk volcanics of Silurian age.

The ore-bearing volcanites consist of basalts and andesites, altered to spilites, as well as much siliceous material now in the form of quartz keratophyre. The basalts and andesites are mildly tholeiitic, with low TiO_2 and incompatible element contents; they are reminiscent of island arc tholeiites. Silurian basic intrusives within the sequence are largely nonspilitic and totally basaltic in composition. It is believed they represent rifting of the island arc sequence in the Silurian (Stephens 1977).

A project of the U. S. Geological Survey to study stratabound sulphide deposits in the Appalachian Mountain belt was started two years ago under the stimulus of IGCP/CCSS. In this project, extensive sampling of ores and host rocks has been completed over a 400-mile-long belt in Virginia, North Carolina, and Tennessee. Compositional data on major elements, trace elements, and rare earth elements, are being obtained for volcanogenic and sedimentary sulphide ores and wall rocks. Evidence for volcanic influences in the nominally sedimentary environment of the Blue Ridge province has been sought and found in the field; the chemical analytical work in progress should provide additional support for the field findings. Perhaps of greater importance, the chemical data will provide the first comprehensive and comparative chemical base level for the sulphide-bearing rocks of the southern Appalachian belt.

Mineralogical Studies

An investigation of the mineralogy, both ore and gangue, of stratabound sulphides is obviously an integral part of many studies now in progress under the auspices of CCSS. In addition, certain research groups have taken up special aspects of the sulphide mineralogy, and are using refined techniques towards the solution of particular problems, such as metamorphic deformation and recrystallisation, geothermometry/geobarometry. Other groups are tackling mineralogical problems of a more directly applied nature, such as the benefication characteristics of the ores and the mineralogical modes of occurrence of their gold and silver values.

A French group within CCSS initiated a particularly interesting project concerned with the textures of sulphide ores in the Norwegian Caledonides. This work is based on a micrographic etching method of studying pyritic mineralization which allows the evolution of sulphide ores during metamorphism and tectonism to be followed in detail.

The first results on Norwegian ores are promising, but more studies will be necessary, initially at least, in collaboration with Norwegian geo-

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logists. However, numerous Cambrian and Ordovician ores have been studied from the margins of the Caledonide zone in western Spain. These are located in different metallogenic districts and in different metamorphic environments (greenschist to granulite facies, as well as at granite contacts). They have been studied by different methods; classical geology, geochemistry, sulphur isotopic studies, as well as by the micrographic etching method.

Apart from their regional applications, these results are providing an excellent basis for further studies on Caledonian ores located in comparable metamorphic and structural environments.

Material for geobarometrical studies on coexisting sphalerite/pyrite/pyrrhotite assemblages from a number of Swedish Caledonian polymetallic sulphides has been sent to workers at the University of Toronto, Canada. Preliminary data for material from the Stekenjokk and Ankarvattnet deposits show that the sphalerite is very homogenous in composition. An equilibration pressure of 6.5 Kbars is indicated for the Ankarvattnet deposit.

Ore deposit studies

Much of the research presently being carried out under the project is in the form of detailed studies of individual deposits of stratabound sulphides, involving more or less all aspects of their geology, mineralogy and geochemistry. Studies of this type are in progress in practically all the participating countries and results of these studies are already appearing in the scientific literature.

It is not possible here to refer in detail to all the deposit studies being carried out under CCSS auspices, but mention may be made of some of the results now appearing in the scientific press, e. g. Halls, *et al.* (1977) and Reinsbakken (1977) on the Skorovass ore and Olsen (1978) on the Joma deposit, both in Norway; Juve (1977 a, b) on the Stekenjokk ore, Sweden; Garson and May (1976) on the Vidlin, Shetland, occurrence; Platt (1977) and Badham (1978) on the Avoca ores, Ireland and Henry, *et al* (1977) (1978) on the great Gossan Lead, Virginia.

Applied studies (Prospecting techniques)

One of the major aims of a project of this nature must be to improve the success of efforts made to find and assess further mineral resources within the Caledonides, and within similar types of geological terrain elsewhere. Obviously, all geological knowledge obtained will, ultimately, have a bearing on this aim, but a limited number of direct applications of prospecting techniques in different environments have formed part of the project's activities.

