Geological Survey of Finland

Bulletin 292

On the geology and geochemistry of the Precambrian iron formations in Väyrylänkylä, South Puolanka area, Finland

by Kauko Laajoki and Risto Saikkonen

Geologinen tutkimuslaitos · Espoo 1977

ERRATA

Fe_2O_3-line in Table 8 on p. 86 should run as follows:

	MOSF			CF			MSOF	SF	Whole formations		
……	……	……	……	……	……	……	……	……	……	……	……
Fe_2O_3[a])	38.96	40.02	39.06	37.99	34.84	37.33	44.87	27.78	38.86	37.93	38.75
……	……	……	……	……	……	……	……	……	……	……	……

Geological Survey of Finland, Bulletin 292

ON THE GEOLOGY AND GEOCHEMISTRY OF THE PRECAMBRIAN IRON FORMATIONS IN VÄYRYLÄNKYLÄ, SOUTH PUOLANKA AREA, FINLAND

BY
KAUKO LAAJOKI and RISTO SAIKKONEN

WITH 73 FIGURES AND 19 TABLES IN THE TEXT AND ONE APPENDIX

GEOLOGINEN TUTKIMUSLAITOS
ESPOO 1977

Laajoki, K. and Saikkonen, R. 1977: On the geology and geochemistry of the Precambrian iron formations in Väyrylänkylä South Puolanka area Finland. *Geological Survey of Finland, Bulletin 292.* 137 pages, 73 figures, 19 tables, one appendix.

The Middle Precambrian (2080 Ma) Väyrylänkylä iron formations, located in northeastern Finland, are of Superior type. They are associated with dolomites, phyllites and black schists of the Marine Jatulian of the Karelidic sequence. The iron formations are composed mainly of banded rocks of the mixed oxide-silicate facies (MOSF) and the carbonate facies (CF). Also present are rocks of the sulphide facies (SF), the silicate facies proper (SFP) and weathered iron-formation rocks of the mixed silicate-oxide facies (MSOF).

A typical rock of the MOSF is a quartz-magnetite-banded rock rich in amphibole and that of the CF a quartz-siderite-banded rock . Both rocks contain phosphorite, noncherty garnet-bearing iron-silicate and pyrrhotite-rich black-schist interbands. The rocks of the SF are iron-rich black schists and phyllites. Typical of the SFP is a chert-mesobanded grunerite rock.

The rocks are intensely deformed, and underwent regional metamorphism under the conditions of amphibolite facies. Nevertheless, primary structures and low-grade metamorphic textures abound. Also retrograde mineral formations are often visible.

Altogether 19 total and 20 partial silicate analyses from different types of iron-formation rocks are given. Trace elements were analysed from 15 rocks and a rare-earth element analysis was made from a garnet-bearing interband. Also given are 23 microprobe mineral analyses. Partial ore analyses of drill cores totalling about 500 metres are also reported. In comparison with iron formations of Superior type in general, the Väyrylänkylä formations are distinguished by exceptionally high P_2O_5 ($\sim$ 2.5 %) and CaO ($\sim$ 3.5 %) due to ubiquitous marine apatite, as well as by comparatively high S ($\sim$ 2 %) and C ($\sim$ 1 %) due to frequent iron-rich metapelite interbands.

The origin of the iron formations is discussed and it is suggested that their extensive development about 2100 Ma ago can be attributed to the specific orogenesis of that time.

ISBN 951-690-075-5

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PREFACE

In 1973 the Geological Survey of Finland carried out prospecting on some small Precambrian (Karelian) iron-formation occurrences in Väyrylänkylä, South Puolanka area, NE Finland (Fig. 1). Owing to their small dimensions, poor grade ($Fe_{tot} \sim 27\ \%$) and detrimentally high phosphorus content ($P_2O_5 \sim 2.5\ \%$) the deposits proved to be uneconomic (Geological Survey of Finland 1974, pp. 22—26; Ervamaa and Laajoki 1977).

In the present paper main geological and geochemical data of the Väyrylänkylä deposits are given and their genesis is discussed from a geochemical point of view. The geochemistry of the rare-earth elements and Pb isotopes in these iron formations were treated in separate papers (Laajoki 1975 b, Sakko and Laajoki 1975). Risto Saikkonen of the Chemistry Department is responsible for most of the total silicate analyses included in the study. He started his part of the work in the winter of 1974. During prospecting, the drill-core material of the iron formations was systematically analysed for Fe_{tot} (total iron), Fe_{HCl} (iron soluble in hydrocloridic acid), Ti, Mn, V, S and P in the Ore Laboratory of the Chemistry Department, Geological Survey of Finland. These analyses are referred to in this paper as ore analyses. The sampling and processing of the analytical results were done by Kauko Laajoki, who alone is responsible for the geological background of the study and for the conclusions drawn in it. He also compiled the manuscript and, unless otherwise stated, took the photographs.

The iron formations in Väyrylänkylä were chosen as an object of detailed prospecting after Laajoki (1973) had completed his general geological study of the South Puolanka area. In connection with this work abundant new stratigraphical data were revealed, thanks to supplementary drilling and detailed mapping. These data, most of which are unpublished, are included in the present study mainly in the form of definitions, maps and profiles.

INTRODUCTION

Iron-rich sedimentary rocks, containing 15 % or more iron of primary (depositional or diagenetic) origin are generally divided into two main subgroups: iron formations and ironstones (e.g. James 1966). The former are typically chert-banded and Precambrian in age. Ironstones, on the other hand, tend to be noncherty, oolitic and of post-Precambrian age.

Lepp and Goldich (1964) approached the origin of the Precambrian iron formations of the Canadian Shield with the aid of a statistical study of their chemical composition. They noted that these occurrences have a surprisingly uniform iron tenor and low contents of Al_2O_3, TiO_2 and P_2O_5. Within one district the iron contents of different facies were surprisingly uniform. In comparison with the post-Precambrian ironstones the Precambrian iron formations were noted to have low contents of Al_2O_3, TiO_2, P_2O_5 and, especially, CaO. However, as pointed out by Trendall (1965 a), the geological features of these two subgroups of iron-rich metasediments are so different that little can be gained from geochemical comparisons between them. Govett (1966) took this fact into consideration when he divided the Precambrian iron formations into banded and oolitic types and compared their chemical composition in terms of $SiO_2 + Al_2O_3 + Fe_2O_3$ with post-Precambrian iron formations (ironstones). He noted that although no post-Precambrian iron formation of any type (non-oolitic, oolitic and recent lake and bog iron formations) falls within the compositional field of the Precambrian banded iron formations, the composition of Precambrian oolitic iron formations is clearly similar to that of post-Precambrian oolitic iron formations.

In his excellent summary of the chemistry of the iron-rich sedimentary rocks, James (1966) verified that typical ironstone (minette type) has no distinctive chemical aspects other than its high content of iron. On the other hand, typical iron formation (Superior type) has a remarkably low content of Na, K, Al, and minor elements; the phosphorus content is generally much lower than in ironstones. Taylor (1969) considered element ratios more significant than absolute percentages for comparison and noted that Precambrian iron formations have, in general, high Fe/Al and low Ca/Mg ratios, whereas the British Jurassic ironstones have the reverse.

Nowadays it is almost generally accepted that iron formations are originally (ortho)chemical sediments. Some authors are, however, disposed to support diagenetic replacement hypothesis (Kimberley 1974, Dimroth and Kimberley 1976). Unanswered questions such as the source and method of transport of iron and silica, depositional environments and importance of diagenetic and metamorphic processed

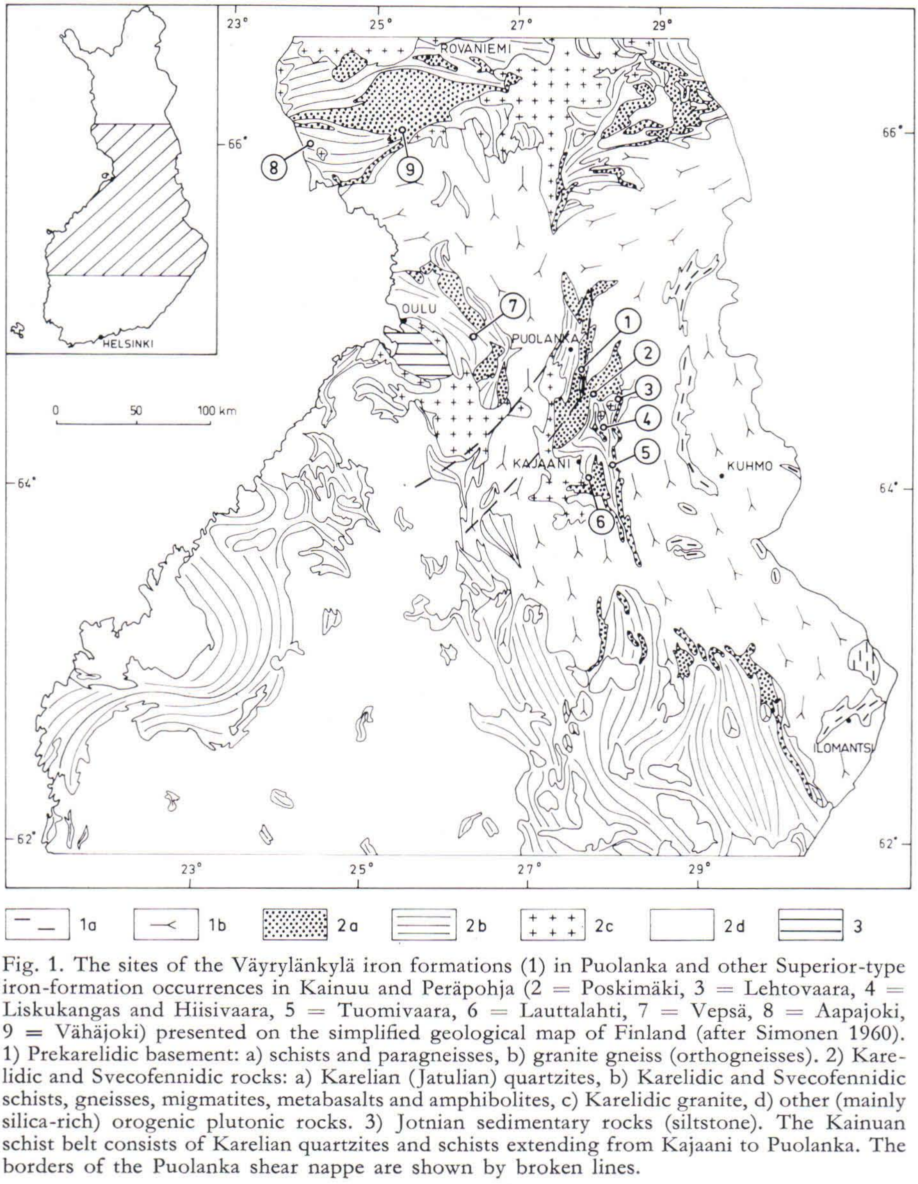

Fig. 1. The sites of the Väyrylänkylä iron formations (1) in Puolanka and other Superior-type iron-formation occurrences in Kainuu and Peräpohja (2 = Poskimäki, 3 = Lehtovaara, 4 = Liskukangas and Hiisivaara, 5 = Tuomivaara, 6 = Lauttalahti, 7 = Vepsä, 8 = Aapajoki, 9 = Vähäjoki) presented on the simplified geological map of Finland (after Simonen 1960). 1) Prekarelidic basement: a) schists and paragneisses, b) granite gneiss (orthogneisses). 2) Karelidic and Svecofennidic rocks: a) Karelian (Jatulian) quartzites, b) Karelidic and Svecofennidic schists, gneisses, migmatites, metabasalts and amphibolites, c) Karelidic granite, d) other (mainly silica-rich) orogenic plutonic rocks. 3) Jotnian sedimentary rocks (siltstone). The Kainuan schist belt consists of Karelian quartzites and schists extending from Kajaani to Puolanka. The borders of the Puolanka shear nappe are shown by broken lines.

constitute major problems that must be solved before the origin is fully understood (James and Sims 1973). From the geochemical point of view, it is worth emphasizing that »the size, position, and duration of basins of deposition, the dimensions of the deposits, and the facies relations both within the iron-rich sedimentary rocks and other rocks formed contemporaneously — all these are of importance to the geochemistry of iron sedimentation, particularly with respect to problems of genesis» (James 1966,

p. W 28). Moreover, iron formations, as discussed in the present paper, can be classified into different types on the basis of stratigraphic associations: e.g. the Algoma and Superior types of Gross (1965) and the volcanic, clastic and chemical stratigraphic associations of Beukes (1973), for which different sources of iron and silica may be postulated. Of great importance with respect to the general time-distribution of iron formations and, especially, to the magnitude of their development about 2100 Ma ± 200 Ma ago is their relationship with post-Precambrian ironstones.

Finland long lacked specific studies on iron formations of pure Superior type, primarily because these formations are comparatively rare and of no economic value in this country, but partly because the scientific significance of these formations was not realized until recently (see footnote). There are two papers which contain descriptions of skarn iron-ores whose origin may well be of Superior type (Borgström 1928, Mikkola 1947). Lately, however, numbers of papers have been published concerning Superior-type occurrences in Finland (Lehto 1975, Niiniskorpi 1975, Mäkelä 1976, Hiltunen and Tontti 1976, Sakko and Laajoki 1975, Laajoki 1975 b). The papers by Lehto and Niiniskorpi include many new chemical analyses of whole-rock samples as well as some published earlier. They also give microprobe analyses of different minerals from known Superior- and Algoma-type iron-formations in eastern and northern Finland. The most valuable of the earlier papers concerning Algoma-type occurrences are those by Hackman (1925), von Knorring (1955), Kaitaro (1949), Rieck *et al.* (1967) and Paakkola (1971). In his paper, von Knorring gives chemical data on the different minerals and whole-rock analyses of the iron-formations of Southern Finland intimately associated with acid (leptite) volcanism.

General definitions

Iron formation [1]*:* In accordance with existing general definitions (James 1954, Lepp and Goldich 1964), iron formation is defined as a rock-stratigraphic unit of mainly chemical sediment typically bedded and commonly laminated, containing 15 % or more iron of sedimentary origin, commonly but not necessarily containing layers of chert. Iron formations associated closely with quartzite, dolomite and black schist and those associated or interbedded with volcanic rocks are considered to represent pure Superior and Algoma types, respectively (*cf.* Gross 1965).

Iron-formation rock: A rock in iron formation containing at least 15 % total iron.

Iron-formation facies: On the scale of macrobands the iron-formation rocks are

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[1] The first Finnish geologists to emphasize the importance of recognising of the concept of banded iron formations were Nuutilainen and Mäkelä (1973). Before that, various names, e.g. iron ore, magnetite-banded quartzite, magnetite-rich phyllite, etc., were used for the Karelian iron formations. This diversity of terms can, at least in Kainuu, be partly attributed to the mixed nature of the iron formations. The similarity of the Kittilä iron ore-bearing quartzites to the iron formations of the Itabira, Krivoy Rog and Lake Superior areas was detected, however, by Hackman as early as 1925. Later, Kaitaro (1949) called them jasper quartzites and Paakkola (1971) manganiferous iron formation.

classified on the basis of their present mineral constituents by different mineralogical facies as defined below.

Mixed oxide-silicate facies (MOSF): An iron-formation rock, in which iron is mainly fixed in oxides, but which also contains an abundance of iron-silicates. This term replaces the term oxide facies used in earlier publications (Laajoki 1975 a, b, 1976 a, Sakko and Laajoki 1975).

Mixed silicate-oxide facies (MSOF): Weathered (oxidized) iron-silicate- and magnetite-rich rocks of the Pääkkö and Iso Vuorijärvi iron formations are classified by the facies of their own named the mixed silicate-oxide facies. It seems probable that this facies is a secondary (post-metamorphic) variety of the MOSF. This term is used instead of the term silicate facies employed in the earlier publications (see MOSF).

Carbonate facies (CF): An iron-formation rock with iron fixed mainly in carbonates. The term was used in the same sense in the earlier publications.

Silicate facies proper (SFP): An iron-formation rock, in which iron is fixed almost solely in iron silicates and which is distinguished by chert mesobands. This facies was not described in the earlier publications.

Sulphide facies (SF): An iron-formation rock with iron fixed mainly in iron sulphides.

Iron formation proper: Of the iron-formation rocks in Väyrylänkylä those of MOSF, (MSOF), SFP and CF are generally chert-banded and their total iron content readily exceeds 15 %. In contrast, the rocks of the sulphide facies are noncherty or contain only few chert bands; chemical analyses are often needed before their classification as iron-formation is warrented. Moreover, they contain abundant indisputably clastic material, which seems to be almost totally lacking from the four facies mentioned first. Thus, iron formation proper is defined as a rock-stratigraphic unit consisting mainly of the more or less chert-banded and true chemical sedimentary iron-formation rocks of MSOF, (MSOF), SFP and CF.

Gigaband: A mappable lithostratigraphic unit (member or formation).

Macro-, meso-, micro- and interbands: All but the last of these terms are used in the same sense as by Trendall (1965 b, p. 56). The term mesoband is, however, restricted to chert-laminated bands in iron formations. For practical reasons a lower core-length limit of about 40 cm had to be chosen for macrobands. Thinner bands of contrasting lithology and origin within the iron formations proper are called interbands or simply bands. Owing to its contrasting and exceptional lithology the phosphorite bands are also named interbands, although they could, if following logically the nomenclature of Trendall, be classified under mesobands.

Black schist: A phyllite or fine-grained mica schist containing enough graphite to impart the distinctive black colour.

Iron-rich black schist and phyllite: Black schist or phyllite with 15 % Fe_{tot} or more, as revealed by chemical analyses.

Chert: In petrology, fine-grained quartzose rock of possible chemical sedimentary origin. In mineralogy, microcrystalline variety of quartz of probable nonclastic origin.

Stratigraphic and structural definitions and redefinitions

The stratigraphic and structural terms used in this study are those defined in the earlier paper by Laajoki (1973) with the exception of the following:

Salmijärvi Dolomite Member (DP_2): In the Pääkkö area only pure dolomite upon Quartzite Formation-III is included in this member.

Pääkkö Quartzite Member (DP_3): The quartzites, phyllites and mica schists, which according to drilling form a more or less uniform horizon between the Salmijärvi Dolomite and iron-formation lenses near the Pääkkö farm, are classified under this new member of the Dolomite-Phyllite Formation (DP). Type quartzite outcrops for the member are those north of drill holes Nos. 356 and 355 (Fig. 3). The member is a clastic metasediment unit the major part of which is composed of quartzite, and its uppermost portions of quartzite and phyllite interbeds. It often shows rythmic features in such a way that biotite-bearing phyllite grades through coarse-grained, subgreywacke-like mica schist and mica-bearing quartzite to almost pure orthoquartzite (Ervamaa and Laajoki 1977, Fig. 13). These rocks are counterparts of the psammitic and pelitic rocks in the Salmijärvi area, which were earlier included in the upper half of the Salmijärvi Dolomite Member (Laajoki 1973, p. 30).

Pääkkö iron formation: The main iron-formation lense (intersected by drill holes Nos. 360 . . . 355, Fig. 3) and the three smaller ones east of it (drill holes Nos. 344, 356, 358) in the vicinity of the Pääkkö farm are treated as an entity called the Pääkkö iron formation.

Pääkkö Iron-Formation Member (DP_4): The lenses which form the Pääkkö iron formation are considered to belong to the same rock-stratigraphic horizon known as the Pääkkö Iron-Formation Member. The member has been excavated in the northern end of the main iron formation lense.

Iso Vuorijärvi iron formation: The main iron-formation lense (drill holes Nos. 345, 364, 361) and the two separate lenses north of it (drill holes Nos. 362 and 363) in the area north of Iso Vuorijärvi form the second iron-formation entity called the Iso Vuorijärvi iron formation. They are informally correlated with the Pääkkö Iron-Formation Member. This occurrence does not outcrop.

Seppola iron formation: In the vicinity of the Seppola farm, at the western margin of the Salmijärvi basin, iron-rich and manganiferous black schists and phyllites were revealed by drill holes Nos. 335, 337, and 341 (Fig. 2). The formations are collectively called the Seppola iron formation. Informally they are correlated with the Pääkkö Iron-Formation Member.

Körölä iron formation: The small iron-formation lenses in the vicinity of the Körölä farm (Fig. 5) are called the Körölä iron formation. They are also correlated informally with the Pääkkö Iron-Formation Member.

Salmijärvi Phyllite Member (DP_5): The definition of this member is unchanged (*op.cit.*, p. 30), except that the iron formations, earlier known as magnetite and/or amphibole-rich phyllite, have now been classified as a specific stratigraphic unit (DP_4).

Väyrylänkylä basement wedge: In Väyrylänkylä, between the western margin of the Hietajärvi anticlinorium and the eastern margin of the Salmijärvi basin (synclinorium), there occurs a rather narrow wedge trending from north to south (Fig. 2). This wedge is composed mainly of Prekarelidic orthogneisses together with strongly sheared metasediments of the lowermost Jatulian horizon (Arkosite Formation) (*op.cit.*, p. 17 and 42). The eastern margin of the wedge can be traced along the fault zone west of Pihlajavaara-Latolanvaara (*op.cit.*, Map 1) in which strongly sheared rocks have been encountered (Laajoki and Ojanperä 1973). In the Pääkkö area, the western contact is not readily determinable (Fig. 3). In the Körölä area, the western contact evidently follows the shear zones, which go via Iso Salmijärvi (*op.cit.*, Fig. 5). This allochthonous rock unit is called the Väyrylänkylä basement wedge.

Pääkkö syncline: The Marine-Jatulian rocks near the Pääkkö farm are interpreted to form a syncline whose core is occupied by black schists and phyllites (Fig. 3). This structural element is called the Pääkkö syncline.

GENERAL GEOLOGICAL SETTING

The study area is in the part of the Karelidic belt known as the Kainuan schist belt (Fig. 1). This NNW-SSE trending schist belt is characterized by the Jatulian at its eastern and southern margins and by the Kalevian in its middle and western parts. The Jatulian is an epicontinental lithosome composed mainly of quartzites with minor basal arkosites. The Kalevian is a flyschoidic lithosome with phyllites and mica schists. These two units are intervened by a minor lithosome that is known as the Marine Jatulian, and which contains phyllites, black schists, dolomites, quartzites and iron formations. Thus, the iron formations of the Kainuan schist belt are of pure Superior type. The most noteworthy iron-formation occurrences are Tuomivaara in Sotkamo (Mäkelä 1976) and those of Väyrylänkylä and Poskimäki in the South Puolanka area (Fig. 2). Karelidic supracrustal rocks are intruded by metadiabases of early Karelidic magmatism, c. 2050 Ma (Sakko and Laajoki 1975) and by Karelidic granite, c. 1900 Ma (Asa 1971).

The general geology of the South Puolanka area has recently been studied by Laajoki (1973). In Väyrylänkylä the iron formations, with an age of 2080 Ma (Sakko ja Laajoki 1975), occur in the synclinorium known as the Salmijärvi basin (Fig. 2). They are encountered within the Dolomite-Phyllite Formation (DP) of the Marine Jatulian [1]). All iron formations within this basin are interpreted to have belonged originally to the same iron-formation horizon (DP_4). In connection with the shear folding of the second deformation stage and the invasion of the Väyrylänkylä basement wedge the stratigraphic relationships between the various occurrences were disturbed, especially in the Pääkkö area, and parts of the iron-formation horizon were evidently sheared off (Ervamaa and Laajoki 1977, Fig. 8).

Fig. 3 shows the general geological setting of the two main iron formations, Pääkkö and Iso Vuorijärvi. As can be seen, the geology of the Pääkkö occurrence is relatively well known, thanks to the many drill holes and outcrops unlike the Iso Vuorijärvi formation, which has been penetrated by a few holes only. Near the Seppola farm the iron-formation horizon was intersected by three drill holes (Nos. 335, 337 and 341 in Fig. 2). The stratigraphy of these three sections is given in Fig. 4.

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[1]) The black schists, phyllites and limestones of this lithostratigraphic unit have been earlier described by Väyrynen. In his classic »Kainuugebiet» study he included them in the middle section of his Mica-Schist Formation (1928, pp. 94—102). Later he considered this sedimentary facies foreign to the area and classified the rocks under the allochthonous series of the Phyllite-Mica Schist Formation (Enkovaara *et al.* 1953, pp. 47—48) or under the western allochthonous Jatulian (Väyrynen 1933, p. 7. and 76; 1954, p. 169).

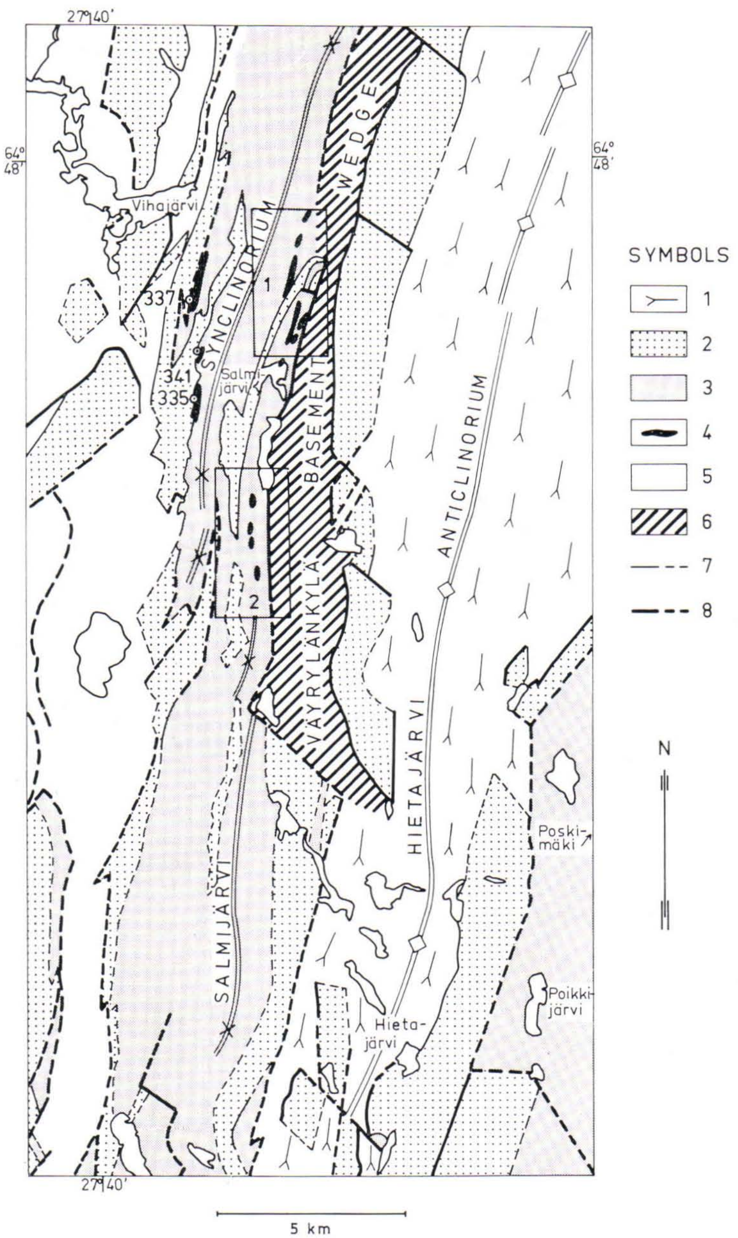

Fig. 2. Iron-formation occurrences of Väyrylänkylä and Poskimäki (on the eastern margin) marked on the generalized geologic map of the South Puolanka area (Laajoki 1973). 1) Prekarelidic basement gneisses. 2) Karelian quartzite (mainly Jatulian). 3) Marine Jatulian and miogeosynclinal Kalevian. 4) Iron formation. 5) Eugeosynclinal Kalevian. 6) Allochtonous formations. 7) Main lithologic-stratigraphic contact. 8) Fault. Areas covered by maps in Figs. 3 and 5 are marked by rectangles 1 and 2, respectively. The numbers 335, 337 and 341 refer to the drill holes which intersected the Seppola iron formation.

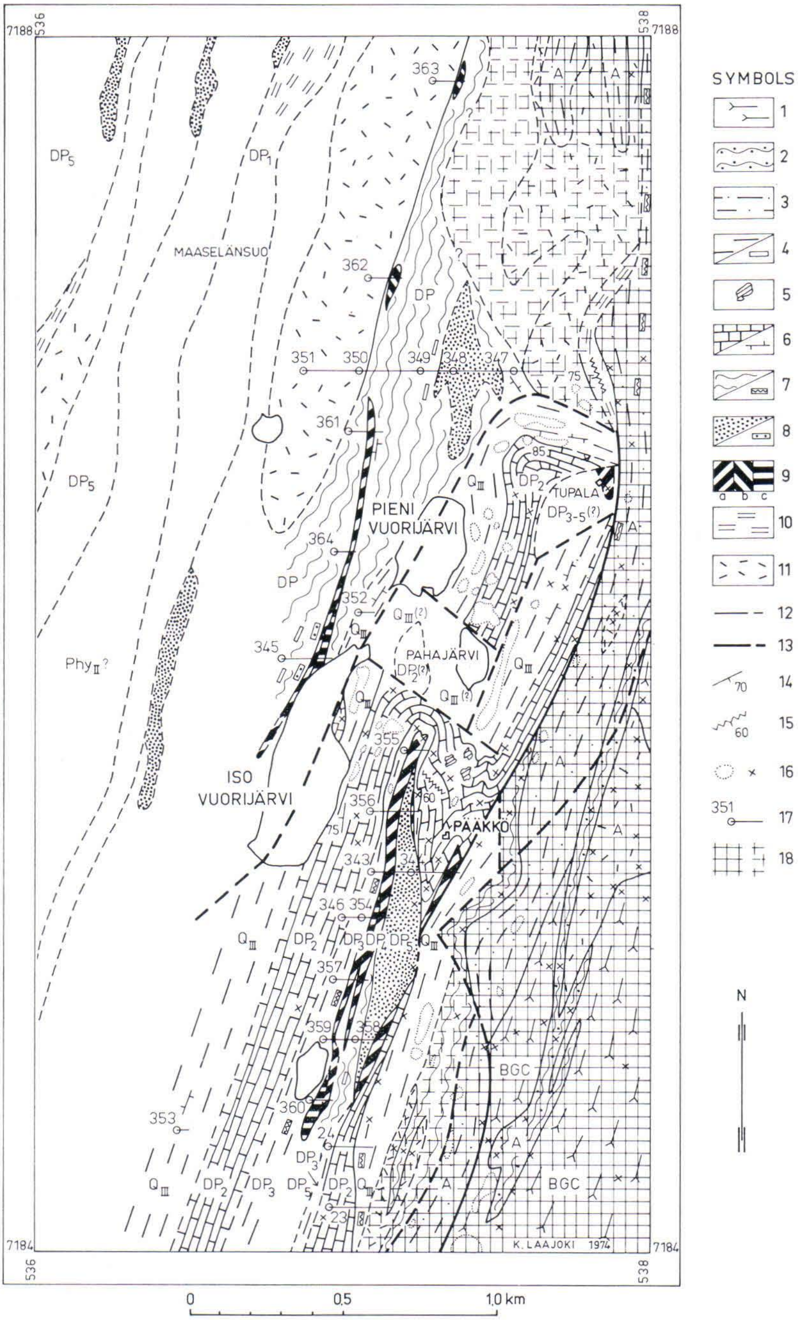

Fig. 3. (Explanation on next page)

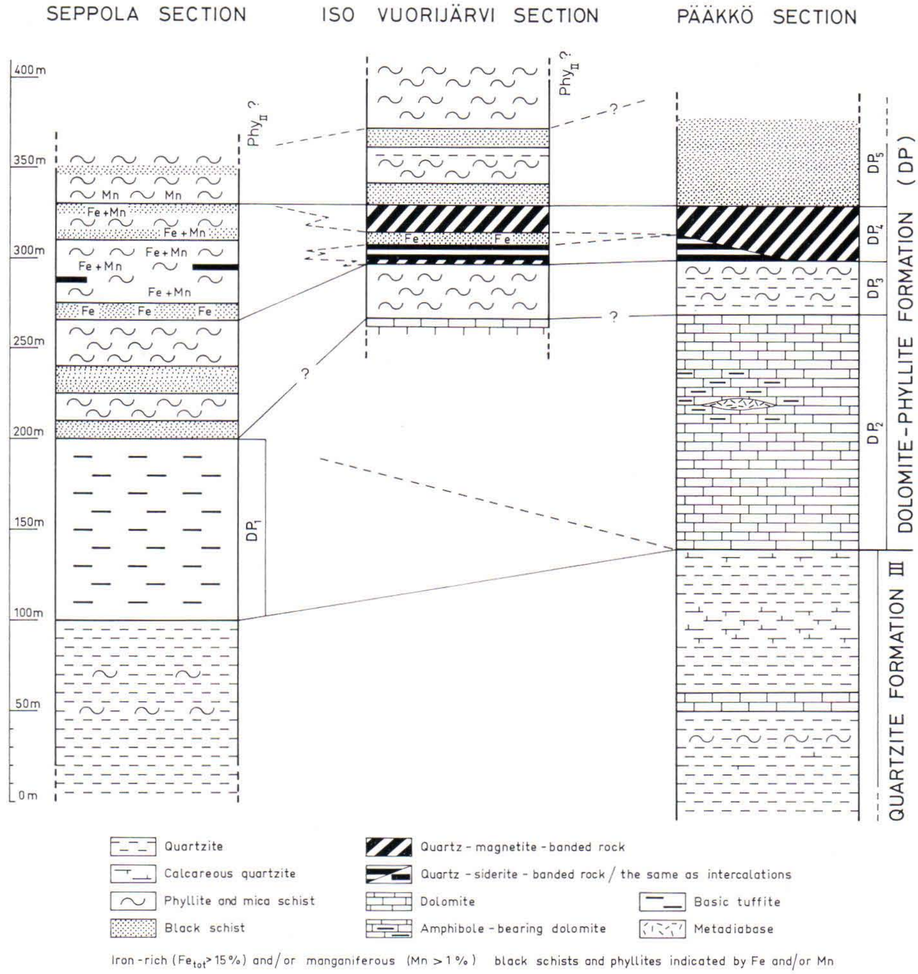

Fig. 4. Stratigraphic columns from the different sections of the Salmijärvi basin.

Fig. 3. Stratigraphic-lithologic map of the surroundings of the Pääkkö iron formation (in the middle lower part) and the Iso Vuorijärvi iron formation (in the middle). Lithologic symbols: 1) Prekarelidic basement gneiss. 2) Conglomeratic mica schist. 3) Arkosite and feldspar- and/or sericite-bearing quartzite. 4) Quartzite-orthoquartzite/the same as intercalations. 5) Tectonic inclusions of quartzite. 6) Dolomite/the same as intercalations. 7) Phyllite/the same as intercalations. 8) Black schist/the same as intercalations. 9) Iron formation: a) MOSF, b) MSOF, c) CF. 10) Basic tuffite. 11) Metadiabase and/or -volcanics. 12) Stratigraphic-lithologic contact. 13) Fault or tectonic contact. 14) Bedding and dip. 15) Approximate strike and dip of shear-folded beds. 16) Outcrop. 17) Drill hole and number. 18) Allochthonous formations. Stratigraphic units: BGC) Basement Gneiss Complex. A) Arkosite Formation, QIII) Quartzite Formation III. DP) Dolomite-Phyllite Formation: DP$_1$) Seppola Tuffite Member. DP$_2$) Salmijärvi Dolomite Member. DP$_3$) Pääkkö Quartzite Member. DP$_4$) Pääkkö Iron-Formation Member. DP$_5$) Salmijärvi Phyllite Member. PhyII) Phyllite Formation II.

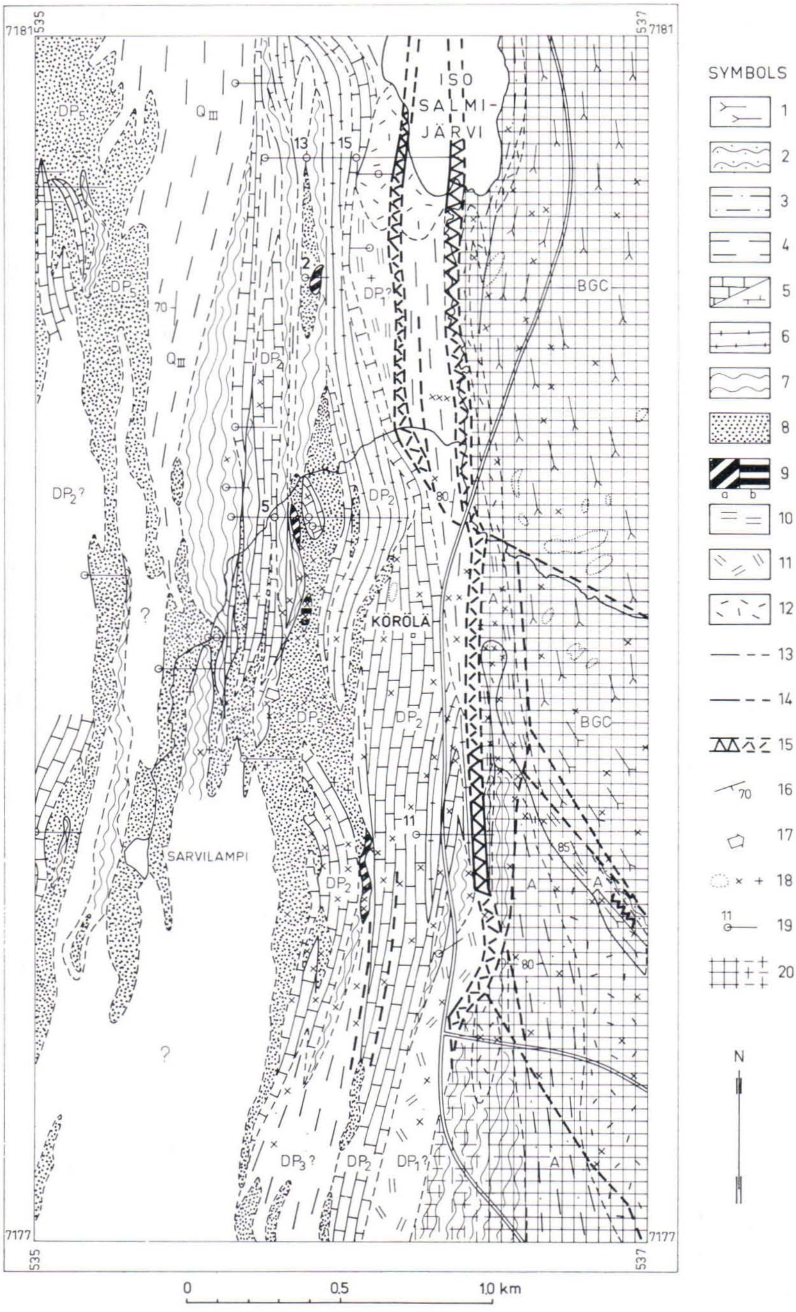

Fig. 5. (Explanation on next page)

The rock succession in the Pääkkö section: dolomite (DP_2), quartzite-phyllite (DP_3), iron formation (DP_4), black schist (DP_5) — in order from bottom to top — is similar to the sequence typical of Superior-type iron formations in all continents (e.g. Gross 1965, p. 91); the thickness of the iron formation in Pääkkö is, however, regrettably small. In the Iso Vuorijärvi section the schists east of the iron formation have been interpreted as underlying the iron-formation horizon. They are, in general, calcareous, whereas the schists west of the iron formation are more argillaceous. Iron formations occur in Iso Vuorijärvi in more pelitic rock association than in the Pääkkö section. The metadiabase-metavolcanics lense west of the northern end of the Iso Vuorijärvi iron formation, called informally the Liejeenjoki metavolcanics (Fig. 3), is composed of basic subvolcanic or volcanic rocks, or both, and often shows amygdaloidal textures. If the stratigraphic interpretation presented is valid these metavolcanics are younger than the iron formation.

In the Seppola section the iron formations are intimately associated with black schists and phyllites; Marine Jatulian dolomites and quartzites are almost totally lacking. Thus, from Pääkkö (the easternmost section) to Seppola (the westernmost section) the dolomites and quartzites within the Dolomite-Phyllite Formation gradually decrease, which indicates that the water was probably shallower in the eastern than in the western margin of the sedimentation basin.

The following regularities can be observed within the iron-formation horizon itself (DP_4). The easternmost parts (Pääkkö iron formation) consist mainly of quartz-magnetite-banded rock of the MOSF. Considerably thick beds of quartz-siderite-banded rock of the CF have been encountered in three drill holes only in the western margin of the main iron formation. Within the schist wedge at the southern end of the Pääkkö iron formation an iron-formation bed has been penetrated. This bed, five metres in thickness, exhibit chert-mesobanded grunerite-rocks of the SFP as a rim round a magnetite-amphibole rock of the MOSF (Ervamaa and Laajoki 1977, Fig. 14).

Somewhat further west, the horizon (Iso Vuorijärvi iron formation) is composed mainly of MOSF rocks. In the eastern margin of the southern end of the main iron-formation body (drill hole No. 345) quartz-siderite-banded rocks of the CF and the iron-rich black schists of the SF are encountered together with the MOSF rocks. Moreover, the two northernmost separate lenses (Fig. 3) are composed mainly of weathered iron-silicate-magnetite rock of the MSOF. In both Pääkkö and Iso Vuori-

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Fig. 5. Stratigraphic-lithologic map of the surroundings of the iron-formation occurrences in Körölä. Lithologic symbols: 1) Prekarelidic basement gneiss. 2) Conglomeratic mica schist. 3) Arkosite and feldspar- and/or sericite-bearing quartzite. 4) Quartzite-orthoquartzite. 5) Dolomite/the same as intercalations. 6) Carbonate-bearing amphibolite. 7) Phyllite and mica schist. 8) Black schist. 9) Iron formation: a) MOSF, b) CF. 10) Basic tuffite or metatuff. 11) »Strahlstein»-amphibolite. 12) Metadiabase. 13) Stratigraphic-lithologic contact. 14) Fault or tectonic contact. 15) Mylonite or breccia zone. 16) Bedding and dip. 17) Top of beds. 18) Outcrop. 19) Drill hole and number. 20) Allochthonous formations. Stratigraphic units: BGC) Basement Gneiss Complex. A) Arkosite Formation. QIII) Quartzite Formation III. DP) Dolomite-Phyllite Formation: DP_1) Seppola Tuffite Member. DP_2) Salmijärvi Dolomite Member. DP_3) Pääkkö Quartzite Member.

järvi the iron formations contain interbeds, 0.5 m — 4 m thick, of phyllite, mica schist, black schist and impure quartzite. These more or less normal clastic units are more common and constitute a greater proportion of the formation in Iso Vuorijärvi than in Pääkkö (Fig. 4). They seem to have divided the Pääkkö formation into at least two (Fig. 71) and the Iso Vuorijärvi formation into two or three different iron-rich units.

In the westernmost parts of the iron-formation horizon (Seppola iron formation) iron-rich, spessartite-bearing black schists and phyllites of the SF are predominant. Only thin macrobands of the CF rocks are sometimes encountered. The iron-rich schists in this section are clearly mesobanded, and relatively rich in carbonate, which further distinguished them from the SF rocks of the Pääkkö and Iso Vuorijärvi sections. Thus presuming that the facies concepts by James (1954) are correct (*cf.* Dimroth 1975), even the changes in facies within the iron-formation horizon itself (DP_4) indicate that the water increased in depth from east to west in the depositional basin.

In the Körölä area (Fig. 5), some 5 km south of the Pääkkö area, small iron formations occur (drill holes Nos. 2 and 5) that consist mainly of quartz-magnetite-banded rock of the MOSF alternating with iron-rich black schists of the SF and with ordinary phyllites (Ervamaa and Laajoki 1977, Figs. 18 and 19). The black schists are often garnet-bearing. One lense of the CF rock has been encountered. Here as well intense shear folding of the second deformational stage has disturbed the primary stratigraphic succession. Both the lithologic association, especially the abundance of dolomite, and structural evidence suggest, however, that these formations occupy the core zone of the southern extension of the Pääkkö syncline and are thus correlative with the Pääkkö iron formation.

The iron formations consisting mainly of the MOSF in the eastern margin of the Salmijärvi basin are clear-cut units with relatively sharp contacts with the country rocks. Thus, in drill hole No. 343 in Pääkkö, the prevailing quartz-magnetite-banded rock grades through a black schist, 1.2 m thick and exceptionally rich in pyrrhotite, in the normal black schist of the overlying Salmijärvi Phyllite (Table 12); in drill hole No. 345 in Iso Vuorijärvi the same rock grades through a 1.4 m thick carbonate-amphibole rock into the black schist to the west (Table 10, see also p. 38). In Seppola, on the other hand, the limits of the iron-rich horizon are much more vague.

LITHOLOGY, PETROLOGY AND DESCRIPTIVE MINERALOGY

Main lithologic and petrologic features of the iron-formation rocks

In this chapter the petrological features that elucidate the geochemistry and origin of the iron formations are briefly described, and some notes on the petrography of the samples analysed are made in Tables 3. On the scale of macrobands the names of the rocks are defined by one or two of the minerals macroscopically most conspicuous and by structure. In the names of the mesobands one or more of the most specific minerals are given in the order of descending abundance.

Table 1

Mineralogical compositions of some iron-formation rocks from Väyrylänkylä. Point-counting method. The areas counted cover the major parts (60—80 %) of the thin sections made from the samples analysed by wet chemical methods. The values for the opaques in brackets were calculated from the chemical analyses. Note the great differences between the latter and the point-counting results in columns 5 and 6 owing to the misleading effects of graphitic matter.

	1	2	3	4	5	6	7
Amphibole	1.8	55.9 a)	40.1	53.1 b)	32.3	49.5 c)	20.1
Biotite	+	+	0.1	0.5	12.0	10.3	58.7 e)
Garnet	—	—	—	—	—	5.9	17.4
Chert and quartz	84.3	+	13.3	14.6	4.3	6.0	+
Plagioclase	—	—	—	—	+		—
Carbonate	7.4	0.9	0.8	—	—	0.2 d)	—
Apatite	—	+	1.7	+	0.1	+	+
Opaques	6.5	36.9	43.9	31.7	51.3	28.1	3.7
(Magnetite)			(< 35.7)	(< 28.0)	(< 16.3)	(< 5.9)	—
(Pyrrhotite)			(0.5)	(0.6)	(4.6)	5.1)	—
(Pyrite)			—	—	—	—	(~ 4.2)
(Chalcopyrite)			—	—	+		+
(As-sulphide)			—	—	—	+	—
(Graphitic substance)			(1.8)	(0.5)	(2.4)	(4.5)	(0.4)
(Total)			(< 38.0)	(< 29.1)	(< 23.3)	(< 15.5)	(~ 4.6)
Limonitic substance	—	6.1	—	—			
Others	—	0.2		0.1			

a) includes 32.2 % weathered and 23.7 % fresh amphibole
b) includes 0.5 % green amphibole
c) includes 0.5 % ferrotschermatic amphibole
d) too low because carbonate was largely outside the calculated area
e) includes 1.1 % chlorite
\+ detected, — not detected

1) Chert mesoband. Table 7, No. 1. Thin section No. 19648.
2) Iron-silicate-magnetite rock. Table 4, No. 4. Thin section No. 18433.
3) Magnetite-amphibole-chert mesoband. Table 7, No. 2. Thin section No. 19649.
4) Amphibole-magnetite-chert mesoband. Table 7, No. 3. Thin section No. 19650.
5) Amphibole-biotite-magnetite-chert mesoband. Table 7. No. 5. Thin section No. 20357.
6) Amphibole-biotite-garnet interband. Table 7, No. 6. Thin section No. 19651.
7) Biotite-amphibole-garnet interband. Table 7, No. 9. Thin section No. 20441.

