texture (Plate IX, 1).

From the above mineral associations it can be deduced that since galena replaces an exsolution texture of chalcopyrite I in sphalerite with an iron content of 9,4 per cent by weight, the galena must have formed at temperatures below 350°C.

11. Silver

Although no silver mineral has been observed under the microscope, the copper sulphides are argentiferous.

The silver present must be hypogene, since silver is present in small amounts wherever the copper minerals are concentrated. The amount of silver present is directly proportional to the amount of copper present and not to the distance below surface. Silver values are not found associated with the galena. The relationship between the quantities of copper and silver present is a simple one, which may be expressed by the algebraic function:

$$y = 3,79 \times 10^{-4}x$$

where: y = Percentage copper (electrolytic analysis)

x = Percentage silver (assay).

The above relationship is derived using the following methods. Firstly all available values obtained from samples which had been analysed for copper, using the electrolytic deposition method, and also assayed for silver, are plotted on rectangular co-ordinates using the copper percentage as the ordinate and the silver percentage as the abscissa. The graph obtained is presented in Figure 6. Values obtained by using the atomic absorption method of analysis are not used in this study due to their inaccuracy. Secondly, by inspection of the graph it is seen that the line of best fit passes through the origin. The equation of the locus of the line of best fit may thus be expressed as :

$$y = mx$$

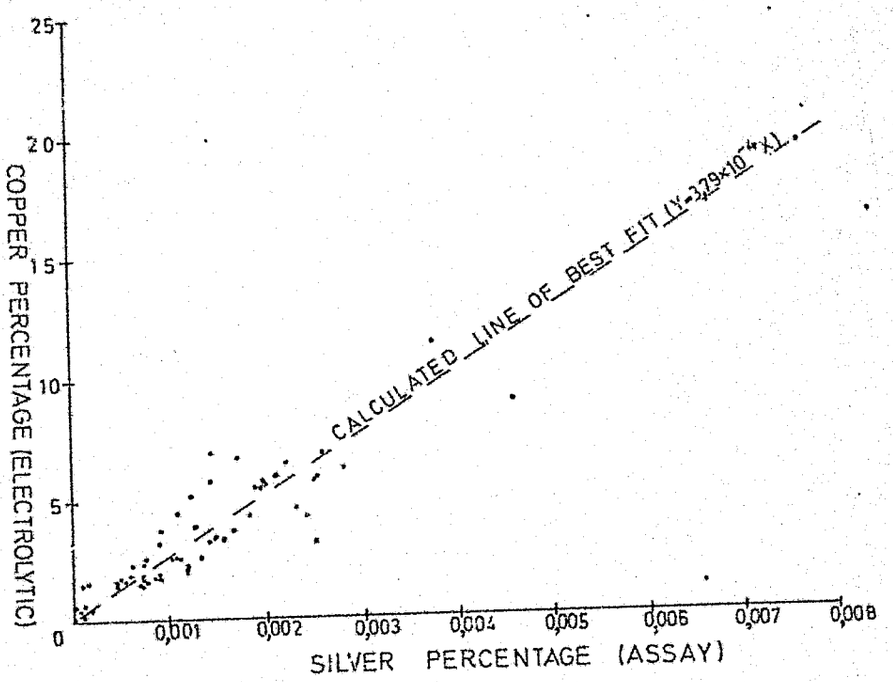
where: m = slope of the line

To determine m the expression:

$$m = \frac{\Sigma xy - \frac{(\Sigma x)(\Sigma y)}{n}}{\Sigma x^2 - \frac{(\Sigma x)^2}{n}}$$

was used.

Fig. 6 Graph showing the relationship between the amount of silver and copper present within the copper sulphides of the Roodekraal Igneous Complex



When the value for m is substituted in the locus of the line of best fit the expression $y = 3,79 \times 10^{-4}x$ is obtained. Expressed in another way it means that for every part of silver present there are 2 638,5 parts of copper.

Due to the fact that the amount of silver present is directly related to the amount of copper sulphides present, and since all the samples in this study were composed mainly of white chalcocite, it can be concluded that the silver is present within the white chalcocite. The silver is probably present as a sulphide since examples of solid solutions between silver sulphides and chalcocite are well known. Edwards (1965) in discussing the cooling of a solid solution of argentite and chalcocite states 'the silver is precipitated not as Ag_2S but as the intermediate compound stromeyerite (Cu_2S. Ag_2S), which develops as blade-like crystals in the (111) directions of the cubic chalcocite'.

From the above it seems probable that the silver is present in the form of hypogene stromeyerite.

12. Native copper Cu

Native copper is present within the upper portions of the mineralized zone extending to a depth of 40,2 metres. It occurs almost exclusively within the porphyritic andesite and the coarsely crystalline andesite.

The native copper was identified by its high reflectivity, distinctive copper-red colour and its Vicker's Hardness. The reflectivity at 590μ using the U.S.S.R. pyrite standard is 82,1 per cent (Folinsbee 1949 gives 67,8 - 72,8 per cent). Vicker's Hardness i.e. VHN_4 is 71,3 (70,2 - 71,9) (Burke 1966 - 1967 gives 50 - 62).

Textural relationships with other minerals

Ore minerals

Blue chalcocite has replaced native copper as a 'rim-replacement' (Plate XII, 2). Thus the native copper must have been deposited before the blue chalcocite.

Gangue minerals

Epidote has been corroded by native copper in a 'caries texture'. The range of temperature within which epidote minerals crystallise is given as 300°C -(400°C or 500°C) (lower limit Stringham (1952), upper limits Ramberg (1949) gives 400°C and Rosenqvist (1952) gives 500°C,

quoted in Deer et al., 1, p. 202, 1967)

The above mineral association would suggest that native copper was formed at temperatures below 300°C. The native copper may be hypogene or supergene in origin since the hypogene native copper of the Lake Superior district of North America was deposited at temperatures of 100° - 250°C, according to Lindgren (1933). The fact that the native copper is restricted to coarser grained (more porous) rock types near the surface seems to indicate that it is supergene in origin (see Plate XX).

13. Goethite $Fe_2O_3 . H_2O$

Goethite has formed as an alteration product of magnetite especially in the vicinity of the mineralized zones. The formation of goethite is considered by the writer to be due to hydrothermal processes involving oxidation and hydration, and not to the weathering processes. The amount of goethite present does not decrease with depth but increases markedly in the vicinity of copper sulphide bodies. Thus goethite, despite its fairly low temperature of formation is a hypogene and not a supergene mineral. Goethite is recognised in polished section by its light yellow internal reflections, moderate anisotropism and iron-grey colour.

The presence of goethite would explain why geophysical surveys using electromagnetic or magnetic methods showed the mineralized zones to be lying along non-magnetic zones. The abundant magnetite gives a high magnetic background and where the magnetite has been altered to goethite, a magnetic low occurs.

Textural relationships with other minerals

The goethite has been formed at the expense of magnetite and for this reason it is uncertain whether textural relationships between goethite and other minerals refer to the goethite or the pre-existing magnetite. The only texture observed between goethite and other minerals which is thought to refer to the goethite itself, is where it encloses bornite and white chalcocite.

The temperature of formation of goethite is given by Edwards (1965) as up to 165°C for alkaline conditions and up to 125°C for neutral conditions. Since the hydrothermal solutions were probably alkaline, containing soda (evidenced by the albitization of the feldspar), as well as a fair amount of potash (necessary for the formation of biotite),

it can be presumed that goethite was deposited at temperatures below 165°C.

14. Rutile TiO_2

Rutile is present as an alteration product of ilmenite. It is clearly seen as a replacement of ilmenite in an 'island and sea' texture in Plate IV, 3 and this same mass of rutile has replaced an epidote grain in a 'caries texture'. Rutile is also observed to enclose epidote completely. From the above it can be concluded that rutile is a supergene alteration product of ilmenite. Ziv (1956) has described the supergene conversion of ilmenite to rutile.

15. Covellite CuS

This mineral is present in minor amounts to a depth of 17,4 metres from surface. Covellite is easily identified by its strong pleochroism from powder-blue to indigo-blue and its strong anisotropism with fiery-orange to red polarization colours.

Textural relationships with other minerals

Ore minerals

Bornite, blue chalcocite and chalcopyrite II have been replaced by covellite along cleavages and cracks (Plate IX, 3) and white chalcocite has been directly replaced by covellite along cleavages and cracks (Plate XII, 1). Owing to the fact that covellite is observed as a replacement of supergene blue chalcocite it can be concluded that covellite was the last copper sulphide to form and that it owes its origin to supergene replacement. A fact which supports a supergene origin for the covellite is that it is only found at shallow depths.

Ramdohr (1955) gave a description of a variety of 'blaubleibender' covellite with an intense blue colour, a description which fits the covellite present at Roodekraal. Moh (1963) has determined that the upper stability limit of this 'blaubleibender' covellite is 157°C ± 3°C.

16. Malachite $CuCO_3 . Cu(OH)_2$, Azurite $2CuCO_3 . Cu(OH)_2$ and Cuprite Cu_2O

These minerals are present as supergene replacement minerals in the zone of weathering.

Cuprite is easily recognised by its red colour in hand specimen, bluish-grey colour in reflected light and parallel etch cleavage developed when treated with a 20 per cent by weight solution of KCN. Malachite and azurite are recognised by their green and blue colours respectively,

and the fact that they effervesce with dilute HCl. These three minerals have replaced the sulphides of copper, and were formed by oxidation and the action of carbonated waters upon the copper minerals at climatic temperatures at or near the surface.

V. DISCUSSION OF RESULTS

A. Formation of the Roodekraal Igneous Complex

1. Tectonic Framework and Age

The Roodekraal Igneous Complex is situated within a well developed tectonic basin in the Pretoria Series of the Transvaal System. The basin was formed by the intersection of two synclinal axes, one developed by the centrifugal forces resulting from the rising of the Vredefort Dome and the other syncline being the 'Eleazar Syncline' (Eriksson, 1971). The 'Eleazar Syncline' was strongly developed running approximately northwest to southeast through the Roodekraal Igneous Complex and intersecting the weakly developed syncline lying parallel to the strike of the sediments at roughly right angles resulting in the development of a tectonic basin. On the map, i.e. Plate XXIV, the Magaliesberg Quartzite which forms the edge of the basin around the Roodekraal Igneous Complex is clearly in evidence on the western and northern sides of the Complex. The Magaliesberg Quartzite consistently dips towards the centre of the Complex, and the contact of the Magaliesberg Quartzite and the amygdaloidal andesite along the northern edge of the Complex is known from drilling operations to maintain its dip of 60° to depths of at least 430 metres (see Plate XXIII). This contact is sharp with no signs of slip movement or brecciation.

A point of great importance is the fact that there has been no significant tilting or folding of the Complex since its emplacement. Evidence supporting this statement is as follows:-

(a) Layers of pyroxene andesite which occur as cap rocks overlying the amygdaloidal andesite are always roughly horizontal, parallel to the present surface (see Plates XVIII, XIX, XX, XXII and XXIII). This demonstrates that no tilting has taken place since the pyroxene andesite flowed out onto the paleo-surface.

(b) From boreholes drilled along the quartzite contact, it appears that no movement has taken place between the quartzite and the andesite.

(c) Contacts between successive flows of amygdaloidal andesite are generally parallel to the present surface.

(d) Where block faulting has taken place due to superincumbent loading, the high angle faults do not appear to have been rotated since their formation (see Plates XVIII and XIX).

From the foregoing it can be concluded that the basin in which the Roodekraal Igneous Complex is situated was formed before the outflow of the andesite. It may however, have been deepened by isostatic sagging under a load of 700 metres or more of andesite and intrusive rocks. Due to the fact that no pillow structures, or other features suggesting extrusion under water, have been observed in the andesite, it can be concluded that the andesite was extruded onto a land surface. The author's conclusions conflict with the findings of Bisschoff (1970) who states firstly that the Complex was emplaced on a 'nearly horizontal surface of the Pretoria Series', and secondly, that the fact that drilling operations show the lava is present to relatively great depths below the present surface 'indicates that the Complex has undergone a certain amount of folding subsequent to its development'. Bisschoff (1972b) concludes that the Roodekraal Igneous Complex was emplaced under shallow water conditions, and does not rule out the possibility that the basin in which the Complex lies was deepened by isostatic sagging.

