GEOLOGY

OF THE SOUTH-EASTERN

RICHTERSVELD

BY

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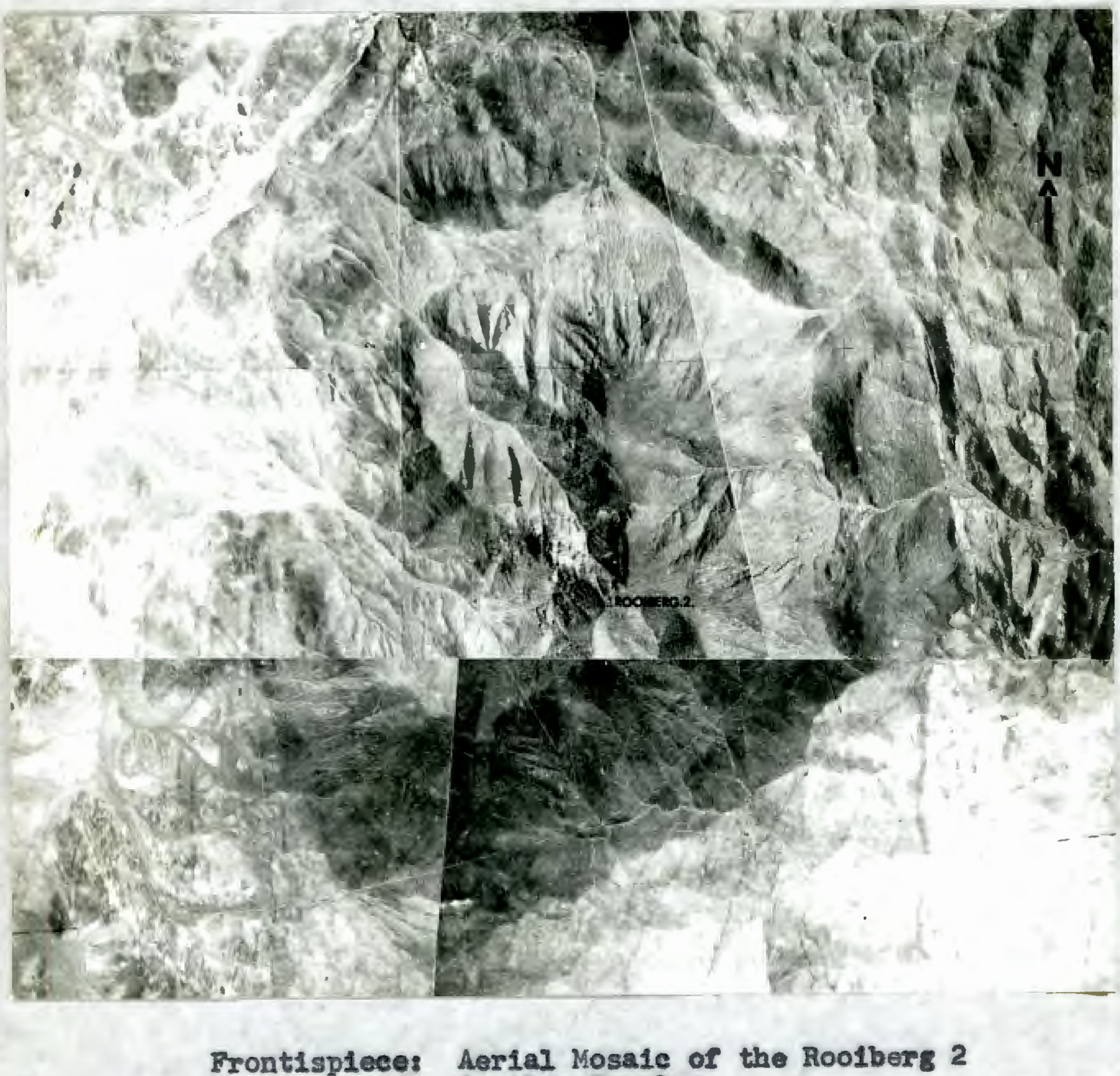

Frontispiece: Aerial Mosaic of the Rooiberg 2 Annular Complex

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SUMMARY

In their recent study of the geology of the Richtersveld, de Villiers and Söhnge (1959) have shown that the south-eastern part of this area is geologically complex, and contains numerous problems that can only be solved by detailed study. The present investigation is an attempt to find solutions to as many of these geological problems as possible, and thus increase our knowledge of the geology of the Richtersveld - an area that forms a link between the geology of the western Cape and South West Africa. During the present investigation a contoured geological map of the south-eastern Richtersveld on a scale of 1:25,000, was constructed. Four-hundred-and-fifty petrographic modal analyses were made; and twenty-five chemical analyses were used, fourteen were new, and the remaining eleven were taken from the Richtersveld Memoir.

It was found that the evolution of the rocks of the south-eastern Richtersveld could be divided into two major episodes. During the first episode the following events took place; (1) The Kheis supacrustal rocks formed; (2) they were intruded by small ultramafic bodies; and (3) all the rocks then found in the area were subjected to the action of the granitization process that resulted in the formation of the Adamellitic Gneiss. During the second major episode the rocks of the Richtersveld Suite, the Stinkfontein Formation and the Nama System are all believed to have formed.

The petrography, structure and geochemistry of the rocks of the south-eastern Richtersveld is described,

their petrogenesis is discussed, and the following conclusions are drawn:

(1) Prior to being metamorphosed the Kheis supracrustal rocks consisted mainly of impure siltstones, impure sandstones, and lavas (mainly acid).

(2) These supracrustal rocks are lithologically more akin to the rocks of the Kaaien Series than those of the Wilgenhoutdrift Series with which they were previously correlated.

(3) The metamorphism of these supracrustal rocks attained the almandine-amphibolite grade, and some metamorphic differentiation took place.

(4) The schistosity of the Kheis supracrustal rocks is believed to have resulted from the stress generated and the shearing produced by increases in volume consequent upon the genesis of the Adamellitic Gneiss.

(5) The metamorphism of the Kheis supracrustal rocks, the formation of the hybrid rocks, and the evolution of the Adamellitic Gneiss are all believed to be part of a single general granitization process. The Kheis supracrustal rocks of the outer zone were subjected to increased temperature-pressure conditions, the hybrid rocks were permeated by granitic solutions, and the Adamellitic Gneiss developed in the central theatre of granitization by the "blowing out" of the schistose texture of the Kheis supracrustal rocks into the gneissose.

(6) The local mobilization of the central gneissic core has in some areas produced significant variations in the usual sequence of metasomatic and metamorphic changes found as one moves out from the centre of granitization.

(7) Twenty-three post-Kheis Hornblendite and Serpentinite bodies, most of them previously unknown, were found

to be emplaced along two east-west trending lines of fracture.

(8) The Hornblendites and Serpentinites were emplaced as gabbroic and peridotitic magmas respectively; and they were later transformed to their present mineral composition by fluids produced during the formation of the Adamellitic Gneiss.

(9) Many of the ultramafic bodies contain previously unrecorded lens- and dyke-like bodies of graphic granite.

(10) Outcrops of plutonic rocks of the Richtersveld Suite form a north, north-west trending belt that extends for 120 miles from Soeties (to the south of the area studied) to Aurus Waterhole (in the eastern Sperrgebiet, S.W. Africa).

(11) Most of the plutonic rocks of the Richtersveld Suite were emplaced into the Epizone as ring-dykes and stocks; and some of these ring-dyke complexes have coalesced to form complex outcrop patterns as in the case of the Xaminxaip batholith.

(12) Petrographic and chemical data necessitated a revision of de Villiers and Söhnge's correlation of the rock units of the Rooiberg 2 outcrop with those of the Xaminxaip batholith.

(13) The parental magma of the Richtersveld Suite is believed to have developed from the selective fusion of the Adamellitic Gneiss.

(14) The syenites resulted from the assimilation of Fe, Mn, Ti and Mg enriched roof and wall rocks, and the escape of silica-rich volatiles.

(15) It is considered significant that the syenites are confined to the southern part of the Xaminxaip batholith as (a) the rocks exposed in this area are believed to represent material emplaced near the roof of the batholith, and (b) ultramafic and mafic country rocks are more common in the area adjoining the southern part of the batholith.

(16) The K/Rb ratios of the syenites ($\bar{x} = 454$), and the Quartz Bostonites (679), are abnormally high; but as Turekian and Wedepohl's (1961) average syenite has a similar abnormal K/Rb ratio, it appears that high ratios are a feature of syenites in general, and this lends support to the hypothesis that syenites are not "normal igneous rocks".

(17) Geochemical, mineralogical and structural evidence all indicates that the Porphyritic Microgranite was emplaced late as a residual magma. Subsequent potash metasomatism accounts for the abnormally high K/Na ratio and the presence of muscovite-rich veins.

(18) The Quartz Bostonites were found to be more salic than previously recorded; and geochemically they are closely akin to the Richtersveld Suite syenites.

(19) The Quartz Bostonites and the Hornblende Diorites are considered to represent the final stage in the igneous cycle of the Richtersveld Suite.

(20) Both of the above dyke rocks are regarded as products of desilicated and contaminated Richtersveld Suite parental magma.

(21) The rocks of the Stinkfontein Formation, the Richtersveld Suite, and the Nama System probably all belong to a single tectogenetic cycle.

(22) The Kuibis Sandstones of the area are mainly felspathic sandstones, and are believed to have been (a) produced by vigorous erosion in a source area in which the Adamellitic Gneiss was the main rock type, (b) transported by means of medium strength traction currents, and (c) deposited in a slightly oxidizing shallow water environment.

(23) The majority of the faults found in the area have a significant strike-slip component, and they are believed to have been active during post-Dwyka times.

CONTENTS

LIST OF ILLUSTRATIONS

Plates | Page

Photomicrographs

LIST OF TABLES.

I. INTRODUCTION

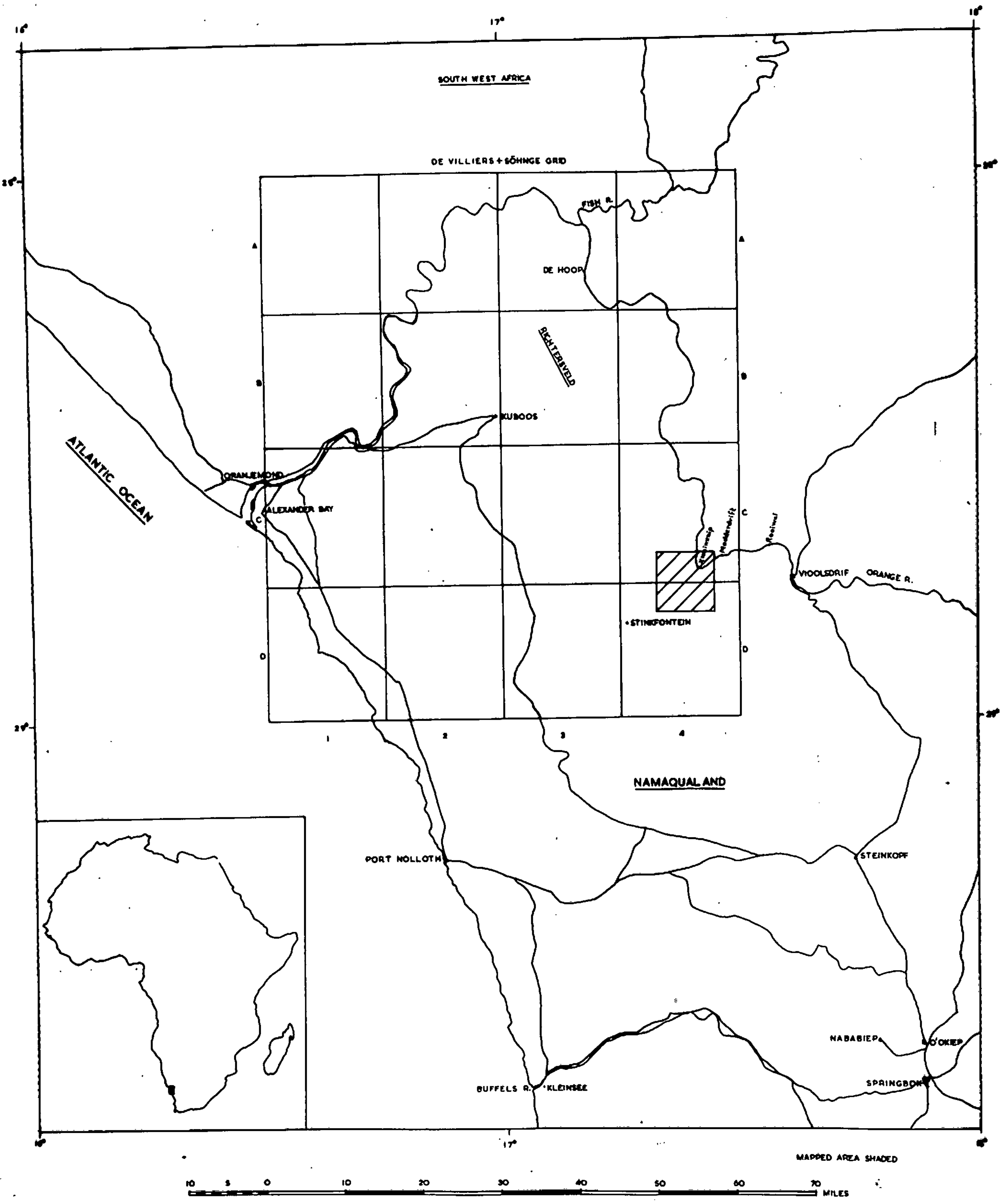
A. Scope of Research:

The aim of the present investigation was to make a detailed geological study of the South-Eastern Richtersveld. The need for such a study became evident with the publication by de Villiers and Söhnge (1959) of their "The Geology of the Richtersveld"; because this memoir not only proved to be a valuable contribution to our knowledge of the complicated geology of north-western Namaqualand, but it also raised many questions that could only be answered by further detailed mapping and laboratory investigation; and the answers to some of the more important of these unresolved questions are to be found in the South-Eastern Richtersveld. The more significant of these problems are: (1) to determine the composition, texture and metamorphic grade, and thence, if possible, the stratigraphic position within the Kheis System, of the Kheis supracrustal rocks of the area; (2) to discover more about the origin and the sequence of compositional and textural changes found to occur in the Adamellitic Gneiss and the transitional acid hybrid rocks; (3) to map all the ultramafic bodies within the area, as many of these bodies were left unmapped by de Villiers and Söhnge, and to determine their composition and origin; (4) to make a detailed study of the different plutonic members of the Richtersveld Suite; (5) to study the petrography and genesis of the two major dyke swarms found within the area and to attempt to discover if they are related to the plutonic rocks of the Richtersveld Suite; and if they are, to see how this fact affects the relative dating of the rock units cut by these dyke swarms; (6) to study the fresh felspathic sediments of the Kuibis Series of the Nama System which overlie all the abovementioned rock units, and which probably carry minerals or other evidence that can be used to determine which of the earlier rock units was the main source rock for these sediments; and (7) to attempt to reconstruct the tectonic environment in which the Richtersveld Suite was emplaced and the Stinkfontein Formation and Nama System were deposited and folded.

As was anticipated the above problems were not found to be separate and unrelated, but were all linked together as the younger rock units tended to have genetic links with the older rock units; that is, (1) the hybrid rocks and the Adamellitic Gneiss appear to have resulted from the granitization and ultra-metamorphism of the Kheis supracrustal rocks; (2) the bulk of the magma that went to form the Richtersveld Suite appears to have been derived from the melting and mobilization of the Adamellitic Gneiss; (3) the genesis of the syenites appears to have resulted mainly from the contamination of the above palingenetic magma by more basic Kheis supracrustal material and perhaps even by the ultramafic bodies; (4) the rocks of the Stinkfontein Formation, Richtersveld Suite, and Nama System all appear to belong to the same tectogenetic cycle; and (5) the Adamellitic Gneiss can be shown to have been the main source of material for the Kuibis sandstones of the area studied.

B. Location and Extent of Area

The area studied in detail during the present investigation is situated in the south-eastern Richtersveld and includes parts of the districts of Namaqualand (South Africa) and Warmbad (South West Africa). It is approximately 57 square miles (148 sq.Km) in extent and is bounded by the meridians 17° 19' 30" E and 17° 27' 00" E and latitudes 28° 42' 00" S and 28° 48' 30" S. The village of Stinkfontein or as it is now sometimes called, Ecksteensfontein, is approximately 5 miles to the south-west of the area.



LOCATION

<u>Figure 1</u> : Location of the Area.

C. <u>Communications</u>

The Richtersveld is one of the most isolated areas in South Africa, and communications are poor. Access to the area studied is easiest by way of the Springbok-Karasburg (S.W.A.) road that crosses the Orange River at Vioolsdrif

18 miles up river from the eastern boundary of the area. A few tracks that are used mainly for donkey-drawn vehicles, are found along the southern perimeter of the area. These tracks consist of (1) the Klein Helskloof pass section of the Vioolsdrif-Stinkfontein track that is located in the extreme south-east of the area, and (2) the ill-defined tracks found in the south-western corner of the area which lead from Stinkfontein to the western side of Rooiberg 2 (3F2).

(Note: 3F2 refers to the location of Rooiberg 2; 3F refers to one of the large squares on the geological map of the area, and the 2 refers to the north central part of the square, as all the large squares of the map have been divided into nine smaller squares numbered from left top corner which is one to the right bottom corner which is nine.).

The absence of adequate roads and public transport within the Richtersveld presented the writer with formidable transport difficulties, as he had no private means of motor transport, and once away from the Orange River water was difficult to obtain.

D. Topography and Drainage

The Orange (or Gariep or Great) River is located in the north eastern corner of the area, and it forms a great loop flowing first west south-west and then turning north at Xaminxaip or Kaminip (6A8)- i.e. the alternative spellings used by de Villiers and Söhnge (1959) and Wellington (1958). The Orange is the major, and only perennial stream found within the area, and all the other streams of the area drain into it. The area is part of a zone of extremely rugged country bordering the Orange River, with the Stinkfonteinberge to its west and the T'Nein - Nababiepsberge to its east. The highest point in the area is the summit of Rooiberg 2 (3F2) of 3066 feet, and Black Face Mountain to the immediate north-west of the area rises to 3,526 feet; thus as the Orange River flows at a level of under 500 feet, the relative relief of the area tends to be great. The ruggedness of the terrain is particularly striking in the zone immediately bordering the Orange River because here the river flows through a gorge with the land rising to above 2000 feet within a short distance of the river.

E. Climate, Vegetation and Human Activities

(i) Climate: Even though the area does not contain a weather station, a fairly accurate assessment of the climate can be obtained by considering data obtained from weather stations within the same "rainfall district" (Weather Bureau 1960). Two stations Steinkopf (to the South) and Goodhouse (to the east), fall within the same rainfall district and are also relatively near the area.

Annual Rainfall.

Station	Ht. in m.	Av. in mm.	(Av. in inches)	Max. m m	Min. m m
Steinkopf	884	132.1	(5.2)	252.2	32.0
Goodhouse	203	54.6	(2.15)	128.5)	15.0

Temperature*

Goodhouse	Year	J	F	M	A	M	J	J	A	S	O	N	D
Air Temp°C (max + min)/2	23.3	30.5	29.9	28.3	24.4	19.1	15.3	14.5	17.2	20.5	24.2	26.4	29.3
No. of days with Max. temp. > 30°C	211.4	30.3	27.8	28.5	20.5	7.1	0.5	0.0	5.3	12.6	22.4	26.5	29.9
No. of days with Min. temp. < 0°C	0.3	0	0	0	0	0	0.1	0.1	0.1	0	0	0	0

* (Air temperature is not measured at Steinkopf).

From the above data it is clear that hot desert conditions prevail. The annual precipitation is small and tends to fluctuate, or to put it in more formal terms the mean deviation as a percentage of the average annual rainfall is approximately 55% (Weather Bureau 1957). The temperature is remarkably high throughout the year as shown by the fact that Goodhouse has 211.4 days annually when the air temperature exceeds 30°C (86°F). From the "Climate of South Africa" part five it is clear that the rainfall district into which the area falls is one with a winter maximum, and that the mean monthly district rainfall as a percentage of its annual mean is as follows:

J	F	M	A	M	J	J	A	S	O	N	D
2.2	4.7	6.5	8.8	13.5	15.6	13.8	13.9	7.8	5.7	4.7	2.8.

The area thus has a BWhs climate according to Köppen (1900) or EB'd climate according to Thornthwaite (Schulze, 1947).

(ii) Vegetation: The vegetation of the area falls into what Acocks (Talbot and Talbot, 1960) calls "Arid Broken Veld", or more specifically (Acocks 1953) "Namaqualand Broken Veld" - a veld type characterized by Aloe dichotoma (Kokerboom). A list containing most of the species of tree, shrub and smaller plant found in this floral district is given on page 110 of Acock's (1953) "Veld types of South Africa". In contrast to the sparse drought-resistant vegetal cover found over most of the area, a ribbon of dense bush is usually found along both banks of the Orange River. Acacias, the Tsabbie or bastard ebony and the willow are the dominant species found along the river. A particularly interesting plant found in the area is the "half-mens" or Pachypodum, and many fine specimens of this plant are to be seen in Donkerkloof (8D1) and on the southern slopes of Rooiberg 2 (3G1). Rooiberg 2 is of considerable microclimatological and botanical interest as its northern slopes are dry, with a very sparse vegetal cover while the southern slopes have a relatively dense vegetal cover with the landscape being dominated by huge Aloe Pillansii. The outcrops of ultramafic rock, particularly the serpentinites, are also of botanical interest in that they produce less soil and support less vegetation than the other rock types surrounding them.

(iii) Human Activities: The eastern half of the area falls within the Vioolsdrif irrigation settlement, while the western half lies within the Richtersveld Coloured Reserve. At the present time the only human activity carried on in either region is goat herding. The goat herding way of life is not very profitable and the "Report of the interdepartmental committee on matters affecting Coloured persons on Coloured Mission Stations, Reserves and Settlements" (1947 p.54) states that the "inhabitants hardly make a living", while their goats do untold damage to the vegetation by selectively eliminating many productive plant species. With the removal of the more palatable species soil erosion tends to increase dramatically, and it is easy to see why the goat is called the "black locust" in the Middle East.

This section on human activities would not be complete without some mention being made of the (Bushman?) petroglyphs found carved on the Schwarzkalk limestone in the Modderdrif area one mile up stream from the area and in the TeCowiep Valley. As these sites, unlike many in Southern Africa, are still in a good state of preservation, it was decided to include plate 1 which gives some idea of the style of the Modderdrif petroglyphs. Most of the petroglyphs take the form of geometrical patterns but others represent animals no longer found in the area, such as giraffe, rhinoceros, ostrich and large buck. It is hoped that steps will be taken to preserve these petroglyphs if, as seems probable at the present time, the area containing them is flooded by one of the proposed Orange River dams.

Plate 1: Petroglyphs from Modderdrif.

F. Previous Literature

"The Geology of part of Namaqualand" by A.W. Rogers (1915) was the first significant paper on the geology of this region. Prior to that cursory observations had been recorded by Capt. J.E. Alexander (1838A and 1838B), W.G. Atherstone (1855), A.W. Wyley (1857) and E.J. Dunn (1872). In 1936 Haughton and Frommurze published their account of the geology of the Warmbad district, South West Africa which includes the north-eastern corner of the area. During the following year (1937) the account by Gevers and others on the pegmatite area south of the Orange River in Namaqualand appeared in print. This dealt with the region to the east of the area. Mabbutt (1950, 1955A and 1955B) published three papers on the geomorphology of Little Namaqualand. Most of his field work was confined to the region to the immediate east of the area. In 1958 Wellington published his magnificent account of the growth and development of the Orange River drainage system. His account of the evolution of the Orange River in the Richtersveld is of particular interest in that critical areas are illustrated by superb oblique aerial photographs. De Villiers and Söhnge's (1959) geology of the Richtersveld is clearly the most substantial contribution to our knowledge of Richtersveld geology published to date. This memoir and the two earlier papers (1946 and 1948) published by the two authors jointly and one paper by de Villiers (1945) alone, represent the published results of a comprehensive field survey undertaken by the two authors during the winter field seasons of 1943, 1944 and 1945.

G. Present Investigation

(i) Field Work: The field work for the present survey was carried out during April, May, July and September, 1960 and May and July, 1961. During July 1960 the writer travelled extensively throughout the Richtersveld paying particular attention to the dyke rocks, and the outcrop of "Richtersveld Granite" north of de Hoop, but during other field trips attention was primarily focused on the south-eastern Richtersveld. As de Villiers and Söhnge (1959 p.11) have noted, no large-scale maps of the Richtersveld exist. Fortunately aerial photographs of the area were available, and the Trigonometrical Survey, Mowbray, Cape was able to provide a principal point lay-down for the north-eastern part of the area, and by fixing additional ground control points it was possible to construct a 1:25,000 principal point lay-down for the whole area. Spot heights were obtained by using an aneroid barometer and correcting the heights so obtained on a diurnal pressure curve constructed for the area. Sample locations and field data were plotted directly onto enlarged vertical aerial photographs of 1:10,000 scale.

(ii) Laboratory Work: The laboratory work can be considered under three headings, (a) petrography and mineralogy, (b) chemistry and (c) map construction.

(a) Petrography and Mineralogy: As the present project was primarily intended to supplement the data given in the Richtersveld Memoir, the study of the petrography of the rocks of the area received high priority. In all some 450 thin-sections were studied. Fortunately most of the rocks of the area are medium grained and it was thus possible to determine their modes. None of the rocks were found to be too coarse for standard modal analytical techniques, and only a few specimens, particularly the finer of the Quartz Bostonite dykes, were found to be too fine grained for quantitative study (Chayes, 1956B, pp. 100-102). The relative proportions of the different mineral species found in the rocks were determined by means of a mechanical stage equipped with a point-counter attachment. In order to eliminate any serial changes in modal composition being introduced by serial variations in operator technique, random number tables were used to randomize the order in which thin-sections were studied. One thousand grains were identified and counted in each thin-section studied. The spacing between consecutive counts was arranged so that the 1000 points counted covered the whole of the large thin-sections used (Chayes, 1955, p. 105). In studying the coarser grained granular syenites 2000 points (1000 on each of two slides) were counted as their grain size was such as to require the study of this larger area according to Chayes' tables (1956B, pp. 79-94).

In the petrographic study of the Nama sediments (Chapter 7) particular attention was paid to the sandstone horizons as it was believed that they would yield the most information for the time expended in their examination. The grain size distributions of these sandstones were determined by measuring the apparent thin-section long axis of 100 grains per slide. A mechanical stage equipped with a point counter attachment was used in the selection of grains to be measured and a micrometer ocular was used in the actual measurement. A hundred counts were considered adequate (Pye, 1943 and Rosenfell, Jacobsen and Fern, 1953) as this number seems to be favoured by present day workers in this field, for example Wiesnet (1961).

Roundness was estimated by studying 50 grains per slide and classifying them visually into four roundness grades - angular, subangular, subrounded and rounded. The degree of roundness was estimated from the visual roundness images provided by Williams and others (1954, p. 282). A very general estimate of sphericity was obtained by studying 50 grains per slide and classifying them visually into the different grades of sphericity using, the visual spericity images provided by Krumbein and Sloss (1951, p. 81). Rock colour was determined by using the "Rock-colour Chart" prepared under the chairmanship of E.N. Goddard (1948). Colour matching was performed in bright indirect sunlight, and final designation of colour was decided upon using dry specimens, but the specimens were moistened while being studied as this merely decreased the value (i.e. makes the specimen darker) but does not change the chroma; and it was found that in some cases the wetting of the specimens made the chroma and hue more readily determinable. The same rock-colour chart was used in determining the colour of the igneous and metamorphic rocks of the area, and it was also used in determining the colour of coloured minerals.

All refractive indices were determined by standard immersion method technique using sodium light and a Leitz-Jelley refractometer. Axial angles and extinction angles were measured on oriented sections using a four-circle universal stage. At least one specimen of each major rock group was crushed and a heavy residue separated. Particular attention was given to the size, roundness and sphericity of the zircons found in the heavy residues. All minerals present in the heavy residue, but not found in the thin-sections were recorded as traces (tr.) in the modes. In naming the various members of the alkali-felspar group, Tuttle (1952A) and MacKenzie and Smith's (1956) curves showing optic angle vs. composition were used. In obtaining the composition of the plagioclase crystals from their refractive indices Chayes (1952B) and Tsubois curves were used. When possible Wright's and Michel-Levy's curves were also used as a further check on the composition of the plagioclase. In naming the various members of the amphibole and epidote groups the nomenclature employed by Winchell and Winchell (4th Ed., 1951) was adopted.

(b) Chemistry: In all, 25 chemical analyses were used in the present study. Eleven were taken from the Richtersveld Memoir (de Villiers and Söhnge, 1959). Three were analysed for the writer by E.C. Haumann of the Division of Chemical Services, Pretoria, S.Afr.; and the remaining eleven were analysed by Mr. A.J. Erlank (of the Department of Geochemistry of the University of Cape Town) and the writer. For details of the analytical methods employed in these latter analyses refer to Appendix I.

(c) Map Construction: As a principal point lay-down of the south and eastern parts of the area was not available, the first operation undertaken was to establish ground control in this part of the area. After this had been done a principal point lay-down of this area was constructed using the "hand templet method" (American Soc. Photogrammetry, 1952, p.p. 419-423). Once a principal point lay-down for the whole area had been established, contouring and the transference of geological field data proceeded as spot heights and field data became available. As relatively small areas were mapped at any particular time it was found more convenient to use a radial intersector than a plotting machine in the transference of data from the aerial photographs to the base map. Contours were considered essential in deciphering the relationships existing between the different members of the Richtersveld Suite, as previous workers (de Villiers and Söhnge 1959 p.p. 74-75) believe the dark syenitic rocks to be roof remnant of a larger mass of granite. The contour interval selected for the final map was 250 feet (English). Smaller intervals were found to obscure geological detail as they tended to be too close together in the area of rugged relief bordering the Orange River.

The area proved to be well suited to photogeological interpretation, because (1) the vegetal cover was sparse except for the narrow strip of vegetation which grows on the alluvium on both sides of the Orange River, (2) the only man-made features that showed up on the aerial photographs were the old and new Klein Helskloof passes, and (3) superficial deposits were generally small in area and few in number. These three factors, together with the marked colour contrast that exists between the major rock units of the area, enable an observer to trace rock unit boundaries between traverses (see plates 4 and 7). The coherent aerial perspective obtained from the stereoscopic study of aerial photographs of the area, proved particularly useful in mapping fault and fracture traces (Lattmann and Nickelsen 1958).

(iii) Rock Classification:

(a) Igneous: It was decided to employ the A.G.I. descriptive model classification of igneous rocks compiled by Peterson (1961), and in those parts where this system is ambiguous (i.e. on grain-size), it was decided to follow the recommendations of the 1936 B.A. committee on petrographic nomenclature. In using this classification it seems important to state explicitly at the outset that (a) all references to mineral percentages refer to volume percentages; (b) the term alkalic is taken to include K-felspar, all types of perthite and albite (An_{0-10}), (c) Colour Index is defined in the Shand (1950.B, p.233) manner, and (d) when mention is made of grain-size, the ferromagnesians in rocks with colour indices of less than 20, and the phenocrysts of porphyritic rocks with less than 50% phyric components, were disregarded.

The wide major dykes of the area belong to Peterson's (1961, p.33) "Diorite Class" although their plagioclase tends to be slightly more sodic than his normal diorite. These dyke rocks also fall within Petterson's (1961, p.31) definition of Lamprophyre partiularly his Spessartite and Odinite groups. It was decided to call the major dyke rocks "hornblende diorite dykes" and to recall their lamprophyric affinities when discussing their genesis.

The other prominent dyke rocks of the area straddle Peterson's (1961, p.32) trachyte and rhyolite classes and grade in grain size into the micro-syenites and micro-granites. It was found that any attempt to fit this swarm of gradational dykes into four rigid classes based on grain size and the percentage of modal quartz present was more misleading than useful, particularly as the modal quartz content of the finer dyke rocks could not be determined with the same precision as that found in the coarser dykes. It was decided to call these dykes quartz bostonites, as the majority of them fall within Peterson's (1961 p.34) definition of bostonite - "aplitic alkalic syenite with sugary to trachytic texture". Hatch, Wells and Wells (1949 p.246) define bostonite as being a dyke rock that mineralogically and texturally includes both microsyenite and intrusive trachyte.

The ultramafic rocks of the area are of two main types - the serpentine-rich and the hornblende-rich rocks. Both groups are believed to have been formed originally by igneous processes though their present-day mineral assemblage has probably been significantly influenced by metamorphic action. As these rocks have both igneous and metamorphic characteristics it has been decided to call them serpentinites (Lodochniko, 1933) and hornblendites and to define these two terms in such a way that neither an igneous nor a metamorphic genesis is implied. Serpentinites and hornblendites are thus defined as rocks containing less than 5% quartz and alkalic felspar, and less than 10% plagioclase (An 10-100), pyroxene, and olivine. Their colour index is greater than 90. In the serpentinites minerals of the serpentine group make up more than 50% of the rock, and in the hornblendites, hornblende constitutes more than 50% of the rock.

(b) Sandstone Classification: Since the publication in 1948 of Krynine's classification utilizing ternary diagrams there has been a renaissance in thought on this subject, and a number of useful and interesting innovations have been introduced. The schemes that have attracted the most attention are those of Pettijohn (1949), Packham (1954), Folk (1956), Gilbert (i.e. in Williams, Turner and Gilbert 1955) and Crooke (1960). The relative merits and demerits of these classifications have been discussed by Crooke (1960) who advocates the use of Packham's classification in a modified form.

Packham's (1954) classification is important in that he is able to show that other classifications based entirely upon mineralogical and textural characteristics are unable to distinguish between the sediments deposited by (1) traction currents and (2) turbidity currents. By using sedimentary structures he divides sandstones into "shallow water sandstones" (or the arkose-quartzose sandstone suite) which are normally deposited by traction currents, and "deep water sandstones" (or the greywacke suite), which are usually deposited by turbidity currents. Earlier mineralogical and textural classifications failed because they did not take into account the overlap that occurs when one plots the matrix percentages of sediments from typical shallow and deep water environments on the same diagram. According to Packham (1954 p.473) the matrix percentage of the sandstones of the arkose-quartzose sandstone suite varies from 30% to zero with a maximum in the region of 5%, while the greywackes range from 10 to 70% matrix with a pronounced maximum about 35%. These data on matrix percentages are of particular interest as all the Nama System sandstones examined in the present study contained less than 10% matrix, and thus belong to Packham's arkose-quartzose sandstone suite. In the present study Packham's classification has been used and those parts of it which are relevant to the present study are reproduced in figure 6.

(c) Metamorphic classification: There is no generally accepted metamorphic rock classification. This is particularly true of hybrid rocks, as they are of dubious systematic position in both igneous and metamorphic classifications. While mapping metamorphic rocks in the field it was found convenient to use well known terms such as slate, schist and gneiss, and to qualify these terms with adjectives describing the colour, grain size and texture. Later when the specimen had been studied in thin-section the field rock name was modified so as to incorporate the name of the dominant mineral species found in the rock (ie. chlorite-schist). If the thin-section study of these rocks revealed relict textures and minerals which clearly indicated the character of the earlier unmetamorphosed rock (particularly when the former rock was igneous) the prefix meta- was added to the former rock name (ie. meta-rhyolite).

(iv.) Acknowledgments. The completion of this project was made possible by the assistance and cooperation of many individuals to all of whom the writer wishes to express his indebtedness. Special thanks are due to Prof. E.S.W. Simpson, Prof. H. Martin, Dr. M. Mathias and

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II. GEOMORPHOLOGY

Most of the land area of south and central Africa has been subjected to sub-aerial denudation since early Jurassic times and this long period of erosion has left its imprint on the present day landscape of the subcontinent. In recent years research by Cahen, Dixey, Fair, King (L.C), Lepersonne, Mabbutt, Pallister and Wellington has made it possible to correlate the major events in the denudational evolution of the subcontinent. These workers have established that the African landscape is characterized by the presence of a number of more or less well defined erosion surfaces, often of continental extent, separated by scarps at various heights. Thus if one wishes to understand the geomorphology of any particular region in Southern Africa it is important to see how it fits into the broader continental picture.

At present some six major erosion surfaces of continental extent are recognized (Dixey 1956, p.4.). They are: (1) the Gondwana cyclic landscape (Jurassic and early Cretaceous), (2) the Post-Gondwana cycle (late Cretaceous), (3) the African Cycle (mid-Tertiary ie. Miocene), (4) the Victoria Falls cycle (end-Tertiary), (5) the Congo Cycle (end-Tertiary), and (6) the Latest Cycle (present day). Cahen and Lepersonne (1952) regard the Congo Cycle, as developed in the Congo basin as being but slightly younger than the Victoria Falls cycle and place both within a group of end-Tertiary cycles.

L.C. King (1951) and Mabbutt (1955B) have indicated that they believe that most of Namaqualand belongs to the African Cycle of erosion with later cycles being found along the coastal belt and as a narrow strip along the Orange River up to the Aughrabies Falls.

The major relief features of Little Namaqualand and the Richtersveld fall into three broad zones as shown by Mabbutt (1955. A., p.77.). The western zone is a low coastal desert plain with ill-defined water courses. This first zone extends approximately 25 miles inland from the coast. The second zone is the dissected rim of the interior plateau and is approximately 30 miles wide. As this zone occurs at greater altitude than the first it tends to have more precipitation and this increased precipitation is reflected in the closer texture of the drainage systems and the greater relative, as well as absolute, relief. East of this zone the landscape opens out to merge into the monotonous Bushmanland Plateau, with its pans and broad sand-filled valleys. The area lies within the belt of rugged relief found along the edge of the interior plateau, but it is not typical of this zone as it adjoins the through-flowing Orange River. The arid nature of the area, and the fact that it is traversed by a through-flowing (or allochthonous river) rising in an area of much higher precipitation, are the two most significant factors in the denudational history of the area studied.

A comprehensive study of the evolution of the Orange River drainage system has been made by Wellington (1958). He regards the last great bend of the river (the Richtersveld bend) as being primarily due to the height and resistance of the Richtersveld rocks and he calls the north-south trending Stinkfontein range the "backbone of the Richtersveld" (p.23). It seems likely that in earlier times the pre-Nama rocks, now exposed in the Xaminxaip (6A8) area, were covered by Nama and possibly also Karroo beds. However, with regard to the Karroo beds, Haughton and Frommurze (1936 p.34) state that some of the present day topography of the Orange River valley between Vioolsdrif and Kwabs Drift is formed of pre-Karroo features particularly the prominent cliff Nama limestone which forms the eastern and northern limits of the Neint Nababeep plateau. This indicates that the present Orange River gorge through the Neint Nababeep plateau existed in pre-Dwyka times ($\pm$ 300 million years.).

As can be seen from the map accompanying this report the Orange River in the area mapped, first flows west south-west and then at Xaminxaip (6A8) - Nama for "water-bend" - executes a right-angled turn to flow northwards through a steep sided gorge cut in the hard rocks of the Richtersveld Complex (See plate 2.). When considered on a regional scale this gorge tract is of considerable interest. At a relatively short distance to the east of the gorge one of the minor tributaries of the Orange River has cut through approximately 1,500 feet of sediments. Thus the present day

Plate 2: The Orange River looking north from Xaminxaip (6 A 8)

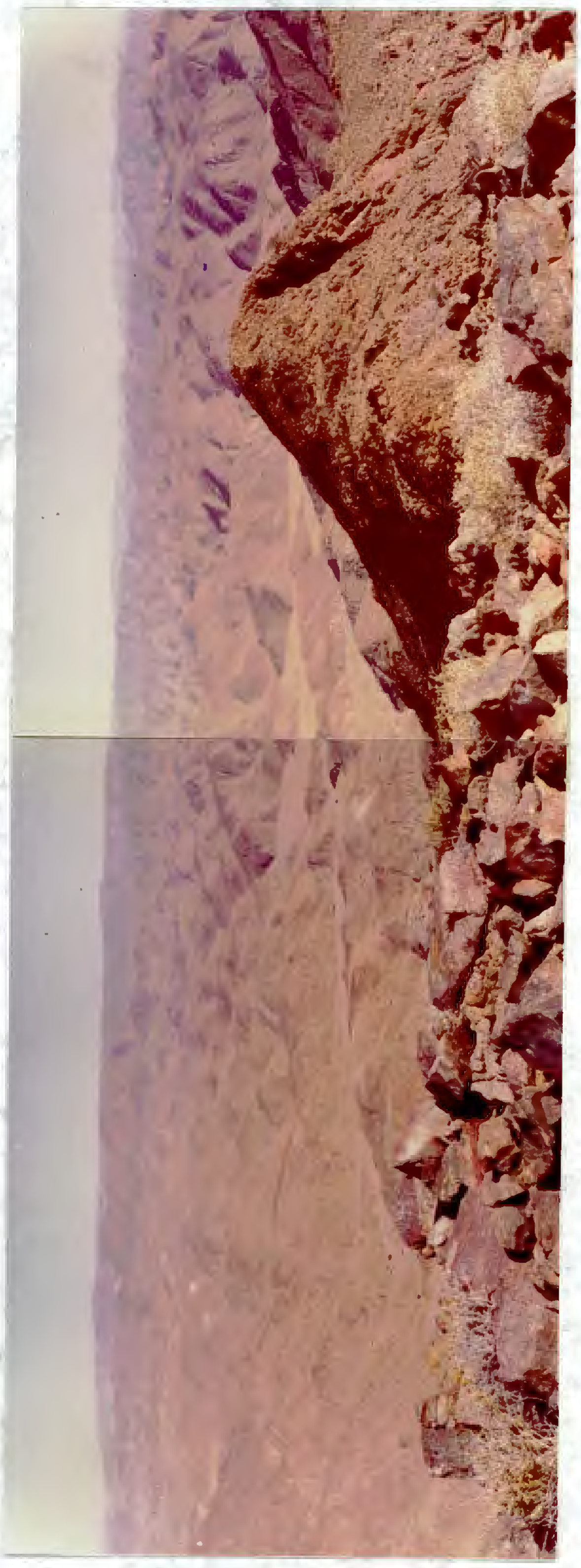
Orange River provides a striking example of superposition (Wellington, 1958, p.23.), as it cuts through the hard rocks of the Richtersveld Complex while softer rocks trending in outcrop in the same direction as it lie but a very short distance to the east (see map 2). The writer believes that this apparent anomaly is due to the former presence of a north-south topographic depression produced by the collapse of an underlying magma chamber. A fuller discussion of this process is given in chapter nine.

Mabbutt (1955 B), in his attempt to unravel the geomorphology of Little Namaqualand, recognized four erosion surface remnants - viz. at ± 3,900 feet his Namaqua Highland surface of Cretaceous age (Gondwana Cycle), at 3,300 feet his Bushmanland surface of Early to Middle Tertiary age (African cycle), and at 2,600 and 2,00 feet two surfaces which are confined to the dissected area bordering the Orange River and are of upper Tertiary and Lower Pleistocene age respectively (Victoria Falls cycle). He also believes that the major rejuvenation that led to the formation of the present-day Orange gorge commenced in Middle Pleistocene times.

De Villiers and Söhnge (1959) speak of a single high-level erosion surface, their Bushmanland Plateau, and they do not distinguish between Mabbutt's Namaqua Highland and Bushmanland surfaces. Below their high-level surface de Villiers and Söhnge (1959, pp.22-23) recognize two more erosion surfaces, viz. the two to three thousand foot "Orange River Mountain Land Surface" and the "two thousand foot surface". Both of these latter surfaces are considered to have evolved during the period between Late Jurassic ~~and Upper Jurassic~~ and Upper Cretaceous times.

In order to reduce as much as possible, the subjective element in the discovery of erosion surfaces, or their remnants, it was decided to subject the area to two well tried methods of morphological analysis. A hypsographic curve of the area was constructed using the random co-ordinate method of Strahler (1956 p.589), and it revealed (1) that the median height of the area was 1,730 feet (528 metres), (2) that there was a slight bevel at and immediately above the present level of the Orange River, (3) that there was a more significant bevel at ± 1,750 feet, and (4) that a very slight bevel occurred just below 2,300 feet (See Hanson-Lowe, 1935, p.180; Raisz, 1938, pp.269-270 and Sparks, 1960, pp.228-231 for a discussion of this method of morphological analysis.).

The second method tried was the projected profile method (see Barrell, 1920). As the Orange River occurs in the area as a single great bend, it was difficult to determine the most representative direction in which to place the projected profile/s. Finally it was decided to construct the profiles at right angles to the river towards the centre of the great bend at Xaminxaip (6A8), as this would enable the profiles to pass through both the Orange River and Rooiberg 2 (the highest point in the area).



Rooiberg 2

Plate 3: Panoramic view of the area from Cone Beacon (located east south east of the area).

It then became apparent that the area that could be legitimately projected was only 2 miles wide, and thus could be projected onto a single profile. Eventually it was decided to construct a projected profile of the highest points in the strip in the normal way, and in addition to construct a projected profile of the lowest points in the strip, so that a comparison between the two profiles would provide an estimate of the relative relief of the strip studied. The low point profile would also give a generalized idea of the grade attained by the tributaries in the strip. The projected profiles revealed (1) a slight planar element in the landscape at and just above Orange River level, (2) a more substantial bevel at approximately 1,750 feet, (3) a vague accord of summits at 2,300 feet and (4) a more steeply inclined plane between 2,800 - 3,000 feet on the northern side of Rooiberg 2 as illustrated in plate 6. The projected profiles also showed the extremely rugged nature in the terrain bordering the Orange River. The slight bevel at, and rising to slightly above, the level of the Orange River is the flood plain of the present day river. The 1,750 foot bevel is a part of de Villiers and Söhnge's 2,000 foot Windvlakte-Ankam surface. The vague 2,300 ft. bevel might be correlated with de Villiers and Söhnges Orange River mountain land surface or Mabbutt's 2,000 or 2,600 ft. surfaces. The 2,800 - 3,000 foot Rooiberg 2 (3F2) bevel (if it is in fact a feature of regional significance), might be part of Mabbutt's 2,600 foot surface though this is uncertain. No new evidence as to the age of these surfaces was found in the area studied, and it seems likely that the unraveling of the full denudational history of Namaqualand and Bushmanland will have to await the publication of accurate large scale topographic maps of these areas.