Table 2

Chemical composition (Wt-%) of amphiboles, biotites and garnets, and partial composition of carbonates of some rock types from the Pääkkö and Seppola iron formation. Electron microprobe analyses by Tuula Paasivirta. The poor quality of the siderite analyses (Nos. 17—21) is apparently due to the low iron and manganese contents of the standard used (dolomite with 4.6 % FeO and 0.6 % MnO).

	1	2	3	4	5	6	7	8	9	10	11	12
SiO_2	47.9	45.0	47.2	$53.1^{a)}$	50.6	38.6	33.4	34.1	37.0	35.6	37.7	37.2
TiO_2	0.1	0.0	0.1	0.0	0.0	1.5	2.0	1.9	2.0	2.1	0.1	0.0
Al_2O_3 ...	0.6	0.2	2.0	0.3	0.2	17.2	18.4	18.8	18.4	17.4	19.5	21.0
FeO *) ..	41.6	43.7	39.4	37.8	29.0	29.7	30.2	31.5	23.8	28.6	39.0	38.5
MnO	0.0	0.7	0.0	0.1	5.1	0.1	0.0	0.0	0.0	0.4	0.1	0.2
MgO	8.1	7.1	10.6	10.7	13.0	2.7	5.1	4.5	8.2	5.9	0.7	1.6
CaO	0.1	0.6	0.2	0.2	0.3	8.7	0.0	0.0	0.0	0.1	3.4	2.3
Na_2O	0.2	0.0	0.0	0.0	0.6	1.6	0.0	0.0	0.0	0.3	0.2	0.0
K_2O	0.0	0.0	0.0	0.0	0.0	0.3	9.0	8.3	9.2	9.6	0.0	0.0
Total	98.6	97.3	99.5	$102.2^{a)}$	98.8	100.4	98.1	99.1	98.6	100.0	100.7	100.8
Fe_{tot}	32.3	34.0	30.6	29.4	22.5	23.1	23.5	24.5	18.5	22.2	30.5	29.9
Mn	0.0	0.5	0.0	0.1	3.9	0.1	0.0	0.0	0.0	0.3	0.1	0.2

*) total iron. a) the excessive value is due to the heterogeneity of the grain analysed in respect of silica.

1) Grunerite. Amphibole-biotite-magnetite-chert mesoband. Thin section No. 20357. (Table 7, No. 5).
2) Grunerite (slightly weathered). Chert-mesobanded grunerite rock. Thin section No. 21071. (Table 4, No. 9).
3) Grunerite. Amphibole-biotite-garnet interband. Thin section No. 19651. (Table 7, No. 6).
4) Grunerite. Biotite-amphibole-garnet interband. Thin section No. 20441. (Table 7, No. 9).
5) Manganoan cummingtonite. Spessartite-bearing iron-rich phyllite. Thin section No. 20358. (Table 6, No. 5).
6) Ferrotschermatic hornblende. The same rock as in 3.
7) Biotite (lepidomelane or siderophyllite). The same rock as in 1.
8) Biotite (lepidomelane or siderophyllite). The same rock as in 3 and 6.
9) Biotite (lepidomelane or siderophyllite). The same rock as in 4.
10) Biotite (lepidomelane or siderophyllite). The same rock as in 5.
11) Almandine. The same rock as in 3, 6 and 8.
12) Almandine. The same rock as in 4 and 9.

Quartz-magnetite-banded rocks of the MOSF

A type rock is a dark coloured and clearly banded rock consisting of alternating mesobands rich in chert or iron minerals (Fig. 6). On the scale of macrobands the main minerals are chert, magnetite and amphibole with lesser amounts of carbonate, apatite, pyrrhotite and graphite. Thin-section studies and microprobe analyses indicate that the amphibole is almost solely grunerite (or iron-rich cummingtonite), and carbonate mainly siderite.

The mesobands vary in thickness from some millimetres to about five centimetres. On the basis of their macroscopically most distinctive minerals they can be divided into three main types: chert, both magnetite- and amphibole-rich and amphibole ±

Table 2. (contd.)

	13	14	15	16	17	18	19	20	21	22	23
SiO_2	37.1	39.0	31.8	39.0							
TiO_2	0.0	0.5	0.3	0.2							
Al_2O_3 ...	20.2	20.4	19.6	18.0							
FeO *) ..	38.5	18.0	15.2	13.1	50.4	44.5	49.8	45.3	22.1	21.8	18.6
MnO	0.4	16.0	24.0	20.9	0.1	0.3	0.1	0.3	24.1	0.9	0.4
MgO	1.8	0.6	2.5	2.0	5.5	9.1	6.3	8.5	6.1	6.0	8.0
CaO	2.2	2.7	3.8	4.4	0.7	0.6	0.7	0.7	4.2	29.4	30.3
Na_2O	0.0	0.1									
K_2O	0.0	0.0									
CO_2 (calc)					37.6	37.8	38.0	37.8	38.5	43.7	44.1
Total	100.2	97.3	97.2	97.6	94.3	92.3	94.9	92.6	95.0	101.8	101.4
Fe_{tot}	29.9	14.0	11.8	10.2	39.2	34.6	38.7	35.5	17.2	16.9	14.5
Mn	0.3	12.4	18.6	16.2	0.1	0.2	0.1	0.2	18.7	0.7	0.3

13) Relic of almandine. Retrograde siderite band in garnet-amphibole-siderite-biotite interband. Thin section No. 21156.(Table 7, No. 7).
14) Almandine-spessartite. The same rock as in 5 and 10.
15) Spessartite-almandine (core). Spessartite-bearing phyllite. Thin section No. 17983.
16) Spessartite-almandine (rim). The same grain as in 15.
17) Siderite. Siderite rock. Thin section No. 19074. (Table 5, No. 8).
18) Siderite replacing almandine in 13.
19) Siderite in a carbonate ball of a phosphorite interband. Thin section No. 21203. (Table 7, No. 12).
20) Siderite in a vug of siderite rock. Thin section No. 21184.
21) Manganosiderite. Quartz-Mn-siderite banded rock. Thin section No. 19084. (Table 5, No. 7).
22) Ferroan ankerite. Quartz-magnetite-banded rock. Thin section No. 21181.
23) Ferroan ankerite. Quartz-magnetite-banded rock. Thin section No. 21201.

biotite-rich mesobands (Fig. 7). In addition to microcrystalline quartz, chert mesobands contain only small amounts (10—20 %) of iron minerals (Table 1, No. 1 and Fig. 8). They show distinct, sometimes wavy, microbanding in which markedly thin magnetite-rich streaks separate thicker chert streaks from each other. It is noteworthy that this iron-poor mesoband type does not contain apatite.

Magnetite- and amphibole-rich mesobands contain two index minerals, magnetite and amphibole, in variable proportions. The ratio of magnetite to amphibole is generally in the region of 1 : 1 (Table 1, Nos. 3 and 4). Moreover, they typically contain chert as thin bands (Fig. 9). Apatite is an essential and almost ubiquitous accessory mineral, whose amount may sometimes rise to some percents. Carbonate and biotite are met with in only a few mesobands. The bands with magnetite as the dominant iron mineral are black in colour, whereas those with amphibole as the most abundant mineral tend to be brownish or yellowish grey. Locally the rock shows distinctive graded bedding when the part rich in iron minerals grades into the chert-rich part of a microband (Fig. 9 A). In this case, as well as in general in the quartz-magnetite-banded rocks the occurrence of apatite is restricted to the iron-mineral-rich parts of the microbands. Similarly, in the Kittilä iron formation apatite occurs in conspicuous abundance in connection with magnetite, but is completely lacking in the parts of the rock free from magnetite (Hackman 1925, p. 27). The magnetite- and amphibole-

Fig. 6. Shear-folded quartz-magnetite-banded rock of the MOSF. Light flakes are amphibole. Grey dense bands are composed of microcrystalline quartz. Diameter of the coin is 15 mm. Outcrop in the northern end of the Pääkkö iron formation.

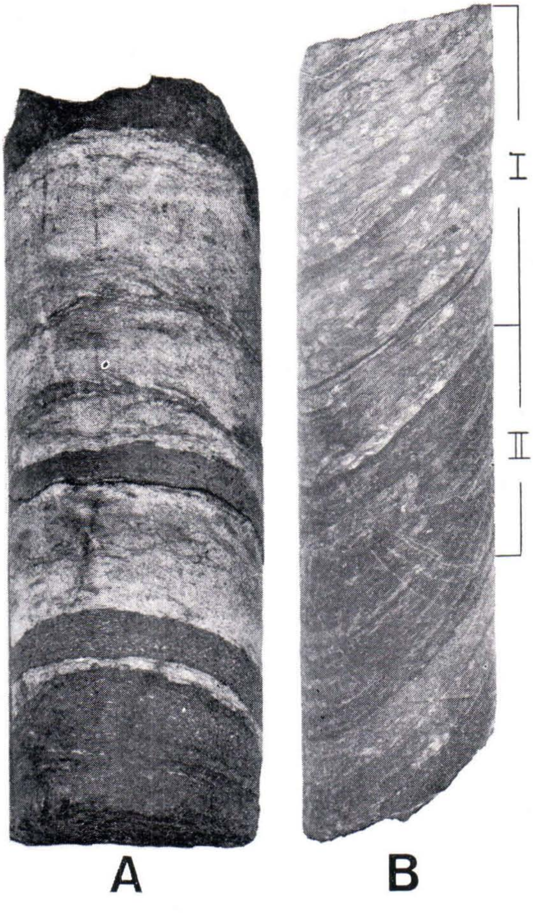


Fig. 7. Mesobanding in the MOSF rocks at Pääkkö. A) White chert mesobands alternating with dark magnetite- and amphibole-rich mesobands (DH 360). B) Amphibole ± biotite-rich mesobands (DH 343). The parts analysed are indicated by I (amphibole-chert-biotite-mesoband) and II (amphibole-biotite-magnetite-chert mesoband). The light spots in I are amphibole bundles; the thin black band is petrographically similar to II. Diameter of the drill cores is 3 cm. Photos by E. Halme.

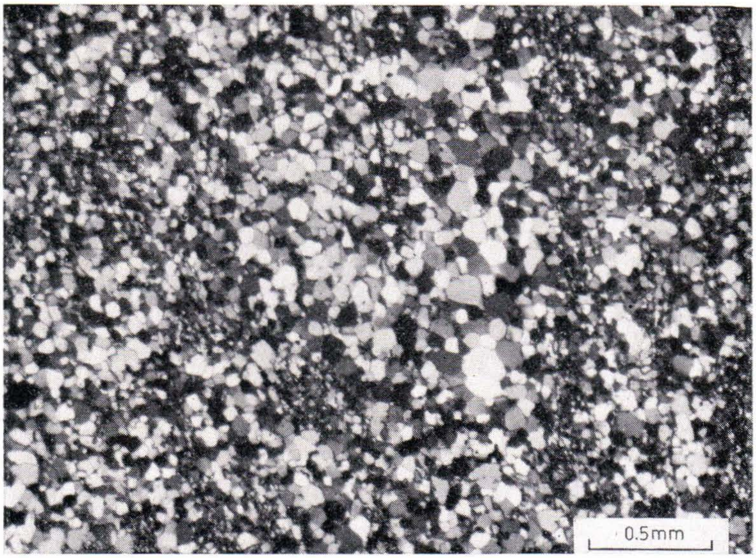


Fig. 8. Texture of a chert mesoband. Thin section No. 19648. Crossed nicols.

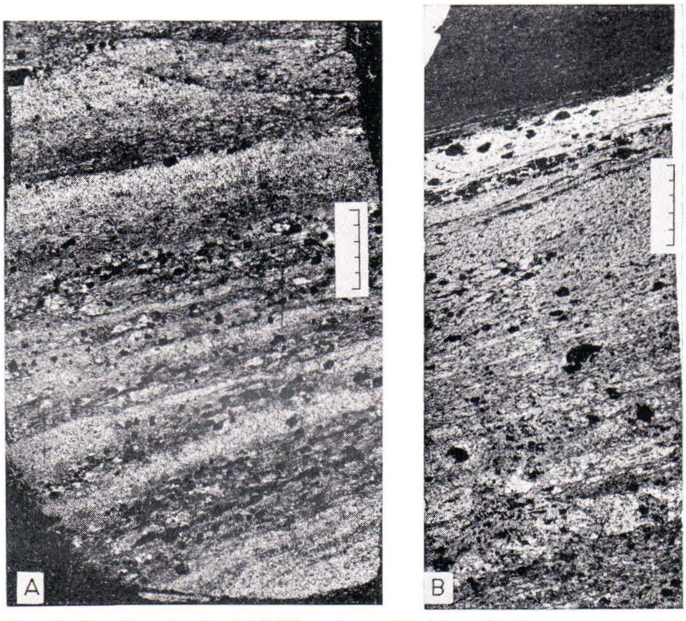


Fig. 9. Banding in the MOSF rocks at Pääkkö. A) Quartz-magnetite-banded rock showing graded bedding. White is chert, black magnetite and grey amphibole. Light spots in the probable lower parts of grades are amphibole. Thin section No. 18577, crossed nicols. B) Amphibole-magnetite-chert mesoband (most of the lower part). Grey is amphibole, black magnetite and white chert. Notice the incomplete chert bands in the lower part of the mesoband. In the upper part a white chert mesoband is separated from a black phosphorite interband by a thin magnetite-amphibole band. Thin section No. 19650. One nicol. Photos by E. Halme. The scales are in mm.

rich mesobands also include varieties which contain little if any chert, in which case the mesobands show massive banding (Fig. 10, also Fig. 45). No features, indicative of undisputed clastic origin, except sporadic biotite, have been detected in these magnetite- and amphibole-rich mesobands and thus, they are considered together with chert mesobands as the purest chemical-sedimentary component of the quartz-magnetite-banded rock.

Amphibole ± biotite-rich mesobands are characterized by the wealth of amphibole and the occurrence of biotite, albeit in highly variable amounts. Moreover they are relatively deficient in magnetite. At least two subtypes can be distinguished (Fig. 7 B): the first is a light-coloured rock consisting mainly of amphibole and chert but containing also small but noticeable amounts of biotite; the second subtype is dark in colour, extremely rich in amphibole and with abundant biotite and magnetite, but relatively poor in chert (Table 1, No. 5, Table 2, Nos. 1 and 7). These bands, named according to their most characteristic minerals, are either amphibole-chert-biotite or amphibole-biotite(-magnetite-chert) mesobands. The former, the chert-rich type, shows distinct wavy microbanding, whereas the latter, the chert-poor variety, has a poikiloblastic texture with lamination (Fig. 11). It is noteworthy that among the chert-banded rocks of the MOSF these bands are the only ones which regularly contain conspicuous amounts of biotite. Furthermore, they are often characterized by green hornblende rimming grunerite.

Besides the mesobands mentioned, yellowish or greyish, nonmagnetic mesobands consisting of siderite and chert are sometimes encountered.

In addition to the forementioned main band components of more or less nonclastic origin, the quartz-magnetite-banded rocks contain thin interbands of pyrrhotite-rich black schist, green noncherty garnet-bearing interbands and black phosphorite interbands, The petrology of these rocks will be described in separate chapters. Also common are thin intersecting quartz-carbonate (siderite) and quartz-amphibole (grunerite-cummingtonite) veins that evidently represent iron-formation material remobilized during folding.

In the Pääkkö iron formation, in drill hole No. 359, a bed exceptionally rich in magnetite and amphibole and deficient in chert has been encountered between the two SFP beds. The maximum original (premetamorphic) thickness of the bed is only 0.5 m (see structural discussion on page 26). The rock is weakly banded, but does not show distinctive mesobanding. Petrographically this magnetite-amphibole rock resembles the chert-poor varieties of magnetite- and amphibole-rich mesobands described above.

Iron-silicate-magnetite rocks of the MSOF

These rocks were regularly attacked by Preglacial weathering processes (Laajoki 1975 a), which greatly hampers the study of their petrology. The most weathered

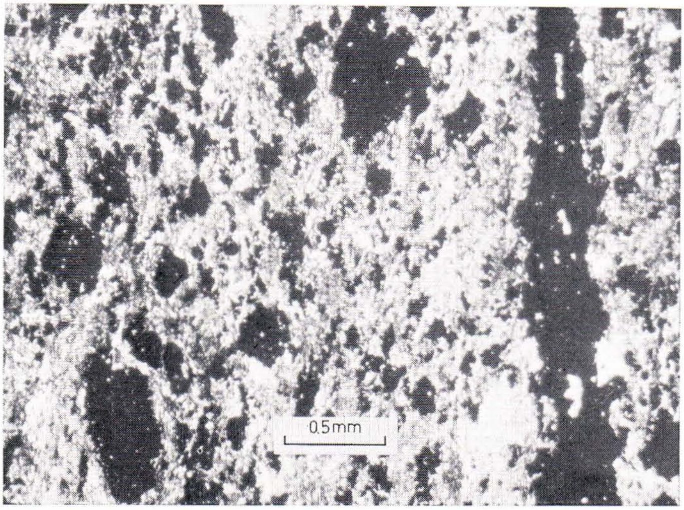


Fig. 10. Texture of a magnetite-amphibole mesoband devoid of chert Black is magnetite, the matrix is composed of amphibole. Thin section. No. 19649. Crossed nicols.

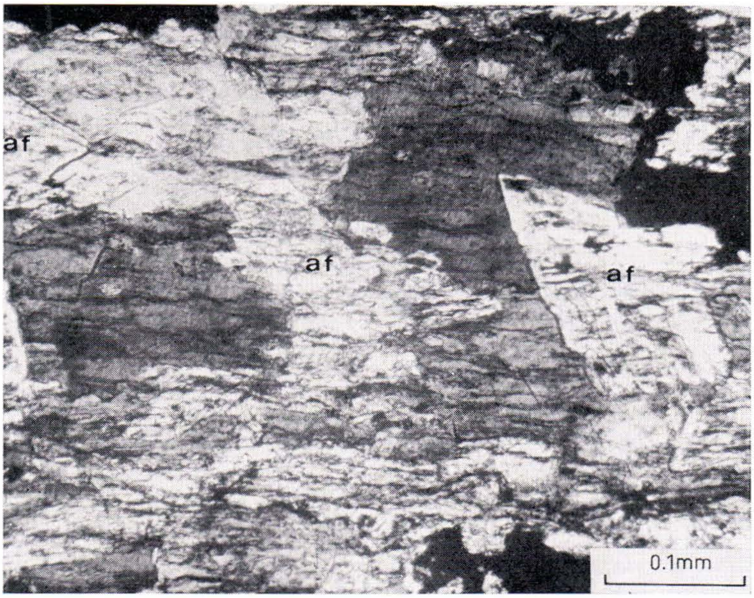


Fig. 11. Texture of an amphibole-biotite-magnetite-chert mesoband showing xenoblastic magnetite (black), idioblastic amphibole porphyroblasts (af) and poikiloblastic biotite (grey). Notice helicitic lamination in amphibole and biotite. Thin section No. 19080. One nicol.

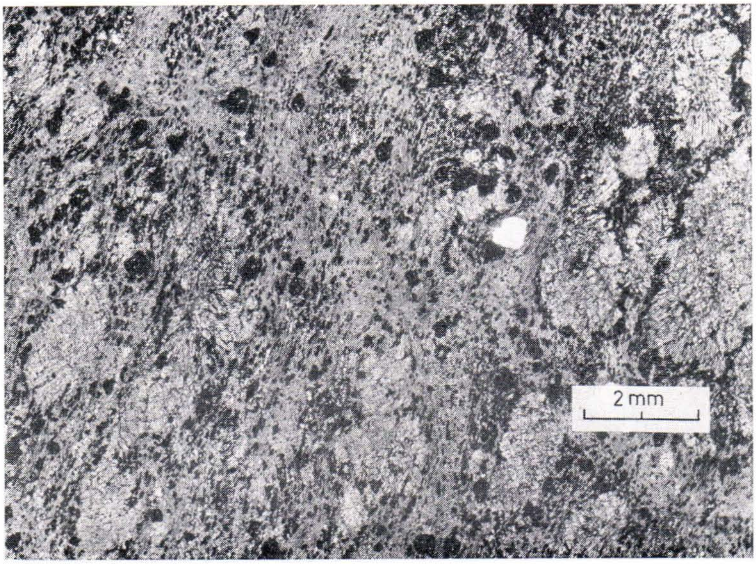


Fig. 12. Texture of an iron-silicate-magnetite rock of MSOF in Iso Vuorijärvi. Grey is amphibole, black magnetite. Grey spots are amphibole porphyroblasts, white is a hole. Thin section No. 18433. Photo by E. Halme.

rocks are rust-coloured amphibole-magnetite rocks containing an abundance of limonitic matter, a weathering (oxidizing) product of amphibole, and a few thin chert bands. It is impossible to say to what degree the abundance of amphibole in the rock is primary or to what degree secondary, caused by the weathering processes. It seems highly probable, however, that these rocks, whose classification into a separate facies stems from economic geologic requirements, were initially little, if any, richer in iron-silicates than is the quartz-magnetite-banded rock of the MOSF. Only exceptionally are relatively fresh iron-silicate-magnetite rocks with porphyroblastic texture (Fig. 12) encountered. Besides grunerite and magnetite the rock contains small amounts of pyrrhotite. In addition to the porphyroblastic texture, the almost complete lack of chert distinguishes this rock from the amphibole- and magnetite-rich mesobands of the MOSF (Table 1, Nos. 2 and 4, respectively).

Chert-mesobanded grunerite rocks of the SFP

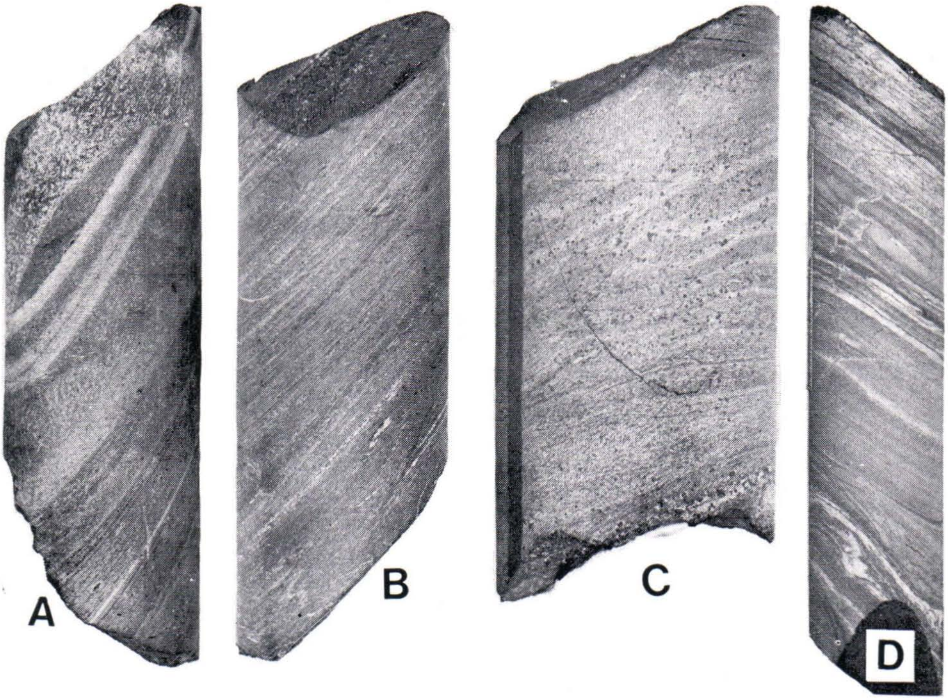
The rocks of the SFP have been met with only in the Pääkkö iron formation, where two beds of a chert-mesobanded rock, separated from each other by the magnetite-amphibole rock (p. 24), were penetrated by drill hole No. 359. The rocks show intense small-scale folding (Figs. 13 and 41), and thus the original (premetamorphic) thicknesses of the beds must have been considerably less than the intersections (2 and 3 metres). It may even be possible that the beds are flanks of a shear fold, and could therefore originally belong to the same metasedimentary unit.

Fig. 13. Chert-mesobanded grunerite rock (chert-rich part) of SFP in Pääkkö showing microfolding of the second deformational stage. Microfolding is revealed by black biotite bands rich in pyrite. White is chert and grey is grunerite. Notice the grunerite bundle in the upper right-hand corner, in which amphibole sprays have grown both parallel and perpendicular to the axial-plane foliation (in vertical position) and that in the lower right-hand corner, which shows a helicitic microfold and axial-plane foliation (thin black striation in vertical position). Thin section No. 21072, one nicol. Photo by E. Halme.

A typical rock is a greenish or greyish grunerite rock showing »strahlstein» - structure and containing white or bluish chert mesobands, and even a few thin dark mesobands rich in biotite (Fig. 13). The rock is somewhat weathered, but to an appreciably lesser degree than the MSOF rocks. The weathering is visible as the greyish colour of some grunerites in a hand specimen, and the obscurity of the amphiboles under the microscope. A microprobe analysis of one grunerite is given in Table 2 (No. 2). An unusually abundant green-bluish green amphibole occurs as a rim around grunerite, especially in close association with biotite-rich mesobands. As a rule, the rock contains small amounts of pyrite both as sporadic dissemination and thin veinlets.

Quartz-siderite-banded and -laminated rocks and siderite rocks of the CF

A typical CF rock is light-coloured, yellowish or greyish, and consists of alternating chert- and siderite-rich bands (Fig. 14, A and E). The rock is mainly composed



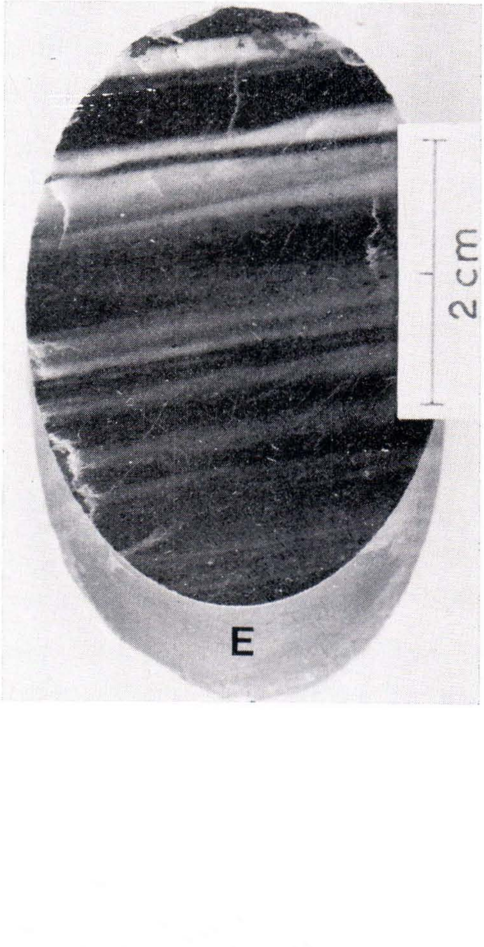


Fig. 14. Drill-core samples (Ø 3 cm) of the main petrographic types of CF rocks in Pääkkö (A-C, and E) and Seppola (D). A) Quartz-siderite-banded rock (DH 360) showing gradation to laminated siderite rock (lower part of the core). Quartz-carbonate vein cuts the banding in the upper part of the core. B) Laminated quartz-siderite rock (DH 354) (Table 5, No. 9). C) Weakly banded siderite rock (DH 360) (Table 5, No. 8). D) Quartz-Mn-siderite-banded rock (DH 337) (Table 5, No. 7). Notice two chert mesobands with marked microbanding at the upper end of the core. E) Polished core of the quartz-siderite-banded rock showing graded bedding (DH 360). White is chert and black siderite. The photograph of the thin section made on the same sample is reproduced in Fig. 15 A. Photos by E. Halme.

of siderite (Table 2, No. 17) and chert with some percents of magnetite and amphibole. Apatite is an essential accessory mineral as are pyrrhotite and graphite. The rocks in the Pääkkö iron formation are darker in colour and conspicuously richer in amphibole and magnetite than are those in Iso Vuorijärvi.

When the thickness of the microbands decreases, probably owing to more intense deformation, the banded rock grades into quartz-siderite-laminated rock (Fig. 14 B), but when the chert vanishes the rock passes into siderite rock which still shows banded structure (Fig. 14 C). This rock type is often markedly richer in amphibole than are the quartz-siderite-banded or -laminated rocks, a fact that may be attributed to silication.

Mesobanding in this facies seems to be relatively simple and consists mainly of chert and siderite-rich mesobands. The former contain, in addition to chert, sparse carbonate, amphibole, magnetite and, occasionally, some biotite. Siderite-rich mesobands may contain thin chert bands or be noncherty, in which case they have appreciably amphibole and variable amounts of magnetite. In this facies too the occurrence of apatite is restricted to the iron-mineral (siderite)-rich bands. The thickness of mesobands is, in general, from 0.5 to 3 cm.

In addition to the main band components described there occur a few magnetite- and amphibole-rich mesobands, amphibole ± biotite-rich mesobands and noncherty garnet-bearing iron-silicate interbands. These are petrographically analogous to those in the quartz-magnetite-banded rocks of the MOSF and are more common and thicker in Pääkkö than in Iso Vuorijärvi. Moreover, in both occurrences the CF rocks frequently contain phosphorite interbands. Owing to the great contrast in colour between the relatively light-coloured host rock and the dark interbands they are more easily detected in this facies than in other facies. They are especially conspicuous in chert-poor siderite rocks (Fig. 33). Moreover, also the CF contains thin interbands of black schist, the abundance of which seems, however, to be much lower than in the MOSF.

The CF rocks of the Pääkkö iron formation include one rock type that shows a well-preserved primary structure and another that exhibits interesting metamorphic textures. The former is a quartz-siderite-banded rock with undisputed graded bedding (Fig. 15). Within a grade, one part (probably the lower, see p. 106) is composed mainly of fine-grained siderite, grain size about 0.01—0.03 mm (Fig. 15 B). It often contains an abundance of apatite, associated with which there are sometimes quartz nodules, generally 0.2—0.5 mm in diameter (Fig. 42 A). This part may also contain magnetite porphyroblasts, which seem to decrease in size towards the chert-rich part of the grade. The amount and size of the porphyroblasts seem to be more or less directly proportional to the thickness of the siderite-rich part of the grade. In some rocks the magnetite porphyroblasts have ragged edges (Fig. 43 A), whereas in more metamorphosed rocks, where graded bedding is no longer visible, they are hypidioblastic or idioblastic with distinct cubic crystal faces (Figs. 16 and 43 B). In the latter rocks relicts of quartz nodules are rare occurrences. In general, the magnetite crystalloblasts

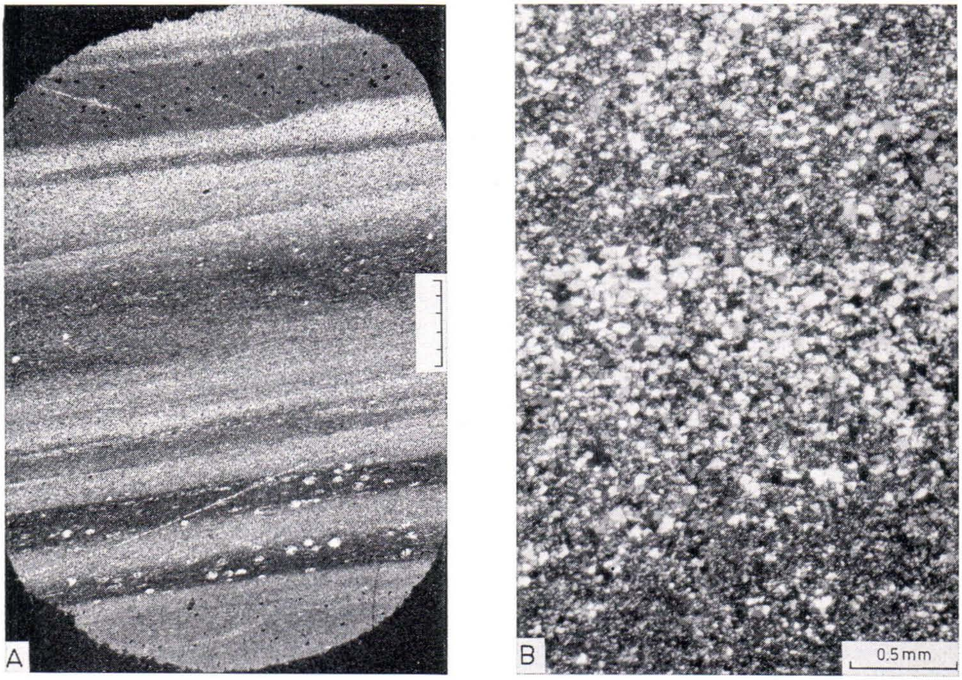


Fig. 15. Quartz-siderite-banded rock showing graded bedding. Thin section No. 18431. A) General view. White is chert, black is siderite and apatite together with graphitic substance and some magnetite. White spots are quartz nodules. Notice the abundant porphyroblastic magnetite in the uppermost thick siderite-rich part of a grade. One nicol. The scale is in mm. B) Close-up of the upper part of A. Grey is siderite and white chert. Notice the fine grain size of both minerals. Crossed nicols. Photo A by E. Halme.

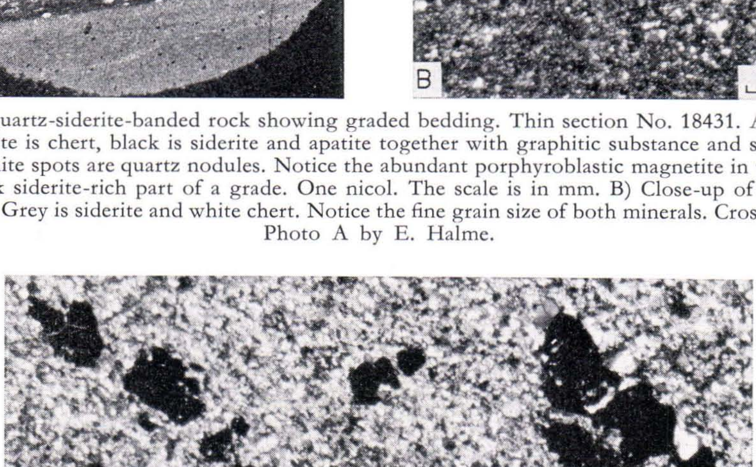


Fig. 16. Sub-idioblastic magnetite porphyroblasts in a siderite rock. Thin section No. 19074. Crossed nicols.

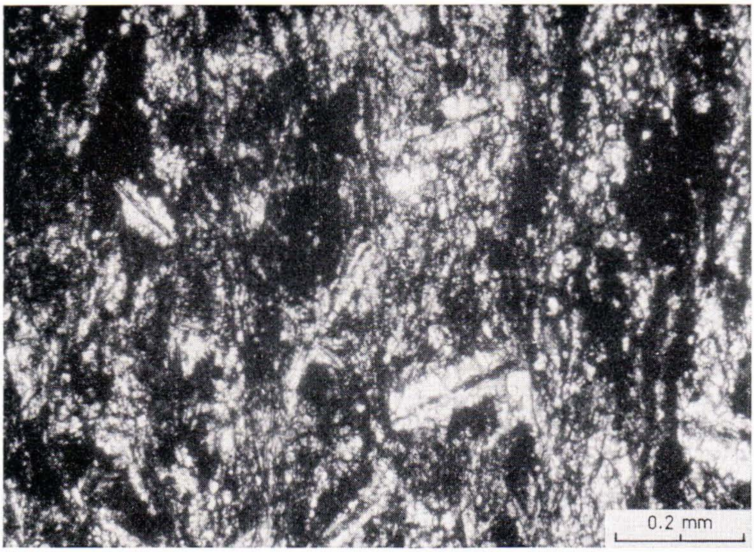


Fig. 17. Siderite rock in Pääkkö showing replacement texture. Foliation vertical. White is siderite, black mainly graphitic substance with minor magnetite, the amount of which is greatly exaggerated because the thin section is slightly too thick. Notice the relationship between the orientation of the replacement vugs and the foliation. Thin section No. 21184. One nicol.

and quartz nodules avoid each other (Fig. 15 A). The origin of the quartz nodules and magnetite porphyroblasts will be discussed in later chapters (p. 53 and 55, respectively). The probable upper parts of the grades are composed solely of chert (Fig. 15 B).

The second interesting rock type is an almost massive, often magnetite-spotted siderite rock showing probable early prograde metamorphic replacement textures. This rock typically occurs as mesobands from 2 to 3 centimetres thick. In addition to siderite and magnetite it contains abundant apatite as thin seams, graphitic substance, quartz and albite (microprobe identification by Jaakko Siivola). Typical accessory minerals are amphibole, biotite, garnet, pyrrhotite and pyrite. The most distinctive textural features are the peculiar occurrence of albite and columnar carbonate (Fig. 17). Moreover, amphibole and biotite occur as reaction products of albite.

Albite is met with as tiny plates, occasionally so tightly intergrown that they form a framework. In thin section they have either needle-like, if single, or net-like optically continuous textures. They show either quite sharp or ragged contacts with the recrystallized matrix carbonate (Fig. 18). The core of the albite lath or network often reveals pared pigment rows or a pigmented rectangular area (Figs. 18 B and 19 B). The significance of this texture is not wholly understood. It is possible that it is a relic of some unknown mineral or texture, of either mechanical or biogenic (algal?) origin.

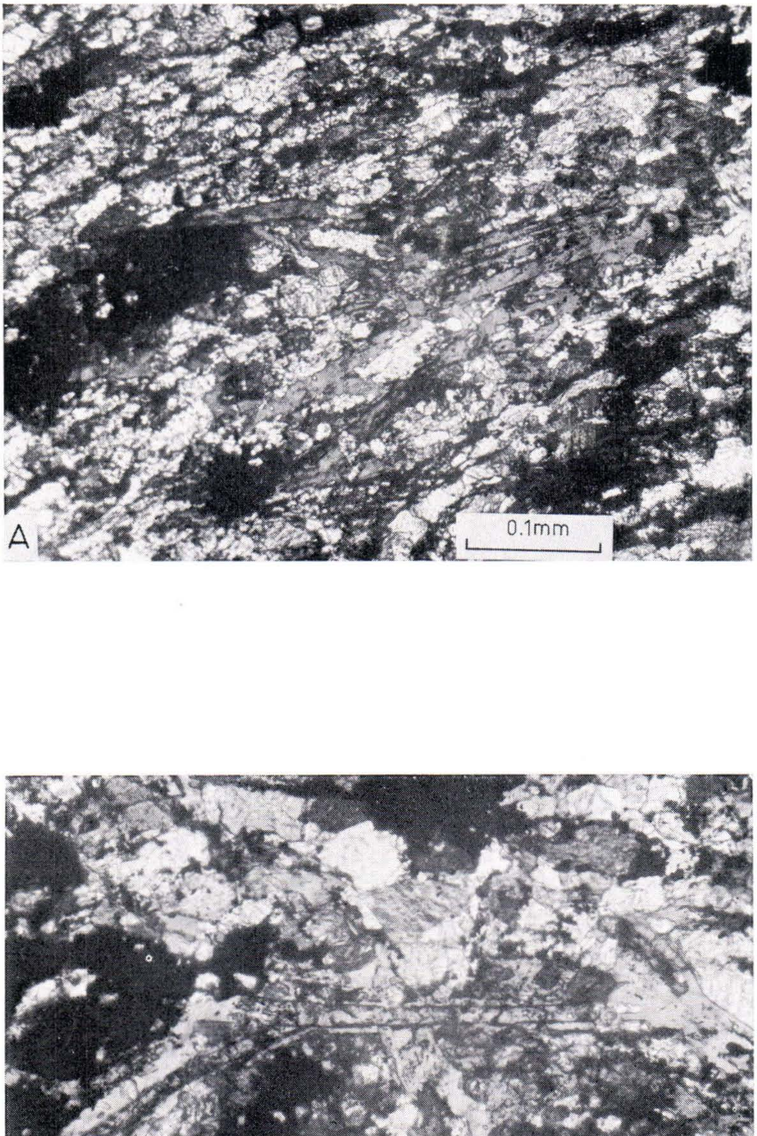


Fig. 18. Occurrence of albite in siderite rock. Thin section No. 17999. Crossed nicols. A) Typical crosscut of albite network (grey). Matrix is mainly recrystallized siderite. B) A close-up of albite network showing relict pigment texture. Albite is grey and optically continuous.

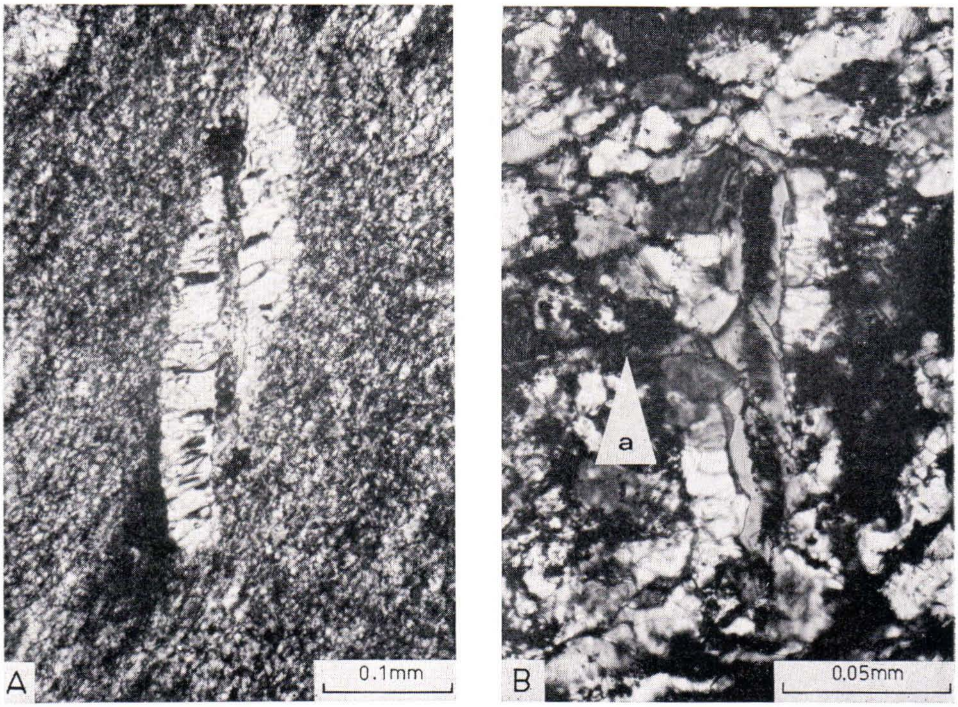


Fig. 19. A) A replacement-vug texture in a siderite rock. Matrix is siderite, the bright margins of the vug are columnar carbonate (siderite). The central »rib» consists of albite (grey, low relief) covered partly by siderite. Foliation runs diagonally to the vug texture. Thin section No. 21212. One nicol. B) A part of albite network rimmed by slightly recrystallized columnar carbonate (siderite). Grey with cleavage and low relief is albite, which extends outside the replacement texture on the left (a). Notice the relict pigment texture in albite. Thin section No. 21184. Crossed nicols.

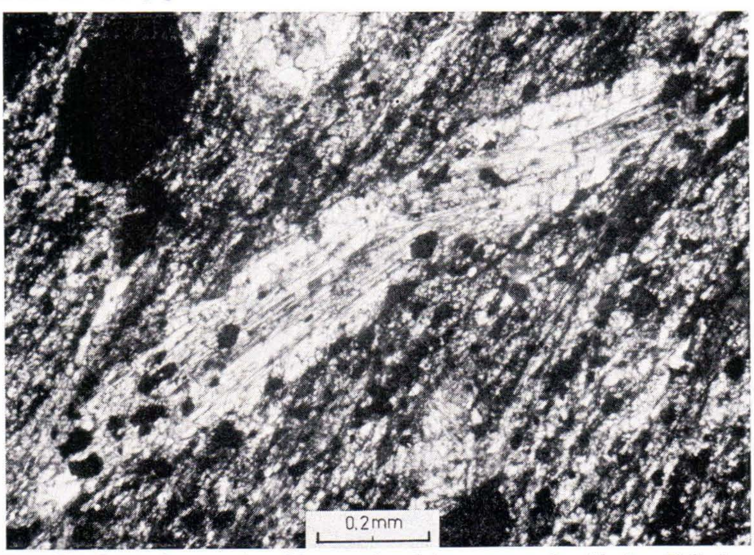


Fig. 20. Grunerite porphyroblast replaced by retrograde siderite. Foliation diagonal. Thin section 21183. Crossed nicols.

The single albite laths are often either totally or partly rimmed by columnar siderite, which has clearly grown from the matrix towards the albite. This has produced filled vug-like textures. In thin section these textures are most readily detected if the section is made somewhat too thick. The vugs seem to be arranged in zigzag fashion and tend to form a rather large angle with foliation (Figs. 17 and 19 A). In texture, the marginal carbonate (Fig. 19, *cf.* Fig. 20) resembles the rim carbonate of open space (porosity) structures of unmetamorphosed sedimentary carbonate rocks or the fibrous calcite cement replacing foraminiferal tests (Kendall 1976, Fig. 1). Thus, it is possible that carbonate was formed by pure early metamorphic replacement or it was deposited in the intestitial spaces between albite and matrix carbonate from which some material was probably removed owing to selective dissolution by metamorphic (or earlier?) fluids.

Because relict prints of albite and marginal carbonate are often detected in both amphibole (Fig. 51) and biotite (Fig. 58) it is concluded that albite and replacement carbonate must have been formed during a relatively early stage of metamorphism. French (1973) has fixed the upper boundary of low-grade metamorphism of iron formations at the appearence of grunerite-rich amphiboles. It therefore seems justifiable to regard albite and the columnar carbonate rimming it to be of low-grade metamorphic origin. Because neither of these minerals shows marks of deformation they are considered post-tectonic. On the other hand, the shear foliation of the second deformational stage cuts the general trend of the replacement vugs (Fig. 17). This indicates that the texture which guided the formation of albite was pre- or syntectonic.

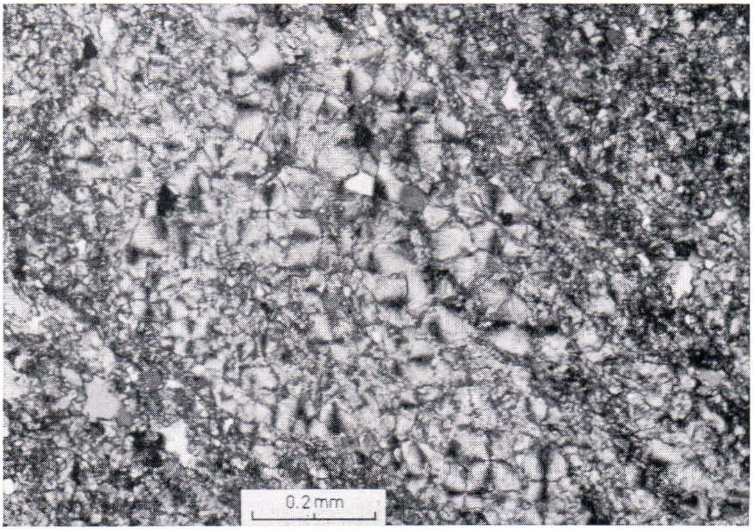


Fig. 21. Calcian rhodochrosite spherites in a Mn-carbonate-chert mesoband in Seppola. Thin section No. 21295. One nicol.