The Danish group within the CCSS, based at the University of Copenhagen, carried out field work in the Caledonides of east Greenland in the summers of 1974 and 1975 and has followed this up with laboratory studies of samples collected in the field. The work has been directed towards determining the presence of stratabound mineralisation in the Late Precambrian and Early Palaeozoic sedimentary succession of the area, using mineralogical and chemical analyses of stream sediments and rocks. A total of some 17 000 m of strata has been investigated; comprising the Lower and Upper Eleonore Bay groups, and the Tillite Group (Precambrian) as well as the overlying Cambrian and Ordovician succession (See Table 1).

The results achieved so far enabled certain generalizations to be made regarding the occurrences of heavy metals in the strata of the region, occurrences which were quite unknown prior to the project. Of particular interest is the presence of copper mineralisations on the Quartzite Series and Multicoloured Series of the Upper Eleonore Bay Group, found in the field and confirmed by emissio nspectrography of the nonmagnetic fractions of heavy mineral concentrates and by AA spectrochemistry of fine-grained stream sediments and rock samples.

Table 1.
Schematic representation of the Latest Precambrian—Lower Paleozoic succession in East Greenland (after Urban *et al.* 1978).

PALAEOZOIC	Ordovician	Middle	225 m
		Lower	700 m
	Cambrian	Middle and/or Upper	580 m
		Lower	270 m
PRECAMBRIAN	Tillite Group	Spiral Creek Formation	25 m
		Canyon Creek Formation	250 m
		Upper Tillite	140 m
		Inter Tillite	340 m
		Lower Tillite	100 m
	Upper Eleonore Bay Group	Limestone—Dolomite Series	1 250 m
		Multicoloured Series	1 170 m
		Quartzite Series	2 270 m
	Lower Eleonore Bay Group	Upper Argillaceous—Arenaceous Series	1 350 m
		Calcareous—Argillaceous Series	150 m
		Lower Arenaceous—Argillaceous Series	8 050 m

It is concluded from the work done so far, that for geochemical prospecting purposes in the region, the use of heavy mineral concentrates is to be preferred, since the analytical results on this material show significant variations for a great number of elements. Fine grained sediments, on the other hand, are more homogenized and do not, except under special circumstances, reveal any distinct anomalies. Although the use of rock samples raises the problem of obtaining really representative samples of the strata investigated, rock analyses have proved to be significant for determining anomalies within certain levels showing enhanced metal contents.

Prospecting for stratabound sulphides, using geochemical, geophysical and geological techniques, has constituted the main activity so far undertaken by CCSS-associated groups in the U. K.

The Institute of Geological Sciences recently completed an initial 5-year Mineral Reconnaissance Programme, funded by the Department of Industry, in this field. Several reports relevant to CCSS have been completed and made accessible for consultation by interested workers.

Due to the relative success of the original programme in locating the presence of stratabound mineralization in Scotland, the contract for the Mineral Reconnaissance Programme has been renewed and a team formed specifically to deal with this type of deposit in Caledonian terrain in the U. K. One of the deposits, at Vidlin in Shetland, has recently been investigated by a mining company. (See Garson and May 1976).

In Ireland, too, the emphasis of prospecting activity for stratabound sulphides, undertaken both by private companies and the Geological Survey of Ireland in the Wicklow-Waterford area, has been influenced by the interest stimulated by CCSS concepts.

Isotopic and other special studies

A limited number of isotopic and other, special, studies are being carried out under the project. In Sweden, these studies are concerned with deposits both along the Caledonian Front and in the metamorphic Caledonides. Work on Laisvall sandstone-hosted Pb-Zn deposit is multidisciplinarian in nature and includes studies of palaeography of the Laisvall area and fluid inclusions in sphalerite and calcite contained in the deposit. The latter indicates that the mineralization occurred at a temperature of approximately 150°C from a concentrated Na-Ca-Cl brine type of solution. (Rickard *et al.* in press).

Within the metamorphic Caledonides, type deposits have been selected within the Seve and from the sediment-hosted and volcanite-hosted deposits in the Köli. Apart from detailed chemical and mineralogical work, these studies so far include sulphur-isotope and geobarometric work.