Hargraves (1970) concluded from measurements of the paleomagnetism of the younger satellitic bodies around the Vredefort Ring Structure that 'the Rietfontein basic complex and its associated alkali granite pluton (and presumably the Vaal River plutons as well) were intruded into essentially flat lying sediments, prior to the development of the Vredefort Ring Structure'. (By 'Vaal River plutons' Hargraves means the alkali granite plutons at Schurwedraai and Baviaanskranz). The author does not agree with this conclusion, and presents the following arguments against it:-

(i) Hargraves tested twelve samples of intrusive rock from the satellitic bodies to see if their magnetic vectors were substantially different from those of the basic granophyre, which was emplaced after the formation of the Vredefort Ring Structure. If these vectors differed substantially it would indicate that the satellitic bodies were emplaced before overturning. Results obtained from ten of the twelve samples used were discounted by Hargraves. The two remaining samples which were taken from the Rietfontein Complex, were used to derive his conclusion that the satellitic bodies were intruded prior to overturning.

(ii) The Losberg Intrusion as Hargraves states, 'gives consistent paleomagnetic data of the same orientation as that from the basic

granophyre'. This would indicate emplacement after overturning.

(iii) The samples taken at Roodekraal were from the coarse pyroxene diorite and the amygdaloidal andesite. The amygdaloidal andesite (sample 14 in Fig. 7) has two ages of intrusives between it and the much younger coarse pyroxene diorite (sample 15 in Fig. 7). However when Hargraves plots the results of the paleomagnetism studies on the 'polar-wander curve' of McElhinny et al. (1968) in Figure 7 it appears that the andesite (14) is much younger than the coarse pyroxene diorite (15) which is incorrect. Hargraves' finding that the magnetic vectors of the two samples from Roodekraal 'are within 35° of the basic granophyre vector' indicates emplacement after overturning.

The only rock formation at Roodekraal which appears to have been tilted after emplacement is the composite coarse pyroxene diorite and olivine gabbro sill. This formation has been tilted towards the centre of the Vredefort Ring Structure, which is exactly the opposite of what would be expected if it had been emplaced before overturning.

(iv) Of the Vaal River plutons, Hargraves himself admits that the results in the alkali granites are widely scattered and that the results from the nepheline syenite, hornfels and dolerite are 'similar to that of the basic granophyre' which would indicate emplacement after overturning.

(v) The Rietfontein Complex has magnetic vectors which indicate that this Complex has been tilted to the south since formation. The writer feels that this tilting of this Complex to the south could have been caused by the rising of the mass of alkali granite which now forms the northern part of the Complex and not by the overturning country rock.

The age of the Roodekraal Igneous Complex is bracketed by two geological events. Firstly the Roodekraal Igneous Complex must be younger than the sedimentary rocks of the Transvaal System, since, as noted by Bisschoff (1950, p. 4), the Complex is emplaced within rocks of the Magaliesberg and Daspoort Stages of the Pretoria Series. Secondly the Complex was emplaced before the introduction of pseudotachylite. The basic granophyre which is regarded as being the same age or younger than the pseudotachylite by Hall and Molengraaff (1925) and others, has been dated by Nicolaysen et al. (1963) at approximately 1 970 million years. It would thus appear that the Roodekraal Igneous Complex was emplaced at a similar time to the Bushveld Igneous Complex. Further evidence suggesting a Bushveld age for the Roodekraal Igneous Complex is that the age datings, carried out on rocks from other centres of intrusion within the Vredefort Ring Structure, give an average which is very close to the age of the Bushveld Igneous Complex as determined by Davies et al. (1970).

Fig. 7 Diagram summarizing paleomagnetic data for intrusive centres within the Vredefort Ring Structure. (taken from Hargraves 1970, p. 260)

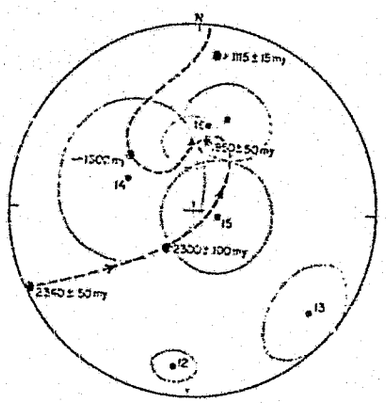


FIG. 3.—Equal-area lower-hemisphere projection summarizing paleomagnetic data in table 1. The heavy broken line linking double solid circles portrays the migration of the earth's magnetic field vector at Vredefort, according to the polar-wander curve of McElhinny et al. 1968. Solid triangle—basic granophyre; solid square—mean of sites 9, 10, 11; results from other sites numbered accordingly. The 95 percent confidence cone for site 16 (9°) has been omitted for clarity. Solid circles are of north-seeking pole on the lower hemisphere; present-day north-seeking pole would be on upper hemisphere.

2. Sequence of Emplacement of the Components of the Roodekraal Igneous Complex

In general the emplacement of the Complex can be divided into three cycles.

a. Andesite/diorite cycle

The oldest rocks of the Complex are the massive andesite and the amygdaloidal andesite, which lie in direct contact with the quartzite and shale of the Magaliesberg Stage and the quartzite, shale and lava of the Daspoort Stage. The andesite is present in great thicknesses, being at least 430 metres thick in the northwest portion of the Complex. The volcanic rocks of the Roodekraal Igneous Complex probably filled a deep tectonic basin within the sedimentary rocks of the Magaliesberg and Daspoort Stages and overflowed its southern edge. Due to the fact that only very small amounts of pyroclastic rocks have been found at Roodekraal the extrusion of the lava probably took the form of a quiet fissure eruption along fractures. The coarsely crystalline andesite horizons probably owe their origin to the extrusion of a particularly thick flow which, although chilled at the margins, would have cooled more slowly in the centre allowing crystals to grow to a larger than average size.

The volcanic agglomerate present does not have the appearance of a pyroclastic rock as the matrix between the rounded rock fragments is not tuffaceous. In the view of the author it probably formed after a period of quiesence, which would result in the development of a land surface strewn with pebbles and boulders of lava and quartzite. When the next outflow of lava took place the rounded rock fragments would have been caught up in it forming the agglomerate. In the closing stages of the volcanism, there was a certain amount of explosive gaseous eruption as evidenced by the presence of narrow zones of volcanic breccia cutting through the andesite. The volcanic breccia which has rock fragments enclosed within a fine-grained andesite matrix may have narrow zones of a later formed breccia, which is cemented by secondary minerals, cutting it. The later formed breccia was probably created when late-stage hydrothermal solutions used the volcanic breccias as channelways, resulting in further brecciation and the introduction of a cement composed mainly of calcite. The second period of brecciation took place after the formation of pseudotachylite which may occur as fragments within the breccia.

The andesite was intruded by the equigranular diorite which in places has an almost gradational contact with the andesite. The diorite has much the same mineralogy as the andesite, a similar chemical composition, and, in certain localities, contains a few widely spaced amygdales which are mainly composed of calcite. As previously mentioned the diorite intrudes the andesite in two northeast to southwest trending zones, which are taken to represent pre-existing fractures along which the magma rose to the surface.

It is concluded that these two rock types were part of the same intrusive cycle and that the equigranular diorite intruded the andesite before it was completely cold.

b. Pyroxene andesite/diorite cycle

This cycle of events was begun by an intrusion, generally within the equigranular diorite, of fine pyroxene diorite. This rock type, when it reached the surface, flowed out as a fine pyroxene andesite to form a layer over much of the surface area of the amygdaloidal andesite. The pyroxene andesite must have been the first rock type of the pyroxene andesite/diorite suite to consolidate since it was extruded as thin flows which were chilled and consequently crystallised before the pyroxene diorite. This layer of pyroxene andesite has to a large extent been removed by erosion. An intrusion of coarse pyroxene diorite then occurred, mainly within the fine pyroxene diorite.

The origins of unusual rock types which are considered by the author to belong to the pyroxene andesite/diorite cycle of intrusion are considered below.

(i) Origin of the olivine gabbro

The olivine gabbro appears to have developed as a differentiate of the coarse pyroxene diorite. The densities of the component minerals of the olivine gabbro, as calculated from their compositions are:-

Plagioclase	-	2,67
Bronzite	-	3,35
Augite	-	3,40
Olivine	-	3,42
Magnetite	-	5,20

These densities are consistent with a gravitational settling origin for the mineral layering in the olivine gabbro, provided that there were no reversals in the normal order of crystallisation. Plagioclase and bronzite being the lightest minerals are concentrated near the top of the olivine gabbro body while augite, the mineral intermediate in density, is concentrated in the centre and the two heaviest minerals, olivine and magnetite, are concentrated along the lower contact. The main texture of the olivine gabbro is a cumulus texture. Cumulus minerals include olivine, magnetite, augite and bronzite while the intercumulus material is provided by plagioclase in the main, with lesser amounts of kaersutite and biotite. The olivine gabbro must have been formed while the coarse pyroxene diorite sill was more or less horizontal, before the sill was tilted to the southwest. The tilting of the sill towards the southwest could have been caused by a subsidence towards the centre of the Complex.

(ii) Origin of the dioritic-pegmatite and dioritic-aplite

The dioritic-pegmatite and dioritic-aplite probably belong to the pyroxene andesite/diorite cycle of activity since the author considers that these rock types are genetically related to the coarse pyroxene diorite and have migrated upwards from it.

Heinrich (1965) considered that pegmatite was formed when a magma, which is saturated with volatile constituents, crystallises. He regarded aplite as a quenched rock formed by a 'pressure quench' owing to an escape of volatiles resulting in a responsive rise in the liquidus temperature. Jahns and Burnham (1969) outlined a genetic model for pegmatite formation which is in agreement with the ideas of Heinrich (1965). This model begins with the normal magmatic stage, which during the course of crystallisation results in a concentration of volatiles within the melt. This stage is succeeded abruptly by a two fluid stage consisting of the rest magma and an aqueous fluid. Next the pegmatite veins are formed from the aqueous fluid, with their lower portions composed of footwall aplite. The aplite is formed by a draining of excess water upwards into the higher portions of the veins in which the pegmatite is crystallising. This model, which has composite veins with aplite in the lower portions, would fit with the field relationships between pegmatite and aplite at Roodekraal. The pegmatite is only found in the equigranular diorite while the aplite is found in both the equigranular diorite and the underlying coarse pyroxene diorite. The high water contents of the pegmatite of 2,17 per cent for RC 23 and 1,13 per cent for RC 25 as compared with the low water content of 0,46 per cent for the aplite RC 27 is also in agreement with this model. Jahns and Tuttle (1963) grouped layered pegmatite-aplite intrusives into two major categories:

I. Composite bodies and complexes representing multiple injection of magma from external sources.

II. Bodies representing a single injection of magma followed by segregation during crystallization and in some occurrences by auto-injection of residual fluid.

The Roodekraal pegmatite and aplite would appear to fall into the second group as the pegmatite is extremely uniform in composition over wide areas and the veins are generally very narrow, the largest vein being 0,30 metres in width. Where pegmatite and aplite occupy the same vein, the pegmatite is fine-grained but has a fairly sharp contact with the aplite. Jahns and Tuttle (1963) divided their group II into four classes. The pegmatite-aplite veins at Roodekraal can be best described by 'class four' which embraces highly assymmetrical bodies, with upper portions consisting mainly or wholly of pegmatite and lower portions which consist mainly or wholly of aplite. Since the pegmatite veins do not contain xenoliths of wall rock and since they may have matching irregularities in the wall rocks on either side they are dilation veins. The aplite may contain

xenoliths of wall rock and for this reason may partly owe its origin to replacement.

c. Late intrusive cycle

During this period several completely dissimilar types of rocks were intruded. Firstly, there was a small localised intrusion of two narrow dolerite dikes within the fine pyroxene diorite in the southern area of the Complex. Secondly, there was the widespread intrusion of narrow syenite dikes which cut through the andesite, diorite and pyroxene diorite. The syenite dikes are generally less than one metre in width, with the largest having a width of five metres. Most of the dikes have a northeast to southwest strike and are composed, essentially of oligoclase feldspar.