Macro- and mesobands of chert- and carbonate-banded rocks are associated with the sulphide-facies rocks of the Seppola iron-formation (Fig. 14 D). The thickness of the bands rarely exceeds 10—20 cm. According to electron microprobe analyses (Table 2, No. 21) the carbonate is a manganosiderite (or highly ferroan rhodochrosite). In addition to its two main constituents, chert and manganosiderite, the rock analysed contains some garnet (spessartite), pyrrhotite, biotite, chlorite, and amphibole.

Another mesoband about 10 cm thick, separated from the massive spessartite-bearing iron-rich black schist (p. 41) by a carbonate-rich phyllite layer about 3 cm thick, is composed mainly of fine-grained (Ø 0.01—0.02 mm) carbonate (Mn-siderite?) and chert or quartz (Ø 0.05—0.1 mm). The carbonate is often granular, like that described by James (1951, Fig. 5), except that in Väyrylänkylä it is cored by graphitic matter. Moreover, the band contains abundant transparent highly manganoan ankerite (MnO ~ 15 %, CaO ~ 25 %) (Ø ~ 0.03 mm) as irregular polygranular patches and spherulitic calcian rhorochrosite (MnO ~25 %, CaO ~15 %) as patches parallel to the bedding (Fig. 21). Furthermore, the band locally exhibits chlorite aggregates, some pyrrhotite and relics of garnet. The diameter of the calcian rhodochrosite spherites is generally about 0.1 mm. They show both spherulitic crosses and concentric layers, and are closely packed. Chlorite occurs rarely as spherites, but mostly as irregular or roundish granule-like aggregates sparsely distributed in the rock (Fig. 22). Sometimes it fills the intestitial spaces between the rhodochrosite spherites. In a very few cases chlorite forms the outermost concentric layer of a calcian rhodochrosite spherulite. The chlorite is brownish, it shows positive elongation and seems to be replaced by spherulitic carbonate, an indication that the spherulitic calcian rhodochrosite may be

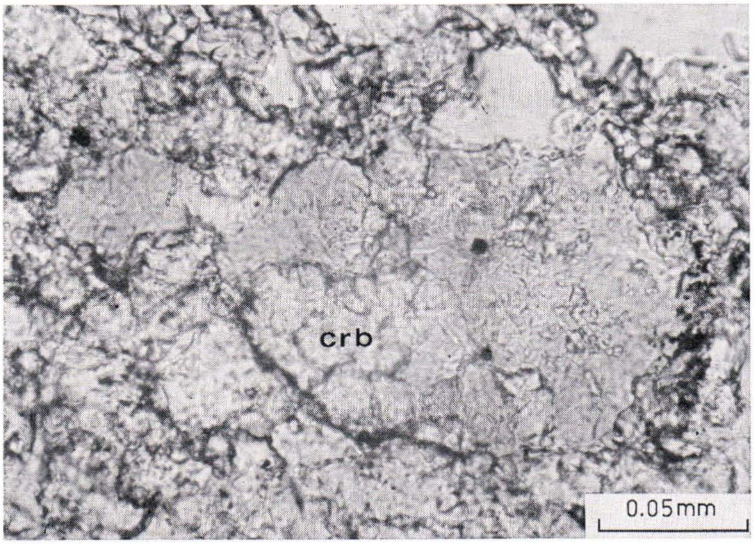


Fig. 22. Chlorite aggregate (greyish) partly replaced by carbonate (crb) in Mn-carbonate-chert mesoband in Seppola. Thin section No. 21295. One nicol.

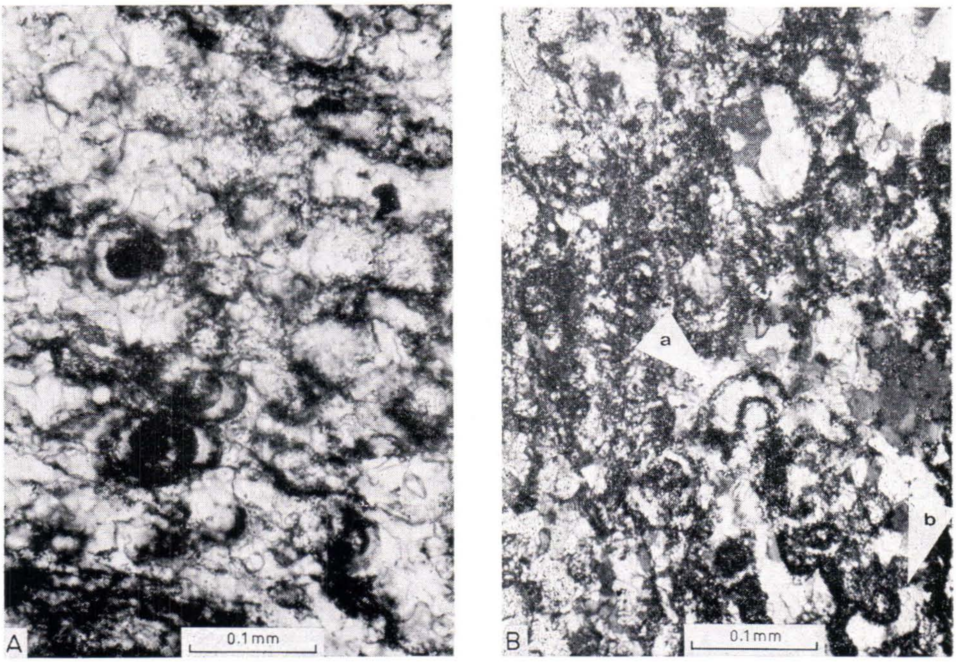


Fig. 23. Oolitic textures in Seppola. A) Carbonate ooids with nucleus of graphitic matter in Mn-carbonate-rich phyllite. Thin section No. 21294. One nicol. B) Carbonate ooids with either one, two or three rings pigmented by graphitic matter. Notice the semicircular ooids (a and b) and the recrystallized ooid in the upper part of the picture. Thin section No. 19083. One nicol.

a replacement product of iron silicate as shown by observations on sphaerosiderites from the post-Precambrian ironstones (e.g. Taylor *et al.* 1952, p. 454). The siderite spherulites are, however, also regarded as products of diagenesis formed by recrystallization of siderite originally disseminated in the groundmass (Carozzi 1960, p. 349).

Ooids or superficial ooids with diameter generally of no more than 0.05 mm or less (Fig. 23 A) abound near the contact of the gradational Mn-carbonate-rich phyllite and in the phyllite itself. Only exceptionally are there some ooids with a diameter of 0.1—0.2 mm. The ooids show asymmetric concentric layers marked by graphitic matter and are cored by the same matter. They are composed mostly of a single carbonate grain showing spherulitic extinction. The graphitic matter of the cores likewise shows concentric arrangement (Fig. 24 A). Carbonate spheres pigmented by graphitic matter are also often encountered. Their diameter is about 0.02—0.03 mm and at a relatively low magnification they appear very much like the 30 μ structures of La Berge (e.g. 1973, Fig. 9). If a greater magnification is applied the balls appear to be composed of one or more siderite grains surrounded by one or two rings of graphitic matter (Fig. 24 B). The outer ring may show angular texture resembling a crosscut of a hexagonal crystal. The inner ring is often poorly developed and spherical but irregular as if etched. Furthermore, the rock contains com-

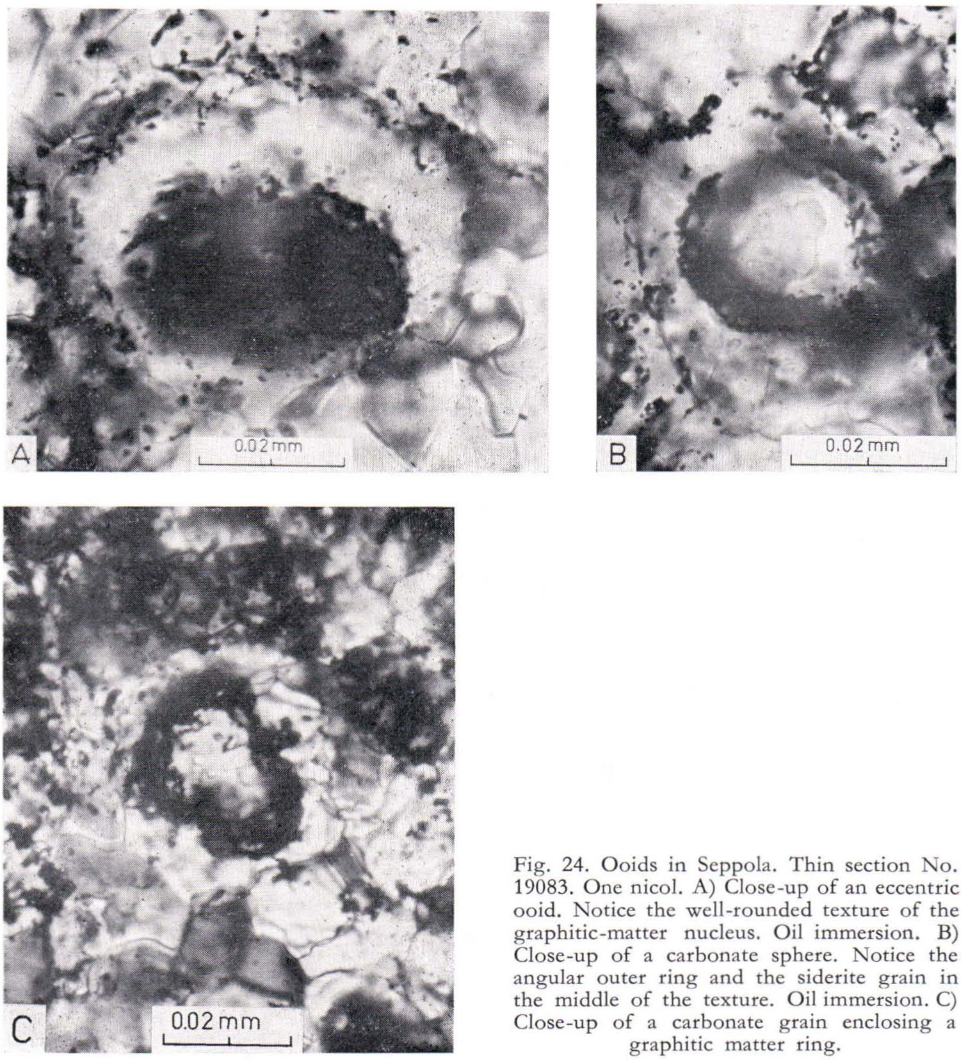


Fig. 24. Ooids in Seppola. Thin section No. 19083. One nicol. A) Close-up of an eccentric ooid. Notice the well-rounded texture of the graphitic-matter nucleus. Oil immersion. B) Close-up of a carbonate sphere. Notice the angular outer ring and the siderite grain in the middle of the texture. Oil immersion. C) Close-up of a carbonate grain enclosing a graphitic matter ring.

paratively coarse-grained (Ø ~ 0.03 mm) transparent carbonate as roundish polygranular fragments markedly greater in size than the oodis.

The carbonate-rich phyllite contains an abundance of spessartite porphyroblasts, pyrrhotite and graphite dissemination, as well as thin biotite-rich bands and aggregates of coarse-grained quartz. The carbonate of the rock is fine-grained (Ø 0.01—0.03 mm) and contains about 20 % CaO, 10—15 % FeO, 12—15 % MnO, and 6 % MgO. The texture of the rock is oolitic. Besides the texture shown in Fig. 23 A, there are also ooids with a double ring of graphitic matter but no nucleus. Also this carbonate shows spherulitic extinction. The space between the two graphite rings is often composed of quartz or fine-grained carbonate that differs optically from the spherulitic carbonate. Both the outer and inner rings are frequently indented, which indicates probable replacement. This type of ooid resembles that of the manganoan siderite in Kittilä (Paakkola 1971, Fig. 46). There are, however, also ooids with interiors composed of polygranular Mn-carbonate (CaO ~ 21 %, MnO ~ 12 %, FeO ~ 10 % MgO ~ 6 %, Fig. 23 B). Some ooids display, in their central part, a large carbonate grain surrounded by fibrous carbonate, whereas some carbonate grains enclose a ring of graphitic matter (Fig. 24 C).

It is significant that these oolitic textures are met with in the contact zone between the Mn-carbonate-chert mesoband and the massive spessartite-bearing iron-rich black schist. This indicates a gradual change from the pure chemical sedimentation stage to the deposition of rocks of mixed clastic-chemical sedimentary origin. As to the origin of the ooids, the well-rounded and concentric graphitic-matter nucleus of the ooids is of aggregationary origin, but the enveloping layers grew at the site of deposition with little if any later transportation. Thus, the Seppola ooids resemble more the Recent off-shore calcareous ooids from Laguna Madre (Freedman 1962) and the chamosite ooids from the Jurassic Winter Gill ironstone (Knox 1970) than the ooids from high energy environments. In accordance with the general concept of oolitic textures, it seems quite probable that the spherulitic carbonate and the quartz of the outer rings are of diagenetic replacement origin. The relatively coarse-grained carbonate of the granule-like aggregates is a recrystallization or replacement product of earlier carbonate mineral.

Carbonate-amphibole rocks of the CF at contact zones

Thin seams of light-coloured rock rich in carbonate (siderite) and containing appreciable, but variable amounts of amphibole (grunerite-cummingtonite) are encountered locally in the upper contacts of the Pääkkö and Iso Vuorijärvi iron formations. The rock is markedly deformed, which appears as intense schistosity and, in comparison with other iron-formation rocks, as relatively coarse grain size. Moreover, the primary banding of the rock has been almost completely destroyed. On the basis of its mineral assemblage (carbonate-amphibole-chert ± apatite ± biotite ± pyrrhotite) the rock belongs to the CF, even though it is generally conspicuously deficient in iron and thus represents a gradation from iron formation to overlying black schists. Owing to its gradational nature and higher grade of metamorphism this kind of rock has been separated from the dominant CF rocks. Fig. 25 shows a carbonate-amphibole rock with well-developed hypidioblastic texture and unusually coarse-grained apatite.

Iron-rich black schists and phyllites of the SF

In Iso Vuorijärvi, Körölä and Pääkkö the SF comprises massive or laminated, iron-rich black schists difficult to distinguish from »ordinary» black schists. Their

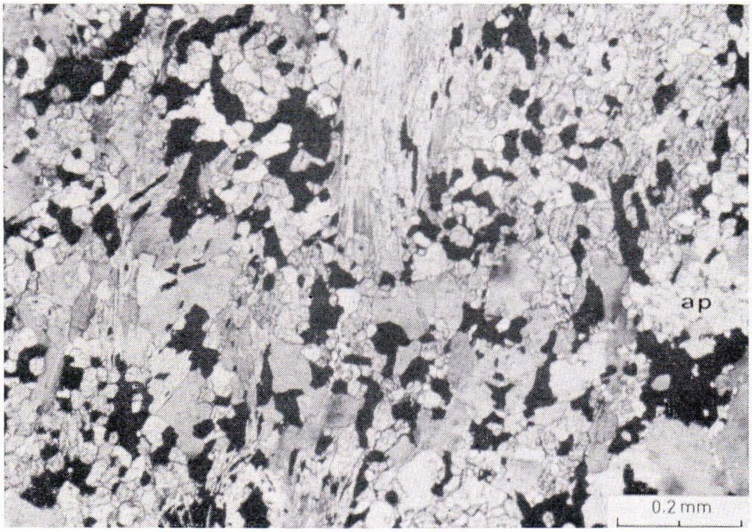


Fig. 25. Biotite- and pyrrhotite-rich part of an carbonate-amphibole rock from Iso Vuorijärvi showing well-developed hypidioblastic texture. Black is pyrrhotite, grey with pleochroic haloes is biotite, white prisms are amphibole, grey grains are carbonate and white grains on the left apatite (ap). Thin section No. 17898. One nicol.

richness in iron can be predicted on the grounds of their occurrence as macrobands in iron formations and of their predominant iron sulphide, which is pyrrhotite and not pyrite as in the »ordinary» black schists (see Väyrynen 1928, p. 98). In thin section the rock is fine-grained, laminated and blastoclastic and may show graded structure. The main minerals are biotite, pyrrhotite, quartz, plagioclase and graphite. Both quartz and plagioclase are clearly of clastic origin. The rocks in the Körölä iron formation are further characterized by an abundance of reddish garnet (almandine) porphyroblasts, and sometimes with amphibole bundles. In Iso Vuorijärvi some sulphide-facies macrobands contain light-grey pellet-like spots a few millimetres in diameter. The spots are composed mainly of almandine (X-ray identification by Pekka Kallio) with some quartz, biotite and a chlorite-like mineral.

The SF of Seppola consists of iron-rich black schists and phyllites, which are distinguished by the fact that the former are a little richer in graphite and are consequently darker in appearance than the latter (see Fig. 26 A and B). A distinction between these two rock types is not always readily made. The two most striking indications that these rocks should be classified as iron-formation rocks are the exceptionally abundant pyrrhotite and white chert mesobands. In comparison with the eastern SF rocks described above the most characteristic feature of the Seppola rocks is the striking banding in grey and black and the ubiquitous white garnet porphyroblasts (Figs. 26 and 27). Moreover, they are also lighter in colour than the former, owing to the relatively much lower graphite content. Yet one more important difference is that no chert bands have been met with in the former rocks.

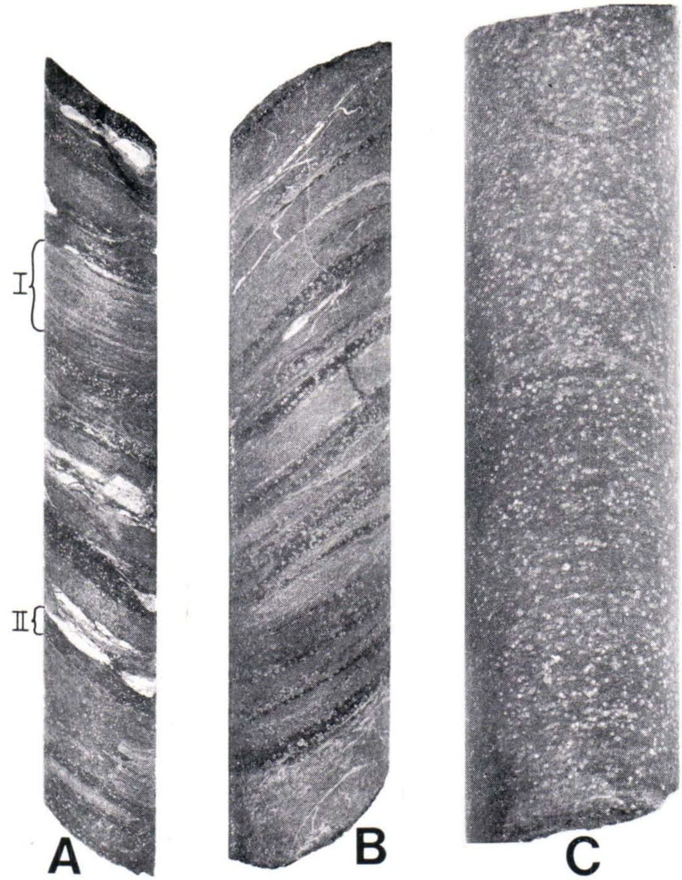


Fig. 26. Drill-core samples (Ø 3 cm) of SF rocks analysed from Seppola (Table 6, Nos. 3—5). A) Spessartite-bearing iron-rich black schist with white chert mesobands. The parts from which partial analyses (Table 6, Nos. 6 and 7) were done are marked by I (carbonate-pyrrhotite-spessartite band) and II (graphite-rich spessartite-biotite band). B) Spessartite-bearing iron-rich phyllite. Notice the enrichment of spessartite into dark graphite- and biotite-rich bands. C) Massive spessartite-bearing iron-rich black schist. Photos by E. Halme.

Thin-section studies and microprobe analyses (Table 2) reveal the manifold mineralogy of the Seppola rocks. The main minerals are biotite, garnet, pyrrhotite, carbonate, amphibole, chert, quartz and K-feldspar, the mutual proportions of which vary greatly in different bands. The following mesoband types can be distinguished: biotite-garnet ± K-feldspar; carbonate-pyrrhotite ± garnet; garnet-graphite ± biotite (Fig. 26 A, II); amphibole ± garnet ± carbonate ± pyrrhotite; carbonate-chert; and chert. Pyrrhotite, the main iron mineral, occurs as both faint dissemination and as veinlets. Its occurrence is mainly controlled by stratification. Microprobe determinations indicate that the garnet contains MnO enough to be called spessartite (Table 2,

Fig. 27. Banding in spessartite-bearing iron-rich phyllite schist from Seppola. On the right there is a garnet-bearing biotite-rich band, in the middle a garnet-bearing amphibole-rich band and on the left a band rich in carbonate, amphibole and pyrrhotite. Notice the occurrence of garnet, which is discussed in more detail in the text (p. 68). Thin section No. 20358. One nicol.

Nos. 14—16). Spessartite regularly exhibits a very unusual skeletal and zoned texture that will be described in detail on page 68. Amphibole contains sufficient manganese to be called manganoan cummingtonite (Table 2, No. 5). Even the biotite is markedly richer in manganese (Table 2, No. 10) than the biotite in Pääkkö (Table 2, Nos. 7—9). In addition to clastic quartz the rock may contain intraclasts of chert. K-feldspar follows strichtly biotite-rich bands. When encountered in greater amounts, it shows weak porphyroblastic features.

The massive spessartite-rich black schists close to the oolitic carbonate-rich phyllite (p. 36) show textural features not met with in other iron-rich metapelites of Seppola. It is comparatively coarse-grained, and both biotite and carbonate (manganoan ankerite, MnO $\sim$ 5 %, FeO $\sim$ 15 %, MgO $\sim$7 %, CaO $\sim$ 26 %) show helicitic relic textures of oolitic rings (Fig. 28). Also abundant are aggregates, black in thin section, that consist mainly of graphitic substance and carbonate, but that sometimes also contain small amounts of quartz and biotite (Fig. 28 B). In addition to garnet porphyroblasts the rock contains yellowish brown pseudomorphes of some unknown mineral which has grown in much the same way as the skeletal garnet.

The carbonate-pyrrhotite-spessartite band analysed (Table 6, No. 6) reveals carbonate-rich parts with many carbonate spheres coated by graphitic matter. The

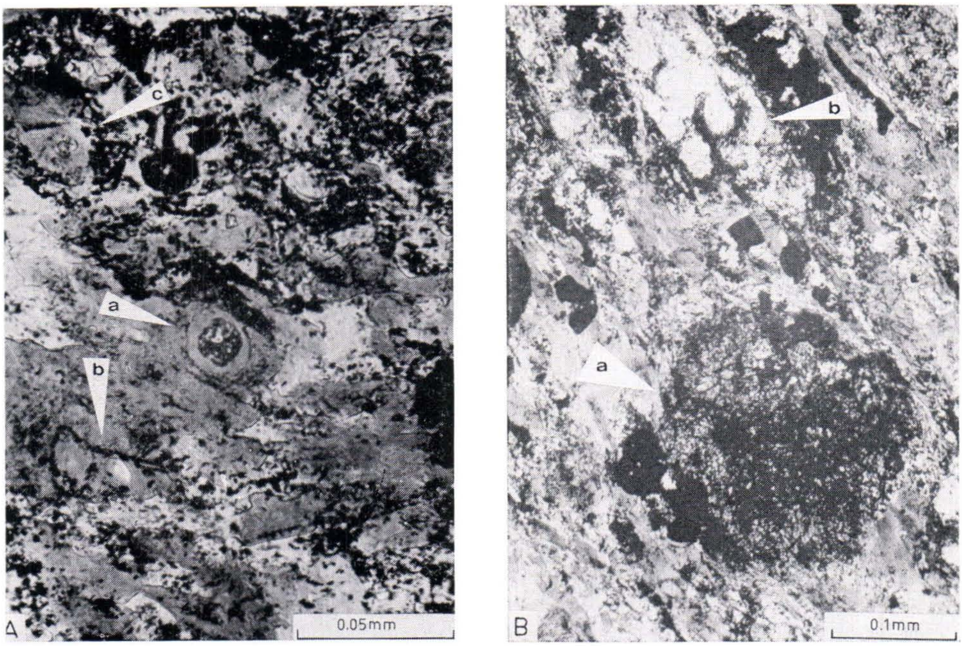


Fig. 28. Textures shown by the massive spessartite-bearing iron-rich black schist from Seppola. Thin section No. 19082. A) Biotite-rich part showing porphyroblastic biotite with helicitic oolitic textures. a = relict ooid with nucleus of graphitic matter, b and c = relict ring textures inside which there is still some carbonate to be seen. White is carbonate and quartz, black pyrrhotite. One nicol. B) Graphitic matter-carbonate aggregate (a). In the upper part there are two relict rings of graphitic matter in manganoan ankerite (b). One nicol.

diameter of the spheres is only 0.01—0.03 mm and they resemble the siderite spheres described on p. 36. The black graphite-rich band analysed (Table 6, No. 7) contains a wealth of porphyroblasts of skeletal spessartite embedded in matrix rich in carbonate, biotite and graphite. The band contains also pyrrhotite but in comparison with its neighbouring carbonate-rich bands it is poor in sulphides.

To sum it up, the lithology, petrography and mineral chemistry of the SF rocks in Seppola differ so drastically from those in the eastern iron formation of the Salmijärvi basin, that simple correlations between them is questionable. As a whole, however, the Seppola iron formation can be considered as a sulphide facies contaminated by mesobands of mixed silicate-sulphide and carbonate facies, whereas in the east the SF has become more strictly separated from the other facies. On the other hand, it is possible that the marked lithologic differences are due in part to the comparatively small contents of graphite in the Seppola rocks which makes their banding so readily discernible.

Noncherty garnet-bearing iron-silicate interbands of the MOSF and CF

Especially within the MOSF rocks in Pääkkö there are numerous green- or brownish-coloured, silicate-rich interbands, which are further distinguished by the abundance of red garnet porphyroblasts, deficiency in magnetite and complete lack of chert bands. The thickness of the bands varies from some centimetres to 30—40 cm. They often show bedded or laminated structure (Fig. 29) or sometimes even graded bedding. They frequently contain black carbonate-rich interbands, 1 to 2 cm thick, of probable retrograde origin (p. 71).

The interbands are composed mainly of amphibole (grunerite), garnet (almandine) and biotite, the mutual proportions of which vary greatly (Table 1, Nos. 6 and 7). The amount of garnet is generally 5 to 20 %, but in extreme cases it can comprise as much as 70 to 80 % of a single band. The two remaining main minerals are amphibole and biotite, either of which may predominate. A typical texture exhibits garnet porphyroblasts, commonly 2 to 3 mm in diameter, and amphibole needles or bundles in a fine-grained biotite matrix (Figs. 30 and 31). The garnet tends faintly, but commonly to have a radial or oblong appearance like the spessarite in the sulphide facies of Seppola. In the amphibole porphyroblasts green-bluish green amphibole often rims grunerite (Fig. 52). Biotite has frequently altered into chlorite and contains abundant minute zircon inclusions. In addition to the three main components, the interbands generally contain some percents of magnetite, pyrite, graphite and carbonate and a little quartz, pyrrhotite and plagioclase. The quartz is fine-grained and of markedly clastic appearance. It may represent intraclastic chert material. One band contains oblong aggregates, 1 to 1.5 mm long, lying parallel to the bedding and consisting of quartz with minor biotite (Fig. 32). They may either represent relict

Fig. 29. Laminated texture of a biotite-amphibole-garnet interband from Pääkkö. Greyish matrix is biotite, white prisms are grunerite, white spots with ragged margins (e.g. in the right upper corner) are porphyroblastic oligoclase. Black lamination is due to graphitic substance. Thin section No. 18054. One nicol. Photo by E. Halme.

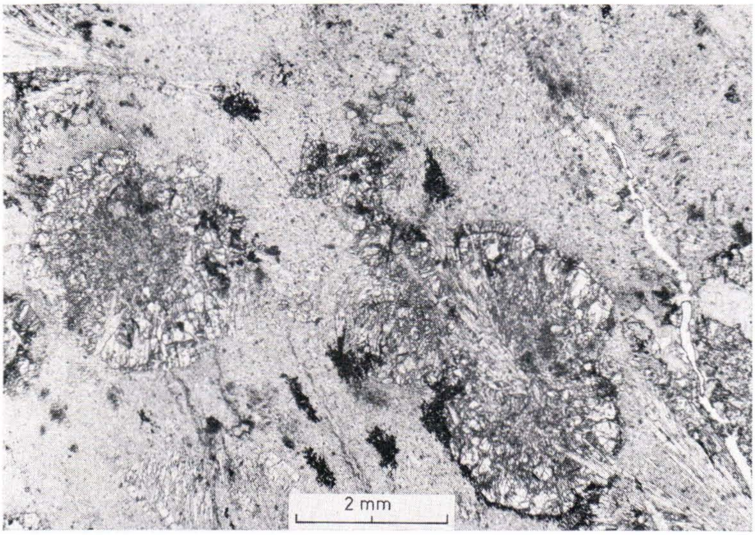


Fig. 30. Texture of a biotite-amphibole-garnet interband from Pääkkö. Matrix is composed of biotite, white prisms are amphibole. Notice the black pyrite rimming roundish garnet porphyroblasts, the cores of which are altered to biotite. Thin section No. 20441. One nicol. Photo by E. Halme.

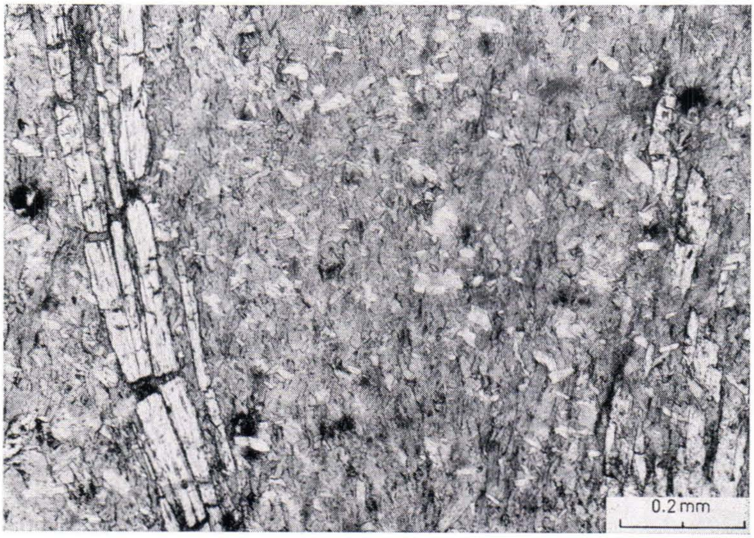


Fig. 31. A close-up of the matrix of the biotite-amphibole-garnet interband in Fig. 30. Matrix is composed solely of biotite. White prisms are amphibole. Thin section No. 20441. One nicol.

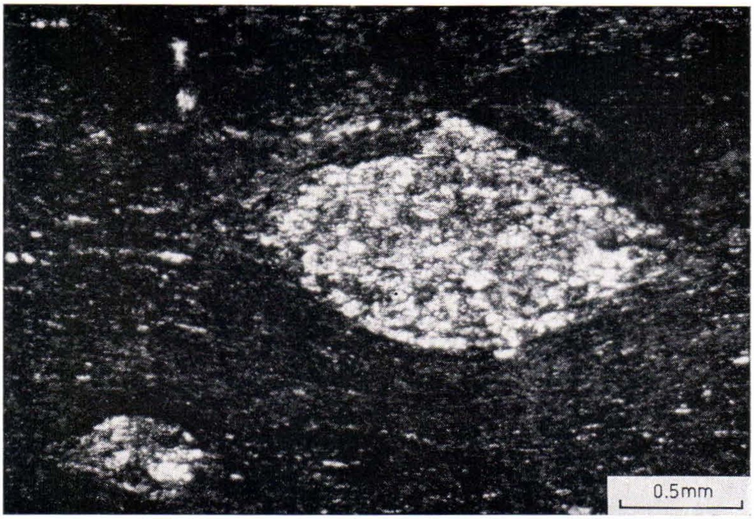


Fig. 32. Probable quartz-biotite intraclasts or relict chert-biotite »islands» in a biotite-garnet-amphibole interband from Pääkkö. Thin section No. 19314. Crossed nicols.

»islands» of primary quartz-rich bands replaced by silicates during diagenesis or be of intraclastic origin. Furthermore, one band contains small porphyroblasts of oligoclase (Fig. 63 A). It is noteworthy that tourmaline is present as a sporadic accessory mineral of the rock. Thus, it is obvious that the interbands described above are of clastic origin, although the quantity of clastic grains is much lower than that of the fine-grained biotite. Especially the interbands rich in both biotite and garnet bear a marked petrographic resemblance to the biotite- and garnet-rich mesobands in the SF of Seppola.

Microprobe analytical data of the main minerals and an accessory green, ferrotschermatic amphibole rimming grunerite are given in Table 2 (Nos. 3, 4, 6, 8, 9, 11 and 12). The plagioclase from one band contains 4.3 % CaO and 9.4 % Na_2O, and the porphyroblastic oligoclase 3.3 % CaO and 9.6 % Na_2O.

If there is no danger of confusion the interbands in question are from now on simply called garnet-bearing interbands.

Phosphorite interbands of the iron formations proper

On a global scale, the Precambrian iron formations in Kainuu are, according to our present knowledge, exceptionally rich in phosphorus [1]). It has been proved (Laajoki 1975 b) that the uniqueness of the Väyrylänkylä formations is due to the presence

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[1]) Berge (e.g. 1973) has described the Precambrian Goe Range iron formation with horizons relatively rich in both phosphorus and aluminium. However, because these horizons are situated in the secondary enrichment zone of the itabirite with phosphorus and aluminium contents normal for those in Precambrian iron formations in general, it seems likely that the high contents of these elements may be due to secondary processes.

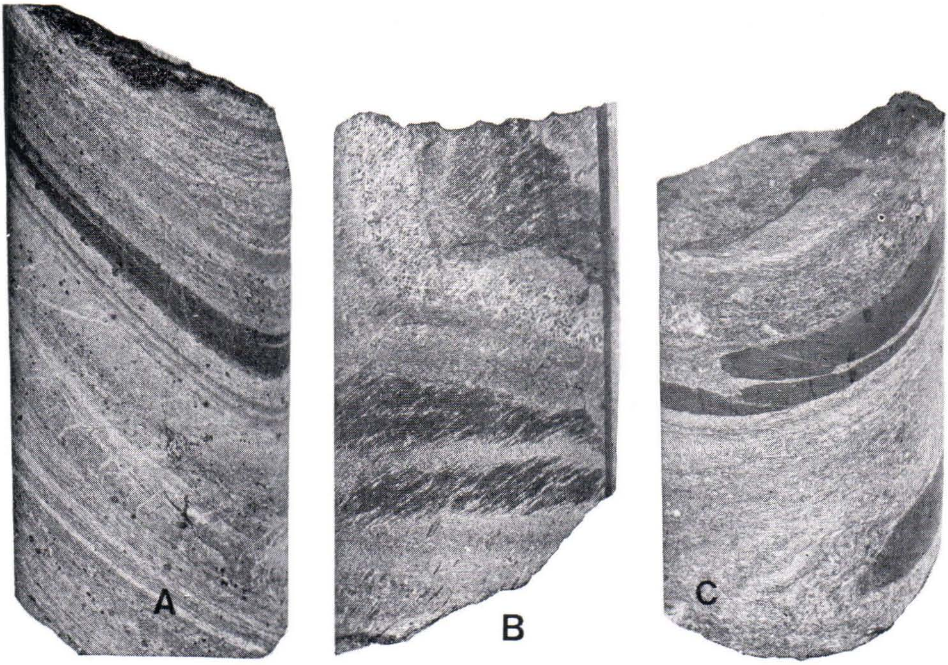

Fig. 33. Drill-core samples (Ø 3 cm) of phosphorite interbands in CF rocks from Pääkkö (DH 360). A) A thin interband consists of two microstrata (depth 53,35 m). B) Two interbands show strong shear foliation (depth 52,40). In the lower part the small grey lenses parallel to foliation are strongly sheared apatite-rich bands that were originally very thin. Thin section No. 21182 (Fig. 34) was made from this rock. C) Massive phosphorite interband sheared into two parts (depth 43,35 m). In the lower right-hand part there is a phosphorite lense at a hinge of a minor fold. Photos by E. Halme.

of marine apatite, which occurs, not only in close association with iron minerals as described earlier in this chapter, but also as phosphorite bands. Owing to the presence of graphitic substance the bands are black in colour and vary in thickness from 0.5 mm to 0.5 cm (Fig. 33). The black colour makes the bands easily distiguishable in light-coloured CF rocks, but in dark-coloured MOSF and MSOF rocks discrimination between them and the black magnetite- and amphibole-rich bands is possible only with the aid of magnet.

Under one nicol the phosphorite bands vary from almost colourless to almost opaque, depending on the abundance of graphitic substance (e.g. Fig. 35). The small grain size (0.005—0.01 mm) of apatite renders the bands almost isotropic under crossed nicols (Fig. 34). Their contacts with the adjacent bands are abrupt (Fig. 33). In general the bands contain both fine-grained apatite and graphitic substance although bands consisting almost solely of apatite are not unknown. They frequently have some percents of magnetite and carbonate (siderite) as well as biotite that has often altered into chlorite, amphibole, pyrrhotite and quartz.

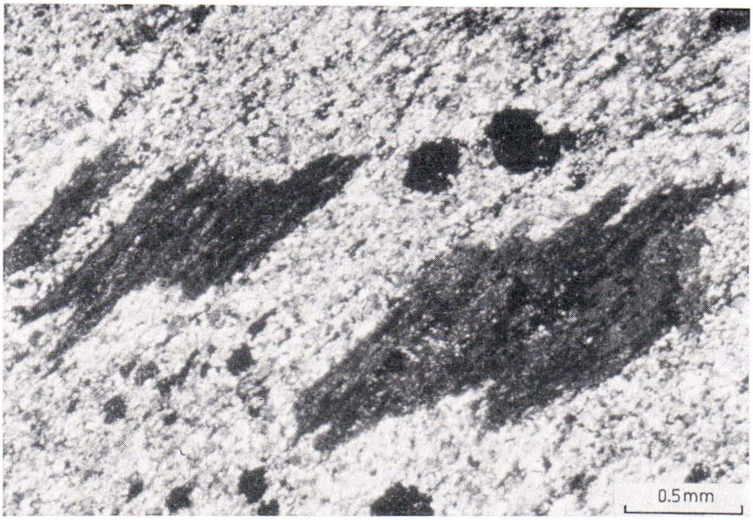

Fig. 34. Fragments of a shear-folded phosphorite interband (grey). Matrix is siderite and opaque is magnetite. Thin section No. 21182. Crossed nicols.

As a rule, the bands show strong foliation, the thin bands being sheared into small phosphorite lenses, and probable earlier textures in the thicker bands being almost completely destroyed (Figs. 33 B and 34). These deformed bands show microgranular texture brought about by small apatite crystals (Ø c. 0.01 mm) and a fine pigment of graphitic substance. Occasionally clear quartz occupies the intestices between apatite crystals. Luckily some relatively mildly deformed bands have been found that show well-preserved textures analogous to those of some siderite rocks (p. 31).

Although sheared throughout, one band, about 0.5 cm thick, in the siderite rock of Pääkkö shows very a peculiar texture of ambiguous origin. It is characterized by two features. The first is a conspicuous abundance of colourless plates, which in thin section exhibit a shard-like appearance, and lie mostly subparallel to foliation (Fig. 35). We have not yet succeeded in separating the plates from their matrix of apatite and graphitic substance, but their shape has been established by etching tests. The plates are mainly apatite, although many of them also contain some quartz, and quite a few of the plates are almost wholly of quartz. In many cases it can be demonstrated that apatite replaces quartz. Thus, it seems apparent that before apatitization the plates were composed of quartz. Sometimes plates can be polybranched (Fig. 36) or may intersect each other (Fig. 37). These latter textures seem to represent cracks filled by quartz which was replaced almost completely by apatite during the progressive metamorphism.

The second characteristic texture of the band is due to the »vugs», mostly oval-shaped (Fig. 36), but not uncommonly angular (Fig. 38). Texturally, they are analogous to the carbonate-albite replacement textures of siderite rocks (p. 34). In this

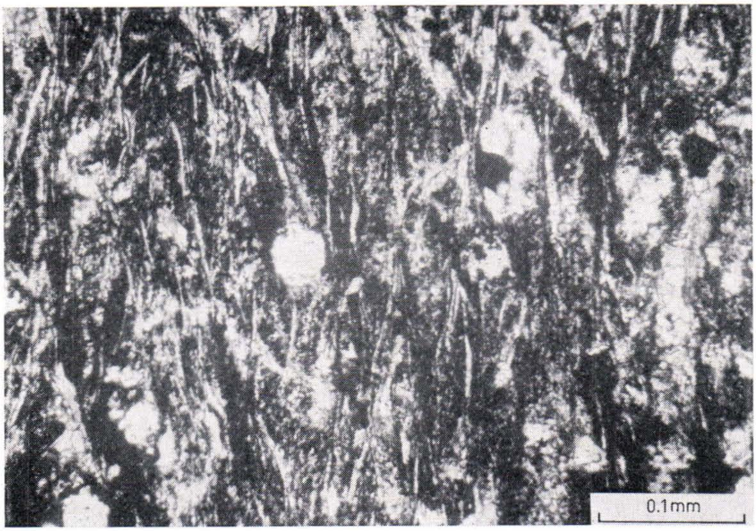

Fig. 35. Typical plate texture in phosphorite interband, White, shard-like plates are almost parallel to foliation (in vertical position), white roundish balls are composed of minute carbonate grains. Matrix is apatite and graphitic substance. Thin section No. 18686. One nicol.

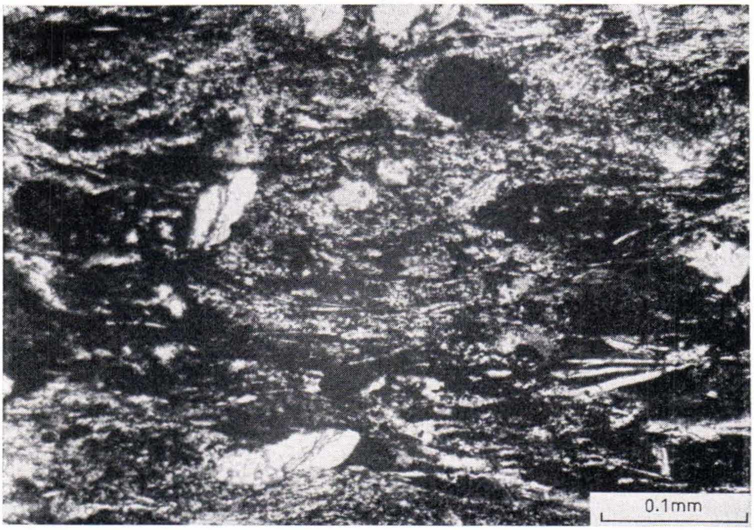

Fig. 36. The same phosphorite interband as in Fig. 35 showing both plate and replacement-vug texture. Notice the three-branched plate texture in the right lower corner. One nicol.

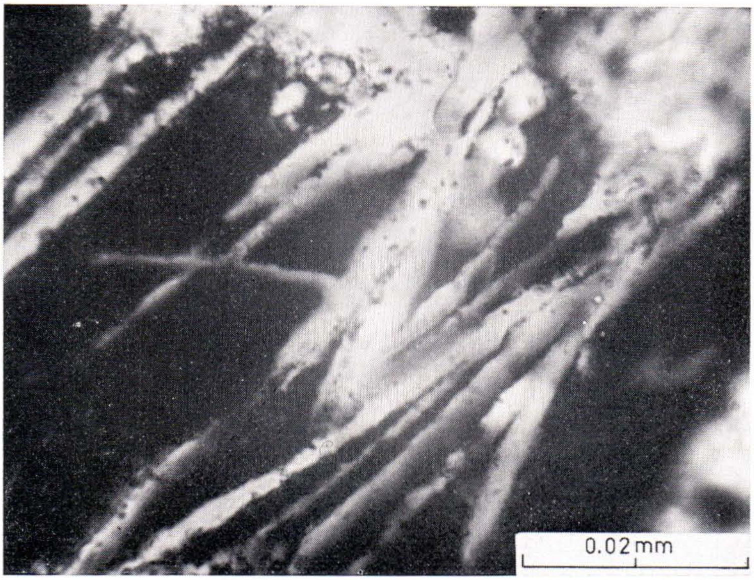

Fig. 37. Close-up of an unusual complex plate texture. The laths are mainly apatite. Notice the very sharp contacts of the black apatite-graphitic substance matrix with the laths. Thin section No. 18686. One nicol. Oil immersion.

case, however, the central part of the »vug» is composed of quartz, instead of albite, and the rims consist of water-clear, undeformed post-tectonic apatite, not columnar siderite. Moreover, the pared pigment row in the »vug» is more regularly in the central quartz. Often the pigment texture extends outside a »vug» into the matrix and is deformed by foliation. Hence, it is very apparent that the pigment row represents a relic of the extensions of primary plate (or rod) margins to the »vug». It can also often be seen how the marginal apatite, if grown long enough, has deformed this pigment row.

The band shows weak pelletal texture owing to the occurrence of small roundish aggregates richer in graphitic substance, and consequently, darker than the band in general. Because they have been enriched in biotite and chlorite, these pellets seem originally to have been richer in clay material than their surroundings. Biotite and chlorite have often formed at the expense of the quartz plates of the pellets, which suggests a relatively early birth for that texture (Fig. 59).

