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The Fiskenæsset complex, West Greenland  
Part IV

Chemistry of sulphide minerals

*by*

*F. C. Bishop, J. V. Smith and B. F. Windley*

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## Abstract

The textures and chemistry of sulphides were investigated in 63 rocks from the Fiskenæsset complex, West Greenland, by light microscopy, electron microprobe, and X-ray diffraction techniques. Sulphides are found disseminated throughout the layered sequence, being especially concentrated in the ultramafic zone and in ultramafic lenses in the higher zones. Sulphide phases are restricted to the Cu-Fe-Ni-Co-S system – pentlandite, pyrrhotite, chalcopyrite, pyrite, millerite, heazlewoodite, cobalt-pentlandite, polydymite, godlevskite, violarite, cubanite, and digenite.

Most sulphides occur interstitially as rounded globules often containing significant quantities of oxides. Some aggregates occur as inclusions in silicate or chromite, as ragged stringers between other phases, or with scalloped intercumulus textures. Subsolidus reactions are common; fine-scale intergrowths of pentlandite-pyrite, chalcopyrite-pyrite, chalcopyrite-cubanite, pentlandite-heazlewoodite and pentlandite-godlevskite-polydymite occur. Euhedral pyrite is found in the anorthosites and chromitites. Pyrite-millerite appears to be an equilibrium pair in some rocks. Weathering affects the major phases in the following order: pyrrhotite > chalcopyrite > pentlandite.

Temperature stability limits for several of the phases, equilibrium phase assemblages, and textures of the sulphide intergrowths, all indicate that the sulphides have re-equilibrated to low temperatures. The common pyrite-pentlandite assemblage places a 230°C upper limit, and the digenite composition places a 125°C lower limit, on the temperature of equilibration.

A marked sulphide stratigraphy which correlates with silicate chemistry is present in the Fiskenæsset complex. There is a systematic progression from Ni,Fe-rich sulphides in the lower ultramafic and mafic zones to Cu,S-rich assemblages in the upper, more felsic zones. Sulphide fractionation exhibiting some of these features is also found in other layered igneous intrusions. The similarity of the Fiskenæsset sulphides with those in other intrusions, coupled with the unlikely possibility of derivation of the sulphides by assimilation or metasomatism, implies that the sulphides were derived from the original magma.

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## INTRODUCTION

The Fiskenæsset complex, West Greenland, is an extraordinary example of an Archaean anorthositic complex in a high-grade gneiss terrain. It comprises a layered, differentiated rock sequence passing upwards from gabbros and ultramafic rocks through gabbros and leucogabbros to predominant chromite-layered calcic anorthosites and minor gabbros (Myers, 1976). The range in rock types, combined with remarkable exposure, makes this an excellent example for study of differentiation of this type of Archaean intrusion. The igneous crystallization pattern is complicated by a metamorphic overprint of amphibolite-granulite grade. The metamorphism has been dated by Pb/Pb at  $2850 \pm 100$  m.y. (Black *et al.*, 1973) and Rb/Sr at  $2810 \pm 70$  m.y. (Alexander *et al.*, 1973).

This paper gives results from the fourth in a series of investigations on the petrology and mineralogy of the Fiskenæsset complex. Particular emphasis has been placed on rocks from Qeqertarssuatsiaq island which is situated in the western part of the Fiskenæsset region. Part I of this series outlined the stratigraphy, structure, and whole rock chemistry of the intrusion (Windley *et al.*, 1973). Part II reported the results of a reconnaissance survey of the chemistry, with particular emphasis on the silicate phases (Windley & Smith, 1974). Part III gave details of the oxide chemistry and further information on the silicates (Steele *et al.*, 1977). This final report is restricted to the chemistry and mineralogy of the sulphide phases.

Sixty-three specimens represent each of the main stratigraphic zones as recognized at Qeqertarssuatsiaq:

Top	Upper Anorthosite (UA) Chromitite (Ch) Lower Anorthosite (LA) Upper Leucogabbro (ULG) Gabbro (G) Lower Leucogabbro (LLG)
Bottom	Ultramafic rocks (including dunites, peridotites and pyroxenites) (UM)

The specimens are listed in table 1 in approximate stratigraphic order. For specimens collected from a single area, relative stratigraphic positions in table 1 are correct. Because of severe tectonic stretching and thinning, it is difficult to stratigraphically correlate specimens from different locations within the same zone. It is,

Table 1. Mineralogy and rock types

Zone	Specimen Number	Rock Type	Pn	Po	Co	Py	Mi	H <sub>z</sub>	CP	Pd	Gd	Vi	Cb	Di
UA	81244	An												
	81235	An												
	81233	An		X		X								
Ch b	132042	Ch				X								
	132041	Ch			X	X								
	132040	Ch			X	X								
	132039	Ch												
	132038	Ch			X	X								
	132037	Ch			X	X								
	132036	Ch				X								
	132035	Ch				X								
	132034	Ch	X											
	Ch a	132055	Ch				X							
132054		Ch												
132053		Ch				X								
132052		Ch			X	X	X*							
132051		Ch				X	X*							
132050		Ch				X								
132049		Ch				X	X							
132048		Ch			†	X	X	X						
132047	Ch				X									
LA	132018	An												
	132016	An												
ULG	86929	LG			X	X								
	86931	LG			X	X								
	86924	LG			X	X								
	132015	G												
G	132014	LG												
	132009	G												
	132008	G												
	132006	G				X								
LLG	84499	G		X	X	X								
	84498	G	X	X		X								
UM	84495	LG												
	84493	LG												
UM	122350	UM	X	X										
	122349	UM	X	X	X									
	86935	UM	X		X	X						X		
	86927	UM	X	X	X	X								
	84490	UM	X	X	X									
	84488	UM	X	X	X									
	84487	UM	X	X	X									
	84485	UM	X	X	X									
	132007	UM	X	X	X								X	
	84472	UM	X	X	X	X								
	84471	UM	X					X	X					X
	84470	UM												
	84469	UM	X	X	X									
	84468	UM	X							X	X			
	84467	UM	X					X	X					
	84466	UM	X	X	X									
	84465	UM	X		X	X								
81216	UM	X				X*								
81209	UM	X	X	X							X	X		
-	132002	Am	X	X	X									
Ultramafic interlayers														
UA	81234	UM	X	X	X	X								
LA	84484	UM	X	X	X	X								
LA	74854	UM	X	X	X	X								
ULG	132012	UM	X	X	X									
G	84497	UM	X	X	X	X								
G	84496	UM	X		X	X								
LLG	84423	UM	X	X	X	X								
LLG	84494	UM	X	X	X	X								
LLG	84492	UM	X	X	X	X								

\*fine-grained Ni-sulphide, probably millerite; Cb cubanite, Cp chalcopyrite, CP Co-pentlandite, Di digenite, Gd godlevskite, Hz heazlewoodite, Mi millerite, Pd polydymite, Pn pentlandite, Po Pyrrhotite, Py pyrite, Vi violarite.

however, relatively easy to determine the zone in which an individual specimen is located, and this information is given in the table. Exact geographical locations are given on the map in figure 1 of Part III in this series (Steele *et al.*, 1977). Also, the appendix of Part III contains petrographic descriptions of all specimens.

The Fiskenæsset complex was intruded into a pile of volcanic rocks which were subsequently metamorphosed (Escher & Myers, 1975). Some of the resulting amphibolites are extremely rich in sulphides; specimen 132002, collected from the amphibolite, has been included in this study.

An important feature of the Fiskenæsset complex is the existence of small ultramafic lenses within the gabbros and anorthosites. These small ultramafic interlayers are in the order of 30 cm – 1 m thick, and usually pinch out within a few tens of metres. Several specimens from these interlayers, which are important in understanding the sulphide chemistry, are included in a separate section of table 1.

Table 1 also gives the sulphide mineralogy of each of the specimens. A similar table of sulphides in Part III (table 1) was of a preliminary nature and is entirely superseded by table 1 of the present report. The twelve sulphide phases found in the present investigation are restricted to the Ni-Co-Fe-Cu-S system, and are especially concentrated in the ultramafic zone and the ultramafic interlayers.

## ANALYTICAL PROCEDURE AND DATA

Polished sections were made of all specimens and studied by incident-light microscopy. Electron microprobe analyses were made of selected grains in most specimens to check optical identifications, obtain exact compositions, and check for homogeneity of phases. Microprobe analyses were obtained using both wavelength- and energy-dispersive techniques on an ARL-EMX electron probe.

For wavelength-dispersive analyses analytical conditions were: accelerating voltage 15 kV, beam current 1  $\mu$ A, analysis time 10 sec, 1–2  $\mu$ m beam diameter and detection level 0.02 wt.%. For energy dispersive analyses, analytical conditions were: accelerating voltage 15 kV, beam current 0.1  $\mu$ A, live time 2 minutes, beam diameter 1–2  $\mu$ m, and detection level near 0.2 wt.%. Standards used for all analyses were synthetic chalcopyrite (Cu, Fe, S), synthetic Fe-Ni sulphide (Ni), Cr-metal (Cr), synthetic alabandite (Mn), Co-metal (Co) and synthetic sphalerite (Zn). The synthetic sulphide standards were kindly supplied by G. K. Czamanske. Matrix corrections were made using the computer program EMPADR VII (Rucklidge, 1967) for wavelength-dispersive analyses and the Reed & Ware (1975) procedure for energy-dispersive analyses. Comparison of analyses of the same grain made by each method showed no systematic differences except for the more sensitive detection level when crystal spectrometers were employed.

Table 2 lists averages of representative analyses for sulphides in the specimens listed in tabel 1. Because analyses done by the two methods outlined above are comparable, the particular analytical technique used in each case is not indicated.