The next rock type to be intruded was the pseudotachylite. The fact that this peculiar rock type occurs within the Roodekraal Igneous Complex immediately links the Complex with other centres of intrusion around the Vredefort Ring Structure which also contain pseudotachylite.

There has been much controversy over the years as to the origin of pseudotachylite. Shand (1916) considered that the pseudotachylite owed its origin to shock phenomena. Hall and Molengraaff (1925) and Nel (1927) saw the pseudotachylite as being formed by shearing under high pressure. Reynolds (1954) concluded that it owed its origin to a 'fluidized' system of gas and solid particles. Philpotts (1964) thought it was formed by the frictional heat or hot gases which are developed in a fault zone during rapid displacement. Bisschoff (1970) considered that the pseudotachylite is a 'tuffisite' formed by the action of gases related to a central pluton. Although the thickest zone of pseudotachylite at Roodekraal follows the contact between an equigranular diorite sill and the amygdaloidal andesite, as mentioned under the section entitled 'Distribution and Structure of the Rock Types', it does not always occur exactly on the contact and has a limited lateral extent. Since the contact between the diorite and andesite elsewhere is free of pseudotachylite this rock type cannot be a true tachylite. As the zone of pseudotachylite follows the contact of the sill, which is an irregular horizontal surface for more than 150 metres, it is hard to imagine how such a feature could be caused directly by shearing although the pseudotachylite itself could have been formed by shearing processes at a considerable distance from its present site, and then injected by gas action. The author thus favours the action of hot gases as the agent of formation of the pseudotachylite as advocated by Reynolds (1954) and Bisschoff (1970). The pseudotachylite from the sedimentary rocks surrounding the Complex has a similar groundmass to the pseudotachylite from within the Complex, but the rock fragments and mineral grains are derived from the sedimentary rocks. The degree to which the included fragments in the pseudotachylite have been rounded

probably depends on the distance the material has been transported by the hot gases, with the more rounded fragments having been transported the furthest.

As mentioned previously Bisschoff (1972b) observed fragments of pseudotachylite occurring within an agglomeratic rock. In the view of the author this does not mean that there are two ages of pseudotachylite emplacement or that the main bodies of volcanic breccia which are cemented by andesite were formed after the pseudotachylite. This structure merely refers to the second period of fragmentation caused by the introduction of hydrothermal solutions slightly prior to, or simultaneous with the emplacement of the sulphides. The pseudotachylite may be replaced by sulphide minerals and thus must have formed before the period of sulphide mineralization.

The next phase of activity was the introduction of the sulphides with their attendant hydrothermal alteration. Finally, probably during Pilandsberg times, a narrow syenite-porphyry dike was intruded in the extreme south of the Complex. This dike has a chill zone up to 50 millimetres in width and, significantly, is the only rock type which has not been intruded by pseudotachylite. As to the age of the syenite-porphyry, it intruded after the pseudotachylite and probably after the sulphide mineralization. The fact that Bisschoff (1950, p. 28) observed copper minerals within this rock does not necessarily mean that it was emplaced prior to the copper mineralization, since it could have collected the copper minerals as it passed through the massive andesite which contains these minerals in fair abundance in the vicinity of the dike..

B. Crystallisation History of the Rock Types

1. Andesite/diorite group

Within the rock types of this group there is a common sequence of mineral deposition although minor variations in the order of crystallisation of the minerals do occur.

The accessory minerals, magnetite and ilmenite (in individual grains or in exsolution with one another) were the first to form. Apatite and sphene are the main accessory minerals present and are frequently poikilitically enclosed within the feldspar and mafic minerals. Plagioclase was the next mineral to form, as evidenced by the large size of individual crystals and the fact that it never encloses mafic minerals. The primary mafic minerals, which include augite, common hornblende and kaersutite, probably crystallised at the same time as the smaller crystals of plagioclase. The amphibole must have, to a certain extent, crystallised after the pyroxene since augite is partially enclosed by kaersutite (Plate XIV, 2). Some of the calcite, or the calcium which is now combined in the calcite molecule, must have been present as a primary constituent of the andesite. Evidence

in support of this statement is the fact that if, when calculating the 'norm' of the lavas from the chemical composition, all the CO_2 and an equivalent amount of Ca are first subtracted, the normative composition derived bears no resemblence to the minerals actually present in the rock. A much better result is obtained if the Ca and CO_2 are included in the calculation. Deuteric-stage minerals are present in abundance in all the rock types of this group, and include albite, which formed as an alteration product of the feldspar, uralite, actinolite, red-brown biotite and olivine pseudomorphs which are alteration products of mafic minerals, and quartz and calcite which were deposited in veins and interstitially to the other minerals. Due to the fact that not a single grain of olivine has been found in equigranular diorite, although olivine pseudomorphs are present, the alteration of the olivine must have been widespread and complete, probably during the deuteric-stage.

The last minerals to be deposited were those formed by hydrothermal alteration associated with sulphide deposition. These minerals are restricted in their occurrence to zones of hydrothermal alteration around sulphide bodies and include replacement-microperthite, olive-green to pale brown biotite, epidote, chlorite, hematite, goethite, sulphide minerals and zeolite (variety phillipsite). Albite is present in greater abundance around sulphide bodies than elsewhere, thus some albite must have been formed by the hydrothermal solutions attending sulphide deposition. Supergene alteration products formed from the sulphides and ilmenite.

Unusual features in the crystallisation history of the individual rock types of this group include:-

a. Amygdaloidal andesite

A notable feature in the crystallisation sequence of this rock is that some of the sphene is late formed being present as an alteration product of the uralite. Sphene may also form within the amygdales. The amygdales may have feldspar crystals aligned parallel to their edges (Plate II, 1), a feature which can probably be ascribed to the surface tension effects of the gas bubble. Sometimes the amygdales are elongated in the horizontal plane, parallel to the direction of flow of the lava. The viscosity of the andesite had probably been lowered by cooling so that the surface tension of the gas bubble was not strong enough to restore it to a spherical, or near spherical shape.

b. Volcanic breccia .

Hematite appears to be the last mineral to have formed within the volcanic breccia, since it occurs as a coating around other mineral grains, and was probably sublimated onto rock fragments by hot gases. In discussing the crystallisation of hematite Deer et al. (1967, 5, p. 24) states that it may be formed 'as a late-stage product of volcanic activities as thin crystals sublimated on to earlier material'.

c. Equigranular diorite

Arfvedsonite is present in this rock type as 'coronas' around grains of kaersutite and since the two amphiboles are in crystal continuity it seems probable that the development of arfvedsonite reflects a change to a more soda-rich composition in the solution from which the amphibole was crystallised. Quartz occurs at one locality within the equigranular diorite as a primary rock forming mineral where it is interstitial to the other primary minerals.

2. Pyroxene andesite/diorite group

The sequence of mineral deposition begins with the same minerals being formed as in the previous group of rocks, i.e. magnetite and ilmenite (in individual and composite grains), apatite and sphene. Augite was the next mineral to crystallise and is present in grains which attain a maximum diameter of ± 60 millimetres and may be partially or completely poikilitically enclosed within plagioclase crystals (Plate XIII, 4). The phenocrysts of plagioclase which may attain lengths of 24 millimetres were then formed, and were followed by smaller crystals of plagioclase. Due to the fact that the primary amphiboles, namely kaersutite and common hornblende, are never enclosed within the feldspar phenocrysts, they must have formed later, probably at the same time as the smaller crystals of plagioclase. Deuteric-stage minerals which are less abundant than in the andesite/diorite group include albite, as an alteration product of the plagioclase, uralite, actinolite and red-brown biotite, as alteration products of the mafic minerals, and finally quartz and calcite. Minerals formed by hydrothermal alteration associated with sulphide deposition include albite, olive-green to pale brown biotite, epidote, chlorite, goethite and the sulphide minerals. Supergene minerals include rutile and leucoxene, formed by alteration of ilmenite, and alteration products of the sulphide minerals.

A discussion of exceptions to this generalized sequence of crystallisation follows.

a. Pyroxene andesite

detailed under the section on 'Petrology of Individual Rock Types' there is a great difference in the relative size of the plagioclase phenocrysts and the plagioclase crystals present within the groundmass of the pyroxene andesite. The simplest explanation of this feature is probably that the phenocrysts formed before the lavas were extruded.

b. Coarse pyroxene diorite

The unaltered plagioclase within the coarse pyroxene diorite has a constant composition of An_{32}, suggesting that the plagioclase was not in contact with a rest liquid for long enough to cause widespread crystal zoning. In addition to this lack of zoning and the fact that the augite phenocrysts have a strikingly uniform composition the coarse pyroxene diorite occupies an anomalous position in the differentiation sequence of the pyroxene andesite/diorite group as derived from their geochemistry. Unusual field characteristics of this rock type are that even the narrowest bodies contain large feldspar phenocrysts which are in places arranged parallel to the contacts of the bodies. The writer feels that all these exceptions to the generalized sequence of crystallisation can be explained by presuming that the coarse pyroxene diorite was intruded as a semi-solid crystalline mush. If in the magma chamber magnetite, olivine, pyroxene and plagioclase had crystallised, and the crystals settled to the lower portion of the chamber, the rest liquid could then have been tapped off and intruded as the fine pyroxene diorite, while the crystal mush left behind subsequently intruded as the coarse pyroxene diorite.

c. Olivine gabbro

This rock type is regarded by the author as being a cumulate formed along the lower contact of a thick sill of coarse pyroxene diorite. Although the magnetite is usually present in euhedral grains, and was generally the first formed mineral, along the lower margin of the body of olivine gabbro it forms an interstitial groundmass around grains of olivine (Plate XIII, 1) and appears in this case to have crystallised after the olivine. The olivine occurs as small elliptical or rounded grains, generally less than 0,25 millimetres in diameter, which are enclosed within bronzite, augite or plagioclase grains. Both olivine and magnetite increase in quantity with depth. Bronzite crystallised before augite, as demonstrated by the fact that it is present as large

anhedral grains up t ,0 millimetres in length which are surrounded by smaller subhedral ains of augite with an average diameter of 0,9 millimetres. The bronzite decreases in quantity with depth while the augite reaches its maximum concentration in the centre of the olivine gabbro body. The plagioclase occurs as an interstitial groundmass between the abovementioned minerals and decreases markedly with depth. Other minerals which may occur as part of the interstitial groundmass are kaersutite and red-brown biotite, although these minerals are also present in small amounts around the edges of magnetite grains. Secondary minerals are completely absent from the olivine gabbro, as the deuteric-stage fluids probably concentrated in the upper regions of the coarse pyroxene diorite sill and there were no large sulphide bodies emplaced in the immediate vicinity.

d. Dioritic-pegmatite

The plagioclase and common hornblende present in the dioritic-pegmatite crystallised simultaneously as evidenced by an intergrowth texture between the two minerals (Plate XV, 2). Red-brown biotite, a green phyllosilicate and actinolite are developed along the contacts between the pegmatites and their wall rocks. Actinolite formed as an alteration product of hornblende while hematite, which is also present, could have formed either as a primary mineral or as an alteration product of hornblende. Deer et al. (1967, 5, p. 24) noted that hematite commonly crystallises from a melt with a low FeO content. The dioritic-pegmatite has a low FeO content due to the fact that it is composed predominantly of plagioclase feldspar.

e. Dioritic-aplite

The aplites contain large and small xenoliths of the wall rock and have a very similar mineralogy and geochemistry to the coarse pyroxene diorite.