One phosphorite interband associated with an amphibole- and magnetite-rich mesoband (Fig. 9 B) shows micropelletal texture (Fig. 39 A). In this case the quartz plates are restricted to opaque graphitic substance — apatite pellets. The plates have needle-like cross-sections giving the false impression that they are big spicules (Fig. 39 B). Probably the texture represents crystal molds of some unknown diagenetic (e.g. gypsum) or early metamorphic mineral filled by quartz. An amphibole-biotite mesoband has similar pellets, which are either sparsely distributed or concentrated

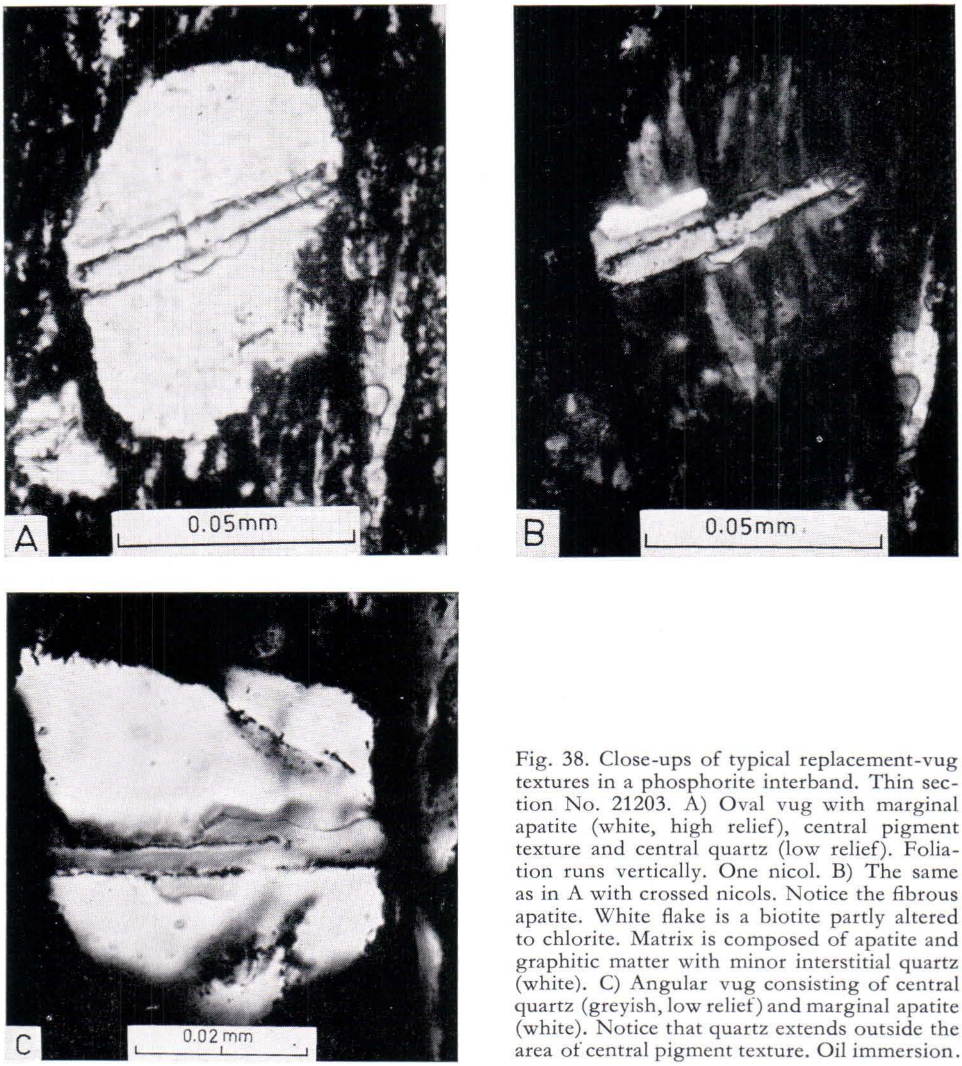

Fig. 38. Close-ups of typical replacement-vug textures in a phosphorite interband. Thin section No. 21203. A) Oval vug with marginal apatite (white, high relief), central pigment texture and central quartz (low relief). Foliation runs vertically. One nicol. B) The same as in A with crossed nicols. Notice the fibrous apatite. White flake is a biotite partly altered to chlorite. Matrix is composed of apatite and graphitic matter with minor interstitial quartz (white). C) Angular vug consisting of central quartz (greyish, low relief) and marginal apatite (white). Notice that quartz extends outside the area of central pigment texture. Oil immersion.

into thin bands. In these pellets as well there are abundant plates and networks of quartz as well as replacement »vugs» (Fig. 40).

The »vug» and framework textures shown by phosphorite interbands are so similar to the corresponding textures in siderite rock (p. 34) that they must have been formed through analogous processes. One unsolved problem is the origin of the mineral or texture of which only pigment relics are to be seen in the carbonate and apatite »vugs». The possibility that the original texture was primary, say, some algal-birth framework, is not quite out of the question especially, since in the apatite vugs these peculiar textures are so clearly restricted to the graphitic substance.

Dr. Risto Tynni kindly made a microfossil study of two phosphorite samples. No undisputed microfossil was found; only one problematic structure was detected. The study revealed that obviously spheromorphic structures of probable biogenic origin are not present in the samples studied.

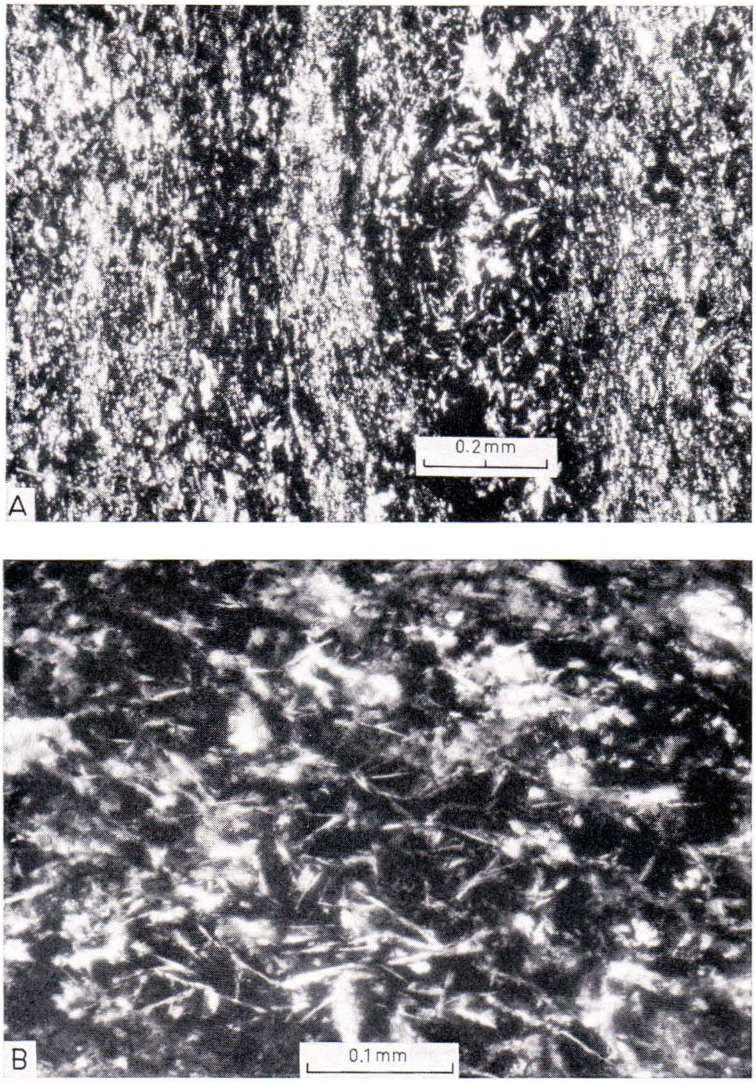

Fig. 39. A phosphorite interband showing pelletal texture. Thin section No. 19650. One nicol. A) General view. Greyish matrix is composed of apatite, graphitic substance and quartz. Black pellet (above the scale) contains plates of quartz (white) embedded in black matrix of graphitic matter and apatite. B) Close-up of a pellet. White quartz occurring as plates with needle-like cross-cuts and tiny vugs. Grey mineral bordering quartz is secondary apatite. Matrix is apatite and graphitic matter.

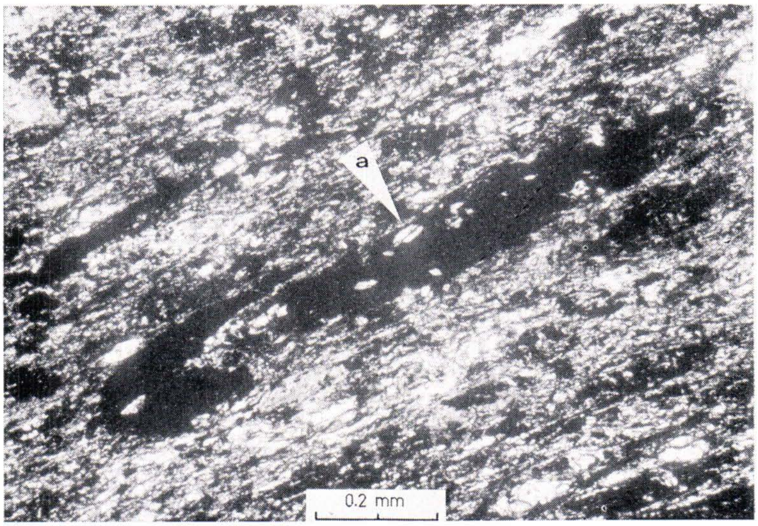

Fig. 40. Strongly elongated graphitic-matter- and apatite-rich pellet parallel to foliation in an amphibole + biotite-rich mesoband. Tiny quartz-apatite vugs (a) are visible in the pellet. Thin section No. 21202. One nicol.

Descriptive mineralogy of the iron-formation rocks

In this chapter the occurrences of the various minerals of the iron-formation rocks of the Väyrylänkylä deposits are decribed and their origins discussed. The description is based on microscopic study by the senior author in which particular stress was laid on non-opaque minerals. A total of 89 thin sections or polished thin sections, of which 52 were from Pääkkö, 16 from Iso Vuorijärvi, 16 from Seppola and 5 from Körölä, and 31 polished sections were examined. The thin sections, photographs of which are published in this study, are listed in Appendix 1. Results from microprobe analyses of some minerals are set out in Table 2. We had these analyses made in order to solve some problems concerning the whole-rock geochemistry and so they are not treated in full in this connection.

The Väyrylänkylä iron formations were metamorphosed throughout under the conditions of amphibolite facies and their minerals are now metamorphic. In the following, however, the metamorphic minerals do not include those minerals, probably primary or diagenetic, in which the effects of regional metamorphism appear only as an increase in grain size. A mineral is assumed to be primary (a recrystallized original precipitate), if it does not replace a pre-existing mineral phase (*cf.* Ayres 1972, French 1973, Floran and Papike 1975, for divergent views see Trendall and Blockley 1970, p. 272, and Dimroth 1976, p. 247).

Quartz

In a scientific sense, silica is a very important and essential component of the Precambrian iron formations, which could, in fact, be called silica-iron formations, and thus is the first mineral to be described. The following generations can be distinguished: 1) chert, 2) diagenetic quartz, 3) metamorphic quartz, 4) vein quartz, 5) retrograde quartz, 6) clastic quartz. Moreover there is still quartz of uncertain origin.

Chert occurs as fine-grained (Ø ~ 0.01—0.1 mm) primary bands (e.g., Figs. 8 and 15). In bands poor in graphite chert is markedly more coarse-grained than in bands containing abundant graphite (Fig. 41). Apparently, graphitic dust inhibited grain growth in chert during metamorphism (*cf.* Eskola 1932, p. 27).

Two of the quartz generations are likely to be of diagenetic origin. First, in quartz-siderite-banded rocks there are nodules consisting of relatively coarse-grained quartz (Ø ~ 0.05—0.2 mm) (Fig. 42 A), which is very similar to the coarse-grained quartz in the Sokoman Formation interpreted as either having replaced or dissolved and occupied the interior of a hematitic iron formation ooid (Dimroth and Kimberley 1976, Fig. 21). In Väyrylänkylä, no indisputable relic of ooid is visible.

Secondly, throughout the chert-rich MOSF and CF rocks at the borders of magnetite crystalloblasts, columnar quartz occurs oriented perpendicular to crystallographic boundaries of magnetite (Fig. 42 B). In the silica-rich matrix, the greater the

Fig. 41. Strongly shear-folded chert mesoband from Pääkkö. Axial-plane foliation vertical. The grey folded band is composed of fine-grained chert and dust of graphitic substance. Thin section No. 21070. Crossed nicols.

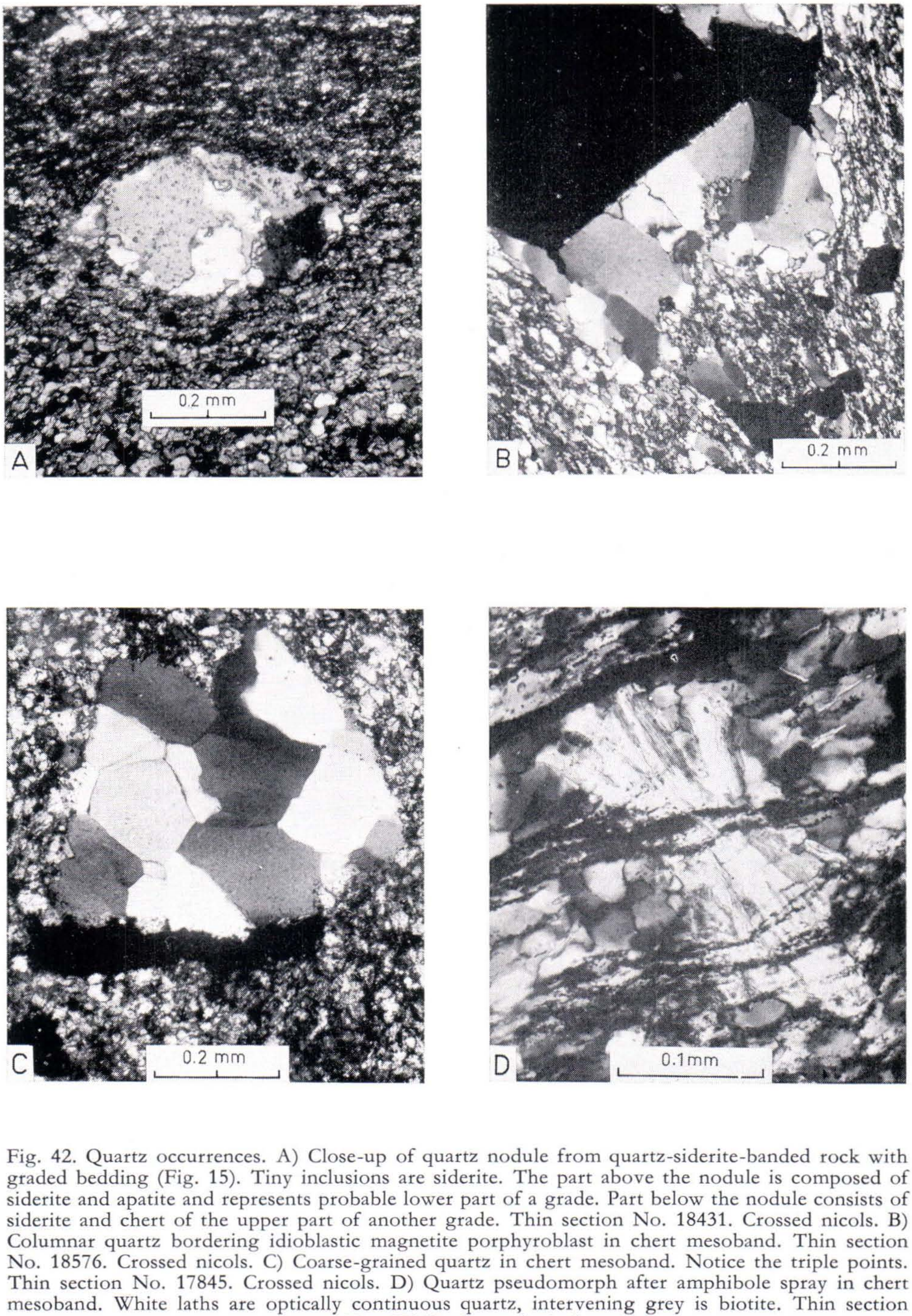

Fig. 42. Quartz occurrences. A) Close-up of quartz nodule from quartz-siderite-banded rock with graded bedding (Fig. 15). Tiny inclusions are siderite. The part above the nodule is composed of siderite and apatite and represents probable lower part of a grade. Part below the nodule consists of siderite and chert of the upper part of another grade. Thin section No. 18431. Crossed nicols. B) Columnar quartz bordering idioblastic magnetite porphyroblast in chert mesoband. Thin section No. 18576. Crossed nicols. C) Coarse-grained quartz in chert mesoband. Notice the triple points. Thin section No. 17845. Crossed nicols. D) Quartz pseudomorph after amphibole spray in chert mesoband. White laths are optically continuous quartz, intervening grey is biotite. Thin section No. 17843. Crossed nicols.

magnetite crystals, the more coarse-grained is the quartz. In its occurrence and, in some measure, in its fabric, this quartz is similar to the primocrystalline or encrusting quartz that is generally considered diagenetic (e.g. Dimroth and Chauvel 1973. Mukhopadhyay and Chanda 1972). Probably in Väyrylänkylä as well, the formation of this kind of quartz begun at the diagenetic stage, because it also rims hypidioblastic magnetite (Fig. 43 B). Apparently, however, its chief formation took place, together with the formation of magnetite porphyroblasts, during the progressive regional metamorphic stage. The pressure minima at the contacts of magnetite seem to have been favourable sites for the growth of this kind of quartz.

As already stated (p. 49) the quartz plates in the phosphorite bands are assumed to be early metamorphic replacement products of some unknown precursor.

The chert mesobands exhibit, especially in markedly deformed rocks, abundant coarse-grained quartz, the formation of which was not controlled by any primary structure (Fig. 42 C) indicating true metamorphic origin. In Seppola, this kind of quartz formed before garnet, because it is met with as helicitic inclusions at the borders of spessartite porphyroblasts.

In two cases, in a chert mesoband and in a phosphorite interband, retrograde quartz pseudomorphs after amphibole bundles (Fig. 42 D) have been encountered.

Clastic quartz grains are common in SF rocks and also, although in much lesser amount, in noncherty garnet-bearing interbands. In the former rocks this quartz may be mainly extraformational (in a narrow sense), whereas in the latter it seems to be intraclastic chert. As discussed earlier (p. 45), the latter rocks also contain quartz-rich balls of unknown origin. Because they are sometimes enclosed by garnet porphyrosblasts, they must have been formed before the crystallization of that mineral.

Magnetite

No magnetite has been found which could be proved to be primary. It may, however, be that the magnetite porphyroblasts in the siderite-rich parts of the grades (Fig. 15 A) are recrystallization products of some relatively early iron-oxide or hydroxide concentrations (granules?) of either primary intraclastic or diagenetic replacement origin. During increased metamorphism they grew more and more idioblastic (Figs. 43 A, B, and 16) and in a few cases became rounded as an outcome of shearing (Fig. 43 D). The logical explanation for the magnetite porphyroblasts in the iron-rich parts of the grades of quartz-magnetite banded rocks (Fig. 9 A) is that they originated in the same way. In this occurrence however, xeno- or subidioblastic porphyroblasts have not been met with, and all of them show well-developed crystal forms, presumably indicating a higher metamorphic grade for the rocks in general. Also in chert mesobands magnetite occurs regurlarly as idioblastic crystals (Fig. 43 C).

On the other hand, in amphibole-rich assemblages magnetite occurs typically as small xeno- or subidioblastic grains mostly arranged parallel to the bedding (Fig. 10). However, not uncommonly magnetite is arranged vertically or obliquely to

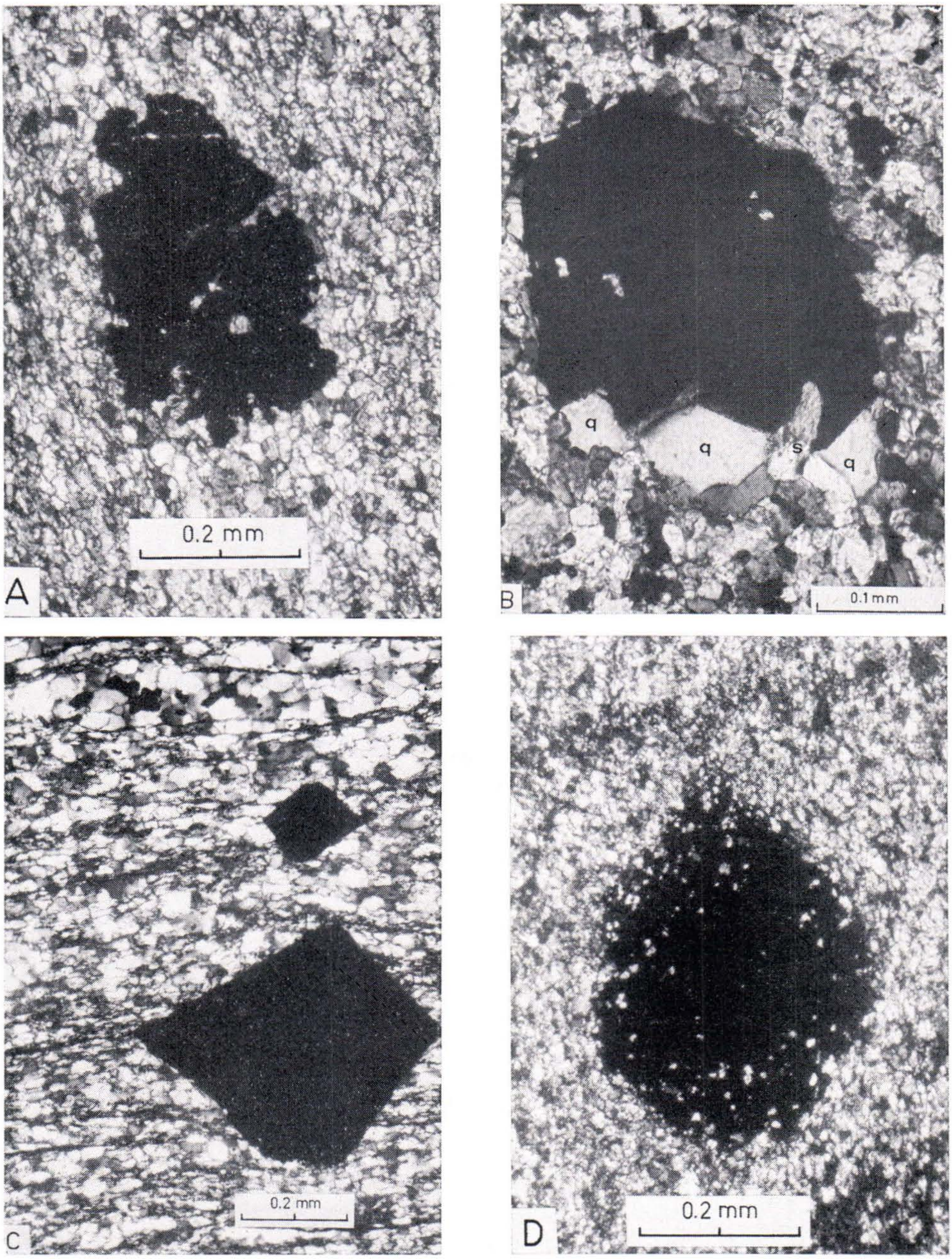

Fig. 43. Magnetite occurrences. A) Xenoblastic magnetite porphyroblast in a quartz-siderite-banded rock. Thin section No. 18431, crossed nicols. B) Almost perfectly idioblastic magnetite porphyroblast in a siderite rock. Notice the occurrence of quartz (q). The grain projecting into magnetite is siderite (s). Thin section No. 19074. Crossed nicols. C) Completely idioblastic magnetite porphyroblasts in a chert mesoband. Thin section No. 17843. Crossed nicols. D) Magnetite porphyroblast curved by shearing with longitudinal axis to foliation. Minute inclusions are siderite. Thin section No. 21212. Crossed nicols.

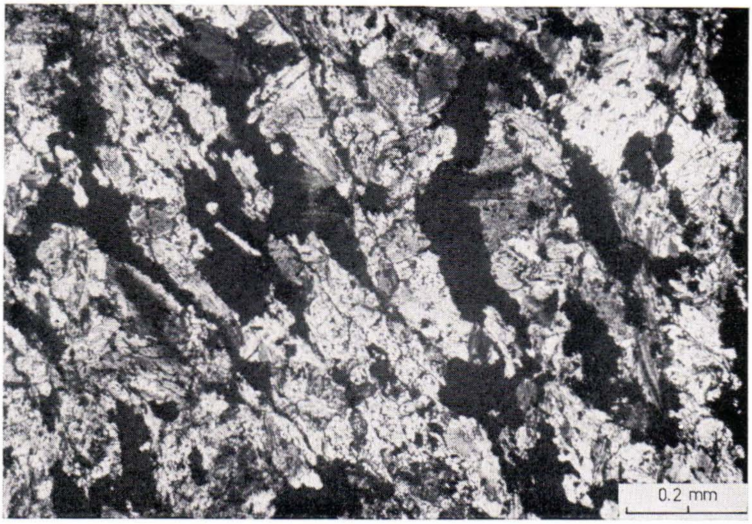

Fig. 44. Magnetite surrounding amphibole crystals. Bedding vertical. Thin section No. 19650. Crossed nicols.

Fig. 45. Magnetite surrounding masses of fine-grained amphibole. Bedding approximately vertical. Thin section No. 19081. Crossed nicols.

primary banding. In some cases this mode of magnetite occurrence is due to the combined effect of foliation and the metamorphic growth of amphibole (Fig. 44). In other cases, especially when magnetite runs vertically to primary banding and is of a relatively irregular appearance, its growth may have been controlled by compression (Fig. 45).

The iron-silicate-magnetite rock contains rather fine-grained hypidioblastic or idioblastic magnetite in the matrix of amphibole porphyroblasts or as inclusions in them (Fig. 12).

Amphibole

All the amphiboles in the Väyrylänkylä iron-formations are of metamorphic origin. Two amphibole species have been identified, the dominant grunerite(-cummingtonite) [1]) and an accessory ferrotschermatic hornblende green in thin section.

Grunerite is most abundant in magnetite- and/or amphibole-rich mesobands as well as in amphibole-rich mesobands of the MOSF, MSOF and CF and, of course, in the chert-mesobanded grunerite rock of the SFP. Moreover, it is a typical mineral in noncherty garnet-bearing iron-silicate interbands as well as in some siderite rocks and in the iron-rich black schists of Seppola. It is also an essential accessory mineral in chert mesobands.

In the SFP rocks grunerite typically occurs as pale yellow »suns», 5 to 10 mm in diameter. Because these porphyroblasts show helicitic microbanding and shear foliation marked by fine graphitic substance and magnetite dust, but are not themselves deformed (Fig. 13), it can be concluded that this grunerite was formed after or during the disappearance of the second main deformation stage of the South Puolanka area.

In the iron-silicate-magnetite rocks of the MSOF, amphibole occurs as porphyroblasts and as fine-grained matrix (Fig. 12). In magnetite- and amphibole-rich mesobands it mainly occurs as relatively small crystals (Figs. 9 B, 10, and 45); amphibole porphyroblasts have been met with only occasionally. In the graded varieties amphibole occurs as small crystals in what is probably the lower parts of the grades (Fig. 9 A). It is impossible to conclude with certainty whether this amphibole had some silicate precursor or whether it is a reaction product of earlier iron carbonate or oxide. When thin chert bands are interbedded with these iron-mineral-rich mesobands it can sometimes be demonstrated that chert bands grade laterally into amphibole (Fig. 46), which has often grown perpendicular to the banding plane (Fig. 47 A). Usually the part composed of amphibole is markedly thicker than the original chert band (Figs. 46 and 47 B). In Fig. 46 the immediate surroundings of the part of the band rich in chert contain chert as well as magnetite, whereas in the surroundings

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[1]) Because we have not determined systematically the exact chemical composition of this mineral and for simplification's sake we have called it amphibole. In addition to those analyses set out in Table 2 we made four check analyses, which gave almost identical results.

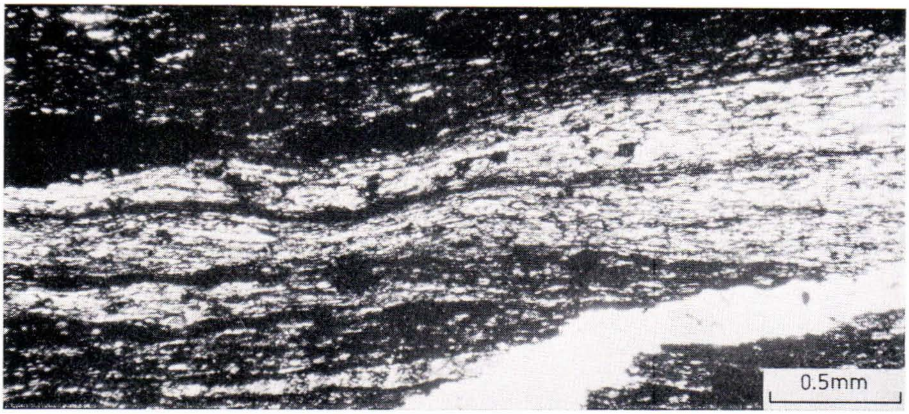

Fig. 46. A chert band almost completely altered to amphibole. At the thinnest point, at the left, the band is chert-rich, whereas at the right it is composed solely of amphibole. Surrounded by black magnetite-rich mesoband. White is quartz vein. Notice the thickening of band from the chert-rich to the silicate-rich part. Thin section No. 17843. One nicol. Photo by E. Halme.

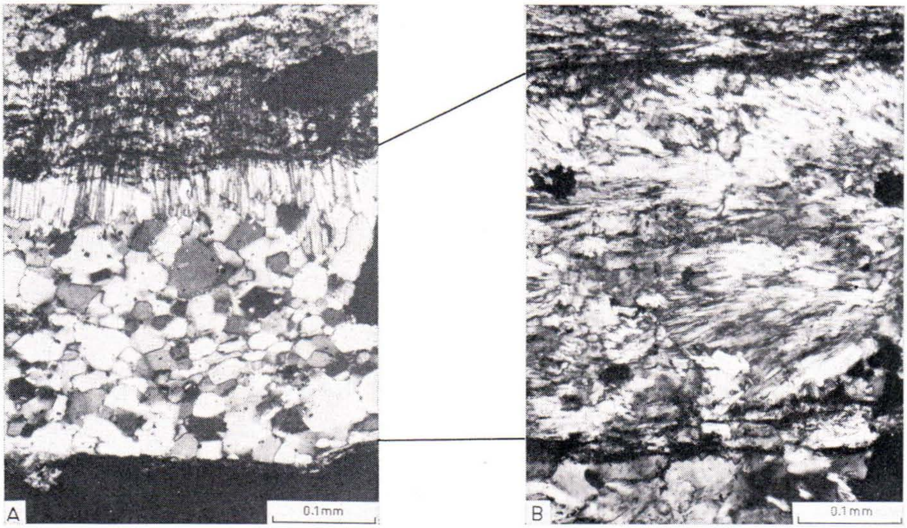

Fig. 47. A chert band showing amphibole. Thin section No. 19081. Crossed nicols. A) Portion composed mainly of chert with minor amphibole grown perpendicular to bedding plane. B) Portion composed of amphibole grown in zigzag fashion. At a distance of 4,5 mm from A.

of the amphibole-bearing part of the same band chert is no more to be seen. These examples indicate that the amphibole in question was formed at the expense of chert.

In chert mesobands proper amphibole occurs typically as porphyroblasts. Especially near the contacts of iron-mineral-rich mesobands there are often sprays of grunerite. Within the band itself porphyroblasts have grown either parallel to the bedding as in Fig. 48 A or perpendicular or oblique to the bedding, in which case

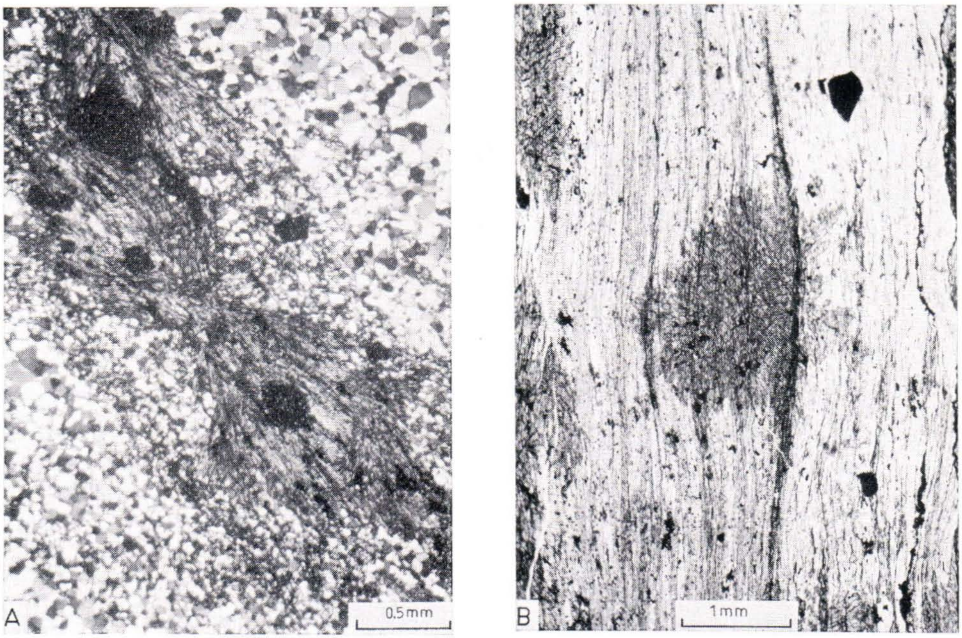

Fig. 48. Amphibole porphyroblasts in chert mesobands. A) An amphibole porphyroblast grown parallel to the bedding plane. Thin section No. 19648. Crossed nicols. B) Amphibole porphyroblast showing bending of microbanding. Thin section No. 17843. One nicol. Photo B by E. Halme.

they have produced deformation in the microbanding (Fig. 48 B). Moreover, in some chert mesobands with unusually large amounts of magnetite porphyroblasts, fine-grained amphibole is to be seen (Fig. 49).

The siderite-rich rocks of the CF that contain abundant chert are generally deficient in amphibole or contain it only in accessory amounts. On the other hand, those with only sporadic chert or quartz often contain ample amphibole. As a rule, this amphibole occurs as thin prisms, parallel or subparallel to the foliation (Fig. 50). In the siderite rock showing the peculiar texture described on page 31 there are amphibole porphyroblasts and shreds of amphibole that exhibit relics of these textures. Fig. 51 A shows a porphyroblast with relics of albite-plate texture with abundant recrystallized vug carbonate. Even the shreds show what seem to be relics of marginal-carbonate texture and albite network (Fig. 51 B). The amphibole in this rock was apparently formed at the expense of network albite.

In amphibole ± biotite-rich mesobands grunerite occurs as well-developed porphyroblasts with helicitic textures (Fig. 11). They are often rimmed by green hornblende. The occurrence of amphibole in noncherty garnet-bearing iron-silicate interbands is very similar to this. Thanks to the fine-grained matrix its crystals are, however, much easier to detect in the latter than in the former (Figs. 31 and 52). It quite frequently happens that grunerite crystals have grown into garnet porphyroblasts (Fig. 30).

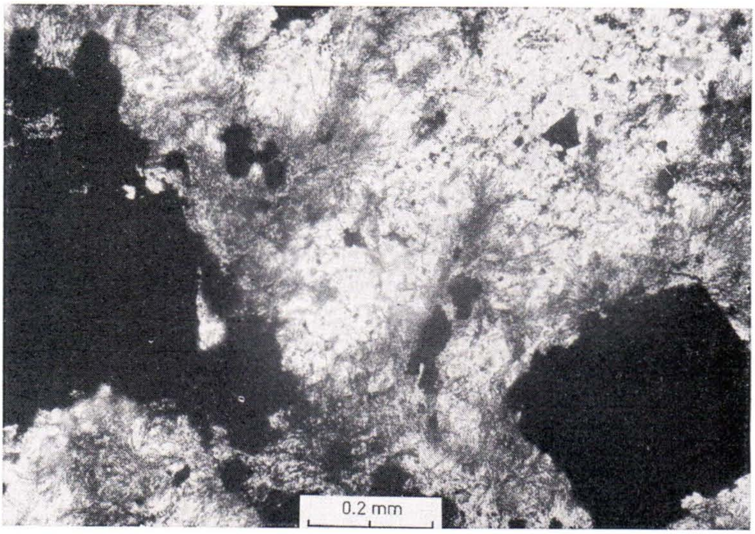

Fig. 49. Fine-grained amphibole (grey) in chert mesoband. Notice the concentration of amphibole crystals near magnetite porphyroblasts. Thin section No. 17845. One nicol.

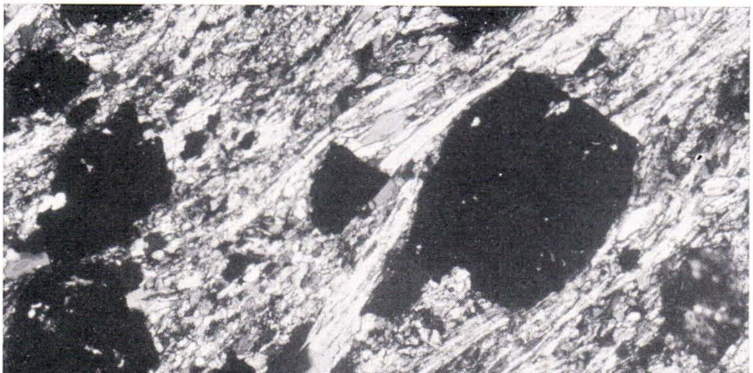

Fig. 50. Amphibole crystals parallel to foliation in magnetite-bearing siderite rock. Thin section No. 18685. Crossed nicols.

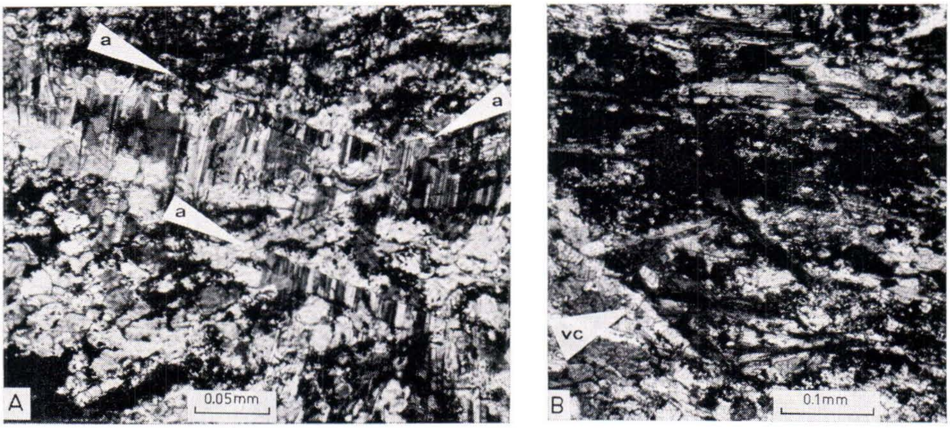

Fig. 51. Occurrence of amphibole showing relict textures in siderite rock. Both with crossed nicols. A) A part of amphibole porphyroblast (twinned) containing relics of albite-plate texture (a) and an abundance of recrystallized replacement-vug carbonate (bright). Thin section No. 21184. B) Amphibole (grey) formed at albite network, but now partly replaced by retrograde carbonate (bright or grey granular). On the left columnar replacement-vug carbonate (vc) is still visible. Thin section No. 17999. The thin section is somewhat too thick.

In Seppola, amphibole occurs in much the same way as in Pääkkö to which the foregoing description applies. It mostly occurs as fine crystals in divers bands or as bands of their own (Fig. 27). Furthermore, it has been met with as porphyroblasts in or near chert mesobands or in biotite- and garnet-rich bands. Both of these types have been replaced completely by retrograde carbonate (Fig. 53).

Carbonates

Carbonate is the most abundant mineral of the CF rocks and an essential constituent of the iron-rich metapelites in the Seppola iron formation. It also occurs as carbonate-rich bands or as an accessory in the other iron-formation rocks. Three carbonate species, siderite, manganosiderite-rhodochrosite and ankerite (ferroan ankerite in Pääkkö, manganoan ankerite in Seppola), have been identified with the aid of a microprobe (Table 2). Siderite is the dominant species in Pääkkö the ankerite occurring only in small amounts. In Seppola manganosiderite is the main mineral in the carbonate- and chert-rich meso- and macrobands; so far manganoan ankerite has been identified from two pelitic samples (p. 37 and 41). There seem to be at least four generations of carbonate: primary, replacement carbonate of diagenetic or early metamorphic age, metamorphic carbonate and retrograde replacement carbonate.

The fine-grained (Ø 0.01—0.03 mm) siderite (manganosiderite in Seppola) of the quartz-siderite-banded and -laminated rocks, as well as most of the carbonates of siderite rocks, which tend to be somewhat more coarse-grained owing to the higher grade of recrystallization, are considered primary. This conclusion is justified by the

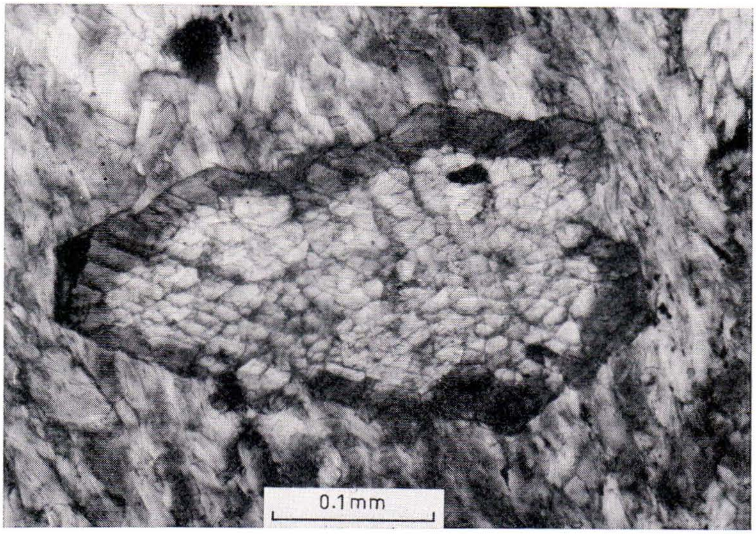

Fig. 52. Amphibole porphyroblast with outer zone of green hornblende in biotite-amphibole-garnet interband. Matrix is biotite. Thin section No. 18054. One nicol with condensing lense.

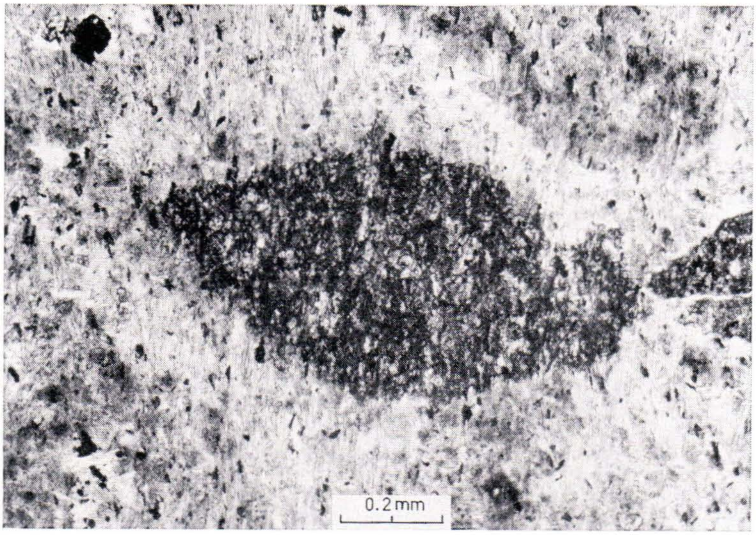

Fig. 53. Carbonate multi-crystal pseudomorph after amphibole porphyroblast. Notice the pale carbonate-chlorite alteration zone around the pseudomorph. White flakes are chlorite. Biotite-rich band in spessartite-bearing iron-rich black schist. Relatively unaltered rock appears in the upper right and lower left parts of the Figure. Thin section No. 18008. One nicol.

following characteristics: 1) Carbonate of this type shows no signs of replacing other minerals. 2) Its occurrence in what is probably the lower part of a siderite-chert microband (Fig. 15 B). It seems questionable that so regular and beautiful a structure could exist if the siderite were, say, a replacement product of pre-existing carbonate as suggested by Kimberley (1974) and Dimroth and Kimberley (1976). 3) This carbonate is often, but not necessarily, pigmented by black graphitic substance, whereas carbonates, as well as other minerals of replacement origin are generally transparent (sparry) and display a much larger grain size.

As was discussed on p. 38, the spherulitic carbonate of both spherulites and ooids in Seppola is a diagenetic replacement product.

The last two features of replacement origin mentioned above are developed in the replacement-vug carbonates (Fig. 19). As was discussed on page 34, this carbonate seems to be of relatively early metamorphic origin, but post-tectonic. In Table 2, analysis No. 20 was done from this type of siderite. The matrix siderite in the same thin section was also analysed. Its composition does not differ significantly from that of the replacement carbonate. During progressive regional metamorphism the »vug» carbonate recrystallized and often lost its columnar appearance.

In some thicker phosphorite interbands, there are abundances of small siderite balls (Table 2, No. 19) whose carbonate is more coarse-grained than that of the adjacent siderite bands (Fig. 54). This carbonate may be late metamorphic or retrograde, because it replaces, albeit seldom, both biotite and »vug» apatite. It often shows spherulitic extinction, but not spherulitic crosses.

The two ferroan ankerites analysed (Table 2) are clearly of metamorphic age. The first is met with fairly frequently, but generally as an accessory in magnetite- and amphibole-rich mesobands of the quartz-magnetite banded rocks. It occurs as either roundish granule-like grains or as a filling in the interstices of magnetite (Fig. 55). The grains are composed of single crystals and resemble texturally the siderite replacing the original silicate(?) granules in the Biwabik Iron Formation (French 1973, Fig. 5). The second type occurs as rhombic porphyroblasts and is typically twinned. Furthermore, it contains a helicitic zone of graphitic material, which apparently indicates an earlier carbonate phase (Fig. 56). This kind of idioblastic ferroan ankerite has been encountered in only one quartz-magnetite banded rock, in its chert-rich parts near magnetite- and amphibole-rich mesobands. Also in Seppola ankerite is true metamorphic, because it encloses helicitically earlier textures (Figs. 24 C and 28 B).

There are two minerals that were readily altered into carbonates during the retrograde stage: the amphibole of the Seppola iron formation (Fig. 53) and the garnet of Pääkkö and Seppola. The latter occurrences will be discussed later. Moreover, amphibole crystals oblique to the foliation in some siderite rocks of Pääkkö have also been replaced by carbonate (Fig. 20). This carbonate is bright like the replacement-vug carbonates, but is mostly equidimensional in appearance. In one case, however, there occurs in a siderite rock relatively rich in amphibole and magnetite

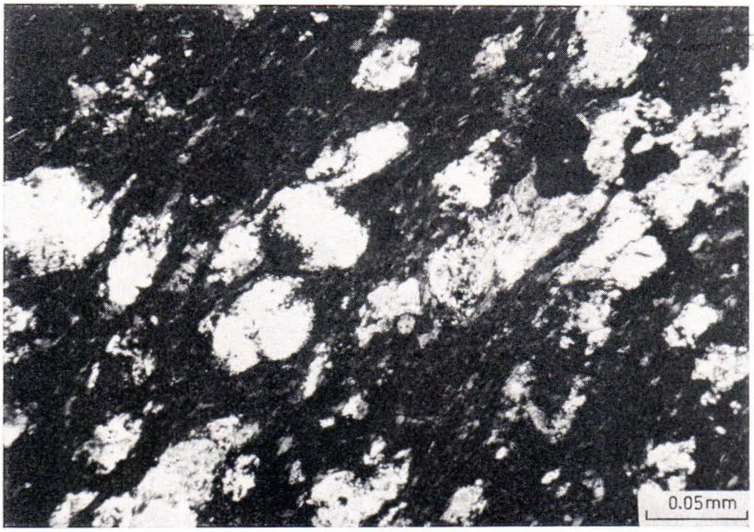

Fig. 54. Occurrence of carbonate (white) in phosphorite interbed. Matrix is composed of apatite and graphitic matter. Thin section No. 18686. Crossed nicols.

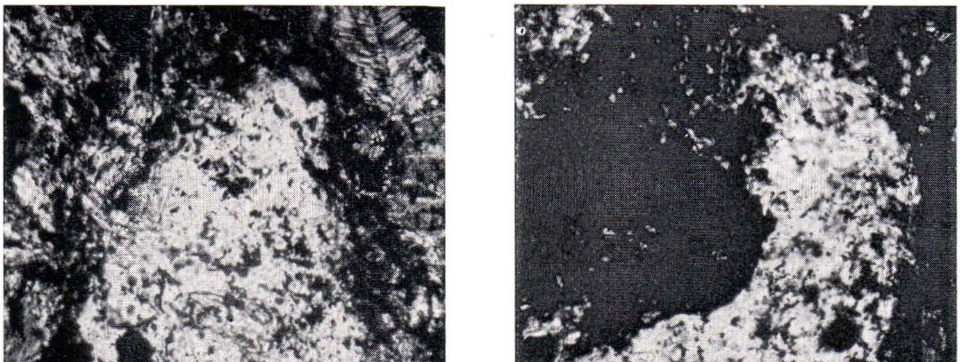

Fig. 55. Occurrences of xenoblastic ferroan ankerite. Thin section No. 18576. Crossed nicols. A) A typical roundish porphyroblast parallel to bedding. B) Xenoblast filling the interstices of magnetite.

