

TABLE XVIII

Comparison between the values obtained for the composition of the plagioclase by the author and by Bisschoff (1950, p. 10)

ROCK TYPE	Composition in per cent Anorthite	
	Values obtained margin - core	Value of Bisschoff (1950)
Olivine gabbro	An ₄₁ (unzoned)	-
Coarse pyroxene diorite	An ₃₂ (unzoned)	An ₃₂
Dolerite	An _{30,5} (unzoned)	-
Fine pyroxene diorite	An ₃₀ - 33	-
Pyroxene andesite	An ₂₈ - 34	-
Dioritic-aplite	An ₃₀ (unzoned)	-
Dicritic-pegmatite	An ₂₇ - 29	-
Equigranular diorite	An ₂₇ - 37	An ₃₇
Andesite	An ₂₄ - 32	-
Syenite	An ₂₈ - 30	An ₂₈ - 30
Syenite-porphry	An ₁₄ - 16	An ₂₀

(ii) Potassium feldspar

Willemsse (Nel et al., 1939, p. 90) observed potassium feldspar (microperthite) within the coarse pyroxene diorite. Bisschoff (1950) and (1972b), however, states that potassium feldspar is not present within the rocks of the Complex, and concluded that Willemsse mistook an abnormal type of albite twinning for microperthite.

In an effort to determine if potassium feldspar is in fact present within the rocks of the Complex, the author examined several selected grains by X-ray diffraction. Two varieties of potassium feldspar grains were X-rayed, firstly turbid, anhedral grains from the syenite-porphry and secondly grains showing exsolution from the coarse pyroxene diorite and equigranular diorite. The 2θ values obtained for the samples X-rayed are given in Table XIX.

TABLE XIX (modified after Wright 1968, p. 89, Table I)

Comparison between the positions of the most important alkali feldspar reflections according to Wright (1968, p. 89) and those obtained for the feldspar samples which were thought to be potassium feldspar

Miller Indices (h k l)	Approximate Intensity	Approximate range of 2θ (Cu K α_1)				Samples from diorite	Samples from syenite-porphyry
		Potassium-rich phases	Anorthoclases	Albite			
($\bar{2}01$)	40	20,8°- 21,2°	21,6°- 21,9°	21,9°- 22,1°		21,85°	-
(002)	100	27,4°- 27,8°	27,8°- 28,0°	27,9°- 28,1°		27,95°; 28,05° & 28,10°	27,80°
($\bar{1}13$)	8	38,6°- 39,0°	not present	not present		39,00°	38,80°
(060)	25	41,6°- 42,0°	41,7°- 42,0°	42,2°- 42,6°		41,95° & 51,20°	41,90°
($\bar{2}04$)	30	50,5°- 51,1°	51,1°- 51,3°	51,2°- 51,5°		50,05°	50,20°

From the data presented in Table XIX it can be noted that, for both sets of samples X-rayed, the ($\bar{2}04$), (060) and ($\bar{1}13$) reflections gave values within the potassium feldspar range. The presence of a ($\bar{1}13$) reflection indicates a potassium-rich phase since this reflection is absent in anorthoclases and albite (Wright, 1968).

b. Pyroxene

Pyroxene is a common rock forming mineral within the Complex occurring in all the rock types except the pegmatite, syenite and syenite-porphyry. The varieties of pyroxene present are limited to augite and bronzite.

(i) Augite

This is by far the most abundant variety of pyroxene and is present in anhedral to subhedral and more rarely euhedral grains that vary in size from 0,5 millimetres to 2,6 millimetres. Generally the euhedral forms consist of stubby crystals with pyramidal terminations and hexagonal cross sections. The augite is commonly twinned parallel to (100) and may contain exsolution lamellae of another pyroxene lying parallel to this plane. These exsolution lamellae are probably composed of orthopyroxene as was suggested by Bisschoff (1950, p. 11). The cleavage of the augite is continuous across the exsolution lamellae.

The optical properties of the augite present in the coarse pyroxene diorite varies slightly from the optical properties of the augite present in the fine pyroxene diorite.

Values for augite from coarse
pyroxene diorite

$$2V_z = 51^\circ$$

$$Z_{Ac} = 40^\circ - 44^\circ$$

Pleochroism

X = colourless

Y = pale greenish-brown

Z = pale greenish-brown

$$\beta = 1,692$$

$$\gamma = 1,725$$

Values for augite from fine
pyroxene diorite

$$2V_z = 47^\circ$$

Pleochroism

As for augite from the
pyroxene diorite

$$\beta = 1,702$$

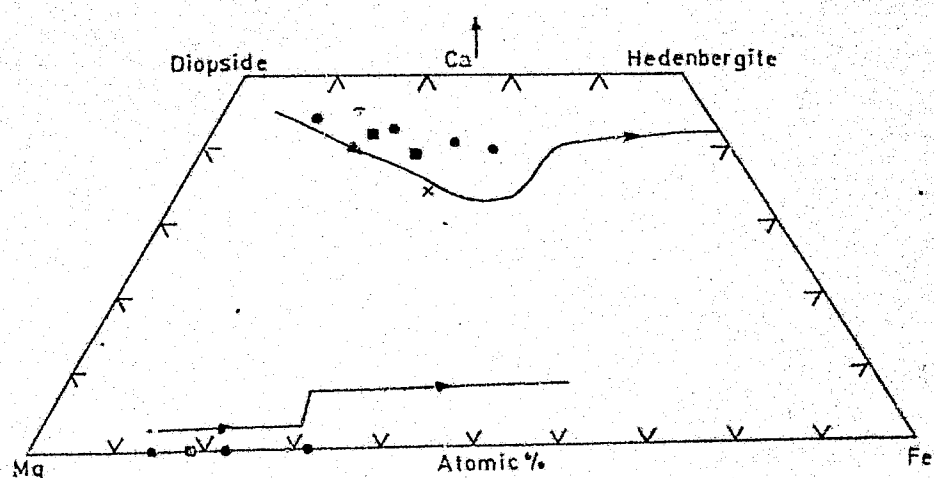
It can be observed from the above that the augites of the fine pyroxene diorite have a smaller $2V$ and a higher β value than the augites of the coarse pyroxene diorite. According to Deer et al. (1967) this indicates that the augite of the fine pyroxene diorite has a higher proportion of iron and less calcium than the augite of the coarse pyroxene diorite. The author obtained compositions of $Wo_{39}En_{37}Fs_{24}$ and $Wo_{42}En_{40}Fs_{18}$ for the augites from the fine pyroxene diorite and coarse pyroxene diorite respectively, by using a diagram showing the variation of optical properties with chemical composition of clinopyroxenes (Deer et al., 1967, 2, p. 132, Fig. 41).

(ii) Orthopyroxene

Orthopyroxene is only present in the olivine gabbro. It occurs as large, elongated and subhedral crystals which may enclose grains of olivine and magnetite poikilitically. The orthopyroxene grains may be up to 5 millimetres in length. In thin section the orthopyroxene is colourless, has $2V_x = 76^\circ$ and $\gamma = 1,687$. Referring these values of $2V_x$ and γ to Deer et al. (1967, 2, p. 28, Fig. 10) a composition of En_{81} is obtained, indicating that the pyroxene is bronzite.

When the compositions of the pyroxenes from the Roodekraal Igneous Complex, Losberg Intrusion and Bushveld Igneous Complex are compared by plotting their positions on the pyroxene triangle (Fig. 3), it can be observed that the pyroxenes of the Roodekraal Igneous Complex have a similar composition to the pyroxenes from the Losberg Intrusion. However, the clinopyroxenes from Roodekraal contain more magnesium in relation to calcium and iron

Fig. 3 Comparison of the compositions of pyroxenes from the Bushveld Igneous Complex, Losberg Intrusion and the Roodekraal Igneous Complex.



- Bushveld Igneous Complex pyroxene differentiation trends according to Atkins (1969, p. 242, fig. 4)
- Composition of pyroxenes from Losberg intrusion according to Abbott and Ferguson (1965, p. 46, fig. 6)
- Composition of pyroxenes from Roodekraal (author's measurements)
- x Composition of pyroxene from Roodekraal according to Bisschoff (1950 p. 12)

and less calcium in relation to magnesium and iron. The clinopyroxenes from both the Roodekraal Igneous Complex and the Losberg Intrusion contain more calcium in relation to magnesium and iron than those from the Bushveld Igneous Complex.

c. Amphibole

Amphibole is present in great abundance, especially within the equigranular diorite and andesite. The varieties of amphibole which have been observed at Roodekraal are listed in Table XX together with the rock types in which they occur.

TABLE XX

Varieties of amphibole present at Roodekraal and their host rocks

Amphibole	Host Rocks
Actinolite	All rock types except olivine gabbro.
Arfvedsonite	Equigranular diorite and syenite.
Barkevikite	Syenite-porphyry.
Common Hornblende	Equigranular diorite, andesite, coarse and fine pyroxene diorite, pyroxene andesite, dioritic-pegmatite, dioritic-aplite and syenite-porphyry.
Kaersutite	Equigranular diorite, andesite, olivine gabbro and, in accessory amounts, in the syenite-porphyry.
Uralite	All rock types except syenite.

Since actinolite and urallite are considered to be secondary rock forming minerals they will be dealt with in a later section.

(i) Arfvedsonite

Arfvedsonite occurs as coronas around kaersutite within the equigranular diorite, and as subhedral rectangular crystals up to 0.4 millimetres in length in the syenite dikes.

The optical properties of arfvedsonite were determined to be as follows:-

$$\begin{aligned} X\lambda c & 30^\circ - 120^\circ \\ Z = b & O.A.P. \perp (010) \end{aligned}$$

Pleochroism

- X = greenish-blue
- Y = lavender-blue to brown-yellow
- Z = light greenish-yellow.

'Arfvedsonite is distinguished from all the other amphiboles except riebeckite, some omphacites and kataphorites by the orientation of the optic axial plane perpendicular to the symmetry plane.' (Deer, et al. 1967, 2, p. 372). Furthermore arfvedsonite is distinguished from riebeckite by the fact that it has a brownish colour in its pleochroic scheme and does not show complete extinction in white light. X-ray analyses were carried out on a composite grain of kaersutite -- arfvedsonite (Table XXI), two d-values are close to the values quoted for arfvedsonite by the American Society for Testing and Materials.

(ii) Barkevikite

Barkevikite occurs solely within the syenite-porphyr as subhedral to euhedral corroded phenocrysts between 0,6 millimetres and 2,4 millimetres long, which may be simply twinned.

The optical properties were determined to be:-

Colour dark brown

Pleochroism

X = light brownish-yellow

Y = red-brown

Z = dark reddish-brown

$Z\wedge c = 14^\circ - 17^\circ$

Bisschoff (1950, p. 14) determined $2V_x =$ approximately 35° .

(iii) Common hornblende

Under this heading only primary hornblende is considered. All hornblende which is interpreted as having been formed by the alteration of pre-existing pyroxene is dealt with under the heading of 'uralite' in the section on secondary rock forming minerals. The primary hornblende occurs within the andesite, equigranular diorite, dioritic-pegmatite, dioritic-aplite, syenite-porphyr and in lesser amounts in the coarse and fine pyroxene diorite and the pyroxene andesite.