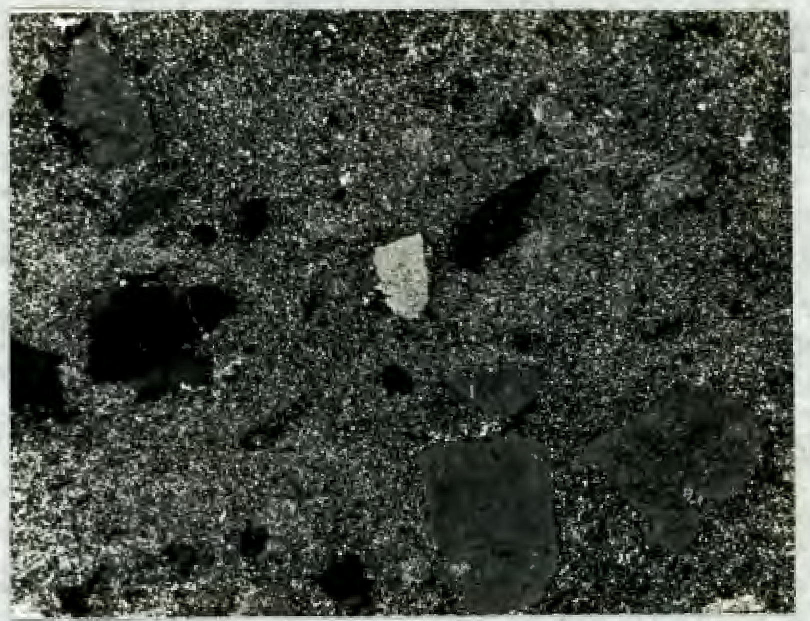
The memoir authors include within this rock group both hybrid rocks in various stages of assimilation and transformation and truly intrusive granite (p.64), and in their description of the grey, gneissic granite of the area they state that it tends to be a severely sheared (p.71.) augen-gneiss (p. 68.). Whilst it is not doubted that the "grey gneissic granite" is a valid rock unit outside the area, it has been found that within the area the separation of the "grey, gneissic granite" from the "hybrid rocks migmatites etc." is more confusing than useful, as it tends to hide the underlying unity existing between these two rock groups. With regard to both composition and texture, it was found that samples collected from the areas mapped by de Villiers and Söhnge as grey gneissic granite and from the areas of hybrid rocks, migmatite etc., were in most cases identical (See Table III.). It has also been found that by combining the two rock units the structure of the area can be much more readily interpreted. Thus in the present study the metamorphosed, metasomatised and ultrametamorphosed Kheis rocks and the central gneissic body, have been divided into three major rock groups, the division being primarily based on field appearance and textures. The three units will henceforward be known as (1) the Kheis Supracrustal Rocks (Kh), (2) the Transitional Acid Hybrid Rocks (Hy) and (3) the Adamellitic Gneiss (Gn).

(B) <u>The Kheis Supracrustal Rocks</u>:

(i) <u>Field Description</u>: The Kheis Supracrustal Rocks outcrop discontinuously along all but the northern borders of the area. The largest continuous outcrop is in the north-east. This outcrop continues until it is unconformably covered by rocks of the Kuibis Series which form outliers of the Neint Nababeep Plateau. In the south only a small tongue of the Kheis Supracrustal Rocks is found, but these rocks are part of a much more extensive outcrop located outside the area. In the west, as in the east, the Kheis Supracrustal Rocks occur along the border of the area, but little of this material outcrops within the area. Many isolated outcrops, too small to be shown on a map of the scale employed in the present study, occur within the zone of hybrids.

(ii) Petrography: The mineral composition of the Kheis Supracrustal Rocks varies greatly, but quantitatively quartz bearing types are much more abundant than might be expected from previous descriptions. Amongst the more important sub-groups recognized are the following: (1) brownish grey (5YR4/1) meta-rhyolites (ie. specimen 39) which have a characteristic appearance both in the field and on aerial photographs (See Photomicrograph 1.): (2) sericite and biotite schists (See Photomicrographs 2 and 3) which have lepidoblastic textures and occasionally contain relict felspar phenocrysts: (3) amphibole schists (ie. specimen 139) in which the dominant amphibole is probably actinolite ($Z = 1.625$, $Z.c = 20^{\circ}$ Z = light blue green (5BG 7/6), Y = moderate yellowish green (10GY 6/4) and X = pale greenish yellow (10Y 8/2)): (4) meta-siltstones (ie. specimen 44) which consist mainly of alternating quartz-rich ($\pm$ 70% of total rock) layers of varying grain size and thin irregular bands and lens - shaped aggregates of epidote, amphibole, albite and chlorite: (5) quartz - chlorite - epidote schists (ie. specimen 118) that show evidence of segregation layering (viz.: in the light grey (N8) bands quartz is dominant, and untwinned albite, chlorite and epidote are found in lesser amounts; in the dusky green (5G 3/2) bands chlorite is dominant, with epidote and an even smaller percentage of quartz being found; and in

Typical Kheis Supracrustal Rocks



Photomicrograph 1: Kheis Meta-Rhyolite (x14), No. 141, from the north-east corner of the area (8B4), x-nicols.

Photomicrograph 2: Kheis Quartz - Sericite - Schist (x15), No. 617, from west of Rooiberg 2, x-nicols.

Photomicrograph 3: As above but photographed in plane polarized light (x15).

the greyish yellow green (5GY 7/2) bands epidote is dominant, and chlorite and quartz are minor constituents): and (6) Meta-Sandstones. that range in colour from greenish grey (5G 6/1) specimen 481 to moderate greyish red (5R 4/4) specimen 482, and have the following range in modal compositions:

	481	482
Quartz	41	72
Sericite	42	20
Alkali-felspar	9	4
Calcite	7	
Biotite and Chlorite	1	3
Heavy Minerals	Tr.	1
	100	100

The quartz grains in specimen 481 have a mean diameter of 0.48 m.m. and are set in a lepidoblastic groundmass of sericite, and the grains in specimen 482 have a mean diameter of 0.83 m.m.

The above rock types which are generally quartz rich and of sedimentary origin, are thus seen to have more in common with the "quartzites and associated schists" of the Kaaien Series than the "amygdaloidal and highly altered associated lavas" of the Wilgenhoutdrift Series (Gevers and others 1937, pp.26-27.). They clearly do not fall into the "mafic lava, usually sheared; agglomerate" rock unit of the Wilgenhoutdrift Series as suggested by de Villiers and Söhnge (1959.). Prof. H. Martin (1963) has stated that he doubts whether the division of the Kheis rocks of the Richtersveld into the three major units found in Griqualand West is valid, as significant lithological changes would be expected over the great distances that separate these areas. The present writer believes that a detailed study of the Kheis System rocks of the Richtersveld will prove Prof. H. Martin to be correct.

(iii) <u>Chemistry</u>: The chemistry of the Kheis Supracrustal Rocks has not been specifically studied as their composition is highly variable and many of the rocks contain some form of segregation layering. It is believed from the petrographic data that some idea of the bulk chemical composition of these rocks can be obtained by adding together two parts graywacke (Pettijohn 1949, p.250.) and one part rhyolite (Nockolds 1954). An analysis of a lava, considered by de Villiers and Söhnge (1959, p.101.) to be similar in composition to the Kheis supracrustal material of the south-eastern Richterveld before it was metamorphosed, was included in the Richtersveld Memoir. Although it is believed that no single analysis can represent the heterogeneous Kheis supracrustal material, this analysis is included in table I as it is found that (1) its norm indicates that it belongs to Johannsen's (1932) Rhyodacite (227) family, and (2) when it is compared with Daly's (1933) average quartz latite and average andesite (see table I) it is found to be of intermediate chemical composition; and thus the chemical composition of this lava is in accord with the present writer's belief that the Kheis Supracrustal Rocks of the south-eastern Richtersveld are richer in silica than previously realized.

(iv) <u>Petrogenesis</u>: A cursory review of the mineral assemblage and texture of the Kheis Supracrustal Rocks studied suggests that these rocks may belong to the Greenschist Facies as defined by Fyfe and others (1958, pp. 217-224.), but more detailed study shows that the total sum of evidence is more consistent with their inclusion in the Almandine-Amphibolite Facies. The absence of minerals such as almandine, staurolite, kyanite and sillimanite which are the usual harbingers of higher metamorphic grade, is probably due to presence of a relatively high K_2O/Al_2O_3 ratio, under which

Table I. - Chemical Data (Kheis Supracrustal)

	Composition, weight per cent: (1)	(2)	(3)	Norm:	
SiO_2	60.77	62.43	59.59	Q	14.04
Al_2O_3	16.14	16.15	17.31	Or	22.24
Fe_2O_3	3.73	4.04	3.33	Ab	28.30
FeO	2.22	1.20	3.13	An	17.79
Mg O	2.81	1.74	2.75	En ($MgSiO_3$)	5.30
Ca O	5.08	4.24	5.80	Di ($Mg Si O_3$)	1.70
Na_2O	3.39	3.34	3.58	Di ($Ca Si O_3$)	1.97
K_2O	3.73	3.75	2.04	Mt	5.34
H_2O^+	0.95	1.90	1.26	Il	1.52
H_2O^-	0.04			Ap	1.01
CO_2	Nil			Hl	0.12
TiO_2	0.80	0.85	0.77	minor constituents	1.03
P_2O_5	0.39	0.27	0.26		100.36
MnO	0.11	0.09	0.18		
SO_3	nil				
Cl	0.07				
F	0.05				
BaO	0.04				
SrO	Tr				
	100.32				
-(Cl+F)=0	0.04				
	100.28	100.00	100.00		

(1) Amygdaloidal lava from the Wilgenhoutdrift Series southeast of Vioolsdrif - de Villiers and Söhnge (1959, p.101).

(2) Quartz latite (Number of analyses 12) - Daly (1933, p.13).

(3) All andesites (Number of analyses 87) - Daly (1933, p.16).

(4) Norm of analysis one.

conditions potassium felspar (Microcline) occurs and the formation of highly aluminous minerals is precluded (Fyfe and others, 1958, p.229).

Turner and Verhoogen (1960, p.553) state that metamorphism in the almandine - amphibolite facies normally covers a temperature range of from 550° to 750° C. and a range of pressures between 4,000 and 8,000 bars. These temperature-pressure conditions are however regarded as tentative as Poldervaart (1953, p.262) has noted that many petrologists have been unable to explain the lack of response shown by some basaltic rocks to metamorphic conditions. From this observation he has argued that the failure of reaction was mainly the result of a low water content in the basalt concerned. Yoder (1955, p.505) and Wilcox and Poldervaart (1958, pp 1323-1368) have expanded this concept and Yoder (1955, p.505) for example, believes that mineralogy may be dependent solely on the water content, and that high water contents can produce all the minerals of the greenschist facies without relevancy to temperature. Recent experimental work by Yoder and Tilley (1962, p.469) "for the most part" supports the concept that basalt metamorphism is dependent on water content rather than changes of total pressure and temperature.

Prior to metamorphism the Kheis Supracrustal Rocks studied appear to have been mainly impure siltstones, impure sandstones, and lavas that ranged in composition from basic to acid, but were predominantly acid. It is, however, possible that the silt-sized quartz grains so characteristic of the Kheis metasediment of the area studied, were originally deposited as sand-size particles and were later reduced in size by intense mechanical granulation. Such a process is thought to have operated in the formation of the schists of the Otago Province, New Zealand (Turner 1941, pp. 1 - 16).

The laminated rocks are considered to be the products of metamorphic differentiation (Turner and Verhoogen, 1960, p. 581). In support of this hypothesis it was found that (1) in most of the laminated Kheis meta-supracrustal rocks studied the same mineral assemblage is present in the different laminae even though the relative proportions of the different minerals in these laminae change radically; this clearly implies that chemical equilibrium was attained throughout the rock and that materials were free to diffuse not only within individual laminae but also between adjacent laminae; and (2) in many rocks one finds nearly pure laminae of epidote and chlorite which seem to be incompatible with any hypothesis postulating their derivation from magmatic sources. Turner (1941, p.4) has given a summary of the many processes that have been proposed to account for these laminated schists, and he concludes that the coarse-grained laminated schists of Otago (which are similar to the Kheis rocks found in the area) are blastophyllonites that have developed from greywackes by a process of intensive mechanical granulation accompanied in the later stages by mineralogical reconstitution under the influence of slowly increasing temperature.

The schistosity of the Kheis meta-supracrustal rocks is believed to result from the stresses generated and the resultant shearing produced by increases in volume consequent upon the development of the Adamellitic Gneiss and the introduction of granitizing fluids (Read, 1957, p.42). Kranck (1957, p.267) in his study of folding-movements in the basement zone has given strong support to this view by his observation that between the unmetamorphosed sediments and the gneisses there is generally a zone with strong shearing.

Plate 4: North of Xaminxaip looking southwards, showing the Kheis meta-supracrustal rocks intruded by the Alaskitic Granite.

(v) Resumé: The petrography of the Kheis Supracrustal rocks of the area indicates that prior to metamorphism these rocks consisted mainly of impure silt-stones, impure sandstones and lavas of predominantly acid composition. These rocks appear to be more akin in composition to the rocks of the Kaaien Series than the Wilgenhoutdrift Series with which they were previously correlated. Metamorphism attained the almandine-amphibolite grade, and some metamorphic differentiation took place. The schistosity of the Kheis Supracrustal Rocks is believed to have resulted from the stress generated and the shearing produced by increases in volume consequent upon the genesis of the Adamellitic Gneiss.

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(C) Transitional Acid Hybrid Rocks (Hy):

(i) Field Description: Field appearance was the main criterion used in recognizing this rock type as it lacks the schistosity of the Kheis Supracrustal Rocks, the broader banding of the banded gneiss, and the granitic to gneissose texture of the more homogeneous "core gneisses". These hybrid rocks are in fact, characterized by the appearance of alkalic felspar porphyroblasts set in a matrix that has retained part at least of its laminated or schistose texture. The occurrence of felspar porphyroblasts is considered to be of particular significance as it indicates that these rocks are probably not only the products of metamorphism but also of metasomatism. In the area to the north, north-west of Rooiberg 2 (3F2) the rocks of this group are frequently more intensely sheared than the normal hybrid rocks of the area, and they tend to have a pseudo-schistose appearance, but a study of their mineral composition indicates that they too belong to the hybrid group as they are found to have been permeated by granitic solutions.

(ii) Petrography: The colour of the Transitional Acid Hybrid Rocks is generally greenish grey (5GY 6/1) to greyish green (10G 4/2) and the rocks often contain very light grey (N8) felspar porphyroblasts. When compared with the Kheis Supracrustal Rocks the composition of these hybrid rocks is found to be much more uniform as can be seen from the following mean mode (Vol.%), and the standard deviations of the major constituent mineral species, of 11 specimens of this rock type:

	Quartz	Alkalic Felspar	Plagioclase	White Mica	Epidote Group	Biotite
$\bar{X}$	36.3	15.3	14.0	13.5	7.6	7.0
s	6.6	8.5	5.9	12.8	6.6	8.6

	Chlorite	Calcite	Opaque Ore Minerals	Apatite	Zircon
$\bar{X}$	4.2	1.8	0.2	Tr.	Tr.
s	7.8	2.4	-	-	-

If the above mean hybrid rock is classified as though it were an igneous rock, it is found to fall within Johannsen's (1932) Adamellite subdivision of the Granite (226") family, or Peterson's (1961, p.32) Adamellite or Quartz Monzonite class. The true systematic position of these hybrid rocks is difficult to ascertain but it is clear that they have much in common with Dietrich's (1960B, p.101) permeation gneiss which he defines as a gneiss, commonly banded, that consists in part of materials introduced by percolating hydrous solutions, gases, molecules, or ions. The grain size of this rock group is variable, but as the mean thin-section grain area is 0.43 sq.m.m., the rock group is considered to be coarse grained.

The quartz found in these hybrid rocks tends to be clear, anhedral, and is frequently strained. Most of the quartz occurs in lenses, laminae and patches of less regular shape. Microcline is the dominant alkalic felspar present. The microcline crystals are generally subhedral and fresh, and in some specimens they can be seen to have replaced earlier altered felspathic material. In contrast to the microcline the plagioclase crystals are generally altered. Epidote, zoisite and sericite are the main alteration products. The plagioclase has a mean composition of An 30. The chlorite tends to occur in anhedral plates that go to form irregular bands. The mineral is pleochroic from colourless (Z) to light green (5G 7/4) (X and Y) and it tends to be found associated with epidote, biotite and the opaque ore minerals. The biotite occurs in similar ragged anhedral crystals and is frequently closely associated with the chlorite. Its pleochroism is variable, but is generally X = very pale orange (10YR 8/2) Z = moderate brown (rYR 4/4). Epidote (pistacite and clinozoisite) allanite and zoisite are all found in this rock group. Apatite and rounded to sub-rounded zircon crystals occur in accessory amounts ~~inxsubhedralxcrystals~~, and they are generally associated with the irregular bands of ferromagnesian minerals. Magnetite is the principal ore mineral found. Some specimens contain patches of leucoxene thus indicating that the original ore minerals contained some titanium mineral - probably ilmenite. The calcite found appears to be secondary.

(iii) Chemistry: The chemical composition of the Hybrid Rocks (column 1, table II) is found to be similar to Nockolds' (1954, p.1014) "average adamellite" (Column 2, table II), to which it is petrographically akin. The Hybrid Rocks do however contain more Fe_2O_3, MgO and CaO and less K_2O than the average adamellite. While it is not possible to contrast the chemical composition of the Hybrid Rocks with the bulk chemical composition of the Kheis Supracrustal Rocks, the chemical composition of the unaltered Kheis lava (column 1, table I) can be compared with that of the Hybrid Rock (column 1, table II). In order to facilitate this latter comparison the Barth (1948, p.54) standard cells of both rock types were calculated and found to be as follows:

(1) the fresh Kheis (Rhyodacite) lava =

$$K_{4.33}\,Na_{5.95}\,Ca_{4.92}\,Mg_{3.81}\,Fe_{4.23}Mn_{0.08}Al_{17.22}Ti_{0.50}\,P_{0.30}$$
$$Ba_{0.02}\,Si_{55.01}\left[O_{148.52}\,(OH)_{11.48}\right]\ 160$$

Table II - Chemical Data (Hybrid Rock)

	Composition, weight per cent: (1)	(2)		Norm: (3)
SiO_2	67.80	69.15	Q	29.76
TiO_2	0.56	0.56	Or	13.90
Al_2O_3	14.51	14.63	Ab	25.68
Fe_2O_3	2.08	1.22	An	17.24
FeO	2.00	2.27	C	0.61
MgO	2.27	0.99	Hy ($MgSiO_3$	5.70
CaO	3.74	2.45	Hy ($FeSiO_3$	1.19
Na_2O	3.27	3.35	Mt	3.02
K_2O	2.34	4.58	Il	1.06
H_2O^+	1.33	0.54	Ap	0.67
H_2O^-	0.04		Hl	0.47
P_2O_5	0.26	0.20	Minor Constituents	1.42
MnO	0.08	0.06		100.72
CO_2	Nil			
SO_3	Nil			
Cl	0.29			
F	0.04			
BaO	0.05			
SrO	Tr.			
	100.32			
-(Cl+F)=O	0.04			
	100.28	100.00		

(1) "Granitized Wilgenhoutdrift lava south of Rooiberg II" - de Villiers and Söhnge (1959, p.102).

(2) Average Adamellite (Av. of 121) - Nockolds, 1954, p.1014

(3) Norm of specimen (1)

(2) the Hybrid Rock =

$K_{2.61}Na_{5.53}Ca_{3.50}Mg_{2.98}Fe_{2.83}Mn_{0.06}Al_{14.92}Ti_{0.38}P_{0.21}$
$Ba_{0.02}Si_{59.25}\left[O_{144.46}(OH)_{15.54}\right]160$

Thus to convert the fresh lava into the granitized lava (Hybrid Rock) there must be additions of 4.24 ions of Si, and 4.06 ions of (OH), and the removal of the other major elements in the following proportions,

$K_{1.72}, Na_{0.42}, Ca_{1.42}, Mg_{0.83}, Fe_{1.40}, Mn_{0.02}, Al_{2.30} Ti_{0.12}$, & $P_{0.09}$.
It seems, however, extremely unlikely that prior to metasomatism specimen 2 was identical to specimen 1, as there is approximately 20 miles separating the localities from which these two samples were collected, and as we have seen the Kheis Supracrustal Rocks are very variable in composition.

(iv) Resumé: The Transitional Acid Hybrid Rocks are more uniform in composition than the Kheis Supracrustal Rocks, and their petrography and chemistry is very similar to that of a normal adamellite. The petrogenesis of this rock type will be discussed in the next section.

(D) The Adamellitic Gneiss (Gn):

(i) Field Description: The Adamellitic Gneiss is not a homogeneous rock but rather the dominant rock type in a group of gradational gneissic rocks having the bulk composition of adamellite. Many of these rocks fall within Turner and Verhoogen's (1960, P.371) definition of a migmatite, viz. "those mixed rocks in which a granitic component (in the broad sense of the phrase) is clearly recognizable". These rocks are, however, the oldest granitic (once again used in the broad sense of the phrase) rocks found in the area and they go to form part of what Gevers and others (1937,p.36) have called the great granite-gneiss massif of Namaqualand and Bushmanland. As to its age, it is clear that the Adamellitic Gneiss is younger than the supracrustal material of the Kheis System which it invades, and older than the rocks of the Richtersveld Suite which were emplaced into it. De Villiers and Söhnge (1959, p.64) state that these rocks (the Adamellitic Gneiss) form part of "the major magmatic cycle loosely known as the "Older granites" which probably denote the end of the Archaean era". Recently Nicolaysen (1962B and C) has indicated that new absolute age determinations on the Namaqualand Gneiss and its associated pegmatites show that this rock type has an absolute age of 980 ± 100 million years. An age but slightly in excess of 980 ± 100 million years is suggested for part at least of the Kheis System (Nicolaysen 1962C, p.586). While the Adamellitic Gneiss as found in the area studied occurs in close association with the Kheis Supracrustal Rocks and this propinquity is believed to be of assistance in investigating the origin of the Adamellitic Gneiss, this very factor has resulted in the Adamellitic Gneiss of the area being heterogeneous. It is thus suggested that the area is not the most suitable place for collecting specimens of the Adamellitic Gneiss for absolute age determinations. Further comments on this topic are to be found in Chapter 6, Section E.

(ii) Petrography: The Adamellitic Gneiss includes de Villiers and Söhnge's (1959) "hybrid rocks, migmatites etc.", and their "grey, gneissic granite". In compiling modal data on the Adamellitic Gneiss an attempt was made to separate these two sub-groups and the results of this modal study are presented in Table III. It is evident from Table III that the two sub-groups are very similar even although Gn_2 (grey, gneissic granite) tends to be richer in alkalic felspar and poorer in chlorite than

Table III - Modes (Adamellitic Gneiss)

	No. of specs.	Quartz	Alkalic Fel-spar	Plagio-clase	White Mica	Bio-tite	Epi-dote Group	Chlo-rite	Cal-cite	Opaque ore minerals	Apatite
$\bar{X}$ Gn.1.	32	35.7	19.5	22.5	7.8	6.3	5.2	2.2	0.4	0.3	0.1
s		9.9	12.0	12.4	7.8	5.9	6.0	4.7	-	-	-
$\bar{X}$ Gn.2.	9	31.4	25.3	19.2	9.6	7.5	6.1	0.1	0.8	Tr	Tr
s		7.3	13.6	10.6	6.3	6.2	7.1	-	-	-	-
$\bar{X}$ Gn.1+ Gn.2.	41	34.7	20.8	21.7	8.2	6.5	5.4	1.8	0.5	0.3	0.1
		9.1	13.0	12.0	7.5	6.0	6.3	4.2	-	-	-

	Hornblende	Fluorite	Sphene	Zircon	Rutile and Leucoxene
$\bar{X}$ Gn.1.	Tr		Tr	Tr	Tr
s	-		-	-	-
$\bar{X}$ Gn.2.	-	Tr			
s	-	-			
$\bar{X}$ Gn1.+ Gn2	Tr	Tr	Tr	Tr	Tr
s	-	-	-	-	-

(Granodiorite (227") of the subdivision Adamellite - Johannsen, 1932.)

Gn_1 = "hybrid rocks, migmatites etc" and Gn_2 = "grey, gneissic granite".

the more strongly banded Gn_1 (migmatite). The large standard deviations of the individual minerals shown in table III are not considered particularly significant, as it is clear from an examination of the hand specimens from which the thin-sections were cut, that as a result of the banded nature of the rock some thin-sections are richer in minerals from the light bands, and others are richer in material from the dark bands of these rocks. The mean modal value of 41 specimens quoted in table III is however, considered significant as it is believed to represent the bulk mineral composition of the Adamellitic Gneiss as it is found in the area.

The texture of these adamellitic rocks ranges from gneissic to banded, and their colour ranges from very light grey (N8) to moderate pink (5R 7/4) for the leucocratic felsic bands, to dark greenish grey (5GY 3/1) or greenish bläck (5 GY 2/1) for the dark mafic bands. Grain size is variable but the mean thin-section grain area of these rocks was found to be 2.04 sq. m.m. When the banding and/or foliation of these rocks is considered on a regional scale it is found that the adamellitic bodies tend to have cores of augen-gneiss, which grade into banded gneiss which in turn pass into the Transitional Acid Hybrid Rocks. In most parts of the area the banding tends to be parallel with the foliation found in the transitional rocks and the laminations found in the Kheis schists, but in a few localities, particularly in the south-east of the area, complicated "wild" or ptygmatic folding is found and the individual bands and veins are highly convoluted. Plate 5 is of a particularly interesting banded gneiss found to the immediate south-west of Rooiberg 2.

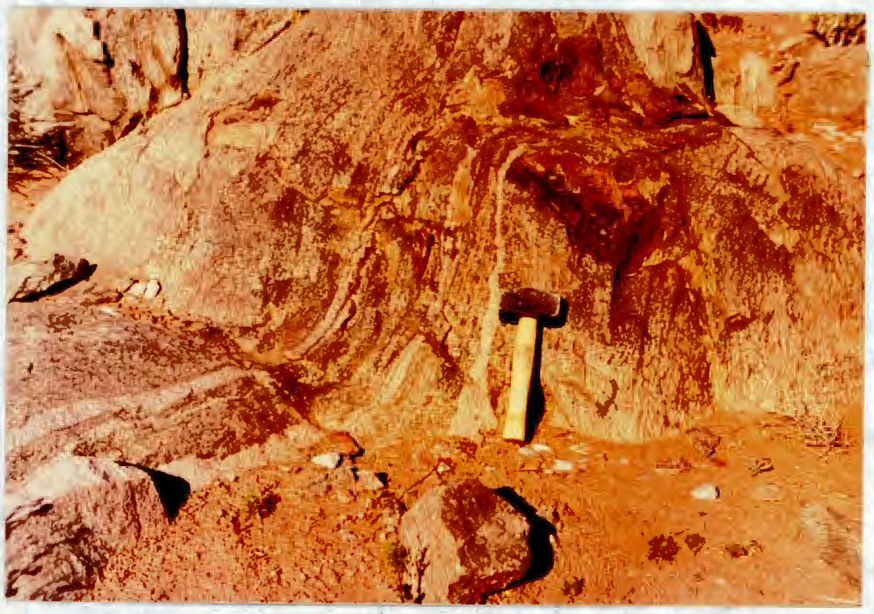


Plate 5: Banded Gneiss from south-west of Rooiberg 2.

A very similar rock type is also found to the east of Rooiberg 2. These rocks differ from the majority of specimens of the Adamellitic Gneiss in that their quartzo-felspathic bands are moderate pink (5R 7/4) in colour, and their alkalic felspar content is clearly higher than the mean of the Adamellitic Gneiss as given in table III.

The anhedral quartz found in the Adamellitic Gneiss is frequently strained and granulated, and occurs (1) in lens-shaped or irregular vein-like aggregates, (2) included within and/or intergrown with the felspar, and (3) in cross-cutting veinlets.

The principal alkalic felspar is microcline and it has the following optical properties X = 1.519, Y = 1.524 and Z = 1.526: 2V = 82°. It occurs in large subhedral plates. Crystals generally show gridiron structure and in many cases microperthitic intergrowths. The plagioclase component in the microcline microperthite crystals is generally too fine to identify.

The plagioclase felspar found in discrete crystals typically occurs in subhedral laths which are more altered than the alkalic felspar. The commonest alteration products are sericite, minerals of the epidote group, and minute irregular grains of calcite.

The composition of fresh specimens of plagioclase was found to vary from acid oligoclase to middle-andesine (An 40). The mean value obtained was An31 (X = 1.545, Y = 1.548, Z = 1.552). Some of the "fresh" plagioclase crystals are so much fresher than other plagioclase crystals in the same thin-section that they appear to have developed later than the bulk of the plagioclase.

The white mica is mainly sericite and occurs in small scaly aggregates and shreds associated with altered felspar. In some of the more strongly banded rocks the white mica is found associated with other micaceous minerals in irregular bands in which the long axes of the crystals tend to lie parallel to the banding. Biotite occurs in anhedral plates and laths that tend to cluster together to form irregular bands. Its optical properties are as follows: X = 1.595 - greyish yellow (5 Y 8/4), Z = 1.637 = moderate olive brown (5 Y 4/4) to light olive (10 Y 5/4): 2Vx = 3° (Siderophyllite).

Minerals of the epidote group occur both as small granules replacing plagioclase and biotite, and in clusters of larger crystals within the dark bands of the Adamellitic Gneiss. Pistacite, zoisite, ~~and~~ allanite and clinozoisite are all found in this rock group. The allanite is mainly pleochroic from X and Y = yellowish grey 5 Y 7/2) to Z = moderate orange pink (5Yr 8/4), but in a few specimens (i.e. specimen 468) it is pleochroic from X and Y = greyish orange (10 Yr 7/4) to Z = dark yellowish orange (10 YR 6/6).

Chlorite is mainly found associated with biotite, particularly in the dark bands of the banded specimens of the Adamellitic Gneiss. Patches of chlorite with indistinct borders are found within some of the biotite plates. The interference colours of many of the crystals is an anomalous "Berlin blue", thus suggestion that much of the chlorite is penninite.

Accessory minerals include apatite, magnetite, zircon, ilmenite, fluorite, sphene, rutile, leucoxene and hornblende [X = pale yellowish orange (10 YR 8/6), Y = moderate yellow green (5 GY 7/4) and Z = dark yellowish green (10 GY 4/4)]

(iii) Chemistry: In table IV the chemical composition of the Adamellitic Gneiss is compared with Nockolds' (1954, p.1014) average adamellite, Poldervaart's (1955, p.135) average quartzo-felspathic gneiss and Turekian and Wedepohl's (1961) average high Ca granite. The Adamellitic Gneiss is found to be normal with regard to SiO_2, TiO_2, Fe_2O_3, CaO, Na_2O, K_2O and MnO slightly high in MgO, slightly low in Al_2O_3 and significantly low in FeO and P_2O_5. When compared with Turekian and Wedepohl's (1961) average high Ca granite the trace element content of the Adamellitic Gneiss is found to be significantly high in Cl and Cs, moderately high in Li, Rb, Sn, Cu, Pb, Ga, and Tl and slightly low in Sr.

As the modal apatite and hornblende content of the Adamellitic Gneiss is low it seems probable that the high Cl content is to be found within the micaceous minerals. The high Cl content of the rock is considered significant as a trend towards high Cl values is found in the rocks of the Richtersveld Suite. The rubidium content is high, but the K/Rb ratio is 163 and thus within the limits of scatter of normal K/Rb ratios (Taylor 1960B, p.318) as can be seen in figure 12 (Appendix 2). The Cs content of the Adamellitic Gneiss is significantly higher than both of Turekian and Wedepohl's (1961) granite averages. In attempting to

Table IV - Chemical Data (Adamellitic Gneiss)

	(1)	(2)	(3)	(4)	(5)	(6)	(7)
SiO_2	71.84	68.20	69.96		69.15	70.7	67.38
TiO_2	0.54	0.78	0.42		0.56	0.5	0.57
Al_2O_3	13.10	13.97	13.82		14.63	14.5	15.49
Fe_2O_3	1.23	1.70	1.92		1.22	1.6)	3.80
FeO	1.04	1.77	1.08		2.27	2.0)	
MgO	1.51	1.89	1.88		0.99	1.2	1.57
CaO	2.24	2.60	2.52		2.45	2.2	3.54
Na_2O	3.54	3.50	3.40	2.28	3.35	3.2	3.83
K_2O	3.80	3.76	3.31	3.70	4.58	3.8	3.04
H_2O^+	0.83	0.75	0.71		0.54		
H_2O^-	0.13	0.15	0.10				
CO_2	0.12	0.09	0.35				
P_2O_5	nil	0.07	nil		0.20	0.2	0.21
MnO	0.04	0.08	0.09		0.06	0.1	0.07
SO_3	0.04	0.02	0.04				
Cl	0.72	0.14	0.15				0.01
F	0.08	0.10	0.05				0.05
BaO	0.05	0.06	0.06				0.05
SrO	nil	nil	nil				0.05
Li(p.p.m)				33 p.p.m			24 p.p.m.
Rb(p.p.m.)				189 p.p.m.			110 p.p.m.
Cs(p.p.m.)				approx 8 p.p.m.			2 p.p.m.
Sn(p.p.m.)				approx 5 p.p.m.			1.5 p.p.m.
Cu(p.p.m.)				45 p.p.m.			30 pp.m.
Pb(p.p.m.)				25 p.p.m.			15 p.p.m
Ga(p.p.m.)				25 p.p.m.			17 p.p.m.
Tl(p.p.m.)				1.45 p.p.m.			0.72 p.p.m
	100.81	99.63	99.86				
-(Cl+F)=O	0.20	0.07	0.05				
Total	100.61	99.56	99.81		100.00	100.0	

(1) de Villiers and Söhnge (1959, p.69): Grey Gneissic Granite: "Sample collected next to canal upstream of Vioolsdrif".

(2) de Villiers and Söhnge (1959, p.69): Grey Gneissic Granite: "Sample collected between Armbank and Kaalbeen".

(3) de Villiers and Söhnge (1959, p.69): Grey Gneissic Granite: "Sample collected northwest of Great Bend (A.2)." (i.e. A2 their grid: Ref. figure 1.)

(4) Specimen 468, an Adamellitic Gneiss from the south east of the area (7 G1). The mode (vol %) of this specimen is 31.4% quartz, 25.0% plagioclase (An 31), 16.7% alkalic felspar, 15.6% biotite, 7.9% epidote, 2.3% white mica, 0.6% chlorite, 0.2% allanite, 0.2% sphene, 0.1% apatite, Tr. Zircon.

(5) Nockolds' (1954, p.1014) average adamellite (av. of 121).

(6) Poldervaart's (1955, p.135) average quartzo-felspathic gneiss (av. of 250).

(7) Turekian and Wedepohl's (1961) high- Ca-Granite: Table 2.

Table V - Norms (Adamellitic Gneiss)

	(1)	(2)	(3)*
Q	32.34	25.92	30.66
Or	22.24	22.24	19.46
Ab	26.72	28.82	28.30
An	10.29	11.68	10.29
C	-	-	0.92
Hy ($MgSiO_3$	-	4.70	-
Hy ($FeSiO_3$	-	0.53	-
En $MgSiO_3$	3.80	-	4.70
Mt	1.62	2.55	2.55
Hm	0.16	-	0.16
Il	1.06	1.52	0.76
cc	0.20	0.20	0.80
Fr	0.07	0.11	-
Hl	1.17	0.23	0.23
Minor constituents	1.09	1.13	0.96
Total	100.76	99.63	99.79

* Numbers refer to the same specimens as those found in Table IV.

account for this high value it is relevant to note that Turekian and Wedepohl's (1961) average shale contains 5 p.p.m. Cs; thus it might well be that specimen 468 was taken from an area of local Cs enrichment resulting from the granitization of shaly material. It would thus seem that a great many more Cs determinations will have to be made on the Adamellitic Gneiss before a significant statement can be made on its average Cs content.

The Barth (1948, p.54) standard cell of the mean of analyses 1, 2 and 3 of Table IV was found to be as follows: $K_{4.10}$ $Na_{5.92}$ $Ca_{2.33}$ $Ba_{0.02}$ $Mg_{2.33}$ $Fe_{2.03}$ $Mn_{0.05}$ $Al_{14.22}$ $Ti_{0.40}$ $Si_{61.98}$ $P_{0.02}$ $\left\{O_{151.06}\ (OH)_{8.94}\right\}_{160}$. If this standard cell is compared with that of the Hybrid Rock which was given earlier in this chapter, it is found that additions of 1.49 K ions, 0.39 Na ions, 0.02 Ti ions and 2.73 Si ions, and subtractions of 1.17 Ca ions, 0.65 Mg ions, 0.80 Fe ions, 0.01 Mn ions, 0.70 Al ions, 0.19 P ions and 6.60 (OH) are required to convert the hybrid rock into Adamellitic Gneiss.

Figure 9 (see Appendix 2) is a modified Simpson (1954A) diagram, on which Green and Poldervaart (1958, p.96.) have plotted and contoured the concentrations of 201 averages of igneous rocks, and by so doing indicate the main or normal differentiation trend of igneous rocks. The mean values of the principal rock types of the area were plotted on this modified diagram and it was found that the rocks of the Richtersveld Suite (Tables IX and X), and the Kheis Lava (Table I) fell within the normal igneous trend, but that the Adamellitic Gneiss plotted off the trend as did the Hybrid Rock (Table II), which plotted in an intermediate position between the unmetamorphosed Kheis Lava and the Adamellitic Gneiss. These data are considered to indicate that the Adamellitic Gneiss is not a normal igneous rock.

(iv) Petrogenesis: It is considered important when discussing the evolution and emplacement of granitic rocks to have a clear picture of the temperature-depth, or intensity zone, in which the rocks developed, and into which they were emplaced. The Adamellitic Gneiss is believed to have been emplaced into the lower "Mesozone" or upper "Catazone" for the following reasons - (1) the grade of metamorphism found in the Kheis Supracrustal Rocks of the area, (i.e. Buddington 1959,p.676, states that the amphibolite facies starts at the top of the "Catazone".), (2) the occurrence of migmatites, (3) the absence of any signs of chilling, (4) the presence of gneissic foliation, and (5) the gradual and generally concordant contacts of the Adamellitic Gneiss. The writer wishes to stress the deep-seated environment in which this granite developed as it is desired, at a later stage, to contrast the physical environment into which it was emplaced (i.e. an environment in which the invaded country rocks would tend to yield by flow or plastic distention) with the epizonal environment into which the later Richtersveld Granite was emplaced (i.e. an environment in which the invaded country rocks would yield by fracture.). In reviewing the origins proposed for granitic rocks emplaced into the Catazone, Buddington (1959, pp. 714-715) states that in the majority of cases "the mechanics of emplacement is at present indeterminate, problematical, or the subject of controversy".

Wegmann (1935) has estimated the minimum depth of the migmatite front to be 10 km (6¼ miles); and Buddington (1959, p.676) believes that erosion has rarely exposed rocks that were ever at a depth greater than 12-15 miles (19-24 Km.); thus it would seem likely that the Adamellitic Gneiss was emplaced at some intermediate depth between these two limits. Evidence from the grade of metamorphism of the Kheis supracrustal material studied suggests a similar depth of emplacement for the Adamellitic Gneiss, because according to Buddington (1959,p.676.)

"... the top of the catazone where the amphibolite facies starts..." ranges ..."from as shallow as 4 miles to as deep as 10 miles."

In an attempt to establish the approximate temperature at which the Adamellitic Gneiss formed, the Barth (1956, pp.3-16) distribution-between-solvents felspar geothermometer method was used. While it is not certain that the felspars in the Adamellitic Gneiss had "simple" (Dietrich, 1960B, p.46) thermal histories, the temperature of ± 450°C, (12% Ab in Alkali-felspar, 69% Ab in Plagioclase; k = 0.174), that this method educed, is believed to be of the correct relative magnitude, as it is approximately half-way between the formation temperatures Barth (1956, p.14) proposed for granodiorites and granitic gneisses. It must, however be stressed that this is an approximate temperature as the Adamellitic Gneiss probably cooled relatively slowly, and the temperature measured may represent any intermediate temperature between the consolidation temperature and the minimum temperature measurable by the geothermometer (Dietrich, 1960B, p.45.).

Many petrologists believe that the "majority of the medium sized and larger zircons" of intrusive granites are euhedral, and that zircons of autochthonous and paraautochthonous granites and migmatites are "frequently not euhedral" (Poldervaart, 1956, p.530). Taking this statement as their premise, Coetzee (1942, pp.91-112) and Poldervaart and von Backström (1949, pp. 433-495) have studied the zircons of the Kheis System supracrustal rocks and the rocks emplaced into them in the Goodhouse-Pella and Kakamas areas of the N.W. Cape. The Kheis supracrustal rocks of the south-eastern Richtersveld are mainly siltstones and lavas, and thus the zircons they contain tend to be small and difficult to study; nevertheless those zircons found support Poldervaart and von Backström's (1949, p.464) statement that "the zircons of the Kaaien beds are generally rounded, and have a prevalent elongation index 1.6 - 1.8". The majority of the zircons found in the Hybrid Rocks and the Adamellitic Gneiss are similar in shape and size to those found in the Kheis Supracrustal Rocks. The small size and rarity of zircons in the Adamellitic Gneiss is well displayed in thin-sections as biotite is a common mineral but pleochroic haloes are rare and when found they generally inclose minute zircon crystals. It appears that locally in the south-eastern Richtersveld fine grained supracrustal rocks dominate and this has resulted in small zircon crystals being found in both the supracrustal material, and in the later hybrid and granitic rocks that replaced them. In contrast to these findings the zircons found in the rocks of the later Richtersveld Suite are generally larger and euhedral. The evidence gleaned from an examination of the shape and size of zircons in the rocks of the south-eastern Richtersveld suggests a non-magnetic origin for the Adamellitic Gneiss and the Hybrid Rocks.

The metamorphism of the Kheis supracrustal Rocks, the development of the Transitional Acid Hybrid Rocks, and the evolution of the Adamellitic Gneiss, are all believed to be part of a single general process. The process that produced these three rock groups was probably analogous to that proposed by Read (1957, p.40) in his description of a typical area that had been subjected to granitization; viz.

"out from the central theatre of granitization there pass waves of metasomatic solutions, changing in composition and in temperature as they become more distant from the core and promoting thereby the formation of zones of metamorphism about it."

The Kheis Meta-Supracrustal Rocks are situated furthest from the central theatre of granitization and are considered to be primarily the products of increased temperature-pressure conditions. Nearer the gneissic core these rocks were permeated by granitic solutions, as evidenced by the appearance of large porphyroblasts of K-felspar of up to an inch in diameter (specimen 312), and the Transitional Acid Hybrid Rocks developed. This is thus a zone of permeation and static granitization (Jansen, 1956, p.172). These hybrid rocks in their turn grade into banded gneisses in which at least part of the rock was mobile during its formation. The core of the gneissic zone contains a more homogeneous gneiss, in which only the nebulous remnants of older formations occur. At the time of its origin this core gneiss probably became more mobile than that of the rocks of the outer zones, and by some process, perhaps kneading, developed its present homogeneity. If, as is suggested above, the Adamellitic Gneiss owes its genesis to a granitization process, then it might be asked whether the Kheis Supracrustal Rocks studied are sufficiently basic in bulk composition to have produced the Adamellitic Gneiss. It is the writer's belief that the rocks studied are probably not sufficiently basic, but that in the production of the Adamellitic Gneiss as found in the south-eastern Richtersveld large volumes of material from the more mafic Marydale Series, and possibly also quantities of the unknown material upon which the Marydale Series rested, were granitized; and that the bulk composition of all this material was sufficiently basic to yield the Adamellitic Gneiss as a granitization product.

In some localities, as for example in the area to the west of Rooiberg 2, and more particularly north and south of the area where the Adamellitic Gneiss outcrops more extensively, the more homogeneous core gneiss was clearly mobile at some stage during its evolution; and these mobile parts appear to have been emplaced by syntectonic permissive injection. In those areas where the core gneiss has become mobile it has tended to produce extremely complex field relationships between the Adamellitic Gneiss, the

Hybrid Rocks and the Kheis Supracrustal Rocks because metamorphic and metasomatic effects have tended to overlap. In terms of Read's (1949, p.143-151) Granite Series most of these gneissic rocks are autochthonous granites, and they were formed essentially in situ; but the central core is believed to have become mobile and developed some of the characteristics of the second member of Read's Granite Series - the parautochthonous granites. Read (1957, p.364.) has in fact stated that

"the old granite of the African continent, displayed so magnificently in Namaqualand, the Eastern Transvaal and the Rhodesias is in my view an autochthonous granite."

He also stated that many of these older granites or gneisses were produced (p.364) by felspathization and permeation with the "blowing-out" of the schistose textures into the gneissose.

The banding found in some of the banded gneisses departs from the north-south foliation so typical of most of the metamorphosed and ultrametamorphosed Kheis Supracrustal Rocks of the area. These aberrant gneisses are found to be folded in a complicated plastic style that has resulted in ptygmatic folding, and this gives the rock the overall appearance of a small folded gneiss (Berthelsen, 1960, p.70. in Sorensen ed. Symposium on migmatite nomenclature.). It is believed that as in the classic migmatite areas of the southern Fennoscandian Precambrian Shield, this wild folding was produced by a large scale swelling of the gneissic and migmatitic rocks due to the introduction of K,Na and Si from activated deeper parts of the crust (Ref. Barth standard cell calculations for these rocks). This increase in volume was probably magnified by volumetric increased produced by fusion (Rich 1951, p.1209).

The banding in the banded gneiss is believed to be mainly the result of the permeating material being preferentially channelled along certain layers - generally parallel to earlier foliation and lamination - in response to favourable chemical and/or physical conditions existing in these layers. This process frequently results in the developing crystals growing so that their direction of greatest crystal growth agrees with old planes of weakness. Differential melting and localized "lit par lit" injection is also believed to have assisted in producing the banding so characteristic of many of the gneissic rocks. Read (1957, p.353) stated, after his visit to Namaqualand, that the biotite-rich streaks and bands found in the Namaqualand gneiss are "reasonably interpreted as basic behinds, the quartzo-felspathic ingredients of the original country rocks have become incorporated in the granitic layers, leaving the residual country-rock relatively enriched in ferromagnesian components".

(v) <u>Resumé</u>: The Adamellitic Gneiss is the oldest granitic rock found in the area, and it forms part of the Namaqualand Bushmanland "granite-gneiss massif". Both petrographic and chemical features indicate that it falls within the adamellite rock group. If it and the Transitional Acid Hybrid Rocks are plotted on a Green and Poldervaart diagram they are both found to fall outside the "normal igneous trend". The Adamellitic Gneiss has a normal K/Rb ratio. This is of interest for comparison with the rocks of the Richtersveld Suite which are thought to have been derived from the Adamellitic Gneiss by partial fusion. In most parts of the area the banding in the Adamellitic Gneiss is parallel to the foliation found in the hybrid rocks and the schistosity of the Kheis schists, but in a few localities, particularly in the south-east of the area, ptygmatic folding occurs. The banding found in much of the Adamellitic Gneiss resulted from granitizing fluids being preferentially channelled along layers in which favourable chemical and/or physical conditions existed. The ptygmatic folding was produced by an irregular swelling of the gneissic and migmatitic rocks upon the introduction of granitizing fluids. The metamorphism of the Kheis Supracrustal Rocks, the development of the hybrid rocks and the evolution of the Adamellitic Gneiss, are all part of a single general process that operated in the Upper Catazone. The Kheis Supracrustal Rocks of the outer zone were subjected to increased temperature-pressure conditions, the hybrid rocks were permeated by granitic solutions, and the Adamellitic Gneiss developed in the central theatre of granitization by the blowing out of the schistose texture of the Kheis rocks into the gneissose. In some areas the local mobilization of the central gneissic core, aided by variations in the susceptability of the country rocks to granitization, produced an overlapping of metamorphic and metasomatic effects which resulted in variations in the normal sequence of changes found as one moves out from the central theatre of granitization.