3. The Late Intrusive group

Rock types within this group differ greatly in both mineralogy and field appearance and thus have no common crystallisation sequence.

a. Dolerite dikes

Magnetite formed first in the dolerite, followed by plagioclase phenocrysts which often enclose magnetite crystals within them and attain lengths of 1,85 millimetres. Augite, the next mineral to form, may be interstitial to the feldspar and intrude it along cracks, or it may partially enclose the feldspar laths in a sub-ophitic texture. Secondary minerals present

include epidote and chlorite.

b. Syenite dikes

The accessory minerals, namely apatite and magnetite, crystallised first in the syenite and were followed by plagioclase, which forms phenocrysts with a length of less than 5,0 millimetres and a groundmass of sub-hedral grains less than 0,40 millimetres in length. Common hornblende, arfvedsonite and red-brown biotite crystallised after the plagioclase phenocrysts, probably at the same time as the feldspar of the groundmass. Quartz was the last primary mineral to form since it is interstitial to the above minerals. Secondary minerals present include albite, epidote, chlorite, olive-green to pale brown biotite and red-brown iron oxide. At the surface the syenite has a distinctive red colour which must be due to the oxidation of iron since the red colour is absent where the dikes are intersected at depth.

c. Syenite-porphyry dike

Magnetite and apatite crystallised first in this rock type and may be enclosed within the phenocrysts of plagioclase and barkevikite. The plagioclase phenocrysts reach a maximum length of 11,0 millimetres and are commonly broken or cracked indicating that the plagioclase phenocrysts were not formed in situ, in addition it is unlikely that such large crystals could form in a dike only 1,3 metres wide with strongly chilled margins. The barkevikite phenocrysts are also cracked and broken, have a maximum length of 4,5 millimetres and were probably transported to their present site. The groundmass of the rock is predominantly plagioclase, with individual grains averaging 0,25 millimetres in length. Subhedral grains of hornblende occur within the groundmass and were probably formed at the same time as the plagioclase. Anhedral grains of orthoclase were probably the last to form, since they are interstitial to the other minerals. Secondary minerals are rare and generally confined to albite.

C. Differentiation Trends present within the Roodekraal Igneous Complex

Various diagrams have been used in the past to demonstrate the presence of differentiation trends in rocks. Several variation diagrams which appear to demonstrate the presence of differentiation trends within the rocks of the Roodekraal Igneous Complex are presented in Appendix II while an explanation of these diagrams is given below.

The variation of the major elements during differentiation will be dealt with first. Since the Roodekraal Igneous Complex probably has the same parent magma as the Losberg Intrusion, and because the mafic rock types from the Roodekraal Igneous Complex have a similar chemical composition to the rock types of the Losberg Intrusion, the chemical changes taking place during differentiation should be much the same for both igneous bodies. Danchin and Ferguson (1970) studied the geochemistry of the Losberg Intrusion and concluded that during differentiation the SiO_2, Al_2O_3, Na_2O and K_2O contents increased while the MgO, CaO, MnO contents and to a lesser extent the (FeO + Fe_2O_3) content decreased. The TiO_2 content showed a steady enrichment until ilmenite was precipitated and thereafter a gradual decrease. Danchin and Ferguson noted that these variations during differentiation were similar to those inferred for the Bushveld Igneous Complex.

Considering the AFM diagram of the Roodekraal rocks (Fig. i, Appendix II) it can be noted that the pyroxene andesite/diorite group lie along a definite line which probably represents a differentiation trend. Presuming that later differentiates contain more alkalis the possible differentiation sequence runs from RC 22→ RC 18 →RC 19→ RC 20→RC 21.

The diagram of percentage Al_2O_3 plotted against C.I.P.W. normative plagioclase composition (Fig. ii, Appendix II) has a possible differentiation sequence running from RC 22→RC 18→RC 20→RC 19 →RC 21 for the rocks of the pyroxene andesite/diorite suite. This sequence is derived presuming that the later differentiates contain more Al_2O_3.

A trend is present in the alkalis-silica diagram (Fig. iii, Appendix II) running from RC 22→RC 18→RC 20→RC 21→RC 19 for increasing alkalis and silica.

The ternary diagram Cpx - Ol - Opx (Fig. iv, Appendix II) will not be considered in this discussion, since only the rocks with (nepheline + quartz)= 0 in their norms are plotted on this diagram.

Nockolds and Allen (1953) used the ternary diagram Na-K-Ca to demonstrate differentiation trends in granodiorite, syenite, diorite and gabbro, thus this diagram would seem suitable for the rocks at Roodekraal. Nockolds and Allen found that when their analyses were plotted on the ternary diagram Na-K-Ca they tended to lie along a smooth curve which represented a differentiation trend, and had a more or less random scattering at the basic end which represented the composition of the parental magma.

Considering the Na-K-Ca ternary diagram for the rocks at Roodekraal (Fig. v, Appendix II) the features discussed below may be observed.

(i) All the analyses lie along a broad curve although those of the andesite

and equigranular diorite are more or less randomly scattered. However this borad curve is probably not a differentiation trend since the andesite which is the oldest rock type, from the evidence of its field relationships, occupies the centre of the curve with younger intrusive rocks on either side of it.

(ii) If we now examine the analyses of pyroxene andesite/diorite group of rocks separately a possible differentiation trend can be inferred running from RC 22→RC 18→RC 19→RC 21→RC 20 for increasing alkalis.

The ternary diagram ($FeO + 0{,}8998Fe_2O_3$)-MgO-TiO_2 (Fig. vi, Appendix II) gives a similar picture, with the analyses of the pyroxene andesite/diorite group lying along a straight line. The differentiation trend in the pyroxene andesite/diorite group runs from RC 22→RC 20→RC 18→RC 19→RC 21 for decreasing MgO content.

The ternary diagram ($FeO + 0{,}8998Fe_2O_3 + TiO_2$)-MgO-$Al_2O_3$ (Fig. vii, Appendix II) is almost identical to the above in the relationships it portrays between the various rock types. The pyroxene andesite/diorite group differentiation trend in this diagram being RC 22→RC 18→RC 20→RC 19→RC 21 presuming later differentiates have a higher Al_2O_3 content.

When the modal analyses of the three rock types of the pyroxene andesite/diorite suite i.e. pyroxene andesite, fine pyroxene diorite and coarse pyroxene diorite, are examined, (see Table XXVII) it can be observed that the differentiation is reflected in the mineralogy.

TABLE XXVII

Comparison of the modal analyses of the rocks of the pyroxene andesite/diorite group

Mineral	Rock types and sample numbers		
	RC 22 & RC 21 Coarse pyroxene diorite	RC 20 & RC 19 Fine pyroxene diorite	RC 18 Pyroxene andesite
Plagioclase	63,5	71,4	70,9
Pyroxene	19,3	9,8	4,5
Amphibole	8,2	7,1	2,6
Other alteration products	1,3	2,2	6,9
Ore minerals	7,1	8,9	15,1
Apatite	0,6	0,6	-
Totals	100,0	100,0	100,0

NOTE: It should be borne in mind during the following discussion that the sequence of formation of the various rock types of the pyroxene andesite/diorite group, as determined by their field relationships, was pyroxene andesite (RC 18) followed by fine pyroxene diorite (RC 19 and RC 20) and finally coarse pyroxene diorite (RC 21 and RC 22).

It can be noted from the data presented in Table XXVII that there is an increase in pyroxene and amphibole content with a progressively more recently formed rock type and a corresponding decrease in the content of the ore minerals and plagioclase. These two opposing trends in the mineralogy must cause two

trends in the chemistry. Magnesium, which is mainly present in the mafic minerals, should increase with a progressively younger host rock, which is the opposite of the normal sequence. Calcium is present in both the plagioclase and the mafic minerals, but since it is present to a greater proportion within the mafic minerals the quantity of calcium present should increase with a progressively younger host rock. Iron should vary in a complex manner since with differentiation in the host rock the increase in mafic minerals causes an increase in the iron content while the decrease in ore minerals present causes a corresponding decrease in the iron content. The sodium which is mainly present within the plagioclase should decrease, as the quantity of plagioclase decreases, with a progressively more recently formed rock type.

If the components of the ternary diagrams are considered in a series of two component diagrams the nature of the variation of single components can be more easily appreciated.

In considering the variation of Fe_2O_3 with respect to FeO (Fig. viii, Appendix II) once again the rocks of the pyroxene andesite/diorite group lie along a curved line. However the oxidation ratio could not have increased progressively as differentiation proceeded since, if it is presumed that rocks which were formed at a later stage have a higher Fe_2O_3/FeO ratio than early formed rocks, this diagram indicates a completely different order of formation for the pyroxene andesite/diorite group than that which is obtained from the ternary diagrams. The differentiation trend which would be inferred from this diagram is RC 20→RC 21→RC 19→RC 18→RC 22 which is the opposite to the trend obtained from all the preceding diagrams.

In Figure ix, Appendix II, where $(FeO + 0{,}8998Fe_2O_3)$ is plotted against TiO_2, the later formed rock types appear to contain less iron and titanium. Thus the differentiation sequence which would be inferred from this diagram for the pyroxene andesite/diorite group should be RC 22→RC 18→RC 20→RC 19→RC 21. The pyroxene andesite/diorite group of rocks appear, on the average, to contain a greater proportion of TiO_2 than the andesite and equigranular diorite.

Figure x, Appendix II, is a graph of $(FeO + 0{,}8998Fe_2O_3)+TiO_2$ against MgO and gives an inferred differentiation sequence for the pyroxene andesite/diorite group of RC 22→RC 18→RC 20→RC 19→RC 21 presuming that later differentiates contain less magnesium, iron and titanium. The andesite and equigranular diorite occur as a scattered group and the syenite and syenite-porphyry were the last rock types to form.

The variation of the minor elements often gives a more reliable result than the variation of the major elements when an attempt is made to derive a differentiation trend. Philpotts and Schmetzler (1969) concluded that fractional crystallisation in a basaltic magma leads to higher K, Ba and Rb

concentrations, slightly higher K/Sr, Ba/Sr and Rb/Sr ratios and slightly lower K/Rb and Ba/Rb ratios in the rest liquid. If Tables XII and XIII are examined, bearing in mind the findings of Philpotts and Schmetzler (1969) and the sequence of formation of the various rock types as determined from field relationships, the points listed below may be noted.

(i) The increase in rubidium, potassium, K/Sr and Rb/Sr values generally agree with an increasingly younger age in the rock types. Exceptions are that the andesite has a proportionally greater content of rubidium than the dolerite and the rocks of the pyroxene andesite/diorite group. The coarse pyroxene diorite contains less rubidium and potassium than the fine pyroxene diorite and the pegmatite and its wall rocks contain very little rubidium or potassium.

(ii) The decrease in K/Rb ratio with an increasingly younger age in the rock types is only very poorly marked. However the decrease in the Ca/Sr ratio with increasingly younger age in rock types gives a good correspondence with the only real exceptions being RC 24 (wall rock to the dioritic-pegmatite), RC 28 (dolerite) and RC 30 (syenite) which have abnormally high values for the Ca/Sr ratio.

In Figures xi, xii, xiii and xiv the author has plotted Rb to Sr, K to Rb, K to Rb/Sr and Ca to Sr respectively. A strong trend in the analyses for the pyroxene andesite/diorite group indicates a differentiation sequence of RC 22 →RC 21→RC 20→(RC 18→RC 19) with only the bracketed pair altering in position.

Probably the best method of determining genetic trends in chemical analyses and the causes of these trends, is by using the technique of Pearce (1968). This method involves dividing the two variables under consideration by a third parameter which is constant and will not effect the relationship between the two variables. In the 'Pearce diagrams' plotted for the rocks at Roodekraal, only the analyses of larger bodies were used and all special cases such as pegmatites and their wall rocks were excluded. In the analyses chosen Al_2O_3 maintains a fairly constant value except in RC 21. In RC 21, which is the analysis of the course pyroxene diorite from the most southerly outcrop in square E 4 of Plate XXIV, the Al_2O_3 content is considerably higher than in the other samples. The ratios used in the comparisons of MgO/Al_2O_3 to SiO_2/Al_2O_3 (Fig. xv (a), Appendix II) and CaO/Al_2O_3 to SiO_2/Al_2O_3 (Fig. xv (b), Appendix II) are all molar ratios. Thus for example the molar ratio of MgO/Al_2O_3 is calculated by dividing weight per cent MgO/molecular weight MgO by weight per cent Al_2O_3/molecular weight Al_2O_3. In Figures xv (a) and xv (b) two nearly parallel straight lines were obtained on each of the diagrams. The lower line represents the crystallisation trend of the andesite and equigranular diorite while the

upper line represents the crystallisation trend of the pyroxene andesite/diorite in both diagrams. The theoretical curves for the minerals olivine, orthopyroxene, augite, apatite and plagioclase are derived from the average value of their composition as determined by optical and X-ray methods (see section entitled 'Mineralogy of Rock Forming Minerals'). The average values used were Fo_{81} for olivine, En_{81} for orthopyroxene, $Wo_{40}En_{38}Fs_{22}$ for clinopyroxene and An_{30} for plagioclase. Since apatite contains no silica the slope for the curve CaO/Al_2O_3 versus SiO_2/Al_2O_3 must be infinity or parallel to the ordinate.