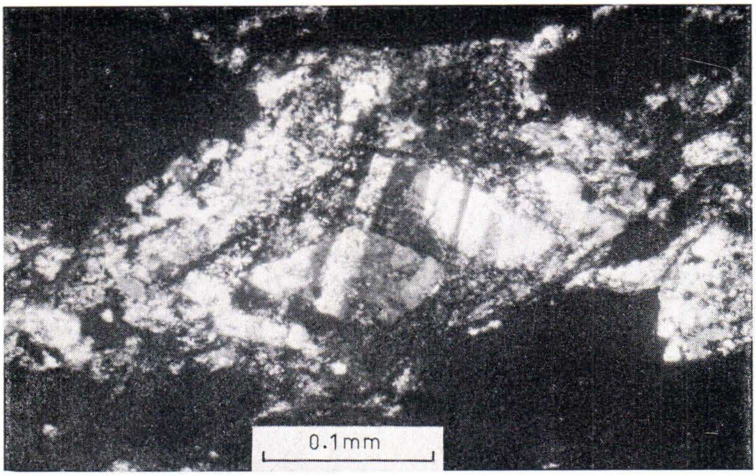

Fig. 56. Idioblastic ferroan ankerite porphyroblast showing twinning and a pigment zone. Thin section No. 21181. Crossed nicols with condensing lense.

columnar carbonate rimming amphibole porphyroblasts. This carbonate has clearly grown from the prism surfaces of the amphibole towards the matrix. In texture it is analogous to the columnar quartz rimming magnetite porphyroblasts (Fig. 42 B).

Biotite and chlorite

Biotite: Biotite occurs mainly in noncherty garnet-bearing interbands and iron-rich black schists. In both rocks biotite is undisputedly of metamorphic origin, but not of the same phase. Biotite also abounds in some amphibole + biotite-rich mesobands. Retrograde biotite has likewise been encountered. Biotite is frequently met with as an accessory in phosphorite interbands.

In amphibole ± biotite-rich mesobands, as well as sporadically in magnetite- and amphibole-rich mesobands, biotite occurs as xenoblastic porphyroblasts filling the interstices of idioblastic amphibole and xeno-idioblastic magnetite. This biotite shows helicitic microbanding and microgranular textures and contains numerous small zircon inclusions (Figs. 11 and 57). It was formed during the final stage of prograde regional metamorphism as a reaction product of chert or quartz and other probably clay minerals. It is only slightly altered into chlorite.

Except for these two most typical cases the porphyroblastic biotite occurs, together with pyrite, as individual bands in the chert-mesobanded grunerite rocks of the SFP (Fig. 13). Moreover, in Iso Vuorijärvi thin biotite-rich bands have been encountered that contain some garnet but no amphibole.

The biotite of the carbonate-amphibole rocks (Fig. 23) should presumably be included in this same phase. However, helicitic textures are not apparent, owing to the higher metamorphic grade of these rocks.

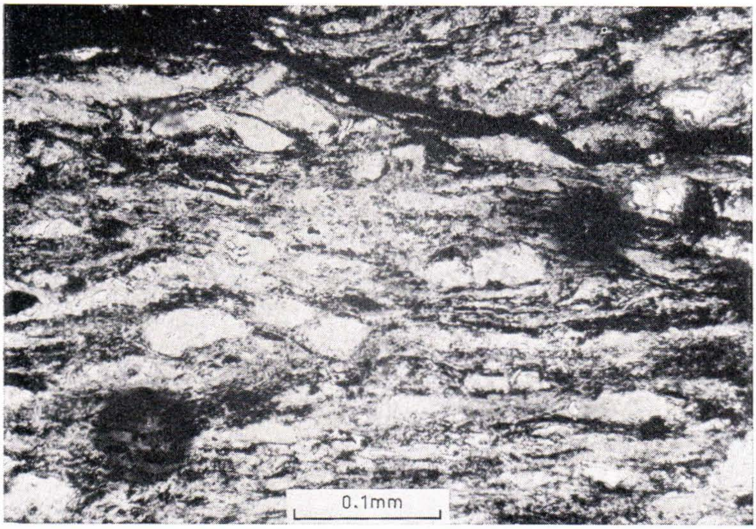

Fig. 57. A part of a biotite porphyroblast showing helicitic granular texture. White granules are relics of clastic quartz or intraclastic chert. Notice dark pleochroic haloes around zircon. Thin section No. 20357. One nicol.

The mode of occurrence of biotite in noncherty garnet-bearing iron-silicate mesobands differs drastically from that described above. It occurs as finely flaked matrix material between the garnet porphyroblasts and amphibole prisms (Figs. 29—31). This biotite contains minute inclusions of zircon. It is often intensely altered into chlorite, which causes the prevailing green colour of the bands. Because this biotite shows no evidence of being a reaction product, it probably derived through recrystallization from some early iron-rich mineral, probably stilpnomelane. This mineral is common in low-grade iron formations and is considered as primary (e.g. Ayres 1972, Floran and Papike 1975), a recrystallization product of a sedimentary precursor (Klein and Fink 1976) or as of late diagenetic origin (Dimroth and Chauvel 1973).

Some authors, however, regard stilpnomelane as a replacement product of volcanic ash or volcanoclastic material (Trendall and Blockley 1970, p. 290, Zajac 1974, p. 64). The supposition that the finely flaked biotite in Väyrylänkylä was a metamorphic recrystallization product of stilpnomelane of probable indirect volcanogenic origin is supported by the occurrence of metavolcanics in the Salmijärvi basin. Furthermore, some biotite-rich bands in the SF rocks in Seppola often contain K-feldspar as their second abundant mineral indicating that the bands are exceptionally rich in potassium (Fig. 63 B).

The iron-rich metapelites in both Seppola and the eastern formations contain biotite of normal metamorphic origin in their iron-poor bands (Fig. 28 A).

Excluding the biotite-rich bands described above, the MOSF, MSOF and CF rocks contain biotite only in accessory amounts, as tiny flakes intimately associated with magnetite porphyroblasts.

In the siderite rocks with replacement-vug textures, biotite commonly replaces the albite of the vugs or the network albite (Fig. 58). The phosphorite bands also contain abundant accessory biotite. It occurs either as tiny flakes, sometimes replaced by carbonate or quartz, or, as porphyroblasts showing helicitic plate texture (Fig. 59). In these two rock types the formation of biotite was clearly controlled by preexisting silica, and it thus belongs to the same group as the biotite in amphibole ± biotite-rich mesobands.

Finally, biotite occurs as a retrograde product of garnet (Fig. 62 A and B). Except when chloritized, biotite is sometimes replaced by carbonate.

Chlorite: Chlorite occurs in variable amounts as a retrograde product of biotite. Especially the finely flaked biotite of the noncherty garnet-bearing iron-silicate interbands as well as the accessory biotite of the phosphorite interbands tend to have altered into chlorite. In Seppola it typically occurs together with carbonate in the alternation zone around carbonitized amphibole crystals (Fig. 53). Sometimes, amphibole and garnet (Fig. 62 C) have also been replaced by chlorite.

The occurrence of chlorite in a Mn-carbonate-chert mesoband in Seppola was described earlier (p. 35).

Garnet

Garnet is an essential mineral constituent in noncherty garnet-bearing iron-silicate interbands and in spessartite-bearing iron-rich black schists and phyllites of Seppola. It also abounds in some iron-rich black schists of Körölä and Iso Vuorijärvi. In accessory amounts it is encountered in some magnetite- and amphibole-rich mesobands as well as in the siderite rock showing replacement-vug textures and even in some phosphorite interbands.

The garnet of the eastern formations is always almandine, whereas in Seppola it is spessartite (Table 2). The texture of both garnets, and spessartite in particular, is unusual in that two phases are present. The older phase is a skeletal garnet forming the core of the porphyroblasts (Fig. 60). This garnet has grown at an exceptionally high rate and typically shows a markedly radial texture of cubic symmetry (Fig. 60 A and B). It often happens that different garnet laths have pushed opaques ahead of them (Fig. 60 C, see also, Fig. 27). Also the laths parallel to the bedding have frequently grown much longer than the laths in other directions (Fig. 60 D). This skeletal garnet does not exhibit helicitic texture, but has deformed the banding in front of it. Neither does it contain helicitic inclusions. The phase seems to have grown explosively owing to extraordinarily favourable physico-chemical circumstances. The occurrence of this garnet is not restricted to iron-formation rocks, but has also been met with in the manganiferous phyllites of Seppola (Fig. 60 D).

The second phase is a »normal» garnet that forms a more or less complete rim around the former (Fig. 60 B and C). It crystallized under normal growth conditions

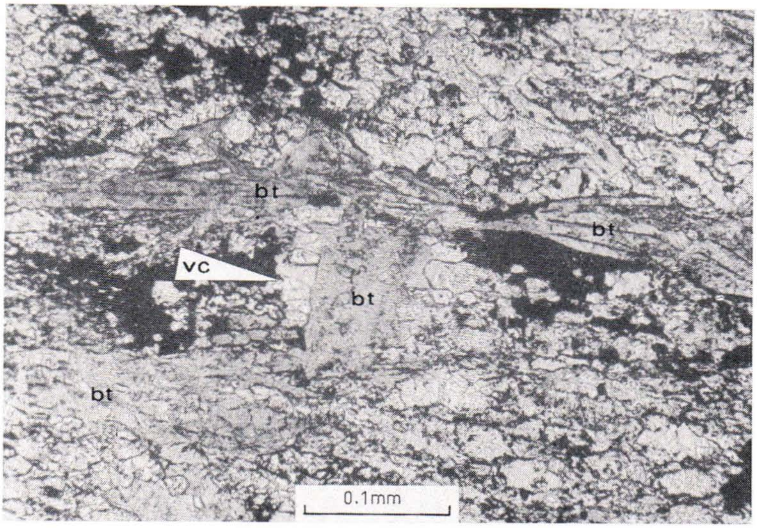

Fig. 58. Biotite (bt) formed at the expense of albite in siderite rock. In the middle columnar replacement-vug carbonate (vc) is still visible. Biotite in the middle shows two helicitic pigment textures. Thin section No. 21184. One nicol.

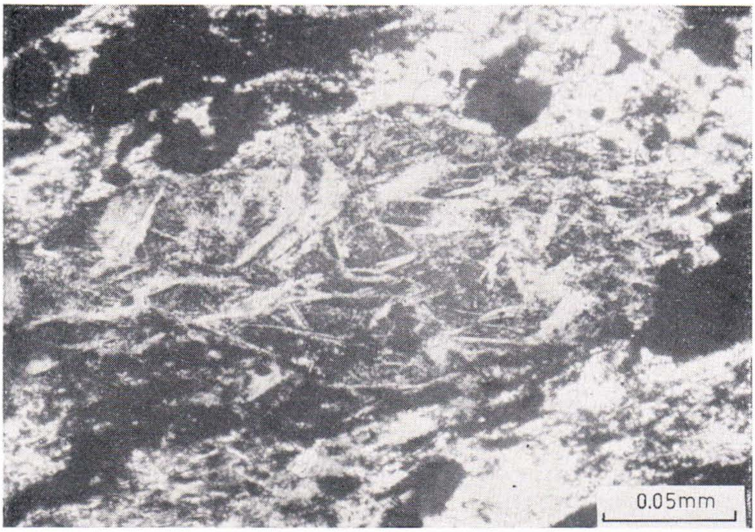

Fig. 59. Biotite porphyroblast showing helicitic quartz-plate texture of a phosphorite pellet from amphibole + biotite-rich mesoband. Thin section No. 21202. One nicol.

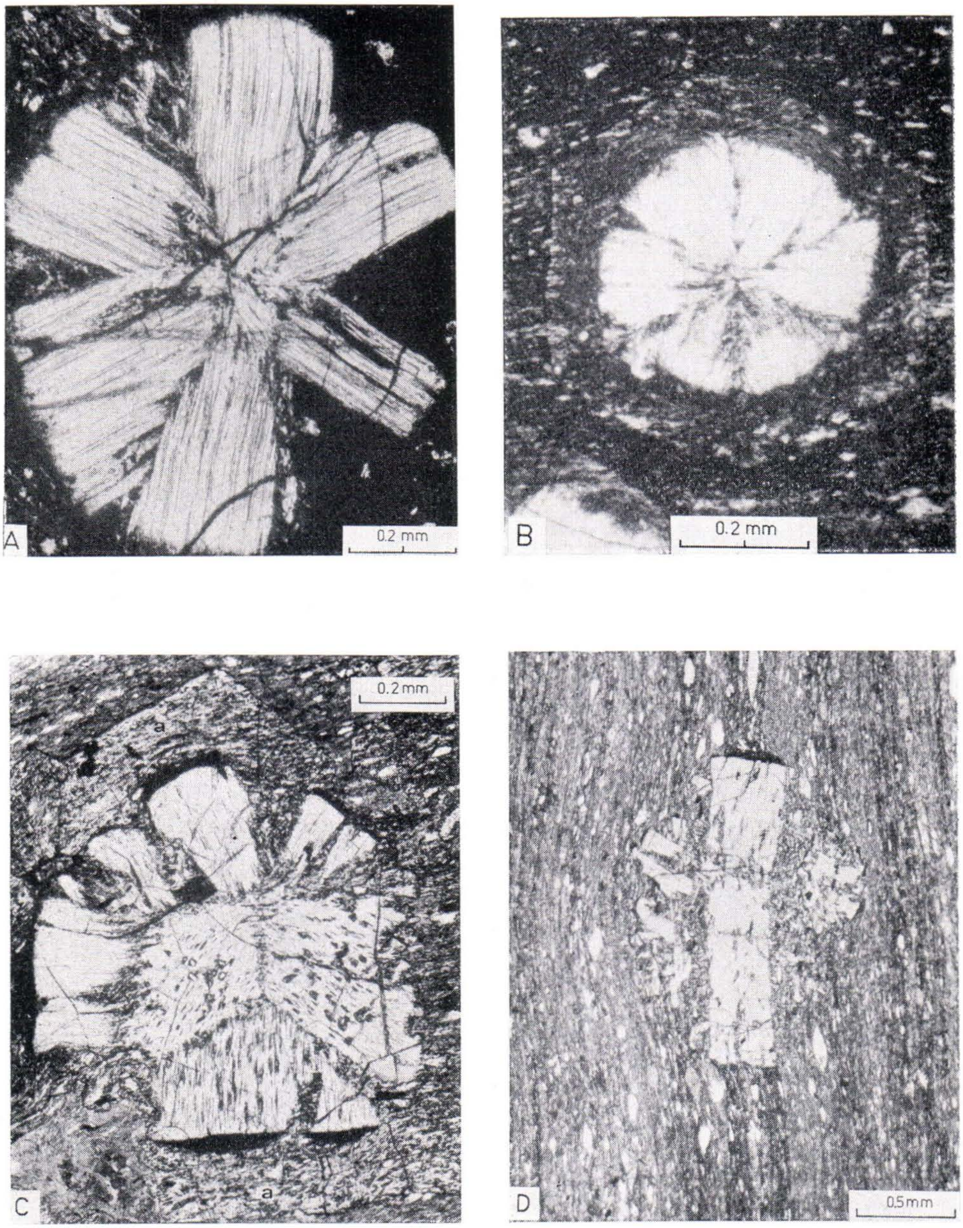

Fig. 60. Garnets from Seppola. A-C from iron-rich metapelites, D from manganiferous phyllite. All with one nicol. A) Skeletal spessartite without second-phase garnet. Thin section No. 17816. B) Spessartite showing well-developed second-phase garnet surrounding skeletal garnet. The same thin section as in A. C) Skeletal spessartite showing only partially developed second-stage garnet (a) and the opaque driven forward by the growth of the skeletal garnet. Thin section No. 17832. D) Skeletal garnet (spessartite?) showing conspicuous growth of the lath parallel to bedding. Notice the way the lath has driven forward the opaque of the band. Thin section No. 18006.

and could not deform the banding, but includes it as helicitic texture and an abundance of helicitic inclusions (Fig. 60 C).

This mode of occurrence of garnet in iron-formation rocks is a regional phenomenon in Kainuu, as is shown by some photographs of iron-formation rocks from Tuomivaara and Poskimäki published by Niiniskorpi (1975, Figs. 112, 113, and 123).

Both almanide and spessartite have often been replaced by retrograde minerals. The most common replacing mineral in Pääkkö is siderite, which has not uncommonly almost completely replaced both skeletal and second-phase almandine. There are two different types of replacement. In the garnet showing only poor skeletal texture, replacement begins along the cleavage planes from which it then spreads outwards. The replacement carbonate thus formed appears as irregular patches (Fig. 61 A). In the skeletal garnet the replacement process took place along the margins of different laths and gave rise to relict garnet strips (Fig. 61 B). This type of alternation is quite common in some siderite rocks and bands of noncherty garnet-bearing iron-silicate interbands, which nowadays are composed almost solely of carbonate but which before replacement were apparently rich in garnet. In Seppola, the latter type is widely distributed; the skeletel spessartite has often been completely replaced, so that only ghost-like relics of garnet are visible (Fig. 61 C).

Sometimes, garnet has been replaced by biotite or chlorite (Fig. 62). These processes are, compared with carbonatization, rather uncommon.

Both the occurrence of skeletal garnet and the ubiquitous evidence of replacement indicate that the PT conditions must have been in some way unusual during the formation of garnet. It seems probable that metamorphism in Väyrylänkylä reached just about the right conditions needed for the formation of garnet. This explains why garnet crystallised very rapidly on the one hand and was so easily attacked by other minerals during retrograde metamorphism on the other. Thus, garnet, together with amphibole, seems to represent the latest prograde mineral in the rocks studied.

Because the garnet porphyroblasts seem to have rotated only slightly if at all and are not deformed they apparently formed after or during the disappearance of the second deformation stage of the South Puolanka area (*cf.* amphibole).

Apatite

Apatite is such an essential mineral constituent of the iron formations proper in Väyrylänkylä that they could be described as apatite-bearing. According to X-ray identifications by Pekka Kallio the mineral is a Ca-OH-apatite ($a_0 = 9.42$ Å, $c_0 = 6.88$ Å, *cf.* McConnel 1973, Tables 5.2 and 5.3).

As a rule, apatite occurs as fine-grained crystals in phosphorite interbands and in close association with iron-minerals in chemical sedimentary band components, but it avoids chert. It also occurs in small pellets of graphitic substance. Its grain size is about 0.01 mm or less and it generally shows well-developed crystal faces.

Fig. 61. Replacement of garnet by carbonate in noncherty garnet-bearing iron-silicate interbands (A and B) and in a Mn-carbonate-rich phyllite (C). A) Almandine (grey) replaced by carbonate occurring as white or greyish patches. Thin section No. 19684. Partly crossed nicols. B) Relics of skeletal almandine in a retrograde carbonate and graphitic substance matrix. Only the bright parts of the laths are still garnet, their obscure parts being carbonate. Thin section No. 18003. One nicol. C) Spessartite replaced by carbonate. Bedding runs diagonally. Thin section No. 21294. Nicols nearly crossed. Photo A by E. Halme.

Such is the mode of occurrence of the apatite generations, say, in the siderite-rich parts of the grades in the quartz-siderite-banded rocks of the CF (Fig. 42 A). Because apatite shows the same evidence on the basis of which it was concluded that the siderite of these microstrata is primary (p. 64), apatite is considered as a recrystallization product of a sedimentary precursor.

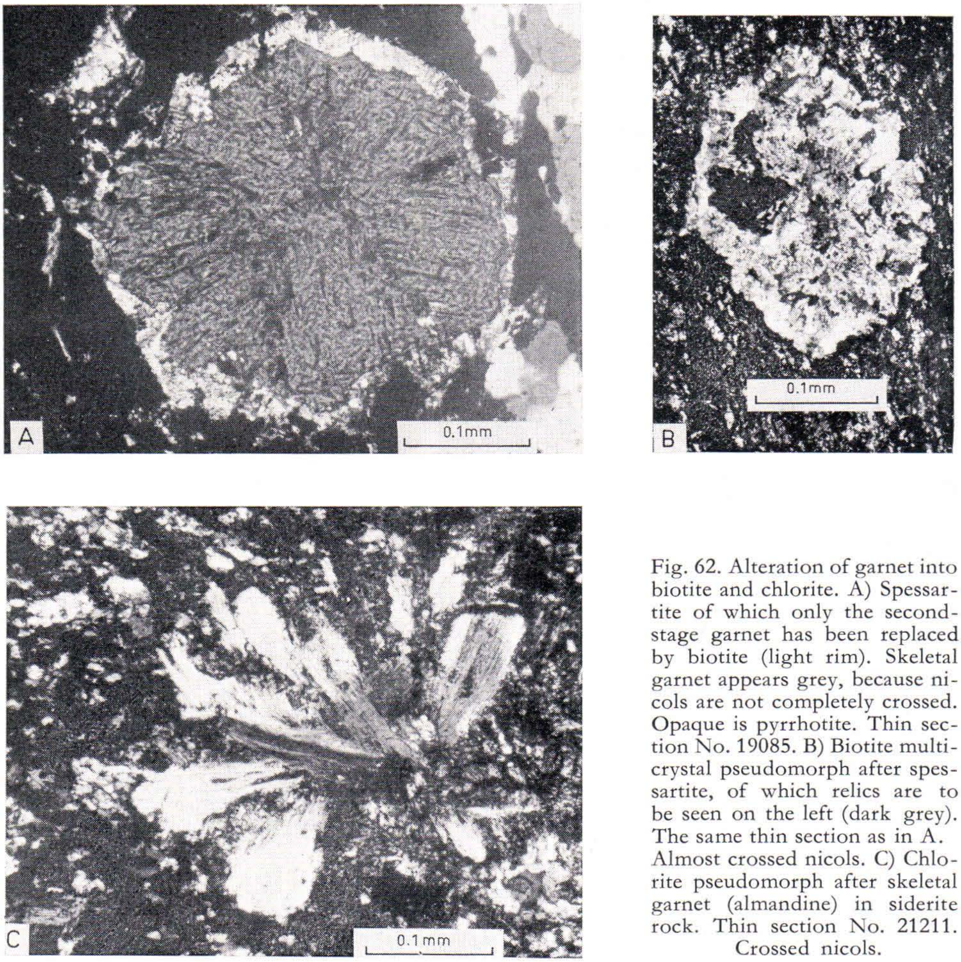

Fig. 62. Alteration of garnet into biotite and chlorite. A) Spessartite of which only the second-stage garnet has been replaced by biotite (light rim). Skeletal garnet appears grey, because nicols are not completely crossed. Opaque is pyrrhotite. Thin section No. 19085. B) Biotite multicrystal pseudomorph after spessartite, of which relics are to be seen on the left (dark grey). The same thin section as in A. Almost crossed nicols. C) Chlorite pseudomorph after skeletal garnet (almandine) in siderite rock. Thin section No. 21211. Crossed nicols.

The second typical apatite phase is the transparent post-tectonic replacement-vug apatite described in detail on p. 49, which occurs, however, in much smaller amounts than the presumably primary apatite. Relict vug textures in biotite porphyroblasts (Fig. 59) are suggestive of relatively early metamorphic age.

In some intensely metamorphosed phosphorite interbands, near the borders of minute biotite flakes there are some thin seams of transparent apatite that have grown out from the matrix perpendicularly towards the biotite flakes replaced by carbonate and quartz. This apatite was apparently formed in the same way as the vug-apatite but during retrogressive regional metamorphism.

Quite often, but in small quantities, the apatite of the iron-formation rocks has been remobilized and thus thin apatite-bearing or -rich veins are frequently to be seen in thin sections.

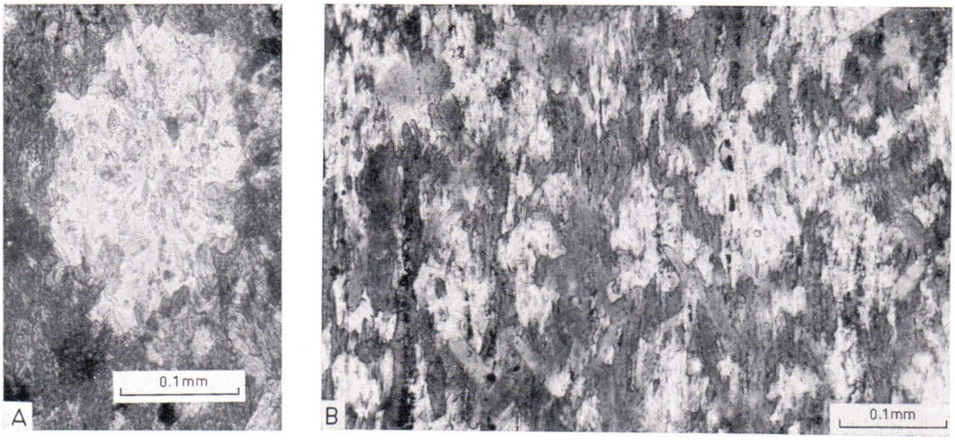

Fig. 63. Occurrences of feldspar. A) Oligoclase porphyroblast in biotite-amphibole-garnet interband containing biotite inclusions. Grey is biotite. Thin section No. 18054. One nicol. B) K-feldspar (white) in spessartite-bearing iron-rich phyllite from Seppola. Grey is biotite. Two tiny grains in the middle with opaque inclusions are tourmaline. Thin section No. 20358. One nicol.

Feldspars

Plagioclase: Occurrences of plagioclase are relatively few. In the iron-formation rocks of pure chemical origin plagioclase has been met with only as network albite in the siderite rock described on page 31.

Small amounts of clastic plagioclase occur in iron-rich metapelites, as well as in noncherty garnet-bearing iron-silicate interbands. Moreover, in the latter rocks one occurrence of porphyroblastic oligoclase has been encountered (Fig. 63 A).

K-feldspar: Except for some microscopic veins, K-feldspar is lacking from the rocks of the iron formations proper. In the iron-rich metapelites of Seppola, K-feldspar (microcline) is an essential mineral constituent of some, but not all of the biotite-rich bands. It then often occurs as small xenoblastic porphyroblasts containing minor biotite inclusions. Fig. 63 B shows a typical occurrence of this kind of feldspar in a K-feldspar- and biotite-rich interband of a spessartite-bearing iron-rich phyllite. The origin of this mineral is unknown. Probably it represents recrystallized volcanic feldspar ejectamenta like the K-feldspar of the black porcelanite of the Joffre Member in Hamersley (Trendall and De Laeter 1972). Davy (1975) believes, however, the K-feldspar of the high-potassic shale intercalations of the Wittenoom Dolomite, Hamersley, to have been formed from illite with potassium extracted from sea water and, possibly, tuffaceous material.

Iron sulphides

On the scale of macrobands, rocks of the iron-formations proper are richer in iron sulphides, especially in pyrrhotite, than are Precambrian iron formations in general. This is due to ubiquitous thin interbands of pyrrhotite-rich black schists or phyllites and faint dissemination of iron sulphides. Moreover, there are abundant thin pyrrhotite-rich veins.

In Seppola, pyrrhotite occurs throughout the formation as dissemination strictly controlled by bedding, indicating either its own or its precursor's primary origin. In contrast, pyrite has mostly been met with in Pääkkö, together with carbonate, amphibole and apatite, as veins representing remobilized iron-formation material. It is also the main, but comparatively rare, iron sulphide in the SFP rocks and in some noncherty garnet-bearing interbands.

Pyrite is generally considered as a primary iron sulphide in Precambrian iron formations and pyrrhotite as a metamorphic product of it. In Väyrylänkylä, however, it is clearly seen that pyrrhotite occurs mainly in the iron-rich metapelites of the SF, which were in fact found with the aid of distinct positive magnetic anomalies caused by pyrrhotite. Ordinary black schists contain pyrite as their main sulphide. Because both rocks have undergone the same metamorphism, it seems reasonable to assume that the difference in their iron sulphide content is due to primary factors, e.g. differences in their relation to probable synsedimentary volcanism (*cf.* Stanton 1972, p. 404).

This line of reasoning is indirectly supported by the occurrence of iron sulphides in the Porkonen-Pahtavaara volcanic complex, where the grade of metamorphism is low, somewhere between that of the greenschist and zeolite facies (Paakkola 1971, p. 72). In that locality the dominant iron sulphide of sulphide schists, apparently representing sulphide facies of the iron formations, is pyrrhotite (*op. cit.* p. 50).

Graphite and graphitic substance

In this study the term graphitic substance is, applied to the opaque black carbonaceous matter that is ubiquitous in all types of iron-formation rocks in Väyrylänkylä. This is because the material cannot usually be identified as graphite under reflected light, which indicates that it has not completely crystallized. Mr. Pekka Kallio kindly performed X-ray diffraction studies on this material from the following five rock types: quartz-magnetite-banded rock (6/A558), magnetite-amphibole mesoband (4/A558), spessartite-bearing iron-rich black schist (A559), siderite mesoband (la/A558) and phosphorite interband (lb/A558) (sample numbers refer to Table 2 in Sakko and Laajoki 1975). The material turned out to be well crystallized graphite in the three first-mentioned rocks, whereas in the latter two rocks it was poorly crystallized.

In addition to iron-rich metapelites and their interbands, graphitic matter is particularly concentrated in some phosphorite interbands and some pellets of apatite-bearing graphitic substance (Figs. 39 and 40) and in some spessartite-biotite bands (Fig. 26 A, II).

Accessory minerals

Tourmaline: Tourmaline has not been detected in pure chemical-sedimentary bands. It is, however, a typical accessory mineral in biotite-rich rocks. Hence it is regularly met with in iron-rich black schists and in noncherty garnet-bearing iron-silicate interbands. In the latter, small tourmaline porphyroblasts are sometimes present. Moreover, tourmaline has been detected in some pyrite-rich veins. As a rule, tourmaline is green-coloured in thin section.

Zircon: This mineral has not been seen in pure chemical-sedimentary bands. In the biotite of garnet- and biotite-bearing interbands there are invariably small zircon inclusions, recognized by their pleochroic haloes (Fig. 57). In addition to this type of zircon, well-rounded grains of clastic zircon are also met with in iron-rich metapelites.

Hematite: The small amounts of martite frequently seen in the margins of magtite crystals are due at least in part to supergene oxidation processes.

Limonite: Limonite occurs as a supergene oxidation product of silicates in weathered iron-formation rocks.

Chalcopyrite: Chalcopyrite is a typical accessory mineral in iron-rich blacks schists. In other types of rocks it is met with only occasionally.

Arsenopyrite: Some sporadic arsenopyrite grains are often found in all types of iron-rich rocks.

Concluding remarks

The following tentative conclusions can be drawn on the basis of the foregoing description:

1) The primary mineral assemblage of the pure CF rocks, excluding foreign meso- and interbands, seems to have been siderite-chert ± apatite with some iron oxide (hematite or magnetite).

2) Most of the apatite of the phosphorite interbands is assumed to be primary.

3) Pyrrhotite is either primary or was formed from primary iron sulphide (pyrite) by metamorphism.

4) Graphitic substance represents organic matter that is only partly recrystallized as graphite.

5) All the grunerite-cummingtonite originated by the reaction of chert or quartz or, to a lesser extent, of albite with primary or low-grade metamorphic iron-minerals (siderite, iron oxides or iron silicates). Thus the pure MOSF (and MSOF) rocks as well as SFP may have been originally iron-silicate-rich or have formed from the CF rocks through metamorphic processes.

6) Progressive regional metamorphism provided ideal conditions for the formation of garnet, and consequently during metamorphism the maximum temperature rose to or slightly exceeded 350—400°C (see discussion in French 1973, p. 1066).

7) Typical metamorphic mineral assemblage in the pure CF rock is siderite-magnetite-grunerite-apatite ± garnet + excess chert or quartz left over after the formation of grunerite.

8) Typical metamorphic mineral assemblage of the pure MOSF (and MSOF) rock is magnetite-grunerite-chert-apatite ± carbonate (ferroan ankerite) ± garnet ± biotite. Except for apatite, this corresponds, as does that in 7, to the mineral assemblage of the lower part of the garnet zone in northern Michigan (James 1955, Table 2).

9) The finely flaked biotite of the noncherty garnet-bearing iron-silicate interbands and of the biotite-rich mesobands have a metamorphic history different from the porphyroblastic biotite of the amphibole ± biotite-rich mesobands and the phosphorite interbands.

10) The temperature peak (~ 400°C) of regional metamorphism was probably reached after the shear foliation of the second deformation stage of the South Puolanka area (*cf.* James 1955, p. 1483).

11) Albite and columnar siderite of the siderite rocks as well as quartz and columnar apatite of some phosphorite interbands seem to have been formed after the second deformational stage, but before the formation of amphibole and biotite porphyroblasts. Also the garnet porphyroblasts are post-tectonic.

12) The abundant formation of retrograde carbonate after garnet is a consequence of the fact that the garnet isograd was only just exceeded during the progressive regional metamorphism.

GEOCHEMISTRY

As a consequence of the complicated structure of the South Puolanka area (Laajoki 1973), it was a difficult task to correlate the different iron-formation lenses. Furthermore, it was not even possible to establish the relationship between different macrobands from adjoining drill holes (Ervamaa and Laajoki 1977, Figs. 14 and 15). For the most part, these difficulties are due to the intense small-scale folding of the second deformation stage, many marks of which have been detected on both outcrops (Fig. 6) and the drill cores (Fig. 13). This feature is well exemplified in the distribution of the CF rocks in Pääkkö, where this facies was intersected by drill hole No. 354, but not by No. 346 drilled below it. Hence, we have not been able to divide the iron formations in question into different macrobands within the limits of a single formation, but we have had to treat them as single geochemical entities. Even so, they originally consisted of at least two or three separate major iron-formation beds as is indicated by the clastic interbands frequently encountered in both the Pääkkö and Iso Vuorijärvi formations.

The absence of outcrops in the surroundings of the Iso Vuorijärvi iron formation raises the theoretical possibility that this and the Pääkkö formation represent different iron-formation horizons. However, not only lithologic evidence, but also their comparable geochemistry, especially their analogous P and S contents, as revealed in this study, and the similarities in their rare-earth elements (Laajoki 1975 b) strongly suggest that the units are contemporaneous.

The sulphide iron formations of Seppola are known only on the basis of three drill holes. Also in this area the folding in the iron-formation and associated rocks has been intense, and our knowledge of the bedrock geology between Seppola and Pääkkö is based only on geophysical data and on very few outcrops (Laajoki 1973). Thus, it is impossible to draw definitive correlations between the eastern iron-formations (Pääkkö and Iso Vuorijärvi) and the Seppola formation. Even so, the stratigraphic position of these iron-rich metapelites largely corresponds to that of the Pääkkö and Iso Vuorijärvi iron formations.

Sampling and results of chemical analyses

Table 3 gives sample codes, locations, rock-stratigraphic units and lithologic-petrologic remarks for the silicate analyses (Tables 4—7). All the analyses were made on drill-core material. The following principles were followed during sampling. First, in order to get statistically representative analyses, pieces of diamond drill core, roughly 1 m long, of fresh MOSF (Table 4, Nos. 1—3) and CF rocks (Table 5, Nos. 1—6) from the Pääkkö and Iso Vuorijärvi iron formations were chosen. From the MSOF only one short intersection of relatively fresh rock was detected (Table 4, No. 4), and so this facies was disregarded in later treatment of the silicate and ore analyses. In order to study the effects of the Preglacial weathering an altered iron-silicate-magnetite rock was analysed (Table 4, No. 5). From the SF of the Iso Vuorijärvi formation drill cores, about 15 cm long, of both the pyrrhotite-rich and -poor parts of a single iron-rich black-schist macroband were collected (Table 6, Nos. 1 and 2). Samples typical of different kind of spessartite-bearing iron-rich rocks were taken from the Seppola iron formation (Table 6, Nos. 3—5, Fig. 26). Also analysed was one carbonate-rich macroband (Table 5, No. 7, Fig. 14 D). The core lengths of the Seppola samples vary from 10 to 20 cm. For the study of geochemical banding on a smaller scale, samples from different kinds of mesobands and interbands from the Pääkkö iron formation were analysed (Table 7, Nos. 1—12). In addition, partial analyses were made on short, homogenous core pieces of siderite rock, which can be considered representative of siderite-rich mesobands, and laminated quartz-siderite rock from Pääkkö (Table 5, Nos. 8 and 9, Figs. 14 B and C) as well as on two contrasting band types (Table 6, Nos. 6 and 7, Fig. 26 A) of a spessartite-bearing iron-rich black schist from Seppola.

The average partial chemical compositions based on systematic ore analyses for each facies of the Pääkkö and Iso Vuorijärvi formations and for their summed formation are set out in Table 8. When calculating these averages only analyses with an iron content of at least 15 % were accepted. Hybrid rocks, such as breccias, veins and contact rocks (e.g. carbonate-amphibole rocks), were rejected. Statistically, these results give the most representative gross information about the distribution of each element in question. However, as to the purity of different lithologic types, the samples selected for silicate analyses are more representative. The average iron contents in Table 8 were used as the basis for calculations of the total chemical composition of each facies and the summed formation of Pääkkö and Iso Vuorijärvi (Table 9).

The SFP rocks in drill hole No. 359 were not detected until the closing stage of the present study when findings were being checked. Owing to the scarcity of magnetite they were overlooked when the core material was sampled for ore analyses by prospectors, and so they are not included in the average contents of the Pääkkö iron formation. Table 4, however, gives one total (No. 10) and three partial (Nos. 8, 9, and 11) silicate analyses of the SFP rocks as well as two partial analyses of the MOSF rock associated with them (Nos. 6 and 7). It should be noted that No. 11 does not contain iron enough to be included in iron formation and, consequently, it is omitted from the SFP in the following discussions.

During sampling of drill cores for ore analyses special attention was directed to the well-developed macrobanding in Iso Vuorijärvi (Table 10) and Körölä iron formation (Table 11). Before that, however, part (depth interval 108.70—125.70 m) of

Tab

Diagnostic data for the rocks of the Väyrylänkylä iron formations, whose ana
»remarks» refer to Laajoki 1975 b, Tab

Anal. No.	Rock name	Location drill hole/depth (m)
Table 4/ 1	Quartz-magnetite-banded rock	343/ 82.00— 83.00
» / 2	» » » »	345/114.70—115.70
» / 3	» » » »	356/ 76.00— 77.00
» / 4	Fe-silicate-magnetite rock	363/ 91.95
» / 5	» » »	362/104.30—105.30
» / 6	Grunerite-magnetite rock	359/ 60.60— 60.70
» / 7	Magnetite-grunerite rock	359/ 60.75— 60.80
» / 8	Chert-mesobanded grunerite rock	359/ 57.30— 59.00
» / 9	» » » »	359/ 59.00— 60.55
» /10	» » » »	359/ 61.10— 62.65
» /11	» » » »	359/ 62.65— 63.65
Table 5/ 1	Quartz-siderite-banded rock	360/ 39.50— 40.50
» / 2	Phosphorite-banded siderite rock	360/ 51.50— 52.50
» / 3	Quartz-siderite banded rock	354/ 53.00— 54.00
» / 4	» » » »	360/ 25.00— 27.30
» / 5	» » » »	345/136.00—137.00
» / 6	» » » »	345/130.20—131.00
» / 7	Quartz-Mn-siderite-banded rock	337/ 77.75— 77.85
» / 8	Siderite rock	360/ 51.65— 51.70
» / 9	Laminated quartz-siderite rock	354/ 64.30— 64.35
Table 6/ 1	Pyrrhotite-rich black schist	345/129.60—129.72
» / 2	Black schist	345/129.74—129.90
» / 3	Spessartite-bearing iron-rich black schist	337/ 64.20— 64.40
» / 4	» » » » » »	337/ 80.35— 80.45
» / 5	Spessartite-bearing iron-rich phyllite	337/ 76.50— 76.65
» / 6	Carbonate-pyrrhotite-spessartite band in No. 3	337/ 64.25
» / 7	Graphite-rich spessartite-biotite band in No. 3	337/ 84.35
Table 7/ 1	Chert mesoband	360/ 70.10— 70.13
» / 2	Magnetite-amphibole-chert mesoband	343/ 83.72— 83.76
» / 3	Amphibole-magnetite-chert mesoband	344/155.45—155.49
» / 4	Amphibole-chert-biotite mesoband	343/ 81.20
» / 5	Amphibole-biotite-magnetite-chert mesoband	343/ 81.25
» / 6	Amphibole-biotite-garnet interband	343/ 63.50— 63.60
» / 7	Garnet-amphibole-siderite-biotite interband	354/ 97.00— 97.10
» / 8	Biotite-amphibole-garnet interband	359/ 57.35— 57.45
» / 9	» » » »	343/ 90.75— 90.80
» /10	Phosphorite interband	360/ 52.50
» /11	» »	360/ 51.40
» /12	» »	360/ 51.50

drill core No. 345 and drill core No. 343 of the Pääkkö formation (Table 12) were sampled by other persons. The iron-rich metasediments in the Seppola area are, however, of such an uneconomic nature that only check ore analyses were performed on them (Table 13).

The results of trace-element analyses are set out in Table 14 and of a rare-earth element analysis in Table 15.

le 3

lytical data are set out in Tables 4—7. Numbers in the brackets in the column
le 1. I.V. = Iso Vuorijärvi iron formation.

Iron formation	Facies	Remarks
Pääkkö	MOSF	Typical sample (No. 1)
I.V.	»	» » (No. 3)
Pääkkö	»	Unusually rich in amphibole (No. 5)
I.V.	MSOF	Slightly weathered, noncherty
»	»	Intensely weathered (No. 7)
Pääkkö	MOSF	Rich in grunerite
»	»	Unusually rich in magnetite
»	SFP	Typical rock
»	»	Typical, thin dark biotite-rich interbands
»	»	» » » » » »
»	»	Iron poor
Pääkkö	CF	Typical sample (No. 9)
»	»	Unusually rich in amphibole and poor in chert
»	»	
»	»	
I.V.	»	Typical sample (No. 10)
»	»	
Seppola	»	Macroband in sulphide facies. Spessartite-bearing. Fig. 14, D.
Pääkkö	»	Typical sample. Fig. 14, C.
»	»	Typical sample. Fig. 14, B.
I.V.	SF	Iron-sulphide-rich part of iron-rich black-schist macroband
»	»	Iron-sulphide-poor part of iron-rich black-schist macroband
Seppola	»	Typical sample, thin chert bands, pyrrhotite $\sim$ 10 vol.%. Fig. 26, A.
»	»	Massive, spessartite $\sim$ 30 vol. %. Fig. 26, C.
»	»	Typical sample, grey in colour, graded bedding, some thin chert bands. Fig. 26, B.
»	»	Fig. 26, A, I
»	»	Fig. 26, A, II
Pääkkö	MOSF	
»	»	
»	»	
»	»	Fig. 7, B, I
»	»	Fig. 7, B, II
»	»	Typical sample
»	»	Rich in carbonate
»	SFP	Typical sample (No. 17)
»	MOSF	Typical sample
»	CF	Typical sample, dark, thickness 0.3—0.5 cm
»	»	Typical sample, black, thickness 0.5—1 cm (No. 8)
»	»	Typical sample, dark, thickness 0.3—0.5 cm

Distribution of elements

In this chapter the distribution of each element within the Väyrylänkylä formations is discussed. Further, the contents of each element are compared with those in other iron-rich metasediments in the world. In this respect, the classic papers by Lepp and Goldich (1964) and James (1966), which inspired Laajoki to start the present study, were of special value. However, as pointed out by Trendall (1965 a) the paper by Lepp and Goldich calls for some critical comments.

Table 4

Chemical composition (Wt-%) of rocks of the mixed oxide-silicate-facies (MOSF), mixed silicate-oxide facies (MOSF) and silicate facies proper (SFP) from iron formations in Väyrylänkylä. Analyses Nos. 2—4 and 10 by Risto Saikkonen, Nos. 1 and 5 by Pentti Ojanperä.

	1	2	3	4	5	A	6 b)	7	8 b)	9	10	11 b)
SiO_2	42.54	37.28	42.52	32.98	33.45	40.78					47.27	
TiO_2	0.10 a)	0.03	0.07	0.06	0.00	0.07	0.23	0.07 b)	0.28	0.08 b)	0.28 b)	0.47
Al_2O_3	0.85 a)	0.56	0.71	0.45	0.19 a)	0.71	1.59 c)	1.18 c)	3.44 c)	0.87 c)	3.27	5.41 c)
Fe_2O_3	17.91	10.46	19.90	26.25	34.98	16.09	21.7	35.2 b)	10.3	5.0 b)	5.1 b)	2.1
FeO	24.70	27.63	21.99	28.87	14.32	24.77	29.8	27.1 b)	23.0	23.9 b)	25.1 b)	14.0
MnO	0.08	0.04	0.13	0.06	0.01 b)	0.08	0.28	0.49 b)	0.21	0.21	0.21 b)	0.04
MgO	3.53	4.78	3.02	5.83	0.83	3.78					4.93	
CaO	3.75	4.96	3.64	0.62	5.65	4.12		3.19 c)		2.60 c)	1.80	
Na_2O	0.16	0.03	0.10	0.05	0.05	0.10		0.05 c)		0.04 c)	0.08	
K_2O	0.18	0.08	0.11	0.02	0.03	0.12		0.40 c)		0.14 c)	0.92	
P_2O_5	2.73	3.11	2.63	0.44	4.41	2.82	3.14	2.89 b)	1.51	1.56 b)	1.24 b)	0.41
CO_2	0.04	7.20	0.64	0.64	0.00	2.63					3.79	
H_2O+	2.32	1.50	1.75	1.90	2.17	1.86					2.55	
H_2O-	0.02	0.04	0.11	0.10	1.61	0.06					0.29	
C	0.96	0.64	1.37	0.08	0.80	0.99					0.64	
S	0.78 b)	1.46 b)	0.83 b)	0.66	0.39 b)	1.02	2.64	3.19 b)	2.72	1.67 b)	2.73 b)	2.87
	100.65	99.80	99.52	99.01	98.89	100.00					100.20	
O = S	0.34	0.64	0.36	0.29	0.17	0.45					1.02	
Total	100.31	99.16	99.16	98.72	98.72	99.55					99.18	
Fe^{3+}	12.53	7.32	13.92	18.36	24.47	11.25	15.2	24.6	7.2	3.5	3.6	1.5
Fe^{2+}	19.20	21.48	17.09	22.44	11.13	19.26	23.2	21.1	17.9	18.6	19.5	10.9
Fe_{tot}	31.73	28.80	31.01	40.80	35.60	30.51	38.4	45.7	25.1	22.1	23.1	12.4
Fe^{3+}/Fe^{2+}	0.65	0.34	0.81	0.82	2.20	0.58	0.65	1.17	0.40	0.19	0.18	0.14
P_2O_5/CaO	0.73	0.63	0.72	0.71	0.78	0.68		0.91		0.60	0.69	

a) Analysed by Ari Puisto (OES), b) Analysed in Ore Laboratory, c) Analysed by Risto Saikkonen.

1) Quartz-magnetite-banded rock. Pääkkö iron formation (Rb 245/73).
2) Quartz-magnetite-banded rock. Iso Vuorijärvi iron formation (Rb 106/74).
3) Quartz-magnetite-banded rock (rich in amphibole). Pääkkö iron formation (Rb 58/74).
4) Fe-silicate(grunerite)-magnetite rock (slightly weathered) Iso Vuorijärvi iron formation (Rb 21/75).
5) Fe-silicate-magnetite rock (intensely weathered). Iso Vuorijärvi iron formation (Rb 246/73).
A) Average of analyses Nos. 1—3.
6) Grunerite-magnetite rock (Ra 2685/76).
7) Magnetite-grunerite rock (Ra 322/74).
8) Chert-mesobanded grunerite rock (Ra 2528/76).
9) Chert-mesobanded grunerite rock (Ra 2529/76).
10) Chert-mesobanded grunerite rock (Ra 2530/76).
11) Chert-mesobanded grunerite rock (iron-poor) (Ra 2531/76).