IV POST-KHEIS ULTRAMAFICS AND GRAPHIC GRANITE

(A) Introduction:

There are 23 Post-Kheis ultramafic bodies in the southern half of the area, and 17 of these bodies were discovered during the present investigation. These ultramafic bodies, called mafic intrusives by de Villiers and Söhnge (1959, pp.45-47), are also found to the immediate south and east of the area; and similar rocks have been discovered by Haughton and Frommurze (1936, p.14-15) in the Warmbad district, and by Gevers and others (1937, p.29) in the Vioolsdrif-Goodhouse area to the east of the Neint Nababeep Plateau. They tend to be elliptical in outcrop pattern and those found in the area range in mean diameter from under 20 yards (18.3 m.) to over half a mile (0.8 Kilometres).

Among the ultramafics and associated rocks found to the east of the Neint Nababeep Plateau, Gevers and others (1937, p.29) mention finding peridotites, hornblendites, quartz hornblende diorites and a very dark and highly mafic hornblende diorite which is often quartz-bearing. This list of rock types includes all the main rock types found in the post - Kheis ultramafic bodies of the area, except that the peridotites of the area have been converted into serpentinites. Gevers and others (1937, p.29) also state that "these rocks are mostly found as remnants of varying size and irregular shape floating in and intruded by the more widely distributed biotite granites and biotite-hornblende granites and associated granodiorites and quartz diorites". This statement is also true of the south-eastern Richtersveld as these ultramafics are seen to cut across the Kheis supracrustal rocks, but the hybrid rocks and the Adamellitic Gneiss appear to have formed after their emplacement. With regard to the distribution of these ultramafic bodies, de Villiers and Söhnge (1959, p.48) state that

"there is some suggestion of an east-west line immediately north of Rooiberg 2 dome and again from the south-eastern edge of this dome eastwards towards Cone trigonometrical station".

The writer agrees with this statement, as (1) most of the newly mapped ultramafic bodies occur along an extension of the southernmost line, and (2) the long axis of some of these bodies is approximately east-west. Gevers and others (1937) and de Villiers and Söhnge (1959, p.46.) believe that these ultramafics were emplaced into the Kheis supracrustal rocks before the emplacement of the Adamellitic Gneiss, and that the vague alignment in the distribution of these bodies probably indicates that the ultramafic magma worked its way up in weakened zones of the earth's crust in irregular dyke like fashion.

Field and petrographic studies reveal that the ultramafic rocks of the area can be divided into two main groups - viz. Serpentinites and Hornblendites. The hornblendites are more numerous than the serpentinites but as both types outcrop along both the northern and southern lines, one is unable to find any definite separate trend in their distributions, and it would thus seem that both rock types were emplaced broadly contemporaneously along the same lines of weakness within the crust.

The large ultramafic body to the immediate south of the area is aberrant and has the following mean modal composition:- hornblende = 36.6%, chlorite = 20.3%; plagioclase = 13.5%, epidote = 12.0%; white mica = 11.2%; quartz = 2.7%; serpentine = 1.6%; zoisite = 1.5%; apatite = 0.2%; augite = 0.2%; calcite = 0.2%; biotite = Tr., and opaque ore minerals = Tr. This ultramafic body also contains a few irregular patches and segregations that differ considerably from the "normal" rock type : viz. (1) medium grained hornblendite patches (specimen 436), (2) irregular lenses of a zoisite-hornblende rock, and (3) segregations containing large ($2\frac{1}{2}$ cms. long) euhedral hornblende crystals.

Another interesting ultramafic body that lies just outside the area is the small hornblendite occurrence situated below and to the west of Cone beacon (see Map 2). Most of this body is composed of normal hornblendite with a mean hornblende content of over 90% (ie. specimens 483 and 488), but located within this material are veins of cross - and mass-fibre asbestos (specimen 499). When viewed under the microscope the asbestos bands are found to be composed of calcite, asbestiform-amphibole and a little serpentine, and the asbestiform amphiboles are seen to grade into the more tabular amphiboles of the host rock. The asbestos fibres were found to be up to 10 cms. long, elastic and variable in colour from colourless to light yellow brown. Each of the thin strands of asbestiform amphibole was found to consist of aggregates of smaller fibres, each of which was elongated parallel to the general fibre axis. There was no preferred orientation of the individual fibres in the bundles, thus the optical properties of individual crystals could not be determined. Vermaas (1952, p.199) has stated that because of this aggregate structure distinction is difficult or impossible optically, between the different varieties of asbestiform minerals, especially between varieties of amphibole asbestos. The mean refractive index for vibrations parallel to the fibre axis of the asbestiform amphibole from the Cone ultramafic body (see Map 2) was found to be in the tremolite range. As much of the rock of the Cone ultramafic body is sheared, it seems likely that in those areas where the shearing was most pronounced the tabular amphiboles became unstable and the asbestiform variety developed. Vermaas (1952, p.229) has stated that such changes can take place without any change in chemical composition as the double silicon-oxygen chains of the amphibole structure lend themselves to 'deformation', and long fibres are formed without a collapse of the structure.

(B) <u>The Hornblendites</u>:

(i) <u>Petrography</u>: The mean mode and the standard deviations of the major constituent minerals of 14 fresh specimens of the hornblendite taken from all over the area are given below:-

	Amphibole	Chlorite	Epidote	Plagioclase	White Mica	Quartz	Biotite
$\overline{X}$	92.9	2.9	1.6	1.1	0.9	0.5	0.1
s	4.2	1.9	2.4	1.1	-	-	-

	Opaque Ore Minerals	Zoisite	Serpentite	Calcite	Apatite	Clino-pyroxene
$\overline{X}$	Tr	Tr	Tr	Tr	Tr	Tr
s	-	-	-	-	-	-

The colour of these fresh specimens is generally medium dark grey (N4) to dark greenish grey (5G 4/1), but on weathered surfaces the rock is a moderate reddish brown (10R 4/6). The texture of the hornblendite varies from granitic to poikilitic. Grain-size is also variable as some large skeletal oikocrysts have a thin-section area of 45 sq.mm, but the mean thin-section area appears to be approximately 0.18 sq. mm. The dominant amphibole occurs in short anhedral to subhedral laths and has the following optical properties:-

$Z = 1.640$, Dusky yellow green (5Y 5/2) to moderate blue green (5 BG 6/6)

$Y = 1.628$, Greyish yellow green (5 GY/ 7/2)

$X = 1.615$, Pale yellow green (5 GY 8/4)

$2Vx$ ($\bar{X}$ of 12) $= 81^{\circ}$

$Z \wedge c = 16^{\circ}$.

Twinning and alteration to chlorite and epidote is common.

The chlorite found in the hornblendites tends to occur in elongated plates which are closely associated with the amphibole. Minerals of the epidote group, mainly pistacite, zoisite and clinozoisite, have an irregular distribution throughout this rock group. They appear to be mainly the result of the alteration of amphibole and plagioclase in those few specimens that contain this latter mineral. In some rock specimens, however, aggregates of fairly large epidote crystals are found which are unrelated in origin to any alteration process in their immediate vicinity. Many of the Hornblendite bodies are crossed by veins of late epidote. The plagioclase found in the hornblendites is invariably highly altered to epidote, zoisite and sericite, and some of the plagioclase laths are found to be rimmed by large epidote crystals. Crystals of clear quartz are found in some specimens. This mineral tends to occur either interstitially or as late cross-cutting veins. In some specimens quartz is found to occur as minute grains associated with chlorite, epidote and calcite, all of which appear to be the alteration products of amphibole. Small irregular granules of opaque ore minerals are found scattered throughout most of the specimens studied. A little biotite is found in a few specimens (i.e. specimen 544) and it is pleochroic from X = pale yellowish orange (10 YR 8/4) to Z = moderate yellowish brown (10 YR 5/4). A few rock specimens contain small quantities of a clino-pyroxene, probably angite, which is found to be altering to amphibole. The calcite found in the hornblendites occurs mainly in minute irregular grains associated with secondary minerals such as epidote and zoisite, but it is also found interstitially between the amphibole crystals. It is of interest to note that within the scree on and about the hornblendite bodies, nodules (often ten centimetres across) of white calcite are frequently found.

(ii) Chemistry: A partial chemical analysis was made of specimen 334 (from 1 E 6.) which is a hornblendite that contains virtually only amphibole, and it was found to contain 7.59% total iron as FeO, 15.5% MgO, 0.15% MnO, 0.61% Na_2O, 0.11% K_2O, 155 p.p.m Ni and 69 p.p.m. Co. This partial analysis clearly shows (1) that the amphibole found in the hornblendite is not a soda amphibole, (2) that both Mg and Fe are present in the amphibole and that Mg is dominant, and (3) that the amphibole is not actinolite as it would require approximately 5% more (Fe, Mg) O to produce this mineral phase. The chemical evidence thus indicates that the amphibole in specimen 334 is intermediate in composition between actinolite $Ca_2\ (Mg, Fe)_5\ (OH)_2\ Si_8O_{22}$ and tschermakite $Ca_2\ Mg_3\ Al_2\ (OH)_2\ Si_6\ Al_2\ O_{22}$. If the above partial analysis is compared with Nockold's (1954) average hornblendite it is found to have a high MgO content, and to be low in total Fe, MnO, Na_2O and K_2O. The Co and Ni content of specimen 334 is much less than the amount found in Turekian and Wedepohl's

(1961) average ultrabasic rock but the values are of the same order of magnitude as their average basalt, (i.e. Turekian and Wedepohl's (1961) ultrabasic contains 2000 p.p.m. Ni and 150 p.p.m. Co, and their basalt contains 130 p.p.m. Ni and 48 p.p.m. Co.). As can be seen from column 4, of Table VI the major element content of Nockolds' (1954) average Tholeiitic Olivine Basalt is also very similar to that of the Hornblendite analysed. This approximate equivalence of hornblendite with basalt has been known for a long time - viz. Lacroix, 1917, p.974 - and will be considered in detail in the section on petrogenesis.

(C) The Serpentinites:

(i) Field Description: Six serpentinite bodies are found in the area; four outcrop to the north-west of Rooiberg 2 (1D9, 1E3 and 1E5), and the other two occur in the south-east of the area (7G3) near the old disused Klein Helskloof track. These bodies, which are circular or eliptical in outline, tend to be more regular in shape than the hornblendite bodies.

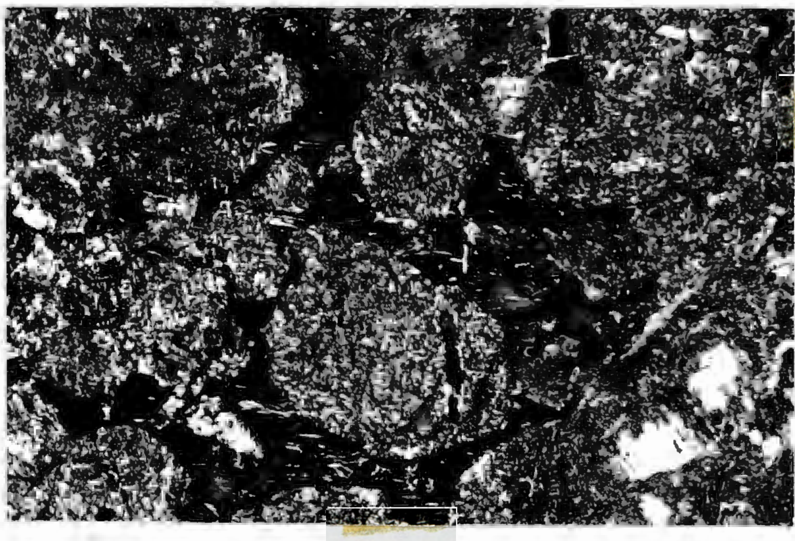
(ii) Petrography: The mean mode and the standard deviations of the major constituent minerals of 11 specimens of the serpentinites (taken from all six serpentinite outcrops) was found to be as follows:-

	Serpentine	Amphibole	Chlorite	Opaque Ore Minerals	White Mica
$\overline{X}$	64.0	11.8	7.8	6.7	5.1
s	11.7	12.3	7.1	3.3	5.5

	Carbonate Minerals	Altered felspar	Brucite	Apatite
$\overline{X}$	4.6	Tr.	Tr.	Tr.
s	4.3	-	-	-

The colour of the specimens is dark grey (N3) with a lustre-mottled appearance on freshly broken surfaces. Weathered surfaces tend to be reddish brown (IOR 4/4). The serpentinites have a mesh texture that probably indicate that the serpentine is derived from olivine (see Photomicrograph 4.). The mean thin-section area of the equidimensional serpentine patches that are believed to be pseudomorphs after olivine is 2.57 sq.m.m.

The serpentine is mainly antigorite and occurs as fine fibrolamellar aggregates. It varies in colour from colourless to pale yellow green (5GY 8/4). Some of the equidimensional pseudomorphs mentioned above contain narrow, generally arcuate, veinlets of opaque ore minerals which seem to trace the course of earlier irregular fractures within the original olivine crystals. The amphibole associated with the serpentine is mainly colourless to very pale green tremolite and it occurs between the equidimensional patches of serpentine. Specimen 326 contains amphibole crystals that enclose minute granules of ore that occur in thin bands or veinlets that intersect one another at approximately 87° and look like the schiller structure characteristic of bronzite. This observation would seem to indicate that at least some of the amphibole is an alteration product of pyroxene. The chlorite found in the serpentinites is predominantly penninite and it tends to show the characteristic anomalous "Berlin blue" interference colour of this mineral species. Some of the chlorite minerals display an anomalous yellowish grey (5X7/2) to dark yellowish orange (10YR 6/6) interference colour; and possess the unusual pleochroism of x = pale green (10G 7/2), Y = Z = moderate orange pink 5YR 8/4). The opaque ore minerals are mainly magnetite, chromite, and ilmenite, and they are found abundantly scattered throughout these rocks.



Photomicrograph 4: Serpentinite (showing antigorite pseudomorphs after olivine ?), x 14, Specimen No. 324, from north west of Rooiberg 2 (i.e. 2.E.I.), X-nicols.

(iii) Chemistry: In table VI the chemistry of a typical Richtersveld serpentinite is compared with the mean of seven serpentinites from Southern Rhodesia because tables containing the average chemical composition of serpentinites could not be found. This comparison revealed that the Richtersveld serpentinite had a significantly high Fe_2O_3, FeO, CaO, CO_2, P_2O_5 and TiO_2 content, was slightly high in SiO_2, Al_2O_3 and MnO, slightly low in H_2O^+ and significantly low in MgO and Na_2O. The high Fe_2O_3, CaO and CO_2 values can be accounted for when allowance is made for Cr_2O_3 (which was not determined) and calcite. The Mg/Fe molecular ratio of specimen 325 is 6.1, and thus, according to Hess (1938, p.341) the serpentinite was probably a crystallization differentiation product of a mafic magma, and was not derived from a magma of the ultramafic magma series.

(D.) Petrogenesis:

Bowen and Tuttle's (1949, pp.439-460) investigation of the system $MgO-SiO_2-H_2O$ showed (p.440) that "pure magnesian serpentine has a maximum temperature of existence of approximately 500°C, varying about 10° in the whole range of pressure 2000 to 40,000 lbs./in^2." These findings left little if any theoretical basis for a belief in the emplacement of a liquid of the composition of olivine and water, or of serpentine, which was unable to effect significant contact metamorphism on the surrounding country rocks. At the present time the majority of petrologists seem to favour Bowen and Tuttle's (1949) suggestion that the alpine serpentinites were emplaced as a mush of olivine, or olivine and enstatite crystals separated by a little interstitial fluid; and that serpentinization was essentially a post-emplacement process. Bailey (1958, p.31) has suggested that the fluid or flux in the crystal mush may have been allied to the magma that has given us carbonatites. The common occurrence of calcite veins and nodules within the Richtersveld ultramafics would seem to support this hypothesis. Bowen and Tuttle (1949, p.453) have also made the interesting suggestion that olivine aggregates are more capable of flow when crystalline than other common anhydrous rock-forming minerals because the crystals are built up of SiO_4 groups without any chain, sheet or space linkage. In support of the crystal mush hypothesis Thayer (1960, p.256) has observed that in order to carry up dense chromite bodies, which are so often found in these ultramafic rocks, and to prevent them from sinking back into the depths, the magma must have been mushy from the very beginning. Francis (1956, p.216) believes that the pod-like shape and the sheared fabric of ultramafic bodies also supports the olivine-mush magma hypothesis.

The temperature at which the olivine-mush magma was emplaced appears to depend on its initial temperature when mobilised and the rate of its movement as the friction effects attendant upon its motion will necessarily be a source of heat (Bowen and Tuttle, 1949, p.453). When movement ceased cooling would be expected to bring the crystals and the vapour phase into the stability field of serpentine, and the serpentinization process would commence. The temperature at which serpentinization began would probably be below 500°C as serpentine would become stable at about this temperature if silica and/or carbon dioxide were present in the vapour phase. As Chattezie (1955, p.101) has observed, opinions were at one time sharply divided as to the origin of serpentinizing solutions. Some writers (Benson, 1918, p.693 and Hess, 1933, p.652.) believed that the solutions represent residual liquids of the ultramafic magma - Hess' autometamorphic - while yet others suggested that the solutions may have been derived from younger acid intrusives (Du Reitz, 1935, p.233). But since the publication of Bowen and Tuttle's (1949) experimental study most writers, including Hess (1955), have accepted Bowen and Tuttle's statement that serpentinization is not an autometamorphic process but tends to be the work of solutions of extraneous origin which have probably reheated the peridotite. It seems likely that serpentinization was brought about in the ultramafics of the area by granitic solutions, or by a process similar to that which Szadeczky-Kardoss (1960, p.262) has called "transvaporization" - that is, the selective migration of volatiles from the relatively wet country rocks into the immobile olivine-rich body, when the partial pressure of the mobile components in the country rock exceeded that of the mobile components in the olivine-rich body. Hess (1938 , p.331) suggested a somewhat similar process.

With regard to the source of the olivine-mush magma, Hess (1955) believes that it is derived from a primary peridotite magma. The evidence from layered intrusions clearly indicates that rocks of dunitic and peridotitic composition can be formed through the action of the processes of segregation and differentiation, and Wager (1958, p.36) has stated that under suitable tectonic conditions these autochthonous olivine cumulates may become mobile. It is also of interest to note that recently, Davis and others (1956) have studied the isotopic composition of lead extracted from olivine nodules found in basalts and have discovered that the isotopic ratios were those of young lead which suggests that the particular olivine nodules investigated were formed in relatively recent times by differentiation from basalt and are not fragments of a primary earth shell.

Table VI - Chemical Data (ultramafics)

	Composition, weight per cent: (1)	(2)	(3)	(4)		Norms: (5)
SiO_2	38.50	33.83	36.59	47.90	Or	1.11
Al_2O_3	2.76	2.14	2.10	11.84	An	6.95
Fe_2O_3	7.96	7.96	4.56	2.32	Hy ($MgOSiO_3$	30.85
FeO	5.47	4.31	4.60	9.80	($FeOSiO_3$	1.26
MgO	32.68	32.03	39.45	14.07	Ol ($2MgOSiO_2$	35.56
CaO	1.57	1.30	0.66	9.29	($2FeOSiO_2$	1.63
Na_2O	0.03	0.00	0.25	1.66	Mt	11.60
K_2O	0.16	0.17	0.14	0.54	Il	0.61
H_2O^+	7.84	11.62	8.95	0.59	Ap	0.34
H_2O^-	0.12	1.06	0.71		CO_2	2.89
CO_2	2.89	0.56	0.44		H_2O^+	7.84
P_2O_5	0.07		0.02	0.19	H_2O_-	0.12
TiO_2	0.29		0.06	1.65		100.76
CrO_3			1.51			
MnO	0.20	0.02	0.10	0.15		
NiO			0.09			
Cu			0.01			
	100.54	100.00	100.24	100.00		

The emplacement of ultramafic bodies also raises a number of problems as it is clear from a study of their densities that such material could not have risen passively through the crust as a result of differences in specific gravity between the ultramafic material and the country rocks.

(1) Specimen 325; Serpentinite from the circular ultramafic body to the north west of Rooiberg 2. Mode = serpentinite 84.1%, chlorite 7.2%, amphibole 3.0%, Opaque ore minerals 3.3%, White mica 1.3%, Calcite 1.1%, Altered felspar Tr. (Analyst - E.C. Haumann.).

(2) Composition of the above specimen (325) calculated from its mode, using 1 $FeO. 4MgO. Al_2O_3. 3SiO_2. 4H_2O$ for the chlorite and Hess' (1960, p.67) "pure serpentine from a vein" for serpentine.

(3) Mean of 7 serpentinites taken from Worst (1958, p.292).

(4) Nockold's, 1954, p.1021, No. VIII Average Tholeiitic Olivine Basalt (av. of 28).

(5) Norm of specimen 325 (i.e. Column 1).

Bowen and Tuttle (1949,p. 453) speak of an olivine-mush magma moving slowly under deformative forces, and Hess (1955,p.395) speaks of alpine serpentines being intruded during the first great deformation of a given mountain system. It is clear that external forces such as those that occur during orogenesis must be invoked to account for the upward movement of a dense relatively cool, quasi-solid ultramafic mass. Most recent writers on the emplacement of ultramafic rocks seem to believe that the ultramafic magma was probably squeezed up along major fracture zones during the onset of folding. In some areas, as for example the coastal ranges of California (Taliaferro, 1943, pp. 159-182) and perhaps also in the Richtersveld, these lines of ultramafic bodies have become the locus of postorogenic faulting, and the ultramafic bodies along the faults have been mobilized for a second time and squeezed up the fault planes as cold intrusions during the post-orogenic phase of the orogenic cycle. Thus to apply this suggestion to the ultramafics of the area, these rocks probably formed at an early stage in the Kheis orogenic cycle and were later emplaced by cold intrusion into their present positions. Read (1957, p.363) has stated that serpentinites emplaced into sandstones "can reasonably be interpreted as tectonic pips squeezed up from the depths" and that " ... these pips may be located".. " in special movement-zones related to their upward migration". The texture of the serpentinites of the area would seem to indicate that solid flow preceded serpentinization. However, as the change of olivine to serpentine involves a volume increase of approximately 25% (Hess, 1955, p.403), and the texture indicates that the serpentinization process was a volume-for-volume exchange, there appears to have been a removal of a great mass of superfluous material. The source of the serpentinizing solutions in the area is not difficult to find as the serpentinite bodies are in close proximity to the Adamellitic Gneiss and its associated migmatites. A review of recently postulated origins for serpentinites would be incomplete without some mention of van Bemmelen's (1940, p.246) suggestion that the ophiolitic suite including the alpine serpentinites are a regional low-level basic front representing a geochemical culmination of cafemic constituents in front of acidification and migmatization of the crustal base. Tyrrell (1955, p.422) has objected to this interpretation "for the sufficient reason that, in all the tectoigneous cycles yet investigated, the ophiolitic phase is separated by a long period of quiescence from the granitization process".

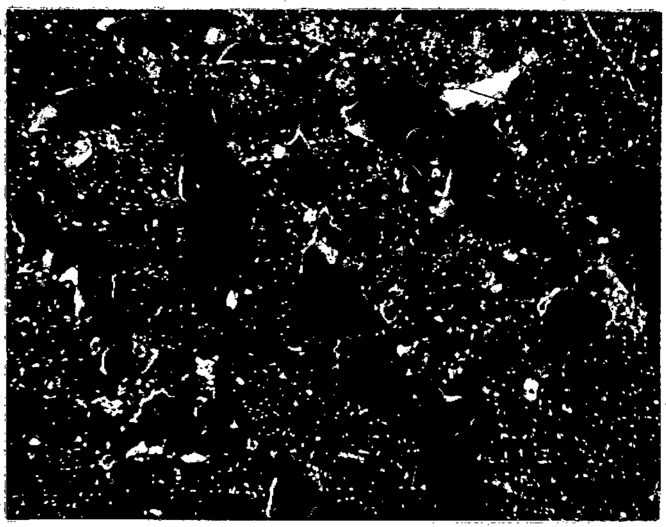
The problem of the genesis of the hornblendites found in the area studied is in many ways similar to that of the serpentinites as they appear to have been emplaced contemporaneously along the same lines of weakness within the earth's crust. The situation in the area might have been similar to that postulated by Noble and Taylor (1960, p.96); that is, the more mobile magma that formed the hornblendites may have been emplaced first, and thus the passage of the later olivine mush magma was assisted by the presence of preheated relatively unimpeded pipelike channels. Recent experimental investigations into the origin of basalt magmas (Yoder and Tilley, 1956, pp. 169-171: 1961, pp. 106-113: and 1962, pp. 455-459) have yielded a considerable amount of new information which is most relevant to any study of the genesis of hornblendites. In their report on the natural tholeiite basalt-water system, Yoder and Tilley (1956, pp. 169-171), state that the stable phases changed drastically and discontinuously with changes in water pressure, and that on quenching the melt at 500 bars and 1125°C the result was a mass that "consisted almost wholly of an amphibole, with small amounts of magnetite and some glass" (p.169). Quenching the melt at a pressure of under 500 bars water pressure at 750°C converted the charge into an assemblage consisting of amphibole, plagioclase, sphene and magnetite with traces of glass (p.170.). They concluded from their study of tholeiite that when such material crystallized as tholeiitic basalt at the surface or gabbro at depth, it must have had a low water content as, "if water had been present and retained, the magma presumably would have crystallized as hornblende or amphibolite" (p.171.). In a more recent paper (Yoder and Tilley, 1962, p.459) they state that in their experiments all the major types of basalt were "metamorphosed" to amphibolite at relatively low temperatures in the presence of water. This experimental study shows that while hornblendite has no true volcanic equivalent, unique processes do not have to be invoked to account for its genesis. It thus seems likely that the hornblendites of the area were emplaced as gabbroic bodies (to which they are chemically similar) and were later transformed during the period of granitization associated with the emplacement of the Adamellitic Gneiss, to the hornblendites found at the present time. This transformation of gabbro into hornblendite involves an addition of water and Yoder and Tilley's (1962, p.470) investigation indicates that this results in a volume increase that is probably slightly in excess of 2%.

(E.) <u>Resumé</u>:

Twenty-three ultramafic bodies of various sizes are found in the area, and seventeen were discovered during the present study. Petrographic and chemical data indicate that they all belong to one of two rock types - the hornblendites and the serpentinites. The hornblendites typically carry over 90% amphibole in the actinolite - tschermankite range, and they are found to be chemically akin to tholeiitic olivine basalts. Bands within some of the hornblendite bodies along which shearing has been particularly strong contain asbestiform instead of the more usual tabular amphiboles. Six serpentinite bodies are found in the area. These rocks generally have a mesh texture and generally contain over 60% antigorite. The serpentinites are believed to have been originally emplaced as an olivine-mush magma that originated at depth, probably early in the Kheis orogenic cycle, as a differentiation product of a mafic magma. The hornblendite is believed to have been originally emplaced as gabbro. As it is clear that some external force must be invoked to account for the upward movement of these dense relatively cool quasi-solid ultramafic magmas, it is suggested that the ultramafic bodies were emplaced along major fractures during the Kheis orogeny and that they were squeezed up into their present positions as "tectonic pips". Later, with the evolution of the Adamellitic Gneiss, the peridotite (olivine-mush) and the gabbro pips were converted into serpentinite and hornblendite respectively.

(F) Graphic Granite:

(i) Introduction: The Post-Kheis ultramafic bodies frequently contain lenses and dyke-like bodies of graphic granite. The largest of the hornblendite bodies to the north-west of Rooiberg 2 is of particular note as it contains a composite dyke-like body composed of a six inch (15 cm.) vein of white milky quartz with bands of graphic granite on either side of it.



Photomicrograph 5: Graphic Granite found within Hornblendite body northwest of Rooiberg 2 (1D9), x 13.5, Specimen No. 344, X-nicols.

(ii) Petrography: The mean modal composition, and the standard deviations of the major constitutent minerals of 6 of these graphic granites is as follows:-

	Perthite	Plagio-clase	Quartz	White Mica	Tour-maline	Calcite	Chlorite
$\bar{x}$	63.2	1.8	34.1	0.4	0.3	0.1	0.1
S	3.9	1.6	3.7	-	-	-	-

On compositional grounds the graphic granites thus fall into Peterson's (1961) Alaskite category and Johannsen's (1932) Kalialaskite or Runite group. As can be seen from their standard deviations the percentages of alkalic felspar and quartz present in this rock type are relatively stable. The relative abundance of the minor mineral species is not as stable, as tourmaline, for example, is found in only one specimen (435), and the white mica content varies greatly. Perthitic felspar (mainly microcline microperthite) is the dominant mineral and it generally forms large crystals in which the quartz occurs as graphic intergrowths. The perthite varies from micro-to vein-perthite, but in some specimens the felspar has the chess-board structure (Anderson, 1937, p.62) that is generally considered to originate in crystals where orthoclase is being replaced by albite. When studied in polarized light the individual angular quartz rods or ichthyolypts within one of these large felspar crystals extinguish at the same time, thus showing that the ichthyoglypts all tend to have a common orientation. Both the felspar crystals and the ichthyoglypts in some specimens are crossed by later quartz veins, but these veins can easily be distinguished as most of them extinguish at a different position to the host ichthyoglypts. The tourmaline in specimen 435 is pleochroic from E = very pale green (10G 8/2) to O = greyish green (5G 5/2), and it generally occurs in irregular anhedral crystals.

(iii) Petrogenesis: Wahlstrom (1939, p. 682.) has stated that the occurrence of graphic granite "serves as a well established criterion for the determination of a rock as pegmatitie;" and it appears that Hauy originally used the term pegmatite as a name for a rock that would today unquestionably be described as a graphic granite. These statements must clearly receive attention when considering the origin of the graphic granites. It is also of interest to note that a number of authors (Holmes 1918, Makinen 1917, Vogt 1931, Wahlstrom 1939 and Spencer 1945) have studied the ratio of quartz to felspar in graphic granites, and most of them have concluded that the ratio tends to be constant. Makinen's (1917) study of the graphic granites of Finland convinced him that the free quartz content of these rocks ranged from 23-32%. Holmes (1918) obtained a quartz range of 24.2 - 27.9% for the graphic granites of Mozambique. Vogt (1930) believed that the amount of quartz in microcline graphic granite is approximately 26%, and is confined within limits ranging from 23% to 28%. He did, however, observe that the amount of quartz in oligoclase graphic granite falls within the limits 33% to 38%. Other investigators (i.e. Wahlstrom, 1939) have, however, questioned the limited range of composition ascribed to these graphic granites, and in so doing have challenged the belief that this texture is the product of simultaneous crystallization. As can be seen from the mean mode the quartz content of the graphic granites of the south-eastern Richterveld is too high to fall within the "ideal" limits.

An interesting feature of the graphic granites of the area is their apparent localization in the ultrabasic bodies, particularly the hornblendites. A recent paper by Olsen

(1961,pp.329-347.) on "high temperature acid rocks associated with serpentinite" has shown that the occurrence of felsic bodies within ultramafic rocks is much more common than formerly believed. The study by Andersen (1931, pp.47-48.) of the relative number of pegmatite bodies per unit area of rock in a 2,000 sq. km. area of Precambrian rocks, is also of great significance as he found that the gabbroic and amphibolitic rocks contained 87% of the pegmatites, the quartzites 7%, the gneiss 4% and the granites 2%. To account for the dominance of pegmatites in gabbroic and amphibolitic country rocks Andersen (1931) suggested that their elastic properties differ radically from those of the other rocks and thus they fracture more readily. Perhaps the most significant physical property of the amphibolites is their large thermal contraction. This property may well have played a significant role in localizing the graphic granites in the ultramafic rocks of the area. It is possible that with the formation of the Adamellitic Gneiss and its associated hybrid rocks the ultramafic rocks were transformed into serpentinites and hornblendites, and the temperature of the whole crustal segment in which these rocks formed was raised. On ccoling the ultramafic rocks contracted to a greater extent than the more acid rocks about them, and this led to the development of fractures up which the late phase pegmatitic fluids associated with the Adamellitic Gneiss could rise and be intruded. The milky quartz veins probably represent the final stage of this process.

V. <u>THE RICHTERSVELD SUITE (PLUTONIC)</u>

(A) <u>Introduction</u>: Rogers (1915, p.97) in his "Geology of part of Namaqualand" was the first to record plutonic rocks younger than the "Namaqualand Gneiss" within the south-eastern Richtersveld. He observed that, "Rooiberg (2), east of Stinkfontein, is made of massive syenite rising from the old schists and gneiss. It is a rather coarse, pink rock, with many dark patches of ferromagnesian minerals; quartz is present, but one has to look for it carefully" (p.97). From Rogers' map it would appear that he believed the rocks of Rooiberg 2 to be of the same age as those at Kuboos, while the more extensive outcrop of rocks of the Richtersveld Suite at, and to the north of Xaminxaip were mapped as "ancient gneiss and schists of sedimentary and volcanic origin".

Söhnge and De Villiers (1946, p.266.) introduced the term "Richtersveld Igneous Complex" for those Richtersveld plutonic rocks younger than their "grey gneissic granite" and its associated metamorphic rocks, yet older than the rocks of the Kuboos Igneous Complex. In their map of the Richtersveld and eastern Sperrgebiet Söhnge and de Villiers (1946, Plate 32) show two main outcrops - viz. (1) the main Richtersveld outcrop that covers an area of approximately 110 square miles and crops out along the Orange River between 28° 46' south and 28° 31' south, and (2) the large outcrop in the Nudabib - Witputs district (See map 2.) of the eastern Sperrgebiet (S.W. Africa) to the north of the Richtersveld. Smaller outcrops of plutonic rocks of the Richtersveld Suite occur (1) at Rooiberg 2 (Ref. frontispiece) within the area studied, (2) astride the Orange River approximately 2 miles north of De Hoop (See map 2), (3) south of Soeties beacon (See map 2), (4) east of the area approximately 12 miles south east of Vioolsdrif (See map 2.), (5) north-west of Nudabib and Witputs (S.W. Afr.) in the area about Aurus Waterhole (See map 2), and (6) to the immediate west of the main Richtersveld outcrop. Thus at the present level of erosion the plutonic rocks of the Richtersveld Suite are seen to form a discontinuous north, north-west trending belt that extends for 120 miles from the Soeties outcrop in the south (28^{o} 58's) to the Aurus outcrop in the north (27^{o} 34's).

De Villiers and Söhnge (1959, p.72) have suggested that rocks from the Warmbad (S.W.A.), Aus (S.W.A.), Springbok, Kamieskroon and Garies districts may also belong to the same magmatic cycle as the Richtersveld Suite. Prof. H. Martin (1963) who is at present directing a mapping project in the area to the immediate north of the Richtersveld, has stated that the only rocks that are known to be comagmatic with the Richtersveld Suite are confined to the general Richtersveld area, and that he doubts whether the outcrops of Richtersveld Suite rocks shown to the north of the Richtersveld in Söhnge and de Villiers' (1946, plate 32) map, are as extensive as shown.

The plutonic rocks of the Richtersveld Suite are clearly younger than the Adamellitic Gneiss and older than the Quartz Bostonite and Hornblende Diorite dykes. Further discussion of the age of the Richtersveld Suite is postponed until after these dyke rocks have been described.

(B) <u>Field Description</u>: The plutonic rocks of the Richtersveld Suite occupy approximately half the area studied and occur in two separate outcrops. The smaller outcrop is in the south-west of the area and is known as Rooiberg 2, after a beacon of that name situated at its summit. The second outcrop is part of the main mass of plutonic rocks of the Richtersveld Suite found in the Richtersveld which commences approximately one mile (1.6 Km.) to the north-east of the Rooiberg 2 outcrop. The Rooiberg 2 outcrop is approximately circular in plan (See Frontispiece), and has a north-south diameter of 2.3 miles (3.7 Km.) and an east-west diameter of 2.2 miles (3.5 Km). Its overall area is $3\frac{3}{4}$ sq. miles (9.7 sq. Km.). The geological map shows that Rooiberg 2 is made up of four concentric rings (from the outside inwards) A, B, C and D, and a central core E. These rings are composed of A = Granular Syenite, B = Porphyritic Syenite, C = Granular Syenite, D = Alaskitic Granite, and E = Porphyritic Microgranite. In detail, however, the picture is more complex, as minor variations in texture, grain size and even composition are found within each unit, and inclusions are common throughout the complex. The five main units that together form the Rooiberg 2 complex have tended to differ in their individual responses to the action of the agents of denudation. Units A and C which are similar coarse to medium grained syenites and unit E which has acted as a channel way for the passage of post emplacement fluids, tend to weather more rapidly than the finer grained rocks of units B and D. Thus in most localities the country rock-unit A boundary is found in relatively low lying areas, and as one proceeds inwards there is an increase in altitude until the A-B boundary is reached. Unit B generally forms the crest of a series of low arcuate hills which rim Rooiberg 2, and are most characteristically developed in the north and eastern part of the complex. Unit C forms a moat-like depression between the arcuate low hills and the main mass of Rooiberg 2. Unit D forms the main mass of Rooiberg 2 and the rocks of the summit belong to this unit. Located at the centre of the complex and of Unit D, is the core of the complex, Unit E. The rock of this central unit is more readily weathered than that of Unit D, but as it is enclosed within Unit D it has been buttressed by the more resistant rocks of Unit D, and in this manner it has been able to retain its present relatively elevated position within the complex as a whole. At a late stage in the development of the complex Unit E acted as a channel-way for the passage of fluids. These fluids have not only aided in the decomposition of Unit E, but by also permeating the rocks adjoining this unit - particularly those to the west of it - they have made these rocks more susceptible to weathering. This has resulted in a break down in the regular pattern of concentric morphological units in the west of the complex where rings B and D have been successfully breached on a large scale.

The relationship between topography and lithology so well displayed in parts of the Rooiberg 2 complex, is generally masked by superimposed drainage - particularly that of the Orange River - in the area of the main Richtersveld Suite outcrop. Even along the Orange River however, the rock analogous to the main granitic member of the Richtersveld Suite (Ring D of the Rooiberg 2 complex) forms steeper sided and generally higher walls in the Orange River canyon than the coarse to medium grained syenite which was found to weather so readily in rings A and C in the Rooiberg 2 complex.

(C.) <u>Petrography of Rooiberg 2</u>:

(i) <u>Ring A</u>: This is the outermost ring of Rooiberg 2, and in the north and east where it is best exposed its mean width is approximately 300 yards (275m). The rock type forming this ring is an allotriomorphic granular syenite. (See photomicrograph 7). In the field it is seen to consist of an aggregate of equidimensional pale red (5R 6/2) to moderate red (5R 5/4) felspar crystals, with occasional greenish black (5 GY 2/1) hornblende and other ferromagnesian crystals which give the rock a speckled appearance. The mean thin-section grain area of the anhedral felspars is 5.2 sq.mm. The subhedral, poorly terminated, hornblende crystals tend to be smaller than the felspars and have a mean thin-section grain area of approximately 2 sq.mm. The mean modal

Plate 6: Panoramic view of Rooiberg 2 as seen from the West.

composition and the standard deviations of the major constituent mineral species of 6 fresh specimens of rock from Ring A is as follows:

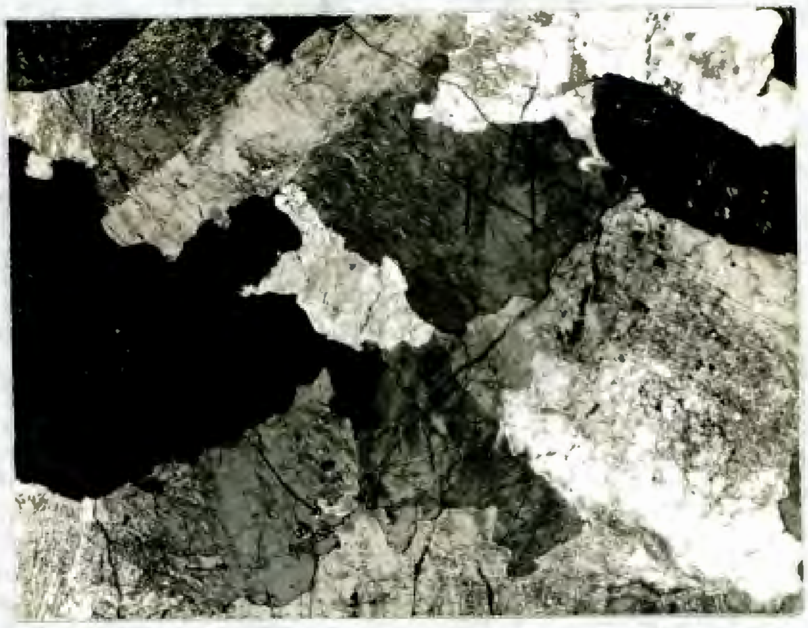
	Alkali-Felspar	Quartz	Amphi-bole.	Bio-tite	(Horn-blende & Bio-tite).	Plagio close	Zircon	Opaque Ore Minerals
$\bar{x}$	83.5	2.7	11.2	1.3	(12.5)	0.4	0.4	0.3
s	3.6	1.7	3.0	1.1	(3.0)	-	-	-

	Fluorite	Calcite	Chlorite	Apatite	Epidote	Sphene	Leucoxene	Allanite
$\bar{X}$	0.1	0.1	Tr.	Tr.	Tr.	Tr.	Tr.	Tr.
s	-	-	-	-	-	-	-	-

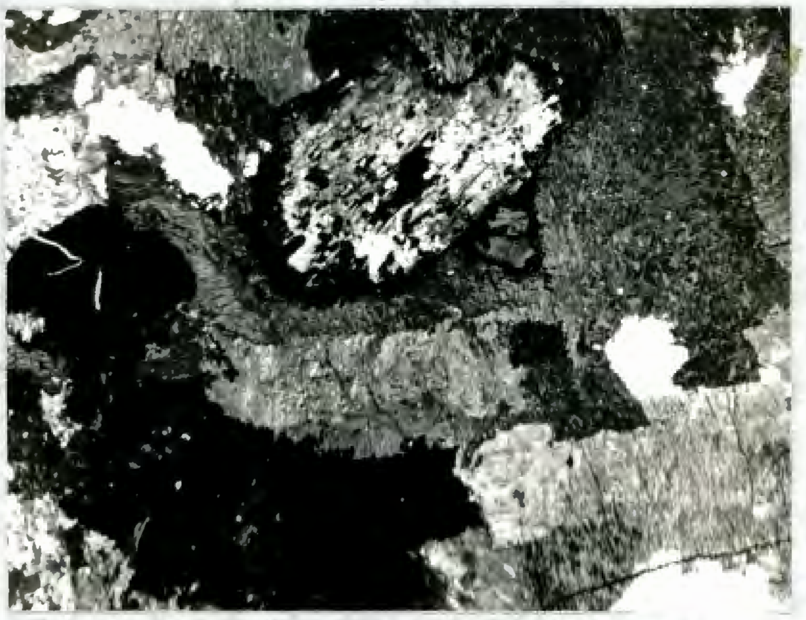
Microcline microperthite in which the K-felspar and plagioclase components are approximately equal in volume, is the characteristic felspar of ring A. Most of the perthite is of the film type, but some patch perthite is found. The 2VX's of a large number of the more homogeneous crystals were measured and the mean was found to be 87° (s = 3.2). The microperthite crystals generally have a slightly clouded pale red (5R6/2) to moderate pink (5R7/4) colour which contrasts strongly with the narrow colourless bands of plagioclase that rim, and frequently grow in optical continuity with the plagioclase components of many of the microperthite crystals. The clear plagioclase which was tabulated separately in the mean mode, has a composition that ranges from An_4 to An_{16} with a mean of approximately An_{10}. In some specimens small crystals of amphibole lie between the microperthite host crystal and the plagioclase rim, thus indicating that the plagioclase rim probably developed late in the evolution of the rock.

The amphibole generally has the following optical properties: X = 1.674, moderate yellowish green (10GY 6/4) to moderate yellow green (5G 7/4); Y = 1.687, brilliant green (5G 6/6) to dark yellowish green (10GY 4/4); and Z = 1.693 dusky yellowish green (10GY 3/2); $Z_{\wedge}c = 19^{\circ}$ and $\bar{X}\ 2V_X = 52^{\circ}$. The optical properties are, however, variable - the refractive indices range from 1. 670 to 1.720, $Z_{\wedge}c$ varies from 5° to 20°, and the 2V measurements are also variable as the strong colour of the mineral produced anomalous extinction at high angles of tilt on the universal stage.

GRANULAR SYENITES FROM 3 DIFFERENT LOCALITIES



Photomicrograph 6: Type Granular Syenite (X15), Specimen No. 28, from Xaminxaip (6A7), X-nicols.



Photomicrograph 7: Granular Syenite (X15), Specimen No. 452, from Ring A Rooiberg 2 (4E7), X-nicols.

Photomicrograph 8: Granular Syenite (X15), Specimen No. 451, from Ring C Rooiberg 2 (4E7), X-nicols.

Amphiboles with a bluish hue were found to rim some of the larger minerals. These rims were found to have lower maximum extinction angles and thus it appears likely that they are more sodic than the core amphiboles. The amphibole is found altered to chlorite, biotite, epidote and minute grains of quartz and calcite. Opaque ore minerals, zircon and apatite sometimes occur included within the amphibole.

Most of the quartz occurs in small clear mildly strained anhedral crystals that typically occupy interstitial positions between the felspar crystals. Some quartz occurs in veinlets and micropegmatitic intergrowths.

Biotite occurs in anhedral to subhedral plates and shreds and is generally found associated with the amphibole crystals. Most of the biotite crystals are pleochroic from X= pale orange (10YR 8/4) to Y and Z = dusky yellowish brown (10YR 2/2).

Accessory amounts of zircon, apatite, the ore minerals (magnetite, ilmenite and leucoxene), fluorite, the epidote group (mainly pistacite and a little allanite), and sphene are found, and they typically occur associated with the ferromagnesiam minerals.

In addition to the main rock type described above, Ring A also contains a number of lenses or vein-like bodies of finer grained micro-syenite, and innumerable inclusions. The finer grained rock generally has a similar composition to its host though some specimens have a slightly higher quartz content. The quartz occurs mainly in micro-graphic intergrowths or veinlets that cut the microperthite crystals. The felspar in these finer grained patches is usually subhedral prismatic and is thus more idiomorphic than the felspar found in the host rock. Specimen 620 is a good example of the normal syenite of Ring A in contact with the finer material. The mode of the finer material of

Specimen 620 is alkalic felspar 75.2%, amphibole 15.6%, quartz 4.8%, biotite 3.0%, ore minerals 1.2% and miscellaneous heavy minerals 0.2%. At its outer contact particularly where it is in contact with the adamellitic gneiss the granular syenite of Ring A becomes contaminated, and is richer in inclusions and its quartz percentage tends to rise.