Table 2. Average electron microprobe analyses of sulphides

	81233				132041		132038			132037			132055		132053		132052			
	Po	Py	Py	Py	Cp	Py	Cp	Py	Py	Cp	Py	Py	Py	Py	Py	Py	Cp	Py	Py	
S	39.0	39.9	53.8	53.6	34.5	53.2	34.9	53.2	53.4	35.5	53.6	52.7	53.5	53.7	53.0	52.8	53.3	34.6	53.3	53.3
Cr	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Mn	0.14	nd																		
Fe	58.5	58.5	46.1	46.7	30.0	46.4	31.4	47.3	45.3	31.5	45.5	47.8	46.2	47.0	46.6	46.5	47.3	30.5	47.2	47.0
Co	nd	0.62	nd	nd	0.44	0.22	0.39	0.42	0.48	nd	nd	0.28	nd							
Ni	0.36	1.03	1.52	0.23	nd	nd	nd	0.29	0.86	0.29	nd	nd	0.21	nd						
Cu	nd	nd	nd	nd	33.9	nd	33.7	nd	nd	32.6	nd	nd	0.15	nd	nd	nd	nd	33.9	nd	nd
Sum	98.00	99.43	101.42	100.53	98.40	99.60	100.00	100.79	100.18	98.89	99.10	100.94	100.32	101.09	100.02	99.78	100.60	99.00	100.78	100.30
	132051				132050		132049			132047		86929			84484				74854	
S	53.1	53.0	53.1	33.7	35.7	53.1	52.8	34.7	53.1	52.8	36.6	35.1	34.9	33.6	33.5	33.2	53.0	53.5	37.4	38.6
Cr	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	0.03	0.01
Mn	nd	nd	0.19	0.24	0.25	0.32	nd	-	-											
Fe	46.2	45.7	46.3	2.49	4.15	46.3	46.3	31.7	41.3	42.5	63.3	30.7	31.0	30.1	30.8	31.2	44.6	45.2	62.1	61.6
Co	nd	1.48	nd	nd	1.10	0.32	0.35	nd	nd	0.27	nd	nd	nd	1.65	1.09	0.52	1.94	1.33	0.02	nd
Ni	nd	nd	0.57	61.0	58.8	0.44	nd	nd	6.61	4.86	nd	nd	0.18	35.2	36.3	35.7	0.22	nd	0.42	0.53
Cu	nd	33.3	0.20	nd	nd	34.4	33.6	nd	nd	nd	0.58	nd	nd	nd						
Sum	99.30	100.18	100.16	97.43	100.00	100.48	99.45	99.70	101.21	100.43	99.90	100.20	99.68	100.55	101.69	100.62	100.34	100.03	99.97	100.74
	74854						86931				86924				132012					
S	38.9	38.6	34.7	34.9	35.2	33.6	33.6	34.6	53.8	54.2	35.0	35.1	35.4	54.6	54.3	54.6	37.3	35.1	34.9	33.2
Cr	0.01	0.02	0.04	nd	0.05	0.04	0.01	0.05	0.04	0.01	nd	-	nd	0.01	0.01	nd	0.01	-	-	-
Mn	-	-	-	-	-	-	-	-	-	-	nd	nd	0.03	0.01	nd	nd	nd	nd	nd	0.14
Fe	61.9	62.4	30.5	31.4	31.3	30.9	30.8	30.5	32.1	45.8	45.5	31.2	28.9	30.5	45.5	45.5	45.4	62.7	30.6	30.6
Co	0.07	0.06	0.07	nd	nd	1.09	0.97	0.88	0.94	2.21	1.04	nd	0.03	0.05	0.21	0.08	0.57	nd	nd	0.23
Ni	0.34	0.33	0.12	0.13	0.09	36.0	35.0	35.7	33.7	0.06	0.44	nd	0.19	0.13	1.59	0.10	0.13	nd	nd	0.29
Cu	nd	0.01	33.7	33.8	34.0	nd	0.03	0.17	0.54	0.14	0.13	33.5	34.6	33.8	nd	nd	nd	nd	33.9	33.4
Sum	101.22	101.42	99.13	100.23	100.64	101.63	100.41	100.90	101.92	102.02	101.31	99.70	98.85	99.90	101.91	99.88	100.71	100.00	99.60	99.42
	132012		132006		84499						84498				84497				84423	
S	32.9	54.8	38.2	40.2	34.8	34.8	36.0	53.3	53.9	53.9	53.6	38.5	53.0	37.0	34.4	32.9	33.0	32.9	51.3	36.4
Cr	-	-	0.02	nd	0.01	0.01	0.02	0.02	nd	nd	nd	-	-	-	-	-	-	-	-	-
Mn	nd	nd	-	-	-	-	-	-	-	-	-	0.37	0.30	nd	nd	nd	0.18	nd	nd	0.03
Fe	30.4	44.3	60.0	60.7	30.7	31.4	31.8	47.7	46.6	47.4	47.3	56.4	44.1	62.4	30.1	29.5	29.9	32.9	46.1	62.0
Co	2.22	0.43	0.08	0.21	nd	0.04	nd	nd	nd	nd	nd	0.31	nd	nd	nd	0.48	0.45	0.49	2.31	0.04
Ni	34.6	2.09	nd	0.35	0.02	0.11	nd	nd	0.33	nd	nd	2.89	3.43	nd	nd	37.3	37.2	32.9	0.25	0.13
Cu	nd	nd	nd	0.03	33.2	32.9	32.2	nd	nd	nd	0.03	nd	nd	nd	33.8	nd	nd	nd	nd	0.07
Sum	100.12	101.62	98.30	101.49	98.73	99.26	100.02	101.02	100.83	101.30	100.93	98.47	100.83	99.40	98.30	100.18	100.73	99.19	99.96	98.67

Table 2 cont.

84423											84494							84492				
Po	Cp	Cp	Pn	Pn	Pn	Pn	Pn	Pn	Pn	Pn	Po	Po	Po	Po	Cp	Pn	Py	Po	Po	Cp	Cp	
S	37.2	34.5	35.2	32.9	33.3	33.1	34.4	33.2	33.2	33.1	36.5	36.4	36.6	36.5	34.4	32.7	51.0	36.5	36.7	35.2	35.2	
Cr	-	-	-	-	-	-	-	-	-	0.02	nd	-	-	-	-	-	-	-	-	-	-	
Mn	nd	nd	nd	nd	nd	nd	nd	nd	nd	0.02	nd	nd	nd	nd	nd	nd	0.13	nd	nd	nd		
Fe	62.7	29.4	29.9	33.7	34.7	35.4	33.4	31.4	31.1	33.1	62.5	63.3	62.0	62.8	30.4	31.9	47.7	63.3	62.8	30.7	30.9	
Co	0.30	nd	nd	0.89	1.19	1.45	1.43	1.30	1.34	1.85	0.32	nd	0.29	nd	0.43	1.39	1.19	nd	nd	nd	nd	
Ni	nd	0.30	0.31	30.7	30.2	29.4	29.8	33.4	33.9	32.5	nd	nd	nd	0.28	nd	34.3	nd	nd	nd	nd	0.20	
Cu	nd	33.9	34.4	nd	nd	0.40	nd	nd	nd	0.20	nd	nd	nd	nd	33.5	nd	nd	nd	33.9	34.1		
Sum	100.20	98.10	99.81	98.19	99.39	99.75	99.03	99.30	99.54	100.79	99.32	99.70	98.89	99.58	98.73	100.29	99.89	99.93	99.50	99.80	100.40	
84492				122350				122349					86935				86927				84490	
Pn	Pn	Po	Pn	Po	Po	Po	Cp	Cp	Pn	Pn	Pn	Pn	Pn	Pn	Py	Po	Cp	Cp	Pn	Py	Po	
S	33.3	33.3	37.1	33.0	36.4	37.1	37.3	34.9	34.7	33.1	33.3	32.0	33.2	32.6	51.3	36.7	34.9	34.7	33.0	49.0	36.2	
Cr	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	0.03	0.04	nd	0.02	nd	-	
Mn	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	
Fe	36.1	35.8	62.8	38.9	62.9	62.9	62.7	30.3	30.9	32.8	32.2	32.9	34.6	32.5	44.7	63.9	30.1	31.7	30.8	46.4	63.3	
Co	1.45	1.34	nd	1.08	0.32	nd	0.46	0.31	nd	1.63	1.93	1.54	1.41	0.21	1.46	0.03	nd	nd	0.71	1.33	nd	
Ni	29.0	29.6	nd	26.5	nd	nd	nd	0.26	nd	32.5	33.2	32.0	31.7	34.4	nd	nd	0.08	nd	35.3	0.21	nd	
Cu	nd	nd	nd	nd	nd	nd	nd	34.0	33.4	nd	nd	nd	nd	nd	0.36	nd	31.9	31.6	nd	nd	nd	
Sum	99.85	100.04	99.90	99.48	99.62	100.00	100.46	99.77	99.00	100.03	100.63	98.44	100.91	99.71	97.82	100.66	97.02	98.00	99.83	96.94	99.50	
84490				84488				84487					84485				84472					
Cp	Pn	Po	Cp	Cp	Pn	Pn	Pn	Po	Po	Cp	Cp	Cp	Pn	Pn	Po	Cp	Pn	Pn	Po	Po		
S	34.9	33.0	36.1	35.0	34.8	32.5	32.4	32.7	36.3	36.3	35.3	34.8	34.7	33.1	32.8	36.3	34.7	33.5	32.9	36.5	37.0	
Cr	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	
Mn	nd	nd	nd	nd	nd	nd	nd	nd	nd	0.13	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	
Fe	31.0	32.6	63.3	31.2	31.6	30.1	33.4	30.5	63.7	64.0	30.9	31.1	31.3	32.9	34.5	63.3	30.8	34.6	34.4	61.3	62.7	
Co	nd	1.60	nd	nd	nd	1.30	1.34	1.28	nd	nd	nd	nd	nd	1.65	1.54	nd	nd	1.73	2.23	0.43	0.45	
Ni	nd	33.1	nd	0.21	nd	35.6	32.1	35.5	nd	nd	0.30	0.24	nd	33.4	31.0	nd	nd	31.7	30.2	0.60	nd	
Cu	33.6	nd	nd	33.0	33.5	nd	nd	nd	nd	nd	33.3	32.8	34.3	nd	nd	0.28	34.6	nd	nd	nd	nd	
Sum	99.50	100.30	99.40	99.41	99.90	99.50	99.24	99.98	100.00	100.43	99.80	98.94	100.30	101.05	99.84	99.88	100.10	101.53	99.73	98.83	100.15	
84472										84471					84469							
Po	Po	Cp	Cp	Pn	Pn	Py	Py	Py	Di	Di	Pn	Pn	CP	CP	Hz	Hz	Po	Po	Cp	Pn		
S	36.8	36.8	34.6	34.4	33.5	32.8	52.4	52.9	53.1	22.1	21.2	32.4	32.4	30.9	31.7	26.3	27.7	36.4	36.7	34.1	32.3	
Cr	-	-	-	-	nd	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-		
Mn	nd	nd	nd	nd	0.02	nd	nd	nd	nd	nd	nd	nd										
Fe	63.0	63.1	31.0	30.6	35.5	34.8	43.3	46.4	46.1	3.07	3.16	30.9	32.8	7.14	14.7	3.25	4.13	61.3	62.0	30.8	31.1	
Co	nd	0.38	0.42	0.36	0.65	1.10	3.41	1.63	2.01	0.26	nd	3.26	2.25	24.1	33.0	0.27	0.29	nd	nd	nd	0.85	
Ni	nd	nd	0.27	nd	31.0	31.7	nd	nd	0.21	1.53	1.11	34.0	32.4	36.0	18.8	70.5	68.5	0.29	0.43	0.44	35.0	
Cu	nd	nd	33.5	33.7	0.20	nd	nd	nd	nd	72.6	72.5	nd	nd	nd	nd	nd	nd	1.22	0.50	32.3	nd	
Sum	99.80	100.28	99.79	98.96	100.87	100.40	99.11	100.93	101.42	99.56	97.97	100.56	99.85	98.14	98.20	100.32	100.62	99.21	99.63	97.64	99.25	