Table 5

Chemical composition (Wt-%) of carbonate-facies (CF) rocks from iron formations in Väyrylänkylä. Analyses by Risto Saikkonen (Nos. 1—7) and Meeri Taug (Nos. 8 and 9).

	1	2	3	4	5	6	7	A	B	8	9
SiO_2	30.87	19.70			37.83		41.10	29.47			
TiO_2	0.03	0.01			0.05		0.00	0.03		0.07	0.05
Al_2O_3 ...	0.48	0.22			0.85		0.12	0.52		0.60$^{c)}$	0.59$^{b)}$
Fe_2O_3 ...	5.79	7.92	4.87	6.67	1.29	3.49	1.50	4.98	5.00	8.22	5.12
FeO	29.80	34.96	26.38	24.25	28.32	27.88	10.99	31.03	28.60	35.7	27.61
MnO	0.03	0.02	0.07	0.09	0.06	0.14	12.81	0.04	0.07	0.01	0.06
MgO	3.15	3.92	2.77	2.39	3.95	3.98	3.59	3.67	3.36		
CaO	4.82	5.96	2.69	2.88	3.39	1.82	5.38	4.72	3.59		
Na_2O ...	0.02	0.01			0.09		0.00	0.04			
K_2O	0.06	0.03			0.18		0.02	0.09			
P_2O_5	3.38	3.53	1.92$^{a)}$	2.20$^{a)}$	2.04	1.33$^{a)}$	0.32	2.98	2.40	0.11	0.16
CO_2	20.06	22.45	17.67	15.65	20.92	18.48	23.38	21.14	19.27	22.60$^{c)}$	19.87$^{c)}$
H_2O+ ..	1.10	1.02			0.93		0.57	1.02			
H_2O- ..	0.00	0.00			0.04		0.01	0.01			
C	0.76	0.65	1.06	0.63	0.26	0.47	0.75	0.56	0.64	0.31$^{b)}$	1.31$^{c)}$
S	1.03	0.13	1.04$^{a)}$	0.30$^{a)}$	0.42$^{a)}$	1.95$^{a)}$	0.31	0.53	0.81	0.20	0.38
	101.38	100.53			100.62		100.85	100.83			
O = S ..	0.45	0.06			0.18		0.14	0.23			
Total	100.93	100.47			100.44		100.71	100.60			
Fe^{3+}	4.05	5.54	3.41	4.67	0.86	2.44	1.05	3.48	3.50	5.75	3.58
Fe^{2+}	23.17	27.18	20.51	18.85	22.01	21.67	8.54	24.12	22.23	27.75	21.46
Fe_{total}....	27.22	32.72	23.92	23.52	22.87	24.11	9.59	27.60	25.73	33.50	25.04
Fe^{3+}/Fe^{2+}	0.17	0.20	0.17	0.25	0.04	0.11	0.12	0.14	0.16	0.21	0.17
P_2O_5/CaO	0.70	0.59	0.71	0.76	0.60	0.73	0.06	0.63	0.67		

a) Analysed in Ore Laboratory, b) Analysed by Ringa Danielsson (OES), c) Analysed by Risto Saikkonen.

1) Quartz-siderite-banded rock. Pääkkö iron formation (Rb 54/74).
2) Phosphorite-banded siderite rock. Pääkkö iron formation (Rb 107/74).
3) Quartz-siderite-banded rock. Pääkkö iron formation (Rb 61/74).
4) Quartz-siderite-banded rock. Pääkkö iron formation (Rb 63/74).
5) Quartz-siderite-banded rock. Iso Vuorijärvi iron formation (Rb 64/74).
6) Quartz-siderite-banded rock. Iso Vuorijärvi iron formation (Rb 62/74).
7) Quartz-Mn-siderite-banded rock. Seppola iron formation (Rb 101/74).
A) Average of analyses Nos. 1, 2 and 5.
B) Average of analyses Nos. 1—6.
8) Siderite rock. Pääkkö iron formation (Ra 601/76).
9) Laminated quartz-siderite rock. Pääkkö iron formation (Ra 600/76).

Table 6

Chemical composition (Wt-%) of sulphide-facies (SF) rocks from iron formations in Väyrylänkylä. Analyses by Risto Saikkonen (Nos. 1—5) and Meeri Taug (Nos. 6—7).

	1	2	3	4	5	A	B	C	D	6	7
SiO_2	27.82	46.98	41.63	45.49	39.89	38.77	42.34	49.47	53.05		
TiO_2	0.36	0.65	0.28	0.58	0.36	0.53	0.41	0.64	0.73	0.03	0.33
Al_2O_3 ...	5.14	8.42	3.48	7.41	4.83	7.01	5.24	9.14	9.51		
Fe_2O_3 ...	1.60	1.65	1.77	1.98	1.15	1.63	1.63	2.84	1.86	1.72	3.83
FeO	40.78	15.69	19.99	20.20	21.45	26.44	20.55	14.06	8.28	24.7	13.2
MnO	0.02	0.01	6.46	4.80	9.37	0.01	6.88	0.04	0.01	5.80	10.24
MgO	1.94	2.88	4.12	2.79	5.01	2.48	3.97	3.45	3.25		
CaO	0.88	1.77	4.62	3.22	4.76	1.39	4.20	1.56	2.00		
Na_2O ...	0.12	0.47	0.01	0.10	0.02	0.32	0.04	0.21	0.53	0.04 a)	0.06 a)
K_2O	1.48	2.86	0.42	2.00	1.34	2.27	1.25	2.63	3.23	0.15 a)	0.46 a)
P_2O_5	0.13	0.83	0.65	0.25	0.70	0.53	0.53	0.23	0.94	0.16	1.35
CO_2	2.22	0.67	7.48	4.12	6.80	1.33	6.13	3.95	0.76	13.87 a)	0.34 a)
H_2O+ ..	2.02	3.12	1.68	1.68	1.44	2.65	1.60	3.59	3.52		
H_2O- ..	0.06	0.08	0.04	0.04	0.06	0.07	0.05	0.11	0.09		
C	4.54	10.83	4.03	2.13	0.85	8.13	2.34	8.07	12.23	4.43 a)	5.12 a)
S	18.87	4.80	7.50	5.97	4.39	10.83	5.95	—	—	10.8	3.54
	107.98	10.171	104.16	102.76	102.42	104.39	103.11	99.99	99.99		
O = S ..	8.25	2.10	3.28	2.61	1.92	4.73	2.60	—	—		
Total	99.73	99.61	100.88	100.15	100.50	99.66	100.51	99.99	99.99		
Fe^{3+}	1.12	1.15	1.24	1.38	0.80	1.14	1.14	1.99	1.30	1.20	0.58
Fe^{2+}.	31.70	12.20	15.54	15.70	16.67	20.55	15.97	10.93	6.44	19.20	10.26
Fe_{tot}	32.82	13.35	16.78	17.08	17.47	21.69	17.11	12.92	7.74	20.40	10.84
Fe^{3+}/Fe^{2+}	0.04	0.09	0.08	0.09	0.05	0.06	0.07	0.18	0.20	0.06	0.06
P_2O_5/CaO	0.15	0.47	0.14	0.08	0.15	0.38	0.13	0.15	0.47		
Na_2O/K_2O	0.08	0.16	0.02	0.05	0.01	0.14	0.03	0.08	0.16	0.27	0.13

a) Analysed by Risto Saikkonen.

1) Pyrrhotite-rich black schist. Iso Vuorijärvi iron formation (Rb 73/74).
2) Black schist. Iso Vuorijärvi iron formation (Rb 97/74).
3) Spssartite-bearing iron-rich black schist. Seppola iron formation (Rb 98/74).
4) Spessartite-bearing iron-rich black schist. Seppola iron formation (Rb 99/74).
5) Spessartite-bearing iron-rich phyllite. Seppola iron formation (Rb 100/74).
A) Average of analyses Nos. 1 and 2 weighted for core lengths.
B) Average of analyses Nos. 3—5.
C) Analysis No. 1 recalculated to FeS-free basis.
D) Analysis No. 2 recalculated to FeS-free basis.
6) Carbonate-pyrrhotite-spessartite band in No. 3 (Ra 607/76).
7) Graphite-rich spessartite-biotite band in No. 3 (Ra 608/76).

Table 7

Chemical composition (Wt-%) of meso- and interbands in the Pääkkö iron formation. Analyses Nos. 1—3, 6, and 10—12 by Risto Saikkonen.

	1	2	3	4	5	6	7	8	9	10	11	12
SiO_2	91.61	35.14	44.17			40.83				11.48	13.03	
TiO_2	0.00	0.05	0.02	0.13a)	0.45a)	0.58	0.77a)	0.90a)	1.19a)	0.00		
Al_2O_3 ...	0.09	0.62	0.48	2.36c)	6.42c)	5.75	7.56b)	9.34e)	10.58b)	0.40	0.91	
Fe_2O_3 ...	1.61	24.66	19.31	7.01a)	11.25a)	4.06			9.0 a)	0.13		
FeO	3.64	26.94	29.04	19.64a)	25.23a)	32.23	38.3 x)	28.0 x)	23.4 a)	11.75		
MnO	0.01	0.09	0.02	0.03a)	0.02a)	0.04	0.09a)	0.05a)	0.06a)	0.10		
MgO	0.45	3.69	3.87	3.61d)	3.77d)	5.51			6.75d)	1.23		
CaO	0.05	3.44	0.11	0.19d)	0.63	2.06		0.91e)	0.99d)	37.0		
Na_2O ...	0.00	0.07	0.03	0.45d)	0.86d)	0.92		0.17e)	0.44d)	0.04		
K_2O	0.02	0.04	0.05	0.29d)	1.43d)	0.60		4.34e)	3.38d)	0.09		
P_2O_5	0.04	2.16	0.02	0.21a)	0.55a)	0.18	0.41a)	0.37a)	0.18a)	27.0	29.4	15.6
CO_2	1.59	0.66	0.10	3.15e)	0.04e)	1.54	5.74e)		0.80e)	7.5		
H_2O+ ..	0.40	1.20	1.39			1.80				1.04		
H_2O- ..	0.00	0.04	0.11			0.01				0.08		
C	0.23	1.80	0.52	0.55e)	2.37e)	4.45	1.69e)		0.44e)	0.17	4.62	
S	0.23	0.20	0.23	0.69a)	1.66a)	1.85	0.12a)	0.20a)	2.25a)	0.08		
F										0.11	0.17	0.12
	99.97	100.80	99.47	(38.31)	(54.20)	102.41			(58.58)	98.20		
O = S ..	0.10	0.09	0.10			0.81				0.03		
O = F ..										0.04		
Total	99.87	100.71	99.37			101.60				98.13		
Fe^{3+}	1.13	17.25	13.51	4.90	7.87	2.84			6.3	0.09		
Fe^{2+}	2.83	20.94	22.57	15.27	19.61	25.05			18.2	9.13		
Fe_{tot}	3.96	38.19	36.08	20.17	27.48	27.89	29.8	21.8	24.5	9.22		
Fe^{3+}/Fe^{2+}	0.40	0.82	0.60	0.32	0.40	0.11			0.35	0.01		
P_2O_5/CaO	0.80	0.63	0.18	1.11	0.87	0.09		0.41	0.18	0.73		

a) Analysed in Ore Laboratory, b) Analysed by Anneli Forsen (OES), c) Analysed by Ringa Danielsson (OES), d) Analysed by Christer Ahlsved (AAS), e) Analysed by Risto Saikkonen, x) Total iron analysed in Ore Laboratory.

1) Chert mesoband (Rb 102/74).
2) Magnetite-amphibole-chert mesoband (Rb 103/74).
3) Amphibole-magnetite-chert mesoband (Rb 104/74).
4) Amphibole-chert-biotite mesoband (Ra 598/76).
5) Amphibole-biotite-magnetite-chert mesoband (Ra 599/76).
6) Amphibole-biotite-garnet interband (Rb 105/74).
7) Garnet-amphibole-siderite-biotite interband (Ra 310/75).
8) Biotite-amphibole-garnet interband (Ra 2187/74).
9) Biotite-amphibole-garnet interband (Ra 321/74).
10) Phosphorite interband (Rb 23/75).
11) Phosphorite interband (Rb 96/75).
12) Phosphorite interband (Rb 22/75).

Table 8

Average partial chemical compositions (Wt-%) of the Pääkkö (P.) and Iso-Vuorijärvi (I.V.) iron formations. Results weighted for drill-core lengths. Analyses were made in the Ore Laboratory of the Chemistry Department, Geological Survey of Finland.

	MOSF			CF			MSOF	SF	Whole formations b)		
	P.	I.V.	Av.	P.	I.V.	Av.	I.V.	I.V.	P.	I.V.	Av.
TiO_2	0.10	0.08	0.10	0.07	0.11	0.08	0.08	0.55	0.10	0.13	0.10
Fe_2O_3 a)	38.96	40.02	39.06	37.99	33.89	37.13	44.87	27.78	38.86	37.74	38.73
MnO	0.09	0.06	0.09	0.05	0.09	0.06	0.04	0.04	0.09	0.06	0.09
P_2O_5	2.52	2.77	2.54	2.68	2.18	2.57	3.41	1.97	2.54	2.58	2.54
S	1.99	1.96	1.99	1.59	1.74	1.62	1.43	7.70	1.95	2.41	2.00
V	0.03	0.04	0.03	0.03	0.03	0.03	0.03	0.07	0.03	0.04	0.03
Fe_{tot}	27.25	27.99	27.32	26.57	24.37	26.11	31.38	19.43	27.18	26.53	27.10
Fe_{HCl}	20.25	21.57	20.37	24.72	23.42	24.45	27.62	18.94	20.69	21.98	20.84
Total core length analysed (m)	385.60	40.10	425.70	42.45	11.30	53.75	65.70	4.80	428.05	56.20	484.25

a) total iron
b) MSOF of Iso Vuorijärvi excluded.

Silicon: The silicon content of the rocks of the iron formations proper is, of course, more or less directly proportional to the abundance of chert bands. Thus, the two quartz-siderite-banded rocks analysed for this element contain 30.87 % and 37.83 % SiO_2, whereas chert-poor siderite rock contains only 19.70 % SiO_2 (Table 5). The average silica content for the MOSF rocks (40.78 % SiO_2, Table 4, A) is markedly higher than that of CF rocks (29.47 % SiO_2, Table 5, A). These values agree with the SiO_2 contents given by James (1954, p. 274) for the two facies. The average value for the MOSF is close to the average silica content of Precambrian iron formations in general (42.9 % SiO_2, Lepp and Goldich 1964).

The only silica determination of the SFP rocks indicates that this facies is richer (47.27 % SiO_2, Table 4, No. 10) in silica than are the rocks of other facies in Pääkkö. The comparatively low iron contents of the other SFP rocks corroborate this concept.

The relatively low silica contents in MSOF rocks are partly due to the primary lack of chert and partly to the leaching out of chert (Table 4, Nos. 4 and 5, respectively).

In the SF rocks, the silica content varies and depends on the mutual proportions of clastic quartz, chert and silica fixed in silicate minerals in each rock.

Titanium: In the rocks of the iron formations proper, except the SFP, the amount of titanium is negligible; in silicate analyses TiO_2 varies from 0.00 % to 0.10 % (Tables 4 and 5). The highest average of ore analyses is 0.11 % TiO_2 for the CF of Iso Vuorijärvi (Table 8). Richer in titanium are the sulphide facies rocks and noncherty garnet-bearing iron-silicate interbands (Table 7, Nos. 6—9) as well as, in some measure, the SFP rocks (Table 4, Nos. 8—11) and amphibole ± biotite-rich mesobands (Table 7, Nos. 4 and 5). The average TiO_2 content of the iron-rich black schists of the Iso Vuorijärvi and Körölä formations is 0.52 % (Table 13). In the garnet-bearing interbands, the TiO_2 content ranges from 0.58 % to 1.19 %. All these values are close to the titanium contents in the ordinary black schists and phyllites in Väyrylänkylä (Tables 10—13). Iron-rich black schists of the Seppola formation seem to be somewhat poorer in titanium than are the sulphide-facies rocks in the eastern margin of the Salmijärvi basin (Table 13). The titanium contents of the SFP rocks (0.08—0.28 % TiO_2) and the amphibole ± biotite-rich mesobands (0.13 % and 0.45 % TiO_2) are intermediary between the high- and low-titanium rocks mentioned (see Fig. 64).

Table 9

Average total chemical compositions of the summed Pääkkö and Iso Vuorijärvi iron formations recalculated to the average iron contents of the formations (Table 8). MSOF and SFP are excluded. Values for P_2O_5 (and for corresponding CaO) and S in brackets are averages of ore analyses (Table 8). For comparison the average chemical composition of the Biwabik Iron Formation, Minnesota, is reproduced (Lepp 1966) as well as that of Russian Proterozoic (Pt_{1-2}) ferruginous rocks (Ronov and Migdisov 1971, Table 2).

	MOSF 1)	CF 2)	SF 3)	Total formations 4)	Biwabik	Av. Russian
SiO_2	43.72	30.36	41.18	42.21	46.12	44.45
TiO_2	0.08	0.03	0.56	0.08	0.04	0.14
Al_2O_3	0.76	0.54	7.45	0.80	0.86	2.10
Fe_2O_3	14.34	4.59	1.46	13.13	19.47	26.42
FeO	22.24	29.46	23.68	23.05	19.26	18.79
MnO	0.09	0.04	0.01	0.08	0.66	0.10
MgO	4.05	3.78	2.63	4.01	2.88	2.27
CaO	4.42	4.86	1.48	4.44 (3.47)	1.79	1.65
Na_2O	0.11	0.04	0.34	0.10	0.05	0.18
K_2O	0.13	0.09	2.41	0.15	0.14	0.50
P_2O_5	3.02	3.07	0.56	3.00 (2.54)	0.07	0.17
CO_2	2.82	21.77	1.41	4.90	6.79	1.24
H_2O+	1.99	1.05	2.81	1.89	1.68*)	1.98
$H_2O—$	0.06	0.01	0.07	0.05		
C	1.06	0.58	8.64	1.00	0.20	0.08
S	1.09	0.55	9.69	1.12 (2.00)		
	100.00	100.82	104.38	100.01		
O = S	0.48	0.24	4.23	0.49		
Total	99.52	100.58	100.15	99.52	100.01	100.07
Fe^{3+}	10.03	3.21	1.02	9.18	13.62	18.48
Fe^{2+}	17.29	22.90	18.41	17.92	14.97	14.61
Fe_{tot}	27.32	26.11	19.43	27.10	28.59	33.09
Fe^{3+}/Fe^{2+}	0.58	0.14	0.06	0.51	0.91	1.26
P_2O_5/CaO	0.68	0.63	0.38	0.68	0.04	0.10

*) Ignition

1) Recalculated from analysis A, Table 4.
2) Recalculated from analysis A, Table 5.
3) Recalculated from analysis A, Table 6.
4) Average of 1, 2, and 3, weighted for drill-core lengths (Table 8).

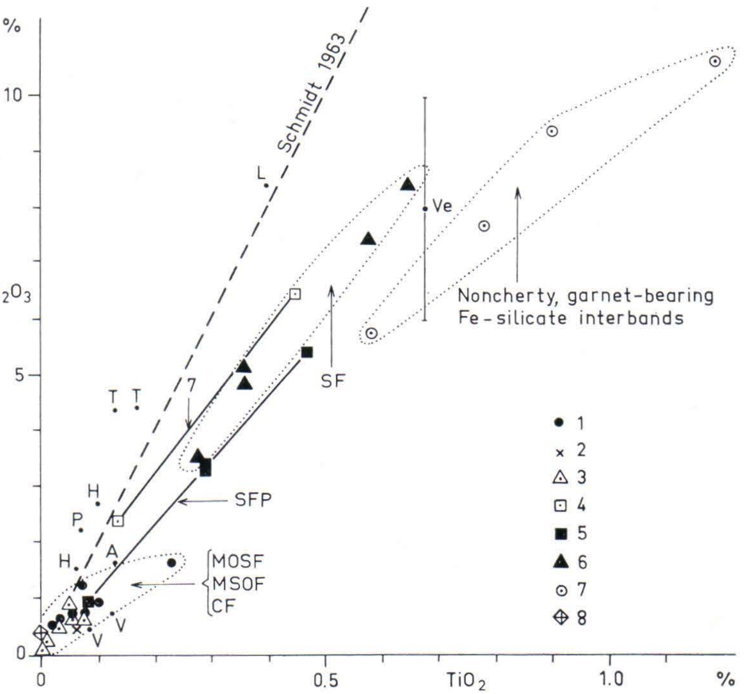

Fig. 64. Al_2O_3/TiO_2 plot of the iron-formation rocks of the MOSF, MSOF, and CF (1, 2, and 3, respectively), amphibole ± biotite-rich mesobands (4), SFP and SF rocks (5 and 6, respectively), noncherty garnet-bearing iron-silicate interbands (7) and a phosphorite interband (8) in Väyrylänkylä. Points marked by letters denote the other Karelian iron-formation rocks from Kainuu and Peräpohja (data from Lehto 1975 and Niiniskorpi 1975); A = Aapajoki, H = Hiisivaara, L = Lauttalahti, P = Poskimäki, T = Tuomivaara, V = Vähäjoki, Ve = Vepsä). The hatched line is from Schmist's (1963, p. 34) diagram and represents the average of TiO_2/Al_2O_3 ratios in »normal» sedimentary rocks; ratios to the right are »abnormal». Erratum: 7 near hatched line should be 4.

Microprobe analyses (Table 2) reveal that the main titanium-bearing mineral is biotite, which is also reflected in the relatively high titanium content of the graphite-rich spessartite-biotite band in comparison with the carbonate-pyrrhotite-spessartite-band in Seppola (Table 6, Nos. 7 and 6, respectively). Thus the comparatively high titanium content of the SFP rocks should apparently be attributed to the presence of biotite-rich interbands. Titanium shows a clear positive correlation with aluminium (Fig. 64). The ratio TiO_2/Al_2O_3 is slightly higher in garnet-bearing interbands than it is in iron-rich black schists.

Aluminium: The rocks of the iron formations proper are, again with the exception of the SFP, low in aluminium as are Precambrian iron formations in general. The average Al_2O_3 contents in the MOSF (excluding Nos. 6 and 7 in Table 4), MSOF, and CF rocks are 0.74 %, 0.32 % and 0.52 %, respectively (Tables 4 and 5). The Al_2O_3

contents, which are one order of magnitude higher in the SF rocks (3.48—8.42 %, Table 6), in garnet-bearing interbands (5.75—10.58 %, Table 7) and in amphibole ± biotite-rich mesobands (2.36 % and 6.42 %, Table 7), distinguish these from the iron formations proper, excluding the grunerite-rich rocks of the SFP, where the Al_2O_3 varies from 0.87 % to 3.44 % (Table 4). It is noteworthy that also the MOSF rocks (Table 4, Nos. 6 and 7) intimately associated with the latter rocks are somewhat richer in this element than are the other MOSF rocks.

Iron: On the basis of systematic ore analyses, the average total iron content for the Pääkkö and Iso Vuorijärvi deposits is 27.10 % (Table 8). This value is surprisingly close to the averages reported for iron formations of the Lake Superior type from the northern hemisphere: the total iron content for Biwabik Iron Formation is 28.59 %, (Lepp 1966, Table 4) and the overall average of the total iron content of 2200 core samples from the Precambrian iron formations of the Canadian Shield is 26.7 % [1]) (Lepp and Goldich 1964). The average total iron content of 670 Russian Proterozoic ferruginous rocks is 33.09 % (calculated from Ronov and Migdisov 1971, Table 2). This slightly higher value may partly arise from the sampling difficulties caused by supergene oxidation, a process which e.g. in Krivoy Rog reached the depth of 2 km or more (Gershoyg and Kaplun 1973).

The amount of total iron in the Pääkkö and Iso Vuorijärvi iron formations decreases with the degree of oxidation in order: MOSF (Fe_{tot} = 30.51 %; Fe^{3+}/Fe^{2+} = 0.58 %, Table 4, A), SFP (Fe_{tot} = 23.4 %, Fe^{3+}/Fe^{2+} = 0.26, Table 4, Nos. 8—10), CF (Fe_{tot} = 22.23 %, Fe^{3+}/Fe^{2+} = 0.16, Table 5, B) and SF (Fe_{tot} = 19.43, Table 8; Fe^{3+}/Fe^{2+} = 0.06, Table 6, A). The differences between the CF and SFP are, however, insignificant. Similar trends appear when the total iron contents are compared with the contents of iron soluble in hydrocloridic acid for different average macrobands (Tables 10 and 11). As a natural consequence of their unusual richness in siderite two of the CF rocks, Nos. 2 and 8 in Table 5, contain conspicuously more iron than do the quartz-siderite-banded types. Owing to its specific mineralogy the quartz-Mn-siderite-banded rock (No. 7) is exceptionally poor in iron. As a result of secondary enrichment due to Preglacial weathering (Laajoki 1975 a) the average MSOF rock is slightly richer in iron than are other facies (Table 8). The iron in the altered iron-silicate-magnetite rock is mainly in ferric state (Fe^{3+}/Fe^{2+} = 2.20, Table 4, No. 5). In fact, of course, the silicate-rich parts of the iron formations proper must originally have been richer in ferrous iron than were the silicate-poor parts, as revealed by the analytical results of the SFP rocks.

The average iron-rich black schist of the Körölä formation is somewhat richer in iron (Fe_{tot} = 23.1 %, Table 11) than is that in Iso Vuorijärvi (Fe_{tot} = 19.4 %, Table 10), which again is slightly richer in iron than are the average iron-rich schists of the Seppola area (Fe_{tot} = 18.0 %, Table 13).

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[1]) This value is evidently an average of analyses made from iron formations of both Superior and Algoma (Keewatin) types, and as such, taking the totally different lithologic associations of these types into consideration, may be somewhat misleading.

Table 10

Partial chemical composition (Wt-%) of the different macrobands of the Iso Vuorijärvi iron formation in drill hole No. 345. The results are weighted for drill core lengths. Analysed in the Ore Laboratory of the Chemistry Department, Geological Survey of Finland. Mineral abbreviations used in Tables 10—12 are as follows: qu = quartz, Fem = magnetite, crb = carbonate, sid = siderite, amph = amphibole. Macrobands included in iron-formation unit (DP_4) are numbered from the bottom upwards. Schists immediately below (DP_3) and above (DP_5) the iron-formation unit are marked with 0 and 00, respectively.

Marcoband	Depth interval analysed	Core length (m)	Fe_{tot}	Fe_{HCl}	TiO_2	MnO	P_2O_5	S	V
Black schist-00	105.35—106.35	1.00	7.15	5.94	0.72	0.05	0.25	6.44	0.07
» »	—107.35	1.00	10.9	9.20	0.58	0.06	0.37	1.49	0.06
(Crb-amph-rock-1)	—108.70	1.35	21.0	14.9	0.05	0.05	4.88	3.98	0.03
Qu-Fem-banded rock-2	—123.70	15.00	31.4	24.2	0.07	0.03	3.14	1.76	0.03
Qu-sid-banded rock-5 ..	—126.00*)	2.20	28.6	26.3	0.10	0.04	2.64	1.75	0.04
Fe-rich black schist-6 ..	—127.70	1.70	21.3	20.9	0.52	0.05	3.67	6.24	0.07
Qu-sid-banded rock-4**)	—129.40	1.70	25.2	23.6	0.13	0.05	2.96	3.60	0.04
Fe-rich black schist-5 ..	—130.20	0.80	20.3	20.3	0.55	0.04	1.15	9.65	0.07
Qu-sid-banded rock-3 ..	—131.00	0.80	24.1	22.5	0.12	0.14	1.33	1.95	0.03
Fe-rich black schist-4 ..	—131.50	0.50	19.1	19.1	0.42	0.03	2.25	9.61	0.07
Qu-sid-banded rock-2**)	—132.00	0.50	25.0	23.2	0.17	0.13	1.54	3.09	0.03
Fe-rich black schist-3 ..	—132.80	0.80	19.5	18.0	0.63	0.04	0.69	9.32	0.06
Qu-sid-banded rock-1 ..	—138.80	6.00	22.5	22.4	0.10	0.10	1.95	1.07	0.03
Fe-rich black schist-2 ..	—139.40	0.60	14.8	14.6	0.60	0.04	0.71	6.20	0.05
Phyllite-1	—140.40	1.00	6.68	6.41	0.48	0.13	0.14	1.45	0.02
Fe-rich black schist-1 ..	—140.80	0.40	16.9	15.8	0.68	0.03	0.60	6.70	0.06
Qu-Fem-banded rock-1	—142.40	1.60	27.3	16.2	0.08	0.13	2.41	1.46	0.03
Phyllite-0	—143.40	1.00	8.56	8.47	1.10	0.06	0.11	0.09	0.03
»	—144.40	1.00	8.31	8.22	1.15	0.08	0.02	0.01	0.03
Phyllite-0 (Fem-bearing)	150.85—151.85	1.00	9.70	9.18	1.17	0.14	0.37	0.05	0.02
Phyllite-0 (crb-banded) .	163.15—164.15	1.00	6.26	5.99	0.78	0.19	0.30	0.06	0.03
Average macroband:									
Qu-Fem-banded rock 1—2		16.60	30.9	23.3	0.07	0.04	3.08	1.73	0.03
Qu-sid-banded rock 1—5		11.30	24.4	23.4	0.11	0.09	2.18	1.74	0.03
Fe-rich black schist 1—6		4.80	19.4	18.9	0.57	0.04	1.97	7.71	0.07
Phyllite-1		1.00	6.7	6.4	0.48	0.13	0.14	1.45	0.02

*) Depth interval analysed 123.70—125.70 m.

**) Contains thin pyrrhotite-bearing black-schist interbands.

The total iron content of the garnet-bearing interbands (21.8—29.8 %, Table 7) is of the same order of magnitude as that in the iron formations proper.

Manganese: The content of manganese in the iron formations proper, except once more the SFP rocks and MOSF rocks associated with them, and in the SF rocks of the Iso Vuorijärvi and Körölä formations, is negligible (MnO $\lesssim$ 0.1 %, Tables 4—6, 8—11). The abundance of this element in the garnet-bearing interbands is similar to that in these rocks (MnO = 0.04—0.09 %, Table 7, Nos. 6—9). In Iso Vuorijärvi

Table 11

Partial chemical composition (Wt-%) of the different macrobands of the Körölä iron formation in drill hole No. 2. The results are weighted for drill core lengths. Analysed in the Ore Laboratory of the Chemistry Department, Geological Survey of Finland.

Macroband	Depth interval analysed	Core length (m)	Fe_{tot}	Fe_{HCl}	TiO_2	MnO	P_2O_5	S	V
Phyllite-00	18.40—20.53	2.13	9.09	8.89	0.88	0.06	0.62	0.23	0.03
Qu-Fem-banded rock-7	—24.00	3.47	26.2	19.0	0.05	0.10	1.99	1.78	0.03
Fe-rich black schist-5 ..	—25.90	1.90	22.9	20.2	0.42	0.06	1.40	7.01	0.04
Qu-Fem-banded rock-6	—27.40	1.50	28.5	19.5	0.27	0.13	1.92	1.99	0.03
Fe-rich black schist-4 ..	—27.90	0.50	23.8	19.3	0.60	0.06	2.61	4.86	0.05
Qu-Fem-banded rock-5	—40.50	12.60	27.1	16.0	0.10	0.09	2.15	1.90	0.03
Fe-rich black schist-3 ..	—43.60	3.10	22.8	17.6	0.37	0.05	2.47	6.21	0.04
Qu-Fem-banded rock-4	—44.60	1.00	30.5	15.8	0.08	0.13	2.36	1.79	0.03
Phyllite-2	—47.95	3.35	9.70	8.80	0.78	0.01	0.30	0.37	0.03
Qu-Fem-banded rock-3	—51.92*)	3.97	25.6	18.0	0.12	0.13	2.47	2.06	0.02
Fe-rich black schist-2 ..	—55.10	3.18	24.3	20.1	0.62	0.10	1.63	5.41	0.06
Qu-Fem-banded rock-2	—58.80	3.70	23.0	19.2	0.17	0.13	2.75	4.43	0.03
Phyllite-1	—60.50	1.70	13.0	12.5	1.47	0.14	0.32	0.95	0.04
Fe-rich black schist-1 ..	—60.97	0.47	17.6	16.8	0.57	0.06	2.20	7.36	0.06
Qu-Fem-banded rock-1	—61.40	0.43	20.0	16.0	0.18	0.21	2.27	6.98	0.03
Phyllite-0	—63.40	2.00	12.0	11.0	1.12	0.14	0.41	2.86	0.04
Average macroband:									
Qu-Fem-banded rock 1—7		26.67	26.3	17.3	0.12	0.11	2.26	2.34	0.03
Fe-rich black schist 1—5		9.15	23.1	19.1	0.49	0.07	1.95	6.08	0.05
Phyllite-1 and -2		5.05	10.8	10.0	1.01	0.05	0.30	0.57	0.03

*) Depth interval analysed 49.50—51.92 m.

the CF rocks are slightly, but distinctly richer in manganese than are the co-existing MOSF and SF rocks (Table 10). On the other hand, in Körölä and Pääkkö, manganese is more enriched in the MOSF than in the co-existing SF and CF, respectively (Tables 11 and 12).

The SFP rocks and MOSF rocks associated with them are regularly about twice as rich in manganese as are the iron-formation rocks described. In them manganese is incorporated in grunerite (Table 2, No. 2).

Against this general low background the SF rocks and, especially, the Mn-siderite macroband of the Seppola iron formation are rich in manganese (Table 6, Nos. 3—5, and Table 5, No. 7, respectively). Their main manganoan minerals are Mn-siderite and spessartite (Table 2).

Magnesium: The magnesium contents of the MOSF and CF rocks are almost constant. For the former the average is 3.78 % MgO (Table 4, A) and for the latter 3.36 % MgO (Table 5, B). These values as well as that of the only SFP rock analysed (Table 4, No. 10) are of the same order of magnitude as are the averages for the Precambrian iron formations and post-Precambrian ironstones (2.8 % and 2.9 % MgO, respectively, Lepp and Goldich 1964). According to microprobe studies, the main magne-

Table 12

Partial chemical composition (Wt-%) of the different macrobands of the Pääkkö iron formation in drill hole No. 343. The results are weighted for drill core lengths. Analysed in the Ore Laboratory of the Chemistry Department, Geological Survey of Finland.

Marcoband	Depth interval	Core length (m)	Fe_{tot}	Fe_{HCl}	TiO_2	MnO	P_2O_5	S	V
Black schist-00	93.20—92.20	1.0	11.9	10.5	0.57	0.08	0.41	4.50	0.07
Fe-rich black schist-00 ..	92.20—91.00	1.2	18.6	17.1	0.47	0.06	1.65	8.55	0.05
Qu-Fem-banded rock-3	91.00—88.00	3.0	28.3	19.9	0.10	0.09	2.88	1.40	0.03
Qu-sid-banded rock-1 ..	88.00—85.00	3.0	28.2	26.5	0.03	0.04	2.82	1.25	0.03
Qu-Fem-banded rock-2	85.00—77.10	7.9	32.1	19.9	0.12	0.09	2.57	2.48	0.03
Black schist-1	77.10—73.20	3.9	8.80	8.39	0.70	0.04	0.23	3.23	0.07
Qu-Fem-banded rock-1	73.20—55.20	18.0	28.6	21.8	0.10	0.09	2.20	2.16	0.03
Iron-rich phyllite-0	55.20—54.20	1.0	16.2	11.7	0.78	0.08	0.62	0.93	0.03
Phyllite-0	54.20—53.20	1.0	9.96	8.36	0.97	0.05	0.16	0.35	0.03
Average macroband:									
Qu-Fem-banded rock-1 —3		28.9	29.5	21.1	0.11	0.09	2.37	2.17	0.03
Qu-sid-banded rock-1 ..		3.0	28.2	26.5	0.03	0.04	2.82	1.25	0.03
Black schist-1		3.9	8.80	8.39	0.70	0.04	0.23	3.23	0.07

Table 13

Average partial chemical compositions (Wt-%) of iron-rich and/or manganiferous metasediments and black schists from the western margin of the Salmijärvi basin. Seppola iron formation was intersected by drill holes Nos. 335 and 337. Analysed in the Ore Laboratory of the Chemistry Department, Geological Survey of Finland.

Drill hole	Rock	Depth interval analysed	Core length (m)	Fe_{tot}	Fe_{HCl}	TiO_2	MnO	P_2O_5	S	V
335	Fe-rich phyllite	36.90— 38.90	2.00	19.2	18.6	0.25	0.92	0.87	10.4	0.05
	Fe-rich black schist	— 41.90	2.00	18.1	17.4	0.27	1.76	0.48	10.0	0.05
	» » »	100.00—102.00	2.00	16.4	15.3	0.42	3.55	0.34	8.00	0.04
	» » »	115.40—117.40	2.00	18.8	17.5	0.25	1.79	0.60	9.49	0.04
			8.00	18.1	17.2	0.30	2.01	0.57	9.47	0.05
337	Fe-rich black schist .	53.00— 63.00	10.00	18.3	17.6	0.33	2.75	0.71	9.36	0.05
	Fe-rich phyllite	93.00—105.00	12.00	17.6	16.8	0.25	1.92	0.85	9.56	0.06
	Fe-rich black schist .	107.00—109.00	2.00	19.3	18.3	0.32	0.49	0.32	12.3	0.05
			24.00	18.0	17.3	0.29	2.15	0.75	9.71	0.05
335+337	Average Fe-rich schist		32.00	18.0	17.3	0.28	2.10	0.69	9.65	0.05
337+341	Average black schist		12.00	10.0	9.60	0.48	0.10	0.44	7.33	0.07
337	Spessartite-bearing phyllite		1.00	9.94	7.89	1.00	3.96	0.30	1.15	0.03
	Average Fe-rich black schist of the Iso Vuorijärvi (Table 8) and Körölä (Table 11) iron formations		13.95	21.8	19.0	0.52	0.06	1.96	6.64	0.06

sium-bearing minerals in Väyrylänkylä are grunerite (MgO ~ 7—13 %) and carbonate (MgO ~ 5—8 %) and, in biotite-rich varieties, biotite (MgO ~ 4—8 %) (Table 2).

Compared with other rocks of the iron formations proper, the rock of the weathered silicate-facies is clearly depleted in magnesium (MgO = 0.83 %, Table 4, No. 5). In this respect it resembles weathered or »altered» Precambrian iron formations in general (e.g. Schmidt 1963, Table 4, Trendall and Blockley 1970, Table 11, No. 13, Gair 1975, Table 26).

Calcium: The MOSF and CF rocks are richer in calcium than are Precambrian iron formations in general (av. 1.59 % CaO, Lepp and Goldich 1964). The average for the former is 4.12 % (Table 4, A) and for the latter 3.59 % CaO (Table 5, B). The positive correlation between CaO and P_2O_5 (Fig. 65) clearly indicates that this element follows phosphorus (apatite). Moreover, their P_2O_5/CaO ratio (0.59—0.78) is almost constant and close to that in the phosphorite interband (0.73, Table 7, No. 10). The MSOF rocks and the SFP rock analysed show a similar correlation between these two elements and also their P_2O_5/CaO ratio is about 0.7 (Table 4).

On the other hand this obvious correlation is lacking in the SF and CF rocks of Seppola and in the garnet-bearing interbands of the Pääkkö formation (Fig. 65). In these rocks calcium is incorporated in carbonates (Seppola, pp. 35 and 41) and garnets (Table 2).

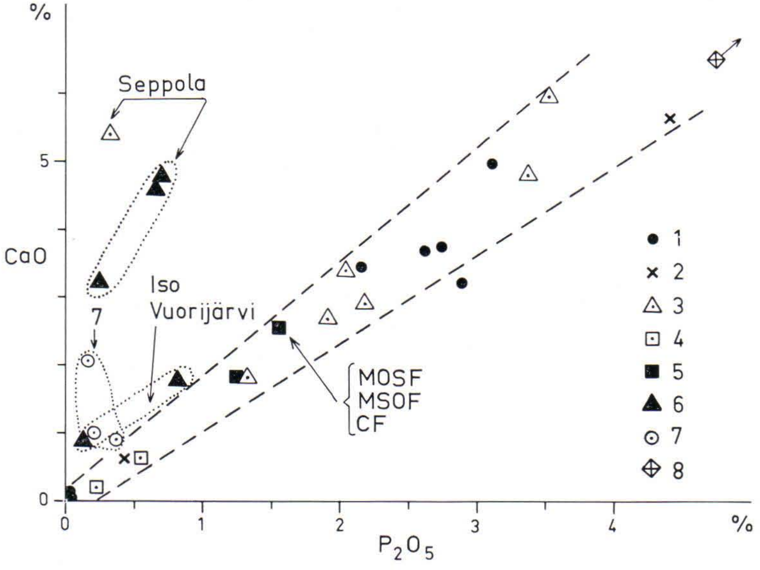

Fig. 65. CaO/P_2O_5 plot of the iron-formation rocks in Väyrylänkylä. The rock symbols are the same as in Fig. 64. Notice the different fields of the SF for Iso Vuorijärvi and Seppola as well as the field for the noncherty garnet-bearing iron-silicate interbands (7).

Sodium and potassium: Rocks of the MOSF, MSOF, and CF are remarkably poor in both sodium and potassium (0.00—0.16 % Na_2O, 0.02—0.18 % K_2O, Tables 4 and 5). Also in the sulphide-facies rocks the sodium content is low (0.01—0.47 % Na_2O, Table 6), whereas that of potassium is relatively high (0.42—2.86 % K_2O) as revealed by the low Na_2O/K_2O ratios in Table 6. The amphibole-rich mesobands and garnet-bearing interbands (Table 7, Nos. 4—9) contain relatively abundant sodium and potassium and thus in this respect they are unique among the rocks analysed from the Väyrylänkylä iron-formations. As in the sulphide-facies rocks, the main potassium mineral is biotite (Table 2), and that of sodium is feldspar and green hornblende. Of the two SFP rocks analysed, one (Table 4, No. 10) is comparatively rich in potassium, which can be attributed to the occurrence of biotite-rich interbands.

Phosphorus: The average phosphorus content (P_2O_5 = 2.54 %, Table 8) of the Pääkkö and Iso Vuorijärvi occurrences is one order of magnitude higher than that in Precambrian iron formations in general (average P_2O_5 = 0.26 %, Lepp and Goldich 1964, see also Barbour 1973, Tugarinov et al. 1973, Gair 1975, p. 133). The Precambrian (Karelian) occurrence of Tuomivaara is, however, also rich in phosphorus (2.06 % P_2O_5, Mäkelä 1976) as are those of Poskimäki and Hiisivaara (P_2O_5 = 1.19 % and 1.27 %, respectively, Niiniskorpi 1975, App. 2) and even Aapajoki (P_2O_5 = 1.24 % Lehto 1975, Table 14). In this respect, these Finnish formations are similar to some post-Precambrian ironstones rich in phosphorus, e.g. the oolitic limonite ironstones and chamositic ironstones described by James (1966, Tables 8 and 13, respectively). In Iso Vuorijärvi, the MOSF rocks are somewhat richer in phosphorus than are the rocks of the CF alternating with them (Table 10). Compared with other facies, the SF rocks are depleted in phosphorus, and in such a way that the eastern occurrences of the Salmijärvi basin in Iso Vuorijärvi and Körölä (average P_2O_5 = 1.97 % and 1.95 %, Tables 10 and 11, respectively) are richer in phosphorus than are the iron-rich schists in Seppola (P_2O_5 = 0.69 %, Table 13). The quartz-Mn-siderite-banded interband of Seppola is comparatively low in phosphorus (0.32 % P_2O_5, Table 5, No. 7). The average P_2O_5 content of the SFP rocks is 1.44 % (Table 4) which is lower than that of other iron formations proper.

The content of phosphorus in the garnet-bearing interbands is low (average P_2O_5 = 0.26 %, Table 7, Nos. 7—9).

Also the Kittilä (Paakkola 1971, Table 5) and Jauratsi (Rieck *et al.* 1967, p. 53) iron formations are unusually rich in phosphorus, while the counterparting quartz-banded iron ores of the Greenstone Group in northern Sweden show normal phosphorus contents, less than 0.1 % (Frietsch 1973, p. 79).

The higher values of phosphorus (27.0 % and 29.4 % P_2O_5, Table 7, Nos. 10 and 11) in the phosphorite interbands are close to the normal phosphorus contents in Paleozoic phosphorite, e.g. the Phosphoria Formation (Gulbrandsen 1966). Supposing that their apatite contains c. 40 % P_2O_5, it can be computed that the apatite content in the phosphorite interbands analysed varies from c. 40 % to c. 75 %.

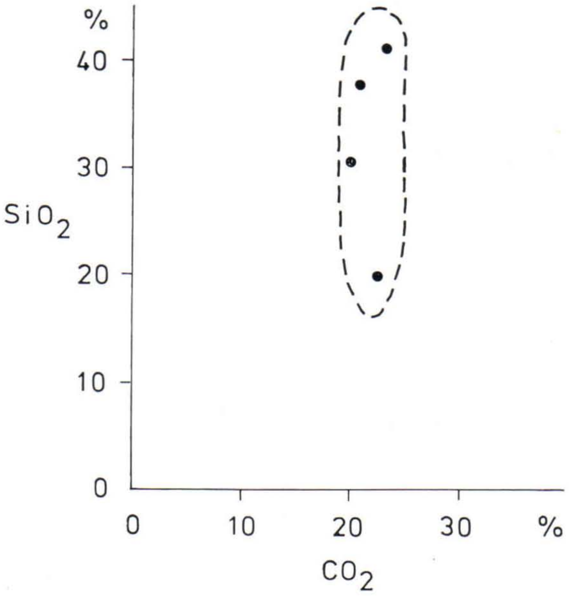

Fig. 66. SiO_2/CO_2 plot of CF rocks in Väyrylänkylä.

The greatly deviating phosphorus content in the three cherty iron-rich mesobands analysed (Table 7, Nos. 2—4), like the occurrence of the phosphorite interbands themselves, clearly indicates that, within the iron formations, phosphorus and iron were controlled by factors independent of one another. The two mesobands analysed from Seppola also contain highly deviating amounts of phosphorus (Table 6, Nos. 6 and 7).

Carbon dioxide: As a natural consequence of their mineral composition, the quartz-siderite-banded rocks of the CF contain an abundance of CO_2 (Table 5). Contrary to the statements by Lepp and Goldich (1966), no marked reciprocal relationship seems to exist between the SiO_2 and CO_2 contents in the Väyrylänkylä iron formations (Fig. 66, see also Trendall 1965 a).

In the other iron-formation rocks of Väyrylänkylä, the CO_2 content varies (Tables 4, 6 and 7) and is evidently directly proportional to the amounts of carbonate minerals in each rock (see Table 6, Nos. 6 and 7).

Carbon: The carbon content of the MOSF and CF rocks ranges from 0.26 % to 1.37 % (Tables 4 and 5). The rocks of the Pääkkö formation seem to be somewhat richer in that element than are corresponding rocks in Iso Vuorijärvi. The carbon content of the iron formations proper in Väyrylänkylä is similar to that in the Tuomivaara occurrence (graphite content 1—2 %, Mäkelä 1976).