The hornblende most commonly occurs as narrow lath shaped or rectangular crystals generally about 0,5 millimetres long, but being up to 2,7 millimetres long within the dioritic-pegmatite. Simple twinning parallel to (100) and a good cleavage parallel to (110) is present. The optical properties were determined to be as follows:-

$Z\wedge c = 16^\circ - 31^\circ$

$2V_x = 73^\circ$

Pleochroism

$$X > Y > Z$$

Pleochroic in various shades of green, yellow-green and brownish-green.

(iv) Kaersutite

This mineral occurs within the andesite, equigranular diorite and olivine gabbro. Kaersutite most commonly occurs as subhedral grains, roughly rectangular in shape with a length of less than 0,5 millimetres.

The optical properties were determined to be :-

$$ZAc = 11^{\circ} - 18^{\circ}$$

$$2V_x = 71^{\circ}$$

Colour yellow-brown to reddish-brown.

Pleochroism

X = pale brownish-yellow

Y = reddish-brown

Z = dark reddish-brown

In an effort to distinguish between kaersutite and basaltic hornblende a small quantity of brown amphibole was drilled from two composite grains of kaersutite-arfvedsonite in a thin section and X-rayed. The interplanar spacings (d-values) obtained are listed in Table XXI.

TABLE XXI

Comparison between the d-values obtained for the composite kaersutite-arfvedsonite grains and the d-values quoted by the American Society for Testing and Materials for kaersutite and arfvedsonite

Values obtained from composite grains				American Society for Testing and Materials values			
Sample K 1		Sample K 2		Kaersutite		Arfvedsonite	
Intensity	d-value	Intensity	d-value	Intensity	d-value	Intensity	d-value
		40	3,7416	8	3,88	18	3,883
20	3,7047						
20	<u>3,3681</u>	20	<u>3,3606</u>	65	<u>3,36</u>	45	3,42
		100	<u>3,1617</u>	50	<u>3,26</u>	100	<u>3,161</u>
30	3,1335			80	3,11		
		30	<u>2,9857</u>	45	<u>2,93</u>	16	<u>2,991</u>
		60	<u>2,7281</u>	12	<u>2,726</u>	80	<u>2,732</u>
100	<u>2,7040</u>			100	<u>2,693</u>		
		70	2,5293	65	2,548	25	2,550
20	2,5224						
15	2,3212	15	<u>2,3363</u>	40	<u>2,337</u>	2	2,339
15	2,2791			45	<u>2,153</u>	35	2,185
		10	<u>1,4750</u>	6	<u>1,470</u>		
20	<u>1,4395</u>			55	<u>1,439</u>		
10	<u>1,3436</u>			Plus 18 lines to 0,778			

The d-values which are correlated in each row, are underlined.

From the data presented in Table XXI it can be noted that most of the interplanar spacings of the brown amphibole samples K 1 and K 2 are close to the values for kaersutite. Another indication of the fact that this amphibole is kaersutite and not basaltic hornblende, is that it is present in a wide variety of rock types and according to Deer *et al* (1967, 2, p. 315) the occurrence of basaltic hornblende is confined mainly to lavas. In a polished section cut from the same sample of equigranular diorite, which a thin section showed to contain kaersutite, the most abundant ore mineral present was almost pure ilmenite (Plate IV, 2) which forms 3 per cent of the volume of the rock. It seems likely that a melt containing an abundance of titanium, as evidenced in the amount of ilmenite present, would crystallise a titanium-rich amphibole such as kaersutite.

Bisschoff, (1950, p. 14) describes a brown amphibole with good (110) cleavage and strong pleochroism (from yellow-brown to red-brown) as occurring within the andesite and diorite. This mineral is probably kaersutite.

d. Olivine

Although pseudomorphs after olivine have been found within the equigranular diorite (Plates XIII, 2 & 3), the only fresh unaltered grains of olivine are found within the olivine gabbro. The olivine occurs in well rounded grains poikilitically enclosed within other minerals. These grains are between 0,1 millimetres and 0,6 millimetres long. One direction of cleavage may be weakly developed.

In thin section the olivine is colourless and has $2V_z = 93^\circ$. The composition of the olivine, determined by X-ray analyses and reference to the diagrams of Yoder and Samaha (1957) and Jambor and Smith (1964), is Fe_{81} .

e. Biotite

Two varieties of biotite are present within the rocks of the Roodekraal Igneous Complex, a red-brown variety which may be either primary or secondary in origin, and an olive-green to pale brown variety which only forms as an alteration product. The primary red-brown variety is dealt with here, and the other varieties are dealt with under the heading 'Secondary Rock Forming Minerals'. The red-brown variety of biotite is present in the equigranular diorite, coarse and fine pyroxene diorite and syenite, as a secondary mineral. In the olivine gabbro, however, it occurs as large grains which show no signs of being an alteration product, and it is these primary grains which are dealt with here. The

biotite grains are anhedral ranging from 0,1 - 2,0 millimetres in length, small grains are often present around the edges of early formed magnetite crystals, while large grains are interstitial to the other rock forming minerals including the feldspar. The biotite has a very prominent (001) cleavage.

The optical properties of the primary biotite are as follows:-

$$2V_x = 0^\circ$$

$$Z\Lambda a = 9^\circ$$

Pleochroism

X = pale yellow

Y = Z = dark brownish-red

f. Quartz

Quartz is only rarely present as a primary mineral, and then only within three rock types, namely the syenite, a small portion of the equigranular diorite lying to the west of the main road from Potchefstroom to Parys and the dioritic-pegmatite. The grains are small, anhedral and generally less than 0,5 millimetres in diameter. The quartz is colourless, but is charged with many inclusions, notable acicular and columnar crystals of apatite and actinolite.

In the rock types in which it occurs the quartz is obviously the last primary mineral to form, since it is interstitial and only present in small grains.

g. Apatite

Apatite is present in all the rock types except the syenite and attains its best development in the coarse pyroxene diorite and the dioritic-aplite. The apatite occurs as small euhedral, columnar crystals, generally only visible under the microscope, which have pyramidal terminations. These crystals are abnormally large in the dioritic-aplite and coarse pyroxene diorite, reaching a maximum length of 5 millimetres in the pyroxene diorite.

In thin section the apatite is colourless to blue-grey. Bisschoff (1950, p. 17) determined the largest refractive index i.e. ω to be 1,632 and therefore concluded that the mineral was fluorapatite. Two varieties of inclusions are present within the apatite crystals. One variety is present in small elongated grains which are reddish-brown to yellow in colour, translucent, have straight extinction and a higher refractive index than the apatite. These grains are considered to be zircon. The other type of inclusion is dark red-brown in colour and appears to be a secondary mineral since it is always present in cracks near the edges

of the grains.

h. Magnetite and ilmenite

Both of these minerals are present in proportionately large amounts in the rock types of the Complex. They are considered by the author to be primary and thus their origin is not related to that of the sulphide bodies. The composition and relationship with other minerals will be dealt with in detail under the section on ore minerals. These two minerals generally occur in subhedral to euhedral grains, enclosed within the silicate minerals. The average grain size is 0.15 millimetres with a maximum grain size of 0.60 millimetres. An exceptional occurrence is where the magnetite forms an interstitial groundmass enclosing the early formed grains of olivine in the olivine gabbro (Plate XIII, 1).

i. Sphene

Sphene is present as a primary mineral in all the rocks of the Complex except the amygdaloidal andesite, dolerite, syenite, syenite-porphry and olivine gabbro. It is present in abnormally large quantities in places, forming 5 per cent of the total volume of the rock in the massive andesite at the southern extremity of the Complex. Sphene forms as a secondary mineral by the replacement of amphibole. The sphene grains are subhedral and rounded, with the typical rhombic cross section only rarely observed.

The optical properties are as follows:-

ZAc	=	51°
Colour		Colourless to pale brown
Pleochroism		
X	=	colourless
Y	=	yellow
Z	=	pale brown

Bisschoff, (1950, p. 18) determined $2V_z$ to be $20^\circ - 25^\circ$.

The primary sphene is seen to enclose ilmenite and magnetite and thus must have been formed after the crystallisation of these minerals.

2. Secondary Rock Forming Minerals

Under this heading all minerals formed by alteration of the minerals already dealt with will be discussed. The secondary minerals may have been formed either by the process of deuteric alteration or by hydrothermal alteration associated with the formation of sulphide deposits.

a. Albite

Albite is present in zones around the margins of primary plagioclase crystals or as individual grains. Albitization of the feldspar has taken place in all the rocks of the Complex except the olivine gabbro and is not confined to the areas around sulphide bodies. The anorthite content of the albite varies between An_0 and An_9 . This variation does not appear to be related to distance from the sulphide bodies.

Samples were X-rayed to determine whether the plagioclase (both primary and secondary) is of the high or low temperature form. The results of the X-ray analyses are plotted in Figure 4. From the measurements of 2θ presented in Figure 4 it appears that at least one of the samples analysed is of the high temperature form. Optical methods used to check this result indicated, however, that all the plagioclase is of the low temperature variety. The anomalous result obtained from X-ray analysis is probably due to interfering peaks of another mineral present in the sample analysed.

b. Uralite

This mineral occurs in all the rock types of the Complex except the olivine gabbro and syenite. In discussing uralite Deer et al. (1967, 2, p. 260) states 'This amphibole is usually considered to be actinolitic in composition and to be derived by the pneumatolytic action of the later water-enriched magmatic fluids on the earlier crystallised pyroxenes'.

The uralite has optical properties which are intermediate between common hornblende and actinolite. Bisschoff (1950, p. 12) called the uralite 'actinolitische hornblende'. The writer, however, prefers to use the term 'uralite' since it has a genetic significance which will distinguish this mineral from primary hornblende which is partially replaced by secondary actinolite. Optical properties as determined by Bisschoff (1950), with which the author is in agreement are:-

$$\begin{aligned} 2V_x &= 68^\circ \\ ZAc &= 18^\circ - 19^\circ \\ \left. \begin{aligned} \beta &= 1.64 \\ \gamma &= 1.66 \end{aligned} \right\} &\text{on (110)} \end{aligned}$$

Colour Green in hand specimen

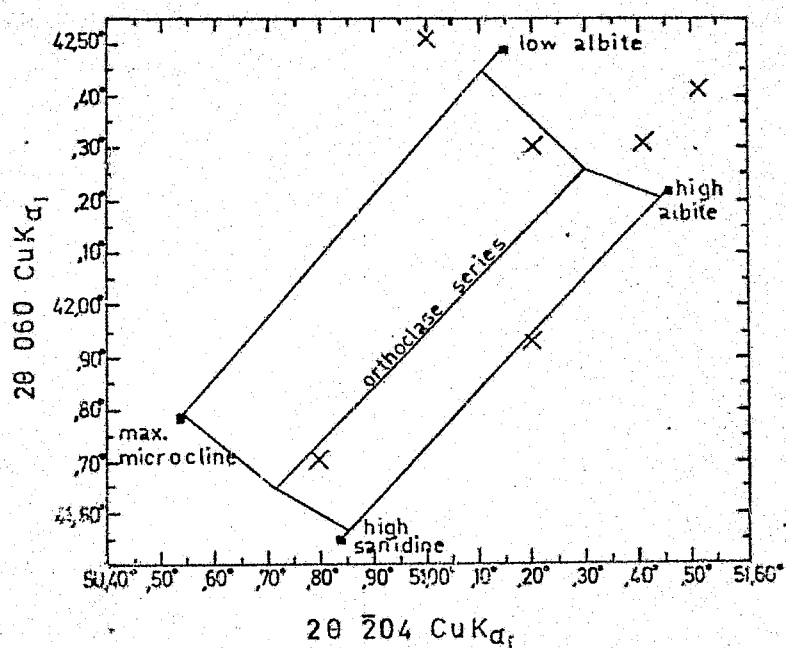
Pleochroism

X = pale brown to yellow

Y = pale green

Z = apple-green

Fig. 4 Diagram showing $2\theta\ 060$ versus $2\theta\ 204$ for the alkali feldspar present at Roodekraal and the various alkali feldspar series. (Modified after Wright 1968, p.91, Fig. 2)



X Alkali feldspars of the Roodekraal Igneous Complex

This amphibole may contain small inclusions of apatite and a mineral regarded by Bisschoff (1950, p. 13) as being xenotime. The uralite alters to secondary actinolite and sphene along its edges and cleavage planes, and in the vicinity of sulphide zones to biotite of the olive-green to pale brown variety.

c. Olivine pseudomorph (bowlingite)

Bowlingite is very variable in composition. Caillère (1935) and Caillère and Hénin (1951) regard bowlingite to be a saponite, whereas Wilshire (1958) considers that it consists of smectite-chlorite together with antigorite and/or chrysotile. The pseudomorphs after olivine are found within the equigranular diorite and pyroxene diorite (Plates XIII, 2 & 3).