(ii) Ring B: This is the second outermost ring of the Rooiberg 2 annular complex. In the north and east of the complex where this ring is best exposed it has an average width of 200 yards (185 metres), though this width decreases as one moves away from this sector to become hardly discernible in the south of the complex. The principal rock type found in Ring B is a dark sempatic (Johannsen, 1931, Vol. I, p.229) porphyritic syenite. This rock is greyish red (5R 4/2) to dark greyish red (5R 3/2) in colour, and usually contains small greenish black (5G 2/1) speckles of ferromagnesian minerals. As the rock is porphyritic hiatal, (see Photomicrograph 10 and figure 4) the grain size distribution is bimodal and thus the arithmetic mean and median are not particularly significant as measures of central tendency. The means of the two (thin-section grain size area) maxima were found to be 10.8 sq. mm. and 0.025 sq. mm. The mean modal composition, and the standard deviations of the major constituent mineral species, of four fresh rock specimens from Ring B is as follows:

	Alkali-Felspar	Hornblende	Biotite	(Hornblende and Biotite)	Quartz	Opaque Ore Minerals
$\bar{x}$	80.4	10.7	6.0	(16.7)	1.5	0.8
s	4.5	3.1	4.0	(5.5)	1.3	0.9

PORPHYRITIC HIATAL SYENITES FROM ROOIBERG 2 AND THE XAMINXAIP BATHOLITH

Photomicrograph 9: Type Porphyritic Syenite (X15), Specimen No. 43, from Xaminxaip (6B6), X-nicols.

Photomicrograph 10: Porphyritic Syenite (X15), Specimen No. 448, from Ring B Rooiberg 2 (4E7), X-nicols.

	Plagio- clase	Zircon	Apa- Tite	Epi- dote	Chlorite	Cal- cite	Alla- nite	Leucox- ene
$\bar{X}$	0.3	0.1	0.1	0.1	Tr.	Tr.	Tr.	Tr.
s	-	-	-	-	-	-	-	-

As in the Granular Syenite, the characteristic felspar is microcline microperthite. The felspars fall into two size grades and the smaller crystals are more idiomorphic as most of them are subhedral prismatic. The optical properties of the microperthite are similar to those found in the dominant felspar of the Granular Syenite. The mean 2Vx was found to be 85^o (s = 3.3). A little clear plagioclase which ranges in composition from An5 - An23 and has a mean composition of An10, is found and it occurs both as rims to some of the large microperthite crystals and as small discrete crystals set in the groundmass. As might be surmised from the colour of the rock the microperthite found in the Porphyritic Syenite is darker than that found in Granular Syenite. De Villiers and Söhnge (1959, p.79) mention finding that anorthoclase was the dominant mineral in the "darkest" specimen of the "darker syenitic rocks". Anorthoclase was not found in the specimens studied. All the alkalicfelspar specimens examined had 2Vx's which were greater than 80^o.

Most of the amphibole occurs in clusters of subhedral to anhedral poorly terminated prisms that are generally confined to the groundmass, but some amphiboles occur as inclusions in the larger alkalic felspars. Their optical properties are as follows:- X = 1.684, pale yellowish orange (10 YR 8/6), Y = 1.700, moderate yellowish green (10 GY 6/4) and Z = 1.704, greyish olive (10 Y 4/2) to greyish green (5G 5/2), $Z_{\wedge}c = 21^o$; and 2Vx = 56^o (s = 5^o). As in the case of the Granular Syenite, the optical properties of these amphiboles are variable and the maximum extinction angle in some specimens with light blue green (5 BG 6/6) rims varies considerably from core to rim. The amphibole is believed to belong to the "common hornblende" group (Winchell and Winchell 1951, pp. 431-444). The alteration products of, and the inclusions in, these amphiboles are similar to those found in the amphibole from the Granular Syenite.

The biotite, which is generally pleochroic from X = moderate yellow (5 Y 7/6) to Y and Z = olive grey (5 Y 3/2), occurs in clusters of small anhedral to subhedral plates and shreds associated with the other ferromagnesian minerals. Most of the quartz is found in small, clear, generally strained, anhedral crystals that occur interstitially and are confined mainly to the groundmass, but a small part of the quartz occurs in micropegmatitic intergrowths. Accessory amounts of opaque ore minerals (magnetite, ilmenite and leucoxene), zircon, apatite and the epidote group (pistacite and allanite) occur.

(iii) Ring C: This is the third outermost ring of the Rooiberg 2 annular complex. Its average width is approximately 300 yards (275 m.), and this width is found to be much more constant than that of Ring B. The characteristic rock type of Ring C appears identical in hand specimen to the principal rock type of Ring A with which it merges in the south of the complex (see Photomicrographs 7 and 8). The main mineralogical difference between the two rock types is the higher biotite/hornblende ratio found in Ring C. While the texture of most of the rock making up Ring C is granular some specimens, particularly those near the C - D boundary, contain patches of rock which have micropegmatitic intergrowth textures. The mean grain size and the grain size distribution of Rings A and C is very similar. The mean modal composition, and the standard deviations of the major constituent mineral species, of four fresh specimens of Ring C is as follows:

	(1) Alkali felspar	Biotite	Amphibole	(Biotite and amphibole)	Quartz	Opaque ore Minerals	Plagioclase
$\bar{X}$	83.9	8.6	2.9	(11.5)	2.5	0.6	0.9
S	8.9	6.6	2.4	(7.3)	0.8	0.4	0.8

	Calcite	Zircon	Apatite	Chlorite	Epidote	Leucoxene	Allanite	Fluorite
$\bar{X}$	0.4	0.1	0.1	Tr.	Tr.	Tr.	Tr.	Tr.
S	-	-	-	-	-	-	-	-

(1) Footnote: Specimen 450, found on the eastern side of the complex near the C - B boundary was found to contain 96% felspar.

The general appearance, optical properties, and the mutual relationships existing between adjacent mineral species, of the minerals found in the rock described above, and those found in the Granular Syenite of Ring A, are identical.

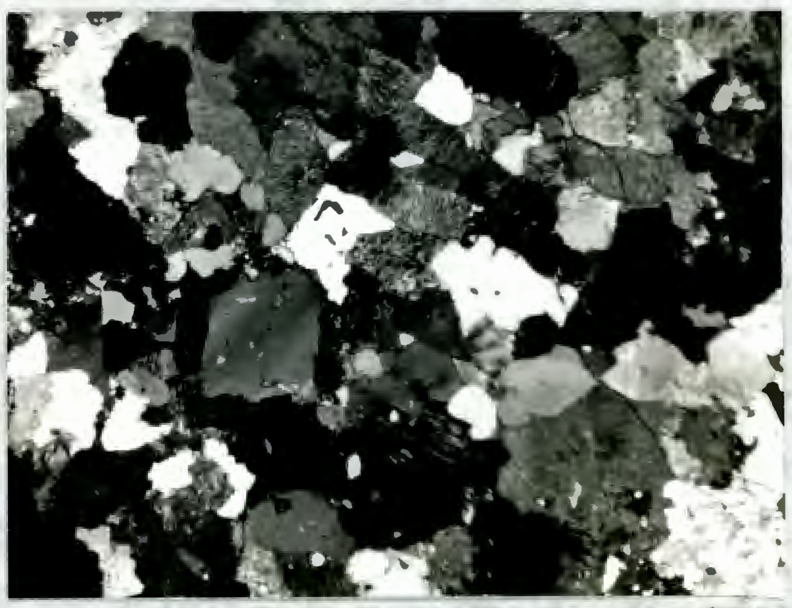
(iv) <u>Ring D</u>: This is the fourth outermost ring of the Rooiberg 2 annular complex. It is approximately 1000 yards (0.91 Km) in width and can be divided on textural grounds into two narrower rings. The outer Ring D_1 is approximately 400 yards (0.37 Km) wide and the inner Ring D_2 is 600 yards (0.55 Km) wide. The texture of the rock comprising D_1 is mainly granitic to seriate, but that of D_2 is characteristically sub-porphyritic to glomeroporphyritic and contains patches of micropegmatite (See Photomicrographs 12 and 13). The textural changes found within Ring D are generally gradational and many specimens display textural features that are intermediate in character between the extreme types of D_1 and D_2. A few porphyritic to sub-porphyritic patches were found in D_1 but in each case they were found to occur in irregular lens, or vein like patches within host material which had a more granitic texture. In the western sector of Ring D another rock sub-type can be distinguished from sub-types D_1 and D_2. This rock type, D_3, shows up clearly on aerial photographs and is found to be more intensely weathered than the remainder of Ring D. As can be seen from the modal data the main mineralogical difference between D_3 and the other rocks of Ring D is its higher fluorite and sericite content. The textures found in D_3 are of interest in that the textural feature of both D_1 and D_2 are retained in the outer and inner parts of D_3, thus suggesting that D_3 represents D_1 and D_2 material that has been permeated by solutions that have led to its more rapid decomposition, and increased fluorite content. If the Titley and Damon (1962, pp. 4491-4495) fluorite dating technique proves satisfactory it may eventually be possible to discover when the fluorite was introduced into unit D_3.

In the field specimens from Rings D_1 and D_2 range in colour from light brown (5 YR 6/4) to pale red (5 R 6/2), and the rocks of Ring D_3 tend to have a light pinkish grey (5 YR 7/1) colour. With regard to grain size the rocks of D_1 have a median thin - section grain area of 0.3 sq.mm. and a mode of 0.25 sq. mm. Thus they just fall on the granite side of the granite-microgranite boundary (B.A. Committee, 1936, p.323). Many of the rocks of D_2 are similar in median gran size to those of D_1, but in some extreme cases phenocrysts of 10.95 sq. mm. are found set in a groundmass which has a mean thin-section grain area of 0.055 sq. mm.; and as the groundmass in such cases tends to occupy a larger volume of the rock than the phenocrysts do, the rock is thus a microgranite. The majority of

THE ALASKITIC GRANITE

Photomicrograph 11: Alaskitic Granite showing dentate interlocking perthite crystals (X15), Specimen No. 122, from north of Xaminxaip (6A3), X-nicols.

Photomicrograph 12: Porphyritic variety of Alaskitic Granite (X11), Specimen No. 564, from Ring D_2 Rooiberg 2 (2F3), X-nicols.



Photomicrograph 13: Type Alaskitic Granite (X13), Specimen No. 581, from Ring D_1, Rooiberg 2 (2F2), X-nicols.

the rocks of D_2 do, however, fall within the granite range, but it is important when considering their genesis to remember that some belong to the microgranite group. The mean modal compositions, and standard deviations of the major constituent mineral species of 19 fresh specimens from rings D_1, D_2 and D_3 are as follows:

Ring	D_1 $\bar{X}$	D_1 s	D_2 $\bar{X}$	D_2 s	D_3 $\bar{X}$	D_3 s	$D_1+D_2+D_3$ $\bar{X}$	$D_1+D_2+D_3$ s
Number of Specimens	3		11		5		19	
Alkali-felspar	67.2	1.1	64.6	3.2	64.0	2.4	64.8	3.0
Quartz	27.9	2.4	30.1	3.0	32.5	3.3	30.4	3.4
Biotite	3.6	0.3	4.9	2.0	1.4	2.0	3.8	2.4
Plagioclase	0.8	1.0	0.1	-	0.1	-	0.2	-
Opaque ore Minerals	0.2	-	0.2	-	0.2	-	0.2	-
Fluorite	0.2	-	0.1	-	0.9	0.2	0.3	-
White Mica	0.1	-	Tr	-	0.9	1.0	0.3	-
Zircon	Tr	-	Tr	-	Tr	-	Tr	-
Apatite	Tr	-	Tr	-	Tr	-	Tr	-
Chlorite	Tr	-	Tr	-	Tr	-	Tr	-
Calcite	-	-	Tr	-	Tr	-	Tr	-
Epidote	-	-	Tr	-	-	-	Tr	-
Leucoxene	-	-	Tr	-	-	-	Tr	-

The characteristic felspar of the rocks of ring D is an equidimensional, subhedral to anhedral light pinkish grey (5 YR 7/1) microcline microperthite. Many of the larger crystals are found to have irregular fretted edges, and in some specimens in which the perthite crystals have grown in close contact with one another, a dentate structure as seen in specimen 122 (Photomicrograph 11) has developed. This dentate structure is similar to that depicted by Jacobson and others (1958) in their plate 5, No. 11. Most of the perthite is of the film type, but string and vein perthite is also found. In some specimens, particularly those from D2, clusters of larger than average microperthite crystals occur and these clusters tend to be devoid of quartz at their centres but the margins of the clusters contain quartz in micropegmatitic intergrowths. The mean 2 V X's of a large number of the more homogeneous microperthite crystals were found to be 81° ($s = 4.7^{\circ}$). The composition of this felspar will be discussed later in this chapter. Narrow rims and discrete crystals of plagioclase were once again found. The plagioclase ranges from An_8 - AN_{30}. The discrete crystals tend to be the most anorthitic. The mean plagioclase composition is An_{15}. De Villiers and Söhnge (1959, p.84) state that in some of the coarser granites at the foot of the central mass of Rooiberg 2 "anorthoclase figures prominently" and that "on unmixing the anorthoclase changes to orthoclase and anorthite (An_{90}- An_{100})". As anorthoclase ($Na, K, Si_3 AlO_8$) is generally considered to be a mixture of albite and orthoclase with the former dominant, the statement by de Villiers and Söhnge (1959, p.84) is not understood, unless instead of anorthite they mean albite; and if this is so then the unmixed anorthoclase is very similar in composition to the majority of the perthite crystals in this rock type, and the presence of anorthoclase supports the hypothesis that the Alaskitic Granite crystallized in the temperature range of the felspars of the high temperature group.

The majority of the quartz grains are clear, slightly strained, equidimensional, anhedral crystals that occur interstitially and in micropegmatitic intergrowths. Many of the quartz grains found in symplektic ingergrowths are oval but others have the shape of ichthyoglypts. In specimen 320 quarts felspar intergrowths are so common that the rock has a granophyric texture.

Biotite occurs in small clusters or aggregates of anhedral plates which are scattered sparsely throughout the rock. In some crystals the biotite is interlayered with white mica and chlorite. Most specimens are strongly pleochroic from X - moderate yellow (5Y 7/6) to Y and Z - greyish olive (10Y 4/2). White mica occurs mostly in small irregular shred-like crystals associated with altered felspar. A few specimens from D_3, however, occasionally contain larger plates of muscovite that appear to be unassociated with altered felspar. Small anhedral, equidimensional grains of fluorite which are generally colourless but often contain reddish purple (5 RP 5/2) to dark red (5 R 2/6) patches, are found scattered throughout the Alaskitic Granite and are particularly abundant in D_3. Accessory amounts of the opaque ore minerals, (magnetite, ilmenite and leucoxene), zircon, apatite, and the epidote group (mainly pistacite) also occur in the Alaskitic Granite. The ore minerals of specimen 284 are of particular interest as they are found to occur as vermicules included within quartz grains.

The calcite found in the Alaskitic Granite occurs mainly in anhedral grains and veinlets and has an irregular distribution.

(V) Unit E: This, the central unit of the Rooiberg 2 annular complex, is a nearly circular stock with a mean diameter of approximately 560 yards (510 metres) and an area of 0.8 sq. miles (200 hectares.). The type rock of this unit is a porphyritic sempatic hiatal microgranite in which the groundmass is volumetrically only slightly more abundant than the phenocryst generation. In the field this rock type is found to consist of a fine grained light grey (N7) groundmass in which are set large enhedral very light grey (N8) felspar phenocrysts and smaller euhedral to subhedral colourless quartz grains. The mean thin-section area of the crystals of the groundmass is 0.0052 sq. mm.; whereas the euhedral felspar phenocrysts have a mean thin-section area of approximately 150 sq. m.m., and some phenocrysts have a thin-section area of 330 sq. mm. The great size of the phenocrysts poses a serious sampling problem in the determination of the mode of this rock type, as some thin-sections contain mainly large felspar phenocrysts, others mainly quartz phenocrysts, and yet others contain only groundmass. In the field fractures are particularly common in this rock unit, and the rock is frequently found to contain zones with cataclastic textures and strained quartz grains.

Because of the sampling difficulties introduced by the large size of the phenocrysts the modes of all the thin-sections of this rock unit studied were added together and the arithmetic mean obtained, as this figure is believed to be the most significant indicator of the mineralogical composition of this rock unit. The arithmetic mean of 6 specimens from unit E is as follows: Quartz = 38.8%, Alkalic felspar = 37.2%, White Mica = 24.0%, Zircon = Tr., Fluorite = Tr., Biotite = Tr. and Opaque Ore Minerals = Tr. The rock is thus a Porphyritic Microgranite.

The felspar phenocrysts are euhedral to subhedral and usually appear in hand specimens as large rectangular prisms which have an average length of 17 m.m. and width of approximately 9 m.m. Most of the crystals show at least slight alteration to sericite. The 2VX's of the phenocryst felspars were found to be large, and as it is known from a chemical study of these rocks that the felspars are extremely K-rich, it seems clear that they fall into Tuttle's (1952.A.), and MacKenzie and Smith's (1956), microcline group. This conclusion appeared to be confirmed by the presence of gridiron structure in many of the specimens. The felspars of the groundmass are similar to the phyric generation except that they are usually less idiomorphic, and more altered.

Specimen 574 is of particular interest as it contains a narrow veinlet of potassic felspar.

The quartz phenocrysts are mainly subhedral broad prismatic crystals, and many have a thin-section diameter of approximately 7 m.m. In some thin-sections which contain basal sections of the quartz phenocrysts each of the six prism faces is found to be equally well developed. Many of the quartz crystals shown strain shadows and some are fractured. The cracks in such fractured quartz grains are sometimes found to be filled with groundmass material, thus indicating the groundmass generation crystallized after the quartz phenocrysts had crystallized and been fractured. Some of the large quartz crystals contain trains of liquid inclusions. The small quartz grains of the groundmass tend to be clear, equidimensional and anhedral.

A great deal of white mica (24%) is found and some of it is a secondary alteration product of the alkalic felspar which is found in all stages of alteration in this rock group; but the larger white mica plates are generally unassociated with altered felspar and thus do not appear to have been derived from felspar alteration.

Extremely small amounts of greyish pink (5R 8/2) zircon, opaque ore minerals, fluorite and biotite are also found in the Porphyritic Microgranite.

(D.) Petrography of the Xaminxaip Outcrop:

(i) Introduction: Three of the major rock types found in the Rooiberg 2 complex are also found in the southern part of the main outcrop of Richtersveld Suite rocks. The three rock types are (i) a rock type that is similar to the principal rock material of rings A and C and which will henceforward be called the Granular Syenite, (ii) a rock similar to the principal rock type of ring B and which will henceforward be called the Porphyritic Syenite, and (iii) a granitic

Plate 7: Alaskitic Granite (light) - Porphyritic Syenite (dark) contact in the Xaminxaip River (6C3)

Plate 8: A close up view of the contact shown in plate 7, and it shows (1) blocks of the Alaskitic Granite engulfed within the Porphyritic Syenite, and (2) narrow lenses of dark rhyolitic material.

and curvilinear. The boundaries of the individual rock units composing the main complex are frequently arcuate, but the overall picture is complex, and a study of the individual outcrop patterns does not produce a simple geometric pattern as is found in the Rooiberg 2 complex; but the presence of numerous arcuate features suggests that arcuate fractures played a significant role in the emplacement of the different rock units of the main outcrop.

(ii) <u>Alaskitic Granite</u>: In the field fresh specimens of the Alaskitic Granite are found to vary considerably in colour, from a moderate orange pink (10R 7/4) as found at De Hoop (see Map 2) the northern most outcrop visited, to a light grey (N7) as found in some of the granites in the southern part of the main outcrop. In most localities, however, the Alaskitic Granite varies from greyish orange pink (10R 8/2) to moderate orange pink (10R 7/4). As these rocks contain less than 5% of dark minerals the colour of the felspar tends to determine the colour of the rock as a whole. With regard to grain size, the median thin-section grain area is 0.27 sq. m.m. and the mode is 0.27 sq. m.m. thus the mode falls on the granite microgranite boundary (B.A. Committee, 1936) and the median falls within the granite grain size grade. For the purposes of the present study the Alaskitic Granites will all be regarded as fine grained granites even though a few specimens were found to fall within the microgranite size grade. The texture of the Alaskitic Granite from the main outcrop is as variable as that found in Ring D at Rooiberg 2. In most specimens the texture is intermediate between granitic and micropegmatitic, but some specimens are sub-porphyritic to porphyritic. The mean mode, and the standard deviations of the individual minerals, of 20 fresh specimens of the Alaskitic Granite collected from the main outcrop is as follows:

	Alkali Felspar	Plagioclase	Quartz	White Mica	Biotite	Opaque Ore Minerals
$\bar{x}$	57.8	1.7	36.5	1.5	1.4	0.5
s	6.1	2.0	3.5	2.0	2.6	0.5

	Chlorite	Calcite	Epidote Group	Zircon	Fluorite	Apatite	Leucox-ene
$\bar{X}$	0.3	0.3	0.1	0.1	Tr.	Tr.	Tr.
s	-	-	-	-	-	-	-

<u>Figure 4</u>: A comparison of the mineral composition and grain size distribution of the syenites of Rooiberg 2 and the main outcrop at Xaminxaip.

The general appearance, mode of occurrence and optical properties of the constituent minerals of the Alaskitic Granite from the main (Xaminxaip) occurrence are similar to those described for the Alaskitic Granite from Ring D.

(iii) Granular Syenite: The Granular Syenite from the main (Xaminxaip) outcrop is generally found to consist of a framework of pale red (5R 6/2) felspar crystals, speckled with greyish black (N 2) hornblende and other ferromagnesian minerals. The median thin-section grain area of the felspar crystals is 5.1 sq. mm., and the mode is 5.4 sq. mm. The hornblende crystals tend to occur as smaller crystals and have a mean thin-section area of approximately 2 sq. mm. Texturally these syenites are allotriomorphic granular, with the majority of the crystals being anhedral though many are subhedral (See photomicrograph 6). The mean mode, and the standard deviations of the individual mineral species, of 11 fresh specimens of the Granular Syenite, is as follows:

	Alkali felspar	Hornblende	Biotite	(Hornblende & Biotite)	Quartz	Opaque Ore Minerals
$\bar{x}$	87.4	6.3	2.3	(8.6)	1.5	0.9
s	5.2	4.7	3.7	(4.0)	1.7	0.5

	Calcite	Plagioclase	Chlorite	Zircon	Epidote Group	Apatite
$\bar{x}$	0.4	0.6	0.3	0.1	0.1	0.1
s	-	0.8	-	-	-	-

The general appearance, mode of occurrence and optical properties of the constituent minerals of the Granular Syenite from the main (Xaminxaip) occurrence are similar to those described when discussing the principal rock types of Rings A and C from Rooiberg 2.

(iv) Porphyritic Syenite: The Porphyritic Syenite from the main (Xaminxaip) outcrop is mineralogically and texturally very similar to the main rock type described from Ring B of Rooiberg 2. It varies from greyish red (5R 4/2) to dark greyish red (5R 3/2) in colour, and its texture is porphyritic sempatic with maxima at 10.8 sq. mm. and 0.025 sq. mm. The mean mode, and the standard deviations of the individual mineral species, of 14 fresh specimens of the Porphyritic Syenite is as follows:

	Alkali Felspar	Biotite	Hornblende	(Biotite & Hornblende)	Quartz	Opaque Ore Minerals
$\bar{X}$	78.8	12.0	3.5	(15.5)	3.1	1.6
s	4.1	8.0	5.3	(3.5)	1.3	0.8

	Calcite	Plagioclase	Zircon	Apatite	Epidote Group	Chlorite
$\bar{X}$	0.7	0.1	0.1	0.1	Tr.	Tr.
x	0.7	-	-	-	-	-

	Fluorite
$\bar{X}$	Tr.
s	-

(E) General Petrographic Relationships:

The foregoing petrographic study as summarized in figures 3 and 4, together with chemical data that will be presented in the next section, are the main reasons for departing from the correlation of the different units of Rooiberg 2 and Xaminxaip that was used by de Villiers and Söhnge (1959). These authors correlated Rings A and C with the Porphyritic Syenite of the main (Xaminxaip) outcrop, ring D with the Granular Syenite of the main outcrop, and unit E of Rooiberg 2 with the Alaskitic Granite of the main outcrop.

(F) Chemistry:

(i) Introduction: The chemical compositions of the different plutonic rock units that comprise the Richtersveld Suite are presented in Table VII and the overall chemical trends are presented in figures 9 and 11 (see Appendix)

TABLE VII - CHEMICAL DATA (RICHTERSVELD SUITE PLUTONIC ROCKS)

Nos.	1	2	3	4	(5)	(6)	(7)	(8)	(9)
SiO_2	68.00	69.92			67.38	74.45	75.01	72.08	73.86
Al_2O_3	14.01	13.98			15.49	13.60	13.88	13.86	13.75
Fe_2O_3	1.73	1.10			3.80	1.83	0.74	0.86	0.78
FeO	1.66	0.86					0.00	1.67	1.13
MgO	0.70	1.06			1.57	0.27	0.09	0.52	0.26
CaO	2.83	3.64			3.54	0.71	1.00	1.33	0.72
Na_2O	4.64	3.94	3.69	3.57	3.83	3.48	3.52	3.08	3.51
K_2O	4.25	3.75	4.87	5.24	3.04	5.06	4.89	5.46	5.13
H_2O^+	0.90	0.67					0.26	0.53	0.47
H_2O^-	0.14	0.09					0.11		
CO_2	0.44	0.15					Nil		
TiO_2	0.31	0.36			0.57	0.20	0.06	0.37	0.20
P_2O_5	Tr	Tr			0.21	0.14	Tr	0.18	0.14
MnO	0.19	0.39			0.07	0.05	Tr	0.06	0.05
SO_3	0.03	0.12							
Cl	0.65	0.32			0.01	0.02			
F	0.51	0.10			0.05	0.09			
BaO	Nil	Nil			0.05	0.09	0.10		
SrO	Nil	Nil			0.05	0.01	Tr		
Rbppm			278	260	110	170			
Lippm			4	4.5	24	40	Tr		
Csppm			<2	<2	2	4			
Cuppm			10	16	30	10			
Gappm			28	34	17	17			
Snppm			Approx. 5	Approx. 5	1.5	3			
Tlppm			Tr.	1.0	0.72	2.3			
Pbppm			25	23	15	19			
	100.99	100.45							
(Cl+F) = 0	0.36	0.11							
Total	100.63	100.34					99.66	100.00	100.00

Analyses 1 to 4 = Alaskitic Granite

() = Analyses introduced for comparison.

TABLE VII - Continued.

Nos.	(10)	(11)	12	(13)	(14)	15	16	17	18
SiO_2	76.52	76.15		73.84	56.96	59.72	56.16		57.82
Al_2O_3	11.93	12.48		14.29	20.24	14.32	15.59		17.17
Fe_2O_3	0.83	0.50		0.34	1.11	2.84	3.24		2.56
FeO	0.50	0.73		0.75	1.18	4.31	2.66		4.88
MgO	0.19	0.19		0.21	0.29	1.38	1.01		1.91
CaO	0.57	0.51		0.69	1.93	2.34	2.61		3.45
Na_2O	3.64	4.06	0.12	3.61	0.18	6.28	6.81	4.60	5.78
K_2O	5.10	4.43	7.91	5.21	14.08	3.46	4.48	5.78	3.59
H_2O^+	0.52	0.33		0.60	1.25	0.91	1.20		0.67
H_2O^-	0.23	0.20			0.36	0.14	0.13		0.18
CO_2		0.02			1.47	0.36	1.32		0.52
TiO_2	0.09	0.06		0.16	0.26	1.08	0.39		1.51
P_2O_5	0.01			0.25	0.06	Tr	Tr		0.08
MnO	0.02	0.01		0.05	0.07	2.45	2.91		0.23
SO_3		0.02			0.26	0.06	0.03		0.07
Cl	0.04					0.65	0.89		0.19
F	0.25	0.35			0.37	0.08	0.12		0.15
BaO					0.13	nil	nil		0.08
SrO					0.01	nil	nil		nil
Rb p.p.m.			677					103	
Li p.p.m.			67					19	
Cs p.p.m.		approx.	3.5					<2	
Cu p.p.m.			5						
Ga p.p.m.			27						
Sn p.p.m.		approx.	6						
Tl p.p.m.			1.7						
Pb p.p.m.			75						
	100.44	100.13			100.21	100.38	99.55		100.84
-(Cl+F)=O	0.11	0.15			0.30	0.18	0.25		0.11
Total	100.33	99.98		100.00	99.91	100.20	99.30		100.73
		(+ZrO_2=0.05)							

Analysis 12 = Porphyritic Microgranite

Analyses 15-17 = Porphyritic Syenite

Analysis 18 = Granular Syenite

TABLE VII - Continued.

Nos.	19	20	(21)	(22)	(23)	(24)	(25)	26
SiO_2	60.59		62.43	59.41	61.86	61.65	60.19	74.86
Al_2O_3	14.99		16.63	17.12	16.91	14.73	16.28	10.57
Fe_2O_3	2.44		4.72	2.19	2.32	4.56	2.74	1.10
FeO	2.30			2.83	2.63	3.68	3.28	4.46
MgO	0.91		0.97	2.02	0.96	0.70	2.49	1.04
CaO	2.55		2.52	4.06	2.54	1.87	4.30	0.78
Na_2O	8.98	5.97	5.45	3.92	5.46	6.69	3.98	3.70
K_2O	3.81	5.93	5.59	6.53	5.91	4.65	4.49	1.78
H_2O^+	0.87			0.63	0.53	0.58	1.16	0.16
H_2O^-	0.13							0.08
CO_2	0.71							0.78
TiO_2	0.58		0.58	0.83	0.58	0.52	0.67	0.56
P_2O_5	Tr		0.18	0.38	0.19	0.17	0.28	0.13
MnO	1.13		0.11	0.08	0.11	0.20	0.14	0.08
SO_3	0.07							
Cl	0.32		0.05					
F	0.21		0.12					
BaO	nil		0.18					
SrO	nil		0.02					
Rb p.p.m.		111	110					
Li p.p.m.		4	28					
Cs p.p.m.		<2	0.6					
Cu p.p.m.			5					
Ga p.p.m.			30					
Sn p.p.m.			x.					
Tl p.p.m.			1.4					
Pb p.p.m.			12					
	100.59							
-(Cl+F)=O	0.16							
Total	100.43			100.00	100.00	100.00	100.00	100.08

Analyses 19 and 20 = Granular Syenite

Analysis 26 = Rhyolitic Vein

TABLE VIII - NORMS (RICHTERSVELD SUITE PLUTONIC ROCKS)

	No.	1	2	15	16	18	19	26
	Q	21.18	25.50	3.00	-	0.72	-	41.87
	Or	25.02	22.02	20.57	26.69	21.13	22.24	10.56
	Ab	37.20	32.49	51.35	49.78	48.21	52.92	31.44
	An	5.84	9.73	1.29	-	10.84	-	-
	Ne	-	-	-	2.56	-	1.70	-
	C	-	-	-	-	-	-	2.55
Di	$CaSiO_3$	1.51	2.90	3.25	1.51	1.16	3.13	2.60
	$MgSiO_3$	0.80	2.10	1.00	0.40	0.60	1.00	6.47
	$FeSiO_3$	0.66	0.53	2.38	1.19	0.53	2.24	-
Hy	$MgSiO_3$	1.00	0.60	2.50	-	4.20	-	-
	$FeSiO_3$	0.79	0.26	5.81	-	4.22	-	-
Ol	$MgSiO_4$	-	-	-	1.40	-	1.05	-
	Fe_2SiO_4	-	-	-	4.49	-	2.34	-
	Ac	-	-	-	-	-	6.93	-
	Ns	-	-	-	-	-	2.56	-
	Mt	2.55	1.62	4.18	4.64	3.71	-	1.62
	Il	0.61	0.76	2.13	0.76	2.89	1.06	1.06
	Cc	1.00	0.30	0.80	3.00	1.20	1.60	1.40
	Fr	0.55	0.11	0.16	0.23	0.16	0.20	-
	Hl	1.05	0.47	1.05	1.52	0.35	0.47	-
Minor Constituents		1.07	0.88	1.11	1.36	1.00	1.07	0.37
Total		100.54	100.49	100.68	99.81	100.92	100.51	99.94

Index for tables VII and VIII

Nos.

1 <u>(86/1) Alaskitic Granite</u> from ring D_3 Rooiberg 2 (de Villiers and Söhnge, 1959, p.86, No.1.)

2 <u>(86/2) Alaskitic Granite</u> from Xaminxaip (de Villiers and Söhnge, 1959, p.86, No.2.)

3 <u>Specimen 122</u>: Type Alaskitic Granite from the main outcrop at Xaminxaip: Mode (vol.%) Quartz 34.8%, Alkali-felspar 63.0%, plagioclase 0.3%, biotite 1.3%, opaque ore minerals 0.3%, zircon 0.1%, fluorite 0.2%.

4 <u>Specimen 565</u>: Type Alaskitic Granite from Ring D Rooiberg 2 : Mode (vol.%) quartz 29.8%, alkali-felspar 66.4%, plagioclase 0.4%, biotite 3.1%, opaque ore minerals Tr, fluorite 0.3%.

5 Turekian and Wedepohl (1961, p.186.) high Ca-Granite.

6 Turekian and Wedepohl (1961, p.186.) low Ca-Granite.

7 Johannsen (1932, p.49.), Vol. 2, Alaskite "type" rock.

8 Nockolds (1954, p.1012.). Average Calc-Alkali Granite (Mean of 72.)

9 Nockolds (1954, p.1012.). Average Alkali Granite (Mean of 48).

10 Jacobsen et.al. (1958, p.14.) - Bargesh biotite granite (L.796.), Mode : quartz 35.6%, perthite 55.9%, albite - oligolcase 4.9%, biotite 3.2%, fluorite 0.2%, iron oxides tr.

11 Jacobsen et.al. (1958, p.14.) - biotite granite from Liruei (X.568), Mode : quartz 37.0%, perthite 54.8%, albite oligoclase 6.3%, biotite 1.4%, fluorite 0.3%, iron oxides 0.2%, zircon tr.

12 Specimen 577: Type Porphyritic Microgranite from Rooiberg 2 : Mode (vol.%) of unit E : quartz 38.8%, alkali-felspar 37.2%, white mica 24.0%, opaque ore minerals tr., zircon tr., fluorite tr.

13 Nockolds (1954, p.1012.). Average Muscovite Granite (Mean of 6.).

14 Von Eckermann (1960, p.527). Borengite: Mode (Wt.%) K-felspar 75.6% ($Or_{98}Ab_2$), sericitic mica 17.8%, carbonate 3.4%, iron hydroxides 1.1%, fluorite 0.9%, ilmenite 0.5%, pyrite 0.6%, apatite 0.1%.

15 Specimen 81/1 : Porphyritic Syenite from ring B Rooiberg II (de Villiers and Söhnge 1959, p.81, No.1)

16 Specimen 81/4 : Porphyritic Syenite from the Southern edge of the main outcrop ½ mile west of the confluence of the Stinkfontein and Tc Osib Rivers (de Villiers and Söhnge 1959, p.81. No.4.)

17 Specimen 43: Type Porphyritic Syenite; Main Outcrop : Mode : Quartz 2.5%, alkali-felspar 81.2%, hornblende 12.5%, opaque ore minerals 3.3%, calcite 0.3%, apatite 0.1%, zircon 0.1%, biotite tr., chlorite tr.

18\. Specimen 81/2: Granular Syenite from ring A, Rooiberg 2, (de Villiers and Söhnge 1959, p.81. No.2).

19\. Specimen 81/3: Granular Syenite from ring A, Rooiberg 2, (de Villiers and Söhnge 1959, p.81. No.3).

20\. Specimen 28: Type Granular Syenite, Main Outcrop: Mode Alkali-felspar 88.4%, hornblende 7.2%, opaque ore minerals 2.1%, plagioclase 1.7%, zircon 0.3%, quartz 0.3%.

21 Turekian and Wedepohl (1961, p.186.). Average Syenite.

22 Nockolds (1954, p.1016.). Average Calc. Alkali Syenite - leucocratic (Mean of 18).

23 Nockolds (1954, p.1016.) Average Alkali Syenite (Mean of 25).

24 Nockolds (1954, p.1016.) Average Peralkaline Syenite (Mean of 47).

25 Daly (1933, p.11.) Average of "all syenites" (Mean of 50).

26 Specimen 285 : A specimen of the dark rhyolitic material found in veins and lenses at the contact between the Alaskitic Granite and the Porphyritic Syenite (veins can be seen in Plate 8.). - Analyst E.C. Haumann.

which show graphically the degree to which the chemistry of these rocks departs from the "normal" igneous trend. These two diagrams are considered useful as unlike most variation diagrams, silica which tends to give undue emphasis to the differences between granites and syenites, is not used as the abscissa.

The Thornton and Tuttle (1960, pp.664-684) diagrams (figure 11 in Appendix 2) reveal that the only significant departures from the normal igneous trend found in the plutonic rocks of the Richtersveld Suite, is the low SiO_2 content of the syenites, the high FeO content of some of the syenites, and the high K_2O and low Na_2O content of the Porphyritic Microgranite. The Green and Poldervaart (1958, pp. 87-122) diagram (figure 9 in Appendix 2) is of interest as it shows (1) that the plutonic rocks of the Richtersveld Suite fall on the "normal igneous trend" and (2) that the Alaskitic Granite is lower down the differentiation series than the syenites. This position still obtains if the Ca associated with F in the Alaskitic Granite is removed. Turekian and Wedepohl's (1961, Table 2) average high and low Ca granites and their average syenite were also plotted on figure 9 and the two average granites were found to plot on either side of the average syenite.

(ii) Chemistry of the Alaskitic Granite: If the analyses in the first eleven columns of table VII are compared it is found that the Alaskitic Granite tends to have a normal Al_2O_3, FeO, Na_2O, K_2O, TiO_2, F, Cs, Cu, Sn and Tl content;

it is slightly high in Fe_2O_3, MgO, Rb, Ga and Pb; significantly high in Cl, MnO and CaO; slightly low in SiO_2; and it is significantly low in Li. While the above statement is true it should be noted that, (1) local enrichment in Na occurs in some specimens, (2) the F content of specimen one is higher than normal as it comes from the fluorite rich unit, D_3, of Rooiberg 2, and (3) the CaO value of specimen one is but slightly higher than normal if the CaO that is required to form normative calcite and fluorite is removed. It is regarded as significant that specimen one which contains 0.51% F, also contains over twice as much Cl as specimen two, and specimen two has a F content of only 0.10%.

If the norms of specimens one and two, as shown in table VIII are compared with the mean modal value of the Alaskitic Granite (mean of 39 specimens) which is given in table IX, it is immediately evident that the normative quartz tends to be slightly lower than the mean modal value. The modal value does, however, have a standard deviation of 5 which indicates that there tends to be a considerable amount of variation in the modal quartz percentage. The normative and modal felspar values are, however, of the same magnitude. With regard to the mafic minerals the modal and normative values are not strictly comparable as we are dealing with different mineral phases; but it is clear that the norms of the dark minerals of specimens one and two and the mean mode in weight per cent of the dark minerals, are quantitatively, broadly similar. The opaque minerals magnetite and ilmenite are clearly more abundant in the norms than in the modes. The presence of over 1% halite in the norm of specimen one is of particular interest, but this high value is to be expected as halite is the only normative mineral which contains chlorine, and as we have seen specimen one, and for that matter most of the rocks of the Richtersveld Suite, contain

Table IX : Mean Modes of the Principal Plutonic Rock Types of the Richtersveld Suite

	Granular Syenite			Porphyritic Syenite			Alaskitic Granite			Porphyritic Microgranite	
	Vol.%		Wt.%	Vol.%		Wt.%	Vol.%		Wt.%	Vol.%	Wt.%
	$\bar{X}$	s	$\bar{X}$	$\bar{X}$	s	$\bar{X}$	$\bar{X}$	s	$\bar{X}$	$\bar{X}$	$\bar{X}$
Alkali-felspar	85.7	5.4	83.08	78.9	4.4	75.73	61.3	5.9	60.46	37.2	35.74
Hornblende	7.2	4.9	8.82	4.9	5.8	4.28					
Biotite	2.9	4.5	3.21	11.3	7.7	12.96	2.6	2.5	2.92	Tr	Tr
(Hornblende and Biotite)	(10.1)	(4.7)	(12.03)	(16.1)	(4.1)	(17.24)					
Quartz	2.0	1.5	1.97	2.6	1.4	3.05	33.6	4.8	33.68	38.8	38.76
Opaque Ore Minerals	0.7	0.5	1.28	1.5	0.8	2.92	0.4	-	0.75	Tr	Tr
Plagioclase	0.6	0.8	0.58	0.1	-	0.07	0.9	1.7	0.90		
Calcite	0.3	-	0.32	0.5	0.6	0.71	Tr	-	Tr		
Chlorite	0.2	-	0.18	Tr	-	Tr	0.1	-	0.09		
Zircon	0.2	-	0.32	0.1	-	0.16	Tr	-	Tr	Tr	Tr
Epidote Group	0.1	-	0.12	Tr	-	Tr	Tr	-	Tr		
White Mica	Tr	-	Tr	Tr	-	Tr	0.9	1.6	0.97	24.0	25.5
Apatite	0.1	-	0.12	0.1	-	0.12	Tr	-	Tr		
Fluorite	Tr	-	Tr	Tr	-	Tr	0.2	-	0.23	Tr	Tr
Leucoxene	Tr	-	Tr	Tr	-	Tr	Tr	-	Tr		
Sphene	Tr		Tr								

significant amounts of chlorine.

As there are only a few quantitatively significant mineral phases present in the Alaskitic Granite, the sodium and potassium present in those phases other than the felspars can be calculated and subtracted from the total alkali content, and with the value so obtained some idea of the overall chemical composition of the felspar can be obtained. In specimen three (table VII) the only mineral other than felspar containing significant amounts of K and Na is biotite. The weight per cent of biotite in this specimen is 1.47%. Using the mean K_2O and Na_2O values of twelve typical biotites from granites and biotite-granites (taken from Foster, 1960B, pp.41-45) it was found that 1.47% biotite contains 0.13% K_2O and 0.01% Na_2O. If these values are subtracted from 4.87% and 3.69% respectively (the total K_2O and Na_2O values of specimen 3) we are left with 4.74% of K_2O and 3. 68% Na_2O. As 4.74% K_2O can form 28.1% Or and 3.68% Na_2O can form 31.2% Ab we arrive at a total of 59.3% Ab+ Or which is 4% short of the 63.3% total felspar content of specimen three. If this 4% remainder is taken to be An, the felspar of specimen 3 is found to consist of 45% Or, 49% Ab and 6% An. If the same procedure is carried out on specimen 4 it is found to contain 43.7% Or, 45.1% Ab and 11.2% An. The mean of the total felspar components of specimens 3 and 4 is Or_{44} Ab_{47} An_9 or in more conventional form Or_{44} $(Ab_{84}\ An_{16})_{56}$.

(iii) <u>Chemistry of the Granular Syenite</u>: Analyses 18, 19 and 20 of table VII are of the Granular Syenite. Analysis 18 is not considered typical of this rock group as de Villiers and Söhnge (1959, p.82) call it "an akerite from the outer ring of Rooiberg 2". While it is clear from the analysis that this rock is not an akerite, it seems evident from the name it was given in the field, that it is probably prudent to regard this specimen as a dark local variation in the Granular Syenite. Specimen 19 which de Villiers and Söhnge (1959, p.82) call a "syenite from the outer ring of Rooiberg 2" is probably more representative of the granular syenites but when its alkalies are compared with those of the "type" specimen (No.20, table VII) it is seen to be more sodic than the "type" specimen.

The mean chemical composition of specimens 18, 19 and 20 is considered to be representative of the chemistry of the Granular Syenite, and when these values are compared with the "average" syenites (columns 21 - 25) of table VII, it is found that the Granular Syenite has a normal SiO_2, Al_2O_3, MgO, CaO, Fe_2O_3, $H_2O^{\pm}$, Rb and Cs content; it is slightly high in TiO_2, FeO, Na_2O and F; significantly high in MnO and Cl; slightly low in K_2O; and it is significantly low in P_2O_5 and Li.

If the norms of specimens 18 and 19 as given in table VIII are compared with the mean mode (wt%) of the Granular Syenite as given in table IX, it is immediately seen that the normative quartz content of specimen 18 is of the same magnitude as the mode, and that specimen 19 is normatively undersaturated. As no mineralogically undersaturated specimens of the Granular Syenite were found during the present study, and the mean quartz content of this rock type was found to be 2.0% (S = 1.5), it is believed that this undersaturated specimen is of but local significance. Specimen 19 is chemically different from the other two Granular Syenite specimens (18 and 20) in that it contains over 3% more Na_2O and 2% more $Na_2O + K_2O$ than their mean value. It appears from the experimental work of Bowen and Tuttle (1950; Tuttle and Bowen 1952) that there is a thermal barrier that lies along the Ab - An felspar join and prevents the transition by simple differentiation from synthetic nepheline syenite to synthetic quartz syenite. It might thus be supposed that this felspar thermal barrier is interposed between specimens 19 and 20. In fact, however, both specimens plot very close together on either side of the Ab - Or join in the system Qz - Ne - Kp; this, together with the homogeneous field appearance of the syenite, seems to indicate that the thermal barrier did not significantly influence the crystallization history of the Granular Syenite. Similar conditions occurred in the evolution of most of Tilley's (1957, p.332) group A alkali complexes which contain rocks that grade from nepheline syenite through syenite to quartz syenite and to granite. The writer agrees with Tilley's (1957, p.332) suggestion that the thermal barrier found in simple experimental melts probably does not operate in complex natural melts which contain mafic materials and various volatile constituents.

The modal and normative felspar values are similar in the case of specimen 18 but the normative value is low in specimen 19. This is to be expected in specimen 19 as Na_2O was used in normative nepheline, acmite and Ns (sodium silicate). The quantities of ferromagnesian minerals found in the modes and norms is broadly similar, though the magnetite and ilmenite content of specimen 18 is found to be high. The mean normative calcite (1.4%) is high when compared with the mean modal value (0.3%). The presence of 0.4% normative halite and 0.2% normative fluorite is considered of interest and will be discussed in a later section on petrogenesis.

Using the same procedure for determining felspar compositions as was employed in the study of specimens 3 and 4 in the section on Alaskitic Granite, it was found that the total felspar composition of specimen 20 was Or_{40} (Ab98 An_2)60.

This calculation ignores the sodium content of the amphiboles, and if it was taken into consideration the An content of the felspar would be slightly increased and the Ab content proportionately lowered.

(iv) Chemistry of the Porphyritic Syenite: When the mean chemical composition of analyses 15, 16 and 17 of table VII which is considered to be representative of the chemistry of the Porphyritic Syenite, is compared with the "average" syenites (columns 21-25 of table VII), it is found that the Porphyritic Syenite has a normal MgO, CaO, TiO_2, K_2O, F, Rb and Cs content, it is slightly high in Fe_2O_3, FeO and Na_2O; significantly high in MnO and Cl; slightly low in SiO_2, Al_2O_3, Li and BaO; and it is significantly low in P_2O_5. While the mean CaO content of the Porphyritic Syenite is similar to the value for the "average" syenites, this value would be considerably reduced if the CO_2 content of the rock was united with the CaO to form calcite. In specimen 16 for example 3.0% normative calcite can be formed and this would require 1.68% CaO.

When the norms and mean modes of the Porphyritic Syenite are compared, the normative and modal quartz values of specimen 15 are in good agreement but specimen 16 which is normatively undersaturated is like specimen 19 very high in Na_2O. The remarks made about specimen 19 in section 5.F.iii are considered applicable to specimen 16. The total amounts of normative and modal felspar is approximately the same. The total normative and modal ferromagnesian mineral content is very similar, although the combined normative magnetite and ilmenite is much greater than the modal (wt%) opaque ore minerals. The normative calcite content of specimen 16 is much higher than the mean modal value, but specimen 15 contains only ¼ as much normative calcite as specimen 16. The mean norm is found to contain 0.2% fluorite while the mean mode contains only traces of this mineral. These differences probably indicate that fluorine was incorporated in the structure of the ferromagnesian minerals of the Porphyritic Syenite. The mean normative value of 1.3% halite is of interest as it reflects the high chlorine content of these rocks.

