STRATIGRAPHY, IGNEOUS PETROLOGY AND EVOLUTION OF THE SINCLAIR GROUP IN SOUTHERN SOUTH WEST AFRICA

by

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ABSTRACT

The investigation of an area fringing the Namib Desert in southern South West Africa has led to the clarification of relationships between several geological units of the Late Precambrian Sinclair Group, and to the formulation of a petrogenetic scheme and geotectonic model of evolution.

In the area studied, the oldest exposed unit of the Group is the hybrid and heterogeneous Haremub granite which is overlain by the predominantly arkosic Kunjas Formation. This is followed conformably by the Barby Formation, consisting predominantly of basic and felsic lava flows and various volcaniclastic beds including ash-flow tuff deposits. Basic lavas constitute the bulk of the succession and are typically highly porphyritic. A preliminary 'member' subdivision of the Barby sequence is suggested, based mainly on easily-recognisable field and petrographic characteristics. Gabbroic and noritic intrusives in the southwestern part of the area relate to the Barby basic magmatic phase. The following *en bloc* emplacement of the northwesterly elongated body of Spes Bona syenite was accompanied by intrusion of monzonitic and dioritic magma along fracture zones marginal to the syenite. The emplacement into high crustal levels of large volumes of the porphyritic/granophyric Nubib granite (1360±50 Ma)took place within a major northwest trending zone of crustal weakness which also marked the site of the developing Nam Shear Belt. This shearing imposed a strong but localised mylonitic imprint on all but the youngest intrusions of Nubib granite and on the Barby basic lavas invaded by the granite. Post-dating the Nubib granite and the main phase of shearing is the mixed sedimentary-volcanic Guperas Formation. Deposition took place mainly within a prominent north-south trending graben structure, and volcanic activity was apparently most intense in three fairly distinct centres. Ash-flows and felsic lavas were extruded from vents, now preserved as plug-like bodies. The intrusion of dense composite swarms of basic and felsic dykes took place toward the end of the Guperas phase. The Rooiberg granite (±1270 Ma) is intrusive into the Guperas. The Auborus Formation, comprising about 2,600 m of red felspathic sandstone and conglomerate represents the youngest unit in the evolution of the Sinclair Group. It is preserved in prominent north-south trending graben structures.

It has been possible to postulate distinct parent magma types on the basis of major and trace element analyses carried out on sixty-seven samples from the Sinclair Group. The felsic rock-types can be differentiated into a relatively high-Ca group and a relatively low-Ca group, the latter being dominant. The basic rock-types comprise both tholeiitic and calc-alkaline types and the latter are characteristically rich in potassium and associated elements, and have been classed as 'shoshonites'. There exists no direct genetic relationship between the basic and felsic rock-types.

It is proposed that the calc-alkaline magmas were derived by 'dry' partial melting of eclogite (transformed oceanic crust) within an ancient subduction zone at depths of about 150-175 km where the incorporation of phlogopite into the melts could take place. Varying degrees of partial melting produced magmas of varying potassium content, the lowest degrees resulting in the production of the Spes Bona syenite magma. Tholeiitic magmas probably resulted from partial melting of upper mantle peridotite overlying the ancient zone of subduction.

The repeated injections of basic magma into the continental crust resulted in a raising of the geotherms allowing partial fusion of the sialic material to take place in the lower crust. At the deepest levels high-Ca felsic magmas were the dominant products whilst at higher levels low-Ca melts were produced.

The evolution of the Sinclair Group is seen to have proceeded within three major cycles. Each cycle, with the possible exception of the first, was initiated by basic magmatic activity which was followed by felsic magmatic activity, vertical tectonism and the formation of local fault-troughs into which immature clastic detritus from local provenance areas was deposited.

The distribution pattern of the Sinclair Group and its probable correlates in central South West Africa and western Botswana describes a prominent and extensive curvilinear feature. Due to the prominence of volcanic, volcaniclastic and plutonic rock-types within this belt the name Rehoboth Magmatic Arc is proposed for this feature.

It is tentatively suggested that the Rehoboth Magmatic Arc represents an ancient magmatic arc developed at the margin of the Kalahari Plate as a result of active consumption of an oceanic crustal plate during Late Precambrian times (±1350-950 Ma).

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STRATIGRAPHY, IGNEOUS PETROLOGY AND EVOLUTION OF THE SINCLAIR GROUP IN SOUTHERN SOUTH WEST AFRICA

1 INTRODUCTION

1.1 GENERAL

The study area is situated immediately north of the Sinclair Mine in southern South West Africa (see locality map on the accompanying geological map) and lies between the Schwarzrand escarpment to the east and the fringes of the Namib desert to the west, approximately 150 km northeast of Lüderitz. The area is very mountainous, particularly in the east, with wide open plains becoming more extensive westward toward the Namib desert, where large dune fields predominate.

The area mapped covers approximately 6400 km^2 and is bounded by latitudes 25°00'S and 25°45'S, and longitudes 16°00'E and approximately 16°49'E. The centre of the area is about equidistant ($\sim$90 km) from the villages of Helmeringhausen and Maltahöhe, which lie to the southeast and northeast respectively. The now abandoned Sinclair copper mine and Duwisib Castle in the northeastern portion of the area, constitute well-known land-marks.

Almost the entire area is divided into large farms which are devoted to the breeding of karakul sheep for pelts.

1.2 PREVIOUS WORK

A concise account of the geological work carried out in the Sinclair-Helmeringhausen area prior to 1967 has been given by von Brunn (1967). The most important of these surveys includes the work of Range (1910, 1912), Beetz (1923, 1924), and Martin (1965).

Range (1912) first proposed the name "Konkipformation" for a series of tuffs, amygdaloidal lavas, porphyries and conglomerates in the vicinity of Helmeringhausen.

Beetz (1923) described these occurrences under the name of "Konkip System" which he regarded as consisting of the following:

Auborus Series	:	Red sandstone and conglomerate
Sinclair Series	:	Acid and basic volcanic rocks, conglomerate, tuff and breccia
Kunjas Series	:	Limestone, clay-slate, conglomerate quartzite

The presence of large stocks of red granite was noted by Beetz (1923) but he was uncertain of their relationship to the units mentioned above.

Beetz (1923) also reported the presence of numerous quartz-porphyry dykes and sheets occurring north of Helmeringhausen on the farms Aruab and Guperas and he considered these to represent a younger part of the Sinclair Series. More recent work (von Brunn, 1967; and this report) has shown that these particular felsic units belong to the newly-defined Guperas Formation. Beetz (1923) also made the observation that the "Sinclair Series" has its maximum development in the neighbourhood of Haremub, Sinclair and Ginas and that the basic lavas or "diabases" range from predominantly black or dark-coloured in the lower horizons to a vivid reddish-brown colour in the upper horizons.

The reconnaissance work of Martin (1965) resolved some of the problems associated with the stratigraphic position of the red granites occurring in the Helmeringhausen and Sinclair areas but the work of von Brunn (1967), in an area south of the present study area, represents the first systematic study on those units considered by Beetz (1923) as belonging to the "Konkip System".

During the course of von Brunn's study the complexity of the geological succession as compared with that proposed by Beetz (1923) necessitated certain name changes and additions, and the stratigraphy was redefined.

The name "Konkip System" was dropped and the major subdivisions were retained as formations. The term Nagatis Formation was introduced for a sequence of felsic lavas and sediments that were found to pre-date the Kunjas beds. These sediments and extrusives were recognised by Beetz (1923) but he was not able to satisfactorily establish their relationship to the "Konkip System".

Von Brunn (1967) included the "Kunjas Series", the "Barby Formation" (the "Sinclair Series" of Beetz 1923) and, provisionally, the "Guperas Series", into the "Sinclair Formation". Due to only very limited outcrop of the "Guperas Series" in the area west of Helmeringhausen and lack of data concerning its true nature and stratigraphic extent, the validity of including this unit into the "Sinclair Formation" was uncertain. Various local names were proposed for the different phases of granitic intrusion.

The geological succession as mapped by von Brunn (1967) can be summarised as follows:

Nama System		Black limestone, quartzite, sandstone, conglomerate
Auborus Formation		Red sandstone, conglomerate
Felsic porphyry dykes		
Rooikam and Tumuab granites		
Sinclair Formation	?(Guperas Series	Conglomerate, sandstone, flagstone, shale)?
	Barby Series	Basic lava, basal felsite, conglomerate, grit, arkose quartzite
	Kunjas Series	Conglomerate, arkose, shale
Tumuab granite		
Kotzērus granite		
Nagatis Formation		Felsic extrusives, basic lava, pyroclastics, arkose
Grey granodiorite		
Kheis System (?)		Various metarocks

FIGURE 1. GENERALISED LITHO-STRATIGRAPHIC SECTIONS OF THE GUPERAS FORMATION

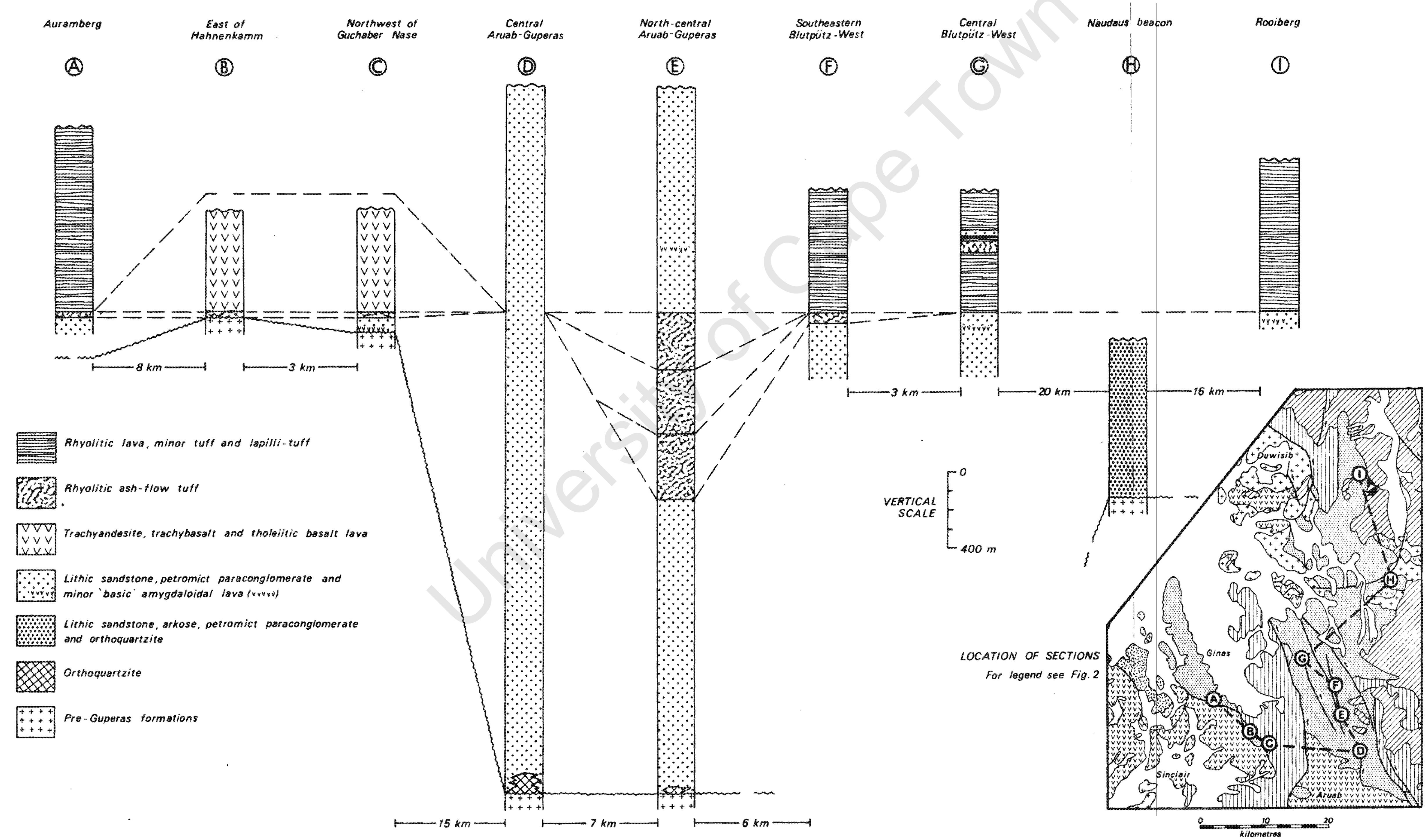
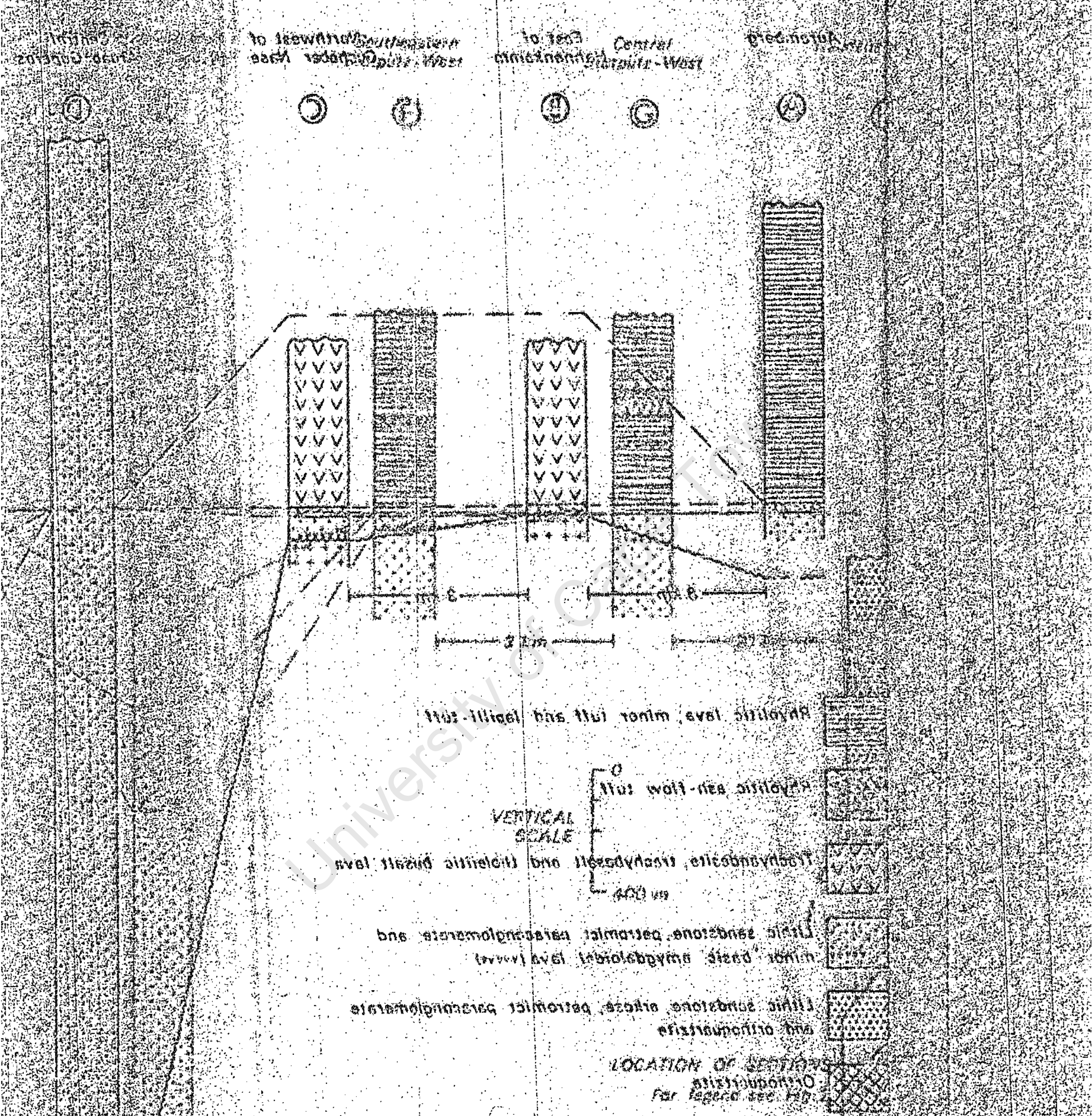


FIGURE 1. GENERALISED LITHO-STRATIGRAPHIC SECTIONS OF THE GUPERAS FORMATION



Miller (1967) has mapped and studied the Auborus Formation in some detail in outcrops occurring south of 25°30'S.

1.3 PRESENT STUDY

The works of von Brunn (1967) and Miller (1967), and the reconnaissance surveys of Martin (1965), brought to light certain problems and aspects of the post-basement/pre-Nama geology in the Sinclair-Helmeringhausen area that required further investigation before the full extent and nature of the local sequence of events could be elucidated and before these could be linked to the regional geological development of southern South West Africa. The prime objectives of the present study were:

(i) to extend the mapping of von Brunn (1967) northward to 25°00'S and to establish the mutual relationships between the major rock units,

(ii) to determine the stratigraphic relationships of the little-known "Guperas Series" (von Brunn, 1967, p. 8) and the rhyolite lavas mentioned by Martin (1965, p. 94-95) as occurring north of the Helmeringhausen area, and possibly constituting part of the "Guperas member" of the "Sinclair Formation",

(iii) to investigate the validity of including all post-basement/pre-Nama formations into a single Group and, more specifically, whether or not the igneous units merely constitute phases of a single prolonged magmatic episode,

(iv) to investigate, in greater detail, the basic units of the "Barby Series" defined by von Brunn (1967),

(v) to expand the petrographic and reconnaissance geochemical work carried out by von Brunn (1967) in order to verify and extend his conclusions concerning the genesis of the felsic rocks, to deduce the exact nature of the basic rocks and to formulate a scheme for the origin of the basic magmas,

(vi) to formulate a model for the evaluation of the post-basement/pre-Nama igneous and related rock units that would be compatible with the regional geological setting.

The area studied had not been previously mapped or systematically examined. The initial task was, therefore, the construction of a geological map. Extensive use was made of aerial photographs (at a scale of 1 : 36 000) during the course of fieldwork (carried out during 1970 and 1971). Geological detail was later transferred onto topographic base maps (at a scale of 1 : 100 000) employing the sketchmaster principle, which proved adequately accurate considering the purpose of the survey and the final scale of the geological maps (1 : 100 000). The final map was drawn on two separate sheets to a scale of 1 : 100 000.

Unfortunately, the geological map was draughted and printed some time before the study was completed and certain minor amendments involving interpretation have since been made to be consistent with the text and final conclusions. These changes are, however, discussed in the text where appropriate.

Major and trace element analyses were carried out using the facilities of the Department of Geochemistry, U.C.T.

1.4 ACKNOWLEDGEMENTS

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A special word of thanks is due to my fiancée, Angela, for continual encouragement, for many hours spent in editing the manuscript and for the draughting of the final diagrams.

2 GENERAL GEOLOGICAL SETTING

Since this thesis is specifically concerned with the intrusive and extrusive igneous and related volcaniclastic rock units of the Sinclair Group, this Section has been included to provide the reader with a somewhat broader view of the geology of the area north of the Sinclair Mine, South West Africa, and to place these igneous and related units into the correct perspective as regards their position in the entire geological succession.

In most cases the following descriptions of the various major rock units are brief, in particular of those units that are dealt with in greater detail in later Sections. The only exception to this is the Guperas Formation which is described in some length. The main reasons for this are:

(i) This work is the first to cover the succession of the Guperas Formation in its entirety and it is felt that at least the essential aspects of the stratigraphy should be recorded.

(ii) Lateral variations are so great that it is often difficult to define members and to provide a satisfactory idea of the relative position of the igneous and related units within the succession, without reasonably detailed descriptions.

(iii) The nature of the sedimentary beds of the formation provides some clue as to the tectonic environment prevailing at the time of deposition.

During the course of this study, the terminology as applied by von Brunn (1967) to the post-basement/pre-Nama rock units has been updated in accordance with the 'South African Code of Stratigraphic Terminology and Nomenclature'. Since all subdivisions coincide with lithological boundaries, a lithostratigraphic nomenclature has been used as suggested by Truswell (1967), Bliss (1968) and Newton (1968).

The geological succession as mapped north of the Sinclair Mine is as follows (see also Table 1):

2.1 KUMBIS FORMATION

The oldest rocks in the area have been grouped together in a formation with a local name since a correlation with other more widely known 'Archean Complex' units of South West Africa or Namaqualand is not possible at this stage.

Martin (1965, plate 1) and von Brunn (1967) have suggested that the old pre-Sinclair 'basement' rocks occurring in the southern half of the Sinclair and the Helmeringhausen areas might be correlated with the Kheis System.