Microscopic features

The mineral aggregates which are thought to be bowlingite are observed to occur either partially or completely enclosed within augite grains (Plate XIII, 3). The pseudomorphs are composed of a pale yellow-green to greenish-brown mineral which often occurs in radiating structures with a fibrous appearance and a moderate birefringence. The pseudomorphs have a similar elliptical or circular shape to the olivine grains of the olivine gabbro and are often rimmed with ore minerals. The rim of ore minerals was probably formed by the iron, which is liberated in the conversion of olivine to bowlingite, and is ejected to the edge of the pseudomorph. It was decided that the pseudomorphs are composed of bowlingite from the facts that more than one component is present, the birefringence is too high for chlorite or serpentine, it is green and has a refractive index which may be less than 1.54, whereas the other common pseudomorph of olivine, iddingsite, is generally reddish-brown with a refractive index which is always greater than 1.54.

d. Actinolite

Occurs in all the rock types of the Complex except olivine gabbro, syenite-porphyry, syenite and dolerite. It is especially abundant in the lavas and is concentrated within the amygdalae. The actinolite is not concentrated in zones of late-stage hydrothermal alteration. Secondary actinolite occurs as fine acicular crystals replacing the mafic minerals along their edges and cleavages, and forming outgrowths from them. Actinolite crystals may be up to 1.8 millimetres long and are enclosed by albite and secondary quartz crystals.

Optical properties of actinolite are:-

$$ZAc = 11^{\circ} - 19^{\circ}$$

Biaxial - ve

Colour Pale green

Pleochroism

X = straw-coloured to colourless

Y = pale yellow-green

Z = pale green

Individual crystals are too small to allow the accurate determination of 2V.

The actinolite was X-rayed using the powder method. The results are given in Table XXII.

TABLE XXII

Comparison between the d-values obtained for a fibrous green amphibole and the d-values which are nearest in magnitude, as quoted by the American Society for Testing and Materials

Fibrous green amphibole		American Society for Testing and Materials values for actinolite	
Approximate Intensity	d-value	Intensity	d-value
80	3,1335	80	3,11
50	2,9702	60	2,94
100	2,7152	100	2,71
80	2,5432	60	2,53
40	2,1761	40	2,16
40	1,6554	40	1,65
30	1,5163	40	1,58
50	1,4442	40	1,43

From the data presented in Table XXII it can be observed that the d-values are close to the values given for actinolite by the American Society for Testing and Materials.

e. Quartz

This mineral is present as a secondary mineral in the andesite, the equigranular diorite, pyroxene andesite, dioritic-pegmatite and aplite, dolerite and syenite-porphyry. The secondary quartz is present as anhedral grains up to 0,25 millimetres in diameter within the ground-mass, generally interstitial to the other minerals. It is also present

as euhedral grains within open spaces and as irregular masses of chalcedony. The quartz is colourless but contains numerous inclusions, mainly acicular crystals of apatite and actinolite. The quartz may fill 'boxworks' of hematite so that the composite structure has the appearance of jasper in a hand specimen.

X-ray analyses were carried out on a grain of secondary quartz from the pyroxene andesite. The grain chosen was 0,15 millimetres in width and contained actinolite crystals. Of the fifteen reflections measured, nine matched α -Quartz when their d-values were compared with the card index file of the American Society for Testing and Materials.

f. Calcite

Calcite is present in all the rocks of the Complex except the dike rocks and the olivine gabbro. It occurs mainly as amygdale and fissure fillings, but may also occur as single isolated grains within the groundmass of the amygdaloidal andesite or the equigranular diorite. The calcite is concentrated within the amygdaloidal andesite, but does not appear to be more abundant in the vicinity of the mineralized zones.

The calcite is often present as a mosaic of interlocking grains which are generally less than 0,4 millimetres in width. The grains are anhedral and vary greatly in size in a single amygdale or fissure. Occasionally, within the larger fissures, rhombohedral crystals of calcite are found with a length of up to 30 millimetres. The calcite has perfect (10 $\bar{1}$ 1) cleavage and commonly exhibits polysynthetic twinning.

g. Mica

Mica has been formed in fair abundance as a hydrothermal alteration product.

(i) Biotite

As detailed under the section entitled 'Primary Rock Forming Minerals', two varieties of biotite are present. There is a red-brown variety which may be of primary or secondary origin and an olive-green to pale brown variety which is exclusively a secondary mineral.

Red-brown biotite

This mineral has been discussed as a primary mineral and only secondary occurrences are discussed here. As a secondary mineral it occurs in the equigranular diorite, the coarse and fine pyroxene diorite and the syenite. The optical properties are the same as for the primary variety and for this reason it must have a similar

composition. The crystals of the secondary variety are, however, much smaller. This mineral is observed to replace the augite and amphibole, and is not restricted in its occurrence to zones of hydrothermal alteration.

Olive-green to pale brown biotite

This is present in fine vermicular intergrowths replacing pre-existing mafic minerals, especially amphibole, and is associated with other alteration products such as actinolite. The vermicular masses are made up of fine-grained flaky aggregates. Individual flakes being less than 0,2 millimetres in length.

Optical properties are :-

$2V_x$ is approximately 0°

Pleochroism

X = olive-green

Y = Z = pale brown

The biotite was X-rayed using the powder method. When the d-values were compared with the card index of the American Society for Testing and Materials, seven of the ten reflections measured were close to those of biotite.

(ii) Green phyllosilicate

This mineral is developed in a limited number of localities, firstly within the equigranular diorite wall rock adjacent to the diorite-pegmatite, secondly in the innermost zone of hydrothermal alteration around the sulphide bodies and finally in calcite veins. The green phyllosilicate occurs in irregularly shaped anhedral grains about 0,3 millimetres in length or as myrmekitic masses of smaller grains which often contain finely disseminated ore minerals. The green phyllosilicate is interstitial to the primary rock forming minerals and calcite, but may be replaced by the olive-green to pale brown biotite.

The optical properties are as follows:-

$2V_x = 43^\circ$

$Y\wedge a = 1,5^\circ$

$X\wedge c = 4^\circ$

Colour apple-green

When the d-values of this mineral, obtained from X-ray photographs, are compared with the card index of the American Society for Testing and Materials in Table XXIII a good match is obtained for muscovite.

TABLE XXIII

Comparison of d-values obtained for the green phyllosilicate and the d-values of muscovite

Green phyllosilicate		Values quoted by the American Society for Testing and Materials of muscovite	
Approximate Intensity	d-value	Intensity	d-value
Broad air diffraction peak		100	3,320
20	2,7281	21	2,789
60	2,5686	54	2,566
40	2,3575	27	2,384
10	2,1994	4	2,189
2	1,8900	2	1,894
10	1,8124	3	1,822
10	1,6908	4	1,699
10	1,5813	3	1,573
5	1,3668	2	1,375
10	1,2858	6	1,292

The mineral has the composition of muscovite, and is always green in colour, it seems probable that it is fuchsite. Whitmore et al. (1946) notes that 'the paragenesis of chromium bearing muscovite has shown that they occur most commonly in hydrothermal carbonate replacement zones', which is in agreement with the environment of the green phyllosilicate at Roodekraal.

(iii) Sericite

This mineral is present in minor amounts within the zones of hydrothermal alteration and may have been formed by two processes. It may

have formed as a hydrothermal alteration product of the feldspars. The fact that this reaction may take place was demonstrated by Norton (1939) who synthesised sericite by heating feldspars and other aluminosilicates in CO_2 charged water at 300°C . (quoted in Deer et al. 1967, 3, p. 19). Sericite may also have been formed at a lower temperature from the potash liberated when biotite was transformed to chlorite by the hydrothermal solutions (Chayes, 1955).

h. Epidote

The epidote is present in all the rock types of the Complex except the olivine gabbro. It is especially abundant in the south of the Complex within the fine pyroxene diorite, pyroxene andesite and amygdaloidal andesite. The epidote occurs either as a fine-grained aggregate coating the primary minerals, especially the aluminous ferromagnesian minerals, or in amygdales as elongated euhedral crystals up to 2.0 millimetres long. The larger grains have a good cleavage which is probably parallel to (001) (Bisschoff, 1950, p. 16).

Other optical properties as determined by Bisschoff (1950, p. 16) and with which the author is in agreement are:-

Pleochroism

X = colourless

Y = greenish-yellow

Z = pale yellow

$$2V_x = 68^\circ - 69^\circ$$

$$\Delta_{Ac} = 4^\circ - 5^\circ$$

$$\beta = 1.765$$

$$\gamma = 1.780$$

From these values Bisschoff derived a composition of :-

38 per cent $\text{HCa}_2 \text{Fe}_3 \text{Si}_3 \text{O}_{13}$

62 per cent $\text{HCa}_2 \text{Al}_3 \text{Si}_3 \text{O}_{13}$

However, this is not in agreement with the value obtained by the author, of 32 Mol. per cent $\text{Ca}_2(\text{Mn}^{+3}, \text{Fe}^{+3}) - \text{Si}_3 \text{O}_{12} (\text{OH})$ from Deer et al. (1967, 1, Fig. 47).

i. Chlorite

Chlorite occurs in all the rocks of the Complex except the olivine gabbro. At least two varieties of chlorite are present.

Variety (i)

A bright green chlorite with abnormal blue to red interference colours is found concentrated in the vicinity of the mineralized zones. It is present as small irregular flakes generally less than 0.2 millimetres in length. Otherwise it may occur in a radiating spherulitic structure.

The optical properties are as follows:-

Biaxial - ve with small 2V. Chlorite flakes are too small to allow an accurate determination of 2V.