Using the same procedure as was employed on specimens 3,4 and 20, it was found that the total felspar composition of specimen 17, the "type" Porphyritic Syenite was Or 45 $(Ab_{93}\ An_{7})$ 55. This calculation ignores the sodium content of the amphiboles, and if this sodium was taken into consideration the An content of the felspar would be slightly increased and the Ab content proportionately lowered.

(v) <u>Chemistry of the Porphyritic Microgranite</u>: The chemical composition of this rock type differs radically from the other members of the Richtersveld Suite and from the "average" granites quoted in table VII. The rock is extraordinarily low in Na_2O, slightly low in Cu, normal in Tl, slightly high in Cs, Ga and Sn, and significantly high in K_2O, Rb, Li and Pb. The K/Na ratio of the Porphyritic Microgranite differs so considerably from that of a "normal" granite that an analysis of a Borengite (specimen 14) was included in table VII, as this is one of the few igneous rocks found in the literature that has a similar K/Na ratio. The Borengite can, however, hardly be regarded as a normal igneous rock type as von Eckermann(1960, p.528) calls it "the most potassic rock of magmatic habit known today in the world", and he adds that it is "not only quite unique but its genesis is .... a challenge to geochemists". As the potassium content of the Porphyritic Microgranite was high, a high rubidium content was to be expected, but when the K/Rb ratio of the Porphyritic Microgranite was determined it was found to be 97 which indicates that there was also strong rubidium enrichment relative to potassium.

The K and Na values of the Porphyritic Microgranite clearly indicate that its felspar is a potassium rich variety. Even if all the Na in the rock went into the formation of felspar, albite would form but 3% of the total felspar. But as most analyses of muscovite contain at least some, and many contain more than 1% Na_2O (Ref. Foster, 1960 A;p. 142 table 6), it is evident that the value of An_3 is a maximum value. This indicates that the felspar of the Porphyritic Microgranite is strongly potassic, a characteristic which is unique among the rocks of the Richtersveld Suite as they generally contain less than 50% of the Or. molecule in their total felspar content.

(vi) Minor Elements: The study made of the trace element content of the different units of the Richtersveld Suite proved most rewarding, because as Butler and others (1962, p.89) have indicated, striking major element compositional changes are not to be expected in rocks which approach the ternary minimum in the system $SiO_2 - NaAlSiO_4 - KAlSiO_4$, but that a knowledge of minor element concentration can be of appreciable significance and that a change in the concentration ratio of a closely allied geochemical pair is generally of petrographic significance. In the plutonic members of the Richtersveld Suite rubidium is found to increase from 103 p.p.m. in the Porphyritic Syenite to 111 p.p.m. in the Granular Syenite, 269 p.p.m. in the Alaskitic Granite, and to 677 p.p.m. in the late Porphyritic Microgranite. The Rb values of the syenites are considered 'normal' (Turekian and Wedepohl, 1961), but the K/Rb ratios are abnormally high (466 and 443) being double the ratio (230) now accepted as 'normal' (Taylor, 1960B, p.318). Figure 12 (in the

Appendix 2) shows that Turekian and Wedepohl's (1961) and Horstman's (1957) "average syenites" have similar anomalous ratios which tend to plot close to the two Richtersveld Suite syenites. This poses the question of whether syenites, which are after all characterized by the unusual combination of properties of alkali enrichment and silica depletion, are 'normal' igneous rocks? Before any attempt can be made to answer this question, however, much more information will have to be obtained on trace element abundances in syenites. Taylor and Heier (1958A, p.p. 202-203) have discussed the question of rubidium depletion in felspars. They emphasise the close geochemical coherence of potassium and rubidium and state that rubidium ratios found to lie beyond the normal limits of scatter (outer diagonal lines in figure 12 in appendix 2) indicate extreme processes capable of fractionating potassium and rubidium (p. 202). They found that specimens with anomalously low rubidium contents tend to fall into area 3 of figure 12; and as can be seen in this figure the Richtersveld syenites and the Quartz Bostonite do fall into this area. In figure 2 of their 1958A paper Taylor and Heier plotted the compositions of a number of felspars with normal and others with anomalous K/Rb ratios on a An - Ab - Or triangular diagram. They also included the cotectic curve that separates the fields in which OrAb - rich and AbAn - rich felspars crystallize in the system Ab-Or-An-H_2O at 500 bars water pressure. They then discovered that "rubidium is preferentially excluded from alkali felspars, which although containing substantial amounts of potassium, have a composition which places them in the Ab An-rich field....close to a composition from which an OrAb-rich felspar would crystallize as a stable phase" (p.202).

They also found that "the AbAn-rich felspars impoverished in rubidium lie close to the maximum of the subsolidus curve for alkali felspars separating the one and two felspar regions, indicating maximum instability for a single felspar phase. When an AbAn-rich phase separates, the number of co-ordinated sites available for potassium must be restricted. In this case the 10% difference in size is liable to become critical, and K^+ may preferentially enter the lattice at the expense of Rb^+. If much Ba^2 (1.34Å) is present this may also occupy the high co-ordination sites to the exclusion of rubidium" (p.202-203). If the above structural explanation was the complete answer to the question of rubidium depletion one would expect this depletion to have occurred in the Alaskitic Granite as well as in the syenites as the felspars of both rock groups are very similar. Taylor and Heier (1958A, p.203) do not, however, believe the above explanation to be the complete one "since plagioclase felspars have normal potassium / rubidium ratios". They believe the "additional factor" to be related to the concentration of potassium in the system from which the felspar is crystallizing, and that if K is present in large enough quantities it could use its size advantage to "force" itself preferentially into the felspar crystal lattice. It is also possible that the difference in the K/Rb ratios of the syenites and the Alaskitic Granite is due to their having formed in different environments, the syenites being the mobilized or remelted products of a felspathization process which acted on the roof rocks of the magma reservoir and the Alaskitic Granite having crystallized from an invading magma. This hypothesis would be valid, only if it could be shown that K is preferentially introduced during felspathization. This might prove to be the case as the ionic radius of Rb^+ (1.47Å) is 10% larger than that of K^+ (1.33Å). Studies by Horstman (1957) and Gast (1960, p.1295), however, seem to suggest that "igneous and metamorphic processes which concentrate potassium generally concentrate rubidium to an equal or greater degree".

The K/Rb ratio of the Alaskitic Granite ($\bar{X}$ = 156) is lower than normal but as we have seen it falls within acceptable limits of scatter. This slight Rb enrichment might well result from the Rb not being readily accepted in the syenites, thus producing a slight enrichment of this element in the magma associated with them.

The Porphyritic Microgranite is found to contain 677 ppm rubidium and to have a K/Rb ratio of 97. This clearly shows the strong rubidium enrichment which is so characteristic of late differentiates, particularly pegmatites (Goldschmidt et.al.,1933; Ahrens and others 1952; Horstman, 1957, p.13; Nockolds and Allen, 1953, p.136; Taylor et al, 1956, p.228; Taylor et al, 1958 A.p.202; Zlobin and Lebedev, 1960, p.105). Volkov and Sovinova (1959, p.638) have stated that the K/Rb ratio may serve as a geochemical index of the age sequence of intrusive phases in magmatic complexes (ie. the ratio decreases with age). The K/Rb ratios of the rocks at Rooiberg 2 follow this rule. The Porphyritic Microgranite is not unique in having a low K/Rb ratio but is rather similar in this respect to the Banks Peninsula rhyolites, and the Mourne and St. Austell granites. Taylor and others (1956, p.228) suggested that

these rocks crystallized from magmas which had "undergone differentiation to a degree typical of the pegmatite phase rather than the less extreme differentiation common in normal acid igneous rocks".

Analyses of these rocks fall within area 2 of figure 12 in appendix 2.Area I of figure 12 represents analyses of alkali felspars from some large pegmatites. The Porphyritic Microgranite is found to have K/Rb values which are intermediate between areas I and II. This is considered particularly significant as this rock type is interpreted as being the product of the crystallization of a highly fractionated magma which after crystallization was permeated by pegmatitic solutions. Recent work on the K/Rb ratios of the younger granites of northern Nigeria (Butler and others, 1962, p.91) has shown that the "biotite-granites" have ratios ranging from 220 to 42, and the "riebeckite-granites" have ratios that range from 270 to 24.

The lithium content of the plutonic rocks of the Richtersveld Suite is found to increase from 4 p.p.m. in the Alaskitic Granite and Granular Syenite to 18 p.p.m. in the Porphyritic Syenite and 67 p.p.m. in the Porphyritic Microgranite. The low Li content of the Alaskitic Granite and Granular Syenite is probably related to their relatively low ferromagnesian and phyllosilicate contents, as Li^{+} tends to replace Mg^{2+}, Fe^{2+} and Al^{3+} (Strock, 1936 and Horstman 1957, p.4). The Porphyritic Syenite which has a higher ferromagnesian content ($\bar{X}$ = 11.3 vol.% biotite and 4.9 vol.% hornblende) has a more normal Li content (see Table VII). The Porphyritic Microgranite once again proves to be of interest as it shows strong Li enrichment, and such enrichment is once again a characteristic feature of the final phase of an intrusive cycle (Strock, 1936; Nockolds and Allen, 1953, p.121; and Horstman, 1957, p.9 etc.). As the Porphyritic Microgranite is conspicuously free of biotite and hornblende it seems evident that the Li is accommodated in the muscovite. Foster (1960A, p.124) has shown that the muscovite structure can accommodate up to 3.3% Li_2O; and the muscovite of the Porphyritic Microgranite is required to accommodate only 0.06% Li_2O to carry the 67 p.p.m. Li in the rock.

The Caesium, lead, thallium and copper values found in the Porphyritic Microgranite are also significantly different from the values found in the Alaskitic Granite. As expected in a late differentiate the Porphyritic Microgranite is enriched in Cs, Pb and Tl (Ref. Shaw, 1957, p.196) and the Cu content falls to half the value found in the Alaskitic Granite. The lead content of the Porphyritic

Microgranite which is three times greater than the Alaskitic Granite value, and four times as large as Turekian and Wedepohl's (1961) average high- and low-Ca granites, is particularly interesting, as the lead appears to have diadochically replaced potassium in the potash felspar. The lead does not appear to be present in the other minerals as the muscovite from a "muscovite vein" found within the Porphyritic Microgranite contained only a trace of lead, and this trace was probably contained in the fluorite which the vein is known to contain. Apatite, a mineral which is known to carry lead, is not a common accessory in the Porphyritic Microgranite. The plumbiferous nature of the potash felspars found in the Porphyritic Microgranite is a further indication that this rock is a late differentiate, as Rankama and Sahama (1950, p.733) state that such lead enrichment in potash felspars is characteristic of late differentiates, pegmatites, and the action of late pneumatolytic and hydrothermal emanations.

(G) <u>Mode and Depth of Emplacement</u>:

The tectonic relationships that exist between the different plutonic members of the Richtersveld Suite will now be considered in order to establish the sequence in which the rock unitswere emplaced. The Rooiberg 2 outcrop which contains all these units, and which also has the simplest outcrop pattern at the present level of erosion, will be considered first. Rooiberg 2 is an "annular intrusive body" and is strikingly similar in plan to many other complexes from different parts of the world which have been interpreted as having been emplaced by ring-fracture stoping or cauldron subsidence. The term "annular intrusive body" is here used as a general term for an intrusive body of circular or oval plan (Billings, 1943, p.131), and thus includes ring-dykes, circular and oval stocks, and cone sheets. Billings' (1943, p.131) definitions of these latter units are also used in the present study.

The geological map of the S.E. Richtersveld indicates that Rooiberg 2 consists of four ring-dykes(from the outside inwards rings A, B, C and D) which are concentrically arranged about a central stock (unit E) which is perhaps more in the nature of a plug. As the composition and texture of the rock forming ring-dykes A and C is the same, and in the South of the complex these two units combine to form a single unit, ring-dykes A and C can be thought of as a single unit. Thus we have a central unit E, and concentrically arranged about it are two major ring-dykes D and AC with a minor incomplete ring-dyke B enclosed within unit AC. Field observations indicate that unit E cuts across unit D. Unit D in its turn appears to cut across unit AC and veins of granitic material from unit D are found to cut across units AC and B. Thus unit E was emplaced last and unit D second last. Unit B was either emplaced between units A and C and is thus younger than unit A, or it was emplaced first, and it and the fragmental remains of a screen of country rock with which it is found associated, together formed a composite screen within unit AC. The second interpretation is favoured as (1) the Porphyritic Syenite of the Xaminxaip complex which is similar in composition and texture to ring B is believed to have been emplaced before the Granular Syenite, (2) the porphyritic texture of unit B is in harmony with the concept that this unit was emplaced first into the relatively cold country rock (of the Epizone), and (3) as unit B is found associated with the fragmental remains of a screen of older country rock the idea that it too is a screen seems more feasible. Consequently unit B is believed to have been emplaced first followed by units A, C, D and E.

When traversing over most of Rooiberg 2 one forms the opinion that the ring-dykes have vertical contacts, but in some localities where the relative relief of the terrain is greatest, and the contacts can be observed in vertical section, the contacts are frequently found to depart from the vertical and dip steeply outwards. The circular shape of the Rooiberg 2 ring-dykes suggests that their emplacement was closely related to a set of fractures produced by a downward point-push (Anderson 1936, p.156).

From what is known of the Rooiberg 2 complex, it is believed that the following sequence of events, which is illustrated in figure 5, occurred: (i) The formation of a magma reservoir beneath where the complex is now found. The magma reservoir shown in figure 5A can be thought of as a cupola of a larger body of magma below and to the north of Rooiberg 2. In the reconstruction of the structural and emplacement history of the Rooiberg 2 complex it seems legitimate to start with a magma body as illustrated in figure 5A, but the reader is clearly justified in asking how the magma made room for itself in the crust and how the roof of the magma reservoir came to have the configuration illustrated in figure 5A. At this juncture it seems sufficient to state that the magma is believed to be composed of remelted crustal material and that the shape of the roof of the magma reservoir was mainly governed by piecemeal stoping and doming of the roof rocks. (ii) There was a falling off in the hydrostatic pressure of the magma in the cupola (probably due to the extrusion of lava to the west of the area - See Chapter 9) so that it became less than the lithostatic pressure in the surrounding rock. This resulted in the development of shear fractures in the roof rocks as shown in figure 5B. (111) Piecemeal stoping proceeded from the bottom upwards along these fractures, and the creation of a cavity was probably aided by a "sagging" (King, 1954A, p.83) of the central block. As the material was removed by stoping its place was taken by the material of unit B which, because it was emplaced as a relatively narrow sheet into cold country rocks, developed a porphyritic texture on cooling rapidly. (1V) A second lowering in the

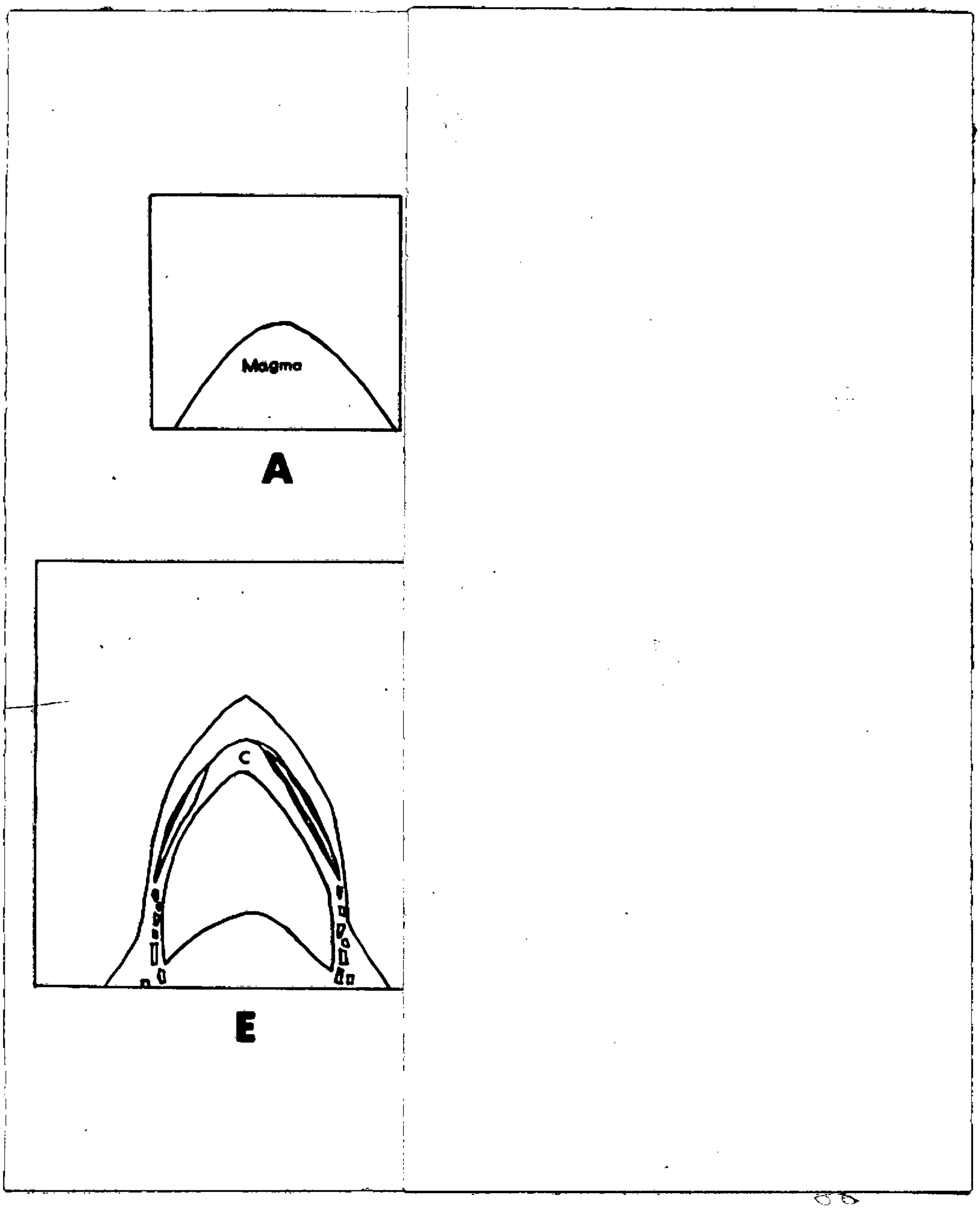


Figure 5: The formation of the Rooiberg 2 rink-dyke complex.

hydrostatic pressure of the magma led to the development of further fractures mainly along the outside of unit B. Piecemeal stoping took place along these fractures and this eventually resulted in the subsidence of a steep sided central paraboloidal block as shown in figure 5D. Ring A was emplaced into the space created by the two complementary mechanisms of piecemeal stoping and block subsidence. (V) While ring A was being emplaced fractures also developed on the inner side of unit B, and piecemeal stoping along these fractures assisted by further sinking of the central paraboloidal block relative to unit B, resulted in the emplacement of unit C. Unit B became a composite screen of porphyritic syenite and country rock as shown in figure 5E. (Vl) After a pause, during which time the sinking of the central block was probably obstructed by a mass of interlocking crystals in the cupola as suggested by Shand (1950A, p.927), cauldron subsidence occurred and the central paraboloidal block was replaced by a stock with the composition of unit D as is shown in figure 5F. **(Vll)** Later, at the close of the plutonic phase of igneous activity at Rooiberg 2, unit E (the Porphyritic Microgranite) punched and stoped its way through the centre of unit D. The removal of a central plug of unit D material was a prerequisite for the emplacement of unit E. The excavation of this material was probably assisted by the sand-blast effect of a fluidized system as the emplacement of unit E is believed to have been accompanied by a large volume of through-flowing gas, and according to Reynolds (1954) the presence of such gas is the main requirement for the development of a fluidized system. Another possibility is that a central plug of

Alaskitic Granite was removed by jet piercing (McBirney, 1959, p.445) which is a process whereby the intense heating of the roof rocks at the apex of a rising column of magma leads to the development of thermal stress within the intruded roof rock and thence to its disintegration. McBirney (1959, p.476) states that "growth will tend to produce cylindrical shapes, since a circular boundary is the most efficient natural conduction surface". In the emplacement of unit E as proposed above, and as shown in figure 5G, the emplacement mechanism is not directly related to the ring-dyke formation or cauldron subsidence that took place earlier in the same general area; thus one is forced to ask why was unit E emplaced at the centre of the Rooberg 2 ring-dyke complex and not somewhere else? As the earlier units of the Rooiberg 2 complex are believed to have been associated with a cupola, it is possible that this cupola may have represented a path of relatively easy access for the later Porphyritic Microgranite. The Rooiberg 2 complex might even represent the plutonic roots of a volcano, with unit E (with its fine grained groundmass) its choked conduit. Another solution to this question may be found in Grout's (1945, p.278) observation that:

"after a syrup has stood quietly for some hours a globule of lighter fluid in it rises rather slowly. If the first globule is promptly followed by another nearby, the second is 'guided' by the path of the first and moves a little more rapidly. The effect is especially strong if the liquids are miscible. This is distinctly reminiscent of field examples of batholiths that have risen into the crust where earlier batholiths must have been".

With the exception of ring-dyke B, the order of emplacement in the Rooiberg 2 complex has been from the outside inwards, which is the sequence King and Sutherland (1960, p.316) regard as normal.

In turning our attention from the ring-dyke complex of Rooiberg 2 to the Xaminxaip batholithic complex we turn from a small complex situated well within the epizone to a batholithic complex which was probably emplaced at the base of the epizone (Buddington 1959, pp. 676-677 and 694). Following Daly (1933) and Buddington (1959, p.675) the dividing line between a stock and a batholith is taken in the present study as being an outcrop area of 40 square miles.

The following is a summary of the main characteristics of granitic rocks emplaced into the epizone (After Buddington, 1959, pp. 677-680). (I) Their contacts are "largely or wholly discordant to the country rock" (p.677). (II) Contacts between plutons and the country rock are generally sharp (p.679). (III) Roof pendants are common (p.678). (IV) Most plutons are "effectively homophanous without lineation or foliation" (p. 678). (V) Most plutons are small in size (p.678). (VI) Small early units of complex plutons, and dykes, apophyses or small satellitic stocks related to larger igneous bodies commonly show chilled contacts with country rocks or porphyritic characteristics (p.679). (VII) "There is often a set of late-stage aphanitic or porphyritic dykes" and "associated lamprophyre dykes are also common" (p.679). (VIII) "Distinct pegmatite veins are typically rare" (p.679). (IX) In some batholiths "aplite or equivalent alaskite may be well developed" (p.679). (X) Miarolitic structure is common (p.679). (XI) Granophyre, in general, occurs exclusively in the epizone (p. 679). (XII) All the well known stocks and batholiths associated with ring-dykes and cauldron subsidence from New Hampshire, Oslo, Nigeria and S.W. Africa are believed to have been emplaced into the epizone (p. 680).

It is clear from the above that the Rooiberg 2 ring-dyke complex has most of the characteristics of an igneous body emplaced into the epizone. The main Xaminxaip complex, although displaying most of these features, has some features which are more characteristic of the Mesozone. For instance, Buddington (1959, p.696), referring to the work of Hans Cloos, states that the outer parts of granite plutons emplaced in the mesozone tend to contain localized areas of subvertical planar foliation, and that such planar structures are elusive or missing in the core. These subvertical planar structures, or schlieren, are clearly displayed in the outer border zone of the Alaskitic Granite in Donker Kloof (8C3). At the Donker Kloof (8C3) outcrop one finds an alternation of quartzo-felspathic and narrow ferromagnesian layers. These schlieren are interpreted as having been produced by an upward welling of magma. The Xaminxaip complex can thus be regarded as a body emplaced into the base of the epizone close to and in parts grading into the "transitional Epizone - Mesozone" (Buddington 1959, p.694). This probably represents a depth of emplacement of some 4 miles beneath the surface. The Rooiberg 2 complex, which is believed to have been emplaced into the true epizone, was probably emplaced at a depth of approximately 3 miles beneath the surface.

Before considering the emplacement of the Xaminxaip complex in detail, it is of interest to note that Precambrian ring-complexes like Rooiberg 2, which is clearly Pre-Nama (late-Proterozoic) have seldom been described in the literature. The Scottish complexes, the Medicine Lake Highlands Caldera of California, the Silverton Caldera of Colorado and the Quitman Complex of the Trans-Pecos, Texas, are all of Tertiary age. The New Hampshire, Oslo region, and South West African complexes are all Post-Tertiary to Devonian. The oldest ring-complex described in the literature is the Ahvenisto complex of Finland which was described by Savolahti (1956) and is believed to have been emplaced 1650 million years ago according to Rb-Sr, K-A and U-Th-Pb dating. A further two ring-dykes of probable Precambrian age have been described by Osborne (1934) from Quebec. The paucity of ring-dyke complexes as one goes back in time is to be expected if they were emplaced into the epizone and are thus relatively near-surface features which are soon attacked by the agents of denudation. The reason for the preservation of the Proterozoic Rooiberg 2 complex is not too difficult to discover as it was spared from the ravages of erosion by the onset of deposition (i.e. the laying down of possibly some of the Stinkfontein Formation, the beds of the Nama System and later the beds of the Karroo System). As a result of this deposition the crustal segment into which the Rooiberg 2 complex was emplaced has probably only been exposed to continuous erosion since late- or post-Karroo times, which is the same length of time as that to which the South West African complexes have been subjected to erosion (See Martin, Mathias and Simpson, 1960, pp. 156-174).

The Xaminxaip batholithic complex as it outcrops at present, was, emplaced at a deeper level in the crust than the Rooiberg 2 complex, and its constituent units have more complex outcrop patterns. Jacobson and others (1958, p.6) have observed that depth of erosion exerts a profound influence on the surface form of the northern Nigerian ring-dyke complexes, and that many complexes which appear as circular bosses at the present level would doubtless appear as ring-dykes at greater depth. The many different rock units of the Xaminxaip batholith are frequently found to have been emplaced along former curvilinear fractures and/or faults as can be seen in plate 7 and in the geological map accompanying this report. In the south eastern part of the batholith, between Donkerkloof and the Xaminxaip river (7C), the Alaskitic Granite is found to have been emplaced first, and it has been cut by the Porphyritic Syenite (see plates 7 and 8). To the west of this outcrop a small stock of Granular Syenite is found emplaced into the Porphyritic Syenite (6C4). Thus in this south eastern part of the batholith the Alaskitic Granite is found to have been emplaced before the syenitic rock types, which is the reverse of the proposed emplacement sequence found at Rooiberg 2 where the Alaskitic Granite cuts the syenitic units. The same sequence as found at Rooiberg 2 is found in the rocks to the immediate north of Xaminzaip (6A); but the reverse sequence in which the Alaskitic Granite is first, is found once again in the north of the batholith where the Alaskitic Granite is the dominant rock type. The field evidence seems to indicate that the Alaskitic Granite has been emplaced first over wide areas in the Xaminxaip batholith; but it has been emplaced after the syenites at Rooiberg 2 and in some parts of the Xaminxaip batholith. The above emplacement sequences are not as enigmatic as they seem at first sight, if one remembers that the Alaskitic Granite (which has by far the largest outcrop area of all the units of the Richtersveld Suite) is probably the parent magma from which the syenites and Porphyritic Microgranite were derived. It is perhaps relevant to note that Oftedahl (1953, p.61) in his study of the cauldrons of the Oslo region has postulated a "series of magmas changing in composition from granite to syenitic and again to granitic".

It seems clear from the evidence gleaned from the present study of the southern part of the Xaminxaip batholith that (at the present level of erosion) the batholith represents the coalescence of a number of smaller stocks and ring-dykes which were individually emplaced mainly by means of piecemeal stoping, large-scale stoping or cauldron subsidence, or more probably a combination of these mechanisms. In the field corroborating evidence in support of the compound nature of the Xaminxaip complex comes not only from the outcrop pattern made by the different component rock units, but also from the many "screens" of older rock found between these rock units. Tectonically controlled "permissive intrusion" (Mayo, 1941, p.1081.), "lateral magmatic wedging" (Pitcher and Read, 1958, p.298.), doming of the roof rocks (as is suggested in Chapter 9, Section C) and even diapiric intrusion, all probably played their part in the emplacement of the Xaminxaip batholith, particularly at deeper crystal levels, as the stoping mechanism is believed capable of explaining the rise of magma for only the last few tens or perhaps hundreds of feet (King, 1954 A., p.85). Read (1957, p.179.) has in face stated that the use of ring intrusion seems to have lowered the room problem by a few thousand feet, but has not solved it.
Because of the above considerations the two geological sections accompanying the map of the area were constructed to a depth of only 500 feet above sea-level, as below this depth the boundaries between the plutonic members of the Richtersveld Suite would be purely conjectural, and a conjectured

interpretation of what happens at depth is given in figure 5. As the two sections do not extend to any great depth they appear simple, but this simplicity is believed to be real as illustrated by the contact shown in plate 7.

When stoping is proposed as an emplacement mechanism it immediately poses the problem of what happens to the stoped material? Daly (1933) suggested that it was "incorporated or assimilated" at depth. Such incorporation or assimilation would have profound effects on the composition of the invading magma, thus it will be considered in the next section which is concerned with the genesis of the Richtersveld Suite.

(H) Petrogenesis:

(i) General Statement: The genesis of the plutonic rocks of the Richtersveld Suite presents the following major problems: (1) Are the rocks of the Richtersveld Suite of magmatic origin?

(2) How was the parental magma produced?

(3) How did the syenites originate?

(4) How is it that subsolvus and hypersolvus granites occur in juxtaposition in the Rooiberg 2 complex?

(5) How did the Porphyritic Microgranite come to have such a remarkably high K/Na ratio, and to contain muscovite veins?

(ii) Magmatic Origin of the Richtersveld Suite: The answer to the first of these questions, as to whether the rocks of the Richtersveld Suite are of magmatic origin, is believed to be in the affirmative. First of all it is clear from the earlier part of this chapter that the granites of the Richtersveld Suite belong to Raguin's (1946, p.254) granites of the "limited massifs" (massifs circonscrits) and such rocks are invariably of magmatic origin; or to put this statement in another way, most petrologists are prepared to admit that magma is required for the formation of ring-dyke complexes. Second, the sharp discordant contacts, the fine grain size of the groundmass generation of crystals in the porphyritic units, the even finer grain sizes of the crystals from the veins, dykes and apophyses that lead out from some of the plutonic rock units, the euhedral nature of the zircons, and the relative homogeneity of the different units, all support a magmatic origin for the rocks of the Richtersveld Suite. Third, the Alaskitic Granite, the Porphyritic Syenite and the Granular Syenite all fall within Tuttle and Bowen's (1958, p.129.) hypersolvus class (Group 1), and can thus be considered high temperature rocks which owe their origin to magmatic processes. The temperatures at which such rocks are believed to form are "higher than 660°C and perhaps as high as 900°C" (p.128). The Barth (1956, pp. 3-16) felspar geothermometer method (see Chapter 3) was also applied to the Alaskitic Granite and the two syenites; and the relative temperatures educed were all very similar, with a mean temperature of ± 850° ($\overline{X}k$ = 52/88). This temperature accords well with the temperature proposed above, and it is clearly in the magmatic range. Buddington (1948, p.41) has also commented on the felspars of magmatic rocks and he stated that "in all the rocks for which field evidence is best interpreted as indicating an origin by consolidation from agma the overwhelmingly predominant felspar is microperthite" (i.e. the main felspar found in the Alaskitic Granite). Tuttle (1952.B, p.113.) has made a similar statement. Fourthly, in recent years the existence of vast amounts of granitic magma in the crust has come to seem more probable as studies by Ross (1955, pp.427-434) and Larsen and Cross (1956, p.94) have indicated that volcanic rocks of rhyolitic, dacitic and quartz latitic compositions are present in many parts of the world in volumes which dwarf many batholiths.

(iii) Parental Magma: With regard to the genesis of the parental magma of the Richtersveld Suite there is a considerable amount of evidence to suggest that it was not the end product of classical fractional crystallization. For example, Waters (in Gilluley ed. 1948, p.107) following a long line of transformists both wet and dry has attacked the Bowen theory of granite formation by indicating that if we are to suppose that basalt results from the partial fusion of peridotite, and granite is the end product of the differentiation of basalt, "the energy problem is truly colossal", and it would be necessary to get "rid of 100 times as much partly fused peridotite and its crystal differentiates as we have granite". This argument is particularly valid in the Richtersveld where we have a large volume of granitic rock and no outcrops of basic rocks of similar age. Chapman and others (1935, p.518) showed in their paper on the evolution of the White Mountain magma series which is broadly similar in composition to the Richtersveld Suite (i.e. 78% granite and 20% syenite), that while it was qualitatively possible to produce the White Mountain magma series by fractional crystallization it was quantitatively impossible, as the production of each cubic mile of White Mountain granite would require the crystallization and differentiation of 16 cubic miles or original gabbroic material. Rastall (1947, p.30) in discussing granites in general has observed that the study of ore deposits associated with granites lends no support whatever to the theory of differentiation of granite magma from basalt, but rather the reverse. As there is no evidence to suggest that the Richtersveld Suite is in any way associated with large quantities of basic material, and as the field evidence suggests that below the present level of erosion the Adamellitic Gneiss is the typical rock type into which the Richtersveld Suite has been emplaced, it seems most probable that the Alaskitic Granite is of palingenetic origin, and has developed mainly as the result of the selective fusion of the Adamellitic Gneiss. Such an origin for the Alaskitic Granite might also help to account for the high Cl content found both in the Adamellitic Gneiss and in the majority of the members of the Richtersveld Suite. Smulikowski (1950, p.137) has discussed this question of the remelting of old granites to form new and states that selective fusion repeated several times, would increase gradually the concentration of granitophile elements in the younger granites. Bowen (1947, p.274) is not in favour of the formation of granite from earlier granitic rocks and states that "the derivation of granitic magma by refusion of granite is merely pushing the problem back into the misty past". This criticism is only true if one believes that there is a general granite problem (Read 1958, p.226), but in the present case the field evidence clearly indicates that there was an earlier more extensive granitic rock outcropping in and beneath the area. The overall picture that emerges is in many ways similar to the Donegal Granite Series (Read 1958, Pitcher and Read 1958, etc.) and Hunter's (1957) interpretation of the Precambrian of Swaziland. The investigators in both areas explicitly state that while the different granite units are genetically related, this does not imply that the emplacement of the different granite units took place in a short space of time. In both the above areas, and in the Richtersveld, it would appear that during the emplacement of the different granite units the crustal segment was rising. Thus to-day the early deep-seated and the later high level phenomena are recorded at the same crustal level.

In Chapter 9, section C an attempt will be made to reconstruct the full tectogenetic cycle of which the Richtersveld Suite is but a phase. The overall pattern that emerges from this reconstruction is found to be similar in broad outline to the cycle postulated by Rich (1951, pp. 1179-1222) for the "origin of compressional mountains and associated phenomena". If Rich's theory is modified and applied to the genesis of the parental magma of the Richtersveld Suite, the magma is seen as having formed as the result of heat (probably generated by radioactivity) concentrating more rapidly than it can be conducted away beneath and to the east of the area. This heat thus tended to concentrate in the north, north-west trending Soeties-Aurus belt that now contains the outcrops of the plutonic rocks of the Richtersveld Suite. At that time this belt is believed to have formed the margin to a cratonic area that extended to its east. The concentration of heat along this less stable marginal belt resulted in anatexis and the genesis of the parental magma of the Richtersveld Suite. The new magma is believed to have occupied a volume approximately 8 per cent greater (Rich, 1951, p. 1209.) than the material it replaced, thus producing a "magma blister" that domed and up-arched the crustal rocks above it. The most interesting effects of this process on the area bordering the blister are explored in chapter 9 section C. The changes in magma pressure required for the emplacement of the ring-dyke complexes along the crestal zone of the blister are believed to have been produced by the extrusion of lava in the manner described in chapter 9.

(iv) <u>The Syenites</u>: A perusal of the literature on syenite occurrences reveals that saturated and oversaturated syenites seldom occur in isolated intrusions but tend to occur as border facies, satellite bodies or apophysal dykes related to larger granitic masses; and that the manner in which the Richtersveld syenites occur associated with a larger mass of granite is typical of syenites the world over. Shand (1951, p.85) states that "the geological occurrence of syenite is such as to suggest that special conditions are required for its formation, and that syenitic magma is not normally present in the earth-crust". The concept has in fact developed that saturated and oversaturated syenites generally result from the desilication of part of a mass of normal acidic or granitic magma (Daly 1933, pp.477-478).

In order to see how the syenites differ from the supposed parental magma of the Richtersveld Suite the mean mode of all the Richtersveld Suite syenites has been compared with the mean mode of the Alaskitic Granite and it was found that the syenites show a marked decrease in quartz, a 37% increase in alkalic felspar and a 48% increase in ferromagnesian minerals. The same picture emerges if the Barth (1948, p.54) standard cell of the mean Alaskitic Granite ($\bar{X}$ 3 and 4, table VII) is compared with the mean standard cell of the syenites ($\bar{X}$ 15, 16, 18 and 19 of table VII).

	K	Na	Ca	Ba	Mg	Fe
$\bar{x}$ Granite	4.54	7.37	3.06	0.00	1.17	1.91
$\bar{x}$ Syenite	4.60	12.66	2.75	0.01	1.83	4.73
	-0.06	-5.29	+0.31	-0.01	-0.66	-2.82

	Mn	Al	Ti	Si	P	OH
$\bar{x}$ Granite	0.23	14.63	0.23	61.24	0.00	9.20
$\bar{x}$ Syenite	1.34	17.14	0.63	54.94	0.01	11.40
	-1.11	-2.51	-0.40	+6.30	-0.01	-2.20

From the above it is clear that to convert the mean Alaskitic Granite into the mean syenite Si must be subtracted K, Na, Fe, Mg, Mn, Al, Ti and water must be added in the relative amounts shown above. The most significant additions to convert the granite into syenite are of total iron (150% increase), manganese (480% increase), titanium (170% increase) and magnesium (60% increase). Shand's (1952, p.82) remarks on the genesis of the Plauen Syenite in which similar chemical features are found are particularly relevant. He states that such features "cannot be explained by settling or flotation of early crystals in the magma of the Meissen Granite, because alkali-felspar, the lightest mineral formed, would tend to float while diopside or hornblende would sink. The end product would then be a variety of granite enriched with alkali-felspar and impoverished with respect to dark silicates. ....In short, we seem compelled, in this and many other cases of similar character, to recognize a special process of differentiation by which the granitic magma was enriched in CaO, MgO and FeO, which combined with some of the available SiO_2 to form diopside or hornblende and this reduced the amount of silica available to form quartz".

The most likely "special process" in the case of the Richtersveld syenites would seem to be assimilation of roof and wall rock materials rich in Fe, Mn, Ti and Mg. The formation of new ferromagnesian minerals as a result of the assimilation process proposed above would also deplete the granitic magma in silica and thus reduce the amount available to form quartz. The above proposal poses three questions. (1) Where does the heat needed for the assimilation process come from? (2) What type of material was assimilated? (3) Is the silica depletion produced by the above process sufficient to account for the quantities of syenite currently found associated with the Alaskitic Granite? With regard to the question of heat Shand's

(1945, p.498) suggestion that the portion of magma which enters into reaction with its wall and/or roof rocks is but a small fraction of the whole, and that its temperature may be maintained for a long time by the outflow of heat from the main body of magma, seems valid in the present circumstances.

The second question which is concerned with the composition of the material assimilated can be tackled in two ways, first by an examination of the xenoliths found in the syenites, and secondly by studying the rocks into which the Richtersveld Suite was emplaced. Many inclusions are found in the syenites and they range in size from large screens to small inclusions only a few centimetres across. Most of the inclusions examined were found to be of metamorphosed Kheis supracrustal material (i.e. quartz sericite schists, metalavas (mainly acid to intermediate) and biotite schists). Inclusions of hornblendite were however found in ring-dyke A of Rooiberg 2. The small outcrop of Granular Syenite in Donker Kloof (8D1) was found to be closely associated with inclusions that were found to contain approximately 55% biotite, 32% white mica, 10% quartz, 3% opaque ore minerals. At Rooiberg 2 and at the southern end of the Xaminxaip batholith where the syenites predominate, the rocks of the Richtersveld Suite were emplaced into Kheis supracrustal rocks, the ultramafic bodies, the acid hybrid rocks and, to a lesser extent, into the Adamellitic Gneiss. In the northern part of the Xaminxaip batholith where the syenites are absent the batholith was mainly emplaced into the Adamellitic Gneiss. It thus seems that, as in the case of the Diana Complex of the Adirondacks (Buddington, 1939, p.105), the development of the Richtersveld Suite syenites was confined to those areas that contain mafic and ultramafic rocks. The depth to which the different parts of the Xaminxaip batholith have been eroded is also believed to have influenced the present-day distribution of the syenites. The rocks now exposed in the southern part of the batholith are believed to have been emplaced very close to its roof, and those outcropping in the northern part are considered to have been emplaced at greater depth within the batholith. As the incorporation of roof rocks (enriched in mafic materials) in the upper layer of the parental magma is considered to be of importance in the genesis of the Richtersveld syenites, erosion may thus be responsible for the absence of syenites in the northern part of the Xaminxaip batholith.

The answer to the third question as to whether the silica depletion produced by the assimilation of country rock is sufficient to account for the quantities of syenite currently found at Rooiberg 2 and in the Xaminxaip batholith, appears to be in the negative. Assimilation can probably account for part of the increase, at least, in Fe, Mg, Mn and Ti as most (if not all) the assimilated material is believed to be higher in these elements than the Alaskitic Granite. The bulk chemical compositions of the Kheis supracrustal rocks and the Adamellitic Gneiss are such that it is believed that while desilication due to assimilation may be locally significant (particularly in areas where amphibole and biotite schists, and hornblendite have been incorporated) in general desilication is quantitatively incapable of producing the volume of syenites now found associated with the Alaskitic Granite. If it is recalled that the syenites were emplaced into the epizone - the zone of fracture - then a start can be made at solving the desilication problem, as it is clear from an earlier section of this chapter that the emplacement of the syenitic rocks was preceded by a great deal of fracturing. This fracturing was accompanied by an increase in volume which is essentially equal to the voids created (Emmons, 1940, pp. 1 - 21). The voids were then filled by the more mobile portion of the intruding magma, and experimental data (Bowen and Tuttle, 1949, p. 459 and 1950, p. 490: Tuttle and Bowen 1958, pp. 90 - 91) indicates that this highly mobile escaping "rest magma" is probably richer in silica than the parent magma beneath it. If this removal of rest magma is on a large enough scale it would probably result, as Buddington (1939, p.160) and Emmons (1940 and 1953) have suggested, in silica depletion and syenite formation in the remaining crystallizing magma. As "the ratio in which the components of granite are transported in the vapour phase"....."favour silica over the felspars and Or over Ab" - Walton (1960, p. 641), the removal of more Or relative to Ab together with the silica is believed to help account for the lower mean K/Na ratio found in the syenites as compared to the parental (Alaskitic) granite.

When an attempt is made to view the genesis of these syenites as part of a much larger tectogenetic cycle (See Chapter 9, Section C), it seems probably that lavas were extruded contemporaneously with the emplacement of the plutonic rocks of the Richtersveld Suite, and thus the silica rich rest-magma may have been removed when these lavas were extruded. Recently, Boone (1962, p. 1470) has used the mechanism of the preferential transport of Si and K to account for the potassic felspar enrichment found in the syenites of the Deboullie district of Maine.

The above explanation of the evolution of the Richtersveld syenites is not the only possible one, and its main weaknesses would seem to be (i) that it requires two mechanisms, assimilation and silica escape to account for the main features observed, and (ii) that it does not really tackle the question of the huge increase in Mn and Ti found as one proceeds from the Alaskitic Granite to the syenites. With regard to these difficulties the paper by Reynolds (1946, p. 411) is particularly interesting as she found that at Portronan (Co. Donegal) a felspathic quartzite becomes strongly felspathized as the granite contact is approached, and that in the felspathized rock Na_2O reaches a geochemical culmination, whilst K_2O drops to a geochemical depression, and in addition TiO_2 and MnO reach geochemical culminations. In this same paper (1946, p.435) she states that many great batholiths are rimmed by "a zone of enrichment in alkalis, one or more of the cafemic constituents and one or more of the minor constituents, TiO_2, P_2O_5 and MnO, which travel ahead of a zone of granitization".

This view has been confirmed by Dennen (1951, p.556) who made a study of igneous contacts and discovered that "all the chemical features of Reynolds' scheme have been found reproduced on a minute scale at the contacts studied when appropriate rock types are considered".

Thus it is clear that the Reynolds' scheme could be applied to the rocks of the Richtersveld Suite if it was postulated that the roof rocks above the reservoir of parental granitic magma were subjected to the processes of granitization and this resulted in the formation of a capping or front low in free silica but enriched in Fe, Mg, Mn and Ti.

During a later episode the syenite capping may have become mobile and been emplaced in two main cycles, the first producing the quickly cooled Porphyritic Syenite, and the second the Granular Syenite which was probably mobile and subject to the physical and chemical influences of the magma chamber environment for a longer time than the rock of the first cycle. In her paper on the Newry batholith Reynolds (1947C, p.214) states that the basic front or zone of molecular desilication formed an outer aureole of enrichment in alkalis, and an inner one of enrichment in Fe, Mg and Ca; thus if two similar fronts surrounded the parental magma reservoir of the Richtersveld Suite, the Porphyritic Syenite would probably contain more mobilized material from the cafemic front than the Granular Syenite. The method of syenite formation postulated above has much in common with King's (1942, p.176) main syenite from the Cnoc Nan Cuilean area of the Ben Loyal complex which was believed to be the rheomorphic magmatic product in depth of the action of a succession of alkalumina emanations on a dominantly sedimentary series.

The writer believes that during the genesis of the Richtersveld Suite syenites all the processes that have been mentioned above operated and that the following generalized sequence of events took place: (1) prior to being assimilated the roof and wall rocks of the Richtersveld Suite magma chamber were subjected to the early stages of the granitization process and this resulted in their becoming relatively enriched in Fe, Mn, Ti and Mg; (2) some of these roof rocks which were enriched in mafic components were then assimilated, resulting in a significant local increase in the mafic content, and a decrease in the silica content of the parental magma; and (3) the escape, along fractures and faults, of a relatively silica-rich and to a lesser extent potassium-rich rest magma.

The main differences between the Porphyritic Syenite and the Granular Syenite are to be found in their textures, and in the volume and type of their dark minerals. The texture of the Porphyritic Syenite can probably be accounted for by its early emplacement into cooler country rocks which resulted in its rapid cooling and the production of its characteristic texture. The larger ferromagnesian content of the Porphyritic Syenite can also be accounted for by its early position in the emplacement sequence, as its parental magma was probably in more direct contact with the country rocks forming the magma reservoir's roof, and thus acquired more ferromagnesian material by assimilation. The incorporation of greater quantities of country rock would also tend to produce a more rapid cooling in the Porphyritic Syenite. The fact that the typical ferromagnesian mineral of the Granular Syenite is hornblende and that the Porphyritic Syenite is biotite is also considered significant, as it probably indicates that the former rock type formed under higher pressure, and probably slightly deeper crustal conditions (Boyd, 1959, p. 383).