Table 2 cont.

	84468					84467													84466				
	Pn	Gd	Gd	Pd	Pd	Pn	Pn	Pn	Pn	Pn	Pn	Pn	Pn	CP	CP	CP	CP	CP	Hz	Hz	Hz	Po	
S	32.5	30.7	30.6	39.8	39.0	32.6	32.6	32.2	32.9	32.8	32.8	32.4	32.6	31.8	32.7	31.6	32.5	26.0	28.1	27.8		37.0	
Cr	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Mn	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd
Fe	30.5	1.38	1.45	3.54	3.63	32.4	29.3	30.6	30.6	31.0	30.3	31.6	27.0	15.1	20.8	19.8	22.0	3.64	3.14	2.69		62.9	
Co	5.25	nd	nd	0.44	nd	4.13	7.57	6.24	4.90	4.68	5.80	4.40	10.6	33.0	23.4	22.0	21.4	0.34	nd	nd		nd	
Ni	31.9	65.6	67.0	54.4	55.5	32.2	31.1	30.4	32.4	31.9	31.2	31.6	29.1	17.8	23.0	23.8	24.4	69.7	68.8	70.8		0.35	
Cu	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd		nd	
Sum	100.15	97.68	99.95	98.18	98.13	101.33	100.57	99.44	100.80	100.38	100.10	100.00	99.33	97.70	99.90	97.20	100.30	99.68	100.04	101.29		100.25	

	84466				84465			81216			81209										
	Po	Pn	Pn	Cp	Cp	Cp	Pn	Po	Po	Po	Cp	Cp	Cp	Cb	Cb	Cb	Cb	Cb	Pn	Vi	
S	37.5	33.2	33.6	34.7	35.4	35.0	35.1	35.3	33.7	38.6	38.7	38.3	34.9	35.1	35.1	35.3	35.7	35.6	35.8	33.7	42.3
Cr	nd	-	-	-	-	-	-	0.04	-	-	-	0.01	-	-	-	-	-	-	-	-	-
Mn	nd	nd	nd	nd	nd	nd	nd	0.04	nd	nd	0.01	-	0.01	nd	0.01	0.01	nd	nd	0.01	nd	nd
Fe	62.5	37.2	37.5	30.5	30.1	29.9	30.0	30.5	37.4	61.4	61.2	62.0	30.7	30.9	30.8	41.6	41.2	41.3	40.9	33.4	31.5
Co	nd	1.66	1.19	0.26	nd	nd	nd	0.05	0.74	0.11	0.14	nd	0.07	0.09	0.07	0.14	0.12	0.05	0.08	2.08	2.20
Ni	nd	27.9	27.8	nd	nd	nd	nd	0.16	28.7	0.28	0.27	0.21	0.01	nd	0.07	0.03	0.06	nd	0.04	31.5	23.6
Cu	nd	nd	nd	34.1	34.3	35.1	34.9	33.5	nd	0.06	0.09	nd	34.2	34.0	34.0	22.9	22.6	23.3	23.9	nd	0.05
Sum	100.00	99.96	100.09	99.56	99.80	100.00	100.00	99.59	100.40	100.45	100.41	100.52	99.89	100.09	100.05	99.98	99.68	100.26	100.73	100.68	99.65

	81209		132002		
	Vi	Vi	Po	Cp	Pn
S	42.2	42.3	39.2	35.3	35.3
Cr	-	-	-	-	-
Mn	nd	nd	0.12	nd	nd
Fe	35.3	31.8	59.8	30.9	37.1
Co	2.23	2.22	nd	nd	6.15
Ni	18.4	22.8	0.56	nd	20.1
Cu	nd	nd	nd	33.2	1.16
Sum	98.13	99.12	99.68	99.40	99.81

Each entry represents the average of from one to fifteen individual analyses. If a significant variation in composition occurs within a specimen, several analyses which span the range in composition are given.

nd: not detected

- : not analyzed

Zn was analyzed in most sulphide specimens, but never found above the minimum level of detection.

Low totals listed for some phases are the result of problems associated with microprobe analysis of minerals with extremely fine grain sizes. Most phases exhibited a high degree of homogeneity (within analytical error) throughout an entire specimen, but notable exceptions are pentlandite grains in 84468 and 84471. Both specimens contain pentlandite exhibiting a wide range in cobalt concentration. Evidence is presented later that this is a result of late-stage events involving the formation of veins in the rocks.

Where necessary, X-ray methods were used to check phase identification. Gandolfi photographs confirmed the identification of many of the minor sulphide phases although fine grain size precluded positive identification of godlevskite, digenite, and heazlewoodite. These identifications are thus based solely on probe and optical data. In addition, identification of the  $\text{Fe}_{1-x}\text{S}$  phases, whether monoclinic pyrrhotite, hexagonal pyrrhotite, or troilite, was not checked by X-ray methods. The complexity of low temperature phase relations and the apparent inability of pyrrhotite to maintain a metamorphic or igneous (higher-temperature) signature (Craig & Scott, 1974) rendered this information of secondary importance to the present investigation. Considering the apparently prolonged annealing of the Fiskenæsset complex at low temperatures, evidence of which is given below, this may make an interesting study in its own right.

For a few specimens, modal analyses were made of the volume percent of sulphide by measuring the area occupied by sulphides and dividing by the total area of the thin section. Other abundance data noted in the text, all qualitative, were obtained by visual estimates. Possible errors attendant in this kind of estimation should be kept in mind.

## MINERALOGY AND TEXTURES

Sulphide minerals in the Fiskenæsset complex occur interstitially, as inclusions in silicates and chromite, and occasionally in small microveins. Sulphides are found in every zone of the intrusion, although the tenor and mineralogy vary significantly and systematically. Although twelve sulphide phases have been identified, no sulphosalts have yet been found in the specimens studied. Special efforts to find platinum minerals (reported by Ghisler, 1976) in the chromitite lenses, and iron-nickel alloys in the rare serpentized specimens, were unsuccessful. Molybdenite was found by Myers (1974), but not in the present specimens.

Evidence summarized in this series of reports indicates that the minerals in the Fiskenæsset complex have undergone retrograde re-equilibration after the granulite-amphibolite metamorphic event. The retrograde metamorphism has particularly affected the sulphides, causing some to re-equilibrate to temperatures below 230°C. Metamorphic effects can also be noted in the fabrics and textures of the sulphide phases.

### UM zone and UM interlayers

The ultramafic zone exhibits the greatest variety in sulphide mineralogy; indeed, all twelve sulphides listed in table 1 occur in this zone. In addition, the concentration of sulphides is, in general, higher in the ultramafic zone than in the overlying gabbros and anorthosites. The concentration of sulphides is not uniform from sample to sample, however, ranging from zero in 84470 to more than 0.2% by volume in 81209. Grain size is also quite variable; although most sulphide globules are in the order of 100  $\mu\text{m}$  or less in diameter, some globules in 81209 approach 2 mm in diameter. Pyrrhotite, pentlandite, and chalcopyrite in that order are the most abundant sulphides in the ultramafic rocks. Pyrite occasionally occurs in the ultramafic rocks, but is particularly characteristic of the overlying zones (table 1).

Restricting consideration to the Fe-Ni-S system, recognized assemblages are: pentlandite-pyrrhotite, pentlandite-pyrite, pentlandite-pyrrhotite-pyrite, pentlandite-heazlewoodite-cobaltian pentlandite, pentlandite-godlevskite-polydymite, pentlandite-millerite, pentlandite-violarite, and pentlandite-pyrite-violarite (pentlandite containing more than 10 wt.% Co is designated as cobaltian pentlandite in this paper). Lack of four-phase assemblages in this system is consistent with equilibrium, but textural evidence indicates that some assemblages may represent late-stage alteration of earlier-formed minerals that may or may not have occurred under equilibrium conditions.