Pleochroism

X = pale greenish-yellow

Y = dark green

Z = olive-green

These properties indicate that the chlorite is an oxidised variety. This chlorite was X-rayed and the d-values obtained are listed in Table XXIV

TABLE XXIV

Comparison between d-values obtained for a bright green chlorite variety (i) and the corresponding d-values of Bavalite (a variety of daphnite) quoted by the American Society for Testing and Materials

Chlorite variety (i)		American Society for Testing and Materials values for Bavalite	
Approximate Intensity	d-value	Intensity	d-value
40	14,0170	60	14,0
100	7,0527	100	7,08
30	4,7061	30	4,681
50	3,5282	50	3,523

From the data presented in Table XXIV it can be observed that a good match is obtained for Bavalite (a variety of daphnite), which is an oxidised chlorite.

Variety (ii)

This mineral has a wider distribution being present as a coating on mafic minerals in all the rock types except the dike rocks and the olivine gabbro. It has a pale green to dark green colour, a low birefringence, green interference colours and straight extinction. This variety of chlorite is not restricted to zones of hydrothermal alteration.

j. Titanium-rich minerals

(i) Sphene

The sphene is present as a replacement of 'uralite' and occurs as sub-hedral crystals aligned parallel to the cleavages of the amphibole. This type of alteration has been mentioned under the section 'Primary Rock Forming Minerals', and was noted by Bisschoff (1950, p. 18). The sphene also occurs within the amygdaloids of the andesite.

(ii) Rutile

This mineral is present as a supergene alteration product of pre-existing ilmenite and will be dealt with in detail under the section on ore minerals. It occurs in all rock types containing ilmenite, but is especially abundant in the pyroxene andesite, and the equigranular diorite.

(iii) Leucoxene

The leucoxene is formed as an alteration product of the ilmenite, and appears to be composed of finely crystalline rutile. The leucoxene is particularly abundant within the dioritic-aplite.

k. Hematite

The hematite is present within the amygdaloidal andesite, fine pyroxene diorite and dioritic-pegmatite. Within the amygdaloidal andesite and pyroxene diorite it occurs as the earthy variety, but within the dioritic-pegmatite it is present as irregularly shaped plates with a width of 0.2 millimetres. The hematite appears to form as an alteration product of the primary hornblende within the dioritic-pegmatite (Plate XV, 4) and as rims around pre-existing mineral grains within the andesite.

The optical properties of the hematite are detailed below.

Colour	blood-red
Pleochroism	orange to blood-red.
Uniaxial negative interference figure.	

A sample of the mineral was X-rayed and gave a fair match with hematite when compared with the card index file of the American Society for Testing and Materials.

l. Zeolite

Zeolite has only been found within the amygdaloidal andesite in the zones of late-stage hydrothermal alteration. The zeolite is present as a

replacement of the feldspar in the 'bleached' zones of amygdaloidal andesite, which will be discussed under the section entitled 'Hydro-thermal Alteration'. Elsewhere it occurs in amygdaloids and calcite veins as an open space filling. The zeolite occurs as fibrous masses with a radiating structure, or as roughly circular grains about 0,3 millimetres in diameter with a radiating extinction.

The optical properties of the zeolite are as follows:-

$$2V_z = 76^\circ$$

$$ZAc = 13^\circ - 17^\circ$$

Colour Pink in hand specimen but colourless in thin section.

The zeolite has a negative relief in thin section.

The mineral was X-rayed and the results obtained are presented in Table XXV.

TABLE XXV

Comparison between d-values obtained for a zeolite and the corresponding d-values of Sodium Aluminium Hydrate (Zeolite, Series P) as quoted by the American Society for Testing and Materials

Zeolite		American Society for Testing and Materials values for Sodium Aluminium Hydrate (Zeolite, Series P)	
Approximate Intensity	d-value	Intensity	d-value
40	3,73	40	3,88
50	3,36	60	3,35
100	3,19	100	3,19
50	2,94	80	2,93
40	2,67	60	2,68
40	2,54	60	2,51
20	2,42	5	2,42

From the data presented in Table XXV it can be noted that the d-values of the zeolite and the quoted values for zeolite, Series P, correspond reasonably closely. The optical properties determined also indicate that this mineral is phillipsite.

D. Distribution and Structure of the Rock Types

Referring to the map (Plate XXIV) it can be observed that the Roodekraal Igneous Complex as a whole is roughly elliptical, with the long axis

aligned in a northeast to southwest direction, parallel to the strike of the surrounding sedimentary rocks.

The Transvaal System rocks surrounding the Roodekraal Igneous Complex are Magaliesberg Quartzite along the northern and western sides, Daspoort Quartzite and Daspoort Shale in the northeast in squares F 2 and F 3 of Plate XXIV, and Ongeluk volcanic rocks in the southeast in squares F 3, F 4, G 3 and G 4 of Plate XXIV. Diabase flanks the Complex for a short distance along its northern boundary in square D 1 of Plate XXIV. The Magaliesberg Quartzite along the northern and western margins of the Complex generally dips fairly steeply towards its centre, suggesting that these sedimentary rocks form a basin in which the rocks of the Roodekraal Igneous Complex lie. The Daspoort Quartzite and Daspoort Shale dip steeply towards the northwest, while the diabase occurs as a concordant intrusion between the Magaliesberg Quartzite and Daspoort Quartzite, as shown in Plate XXIII.

Within the Complex itself it can be observed that the lavas completely encircle the intrusive rock types. Massive andesite is found in the southern and southeastern portions of the Complex while amygdaloidal andesite occurs over most of the remaining predominantly soil covered areas, and the pyroxene andesite occurs in many scattered outcrops which, taken individually, are smaller than the outcrops of massive and amygdaloidal andesite. The lava flows are roughly parallel to the present day surface, as shown by the attitude of the pyroxene andesite and the coarse-grained andesite in Plates XVIII, XIX and XX. Since the lava flows are more or less horizontal and the amygdaloidal andesite occupies the highest point of the Complex while the massive andesite occupies the lowest portions, it can be concluded that the amygdaloidal andesite probably overlies the massive andesite. Where the amygdaloidal and massive andesite are in contact in the field, as for example in square E 4, it does appear that the massive andesite is overlain by the amygdaloidal andesite. The lavas lie directly on top of and in contact with the sedimentary rocks of the Transvaal System. The massive andesite is in contact with the Daspoort Quartzite in the southeast while the amygdaloidal andesite lies against the Magaliesberg Quartzite along the northern and western edges of the Complex (Plate XXIII).

In the northern portion of the Complex where the amygdaloidal andesite attains its greatest thickness, the drilling program has shown the existence of certain lava flows which are coarser grained than the average, and may in certain cases be quite as coarse as the equigranular diorite. In general these zones are discontinuous and vary in width, although the

two best developed zones are of a fairly uniform thickness and can be traced over several hundred metres. These two zones have been used as marker horizons to determine the structure within the andesite. The most important structure within the andesite is the development of a fairly large normal fault which has a curved fault plane dipping steeply to the south. A smaller antithetic fault dipping steeply to the north is developed so that the block between the two faults is downthrown. These faults are of prime importance since the major sulphide bodies are localised along them. In general only the two faults mentioned above are developed, with the larger sulphide body lying along the larger fault. This structure can be seen on Plates XVIII and XIX. On Plate XX however, a slightly more complex structure exists where a synthetic fault is developed between the main fault and the antithetic fault. As a result of the edges of the block between the main fault and the synthetic fault tending to be moved in opposite directions the block has been rotated (see Plate XX). This rotation imparted on a block lying between two similar faults has been described by De Sitter (1964).

A narrow zone of lava which is present in boreholes drilled in the northern portion of the Complex contains many irregularly shaped calcite filled amygdaloids, and wheel like structures of cryptocrystalline quartz and hematite (Plate II, 2). This structure probably represents a pumiceous zone which had secondary minerals deposited within it.

The volcanic agglomerate is found in the northern portion of the Complex within the amygdaloidal andesite in square D 2 of Plate XXIV. The volcanic breccia outcrops are in the extreme north (in square D 1 of Plate XXIV) and in the extreme south (in square D 5 of Plate XXIV) of the Complex. There are two main occurrences of the volcanic breccia in the northern portion of the Complex. Firstly it outcrops as a narrow lenticular body striking approximately northeast to southwest in square D 1 of Plate XXIV and secondly it occurs below the surface in boreholes drilled on section 3 (Plate XX) where it is up to fifteen metres in width. The underground intersection of the breccia body is 775 metres to the southwest of its outcrop. The breccia body in the extreme south of the Complex occurs within amygdaloidal andesite and has a syenite dike intruded along its southern contact. These two occurrences of volcanic breccia are composed of rock fragments present within a matrix of fine-grained andesite. The other variety of volcanic breccia which is cemented by secondary minerals is present as narrow zones mainly within the breccia bodies discussed previously.

Outcrops of intrusive rock types are mainly confined to two roughly parallel axes which trend northeast to southwest and which are marked on

the map (Plate XXIV). The northern axis is well developed and has the greater proportion of the intrusive rocks of the Complex present along it, while the southern axis is less well defined (see Plates XXIII and XXIV).

The intrusive rock type which outcrops over the greatest area is the equigranular diorite, which has sharp contacts with both massive and amygdaloidal andesite and is present as narrow sills intruded into the amygdaloidal andesite on either side of the northern axis of intrusive rocks (see Plates XVIII, XIX, XXII and XXIII).

An unusual feature is the presence of rheomorphic veins, as recorded by Bisschoff (1950, p. 33), within the equigranular diorite to the north of the most southerly outcrop of coarse pyroxene diorite in square E 4 of Plate XXIV.

The fine pyroxene diorite is present as a continuous outcrop in the northern zone of intrusive rocks, stretching along its southern margin, and also occurs as isolated outcrops within the southern zone of intrusive rocks. The fine pyroxene diorite has sharp contacts with the equigranular diorite and is chilled along these contacts. A good example of such a chilled contact is present on the eastern portion of the quarry face in square C 3 of Plate XXIV.

Pyroxene andesite is present as irregularly shaped outcrops scattered over almost the entire Complex. It forms a flat lying 'cap rock' over the dioritic rocks and grades imperceptibly into the fine pyroxene diorite which is the intrusive phase of this rock type. Boreholes drilled in squares D 1 & 2 of Plate XXIV demonstrated that these two rock types are interconnected and have gradational contacts.

The coarse pyroxene diorite intrudes all the rock types described above, the main outcrops being in the northern axis of intrusive rocks. Of the intrusive rock types the coarse pyroxene diorite has the smallest outcrop area. Although the coarse pyroxene diorite has sharp contacts with the rocks it intrudes it invariably contains large feldspar phenocrysts up to 24 millimetres in length which are often aligned parallel to the contacts of the bodies, even where it forms narrow dikes as for example in square D 3 of Plate XXIV. In the northeast of the Complex in square F 2 a sill of coarse pyroxene diorite, which has a zone of olivine gabbro along its lower contact, is present below the surface (see Plate XXI). There is no sharp contact between the coarse pyroxene diorite and the olivine gabbro.

Several types of dike rock are present which cut across the other rock types of the Complex. Two narrow dolerite dikes, both less than one metre in width, occur within the fine pyroxene diorite and exhibit strongly chilled margins. These dikes strike northeast to southwest and lie a short

distance to the north of the most southerly outcrop of coarse pyroxene diorite in square E 4 of Plate XXIV.