The 'metarocks' of the northern part of the Sinclair area have been included in the Marienhof Formation by Martin (1965, plate 1) but it will be demonstrated later in this work that most, if not all of these rocks belong,

Table 1. Comparison of the geological succession and terminology between the area of the present investigation and the adjoining area west of Helmeringhausen (von Brunn, 1967)

<table>
<tr><td colspan="2">Area of present investigation</td><td colspan="2">Area west of Helmeringhausen (modified after von Brunn, 1967)</td><td colspan="2">Area west of Helmeringhausen (after von Brunn, 1967)</td></tr>
<tr><td colspan="2">Nama Group</td><td colspan="2">Nama Group</td><td colspan="2">Nama System</td></tr>
<tr><td>Auborus Formation</td><td rowspan="11">SINCLAIR GROUP</td><td>Auborus Formation</td><td rowspan="11">SINCLAIR GROUP</td><td colspan="2">Auborus Formation</td></tr>
<tr><td>Rooiberg granite</td><td>-</td><td colspan="2" rowspan="2">Linear-granite-porphyry dykes
Quartz-porphyry dyke swarm</td></tr>
<tr><td>Guperas Formation</td><td>Guperas Formation including linear granite-porphyry and quartz-porphyry dykes</td></tr>
<tr><td>Nubib and Barby 'metarocks'</td><td>Tiras gneiss (?)</td><td colspan="2" rowspan="2">Tumuab and Rooikam granites</td></tr>
<tr><td>Nubib granite</td><td>Rooikam granite (and Tumuab granite)</td></tr>
<tr><td>Spes Bona syenite</td><td>-</td><td>(Guperas Series) ?</td><td rowspan="3">SINCLAIR FORMATION</td></tr>
<tr><td>Barby Formation</td><td>Barby Formation</td><td>Barby Series</td></tr>
<tr><td>Kunjas Formation</td><td>Kunjas Formation</td><td>Kunjas Series</td></tr>
<tr><td rowspan="2">Haremub granite</td><td rowspan="2">Kotzērus hybrid granite</td><td colspan="2">Tumuab granite (and Tiras gneiss)</td></tr>
<tr><td colspan="2">Kotzērus hybrid granite</td></tr>
<tr><td>-</td><td>Nagatis Formation</td><td colspan="2">Nagatis Formation (including Okarus granite porphyry)</td></tr>
<tr><td colspan="2" rowspan="2">Single, small intrusion of grey granodiorite into Kumbis Formation</td><td colspan="2" rowspan="2">Grey granodiorite and gneissic granodiorite</td><td colspan="2">Tumuab granite intrusion (?)</td></tr>
<tr><td colspan="2">Grey granodiorite and gneissic granodiorite</td></tr>
<tr><td colspan="2">Kumbis Formation</td><td colspan="2">Kheis (?) Group</td><td colspan="2">Kheis (?) System</td></tr>
</table>

in fact, to the younger Sinclair Group and are, therefore, not 'basement' as was previously thought.

No detailed mapping program was carried out on the Kumbis Formation and, consequently, the relationships between the various units are not always clear. At present it is felt that a correlation with the Kheis System or the Marienhof Formation would be highly speculative and probably meaningless due to the lack of detailed information concerning the oldest rocks in the Sinclair area.

Large areas of Vergenoeg, Kumbis and to a lesser extent, Sonop, are underlain by a variety of rock-types included in the Kumbis Formation. The principle rock-type here is a fine-grained adamellite which may be weakly foliated and which intrudes various fine-grained amphibolites, and a "porphyroblastic metabasalt" (von Brunn, 1967). Cutting the adamellite on Vergenoeg is a small body of grey granodiorite.

To the northeast, mainly on the southern end of Kameelhof, faintly layered amphibolites together with overlying quartzites are included in this formation.

A sequence of fine-grained amphibolitic basic lavas and felsitic lavas occur at the Nagel beacon outcrop. These flows may form part of a succession including the amphibolites and quartzites of the previously described occurrence.

A narrow fault-bounded block of schistose, amphibolitic and gneissic rocks occurs trending northwestwards on Betta, Kameelhof and Vrede.

At the D 146 beacon a small inselberg consists of amphibolitic rocks which are cut by granitic material.

2.2 SINCLAIR GROUP

2.2.1 HAREMUB GRANITE

A small pluton of this granite with a total outcrop area of approximately 22 km^2 occurs on the farms Klein Haremub and Spes Bona, where it is overlain by the Kunjas Formation. The granite is typically red or grey in colour, granitic or granodioritic in composition and of a hybrid nature, and contains basic xenoliths. It intrudes amphibolites, presumably of the Kumbis Formation.

The reasons for including this granite in the Sinclair are based on the conclusion that it is almost certainly the correlate of von Brunn's (1967, p. 80-89) "Kotzērus hybrid granite" (Section 4), and since the younger Nagatis Formation (von Brunn, 1967) is regarded in this thesis as being an essential part of the Sinclair Group (Section 4) the inclusion of the Kotzērus granite and hence, the Haremub granite in it, is an obvious consequence.

2.2.2 KUNJAS FORMATION

This sequence comprises a variable thickness of clastic rocks consisting of a basal pebble conglomerate or grit, arkose, shale and dark brown quartzites overlying the Kumbis Formation and Haremub granite uncomfortably. Toward the top of the succession there is a progressive development of beds of white orthoquartzite.

Maximum thickness of the basal conglomerate in the extreme south is 120 m, but when traced along strike northwestward through Vergenoeg to Sonop it degenerates into an arkosic grit of only 5 m thickness. Total thickness of the formation in the northern part of Vergenoeg, where it attains a maximum development for the area, is approximately 2500 m. On southern Vergenoeg and Campbells Valley the sequence thins to 1500 m.

On Klein Haremub, the formation transgresses onto the Haremub granite pluton and here the total thickness varies between 700 m and 1400 m, indicating that the granite area was probably a topographic high at the time of deposition. There is no basal conglomerate or grit at this locality and the lowest part of the sequence here is represented by shales which are highly ferruginous in places and may contain a few very thin limestone lenses immediately above or at the granite-shale contact. These limestone lenses may be the stratigraphic equivalent of those observed by Beetz (1923) on the farm Korais in the area west of Helmeringhausen which were not located by von Brunn (1967) during subsequent mapping and must, therefore, be a very minor element, as they are on Klein Haremub.

The northernmost occurrence of the Kunjas Formation is on the farms Vrede, Kameelhof and Betta where 500 m of the upper and predominantly orthoquartzitic part of the sequence is exposed in a range of northwest trending hills.

2.2.3 BARBY FORMATION

(i) Volcano-clastic succession

The contact between this succession and the Kunjas Formation is rather obscure in the area north of Sinclair. Von Brunn (1967) followed Beetz (1923) in suggesting a disconformable relationship between the two units with the development of a conglomerate at the base of the Barby 'Series' (Formation in this thesis) overlain by grits, quartzitic sandstones, shales, arkoses, various felsic volcanics, volcaniclastic rocks and basic lava. In the area under discussion, however, no trace of the disconformity conglomerate was found and, furthermore, there exists a transition from typical Kunjas arkoses and shales into white quartzitic sandstones and grits of the basal Barby 'Series' of von Brunn (1967), thus indicating a conformable relationship. Overlying these quartzites and grits are the various volcanic rocks mentioned above.

It is proposed that the boundary between the Kunjas and Barby Formations be placed arbitrarily at the lowermost volcanic/sedimentary contact, a contact that is invariably well-defined and easily recognisable in the field, thereby making the Barby Formation a predominantly volcanic succession with certain exceptions to be mentioned later.

The most prominent and continuous outcrops of the Barby Formation occur south of latitude 25°30'S where an estimated thickness of at least 8500 m is exposed. Distinct lava types appear in fairly well-defined units and have a very local distribution, with limited lateral extent. An exception to this is the 'Basal volcaniclastic member' which is thin, very widespread and fairly constant in its composition, consisting of felsic ash-flow tuff, stratified tuffite and volcanic conglomerate. Overlying the basal sequence in the south are the various localised lava flow units mentioned above, consisting of highly characteristic felspar and/or pyroxene porphyritic basaltic-andesite, andesite, trachyandesite and rhyolite. Only relatively minor volcaniclastics appear intercalated with these flows and only the uppermost 400 m of the succession, as exposed on Aubures, consists of both normal clastic and volcaniclastic beds.

North of latitude 25°30'S the Formation consists of altered and frequently porphyritic basaltic lava, with an intercalated and intertongueing 'wedge' of felsic volcanic, volcaniclastic and minor basaltic lava and clastic beds. This 'wedge' thickens in a northerly and northwesterly direction, and in the Nubib mountains it constitutes approximately 3700 m of a succession that has an exposed thickness of 5100 m. True clastic beds, quartzite, arkose and sandstone, make up less than 400 m of this. The stratigraphy of these felsic volcanic and volcaniclastic units is described in more details in Section 3.1.3.1.

(ii) Intrusive units

Intrusives belonging to the Barby Formation take the form of:

(a) Rare felspar-porphyritic trachyandesite dykes.

(b) Felspar-porphyritic trachyandesite intrusives occurring mainly on Aruab.

(c) 'Gabbro' bodies that occur almost exclusively along the southwestern boundary of the area where they intrude Kumbis and Kunjas rocks. The intrusions range from small irregular masses and sill-like forms to large coarse-grained bodies which frequently display an igneous lamination. A wide variety of rock-types such as gabbro, quartz-gabbro, norite, anorthosite, olivine-gabbro and picrite are present.

2.2.4 SPES BONA SYENITE AND ASSOCIATED BASIC INTRUSIVES

Post-dating the various units of the Barby Formation and pre-dating the Nubib granite is a very coarse-grained grey syenite which forms a roughly oval-shaped body in the west-central part of the area. Almost without exception the syenite intrusion is surrounded by a zone of monzonitic and dioritic intrusions. The syenite post-dates the basic rocks in at least one locality and available evidence points to a genetic relationship between the syenite and these basic intrusives.

2.2.5 NUBIB GRANITE

Varying textural types of this granite, which forms large high-level masses, dyke- and sill-like bodies, intrude all afore-mentioned rock-types and are particularly prominent in a broad northwest-trending strip in the northern part of the area. Multiple intrusion is evident and typically the rock is a fine- to medium-grained porphyritic and/or granophyric granite of prominent red colour. Total outcrop area is in the order of 400 km^2. Lesser amounts of flow-banded quartz-felspar porphyry and felsite are related to the same phase of intrusion.

2.2.6 NUBIB AND BARBY METAROCKS OF THE NAM SHEAR BELT

In the north-eastern and north-central parts of the area basic Barby lava and the intrusive Nubib granite have been foliated, mylonitised in parts and metamorphosed, within a definite belt, named the Nam Shear Belt. There exists a complete transition over a distance of only a few kilometres from virtually undeformed and unmetamorphosed basalt and granite into gneissic granites and amphibolites, the latter frequently displaying a remnant felsparphyric and/or amygdaloidal texture so typical of much of the Barby basic lavas. In the central parts of the belt, however, where mylonitisation has been particularly severe, banded amphibolitic rocks and true gneisses occur. Minor quartz-felspar segregations and pegmatites (quartz, felspar and rare mica) have also been sporadically secreted in the central zones. The foliation is generally vertical, striking 130°-160°.

It is likely that this cataclastic deformation and metamorphism took place mainly during the late stages of intrusion of the Nubib granite. The nature of the granite bodies suggests predominantly pre-tectonic emplacement but also some syn- and post-tectonic intrusion. The main period of shearing pre-dates the Guperas Formation, or at least the upper, volcanic part of the sequence, by virtue of the fact that quartz-porphyry dykes of the Guperas cut the foliated gneisses and amphibolites.

2.2.7 GUPERAS FORMATION

The name 'Guperas Member' (of the Sinclair Formation) was first proposed by Martin (1965, pp. 94-95) for a succession of red rhyolitic flows overlain by coarse conglomerates and boulder beds with some intercalated amygdaloidal lava flows, occurring typically on the farm Guperas. The possibility of an unconformable relationship to the lower part of the 'Sinclair Formation' (Martin, 1965) was also mentioned, as was the possibility of a still higher member of felsic lavas - the supposed eruptive products of a northeasterly striking swarm of red quartz-porphyry dykes cutting the Guperas beds.

Beetz (1923) makes mention of 'red to reddish-brown quartz-porphyry' which he found to be younger than certain 'diabases' of the 'Ginasberge' (situated in the southern part of Ginas). The latter are almost certainly the lavas of the Barby Formation, whereas the former must represent felsic flows of the Guperas Formation.

Von Brunn (1967), mapping to the south, encountered limited outcrops of conglomerate, sandstone, flagstone and purple shale overlying the Barby 'Series' unconformably. These various sedimentary beds were included in the Guperas 'Series' of the Sinclair 'Formation' by the above author.

As a result of the present investigation it is possible to define the extent and stratigraphy of the Guperas Formation rather more fully. The formation is preserved within two approximately north-northeast and north trending graben structures of post-Guperas development. Geographically these are located in the south-central part of the area extending from Kronenberg through Auramberg to Guchaber Nase, and in the southeastern and eastern part of the area extending in a broad strip from Wittmanshaar northward to Rooiberg. Scattered basal remnants occur on Aruab and Wittmanshaar (reported by von Brunn, 1967).

The stratigraphy of the Guperas is extremely variable as can be seen from Figure 1, in which an attempt has been made to illustrate such variations by presenting sections through the sequence at various localities.

Essentially, the succession, which overlies the Barby Formation and Nubib granite unconformably, consists of a variable thickness (0-3700 m) of lithic sandstone of subgreywacke class (Pettijohn, 1957, p. 291), and petromict paraconglomerate (Pettijohn, 1957, p. 255) in part overlain by, or with an intercalation of, rhyolite lava and lesser amounts of ash-flow tuff. The thickness of the eruptives varies from 0-1000 m. Basic lava is also prominent in some areas. It can also be mentioned that the sandstone and conglomerate form only the end members of a spectrum of immature sedimentary rock-types including for example 'pebbly lithic sandstones'.

When dealing with such a variable sequence, however, generalisations are not very meaningful and many of the features observed in exposures can be adequately described with reference to particular stratigraphic sections. Hence each major 'facies', as represented by a section or sections in Figure 1 will be described briefly and minor variations not represented in the columns will be mentioned in passing and are discussed in greater detail in Section 3.1.6.

Basic and felsic dykes and other intrusive bodies related to the extrusives constitute an important part of the Guperas Formation and will also be discussed in later Sections (Sections 3.1.6.3 and 3.1.6.5).

All members of the Guperas Formation are unconformably overlain by the Auborus Formation or Nama Group and the sandstone, conglomerate, pyroclastics and rhyolitic lava flows east of Duwisib have been invaded by small bodies of Rooiberg granite.

(a) Auramberg section

In the southern part of Ginas, in the vicinity of the Auramberg trigonometrical beacon, the Guperas Formation has been downfaulted and brought into tectonic contact with the older Barby basic lavas; the base of the succession is, therefore, not exposed. Approximately 100 m of cross-bedded, ripple-marked and mud-cracked brown lithic sandstone and petromict paraconglomerate is exposed, overlain by an estimated 1000 metres of rhyolitic lava which is frequently flow-banded and has a thin ash-flow tuff unit at the base. Northward, these extrusives are exposed continuously for 20 km forming a very prominent mountain chain.

(b) Section approximately 2 km east of Hahnenkamm trigonometrical beacon

About 8 km from the previous section in a southwesterly direction, the sequence has changed considerably. No sediment occurs at the base of the succession and the thin (approximately 50 m) rhyolite ash-flow tuff unit rests directly and unconformably on the Barby Formation lava. Conformably overlying these felsic eruptives is a minimum exposed thickness of 520 m of basic amygdaloidal lava.

(c) Section approximately 2 km northwest of Guchaber Nase

This section is essentially similar to the previous one with the exception of a petromict paraconglomerate (maximum thickness 80 m) with a volcanic sandstone matrix that is present at the base immediately below the ash-flow tuff. In a few places a thin (up to 10 m) basic amygdaloidal lava band forms the basal part of the sequence underlying the conglomerate.

(d) Central Aruab-Guperas section

The sequence here is almost entirely sedimentary, consisting of an estimated 3700 m of red to brown and greenish-brown lithic sandstone, pebbly lithic sandstone and petromict paraconglomerate with a thin (30 m) bed of tuffite and a thin basic amygdaloidal lava flow of very limited lateral extent.

Cross-bedding in the sandstone is ubiquitous and ripple-marking and mud-crack features are common in the finer and more flaggy sandstones. The conglomerate beds have a restricted lateral extent and are of lenticular or wedge form.

Unique, lenticular, basal developments of cross-bedded white, pale pink and minor red-coloured orthoquartzite occur within this area. Two such occurrences appear at the Guperas-Aruab boundary and a third forms the very prominent Aruab mountain which is an outlier forming a plunging synclinal structure, truncated to the west by post-Auborus faulting (Plate 1). Maximum thickness attained by these orthoquartzite lenses is 150 m and lateral extent is limited to the range 1-3 km. True shales within the sequence are extremely rare and occur only as thin discontinuous beds. The flaggy sandstones are not very prominent in the section described but increase in importance toward the Aruab-Ganaams boundary i.e. about mid-way up in the exposed succession.

(e) North-central Aruab-Guperas section

The major difference between this and the previous section is the appearance of a volcaniclastic intercalation composed of three successive ash-flow tuff units situated approximately 1600 m above the base. The volcanic units constitute a tongue which thickens north-northwestward to a total maximum of about 900 m and pinches out in the opposite direction. A fairly persistent band of basic amygdaloidal lava which is approximately 10 m thick occurs about 350 m above the upper surface of the ash-flow intercalation.

(f) Southeastern Blutpütz West section

Between this section and that of the north-central Aruab-Guperas area, the ash-flow tuff wedge once again thins out drastically to about 50 m but is now overlain by rhyolite lava flows, with a minimum thickness of some 650 m, rather than the sandstone and conglomerate that overlie the ash-flow tuff to the southeast.

Underlying the tuff bed is a minimum thickness of 250 m of lithic sandstone and conglomerate. Once again basic amygdaloidal lava flows are intercalated with the sediments, in this case immediately below the ash-flow tuff/ sediment contact and invariably within the uppermost 150 m of the latter. In rare instances the tuff rests directly on the basic lava.

(g) Central-north Blutpütz West section

The base of the succession is nowhere exposed here and a minimum of 250 m of lithic sandstone, pebbly lithic sandstone and petromict paraconglomerate is overlain by at least 650 m of rhyolite lava with an intercalation of highly lithic ash-flow tuff (approximately 60 m thick) and a thin lithic sandstone/ conglomerate bed (approximately 30 m thick), about two-thirds of the way up from the base of the extrusives. No ash-flow tuff appears at the base of the rhyolite extrusive succession because the thick units of the north-central Aruab-Guperas area have lensed out completely in this direction.

(h) Naudaus beacon section

Proceeding from Blutpütz West toward the Naudaus trigonometrical beacon the stratigraphy of the Guperas Formation changes once again. In this section no lava is in evidence overlying the sediments, having either been removed by erosion or not deposited at all. Basic lava flows also do not appear intercalated within the succession. The typical red to brown sandstone and conglomerate of the preceding sections is notably absent and replaced by a succession of grey to brown lithic sandstones, arkoses and minor conglomerates and orthoquartzites with a minimum thickness of approximately 850 m. Generally the sediments here are richer in felspar and poorer in lithic fragments than in other areas.

(i) Rooiberg section

In this area the sediments have reverted to the more typical red-brown lithic sandstone and petromict paraconglomerate as occurring on Blutpütz West and further southward. Only a thickness of 70 m is exposed, grading into lapilli-tuff which constitutes the upper 100-150 m of the succession immediately below an 800 m thick pile of rhyolite lava. A thin (5 m) basic amygdaloidal lava flow is present about 40 m below the base of the rhyolite sequence.

Mention should be made here of the appearance of lithic sandstone and petromict conglomerate to the west of the area mapped, i.e. in the Awasib Mountain area. These sediments overlie, unconformably, basic lava, often amygdaloidal and/or felspar phyric, and various intrusive granites. The sediments bear a remarkable resemblance to the rocks of the Naudaus beacon area but are coarser-grained in general. No volcanic activity is in evidence but the sequence is cut by dykes and large bodies of quartz porphyry. A correlation of these lithic sandstones, conglomerates and quartz-porphyry intrusives with the Guperas Formation seems most likely.

Nature of the sediment

As already mentioned, the vast majority of the Guperas sedimentary succession consists of lithic sandstones, almost entirely of the subgreywacke class (Pettijohn, 1957, p. 291), and petromict paraconglomerate (Pettijohn, 1957, p. 255). The sandstone constituents are fairly well sorted, not very well rounded, lack any significant amount of matrix and the lithic fragments are predominantly of volcanic origin. The conglomerates typically have a disruptive framework, pebbles are generally well rounded and composed of igneous and sedimentary rock-types that can be matched with older formations in the immediate vicinity, e.g. felspar-phyric basic lava of the Barby Formation, porphyritic-granophyric granite of the Nubib intrusive phase and quartzite possibly derived from the Kunjas Formation. Very minor amounts of pre-Sinclair 'basement' rocks are also represented in the pebble fraction. Occasionally the matrix of the conglomerate becomes virtually 100 per cent volcanic material (e.g. in the section 2 km northwest of Guchaber Nase) and the rock may be descriptively called a 'petromict paraconglomerate with a volcanic sandstone matrix'.

The orthoquartzite lenses occurring at the base of the succession in the Aruab area are unique in that they represent the only very mature sediments in the entire sequence. The rock consists of very well rounded quartz and minor chert grains which have been well cemented by authigenic silica resulting in well marked overgrowth features.

The lithic sandstones of the Naudaus beacon are also subgreywackes, despite their more mature and indurated appearance when compared in hand-specimen with the sandstone of the main Aruab-Guperas-Blutpütz sequences. Lithic fragments in this case consist almost exclusively of granophyric intergrowth and fine-grained quartz-felspar porphyry material derived directly from the underlying and adjacent Nubib granite and porphyry. The sandstones are quite well cemented, possibly due to their generally more silica-rich nature. The fine sediments in this sequence tend to be more arkosic and subarkosic which is most probably due to the more advanced mechanical breakdown of the quartz-felspar intergrowth grains into individual mineral components. The conglomerates tend to be 'cleaner' than in other sections and lack an abundance of volcanic rock pebbles. They do, however, contain many granite and quartz-felspar porphyry inclusions and are therefore still petromict.