Pentlandite and pyrrhotite occur as discrete grains and as intergrowths with each other and chalcopyrite. The sulphides in specimens that do not show evidence of late-stage alteration and oxidation typically have sharp, rounded contacts with neighbouring silicates and oxides (plate 1 A,B,C), although a few grains with irregular boundaries have been noted. Quite commonly, the oxides and sulphides are associated together in the rounded globules (plate 1 B,C). When this occurs, the sulphides are segregated at one end of the globule and the oxides are at the other, i.e., there appears to have been polarization of the anionic species. This occurs even though the sulphides may exhibit complex intergrowth textures with each other, and many of the oxides may show complex intergrowth textures, but at opposite ends of the same opaque globule. In these globules, either magnetite or Cr-magnetite is the dominant oxide (see part III of this series for details), but exsolved spinel *sensu stricto* forms an interface with the sulphide portion of the globule. Apparently the sulphide phases either form a suitable substrate for exsolution of spinel or otherwise help to catalyze the exsolution reaction. Unlike this apparent affinity between spinel and sulphide, no specific sulphide phase shows a preference for the oxide-sulphide boundary. Graphic evidence of this is found in specimen 84423, in which pyrrhotite, pentlandite and chalcopyrite can all be seen in contact with spinel in one globule (plate 1 A). The nearly complete segregation of oxides from sulphides is taken as evidence of annealing since oxide-sulphide intergrowths in quickly cooled lava flows and pillow lavas show little evidence for

segregation (Desborough *et al.*, 1968; Skinner & Peck, 1969; Czamanske & Moore, 1977).

Euhedral sulphide grains are not found in the ultramafic zone. Besides the globules mentioned above (with or without oxides), sulphides occur rarely as inclusions in silicates or chromite, as ragged stringers between other phases (often with oxides as well as sulphides), or with scalloped, intercumulus-appearing textures (e.g. as in the MuskoX intrusion; Chamberlain, 1967). Liebenberg (1970) noted in the Bushveld igneous complex that sulphide grains less than 50  $\mu\text{m}$  in diameter were too small to attain a scalloped texture. There appears to be some size control in the Fiskenæsset complex as well, but only a few of the larger sulphide globules have acquired this texture (plate 1 A). If scalloped textures were commonly formed in the initial crystallization of the Fiskenæsset complex, they have been substantially altered in most specimens during the subsequent history of the intrusion.

Subsolidus reactions played an important role in the formation of the present-day mineralogy. In several specimens pentlandite and pyrite with or without pyrrhotite, are intergrown in a fine-grained texture (plate 2 E). Because of solid solution in Mss (a  $(\text{Fe,Ni})_{1-x}\text{S}$  solid solution which spans the Fe-Ni-S system between 300° and almost 1000°C), pentlandite and pyrite cannot coexist in equilibrium above 230°C (Misra & Fleet, 1973; Craig, 1973). This texture is on a much finer scale than the unmixing between say, chalcopyrite and cubanite (compare plates 2 B and 2 E), and is further evidence for extended annealing at low temperatures.

The four nickel-rich sulphides – heazlewoodite, godlevskite, millerite and polydymite – occur as lamellae in pentlandite (plate 2 D) or as complex intergrowths within a pentlandite host (plate 2 A). Some of the cobalt-rich pentlandite is also intergrown with pentlandite of a more moderate cobalt content. Because the optical properties of these two phases are similar (Burns & Vaughan, 1970), identification could only be made with the microprobe. Specimen 84467 contains both globular intergrowths and veinlets of pentlandite and cobaltian pentlandite. The width of these veinlets is only a few micrometres, and many intersect sulphide globules.

Cubanite occurs as lamellae in host chalcopyrite. Plate 2 B shows a typical cubanite lamella extending across the entire chalcopyrite grain. Chalcopyrite also exhibits a lamellar intergrowth with pyrite in some specimens (plate 2 F).

Some sections in addition to 84467 are cut by small veinlets. In general compositions of minerals in the veinlets are similar to those within the globules although relative proportions of the minerals may change. The veinlet in plate 1 B contains po, cp, and py, whereas the globule also contains pn.

An iron-rich violarite containing up to 35.3 wt.% Fe is associated with pentlandite in a few sections. Most violarite appears to have invaded pentlandite along ragged irregular margins. It probably formed as a late-stage alteration of pentlan-

dite, as for other associations of Fe-rich violarite and pentlandite (Misra & Fleet, 1974; Craig & Higgins, 1975). Because violarite is secondary, the pentlandite-violarite association is not necessarily in equilibrium, and does not preclude a low-temperature tie-line between pyrite and millerite. Violarite in 81209 is not as irregular, however, and appears in a lamellar intergrowth with pentlandite (plate 2 C). Although this could represent an equilibrium solution, it might also represent alteration of pentlandite along parallel cleavage cracks. The latter suggestion is preferred because experimental evidence for a narrowing of the range of sulphur content in pentlandite with decreasing temperature is equivocal (Misra & Fleet, 1973, fig. 2). In addition, if exsolution is possible, the associated phase should be pyrite, as in many other specimens studied here.

The one occurrence of low digenite also has a ragged, irregular appearance (plate 3 C), and the digenite grains do not exceed 10  $\mu\text{m}$  diameter. It is most closely associated with pentlandite, but may represent replacement of an earlier Cu-sulphide.

A few specimens exhibit clear-cut examples of weathering (plate 3 A,B). Plate 3 B shows the general pattern of sulphide weathering in the Fiskenæsset complex. Pyrrhotite is oxidized most, and is being replaced by an Fe-oxide (probably goethite). Chalcopyrite is also weathered, and grains in 132007 show oxidation proceeding along a trellis pattern which is probably crystallographically controlled. Pentlandite is apparently not affected by the weathering. The different resistance to weathering of pyrrhotite and pentlandite is exactly the opposite of that observed by Chamberlain (1967), but no explanation is offered.

The sulphide mineralogy of the ultramafic interlayers is similar to that in the ultramafic zone. Pentlandite-pyrrhotite-chalcopyrite intergrowths are common, and frequently are associated with oxides. Pyrite is sparse but widespread. It usually forms a fine-scale lamellar intergrowth with pentlandite or chalcopyrite.

### **Gabbros and anorthosites**

There is a marked decrease in the abundance of sulphides in the gabbros, leucogabbros, and anorthosites as compared to the underlying and interlayered ultramafic rocks. Pyrite is the major sulphide phase in the mafic rocks. It commonly occurs as euhedral grains, sometimes containing small blebs of chalcopyrite (plate 1 D). Pentlandite was found in only one gabbro sample and one chromitite sample. Pyrrhotite also becomes less common, being found in only one anorthosite and two gabbro samples.

The sulphide content increases somewhat in the chromitite seams, although it does not reach the levels found in the ultramafic zones. Inclusions of pyrite in chromite are common in some specimens, but euhedral pyrite is the dominant phase, and again it often contains blebs of chalcopyrite. Millerite also occurs as fine-grained blebs and lamellae in the pyrite. These blebs of millerite have sharp

straight borders with pyrite suggesting an equilibrium texture (plate 1 E). Partial oxidation of the pyrite is ubiquitous, and magnetite pseudomorphs after pyrite with residual pyrite cores are common (plate 1 D).

### Pyroxene amphibolite

The pyroxene amphibolite into which the Fiskenæsset complex was intruded has lenses which contain abundant sulphide. Chamberlain (1967) noted that the country rocks near the contact of the Muskox intrusion are also extremely rich in sulphides. Specimen 132002 contains 5.5 modal % sulphide, consisting almost entirely of large, globular pyrrhotite grains. These grains sometimes show small segregations of chalcopyrite and pentlandite with 'flame' exsolution textures. Evidently the amphibolite was quenched more rapidly than the rocks within the main body of the Fiskenæsset complex.

## CHEMISTRY AND GEOTHERMOMETRY OF THE SULPHIDES

The present-day mineralogy of the Fiskenæsset complex gives abundant evidence of modification at low temperature. If the present-day sulphides were derived from the original magma, some changes would have taken place during the initial cooling of the intrusion. Nevertheless, the subsequent metamorphic event must have had a profound effect on the primary sulphide mineralogy. Experimental studies of bulk compositions in the Cu-Fe-Ni-S system indicate extensive recrystallization in geologically short periods of time at temperatures over 300°C (e.g. Kullerud *et al.*, 1969). During granulite-facies metamorphism, temperatures throughout the complex may have reached 850°C (based on two-pyroxene geothermometry; see Part III for details). Although those temperature estimates may be as much as 100°C too high, the present-day sulphide mineralogy of the Fiskenæsset complex must result from subsolidus reactions and recrystallization during metamorphism.

Most specimens contain sulphide assemblages which are dominantly pyrrhotite + pentlandite ± chalcopyrite ± pyrite. At 850°C, bulk compositions in this portion of the Cu-Fe-Ni-S system consist of  $Mss + Iss \pm (Ni,Fe)_{3\pm x}S_2$  (Craig & Kullerud, 1969; Cabri, 1973), where *Iss* is the centrally-located higher-temperature solid solution in the Cu-Fe-S system, usually more sulphur-deficient than ideal chalcopyrite stoichiometry. Some sulphide liquid could be present at those temperatures, especially because magnetite, often present in mineral assemblages, drastically lowers the solidus temperatures of comparable sulphide assemblages (Craig & Kullerud, 1969; Naldrett, 1969). Several specimens from the ultramafic zone con-

tain assemblages rich in Ni and Co. These specimens, 84467, 84468 and 84471, could have reached an assemblage of  $\text{Mss} + (\text{Ni, Fe})_{3\pm x}\text{S}_2 \pm \text{liquid}$  during the metamorphism.

As cooling proceeded after culmination of the metamorphic event, any liquid would quickly crystallize, followed by a series of subsolidus recrystallizations of the sulphide minerals. These subsolidus reactions have been summarized by a number of workers who have reproduced the reactions in the laboratory. This section discusses the chemical evidence for low temperature equilibration of the Fiskensæset sulphides, and then considers other phase relationships, apparently stable below 200°C, which may represent equilibrium assemblages below the lower temperature limit of practical experimentation.

### Fe-Ni-Co-S system

Compositions of all phases in the Fe-Ni-Co-S system except pyrrhotite are plotted in figure 1. Maximum thermal stabilities for most phases in the pure Fe-Ni-S system are below 700°C: godlevskite, 573°C; polydymite, 356°C; violarite, 461°C; pentlandite, 610°C (cf. Craig & Scott, 1974). Recent results indicate that the stability of pentlandite may be increased by Co substitution. Pentlandite with 40.8% Co has a maximum thermal stability of 746°C (Vaasjoki *et al.*, 1974) compared to 835°C for pure  $\text{Co}_9\text{S}_8$  (Rosenquist, 1954). This stabilizing effect, due to the greater crystal field stabilization energy gained by divalent low spin Co as compared to either  $\text{Fe}^{2+}$  or  $\text{Ni}^{2+}$  (Rajamani & Prewitt, 1973), cannot be discounted for other Fe-Ni sulphides which accept Co. Because table 2 and figure 2 show that few of the sulphide phases in the Fiskensæset complex except for pentlandite in specimens 84467 and 84471 contain appreciable Co, the above thermal maxima are reasonable.