An interesting pair of rock types are the dioritic-pegmatite and dioritic-aplite. The pegmatite is restricted in its occurrence to the equigranular diorite and the coarse pyroxene diorite. The largest pegmatite veins have been shown by drilling operations to be present a short distance above the contact of the equigranular diorite with the underlying coarse pyroxene diorite, in the northeastern portion of the Complex. Small irregular pegmatite and aplite veins occur within the equigranular diorite on the northeastern side of the road from Potchefstroom to Parys in square D 5. Narrow aplite veins also occur in the most southerly outcrop of the coarse pyroxene diorite in square E 4. The pegmatites and aprites are not present as pipe-like bodies and appear to be fissure fillings. Since the pegmatite veins do not contain xenoliths of the wall rock, and since they have matching irregularities in the wall rocks on either side, they probably represent dilation vein fillings.

Syenite dikes are concentrated along the northern zone of intrusive rocks. They generally have a northeast to southwest strike, and may dip steeply to the northwest or southeast. The dikes are usually less than one metre in width, but the largest dike which outcrops in square D 2 of Plate XXIV has a width of five metres. The syenite dikes cut across all the abovementioned rock types but were emplaced before the period of sulphide mineralization.

Pseudotachylite is present throughout the Complex and intrudes the surrounding sedimentary rocks. Most commonly it occurs as a network of veins which constantly vary in both direction and width, with a maximum width of 150 millimetres, but it also occurs, as shown by drilling operations (in square D 2, Plate XXIV), as a solid band, or series of bands, in places up to 450 millimetres in thickness, intruded along the top contact of the diorite sill within the amygdaloidal andesite. The pseudotachylite bands do not always occur exactly on contact of the two rock types, have a limited lateral extent, since the diorite to andesite contact elsewhere is free of pseudotachylite, and have a network of fine pseudotachylite veinlets extending for up to a metre into the wall rock on either side of them. Bisschoff (1972b) noted the presence of fragments of pseudotachylite within an agglomeratic rock in which the fragments are cemented by lava. The author, however, has only observed fragments of pseudotachylite in volcanic breccias where the fragments are cemented by secondary minerals.

The syenite-porphry dike is about 1,3 metres in width and is present in the extreme south of the Complex, on the west of the main road from Potchefstroom to Parys, in square E 6 of Plat XIV. It is emplaced within massive andesite, has an east to west strike and dips steeply to the south. It has a chill zone of up to 50 millimetres in width and is the only rock type which has not been intruded by pseudotachylite.

E. Hydrothermal Alteration

A distinction must be made between deuteric alteration which was produced by residual post-magmatic solutions as a direct consequence of the consolidation of the parent magma, and strictly secondary changes caused by a later period of alteration associated with ore deposition. The varieties of alteration which are regarded as deuteric are not limited in their distribution to zones around sulphide bodies, but occur in all the rock types except the olivine gabbro and syenite-porphry, and are especially abundant within the lavas. The hydrothermal alteration, which is thought to be strictly related to the phase of sulphide mineralization, includes all alteration effects which occur solely in the vicinity of the sulphide bodies, usually forming an envelope around the zone of sulphide enrichment.

1. Deuteric Alteration

This form of alteration takes place at high temperatures and is in effect an integral part of the formation of the rock, caused by the rock 'stewing in its own juice' (Hess, 1933). Deuteric-stage fluids have been very active at Roodekraal, especially in the lavas. Most of the rock forming minerals have been altered and open spaces such as vesicles, fissures and fault zones have been filled with secondary minerals. Alterations undergone by the rock forming minerals include:-

- (i) Extensive albitization of the pre-existing feldspar, with the formation of the low temperature (ordered) variety of albite. Alteration has in many cases gone further, with the formation of sericite and epidote.
- (ii) The augite has to a great extent been altered to urallite probably by pneumatolytic action associated with the later fluids. Uralite replaces the augite in every gradation between replacement along the edges and cracks (Plate XIV, 3) and complete pseudomorphous replacement.

- (iii) Olivine has been converted to a green pseudomorph (Plates XIII, 2 & 3) which is probably 'bowlingite' (see section entitled 'Mineralogy of Rock Forming Minerals').
- (iv) The amphiboles may alter to actinolite, reddish-brown biotite, sphene and chlorite. An unusual feature is present where the kaersutite has coronas of arfvedsonite around it (Plate XV, 1). This however need not be an alteration but could have been caused by the amphibole crystallising during the deuteric-stage and its composition being modified by the presence of soda-rich fluids.

The open spaces which may be filled by secondary minerals comprise vesicles within the lavas, miarclitic cavities within the intrusive rocks, fissures and cracks. Minerals which have been deposited in open spaces within the rocks by deuteric-stage fluids include albite, calcite, quartz and zeolite.

The calcite present in the lavas and the CO_2 content of the lavas show no systematic variation with distance from the sulphide bodies. This leads to the conclusion that the majority of the calcite was emplaced by deuteric-stage fluids and not by late-stage hydrothermal solutions. The quartz is present in several forms ranging from perfect euhedral crystals formed along the walls of fissures and amygdaloids to cryptocrystalline aggregates formed along fine cracks. The zeolite (var. phillipsite) has only been found within the lavas except for one locality, namely on the contact of the Daspoort Shale and the equigranular diorite.

As far as the composition of the deuteric-stage fluids is concerned, large amounts of Na_2O and smaller amounts of SiO_2 must have been present to account for the widespread albitization. During the process of albitization CaO and Al_2O_3 would be released. For the conversion of augite to hornblende Na_2O , Al_2O_3 and H_2O must be added and CaO removed. CaO is required to alter hornblende to actinolite and K_2O must be added to convert the hornblende to biotite. Because many of the alteration products of hornblende, e.g. biotite and chlorite, contain more hydroxyl groups than the hornblende, H_2O is necessary for their formation.

The CaO required could be provided by that released when augite is converted to hornblende or when a more basic plagioclase is altered to albite. Some of the K_2O might have been provided by the process of albitization, since plagioclase frequently contains small amounts of K_2O which would be released if the plagioclase were converted to low temperature albite, which would have a more ordered state than the original feldspar. Any Al_2O_3 and CaO excess derived from the alteration of the primary rock

forming minerals, would combine with the remaining solutions to form the secondary minerals which fill open spaces. The Al_2O_3 may have combined with Na_2O , SiO_2 and H_2O to form zeolites, while CaO probably combined with other compounds to form minerals such as epidote and calcite, and the excess SiO_2 could have been deposited as quartz.

From the above it can be concluded that the deuteric-stage fluids were probably water rich and contained large amounts of Na_2O and smaller amounts of SiO_2 and K_2O .

2. Late-stage Hydrothermal Alteration

This includes all alteration of gangue minerals by the solutions attending the sulphide mineralization. The andesites in the vicinity of the mineralized zones have two varieties of alteration, narrow bleached zones which are scattered within the mineralized area, and the zones of concentric alteration around the sulphide bodies.

a. Narrow bleached zones

The bleached zones vary from 25 millimetres to three metres in width and occur in or near the sulphide body, generally being below it. The most important process involved in the bleaching of the andesite is the introduction of secondary quartz and calcite and the breakdown and almost complete removal of the mafic minerals such as uraltite. Other minerals formed within the bleached zones are sericite and plumose aggregates of zeolite. Chlorite and actinolite are present in fair proportion with the chlorite being concentrated within the amygdalae. The bleached zones are often concentric about a vein which is filled with chlorite or calcite (see Plate III, 2). This leads to the impression that these zones were formed as a result of wall rock alteration around fissures which acted as channelways for hydrothermal solutions.

b. Concentric zones of mineral alteration around sulphide bodies

The zones of hydrothermal alteration are concentric about the sulphide body and will be dealt with in sequence from the centre outwards.

- (i) Mineralized zone :- The most important differences between the rock within the mineralized zone itself and that surrounding it, are that within the mineralized zone it is a darker grey and has a smaller grain size. The darkening is mainly caused by the presence of finely disseminated sulphides and secondary minerals such as olive-green to pale brown biotite, actinolite and chlorite. The feldspar has been recrystallised to small laths of albite,

generally between 0,15 and 0,25 millimetres long. Virtually no primary minerals have been preserved in this zone.

- (ii) Biotite-chlorite zone :- This zone has a width of about sixteen metres and forms an envelope around the mineralized zone. Both the original mafic minerals and the feldspar have been converted to secondary minerals. Biotite is abundant within this zone, which contains small amounts of a green-phyllisilicate up to about twelve metres away from the sulphide body. Chlorite is present in fair proportion and has abnormal blue to red polarization colours. It occurs within amygdalae replacing calcite and biotite and may replace mafic minerals and biotite present in the groundmass of the rock. Actinolite, secondary quartz, calcite, sericite, epidote and zeolite are also present within this zone.
- (iii) Biotite zone :- Biotite of the olive-green to pale brown variety is by far the most abundant product of hydrothermal alteration in the vicinity of the sulphide bodies. Roodekraal thus has a similar type of hydrothermal alteration to the porphyry copper deposits of Bingham, Ely, Bagdad and Ajo (Burnham, 1962), where muscovite is absent locally and biotite abundant.

The biotite zone exists from about sixteen metres from the sulphide body to more than 150 metres away. Actinolite is abundant in this zone in acicular crystals up to 0,30 millimetres long. Amphibole is preserved at distances of greater than sixty metres from the sulphide body where the amount of biotite present has diminished considerably. Pyrite occurs in fair abundance and probably owes its origin to the action of sulphur bearing fluids on iron originally present in the rock. Secondary quartz, calcite and zeolite may be present within veins and amygdalae.
- (iv) Outer zone :- Burnham (1962) divides hydrothermal alteration into two main facies, namely the phyllic and argillic facies. The phyllic facies of alteration is a high temperature facies, with muscovite and/or biotite being developed within it, while the argillic facies has a lower temperature of formation and abundant presence of clay minerals. At Roodekraal since the phyllic facies is well developed around the sulphide bodies we would expect the alteration to change gradually into the argillic

facies with increasing distance from the sulphide bodies. Unfortunately due to the fact that no systematic grid of boreholes has been drilled outside the phyllic facies of alteration it is not certain whether hydrothermal alteration of the argillic facies is present.

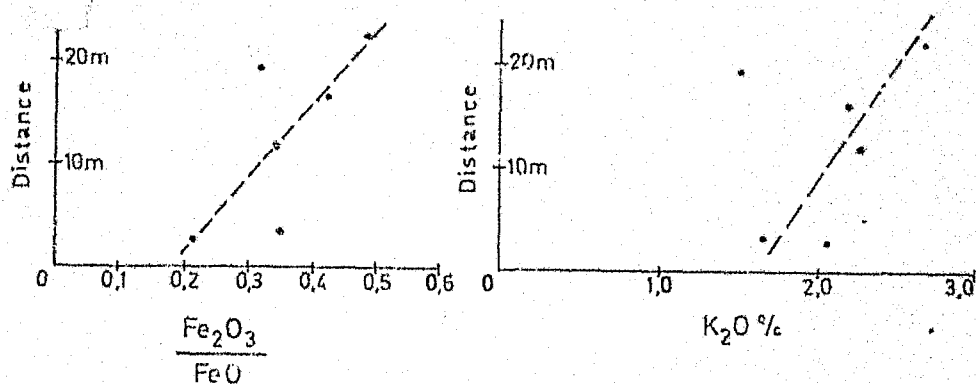
Sericite occurs in the outer fringes of the biotite zone and forms as an alteration product of the feldspar. A temperature of 340°C to 360°C is considered by Gruner (1944) to represent the boundary of stability between kaolinite and sericite. As kaolinite has not been identified at Roodekraal, since the mineral taken to be sericite is insoluble in HCl, it must be assumed that the sericite formed above 360°C. This is in agreement with the fact that the pyrrhotite in the immediate vicinity of sericite has an iron content which indicates a temperature of formation of approximately 430°C (see under section entitled 'Mineralogy of the Ore Minerals').