The volcaniclastic deposits which are always stratigraphically related to the extrusives, are discussed in detail in Section 3.1.6.4.

2.2.8 ROOIBERG GRANITE

Intruding the Guperas lithic sandstones, conglomerates, pyroclastics and rhyolite lava flows east of Duwisib Castle is a fine- to medium-grained porphyritic and granophyric granite which is petrographically indistinguishable from much of the Nubib granite. Only three small plutons with an approximate total outcrop area of 2 km^2 are present. This granite probably represents the youngest igneous unit of the Sinclair Group and hence the youngest igneous rock of the entire geological succession.

2.2.9 AUBORUS FORMATION

On the accompanying geological map this formation appears excluded from the Sinclair Group. However, since the map was completed during the early stages of this investigation, subsequent consideration of the evolution and tectonic setting of the Group as a whole has indicated the desirability of including the Auborus beds into the sequence (Section 5).

It is proposed, therefore, that the Auborus Formation be regarded as the youngest unit of the Sinclair Group.

Elongate block-faulted exposures of the Auborus Formation occur in the south-central, southeastern and northeastern parts of the area where it overlies various other units of the Sinclair Group unconformably and is in turn overlain in a similar manner by strata of the Nama Group.

According to Miller (1967) who mapped and described the outcrops south of latitude 25°30'S, an estimated thickness of about 2600 m is attained. The Auborus Formation outcropping north of this line does not appear to be any different, consisting of bright red felspathic and arkosic sandstones, and poorly-sorted conglomerate, which is mainly developed in the basal part of the sequence.

The immature nature of the sediment and the poorly sorted nature of the basal parts of the sequence indicate rapid erosion in the source areas and for the lower parts of the formation, rapid deposition. This is further indicated by the extremely local provenance of inclusions in the conglomerates (Miller, 1967, p. 5-7). Pebble-types are invariably recognisable as having been derived from older units of the Sinclair Group outcropping within a few kilometres of the Auborus occurrences. As the sedimentary basin, which was probably not very much larger than the area covered by present day outcrops, filled up, "... the duration of transport must have increased as the pebble size diminishes higher up in the basal conglomerate and sorting and roundness improve. New material was also introduced into the basin from sources further afield (cf. the appearance of granite and gneiss pebbles)" (Miller, 1967, p. 22). The shales at the top of the succession in the south are probably representative of the closing stages of this sedimentary cycle when the influx of sediment into the basin was at a minimum as regards quantity and rate.

Evidence for extrusive or intrusive igneous activity is conspicuously lacking in the Auborus Formation.

2.3 NAMA GROUP

Essentially horizontally-bedded sedimentary rocks of the Nama Group (Germs, 1972) cover much of the eastern part of the area in addition to numerous small outliers which become increasingly less frequent in a west-southwesterly direction. Pre-Nama formations in the vicinity of Osis form a prominent ridge and many hills composed of these older Sinclair Group rocks reach heights of up to a few hundred metres above the base of the Nama. According to Germs (1972) the Osis ridge has had a marked effect on the deposition of the Kuibis Formation (including the Clastic Member and Schwarzkalk Limestone Member) and to a lesser extent on the deposition of the Schwarzrand Formation (including the Basal Clastic Member, Nasep Quartzite Member and Nomtsas Clastic Member). Facies differences to the north and south of the ridge, as well as progressive changes in facies and general increase in thickness of the various units away from the ridge, are characteristic of the Kuibis and Schwarzrand Formations, the former in particular.

Although variable, the Kuibis Formation in the area generally consists of conglomerates and quartzitic-felspathic sandstones (Clastic Member) and black limestones (Schwarzkalk Limestone Member). The Schwarzrand Formation is essentially a shale and quartzite sequence.

The reader is referred to the work of Germs (1972) for further details of the lower Nama Group in the area since no detailed study of these rocks was carried out during the present investigation.

2.4 STRUCTURE AND METAMORPHISM

Large-scale and regional structural features of the area are controlled by two dominant trends, one generally northwest and another north to north-northwest.

The northwesterly structures display both compressional and tensional features and appear to have been dominant up to Guperas Formation times. This latter formation and all younger units display only minor effects of this regional lineament.

Evidence for this structural trend is provided by the following features:

(i) Strike of foliation and bedding (where recognisable) in Kumbis volcanics and metasediments in the Kameelhof, Vrede and Betta area.

(ii) Strike of the Kunjas and Barby Formation in exposures on Vrede-Betta, Duwisib, Rooiberg Süd, Osis, Aruab, Wittmanshaar.

(iii) The elongate shape of the Spes Bona syenite body (Fig. 5).

(iv) Normal faulting in the Vrede-Betta area.

(v) A prominent mylonite belt, the Nam Shear Belt, occurring in the north-central part of the area. Shearing in this belt has had a considerable effect on the Barby Formation rocks and Nubib granite in this area and these effects have been described in Section 3.1.5.

(vi) Distribution of the main Nubib granite masses.

(vii) Strike of the Guperas Formation sedimentary beds on Aruab, Guperas and Wittmanshaar.

(viii) Post-Nama faulting, generally with only minor displacement.

This trend is also prominent south of the Sinclair area. Similar evidence has been listed by von Brunn (1967).

The north to northwesterly trend mainly affects the Guperas Formation and younger units including the Auborus Formation and Nama Group and is the result of predominantly tensional deformation. This trend is represented by the following structural features:

(i) Strike of the Barby Formation on Nubib West. This represents the oldest evidence for this trend.

(ii) Dense quartz-porphyry and basic dyke swarms of the Guperas Formation occurring mainly in the eastern part of the area.

(iii) Dip of Guperas Formation ash-flow tuff units and sedimentary beds on the farm Guperas.

(iv) Approximate north-south distribution of the ash-flow tuff units (Section 3.1.6.4).

(v) Extensive normal faulting of post-Guperas/pre-Nama age mainly in the southeastern part of the area, resulting in north to northwesterly trending horst- and graben-type structures. Many of these faults, however, also display minor post-Nama movement with variable direction of displacement.

(vi) Shape of the present-day outcrop areas of the Guperas and Auborus Formations. This is largely the result of prominent faulting discussed above but there is evidence that at least the southern Auborus basin of deposition was elongated approximately north-south (Miller, 1967, p. 20).

This trend is once again represented to the south of the present area of investigation (von Brunn, 1967) in north-south and northeast trending faults and late felsic dykes, post-dating the main granite bodies (Rooikam and Tumuab granites).

In addition to these two main trends, numerous localised structures are present:

(i) Arcuate faulting possibly associated with the Rooiberg and Auramberg-Kronenberg fields of Guperas volcanic activity (Section 3.1.6.4).

(ii) Northeastward-plunging shallow anticlinal structures affecting the Kunjas and Barby Formation in the southwestern part of the area.

(iii) Kunjas and Barby strata dipping radially away from the Haremub granite pluton on Klein Haremub and adjoining farms Aubures and Sinclair.

Bedding dips are generally low to moderate, i.e. 20°-45° and only occasionally attain higher angles. Dips are nearly always the result of tilting rather than folding.

Metamorphic effects only become significant in the Nam Shear Belt and these are discussed more fully in Section 3.1.5.

3 VOLCANO-PLUTONIC ROCK UNITS OF THE SINCLAIR GROUP

3.1 DISTRIBUTION, FIELD CHARACTER AND PETROGRAPHY

3.1.1 HAREMUB GRANITE

Distribution and Field Character

A small pluton of this granite, with a total outcrop area of approximately 22 km^2 occurs on the farms Klein Haremub and Spés Bona.

Several other far smaller bodies of granite, intrusive into various rocks of the Kumbis Formation, have been attributed to the same intrusive phase that gave rise to the main body of Haremub granite. These smaller bodies occur in the central part of Vergenoeg, at the D 146 beacon, in the vicinity of the Nagel beacon, on southern Kameelhof and southwestern Sonop. In the case of the latter locality, a northwest-trending foliation is developed to a varying extent and in places granite-gneisses have resulted.

Typically, the Haremub granite is hybrid and heterogeneous with respect to colour, gross texture, modal mineral proportions and xenolith content. The colour is variable from red to grey and frequently the reddish colour has a patchy distribution when viewed in hand-specimen. Composition varies with colour, the reddish variety tends to be granitic whereas the grey rocks, which also have a higher proportion of dark minerals, tend to be granodioritic. These various types do not form distinctly different intrusions but rather appear to be constituents of a single inhomogeneous mass. The texture as seen in hand-specimen is normal granitic (hypidiomorphic granular).

Characteristically, the granite contains an abundance of basic xenoliths which vary greatly in size, ranging from less than 1 cm up to several metres in diameter. With the exception of the smaller inclusions, the xenoliths have sharp outlines and subangular shapes. This seems to suggest that assimilation has been minimal, thus further indicating that the hybrid and heterogeneous nature of the granite results from contamination prior to its emplacement into the present country-rocks. The xenoliths appearing in the granite were derived from neighbouring country-rock as is indicated by the fact that where the granite intrudes amphibolites of the Kumbis Formation (on Klein Haremub and Spes Bona), basic amphibolitic inclusions are abundant whereas where the granite appears in Kumbis Formation adamellites (on Vergenoeg), basic xenoliths are very rare but the hybrid and heterogeneous nature still persists.

Von Brunn (1967), in discussing the Kotzērus granite west of Helmeringhausen, which is considered to be the correlative of the Haremub granite (Section 4), has arrived at similar conclusions concerning pre-emplacement contamination.

Relationships of the Haremub granite to other rock-types in the area are as follows:

(i) On Klein Haremub the granite is overlain by sedimentary strata of the Kunjas Formation. The relationship appears to be a transgressive one and indicates that the pluton formed a topographic high at the time of deposition since the Kunjas beds thin considerably toward the granite. Furthermore, the trace of the Kunjas-Haremub granite contact has a well-defined semicircular form with the sediments dipping radially away from the granite mass. It would seem, therefore, that the granite was not only a topographic high but was actively rising at the time of deposition of the Kunjas.

(ii) The Haremub granite is cut by a felspar-porphyry dyke of the Barby Formation and also by quartz-porphyry dykes of the Guperas Formation.

(iii) Near the homestead on Spes Bona the granite intrudes amphibolites presumably belonging to the Kumbis Formation. At the Nägel and D 146 beacons and on southern Kameelhof the Haremub granite also intrudes rocks of the Kumbis Formation.

(iv) On Vergenoeg a small body of the granite intrudes adamellite of the Kumbis Formation.

Petrography

The texture of the Haremub granitic/granodioritic rocks is hypidiomorphic granular with very occasional subporphyritic tendencies. Essential mineral constituents are alkali felspar, plagioclase felspar, quartz and altered ferromagnesian minerals. The relative proportions of these minerals varies considerably thus illustrating the heterogeneous nature of the 'granite' (Table 2).

Table 2. Modal analyses of Haremub granite

Sample	BW 128	BW 160	BW 1621	BW 1624
Quartz	36	29	22	38
Alkali felspar	40	49	14	35
Plagioclase felspar	16	14	48	22
Altered ferromagnesian minerals	7	8	15	5
Accessory minerals	<1	trace	<1	<1

Quartz occurs as irregular mosaic-like aggregates in which the individual crystals are always anhedral and show pronounced undulatory extinction. Contacts between the individual grains are ragged and interlocking with some evidence of recrystallisation. The margins of the aggregates display even greater amounts of recrystallisation, and redistribution throughout the rock is also prominent. The size of the crystals ranges from less than 0,2 mm up to 0,5 cm in diameter.

Alkali felspar is typically perthitic orthoclase with fine string-type exsolution lamellae and frequent carlsbad twinning. The very fine cross-hatch twinning of microline is also displayed in some specimens. In general the crystals are anhedral, measuring about 4 mm, but rarely, large (up to 2 cm) subhedral crystals are present.

Plagioclase felspar occurs as anhedral to subhedral and less frequent euhedral crystals. Saussuritisation is fairly well advanced in most samples. Multiple twinning is present but is not ubiquitously well developed. Crystals range in size from about 0,4 to 4 mm in length and are stumpy in outline. Small crystals are frequently enclosed by alkali felspar.

Patchy aggregates of chlorite, pale green fibrous amphibole, opaque ore minerals and epidote mark the presence of highly altered ferromagnesian minerals. Occasionally, flakes of biotite, only partly altered to chlorite are present.

Accessory minerals are sphene, zircon and apatite, and minor carbonate is secondary.

3.1.2 BARBY FORMATION

The Barby Formation, as defined in Section 2.2.3 and occurring within the area under discussion, consists of both extrusive and intrusive igneous rock-units, the former being the dominant phase, For convenience the two are described separately and volcaniclastic beds have been included in the extrusive units.

3.1.2.1 Extrusive Units

The products of extrusive volcanic activity, namely lava flows and volcaniclastic (Fisher, 1961) deposits, constitute by far the greatest proportion of rocks included in this formation. In terms of present-day outcrop , volcanic units of the Barby cover a total area of about 400 km^2.

The largest and most continuous exposures are in the southern part of the area, between latitudes 25°00'S and 25°45'S, forming a broad strip extending from Sonop and Vergenoeg in the west to Aruab and Wittmanshaar in the southeast. The outcrops extend farther into the southern adjacent area where, according to von Brunn (1967, p. 18) an additional 300 km^2 of exposure is present.

There are considerable outcrops of Barby extrusives between latitudes 25°00'S and 25°30'S but block-faulting, intense intrusion of Nubib granite and erosion has resulted in many disconnected exposures. These factors, together with strong alteration, mylonitisation in part (Section 3.1.5) and the lack of suitable marker beds, creates difficulties in regional correlation and accurate determination of the stratigraphic sequence.

In the southern outcrops definite differences in modal and chemical composition, texture and other physical characteristics made it possible to determine, with some degree of confidence, the stratigraphy of the Barby extrusives for this area, which is represented by lithostratigraphic columns A-H in Figure 2. Lava types tend to form such distinct units that the sequence has been subdivided into various informal units, which have been provisionally called 'members', until such time as the distribution and stratigraphic limits of each unit making up the Barby Formation can be defined in detail for the Sinclair and adjacent areas. When this more detailed information is available it should be possible to assign true 'Member' status to these units.

As is to be expected in such volcanic terrains, lateral variations are tremendous and, when viewed in cross-section, members have lense-shapes with stratiform emplacement of lava flows and invariably a certain amount of interfingering. Indications are that each member i.e. each distinctive lava type constituted a large individual volcano.

The stratigraphic sequence north of $25^{\circ}30'$S has also been determined, but with less detail than in the south. The sequence is represented in columns I-L of Figure 2. Fortunately, the presence of felsic volcanics and normal clastics forming a well-defined unit (Rhyolite member) within an otherwise monotonous succession of 'basic' lava, has to some extent facilitated the determination of stratigraphy and regional variation. The base of the succession is exposed in only one locality in the north, i.e. on Betta-Vrede, where it appears in outcrops isolated from the rest of the sequence by faulting and lack of exposure. As a result of this, the correlation between the upper parts of the respective sequences north of $25^{\circ}30'$S is rather uncertain. Certainly the lava types of the north were derived from different eruptive centres than those of the south.

The total thickness attained by the Barby Formation is considerable. The excellent exposures in the Klein Haremub-Aubures area indicate an accumulation of approximately 8500 m of volcanic, minor volcaniclastic and clastic material. The top of the succession is nowhere exposed but the appearance, in the uppermost exposed part of the sequence, of a 400 m thick pile of well-bedded clastics and volcaniclastics, which become increasingly mature upwards, seems to indicate that volcanic activity was waning at this stage of deposition and the figure of 8500 m may thus be fairly close to the actual maximum development.

Thicknesses in excess of 5000 m have also been determined for the Nubib mountain area and since the lower and upper limits of the sequence are not exposed, this may be a minimum figure.

Von Brunn (1967, p. 19) has estimated a thickness of 2000-2500 m for the lavas of the Barby 'Series' and although this may hold true for the area west of Helmeringhausen, this estimate is certainly very conservative for the Barby Formation north of latitude $25^{\circ}45'$S. Von Brunn (1967, p. 139) is of the opinion that *en toit d'usine* faulting has resulted in repetition of various parts of the sequence and he provides good evidence that such step-faulting has in fact taken place in the Rooikam-Lovedale area. Such stratigraphic repetition, if not recognised, would result in a gross over-estimate of the thicknesses involved.

The possibility of repetition due to faulting in the Klein Haremub-Aubures and Nubib mountain areas has been considered but no evidence for displacement could be found in the field. The various members making up the respective sequences, in the south in particular, are easily recognisable and nowhere repeated in the succession. Duplication would have to have taken place within individual members only and although this might have occurred it is felt that strike faulting on a scale large enough to appreciably affect the total thickness would have resulted in at least some repetition of individual members. Furthermore, contacts between members are generally of a transitional nature and not sharp as would be the case with extensive faulting.

The present-day outcrop distribution suggests that most, if not all, of the area mapped was once underlain by the Barby Formation and if a total thickness of 5000-8500 m is accepted for the sequence, the entire volume of magma erupted within the area covered by the present investigation could have been in the order of 40-50,000 km^3. When it is further considered that, (i) the Barby Formation covers at least 300 km^2 immediately to the south of the Sinclair area (von Brunn, 1967, p. 18), (ii) great thicknesses of Barby basic lava correlates occur west of the Sinclair area in the Awasib mountain area (Section 4) and, (iii) Schalk (pers comm.) reports the presence of Barby Formation rocks, largely volcanic, north of latitude 25°00'S, the total volume of magma brought to the surface during this eruptive episode must have been enormous.

The major proportion of the extrusive succession is made up of 'basic' lava, very often of distinctive plagioclase felspar-phyric character, but felsic lava and ash-flow tuff constitutes between 20 and 45 per cent with an average of 25 per cent of the total thickness, the greater proportion being present in the Nubib mountain sequence. This is a considerably higher proportion of felsic material than that occurring in the succession west of Helmeringhausen, where, according to von Brunn (1967, p. 18), felsic volcanics make up only 1-15 per cent of the sedimentary succession underlying the basic lavas and are considered by him as belonging to the Barby 'Series'. The present writer prefers to exclude these sedimentary beds from the Barby Formation (Section 2.2.3), and since they are much subordinate in thickness to the lavas, these felsic volcanics would certainly constitute less than 5 per cent of the total Barby Formation (as redefined in Section 2.2.3) south of the Sinclair area. There is, therefore, a marked increase in the 'felsic volcanic' to 'basic volcanic' ratio as one proceeds northwards.

With the exception of those occurrences that have been involved in the Nam Shear Belt, the Barby Formation has not suffered any extensive deformation. Beds dip at angles between 15° and 90° but are most frequently in the range 25°-45°. At most places the sequence has merely been tilted without any complex folding. In the Sonop-Vergenoeg-Campbell's Valley area the direction of dip varies from northward through eastward to southeastward as a result of broad folding about a northeasterly plunging anticlinal axis. If traced southwestward out of the area, the Barby Formation sequence occupies the northwestern limb of a broad synclinal structure (von Brunn, 1967, p. 19) which apparently also has an approximately northeasterly plunging axis paralleling that of the adjacent anticline.

On Klein Haremub the beds dip radially eastward through southward to southwestward away from the Haremub granite pluton. The most extensive exposures are toward Aubures and here the beds dip fairly constantly at 25°-40°. Dips on Aruab, Guperas and Wittmanshaar are steep, 60°-90° toward the south and southwest.

The strike of the main outcrops north of latitude 25°30'S is northwest in the Betta-Vrede-Gorab area swinging to north in the Nubib mountains. Dips are to the northeast and east respectively at moderate angles, i.e. 25°-45°.

The topography associated with Barby Formation outcrops is generally mountainous and rugged with steep debris-strewn slopes. The Hahnenkamm trigonometrical beacon, for instance, appears within Barby volcanic terrain at an altitude of 1864 m, some 600 m above the outwash plains immediately to the south. Occasionally, as on Aubures, the topography strongly reflects the varying resistance to erosion of alternating flows (Plate 2). On Aruab the topography can be more aptly described as "monotonous and undulating" as noted by von Brunn (1967, p. 18) west of Helmeringhausen.

Field relationships between the Barby Formation and other rock units are as follows:

(a) Overlies the Kunjas Formation conformably in exposures on Klein Haremub, Sonop, Vergenoeg and Campbell's Valley.

(b) Intruded by 'basic' bodies which are included at the top of the Barby Formation (Section 3.1.2.2).

(c) Intruded by Spes Bona syenite on Sonop and Spes Bona.

(d) Intensely invaded by all varieties of Nubib granite, particularly in the northern part of the area.

(e) Unconformably overlain by Guperas lithic sandstone, conglomerate and rarely pyroclastics at various localities from Aruab-Wittmanshaar northward to Duwisib-Rooiberg.

(f) Intersected by quartz-porphyry and basic dykes of the Guperas Formation.

(g) Unconformably overlain by the Auborus Formation and by Nama Group strata.

(i) Basal volcaniclastic member

Distribution and Field Character

This thin but remarkably persistent volcanic member has been assigned to the base of the Barby Formation (see also Section 2.2.3) and rests conformably on sedimentary rocks of the Kunjas Formation. Von Brunn (1967, p. 18), mapping to the south, found stratigraphically equivalent felsic volcaniclastic beds interstratified with sandstones and quartzites immediately underlying the base of the main basic lava sequence.