Over most of the area in which they outcrop the Granular Syenite and Porphyritic Syenite tend to be relatively homogeneous; and this probably indicates that some mixing or homogenizing process was operative prior to their emplacement. The homogenization process was probably "mechanical kneading" (Reynolds, 1958, p. 382). A notable exception to the general homogeneity of the syenitic members of the Richtersveld Suite is the leucocratic border phase of the Porphyritic Syenite found along the northern banks of the Stinkfontein River (i.e. specimens 528 and 532). Not only is the colour index of this border phase low, but its felspar crystals tend to occur as euhedral to subhedral laths, being elongated and generally showing better crystal outline than is normally found in the plutonic members of the Richtersveld Suite. The low ferromagnesian content of this leuco-syenite can be accounted for in a number of ways, but the most probable appears to be the operation of a local winnowing process which retained the platy felspar crystals while the heavier ferromagnesian minerals sank to lower levels. Such a process would also account for the better crystal outline displayed by the accumulated felspar crystals as they would have grown suspended in a fluid medium. It also seem probable that the felspar crystals, like those of the Salem (Toulmin, 1960 pp 275 - 286) Syenite, collected on an outer lip of the batholith

Before considering the evolution of the granitic members of the Richtersveld Suite, the Alaskitic Granite-Porphyritic Syenite contact portrayed in plates 7 and 8 will be discussed as this contact was studied in detail and it is believed to provide valuable supplementary evidence on the petrogenesis of the syenites. Plate 7 shows the sharp curvilinear contact as it appears from a distance, and plate 8 shows an interesting part of the contact in more detail. Plate 8 is particularly interesting in that it is similar to figure 18 (p. 56) in Oftedahl's (1953) paper on the cauldrons of the Oslo region, which shows a contact between quartz porphyry and granite. The similarity is mainly the result of both contacts being crowded with small dark vein-like bodies. In the Oslo region the veins are of felsite and are believed (p. 56) to "represent crack fillings introduced after the consolidation of the quartz porphyry". In the Richtersveld the veins are dark grey (N3) and are found to be composed of alternating bands of fine grained crystalline and cryptocrystalline material. In the crystalline bands, microperthite, quartz, biotite and a little opaque ore were the only minerals to be identified. Analysis 26 of table VII is of one of these veins, and it is found to contain the following proportions of normative minerals 42% quartz, the same amount of felspar (42% Or_{25} Ab_{75} An_0) 2½% corundum, 9% diopside, 2½% ore minerals, and 1½% calcite. The composition of this specimen is of particular interest as if it, like its Oslo counterpart represents a crack filling introduced after the consolidation of the Porphyritic Syenite at that level in the crust, the material of which it is composed possibly represents the fluid which escaped from the magma crystallizing at depth. The texture of the veins supports this interpretation as it typically displays crustification and in parts a vague comb structure, both features being typical of cavity or fracture fillings (Bateman, 1950, p. 108). If, as suggested above, the dark acidic veins are indicative of the composition of the fluids escaping from the crystallizing magma then their silica rich nature is similar to that which was postulated as escaping from these magmas earlier in this chapter. The dominance of Na over K in this material is unexpected, as K is believed to be transported in preference to Na in the vapour phase (Walton, 1960, p. 641), but Orville (1960, pp. 104 - 108) has shown that the relative movement of K and Na is governed by thermal gradients if the alkali ions are free to diffuse between high and low temperature parts of a rock mass. The cracks which the dark rhyolitic veins fill tend to run parallel to the contact and are believed to result from the tendency of syenite magma, like granite magma, on crystallizing, to occupy less space than its melt (Johannsen 1932, p. 130). Plate 8 also shows a block of Alaskitic Granite engulfed within the Porphyritic Syenite.

(V) The Granites: We now turn to discuss the granites of the Richtersveld Suite. This topic has been left until after the discussion on the genesis of the syenites, because (as there is such a vast literature on the origin of granites) it was believed that our field of enquiry could be narrowed considerably if the associated syenites were considered first. The two main granites differ fundamentally from one another in that the Alaskitic Granite is a high temperature "hypersolvus" granite, and the Porphyritic Microgranite is a low temperature (IIC) "subsolvus" granite (Tuttle and Bowen, 1958, pp. 126 - 130). The evolution of the more common Alaskitic Granite will be considered first, and figure 10 (in Appendix 2) shows that it falls into the low temperature trough of the system $Na\ Al\ Si_3O_8$ - $KAl\ Si_3O_8$ - SiO_2 (petrogeny's residua system). This indicates that equilibrium between liquid silicate melt and crystalline phases played a significant role in the evolution of this rock type. This, however, does not imply that the Alaskitic Granite necessarily crystallized from the liquid residues of fractional crystallization, because as Turner and Verhoogen (1960, p. 348) have stated precisely similar chemical characters may be expected in acid melts formed by partial fusion of common sediments or igneous rocks. Walton (1960, p. 635) has expanded on this latter concept and from it he has developed his anatectic model. Recent experimental studies by Orville (1960, p. 104) have revealed that there is a strong possibility that rocks which have bulk felspar compositions corresponding to minimum melting compositions in the synthetic granite and ternary felspar system and are therefore commonly regarded as late magmatic differentiates, might also be produced by transfer of felspar compounds in solution through a vapour phase . The concept of petrogeny's residua system is still believed to lead to fruitful results even though the number of paths leading to it have been increased since Bowen proposed the concept in 1937. Recently Thornton and Tuttle (1960, pp. 664-684) have re-examined the concept in the light of new experimental data, and they have been able to state (after studying the results of nine new experimentally investigated systems) that Bowen's deductions regarding the consequences of crystallization hold for all of these systems (p. 669).

In section (iii) of this chapter the Alaskitic Granite was found to be of palingenetic origin, and the newly formed magma was considered to have made room for itself by distending its country-rock envelope, and later stoping and block subsidence assisted in its emplacement into higher crustal levels. The local textural variations found within the Alaskitic Granite are believed to be the result of local differences in its cooling history. Turner and Verhoogen (1960, p.65) have shown that "continual slow cooling essential for the development of the plutonic fabric is controlled only partly by depth. Of equal or greater importance, especially in the shallow plutonic masses, are shape of the magma body (particularly the ratio of volume to area of cooling surface) and temperature of host rock" The textural differences between D_1 (the outer part of Ring D of Rooiberg 2) and D_2 (inner ring) seem to indicate that D_2 cooled more rapidly than D_1. It is of interest to note that Kapp (1961, p.586) has stated that frequent changes of grain size and texture, porphyritic and brecciated facies, as well as pegmatitic patches and schlieren which are all features of the Alaskitic Granite, are signs of a near surface intrusion. Unit D_3 and the rocks immediately about the Porphyritic Microgranite are believed to have been modified by fracturing, crushing, and pneumatolytic and hydrothermal effects associated with the emplacement of the Porphyritic Microgranite. The glomeroporphyritic aggregates that occur in some of the porphyritic parts of the Alaskitic Granite are regarded as having been formed "during an early phase of crystallization afterwards rising or sinking through the magma or being broken away by the latter from their first place of attachment and carried up or along during a later stage of magmatic movement" (Holmes, 1930, p. 358).

The texture, mineralogy and geochemistry of the subsolvus Porphyritic Microgranite clearly sets it apart from the Alaskitic Granite. As this rock was emplaced after the main phase of Richtersveld Suite activity, and as its outcrop area is much less than that of the Alaskitic Granite, it seems probable that it crystallized from a late residual liquid derived from the crystallizing Alaskitic Granite. The marked increase in the Rb, Li, Pb and Tl content, the decrease in Cu, and the low K/Rb ratio (97) of the Porphyritic Microgranite are all geochemical trends that support this hypothesis (Nockolds and Allen, 1953; Taylor and others, 1956; Horstman, 1957; Shaw, 1957; Taylor and Heier, 1958A; Heier and Taylor, 1959; Volkov, 1959; Zlobin and Lebedev, 1960; and Butler and others, 1962). The Porphyritic sempatic texture of the Porphyritic Microgranite is of considerable interest, as (1) the large and often euhedral quartz and K-felspar phenocrysts appear to have grown for a considerable while suspended in a medium that did not inhibit the development of crystals with good crystal outline, and (2) the fine groundmass suggests rapid cooling. The Porphyritic Microgranite occurs in a stock with a mean diameter of 560 yards (510 m.) thus if the material composing the stock was emplaced into cold country rock it would probably have cooled rapidly. As the mean grain diameter of the groundmass crystals is 0.07 mm, only just greater than the 0.05 mm. diameter which the B.A. Committee (1936) recommended as the dividing line between rhyolites and microgranites, it can be seen that with regard to grain size this rock type is perhaps more akin to a porphyritic rhyolite than to a normal granite. The Porphyritic Microgranite is in fact similar to many of the "rhyolites" of the ring complexes from the younger granite province of northern Nigeria (Jacobson and others 1958). Jacobson and others (1958, p. 13) in discussing these rocks state that the late rhyolites are invariably porphyritic and usually the proportion of phenocrysts is of the order of 50% and sometimes greater. At the present level of erosion the Porphyritic Microgranite outcrops as a stock thus the designation microgranite is probably correct, though it is certainly within the realms of possibility that the present day stock was a conduit up which magma once flowed and crystallized at the surface as rhyolite. This suggestion would be in harmony with much of what is known about ring-dyke complexes from other areas, as many such areas contain subvolcanic plutons which are directly associated with volcanic rocks. The striking increase in the fluorine content of the rocks about the outcrop of the Porphyritic Microgranite probably indicates that volatiles from this source permeated the rocks about the central stock and this too could be most satisfactorily explained by the conduit hypothesis. It thus seems likely that the Porphyritic Microgranite evolved in the following manner. Towards the close of the crystallization of the Alaskitic Granite, a conduit formed, or was re-opened, by one of the methods described earlier in this chapter. This resulted in the fracturing of the rock about the conduit. As the conduit developed it filled with a residual granitic-rhyolitic magma which was probably charged with volatiles and carried euhedral alkali-felspar and quartz crystals. The volatites permeated the fractures about the conduit, particularly on its western side where the fractures are believed to have been more numerous and this permeation of the Alaskitic Granite by volatiles probably led to the formation of the distinctive fluorite rich granite of unit D_3. Eventually magmatic activity ceased and the conduit became choked with material which crystallized rapidly, because by this stage in the evolution of the Richtersveld Suite the events being described probably took place very close to the surface. On crystallizing the Porphyritic Microgranite probably contracted slightly particularly near its outer contacts, (i.e. crystallized granite magma tends to occupy less space than its melt - Johannsen, 1932, p. 130) and this produced a network of fractures linking the Porphyritic Microgranite with the great bulk of crystallized and crystallizing material beneath Rooiberg 2. These fractures are believed to have assisted in the development of the unusual mineralogy and chemistry that characterizes the Porphyritic Microgranite, as late solutions are believed to have travelled along these fractures and permeated the Porphyritic Microgranite.

As can be seen from Table IX the mineralogy of the Porphyritic Microgranite is very simple, with K-felspar, quartz and white mica being the only significant phases. The mean modal quartz content (wt.%) is 38.8% and thus falls within the limits set by Chayes (1950B, p. 148 and 1957, p. 58) for "normal" granites. This quartz percentage is however, slightly higher than that of the "average" New England Granite. The 24.0% white mica is also considered significant as such high percentages are decidedly uncommon in normal granites. Chayes (1952A, p. 212) has however, observed that "there does seem to be a relationship between mica and quartz .... quarts rich masses being relatively rich in mica". The largest amount of modal muscovite observed by Chayes (1952A) in a single specimen of granite was 12.2%. Yoder and Eugster (1955, p. 225) in their study of synthetic and natural muscovites have noted that muscovite is absent in those igneous rocks thought to have formed at high temperatures, is not common in hypabyssal salic igneous rocks, and is absent in extrusive rocks. By superimposing their stability curve of muscovite on Tuttle and Bowen's minimum melting curve of "granite" they (Fig. 16, p. 267) are able to show that above approximately 1500 atm. it is possible for muscovite to form on the liquidus surface of the granite system, but below that pressure muscovite can form only in the solid state. As the Porphyritic Microgranite is believed to have crystallized in a near surface environment, these data would seem to indicate that the white mica formed in the solid state. This evidence accords well with that gleaned from the felspar (i.e. the felspar showed that the Porphyritic Microgranite was a low temperature subsolvus, IIC, granite), thus the present mineral assemblage of the rock is considered to have developed at sub-magmatic temperatures. Yoder and Eugster (1955, p. 267) also draw attention to the fact that normative corundum can be used to estimate the amount of muscovite to be found in a particular granite. By constructing a histogram of the weight % of normative corundum in the 1640 analysed rocks of granitic composition given by Washington (1917), they (p. 267) found that the average normative corundum of all such rocks is 1.23% which amounts to about 4.6% muscovite. Approximately 9% of the analyzed granites have estimated normative anhydrous muscovite exceeding the highest value found by Chayes (1952A), and approximately 1% have estimated normative anhydrous muscovite contents exceeding that found in the Porphyritic Microgranite. Yoder and Eugster (1955, p. 269) also found that " perusal of Washington's compilation of chemical analyses of igneous rocks indicates that those granites in which K_2O is greater than $Na_2O + CaO$, are rich in muscovite". This observation is certainly true in the case of the Porphyritic Microgranite.

The high K/Na ratio of the Porphyritic Microgranite, and thus of the muscovite it contains, is of considerable interest in that Na is known to replace K in most muscovites and this replacement increases at increased temperature of muscovite formation (Berry and Mason, 1959, p. 612). This indicates that the muscovite found in this rock is probably a low temperature variety, though the low sodium content may indicate no more than that the muscovite formed in an environment deficient in sodium.

(vi) <u>Late Magmatic Features</u>: A number of most interesting irregular veins, or narrow dykes, were found within the Porphyritic Microgranite. The veins which are volumetrically insignificant, tend to be about 4 inches (10 cms.) wide, and are composed almost entirely of white mica, but carry accessory amounts of fluorite and quartz. The texture of this vein material is granular and not at all schistose as can be seen from specimen 571. A semiquantitative partial analysis of this material was made and it was found to contain ± 10% K, less than 1% Na, Li ± 8 p.p.m., Rb ± 300 p.p.m., Sn 20 p.p.m., Cu 3 p.p.m. Pbtr., Ga 60 p.p.m., Tl 3.7 p.p.m. The Sn, Ga and Tl values were all significantly higher than those found in the granites of the Richtersveld Suite. Descriptions of similar veins or dykes are seldom found in the literature but Morozewicz (1899, p. 215) has described a corundum-orthoclase pegmatite. Yoder and Eugster (1955, p. 269) believe that "on the basis of this occurrence, it seems reasonable to expect, in nature, dykes consisting mainly of muscovite". Schermerhorn (1956, p. 347) has found such muscovite veins in one of the Transconso Granites of Portugal. He believes that these veins developed from "rest fluids of magmatic derivation". A rest-magma of similar composition was proposed by Read (1931, pp. 162-163) to account for the muscovite and sillimanite in the Sutherland Granite.

Another interesting rock type is also found associated with the Porphyritic Microgranite. This rock type occurs on the knife-edge ridge to the immediate west of the Rooiberg 2 trigonometrical beacon, and is at the contact between the Porphyritic Microgranite and the Alaskitic Granite. In this locality a small patch of granite is found "with an elementary graphic structure" and within which "cassiterite is present as an accessory" (de Villiers and Söhnge, 1959, p. 253). The authors of the Richtersveld Memoir had some of this material spectrographically analysed for tin and it was found to contain

0.35% Sn. De Villiers and Söhng (1959, p. 253) also state that "further investigations of this locality .... indicated that there were no local concentrations of the mineral (cassiterite) at the surface and that there was no justification for further expenditure on exploration for commercial tin ore". The evidence obtained during the present investigation is in complete agreement with the above statement. The tin contents of both the Porphyritic Microgranite and the Alaskitic Granite (from ring D and from the Xaminxaip batholith) were determined, as it was believed that if a significantly higher tin content was found in the Porphyritic Microgranite it would enable any future prospecting to be limited to this rock type. The results, however, proved negative as the tin content of all the granites studied including the much earlier Adamellitic Gneiss, was found to be similar. The occurrence of 0.35% tin even though in a very localized area is of considerable geochemical significance as Ringwood (1955, p. 198 and 244) and Onishi and Sandell (1957, p. 263) have shown that Sn^{2+} and Sn^{4+} both tend to accumulate in residual magmas. To summarize, the presence of fluorite, muscovite veins and the cassiterite bearing micropegmatite within the associated with the Porphyritic Microgranite all support the hypothesis that the Porphyritic Microgranite was fractured and permeated by late volatile rich (pegmatitic) solutions. The presence of F and OH in the late solutions is of particular interest as they are both well known mineralizers (Buerger 1948, p. 746) which increase the fluidity of the silicate melts by breaking down (Si-o-Si) oxygen bridges and causing "the magma to undergo a transition from a glassy state to a liquid state".

The genesis of the unusually high K/Na ratio (66) of the Porphyritic Microgranite poses a number of questions. Von Eckermann's (1960, pp. 519-528) Borengite which was discussed earlier in this chapter is one of the few igneous rocks with a K/Na raio greater than that found in the Porphyritic Microgranite. A bostonite from New Haven, Orkney was found to contain 11.34% K_2O and 1.66 Na_2O (See Summary of the Progress of the Geological Survey of Great Britain, 1928); and Stringham (1953, pp. 945 - 991) has described a small stock from Bingham, Utah where some of the granite is exceptionally K-rich. Kathove (1949, pp. 467 - 470) Kennedy (1955, p.499) and Schermerhorn (1956, pp. 329-348) have shown that many potash-rich rocks, particularly granites, have acquired some of their potassium from post-crystallization reations (e.g. microclinization). A perusal of the literature, however, indicates that the rhyolites of the Permian and Triassic Koipato group of north-western Nevada have the most in common with the Porphyritic Microgranite. Tatlock (1962) in his description of these rocks states that they were originally derived from a magma which crystallized as a leucogranite containing 5.8% K_2O and 3.0% Na_2O. He then states that the upper half of the volcanic pile was derived from a later potassic differentiate (6.8% K_2O, 1.2% Na_2O) of the leucogranite magma. Later, "K-rich fluids associated with the potassic differentiate infiltrated the pile including previously albitized zones as well as parts of the leucogranite pluton, accomplishing a nearly pervasive replacement of albite by K-felspar. The effect was to greatly enrich the pile in K at the expense of Na; an estimated 40 percent of the pile contains 6.0% to 13.0% K_2O and $<$ 0.5% Na_2O". As previously stated, it is believed that broadly similar processes were at work in the formation of the Porphyritic Microgranite (as it occurs today).

The second part of the above hypothesis is not without experimental backing, as recent studies by Orville (1960, p. 104) have established that "ion - exchange reactions take place rapidly and reversibly between alkali-bearing hydrous solutions and alkali felspars at temperatures greater than 300°C and pressures ranging from a few to thousands of atmospheres". His (p. 105). investigations of the temperature dependence of the alkali felspar - vapour ion-exchange reaction indicates that "in the simplest instance, alkali-bearing vapour in equilibrium with two alkali felspars at high temperature will, on cooling, be capable of replacing a certain amount of Na felspar by K felspar in accordance with the reaction Na felspar + $K^+ \rightarrow$ K felspar + Na^+"

As Orville shows his data are in agreement with observations made in thermal hot spring areas where Na felspar is commonly replaced by K-felspar.

The Porphyritic Microgranite is thus believed to be the product of two processes: (1) the rapid crystallization of a late granite magma that contains euhedral quartz and felspar crystals, and (2) the infiltration into the crystalline Porphyritic Microgranite of K-rich fluids that generated the K-felspar and muscovite rich rock found today.

(I) Resume: Two outcrops of plutonic rocks belonging to the Richtersveld Suite are found in the area. They are the Rooiberg 2 and Xaminxaip outcrops and together with other outcrops of the Richtersveld Suite they form a north, north-west trending belt that extends for 120 miles from Soeties in the south to Aurus in the North. The Rooiberg 2 outcrop is made up of four concentric ring-dykes, A, B, C and D, and a central stock (unit E). The Precambrian age of this ring-dyke complex is of interest as similar complexes from other parts of the world are generally much younger in age. Rings A and C are found to be composed of Granular Syenite, Ring B of Porphyritic Syenite, Ring D of Alaskitic Granite of three sub-types, D_1, D_2 and D_3, and unit E is found to be composed of Porphyritic Microgranite. Rock types similar to all the above units, except the Porphyritic Microgranite, are found in the Xaminxaip batholith. Petrographic and geochemical data necessitated a revision of the earlier correlation of the units of Rooiberg 2 with those of the Xaminxaip batholith. Contacts between the different rock units are sharp and generally curviliner and thus appear to indicate that arcuate fractures played a significant role in the emplacement of these rock units. The Alaskitic Granite, Granular Syenite and Porphyritic Syenite were all found to belong to Tuttle and Bowen's (1958) hypersolvus salic rock group and are thus interpreted as being high temperature magmatic rocks. Evidence from rock textures, grain sizes, mode of emplacement and the nature of contacts all support the magmatic hypothesis. The type of felspar and the muscovite content of the Porphyritic Microgranite indicate that its present mineral assemblage developed at low temperatures. The Alaskitic Granite, Granular Syenite and Porphyritic Syenite were all found to be significantly high in Cl and Mn and low in Li. Both syenites are also significantly low in P_2O_5 and have abnormally high K/Rb ratios ($\bar{X}$ = 454). Their ratios are double those (230) now accepted as normal. The Porphyritic Microgranite is significantly enriched in K, Rb, Li, Cs, Pb and Tl and is depleted in Na and Cu. The emplacement sequence at Rooiberg 2 is believed to have been Porphyritic Syenite — Granular Syenite — Alaskitic Granite — Porphyritic Microgranite, but in the Xaminxaip batholith the Alaskitic

Granite, which is considered to be similar in composition to the parent magma from which these rocks are derived, was emplaced first over wide areas. At the present level of erosion the Xaminxaip batholith is believed to represent the coalescence of a number of smaller stocks and ring-dykes. The Alaskitic Granite is interpreted as being of palingenetic origin, and as having developed from the selective fusion of the Adamellitic Gneiss. The textural variations found in the Alaskitic Granite are believed to have resulted from local differences in its cooling histroy. The development of the syenites from the granitic parent magma required the removal of Si, slight additions of K, Na and Al and significant increases of Fe, Mg, Mn and Ti. Desilication was produced by both the assimilation of the more basic roof and wall rocks and the escape of silica rich volatiles. Prior to assimilation the roof and wall rocks are believed to have been subjected to the early stages of the granitization process and this has resulted in their containing higher than normal concentrations of Fe, Mn, Ti and Mg. It is considered significant that the syenites are confined to the southern part of the Xaminxaip batholith as (1) the rocks exposed in this area are believed to represent material emplaced near the roof of the batholith, and (2) ultra-mafic and mafic country rock is more common in the south of the batholith. The petrography and geochemistry of the Porphyritic Microgranite clearly set it apart from the other rocks of the Richtersveld Suite and they indicate that it probably crystallized from a residual granite-rhyolitic magma that was charged with volatiles and carried alkali-felspar and quartz crystals. This magma cooled quickly in a near surface environment. It then fractured, and the rock was permeated by K-rich fluids that gave it its unique K-rich character and also resulted in the formation of muscovite rich veins.

VI QUARTZ BOSTONITE AND HORNBLENDE DIORITE DYKES

(A) Introduction:

Rogers (1915, p. 97) was the first to mention the numerous dykes found in the Richtersveld. In the south-eastern Richtersveld these dykes can be divided into two main swarms on a basis of their width, composition, and orientation. Both of the dyke swarms cut the Alaskitic Granite and the Porphyritic and Granular Syenites and were thus emplaced after the main phase of Richtersveld igneous activity. The rocks of the Nama System lie unconformably above these dykes and thus were laid down after their emplacement. Haughton and Frommurz's (1936) map, however, shows bostonitic dykes cutting the "Numees Beds" at Nabas (or Chamgabmund: See map 2). De Villiers and Söhnge (1959, p. 181) believe that sediments similar to these Numees Beds which are found on the western side of the Orange River belong to the Nabas Stage of the Kuibis Series of the Nama System. The bostonitic dykes are thus seen to cut, and are younger than, the lower beds of the Nama System. The present writer examined a specimen of one of these dykes from Nabas (which Prof. H. Martin kindly supplied) and it was found to differ considerably in appearance from the quartz bostonites of the area studied. It is thus suggested that the Nabas bostonites are probably genetically related to the Tatasberg Pluton (see Söhnge and de Villiers, 1948) which outcrops very close to them and that they do not belong to the Richtersveld Suite.

The majority of the wide Hornblende Diorite Dykes strike due north and are most densely clustered about the central meridian of the area (5A, 5B, 5C, 5D, 5E, 5F, 5G, 4F, 4G). These "wide" dykes were studied throughout the whole of the Richtersveld and all were found to have similar strikes. The main direction of strike of the narrow Quartz Bostonite Dykes is found to vary between N 10°E and N 30°E with a slight maximum at N 30°E. These dykes are most common in the south and particularly the south-eastern part of the area. The mean width of the Hornblende Diorite Dykes, as found in the area, is approximately 12 ft. (4 m.), but if the whole Richtersveld is taken into consideration they are found to range in width "from a few feet (or less) to over 1 mile" (de Villiers and Söhnge, 1959, p. 139). The Quartz Bostonite Dykes are generally less than a yard (1 metre) wide, many are but a few inches wide, and the mean width of these dykes is approximately 18 inches (46 cms.). From their width it is clear that many of the Quartz Bostonite Dykes do not show up on aerial photographs. Thus while an attempt has been made to map as many of these dykes as possible, it has not been possible to map them all. De Villiers and Söhnge (1959, pp. 138 - 145) include both of these dyke swarms in their "dykes older than the Kuboos Igneous Complex" category. The Hornblende Diorite Dykes were included in their "mafic and ultramafic" dyke group, and the Quartz Bostonite Dykes in their "diabasic and felsic dyke group". The latter group is particularly large and all embracing, and it contains rocks ranging in composition from aplite to dolerite and ranging in time from pre-Stinkfontein formation to post-Kuboos. The Richtersveld Memoir authors (p.141) state that "there was neither opportunity nor need to assign the countless dykes to their respective groups".

With regard to the relative time of emplacement of the Quartz Bostonite and Hornblende Diroite Dykes, de Villiers and Söhnge (1959, p. 138) state that the "diabasic and intermediate dykes" .... "cut the mafic dykes". During the present study a number of locaities at which the two swarms intersect were visited. In all localities exposures were poor, but at none could a Quartz Bostonite be found cutting the Hornblende Diorite Dykes. As de Villiers and Söhnge's (1959) "diabasic and intermediate dykes" included dykes of many different compositions and ages, it seems likely that the examples they found of rocks of this group cutting the dioritic dykes were taken from outside the area being studied here in detail, and that the intersecting rock was not a Quartz Bostonite.

(B) <u>Quartz Bostonite</u>:

(i) <u>Field Description</u>: The Quartz Bostonites (Bo.) are here defined as dyke rocks which mineralogically and texturally include both microsyenites and intrusive trachytes (quartz less than 20%, alkalic felspar greater than 2/3 total felspar, colour index generally between 0-20 i.e. Hatch, Wells and Wells, 1949, p. 246). De Villiers and Söhnge (1959, p. 144) did not make a petrographic study of these dyke rocks but after considering their characteristic colour of dark grey (N3), speckled with pinkish grey (5YR 8/1) and medium dark grey (N4) phenocrysts, they concluded that "the vast majority are fine-grained dark rocks of diabasic appearance". This diabasic appearance is highly deceptive as petrographic study invariably reveals the rocks to be Quartz Bostonites. (It might perhaps be noted that the term dolerite means "a deceptive rock" - Tyrrell 1929, p. 120). Most of the Quartz Bostonite Dykes have a vertical dip, but some dip steeply to either east or west as can be seen from their outcrop traces on the contoured geological map accompanying this report. In contrast to the Hornblende Diorite Dykes the finer grained Quartz Bostonite Dykes tend to be more resistant to weathering than their host rocks. Characteristically the Quartz Bostonites break into flaggy bladed fragments, on weathering.

(ii) Petrography: The grain size and texture of the Quartz Bostonite Dykes varies considerably and these variations have their repercussions on the determination of the modal composition of these rocks as (1) the smaller sized minerals in the finer grained specimens pose problems of identification, and (2) the Holmes and Elliott effects become significant, particularly when dealing with the minute ore minerals. Another problem was posed by the frequent appearance of extremely narrow veinlets of secondary quartz and calcite. The mean modal (vol.%) composition, and the standard deviations of the individual mineral species, of 16 fresh specimens of the Quartz Bostonite was found to be as follows:-

	Alkalic Felspar	Quartz	Opaque ore Minerals	Biotite	Calcite	Hornblende
$\bar{X}$	74.2	4.5	3.0	13.3	2.4	0.9
S	10.1	3.5	1.8	9.6	2.1	-

	Epidote Group	White Mica	Zircon	Chlorite	Plagioclase	Apatite
$\bar{X}$	0.8	0.7	0.1	0.1	Tr.	Tr.
S	-	-	-	-	-	-

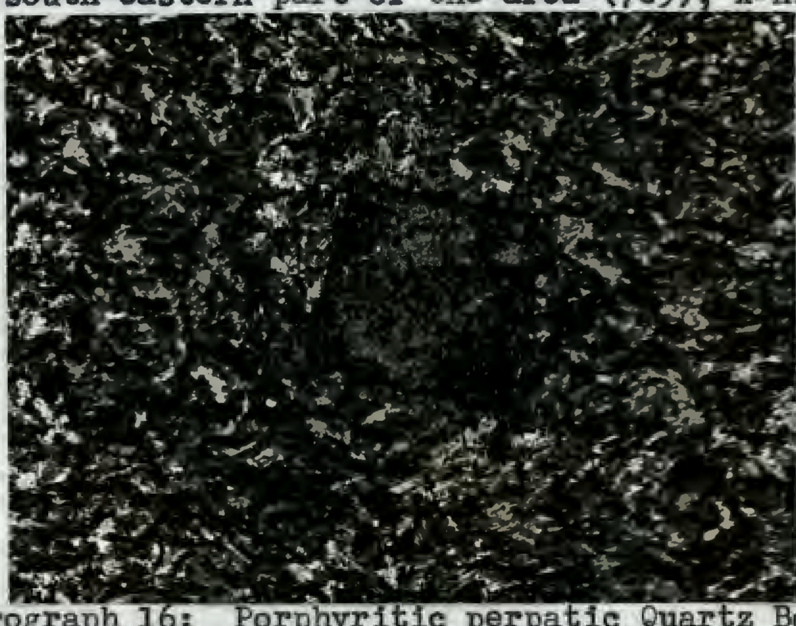
	Fluorite
$\bar{X}$	Tr.
S	-

From the above it can be seen that the Quartz Bostonites are more variable and tend to be richer in quartz, the opaque ore minerals and secondary minerals like calcite, than the plutonic syenites of the Richtersveld Suite (i.e. the mean syenite contains 2.3% quartz, 1.1% opaque ore minerals and 0.4% calcite). In photomicrographs 14 - 17, which are typical of the different textural types of Quartz Bostonites discovered in the area, it is found that as the minerals in the groundmass increase in size the texture of the rock changes from porphyritic perpatic with a pilotaxitic groundmass in which the microlites have a mean thin-section area of 0.0016 sq. mm., to a similar rock with a sub-trachytoid groundmass texture. This second rock type grades into a rock with a granitic groundmass texture and finally, in the coarsest grained specimens, the texture is uniform and granular with a mean thin-section grain area of 0.71 sq. mm. (see Photomicrograph 17). The phenocrysts, which are generally euhedral to subhedral perthite crystals, have a mean thin-section area of 1.89 sq. mm. Microperthite is both the most common phenocrysts and the dominant groundmass mineral. In the coarser specimens the perthite crystals frequently show simple Carlsbad twinning. The optical properties of some of the more homogeneous microperthite crystals were determined. The mean 2Vx was found to be 83°; and refractive indices of X = 1.520, Y =1.524 and Z = 1.529 were found to be typical of these crystals. Fractures found in some of the phenocrysts were observed to contain material similar to that of the groundmass, thus indicating that the phenocrysts fractured prior to the completion of crystallization of the groundmass.

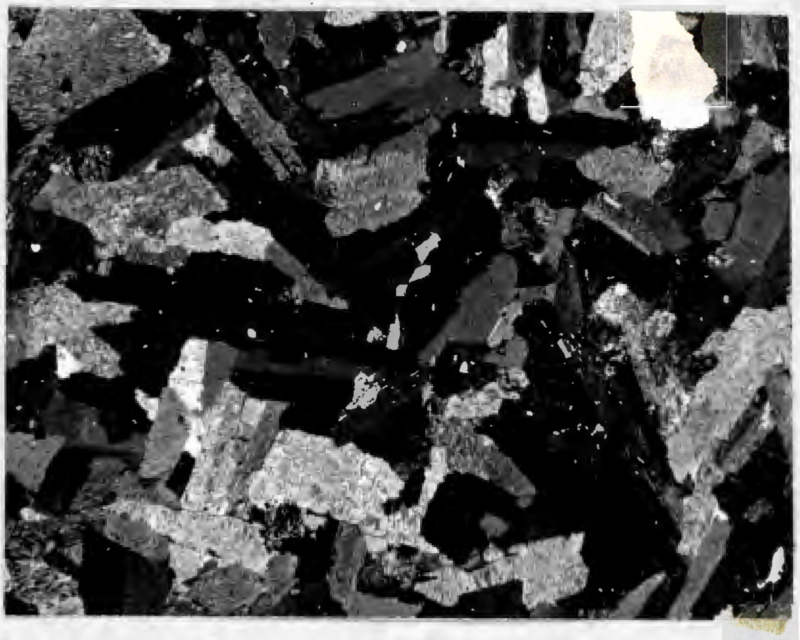
The Variations in Texture and Grain Size found in the Quartz Bostonites

Photomicrograph 14: A Quartz Bostonite showing its pilotaxitic groundmass texture (X15), Specimen No. 516, from north of Rooiberg 2 (3D7), X-nicols.

Photomicrograph 15: A Quartz Bostonite showing its subtrachytoid groundmass texture (X13.5) Specimen No. 417, from the south-eastern part of the area (7G5), x-nicols.



Photomicrograph 16: Porphyritic perpatic Quartz Bostonite with a granitic groundmass texture (X13.5), Specimen No. 467, from the eastern part of the area (7G1), x-nicols.



Photomicrograph 17: Quartz Bostonite (X13.5) with a granular texture, Specimen No. 275, from Xaminxaip (6A8), x-nicols.

Biotite and hornblende, the dominant ferromagnesian minerals, were in most specimens too small to study, but in the coarser grained specimens the hornblende was found to have the following optical properties X = dusky yellow (5Y 6/4) Y = moderate yellowish green (10GY 6/4), Z = 1.673 = greyish green (10G 4/2), and $Z_{\wedge}c = 15^{o}$. In most specimens a little clear quartz is found to occur in the interstices between the felspar grains. The groundmass of the finer grained specimens is generally "peppered" with minute opaque ore minerals which tend to have a fairly uniform distribution. Similar fine grains of ore minerals are found included (sometimes in regular arrangements) within the microperthite crystals of some of the coarser grained specimens. Veinlets of calcite, quartz and very rarely fluorite cut across the Quartz Bostonites. Accessory amounts of pistacite, zircon and apatite are found sporadically distributed throughout these rocks.

The epidote and calcite found in some specimens appears to fill vesicles. If this interpretation is correct .this indicates that these particular dykes were emplaced in a near surface environment.

If the grain-size of the different Quartz Bostonite Dykes is considered in relation to their field distribution it is found that grain-size tends to be related to proximity to the plutonic members of the Richtersveld Suite; that is, those dykes that are believed to have been emplaced deepest within the Xaminxaip batholith (i.e. Specimen 275) are the coarsest grained, and those emplaced into earlier rocks away from contacts with the plutonic rocks of the Richtersveld Suite, are the finest grained. These observations, together with the fact that the coarser grained Quartz Bostonite Dykes do not decrease in grain size at their margins, would seem to indicate that the dykes were emplaced at a time when the Richtersveld rocks were still relatively hot. The curvilinear outcrop traces of some of the coarser grained dykes would seem to indicate that the fracture planes into which they were emplaced developed in embryo before the plutonic units of the Richtersveld Suite had completely consolidated, and the dykes were dragged from their initial position by the final movements in the plutonic rocks of the Richtersveld Suite. A similar suggestion has recently been made by Pitcher and Read (1960, p. 53) in their study of the dykes associated with the Donegal granites, but the idea appears to have been first proposed by Balk and Grout (1934) in their study of the embyonic joints in the plutonic rocks of Montana. It might be postulated that the coarser grained Quartz Bostonite Dykes which are characteristically found emplaced into the Richtersveld

Suite and the finer grained dyke rocks, are not of the same age, but this suggestion does not have much to recommend it as both dyke groups have similar compositions and orientations.

(iii) <u>Chemistry</u>: By comparing the Quartz Bostonite (analyses 1 and 2 of Table X) with Turekian and Wedepohl's (1961) average syenite (analysis 3) and Daly's (1933, p. 27) average bostonite (analysis 4), it is apparent that the Quartz Bostonite is "normal" with respect to SiO_2, TiO_2, CaO, H_2O^+, P_2O_5 and Cs (Na and K are variable but normal). The Al_2O_3 content is slightly low, Fe_2O_3 and MgO are slightly high, and FeO and MnO are significantly high. The high iron and manganese content of these rocks is a feature they share with the syenites of the Richtersveld Suite. The Lithium content of the Quartz Bostonite is significantly low, but this is to be expected as the specimen analysed contained only 5.8% (vol. %) hornblende and no phyllosilicates. Low Li values are also characteristic of most plutonic members of the Richtersveld Suite. The rubidium content of the Quartz Bostonite is also significantly low. Its K/Rb ratio of 679 is higher than that found in any of the plutonic rocks of the Richtersveld Suite and it falls without "the limits of normal K/Rb ratios" (Taylor, 1960,B, p. 318). In Figure 12 (in Appendix 2) the Quartz Bostonite is found to fall in the centre of "area 3", (Taylor and Heier 1958A, pp. 202-203) and, as we have already discovered, the syenitic rocks of the Richtersveld Suite are also found to be associated with "area 3". All that was said about the abnormal K/Rb ratios of the Richtersveld Suite syenites is thus equally applicable to this rock type.

TABLE X - CHEMICAL DATA (QUARTZ BOSTONITE)

Composition, weight per cent: Norm:

No.	1	2	3	4		5
SiO_2	61.56		62.43	61.32	Q	16.20
TiO_2	0.82		0.58	0.89	Or	26.13
Al_2O_3	15.25		16.63	18.43	Ab	34.58
Fe_2O_3	4.29		) 4.72	3.84	An	4.17
FeO	3.89		)	1.60	C	2.14
MnO	0.25		0.11	0.01	Hy($MgSiO_3$	3.20
MgO	1.28		0.97	0.46	($FeSiO_3$	2.64
CaO	1.68		2.52	1.45	Cc	1.50
Na_2O	4.10	6.45	5.45	5.75	Mt	6.26
K_2O	4.40	5.80	5.79	4.94	Il	1.52
H_2O^+	1.12			1.31	Ap	tr.
H_2O^-	0.12				Misc.	1.24
P_2O_5	0.16		0.18		Total	99.58
CO_2	0.63					
Li (p.p.m.)		3 p.p.m.	28 p.p.m.			
Rb (p.p.m.)		71 p.p.m.	110 p.p.m.			
Cs)p.p.m.)		<2 p.p.m.	0.6 p.p.m.			
Total	99.55			100.00		

1. Analysis of Quartz Bostonite Dyke 417: Photomicrograph 15; Mode (Vol.%), Perthite 73.0%, Quartz 8.9%, Hornblende 7.9%, Bioti e 3.0%, Opaque Ore Minerals 5.1%, Calcite 0.3% Epidote Group 1.7% and Zircon 0.1% - Analyst E.C. Haumann.

2. Partial analysis of Quartz Bostonite Dyke 275: Photomicrograph 17; Mode (Vol.%). Perthite 92.4%, Hornblende 5.8%, Opaque Ore Minerals 1.6%, Zircon 0.1% and Calcite 0.1%.

3. Turekian and Wedepohl's (1961) Average Syenite.

4. Daly (1933, p.27) Bostonite (Average of 5).

5. Norm of specimen 417 (analysis 1).

From the norm of specimen 417 (Table X, column 5) the composition of its perthitic felspar was found to be Or_{40} $(Ab_{89}\ An_{11})60$. The composition of the perthite of specimen 275 was calculated from its mode and the partial analysis, and was found to be Or_{38} $(Ab_{98}\ An_{2})62$. The mean of the two above values is Or_{39} $(Ab_{93}\ An_{7})61$ and is a figure that agrees closely with the mineralogical data presented earlier in this chapter.

(C) Hornblende Diorites

(i) Field Description and Petrography: The wider Hornblende Diorite Dykes were sampled and examined not only in the area mapped but all over the Richtersveld - from de Hoop and the Pokkiespramberge in the north to Stinkfontein in the south. As indicated by their linear outcrops shown on the accompanying geological map the dip of these dykes is essentially vertical. They also tend to persist over considerable distances in their strike direction. De Villiers and Söhnge (1959, p.139) traced one of these dykes along strike for a distance of 36 miles. The mean mode and the standard deviations of the individual mineral species, of 8 specimens of the Hornblende Diorite is as follows:

	Plagioclase	Alkali Felspar	Hornblende	Epidote Group	Biotite	Chlorite
$\bar{x}$	37.3	13.6	17.0	6.8	8.3	7.6
s	9.2	8.8	4.6	4.1	9.0	3.7

	Opaque Ore Minerals	Leucoxene	Apatite	Serpentine	Calcite
$\bar{x}$	5.5	1.7	0.9	0.6	0.4
s	2.8	1.8	-	-	-

	Quartz	White Mica	Sphene
$\bar{X}$	0.3	Tr	Tr
s	-	-	-

(Titanaugite occurs in a few of the specimens examined but not in the eight specimens mentioned above)

The high ferromagnesian content, the presence of carbonate minerals and shearing have all assisted in making these rocks highly susceptible to weathering. The relative ease with which this rock type weathers has had a definite effect on the topography of the area, as these dykes tend to produce narrow furrow-like depressions in the areas of subdued topography associated with the pre-Richtersveld Suite rocks, and notches in the crestlines of the hills and mountains of the more resistant Richtersveld Suite plutonic rocks. The mineral composition of some of these dyke rocks is found to vary considerably from the mean mode but a closer examination of such anomalous rock types generally reveals that they are closely associated with partly digested xenoliths. The overall picture obtained is that these dyke rocks are essentially very similar throughout the Richtersveld. Specimen 389 from the Pokkiespram hills north of Sendlingsdrif (See Map 2, North of Numees) showed the greatest variation from the mean mode.

An interesting textural feature observed in the Hornblende Diorites cutting rocks older than the Richtersveld Suite, was the variation in grain size as one proceeded inwards from their margins. This gradation in grain size is particularly well displayed by the largest dyke in the Richtersveld which was sampled at De Hoop (See Map 2). Specimens 358, 359, 360 and 392 were collected in a traverse across this dyke, and their mean modal composition (Vol.%) was found to be: Plagioclase 35.3%, Alkalic felspar 18.1%, Hornblende 18.9%, Chlorite 9.1%, Opaque Ore Minerals 7.8%, Minerals of the Epidote Group 5.1%, Biotite 2.5%, Apatite 1.7%, Serpentine 0.9%, Calcite 0.3% White Mica 0.1%, Sphene 0.1%, Leucoxene 0.1%, Quartz Tr. The mean thin-section grain area of the felspars in these specimens was found to vary from 0.179 sq. mm. (Specimen 358) near the dyke's outer contact to 4.2 sq. mm. towards the centre of the dyke (Specimen 392). This grain size variation indicates that chilling took place, and this in turn suggests that these are dilation dykes that crystallized from magmatic material. The discovery that this chilling effect is more clearly displayed in dykes that outcrop away from the main masses of Richtersveld Suite plutonic rocks probably indicates that these dykes were emplaced before the plutonic rocks were able to cool to any appreciable extent.

Further evidence in support of such a magma is provided by a dyke found near the summit of Klein Helskloof Pass in which the felspars have a mean thin-section grain area of 0.095 sq. mm. Another interesting observation was made by de Villiers and Söhnge (1959, p.142) who discovered that in the Mount Erebus (See figure 1, de Villiers and Söhnge grid 3B) area the Hornblende Diorite Dyke which persists over the greatest strike distance (the Gannakouriep dyke) grades into "quartz diorite" and

"locally, especially in the western part of the dyke, into quartz syenite over a width of about 100 yards, the whole dyke measuring about 300 yards across".

This probably means that the Hornblende Diorite and Quartz Bostonite dykes formed during approximately the same period.

The original texture of the Hornblende Diorite Dykes is often found to have been obliterated by the growth of secondary minerals but in the fresher specimens the texture varies from intergranular to panidiomorphic, though some specimens are patchily subpoikolitic. The colour of the rock is characteristically dark greenish grey (5G 4/1) with light greenish grey (5G 8/1) speckles, and on weathering it tends to develop a moderate brown (10R 4/6) crust. The mean thin-section area of the felspar crystals found in a typical specimen of this rock type is approximately 4.2 sq. mm., but as we have observed the border phase of these dykes is often much finer grained. The dominant felspar in the Hornblende Diorites is plagioclase which occurs in subhedral laths with a composition that ranges between An_{24} and An_{38}, with a mean at approximately An_{30}. The plagioclase laths are often altered to sericite and epidote. Most of the alkalic felspar appears to be microcline as can be seen from its grid-iron structure, but the crystals are generally too altered and clouded to yield useful optical data. In contrast to the prismatic shape of the plagioclase crystals the alkalic felspars tend to be more equidimensional in form.

Amphibole occurs in a number of different ways within these dyke rocks; it occurs (i) interstitially between the felspar crystals, (ii) in large crystals that tend to have felspar laths included within them, (iii) some crystals are euhedral to subhedral and occur as discrete individuals that appear to have formed early in the crystallization sequence, and (iv) the amphiboles in some specimens occur in patches of randomly orientated acicular crystals. Many of the amphibole crystals are twinned. The amphibole characteristic of the Hornblende Diorite has the following optical properties: X = greyish yellow (5Y 8/4), Y = moderate yellow green (5GY 7/4) and Z = 1.667 = moderate green (5G 5/6); $\bar{x}$ 2Vx = 78°; and $Z_{\wedge}c$ = 16°.

Minerals of the epidote group occur mainly in one of two ways, either as minute crystals scattered throughout altered plagioclase crystals, or in larger crystals that form clusters. In the more altered and sheared rocks these patches grade into veins. Many of the larger crystals are patchily pleochroic from moderate yellow (5Y 7/6) to greyish yellow (5Y 8/4), and are believed to be pistacite. Some rock specimens carry nonferriferous zoisite which shows its characteristic anomalous interference colours. Clino-zoisite also occurs in some rocks.