Epidote is present in the outer zone as a minor constituent, except where it is localized along fissures. The epidote is considered to have been formed by late-stage hydrothermal fluids associated with ore deposition. In certain localities (marked on the map i.e. Plate XXIV) the epidote has replaced the andesite and fine pyroxene diorite along narrow fissures. Epidote minerals probably crystallise between 300°C - (400°C - 500°C) (lower limit - Stringham (1952), upper limits - Ramberg (1949) gives 400°C and Rosenqvist (1952) gives 500°C, quoted in Deer et al. 1967, 1, p. 202).

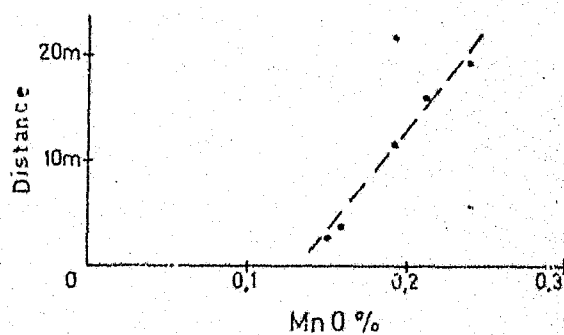
4. Concentric chemical alteration

Clear evidence of the existence of concentric alteration effects around the sulphide bodies is presented in Figure 5 where the percentage of MnO and K₂O and the ratio Fe₂O₃/FeO are plotted against distance from the centre line of the sulphide body. All these functions show an increase with distance from the centre line of the sulphide body. The increase in the 'oxidation ratio' i.e. Fe₂O₃/FeO with distance from the sulphide zone may be explained by the fact that over the distance plotted on the graphs, the biotite which contains ferric iron becomes more abundant with increasing distance from the sulphide body, while the sulphides contain iron in a low oxidation state. The increase in K₂O with increasing distance from the sulphide body is due to a

Fig. 5 Chemical variations with distance from sulphide body.



Distance is quoted in metres from the centre line of the sulphide body.



corresponding increase in the amount of biotite present. Finally the increase of the MnO content with distance from the sulphide body can be attributed to the fact that the mafic minerals which contain the most Mn have undergone intense alteration in the vicinity of the mineralized zone.

d. Chemical alteration caused by late-stage hydrothermal fluids

A comparison between the geochemistry of a sample of unaltered andesite and one in which extensive hydrothermal alteration has taken place should give an indication of the composition of the mineralizing fluids. In Table XXVI the analysis of RC 14, the sample furthest from the mineralized zone, and thus the least altered one, is compared with RC 5 which is the most highly altered sample and is in the mineralized zone.

TABLE XXVI

Comparison between standard cells of unaltered andesite and andesite which has undergone hydrothermal alteration

Cation	1 RC 14 (unaltered)	2 PC 5 (altered)	3 Difference 2 - 1
Si	49,46	47,61	- 1,85
Al	13,09	12,62	- 1,27
Fe ³⁺	5,64	1,83	+ 1,19
Fe ²⁺	9,44	9,49	+ 0,05
Mg	2,19	4,16	+ 1,97
Ca	3,48	4,44	+ 0,96
Na	6,96	7,86	+ 0,90
K	1,84	2,48	+ 0,64
H (hygroscopic)	0,96	0,92	- 0,04
H (combined)	10,82	9,28	- 1,54
C	5,53	5,52	- 0,01
Ti	1,00	1,34	+ 0,34
P	0,30	0,36	+ 0,06
Mn	0,06	0,12	+ 0,06
Algebraic Totals	106,57	108,03	+ 1,46
Proportion of minor elements present			
Cu %	0,02	0,31	+ 0,29
Sr ppm	436	283	- 153
Rb ppm	44	72	+ 28

ppm = parts per million

Copper analyses quoted were determined by Rio Tinto Exploration laboratories in parts per ten thousand.

It can be observed that Fe^{3+} , Mg, Ca, K, Na, Ti and to a lesser extent Fe^{2+} , P, Mn, Cu and Rb are present to a greater proportion in the altered rock than the unaltered rock. The higher content of Fe^{2+} , Fe^{3+} and Ti is thought by the author to be due to the presence of greater amounts of magnetite and ilmenite in sample RC 5 before alteration took place. The Mg, Mn, Ca, Na, K, P and Rb could all be derived by high temperature leaching of the wall rocks around the ore conduit by water-rich solutions. The fact that there has been no increase in water and a relatively large loss in Si is attributed to the high temperature mineralizing fluids causing the existing water in the rock to dissolve a certain amount of Si and migrate to cooler regions. The fact that there is no increase in the amount of CO_2 present could be due to gases such as H_2S and SO_2 being the dominant gas phases within the mineralizing fluids.

This leaves unexplained increases in Cu content, and of course sulphur since the copper is present as copper sulphides. These constituents can be regarded as having been emplaced by the mineralizing fluids.

IV. MINERALIZED ZONES

A. Structure of the Sulphide Bodies

The location of the sulphide bodies appears to have been controlled by structural features. The main occurrences of sulphides are found within the amygdaloidal andesite and generally lie between the pyroxene andesite above and the main coarsely crystalline andesite layer below. An equigranular diorite sill has been intruded along the contact of the lower coarsely crystalline andesite and the amygdaloidal andesite. It is intruded along both the lower contact, as in the case in Plates XVIII and XIX, and the upper contact as is the case in Plate XXII. Since this sill is not displaced it was intruded after the faulting took place and since it acts as a lower limit to the sulphide bodies, it was emplaced before the period of mineralization.

There are two distinct types of sulphide bodies. The more massive variety consists of a narrow central zone of bornite-chalcocite present along high angle normal faults (see Plates XVIII and XIX), in this zone actual replacement of the wall rock by the sulphides may have taken place. Surrounding the bornite-chalcocite zone is a more diffuse bornite-chalcopyrite zone and finally a chalcopyrite-pyrite-pyrrhotite zone, in these zones the sulphides have simply replaced calcite and filled available open spaces within the andesite. This variety of sulphide body exhibits 'reversed zoning' and is discussed in detail in the section on hydrothermal zoning. The second variety of sulphide body is less massive, the bodies being irregularly shaped and elongated parallel to the contacts between the amygdaloidal andesite, the overlying pyroxene andesite and the underlying equigranular diorite (see Plates XX and XXII).

In Plate XX the sulphide body is not localised along the faults but occurs in two distinct zones. The northern zone is concentrated along a volcanic breccia which probably provided an area of low pressure, while the southern zone is confined to the amygdaloidal andesite lying between the lower coarsely crystalline andesite and the pyroxene andesite. In this section it can be observed that the best copper sulphide mineralization is present in a 'halo' around the pyrrhotite body.

B. Mineralogy of the Ore Minerals

1. Ilmenite FeO . TiO_2

This mineral is recognised by its moderate anisotropism, pale brown colour, reflectivity and Vicker's Hardness. Reflectivity at 590μ using a Sika (Henry) standard is 19 per cent (Folinsbee 1949 gives 19,2 - 19,6 per cent). Vicker's Hardness i.e. VHM_6 is 699 (613 - 732) (Burke 1966 gives 585 - 739).

Ilmenite is present as an early formed, primary mineral, and occurs as skeletal grains within the gangue (Plate IV, 1) or as discrete grains which may contain exsolution lamellae of magnetite (Plate IV, 2). The discrete crystals of ilmenite are generally quite large (up to 0,60 millimetres) and are rounded. The many skeletal grains of ilmenite (Plate IV, 1) probably represent exsolution lamellae which formed in magnetite and which were more resistant to replacement by hydrothermal solutions than the magnetite portion of the crystal. The average grain size of the discrete grains of ilmenite is 0,15 millimetres which is much larger than the average grain size of the magnetite, and suggests that ilmenite might have been the first mineral to form.

Textural relationships with other minerals

Ore minerals

Ilmenite occurs in an 'exsolution texture' with magnetite. Narrow exsolution lamellae of magnetite may be present in ilmenite (Plate IV, 2), or lamellae of ilmenite may be present within a grain of magnetite. Euhedral grains of ilmenite are enclosed poikilitically by a later formed magnetite crystal which has since altered to goethite (Plate V, 1). Ilmenite has not been observed to have replaced magnetite. The ilmenite has commonly undergone alteration to rutile as in Plate IV, 3, where remnants of ilmenite occur in rutile in an 'island and sea' texture. The ilmenite may also have been altered to leucoxene as in Plate IV, 4, where a skeletal grain of goethite is surrounded by leucoxene and has been partially replaced by bornite and white chalcocite. Originally the crystal was probably composed of ilmenite containing exsolution lamellae of magnetite, the ilmenite having subsequently altered to leucoxene and the magnetite to goethite.

As far as the temperature of formation of the ilmenite is concerned, the 'exsolution texture' between ilmenite and magnetite is taken by Ramdohr (1955) to indicate a temperature of formation of between 400°C and 700°C , generally at about 600°C . Buddington and Lindsay (1964)

came to the conclusion that the magnetite-ilmenite intergrowths owe their origin to oxidation of a pre-existing intergrowths of magnetite and ulvöspinel. The magnetite-ulvöspinel intergrowths were thought by Buddington and Lindsay to form at 600 - 700°C in dioritic rocks but possibly as low as 400 - 550°C, the exact temperature depending mainly on the oxygen fugacity.

2. Magnetite $\text{FeO} \cdot \text{Fe}_2\text{O}_3$

Magnetite is present in fairly large quantities in all the rock types of the Complex. It occurs in subhedral and anhedral grains up to 0,25 millimetres in width, although the crystals are generally extremely small being less than 0,02 millimetres in width.

Magnetite is recognised by its isotropic character, iron-grey colour, reflectivity and Vicker's Hardness. Reflectivity at 590 μ using a Sika (Henry) standard is 18,9 per cent (Folinsbee 1949 gives 20,7 - 21,4 per cent). Vicker's Hardness i.e. VHM_7 is 718 (Burke 1966 - 1967 gives 489 - 734).

Textural relationships with other minerals

Ore minerals

As mentioned under the ilmenite section magnetite forms an intergrowth texture with ilmenite (Plate IV, 2). Pyrite has replaced magnetite (which is partially altered to goethite) in an 'island and sea' texture (Plate V, 2). Chalcopyrite has replaced a euhedral magnetite grain (which has subsequently been altered to goethite pseudomorphously), and is in turn replaced along its edges by bornite (Plate V, 3). The magnetite has been extensively altered, especially in the region of the sulphide bodies. The process of alteration of the magnetite was probably initially oxidation with the formation of hematite, which was then altered to goethite by hydrothermal solutions containing CO_2 . Thus many of the textural relationships displayed by goethite relate to pre-existing magnetite or hematite.