Conclusion

To conclude it would seem that CCSS by investigating an orogen of the type constituted by the Caledonides/Appalachians (one which is not too ancient to make the application of plate tectonic notions of doubtful value), stands a good chance of arriving at results which will be applicable, not only to sulphide-bearing orogens of a similar age, but also, hopefully, to the more ancient orogens which are the concern of Project 91 where the relations to presentday plate boundaries are not so clear.

Acknowledgements

Those involved in IGCP Project No. 60, CCSS, are grateful to their colleagues of Project 91 for being invited to provide this review for the present volume. The author is indebted to his colleagues in CCSS who actually carried out the work reported here and he apologises for the undoubtedly imperfect coverage of their activities.

References

Badham, J. P. N., 1978. Slumped sulphide deposits at Avoca, Ireland, and their significance. Inst. Min. Metall. London. Trans. 87, Sect. B, 21—26.

Bjørlykke, A., 1977. Om blyforekomster i eokambriske-kambriske sandsteiner på det Baltiske skjold. *In* Kaledonske Malmforekomster, ed. by A. Bjørlykke, I. Lindahl, and F. M. Vokes, Trondheim, BVLI's Bergforskning, 72—75.

Garson, M. S. & May, F., 1976. Copper mineralization at Vidlin, Shetland. Inst. Min. Metall., London, Trans. 85, Sect. B, 153—157.

Grenne, T., 1977. Metavulkanitt-geokjemi i Fundsjøgruppen/Hersjø-Formasjonen og Støregruppen, østlige og vestlige del av Trondheimsfeltet. *In* Kaledonske Malmforekomster ed. by A. Bjørlykke, I. Lindahl and F. M. Vokes. Trondheim, BVLI's Bergforskning, 62—63.

Halls, C., Reinsbakken, A., Ferriday, I., Haugen, A. & Rankin, A., 1977. Geological setting of the Skorovass ore body within the allochthonous volcanic stratigraphy of the Gjersvik Nappe, Central Norway. *In* Volcanic processes in ore genesis. London, Inst. Min. Metall., 128—151.

Henry, D. K., Staten, W. T., Craig, J. R., Gilbert, M. C. & Hewitt, D. A., 1977. Ore minerals and silicate gangue variations in the Great Gossan Lead, Virginia. Trans. Am. Geophys. Union, 58 (6), 524.

Henry, D. K., Craig, J. R. & Gilbert, M. C., 1979. Ore mineralogy of the Great Gossan Lead. Econ. Geol. 74 (3), 645—656.

Juve, G., 1974 a. Metal distribution at Stekenjokk: Primary and metamorphic patterns. Geol. Fören. Stockh., Förh. 99, 149—158.

Juve, G., 1977 b. Paragenesis — »metaparagenesis» in the Stekenjokk deposit, central Scandinavian Caledonides, Sweden. Procs. 4th. IAGOD Symposium, Varna 1974, (in press.).

Moorbath, S. & Vokes, F. M., 1963. Lead isotope abundance studies on galena occurrences in Norway. Norsk geol. tidsskr., 43, 283—343.

Olsen, J. (in press). The Joma Deposit, central Scandinavian Caledonides. 5th IAGOD Sympos., Alta Utah. Procs.

Platt, J. W., 1977. Volcanogenic mineralization at Avoca, Co. Wicklow, Ireland and its regional implicarions. *In* Volcanic processes in ore genesis. London, Inst. Min. Metall., 163—170.

Reinsbakken, A., 1977. Geology of the Skorovas massive sulphide ore body. *In* Kaledonske Malmforekomster, 37—50.

Rickard, D., Willden, M. Y., Marinder, N.-E. & Donnelly, T. H., 1978. Studies on the genesis of the Laisvall sandstone lead-zinc deposit, Sweden. Parts I—IV, Econ. Geol. 74 (5), 1255—1285.

Stephens, M. B., 1977. The Stekenjokk volcanites — segment of a Lower Paleozoic Island Arc complex. *In* Kaledonske Malmforekomster, 24—36.

Urban, H., Ghisler, M., Jensen, A. & Stendal, H., 1978. Denmark. *In* CCSS Newsletter No. 3, ed. by F. M. Vokes. Trondheim, 6—12.

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