Minerals of the chlorite group occur in plates and shreds throughout most of the specimens examined and are believed to be mainly secondary alteration products of hornblende. Many of the crystals have the anomalous "Berlin blue" interference colours characteristic of penninite. Biotite has a similar mode of occurrence and distribution to the chlorite. Specimens 455 and 278 are particularly interesting in that they contain composite crystals composed of titanaugite cores rimmed with hornblende. A little antigorite is found in some specimens and it is generally associated with chlorite. It probably formed as an alteration product of hornblende, though it is possible that a little olivine was originally present in some of the quartz free dykes. In many dykes apatite exceeds 1% (by volume) and in these rocks it tends to occur in long acicular, euhedral crystals. Calcite is a common constituent of many specimens, particularly those that are sheared, and it is mainly found in veinlets that appear to be of secondary origin. Sphene and leucoxene are both found associated with the opaque ore minerals. The leucoxene is mainly an alteration product of ilmenite; and the ilmenite is often found forming symplektic intergrowths with magnetite. In the fresh rocks the proportion of sphene and leucoxene tends to be low but increases with increased alteration. The opaque ore minerals consist mainly of magnetite and ilmenite, and the ilmenite often occurs in the form of "skeleton crystals". Accessory amounts of anhedral quartz occurs interstitially in many specimens. In a few specimens it exceeds 1%, but in most of these specimens the rock appears to have been contaminated by quartzose xenoliths.

(ii) Chemistry: A partial chemical analysis was made of Specimen 360 (Table XI, column 1) which was considered a typical specimen of the Gannakouriep Dyke - the dyke which outcrops over the greatest distance in the Richtersveld. The specimen was collected at De Hoop (See Map 2). An attempt would have been made to obtain a complete chemical analysis of this rock type if it had been realized at an early date that the "analysis of mafic dyke-material", Table 14, p. 143 (de Villiers and Söhnge, 1959) was not of dyke material at all. It appears that the analysis of the dyke material was mislaid and the analysis quoted on page 143 is a repeat of the syenite analysis quoted on page 81, column 4. As this Hornblende Diorite falls into Nockolds' (1954) mangerite class and is also similar to the hornblende lamprophyres, Nockold's

Table XI - Chemical Data (Hornblende Diorite)

	Composition weight per cent					norm:	
Nos.	1	2	3	4	5		6
SiO_2		52.85	50.00			Or	14.4
TiO_2		1.03	2.29			Ab	29.9
Al_2O_3		15.74	16.31			An	18.1
Fe_2O_3 (		3.04	3.99			Ne	3.4
FeO (	8.71^{x}	4.81	6.26			Di ($CaSiO_3$	8.2
MnO	0.21	0.06	0.20			Di ($MgSiO_3$	5.3
MgO	2.88	6.24	4.46			Di ($FeSiO_3$	2.4
CaO		7.58	8.33			Ol (Mg_2SiO_4	4.0
Na_2O	4.14	3.58	4.27			Ol (Fe_2SiO_4	1.8
K_2O	2.53	2.39	2.43			Mt	5.8
H_2O		2.05	0.83			Il	4.4
P_2O_5		0.07	0.63			Ap	1.5
CO_2		0.12					
Li(p.p.m.)	13 p.p.m.			17 ppm	28 ppm		
Rb(p.p.m.)	67 p.p.m.			30 ppm	110 ppm		
Cs(p.p.m.)	<2 p.p.m.			1.1 ppm	0.6 ppm		
Ni(p.p.m.)	17 p.p.m.			130 ppm	4 ppm		
Co(p.p.m.)	29 p.p.m.			48 ppm	1 ppm		
Total		99.58	100.00				

^{x}Total Fe as FeO

1. Specimen 360, Hornblende Diorite from De Hoop. Mode (vol.%): Plagioclase (An_{25}) 50.9%, alkalic felspar 13.6%, hornblende 16.0%, opaque ore minerals 6.4%, epidote group 4.3%, Chlorite 4.2%, biotite 3.0%, apatite 1.6%, quartz tr., white mica tr.

2. Johannsen (1937), Vol. III, p. 192 - Average chemical composition of 14 rocks called Spessartite.

3. Nockolds' (1954) p. 1018 - Average alkali mangerite, average of 53 analyses.

4. The Li, Rb, Cs, Ni and Co content of Turekian and Wedepohl's (1961) average basaltic rock.

5. The Li, Rb, Cs, Ni, and Co content of Turekian and Wedepohl's (1961) average syenite.

6. Norm of Nockold's (1954) - Average alkali mangerite (i.e. column 3).

average alkali mangerite (column 3) and Johannsen's average spessartite (column 2) have been included in in Table XI for purposes of comparison. The norm of the average alkali mangerite (column 6) is observed to be broadly similar to the mode of Specimen 360. As Nockolds and Johannsen do not include trace element abundance data in their tables, selected elements from Turekian and Wedepohl's (1961) average basalt and syenite are also included in Table XI. Inspection of Table XI reveals that the Hornblende Diorite is normal with regard to total Fe, Na_2O, K_2O, Cs, Ni and Co, slightly high in MnO, low in Rb and significantly low in MgO and Li. The occurrence of high MnO and low Rb and Li values is of particular interest as the same trend was observed in the Quartz Bostonites, and high MnO and low Li was the characteristic geochemical feature of the bulk of the Richtersveld Suite plutonic rocks. The Mg content of the Hornblende Diorite is clearly lower than the value found in the average alkali mangerite and average spessartite, but the concentration of this element appears to be sufficient to construct the (modal) minerals found in it. The K/Rb ratio of the Hornblende Diorite is 314 which is a little high, but this ratio does, however, fall within the accepted limits of scatter (See Figure 12 in Appendix 2 and Taylor 1960B, p. 318). It is of interest to note that if all the Na_2O of Specimen 360 was used to form plagioclase, it can be calculated from the weight per cent mode of this specimen that the plagioclase would have the composition of Ab_{76} An_{24}, which is very close to the value determined by optical study. It does, however seem likely that a small part of the Na_2O found in the rock was incorporated in the K-felspar and/or amphibole. The K_2O content of the rock is of the right order of magnitude to satisfy the modal biotite and modal alkalic felspar.

(D) Petrogenesis:

In the foregoing discussion of the Quartz Bostonite and the Hornblende Diorite dykes, the geochemical, textural and structural evidence presented (particularly in the case of the Quartz, Bostonite) suggests that these dyke rocks are compositionally akin to the plutonic rocks of the Richtersveld Suite and that they were possibly emplaced into the Richtersveld plutonic rocks before these bodies had completed their cooling and, in the case of the Quartz Bostonites, before they had become immobile. Circumstantial evidence in support of the hypothesis that both the dykes and the Richtersveld Suite plutonic rock are comagmatic is found in the frequent association of bostonites and lamprophyres with the "granites of the limited massifs" (Raguin, 1946). The occurrence of both bostonitic and dioritic material together in the same dyke at the foot of Mount Erebus would also seem to indicate that both dyke swarms were emplaced broadly contemporaneously, though the syenitic border zone at Mount Erebus may have resulted from contamination. There does, however, appear to have been a break in time between the emplacement of the Quartz Bostonites and the Hornblende Diorites as both swarms have their own characteristic orientations, and the Hornblende Diorites appear to have a wider distribution.

It would seem that with the crystallization of the Richtersveld batholith, particularly its upper and outer parts, there was a slight decrease in volume (Johannsen, 1932, p.130.) which resulted in the formation of embryonic fractures in the plutonic rocks, and probably more definite fractures in the less flexible envelope of country rock. As fractures can only form in fairly solid rock, the embryonic fractures within the batholith would tend to reach down into a zone where the rock was too mobile, under the temperature-pressure conditions prevailing, to fracture. It is believed that when the first of these fractures opened the great bulk of the plutonic material of the Richtersveld Suite had already crystallized though much of it was still relatively hot, and that the fractures thus tapped the top fraction of the remaining potentially mobile material in the magma chamber. The mobile material is believed to have consisted of two fractions; (1) an upper syenitic layer, and (2) a lower more basic layer consisting of highly contaminated material that formed through the settling, accumulation and assimilation of xenoliths during the final stoping phase in the emplacement of the plutonic rocks of the Richtersveld Suite. The upper syenitic layer is believed to have resulted from the parental Alaskitic Granite magma being contaminated and desilicated by contact and reaction with the lower more basic layer. The composition of the magma injected into the fractures probably differed slightly from area to area as cooling in the plutonic rocks was probably unequal, thus resulting in the fractures tapping slightly different levels in the magma reservoir. The escape of silica-rich volatiles also probably assisted in producing the extremely low percentages of free quartz found in some dykes (as proposed by Emmons 1940, p.12.). In those areas where the magma was injected into relatively hot plutonic rocks the dykes cooled slowly to become medium grained and in other parts, where the magma was injected into cooler host rocks, the magma crystallized rapidly and gave rise to finer grained dykes. Later movements within the plutonic mass dragged some of the dykes emplaced within the plutonic rocks of the Richtersveld Suite from their original positions producing their present day curvilinear outcrop traces.

After a further period of cooling and concomitant shrinkage a second group of fractures may have formed, but most probably the event was more spectacular and the magma chamber that probably extended from Soeties in the south to Aurus in the north (See chapter 5 Section A) collapsed forming large north trending vertical tension cracks which developed parallel to the long axis of the magma chamber. As there had been further crystallization of the plutonic rocks since the last fracturing episode, the magma, tapped by the second set of fractures, was at greater depth in the magma chamber and its composition was that of the more basic contaminated material which probably existed at the base of the magma chamber. The release of pressure caused by the opening up of these fractures probably made the contaminated magma more mobile and it welled up the fractures and crystallized to form the Hornblende Diorite Dykes. When intruded into cold country rock these wide dykes developed finer grained border phases. Later the dykes became sheared and a fluid probably rich in CO_2 percolated through the dykes resulting in their present altered appearance and the ubiquitous presence of calcite. The above interpretation of the genesis of the lamprophyric Hornblende Diorite is supported by Grout's (1937, p. 1549) observation that there is a great similarity in the texture and/or composition of lamprophyres and the inclusions found in plutonic bodies. In fact he (1937, p. 1549) stated that

"early inclusions may be completely assimilated or disintegrated and mixed into the host; later ones may be so altered as to form unrecognizable inclusions or lamprophyres; and very late ones remain recognizable xenoliths".

As these dyke swarms are believed to represent two closely related phases of the same igneous cycle, and as the Hornblende Diorites are found to intrude the Stinkfontein Formation, it seems credible to correlate the Quartz Bostonite with the "many lenticular veins consisting of pegmatitic quartz and subordinate pinkish to white felspar" (de Villiers and Söhnge, 1959, p. 116) that occur "over the entire Stinkfontein Mountain land". The veins possibly represent the most fluid fraction of the crystallizing Quartz Bostonite magma which moved the greatest distance from the magma source. The higher quartz content of these veins would be in keeping with the hypothesis because, as we have noted before, silica tends to be preferentially transported in the vapour phase (Bowen and Tuttle 1949, p. 459; and Walton 1960, p. 641).

(E) <u>Age and Correlations</u>:

If the above interpretations are correct, it is found that the Stinkfontein Formation which is cut by both the Hornblende Diorite Dykes and the veins that are correlated with the Quartz Bostonites, is older than both dyke swarms. As the dykes belong to the closing stages of the Richtersveld igneous cycle, the Stinkfontein Formation is either older than, or was laid down contemporaneously with, the emplacement of the plutonic phase of the Richtersveld Suite. It would also seem likely that if the Stinkfontein Formation was laid down contemporaneously with the emplacement of the plutonic members of the Richtersveld Suite, then the andesitic and/or trachytic lavas found intercalated with the arenaceous sediments of this formation might be genetically related to the Richtersveld Suite. A comparative geochemical study of the Stinkfontein lavas and the Richtersveld Suite material would probably be most rewarding, as would a study of the pebbles found in the conglomeritic horizons of the Stinkfontein Formation. These problems were not tackled during the present investigation because at the time when the area was being studied in the field, de Villiers and Söhnge's(1959, p. 26) view that the Stinkfontein Formation was younger than the "Richtersveld Igneous Complex" was accepted, and thus the Stinkfontein Formation was considered to have no bearing on the investigation being made.

The above statements on the relative age of the Stinkfontein Formation open up a wide field for speculation on the age of this formation and the possibilities of correlating it with other formations occurring outside the Richtersveld. As the rocks of the Richtersveld Suite intrude, and those of the Stinkfontein Formation overlie, the Adamellitic Gneiss, both of the former rock groups are younger than the Adamellitic Gneiss and thus younger than 980 ± 100 million years (Nicolaysen 1962C) - the age now generally attributed to the Namaqualand Granite-Gneiss. It is thus seen that the correlation of the 2100 million year old Witwatersrand System with the younger than 980 million year Stinkfontein Formation, as proposed by de Villiers and Söhnge (1959) and Truter (1959), is most unlikely.

De Villiers and Söhnge (1959, p. 88) claim to have identified inclusions of "Richtersveld Granite" within both the Numees and Kaigas Formations. As the Numees and Kaigas Formations overlie the Stinkfontein Formation along the western flank of the Stinkfontein Mountains, these two formations are thus both younger than the Richtersveld Suite and the Stinkfontein Formation. The Nama System which is younger than the Richtersveld Suite and Stinkfontein Formation, is as will be seen in Chapter VII, of Late Proterozoic age and is thus approximately 600 million years old (Holmes, 1960, p. 204). Thus there is an upper age limit of 980 million years, and a lower limit of 600 million years, for the age of the Richtersveld Suite and the Stinkfontein Formation.

As tentative circumstantial evidence of the age of the Richtersveld Suite it should be noted that ring-dyke complexes are near-surface epizonal features that soon succumb to denudation and thus (as was shown in Chapter V) they are very seldom of Precambrian age, hence by analogy it would seem that the age of the Richtersveld Suite is unlikely to be greatly in excess of the 600 million year date quoted for the commencement of the Palaeozoic.

The question of the age of the Richtersveld Suite and the Stinkfontein Formation can be approached from yet another direction if a search is made for rock units outside the Richtersveld with which the Stinkfontein Formation can be correlated, though as will be seen in Chapter IX, it is possible that the Stinkfontein Formation developed under special conditions and is of limited distribution. The formation that is most suited to fill this role on both lithological and relative age grounds, is the Tsumis Formation of South West Africa. If attempts are made to correlate the Richtersveld rocks with rock units to their south it is soon discovered that the Cape Granite and the Richtersveld Suite appear to be of broadly similar age, and the small Klein Kogelfontein Complex of the Bitterfontein area (Jansen, 1960, pp. 34-55) and the plutonic rocks of the Springbok, Kamieskroon and Garies districts mentioned by de Villiers and Söhnge (1959, p. 72) may represent the link between these two igneous bodies. As the Cape Granite and the Malmesbury System that it intrudes, are believed to belong to the same tectogenetic cycle (Sholtz, 1946), the relative age of the Malmesbury as compared to the Cape Granite, and of the Stinkfontein Formation as compared to the Richtersveld Suite, appear to be similar; thus the Malmesbury and Stinkfontein Formations may be equivalent to one another. If, however, lithology is taken into consideration the correlation of the Malmesbury beds and the Kaigas Formation seems more probable, though it must be noted that Jansen (1960, p. 12) has divided the Malmesbury Formation of the Bitterfontein area (to the south of Namaqualand) into three suites, the uppermost of which, his quartzite suite, is lithologically similar to the rocks of the Stinkfontein Formation.

In a recent synthesis of the Upper Proterozoic geology of South West Africa, Martin (1962) has proposed that the Otavi, Damara and Nama Formations of South West Africa are miogeosynclinal, eugeosynclinal and cratonic platform facies respectively, of a single great geosynclinal belt which formed the western fringe of the older cratonic nucleus of the Southern African subcontinent. Martin (1962) has also suggested that the Stinkfontein, Kaigas, Numees, Grootderm and Bogenfels Formations were deposited in the "tectonically active zone between the main geosyncline and the cratonic foreland" and that these formations belong to the "same tectogenetic cycle" as the Damara-Otavi System of South West Africa. All the data gleaned during the present study supports Martin's hypothesis and it is felt that the Richtersveld must in future be viewed, not as being peripheral to both South West Africa and South Africa, but rather as being a bridge between the two areas. If the Richtersveld is used as a bridge in this way it might enable future stratigraphers to establish a West Coast System embracing the Malmesbury, Stinkfontein, Kaigas, Numees, Nama, Otavi, Damara, and Katanga Formations and Systems.

If the rocks of the Richtersveld are to play a significant role in disentangling the ravelled stratigraphic and igneous history of the Late Proterozoic of South and Central Africa, the absolute age of a number of rock units of different relative ages must be established. It is suggested that the absolute ages of the following rock types should be obtained: (1) the Adamellitic Gneiss from the area below the Stinkfontein beacon where the Stinkfontein Formation is seen to overlie the Adamellitic Gneiss, (2) the lavas from the Stinkfontein Formation, (3) the main Alaskitic Granite of the Richtersveld Suite, (4) the Hornblende Diorite Dykes from a locality where they are found to cut the Stinkfontein Formation, and (5) the Kaboos Granite which is younger than the Numees Formation. If geochronologists need targets to shoot at it is proposed that the age of the Adamellitic Gneiss is approximately 1000 million years, the Stinkfontein lavas and the Alaskitic Granite approximately 650 million years, and the Hornblende Diorite Dykes approximately 630 million years. The Kuboos Granite is believed to be unrelated to and younger than the Stinkfontein-Richtersveld-Nama tectogenetic cycle (Chapter IX). All that can be said about its age is that it is younger than 600 million years - possibly a great deal younger.

Type specimens of the Alaskitic Granite and the high Rb Porphyritic Microgranite have been supplied to Dr. L.O. Nicolaysen of the Bernard Price Institute of Geophysical Research, Johannesburg. Work on determining the absolute ages of these rocks has commenced.

(F) Resumé:

The Quartz Bostonites have a mean width of 18 inches (46 cms.) and they tend to dip vertically and strike between N 10° E and N 30° E with a slight maximum at N 30°E. Their grain size and texture varies considerably. If these grain size variations are considered in relation to the field distribution of the dykes, it is found that grain size tends to increase as one moves away from the pre-Richtersveld Suite rocks and into the heart of the Richtersveld plutonic rocks. This observation together with others such as the absence of cooling at the margins of the coarser dykes which cut the Richtersveld Suite plutonic rocks, and the curvilinear outcrop traces of some of the dykes emplaced into the plutonic rocks, all suggest that the dykes were emplaced before the complete consolidation and cooling of the plutonic rocks of the Richtersveld Suite. The Quartz Bostonites, which were formerly regarded as dolerites, are petrographically and chemically normal bostonites. Chemically they are of interest in that they have high total Fe and Mn contents, low Li and Rb contents, and abnormally high K/Rb ratios, which are all features they share with the syenites of the Richtersveld Suite.

The Hornblende Diorite Dykes which persist over considerable distances in their strike direction, strike approximately due north, have a dip that is essentially vertical, vary greatly in width, and have a mean width of 12 feet (4.m.). They are generally altered and often sheared. Some of the dykes, particularly those intruded into rocks older than the Richtersveld Suite, become finer grained towards their margins, and thus appear to be dilation dykes that crystallized from magmatic material. The texture of the Hornblende Diorite varies from intergranular to panidiomorphic and some specimens are partly subpoikolitic. Petrographically these rocks fall into Peterson's diorite and lamprophyre rock groups. Chemically the Hornblende Diorite is high in Mn, low in Rb and significantly low in Mg and Li. The high Mn and low Rb and Li values are of particular interest as the same trend was observed in the Quartz Bostonites.

The Hornblende Diorite and Quartz Bostonite dykes are believed to represent the final phase in the Richtersveld igneous cycle. It is suggested that towards the close of this cycle the base of the magma chamber in which the plutonic rocks had been generated, contained two layers of magma, (1) a basal layer of highly contaminated dioritic magma, and (2) an upper layer of bostonitic magma which had formed as the result of the desilication of the parental alaskitic magma by contact with the basal contaminated magma. With the crystallization of the plutonic rocks above these layers, contraction fractures developed, and material from the upper bostonitic magma layer was injected into them. After further cooling the plutonic mass collapsed and long wide fractures with strikes parallel to the long axis of the magma chamber and extending from at least

Soeties in the south (28°58'S.) to Aurus Waterhole in the North (27°34'S.), opened up. The contaminated dioritic magma which had been produced by the settling and accumulation of xenoliths during the final phase in the emplacement of the plutonic rocks of the Richtersveld Suite, was emplaced into them. Later the Hornblende Diorites were fractured and percolated by fluids rich in CO_2.

As the Hornblende Diorites are considered to be the final phase in the Richtersveld igneous cycle, the Stinkfontein Formation which they cut is regarded as being either older than, or of similar age to, the plutonic rocks of the Richtersveld Suite. It is also tentatively suggested that the Richtersveld Suite, the Stinkfontein, Kaigas and Numees Formations and the Nama System all belong to a single Late Proterozoic tectogenetic cycle (The West Coast System?) that is equivalent to the Katanga, Damara-Otavi and Malmesbury Systems.

VII THE NAMA SYSTEM

(A) Introduction:

The Nama System sediments of the Richtersveld were first mentioned by Atherstone in 1855. He correlated them with the Table Mountain Series of the Cape System. A year later in 1856 (published in 1857) Wyley studied these same rocks and stated that he "scarcely knew whether to refer them to the age of the Table Mountain strata or not", but when he constructed his map the beds were shown as belonging to the Table Mountain Series. In Dunn's famous maps of the geology of South Africa the beds of the Nama System of Namaqualand south of the Orange River, first appeared as belonging to the Table Mountain Series (1875), but later (1887), they were considered to belong to the Witteberg Series. The term "Namaformation" was introduced by Schenk in 1885, and in his paper of 1888 he employed the Nama beds of Namaqualand south of the Orange as a connecting link between the Table Mountain Series of the southern Cape and the Nama beds of South West Africa, and all these sediments were collectively called the "Cape Formation". In 1912 Range divided the Nama System of South West Africa into the following divisions:

Fish River Series Schwarzrand Series	Upper Nama
Schwarzkalk Series Kuibis Series Basal Beds	Lower Nama

Haughton and Frommurze (1936 p.21) recognized the same divisions in the Nama of the Warmbad district, but they found it convenient to group the Basal and Kuibis beds together. The same procedure was followed by de Villiers and Söhnge (1959, p181).

Beds of the Nama System are the youngest consolidated rocks in the area and they rest unconformably over all the other consolidated rocks they come into contact with, including the dyke rocks mentioned in the previous chapter. A number of primitive fossils have been found in the Kuibis beds (Gurich 1930 and 1933 Haughton 1929, Haughton and Frommurze 1936 and Richter 1955 Haughton 1960) but their systematic position and value in dating of these beds remained dubious until large numbers of similar fossils were found in the Pound Quartzite of the Adelaide Series of South Australia (Glaessner, 1958 and 1961) and in Charnwood Forest, Leicester (Ford, 1962). Glaessner (1958, p.526) found that besides large numbers of medusae and Dickensonia the South

Australian fossil fauna "includes specimens closely resembling Pteridinium simplex Gurich (particularly the nesotype figured by Richter 1955, pl.1, but not so much the other specimens figured by him) and Rangea schneiderhöhi Gurich (pl.7, fig.12) from the Kuibis quartzite of the Nama Series of South West Africa".

Richter (1955) considers Pteridinium and Rangea to be impressions of gorgonarian coelenterates but Glaessner (1958, p.526.) considers the frond-like Rangea to be similar to the modern Pennatulids (sea-pens). Ford (1962, p. 194) has warned that it is "dangerous to compare Pre-Cambrian fossils with modern animals (or plants) when there are no intervening fossils".

The late Precambrian fossil remains that have been

found to date, and they include "a dozen or so different jelly-fish, a segmented Annelid-Arthropod Spriggina, several frond-like organisms such as Charnia and Rangea, a number of impressions of uncertain nature ..... numerous worm-trails, algal limestones, spore-like bodies, and a few shells such as Lingulella montana" (Ford 1962, p. 194.), are composed mainly

of fragile soft tissues which require "unusual" (Glaessner, 1961, p.73) conditions for their preservation. This is believed to account for their infrequent occurrence in beds of the Nama System. Recently, Haughton (1960, pp.57-59) has restudied some of the Nama fossils from the Kuibis beds of the Ham River, South West Africa, and he now believes that he can recognize double-walled Archaeocyathids of lower Cambrian age.

In South Australia the position of the lower limit of the Cambrian is problematic in that it appears that sedimentation continued uninterrupted from definite Precambrian to definite Cambrian times. In 1955 Glaessner discussed this question in his paper on the "Time-stratigraphy of the Late Pre-Cambrian" and he proposed the term 'Late Proterozoic' for the 80 million years preceding the beginning of the Cambrian. He was not in favour of the terms 'Eo-or Infracambrian' or 'Late Algonkian' though he also favours the term 'Riphaean'. He (1961 p.73) also seems to favour the characterization of the Late Proterozoic as the "age of the jellyfish". Korn and Martin (1959) believe that statigraphic as well as fossil evidence favours a Late Proterozoic age for the Nama System. They state that field evidence from the Vanrhynsdorp district indicates that the Nama sediments are older than the Cape System and thus older than the Upper Silurian, while in the Witpütz area the Nama beds overlie the highest members of the Kaigas System which (as we saw in chapter 6E) they consider to be an equivalent of the Damara-Otavi System. Recently Martin (1962) has suggested that the Nama beds are the neritic deposits of the cratonic shelf (platform) which formed the southern and eastern foreland of the Otavi-Damara geosyncline, and that the Nama beds of the Naukluft area (S.W.A.) were involved in the latest phase of the Damara orogeny.

(B) Stratigraphy and Petrography:

The strata of the Kuibis Series of the Nama System are approximately 380 feet thick in the area along the western flank of the Neint Nababeep Plateau (de Villiers and Söhnge 1959 p.183), but only the beds of the lower 150 feet fall within the area mapped. In plate 9 beds of the lower Kuibis Series can be seen resting unconformably on Kheis

Plate 9: Nama beds resting unconformably on Kheis Supracrustal Rocks at Modderdrif-Suid where the stratigraphic section was measured.

meta-supracrustal rocks in the area where the following succession was recorded:

Table XII - Stratigraphic Section

Late Proterozoic (Riphaean): Kuibis Series: Nama System: Modderdrif-Suid.

		Feet	Inches.
Unit 1	Felspathic Sandstone Mineral Composition: Quartz 80.6% K-felspar 14.2%, Plagioclase 0.1%, White Mica 4.6% Biotite 0.1%, Ore Minerals 0.1%, Rock Fragments 0.3%. Gross Character : Yellowish grey (5Y 7/2) to Very light grey (N8) in colour, compact, bedding irregular to massive. Texture: Sorting generally good, So = 1.51, but occasional milky to clear spherical quartz pebbles occur. The corrected median grain diameter is 0.44 mm; Particles are predominantly sub-angular though the felspar grains tend to be sub-rounded.	2 -	0
Unit 2	Olive grey (5Y 3/2) silty shale containing moderate red (5R 4/6) patches; micaceous along partings. Bedding shaly to flaggy.	0 -	8
Unit 3	Felspathic Sandstone (As unit 5)	1 -	8

		Feet		Inches.
Unit 4	Olive grey (5Y 3/2) silty shale. Bedding shaly to flaggy.	0	-	3
Unit 5	Felspathic Sandstone Mineral Composition: Quartz 81.8%, K-felspar 16.4% Plagioclase 0.1%, White Mica 0.7%, Ore Minerals 0.1%, Rock fragments 0.6%, Clay 0.3%. Gross Character : Very light grey (N9) in colour, compact, bedding irregular to massive. Texture: Sorting generally good, So = 1.65, but occasional spherical milky quartz pebbles are found. The corrected median grain diameter is 0.46 m.m. Particles are predominantly sub-angular, but the felspar grains are mainly sub-rounded.	2	-	3
Unit 6	Moderate red (5R 5/5) fractured shale	0	-	8
Unit 7	Felspathic Sandstone Mineral Composition: Quartz 80.0%, K-felspar 17.5% Plagioclase 0.1%, Rock fragments 0.2%, Clay 2.2% Gross Character : Light grey (N7) in colour, compact, bedding massive. Texture : Sorting good, So = 1.20. The corrected median grain diameter is 0.37 m.m Particles are predominantly sub-angular, but the felspar grains are mainly sub-rounded.	2	-	9
Unit 8	Felspathic Sandstone Mineral Composition : Quartz 79.4%, K-felspar 20.1%, Plagioclase 0.1%, Clay 0.4%. Gross Character : Light grey (N7) in colour, compact, bedding massive with vertical joints. Texture : Sorting good, So = 1.16. The corrected median grain diameter is 0.44 m.m. Particles are predominantly sub-angular, but the felspar grains are mainly sub-rounded.	4	-	4
Unit 9	Deeply weathered greenish grey (5GY 5/1) moderate red (5R 5/5) and light grey (N7) shales, mainly covered by scree; some of the shales have changed to a powdery mud; patches of ochre occur.	±12	-	6
Unit 10	Felspathic Sandstone Mineral Composition: Quartz 72.5%, K-felspar 17.9% White Mica 0.7%, Ore Minerals 2.9%, Zircon 0.1%, Rock fragments 0.1%, Clay 5.8%. Gross Character : Pale red (5R 6/2) in colour, compact, iron stained with a few plates of white mica being visable in the hand specimen. Texture : Sorting tends to be good, So =1.71; particles are predominantly sub-angular, but the felspar grains tend to be sub-rounded.	1	-	7

		Feet	Inches.
Unit 11	Olive grey(5Y 3/2) Shale containing small silty-sandstone lenses, bedding irregular, shaly to flaggy.	3 -	3
Unit 12	Felspathic Sandstone Mineral Composition : Quartz 74.5%, K-felspar 21.6%, Plagioclase 2.3%, White Mica 0.7%, Ore Minerals 0.8% Hornblende 0.1%. Gross Character : Pale Red (5R 6/2) in colour, compact. Bedding irregular, as this rock occurs in lense shaped beds. Texture : Sorting good, So = 1.24; the majority of the particles are sub-angular, but the felspar grains tend to be sub-rounded.	3 -	4
Unit 13	Dark grey (N3) finely laminated shale.	1 -	4
Unit 14	Limestone Two varieties are present which lense one into the other. Variety A is a sandy limestone and is Greyish red (5R 4/2) in colour on fresh surfaces. Variety B is a true limestone and is dark bluish grey (5B 4/1) in colour and breaks with a concoidal fracture. Mineral Composition: (A) Quartz 36.2%, K-felspar 7.0%, White Mica 0.1%, Calcite 56.3%, Ore Minerals 0.2% Rock Fragments 0.2% (B) Carbonate minerals 99.8% Quartz 0.2%, K-felspar Tr., White Mica Tr. The corrected median grain size of the quartz and felspar particles of 14A is 0.44 m.m., and that of 14B is 0.06 m.m.	1 -	4
Unit 15	Felspathic Sandstone Mineral Composition: Quartz 87.6% K-felspar 12.3%. Zircon 0.1%. Gross Character : Greyish Pink (5R 8/2) in colour, compact, with massive bedding. Texture : Well sorted, So = 1.18; particles are mainly sub-angular with the felspar being sub-rounded.	11 -	3
Unit 16	Shales and fine siltstones, mainly scree covered. The shales are moderate red (5R 4/4) in colour and finely laminated and contain hematite pellets. The fine siltstones are mainly light grey (N7) in colour.	± 22 -	0
Unit 17	Arkose * Mineral Composition : Quartz 46.1%, K-felspar 44.7% Plagioclase 4.2%, White Mica 2.8%, Biotite 1.4%, Ore Minerals 0.6%, Zircon 0.2%.	6 -	0

* As this rock was found to have such a high felspar content, it was decided to determine its alkali content to see if the high modal felspar content was correct. The results of this chemical analysis will be discussed later.

		Feet	Inches
	Gross Character : Pinkish-grey (5YR 8/1) in colour, compact, thickly bedded, though the bedding thins towards the top of this unit. Texture : Sorting Good, So = 1.09, particles are mainly sub-angular but the felspars tend to be sub-rounded.		
Unit 18	Limestone Mineral Composition: Carbonate minerals 91.0%, Quartz 7.2% K-felspar 1.6%, Plagioclase 0.1%, Zircon 0.1%, Ore Minerals Tr. Brownish grey (5YR 4/1) in colour. The corrected median grain diameter of the quartz particles found in this unit is 0.18 m.m.	0 -	5
Unit 19	Moderate red (5R 4/4) and light grey (N7) shales	0-	4
Unit 20	Limestone Carbonate minerals 99.0%, Quartz 0.5%, Ore 0.3% Plagioclase 0.1%, White Mica 0.1%. Greyish red (5R 4/2) in colour. The Corrected median grain diameter of the quartz particles found in this unit is 0.05 mm.	1 -	0
Unit 21	Moderate red (5R 4/4) and light grey (N7) interbedded shales and mudstones.	8 -	0
Unit 22	Quartzose Sandstone. Mean Mineral Composition: Quartz 92.2% K-felspar 5.0%, Plagioclase 0.2%, White Mica 0.4% Biotite 0.1%, Ore 1.0%, Zircon 0.1%, Hornblende 0.1% Rock fragments 0.9%. Gross Character : Mainly light grey (N7) in colour as in specimen 22A, but there are some lenses of coarser grained moderate red (5R 4/4) sandstone (ie. specimen 22B). As a whole this unit is a thickly bedded, scarp forming sandstone which contain in some localities a thin discontinuous shaly horizon. Texture: Sorting good, So = 1.72, though a few pebble sized particles are found in the lenses of 22B material. Particles are mainly sub-angular with the felspar being sub-rounded.	±44 -	0
		141 -	4

As "fine sediments are difficult to study in thin-section with ordinary equipment and texural and compositional parameters of fine silt and clay raise many unresolved problems of interpretation" (Cadigan 1959, p.530.), it was believed that the greatest amount of pertinent data could be obtained in the time available from a petrographic study of the sandstone horizons. Four textural concepts were used in this investigation -

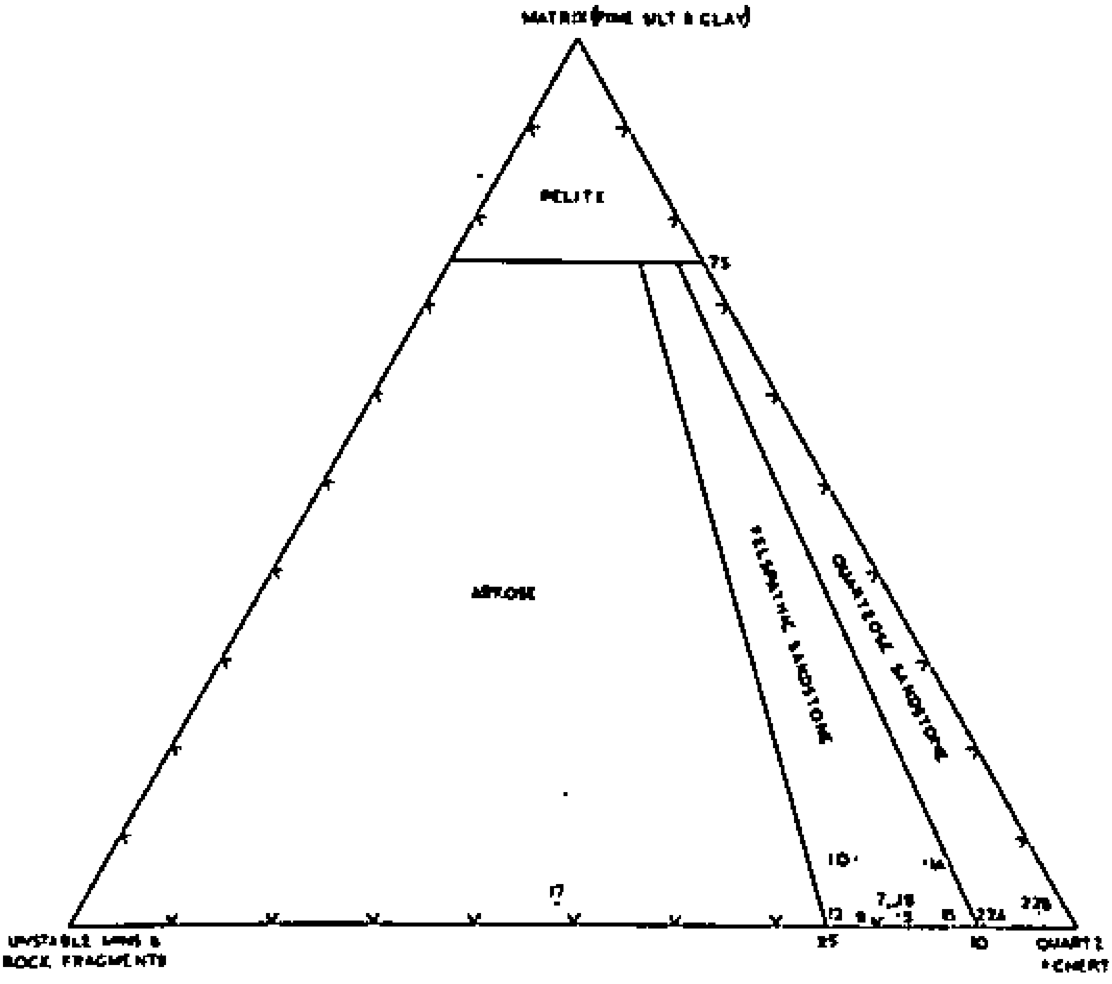


Figure 6: Sandstone Classification (After Packham 1954, p. 475). (The numbers refer to the sedimentary units described in table XII.)

median grain size, sorting, skewness, and kurtosis. The properties of the different sandstones which may be said to define their grain-size distribution are summarised in Table XIII and Figure 8.

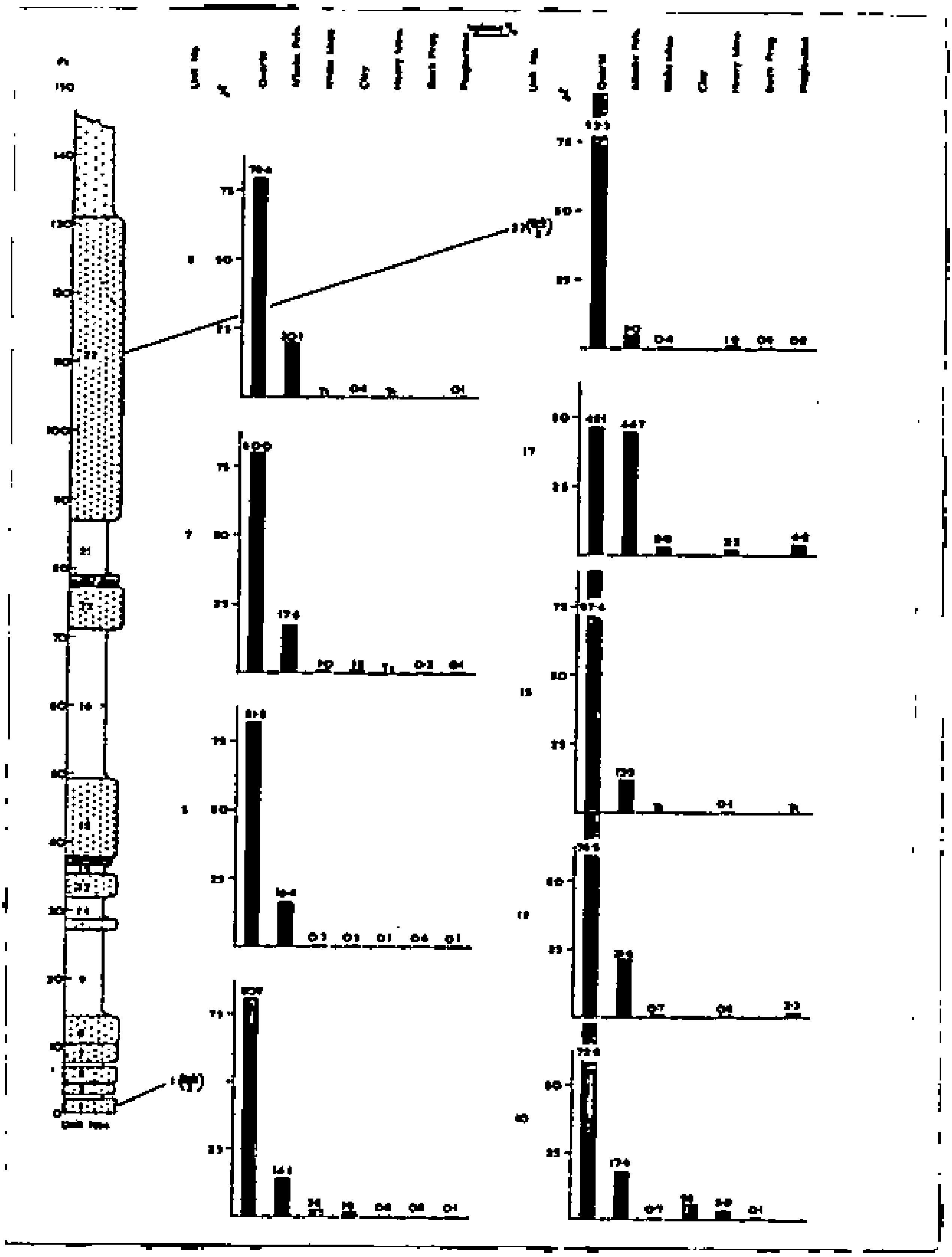


Figure 7: The mineral composition and stratigraphic position of the Kuihis Sandstones.

Table XIII Grain Size Distribution of Kuibis Sandstones.

Specimen Number	Md. (Median) m.m. *	Md. XI.27 ∅	Friedman (1958) corrected Md. m.m.	Mo. (Mode) m.m	So. Cofficient of sorting.	SK. Skewness	K. Kurtosis
1(a b)	0.50	0.63	0.44	0.54	1.51	0.94	0.199
5	0.53	0.67	0.46	0.61	1.65	0.88	0.265
7	0.44	0.56	0.37	0.38	1.20	0.90	0.234
8	0.50	0.63	0.44	0.54	1.16	1.00	0.197
10	0.33	0.42	0.28	0.38	1.71	0.87	0.278
12	0.19	0.24	0.17	0.18	1.24	0.96	0.235
15	0.50	0.63	0.44	0.54	1.18	0.99	0.167
17	0.13	0.16	0.12	0.15	1.09	0.85	0.133
22(a+b)	0.47	0.60	0.39	0.42	1.72	1.13	0.269
Arithmetic mean	0.40	0.51	0.34	0.42	1.38	-	0.220

*Median of the apparent long axis of the grains in thin section.

∅Median of the estimated maximum long axis (or diameter) of the grains.

The median values of these sandstones after applying the Friedman (1958) correction range from 0.12 m.m to 0.46 m.m. with an arithmetic mean of 0.34 m.m. Thus they all fall within Wentworth's (1922) medium and fine sands (ie. between $\frac{1}{2}$ and $\frac{1}{8}$m.m). Specimens 1,5,7,8,10,15 and 22 fall within the medium sand size grade, and specimens 12 and 17 are fine sands. As the largest corrected median grain diameter was only 0.46 m.m. it would appear from this data that the strength of the transporting current which moved this material to the site of deposition, was probably not very great. Modes, or the size value (or values) around which grains tend to concentrate, were also calculated and the sandstones, which are all unimodal as can be seen from figure 8, were all found to have very similar modal and median values thus indicating that their size distributions are likely to be normal. It is of interest to note that Krynine (1948,p.154) states that the average grain size of a typical arkose is 1.0-0.18 m.m.

The sorting coefficients ($So = \sqrt{Q3/Q1}$) of the sandstones which are considered to be indices of the range of conditions present in the transporting fluid range from 1.09 in specimen 17 to 1.72 in specimen 22, thus indicating that the sandstones are all within Trask's (1923) well sorted category which includes all sediments with an So value of less than 2.5. The arithmetic mean of these So values is 1.38 which indicates that the sandstones are better sorted than the majority of "near-shore marine sediments of the sand grade" (Pettijohn 1949, p.24) which have an average So value of 1.45. It must, however, be remembered when using these sorting coefficients in reconstructing the environment in which the sediments were deposited and in determining the selective action of the transporting medium, that this method of assessing size sorting ignores the occasional pebbles which are found in units 1, 3 and 5 though it is clear that they are significant criteria in determining past conditions of transport and deposition.

The skewness (or the tendency of the size distribution of the sandstones to depart from the theoretical symmetrical normal form) was measured by using $Q_1 . Q_3/(Md^2)$. It was found that the size distributions of specimens 8,12 and 15 were extremely symmetrical, while the other specimens with the exception of specimen 22 show a slight negative skewness (ie. poorer sorting in the coarser size grades). Specimen 22 shows a slight positive skewness (ie. poorer sorting in the finer size grades).

Kurtosis or peakedness measures the relative degree of sorting that the centre of a distribution bears to the two ends. The kurtosis $K = (Q_3 - Q_1)/ 2(P_{90} - P_{10})$ of the size distribution of the Kuibis sediments ranged from 0.133

to 0.278 with an arithmetic mean of 0.220.

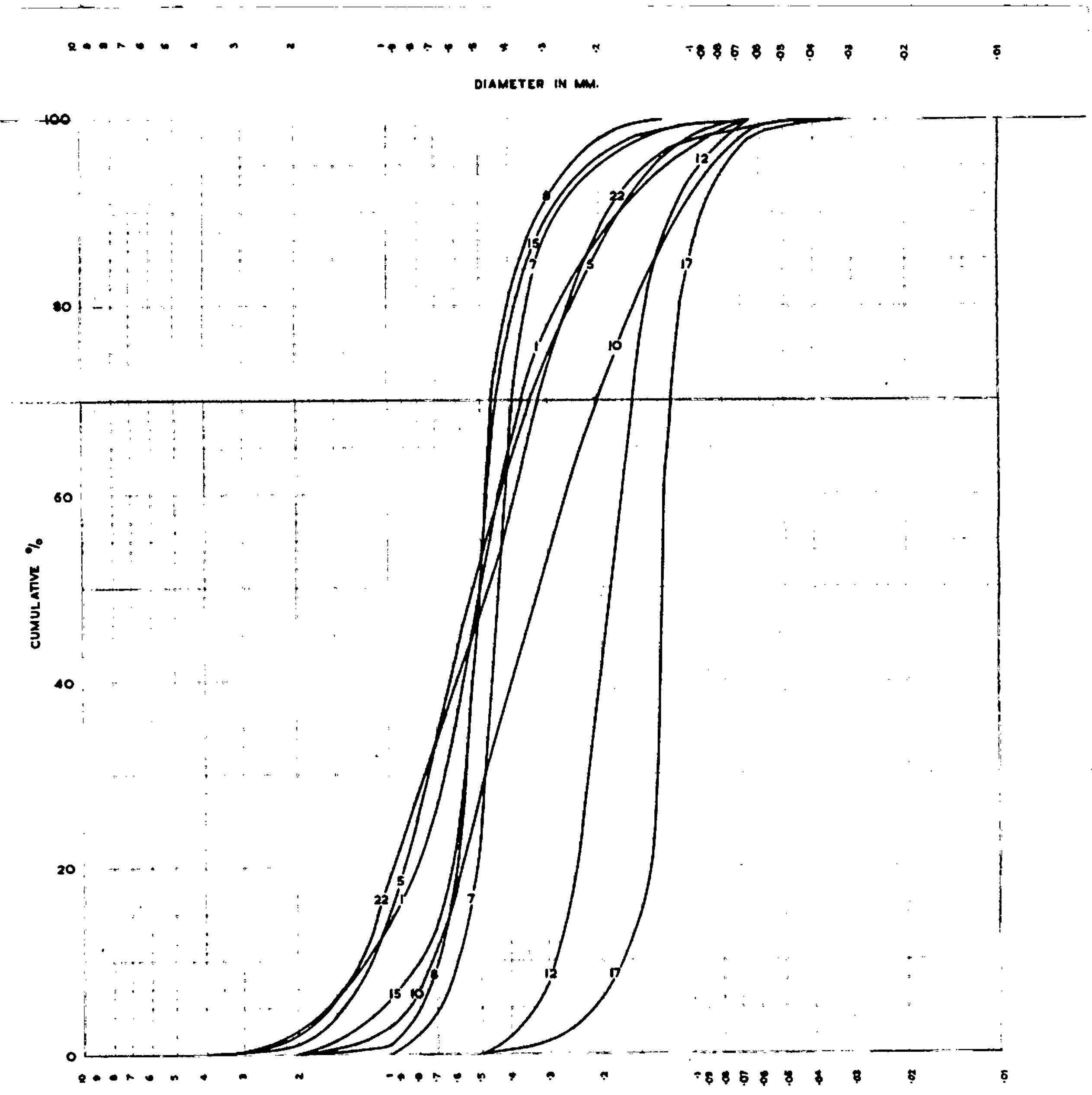


Figure 8: Grain size distribution of the Kuibis Sandstones.