From the above relationships it is evident that ilmenite and magnetite were the first 'ore minerals' to crystallise and probably crystallised before the arrival of the hydrothermal solutions. Bischoff (1950, p. 17) in discussing the magnetite present throughout the Complex states; 'In Paar gevalle is in die intrusiewe gesteentes aangetref waar die magnetiet groot geraamte - kristalle vorm met oorblyfsels van amfibool of veldspaat daarin (Fig. 11.). Soms vorm die magnetiet gedeeltelike pseudomorfe na amfibool. Dit is dus taamlik seker dat meeste van die

magnetiet sekondêr is, en dat daar min primêre magnetiet teenwoordig is'. The writer considers, however, that it is unlikely that magnetite formed as a secondary mineral and concludes that the magnetite was poikilitically enclosed within the amphibole. Hematite has been observed to form as an alteration product of the amphibole. Magnetite, however, is not observed as a replacement of amphibole, also the exsolution texture of magnetite with ilmenite (Plate IV, 2) points to a primary origin.

As discussed under the ilmenite section, the exsolution texture between ilmenite and magnetite indicates a temperature of formation for the two minerals of between 400°C and 700°C. It can be concluded from the fact that later grains of magnetite (now altered to goethite) enclose euhedral grains of ilmenite which contain exsolution lamellae of magnetite (Plate V, 1), that the deposition of magnetite must have continued after the deposition of ilmenite ceased.

3. Hematite Fe_2O_3

Hematite is formed as an alteration product of amphibole. It is also possible that hematite is formed as an intermediate phase in the alteration of magnetite to goethite. However, the hematite is only observed as an alteration product of magnetite where the magnetite forms exsolution lamellae within the ilmenite. The reaction which took place could have been that of oxidation of magnetite to hematite and finally the alteration of hematite to goethite by hydrothermal solutions in which CO_2 has been concentrated.

4. Pyrite FeS

Two generations of pyrite having similar properties under reflected light are present at Roodekraal. Differences in paragenesis are however indicated by a disparity in grain size and interrelationships with other minerals as detailed in the following section. These two varieties will be termed pyrite I and pyrite II.

Pyrite I

The most abundant variety is pyrite I which occupies the normal place in the paragenetic sequence, and occurs in cubic crystals with sides up to 3 millimetres in length which have characteristic striations on them.

Textural relationships with other minerals

Ore minerals

Magnetite is seen to be almost completely replaced by pyrite I in an 'island and sea' texture (Plate V, 2). Pyrrhotite has replaced pyrite I leaving corroded remnants of pyrite I grains in a groundmass of pyrrhotite (Plate VI, 2 and Plate VI, 3). Sphalerite has replaced pyrite I in a 'caries texture' (Plate VII, 1) and chalcopryite has replaced pyrite I, leaving remnants of pyrite I completely encircled by chalcopryite in an 'island and sea' texture (Plate VI, 4).

Gangue minerals

Pyrite I has replaced the andesite host rock in large euhedral crystals (Plate V, 4). Quartz formation seems to have taken place prior to the formation of pyrite I as quartz does not cut across crystals of pyrite I, also pyrite I is observed surrounding quartz in amygdales, forming 'atoll like structures' (Plate VI, 1).

From the above mineral associations it can be concluded that pyrite I was formed after magnetite but before pyrrhotite.

Pyrite II

Pyrite II occupies an anomalous position in the paragenetic sequence. It consists of extremely small euhedral grains, generally less than 0.01 millimetres in diameter occurring within sphalerite, pyrrhotite and chalcopryite. The small grains of pyrite II are arranged along curved lines within host minerals which would all normally be younger than pyrite (Plate VII, 2 and Plate VII, 3).

Mineral associations suggesting that pyrite II crystals were not formed before the sphalerite, pyrrhotite and chalcopryite are detailed below.

- (i) The lines along which the pyrite II grains lie are curved, not straight, making it appear that they occupy cracks not crystallographic directions. Thus it cannot be argued that the pyrite II grains were expelled by the crystal lattice of the growing host minerals, because they should then lie along crystallographic directions.
- (ii) It is difficult to imagine that small grains of pyrite II would remain in position while a pre-existing host was completely replaced by pyrrhotite, sphalerite or chalcopryite.
- (iii) None of the pyrite II grains are corroded.
- (iv) Pyrite II replaced grains of pyrite I (Plate VII, 4).

The following facts indicate that pyrite II is not an exsolution product of the host minerals.

- (1) The pyrite II crystals do not lie along the crystallographic directions of the host minerals.
- (2) The lines of pyrite II crystals are observed to cut through gangue minerals (Plate VII, 4).
- (3) It would be strange if pyrite II crystals of the same size were developed in sphalerite, pyrrhotite and chalcopyrite at different stages in the paragenetic sequence.

From the above mineral associations it would appear that pyrite II was formed later than the minerals in which it is now situated. It appears from Plate VII, 4 that pyrite II is at least partially a re-crystallisation of pyrite I since half of a pyrite I crystal is missing and the space is occupied by fine-grained pyrite II crystals.

According to Smith (1948) there are two separate periods of pyrite mineralization, the first extending over the temperature range 400°C to 630°C at high pressures, and a second at 150°C at much lower pressure.

5. Pyrrhotite $\text{Fe}_n\text{S}_{(n+1)}$

Pyrrhotite is a fairly common ore mineral at Roodekraal, and is usually associated with pyrite and chalcopyrite. Reflectivity at 590 μ using the U.S.S.R. pyrite standard is 42,0 per cent (Folinsbee 1949 gives 39,5 - 40,2 per cent). Vicker's Hardness i.e. VHM_7 is 255 (248 - 262) (Burke 1966 - 1967 gives 247 - 390).

Two varieties of pyrrhotite, hexagonal and monoclinic, are present. This fact was established by etching pyrrhotite with dilute chromic acid, the monoclinic variety being strongly etched while the hexagonal variety was left virtually unaltered. In Plate VIII, 1 the hexagonal pyrrhotite can be seen as rounded grains present within a groundmass of later formed monoclinic pyrrhotite in an 'island and sea' texture. Hexagonal pyrrhotite also occurs as narrow lamellae within the monoclinic variety.

Textural relationships with other minerals

Ore minerals

Pyrrhotite is observed to have replaced pyrite I in an 'island and sea' texture (Plate VI, 2). Chalcopyrite has clearly replaced pyrrhotite in a modified 'caries texture' in Plate VI, 3 where pyrite I inclusions

within the pyrrhotite acted as barriers to replacement by chalcopyrite, and replacement proceeded at a greater rate where the pyrite I was not present, so that tongues of chalcopyrite extended through gaps in the pyrite barrier and replaced the pyrrhotite beyond.

In geothermometry the d_{102} spacing of hexagonal pyrrhotite has been used by Arnold and Reichen (1962) to establish the metal content of the pyrrhotite and thence the temperature of formation. Arnold (1966) has demonstrated however that the lower iron content runs were in fact mixtures of monoclinic and hexagonal pyrrhotite. Kullerud et al. (1963) have shown that monoclinic pyrrhotite cannot be used for temperature determination unless it inverted from hexagonal pyrrhotite. In selecting a grain of hexagonal pyrrhotite for measurement of the d_{102} spacing care was taken to use only pyrrhotite grains which had partially replaced pyrite crystals, so that sufficient sulphur was present for the formation of a pyrrhotite crystal with an equilibrium composition at the temperature of formation. The d_{102} spacing was measured on four such grains and all gave the same result of $2,061 \text{ \AA}^0$, from which it was calculated that, the percentage of atomic iron present is 46,9 and the temperature of formation was 430°C . These results can, however, only be used as an approximate guide since the pyrrhotite geothermometer is not yet fully understood.

6. Sphalerite ZnS

Sphalerite occurs in calcite bearing veins associated with chalcopyrite, galena and early formed pyrite.

Identification of the sphalerite was confirmed by reflectivity, Vicker's Hardness, its characteristic internal reflections and by X-ray analyses using the powder method. Reflectivity at 590μ using the Sika (Henry) standard is 17,9 per cent (Folinsbee 1949 gives 17,8 - 18,5 per cent). Vicker's Hardness i.e. VHM_5 is 221 (Burke 1966 - 1967 gives 147 - 276).

Textural relationships with other minerals

Ore minerals

Sphalerite is seen to have replaced earlier pyrite I in a 'caries texture' (Plate VII, 1). It has an exsolution texture with chalcopyrite having blebs and laths of chalcopyrite distributed parallel to the (111) and (100) crystallographic directions of the sphalerite (Plate VIII, 2). This indicates that both minerals were deposited at the same time. Strings of tiny euhedral pyrite II crystals cut across the sphalerite and gangue

(Plate VII, 2). This relationship has been taken to indicate that the pyrite II formed after the sphalerite. Galena has replaced the intergrowth of sphalerite and chalcopyrite in a 'vein texture' (Plate VIII, 3) and a 'caries texture'.

The exsolution texture between chalcopyrite and sphalerite indicates an unmixing in solid solution, and a temperature of formation of 350°C - 400°C (Buerger, 1934) for the two minerals. X-ray diffraction was carried out on sphalerite which is in exsolution with chalcopyrite. From thirteen readings the average value of the unit cell dimension was derived as 5,4141 Å. Using the graphs of Coleman (1957) the author derived the following values for the sphalerite:-

Weight percentage of FeS present	= 9,4
Temperature of formation	= 350°C.

Due to inaccuracy in the fourth decimal place caused by observation errors in measuring the photographs, these values must be regarded as approximate.

7. Chalcopyrite CuFeS

Chalcopyrite has the most widespread occurrence of all the copper minerals. It is found associated with pyrite, bornite, sphalerite and more rarely chalcocite, galena, pyrrhotite and magnetite. It is present in two distinct varieties termed chalcopyrite I and chalcopyrite II. Chalcopyrite I is by far the most abundant and is the early formed variety, while chalcopyrite II is present in small amounts, associated with bornite and blue chalcocite.

Chalcopyrite I

This is the common variety of chalcopyrite and has the typical brass-yellow colour and weak anisotropism with blue to green polarization colours.

Textural relationships with other minerals

Ore minerals

Magnetite (which has subsequently been altered to goethite) has been replaced by chalcopyrite I pseudomorphously, which has in turn been replaced by bornite (Plate V, 3). Pyrite I has been replaced by chalcopyrite I in an 'island and sea' texture (Plate VI, 4). Pyrrhotite has been replaced by chalcopyrite I in a modified 'caries texture' in Plate VI, 3, but may also exhibit a 'mutual texture' with the chalcopyrite I (Plate VIII, 4) where the chalcopyrite I is embayed by 'caries'.

of pyrrhotite and occurs as an 'island' within a 'sea' of pyrrhotite, yet it appears to have replaced the pyrrhotite in 'caries' itself. Sphalerite contains exsolution blebs and lamellae of chalcopyrite I within it (Plate VIII, 2) indicating that these two minerals were deposited simultaneously. Bornite is seen to have replaced chalcopyrite I in a 'caries' texture in Plate V, 3. Bornite is however not observed to have any variety of mutual texture with chalcopyrite I, and it thus appears that the deposition of chalcopyrite I was completed before any bornite was deposited. Galena replaces chalcopyrite I in an 'island and sea' texture and the chalcopyrite I surrounds a corroded calcite crystal (Plate IX, 1).