One of the garnet-bearing interbands analysed is conspicuously rich in carbon (Table 7, No. 6). In the phosphorite interbands, the content of carbon varies, the darkness of the band evidently depending on the richness in carbon (Table 7, Nos. 10 and 11).

The SF rocks in Iso Vuorijärvi contain abundant carbon (Table 6). The iron-sulphide-poor part of the iron-rich black schist of Iso Vuorijärvi is markedly richer in carbon than is the iron-sulphide-rich part of the same rock (Table 6, Nos. 1 and 2, also C and D). The iron-rich metapelites in Seppola are comparatively carbon-poor, which is consistent with lithologic observations.

Sulphur: The rocks of the iron formations proper commonly contain small amounts of sulphur (0.13 %—3.19 %, Tables 4 and 5). Compared with the sulphur contents in the Precambrian iron formations in general (e.g. James 1966), the average sulphur contents in the MOSF and CF in Väyrylänkylä (Tables 8—12) as well as in the SFP (Table 4) are high. This is due to the thin pyrrhotite-bearing black-schist interbands and pyrrhotite dissemination and veins so frequent in the Väyrylänkylä rocks. There is no difference in the content of sulphur between coexisting MOSF and CF rocks (Table 10). The averages of sulphur for these facies in Väyrylänkylä (1.99 % and 1.62 % S, Table 8, respectively) are of the same order of magnitude as are those for the Tuomivaara iron formation (1—2 % S, Mäkelä 1976).

In the SF, the sulphur content varies from 4.39 % to 18.87 % (Table 6). The average sulphur contents for this facies of Körölä and Iso Vuorijärvi are slightly lower (7.71 % S and 6.08 % S, Tables 10 and 11, respectively) than those in Seppola (9.65 % S, Table 13). The same trend seems to exist in the »ordinary» black schists within the Salmijärvi basin (Tables 10, 12 and 13).

Vanadium: The average vanadium content for the iron formations proper is 0.03 % (Table 8). For the sulphide facies it is slightly, but regularly higher (0.05—0.07 % V, Tables 10, 11 and 13). The graphite-rich spessartite-biotite band and the carbonate-pyrrhotite-spessartite band analysed from Seppola (Table 7) contain 0.05 % and 0.02 % V, respectively.

Fluorine: This element was analysed only from three phosphorite interbands (Table 7, Nos. 10—12). The low abundances (0.11—0.17 % F) confirm that the apatite in question is free from F (*cf.* p. 71). These fluorine contents are lower than those in Paleozoic phosphorites.

Trace elements: Like iron formations in general (Landergren 1948, Schmidt 1963, James 1966, Barbosa and Grossi Sad 1973), the Väyrylänkylä iron formations proper as also the amphibole + biotite-rich mesoband and the garnet-bearing interband are poor in trace elements (Table 14, Nos. 1—7, 13 and 14). The latter two rocks seem to be slightly richer in Cr and Ni than are the former. In comparison with them, the trace-element contents in the SF rocks (Table 14, Nos. 8—12) are markedly higher and close to those in normal Precambrian black schists and phyllites in Finland (Peltola 1960, Lonka 1967).

Table 14

Trace-element contents (ppm) of the iron-formation rocks in Väyrylänkylä. Analysed by Ringa Danielsson, Maija-Leena Hagel-Brunnström, Ari Puisto and Paula Lindström.

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15
As	73	90		200	26		96	180		210		100	47		
Ba	< 100	< 100	< 100	< 100	< 100	< 100	< 100	< 100	300	< 100	200	200			570*)
Co	< 20	< 20	< 20	< 20		< 20	47	< 20	< 20	83	170	130	< 20	< 20	< 20
Cr	66	50	80	53		75		290	290	150	190	150	270	240	180
Cu	< 20	< 20	< 20	28		64	< 20	220	110	580	280	160	240	47	22
Ga								29	18	13	15	15	16	19	< 15
Ge	< 10	< 10	< 10	< 10	< 10	< 10	< 10	< 10	< 10	< 10	< 10		22	24	
Ni	< 20	< 20	< 20	< 20		< 20	66	380	280	160	150	120	170	60	< 20
Pb	15	9	8	42		6	70	35	27	22	31	27	< 10	31	
Sb	< 200	< 200		< 200	< 200	< 200	< 200	< 200	< 200	< 200	< 200	< 200	< 200	< 200	
Sc	< 30	< 30	< 30	< 30		< 30		< 30	30	< 30	< 30	< 30	< 30	< 30	< 30
Sn	< 3	< 3	< 3	< 3		< 3		< 3	< 3	< 3	< 3	< 3	< 3	< 3	
Sr	< 30	55	100	60		72		< 30	< 30	< 30	< 30	< 30		100	
Th															0.084*)
U		< 1	< 1							2		2			3.2*)
V								210	800	180	310	180			31.5*)
Y								95	98	41	49	75	43	25	500
Zn	< 100	< 100	< 100	< 100		< 100	< 100	220	540	100	150	270	< 100	100	
Zr	< 30	90	< 30					< 30	70	< 30	< 30	< 30	97	60	

*) Analysed by Riitta Zilliacus in the Reactor Laboratory of the Technical Research Centre of Finland.

1) Quartz-magnetite-banded rock, Table 4, No. 1.
2) Quartz-magnetite-banded rock, Table 4, No. 3.
3) Fe-silicate-magnetite rock, Table 4, No. 5.
4) Quartz-siderite-banded rock, Table 5, No. 1.
5) Phosphorite-banded siderite rock, Table 5, No. 2.
6) Quartz-siderite-banded rock, Table 5, No. 5.
7) Quartz-Mn-siderite-banded rock, Table 5, No. 7.
8) Pyrrhotite-rich black schist, Table 6, No. 1.
9) Black schist, Table 6, No. 2.
10) Spessartite-bearing iron-rich black schist, Table 6, No. 3.
11) Spessartite-bearing iron-rich black schist, Table 6, No. 4.
12) Spessartite-bearing iron-rich phyllite, Table 6, No. 5.
13) Amphibole-biotite-magnetite-chert mesoband, Table 7, No. 5.
14) Amphibole-biotite-garnet interband, Table 7, No. 6.
15) Phosphorite interband, Table 7, No. 10.

As by spectrophotometry. Ba, Sr and Zr by X-ray fluorescence spectrometry. U by fluorometry. Other elements by optical emission spectrography.

The manganiferous rocks of Seppola (Nos. 10—12) seem to be markedly richer in Co and somewhat poorer in Ni than do the SF rocks of Iso Vuorijärvi. The phosphorite interband (No. 15) is characterized by relatively high Ba (570 ppm), Cr (180 ppm) and Y (500 ppm) and by extremely low U (3.2 ppm) and Th (0.1 ppm).

Rare-earth elements (REE): The REEs in the Väyrylänkylä iron formation have recently been discussed in another publication (Laajoki 1975 b). Nevertheless, in Table 15 unpublished analytical data are given for an amphibole-biotite-garnet interband. The REE contents normalized to the composite of 40 North American shales (NAS) (Haskin *et al.* 1968, Table 2) are presented in Fig. 67. The low total REE content, the marked relative depletion in the lighter REE, and the lack of Ce depletion distinguishes this rock from the other iron-formation rocks in Väyrylän-

Table 15

Rare-earth content (ppm) of the amphibole-biotite-garnet interband (Table 7, No. 6) from the Pääkkö iron formation. Analysed by Riitta Zilliacus in the Reactor Laboratory of the Technical Research Centre of Finland.

La	4.7	Eu	0.39
Ce	9.8	Yb	1.1
Nd	4.1	Lu	0.25
Sm	1.3	Σ	21.64

Eu: Sm = 0.30

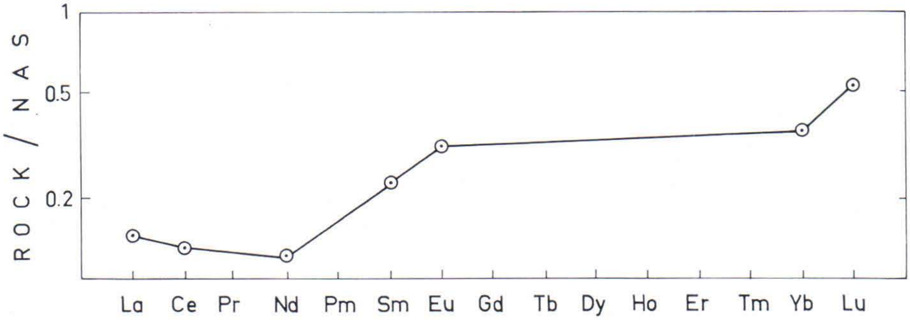

Fig. 67. NAS-normalized REE distribution pattern for amphibole-biotite-garnet interband (Table 7, No. 6).

kylä and the associated metapelites (*op. cit.*). On the other hand, its REE distribution and its mineralogical composition and chemistry in general, clearly resemble the amphibole-biotite-garnet rock from the schist wedge in the Pääkkö iron formation (No. 17 in *op. cit.*, No. 8 in Table 7 of this study).

Geochemical pecularities of the iron-formation rocks and general comparisons

Iron formations proper

Representative silicate analyses for the MOSF and CF rocks from the Pääkkö and Iso Vuorijärvi iron formations are set out in Table 4 (Nos. 1—3) and in Table 5 (Nos. 1, 2 and 5), respectively. The averages for each facies were used in calculating the total compositions for the summed formation of Pääkkö and Iso Vuorijärvi (Table 9).

As can be seen from Table 4 the quartz-magnetite-banded rock of the MOSF is characterized by high silica and iron oxides, which constitute about 80 % of the total oxides. In addition to them, the rock contains 3 to 4 % MgO, CaO, P_2O_5, CO_2, about 2 % S and about one percent Al_2O_3 and C. The contents of TiO_2, MnO, Na_2O and K_2O are negligible. The CF rocks (Table 5) are distinguished from these rocks only by their high FeO and CO_2 contents and relatively lower SiO_2 content. These differences are attributed to the presence of siderite instead of magnetite. The CF rock of the Seppola formation is distinguished from those in Pääkkö and Iso Vuorijärvi by its high Mn and, consequently lower Fe and low P contents.

If it is supposed that all Fe_2O_3 is incorporated in magnetite, all S in pyrrhotite, all CO_2 in siderite, all P_2O_5 in apatite (containing about 40 % P_2O_5), that the iron not soluble in hydrocloridic acid is in grunerite (containing ~ 40 % FeO) and that the remainder, excluding C calculated as graphite, is quartz, the following very rough estimates for the mineral compositions (weight percent) of the MOSF and CF based on their average chemical compositions (Tables 8 and 9) can be made. Mixed oxide-silicate facies: 38 % quartz, 21 % magnetite, 22 % grunerite, 7 % siderite, 6 % apatite 5 % pyrrhotite, 1 % graphite (*cf.* Lepp 1972). Carbonate facies: 51 % siderite, 26.4 % quartz, 7 % magnetite, 6 % apatite, 5 % grunerite, 4 % pyrrhotite, 0.6 % graphite. The amount of quartz is evidently too high at the expense of carbonate. As can be seen, in the MOSF the ratio of magnetite to grunerite is approximately 1 : 1. The ratio of quartz in the CF to quartz in the MOSF is about 0.69, which is close to the ratio of iron in siderite to iron in magnetite (0.67). This indicates that the lower SiO_2 content of the former facies is simply due to the occurrence of siderite instead of magnetite as its main iron mineral.

If the secondary effects of the Preglacial weathering (p. 86, 89, and 91) are filtered away, the chemical composition of the MSOF is apparently very close to that of the MOSF. Owing to the complete lack of chert, the relatively fresh iron-silicate magnetite rock (Table 4, No. 4) has the lowest silica content of the rocks of the iron formation proper, mesobands included, and its total iron is the second highest.

The comparatively high titanium, aluminium and potassium contents as well as the somewhat higher manganese content distinguish the SFP rocks from the other facies of the iron formations proper. Moreover, they are characterized by a high ferrous iron content. The exceptional high contents of Ti, Al, and K are attributed to biotite-rich mesobands and there is in fact one chert-mesobanded grunerite rock (Table 4, No. 9) that contains these elements in as negligible quantities as in the MOSF and CF rocks. On the basis of whole-rock and mineral analyses (Table 2), it can be estimated that this facies contains about 60 % grunerite, 10 % chert, 10 % biotite, 10 % carbonate (siderite?), 4 % pyrite, and 4 % apatite with 1 % graphite. The SFP generally shows slight geochemical gradation towards the SF (see Fig. 64).

As a whole, the geochemistry of the iron formations proper is similar to that of Precambrian iron formations in general (see e.g. Schmidt 1963, Lepp and Goldich 1964, Lepp 1966, James 1966, Trendall and Blockley 1970, and Gair 1975). They are, however, characterized by exceptionally high phosphorus and calcium contents owing to the presence of apatite. Furthermore, their sulphur and probably their carbon contents, too, are fairly high (Table 9). These geochemical features (high P, Ca, and relatively high S and C) seem to be characteristic of the Precambrian iron formations of Superior type in Kainuu. However, the MOSF and CF, the two prevailing facies of the Väyrylänkylä formations, are distinguished from the other Kainuan Superior type iron formations by the comparatively low aluminium and magnanese contents. In this respect, the SFP rocks resemble more closely the latter formations, in which iron-silicate-rich rocks are prominent lithologic units. Thus, in Tuomivaara for instance iron-silicate-rich rocks contain 3—6.2 % Al_2O_3 and 0.3—4.2 % MnO (Mäkelä 1976, Table 2) and according to Niiniskorpi (1975, App. 2), in Hiisivaara, Liskunkangas and Poskimäki the Al_2O_3 content varies from 1.44 % to 4.30 % and the MnO content from 0.11 % to 0.83 %.

Sulphide-facies rocks

The SF rocks are distinguished from the iron formations proper by high S and C contents, by distinctly higher Al, Ti, K and V contents and relatively lower Fe and P contents. Geochemically, they lie between the iron formations proper and ordinary black schists, e.g. those of the Outokumpu region (Peltola 1960). They differ clearly from the latter rocks by higher Fe and P contents and the low ratio of sodium to potassium (Tables 16 and 17). Table 17 reveals that the sulphide facies rocks of Precambrian iron formations are in general characterized by both low Na_2O content and low Na_2O/K_2O ratio. The only exception found in the literature is the pyrrhotite-bearing graphite schist from Kittilä. The SF of Seppola is characterized, moreover, by relatively high Mn and CO_2 contents, and by low P_2O_5/CaO ratios.

Chert, magnetite- and amphibole-rich mesobands, and amphibole ± biotite-rich mesobands

The chert mesoband, which, in addition to quartz, contains only some percents of carbonate, magnetite and grunerite (Table 1), is composed almost solely of silica with small amounts of iron oxides, MgO, and CO_2.

Table 16

Na_2O and K_2O contents of some black schists and one phyllite from the Salmijärvi basin. Analyses by Mervi Wiik (AES).

No.	Rock.	Stratigraphic unit	Na_2O	K_2O	$Na_2O : K_2O$
1	Black schist	DP_5	0.62	3.27	0.19
2	» »	DP_5	0.67	2.82	0.24
3	» »	DP_5	0.84	2.95	0.28
4	» »	DP_5	0.30	3.05	0.10
5	» »	DP_{3-4}	0.99	2.74	0.36
6	Phyllite	DP_3	0.93	1.74	0.53

1) Seppola, drill hole No. 341/ 72.00— 74.00 m.
2) Körölä, drill hole No. 327/ 48.45— 49.45 m.
3) Pääkkö, drill hole No. 343/103.40—104.40 m.
4) Pääkkö, drill hole No. 343/ 92.20— 93.20 m.
5) Pääkkö, drill hole No. 343/ 73.20— 75.20 m.
6) Pääkkö, drill hole No. 343/ 53.20— 54.20 m.

Table 17

Na_2O contents and Na_2O: K_2O ratios in sulphide-facies rocks from Väyrylänkylä and some other Precambrian iron formations compared with some ordinary carbonaceous schists.

Rock	Occurrence or region	Anal.	Na_2O	Na_2O/ K_2O	Reference
Iron-formation rocks ($Fe_{tot} \gtrsim$ 15 %)					
Iron-rich black schist	Väyrylänkylä	2	0.32	0.14	This study, Table 6, A
Mn-ferous iron-rich black schist	»	3	0.04	0.03	This study, Table 6, B
Black shale	Sokoman	4	0.10	0.02	Zajac 1974, Table 4, Nos. 1—4
Graphitic slate	Iron River	1	0.26	0.14	James 1951, Table 2
Dark green or black shales of Mount McRae Shale	Hamersley	5	0.30	0.26	Trendall and Blockley 1970, Table 15, Nos. 7—11
Argillite (Fe_{tot} = 13.9)	Ironwood	2	0.05	0.02	Huber 1959,Table 9, A and B
Pyritic rock	Soudan	1	0.16	0.06	Cloud *et al.* 1965, Table 1, No. 2
Pyrrhotite-bearing graphite schist	Kittilä	1	1.82	4.33	Lehto 1975, Table 6, No. 24
Ordinary black schists ($Fe_{tot} \ll$ 15 %)					
Black schist	Väyrylänkylä	5	0.68	0.23	This study, Table 16, Nos. 1—5
Calcareous black schist	Outokumpu	6	1.67	0.93	Peltola 1960 Table 8
Argillaceous black schist	»	14	2.06	0.95	» » » 7
Black schist	Hammaslahti	2	1.09	0.22	Nykänen 1971, Table 4, Nos. 3—4
Graphitic argillite (Fe_{tot} = 4.89 %)	Biwabik	1	1.62	0.28	White 1954, Table 3 No. 2
Carbon-rich rock	Soudan	1	0.17	0.07	Cloud *et al.* 1965, Table 1, No. 1

Like the quartz-magnetite-banded rocks, the magnetite- and amphibole-rich mesobands (Table 7. Nos. 2 and 3) are composed mainly of iron oxides and silica with c. 4 % MgO, and variable amounts of phosphorus with incorporated calcium. They are also distinguished by negligible amounts of Ti, Al, Mn, Na and K. Owing to their richness in iron minerals, especially in magnetite, the total iron content is some 10—20 % higher and the silica content correspondingly lower than the average in the representative MOSF rocks.

Owing to their deficency in magnetite the amphibole ± biotite-rich mesobands (Table 7, Nos. 4 and 5), for which total silicate analyses are unfortunately lacking, are more depleted in total and ferric iron than are the magnetite-rich types. They also differ from the latter, as well as from quartz-magnetite-banded rocks in general, by their relatively high Ti, Al, and K contents, attributed to the presence of biotite, and by the higher Na content evidently due to small amounts of plagioclase and green hornblende. In the latter respect, the rocks resemble the garnet-bearing interbands. In fact, the amphibole ± biotite-rich mesobands seem to lie geochemically and petrographically somewhere between these interbands and other MOSF rocks.

Noncherty garnet-bearing iron-silicate interbands

The garnet-bearing interbands analysed (Table 7, Nos. 6—9) are characterized by high Ti, Al, Na and K contents, by low P contents and by a low P_2O_5/ CaO ratio in comparison with the iron formations proper. Their total iron contents are, however, close to the average total iron contents of the MOSF and CF (Table 8). As a whole the geochemistry of these rocks is very similar to that of the SF rocks, except that these have much lower sulphur, carbon and phosphorus contents, and a high Na_2O/ K_2O ratio. In the light of the chemical composition of the main minerals of the bands (Table 2), the geochemical pecularities mentioned are due to the presence of grunerite (high Fe), almandine (high Fe, Al and relatively high Ti and Ca), biotite (high Fe, Al and Ti) and plagioclase (high Al, Ca and Na). In general, the rock is geochemically comparable with some Precambrian nongranular silicate iron-formations showing clastic features (e.g. James 1954, Table 8, Nos. 7—9) and even with some post-Precambrian chamositic ironstones, which exhibit almost identical chemical compositions (e.g. James 1966, Table 13, Stanton 1972, Table 13—5). They also correlate well both petrologically and geochemically with the almandine-biotite-amphibolite macrobands of the Precambrian ironstones (iron formations) from the Tonkolili district (Gaskin 1975, Table 5), and reveal striking similarities with some stilpnomelane mesobands from the Brockman Iron Formation (Trendall and Blockley 1970, Table 12).

As a whole the geochemistry of the amphibole-garnet mesoband is so unlike that of the iron formations proper, that their genesis must be totally different. This problem will be discussed in a later chapter (p. 108).

Phosphorite interbands

These bands are mainly characterized by high P and Ca incorporated in apatite. In addition to these two main elements the rock analysed (Table 7, No. 10) contains appreciable amounts of silica, ferrous iron and CO_2. Like the iron formations proper, it is characterized by negligible contents of Ti, Mn, Na and K. Also the Al content is low, which further indicates that the bands are of chemical-sedimentary origin. Supposing that apatite contains some 40 % P_2O_5 and that ferrous iron is bounded in siderite, the rock contains about 68 % apatite and 20 % siderite. The remaining 12 % consists of chloritized biotite, quartz, magnetite and graphitic substance. Sulphides are almost completely lacking from this kind of band.

In comparison with more recent phosphorites, e.g. those of the Phosphoria Formation (Gulbrandsen 1966), the Pääkkö phosphorite interbands are distinguished, not only by the high iron content attributed to their co-precipitation with sedimentary iron, but above all by the low F content due to the presence of hydroxyl-apatite. As far as the senior author knows the species of apatite has been identified only from two Precambrian iron formations: hydroxyl-apatite from the Brockman Iron Formation, Hamersley Basin (Morris 1973) and carbonate-apatite from a cherty iron formation in Michigan (Mancuso *et al.* 1975). These few determinations indicate that, in general, apatite in Precambrian iron formations may be poor in fluorine (*cf.* Påläng below).

A second geochemical feature that distinguishes the Pääkkö phosphorite interbands from the younger phosphorites is the extremely low U (Table 14, No. 15), even though the latter are known to contain potential amounts of this element (Altschuler *et al.* 1958). In the Phosphoria Formation, uranium is positively correlated with fluorine (Sheldon 1959), which indirectly suggests that the lack of U in Väyrylänkylä is due to the appearence of hydroxyl-apatite and not fluor-apatite. This conclusion seems to be corroborated by the chemistry of the apatite from Påläng, northern Sweden where thin apatite-rich intercalations occur in a Precambrian metasedimentary sequence consisting of graphite- and sulphide-bearing schists, dolomites, chemical sedimentary quartzites and low-grade iron ores (Frietsch 1974, p. 17). The apatite contains 1.9 % F and 0.2 % U (*op.cit.*, Table 1).

Lateral geochemical variations within the iron-formation horizon

In this chapter, the geochemical variations within the different facies of the iron-formation horizon (DP_4) of the Salmijärvi basin and the bulk geochemical variation within the horizon itself are discussed. As was stated in an earlier chapter (p. 9), the depth of the water in the sedimentation basin may have increased from Pääkkö and Körölä through Iso Vuorijärvi to Seppola. The discussion in the present chapter is based mainly on systematic ore analyses. It should be kept in mind that we had at our disposal highly variable amounts of geochemical data from each iron formations, as is revealed by the core lengths given in Tables 8, 10—13.

Mixed oxide-silicate facies (MOSF): The MOSF of Pääkkö and Iso Vuorijärvi are almost identical (Table 8). In the former, however, a comparatively greater part of iron seems to be incorporated in silicate minerals as is indicated by the somewhat greater difference between total iron and the iron soluble in hydrocloridic acid. Of the MOSF, that of Körölä seems to be richest in silicate iron (Table 11, average quartz-magnetite-banded rock).

Carbonate facies (CF): The carbonate facies in Pääkkö is slightly richer in total iron, silicate iron, phosphorus, carbon, and probably manganese than the same facies in Iso Vuorijärvi. Thin-section studies have revealed that the quartz-siderite-banded rock of the latter formation contains only small amounts of biotite and chlorite, and amphibole seems to be lacking completely. In contrast, the same rock in Pääkkö regularly contains a few percent of grunerite. Also the amount of magnetite is clearly greater here than in Iso Vuorijärvi, a fact that is geochemically indicated by the relatively higher Fe_2O_3 content in the carbonate-facies rocks in Pääkkö (Table 5).

Silicate facies proper (SFP): Because the only known occurrence of the rocks of the SFP is limited to the southern end of the main iron formation lense of Pääkkö, this facies seems to be a part of the tail of the Pääkkö formation deposited in somewhat deeper water than was the bulk of the formation. Furthermore, because it has been met with in only one drill hole, no internal geochemical comparisons of the SFP can be made within the Salmijärvi basin. However, as stated earlier (p. 100), it resembles both geochemically and petrographically the other iron formations in Kainuu, e.g. Poskimäki and Tuomivaara, that lie in metapelite-dominent surroundings in which dolomite occurrences are rare or altogether lacking.

Sulphide facies (SF): From Körölä (Table 11) through Iso Vuorijärvi (Table 10) to Seppola (Table 13) the content of total iron decreases, whereas that of sulphur increases. This indicates that in the SF of Seppola a relatively higher proportion of iron is bounded by sulphur than it is in the two more eastern facies. The phosphorus content in Körölä and Iso Vuorijärvi is higher than in the Seppola formation. In Körölä, where the iron-rich black schists are often garnet-bearing, the content of silicate iron is remarkably high. On the basis of colour observations and chemical analyses (Table 6), the SF in the eastern part of the Salmijärvi basin contains conspicuously larger quantities of carbon than do the SF of Seppola. In Seppola the manganese and CO_2 contents of the sulphide facies are distinctly high. As to trace elements, the Co content is higher and the Ni content lower in the Seppola rocks than in the SF rocks of Iso Vuorijärvi (Table 14). The Pääkkö iron formation is overlain by a pyrrhotite-rich black schist that contains enough iron to be classified as iron formation (Table 12). This rock represents, however, a transition between iron formation and ordinary black schists and has been included, for ore-inventory reasons, in the Salmijärvi Phyllite Member.

Total iron-formation horizon: On the basis of the general facies relationships in the Salmijärvi basin (p. 18) and the geochemical aspects discussed on the foregoing pages, the following geochemical regularities can be drawn for the iron-formation horizon itself: From the part deposited in the shallowest water (Pääkkö and Körölä) by way of a slightly deeper-water formation (Iso Vuorijärvi) to the formation deposited in the deepest or at least most off-shore part of the Salmijärvi basin (Seppola) the contents of total iron, silica fixed in chert and phosphorus decrease, whereas the sulphur content, and, consequently, the content of iron incorporated in sulphides, as well as the carbon, manganese, titanium, aluminium and potassium increase. These observations are consistent with the general facies principles presented by James (1954). Even so, there is always the possibility that in Väyrylänkylä some of the MOSF derived from the original CF by metamorphic processes (p. 76).

Geochemical banding

Trendall has recently re-emphasized the genetic significance of the specific banded structure or, in other words, the heterogeneity of Precambrian iron formations. In

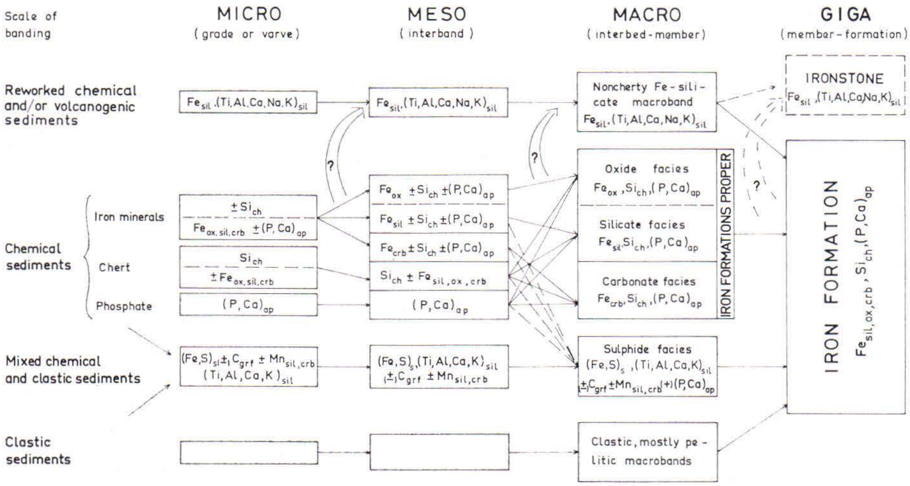

Fig. 68. Components of geochemical banding in the Väyrylänkylä iron formations. The hatched line separating the fields of the oxide and silicate facies denotes mixing. The most distinctive elements and the minerals in which they are now incorporated are indicated. The mineral abbreviations are: ap = apatite, ch = chert, crb = carbonate, grf = graphite, ox = oxides, s = sulphides, sil = silicate. Plus-minus before an element indicates that the element is less characteristic, but still significant. Manganiferous bands refer to the Seppola formation.

one of his earlier papers he stated: »Banding is a complex and puzzling feature of Precambrian iron formations, and explanation of it must be included in any serious attempt to explain their origin» (Trendall 1965 a, p. 1069). In this chapter the geochemistry of bands on different scales, that is, the geochemical variations in a vertical direction of the Väyrylänkylä iron formations are discussed and some genetic reasonings given for them. It is believed that metamorphism was isochemical, and so the different kinds of geochemical banding were formed through primary sedimentary processes. Fig. 68 gives a schematic presentation of different bands in the Väyrylänkylä iron formations with the elements that distinguish them most markedly from each other and from the associated metasediments with normal iron content. The treatment of the banding on the two smallest scales, micro- and mesobanding, is restricted to iron-formation rocks.

Microbanding

According to petrographic observations, a microband in the quartz-magnetite-banded and quartz-siderite-banded rocks can be regarded as a grade or varve ¹) composed of chert-prominent and iron-mineral-prominent parts, of which the latter may contain variable amounts of apatite. Similar graded structure has previously been described as an unusual feature from the Dales Gorge Member and the Weeli Wolli Formation (Trendall and Blockley 1970, Figs. 24 C and 26 C, respectively). Later,

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¹) Used without any glacial implication.

Trendall (1973, p. 1091) stated that in the striped facies of the Weeli Wolli Formation the boundaries between darker and paler stripes of microbands are gradational, rather than sharp. Although Trendall and Blockley considered that microbands are annual seasonal varves, they were not able to explain the graded structure (*op. cit.* p. 268). Later, Trendall (1972) argued that microbands are chemical evaporitic varves.

Conclusive evidence is still lacking as to which part of the grades in the Väyrylänkylä iron formations was deposited first. The restriction of quartz nodules to the iron-mineral ± apatite-rich parts of the grades suggests that this part was deposited first. If this hypothesis is accepted, the top-and-bottom-determinations made on the western flank of the Pääkkö syncline on graded iron-formation rocks and rhythmically deposited quartzites and phyllites of the Pääkkö Quartzite Member uniformly exhibit mainly eastern top directions. Indirect support comes from the Negaunee Iron Formation, where clastic sandy grains are restricted largely to ferruginous laminae (Gair 1975, p. 112). Moreover, almost 50 years ago Moore and Maynard (1929, p. 518) demonstrated experimentally that from the solution containing ferric hydroxide and silica the silica required a longer period than the iron oxide to settle under the force of gravity. Hence, it seems quite possible that the iron-rich parts of the graded microbands are older than the chert parts.

On the basis of petrographic observations it can be concluded that in terms of geochemistry a microband is a band consisting of a stratigraphically lower part enriched in iron that is fixed in either oxide and/or silicate, or carbonate and of an upper part consisting merely of silicon incorporated in chert. The lower part may also contain variable amounts of phosphorus and calcium fixed in apatite. It seems probable that the deposition of silica was a continuous process but that it was overtaken by the deposition of iron at the onset of the sedimentation cycle, as indicated by experiments by Moore and Maynard (1929, see also Belevtsev and Melnik 1976).

Meso- and interbanding

In addition to the normal clastic (phyllite and black-schist) and mixed chemical sedimentary — normal clastic (iron-rich phyllite and black-schist) interbeds which will be discussed in the following chapter, two main types can be distiguished among the meso- and interbands: 1) those extremely poor in Ti, Al, Ca not fixed in apatite, Na and K, which thus resembles the general geochemistry of iron formations proper and 2) those showing roughly as high Ti, Al, Ca, Na and K contents as the normal clastic pelitic rocks in general. The bands of the first group form the bulk of the iron formations proper and can further be classified into three mesoband classes, namely oxide- and iron-silicate-rich, iron-carbonate (in Seppola manganoan iron-carbonate) and chert mesobands, and into phosphorite interbands. The first two mesobands are geochemically characterized by the same elements as are the corresponding macrobands (p. 112) except that they are enriched in iron at the expense of silica; the chert bands consist mainly of silica; the phosphorite interbands consist mainly of phosphorus and calcium incorporated in apatite.

The second group includes two petrographically distinct types: 1) chert-bearing iron-silicate (amphibole ± biotite)-rich mesobands (Table 7, Nos. 4 and 5) and 2) noncherty iron-silicate (amphibole + biotite + garnet) interbands (Table 7, Nos. 6—9). The chemical differences between the types are slight. Especially the chert-poor variety of the former type cannot be distinguished from the second merely on the basis of chemical analysis. A hint of their different petrology is, however, given by the P_2O_5/ CaO ratios, which reveal that the cherty types are genetically related to the iron-silicate-rich mesobands of the first group. The chemistry of the noncherty types resembles in many respects that of the SF rocks, although in them iron is fixed by silicates rather than by sulphur.

Trendall and Blockley (1970) attributed the development of mesobands in the Hamersley Basin to probable diagenetic compaction controlled by astronomical factors. The analogous banding in Griquatown was considered by Cullen (1963) to be a product of intermittent tectonic activity regulating a differential influx of materials into the sedimentation basin. It is important to remember that, in the Väyrylänkylä formations, in addition to the »normal» mesobanding of iron- and silica-rich minerals, there is a second banding on the same scale superimposing the first. The second banding is composed of phosphorite interbands. The independence of phosphate deposition from iron deposition is shown by the poor correlation between the P_2O_5 and Fe_{tot} peaks in Figs. 69 and 71 [1]). On the scale of macro- and gigabands though, phosphate and iron follow each other. A similar mutual correlation was detected by Morris (1973) for phosphate banding and mesobanding in Hamersley and he maintained that phosphate bands are essentially a primary depositional feature. Thus, before turning our attention to the controls of mesobanding, we have to discuss why some of the phosphate in Väyrylänkylä deposited together with iron as indicated by microbanding observations, and why some deposited alone.

Unfortunately, we have not yet found enough evidence to establish whether the phosphorite bands in Väyrylänkylä are of organic or inorganic origin. Consequently, any statements about their controls are more or less speculative, and we can do no more than list some possible explanations for the mutual independence of mesobanding and phosphorite interbanding: 1) Phosphate and iron (together with silica?) have different sources. 2) Phosphate and iron (together with silica?) derive from a common source but in varying mutual proportions and at different times. 3) Phosphate and iron deposited at different times. 4) Some erratic external factor caused random abundant deposition of phosphate (e.g. a sudden influx of volcanic ash as discussed by Lowe 1972). Of these possibilities, the third is ruled out by the fact that apatite, like siderite, is considered primary. Consequently, phosphorite interbands, in agreement with current general opinion of the process of phosphate sedimentation (e.g. Gulbrandsen 1969, *cf.* Cook 1976), are to be regarded as primary sedimentary components. Furthermore currents, either upwelling ocean or estuarine cur-

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[1]) The seemingly good parallelism of these peaks in Körölä (Fig. 70) is evidently due to the comparatively long drill-core intervals analysed.

rents, are generally accepted as having controlled post-Precambrian phosphate formations.

To return to mesobanding of iron and silica, the occurrence of phosphorite interbands gives a certain right to maintain that currents were probably also the predominant regulator of the alternation in silica- and iron-rich mesobands (*cf.* Borchert 1960). The quantity and quality of their load was probably controlled by rhythmic climatic changes.

Finally, it has been contended that owing to the distinctive geochemistry of the noncherty garnet-bearing iron-silicate interbands they are of different origin than the other iron-formation rocks (p. 102). In the opinion of Gaskin (1975), rocks with a similar chemical and mineral composition are products of a pelitic phase with some chemical precipitation. Trendall and Blockley (1970, see also LaBerge 1966 a and b) describe stilpnomelane meso- and macrobands (shales) from the Hamersley Group, which greatly resemble chemically as well as texturally and, as to minor constituents, even mineralogically, the garnet-bearing interbands in Väyrylänkylä. They suggest that the environment of shale deposition was controlled jointly by increased volcanic activity and tectonic instability (Trendall and Blockley 1970, p. 290). Moreover, they describe shard bands of direct volcanic origin. On the basis of one REE analysis, Laajoki (1975 b, p. 102) concluded that the garnet-bearing interband analysed could not be of normal clastic origin. He considered that the band contains tuffaceous material or material derived from basic rocks or consists of reworked iron-formation material. At this time no total silicate analyses were available from either these interbands or the SF rocks of Seppola. Because, in Seppola there are biotite- and K-feldspar-rich bands that seem petrographically to be almost analogous to the stilpnomelane tuffs of the Joffre Member (Trendall and Blockley 1970, p. 78) and because the second REE analysis from the garnet-bearing interbands (Table 15) confirms the unusually low (compared to contemporaneous »normal» metapelites) REE content of the rock and, furthermore, because their Al_2O_3/TiO_2 ratios (Fig. 64) strongly suggest »abnormal» origin, Laajoki is now disposed to accept a more direct volcanic origin for these interbands. However, additional work is needed before the final conclusions can be drawn. In this respect the rocks of the Seppola area appear to play a key role.

Macrobanding

Ore analyses of the well-developed macrobands in the Iso Vuorijärvi and Körölä formations are given in Tables 10 and 11, respectively. The phyllite macrobands clearly represent periods of normal mechanical sedimentation as is indicated by the low total iron, phosphorus and sulphur contents and the high titanium content (Figs. 69 and 70). The black-schist macroband in Pääkkö is discerned from its iron-rich surroundings by its low total iron and phosphorus and its high vanadium and titanium (Fig. 71).

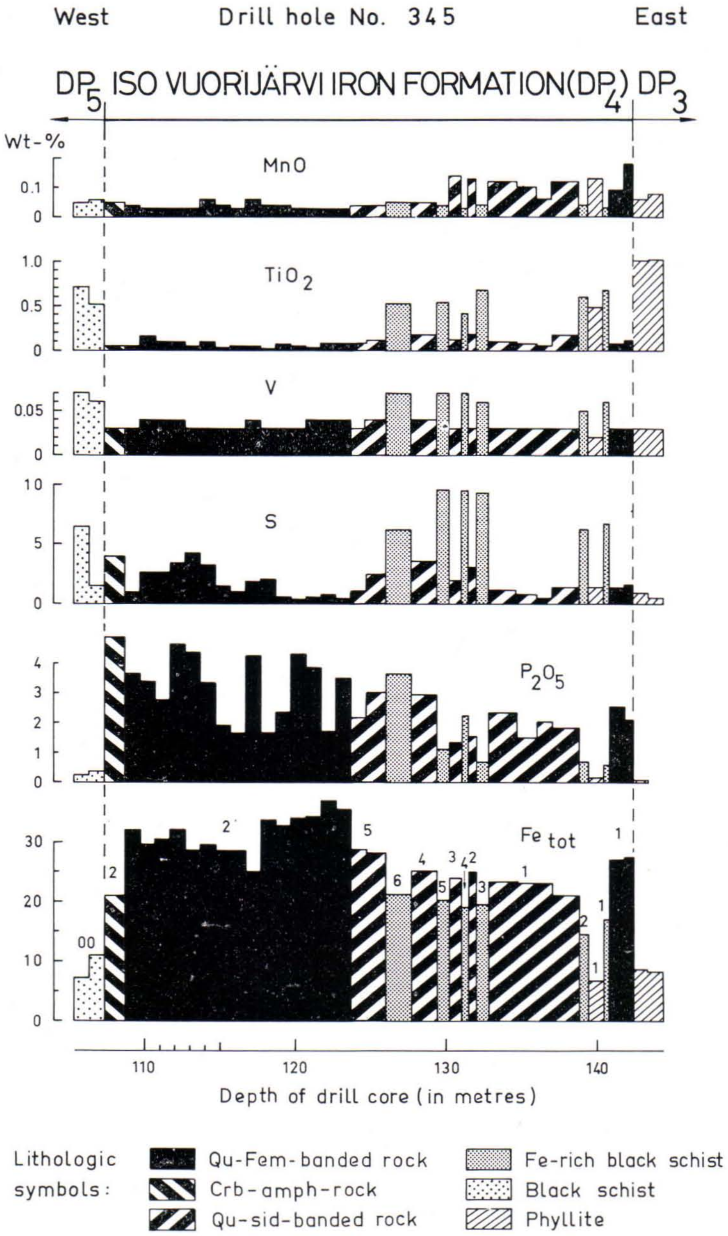

Fig. 69. Lithologic-geochemical macrobanding in the Iso Vuorijärvi iron formation. Probable stratigraphic top to the left.

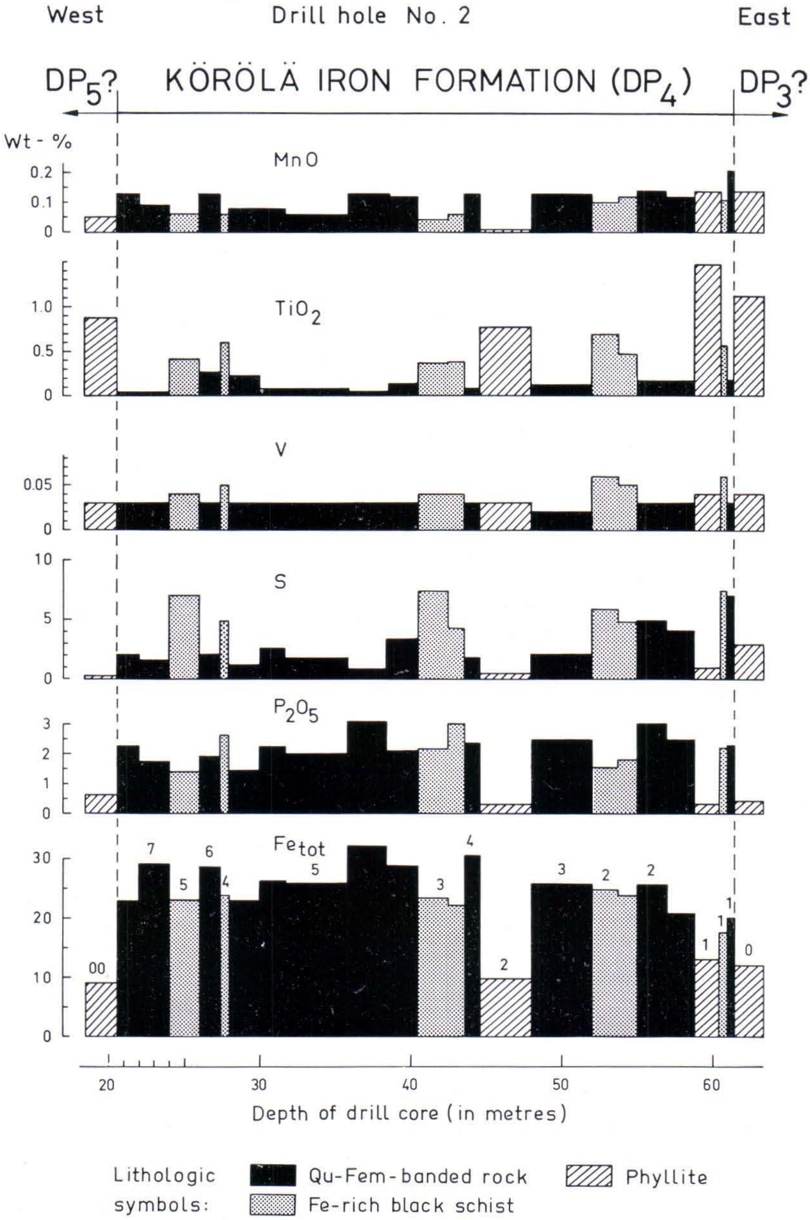

Fig. 70. Lithologic-geochemical banding in the Körölä iron formation. Probable stratigraphic top to the left.

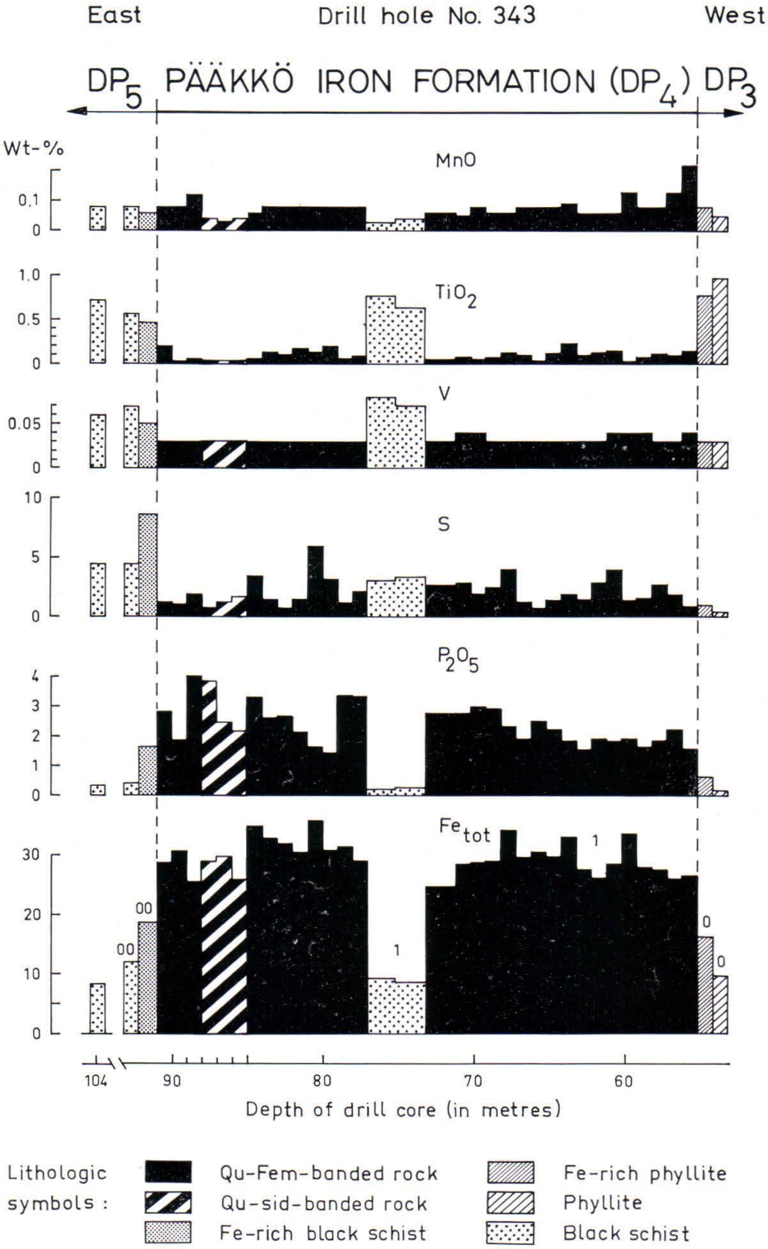

Fig. 71. Lithologic-geochemical banding in the Pääkkö iron formation. Stratigraphic top to the left.

In Iso Vuorijärvi, the iron-rich black-schist macrobands are distinguished from the associated CF macrobands mainly by high sulphur, vanadium and titanium peaks (Fig. 69). Their total iron contents are regularly slightly, but unmistakably lower than those in the adjoining macrobands. Differences in phosphorus contents vary, sometimes the SF macroband is richer than the CF macroband in that element, sometimes the reverse is true. The chemical differences between the co-existing SF and MOSF macrobands in Körölä are relatively small (Fig. 70). Once again the best indicators of difference are sulphur, vanadium and titanium, whereas differences in total iron and phosphorus contents are insignificant. Here, too, the iron-rich black schists seem to be slightly depleted in manganese.