In the area under discussion, however, volcaniclastic beds of this member form a distinct and sharply-defined unit separating the sedimentary rocks of the Kunjas Formation from the main basic lava sequence of the Barby Formation. This widespread and persistent horizon forms a valuable and easily identifiable stratigraphic marker which, as a depositional unit, was probably deposited almost simultaneously over its entire extent, and for this reason the base of this unit is regarded as defining the base of the Barby Formation.

Rocks of this member are exposed in three localities:

(a) Sonop-Vergenoeg-Campbell's Valley, where the beds dip at angles of 30^{o}-50^{o} in a generally northerly, easterly and southeasterly direction as a result of shallow large-scale folding about a northeast-plunging axis. Total thickness on Sonop is 15 m, thickening to 150 m at Vergenoeg and thinning again to about 10 m on Campbell's Valley.

(b) Klein Haremub, where the Kunjas Formation sedimentary strata and overlying volcanics dip radially away from the Haremub granite mass. Total thickness of the felsic volcaniclastic member varies in this outcrop from 35 m in the west to 80 m in the east.

(c) Kameelhof and the adjoining portions of Vrede and Betta where, in a chain of northwesterly trending hills, the volcanic beds dip northeastward at about 30^{o}, sandwiched between Kunjas quartzites and Barby 'basic' lavas. Total thickness here is 200 m, representing the greatest thickness attained over the entire area.

By far the greater part of this member is made up of ash-flow tuff beds with recognisably distinct flow units varying in thickness between 2 and 100 m. In a particularly good exposure 2 km southeast of the farmhouse on Klein Haremub, as many as nine individual flows can be recognised building up a succession with a thickness of about 80 m. The number of flows in this locality is exceptional, however, and generally only one or two units are recognisable.

The tuff is typically red to grey or greenish-grey in colour on fresh surfaces, weathering to a pale red-brown or yellow colour. This latter property together with an extremely resistant and fairly massive nature, results in prominent outcrops contrasting strongly with the darker coloured overlying basic lavas and well-bedded underlying sedimentary strata. The Klein Haremub volcaniclastics are particularly prominent in outcrop since they have an exceptionally pale yellow to white colour. Rock fragments measuring up to 5 cm in diameter and consisting chiefly of fine-grained felsic material, are a common constituent of these ash-flow tuffs, and were it not for the inclusions extreme difficulty would be experienced in recognising the rock as being of pyroclastic flow origin when viewed within the confines of any limited outcrop. Compaction structure is not well developed but nevertheless imparts a faint fabric to the rock particularly in the Klein Haremub area. There is a complete lack of sorting and bedding within individual flow units and a vertically layered or zonal character has not been observed. The lack of pumice fragments is also notable and characteristic.

In some localities, in the eastern part of Vergenoeg in particular, reworking of the upper and presumably non-welded 2-3 m of the uppermost ash-flow unit has resulted in the formation of crudely-bedded tuffs and lapilli-tuffs (definition after Fisher, 1966, p. 292). The coarser lithic and crystal fraction is dominant in these rocks and apparently the finer vitric and crystal fragment groundmass of the original ash-flow deposit has been removed during reworking.

Less common, and occurring in the Vrede-Kameelhof-Betta outcrop only, are beds of stratified tuffite with a volcanic conglomerate (Fisher, 1961, p. 1410) interbed (Fig. 3). Stratified tuffite and interbedded volcanic

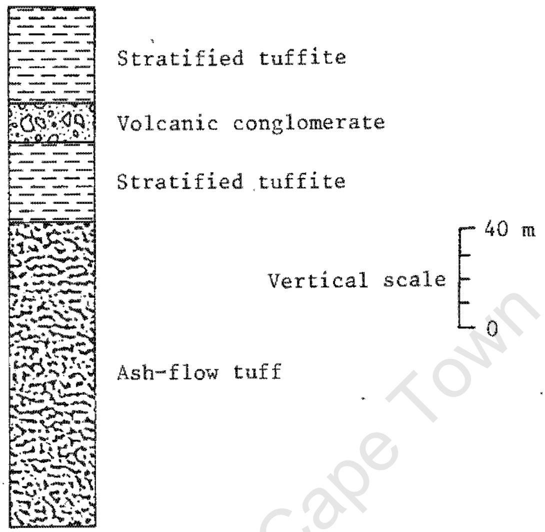


Fig. 3 Stratigraphic section of the Basal volcaniclastic member of the Barby Formation in the Kameelhof area

conglomerate make up a total thickness of 80 m and, overlying the ash-flow tuff, they constitute the uppermost beds of this member. West of Helmeringhausen, i.e. south of the present area, where the stratified tuffite is apparently more consistently developed, it persistently occupies the same stratigraphic position (von Brunn, 1967, pp. 139-142). The tuffite is a brown coloured rock with fine stratification or bedding due to alternating bands of varying grain size. Rare cross-bedding was noted.

The interbedded volcanic conglomerate is a deep red-coloured rock of an extremely poorly-sorted and non-bedded character. Angular fragments and blocks measure up to 10 cm in diameter but have an average size of about 1-2 cm and are set in a fine-grained matrix to form a disrupted framework. The term volcanic conglomerate is used here in a non-genetic sense as suggested by Fisher (1961). The original deposit was almost certainly pyroclastic and hence the rock could be called a lapillistone (Fisher, 1966), but in the absence of definite evidence for an airfall origin the non-genetic term is preferred.

Petrography

(a) Ash-flow tuffs

Pyroclastic characteristics are generally easily recognised in these rocks, but are not as well preserved as those of the younger Guperas Formation ash-flow tuffs (Section 3.1.6.4). Extensive devitrification, recrystallisation and redistribution of quartz has frequently resulted in partial or complete obliteration of shard structure in the groundmass. Lithic fragments are always present and, in the absence of obvious vitroclastic properties, give some clue as to the true nature of the rock. Phenocrysts of quartz, alkali and plagioclase felspar are important constituents which may have angular chip-like forms providing further evidence for a fragmental character (Plate 3).

The rocks range from vitric tuffs with low crystal and lithic fragment contents to lithic-crystal tuff rich in rock fragments, moderately rich in crystal material and relatively poor in a glassy groundmass component. The two extremes occur on Klein Haremub and Kameelhof respectively.

As a phenocryst mineral, quartz has the form of rounded and strongly embayed crystals measuring up to 2 mm in diameter, which frequently retain euhedral hexagonal or bipyramidal outlines, and smaller fragmented and highly angular grains. The volume of quartz phenocrysts and phenocryst fragments present varies between 2 and 10 per cent and quartz is thus the dominant phenocryst mineral. The proportion of fragmented to non-fragmented grains is variable, but of all the phenocryst minerals, the fragmented character is best developed in the quartz.

Alkali felspar phenocrysts are the dominant felspar insets and constitute less than 5 per cent by volume of the total rock. The crystals are cloudy, have a deep reddish-brown colour and are subhedral when unfragmented, with a maximum size of 2,5 mm. The felspar is orthoclase and is invariably finely perthitic. Carlsbad twinning is common. Replacement by albitic plagioclase, proceeding from the margins of the crystals was noted in one specimen only (BW 228). Alkali felspar phenocrysts may display considerable fragmentation and unbroken phenocrysts are generally well rounded and even embayed by resorption.

Plagioclase felspar is the least common phenocryst mineral and makes up less than 1 per cent of the total rock volume. Polysynthetically twinned unzoned crystals of stumpy subhedral form are typical. No glomeroporphyritic clustering occurs. Alteration to saussurite is usually well advanced. The highly shattered character of the quartz does not extend to the plagioclase phenocrysts and resorption effects are also less marked than in the case of the other two phenocryst minerals.

Rock fragments are an ubiquitous and diagnostic constituent of the ash-flow tuffs but vary considerably in the relative proportions present. The Kameelhof flows contain the highest proportion and those of Klein Haremub the lowest. Overall range is $\sim$ 1-35 per cent by volume and grain sizes range from less than 0,2 mm to 5 cm in diameter. The fragments are typically of a highly angular nature and are, in the vast majority, composed of cryptocrystalline devitrified felsic volcanic material, most commonly of a non-porphyritic habit. A very small proportion of the lithic fraction is composed of very fine-grained basic material and even more rarely of quartzite.

Pumice fragments, which, according to Ross and Smith (1961, p.18), constitute the most important single criterion for the recognition of the pyroclastic nature of ash-flow tuffs, are apparently completely absent from the pyroclastic flows of this member.

The groundmass was probably originally composed of glass shards and lesser amounts of fine-grained crystal fragments and dust. The glass has been completely devitrified to crypto-crystalline quartzo-felspathic intergrowths, and in some cases alteration to sericite is in an advanced state. The quartz is commonly recrystallised and redistributed.

Shard structures can be recognised in most of the flows (Plate 4), but the state of preservation is variable. In the best preserved examples devitrification has resulted in extremely fine fibrous structures, axiolites (Ross and Smith, 1961), developing within the boundaries of each individual shard. Thus, although crystallisation has been complete in these cases, the original texture is pseudomorphically preserved.

Welding and compaction has not been severe at all and the shards show very little (BW 228) or no (BW 508) distortion. In the former case, only a slight 'moulding' of the shards about inclusions and phenocrysts is observable. Only one specimen (BW 204), from eastern Vergenoeg, exhibits any indication of marked compression (and severe welding ?) and a faint pseudo-flow-banding is developed (Plate 3). Compaction and devitrification has all but destroyed any evidence of shard structure; only very vague and sparse,highly flattened Y- and U-shapes remain.

A granular type of devitrification is dominant in the ash-flow tuffs and this has to a great extent reduced the original shard-structured groundmass to a massive cryptocrystalline mass. Specimens BW 227, 218, 219 and 204 respectively, show various stages of shard preservation from near complete to virtually non-existent. Shards range from colourless to pale brown colour.

Ferromagnesian minerals or their alteration products are conspicuously absent. Opaque ore minerals in the form of small to minute granules are typically scattered throughout the groundmass.

(b) Stratified tuffite

These rocks are distinctly bimodal consisting of a dominant fine fraction of highly angular grains of quartz, felspar, opaque minerals and rock fragments, and a coarse fraction of angular to sub-rounded grains of felspar, quartz and rock fragments. Average sizes of these fractions and relative proportions present are about 0,08 and 0,75 mm, and 75 and 25 per cent respectively.

Alteration of relatively coarse- and fine-grained bands produces the conspicuous finely stratified character. The felspar of the coarse fraction is predominantly sodic plagioclase which may display subhedral to euhedral outlines. Rock fragments are invariably devitrified glass with lesser amounts of fine-grained basic material. Generally, the degree of rounding of the coarser fraction is only slight but noticeably more pronounced than that of the finer grains. The rock has been stained by red-brown iron oxide resulting in a brown colour in hand specimen. The weathered colour is a paler shade of brown.

(c) Volcanic conglomerate

The gross characteristics of these rocks have been described in the preceding Section. The coarser grain fraction is composed predominantly of red-coloured felsic igneous rock fragments, frequently of a porphyritic and/or flow-banded nature and welded ash-flow tuff fragments were also noted. Basic rock fragments are always present but never in great abundance. The finer fraction consists of quartz and felspar grains, rock fragments and an extremely fine component, probably a glassy volcanic dust, which is dark coloured due to the presence of dust-fine granules of opaque ore minerals. Felspar is highly altered and cloudy, and is predominantly plagioclase with lesser alkali-felspar. All grains are highly angular with no apparent rounding.

(ii) High-Ca rhyolite member

Distribution and Field Character

Rhyolite lava of a relatively high-Ca character (see Section 3.3.3) is prominent and persistently developed near the base of the Barby succession immediately above the Basal volcaniclastic member in the southern part of the area i.e. south of latitude 25°30'S. This particular rhyolite appears to be absent in the north, but lack of continuous exposure throughout the lower part of the succession casts some uncertainty on this supposition.

This rhyolite sequence is generally split up into two or three stratigraphic units by minor intercalations of 'small-felspar trachyandesite' and 'basaltic andesite', and where both upper and lower limits of the rhyolite 'zone' are exposed, its total maximum thickness varies between 800 and 1550 m. The greatest development was found on Vergenoeg.

A sparsely porphyritic texture is characteristic of the lava with phenocrysts or glomeroporphyritic aggregates of plagioclase felspar laths set in an aphanitic groundmass (Plate 5), ranging in colour from black through dark maroon to red. Flow-banding is always present but is never very strongly developed, hardly ever highly contorted and is generally planar, paralleling the presumed base of the flow fairly constantly. Amygdales are occasionally present, and appear as thin lenticular or discoid forms flattened within the planes of flow and therefore parallel to the flow banding.

Numerous bands of agglomerate up to 50 m thick were found intercalated within the rhyolite lavas only in the central part of Aruab. These pyroclastics consist of non-bedded accumulations of poorly-sorted and angular blocks or fragments of flow-banded rhyolitic material, of maximum size 15 cm. An occasional slight rounding of the blocks is attributed to attrition during transport within a volcanic vent. The lava in proximity to the pyroclastic beds is noticeably flow-banded, invariably in a contorted fashion.

Petrography

Plagioclase felspar and opaque ore minerals constitute the only phenocrystic mineral phases. The plagioclase felspar insets make up between <1 and 8 per cent of the rock and have a composition within the range An_{33-35}. Crystals measure up to 2,5 mm in length and most frequently occur as glomeroporphyritic clusters (Plate 5) which often consist of a large number of individual grains. The crystal form is subhedral to less commonly euhedral and anhedral, and outlines appear somewhat rounded by resorption. Alteration to saussurite is incipient to complete.

Opaque ore mineral phenocrysts are smaller (0,3 mm) and are present in trace amounts only. The grains are mainly anhedral and display resorption features, but subhedral to euhedral forms are also occasionally encountered.

Groundmass textures range from cryptocrystalline devitrification intergrowths of quartz and felspar to microcrystalline felted or trachytoid masses of poorly-developed felspar crystals and quartz. Devitrification products tend to be somewhat granular with a patchy 'poikilitic' relationship between the mineral phases. Minute ore granules occur scattered throughout and, when present in high proportion, result in the black colour of some of these rocks. The red colour is due to a red cloudiness that has developed in the alkali felspar, and maroon coloured rocks are those in which clouded felspar occurs together with a relatively high proportion of dust-fine opaque mineral grains.

Visible flow structure is due to a subparallel alignment of groundmass felspar crystallites, and/or varying concentrations of opaque mineral granules within alternating bands. The latter are generally fairly straight even when viewed in thin-section but may 'wrap around' phenocrysts or phenocryst aggregates (Plate 5).

Minute epidote granules occur ubiquitously throughout the groundmass and are the product of felspar alteration. A very sparse and patchy distribution of chlorite is present. From the nature of the occurrence it is not possible to say whether or not this mineral has resulted from the breakdown of fine-grained ferromagnesian minerals. Certainly, no pseudomorphic outlines are present.

Amygdales are composed predominantly of quartz with lesser amounts of calcite, epidote and chlorite.

(iii) Small-felspar trachyandesite member

Distribution and Field Character

Maximum development of this member occurs in the Vergenoeg-Sonop area, where it is composed entirely of flows of felspar porphyritic trachyandesite. Stratigraphically the member does not constitute a well-defined unit and much of the lava appears as intercalations in the underlying rhyolite and overlying basaltic andesite succession. The contact with the latter is transitional with a resultant zone of intercalations. The total exposed thickness is approximately 2000 m, of which 1100 m occur as an uninterrupted succession, the remaining 900 m constitutes the total thickness of the lava interbedded with the stratigraphically adjacent lava types mentioned above.

The most easterly exposures of this member are on eastern Klein Haremub and western Ginas where thicknesses are never in excess of 250 m. Lack of exposure does not permit delineation of the full extent of the member, but from the account of von Brunn (1967, p.42-44) it would appear that this lava type is also prominent and possibly dominant in the Barby succession to the south of the present area, i.e. west of Helmeringhausen.

The lava forms flows up to 250 m thick, and is typically a greenish-grey-coloured, medium-grained rock with abundant plagioclase felspar phenocrysts, and less frequent and smaller ferromagnesian mineral insets set in an aphanitic groundmass of variable proportions. The porphyritic character is very well displayed on weathered surface where pale yellowish-brown plagioclase insets and dark rust-red ferromagnesian mineral crystals contrast strongly with a light brown-coloured matrix. Amygdaloidal texture is frequently developed and the amygdales are invariably composed of quartz.

Petrography

Dominant textural features are phenocrysts of plagioclase felspar and clinopyroxene set in a fine-grained intergranular groundmass.

Plagioclase felspar phenocrysts are lath-shaped or tabular, have subhedral and anhedral to less frequent euhedral outlines, and make up between 25 and 37 per cent of the rock volume. Crystal size varies considerably, ranging from 0,4 to 4 mm and both extremes may occur in the same rock. Resorption effects are evidenced by a slight but common rounding of the crystals. Glomeroporphyritic aggregates are fairly common but not as frequent as might be expected. Zoning is not very conspicuous and is barely discernable in many flows; when present it is confined mainly to the extreme marginal parts of the phenocrysts. Composition ranges from calcic andesine (An_{43}) to sodic labradorite (An_{56}). Saussuritisation is incipient to complete, often preventing the microscopic determination of the felspar composition. Twinning is universal and according to the albite, carlsbad and pericline laws.

As a phenocryst mineral, clinopyroxene is subordinate to the plagioclase felspar in both amount ($\sim$3 to 15 per cent by volume) and size, measuring between 0,2 and 32 mm in length. The crystal form is subhedral to anhedral and occasionally euhedral. Partial alteration to pale green uralite amphibole is ubiquitous and generally proceeds inward from the margins of the crystals or cracks with a characteristic exsolution of fine granular opaque ore in the marginal parts of the crystals. Unaltered pyroxene is pale brown to neutral augite. Zoning is either absent or only barely discernible, but twinning is common and is both simple and repeated. Glomeroporphyritic aggregation is not at all common and although corrosion features were noted these are not typical.

Opaque ore minerals occasionally form subhedral to euhedral microphenocrysts measuring up to 0,8 mm in size.

The groundmass, which makes up between 12 and 48 per cent of the rock, is composed essentially of a second generation of anhedral to subhedral plagioclase crystals with subordinate amounts of intergranular altered clinopyroxene and accessory opaque ore mineral grains. Plagioclase crystals may be up to 0,4 mm in size. The presence of interstitial alkali felspar, as suggested by the chemical analysis and norms (Section 3.3.1) could not be microscopically verified due to the fine grain size. For the same reason and also because of alteration effects, the composition of the groundmass plagioclase could not be determined. The presence of small amounts of interstitial biotite (usually chloritised to varying degrees) is erratic.

Amygdales are composed of quartz, accicular uralitic amphibole and occasional epidote, and due to their small size are frequently only recognisable in thin-section.

(iv) Large-felspar trachyandesite member

Distribution and Field Character

Lava flows of this member are exceptionally conspicuous and unmistakeable in the field. Typically, the lava is plagioclase felspar-porphyritic (Plate 6) with large greenish-coloured crystals of the mineral, measuring up to 2,5 cm in length and set in a red-brown to bright red aphanitic groundmass. Relatively smaller dark green ferromagnesian mineral phenocrysts are also invariably present in lesser amounts. The large size of the felspar phenocrysts and the red to red-brown colour of the matrix are the most diagnostic properties from the point of view of field recognition. On weathered surface the rock takes on an even more conspicuous appearance with the plagioclase crystals becoming white, and the groundmass accquiring an even brighter shade of red.

Sparsely distributed and irregularly-shaped vesicles of a wide range in size are not uncommon, and these may be partly or completely filled with epidote and quartz. A subparallel alignment of the phenocrystic plagioclase laths is also common and is undoubtedly the result of flow, since it is generally parallel to the base of the units.

Recognition of individual flow units in the field is rendered rather difficult by the generally homogeneous nature of the lava, and weathering characteristics that result in boulder-strewn outcrops. However, flows appear to be rather thick, in the order of 50-100 m and possibly even thicker.

The occurrence of this member is restricted to eastern Klein Haremub/ southwestern Ginas. In this area the lava builds up to a fairly well-defined unit consisting of up to five sub-units of a total maximum thickness of 1300 m. Upper and lower parts of the member contain intercalations of stratigraphically adjacent lava types, illustrating the transitional relationships between the various members of the Barby Formation. The flows dip at moderate angles (30^{o} - 40^{o}) to the east.

This lava type is absent in sections on Vergenoeg, 15 km to the west and on Aruab, 29 km to the east-southeast. It has, therefore, a fairly restricted distribution but is nevertheless a locally very important and thick part of the succession.

Intrusions of felspar-porphyritic trachyandesite material into flows of the pyroxene trachyandesite member on Aruab bear an identical appearance to the flows discussed above (Section 3.1.2.2) and are also chemically very similar (Section 3.3.1).

Petrography

The texture is porphyritic with phenocrysts of plagioclase felspar, clinopyroxene, olivine and opaque ore minerals set in a microcrystalline groundmass (Plate 7).

By far the most dominant phenocryst mineral is plagioclase felspar constituting 12 to 48 per cent of the total rock volume and about 85 to 90 per cent by volume of the phenocryst fraction. The crystals are exceptionally large, ranging in size from 1,0 mm to 2,5 cm. The crystal form is generally subhedral to euhedral, and in a few cases magmatic corrosion has resulted in varying degrees of rounding. Glomeroporphyritic aggregates occur, but are not at all common and are composed of only a few individual crystals. Saussuritisation is well advanced, inhibiting to some extent the determination of felspar composition. A limited number of reliable determinations indicate a compositional range from An_{38} to An_{45}, i.e. calcic andesine. Thin rims of untwinned alkali felspar, which is in optical continuity with the adjacent groundmass felspar, are common phenomena. Twinning of plagioclase is according to the albite and carlsbad laws and, rarely, the pericline law. No compositional zoning is apparent. Dark altered inclusions of glass may be common.