Considering Figure xv(a), Appendix II, the slope of the trend for the pyroxene andesite/diorite group running from RC 22→RC 18→RC 20→RC 19, but excluding RC 21 (which has an abnormally high Al_2O_3 content), lies between the theoretical slopes of olivine and orthopyroxene of the compositions present at Woodekraal. Thus the trend in the pyroxene andesite/diorite group could have been caused by the fractionation of olivine and orthopyroxene. The observed trend in the andesite and equigranular diorite appears to be due to the fractionation of orthopyroxene. In Figure xv(b), Appendix II, the trend for the pyroxene andesite/diorite group could have been caused by crystal fractionation of the primary calcium-rich minerals, namely apatite, augite and plagioclase, since the slope of the pyroxene andesite/diorite trend lies between the slopes determined for apatite and the other two minerals. The trend for the andesite and equigranular diorite could have been caused by the fractionation of the same minerals as in the case of the pyroxene andesite/diorite group. The fact that the slope of the pyroxene andesite/diorite group is steeper than that of the andesite and equigranular diorite can be attributed to the fact that the former group of rocks contain more apatite than the latter group.

The most important conclusion which can be drawn from the 'Pearce diagrams' is the fact that there are two distinct differentiation trends, one for the pyroxene andesite/diorite group and another for the andesite and equigranular diorite. This underlines the necessity of distinguishing between these two groups of rocks. It can also be observed that the pyroxene andesite/diorite group of rocks may owe their origin to crystal fractionation.

Summary of the differentiation sequence in the pyroxene andesite/diorite group

Several conflicting differentiation trends have been derived from the foregoing variation diagrams. The most important sequences being:-

(i) RC 22→RC 18→RC 20→RC 19→RC 21 sequence obtained from the variation of the major elements in Figures ii, vii, ix, x, and xv(a) and (b).

(ii) RC 20→RC 21→RC 19→RC 18→RC 22 sequence obtained from the oxidation ratio (Fe_2O_3/FeO) in Figure viii.

(iii) RC 22→RC 21→ RC 20→(RC 18→RC 19)sequence obtained from the variation of the minor elements in Figures xi, xii, xiii and xiv.

A possible explanation of the fact that the last rocks to be emplaced occur first in the differentiation sequence is that the coarse pyroxene diorite could have intruded as a semi-solid crystalline mush as explained under the section entitled 'Crystallisation History of Rock Types' and that the crystallisation of this rock type, within the magma chamber, could well have begun before that of the fine pyroxene diorite even though the coarse pyroxene diorite was intruded at a later date. If in the magma chamber, magnetite, olivine, pyroxene and plagioclase had crystallised and the crystals had settled to the lower portion of the chamber, the rubidium would have been enriched relative to strontium in the rest liquid, due to the fact that rubidium cannot easily enter the crystal lattices of these minerals, but plagioclase may act as a host to strontium. If it is now presumed that the rest liquid was tapped off to be intruded as the fine pyroxene diorite, this rock would have a higher rubidium to strontium ratio than the crystal mush left behind which was subsequently intruded as the coarse pyroxene diorite.

If the differentiation sequences as established from the variation diagrams are reconsidered in the light of this new information, the sequence obtained from the variation of the minor elements seems perfectly logical, since it shows that the coarse pyroxene diorite formed first and the fine pyroxene diorite and pyroxene andesite formed last. The sequence obtained from the variation of the major elements differs notably from that of the minor elements in that RC 21 (coarse pyroxene diorite) occurs last, which can be explained by the fact that the RC 21 sample contains much more feldspar than the other samples and consequently has higher contents of Al_2O_3, Na_2O and CaO. The explanation of the differentiation sequence determined from the oxidation ratio is that the premise that the oxidation ratio increased progressively during differentiation is invalid, due to the fact that the partial pressure of oxygen varied throughout the cycle of differentiation.

D. Nature of the Parental Magma

1. Parental Magmas of Mafic Igneous Rocks

A basaltic magma has long been regarded as the parent of the majority of mafic igneous rock series. Two principal basalt magmas have been recognised, namely the alkali-olivine basalt and the tholeiite. The two parent magmas are distinguished primarily on a basis of their mineralogical characteristics. Kennedy (1933), Tilley (1950) and Turner

and Verhoogen (1960) broadly defined the tholeiite to be composed of two pyroxenes, generally augite together with hypersthene or pigeonite, basic plagioclase, iron ores and possibly small amounts of magnesium-rich olivine. The alkali-olivine basalt was defined as being composed mainly of olivine, one pyroxene, generally augite, which is commonly of a titaniferous variety, basic plagioclase and iron ore. Apart from mineralogical considerations an important distinction between the tholeiite and the alkali-olivine basalt is the fact that, as noted by Chayes (1965, 1966), mafic volcanics with nepheline in the norm are alkaline while those with quartz in the norm are sub-alkaline.

The relationship between these two parent magmas has been the subject of some controversy. Bowen (1928) regarded the alkali-olivine basalt as being the parent and suggested that tholeiite was derived from it by the removal of early formed olivine. Daly (1933) also decided that the alkali-olivine basalt was the parent and thought that tholeiite was developed when the alkali-olivine basalt assimilated granitic material. Kennedy (1933) postulated a difference in composition in the earth's crust beneath continents on the one hand, giving rise to tholeiite, and beneath oceans on the other hand, giving rise to alkali-olivine basalt. He decided that the alkali-olivine basalt was composed essentially of olivine, augite, calcic plagioclase and iron ore, while the tholeiite was composed essentially of pyroxene, calcic plagioclase and iron ore. He came to the conclusion that the calc-alkaline suite of igneous rocks are differentiates of the alkali-olivine basalt. Tilley (1950) agreed broadly with ideas set out by Kennedy (1933) but suggested that the tholeiite might be the source magma for both lines of descent. Turner and Verhoogen (1960) suggested that the two varieties of basalt were generated by fusions at different levels in the peridotite mantle below the Mohorovičić discontinuity. Yoder and Tilley (1962) concluded that the tholeiite and the alkali-olivine basalt may be generated from a single parent, and that the basalt type generated is probably dependant on pressure.

Yoder and Tilley (1962) use a basalt tetrahedron to sub-divide the basaltic rocks. This tetrahedron has two important planes within it, namely the clinopyroxene - olivine - plagioclase (plane of critical undersaturation) and the clinopyroxene - orthopyroxene - plagioclase (plane of quartz saturation), which divide the basalts into five unique groups according to their normative components, these groups are:-

1. Tholeiite (oversaturated) normative quartz and hypersthene.
2. Tholeiite (saturated; hypersthene basalt) normative hypersthene.
3. Olivine tholeiite (undersaturated) normative hypersthene and olivine.
4. Olivine basalt : normative olivine.
5. Alkali basalt : normative olivine and nepheline.

Kuno (1968) distinguishes between three different parental basalt magmas, namely tholeiite, alkali-olivine basalt and high-alumina basalt. His high-alumina basalt contains more than 16,5 per cent Al_2O_3 in aphyric rocks and the $Na_2O + K_2O$ contents lie between those of the other two basalt types for a given SiO_2 content. The high-alumina basalt is transitional between the other two types.

2. The Parental Magma of the Roodekraal Igneous Complex

The Vredefort Ring Structure is intruded by two suites of igneous rocks. The older suite which lies closest to the centre of the Vredefort Ring Structure is tholeiitic being composed of dolerite, hyperite and norite sills (Bisschoff, 1972a). The younger suite includes three bodies of igneous rock with alkaline chemical compositions. The satellitic bodies at Baviaanskranz and Schurwedraai are composed of alkali granite while the Rietfontein Complex is only partly composed of alkali granite. Nepheline-syenite dikes cut all the bodies of alkali granite.

The author is in agreement with the widely held opinion that the Roodekraal Igneous Complex, together with the Schurwedraai and Baviaanskranz Plutons, the Rietfontein Complex and the Lindequesdrift body, is a soda-rich differentiate of a mafic magma body lying beneath the Vredefort Ring Structure (Hall and Molengraaff, 1925; Nel, 1927; Bisschoff, 1950, 1970 and 1972b). The abovementioned authors do not suggest to what magma type the concealed body below the Vredefort Ring Structure may belong. Bisschoff (1972a) and (1972b) suggests, and the author agrees that the mafic magma body was the source of both the older and the younger suites of intrusive igneous rocks.

The rock types at Roodekraal do not readily lend themselves to definition by geochemical means since deuteric and late-stage hydrothermal alterations have modified their original chemistry. Despite this limitation an attempt is made to define the andesitic and dioritic rocks of the Roodekraal Igneous Complex using the following methods:-

I. The average composition of the andesite of the Roodekraal Igneous Complex is compared with well known average compositions of andesite. The andesite flows should give the best indication of the nature of the parent magma since 'the quenching to which they have been subjected on extrusion, reveals their nature prior to extrusion, whereas in granular rocks the continued outgrowth of crystal boundaries obscure the stages of their development' (Bowen, 1928).

With the exception of a few minor variations, the Roodekraal andesite has an almost constant composition. To get a preliminary indication as to which suite of igneous rocks the andesite at Roodekraal belongs the chemical analyses of the Roodekraal andesite are compared with well known analyses (Table XIV, p. 29). This method can be used, since most methods of distinguishing between the various suites of basaltic rocks involve the use of chemical analyses or normative minerals. From an examination of Table XIV it can be concluded that the andesite present at Roodekraal is closest in composition to the 'average tholeiitic andesite' of Nockold's (1954). The average andesite from Roodekraal however, has lower MgO, CaO, TiO_2, Fe_2O_3 and Al_2O_3 contents and higher Na_2O, K_2O, FeO and CO_2 contents. The high K_2O content is probably due to late-stage hydrothermal alteration which resulted in the development of large amounts of olive-green to pale brown biotite.

2. The mineralogy and chemistry of the late-stage differentiates present within the Roodekraal Igneous Complex are compared with the mineralogy and chemical analyses for late-stage differentiates of tholeiite and alkaline-olivine basalt. Kennedy (1933) differentiates between the segregation veins, 'pegmatitoides', associated with alkali-olivine basalt and those associated with tholeiite, on a basis of their mineralogy. The veins associated with the alkali-olivine basalt magma contain no quartz, some feldspathoid, abundant potash and alkali feldspar, soda bearing pyroxene and amphibole and iron ore. Those veins associated with the tholeiite contain quartz, no feldspathoid, acid plagioclase, subordinate orthoclase mainly present as micropegmatite, augite or less frequently amphibole and a little iron ore. Except for the fact that an amphibole is the chief mafic mineral of the dioritic-pegmatite at Roodekraal, the mineralogical composition of this rock type fits well with that of the late-stage differentiates of the tholeiite.

The chemical composition of the pegmatite veins at Roodekraal is compared with the chemistry of the late-stage differentiates of olivine-basalt and tholeiite in Table XXVIII below.

TABLE XXVIII

Comparison between the average chemical compositions in weight per cent of late-stage differentiates of olivine-basalt, tholeiite and the Roodekraal Igneous Complex

	1. Pegmatitoides of olivine-basalt magma	2. Pegmatitoides of tholeiitic magma	3. Dioritic-pegmatite from Roodekraal Igneous Complex
SiO_2	49,19	68,96	55,28
Al_2O_3	14,75	12,55	16,67
Fe_2O_3	5,13	1,64	-
FeO	5,61	2,73	7,29
MgO	3,22	1,49	2,37
CaO	7,83	2,76	6,25
Na_2O	4,22	5,24	7,19
K_2O	2,85	1,74	0,34
H_2O+	2,35	1,41	1,50
H_2O-	1,00	0,27	0,15
TiO_2	3,22	0,60	1,75
P_2O_5	0,60	0,31	0,18
MnO	0,15	0,04	0,06
Rest	0,00	0,72	0,08
Total	100,12	100,46	99,12

Values quoted in columns 1 and 2 are arithmetic means of the values given in Kennedy 1933, p. 245, Table III.
Values given in column 3 represent the arithmetic mean of analyses RC 23 and RC 25 (Analyst Gold Fields Laboratories (Pty) Ltd.)