As previously noted, the assemblage pyrite-pentlandite is unstable above 230°C. The pentlandite in 84497 associated with unmixed pyrite (compare fig. 1 and fig. 3)

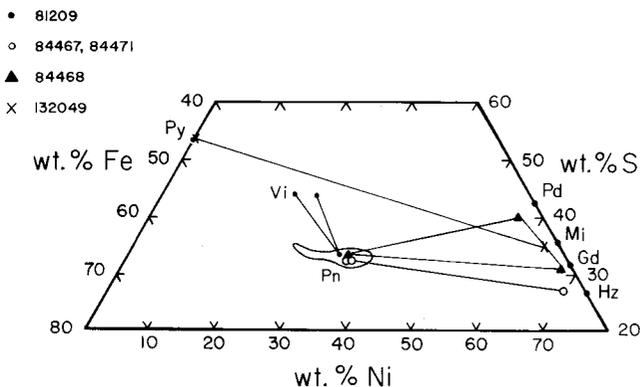
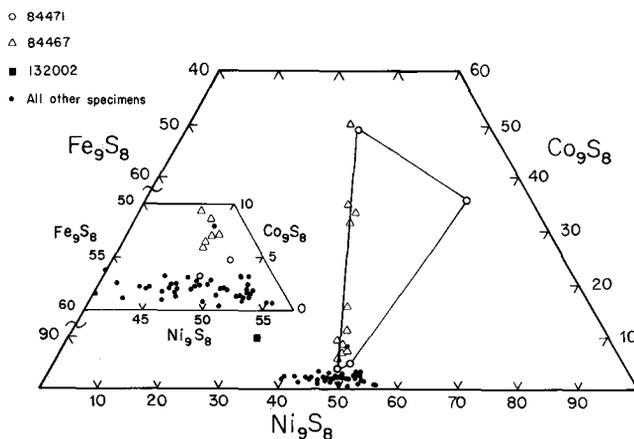


Fig. 1. Compositions of Fe-Ni-S phases plotted on a portion of the Fe-Ni-S ternary diagram. Wt.%,  $\text{Fe} + \text{Ni} + \text{S} = 100\%$ . Irregularly-shaped region near centre of diagram indicates the range in composition of all pentlandite grains analyzed. Stoichiometric compositions of polydymite, millerite, godlevskite and heazlewoodite have been plotted for reference.

Fig. 2. Weight percent ternary plot of all pentlandite analyses in table 2 recalculated to  $(\text{Fe,Ni,Co})_9\text{S}_8 = 100\%$ . Tie-lines join pentlandite compositions from a single specimen (GGU 84471). Inset shows low-Co pentlandite compositions in more detail.



is richer in Ni than pentlandites associated with heazlewoodite, godlevskite, millerite, and polydymite. The maximum Ni concentration in pentlandite has been experimentally shown to be an irregular function of temperature with maxima occurring at 600°C (Misra & Fleet, 1973, fig. 3). Another maximum could exist below 230°C, and pentlandite equilibrating with pyrite at this temperature could contain more Ni than a pentlandite-heazlewoodite assemblage equilibrating at another temperature. Another possible explanation could lie in the fact that pentlandites coexisting with Ni sulphides all contain > 4 wt.% Co. In the quaternary system, Fe-Co-Ni-S, tie-lines between Co-rich pentlandite and heazlewoodite need not cross tie-lines between pyrite and Ni-rich pentlandite.

Figure 1 shows that some specimens contain pentlandite-violarite and others pyrite-millerite assemblages. It is questionable whether pentlandite-violarite or pyrite-millerite form equilibrium assemblages at low temperature (Kullerud *et al.*, 1969; Graterol & Naldrett, 1971; Craig, 1973). Strong support for a pyrite-millerite tie-line in some temperature interval is given by the texture of this association in

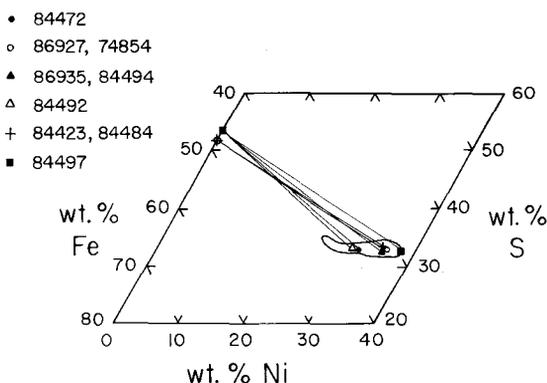


Fig. 3. Ternary plot of compositions of coexisting pyrite and pentlandite. Notice that pyrite coexists with several Ni-rich pentlandites including the most Ni-rich pentlandite found in this study.

132051 (plate 1 E) where the border between the two grains is straight and sharp. Neither grain appears altered or weathered, which is to be expected since the grain is an *inclusion* in a chromite crystal. This armouring apparently allowed formation of an equilibrium assemblage.

A novel explanation for the existence of the two assemblages (pentlandite-violarite and pyrite-millerite) is suggested by figure 1. The tie-lines for these two assemblages do not cross, and pyrite-millerite may be a stable assemblage in the same temperature range as pentlandite - Fe-rich violarite. There are several lines of evidence arguing against this position, however: (1) Fe-rich violarite apparently forms a continuous solid solution with polydymite at low temperatures (Misra & Fleet, 1974; Craig & Higgins, 1975); (2) the textural evidence indicates violarite formed by alteration of pentlandite under possibly non-equilibrium conditions (see previous section); and (3) the tie-lines for pentlandite-violarite cross with tie-lines for coexisting pentlandite and pyrite from the same zone (fig. 3).

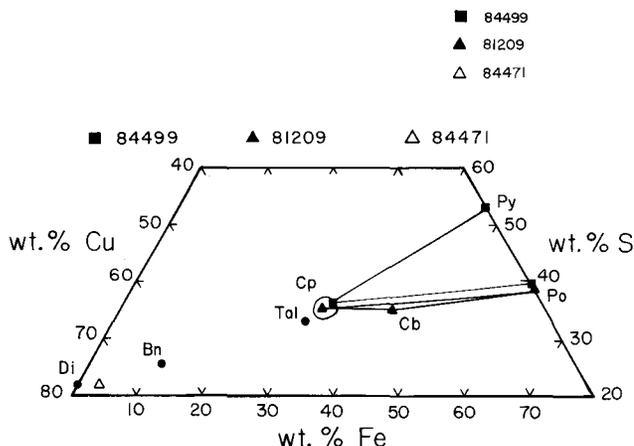
A surprise is the assemblage pentlandite-godlevskite-polydymite in 84468. Although identification of godlevskite and polydymite could not be checked by X-rays because of the fine-grained nature of the phases (plate 2 A), compositions are nearly stoichiometric except for a few wt.% Fe (table 2 and fig. 1). Coexistence of these three phases does not violate the phase rule, but the lack of millerite is disturbing. If millerite were present in undetectable amounts, disequilibrium would be indicated by the coexistence of four Fe-Ni-S phases. Although the intergrowth does not give any textural indication of being a non-equilibrium assemblage, this possibility cannot be ruled out. The alternatives include the possibility that millerite becomes unstable at low temperatures or that a tie-line is established between Fe-bearing polydymite and Fe-bearing godlevskite. This latter explanation is unlikely because millerite accepts much more Fe than any other Ni-sulphide at higher temperatures (Kullerud & Yund, 1962). Interestingly, Misra & Fleet (1973) reported that godlevskite may have a lower temperature limit of stability. The low-temperature phase relations of the Ni-sulphides are complex and nothing definitive can be stated at this time.

Specimens 84497, 84467 and 84471 contain pentlandite of variable composition. Pentlandites from 84467 and 84471 contain substantial cobalt. Figure 2 shows pentlandite compositions as a function of variations in Fe, Ni, and Co. Tie-lines connect coexisting pentlandite compositions in specimen 84471. The wide range in composition exhibited in 84467 and 84471 suggests disequilibrium, a conclusion strengthened by the presence of veinlets containing high-Co pentlandite.

### **Cu-Fe-S system**

Compositions of the Cu-Fe-S phases are plotted in figure 4. Cubanite is nearly stoichiometric  $\text{CuFe}_2\text{S}_3$ . Its X-ray powder pattern matches that of orthorhombic

Fig. 4. Compositions of Cu-Fe-S phases plotted on a portion of the Cu-Fe-S ternary diagram. Wt.%, Cu+Fe+S = 100%. Chalcopyrite field indicates range in composition of all grains analyzed. Stoichiometric bornite and digenite and talnakhite compositions are plotted for reference.



cubanite, indicating an inversion has taken place from cubic cubanite below 210°C (Cabri, 1973). In both 81209 and 132007, cubanite is associated with pyrrhotite and chalcopyrite, consistent with equilibration at about 350°C or below (Kullerud *et al.*, 1969; Vaughan & Craig, 1978; Sugaki *et al.* 1975).

All of the chalcopyrite analyses are within analytical error of M/S = 1. However, a small variation in Cu/Fe slightly exceeds the analytical error (fig. 4). Nevertheless, the compositions do not approach that of talnakhite, the phase most similar chemically. The chalcopyrite assemblages, chalcopyrite-pyrrhotite-pyrite ( $\pm$  pentlandite), chalcopyrite-pyrrhotite ( $\pm$  pentlandite), chalcopyrite-pyrrhotite-cubanite-(pentlandite), and chalcopyrite-pyrite ( $\pm$  pentlandite  $\pm$  millerite), are all consistent with known and predicted phase relations at 25°C (Craig & Scott, 1974).