Clinopyroxene phenocrysts are colourless to very pale green diopsidic augite which is partially or completely altered to pale-green fibrous amphibole, chlorite and epidote. The crystal shape is variable from anhedral to euhedral, and corroded forms, some even displaying embayments (e.g. BW 391), are not uncommon. As a phenocryst mineral clinopyroxene constitutes between 6,2 and 7 per cent by volume of the total rock. Twinning is common and may be relict in the altered crystals. The size of the clinopyroxene phenocrysts is generally much smaller than that of the plagioclase phenocrysts, ranging from 0,1 to 5 mm. As with the plagioclase, crystal aggregates are occasionally found.

Olivine occurs as small early-formed, resorbed and even embayed (BW 387) phenocrysts, which have in all cases been completely serpentinised with the exsolution of an opaque ore mineral (Plate 7). The pseudomorphs may also contain appreciable amounts of actinolite, calcite, chlorite, quartz and epidote which probably represent both alteration and replacement products. A relict coarse and irregular parting is also characteristic. The physical characteristics and nature of the alteration readily distinguishes olivine from clinopyroxene. Olivine makes up less than 2 per cent of the total rock volume and the anhedral to subhedral crystals may measure up to 4,5 mm in diameter. Glomeroporphyritic clusters occur fairly frequently and crystals are often partially or completely enclosed by plagioclase or clinopyroxene phenocrysts.

Small opaque ore mineral grains may also form phenocrysts. These are very well rounded by resorption, are present in minor amounts only and their size is in the range 0,1 to 0,7 mm.

The groundmass, which constitutes 43 to 86 per cent by volume of the total rock, is composed of very small sodic plagioclase felspar laths, which form cores to anhedral alkali felspar grains. Growth of the alkali felspar about the plagioclase is completely gradational, and no sharp distinction between the two exists. The plagioclase is, however, evidenced by vague polysynthetic twinning and lack of a conspicuous strong red-brown cloudiness that appears ubiquitously within the alkali felspar. Relative proportions of the two felspar phases present are difficult to estimate but are probably about equal, with the alkali phase dominating occasionally. An opaque ore mineral is a universal constituent of the groundmass and occurs as minute granules scattered throughout.

Interstitial quartz may be present in small amounts (e.g. BW 391), and is considered primary in spite of the presence of olivine. As demonstrated by the extensive resorption, the latter mineral is so obviously out of equilibrium in the rock that the presence of small amounts of primary quartz is not considered unusual.

The mode of occurrence of alteration and secondary products such as saussurite, pale-green fibrous amphibole, chlorite, serpentine, epidote, opaque ore and calcite has already been mentioned.

Apatite is an important accessory mineral and occurs as well-formed hexagonal crystals of prismatic habit. Amygdales are composed of epidote, quartz, calcite, fibrous amphibole and chlorite.

(v) Basaltic andesite member

Distribution and Field Character

Overlying, and in part intercalated with, flows of the previous member in the Spes Bona-northwestern Klein Haremub area, is a succession of pyroxene-porphyritic basaltic andesite lava flows. The total thickness of this rather local development is 1400 m and the only other occurrence of lava of this type is on eastern Klein Haremub where a thickness of 150 m appears to be intercalated within the uppermost flows of the High-Ca rhyolite member. The stratigraphic position of this lava band is, therefore, lower in the total succession than that of the 'type' area on Spes Bona and northeastern Klein Haremub. However, the lavas of the two occurrences are identical and have been included in the same unit since the source material must have been the same in both cases.

The lava has a characteristic greenish-grey colour on fresh surface, and is spotted with relatively sparse dark-green ferromagnesian mineral phenocrysts. The groundmass is aphanitic. On weathered surface the rock takes on an even more conspicuously spotted appearance. Phenocrysts are distinctly rounded, and close examination (on weathered surface) often reveals a very faint flow lineation within the groundmass. Amygdales are very rare and always small.

Petrography

Pyroxene phenocrysts or, more commonly, the altered pseudomorphs thereof, and erratic plagioclase phenocrysts occur in a crypto- to microcrystalline pilotaxitic-textured groundmass of plagioclase felspar laths. The presence of alkali felspar could not be ascertained due to the very fine-grained nature of the groundmass, and for the same reason no plagioclase compositions were determined.

Pyroxene phenocrysts are fairly sparsely distributed and make up between 6 and 43 per cent of the rock and in its unaltered state the pyroxene is pale brown augite. Alteration of the phenocrysts to pseudomorphic aggregates of pale green uralitic amphibole and minor epidote is extensive and only occasionally are the cores unaffected. The alteration has been accompanied by the exsolution of opaque ore granules which appear as part of the alteration aggregate. Twinning or relict twinning of the pyroxene phenocrysts or pseudomorphs is evident. The crystal outline is anhedral to subhedral but has been considerably modified by corrosion, resulting in characteristic rounded forms ranging in size between 0,4 and 3,2 mm.

Plagioclase phenocrysts (e.g. BW 253) are rather erratic and may be absent altogether. They are always subordinate to pyroxene as a phenocryst mineral in both relative amounts present and size, ranging from 0 to 12 per cent by volume and from 0,4 to 4,3 mm respectively. Saussuritisation is invariably complete and no compositions could, therefore, be determined. Crystal form is anhedral to less frequently subhedral, the phenocrysts having suffered the effects of corrosion.

A directional texture in the groundmass varies from barely discernible (e.g. BW 253) to well-pronounced (e.g. BW 252). Minute crystals of uralite appear as flecks throughout the groundmass, probably as a result of alteration of small pyroxene granules. Minute ore grains constitute an important accessory mineral of the groundmass. Amygdales may be present and are composed of quartz, uralite amphibole and minor epidote and carbonate.

(vi) Pyroxene trachybasalt member

Distribution and Field Character

Flows of this member constitute a well-defined unit which occurs only in the vicinity of the Hahnenkamm trigonometrical beacon in the southern part of the area. The total thickness of these basalts is approximately 3150 m, unbroken except for three intercalated flows of 'large-felspar trachyandesite' near the base and two prominent bands of andesite agglomerate/lapillistone about two-thirds of the way up in the member.

The basalt is a fine-grained, usually moderately light-coloured rock, i.e. pinkish-brown to greenish-grey, with small phenocrysts of ferromagnesian minerals. In hand-specimen the rock is devoid of any obvious flow texture and only rarely are small rounded amygdales present. The weathered surface is of a light red-brown colour and appears distinctively pitted due to the weathering-out of ferromagnesian phenocrysts.

This member does not appear in the Barby sequence on Aruab or on Vergenoeg, and thus indicates a fairly limited distribution despite the enormously thick local development. Any possible southward and northward extensions are obscured by the intrusion of Nubib granite and by younger cover.

Petrography

The porphyritic nature of the rock is typical although it is not always very pronounced. Phenocrysts and microphenocrysts of altered olivine and relatively unaltered clinopyroxene are set in an intergranular or intersertal matrix which often displays a well-developed flow fabric visible only in thin-section (Plate 8).

Olivine usually forms a phenocryst phase only, although due to resorption the range in grain size is wide, i.e. < 0,1 to 4,5 mm. The crystals have been completely altered to serpentine, lesser chlorite and pale-green fibrous amphibole with the accompanying exsolution of an opaque ore mineral, giving the pseudomorphs a distinctive appearance. The crystals are early-formed and may be partly enclosed by clinopyroxene. Corrosion features are common and embayments also occur. The crystal form is anhedral to euhedral and has in many cases been somewhat modified by corrosion. A glomeroporphyritic habit is not as well developed as in the case of the clinopyroxene. Olivine makes up less than 4 per cent of the total rock volume and is subordinate in amount to clinopyroxene.

Clinopyroxene phenocrysts constitute 1 to 12 per cent of the rock volume and vary greatly in size from about 0,2 to 3,4 mm. The mineral is augite which is neutral to very pale-green in colour. Compositional zoning is present but is very poorly developed. The state of alteration varies but is rarely complete the alteration products being chlorite, pale-green fibrous amphibole, epidote and calcite. The crystal form is subhedral to less frequent anhedral and euhedral. Glomeroporphyritic aggregates are common, frequently consisting of large numbers of small individual crystals. No corrosion effects were noted.

The groundmass, which constitutes 84 - 99 per cent of the rock, consists essentially of small plagioclase felspar laths with a mesostasis of reddish cloudy glass or alkali felspar, and in some cases small intergranular crystals of clinopyroxene. Minute granules of an opaque ore mineral are ubiquitously scattered throughout the groundmass. Extensive alteration of all components, the plagioclase in particular, has resulted in much sericite, epidote, chlorite and calcite in the groundmass replacing most mineral phases. Much of the ore may also be secondary. A limited number of optical determinations indicate a fairly restricted compositional range of An_{32} to An_{35} for the plagioclase felspar. The size of the plagioclase laths varies from microlites to crystals of 0,3 mm in length. In many flows a subparallel alignment of the plagioclase laths has resulted in a pilotaxitic texture (Plate 8) or, when interstitial glass is present, in a hyalopilitic texture. Amygdales, which are frequently of such a small size as to be observable under the microscope only, are composed of quartz, epidote, pale-green fibrous amphibole and calcite.

(vii) Pyroxene trachyandesite member

Distribution and Field Character

Exposures of this very conspicuous member are confined to the area adjacent to the Ganaams-Aubures boundary and the southern Aruab-Wittmanshaar area. Both localities must once have been part of the same volcanic field but are now separated by a 3-6 km wide cover of down-faulted Auborus strata.

This member is composed predominantly of bright red-brown flows of porphyritic trachyandesite with irregularly interspersed beds of tuff and lapilli-tuff, also of a conspicuous red-brown colour. Grey-coloured lava constitutes a relatively minor part of the sequence. In the Ganaams-Aubures area the uppermost 400 m are composed of well-bedded red-brown volcanic sandstone and conglomerates (Fisher, 1961, p.1412), and lithic sandstone. These volcaniclastic beds, which probably represent epiclastic volcanic deposits, apparently constitute the uppermost stratigraphic horizons of the Barby Formation as exposed within the southern part of the area.

The lava is invariably porphyritic, with phenocrysts of ferromagnesian minerals (predominantly green pyroxene) set in an aphanitic groundmass of a characteristic red colour. Only rarely is the groundmass a grey colour. Much of the lava is often very highly amygdaloidal, the original lava having been of a scoriaceous nature. Thin tuff bands are often associated with this highly amygdaloidal lava.

The total thickness of the unit in the Aubures area is 2500 m, of which lava makes up about 2000 m. The flows here dip east-southeastward and individual flows are fairly easily recognised due to their varying resistances to erosion. The more amygdaloidal and, to a lesser degree the more highly porphyritic lavas, weather in a characteristic crumbly fashion, contrasting strongly with the denser and less porphyritic varieties which do not weather as readily. Due to the ready disintegration of the vesicular flows, their outcrop is very often completely obscured by a gravelly cover of scree. The thickness of individual flows varies from a few meters to about 100 m.

On Aruab, lava flows of this member lie directly on felsic flows of the High-Ca rhyolite member; both the Large-felspar trachyandesite and Pyroxene andesite members are absent here. The total thickness of the Pyroxene trachyandesite member is difficult to estimate in this area, but could be as much as 5800 m, and is certainly considerably more than that of the succession on Aubures to the northwest, indicating more intense and prolonged volcanic activity of this type in the vicinity of Aruab.

The presence on Aruab of large-felspar trachyandesite material intrusive into the flows of this member (Section 3.1.2.2) about one-third up in the sequence from the base, further indicates that the emplacement of the lower two-thirds of this member was in all probability contemporaneous with the emplacement of the large-felspar trachyandesite lava flows occurring approximately 15 - 20 km to the northwest (Fig.2). The implication is, therefore, that only the upper 2000 m or so of this member on Aruab-Wittmanshaar are the approximate time equivalents of the pyroxene trachyandesite succession as exposed on Ganaams-Aubures. Beds of stratified tuff, up to 100 m thick, also occur on Aruab, sporadically intercalated with the lavas and constituting a relatively minor part of the succession. Coarser volcaniclastics are rare from this locality.

The flows of this member are undoubtedly the "diabases of a vivid reddish-brown colour" observed by Beetz (1923) and recognised by him as belonging to the upper part of the "Sinclair Series".

Petrography of the lava

Under the microscope the trachyandesites are even more striking than in hand-specimen. Phenocrysts of clinopyroxene are prominent and, together with more erratic olivine and hornblende , they are set in a microcrystalline intergranular or intersertal to pilotaxitic or hyalopilitic groundmass, which is invariably heavily charged with hematite. Relatively small plagioclase phenocrysts are only occasionally present.

Clinopyroxene is pale yellowish-green diopsidic augite which is never altered and makes up between 8 and 21 per cent of the rock. Practically all the crystals are phenocrystic or microphenocrystic, ranging in size from less than 0,1 to 4 mm (Plate 9). The smallest and quantitatively most insignificant crystals have similar dimensions to the felspar crystals in the groundmass and are not truly phenocrysts, but their early-formed character with respect to the felspar is obvious. The crystal form is euhedral to subhedral to less frequent anhedral and the grain boundaries have only very rarely been affected by slight resorption, the crystal faces being nearly always sharply defined. A glomeroporphyritic habit is well-developed and clusters of large numbers of individual crystals measuring up to 3,5 mm in diameter are not uncommon. Twinning,often of polysynthetic type,is common. Some degree of zoning is occasionally displayed and this may have normal or oscillatory patterns (Plate 9). In most cases the zoning is visible only under crossed nicols, but in a few flows (e.g. BW 456 and BW 1558) zonal growth is indicated by alternating brown and yellow-green coloured bands.

Altered olivine is a common phenocryst mineral which constitutes up to 23 per cent of the rock and is only absent when hornblende is present. A mutually exclusive relationship between these two phenocryst minerals exists for the vast majority of flows but is apparently not universal since in one specimen, BW 1575, small resorbed grains of altered olivine appear in aggregates together with large altered hornblende crystals (Plate 10). The olivine crystals show evidence of having had well-formed outlines, and euhedral shapes can still be observed, but on the whole resorption has affected the mineral to a great extent and rounded and embayed forms are most common. Alteration and replacement products are serpentine, chlorite, carbonate and opaque ore minerals, the latter being typically concentrated in irregular cracks and around the margins of the pseudomorphs. The size of the crystals is variable from 0,1 to 5 mm and glomeroporphyritic aggregates may occur.

Amphibole phenocrysts form an important constituent mineral of some of the lava flows in the upper parts of the member, i.e. on eastern Aubures and southern Aruab. The mineral may make up as much as 17 per cent of those lava flows in which it occurs, and is present as corroded subhedral to euhedral crystals measuring up to 2,6 mm in length. Typical prismatic and basal forms may be well displayed, but in the majority of cases the effects of resorption are ubiquitous. Complete replacement by hematite, granular opaque ore and pyroxene (?) has taken place, and the pseudomorphs are probably representative of late magmatically altered and replaced oxyhornblende or basaltic hornblende (Plate 10). Replacement by iron oxides has been so intense that in some instances (e.g. BW 1557) the crystals might easily be mistaken for primary opaque ore mineral phenocrysts, were it not for the characteristic crystal outline and remnant prismatic cleavage.

Rare plagioclase phenocrysts (andesine) measuring up to 2 mm in length may be present (e.g. BW 1623), but never constitute greater than 2 per cent of the total rock volume.

The groundmass, which constitutes 56 - 91 per cent of the total rock volume, is composed of small plagioclase laths, less than 1 mm in length, with a high proportion of interstitial alkali felspar or glass. In many cases the plagioclase crystals are too small and the twinning too indistinct to allow determination of composition by normal optical methods, but those that are large enough indicate a range of An_{32} - An_{38}. Most crystals are slightly and normally zoned. The laths may be randomly oriented but more commonly they display a preferred flow-induced orientation resulting in a pilotoxitic or hyalopilitic texture. The entire groundmass is invariably heavily charged with small grains of opaque iron ore and flecks of hematite. In the grey lavas the hematite is of only minor importance. Interstitial glass and alkali felspar have been severely affected by the hematite staining and may appear nearly opaque. The plagioclase has been affected by incipient saussuritisation. Amygdales are usually irregular in shape and are composed of quartz, calcite, zeolites, chlorite and epidote.

Petrography of the volcaniclastics

Of the true pyroclastics the dominant rock types can be classified as tuff, grading into lapilli-tuff or lapillistone as the grain size increases (grade size categories after Fisher, 1961). Lithic fragments are the dominant constituents and are entirely volcanic in nature, being identical in appearance to the lavas of this member. Crystal fragments are also prominent constituents but are always subordinate to the rock fragments, and all the characteristic minerals of the pyroxene trachyandesites are represented, i.e. diopsidic augite, plagioclase felspar, altered olivine and amphibole, and ore. Calcite and a zeolite mineral form ubiquitous cements. Sorting is generally poor but may become moderate in some of the finer-grained deposits. Grains are invariably angular except for some degree of rounding occasionally developed in the tuffs which are generally well-laminated as a result of alternating coarser and finer bands clearly discernible in thin-section. Lapilli-tuffs display less well-developed bedding and they are often completely unstratified.

The rounding of volcanic fragments and the incorporation of varying amounts of well-rounded normal clastic material, predominantly quartz grains, has occurred mainly, but not exclusively, in the uppermost part of this member, resulting in rocks which can generally be classified as volcanic sandstones and conglomerates (Fisher, 1966). The main difference between the latter rock-types and the aforementioned intercalated pyroclastics is that the volcanic sandstones and conglomerates are almost certainly of epiclastic origin, whereas the intercalated tuffs and lapilli-tuffs are of pyroclastic nature, probably reworked (secondary) in part.

(viii) Basalt member

Distribution and Field Character

The difficulties encountered in assigning meaningful member subdivisions to much of the Barby basic lava sequence in occurrences north of latitude 25°30'S has, to some extent, already been discussed at the beginning of this Section, particularly with respect to the lack of continuous exposure. Furthermore, the basic lavas of this area have suffered such considerable alteration that any subdivision would be based purely on gross textural features, which do not appear to vary as significantly as those of the lavas in the south. The basic lavas of the Barby Formation north of 25°30'S will be regarded as belonging to a single member, referred to as the 'basalt member'. Total maximum development is on Duwisib, southeast of th Duwisib trigonometrical beacon where an approximately 4200 m thick uninterrupted succession of basaltic lavas was encountered. Other prominent outcrop areas are on Osis, Duwisib (west of the Duwisib trigonometrical beacon), Vrede, Nubib West and northeastern Nam. In all but the Osis outcrops, the lavas of this member are associated with clastics, pyroclastics and felsic lavas of the Rhyolite member (Section 3.1.2.1(ix)).

Large-scale intrusion of Nubib granite has disrupted the sequences and the lack of marker horizons frequently places some uncertainty on the exact stratigraphic position of many of the occurrences.

The lavas are of a uniform dark green to greenish-grey colour on fresh surface and texturally they may be divided into the following dominant types:

(a) Non-porphyritic and very fine-grained (e.g. BW 1344)

(b) Porphyritic with large plagioclase felspar phenocrysts measuring up to 2,5 cm in length, set in an aphanitic groundmass (e.g. BW 1136)

(c) Porphyritic with relatively small plagioclase felspar phenocrysts measuring up to 5 mm in length and very rare, smaller pyroxene phenocrysts, fairly sparsely distributed throughout an aphanitic groundmass (e.g. BW 549).

A few thin and discontinuous agglomerate horizons appear intercalated with the lava on Vrede and Osis. These consist predominantly of highly angular basic lava fragments and fewer slightly rounded fragments of light-coloured felsic volcanic material, ranging in grain size from less than 0,5 mm to 10 cm. The Osis agglomerates do not have the felsic fragment component and are entirely basic in composition. There is a complete lack of sorting and bedding in both occurrences.

Petrography of the basalts

In thin-section it is apparent that the greenish colour of the rock, as seen in hand-specimen, is due to extensive alteration to saussurite, uralitic amphibole, chlorite, epidote, carbonate, opaque ore minerals and leucoxene. Original textures may be fairly well preserved despite the altered nature.

Plagioclase felspar phenocrysts are often completely decomposed to pseudomorphic aggregates of saussurite, but occasionally relatively well-preserved crystals or portions of crystals are encountered whose compositions vary considerably from An_{37} to An_{64}, and which are generally more calcic than the plagioclases in lavas of the southern occurrences. Zoning is fairly common and of normal type. The crystal form is anhedral to euhedral, the outlines occasionally displaying signs of resorption and/or modification by alteration.

Ferromagnesian mineral phenocrysts have, in most cases, been completely altered to uralitic amphibole and granular ore with less common epidote and chlorite. Rare unaltered patchy cores indicate pale brown clinopyroxene, probably augite.

The nature of the groundmass depends on the state of alteration, and varies from moderately altered with plagioclase laths of random or slight preferred orientation with intergranular uralite, chlorite, epidote and opaque ore minerals, to completely decomposed granular and fibrous aggregates of sericite, epidote, chlorite, uralite, opaque ore minerals, carbonate and quartz. Primary ore is present in accessory amounts in all of the lavas and may be altered to leucoxene.