From Table XXVIII it can be observed that dioritic-pegmatite from Roodekraal has a chemical composition which lies between those of the pegmatitoides from olivine-basalt and tholeiite but closer to those from the olivine-basalt.

3. The procedures employed by Irvine and Baragar (1971) are used to classify the rock types of Roodekraal according to three parental magma groups established by Kennedy (1933) and Tilley (1950), namely, tholeiite, alkaline-olivine basalt and calc-alkaline basalt. Irvine and Baragar divide the basalts into alkaline and sub-alkaline groups, where the term 'sub-alkaline' is used in the same manner as Wilkinson (1968) to include calc-alkali basalt and tholeiite and to distinguish them from alkaline and peralkaline rocks.

Firstly it must be decided whether the rocks are members of the alkaline group or sub-alkaline group. Considering the alkalis-silica diagram (Fig. iii, Appendix II) the majority of the rock types lie in the alkaline field, however Irvine and Baragar (1971) themselves admit that the alkalis-silica diagram has shortcomings that result in the composition of some tholeiitic rocks lying within the alkaline field, and conclude that 'the diagram has definite limitations, and should be used with care'. The second variation diagram used by Irvine and Baragar (1971) was the ternary system olivine - nepheline - quartz. This diagram is totally unsuitable for the rocks from Roodekraal, which in the majority of cases only have one of the three end members olivine, nepheline and quartz in their norms.

The method used by Irvine and Baragar (1971) which is the most effective in classifying the Roodekraal rocks is the procedure of Chayes (1965, 1966). Chayes noted that basic volcanics with nepheline in the norm are alkaline, while those with quartz in the norm are sub-alkaline and he suggests the use of the clinopyroxene - olivine - orthopyroxene ternary diagram in deciding whether rocks with (nepheline + quartz) = 0 in their norms are alkaline or sub-alkaline. Chayes derived two dividing lines for his ternary diagram (Fig. iv, Appendix II), a linear discriminant function and a curve which is a combined linear and quadratic function. He found that the curve gave the best discrimination between the alkaline and sub-alkaline fields.

Of the rocks at Roodekraal, considering first the lavas of which there are sixteen analyses in all:-

Seven samples have quartz in the norm and are therefore sub-alkaline.

One sample has nepheline in the norm and therefore cannot be tholeiitic.

Eight samples have normative (nepheline + quartz) = 0 (see Fig. iv). Five of these fall within the curved sub-alkaline field of Chayes and three outside it.

Thus of the sixteen samples of lava, twelve indicate a sub-alkaline composition.

If the same rules are applied to the intrusive rocks we find that four have quartz in the norms and are sub-alkaline, five have nepheline in the norms and are alkaline while the remaining six rocks with normative (nepheline + quartz) = 0 fall into the alkaline field of Figure iv. It must be remembered however that none of the rock types present at Roodekraal contain nepheline in their modes, and the nepheline present in the norm could be a result of the presence of secondary amphiboles.

A final factor used by Irvine and Baragar (1971) to distinguish between sub-alkaline and alkaline rocks is that sub-alkaline rocks typically show orthopyroxene in the norm. Orthopyroxene is present in the norms of all the samples taken, except the equigranular diorite, massive andesite, coarse pyroxene diorite, the dioritic-pegmatite and its wall rock (see Appendix I). Thus orthopyroxene is present in the norms of twenty five of the thirty one samples analysed.

Where the rock types have been designated as sub-alkaline a distinction must be made between the tholeiitic and calc-alkaline series. Considering the AFM plot of the Roodekraal rocks (Fig. i, Appendix II) it can be noted that twenty eight of the thirty one rock types analysed fall within the tholeiitic field.

In Figure ii, Appendix II where the weight percentage of Al_2O_3 is plotted against the normative plagioclase composition, it is observed that the majority of the rock types lie in the tholeiitic field. The rock types lying in the calc-alkaline field contain abundant plagioclase phenocrysts and include coarse pyroxene diorite, dioritic-pegmatite and aplite, syenite and syenite-porphyry.

4. If the sixteen samples of lava which have been analysed from Roodekraal are defined according to the basalt tetrahedron of Yoder and Tilley (1962) we have :-

Seven samples have quartz and hypersthene in the norm and are thus tholeiitic.

Eight samples contain olivine and hypersthene and are thus olivine tholeiites.

One sample contains olivine and nepheline and is thus an alkali basalt.

However, due to the fact that the rocks which fall into the olivine tholeiite group contain no olivine when examined under the microscope, and since the nepheline in the norm could be a result of the presence of secondary amphibole, the author feels it would be more reasonable to name these lavas tholeiites.

To draw a conclusion as to the nature of the parental magma of the Roodekraal Igneous Complex, a resumé of the findings from the classification methods described above is helpful.

1. From the direct comparison of the average composition of the Roodekraal andesite with well known world averages it was found that the andesite present at Roodekraal is closest in composition to the 'average tholeiitic andesite' of Nockolds 1954.

2. The comparison between the mineralogy and chemistry of the late-stage differentiates of the Roodekraal Igneous Complex and alkali-olivine basalt and tholeiite shows that although the mineralogy of the pegmatite is clearly similar to that of the tholeiitic differentiates, the chemical composition is closest to that of the alkali-olivine basalt.

3. Using the methods of Irvine and Baragar (1971), the alkalis-silica diagram (Fig. iii, Appendix II) indicates that most of the rock types at Roodekraal lie in the alkaline field. The methods of Chayes (1965, 1966) indicate that sixteen of the rock types are tholeiitic and fifteen alkaline. Orthopyroxene is present in the norms of twenty five of the thirty one samples analysed indicating that the vast majority of rock types have a sub-alkaline composition.

4. When the sixteen analyses of lavas are defined according to the basalt tetrahedron of Yoder and Tilley (1962) all but one indicate a tholeiitic composition.

It can be concluded from the above summary that the majority of evidence points to the parental magma of the Roodekraal Igneous Complex having a tholeiitic composition. An explanation of the fairly frequent exceptions to this rule is present in the fact that the rock types of the Roodekraal Igneous Complex do not really lend themselves to strict tholeiitic or alkaline-olivine basalt classification. The rock types at Roodekraal have an andesitic and dioritic rather than a basaltic composition and have undergone secondary hydrothermal alteration which has modified the chemical composition by rendering it more alkaline (see section entitled 'Hydrothermal Alteration'). Evidence supporting the conclusion that the parental magma of the Roodekraal Igneous Complex has a tholeiitic composition is that, as previously stated, the parental magma of the Roodekraal Igneous Complex was also the source of an earlier suite of tholeiitic rocks.

E. Classification of the Rock Types

The classification according to Streckeisen (1967) was designed to fit in with the current usage of the names for the various rock types, and defines the rock types according to their position in the double triangle; quartz - alkali feldspar - plagioclase - feldspathoid.

The majority of the intrusive and extrusive rock types at Roodekraal should be called diorite and andesite respectively, according to the classification schemes of Holmes (1920), Johannsen (1937), Shand (1950) and Streckeisen (1967). The author feels however that a division must be made between the two groups of dioritic rocks present at Roodekraal which have different macroscopic appearances, mineralogy and chemical compositions. The nature of the dark minerals present was chosen as the criterion for distinguishing between the two groups of rocks since, as is detailed in the section entitled 'Petrology of Individual Rock Types', pyroxene is the chief dark mineral present wit' in the porphyritic non-amygdaloidal rocks which are accordingly called pyroxene andesite and pyroxene diorite. Amphibole is the most abundant dark mineral in the equigranular, amygdaloidal rocks which are simply termed diorite or andesite depending on whether they are intrusive or extrusive. As Bisschoff (1970) points out, the change from amphibole crystallization to pyroxene crystallization marks the crossing of an important physio-chemical boundary, so that it is desirable to distinguish between pyroxene rich rocks and rocks in which amphibole is the predominant mafic mineral.

The pyroxene diorite would correspond to the soda gabbro of Bisschoff (1970), which he defined as being composed predominantly of clinopyroxene, and plagioclase with a composition of (An $<$ 50).

Where important textural differences are present they are used to sub-divide rock types, for example the massive andesite is distinguished from the amygdaloidal andesite by its lack of amygdales.

Several rock types are present in the Roodekraal Igneous Complex which are neither andesite nor diorite.

As far as the naming of segregation veins present within the Roodekraal Igneous Complex is concerned, the classification scheme of Streckeisen (1967) is not detailed enough to include these rock types. Originally it was suggested that the term 'pegmatite' only be used for coarse-grained rocks of granitic composition with a graphic texture, and that the term 'pegmatoid' be used for those without a graphic texture, Holmes (1920). Contemporary usage (Jahns and Burnham, 1969) seems to be to employ the term 'pegmatite' for a wider range of coarse-grained rock types without graphic texture, and to use a prefix to qualify the rock type more closely e.g. syenite-pegmatite. The writer uses the term dioritic-pegmatite to describe the coarse-grained segregation veins present at Roodekraal since although they have a silica content of up to 58,80 per cent (as in sample RC 23) and contain primary quartz, they have plagioclase as the feldspar and also contain hornblende and actinolite. Grout (1932) used the term aplite to denote a granitic rock with an average grain size of less than one millimetre. The author uses the term dioritic-aplite for the fine-grained segregation veins present at Roodekraal.

The olivine rich rocks present at Roodekraal do not have a constant mineralogy, however all samples examined have dark minerals making up the majority of the rock, and plagioclase with a composition of An_{41} as the only feldspar and all contain olivine. The author decided to call these rocks 'olivine gabbro' since they fulfill all the conditions required by Streckeisen (1967) for classification as a gabbro, except that the plagioclase present does not have an anorthite content of greater than 50 per cent.

Two narrow dikes having a dioritic composition were classified as dolerites by Bisschoff (1950, p. 30). Since these rocks have a sub-ophitic texture, occur as dikes and are predominantly composed of plagioclase and augite, the author agrees with this classification.

The pale dike rock which is widespread within the Complex and is composed mainly of plagioclase feldspar is classified by Bisschoff (1950) under two names, firstly as an oligophyre, when it is porphyritic and

contains little quartz, and secondly as plagi-aplite, when it is less porphyritic and contains more quartz. Bisschoff (1972b), however, classifies this pale dike rock, together with the rock type which contains abundant phenocrysts of plagioclase and barkevikite and forms a narrow dike in the extreme south of the Complex, as albitite. The author regards that a simpler classification than that of Bisschoff (1950), should be used, so that the pale dike rock, which is uniform in appearance and composition, may be referred to by a single name. The author is further of the opinion that the term albitite employed by Bisschoff (1972b) is too general in that it includes two distinct rock types with differing petrology, texture and field appearance. From the petrology of the dike rocks in question it does not appear that albite is the dominant variety of plagioclase present, except in the vicinity of sulphide bodies where late-stage hydrothermal alteration has resulted in the conversion of more basic plagioclase to albite. Bisschoff (1950) in discussing the narrow dike in the extreme south of the Complex states 'Die veldspaat is min verander behalwe vir 'n bietjie verwering na klei minerale', and as previously recorded, Bisschoff (1950) had determined the composition of this plagioclase to be An_{20}.

According to the classification of rock types of Streckeisen (1967) the pale dike rock described in the preceeding paragraph would be called a syenite. Morse (1968) in a discussion on syenite describes a specific variety of syenite which appears to be very similar to the rock type present at Roodekraal. He calls this variety 'hydrous-syenite' and states that it contains amphibole or biotite and generally has a silica oversaturation trend. The high water content of the syenite at Roodekraal, 3,76 per cent for RC 30, suggests a hydrous nature. The other analysis for syenite i.e. RC 27 has a low water content but since this specimen was taken from outcrop the water content is not reliable, whereas RC 30 analysis should be reliable since it is a channel sample of a dike intersected at depth in a borehole. The author, however, for the sake of simplicity, uses the term syenite for the rock type at Roodekraal.