The chalcopyrite-pyrite intergrowth in 84494 could not have formed above 230°C because of the presence of a large amount of pentlandite in the specimen. Below 300°C, Iss should not exsolve pyrite (Craig & Scott, 1974, figure CS-31) which suggests that the chalcopyrite, subordinate in volume, also may be exsolved from a Cu-bearing precursor to pentlandite and pyrrhotite.

In specimen 84471, low-digenite with 3 wt.% Fe was found in association with Fe-Ni sulphides (fig. 4). From the experimental work of Kullerud (1960) on the digenite-bornite miscibility gap, a minimum temperature of 185°C can be established for the one-phase assemblage. The microprobe analysis of 1.3% Ni may result in part from finely-intergrown pentlandite. Removal of an appropriate amount of Fe from the digenite analysis would lower the minimum temperature to below 125°C, which probably represents the temperature range at which many of the late-stage alteration products were formed.

Table 3. Cobalt concentrations in sulphides and whole rock

Zone	Specimen Number	Py (wt% Co)	Pn (wt% Co)	Rock (ppmw Co) (from Part I)
UA	81233	n.d.		
Ch b	132041	n.d.		
	132038	0.31		
	132037	0.22		
Ch a	132055	0.40		
	132053	0.24		
	132052	0.14		
	132051	n.d.		
	132050	1.24		
	132049	0.16		
	132047	0.18		
LA	86929	0.14		30
ULG	86931			50
	86924	0.60		50
G	132006	0.22		
	84499	n.d.		
	84498	n.d.		
UM	122350		1.08	
	122349		1.50	
	86935	1.46	0.21	170
	86927	1.33	0.70	200
	84490		1.60	
	84488		1.30	
	84487		1.50	
	84485		2.00	
	84472	2.0	0.85	
	84471		20.00	
	84469		0.85	
	84468		5.25	
	84467		20.00	
	84466		1.50	
	81216		0.74	
	81209		2.08	
-	132002		6.15	
Ultramafic Interlayers				
LA	84484	1.60	1.05	
LA	74854	1.50	1.00	
ULG	132012		2.22	
G	84497	2.30	0.46	
LLG	84423		1.30	
LLG	84494	1.2	1.40	
LLG	84492		1.40	

Abbreviations as in Table 1.

### Minor elements

Low temperature of equilibration has reduced the minor element substitutions in the sulphides. For example, the highest Ni concentration in chalcopyrite is 0.44 wt.% as compared, say, to Ni contents of up to 2.31 wt.% in quenched Iss from submarine basalts (Czamanske & Moore, 1977).

Two notable exceptions exist. The first is the Ni concentration in pyrite, which reaches as high as 6.61 wt.% in 86929. None of the specimens containing pyrite with Ni > 0.44 wt.% contain olivine, and only one, 84498, contains pentlandite. Apparently when bulk compositions prevent formation of the more common Ni-bearing phases, Ni is accommodated by pyrite.

Cobalt is a common minor constituent of pentlandite, pyrite and pyrrhotite, but exceeds 5 wt.% in pentlandite from the three specimens containing Ni-rich assemblages (84467, 84468 and 84471) and 6 wt.% in pentlandite from the pyroxene amphibolite, 132002. A decrease in whole rock Co with stratigraphic position was noted in Part I of this series. There is a concomitant decrease in Co concentrations in pyrite and pentlandite from these rocks (table 3). Striking exceptions to the general trend are the ultramafic interlayers in which Co concentrations are similar to those from the ultramafic zone.

## SULPHIDE STRATIGRAPHY AND ORIGIN OF THE SULPHIDES

Variation in sulphide mineralogy as a function of stratigraphic placement within the Fiske­næsset complex is plotted in fig. 5. Late-stage exsolution or alteration products have not been plotted because the emphasis here is on the major phase relations. In general, there is a systematic progression from Ni, Fe-rich sulphide assemblages in the lower ultramafic and mafic zones to Cu,S-rich assemblages in the upper, more felsic zones. There is also a strong correlation between host rock type and sulphide assemblage, indicating that silicate chemistry may relate strongly to sulphide chemistry.

The extent of this effect is most clearly indicated by the sulphide mineralogy of the ultramafic interlayers in plagioclase-bearing zones of the Fiske­næsset complex. Typically the interlayers are around a metre thick and seldom exceed ten metres in length. Contacts with enveloping rocks are sharp and conformable. We regard the interlayers either as minor cumulate units produced during progressive crystallization of the main magma, or as the products of minor influxes of new magma, or perhaps as the products of rafting by convection currents from the walls of the complex.

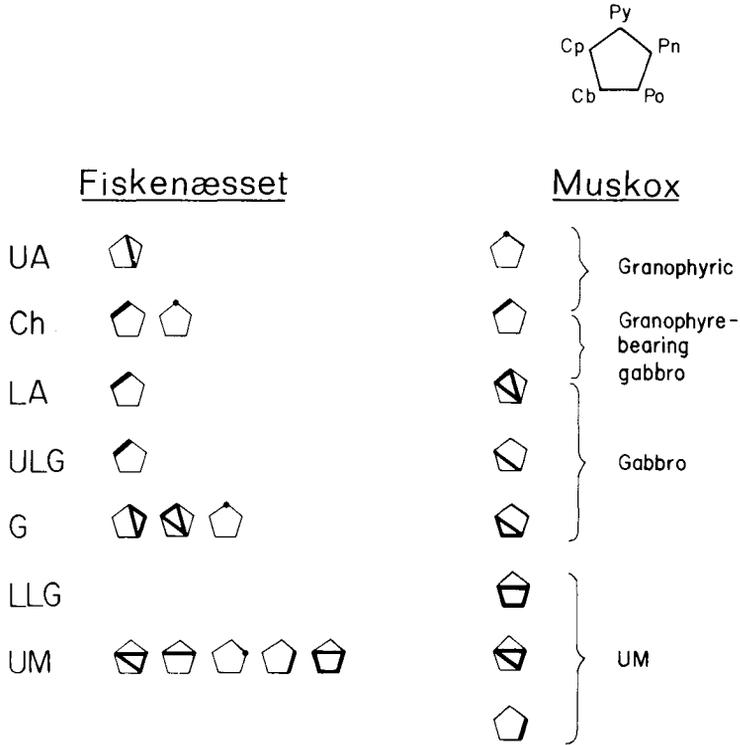


Fig. 5. Sulphide mineralogy of the Fiskeenæsset complex as a function of stratigraphic position in the intrusion. UM interlayers not plotted. Sulphide mineralogy of the Muskox Intrusion, Canada, is shown for comparison.

The character of the sulphide mineral assemblages in the ultramafic interlayers is distinctly different from that in the surrounding plagioclase-rich rocks, and the higher abundance of sulphide in the ultramafic interlayers even approaches that found in the ultramafic zone. Whereas pyrite is relatively important in the plagioclase-bearing rocks, it is subordinate to pentlandite, pyrrhotite and chalcopyrite in the interlayers. Wherever pyrite occurs in the interlayers, it is less abundant than the other sulphides and forms a lamellar intergrowth with pentlandite (e.g. plate 2 E). Because these fine-scale intergrowths of pyrite and pentlandite are surrounded by massive pyrrhotite, it is implausible to attribute the pyrite to replacement of pyrrhotite in response to change of sulphur fugacity. Apparently the pyrite results from late-stage reaction of  $Mss$  to pentlandite + pyrite. It is evident from the extreme contrast in sulphide assemblages between the ultramafic interlayers and the plagioclase-bearing rocks less than a metre away that the nature of the sulphide mineralogy in the Fiskeenæsset complex reflects the composition of the immediate host rock.

Late-stage recrystallizations and alterations accompanying the metamorphism in

the Fiskenæsset complex have acted to obscure evidence on the origin and early history of the sulphides. Nevertheless, some tentative conclusions about the origin of the sulphides can be made based on comparison with sulphide mineralogy previously studied in other layered complexes. Those complexes which have been subjected to little post-emplacement recrystallization provide a valuable guide to the original mineralogy and textures expected in the Fiskenæsset complex prior to extensive metamorphism and deformation.

The sulphides in the Muskox intrusion fall within the Fe-Ni-Cu-S system (Chamberlain, 1967), and their assemblages vary with stratigraphic position in a similar manner to the assemblages in the Fiskenæsset complex. Ultramafic members of the Central Layered Series of the Muskox intrusion contain the nickel-rich assemblages pyrrhotite-pentlandite, (chalcopyrite)-pyrrhotite-pentlandite and chalcopyrite-pyrrhotite-cubanite-pentlandite. In contrast, the upper gabbroic and granophyric members of the Muskox intrusion contain the nickel-poor assemblages chalcopyrite-pyrrhotite-cubanite, chalcopyrite-pyrrhotite, and the sulphur-rich assemblages pyrite-chalcopyrite-pyrrhotite, and pyrite-chalcopyrite. The similarity in the stratigraphic sequence between the Fiskenæsset complex and the Muskox intrusion is striking (fig. 5), and the strong correlation between rock type and sulphide assemblage is also apparent in the Muskox intrusion.

The initial magma of the Skaergaard intrusion was undersaturated in sulphur at the time of emplacement because (1) sulphides are absent in the lower cumulate rocks, and (2) the S concentration of the chill zone is only 50 ppm. By the time sulphides began forming, most of the Ni had been removed from the melt by olivine, thereby explaining the notable lack of nickel sulphides (Wager *et al.*, 1957). The Layered Series, which is dominated by gabbros and ferrogabbros, contains a Cu-rich suite of sulphides including bornite, chalcopyrite, digenite, and covellite. Towards the top of the ferrogabbros a pyrrhotite-rich assemblage appears, and, finally, a pyrite-chalcopyrite assemblage is found in the acid granophyres.