Amygdales are usually small, up to about 0,8 cm in diameter, are rounded and composed of quartz, chalcedony, epidote, carbonate, uralite, opaque ore minerals and hematite.

Petrography of the agglomerates

These very poorly-sorted pyroclastics are made up of angular fragments of highly altered basic lava as described above, which, in some instances (e.g. Vrede occurrence), occur together with slightly rounded fragments of devitrified felsic material set in a dark matrix of basic lava fragments, many of which are highly amygdaloidal. The felsic fragments have very jagged outlines and must therefore have originated from an explosive shattering of solid material. Most of the finer material is, however, very highly decomposed and replaced by chlorite, epidote, opaque ore minerals and calcite so that discrete fragments are not always readily discernible.

(ix) Rhyolite member

Distribution and Field Character

This well-defined member is confined to areas north of latitude 25°20'S and consists essentially of felsic lava, various volcaniclastic and normal clastic deposits, and basic lava. The maximum development occurs in the Nubib Mountain area, where a thickness of about 3700 m is attained. The succession here dips eastward at angles of 30° to 40° and has been isolated by extensive invasion of Nubib granite, which locally builds up mountains of towering proportions. Outcrops of the Rhyolite member appear on the western flank of

Nubib Mountain, on the eastern flanks of Losberg and in the intervening valley. The fact that none of these beds are present in the extensive outcrops of Barby rocks south of latitude 25°30'S is suggestive of a southward wedging-out of the unit, and this can be observed on Duwisib where beds representative of the member wedge out completely in a southeasterly direction. The sequence in the Nubib mountains (Fig. 4), being unbroken and having the greatest development, is regarded as the 'type' locality and is described below.

By far the greater part of the sequence, up to 2200 m, is made up of dark blue-black to rarer maroon-coloured rhyolitic extrusives, i.e. lava flows and lesser amounts of ash-flow tuff.

Rhyolite lava constitutes flows of limited lateral extent and abrupt distal terminations. Flow-banding is very prominent and frequently highly contorted, being particularly evident on the typical yellowish-brown weathered rock surface (Plate 11). The flow-banding is often, but not always, defined by light coloured quartz-rich laminae which may be as much as 2 cm thick. An original vesicular texture, while not abundant, may be present in flows in close stratigraphic association with volcaniclastic horizons and is evidenced by large (up to 4 cm) rounded amygdaloidal 'clots' of quartz or cavities partly filled with quartz, generally closely associated with the thick flow laminae mentioned above. The lava is frequently felspar-porphyritic with less common quartz phenocrysts in a felsitic groundmass, but phenocrysts are always small and never present in abundance, often being completely absent. A conchoidal fracture is typical and reflects the extremely fine grain of the rocks. Lithophysal structures were noted in only a few of the thinner flows, the spherical forms having a maximum diameter of about 2,5 cm.

Rhyolitic ash-flow tuff is subordinate to the felsic lava and forms relatively thin deposits (about 10 - 100 m), ranging from highly lithic to highly vitric varieties. The former contain a very large proportion of angular rock fragments measuring up to 4 cm and set in a dark-coloured matrix. Compositionally, the foreign fragments are predominantly of dark felsic volcanic and less abundant basic volcanic and clastic material. Pumice fragments were not noted.

On fresh surface the rock may have a uniform black colour which often makes recognition of its fragmental nature difficult. On weathered surface, however, the fragmental nature becomes obvious by virtue of the fact that the matrix is yellowish-brown in colour whilst the fragments tend to remain dark. Compaction banding, frequently wrapping about lithic fragments, may also become more noticeable on weathered surfaces. There is a complete lack of sorting and bedding in these pyroclastic units.

The vitric tuffs contain only very few or no rock fragments, and are very similar in appearance to the rhyolite lavas. Compaction banding is present and is aligned fairly consistently parallel to the base of the unit as compared with the highly contorted flow-banding of some of the lavas. In hand-specimen, and in the absence of rock fragments, it is virtually impossible to distinguish between lava possessing a planar flow-banding and a highly vitric, thoroughly welded and compacted ash-flow tuff and the complete lack of pumice fragments in the ash-flow tuff of this member magnifies this problem. On the other hand it is also not always possible in hand-specimen and/or limited outcrop to distinguish between ash-flow tuff beds and primary (Fisher, 1961,

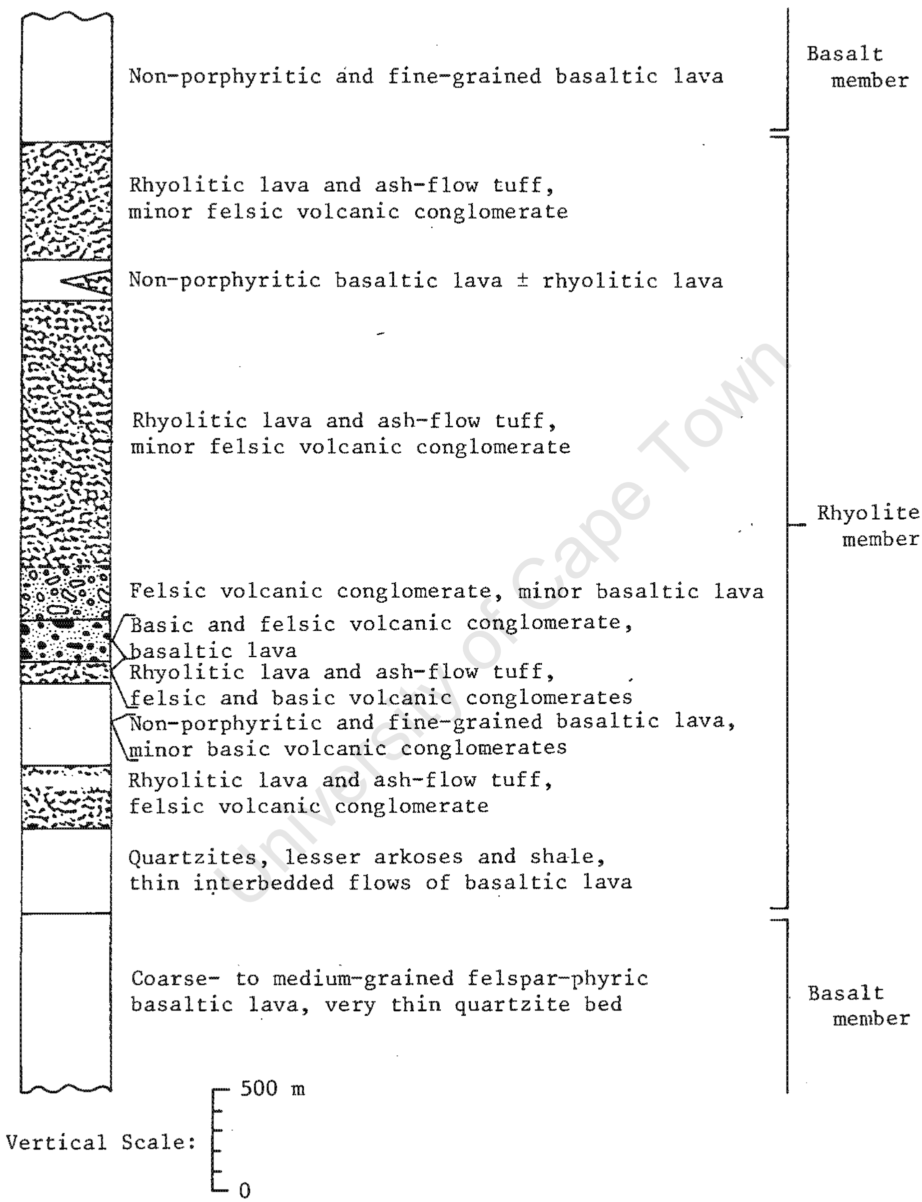


Fig. 4. Stratigraphic succession of the Barby Formation in the Nubib Mountains

p.1412) pyroclastic deposits of non-flow origin. These aspects will, however, be discussed in more detail in the following section on petrography since in thin-section a correct identification can generally be made.

An estimated thickness of 500 m of volcanic conglomerate occurs mainly within the lower half of the succession. Individual beds may be up to 50 m thick but are generally less than half this thickness. These coarse-grained volcaniclastics are composed almost entirely of felsic and basic volcanic fragments. Boulders of granite and quartzitic material often constitute a small component of these deposits but are conspicuously rare. The majority of grains or fragments are angular,but subrounded to well-rounded examples are also common. As a general rule a higher proportion of the larger fragments, in particular those of basic volcanic composition, display some degree of rounding as compared with the finer material. Sorting is extremely poor and grain or clast sizes vary from less than 1,0 mm to a maximum of about 0,7 m in some cases. More commonly, however, the maximum boulder dimension of these deposits is in the order of 15 cm. Bedding is never well developed but occasionally individual units may display a crude grading from very coarse at the base to relatively fine towards the top.

The ratio of basic to felsic fragments or boulders in the conglomerates varies considerably between two virtually pure end members, but usually deposits closely associated with felsic extrusives contain an overwhelmingly large proportion of felsic material whereas those associated with basic lava units, although containing dominant amounts of basic fragments and boulders, also include appreciable proportions of felsic material.

The non-genetic term 'volcanic conglomerate' (Fisher, 1961, p. 1412) has been applied to these rocks since the nature of the deposits does not allow reliable determination of the fragmentation process giving rise to the volcanic rock debris. If the rocks are of a purely pyroclastic nature (agglomerates) then reworking has taken place to produce the rounding observed in many of the fragments and blocks, but if purely epiclastic (epiclastic volcanic conglomerate) the transport could not have occurred over any great distance. Either of these genetic derivations is not completely compatible with the observed ubiquitous mixing of highly angular and well-rounded particles of all size grades, and it is therefore probable that the deposits represent mixtures of both epiclastic and reworked pyroclastic material. Furthermore, the inclusion of pyroclastic material of a primary nature (Fisher, 1961, p.1412) is also considered most likely.

Basic lava flows comprise about 600-700 m of the total sequence and usually form fairly distinct units. In one part of the succession, however, exposed 8 km northwest of the Eckberg trigonometrical beacon, the flows alternate with, and are intimately associated with, felsic extrusives and volcaniclastics as described above. The basic lava is identical to that of the Basalt member,undoubtedly belonging to the same basic extrusive phase, and should therefore be excluded from the 'Rhyolite' member. For convenience of subdivision and description, however, it has been included in the latter. Appreciable volumes of basic lava of the Basalt member underlie and overlie the Rhyolite member with an exposed thickness of 900 m and 600 m respectively. The lowermost 400 m of the Rhyolite member consist of well-bedded quartzites with thin intercalations of basic lava.

On the farm Duwisib, 45 km southeast of the Nubib Mountain sequence, the Rhyolite member is represented by a wedge of clastic rocks with two intercalated rhyolite flows, 150 m and 80 m thick respectively, and a single 30 m thick basic lava flow. The clastics are predominantly quartzites and lithic sandstones with two bands of conglomerate, one at the base and another near the upper limit of the unit. The conglomerate is probably an epiclastic volcanic rock consisting almost entirely of light-coloured and well-rounded rhyolite pebbles up to about 10 cm in diameter and set in a volcanic sandstone-type matrix. This part of the succession in the Duwisib area has a maximum exposed thickness of 1600 m, wedging out completely over a distance of 5 km. The wedge is stratigraphically associated with basic lava of the Basalt member. The rocks in this area display considerable effects of shearing related to cataclastic deformation in the Nam Shear Belt (Section 3.1.5).

Intrusion of Nubib granite has almost, but not entirely, obliterated outcrops of the Rhyolite member in the area between the Duwisib and Nubib Mountain occurrences. On the farm Betta, alternating basic and felsic flows, some of recognisable ash-flow tuff nature, constitute an isolated fault-block, and are regarded as belonging to the Rhyolite member. The total thickness is in the order of 1100 m.

On Gorab, isolated outcrops of quartzite with a minimum thickness of 250 m are probable correlates of the basal quartzites of the Rhyolite member as exposed in the Nubib Mountains, and on Duwisib.

In the northeastern portion of Nam, 700 m of conglomerate and sandstone with a single 20 m thick intercalation of basic lava are exposed in a narrow outcrop isolated by younger cover and intrusion of Nubib granite. These rocks are underlain by 1200 m of basic lava of the Basalt member. The conglomerate is identical to that occurring in the Duwisib sequence, consisting almost entirely of well-rounded rhyolite pebbles set in a volcanic sandstone matrix and is interpreted as an epiclastic volcanic conglomerate. The sandstones are lithic types and are also rich in volcanic fragments.

Recent work by Schalk (pers. comm.) in the area north of latitude 25°00'S has revealed the presence of great thicknesses of the rhyolite pebble conglomerate and rhyolitic lava of this member at the northern limit of the Nubib Mountain range. The indications are, therefore, that this particular member continues to thicken northward and may even become predominant in the Barby Formation as a whole.

Petrography of the rhyolite lava

A porphyritic texture is common but not characteristic of these lavas, the presence of phenocrysts being rather erratic and certainly never abundant. Phenocryst minerals, which in total amount never constitute more than 8 per cent of the rock, are plagioclase felspar, minor quartz and infrequent alkali felspar. Non-porphyritic lava types are not uncommon.

Plagioclase felspar is the dominant phenocryst mineral and may make up a maximum of 8 per cent of the rock. Its composition is in the range An_0 - An_{11}. The crystal form is most commonly subhedral to anhedral and may be considerably rounded, embayed and generally modified by corrosion. Twinning is according to the albite, pericline and carlsbad laws. Alteration to saussurite, although never extensive, is always present. In a few specimens (e.g. BW 759) rare secondary albitisation has resulted in relatively clear patches of chess-board albite being developed within the cloudy and incipiently-altered phenocrysts. The crystals are always small, ranging in size from 0,1 to 4,0 mm, and even the ubiquitous glomeroporphyritic aggregates never attain dimensions greater than 5,0 mm.

Quartz phenocrysts may constitute up to 5 per cent of the rock volume but are often completely absent. The grains have highly corroded forms, and are always well rounded and embayed. Only rarely are original crystal forms or faces preserved. The size range is from less than 0,1 to 2,3 mm.

The presence of original alkali felspar, phenocrysts, now highly corroded and albitised, is suspected. The mineral has been replaced by chess-board albite, resulting in the uncertainty in recognition.

Opaque ore mineral grains, well-rounded and never larger than 0,5 mm, constitute rare microphenocrysts.

The groundmass consists almost entirely of cryptocrystalline quartzo-felspathic intergrowths resulting from the devitrification of an original glass. Recrystallisation has further resulted in relatively coarse, granular, quartz-rich patches (e.g. BW 852).

Devitrification products vary considerably in texture

a) Granular, which may rarely be superimposed on a perlitic texture.

b) Small and mostly vague radiating and plumose structures which usually constitute the entire groundmass of those rocks in which they appear.

c) Large radiating fibrous structures - spherulites - up to 6 mm in diameter (Plates 12 and 13). These do not often constitute the entire groundmass but usually occur in combination with (a), the granular products filling the spaces between spherulites. Spherulite growth may or may not proceed from a phenocryst nucleus and usually cuts any flow structure that may be present.

d) Extremely fine fibrous growths, the fibres all being aligned approximately parallel to the flow-banding and not strictly radiating.

e) Axiolitic growths developed parallel to and in part defining the flow structure.

f) Massive, patch, non-flow oriented poikilitic growths.

Flow-banding in these rocks is often not as obvious in thin-section as it is in hand-specimen, particularly on weathered surfaces. The flow structure is invariably defined by varying grain sizes of crystalline products and/or varying concentrations of opaque ore mineral granules within alternating bands. In thin-section these bands are often fairly linear, but may less frequently be highly contorted. Of lesser importance in defining a flow-banding are, coarse quartz-rich lenticles which are probably the result of a vapour-phase crystallisation, axiolitic structures mentioned above, lenticular glomeroporphyritic clusters, and preferred orientation of felspar phenocrysts.

The presence of dust-fine opaque ore minerals in the groundmass is ubiquitous and trichites may occur in association with large spherulites (Plate 13).

Ferromagnesian minerals are represented by blue-green amphibole which may occur as small ragged crystals or scattered granules and granule aggregates. To what extent this represents a primary ferromagnesian mineral is not certain, but in most cases some redistribution has certainly taken place. Zircon is an important accessory mineral, forming rare but often quite large (up to 0,2 mm) subhedral to euhedral crystals.

Petrography of the ash-flow tuffs

These rocks range from highly lithic to highly vitric varieties, and the vast majority of flows correspond to either one or the other of these two extremes. The descriptions will therefore be confined to these two end member types.

(a) Vitric ash-flow tuff of the Nubib mountain sequence is virtually indistinguishable under the microscope from much of the rhyolite lava described in the preceding pages. Extreme compaction, welding and complete devitrification and partial crystallisation has all but eliminated the original shard structure. The rock is composed of more than 94 per cent devitrification and recrystallisation products, thus giving evidence of the highly vitric nature of the original ash. The remaining 6 per cent or less is made up mainly of phenocrysts and angular crystal fragments of plagioclase felspar, quartz, alkali felspar and rare angular rock fragments. The latter are nearly always composed of devitrified felsic volcanic material, differing very little in appearance from the cryptocrystalline matrix. In the absence of recognisable shard textures it is the occasional presence of fragmented crystal and lithic material that provides clues as to the pyroclastic nature of the rock.

Frequently, examination in thin-section will also not reveal the presence of lithic fragments or obviously fragmented crystals and it is only in hand-specimen or in outcrop that the presence of foreign rock fragments can be recognised. The size of crystals and crystal fragments ranges from less than 1,0 mm to 2,0 mm and that of the lithic fragments up to 2 cm. Opaque ore minerals may constitute rare microphenocrysts (< 0,8 mm in diameter).

Compaction banding has developed, moulded about phenocrysts and rock fragments, and closely resembles some of the flow-banding seen in the lavas. Evidence of shard structure is only poorly and erratically preserved as highly flattened Y- and U- shapes, and the resulting lense-like forms are of a discontinuous nature as opposed to the more continuous development of banding resulting from flow processes in a magma.

Under crossed-nicols the devitrification growths completely mask any primary vitroclastic textures. Devitrification textures are very similar to those developed in the lavas, i.e. predominantly granular and fibrous non-spherulitic, spherulitic and patchy poikilitic growths. Relatively coarse-grained quartz-rich lenticular aggregates are interpreted as vapour-phase crystallisation products. Secondary recrystallisation and redistribution of quartz is a common phenomenon and takes place in an irregular patchy fashion or in a 'net-veining' pattern.

Pumice fragments were not observed in these welded tuffs, but the extreme devitrification which is presumed to have followed thorough welding, is likely to have obscured their presence (Ross and Smith, 1961, p. 18).

Vitric welded tuff units occurring in the outcrops on Betta (e.g. BW 598), were apparently thinner than those on Nubib and have probably not undergone the same degree of welding and distortion of shards since the vitroclastic texture in these rocks is well preserved. Compositionally the flows are identical to those described above, i.e. highly vitric with very low crystal and lithic fragment content, but the devitrification, although complete, has not entirely obliterated the shard structure. Moderate distortion of shards has taken place as evidenced by a faint 'pseudo-flow texture' which is particularly noticeable where shards have been bent and stretched about phenocrysts and foreign rock inclusions.

Accessory amounts of opaque ore dust are distributed ubiquitously throughout the rock and rare scattered aggregates of chlorite indicate the possible presence of altered ferromagnesian phenocrysts. Zircon is also a common accessory mineral.

(b) Lithic ash-flow tuff. Characteristically these rocks contain a high proportion of rock fragments, up to 48 per cent by volume, and the crystal content is also appreciably higher (up to 12 per cent) than that of the vitric tuffs. The mineralogy of the phenocrysts is essentially the same as in the vitric tuffs but the crystals display a less resorbed and more highly fragmented character. Highly angular lithic fragments, measuring up to 4 cm in diameter, are composed predominantly of devitrified felsic volcanic material, minor basic volcanic and rare clastic material. The basic volcanic fragments are often highly vesicular and have jagged, completely unabraded outlines.

The 'matrix' is composed predominantly of devitrified glass with lesser amounts of small lithic and crystal fragments, 'volcanic dust' and pumice fragments. Shard structure is barely discernable. Welding and distortion of the shards was apparently fairly severe, resulting in a pseudo-flow-banded matrix, the lines of which wrap around the crystals and inclusions. As with the vitric tuffs, however, the 'pseudo-flow texture' is defined by discontinuous lenses (stretched and flattened shard units) rather than the continuous flow lines typical of lavas. Complete devitrification of the glass fraction has further obscured the vitroclastic texture, resulting in an almost featureless granular mass.

The presence of highly compressed pumice fragments was positively identified in only a few of these tuffs. The overall devitrification has rendered the outlines of the fragments very vague but the typical flattened shapes with ragged ends are still recognisable. A faint fibrous structure perpendicular to the direction of compaction is notable and probably reflects compressed and laterally extended vesicles of the original undeformed pumice. The pumice also has a slightly higher concentration of finely distributed opaque ore-dust than the matrix, a property which facilitates its recognition. Pumice fragments probably constitute less than 1 per cent of the total rock volume.

(c) Volcanic conglomerate. Due to the very coarse grain size of the components making up these rocks, microscopic examination does not reveal much more than observation made in hand-specimen. The minimum grain size is about 0,1 mm, and dust-fine material is conspicuously absent. No preferred orientation of elongated or flat grains/fragments is evident and sorting remains extremely poor when viewed under the microscope. Grain shapes vary from predominantly angular to subrounded and well-rounded examples. Felsic and basic rock material is dominant, the relative proportions of each varying considerably from bed to bed, and clastic rock material invariably constitutes a minor proportion of the grains/fragments.