The rock type which is present as a narrow dike in the extreme south of the Complex, which cuts the massive andesite and has not been intruded by pseudotachylite was originally classified by Willemse (Nel et al. 1939, p. 9) as a syenite-porphyry. Bisschoff (1950) felt, however, that since it contained no K-feldspar it should be called a diorite-porphyry. As previously discussed Bisschoff (1972b) classifies this rock type, together with the syenite, as albitite although albite is not the dominant plagioclase present in this rock type. It has been stated by

Morse (1968) that a K-rich feldspar may occur in very late-stages, but is not necessary for a syenite which is always dominated by a lime poor plagioclase. The author feels that for the above reason and because of the fact that a little orthoclase is present, the rock should be classified as a syenite-porphyry according to the original suggestion by Willemse. The author concludes from the field evidence, the similar geochemistry to the Pilandsberg syenite at Leeufontein and the fact that a Pilandsberg syenite dike cuts the Rietfontein Complex, that this dike is in fact a Pilandsberg syenite. A notable feature in the field evidence is the fact that this dike has margins which have been more strongly chilled than the margins of the other dike rocks of the Complex, indicating that the syenite-porphyry intruded at a later stage than the other dikes, when the host rocks had cooled to a greater extent.

F. Genesis and Classification of the Sulphide Bodies

1. Paragenesis of the Ore Minerals

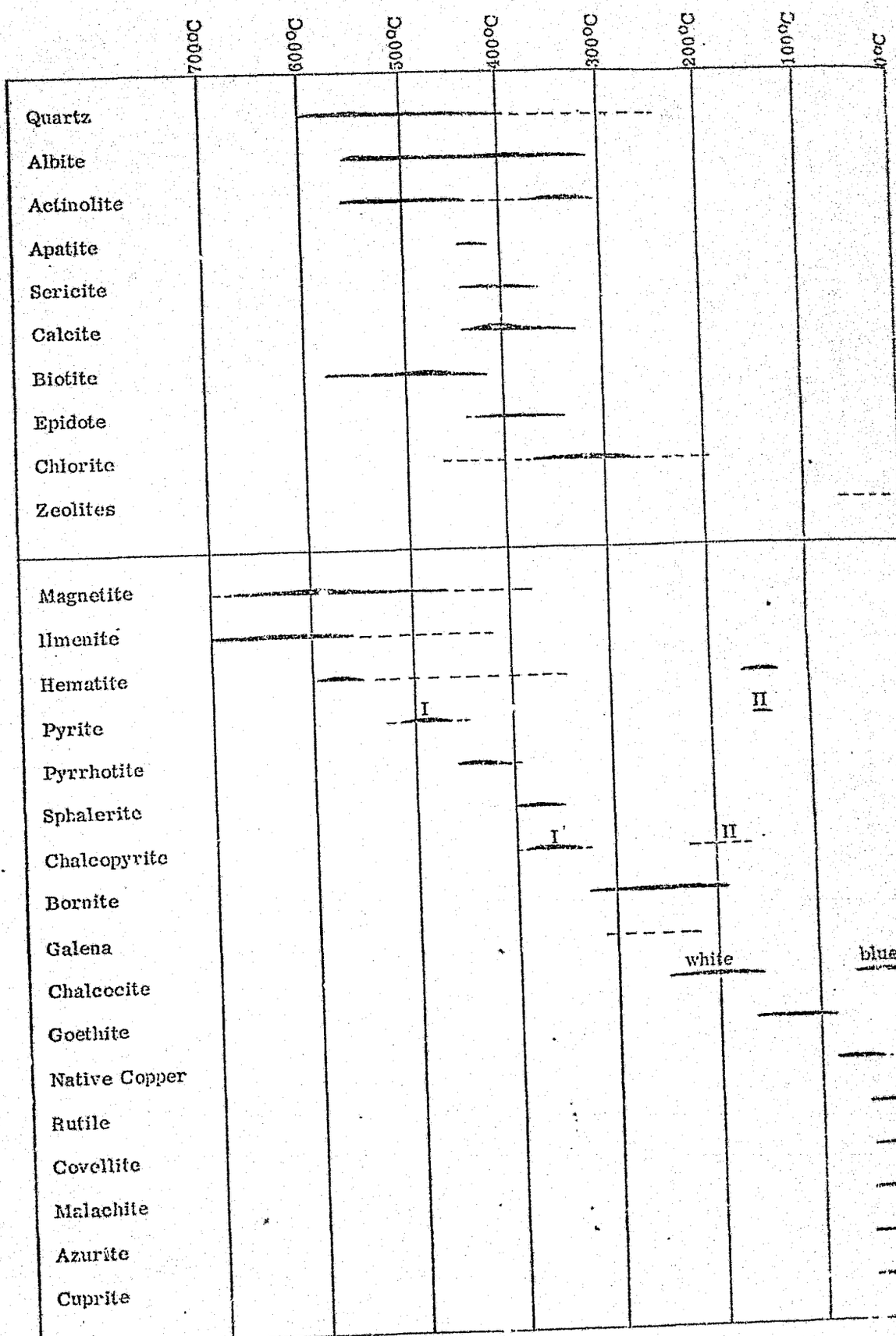
The sequence of deposition of the ore minerals and the hydrothermal gangue minerals is graphically represented on the paragenetic chart (Fig. 8). The minerals are considered in their order of crystallisation as determined by their textural relationships with other minerals.

Magnetite and ilmenite are regarded by the writer as being primary minerals formed by igneous processes. Ramdohr (1955) and Buddington and Lindsay (1964) give the temperature of formation of the exsolution texture between ilmenite and magnetite, which is commonly observed at Roodekraal, as between 400°C - 700°C.

Hematite occurs as an early formed mineral within the pegmatite (Plate XV, 4), where it may have formed either as an alteration product of hornblende or as a primary mineral. A second generation of hematite may be formed at lower temperatures above 135°C as an intermediate step in the conversion of magnetite to goethite, although magnetite may be altered directly to goethite. Hematite appears to have formed before secondary quartz since 'boxworks' of hematite are filled with quartz (Plate XVI, 1).

The position of quartz in the paragenetic sequence is difficult to assess since it was probably formed both as a deuteric-stage mineral and during hydrothermal alteration accompanying the deposition of the sulphide minerals. The quartz formed after hematite, but it must have formed before pyrite I, since pyrite I encircles the quartz in 'atoll' like structures (Plate VI, 1).

Fig. 8 Paragenetic chart of gangue and ore minerals present within the sulphide bodies at Roodekraal.



Albite, the majority of which is considered to have formed as a high-temperature, deuteric, modification of the pre-existing feldspar, is present in great abundance especially in the lavas. However the original feldspar has been more completely altered to albite in the vicinity of the mineralized zones than elsewhere in the rocks of the Complex. Thus a certain amount must have been formed by the late-stage hydrothermal solutions associated with the period of sulphide mineralization.

Actinolite is present in fair proportion as a secondary mineral in most of the rocks of the Complex and occurs as fine acicular or columnar crystals within the groundmass and amygdales. It is observed to have formed earlier than the chalcopyrite I which surrounds corroded crystals of it (Plate IX, 4), and earlier than bornite and chalcocite, since the actinolite needles act as partitions between the bornite and white chalcocite (Plate XI, 4). As the actinolite is evenly distributed throughout the lavas and other main rock types of the Complex, it probably owes its origin to deuteric alteration. It is unlikely that actinolite was produced by a later metamorphism since the syenite and syenite-porphyry dike rocks do not contain actinolite.

The bulk of the sericite is thought to have formed at temperatures above 350°C during the deuteric-stage alteration. However, some sericite could have been formed at a later stage when potash was released during the conversion of biotite to chlorite.

Calcite apparently formed before the sulphide minerals. It was formed before chalcopyrite I and sphalerite, since it has been surrounded and corroded by chalcopyrite I (Plate IX, 1), and the chalcopyrite I in the same polished section occurs in exsolution with sphalerite. The sphalerite has an iron content of 9,4 per cent indicating a temperature of formation of 350°C which is in accordance with the findings of Buerger (1934) who stated that the exsolution of chalcopyrite within sphalerite took place between 350°C and 400°C. From this it can be concluded that the calcite must have formed at temperatures in excess of 400°C. It is probable that the majority of the calcite was formed early, during the deuteric-stage of alteration.

Biotite is considered by the writer to have formed by hydrothermal potassium metasomatism of pre-existing mafic minerals. It is also concluded that the biotite was formed by late-stage hydrothermal solutions associated with sulphide deposition.

Pyrite I was deposited over two temperature ranges, probably within

the limits suggested by Smith (1948) of 630°C - 400°C at high pressure and 150°C at low pressure.

Pyrrhotite was the next mineral to be deposited and it has replaced both pyrite and quartz. The pyrrhotite has an iron content of 46,9 per cent atomic Fe, giving a temperature of formation of 430°C at pressures greater than one bar. Pyrrhotite replaced quartz but its relationship to sericite is uncertain.

Epidote occurs in association with pyrrhotite in Plate VIII, 4, where the epidote appears to have replaced the pyrrhotite in a 'caries texture', but has itself been replaced by chalcopyrite I. Epidote formed as a late-stage hydrothermal mineral by the alteration of aluminous ferromagnesian minerals. It may also be formed in small quantities as an alteration product of the feldspar when the feldspar is albitized with the liberation of calcium. It has been noted under the section on the 'Mineralogy of the Ore Minerals' that chalcopyrite I has replaced epidote and that epidote in turn replaced pyrrhotite. Since the temperature of formation of the pyrrhotite has been determined as 430°C and that of chalcopyrite as about 350°C it follows that epidote was formed between these two temperatures. The above conclusion fits well with the range of the temperature of formation of epidote as quoted by Deer, et al. (1967) of 300°C - (400° - 500°C).

Chlorite occurs in most of the rocks of the Complex in small amounts. It is most abundant in the vicinity of the mineralized zones. One of the varieties of chlorite present is possibly bavalite (a variety of daphnite) which is generally formed between 200°C and 500°C. The chlorite probably formed at the expense of the olive-green to pale-brown biotite, with the K_2O released during the reaction being used in the formation of sericite.

Sphalerite is present in small quantities mainly as vein fillings. Since it occurs in exsolution with chalcopyrite I, it must have been deposited at the same time as the chalcopyrite. The sphalerite has an iron content of 9,4 per cent which indicates a temperature of formation of 350°C.

Chalcopyrite I was formed at about 350°C where it is in exsolution with sphalerite and deposition probably did not continue to much lower temperatures since chalcopyrite I is not found in exsolution with bornite.

Bornite and white chalcocite must have formed at temperatures below 350°C because they do not occur as intergrowths with chalcopyrite I. The 'myrmekitic' intergrowth of bornite and white chalcocite was, according

to Brett (1963), probably formed at between 200°C and 270°C.

Chalcopyrite II is present as a minor exsolved phase along the cleavage planes of bornite, and according to Brett (1963) it must have formed at temperatures below 228°C.

Goethite which is formed by the oxidation and hydration of magnetite is considered to be hypogene in origin due to the fact that it is concentrated in the vicinity of sulphide bodies and because the amount of goethite present does not decrease with distance from the surface. As mentioned under the section entitled ' Mineralogy of the Ore Minerals' the author considers that the goethite at Roodekraal formed under alkaline conditions. Edwards (1965) concludes that goethite forms at temperatures below 165°C under alkaline conditions.

Zeolites are low temperature minerals and probably formed after all the hypogene sulphides had been deposited. The variety of zeolite present was determined by X-ray analysis to probably be a member of the phillipsite group which Daubrée (1879) quoted in Deer et al. (1967, 4, p. 399) noted as forming at temperatures below 70°C.

Native copper replaces epidote but is rimmed by later blue chalcocite. The native copper appears to be supergene in origin since it is only present in porous rocks near the surface, where it could easily have been deposited by meteoric solutions. Rutile formed as a supergene alteration product of ilmenite. Blue chalcocite is present as a supergene alteration in a 'rim texture' around bornite and native copper.