The sulphide mineralogy of the Stillwater intrusion also exhibits parallels with that of the Fiskenæsset complex. Howland (1933) found irregular bodies composed predominantly of sulphides scattered along the base of the Stillwater complex where they apparently concentrated in depressions on the floor of the magma chamber. Some of these bodies continue a few metres into the country rock, a quartz-biotite-cordierite-hypersthene hornfels. These bodies are composed principally of pyrrhotite, pentlandite and chalcopyrite with secondary marcasite, violarite(?) and limonite. In contrast to the Skaergaard intrusion, sulphides, probably in the form of an immiscible liquid, must have been present in the initial Stillwater magma or have formed immediately after the intrusion. Other sulphides are found disseminated in the basal zone (Page & Simon, 1978), in the chromitite zones of the Ultramafic Zone (Page & Jackson, 1967), and in thin horizons stratigraphically higher in the intrusion (Howland, 1933). Disseminated sulphides of the Basal Zone

consist predominantly of pentlandite and chalcopyrite (Page & Simon, 1978). In the chromitite zones, pentlandite is the most abundant sulphide followed by pyrrhotite, chalcopyrite and several minor phases, but pyrite is not present (Page, 1971). One sulphide segregation was described from an anorthosite higher up in the intrusion at the Picket Pin Creek locality. Although pentlandite, chalcopyrite and pyrrhotite are still abundant there, it is significant that pyrite, with both euhedral and irregular outlines, has become a common phase (Howland, 1933).

Sulphides are found disseminated throughout the sequence of layered rocks in the Bushveld igneous complex (Liebenberg, 1970). Differentiation of the sulphides is well documented and has been summarized by Liebenberg (1970):

“The sulphides in the ultramafic rocks of the Basal Zone are rich in pentlandite and poor in Cu sulphide. Towards higher horizons the percentage of pentlandite decreases and the percentage chalcopyrite increases to a maximum below the Main Magnetitite. In the Upper Zone the chalcopyrite becomes progressively less towards higher horizons. The sulphides near the top of the Upper Zone are very rich in pyrrhotite and consist only of pyrite in the diorites.”

The above-mentioned complexes differ in initial magma composition, initial concentrations of sulphur and chalcophile elements, oxygen fugacity, history of multiple intrusions, post-emplacement alteration, volume of magma, and age. Yet they all exhibit progressive changes in sulphide mineralogy with stratigraphic position, and the mineralogy correlates well with host rock; pentlandite is usually an important mineral in ultramafic rocks whereas chalcopyrite and pyrite become more important in gabbro and anorthosite.

A change in  $\text{Cu}/(\text{Cu}+\text{Ni})$  in sulphide ore deposits as a function of the host rock chemistry has long been recognized (Wilson & Anderson, 1959; Naldrett, 1973). Nickel-copper ore bodies show an increase in  $\text{Cu}/(\text{Cu}+\text{Ni})$  from 0.3 to 7.5 as MgO falls from 33 wt.% to 4 wt.% in the silicate magma from which the sulphides segregated (Naldrett & Cabri, 1976). In a crude parallel, a similar change in composition of disseminated sulphides seems to occur in layered intrusions, where the change in whole rock composition is due to fractionation of a single magma. Olivine fractionation (Wager *et al.*, 1957), distribution of Cu, Ni, Co and Fe between immiscible silicate and sulphide liquids (Rajamani & Naldrett, 1978), Fe-depletion, and possibly even segregation of a sulphur-rich liquid (Naldrett & Gasparrini, 1971) probably all combine in a complex interplay to affect the evolution of sulphide compositions. Sulphide stratigraphies within layered intrusions would be difficult to reproduce by assimilation of country rocks or a post-crystallization, sulphide metasomatism.

Because of the similarity of the sulphide mineralogy in the Fiskensæset complex with that of the Muskox (fig. 5) and other layered intrusions (Myers, 1976), it is reasonable to conclude that most sulphides present in the Fiskensæset complex were originally derived from the magma. Other lines of evidence support this conclusion. It is unlikely that sulphur or sulphides were incorporated into the

magma by assimilation of wall rocks because only one inclusion (approximately 50 cm in diameter) of such a rock has been recorded in the intrusion (Part III, p. 25). Occasionally, rafts of roof marbles are found, and while these may affect silicate-oxide relations (Part III) they lack sulphides. Quartzo-feldspathic gneisses in the area were probably derived from intrusive tonalites and granodiorites, and post-date the complex.

The possibility of sulphide metasomatism also is remote. The mineral chemistry of silicates and oxides in Parts II and III indicates that chemical migration of elements within and amongst these minerals was small during metamorphic re-equilibration. The contrast in sulphide assemblages within ultramafic interlayers and surrounding plagioclase-bearing rocks is a strong indication that migration of sulphur was also minimal. Even Co concentrations within the sulphide minerals remain distinct between the two rock types (table 3). Thus, the sulphide stratigraphy is best viewed as being similar to that of the complex before metamorphism.

It is now commonly recognized that sulphur-bearing phases are widely distributed throughout the upper mantle (Frick, 1973; Bishop *et al.* 1975; Tsai *et al.*, 1979). Sulphides occur in many peridotite xenoliths, as inclusions in silicates, as interstitial globules, and in veinlets (Harte *et al.*, 1975). Sulphide inclusions also occur in diamond (Gurney *et al.*, 1979). Available experimental evidence on systems free of Ni and Cu indicate that partial melting of mantle material would preferentially concentrate iron sulphide into the initial melt (MacLean, 1969; Shamazaki & Clark, 1973). Some recent magmas which have reached the crust have been shown to be saturated with respect to sulphide liquids (Mathez & Yeats, 1976; Anderson, 1974; Czamanske & Moore, 1977). The great age of the Fiskensæset complex increases the likelihood that the initial magma would contain appreciable sulphide if there has been any sulphur depletion in the mantle through geologic time (Naldrett, 1973). The maximum in sulphide concentration in the lowest zone in the intrusion could result from immiscible sulphide droplets at the time of emplacement. Liquid immiscibility would be enhanced by the relatively high oxidation state of the magma (Steele *et al.*, 1977).

Oxygen has been shown to be an integral component of immiscible sulphide liquids coexisting with basaltic liquids (Desborough *et al.*, 1968; Naldrett, 1969). The amount of oxygen dissolved in the sulphide liquid, however, is expected to be too small for this mechanism to account for the sulphide-oxide grains in the Fiskensæset complex. The effects of pressure, temperature and liquid composition on oxygen solubility in sulphide liquids, although unknown, are unlikely to account for the observed relationships. During crystallization, oxides could have adhered to trapped sulphide melt.

## CONCLUSIONS

This paper completes the reconnaissance of the mineralogy of the Fiskenæsset intrusion on Qeqertarsuaq. In this series petrogenetic histories of 29 mineral species (10 silicates, 7 oxides, and 12 sulphides) in key lithologies of the complex have been elucidated. A unique feature of this complex is that it retains much of its original, igneous, chemical pattern, indicating relatively restricted diffusion during metamorphism. The sulphide chemistry presented in this paper records subsolidus reaction products at extremely low temperatures of an original fractionated igneous paragenesis.

Although this study was initiated in the belief that Archaean crust might show similarities to lunar crust, and that, in particular, Fiskenæsset anorthosites may be analogous to rocks from the lunar highlands, we have shown that the Fiskenæsset complex is quite distinct. It was derived from a hydrous, somewhat oxidized magma which contrasts with the dry, reduced state of lunar rocks. The Fiskenæsset complex has provided insight to igneous and metamorphic processes on the Earth, however, as well as important evidence on the nature of Archaean tectonics (Windley & Smith, 1976).

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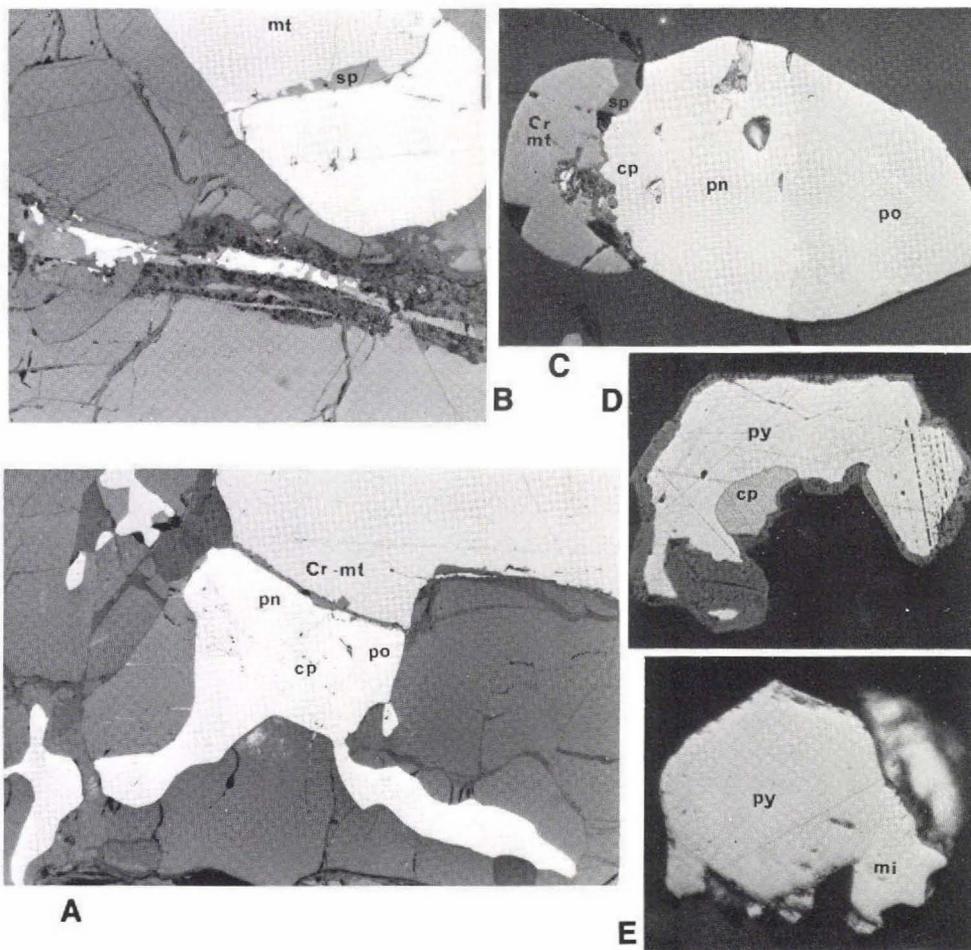
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## **Plates**