The shape as well as the size of the sandstone grains was studied, but as will be seen in the discussion of texture, it was not always possible to determine the original or primary shape of the grains. With regard to roundness it was found that the quartz grains tend to be sub-angular while the felspar grain tend to be more rounded and tend to fall into the sub-rounded category. A rough estimate was made of the second main geometrical aspect of particle shape, sphericity, and it was found that the median sphericities of the sandstones ranged from 0.5 to 0.7 with an arithmetic mean of 0.56 (ie. On this scale (after Krumbein and Sloss 1951, p.81) a sphere has a sphericity of 1.0 and a thin, needleshaped particle a sphericity of 0.0.).

As can be seen from table XII specimens 14B, 18 and 20 are classified as limestones as they contain over 90% (by volume) of carbonate minerals, while specimen 14A is considered to be a sandy limestone as it contains between 50% and 90% of carbonate minerals. (Ref. Pettijohn (1949 p.290) for his nomenclature and classification of sand-limestone mixtures.). The limestones appear to have a heterogeneous texture (Carozzi, 1960, p.207) being mainly composed of small diffusely bordered, light-absorbing carbonate clots associated with occasional clearer and brighter calcite aggregates. In this carbonate material a few small irregular shaped quartz and felspar grains occur. Specimen 14A contains 36.2% quartz and 7.0% felspar, and the quartz and felspar particles of this specimen are much larger than those found in the limestones (14B, 18 and 20) as their corrected median size is 0.44 m.m., as compared to 0.09 m.m which is the mean of the corrected medians of the quartz and felspar grains of the other limestones. The quartz and felspar grains of the limestones and sandy limestone are of particular interest in that they have the scalloped and serrated shapes of grains which have undergone partial replacement by carbonate minerals. (See Photomicrographs 21 and 22.). No organic remains were found in these limestone horizons, but this does not in any way rule out the possibility that these deposits are the product of some organic process. These thin limestone horizons when viewed in relation to the full Nama succession are seen as the harbingers of the thick limestone horizons lying conformably above them and which contain "well preserved" algal remains. (Haughton and Martin 1956, p. 324.). In fact algal domes can be seen in the limestone horizons of the Neint Nababeep Plateau to the east of the area.

A detailed study of the many limestone horizons of the Nama System appears to be a fruitful field for further study as these beds are believed to be broadly contemporaneous with the Archaeocyathus limestones of Australia which are the most extensive Paleozoic reefs discovered (Gignoux, 1955, p.46.).

As was shown in section IG iii b all the sandstones studied fall into Packham's (1954) arkose-quartzose sandstone suite. A closer examination of the composition of these Kuibis Sandstones, reveals that they form a fairly closely related group of rocks which tend to grade one into another, and that they all have a low "matrix" percentage (under 10%). Specimen 17 is an arkose, and all of the remainder of the sandstones but for the specimens from unit 22, are felspathic-sandstones. The occurrence of a relatively high proportion of detrital felspar in this suite of Kuibis Sandstones is considered a most significant fact, as Pettijohn (1949, p.94) has indicated that "detrital felspar is an index of the intensity of diastrophism", and appreciable amounts of felspar are likely to occur "where the erosion is vigorous and sedimentation rapid". The occurrence of over 10% alkalic felspar, mainly microcline, in all the sandstones below unit 22 is also probably indicative of a granitic or gneissic source area for these sediments. The freshness of the felspar is also of interest in that it seems to indicate that the decomposed mantle rock in the source area had been removed prior to the erosion of the material now found in the Kuibis Sediments. This decomposed mantle rock might have been deposited closer to the source area, or have been removed by ice during the Nabas glaciation. Packham (1954) states that sandstones belonging to his arkose-quartzose sandstone suite are invariably deposited by traction currents in shallow water environments. The occurrence of ore minerals in most of the sandstones (the percentage increases to over 2% in specimen 10) and the presence of a little detrital biotite and hornblende in some of the sediments is also of mineralogical and genetic significance. The occurrence of ferromagnesians is of particular interest as their presence is also indicative of conditions of rapid erosion and deposition.

The almost complete absence of matrix in the lower Kuibis sandstones is reflected in their textures that tend to consist of a mosaic of interlocking quartz and felspar grains (See Photomicrographs 18, 19, 20 and 23). Dust-rings (Carozzi, 1960, p.21.) which usually indicate the boundaries between overgrowths of sedimentary silica and the original detrital grains are not common in these specimens, but this does not rule out the possibility of secondary overgrowth having occurred. Upon closer inspection the texture of the sandstones appears to have formed as a result of a combination of secondary overgrowth around detrital cores, and a reciprocal interlocking of grains (mainly quartz) produced by dissolution at their margins, probably during compaction. When considering the origin of this interlocking mosaic texture one begins to wonder if the original sorting of these rocks was as good as it seems today, and whether smaller particles once existed in the voids between the original grains, and being small with relatively large surface areas were dissolved and now form part of the secondary overgrowths. Carozzi (1960, p.56.) does, however, state in his discussion on the felspathic sandstone series that the quartz and felspar grains are often closely packed so that an original matrix is not necessarily required.

(C) Chemistry:

The identification of the felspar as being mainly potash felspar was verified by the partial chemical analysis of unit 17 - the arkose. The alkali metal compnents of this rock are given in column 3 of table XIV. An interesting feature found in the analyses quoted in table XIV is the dominance of K with respect to Na. The K/Na ratio of the arkose from unit 17 is 9.4(:1). Pettijohn's average arkose (column 1) has a K/Na ratio of 3.0 and the Torridon sandstone (column 2) has a K/Na ratio of 11.5. The K/Na ratio of the "average" granite is 1.24 (Turekian and Wedepohl, 1961, p.186, Table 2) and that of an "average" gneiss has a similar ratio 1.33 (ie. average quartzo-felspathic gneiss - Poldervaart, 1955, p.135). Ramberg's (1951,p.31.) average Greenland gneiss has a K/Na ratio of 0.63. From the above data it can be seen that arkosic sandstones differ greatly with regard to their K/Na ratios from their supposed granitic parent rocks. The depletion of Na relative to K in the Kuibis sandstanes is clearly not unique but it is of interest that this fractionation process should have been so efficient while the potassic felspars have remained fresh and clear.

As can be seen from table XlV column 3, unit 17 of the Kuibis Series contains 248 p.p.m. rubidium, and has a K/Rb ratio of 242 (See figure 12 in app. 2) which falls well within the limits of scatter of "normal K/Rb ratios" (Taylor, 1960 . B., p.318). The K/Rb ratio of unit 17 is also very close to the value of 230 which is now taken as representing the average K/Rb ratio "for igneous and sedimentary rocks and minerals and meteorites" (Taylor 1960 B, p. 318.). The absolute rubidium content of unit 17 is high when compared to Turekian and Wedepohl's (1961, Table 2.) value of 60 p.p.m. for their average sandstone, but this high value is to be expected when it is noted that the K value of unit 17 is

5.99%, and Turekian and Wedepohl's average sandstone has a K value of only 1.07%. The Li value of unit 17 (7 p.p.m.) is low in comparison to Turekian and Wedepohl's (1961, Table 2.) average sandstone value (15 p.p.m.) and the 33 p.p.m. value of its postulated source rock the Adamellitic Gneiss. This discrepancy is easily explained when it is recalled that lithium tends to accumulate in the micas, pyroxenes and amphiboles of igneous rocks and that these minerals are less abundant in unit 17, than in the Adamellitic Gneiss, and are probably also less abundant in this unit than in Turekian and Wedepohl's average sandstone which includes many sandstones with relatively high, phyllosilicate contents.

Table XIV - Chemical Data (Kuibis Sandstone)

	1	2	3	4	5
SiO_2	76.37	82.14		79.46	
TiO_2	0.41	-		-	
Al_2O_3	10.63	9.75		10.50	
Fe_2O_3	2.12)	1.23		0.78	
FeO	1.22)			0.57	
MnO	0.25	-		-	
MgO	0.23	0.19		0.09	
CaO	1.30	0.15		0.25	
Na_2O	1.84	0.50	0.86	0.75	0.44
K_2O	4.99	5.27	7.22	7.19	1.25
H_2O	0.83	0.64		0.18	
P_2O_5	0.21	0.12			
CO_2	0.54	0.19			
ZrO_2	-	-		0.23	
Li (p.p.m.)	-	-	7.p.p.m.		15 p.p.m.
Rb (p.p.m.)	-	-	248 p.p.m.		60 p.p.m.
Cs (p.p.m.)	-	-	<2 p.p.m.		0.X p.p.m.
	100.94	100.18		100.00	

(1) Pettijohn (1957, p.324). Average Arkose (Mean of 5).

(2) Mackie (1905, p.58.) Torridon Sandstone from Scotland (quoted by Pettijohn, 1957, p.324.).

(3) Unit 17: Arkose from Modderdrif Suid.

(4) Unit 17: Arkose from Modderdrif Suid: chemical composition

determined from the modal composition taking the alkalic felspar composition to be Or_{92} Ab_8, the plagioclase to be Ab_{70} An_{30} and the ore minerals to all be of magnetite composition.

(5) Turekian and Wedepohl (1961, Table 2). Average alkali-metal content of sandstones.

Table XV - Modes of Nama Sandstones and possible Source Rocks Compared

	$\bar{X}$ Ad.Gneiss	$\bar{X}$ Nama Sandstone	Unit 17 Arkose	$\bar{X}$ Alaskitic granite.
Microcline	20.8	18.9	44.7	-
Perthite	-	-	-	61.3
Hornblende	Tr	Tr	-	-
Biotite	6.5	0.2	1.4	2.6
Quartz	34.7	77.1	46.1	33.6
Ore Mins.	0.3	0.6	0.6	0.4
Plagioclase	21.7	0.8	4.2	0.9
Calcite	0.5	-	-	Tr
Chlorite	1.8	-	-	0.1
Epidote	5.3	-	-	Tr
White Mica	8.2	1.1	2.8	0.9
Apatite	0.1	-	-	Tr
Orthite	0.1	-	-	-
Zircon	Tr	0.1	0.2	Tr
Fluorite	Tr	-	-	0.2
Rock fragments	-	0.2	-	-
Clay	-	1.0	-	-
(No. of specs.)	(41)	(11)	(1)	(39)

(D) Source Rocks:

Table XV which shows the mean modal compositions of the Adamellitic Gneiss, the Kuibis Sandstones, and the Alaskitic Granite, reveals that on mineralogical grounds the Adamellitic Gneiss with its high microcline content seems the most likely source rock of the Kuibis sandstones of the area. If this is indeed so, it is clear that most of the plagioclase and the ferromagnesian minerals were removed from the sediments before deposition. Column 3 of table XV shows that the arkose from unit 17 is unique among the sandstones of the area in that its microcline content is so high relative to the its quartz content as to indicate that some special sedimentary process, perhaps winnowing, operated during its formation. It is also possible that its source rock or rocks were slightly more microcline rich than the average Adamellitic Gneiss. As the processes of microcline enrichment would tend to be of limited magnitude, it is clear that the importance of unit 17 must not be over-emphasized when attempts are made at reconstructing the paleogeography of early Nama times. Table XVI shows the ratio of the mean modal values of the major phases found in the Kuibis sandstones over the means modal values of the same minerals found in the Adamellitic Gneiss; thus values greater than one show that sedimentary processes have tended to concentrate that particular mineral while values less than unity show relative depletion in that mineral. These figures again show the greater stability of microcline relative to plagioclase in the sedimentary environment in which the early Nama beds formed.

The occurrence of 0.2% of fine grained rock fragments in the Kuibis sandstones studied is also of considerable significance as it clearly indicates that granitic rocks (the Adamellitic Gneiss) were not the only rocks outcropping in the source area of these sediments. These rock fragments are probably fragments of Kheis System meta-supracrustal material.

<u>Table XVI - $\bar{X}$ Kuibis Sandstone and $\bar{X}$ Adamellitic Gneiss Compared</u>

Quartz	Ore Minerals	Microcline	White Mica	Plagioclase	Biotite
2.22	2.0	0.91	0.13	0.04	0.03

(E) <u>Paleogeography</u>:

In the foregoing description of the sediments found in the lower 150 feet of the Nama System at Modderdrif-Suid, a number of deductions were made concerning their source, mode of transport and deposition. These inferences will now be linked together in an attempt to reconstruct the paleogeographic environment, but it must be stressed that many more stratigraphic sections in different localities, and a statistical study of sedimentary structures will be needed before a balanced picture of the Nama depositional environment can be obtained. Thus the present investigation does not pretend to be a comprehensive study of the Nama sediments of Namaqualand but rather it is seen as a pointer to further study as it has now been established that the Kuibis sandstones are ideal for petrographic study and are capable of yielding a great deal of significant quantitative data. It is also clear that a detailed petrographic and structural study of the Nama sediments of Namaqualand would be a major project in itself and that such an investigation would take a number of years to complete.

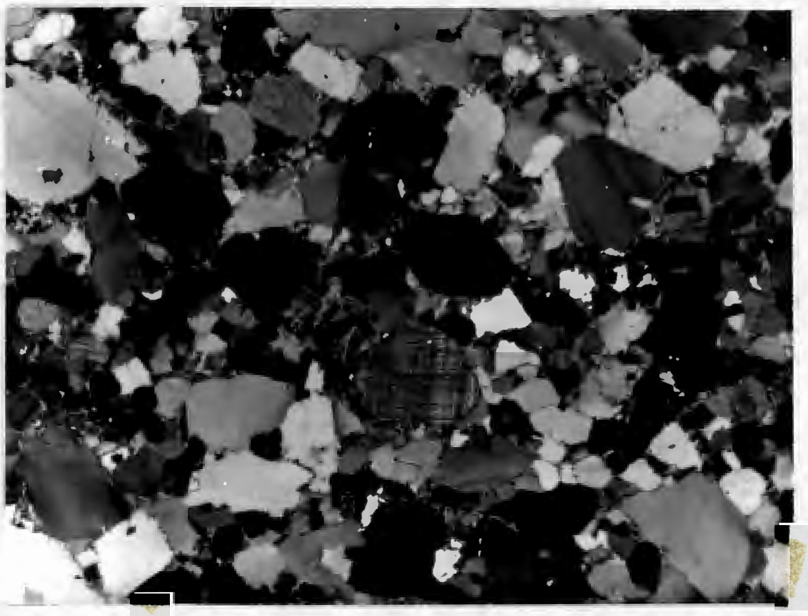
No direct evidence pertaining to climate prevailing during Nama times was found in the sediments studied, though frigid conditions during at least part of Lower Nama times would seem likely as Schwellnus (1942) has found evidence of glacial conditions in the Lower Schwarzkalk beds of the Klein Kharas mountain area (S.W.A.); Söhnge and de Villiers (1946), and de Villiers and Söhnge (1959) have found evidence of glacial conditions in the Nabas Stage of the Kuibis Series in the north eastern Richtersveld (ie. 1½ miles north of Nabas, See Map 2); and Haughton (1961, p.73) has stated that during the Late Precambrian, Central and West Africa experienced a widespread glaciation of "semicontinental dimensions and existing for a long period with fluctuating degrees of intensity". The fresh condition of the K-felspar found in the Kuibis sediments might also be taken as indicative of glacial conditions though Krynine (1941 A., p.1918) has stated that "a restudy of arkoses in the geological column would assign most of them to ultrahumid climates operating on very steep (not necessarily high) relief". The pebbles found scattered throughout the lower sandstone horizons, in particular the first unit, might be taken as further evidence in support of the occurrence of glacial conditions during Lower Nama times, but Martin (1962) has indicated that the presence of such pebbles in S.W. African formations is probably indicative of an off-shore depositional environment, as he states that away from the shore conglomerates change not by a uniform decrease in pebble size but rather that the pebbles retain their size and decrease in frequency of occurrence. This latter proposal seems the more likely in the area studied as the pebbles are generally found to have been rounded. If, however, at some future date it is found that the Nama and Numees beds (See Chapter 6, Section E) are of the same age it will be clear that glacial conditions were prevalent during Nama times, particularly to the west and north west of the area.

The presence in the Kuibis sandstones of fresh K-felspar, indicates a source area rich in relatively unweathered granitic or gneissic rock. Krynine (1941.A, p.1919) in his paper on the paleogeographic and tectonic significance of arkose states that "tectonically arkoses are par excellence related to granitic terranes in regions of steep youthful topography. This places them at the very end of the geosynclinal stage after magmatic intrusion, uplift and block faulting, and before peneplanation".

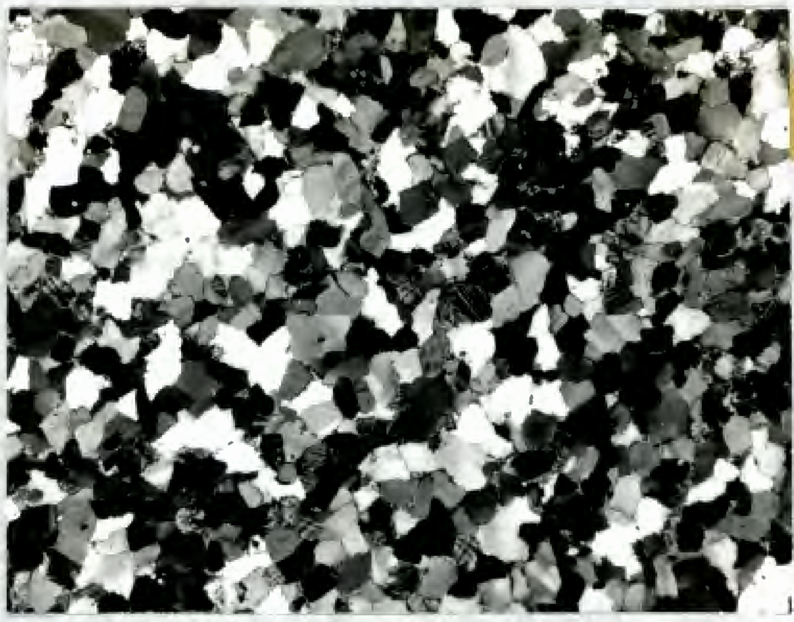
It is also significant that of the 22 units studied over half (13) are wholly or partly red or pink in colour. Although widely divergent views have been expressed about the origin of the so called "red beds", the statement by Dunbar and Rodgers (1957, p.209) that with minor exceptions the red beds consist of non-marine terrigenous, fragmental sediments, is probably a fair summary of current opinion on the redbed problem. As Weller (1960, p.324) has emphasised colour more than any other rock feature, provides information about the oxidizing or reducing conditions of a sedimentary environment, and strong red colours are sure signs of oxidation. The pigment in the red beds of the Kuibis Series appears to be mainly finely divided hematite, which from the foregoing statement can be taken as indicative of an oxidizing sedimentary environment. Interbedded and alternating with the red beds of the Kuibis Series, one finds mainly light grey beds which are indeterminate as indicators of depositional environment; but which together with the red beds seem to indicate alternations in the conditions of sedimentation. Most authors who have written on the red bed problem have stressed that an oxidizing environment is seldom found at the site of marine deposition; but they usually qualify this statement as Weller (1960,p.35) has done by stating that an oxidizing environment could be found in some very shallow coastal areas and in the deepest parts of the ocean where sedimentation is slow and bottom life scant. In areas of present day marine sedimentation reducing conditions are generally created by the presence of decaying organic matter which tends to exhaust the environment of its free oxygen. If, however, in early Nama times the primative life forms were unable to survive in the conditions of rapid sedimentation, indicated by the abundant fresh felspar, oxidizing conditions might have been able to prevail in the depositional environment; but even if the environment was slightly reducing in character the Kuibis red beds might well have been deposited sufficiently rapidly for some of the hematite to have survived and given the Kuibis beds their present colour.

Grain-size distribution data seems to indicate that the constituent particles of the sandstones were transported by medium strength traction currents, and deposited in a shallow water environment. The alternation of fine and coarse sediments might be considered to represent oscillations in water depth during deposition, though this in only one possibility as changes in any of the other variables in the depositional environment might lead to like results. While the felspar percentages of the sandstones seem indicative of conditions of rapid erosion and deposition, the good sorting of the sandstones and their grain shapes, would seem indicative of longer transport. Perhaps, there was at first a period of rapid erosion during which the constituent materials of the sandstones were laied down followed by an oscillation in depth resulting in a reworking of some of the material with a winnowing action operating. If such a reworking did occur it would have had to take place in an oxidizing environment, so as to retain the hematite. It does, however, seem more likely that the size and shape of the quartz grains as found in the sandstones is directly related to the quartz grain-size and shape of the granitic rocks of the source area. The limestone horizons, particularly the great thickness of limestone found above the Kuibis sandstones, probably represents a period of quiescence after the tectonic activity that produced the felspathic sandstones of the Kuibis Series.

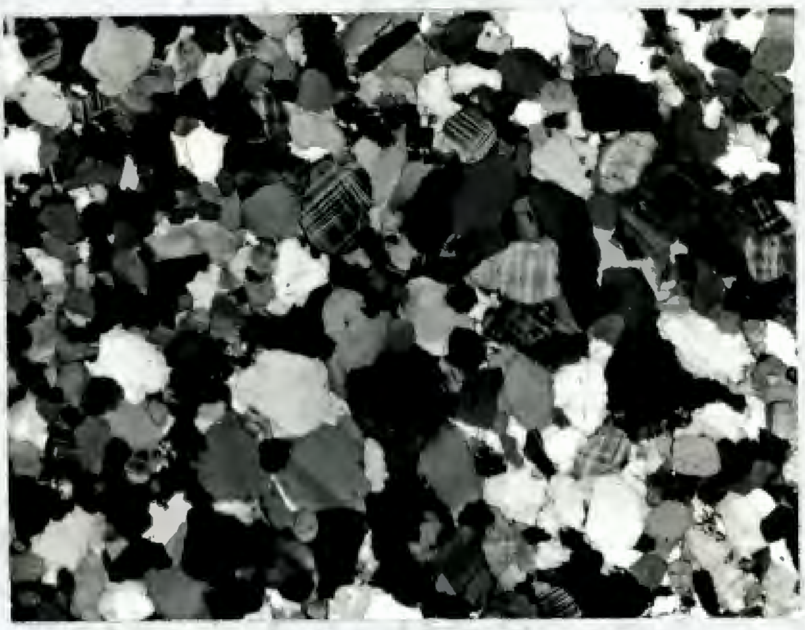
After deposition the sediments were compacted, and partial recrystallization took place giving rise to sediments with the textures found in them to-day.



Photomicrograph 18: Nama Sandstone ie. from Unit 1 (X13.5), Specimen No. 152, X-nicols.

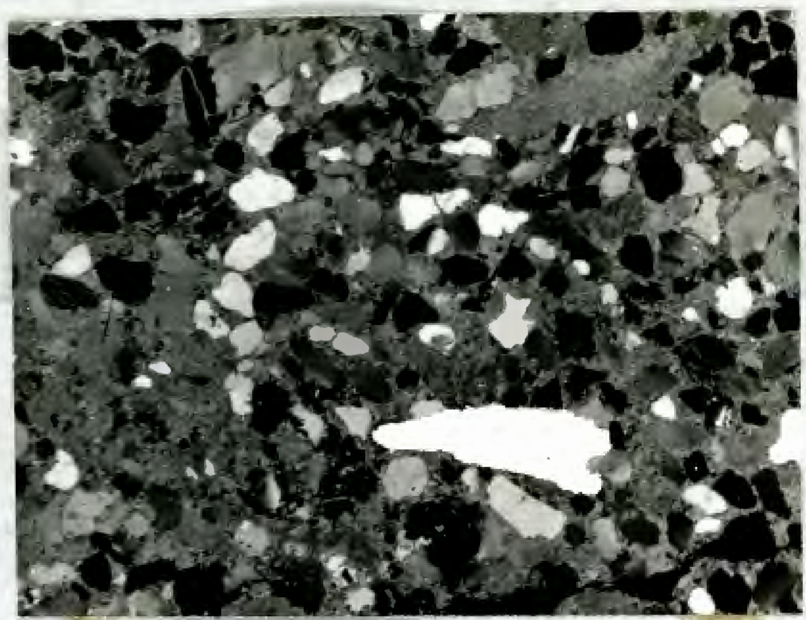


Photomicrograph 19: Nama Sandstone ie. from Unit 7 (X 13.5), Specimen No. 176, x-nicols.

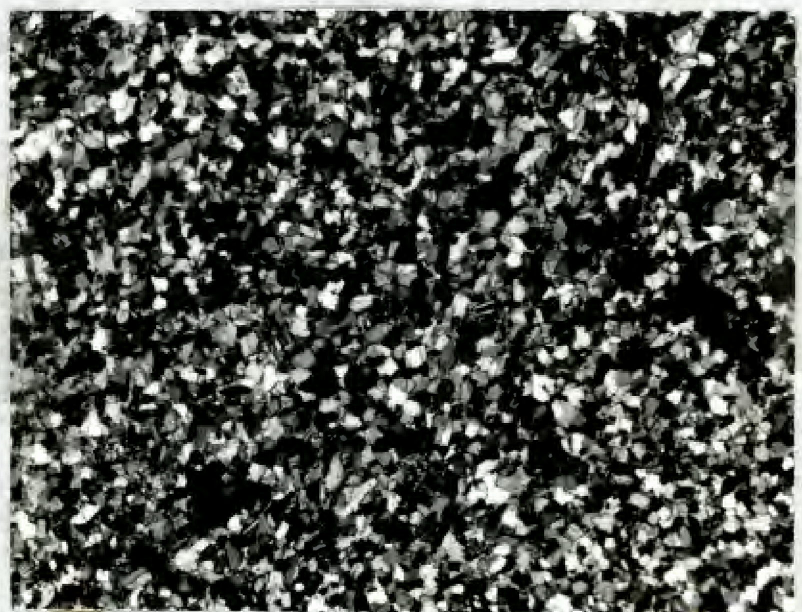


Photomicrograph 20: Nama Sandstone ie. from Unit 8 (X 15), Specimen No. 177, x-nicols.

Photomicrograph 21: Nama Limestone ie. from Unit 14 (x 12), Specimen No. 181, x-nicols.



Photomicrograph 22 : Another part of the thin-section of limestone from Unit 14 (X 12), Specimen No. 181, x-nicols.



Photomicrograph 23: Nama Arkose, ie. from Unit 17 (X 15), Specimen No. 143, x-nicols.

(F) Resumé:

The Late Proterozoic Nama sediments rest unconformably above the other consolidated rocks in the area. A section through the lower Kuibis Series was studied and found to contain 9 shale horizons, 8 felspathic sandstone horizons, 3 limestone horizons, 1 arkose horizon and 1 quartzose sandstone horizon. All the sandstones belonged to Packham's arkose - quartzose sandstone suite. Petrographic data seems to indicate that (1) the sandstones are products of vigorous erosion in a source area in which the Adamellitic Gneiss was the main rock type, (2) transportation was by means of medium strength traction currents and (3) deposition took place in a slightly oxidizing shallow water environment. The limestone horizons particularly the thick limestone horizons above the Kuibis sandstones, probably represent a period of quiescence after the tectonic activity that produced the felspathic sandstones. All the evidence obtained was considered to be in agreement with Martin's (1962) statement that the Nama sediments are cratonic platform deposits, though as will be seen in Chapter 9, the Nama sediments of the area have some unique features as they probably developed on the very edge of this platform.

VIII SURFACE DEPOSITS.

These surface deposits which are believed to range in age from early Tertiary to the present, can be divided into three main groups : (1) deposits associated with remnants of river terraces, (2) unconsolidated sand and rock waste and (3) flood-plain alluvium. In the area the only significant remnant of a river terrace is found at Xaminxaip where a layer of rounded to sub-rounded pebbles and boulders is found cemented by calcareous material on a truncated surface some 30 feet (9.1m.) above the approximate mean level of the Orange River. No fossil remains or diamonds were found in this terrace deposit. To the immediate east of the area, at the confluence of the Tc Cowiep and Orange rivers a number of boulder strewn terraces are to be found. The orange coloured tent in Plate 9 is pitched on one of these terraces. Larger terraces which are usually in a better state of preservation than those found in the area, are to be seen both up and down the Orange River from the area. Magnificent terraces, that generally contain many water polished semi-precious stones (agate, jasper, agate-jasper and other varieties of chalcedony), are found at Rooiwal ($\pm$ 50 feet: 16.4 m.), 5 miles upstream from the area, and at De Hoop, 45 miles down stream from the area. The unconsolidated sand is mainly confined to the dry river beds of the tributary streams of the area, and it grades into rock debris and scree as the inclination of slope increases away from the river beds. The sand is found to be composed mainly of quartz, felspar grains and rock fragments which have been produced by the mechanical breakdown of the surrounding country rocks. The alluvium occurs as a narrow ribbon along both banks of the Orange River, and consists mainly of a fine, generally dark, silt. At those localities where tributary streams have cut through the silt exposing a vertical profile, the silt is found to be bedded and sometimes cross-bedded, and a few thin irregular gravel layers are occasionally found. In those parts of the flood plain which are devoid of vegetation the silt is often blown into small dunes, and as can be seen in Plate 9, wind ripples are a common feature.

IX STRUCTURE.

(A) Introduction:

The key to an understanding of de Villiers and Söhnge's interpretation of the structure of the Richtersveld is found in their (1959, p.197) statement that the "structures of the Richtersveld are remarkable not so much for their intensity, except in the oldest rocks, as for the exceptional parallelism exhibited by structures of widely different ages". This parallelism of structures of different ages is clearly to be seen in the area, but it is not considered to be as remarkable as de Villiers and Söhnge believe, as the account of the stratigraphy of the Richtersveld proposed in

Chapter 6 suggests, that most of the rock units that display this parallelism belong to the same tectogenetic cycle.

(B) Kheis-Adamellitic Gneiss Cycle:

In the Kheis Supracrustal Rocks of the area the most characteristic structural feature is the presence of considerable shearing, foliation and schistosity. Most of these structural features (the trends of which are shown on the geological map accompanying this report) while not being uniform in intensity or distribution, tend to strike north-south and run approximately parallel with the border of the Adamellitic Gneiss. Early faults and folds within this material have for the most part been obliterated by later shearing and metamorphism. The Adamellitic Gneiss and its associated hybrid rocks, while displaying "wild folding" in some parts, frequently contains foliation and banding which has a north-south trend similar in orientation to that found in the Kheis meta-supracrustal rocks. In contrast to the later plutonic rocks of the Richtersveld Suite, the Adamellitic Gneiss has gradual and generally concordant contacts that support the view that it was emplaced at greater depth than the Richtersveld Suite plutonic rocks.

The Post Kheis Ultramafics are believed to have been emplaced along two east-west trending zones of crustal weakness that probably developed during the Kheis orogeny as it is clear that some external force must be invoked to account for the upward movement of dense relatively cool quasi-solid ultramafic magma.

(C) Richtersveld Suite and Nama Cycle:

In Chapter 6 it was suggested that the Stinkfontein, Kaigas, Numees and Nama Formations together with the Richtersveld Suite all belong to a single tectogenetic cycle. If this hypothesis is accepted then the structure of all these rock units and their correlates (that probably extend from the South Western Cape to Angola) must be viewed as a single unit. In relation to the huge area mentioned above the area studied is very small indeed, hence the following descriptions of the structure of the Richtersveld Suite and the Nama sediments of the area must be seen as descriptions of a small part of a very much larger whole.

As was stated in Chapter 5 the known outcrops of the plutonic rocks of the Richtersveld Suite fall within a gently curved north, north west trending crescent that extends for at least 120 miles from Soeties in the south to Aurus Waterhole in the north. To the west of these rocks the outcrop of the Stinkfontein Formation and its boundary faults (Söhnge and de Villiers 1946, plate 32) trace a trend that runs parallel to the outcrops of the plutonic rocks. East of the plutonic Richtersveld Suite rocks of the area lie the younger Nama sediments of the Neint Nababeep Plateau.

If an attempt is made to reconstruct the sequence of events that occurred at the time of the development of the Stinkfontein Formation and the Richtersveld Suite, Rich's (1957,pp.1179 - 1222.) "magma blister" theory (Chapter 5 section H.) is found to be a most useful guide to the possible sequence of events. Recast in terms of Rich's theory the Richtersveld Suite magma is seen to have formed by palingenesis in the zone of relative weakness found along the western edge of a craton; where because of the increase in volume produced by the melting of the preexisting rocks during palingenesis, a "magma blister" or geanticline formed. The surface rocks above the magma blister which probably consisted mainly of quartzites and schists, (Kaaien) on being elevated began to be eroded thus lightening the raised central mass. As a result of this isostatic equilibrium was disturbed and further uplift followed (Rich, 1951,p.1182, figs. 1 and 2.). Subcrustal matter flowed inward from beneath the western marginal zone and crustal sinking started along the blisters western flank. The Stinkfontein sediments accumulated along this western flank first as piedmont deposits and later as the geosyncline developed as lacustrine and then deeper water deposits (de Villiers and Söhnge 1959,p.204.). The Stinkfontein lavas were probably extruded from the magma blister during times of isostatic uplift when faulting and fracturing occurred along the inner edge of the developing geosyncline. Lava was possibly also extruded through the tensional fractures and faults that probably developed in the stretched surface that occurred over the top of the magma blister. Later when the plutonic rocks were emplaced into their present positions and they had crystallized, the blister contracted resulting first in the emplacement of the Quartz Bostonites in the manner explained in Chapter 6; and later further cooling and shrinkage is believed to have resulted in the collapse of the Richtersveld Suite magma chamber (ie. the blister) that lay beneath the 120 mile long Soeties - Aurus outcrop crescent. As a result of the collapse large vertical tension cracks developed parallel to the outcrop crescent and the Hornblende Diorites were emplaced into some of them. The north-south boundary faults of the Stinkfontein Formation are also believed to have been active at this time, and the Stinkfontein beds may have acquired their present westerly dip as the result of the block between the boundary faults tilting westwards. After a period of erosion, deposition appears once again to have taken place and the Kaigas and later the Numees and Nama beds were laid down.

The folding of the Nama sediments is believed by de Villiers and Söhnge (1959,p.208) to be the result of the already consolidated sediments being pinched in the sinking trough and deformed along the edges of the gently pitching syncline produced. Not much can be said about the structure of the Neint Nababeep plateau as the area contains but the outliers of these rocks along its eastern border, but one fact that does emerge, and can be seen on the map accompanying this report, is that the Nama beds that are found dip eastwards or inwards towards the centre of the main mass of Nama sediments of the Neint Nababeep plateau. The Neint Nababeep plateau when considered as a single unit, appears to form a broad shallow syncline with a north-south axis, and contained within this major structural unit there are a number of smaller local folds which though small in size are often strongly folded and even overfolded. An example of Nama sediments overfolded towards the east is found at Modderdrif-Suid (See Map 2.).

As was stated earlier it is believed that the Nama sediments of the Neint Nababeep Plateau represent a late stage in the Stinkfontein - Richtersveld Suite tectogenetic cycle. It would seem that while the Stinkfontein geosyncline was developing the Neint Nababeep area was part of the "magma blister" (ie. two very small outcrops of Richtersveld Suite plutonic rocks are found east of the Neint Nababeep Plateau: Söhnge and de Villiers, 1946, plate 32.) and thus no deposition occurred. Later as the blister contracted a depression formed in the Neint Nababeep area and the Kuibis sandstones developed. During the next stage an epicontinental sea covered much of the western part of southern Africa, and in the Neint Nababeep area the depression, that had now become a basin, continued to subside and fill with a great thickness of limestones. Eventually the rocks of the basins margin slumped and flowed plastically under the influence of gravity to produce most of the folds found today. This is, however a simplification of the structure found in the Neint Nababeep area, as the pre-Nama surface on which the Nama sediments lie is itself folded in some localities, and this leads one to suggest that the Nama sediments of this area, like those of the Naukluft area, may have been caught up in the final phase of the Damara orogeny.

The movements associated with the deposition and folding of the Nama beds probably also left their imprint on the older rocks of the area, but the fracturing and faulting produced appears to have been parallel to the axis of the Neint Nababeep syncline and thus parallel to earlier fractures, faults and shearing.

(D) Post-Nama Events:

The majority of the faults of the area have similar trends to faults that have Post-Nama displacements. The commonest strike direction is north-west, and faults with a similar general trend are found cutting the rocks of the Dwyka Series in the Vioolsdrif area (to the east of the area - See Map 2). It thus seems likely that these faults came into being, or that large displacements occurred along them, during Post-Dwyka, Karroo times. In the area some of the north-west trending faults are displaced by faults with a strike that is approximately due north. These north trending faults are either post-Karroo, or as this trend is common in many earlier fractures and faults, further movements took place along these fault planes in post-Karroo times. The principal direction of displacement found along many of the faults of the area is in the strike direction. As most of the dykes and plutonic contacts found in the area are essentially vertical, the dominance of the strike-slip component might be more apparent than real, but it does seem that the faults of the area are primarily wrench faults with the dominant relative movement of one block to the other being horizontal. Most of the fault planes are essentially vertical as would be expected of wrench faults (Moody and Hill, 1956,p. 1208.).

As can be seen from the geological map accompanying this report milky quartz and to a lesser extent calcite and greyish yellow (5Y 8/4) fluorite are frequently found along the fault and fracture planes of the area. Less frequently deposits of siderite and galena are found secreted along fractures. The siderite is dark yellow brown (10YR 3/2) and occurs associated with milky quartz veins, in a small lenticular body in fractured hybrid rocks found to the north-west of Rooiberg 2 (at 1D6). On the western side of the Tc Cowiep valley at its confluence with the Orange River and approximately half a mile up the Orange River from the area, and also along the south eastern border of the area, a number of galena-bearing quartz veins are found associated with minor faults and fractures. These veins are of no commercial value, but the Tc Cowiep veins are of geological interest as they are found emplaced within the shales found at the top of the Kuibis Series of the Nama System. The veins occur along fractures, minor faults and the bedding planes in the Nama sediments. The main galena vein at the Tc Cowiep - Orange confluence, reaches 12 inches (30.5 cms.) in width and strikes in a north easterly direction and has a variable dip.

POSTSCRIPT:

Dr. L.O. Nicolaysen (Private Communication 17th April, 1963) states that in addition to determining the absolute age of the plutonic rocks of the Richtersveld Suite he now plans to determine the absolute ages of (1) the Pre-Nama dykes of the S.E. Richtersveld and (2) the dykes that cut the Numees (or Nabas Stage, Nama) Beds at Nabas. He hopes that this geochronological program will provide "unequivocal maximum and minimum limits for the time of deposition of the Nama."

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Appendix I : Chemical analytical methods employed.

1) Sodium and Potassium: These elements were determined by Eel (model A) flame and photometer. A full description of the procedure employed is found in Taylor (1960.A.). The following standards were used to establish the working curves:

		Na%	K%
G-1*	Granite	2.46	4.51
W-1*	Diabase	1.54	0.53
U.S.B.S.,89+	Lead-Barium Glass	4.299	6.97
U.S.B.S.,89+	Plastic Clay	0.193	2.63
U.S.B.S.,99+	Soda Felspar	7.96	0.34

* Stevens et al (1960)
\+ U.S. Bureau of Standards Samples.

These silicate standards were used in order to minimize possible matrix effects. Precision was found to be good in the case of both elements, and the relative deviation of the analytical method is believed to be 1% (Coefficient of Variation).

2) Lithium and Rubidium: These elements were determined by optical spectrography, using the variable internal standard procedure for the alkali elements (Ahrens and Taylor, 1960, Section 13- 3). Operating conditions were as follows :- "Hilger" large quartz and glass spectrograph (E.478); wavelength 4600-9600 Å glass optics; slit width 10μ; seven-step sector (2:1 ratio); "Kodak" IN plate; current 4 amps d.c. (short-circuit setting); anode excitation; lower electrode (anode); "National Carbon Co." electrodes; internal diameter of crater 2.4 mms., depth 2.5 mms.; and the upper electrode (cathode) was of "ship Carbon". The unmixed sample powders were arced till the end of the alkali metal distillation period. Plates were developed for 4 mins. in Kodak D-19,b.developer and the lines read with a Hilger non-recording microphotometer. Intensities were determined using the self calibration procedure of Ahrens and Taylor (1960). Sodium was used as the internal standard in all but specimen 577, and the following lines were used Na 5688, Li 6707, Rb 7800 and Rb 7947. Natural silicate standards were again used and produced satisfactory working curves. The following values were used:-

		Li (p.p.m.)	Rb (p.p.m.)
G-1*	Granite	24	220
W-1*	Diabase	9	22
O.G.19+	Granite	-	67

* Stevens et al (1960)
\+ Nicholaysen (1962.A.)

A relative deviation of 5% is generally obtained for these elements using this method. While reading the above plates for Rb and Li, the Cs 8521 line was also examined, and in most cases the amount present was below the limits of detection (2 p.p.m.), but in specimens 577 and 468 a Cs content of > 2 p.p.m. was detected.

3) <u>Tin, Copper, Thallium, Gallium and Lead</u>: As these elements are all relatively volatile in the d.c. arc, the principle of alkali metal distillation was once again employed. A semi-quantitative method, employing no internal standards, was used, as the main purpose of the study was to establish the presence or absence of significant quantities of tin in the granite specimens. To reduce excessive background due to cyanogen bandheads all samples were mixed with 5% Johnson-Mathey "Specpure" NaCl, to act as a buffer. Operating conditions were as follows: wavelength range 2750- 4680 Å quartz optics; slit width 10μ; seven-step sector (2:1 ratio), "Kodak" 103-0 plate; current 4amps d.c. (short circuit setting); lower electrode (anode) with crater of 4 mms.,internal diameter and 5 mms. in depth; and upper electrode (cathode) of "Ship Carbon". All samples were exposed for exactly 60 seconds. Plates were developed and read as described above. The following lines were read Sn 3175, Cu 3274, Tl 3775, Pb 4057, Ga 4172. Synthetic standards were employed for the estimation of tin as the detection of tin in G-1 and W-1 is not possible with the methods used. The standards were prepared by diluting Cassiterite with quartz and NaCl to give samples with concentrations of 300, 100, 30, and 10 ppm tin (Sn.). The working curve for thallium was derived from G-1 and two other granites that had previously been determined. Although internal standards were not used, the working curves and the duplicate values for the samples analysed were found to be most satisfactory.

The value of the standards used were:

	Tl p.p.m.	Cu p.p.m.	Pb p.p.m.	Ga p.p.m.
G-1 Granite	1.30 (1)	13 (2)	50 (2)	20 (2)
W-1 Diabase	-	110 (2)	10 (2)	18 (2)
O. G.16 Granite+	1.33			
Mbabane Granite+	1.63			

1 = Stevens and Fleischer (in press)
2 = Stevens et al (1960)
+ = Kaye (1962)

4) <u>Total Iron, Magnesium, Manganese, Cobalt and Nickel</u>: Following Ahrens and Taylor (1960), Palladium was used as internal standard for the determination of these elements. One part of sample was mixed with two parts of National Carbon Co. grade SP-2 graphite powder which contained 1% $[(NH_3)_4Pb](NO_3)_2$ by weight. Operating conditions were as before. The lines used were Pd 3258, Fe 2929, Mg 2783, Mn 2593, Ni 3414 and Co 3453. Plate calibration was by the self calibrating method of Ahrens and Taylor (1960).

			<u>G-1</u>	<u>W-1</u>	<u>U.S.B.S.</u>
Standards Used	G-1	W1	2	2	Plastic Clay 98
Total Fe %	1.37x	7.76x		4.56x	1.43+
Mg %	0.235x	3.99x		2.11x	0.434+
Mn p.p.m.	230x	1320x		775x	-
Ni p.p.m.	1.2x	80x		41x	-
Co p.p.m.	2.4x	52x		27.2x	16.5≠

x Stevens and others (1960)
+ U.S.B.S. Standard Sample
≠ Carr, H.M. and Turekian, K.K. (1961)

Although background corrections were necessary for cobalt and nickel, the working curves obtained for all the above elements were most satisfactory.

Table XVII : Duplicate Values

Specimen	K%	Na%	Rb p.p.m.	Li p.p.m.	Sn p.p.m. Approx.	Cu p.p.m.	Pb p.p.m.	Ga p.p.m.	Tl p.p.m.	Mg%	Fe%	Mn p.p.m.	Ni p.p.m.	Co p.p.m.
122a	4.024	2.713	292	3.9	5	11.5	27.0	28.0	Tr					
122b	4.050	2.771	263	3.5	5	8.9	22.4	28.2	tr					
143a	6.025	0.645	233	7.2										
143b	5.959	0.639	264	6.3										
565a	4.372	2.649	251	4.6	5	14.7	21.8	33.5	1.02					
565b	4.332	2.646	270	4.4	5	19.0	24.0	34.0	1.07					
577a	6.791	0.080	641	72	6	5.8	80	27.5	1.82					
577b	6.942	0.075	713	62	6	5.1	70	25.5	1.60					
43a	4.822	3.476	110	17.5										
43b	4.754	3.495	96	19.5										
43c	4.827	3.495												
28a	4.918	4.425	113	3.9										
28b	4.930	4.430	108	4.4										
275a	4.783	4.744	72	2.8										
275b	4.856	4.818	69	3.0										
468a	3.082	1.685	194	32.5	5	48.0	24.8	25.5	1.38					
468b	3.054	1.697	194	32.5	5	42.0	25.8	25.0	1.48					
360a	2.086	3.040	70	13.6						1.82	6.95	1180	15.6	27.3
360b	2.111	3.104	65	13.0						1.66	6.60	1120	18.6	31.1
334a	0.103	0.434								9.50	5.90	1680	154	67.0
334b	0.116	0.469								9.20	5.90	1620	156	71.0
571a					19.5	3.4	tr	64	4.05					
571b					20.5	3.1	tr	54	3.25					