Gangue minerals

Actinolite needles are surrounded and have been replaced by chalcopyrite I and must have formed before the chalcopyrite I (Plate IX, 4). Calcite is surrounded and has been corroded by chalcopyrite I in Plate IX, 1. Epidote is commonly present as 'islands' within a 'sea' of chalcopyrite I, and has been replaced by chalcopyrite I in a 'caries texture' (Plate X, 1).

Chalcopyrite II

Chalcopyrite II occurs as exsolution lamellae within bornite. It is present only in very small amounts and although it has the typical brass-yellow colour it has a more intense anisotropism than chalcopyrite I and blue-green to yellow-green polarization colours.

Textural relationships with other minerals

Bornite contains exsolution lamellae of chalcopyrite II (Plate IX, 2), which is also observed to have a 'mutual boundary' texture with the bornite. Blue chalcocite is observed to replace chalcopyrite II and bornite (Plate IX, 3). Lamellae of chalcopyrite II cross the grain boundary of bornite and blue chalcocite and occur as corroded remnants within the blue chalcocite, demonstrating that blue chalcocite replaced the bornite more readily than the chalcopyrite II. Covellite, in irregularly shaped masses, has replaced the blue chalcocite and the included chalcopyrite II (Plate IX, 3). It is clearly evident that chalcopyrite II did not form as a reaction product of the replacement of bornite by covellite, as was the case in the copper ores of the Artonville Mine, Messina district (Jacobsen, 1967). Evidence against such a mode of origin is the fact that chalcopyrite II is present in bornite which has not been replaced by covellite.

Although chalcopyrite II forms an 'exsolution texture' with bornite, this texture cannot be regarded as indicating a high temperature of formation since there is less than 8 per cent chalcopyrite present. Brett (1964) has shown that this exsolution could take place between 0° and 228°C. The exsolution was merely due to an excess of iron in the bornite resulting in the formation of chalcopyrite II, which diffused to the grain boundaries of bornite.

8. Bornite Cu_5FeS_4

Bornite is present in approximately the same proportion as chalcopyrite but whereas chalcopyrite is widespread in its occurrence, the bornite is concentrated together with chalcocite in the central and upper portions of the mineralized zones. When freshly polished the bornite is pinchbeck-brown in colour, but the bornite in contact with chalcocite rapidly tarnishes to a pink or even a purple colour.

Textural relationships with other minerals

Ore minerals

The relationship between the two varieties of chalcopyrite and bornite has already been discussed. It was concluded that chalcopyrite I was deposited before bornite, and that chalcopyrite II was deposited together with bornite at fairly low temperatures between 0°C and 228°C. Chalcocite is the mineral which is most commonly associated with bornite. Two varieties of chalcocite are associated with bornite, a steely-white hypogene variety, and a pale blue variety which is probably supergene. White chalcocite is present in a 'myrmekitic' intergrowth texture with bornite (Plate X, 2). The bornite and chalcocite intergrowth is composed of irregular bands which are alternately coarse and fine.

There has been much controversy over the interpretation of the 'graphic', 'sub-graphic', 'pseudo-eutectic' or 'myrmekitic' texture as it has been called by various writers. Schwartz (1939) states 'It seems probable that graphic textures may originate in more than one way. It is now accepted that the eutectic explanation is untenable. Investigations published in the last ten years seem equally divided between the theory of origin by replacement or by contemporaneous deposition.' More recently Brett (1964) has interpreted the 'myrmekitic' intergrowth as resulting from the breakdown of exsolution 'basket-weave' textures. Edwards (1965) states 'Graphic textures resulting from replacement can frequently be recognised from their association with veins of the metasome traversing the host'. It thus seems reasonable to assume that

we can distinguish replacement of bornite by chalcocite from exsolution textures by noting whether veins of chalcocite cut the bornite or not. Veins of white chalcocite are seen to cut through the gangue and unite with the chalcocite of the intergrowth (Plate X, 3), thus it would appear that the texture resulted from the replacement of bornite by chalcocite. However it is important that although narrow veins of white chalcocite may cut bornite grains, white chalcocite does not occur as replacements following cracks and fissures within the bornite. The coarser grained portion of the bornite-chalcocite intergrowth commonly displays a 'mutual boundaries' texture, with 'islands' of bornite which are embayed by chalcocite, themselves containing irregularly shaped 'islands' of white chalcocite (Plate XI, 1). 'Vein-replacement' and 'rim-replacement' give clear evidence that the white chalcocite was in part later than the bornite (Plate XI, 2). From the above it would appear that bornite was deposited at the same time as white chalcocite but that the deposition of white chalcocite continued after bornite formation had ceased.

Schwartz (1928) derived the temperature of homogenization of bornite and chalcocite as being 225°C. If we accept this as being correct it means that the white chalcocite must have formed at temperatures below 225°C.

Blue chalcocite is observed to have replaced bornite in a 'rim-replacement' texture (Plate XI, 3), it is thus clearly later than the bornite. Goethite has been observed to contain small 'islands' of bornite and chalcocite within it, which probably demonstrates that the goethite formed later than the bornite and white chalcocite. In a skeletal grain of goethite (Plate IV, 4) the bornite and white chalcocite are observed to have replaced the goethite pseudomorphously, but the writer believes that this texture refers to the original magnetite grain which has been subsequently altered to goethite.

Gangue minerals

Actinolite needles were formed before bornite and white chalcocite, since they act as partitions between adjacent areas of bornite and white chalcocite (Plate XI, 4). Actinolite has also been corroded by the bornite and white chalcocite.

9. Chalcocite Cu_2S

Chalcocite is generally concentrated with bornite in the central and upper portions of the mineralized zones where it is the preponderant

copper mineral. As discussed under the hornite section, two distinct varieties of chalcocite are present, an early formed white variety and a late formed blue variety.

White chalcocite

The white chalcocite is present in the greatest abundance. It is slightly anisotropic and coarse-grained having grains below 0,5 millimetres in diameter and a prominent etch cleavage when etched with a 20 per cent by weight solution of KCN (Plate X, 4).

Textural relationships with other minerals

Ore minerals

Chalcopyrite I has been replaced by white chalcocite along its boundary with a grain of epidote (Plate X, 1). The 'myrmekitic' texture between bornite and chalcocite has been discussed in detail under the bornite section. It was noted that there has been much controversy as to whether the texture owes its origin to exsolution of the two minerals or to replacement of the bornite by chalcocite. Irrespective of its mode of origin this texture points to a hypogene origin for the chalcocite, since Lindgren (1933) holds that 'myrmekitic' intergrowths are indicative of hypogene replacement, in that they only develop in compact host rocks that were under considerable pressure at the time of replacement. The host rock at Roodekraal is a compact andesite in which there are very few channelways for solutions, yet the white chalcocite is known to exist to depths of 247 metres below the surface.

The writer feels that there can be little doubt that the great majority of white chalcocite is of hypogene origin. Facts which point to a hypogene origin for white chalcocite are summarized below.

- (i) The fact that white chalcocite is found in a 'mutual boundaries' intergrowth (Plate XI, 1) and a 'myrmekitic' intergrowth (Plate X, 2) with bornite.
- (ii) The large grain size and strong etch cleavage displayed when white chalcocite is etched with a 20 per cent by weight solution of KCN (Plate X, 4).
- (iii) The anisotropism displayed by the chalcocite. According to Short (1940) fine-grained supergene chalcocite is apparently isotropic.

- (iv) The fact that white chalcocite does not occur as replacements originating from cracks or fissures within the bornite.
- (v) The depths to which it is found in a compact host rock.

As mentioned under the bornite section, goethite encloses bornite and white chalcocite in an 'island and sea' texture. Covellite is observed to have replaced white chalcocite along cleavage planes (Plate XII, 1).

When discussing the bornite-chalcocite 'myrmekitic' intergrowth it was concluded that the white chalcocite formed at temperatures below 225°C.

Blue chalcocite

Blue chalcocite is present in small amounts near the surface. The greatest depth at which it is found is 17,4 metres below the surface. It is apparently isotropic. An anomalous occurrence of blue chalcocite is where it occurs at greater depths as a narrow rim along the edges of bornite grains which are in contact with white chalcocite (Plate XI, 1). This feature has no connection with the blue chalcocite found near the surface and is artificially formed by the heat produced when mounting the specimen in molten plastic. The heat is sufficient to cause a little of the bornite to dissolve in the surrounding chalcocite giving it a blue colour. A similar effect has been noted by Edwards (1965) who states 'Such "blue chalcocite" is formed readily by heating natural chalcocite, containing a fine intergrowth of bornite to 100°C'.

The blue colouration of the chalcocite may be caused by dissolved covellite or bornite. In an effort to establish if any of the blue chalcocite is in fact digenite, a quantity of powder was drilled from two specimens of blue chalcocite, one situated between bornite and covellite and the other within white chalcocite. The powders were examined by X-ray diffraction methods and both specimens matched the values quoted by the American Society for Testing and Materials for alpha (low) chalcocite. Since digenite is not present the blue colouration must either be due to dissolved bornite, or dissolved covellite in amounts of less than 8 atomic per cent.

Textural relationships with other minerals

Ore minerals

As mentioned under the bornite section, the blue chalcocite has replaced the bornite as a 'rim-replacement' (Plate XI, 3), or it forms an

intermediate zone between covellite and bornite (Plate IX, 3). When white chalcocite was replaced by covellite the blue chalcocite may have formed as an intermediate zone, but more commonly the white chalcocite has been replaced directly by covellite. The blue chalcocite occurs as a 'rim-replacement' around native copper (Plate XII, 2). It is concluded by the writer that blue chalcocite is of supergene origin. Evidence supporting the assumption of such an origin for the blue chalcocite is summarized below.

- (i) It forms as a 'rim-replacement' around native copper which is a late-formed mineral.
- (ii) It has at the time of writing, only been found to a depth of 17,4 metres below the surface.
- (iii) It is commonly associated with covellite, a mineral which is usually supergene.
- (iv) It is isotropic.
- (v) It is fine-grained and has no etch cleavage, and turns black almost immediately when treated with a 20 per cent by weight solution of KCN.
- (vi) X-ray examination gives a best fit for alpha (low) chalcocite.

10. Galena PbS

Galena is found associated with sphalerite and chalcopyrite I and is recognised by its reflectivity, hardness and characteristic development of triangular pits on a polished surface due to its perfect cubic cleavage. Reflectivity at 590μ using the U.S.S.R. pyrite standard is 43,9 per cent (Folinsbee 1949 gives 41,6 - 43,2 per cent). Vicker's Hardness i.e. VHM_{10} is 80,5 (Burke 1966 - 1967 gives 60 - 116).

Textural relationships with other minerals

Ore minerals

Galena has clearly replaced an exsolution texture of chalcopyrite I and sphalerite in a 'vein texture' (Plate VIII, 3) and in a 'series texture'. Chalcopyrite I was replaced by galena in an 'island and sea' texture (Plate IX, 1).

Gangue minerals

Calcite has been replaced by chalcopyrite I and both these minerals have subsequently been replaced by galena in an 'island and sea'

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