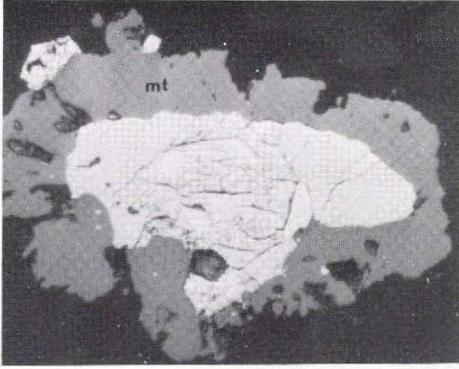
**Plate 1**

- A: Large sulphide-oxide aggregate showing a scalloped margin. Pyrrhotite, pentlandite and chalcopyrite are all abundant with minor, exsolved pyrite. Light grey Cr-magnetite contains fine lamellae of spinel (*sensu stricto*). Spinel also forms the dark border between the sulphide and oxide portions of the grain. Reflected light. GGU 84423. 2 mm across.
- B: Sulphide-oxide globule and nearby veinlet containing sulphides. Light-coloured sulphides separated from grey magnetite by dark grey spinel (*sensu stricto*). Veinlet contains pyrrhotite and chalcopyrite of the same composition as that found in nearby sulphide globules. Reflected light. GGU 84423. 1.5 mm across.
- C: Sulphide-oxide globule. Light grey phase in middle is pentlandite containing a small, light-coloured crescent of chalcopyrite. Medium grey phase is pyrrhotite, on opposite end of globule from dark grey Cr-magnetite. Very dark oxide between pentlandite and Cr-magnetite is spinel (*sensu-stricto*). Reflected light. GGU 84488. 0.3 mm across.
- D: Euhedral hopper crystal of pyrite. Dark grey rim is incipient oxidation to magnetite. Small grey bleb is chalcopyrite. Reflected light. GGU 86924. 0.5 mm across.
- E: Pyrite-millerite grain. Darker phase is pyrite. Note sharp border between the two phases. Reflected light. GGU 132051. 0.1 mm across.

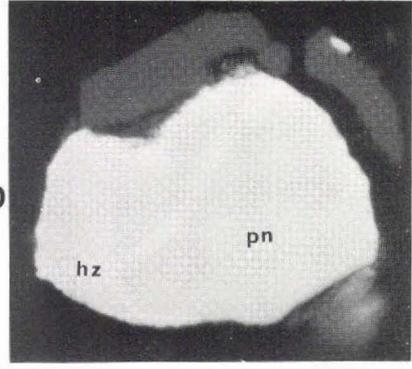


**Plate 2**

- A: Pentlandite grain containing fine-scale intergrowth of polydymite and godlevskite, surrounded by oxidation rim of magnetite. Reflected light. GGU 84468. 0.3 mm across.
- B: Chalcopyrite crystal containing band of cubanite. Both minerals abut against a dark crystal of pyrrhotite. Partially crossed polars, reflected light. GGU 81209. 0.3 mm across.
- C: Pentlandite (light grey) containing dark grey bands of violarite. Reflected light, partially crossed polars. GGU 81209. 0.3 mm across.
- D: Pentlandite (grey) containing light grey bands of heazlewoodite. Small adjacent grains of magnetite (dark grey). Reflected light. GGU 84467. 0.1 mm across.
- E: Pyrite-pentlandite intergrowth. Lighter phase is pyrite. Pentlandite also forms massive areas in same globule. Reflected light. GGU 84494. 0.15 mm across.
- F: Chalcopyrite-pyrite intergrowth. Near area of plate 2E. Chalcopyrite is dark grey phase. Reflected light. GGU 84494. 0.15 mm across.

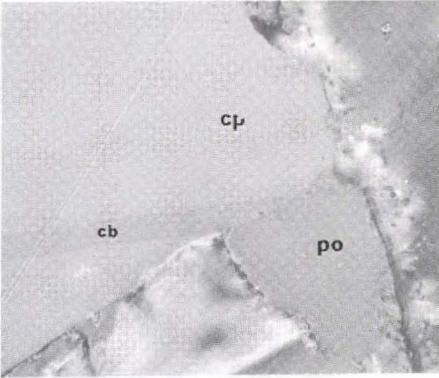


A

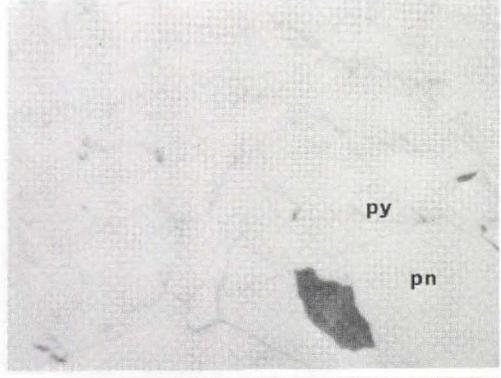


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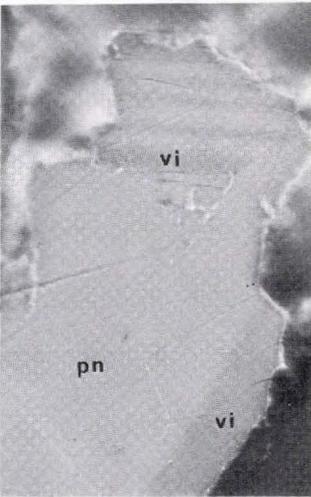
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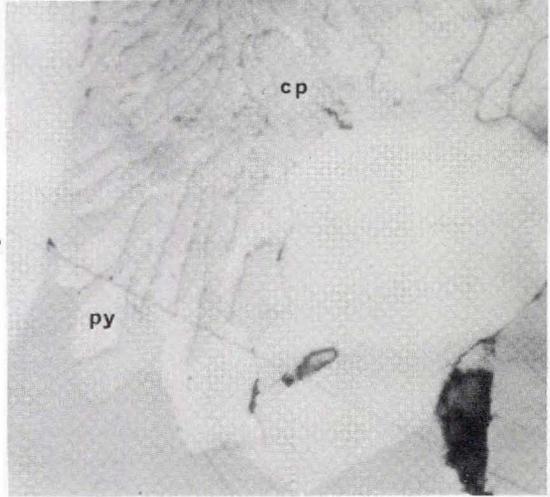
B



F



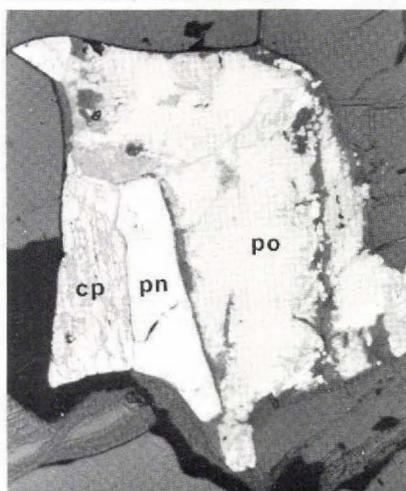
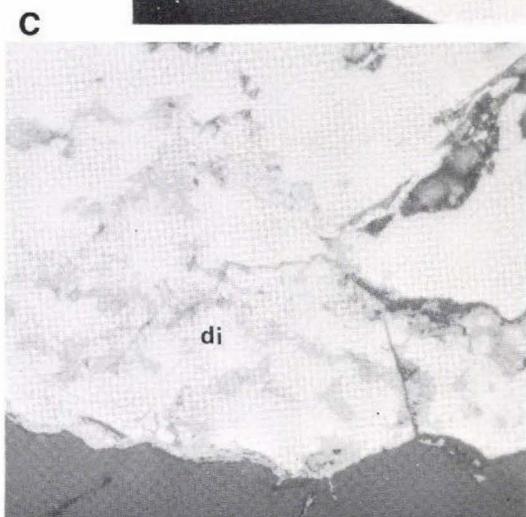
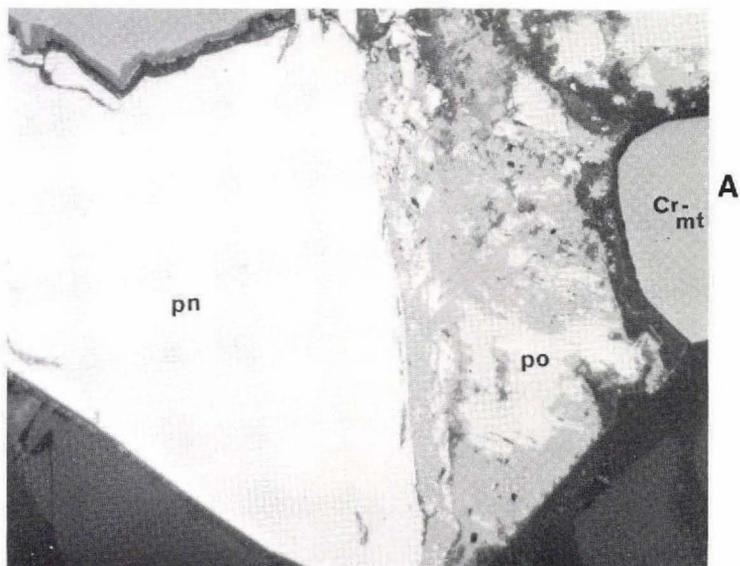
C



E

**Plate 3**

- A: Sulphide globule containing heavily oxidized pyrrhotite and a small grain of dark grey Cr-magnetite. Light, unaltered phase is pentlandite. Reflected light. GGU 132007. 0.4 mm across.
- B: Large sulphide globule containing pentlandite, pyrrhotite and chalcopyrite. Heavily weathered pyrrhotite contains fine lamellae of pentlandite and chalcopyrite. Chalcopyrite is oxidized with a trellis pattern and pentlandite is relatively unweathered. Reflected light. GGU 132007. 1.5 mm across.
- C: Coarse pentlandite grain showing alteration rim comprised of fine-scale intergrowth of digenite and magnetite. This may represent alteration of pre-existing chalcopyrite. Reflected light. GGU 84471. 0.2 mm across.



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