3.1.2.2 Intrusive Units

(i) Gabbros and Norites

In the southwestern part of the area, on the farms Kumbis, Sonop, Saffier, 166 and Aandster, numerous bodies of gabbro, norite and related rocks are present (Fig. 5). The inclusion of these rocks into the Barby Formation is based mainly on the field relationships between the intrusives and other rock-units. These relationships may be listed as follows :

a) The gabbros and norites post-date the Kumbis and Kunjas Formation by intrusion.

b) Red granite and quartz-porphyry dykes, belonging to the Nubib intrusive phase and the Guperas Formation respectively, both cut the basic intrusives.

c) Monzonitic and dioritic intrusives, found in close proximity to the Spes Bona syenite body, post-date the gabbroic/noritic intrusives. On the basis of similar field relationships, it is apparent that the emplacement of both sets of basic intrusives was closely spaced in time and this fact, together with a similar restriction in distribution mainly to the southwestern part of the area (Fig. 5), might suggest that the two sets of intrusives represent a single intrusive phase. However, definite textural and chemical (Section 3.4) differences exist between the two and on this basis as well as the obvious close areal association of the monzonites and diorites with the Spes Bona syenite, the two sets of intrusives are described separately.

The gabbro/norite outcrops appear most frequently as black, steep-sided hills sticking up through the desert plain and contact areas are, therefore, not well exposed. Within individual outcrops, contacts between successive phases are also not readily apparent due to the weathering characteristics of the rocks which tend to produce a thick covering of rubble on the exposures.

A pronounced igneous lamination, the result of crystal settling, is apparent in nearly all outcrops and a layered character, with bands relatively richer or poorer in mafic minerals, has resulted from the same process of crystal settling.

The largest single body occurs at the boundary between the farm Kumbis and State Land, and although sand-cover obscures most of the contacts, the shape of the body is regarded as being roughly circular with a diameter of about 5 km. Numerous phases of intrusion are evident:

a) A main layered phase consisting of very well- to poorly-laminated gabbros and norites. Thin anorthositic bands are also present as a result of crystal settling and rarely these may form fairly large 'pods' of up to 1 metre thick, which are typically spotted with chloritic patches. The igneous lamination dips radially toward the presumed centre of this body, at angles of up to 62^0, the angle of dip decreasing inward.

b) Intruding rocks of the main layered phase are veins of a gabbro pegmatite which is extremely coarse-grained (up to 4 cm) and is probably a late stage differentiate of the main layered phase.

c) Cutting both the above, is a picrite sill which is homogeneous and displays no sign of crystal settling. The maximum thickness of the sill is about 12 m.

Rock-types of all the above phases have been invaded by small veins of a fine-grained leucocratic granite which probably represents remobilisation products of the surrounding basement granite country-rock. Numerous very fine-grained basic dykes, of width less than 1 metre, cut the gabbros and norites of this body and it is not certain whether they represent a late and final phase of the intrusion or not since they could well be related to the Guperas phase of basic dyke injection (Section 3.1.6.3).

A small body of well-laminated gabbro occurs on the farm Kameelhof intruding pre-Sinclair basement rocks and is correlated with the Barby basic intrusives described above.

(ii) Monzonite

Intruding the Barby basic lavas on Aruab are small bodies of plagioclase-porphyritic monzonite which is indistinguishable in hand-specimen from the highly characteristic lavas of the 'Large-felspar trachyandesite member' described earlier in this Section. Only the obvious discordant nature of the bodies and lack of amygdales distinguish this rock from its extrusive equivalent. Two narrow ($\sim$ 5 m) dykes cutting rocks of the Kumbis Formation on Spes Bona and Kameelhof have a plagioclase-porphyritic monzonitic character and are included in the Barby Formation.

Petrography

Main layered phase

In thin-section it is apparent that a complete range of rock-types exist from gabbros on the one hand to norites on the other. Furthermore, these rocks range from olivine- to quartz-bearing and there is also a tendency toward anorthositic types as the ferromagnesian mineral content decreases.

Plagioclase felspar makes up between 35 and 98 per cent of these rocks and has the form of well-twinned, subhedral to euhedral laths with an average length of 1,5 mm and a maximum of 4 mm. Composition is variable but always falls within the labradorite range. Zoning of the normal type is common and compositional differences betwen crystal cores and margins may be up to An_{12}. The plagioclase is frequently a cumulus mineral and crystals commonly display a pronounced preferred orientation due to settling and this is particularly well developed in the anorthositic bands. With the exception of the more highly uralitised examples, saussuritisation is generally not well advanced.

Clinopyroxene is augite which constitutes 0 to 35 per cent of these rocks. It occurs most commonly as intergranular or intercumulus, anhedral to subhedral crystals of average size 1 mm. Large (7 - 8 mm maximum) poikilitic plates may also occur but are much less common. Exsolution lamellae of orthopyroxene and schiller structure are common.

Orthopyroxene may make up as much as 50 per cent of these rocks and is brilliantly pleochroic hypersthene. Crystals have similar dimensions and habit to the clinopyroxene and exsolution lamellae of clinopyroxene are characteristic.

Both pyroxenes are frequently partly or completely altered to a pale green fibrous uralitic amphibole, probably as a result of a late-magmatic process. Usually the amphibole retains the outlines of the pyroxene crystal that it is replacing but this is not always the case. Those rocks that display extensive uralitisation almost invariably contain interstitial quartz (up to 5 per cent).

Olivine may be present (up to 14 per cent) as anhedral to subhedral, early-formed crystals measuring up to 2 mm in size. Alteration of this mineral is never extensive.

Biotite (< 4 per cent) and opaque are minerals (1 - 3 per cent) constitute important accessory minerals although the former is not ubiquitous. When present, biotite usually occurs in close association with the opaque mineral.

Texturally, these rocks range from intergranular with no evidence of crystal settling to cumulus varieties with a well-laminated character, displaying primary precipitate and interprecipitate (adcumulus and intercumulus) growth.

Gabbro pegmatite

Typically, these rocks are very coarse-grained with highly saussuritised plagioclase felspar laths (up to 4 cm in length) making up about 55 per cent by volume, and intergranular growths of pale green fibrous amphibole after pyroxene. Where it could be determined, the plagioclase composition was in the labradorite range.

Picrite

Olivine is the dominant mineral constituent of this rock, making up between always small (up to 1 m) and of dioritic or gabbroic composition.

Relationships between the syenite and other rock units are as follows:

(i) It post-dates the amphibolites and quartzites of the Kumbis Formation as shown by the intrusion at the Excelsior-Kameelhof boundary.

(ii) It intrudes, with sharp contact, lavas of the Barby Formation on Sonop and Spes Bona.

(iii) In the field, a definite spatial relationship is apparent between the Spes Bona syenite, and monzonitic and dioritic intrusives to be discussed below. The syenite, the inferred immediate subsurface and outcrop extent of which has been outlined in Figure 5, forms the youngest unit and is easily recognisable as a single fairly homogeneous body.

The more mafic 'monzonite' rock-types (including diorites) occur almost exclusively in a zone marginal to the syenite. In outcrop these monzonites are also closely associated with the gabbros and norites described in Section 3.1.2.2 but generally, in areas adjacent to the syenite, monzonite is the dominant rock-type and is certainly not located at any great distance from it,as the gabbros might be.

Outcrops of monzonite have the same characteristics and are identical in gross appearance to those of gabbroic constitution. Steep, debris-strewn and frequently conical-shaped hills of a black and streaky appearance, rising from the Namib desert plain, are typical. The rocks are dark- to medium-grey coloured in hand-specimen and the textural and other petrographic characteristics are described later in this Section.

Contact relationships between the monzonite and the syenite and gabbro are not clear due to lack of good exposure in critical areas, but there are indications on the farm Saffier that the syenite intrudes the monzonite and that the gabbro is cut by the monzonite.

The gabbro, as mentioned previously (Section 3.1.2.2) occurs mainly to the southwest of the syenite with a particularly large occurrence on Kumbis and the adjoining State Land. Minor occurrences appear adjacent to the main mass of syenite and also to the marginal monzonite outcrops on Excelsior, Dina and Aandster.

On the accompanying geological map:

(a) No distinction has been made between monzonite and gabbro, but reference to Figure 5 indicates the presence of the monzonite within outcrops denoted as 'gabbro'.

(b) The 'syenite' outcrops on the farm Dina and in the extreme northern and northeastern parts of Excelsior in fact consist of coarse-grained diorite and monzonite (Fig. 5) and are, therefore, incorrectly represented on the map. These minor misrepresentations are the result of the completion of the geological map prior to detailed microscopic examination of all rock specimens.

(iv) Granophyric granite and aplite, relating to the Nubib granite phase of intrusion, post-date the syenite by intrusion, notably on Saffier. Intrusions tend to be small, irregular in shape and, in the case of the aplite, vein-like in form. Thin chill margins within the granite at the contacts are not uncommon.

(v) Quartz-felspar porphyry and variously textured basic dykes relating to the Guperas Formation cut the syenite in most outcrops.

Petrography of the syenite

The Spes Bona syenite is a very distinctive, coarse-grained to very coarse-grained grey-coloured rock displaying hypidiomorphic granular texture. The uniformity of the syenite over most of its outcrop area has already been mentioned, and is further evident in thin-section by negligible variations in texture and modal composition. Essential mineral constituents are alkali felspar, plagioclase and the ferromagnesian minerals clinopyroxene and biotite.

The alkali felspar constitutes 85 per cent of the rock and occurs as large subhedral and anhedral crystals with a maximum size 5 cm and an average of about 2 cm. Exsolution of the soda component has resulted in a very sparse perthitic character that is classed as 'bead' type (Deer, Howie & Zussman, 1966, p.313). The exsolution blebs are generally fairly large and display typical polysynthetic twinning.

Twinning of the large alkali felspar crystals is almost universal and is according to the carlsbad law. An extremely fine and often barely discernible crypto-cross-hatching is present.

Plagioclase felspar, other than the much smaller exsolution blebs, is present as relatively small (up to 5 mm) subhedral to euhedral crystals, usually completely saussuritised and making up 9 per cent of the total rock volume. Where determinable, the composition is in the andesine range. Frequently the crystals have bent twin lamellae and a wavy extinction. Inclusions of this mineral may also occur in the K-felspar, but these are distinct from the exsolved component.

Ferromagnesian constituents are almost entirely clinopyroxene (4 per cent) and biotite (2 per cent). Very minor amphibole was observed in one specimen (i.e. BW 1246) where it appears to be replacing clinopyroxene.

The clinopyroxene has the form of pale greenish-brown unzoned subhedra (and occasional anhedra), frequently twinned and possessing a coarse cleavage. Schiller structure due to needle-like inclusions of opaque ore minerals is very common. The mineral is probably of diopsidic-augite composition. Alteration of the pyroxene is not extensive but in some sections chlorite and a fibrous pale green tremolitic product is notable.

Biotite occurs as anhedral to subhedral, ragged, prismatic shapes or irregular basal sections and is clearly associated with the opaque ore minerals, frequently enveloping them. It is generally of an intergranular habit with respect to the felspars, and often includes earlier-formed pyroxene crystals. Biotite of a different habit also occurs associated with the opaque ore minerals, and consists of smaller and usually pale-coloured crystals radiating from, and surrounding, the ore grains in the form of corona structures. Stringers of epidote (as seen in two dimensions) are frequently present in the biotite, aligned parallel to the cleavage traces. Opaque ore minerals may be as abundant as the biotite whereas apatite, although fairly common, is an accessory. The ferromagnesian and accessory minerals typically form aggregates.

In a few sections completely serpentinised pseudomorphs of very small amounts of what was probably olivine are present, and the fact that the syenite is undersaturated in the norm substantiates this observation.

Petrography of the monzonites and diorites

Essentially, these rocks are composed of plagioclase felspar, variable proportions of alkali felspar, clinopyroxene, biotite and olivine, and two different textures are notable:

(i) Coarse-grained hypidiomorphic textures occurring in outcrops along the northeastern, northern and western margin of the syenite. These textures are dominant in the dioritic rock-types.

(ii) Medium-grained porphyritic textures appearing in the outcrops bordering the syenite body on its southwestern side. Phenocrysts are plagioclase and very rarely clinopyroxene. Groundmass textures are typically medium-grained hypidiomorphic granular to intergranular (and poikilitic in the case of BW 1256). These textures are dominant in the monzonitic rock-types.

Plagioclase is present to the extent of 35 to 65 per cent by volume and is typically, but not elaborately, zoned in a normal fashion. In the larger crystals, and particularly in the phenocrysts (up to 1,5 cm in length) the zoning is confined to the borders (of the crystals) and is once again normal. The smaller grains (average of $\sim$1,0 mm) are generally zoned throughout. Zoning in the hypidiomorphic granular varieties (e.g. BW 1606) is not as well developed as in the porphyries (e.g. BW 1256 and BW 1613). The average size of the plagioclase crystals in the former is 0,4 cm with a maximum of 1,0 cm.

As a result of the zoning, an average composition for the plagioclase felspar is not easily obtained. Core and margin composition varies considerably, i.e. An_{50} and An_{43} respectively, with an average of An_{47}. A notably persistent characteristic of these rocks is the frequent presence of bent plagioclase crystals. Often the bending has been of such an intensity as to fracture the grain. Similar features can be observed in the syenite.

The form of plagioclase crystals is usually subhedral to euhedral and well-formed laths frequently merge into anhedral plates of intergranular or poikilitic alkali felspar. Alteration of the plagioclase is not extensive.

Alkali felspar is often a relatively minor but essential component (5 to 35 per cent) and is always interstitial and anhedral. Optical properties are identical to those of the alkali felspar forming the bulk of the syenite, with the exception that in the case of the monzonites and diorites the mildly perthitic character is not present. Distinctive properties are the barely discernible very fine cross-hatching, and a faintly brown cloudy character.

Clinopyroxene, the major ferromagnesian mineral (22 to 36 per cent by volume), is a pale greenish-brown diopsidic augite. Only rarely euhedral, the crystal habit is usually subhedral to anhedral. Schiller structure due to minute needle-like inclusions is common, and the pyroxenes are characteristically flecked with small inclusions of biotite.

Minor amounts of orthopyroxene are present in two forms:

(i) Discrete crystals with no marked pleochroism.

(ii) Thin and usually incomplete reaction rims on olivine. This occurrence has the typical pale pink to green pleochroism of hypersthene.

The pyroxene may display very slight alteration to chlorite.

Biotite generally occurs as fairly large interstitial anhedral plates (BW 1606 in particular) and is usually, but not always, associated with the opaque ore minerals, moulded onto them as late reaction products. Fairly fine-grained radiating corona aggregates are also common in some specimens. This mineral makes up between 3 and 8 per cent of the rock volume.

Olivine is an important constituent (to the extent of 4 to 11 per cent by volume), appearing mainly as early-formed anhedral to subhedral grains and displaying a slight amount of alteration to serpentine and to opaque ore minerals. Associated with some of the olivine is the orthopyroxene already mentioned. The crystals have maximum size of 0,5 cm.

Accessory minerals are opaque ore, apatite and minor secondary serpentine and chlorite as alteration products of olivine and pyroxene

3.1.4 NUBIB GRANITE

Distribution and Field Character

Considerable volumes of granitic material, relating to what is considered to be a single magmatic event, occur in the area north of Sinclair. The occurrences have been collectively called the Nubib granite - the term 'granite' being applied in a broad sense to include many different textural types. Compositionally, the granite is remarkably uniform (Section 3.3.3) regardless of texture, and hence it is felt that the general usage of the term is justified.

Exposures of this granite cover an area of approximately 400 km^2, with the vast majority occurring in the northern and northeastern parts of the area. The outcrop pattern here is such that the granite appears as a broad strip extending from Osis in the east to Nubib in the northwest following the trend of, and partly coinciding with, the prominent Nam Shear Belt (Section 3.1.5), i.e. approximately northwest.

The Nubib granite gives rise to a very rugged topography, resulting in some of the highest peaks in the area. Nubib Mountain, from which the name of the granite has been derived, and neighbouring Losberg, both tower to heights of 1979 m, approximately 900 m above the surrounding Namib plain and are extremely inaccessible.

A large part of the pre-Nama 'Osis Ridge' (Germs, 1972), in addition to many small pre-Nama 'islands' immediately to the north of the ridge, are composed of Nubib granite, thus providing evidence of a persistently rugged terrain.

Typically, and with only minor exceptions, the Nubib granite has a reddish-brown colour making it a very conspicuous rock in the field. On weathering, the colour fades to a light reddish- or brownish-yellow. Exceptions to the above are quartz-felspar-porphyry types with a very fine-grained groundmass which may, probably as a result of the finer grain, have a deep red or maroon colour, and a small occurrence of granophyre on the southeastern flanks of Losberg, which has a grey colour. The relationship of the latter to red granite is not clear due to poorly exposed but apparently transitional contact relationships.

The texture of the Nubib granite is dealt with in the next subsection but it may be mentioned here that the most common textural variety is a fine- to medium-grained porphyritic and granophyric rock. Also important, on a macroscopic scale, is the presence of a miarolitic texture which appears sporadically throughout the granite, and is best developed in the vicinity of Naudaus beacon on Osis.

The relationships of the Nubib granite to other major rock units are as follows:

(i) Small dyke-like bodies of granitic material tentatively ascribed to the Nubib intrusive phase intrude the quartzites and amphibolites of the Kumbis Formation on southern Kameelhof. The presence of these minor bodies has not been indicated on the accompanying geological map due to their small size.

(ii) On Sonop, quartzites and arkoses of the Kunjas Formation have been intruded by a body of quartz-felspar-porphyry, which is a probable correlate of the Nubib granite since it also intrudes the Spes Bona syenite and is in turn cut by quartz-porphyry dykes of the Guperas Formation.

(iii) The vast majority of Nubib granite has invaded the lavas and clastic or volcaniclastic beds of the Barby Formation.

(iv) On Saffier, Aandster and Sonop the Spes Bona syenite has been intruded by Nubib granite.

(v) On Saffier the granite is once again the younger unit by intruding gabbroic rocks of the Barby Formation.

(vi) The relationship between the gneisses developed within the Nam Shear Belt and the undeformed Nubib granite is discussed in detail in Section 3.1.5 where evidence will be presented for the conclusion that the gneiss is a sheared and mylonitised product of the granite.

(vii) A normal sedimentary contact relationship between the Nubib granite and the younger Guperas Formation can be observed in many places, e.g. Persia, Naudaus, Duwisib, Rooiberg Süd and Osis.

(viii) Nubib granite (and gneiss) is cut by, and therfore pre-dates, basic and quartz-porphyry dykes and rhyolitic plugs of the Guperas Formation.

(ix) The granite is overlain by the Nama Group at many places in the vicinity of Osis and further to the northwest in the vicinity of Nam. Also in the latter area, the Auborus basal conglomerate rests on the granite with a normal sedimentary contact.

Contact relationships between the intrusive granite and country-rocks are always sharp, but with little evidence of chilling except in the case of a few small felsitic or fine-grained porphyritic apophyses where thin (< 5 cm) chill margins are developed. The mechanical effects of intrusion are dominant and ubiquitously well developed. Intrusion breccias at the contacts are common and particularly well displayed in the Nubib mountains. Blocks of country rock, mainly basic lava, within the granite range up to many meters in size, but are generally less than 20 cm in size. Usually the blocks have sharp outlines and assimilation has apparently been minimal, but granite in intimate association with the basic lava blocks may be grey in colour as a result of contamination, in contrast to the typical red colour of 'normal' Nubib granite.

In the immediate vicinity of the Eckberg trigonometrical beacon mainly, basic lava inclusions have rounded and diffuse outlines indicating some amount of assimilation. The presence of dioritic rocks is further evidence for fairly extensive contamination. No detailed petrographic or chemical study has been carried out on these hybrid rocks but the effects of hydridisation by basic material of the Rooikam granite, the correlate of the Nubib granite (Section 4) occurring south of latitude 25°45'S, have been treated in some detail by von Brunn (1967). These effects are considerable, but very local in their distribution, and the above author attributes the hybridisation to intrusion at a somewhat deeper level (in the "mesozone", *after* Buddington, 1959) than in the case of the majority of the Rooikam granite plutons.

All stages of hybridisation by assimilation of basic material as described by von Brunn (1967, pp.102-107) were noted in the present area, i.e.

(i) "Granite, modified by contamination of the basic material it has intruded" (von Brunn, 1967, p.103). Representative of this, first stage of hybridisation, is the grey-coloured granite mentioned above in association with sharply-defined blocks of basic breccia.

(ii) "'Granitised' basic lava, adamellitic in composition and ranging in texture from porphyroblastic/porphyritic (felspar insets) to micrograniti c. The colour of the rock is reddish-grey"(von Brunn, 1967, p.104). In the area north of the Sinclair Mine, such rocks are confined to the vicinity of the Eckberg trigonometrical beacon.

(iii) "Diorite (hybridised basic lava)" (von Brunn, 1967, p.104). Rocks of this type are also restricted to the Eckberg area and are associated with, and grade into, the material described under (ii) above.

Since the exact extent of contamination in all cases is unknown, no distinction has been made on the accompanying geological map between pure mechanical mixtures of granite and country-rock (mainly basic lava) and any of the above three recognised hybrid types.

Contact metamorphic effects, other than localised hybridisation are negligible in the basic lava, which almost invariably constitutes the country-rock. There is, however, minor induration of Kunjas sediments on Sonop and von Brunn (1967, p. 91) does note the development of andalusite-bearing hornfels in the Kunjas shales intruded by Rooikam granite.