Covellite was the last supergene sulphide to form. It is present as an alteration of chalcopyrite II, bornite and white chalcocite. Since it is of the 'blaubleibender' variety it must have formed at temperatures below 157 ± 3°C (Ramdohr, 1955). Malachite, azurite and cuprite are formed at or near the surface by the processes of weathering. These minerals form at the expense of any of the pre-existing copper minerals.

An inspection of the paragenetic chart (Fig. 8), shows that the ore minerals follow a perfectly logical sequence with the high temperature minerals having been deposited first and the low temperature minerals last. This decrease in the temperature of formation is accompanied by a parallel decrease in the sulphur content. This would seem to indicate that there was only one period of mineralization, since if there was more than one period of mineralization we might expect local reversals of the sequence with a high temperature mineral replacing a low temperature one.

The gangue minerals do not follow a simple sequence since they were not all formed at the same time. Two distinct sequences are present:-

(i) Secondary rock forming minerals formed during the deuteric-stage alterations. These include quartz, albite, actinclite, sericite, apatite and calcite.

(ii) Secondary rock forming minerals formed by the hydrothermal solutions which accompanied the sulphide mineralization. Minerals included here are biotite, epidote, chlorite and zeolite.

2. Hydrothermal Zoning

Zoning is a subject which due to its variable nature is not adequately explained by any single hypothesis. Two types of zoning are present at Roodekraal, a depth zoning, and a concentric zoning within the replacement bodies themselves.

a. Depth zoning

This feature which is used to classify hydrothermal ore deposits has been discussed by many writers, notably Lindgren (1933), Graton (1933) and Schneiderhöhn (1941). The essence of the classification schemes put forward is that ore deposits formed at different depths have distinctly different suites of ore minerals and conditions of formation. Unfortunately it is very difficult to apply these schemes in practice since few deposits comply with all the requirements of any particular scheme.

The widely used classification of Lindgren (1933) is taken as an example. In his sub-division of deposits formed by chemical processes of concentration from hot ascending water charged with igneous emanations, he makes three sub-divisions, namely:-

(i) Deposition and concentration at slight depth. Epithermal deposits (approximately 50 - 200°C, moderate pressure).

(ii) Deposition and concentration at intermediate depth. Mesothermal deposits (approximately 200° - 300°C, pressure high).

(iii) Deposition and concentration at great depth or at high temperature and pressure. Hypothermal deposits (300°C - 500°C, pressure very high).

The sulphide deposits present at Roodekraal would not satisfy the conditions for inclusion in any one of these three sub-divisions, since although it would be epithermal by its depth, the temperature range of the sulphide mineral deposition from above 430°C to about 100°C stretches

over all three sub-divisions from epithermal to hypothermal.

Although the depth of emplacement was very shallow, the writer does not favour the use of the term 'telethermal' since this indicates that H_2O, CO_2, NaCl and other compounds escaped into the hydrosphere and atmosphere and that the deposits were formed at low pressure by open space filling (Brown, 1950). The writer feels that the main sulphide deposits at Roodekraal were trapped beneath a cap rock of pyroxene andesite, and regards the extensive disseminated replacement bodies as being indicative of formation under at least moderate pressure. Many of the requirements for inclusion as a mesothermal deposit are met since many of the sulphide minerals were deposited between 200°C - 300°C and the sulphide bodies have a tenuous connection with the surface.

The writer is of the opinion, however, that the sulphide deposits at Roodekraal must be regarded as epithermal since they are emplaced in lavas which were probably only covered by subsequent lava flows of the Roodekraal Igneous Complex, if it is assumed that the period of mineralization was pre-Karroo. The depth of burial must be the prime criterion of a 'depth' zone classification, even though the word thermal with several prefixes is used to describe the various sub-divisions.

b. Concentric zoning

This is the type of zoning which is present within a particular body and has received less attention than the depth zoning. Authors who worked on the problem of zoning in hydrothermal deposits such as Emmons (1924), Blanchard (1927), Graton and Bowditch (1936), Parks (1955), Sales and Meyer (1950), Schwartz (1956) and Parks and McDiarmid (1964) concluded that with distance from the source of the mineralizing fluids successively lower temperature minerals are deposited. However this generalization applies more to an open space, fissure filling type of deposit than to a disseminated replacement deposit. So many reversals of this zonal sequence take place that it cannot be regarded as more than an idealised arrangement.

At Roodekraal, there is a distinct zoning in the sulphide bodies which occur around the high angle normal faults. This zoning is a complete reversal of the normally accepted sequence, having an inner zone of bornite-chalcocite followed outwards by a bornite-chalcopyrite zone and finally a chalcopyrite-pyrite-pyrrhotite zone (see Plates XVIII and XIX). This relationship can also be seen in Plates XXI and XXII where although no faults occur, the best copper values are always within the bornite zones becoming poorer in the chalcopyrite-pyrite-

pyrrhotite zone. The white chalcocite which is present in greatest abundance in the centre of the sulphide bodies must have been formed at fairly low temperatures since it invariably contains silver. Native copper is not involved in the zoning since it is supergene in origin and only occurs near the surface.

An exception to this general zonal sequence is found in Plate XX, where a large irregularly shaped pyrrhotite body, elongated in the horizontal plane, is surrounded by a chalcopyrite zone which gives way to a bornite-chalcocite zone above, and a bornite-chalcopyrite zone below. The northern limb of the pyrrhotite body occupies an irregularly shaped volcanic breccia which acted as a favourable passageway for the mineralizing fluids. This breccia zone is possibly connected to the similar breccia which outcrops to the northeast and which would have allowed the gases to escape to the surface bringing about a great reduction in pressure and causing the classical low pressure, fissure type of zoning to develop with the high temperature sulphides on the inside. It thus appears that there are two distinctly different zonal sequences:-

(i) The normal zonal sequence which occurs in low pressure, open space filling deposits.

(ii) The reversed zonal sequence which occurs in moderate to high pressure replacement deposits.

Jacobsen (1967) has proposed a model to explain the occurrence of reversed zoning in replacement bodies formed at moderate to high pressures. He considers that the copper ore bodies of Artonvilla, which exhibit reversed zoning, were formed in a 'nearly or totally closed system in the physical sense'. He postulates that only one influx of mineralizing solutions entered a structural trap which was further sealed by the formation of a silicified and slightly sericitized pyrite zone. He then reasons that the early precipitation of pyrite resulted in a depletion of sulphur in the rest solutions so that minerals containing progressively less sulphur were deposited on the outside of the 'bubble' of mineralizing fluids as it cooled inwards. The gangue minerals were progressively enriched in water towards the centre of the body. Jacobsen considers that the mineralizing fluids were transported as pneumatolytic solutions above the critical temperature of water, which however condensed as temperatures and pressures became lower.

The writer feels that the concentric zoning within the main sulphide bodies at Roodekraal is adequately explained by a direct application of Jacobsen's model. As far as the exact mode of deposition of the ore minerals is concerned the writer would like to refer to the work of Brown (1971) who studied reversed zoning in the White Pine Copper district, where the copper deposits were formed by replacement under pressures of 240 to 650 bars. Brown suggests that dissolved copper diffused through a stationary pore solution due to a chemical potential gradient generated by the estimated differences in the copper contents of the mineralizing solutions and the interstitial solutions in the unmineralized rock.

3. Classification of the Ores

This has to a large extent been discussed in the section entitled 'Depth Zoning' where it was concluded that due to their shallow depth of formation, the sulphide bodies at Roodekraal are epithermal in origin. These sulphide deposits exhibit the following mesothermal characteristics:-

(i) A temperature of formation in the mesothermal range of Lindgren (1933).

(ii) A tenuous connection with the surface (Parks and McDiarmid, 1964).

(iii) Extensive replacement of pre-existing minerals as well as open space filling.

The wide range of temperature over which the sulphides formed would imply that the mineralized body is a 'telescoped deposit' (Spurr, 1922), but the logical and well developed zoning present appears to refute this.

If a simple genetic classification such as the one suggested by Bateman (1964) is used to classify the Roodekraal deposits, they would be included in deposits formed by hydrothermal process and would belong to both of the following two divisions:-

A. Cavity filling, sub-section f. Vesicular fillings

B. Replacement, sub-section c. Disseminated

There is a striking similarity between the sulphide replacement bodies of Roodekraal and Artonvilla. The differences which do occur, mainly in the type of gangue minerals present, are caused by the completely different types of host rock. A comparison between these two

deposits is presented in Table XXIX.

TABLE XXIX

A comparison between the sulphide replacement bodies of Roodekraal and Artonvilla

Property	Artonvilla	Roodekraal
Location	In structural traps generally at the intersection of two cross folds.	In structural traps within the amygdaloidal andesite between the pyroxene andesite above and the equigranular diorite sill below.
Immediate Host rock	Constantly in the 'Melanocratic' zone which has a favourable chemical reactivity.	Always within the amygdaloidal andesite which is chemically favourable to replacement.
Zoning	Reversed zoning with high Fe-S minerals at outer extremities and high Cu, H_2O concentrations at the centre of the ore body.	As for Artonvilla.
Formed by	Ascending hydrothermal fluids of magmatic origin.	As for Artonvilla.
Temperature range of sulphide deposition	From temperatures of above 417°C to below 100°C.	From temperatures of above 430°C to below 100°C.
Type of Deposit	Disseminated replacement type with intense hydrothermal alteration and introduction of H_2O, K_2O, Na_2O, Al_2O_3, CaO, MgO and minor SiO_2.	Disseminated replacement and open space filling type with intense hydrothermal alteration and probable introduction of H_2O and K_2O.
Mode of Formation	Ore solutions entered traps in one distinct upsurge and then crystallised from the margins inwards.	As for Artonvilla
Type of Mineralization	Eleven of the ore minerals present at Artonvilla are also present at Roodekraal. Sulphides which are common to both deposits include pyrite, pyrrhotite, sphalerite, chalcopyrite, bornite, chalcocite and covellite. Eight of the silicate gangue minerals, which were deposited by the same hydrothermal solutions as the ore minerals, are common to both deposits. These minerals are albite, quartz, apatite, sericite, calcite, biotite, chlorite and epidote.	

From the comparison presented in Table XXIX it can be concluded that the sulphide bodies at Roodekraal are best described as hydrothermal disseminated replacement bodies of the 'Artonvilla type'.

4. Nature of Mineralizing Solutions

a. Physical state of the mineralizing fluids

As regards the actual physical state in which ore bearing fluids leave their source, earlier authors such as Fenner (1933), Graton (1940) and others concluded that ideal gases were incapable of transporting metals. In support of this contention Smith (1954) suggests that hydrothermal deposition is confined essentially to the liquid phases.

More recently, however, Ovchinnikov (1963) conducted experiments which indicated that metals could be moved from an artificial magma by gases, mainly composed of water vapour. Krauskopf (1967) also concludes that hydrothermal fluids may be partly or completely in the gas phase.

The hydrothermal solutions at Roodekraal must have been capable of penetrating dense wall rocks, judging by the extensive wall rock alteration.

b. Transport of metal in solution

Due to the fact that sulphide compounds are extremely insoluble, it seems probable that some form of chemical bonding takes place which could increase the solubility of these compounds by several orders of magnitude. Transport of metals in chloride complexes was postulated by Helgeson (1964). This hypothesis is substantiated by high concentrations of Cl^- in most hydrothermal environments.

Spurr (1922) put forward the idea that the mineralizing fluids are composed essentially of sulphides in an 'ore magma'. A more sophisticated theory along similar lines is that of Brown (1950) who feels that the mineralizing fluids were in the form of a vapourized (volatilized) sulphide melt. He suggests that a mixed sulphide melt accumulated deep within the earth, which was stratified in accordance with metallurgical principles. Vapourization of the substances having access to the avenues of escape then proceeded in accordance with their relative volatilities at the particular temperature of the melt, the probable order of vapourization being:-

pyrite → chalcopyrite, sphalerite → enargite, tennantite, tetrahedrite, jamesonite → chalcocite, galena, argentite.

He concludes that although water is of little value as a liquid or vapour in the transport of metal sulphides it could be of great utility in the elimination of carbonaceous or siliceous gangue to make way for the deposition of sulphides. In support of his idea he quotes the