Macrobands of iron formation proper (quartz-magnetite- and quartz-siderite-banded rocks) are characterized by high total iron, high, but greatly variable phosphorus, and extremely low titanium. In the CF macrobands, the iron is almost totally in a ferrous state as is indicated by the minor differences between their total iron and the iron soluble in HCl (Table 10). Both macroband types are markedly richer in manganese than is the co-existing SF rock.

Unfortunately there are no systematic ore analyses for the macrobands of the Seppola iron formation. However, the silicate analyses (Table 5, No. 7 and Table 6, Nos. 3—5) indicate that the relationships between the co-existing CF and SF macrobands are analogous to those in Iso Vuorijärvi. Here as well the relative enrichment of manganese into the carbonate facies seems to be marked.

The foregoing and earlier discussions on the results of silicate analyses indicate that the macrobands within the Väyrylänkylä iron formations can be classified in terms of their element contents as follows: 1) $Fe + Si + (P + Ca)_{apatite}$-rich and $Ti + Al + Ca + Na + K$-poor macrobands (iron formations proper), 2) $Fe + S + C \pm Mn \pm (P + Ca)_{apatite}$-rich macrobands with almost normal distribution of Ti, Al, Ca and K, but poor in Na (sulphide facies), 3) macrobands with normal elemental distribution (phyllites and black schists). The first-mentioned can be further divided according to their main iron-bearing mineral into MOSF, SFP and CF macrobands, of which the first is chemically characterized by comparatively larger amounts of ferric iron, the second by a comparatively high ratio of ferrous iron to ferric iron, and the third by high ferrous iron and CO_2.

The geochemical variations between the different macrobands, especially the MOSF versus the SF in Körölä and the CF versus the SF in Iso Vuorijärvi, are so abrupt and so distinct that their formations are believed to have been controlled by tectonic processes. This banding is much like the alternation of sericite-phyllite beds with quartzite beds in the Quartzite Formation-I in the South Puolanka area, which was explained as having been controlled by faulting (Laajoki 1973, p. 44). Without going into details, we may be justified in noting that also these rocks contain markedly more iron fixed in hematite than do other Jatulian rocks in the environment. Gair (1975, p. 113) concluded that the many interlayered beds of clastics in the Negaunee Iron Formation are difficult to attribute to minor episodes of transgression and regression (*cf.* Button 1976 a), but that »continuing fault displacements during deposition and adjacent to the site of iron-formation deposition seems to offer the best explanation for the difference between »normal» clastic-poor iron-formation and the mixed rock of the Palmer area» (*op.cit.* p. 114). Furthermore, Cullen (1963) suggested that the banding of the Griquatown iron formations is related to recurrent isostatic readjustments. Trendall and Blockley (1970, p. 291) suggested that primary controls of major shale/iron formation alternation as well as the macrobanding of the Dales Gorge Member in the Hamersley Basin were tectonic adjustments followed, in the former case, by increased volcanic activity.

Thus, it is concluded that the macrobanding in Väyrylänkylä was very likely caused by intermittent block faulting due to dominantly vertical tectonism.

Gigabanding

On the scale of gigabands the eastern iron formations in the Salmijärvi basin (Tables 10—12, Figs. 69—71) constitute a stratigraphic unit (member) which, in comparison with the country rocks, is enriched in iron incorporated in oxide, silicate and carbonate minerals, as well as in silica of chemical sedimentary origin (chert) and in phosphorus and calcium fixed in apatite. On the other hand, they are strongly depleted in Ti, Al, Ca not fixed in apatite and in alkalies, and slightly depleted in Mg. The sulphide iron-formation of Seppola represents the western extension of this iron-rich unit. Here, however, owing to the mixed chemical sedimentary — clastic sedimentary origin of the prevailing sulphide facies (p. 100), the chemical contrasts between the formation itself and the country rocks are reduced and the formation is distinguished from its surroundings only by the higher iron content, which in this case is due almost solely to iron sulphides, and by the somewhat higher phosphorus content (Table 13). Although the Seppola formation is considerably more enriched in manganese than are the eastern iron formations it is not in this respect unique within the Seppola area itself, since manganiferous country rocks with almost normal iron distribution (spessartite-bearing phyllite in Table 13) are also encountered.

In short, the Pääkkö, Iso Vuorijärvi and Körölä iron formations are Fe-Si-P-units exhibiting clear-cut contacts with country rocks. The Seppola iron formation represents a deeper-water (off-shore) tail or part of the same iron-rich horizon with less distinctive geochemical pecularities and with more gradational contacts. The abundance of P distinguishes the Väyrylänkylä iron formations from other Precambrian iron formations elsewhere in the world (see Table 9). In this connection it is appropriate to recall that also in the Kittilä iron formations phosphorus and iron follow each other (Kaitaro 1949, p. 144).

The conditions that prevailed at the formation of gigabanding will be discussed in the next chapter.

Concluding remarks

The Väyrylänkylä iron formations show manifold lithologic-geochemical banding. They can be regarded as a composite of chemical sediments contaminated by normal clastic and probably also by volcanic sediments. Chemical sediments are chert, iron minerals and phosphate, of which the first two predominate. Moreover, they contain an abundance of organic matter as graphite or graphitic substance. The Väyrylänkylä iron formations demonstrate the close association of phosphate and organic matter with iron minerals and chert that prevailed as early as the Middle Precambrian, which suggests that the environment of Precambrian iron formation of Superior type and phosphate deposition have many factors in common (compare the geochemical model of Drever, 1974 to that of Gulbrandsen, 1969).

PALEOGEOGRAPHIC RECONSTRUCTION OF THE SEDIMENTATION BASIN. A FIRST APPROACH

The paleogeographic basinal conditions that regulated the genesis of iron-formation and ironstone and even the deposition of phosphorite have been much debated. Most authors agree at least that sedimentation took place in a marine basin sheltered or cut off from the open sea. James (1954) connected the basin evolution to orogenic cycle and attributed its restriction to structural development of off-shore buckles or swells that subsequently developed into island arcs characterized by volcanism. Trendall and Blockley (1970) compared the iron formations in Hamersley to evaporites and suggested deposition in a closed evaporitic basin about 200 metres deep. Recently, Chauvel and Dimroth (1974) have published a comprehensive study of paleofacies in which they conclude that the distribution of facies types of the Sokoman Iron Formation suggests deposition in two basically different environments: a lagoonal platform and a basinal environment separated by a narrow domain of oolite shoals. A not so very different environment was envisioned by Button (1976 a) for the deposition of the iron formations of the Transvaal Supergroup. The recently found Nabberu iron formations are similar to those in the Animikie Basin and Labrador Trough and are considered by Goode and Hall (1976) to represent the shallow water marginal equivalents of the Hamersley iron formations.

Väyrylänkylä, with complicated structure and sparse outcrops, poses numbers of problems in the reconstruction of the paleogeographic environments. Even so, on the basis of general stratigraphic-lithologic observations it can be concluded that the deposition of the Marine Jatulian took place in a restricted marginal basin isolated from eugeosyncline by a geanticline (Laajoki 1973, p. 44). The present study confirms this concept. It should be emphasized, however, that eugeosyncline refers to a relatively deep water regimen and miogeosyncline to a shallow-water regimen close to the continent without any bearing on plate-tectonic regimens. Because the present study indicates that vertical tectonism had a greater significance during the evolution of the Karelidic sedimentation basins in the study area than previously realized, the term geanticline is used in the sense of uplifted land with no reference to lateral compressions.

Moreover, in the light of recent mapping carried out by the Exploration Department of the Geological Survey of Finland north and west of the South Puolanka area, it seems that the Karelidic formations of the northern end of the Kainua Schist Belt were originally situated somewhere between Oulujärvi and the Utajärvi-Ylikiiminki schist area (Laajoki 1976 b). In other words, the major thrusts took place from the southwest to the northeast and not from the west to the east as Wegmann (1928, p. 12) concluded in the Melalahti area. He further concluded that the Karelidic schists in Kainuu are preserved in the so-called Oulujärvi axial depression and that the axial culmination zone of Nurmes lies between this and the Karelidic schists of the Kuopio area. In the light of our present knowledge, however, the Kainuan schists of the Nuasjärvi basin are relatively autochthonous and thus the environment of Oulujärvi represents an erosion remnant of sedimentation basin overturned slightly towards the northeast and originally separated from the Kuopio area by a topographic high. This interpretation is corroborated by the absence of Superior type iron formations of Karelidic age from the latter area and the Karelides southeast of it.

On the basis of these observations and interpretations it is concluded that the Karelian iron formations were deposited in Kainuu into a paleobay or -lagoon that opened towards the northwest and was isolated from the open sea by a strip of dry land between Kainuu and the Salahmi schist area (Fig. 72). This reconstruction differs from Väyrynen's concepts only in that the approximate Jatulian paleoshoreline is drawn from the North Karelia area via the Salahmi, described excellently by Savolahti (1965), and Kemi areas to Kuusamo, whereas Väyrynen (1954, Fig. 55) drew it via Puolanka to Kuusamo.

As to the distribution of iron-formation facies within the Kainuan paleobay, the comparatively Al-poor MOSF and CF of the Pääkkö, Iso Vuorijärvi, and Körölä formations apparently deposited somewhat closer to the northeastern shore of the bay than did iron-silicate-rich formations of Tuomivaara, Poskimäki and other smaller occurrences distinguished by their slightly higher Al and Mn contents. The SF of Seppola very probably deposited close to the southwestern shore of the bay. This is confirmed by the occurrence of Marine-Jatulian black schists with cummingtonite-bearing intercalations in Losonvaara, in the western margin of the Nuasjärvi basin (Äikäs 1975, p. 41). The relatively low carbon content of the SF of Seppola indicates that conditions here were more turbulent than further in the northeast.

It is not easy to establish the depth of the water in the basin. As pointed out by Dimroth (1975) special care must be taken if the mineralogy of iron-formation is used for bathymetric purposes. The most striking stratigraphic features of the eastern iron formations of the Salmijärvi basin are the alternation of iron formations proper with the carbonaceous and pyrrhotite-rich black schists of the SF and the relatively small size of the iron-formation bodies as well as the lack of mutual correlation between adjacent occurrences. These features are at their best in the Körölä area (Fig. 5), possibly because of the original basin topography (*cf.* Väyrynen 1928, p. 102).

If this is true it means that in the sedimentation basin there were a number of small topographic depressions into which iron-rich mud and material of iron formations proper were deposited alternately. The sedimentation environment, as a whole, seems thus to correspond to the lagoonal platform facies of the iron formations

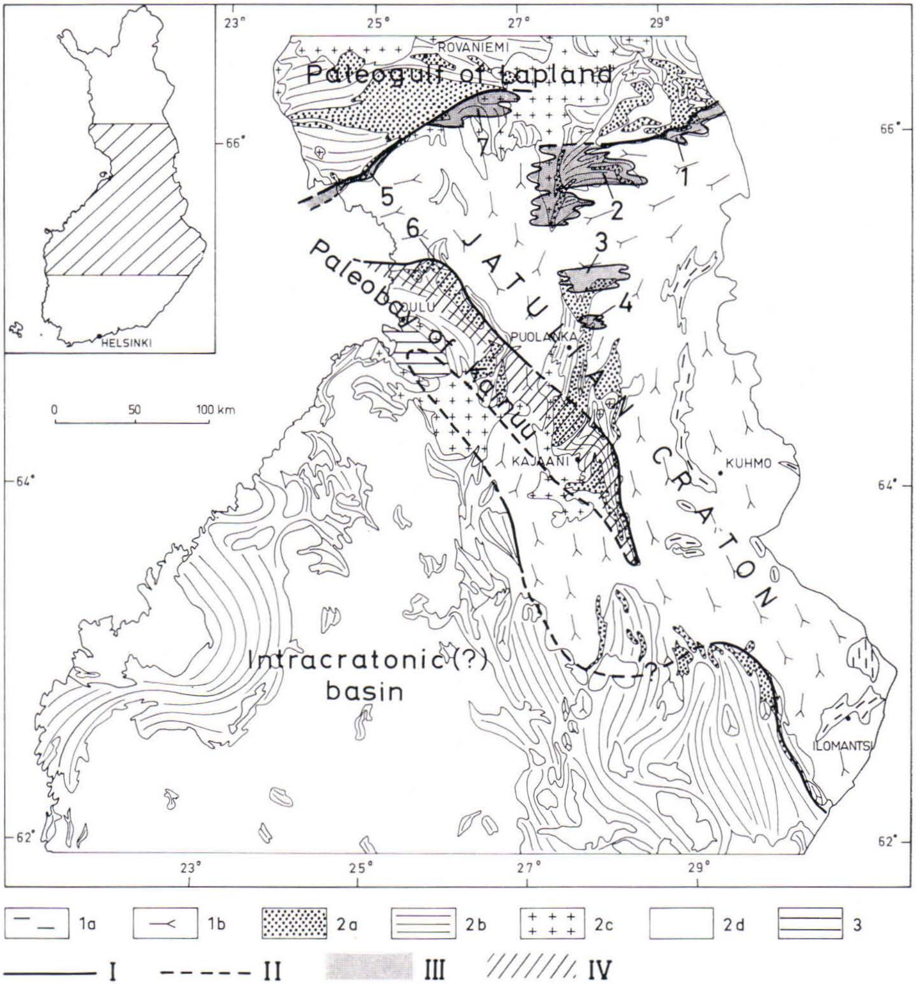

Fig. 72. Schematic representation of the main paleogeographic conditions during the Jatulian — Marine Jatulian in Finland. Not corrected for the flattening caused by compression from southwest to northeast neither for the N-S fault movements. I) Fairly allochthonous discordance surface between the Prekarelidic basement or Prejatulian rocks and the Jatulian. II) The discordance surface strongly deformed or destroyed. III) Prejatulian basic metavolcanics and intrusive rocks whose Prejatulian distribution is represented schematically (1 = Greenstone-I of Kuusamo, 2 = basic rocks of Posio-Taivalkoski, 3 = Kurkikylä metavolcanics, 4 = Kolkonkangas metavolcanics, 5 = greenstones of Kemi area, 6 = greenstones of Yli-Ii, 7 = greenstones of Suhanko). IV) Area of iron-formation deposition. The base map is the same as in Fig. 1.

of Superior type described by Dimroth (1975, p. 360). However, the manifold banding of the Väyrylänkylä formations proves that the sedimentation conditions varied conspicuously. Hence, for instance the thin phosphorite interbands (Fig. 33 A) must have formed under quite low-energy conditions, whereas the phosphorite pellets in amphibole + biotite-rich mesoband (Fig. 40) apparently originated under agitated water conditions. More detailed work is needed before the exact sedimentary environments can be established.

So few observations have been made on relict oolitic textures in the eastern iron formations that these formations can be regarded as almost nonoolitic. They were deposited under less agitated water conditions and probably also in deeper water than the iron formations of Superior type in general in the northern hemisphere, in which oxide facies predominate and oolitic and other shallow-water textures and structures are more common (e.g. Gross 1972, Mengel 1973). This is corroborated by the unusually high carbon and sulphur contents, and apparently also by the phosphate content in the iron formation proper as well as by the CF or MOSF-/the SF macrobanding in Väyrylänkylä. So it seems quite probable that the Karelian iron formations in Kainuu represent a section of the iron formations deposited in comparatively stagnant or deep water, or both, and that the economically more promising shallow-water facies were eroded away.

The SF of Seppola resembles both geochemically and, in some measure, lithologically the manganiferous iron formations of the Kittilä area (Paakkola 1971). More detailed descriptions are, however, needed for both occurrences before the significance of these similarities can be properly interpreted.

GENERAL DISCUSSION

The literature dealing with Precambrian iron formations is manifold and a reader interested in general problems of these facinating metasediments is referred to a wealth of well-known publications by several authors (e.g. James 1954, 1966, Gross 1965, Trendall and Blockley 1970, Stanton 1972) and to special issues devoted recently to this subject (UNESCO 1973, Earth Sciences 9, Economic Geology 1973, vol. 68, No. 7, and Lepp 1975) as well as to the papers by Dimroth (1976) and Eichler (1976) in the new Handbook of strata-bound and stratiform ore deposits. In this connection, only those questions on which the present study throws more light will be discussed.

Source of iron and silica

The iron of the Precambrian iron formations is generally attributed to continental erosion, volcanism, or leaching of sea-floor detritus. Holland (1973) recently proposed the oceans as the possible source of iron.

The bulk chemical composition of the Prekarelidic Basement Gneiss Complex in Kainuu is granitic-granodioritic and, at first glance, it seems difficult to find any rock formation within it from which iron could readily have derived. Not so long ago, however, a flat-lying metavolcanics formation was found in Kurkikylä North Puolanka area, underlying Jatulian quartzites (Laajoki 1976 b). The formation seems to have been folded before the Jatulian and is correlative with Greenstone-I (early Karelidic volcanogenous rocks in the classification of Piispanen, 1972) in the Kuusamo area, which is likewise flat-lying and seems to have folded gently before the deposition of the overlying Jatulian rocks (Silvennoinen 1972, p. 40). Also in the Kemi area there occur basic amygdaloidal rocks lying below the Jatulian quartzites (Perttunen 1975). Furthermore, the conglomerate of Pikku Rytivaara, Suhanko area, with the abundant clasts of amygdaloidal rock (Isohanni 1971, p. 31) proves that also there extrusions of basic rocks took place before the sedimentation of the Jatulian. Thus it seems plausible that the area between Kainuu and Lapland could have been extensively covered by basic rocks before or even during deposition of the Väyrylänkylä iron formations (Fig. 72). Moreover, the Prekarelidic paragneisses in the North Puolanka area were folded into open, gently plunging antiforms and synforms, suggesting low relief for the Prekarelidic basement (Laajoki 1976 b).

Table 18

Fe_2O_3, TiO_2, MnO and P_2O_5 contents in some Marine Jatulian metadiabases and basic tuffites from the Salmijärvi basin. Analysed in the Ore Laboratory of the Geological Survey of Finland.

	1	2	3	4	5	6
Fe_2O_3 *)	24.3	11.5	26.7	20.0	14.3	10.2
TiO_2	2.24	0.50	2.02	1.32	1.20	1.15
MnO	0.31	0.19	0.04	0.08	0.18	0.21
P_2O_5	0.25	0.23	0.21	0.32	0.21	0.30

*) total iron

1) Metadiabase, Seppola (119 a/PE/65).
2) Metadiabase, Liejeenjoki (R 351/167,80 m) (Laajoki 1975 a, Table 2 No. 19)
3) Magnetite-bearing tuffite, Huhtalampi (209 h/KL/71).
4) Magnetite-bearing tuffite, Huhtalampi (209 i/KL/71).
5) Tuffite, Seppola (118/PE/65) (Laajoki 1975 a, Table 2, No. 18).
6) Carbonate-rich tuffite, Tupala (2 b/KL/73).

The occurrence of metavolcanics and metadiabases in the Salmijärvi basin (Table 18), and the noncherty garnet-bearing iron-silicate interbands in the iron formations themselves are suggestive of a volcanic source, as is the Al_2O_3/ TiO_2 ratio of the iron-formation rocks (Fig. 64). This was checked by plotting Al_2O_3/TiO_2 for some iron formations of both Algoma and Superior-type (Fig. 73). The North Karelian occurrences, which lie in pelitic metasediments (Lavikainen 1973) show ratios similar to »normal» sediments as does also Trommald Formation (Schmidt 1963, Fig. 11). The iron-formation rocks proper of Kuhmo, intimately associated with basic metavolcanics (Papunen 1960, Hyppönen 1973), and Hamersley and Väyrylänkylä are extremely poor in these two elements and so their points fall near to origin in the diagrams. Surprisingly, the points of the Kittilä occurrences, which are closely associated with basic metavolcanics (Paakkola 1971), show ratios that, although somewhat scattered, are close to the average Precambrian slate. The rocks of Kostamus (Kostamuksha) show rather random Al_2O_3/ TiO_2 ratios. All these Al_2O_3/ TiO_2 ratios, in one way or another rather illogical, can be explained as giving information on the origin of the material that contaminates the iron-formation material originally low in both Ti and Al. Fig. 73 also plots the famous Kiruna Ores. Except for the Hauki Hematite and some of the Per-Geiger Ores, their Al_2O_3/ TiO_2 ratios differ markedly from those of the iron formations discussed.

As to the third possible iron source, if it is assumed that the phosphate deposition in the Salmijärvi basin was caused by upwelling currents, then the occurrence of phosphorite interbands in Väyrylänkylä indicates that iron could have been carried from deeper parts of the ocean floor by currents. This is, however, indirect evidence of little, if any, value.

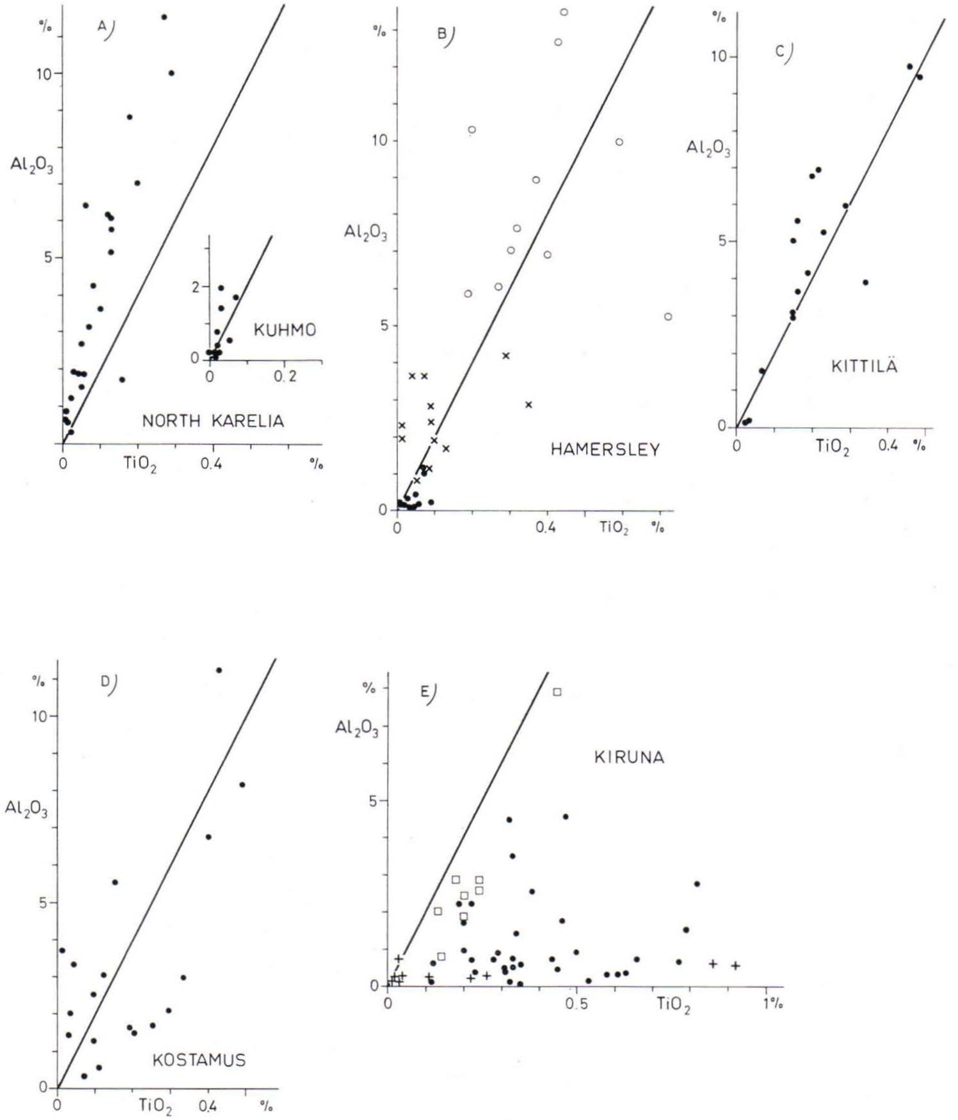

Fig. 73. Al_2O_3/TiO_2 plots (see Fig. 64) for some Precambrian iron formations. A) North Karelia (data from Niiniskorpi 1975, App. 2) and Kuhmo, Finland (*op. cit.*, Laajoki, unpublished data). B) Hamersley, W Australia; points = Brockman Iron Formation, crosses = thin green mesobands of the Dales Gorge Member, circles = Mt McRae Shale (Trendall and Blockley 1970, Tables 11, 12 and 15). C) Kittilä, Finland (Lehto 1975, Table 6). D) Kostamus, USSR (Chernov et al. 1970, Tables 3 and 26). E) Kiruna, Sweden; crosses = Main Ores, points = Per Geiger Ores, open squares = Hauki Hematite (Parák 1975, Tables 17—23).

The source of silica like that of iron has been much debated and is largely attributed to either continental weathering or volcanism. Eugster (1969) suggested that Precambrian iron formations formed in an alkaline lake and that magadiite was a chert precursor. This model does not fit the Väyrylänkylä formations, because the REE data (Laajoki 1975 b) rule out lacustrine deposition.

Role of micro-organisms in deposition of iron

It has been proposed by many authors that micro-organisms played a key role in the deposition of iron, a supposition for which there is, unfortunately, no concrete evidence to date. In Väyrylänkylä, where phosphate occurs in unusually great quantities, it could be expected that microfossils would be detected in abundance. So far, however, we have found no trace of them in phosphorite interbands (p. 51), and all suspicious spherical textures in Seppola (p. 38) have been interpreted as of inorganic origin. One of the most serious objections to biological deposition of iron is the occurrence of iron formations in lithologic associations of highly variable derivation, e.g. associated with metavolcanics (Helen, Goodwin 1964), metaturbidites (Mt. Belches, Dunbar and McCall 1971), epicontinental quartzites and dolomites (Superior-type in general), and even with glaciogenic rocks (Rapitan, Young 1976).

On the other hand, carbon and sulphur isotope data indicate that organisms played an important role in (or during) the deposition of iron formations of Algoma type even in Archean times (Goodwin *et al.* 1976).

Significance of phosphorite interbands

The petrographic characteristics of the phosphorite interbands confirm their marine origin, also proved by an earlier REE study (Laajoki 1975 b). Their presence proves that also the iron formations themselves are marine. Excluding some specific characteristics related to the simultaneous deposition of iron, the petrography and chemistry, especially the REE distribution of the phosphorite bands and their gross lithologic association in Väyrylänkylä are so similar to those of younger phosphorites that it is reasonable to suggest that they originated by analogous processes (see, e.g. Bushinskii 1969, Gulbrandsen 1969, deKeyser and Cook 1972, Cook 1976). This concept is further supported by paleomagnetic data which suggest a paleolatitude of 22°N for the Fennoscandian protocontinent during the Svecokarelidic orogeny (Neuvonen 1974.) Morris (1973) suggests an organic origin for much of the phosphate on the basis of his study on the qualitive distribution of phosphate in the Brockman Iron Formation. To explain the relationships between the iron, silica and phosphate deposition and to determine the roles of biogenic activity and the diagenetic processes below the sediment/water interface during accumulation of these three chemical sedimentary components in the Väyrylänkylä iron formations, more detailed work is needed.

Role of diagenesis

Many of the recent excellent petrographic and mineralogical studies of well-known low-metamorphic Precambrian iron formations referred to in this study have proved that diagenesis has greatly altered the primary mineral compositions as well as, but probably in a lesser degree, the texture of iron-formation rocks. Mostly, these phenomena, excluding iron reduction and the decrease in H_2O and CO_2 contents, can be considered as isochemical rearrangements within the iron formation. Kimberley (1974) and Dimroth and Kimberley (1976), however, favour the interpretation that iron formations were as a whole early diagenetic replacements of dominantly aragonitic sediment much like the local replacement of Quaternary Bahaman aragonite. Epigenetic processes, e.g. dolomitization, generally show discordant relationships with earlier structures. Except for the isochemical replacement textures mentioned above (Han 1971), such evidence has not been reported from Precambrian iron formations. Furthermore, banding, especially the persistence of microbands in space in Hamersley (Trendall 1972) and the graded bedding of iron and silica in Väyrylänkylä can be explained by this mode only with difficulty. The same seems to hold good of many even-grained Phanerozoic siderite mudstones showing graded bedding (Taylor 1967, p. 180). Since iron-bearing oolites have been forming in Recent time (Lemoalle and Dupont 1973), even in a lacustrine environment, there does not seem to be any reason why they could not have been forming during the Precambrian.

On the other hand, diagenetic processes may have played a significant role in Seppola, especially as regards the oolitic calcareous iron-rich metapelites. Also there, however, pure even-grained biotite- and graphite-poor Mn-siderite meso- and macrobands seem to be orthochemical in origin.

Role of metamorphism

In keeping with the tendency in Precambrian iron formations (e.g. Klein 1973) and other rocks (e.g. Ronov and Migdisov 1971, p. 162, Stanton 1976 a), prograde regional metamorphic processes were isochemical in the iron formations of Väyrylänkylä as well. In Pääkkö and Seppola particularly, the retrograde metamorphism was highly effective. During this stage, whenever garnet was replaced by carbonate, migration of silicon, aluminium, and magnesium took place (Table 2, Nos. 13 and 18). These elements do not seem, however, to have left the iron-formation system, but are encountered within it as quartz- and chlorite-rich veins.

As discussed earlier in this paper, it is probable that the MOSF and possibly also the SFP were derived from the primary CF by regional metamorphism.

The factors that caused the huge deposition of iron formation about 2100±200 Ma ago

When the present study was started it was hoped that with the aid of geochemical methods some new light could be thrown on the interesting problem of whether the typically chert-banded Precambrian iron formations and the noncherty post-Precambrian ironstones were the counterparts of lithologic units or whether they were totally different entities. In the course of the study it soon became evident that the latter is the more plausible alternative, because it was found out that the Väyryläkylä iron formations contain abundant garnet-bearing interbands which can be regarded, on the basis of their chemistry, as an ironstone component (Fig. 68) and because there are also chert-banded iron-rich sediments of relatively young age (Schultz 1966, O'Rourke 1961, Beukes 1973, Russel 1975; see also discussions in James 1969 and Cloud 1973).

Furthermore, many recent studies show that both the extra- and intrabasinal physico-chemical conditions of the Middle Precambrian may have been almost identical to those of younger eras and the Recent. For example, the relative portion of organic carbon in the total sedimentary carbon reservoir since about 3300 Ma seems to have been constant (Schidlowski *et al.* 1975). The Precambrian sea water does not seem to have been very different from the sea water of today (Holland 1976, *cf.* Veizer and Compston 1976). The marine phosphorites were deposited in greater quantities during the Precambrian than realized in general (Davidson 1963, Bushinskii 1969, also this study). The oxygen-free atmosphere revived recently by Cloud (1972) in order to explain the extensive formation of iron formations in the Precambrian is not supported by any unambiguous geological or geochemical evidence as discussed by Dimroth and Kimberley (1976, see also Pettijohn 1975, p. 422). Also the results of the comprehensive carbon-isotope study by Schidlowski *et al.* (1975) indicate, if the interpretation mode of Broecker (1970) is accepted, that the oxygen content of atmosphere must have been comparable to its present value in the Precambrian. The authors themselves argue, however, that the amount of organic carbon in the ancient sedimentary reservoir as derived from their isotope data is just a measure of the gross amount of photosynthetic oxygen produced, withholding any information as to how this oxygen was partitioned between the principal geochemical reservoirs.

Thus, it is more sensible to reason why the iron formations evolved in such enormous quantities during the Middle Precambrian without considering them as the antithesis of post-Precambrian ironstones (*cf.* Trendall 1965 a). As a logical consequence of the above evidence, this contrast may be solely due to the specific tectonism and structural features of the earth at that time concerning which some geologic and geochemical information has recently been gained (Dimroth 1972, Kröner *et al.* 1973, Young 1973, p. 5, Engel *et al.* 1974, Veizer and Compston 1976, Stanton 1976 b see also Miyashiro 1972, p. 152). Geological correlations between the major basins containing iron formations in the southern hemisphere (Trendall 1968, Dorr 1973, Button 1976 b) indicate that the vast Middle Precambrian iron formations of Africa, Australia, India, and South America formed when these continents were contiguous or that the evolution of the basins were dependent on each other in some other way and that the iron formations were deposited, at least in Minas Gerais, Transvaal, Hamersley, and Nabberu, in intracratonic basins. In Canada, it has been demonstrated that at least certain segments of the famous Labrador Trough have their bases on the continental crust (Dimroth 1972, Fig. 1, Kearey 1977). Also in the Lake Superior area, iron formations are considered to have deposited in an epi- or intracratonic basin (Sims 1976, p. 1103).

The Väyrylänkylä iron formations are very similar to the Animikie iron formations and so is the general geology of both areas (Table 19). This suggests a uniform tectonic history for both basins (*cf.* Van Schmus 1976 b). Paleomagnetic data, however, suggest that North America and Finland were widely separated before the Svecokarelidic orogeny (Neuvonen 1974, see also Donaldson *et al.* 1973). In any case it is, however, important to realize that in the Labrador Trough the basin filling represents two, or possibly three, tectonic cycles, each initiated by an orthoquartzite-limestone sequence (including iron formation in the second and third cycles), and culminating in deposition of shales and of flysch-type deposits in the external zone of the geosyncline, and in intense volcanic activity in the central and internal zones (Dimroth 1970, p. 2717). Indications of cyclicity of the same kind have been found also in Puolanka, where the main iron-formation occurrences belong to the second major sedimentary-tectonic cycle (Table 19). The isotopic compositions of galenas of the Karelidic province proper (South and North Karelia, the Kuopio area) suggest that the Karelian rocks also there may have undergone more than one orogenic episode as discussed by Kouvo and Kulp (1961).

At this stage it is fascinating to discuss the roots of the Svecofennides and Karelides in Finland. The archean (> 2500 Ma) cratonic basement of the Karelides is well-exposed in eastern and northern Finland and comprises mainly granodioritic orthogneisses. The basement of the Svecofennides has, however, still not been found. The most common explanation is that this basement remelted during orogeny (Väyrynen 1954, p. 243, Simonen 1964, p. 201). Simonen (1964) wonders whether the basement was quartz dioritic or granodioritic in composition. Kouvo and Tilton (1966) proved the source area of the greywacke of the Tampere sequence to be at least 2300—2400 Ma old on the basis of detrital zircon age determinations. However, the complications set in when the age of 1900 Ma was obtained for dioritic and quartz-syenitic boulders in the upper part of the same sequence. Moreover, Matisto (1968) explained the meta-arkose of Mauri, which occurs as interbeds in the lower part of the Tampere sequence as having received its material from an orogenic aplite granite. This concept was confirmed by the age of 1900 Ma obtained for a detrital zircon from the meta-arkose. It has therefore been concluded that the Tampere group represents an orogenic greywacke-basalt association with a maximum age of 1900 Ma (Simonen 1971, Kouvo 1976).

In Finland there is no proof of the primary rock crust which acted as the basement for the Svecofennian sedimentation. But on the basis of the comparatively insignificant quantity of basic and ultrabasic rocks within the Svecofennides (Simonen 1962, Table 2), the almost complete lack of ophiolites in them, and the results obtained in the Labrador Trough (Dimroth 1970, 1972) it is logical to suppose that

Table

Comparison of the stratigraphy of the Middle Precambrian rocks in the Salmijärvi

Lake Superior Region (Sims 1976)			Lake Huron Region (Van Schmus 1976 a)
Animikie Group	Baraga Group		
— turbidites, slates and greywackes	— slate, volcanics, iron formations — quartzite	1950 Ma 1)	
	— unconformity —		
	Menominee Group — iron formation — local slate — quartzite	c. 2000 Ma 2)	
unconformity			
Chocolay Group — local slate — dolomite — quartzite — local tillite			
unconformity		2150 Ma 3)	The Huronian supergroup
Lower Precambrian gneisses			Cobalt Group — quartzite, red siltstone — varicoloured siltstone — conglomerate (tillite), arkose, argillite
			major unconformity
			Quirke Lake Group — arkose, sub-greywacke — dolomite, siltstone, greywacke, limestone — conglomerate (tillite)
			Hough Lake Group — subarkose — argillite, siltstone — conglomerate (tillite)
			Elliot Lake Group — argillite, sub-greywacke — subarkose, uraniferous conglomerate — mafic to felsic volcanic rocks
		2400 Ma 3)	major unconformity
			Archean Basement Complex

1) Age of a felsic volcanic rock from the Hemlock Formation by Banks and Van Schmus, (cit. in Sims 1976).

2) Estimate by Sims.

3) Estimate by Van Schmus (1976 a).

4) Age of the Pääkkö Iron Formation (Sakko and Laajoki 1975).

5) Age of the Leipivaara metavolcanics-metadiabase in contact with the Mäntykangas conglomerate (zircon-age determination by M. Sakko in Laajoki 1976 b).

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Basin with the contemporaneous rock sequences in the Great Lake area, North America.

Salmijärvi Basin
(Laajoki 1973, 1976 b)

	(western margin)	(eastern margin)
		Pyssykulju Quartzite Fm — quartzite — conglomerate
	Phyllite-II	——— unconformity ———
	Dolomite-Phyllite Fm	
	— black schist and phyllite	
2080 Ma 4)	— iron formation	
	— quartzite and phyllite	
	——— unconformity? ———	
	— dolomite — local tuffite	
	Quartzites of Leipivaara-Huuhkajavaara	Quartzite-III
	Mäntykangas quartzite	Quartzite-II
≳ 2140 Ma 5)	Mäntykangas conglomerate	Quartzite-I
	——— major unconformity ——— — — —	
	Phyllite-I	
	Feldspar-bearing quartzites and arkosites of Akanvaara, Ahvenvaara, Roninkangas, Aittovaara, and Pärekangas	Arkosite Fm — feldspar-bearing quartzite, arkosite — local metavolcanics — basal arkosite and conglomerate

——— great discordance ———

Prekarelidic Basement Gneiss Complex

the basement was continental rather than oceanic in composition. The usefulness of comparisons between geological features on either side of the Atlantic has long been realized in Finland. The great similarity between the Huronian and the Karelian sequences was noted by Wiik as early as 1876 (p. 42, see also Sederholm 1893, p. 150). Bowes (1975) correlated Archaean rocks in Scotland with the Presvecokarelidic episode. These, as well as the Archaean rocks within the Variscides, in France and the Channel Islands, (Bishop *et al.* 1975, quoted in Bowes 1975) may represent the hinterland of the Svecokarelides. Furthermore, the tectonic style of the Svecofennides differs drastically from that of Alpine orogeny as stressed by Wegmann (1961). According to newer geological maps by the Geological Survey of Finland the Svecofennian metasediments in Central and western Finland are in general flat-lying (e.g. Salli 1964 and 1971). This indicates that the rocks passed through only a relatively mild lateral compression during the Svecokarelidic orogeny.

Trendall (1968) stressed the point that the medial iron formations mark a change in the type of sedimentation in all three of the basins he compared. This is also very obvious in Finland. The Karelian iron formations lie in both Kainuu (Mäkelä 1976, Laajoki 1973) and Kittilä (Paakkola 1971) between craton-derived quartzites and phyllites, shallow-water dolomites etc. and the metapelites of flyschoidic character. They and the associated black schists correspond to the preflysch facies in the sense of Aubouin (1965) and it is worth mentioning that radiolarian jasper is persistent in space (*op. cit.*, p. 117) very much as are the iron formations in Hamersley. Against this background, time-stratigraphic correlations between different iron-formation horizons are of marked importance as has been stressed e.g. by Fryer (1972). For dating the deposition (primary) age of metamorphosed iron formations the whole rock Pb-Pb isochron method seems to be, however, more reliable (Moorbath *et al.* 1973, Sakko and Laajoki 1975) than the Rb-Sr method (see discussion in Goldich 1973, p. 1131). The age of the iron formations in SW Finland, probably correlative with the quartz-banded iron ores in Central Sweden associated with leptites, is not yet known. Poor evidence in favour of their inclusion in a different, probably younger sedimentation-volcanism cycle is given by the comparatively low age of the Kytäjä limestone (1950 ± 100 Ma, Wampler and Kulp 1964). Further, the age of the Svecofennidic metavolcanics (Kouvo 1976) is so much lower (c. 300 Ma) than the age of the intracratonic Karelidic magmatism (Sakko 1971, Sakko and Laajoki 1975) that the Svecofennidic sedimentation and volcanism can be suspected of having taken place in two or more cycles as in the Karelides. The open folding and ubiquitous migmatization and plutonism, however, make this difficult to prove. If the model of the Labrador Geosyncline turns out to fit for the Svecokarelides, then the cycle that contains the iron formations of SW Finland should correspond to the third cycle in the model.

Finally, it is suggested that the huge iron-formation deposition can possibly be attributed to the much higher proportion of continents to seas in the Middle Precambrian than in later eras (*cf.* Kröner 1977). As a consequence, the oceanic current system was comparatively restricted and sedimentation took place in oceanic basins of intracratonic nature (*cf.* Woolnough 1941).

ACKNOWLEDGEMENTS

The authors are indebted to the following staff members of the Geological Survey of Finland: the personnel of the Ore Laboratory for the ore analyses; the members of the Chemistry Department mentioned in the captions of the Tables for complementary chemical analyses; Miss Tuula Paasivirta, M.Sc., and Dr. Jaakko Siivola for microprobe analyses and indentifications; Mr. Pekka Kallio, M.Sc., for X-ray determinations; Dr. Risto Tynni for microfossil studies; Mrs. Rauha Tapanainen, Mrs. Marjo-Riitta Kujala-Tammi and Mrs. Aino Lehti for drawing the figures and Mr. Erkki Halme for many of the photographs. Dr. Ilmari Haapala read the manuscript critically and Dr. Atso Vorma selected parts of it. The suggestions made by them as well as the valuable editorial help given by Dr. Marjatta Okko are acknowledged with sincere thanks. The English of the manuscript was corrected by Mrs. Gillian Häkli.

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Appendix 1

List of the thin sections referred in the study.

Number of thin section	Rock name	Occurrence	Location Drill hole/depth	References
17816	Spessartite-bearing iron-rich black schist	Seppola	335/40.00	Figs. 60 A and B
17832	Spessartite-bearing phyllite	Seppola	337/27.95	Fig. 60 C
17843	Quartz-magnetite-banded rock	Pääkkö	343/57.25	Figs. 42 D, 43 C, 46 and 48 B
17845	Quartz-magnetite-banded rock	Pääkkö	343/78.45	Figs. 42 C and 49
17898	Amphibole-carbonate rock	Iso Vuorijärvi	345/107.80	Fig. 25
17983	Spessartite-bearing phyllite	Seppola	335/40.00	Table 2, Nos. 15 and 16
17999	Siderite rock	Pääkkö	354/73.20	Figs. 18 A + B and 51 B
18003	Retrograde carbonate-biotite-garnet band	Pääkkö	354/63.25	Fig. 61 B
18006	Spessartite(?)-bearing phyllite	Seppola	341/48.80	Fig. 60 D
18008	Spessartite-bearing iron-rich black schist	Seppola	341/45.80	Fig. 53
18054	Biotite-amphibole-garnet interband	Pääkkö	355/72.45	Figs. 29, 52 and 63 A
18431	Quartz-siderite-banded rock	Pääkkö	360/39.60	Figs. 15 A + B, 42 A and 43 A
18433	Iron-silicate-magnetite rock	Iso Vuorijärvi	363/91.95	Table 1, No. 2, Fig. 12
18576	Quartz-magnetite-banded rock	Pääkkö	343/60.55	Figs. 42 B, 55 A and B
18577	Quartz-magnetite-banded rock	Pääkkö	343/82.25	Fig. 9 A
19074	Siderite rock	Pääkkö	360/51.65	Table 2, No. 17 Figs. 16 and 43 B
18685	Magnetite-bearing siderite rock	Pääkkö	360/51.40	Fig. 50
18686	Phosphorite interband	Pääkkö	360/51.50	Figs. 35, 36, 37, and 54
19080	Amphibole-biotite-magnetite-chert mesoband	Pääkkö	360/51.25	Fig. 11
19081	Amphibole-rich quartz-magnetite-banded rock	Pääkkö	360/63.70	Figs. 45, 47 A and B
19082	Spessartite-bearing iron-rich black schist	Seppola	337/80.35	Fig. 28 A and B
19083	Mn-carbonate-rich phyllite	Seppola	337/80.45	Figs. 23 B and 24 A-C
19084	Quartz-Mn-siderite-banded rock	Seppola	337/77.00	Table 2, No. 21
19085	Spessartite-bearing iron-rich black schist	Seppola	337/94.00	Figs. 62 A and B
19314	Biotite-garnet-amphibole interband	Pääkkö	359/57.35	Fig. 32

Number of thin section	Rock name	Occurrence	Location Drill hole/depth	References
19648	Chert mesoband	Pääkkö	360/70.10	Table 1, No. 1, Figs. 8 and 48 A
19649	Quartz-magnetite-banded rock	Pääkkö	343/83.72	Table 1, No. 3, Fig. 10
19650	Amphibole-magnetite-chert mesoband + chert mesoband + phosphorite interband	Pääkkö	344/155.45	Table 1, No. 4, Figs. 9 B, 39 A + B and 44
19651	Amphibole-biotite-garnet interband	Pääkkö	343/63.50	Table 1, No. 6 Table 2, Nos. 3, 6, 8 and 11
19684	Garnet-amphibole-biotite interband	Pääkkö	354/97.00	Fig. 61 A
20357	Amphibole-biotite-magnetite-chert mesoband	Pääkkö	343/81.25	Table 1, No. 5 Table 2, Nos. 1 and 7 Fig. 57
20358	Spessartite-bearing iron-rich phyllite	Seppola	337/76.50	Table 2, Nos. 5, 10 and 14, Figs. 27 and 63 B
20441	Biotite-amphibole-garnet interband	Pääkkö	343/90.75	Table 1, No. 7 Table 2, Nos. 4, 9 and 12, Figs. 30 and 31
21070	Chert mesoband	Pääkkö	359/59.00—60.55 B	Fig. 41
21071	Chert-mesobanded grunerite rock	Pääkkö	359/59.00—60.55 A	Table 2, No. 2
21072	Chert-mesobanded grunerite rock	Pääkkö	359/61.10—62.65	Fig. 13
21156	Garnet-amphibole-siderite-biotite interband	Pääkkö	354/97.00	Table 2, Nos. 13 and 18
21181	Quartz-magnetite-banded rock	Pääkkö	343/88.60	Table 2, No. 22 Fig. 56
21182	Phosphorite-banded siderite rock	Pääkkö	360/52.40	Figs. 33 B and 34
21183	Amphibole- and magnetite-bearing siderite rock	Pääkkö	360/51.40	Fig. 20
21184	Siderite rock	Pääkkö	354/73.20	Table 2, No. 20, Figs. 17, 19 B, 51 A and 58
21201	Quartz-magnetite-banded rock	Pääkkö	343/60.55	Table 2, No. 23
21202	Amphibole + biotite-rich mesoband	Pääkkö	360/49.35	Figs. 40 and 59
21203	Phosphorite interband	Pääkkö	360/51.50	Table 2, No. 19 Figs. 38 A-C
21211	Magnetite-spotted amphibole- and chert-bearing siderite rock	Pääkkö	343/87.87	Fig. 62 C
21212	Siderite rock	Pääkkö	343/86.28	Figs. 19 A and 43 D
21294	Mn-carbonate-rich phyllite	Seppola	337/80.45 B	Figs. 23 A and 61 C
21295	Mn-carbonate-chert mesoband	Seppola	337/80.45 C	Figs. 21 and 22