Multiple intrusion is apparent from cross-cutting features within many of the outcrops, but since the various intrusions are apparently closely-spaced in time and display only slight textural differences in most cases, no attempt has been made in this study to differentiate between them.

The form of the intrusions is such that although the total outcrop area is large, the mass is composite and made up of many smaller stock-sized bodies (i.e. areal extent less than 100 km^2 according to Gary *et al.*, 1972. Furthermore, erosion and younger cover have resulted in outcrops of relatively small areal extent placing some uncertainty on their continuity as larger complexes. The Nubib mountain granite mass although composite, is continuous in outcrop and of batholithic dimensions (i.e. areal extent greater than 100 km^2 according to Gary *et al.*, 1972). Dyke- and vein-like intrusions are present and nearly always form small apophyses of larger bodies.

Pegmatitic phases of the Nubib granite are completely absent, and only very minor aplite is present, generally forming small isolated intrusions.

Petrography

On the basis of texture, five varieties of this granite can be distinguished:

(i) Fine- to coarse-grained (Whitten and Brooks, 1972) hypidiomorphic granular.

(ii) Porphyritic with a fine- to medium-grained hypidiomorphic granular groundmass. It can be emphasised here that the size distinction between phenocrysts and groundmass crystals is never very marked. The early-formed crystals generally display subhedral to euhedral form and are resorbed to varying degrees, the quartz in particular.

(iii) Granophyric. In general there is a complete range between both the afore-mentioned textural types and true granophyre. The vast majority of (i) and (ii) display granophyric texture to some degree and pure end members in the two ranges are not common (Plates 14 and 15).

(iv) Porphyritic with a devitrified or micrographic quartzo-felspathic groundmass and rare, vague fluidal banding.

(v) Very fine-grained, felsitic or micrographic.

The mineralogy of the different textural varieties is essentially similar and will, therefore, be described collectively.

Quartz in the granitic-textured varieties forms an inter-locking mosaic with alkali felspar and plagioclase felspar. The crystal form is anhedral to subhedral in this case and sizes range from less than 0,2 mm to 4,5 mm.

Quartz of the first porphyritic type discussed above (ii) is of two generations; early-formed subhedral crystals which are usually resorbed and embayed to varying degrees (Plate 15) and a second generation comprising part of the groundmass. The size of the early-formed quartz crystals is very variable, ranging from 0,2 to 4,0 mm in diameter, and it is the larger grains of this type that partly define the porphyritic character of the rock.

Spherulitic and granular devitrification intergrowths of quartz and felspar, as well as relatively large,resorbed,quartz crystals (up to 2,5 mm in diameter), occur in the second porphyritic type (iv).

Quartz of the granophyric varieties is intimately associated with potash felspar in the form of plumose and cuneiform intergrowths. Not uncommonly, an early-formed crystal acts as a nucleus to such intergrowths. Undulatory extinction is frequently observed but, as will be discussed in Section 3.1.5.2, such strain phenomena are restricted almost entirely to granites and their gneissic derivatives occurring in or immediately adjacent to the Nam Shear Belt in the northern part of the area (Fig. 6). Quartz makes up between 29 and 48 per cent of the total rock volume.

Alkali felspar makes up 32 to 57 per cent of the rock volume and is predominantly orthoclase perthite in the undeformed granites and porphyries, and microcline perthite in areas where these rocks have been affected by shearing. In a few specimens both types of alkali felspar are present. Much of the porphyritic granite of Losberg is anomalous in that it is apparently undeformed but contains microcline felspar only. Perthite is of the patch and string type and the exsolved sodic component may constitute up to 40 per cent of the total crystal volume.

A notable, but not very common feature, is the incomplete replacement of alkali felspar by sodic plagioclase progressing from the crystal margins inwards. The replacement plagioclase, which has a composition An_0 to An_5 frequently displays a chess-board twin pattern and is usually in optical continuity with the perthitic plagioclase lamellae of the host crystal.

A faint zoning representing successive stages in the crystallisation was observed in a few specimens (e.g. BW 350) and twinning, when present, is according to the carlsbad law. Characteristically, the alkali felspar of this granite has a very well-developed reddish-brown cloudiness which imparts a strong reddish-brown hue to the rock.

As with quartz, two generations of alkali felspar are frequently present. Early-formed, relatively large (up to 5 mm in diameter), somewhat resorbed, subhedral to euhedral crystals of a porphyritic or glomeroporphyritic nature, can be distinguished from later-formed anhedral plates, which may or may not be granophyrically intergrown with quartz. The latter range in size from $<1,0$ to 4,5 mm.

Plagioclase felspar, constituting 4 to 21 per cent of the rock, forms subhedral to euhedral crystals of early development, and the exsolution and replacement types already discussed. Frequently it bears a poikilitic relationship to the plates of anhedral alkali felspar. Some degree of saussuritisation is always present, but generally this is not a prominent feature. Microscopic determination of the plagioclase composition (not including exsolved plagioclase) indicates a range An_7 - An_{11}. Crystals vary in size from 0,2 to 9,0 mm and are of stumpy proportions.

Ferromagnesian minerals are relatively minor constituents (1 to 6 per cent), the most important being biotite, which has often been partially or completely altered to chlorite, epidote and opaque ore. In its rare fresh state the biotite is pleochroic in shades of brown, whilst pleochroic colours from brown to green are typical of partially altered examples. Chlorite often displays the typical anomalous deep blue interference colour of penninite. Dark radiation haloes within biotite and chloritised biotite indicate the presence of minute zircon inclusions. Rarely, hornblende is the dominant or only mafic mineral. Muscovite in addition to biotite was observed in specimens from a small pluton on the farm Nam (e.g. BW 1033).

Accessory amounts of sphene, apatite, fluorite, calcite and prehnite may be present, the latter two almost certainly being secondary. Opaque ore minerals always seem to be associated with the breakdown of biotite and are, therefore, mostly of secondary origin.

3.1.5 NUBIB AND BARBY METAROCKS OF THE NAM SHEAR BELT

Distribution and Field Character

A prominent zone of shearing occurs on the farms Duwisib, Eldorado, Gorab, Steinfeld and Nubib in the northern part of the area mapped. The effect of this deformation was to impose upon the rocks certain metamorphic effects and a foliation which strikes generally 120° - 160°, with dips slightly variable about the vertical. On the northwestern and northern parts of Gorab and Eldorado respectively, the foliation has been folded into an anticlinal structure the axis of which plunges from 0° to 45° to the south and the dip of the foliation planes here varies from 10° to 90° toward the southwest, south and southeast.

The shearing is confined to a belt approximately 15 km wide (where determinable). The southwestern margin of the belt, although gradational, is well-defined and easily recognisable in the field; the northeastern margin is to a large extent covered by younger strata of the Auborus Formation and the Nama Group. However, in the northeastern corner of the farm Nam, relatively undeformed rock-types can be found which become progressively more foliated when traced southwestward into the shear belt.

Rock-types involved in the shearing are basic lavas and intrusives, minor quartzites, conglomerates and rhyolites of the Barby Formation, various phases of Nubib granite which intrude the Barby extensively. Completely undeformed granite passes gradationally into highly sheared, foliated, gneissic granite and the progressive development of the macroscopic foliation is due to the fragmentation and elongation of mineral grains, particularly quartz. Dark mineral aggregates have become aligned parallel to the foliation direction and appear as streaky patches. Felspar phenocrysts are generally preserved as porphyroclasts up to a fairly high degree of deformation.

The general effect of shearing on the granite is, therefore, to impress a fabric upon the rock and to reduce the grain size by fragmentation, thus producing gneisses of a mylonitic character. In extreme cases, recrystallisation has occurred resulting in fine-grained glassy-looking rocks with few or no porphyroclasts. The red colour so typical of the majority of the Nubib granite persists through all stages of gneissification.

As will be shown later in this Section, the effects of the shear stresses can be detected microscopically even beyond the outer limits of areas with a macroscopically recognisable foliation.

Where the foliation has been folded, on the farm Eldorado, pegmatites have formed parallel to it near the axis of the fold. These pegmatites are unfoliated late syntectonic or post-tectonic secretions of quartz and felspar, with or without muscovite. A crude zonal arrangement of minerals is frequently developed showing a quartz core or central zone with a dominantly felspathic marginal zone. Narrow segregations which are less than 1 m wide, and are also composed of quartz, felspar and minor amounts of muscovite, occur in the extreme northern part of Steinfeld.

Most of the granite occurring within the shear belt has been affected by tectonism as described above, but a few intrusions of granitic material in both the marginal and central parts of the belt are exceptional in that either, (i) intrusion has been influenced by the shearing resulting in elongated bodies which intruded parallel to the foliation, and although usually sheared to some extent this is not a typical feature, or (ii) intrusion has not been influenced by shearing and the bodies are not visibly foliated, although, as will be described presently, microscopically it is obvious that the rocks have in fact been deformed to some degree.

Basic lavas of the Barby Formation which were intruded by Nubib granite within the shear belt have become amphibolitised and foliated. Along the margin of the belt, the lavas are frequently amygdaloidal and may possess the typical felspar-phyric character, and, although they are unfoliated, the ferromagnesian minerals have been altered to amphibole (e.g. BW 1196).

An intrusive body of amphibolitised gabbro, probably relating to the Barby basic extrusive and intrusive episode, occurs immediately to the south-southeast of the Steilabfal beacon. Contacts with the country-rock have been obscured by subsequent shearing, but the major portion of the body is unfoliated and possibly acted as a resister during deformation. Due to the rather vague nature of the contacts and the generally similar amphibolitic character of all the rocks in this area, this body was not mapped as a distinct unit but has been included in the meta-Barby Formation of the shear belt.

The granitic rocks were more easily fragmented than the basic rocks and, consequently, unfoliated examples of the latter are frequently found in contact with gneissic granites in the marginal parts of the belt (e.g. on north-central Gorab).

Immediately west and southwest of the Duwisiberplatte, basic lavas, intercalated sediments (quartzites and conglomerates) and intrusive Nubib granite have also been affected by a shear trending 125° and dipping 60°– 70° to the northeast. This apparently isolated 'patch' of shearing is not a direct southeasterly continuation of the Nam Shear Belt since a zone of unfoliated, but nevertheless strained, granite (Fig. 6) separates the two domains. The basic lavas, although displaying shear-stepping of felspar phenocrysts and fragmentation, have not been subjected to extensive recrystallisation. Granitic rocks in this area are mainly quartz-felspar porphyries which have developed a pseudo-flow-banding parallel to the foliation, i.e. porphyroclastic mylonites.

The contact relationships between the shear belt and rock units younger than the Nubib granite and Barby Formation are as follows:

(i) Quartz-felspar porphyry and basic dykes, relating to the Guperas igneous phase, cut the foliation imposed by the shear and are themselves not affected by it. Basal Guperas sediments are nowhere in contact with the sheared rocks.

(ii) Both the Auborus and the Nama sediments overlie the sheared rocks unconformably, the contact with the former is frequently a faulted one.

The sheared, foliated and metamorphosed rocks of the Nam Shear Belt including those to the southwest of the Duwisiberplatte, appear as 'undifferentiated Archean Complex' on the official Geological Map of South West Africa (1963), but the evidence presented here invalidates this interpretation to a great extent.

Schalk (pers. comm.) has found that north of latitude 25 00'S undoubted 'basement' rocks do occur and these would, therefore, form the continuation of those occurring south of latitude 25°00'S and interpreted here as sheared Sinclair Group rock-types. It is possible that the Nam Shear Belt represents a zone of relative crustal movement, with an uplifted block to the northeast of the zone, resulting in exposed basement rocks in that area. It is also possible, therefore, that some of the more highly deformed basic rocks occurring in the extreme northern part of the Sinclair area only on Steinfeld, immediately south of latitude 25°00'S, could represent pre-Sinclair basic rocks that have become involved in the shearing (Fig. 6). However, far more work of a more detailed nature than the present study would be necessary to establish whether these basic rocks are basement or not.

In any event, very little of what appears as undifferentiated Archean Complex south of latitude 25°00'S on the official Geological map of South West Africa (1963) should still be regarded as such.

Petrography of Nubib granite and gneiss

The texture of granitic rocks related to the Nubib intrusive episode, that have been affected by the shear deformation in the Nam belt, varies considerably depending on the degree of deformation. The mineralogy is also variable, but differences reflect changes in mineral phase rather than composition.

strained crystal aggregates are set in a fragmented matrix that may make up as much as 50 per cent of the total rock volume (Plate 16). Quartz in the matrix tends to form highly strained and elongated crystals or aggregates, imparting a foliated structure to the rock which is apparent on both macroscopic and microscopic scales. At this state of deformation all orthoclase has inverted to microcline but the perthitic texture is still present. Plagioclase laths usually have bent and/or shear-stepped twin lamellae. Clouding due to saussuritisation is ubiquitous but variable. Ferromagnesian minerals and other accessory minerals are very minor in amount, forming streaky aggregates of chloritised biotite, opaque ore, sphene, zircon and apatite.

(v) Streaky gneissic granite with microcline perthite, strained quartz and mylonitic texture: At this stage the rock consists of porphyroclasts or relatively unfragmented but highly strained crystal aggregates set in a matrix of highly strained and elongated fragmented material comprising 50-90 per cent of the total rock volume. Microcline and cloudy plagioclase form the porphroclasts which are always fractured and enclosed by a foliation imparted by elongated quartz crystals and aggregates of the matrix. Recrystallisation at this stage is limited and only involves quartz. Dark minerals, chloritised biotite and opaque ore, are oriented in, and therefore partly define, the general foliation of the rock. Sphene, zircon and apatite are accessories.

(vi) Streaky gneissic granite with microcline, slightly to moderately strained quartz, and blastomylonitic texture. Macroscopically, the foliation is still prominent mainly as a result of streaky, very fine-grained aggregates of dark minerals. Microscopically, however, the rock displays varying degrees of recrystallisation and a consequent loss of the marked directional texture. At the most advanced stage of recrystallisation observed (blastomylonite) the rock consists entirely of small granoblastic-polygonal crystals of relatively clear plagioclase, weakly perthitic microcline and slightly to moderately strained quartz. The microcline perthite lamellae, where present, are of the band or ribbon type (Spry, 1969, p. 182), and are always well-defined in contrast to the predominantly irregular patch and string perthite lamellae of the afore-mentioned granite and gneiss types. Small anhedral crystals of pale pink to colourless garnet commonly develop in accessory amounts and may form small aggregates which are elongated parallel to the foliation or relict foliation. Dark minerals are hornblende, chlorite and opaque ore, and zircon and sphene are always present as accessories, the former in noticeably larger amounts than in the undeformed granites. Muscovite is rarely developed. Dark minerals retain a streaky distribution and the polygonal recrystallisation texture may be superimposed on trains of ore granules representing the relict foliation.

The distribution of the various textural and mineralogical varieties of Nubib granite and gneiss in the northern part of the area has been plotted and contoured on the generalised geological map of Figure 6. Unfortunately, the samples examined are too few for the contouring to stand up to a rigorous statistical evaluation but the isopleths serve to illustrate the gradational development of the mylonitic gneisses from undeformed granites and the fact that the deformation is confined to a belt.

Furthermore, it is obvious from Figure 6 and the afore-mentioned petrographic descriptions that the effects of shearing extend beyond the limits of visible foliation.

Petrography of the basic lavas of the Barby Formation

The progressive intensification of dislocation metamorphism and foliation of the basic rock of the Barby Formation within the Nam Shear Belt compliments the same trend observed in the granitic rocks. The basic rocks have, however, responded to the shear stresses in a rather different manner to the granites. The first signs of macroscopic foliation are developed deeper in the zone and the mechanical effects of granulation or mylonitisation are to a large extent obscured by the growth of amphiboles after pyroxene and by alteration products of the original lava.

For convenience, the basic lavas and their amphibolitic derivatives occurring within the shear belt will be described as distinct 'types', but once again it must be realised that gradations exist between these 'types'.

(i) Marginal unfoliated, amphibolitised lavas and intrusives. No foliation is developed in the lavas occurring in the marginal parts of the shear belt. All ferromagnesian minerals (pyroxenes and/or their alteration products) have been converted to a blueish-green to pale-green amphibole with prismatic habit and a low extinction angle indicative of actinolite. Characteristically, no preferred orientation is displayed by the amphibole crystals at this stage (by contrast, the granitic rocks in contact with these lavas show strong foliation).

In many instances the lavas are amygdaloidal, or have the typical plagioclase-phyric character of the Barby lavas to the south. The plagioclase phenocrysts are always cloudy with saussuritisation which in most instances inhibits the determination of optical properties. A few measurements do, however, indicate compositions in the andesine-labradorite range. Groundmass plagioclase is also highly altered and no compositions could be determined. The plagioclase of the amphibolitised lavas is saussuritised to a far greater degree than the plagioclase of granites in the immediate vicinity, probably due to the more calcic nature of the former. Outcrops in the north-central part of Gorab and those of central Nubib fall into this stage of deformation/metamorphism. Certain ill-defined parts of the Gorab exposure have the appearance of gabbros rather than lavas, consisting of large moderately altered laths of plagioclase of labradorite composition and prismatic crystals of actinolitic amphibole as described above. Biotite which has been wholly or partially altered to chlorite is a minor, but nevertheless ubiquitous, constituent. Unstrained (recrystallised) quartz may also be present, and opaque ore is an accessory.

(ii) Moderately to strongly foliated, amphibolitised lavas and intrusives. The effects of shearing are well developed in these rocks. Plagioclase is very cloudy with alteration making the determination of composition virtually impossible except for small, less altered patches within the crystals, which provide indications of a composition in the andesine-labradorite range. Plagioclase phenocrysts or glomeroporphyritic aggregates of phenocrysts have been drawn out into lenticular shapes paralleling the foliation (Plate 17). Groundmass plagioclase is very cloudy, and also displays a tendency to be elongated parallel to the general foliation.

Amphibole is darker in colour than that of the previous stage, the extinction angle is higher and pleochroism is dark greenish-brown to lighter yellowish-green. Two textural forms of the amphibole are present:

a) Prismatic 'groundmass' crystals aligned in, and defining the foliation, and

b) Large unoriented plates which are enclosed by the foliation and have 'plucked' ends. These have probably developed from phenocrysts in the original lava and impart a blastoporphyritic textural element to the rock.

Amygdales, when present, are seen as stretched lenticular aggregates of slightly strained, but recrystallised, quartz crystals (Plate 17). Small slightly strained quartz crystals may form an important constituent of the groundmass.

(iii) Moderately foliated amphibolites. Examples of this extreme stage of deformation are not common and are confined to a narrow central part of the shear belt in and around the pegmatite zone on Eldorado, on north-central Nam and on Steinfeld. In these rocks the strong preferred mineral orientation of the previous stage has been somewhat weakened by complete or almost complete recrystallisation of all mineral constituents. Amphibole remains much as in stage (ii), but frequently displays considerable alteration to anomalously birefringent chlorite. Plagioclase forms small, relatively clear, polygonal crystals of composition An_{34}. Quartz is recrystallised and displays only slight undulatory extinction. Small amounts of epidote are commonly present. These amphibolites could represent basement amphibolites, as discussed earlier.

The basic lavas occurring within the small zone of shearing to the southwest of the Duwisiberplatte show considerable granulation but no development of amphibole. The effects of shearing are particularly evident in the plagioclase-phyric varieties. Here, the plagioclase phenocrysts are fragmented and excellent examples of shear-stepping of the twin lamellae are present (Plate 18), and saussuritisation is only slight. All original ferromagnesian minerals have been altered to chlorite and minor epidote and ore.

3.1.6 GUPERAS FORMATION

3.1.6.1 Basic Lava

Distribution and Field Character

Over the major part of the outcrop area, of the Guperas Formation, basic lava does not feature very prominently within the sequence. The exception to this is a local pile of basic lava at least 520 m thick, on the southwestern part of Ganaams, which conformably overlies a very thin basal sequence of ash-flow tuff, conglomerate and a thin band of basic lava (see stratigraphic sections B and C of Fig. 1). Elsewhere, basic lava forms thin horizons intercalated within the Guperas sandstone and/or conglomerates, and invariably, but not exclusively, they occur within the uppermost 150 m of the sequence below the sediment/rhyolite contact. Occasionally the felsic volcanics may rest directly on the upper surface of the basic lava with no intervening clastic material (e.g. Fächerberg area) or, as in the central Aruab-Guperas area the acid volcanics form a relatively thin wedge within the clastics and the basic lava occurs about 350 m above this unit.

The thick pile of basic lava immediately east of the Hahnenkamm trigonometrical beacon is unique in the Guperas succession constituting a local phenomenon since stratigraphic sections 8 km to the northwest and 6 km to the east, are barren of such lava.

The average thickness of individual basic lava units varies from 3 m to 25 m except for the Hahnenkamm-Guchaber Nase occurrence, and the lateral extent is limited to a maximum of 4 km. Units are usually composite, consisting of numerous distinct flows, but more rarely, they may be comprised of a single flow.

Although basic lava is not entirely confined to areas where felsic volcanics are prominent in the sequence, there is a definite sympathetic relationship between the distribution of these two extrusive units. The band on southern Guperas-Wittmanshaar is the only known occurrence not associated with nearby (stratigraphically and geographically) felsic volcanic horizons.

The lava is very fine-grained, amygdaloidal and slightly porphyritic. The degree of vesicularity varies greatly and in the highly amygdaloidal lavas the amygdales are very irregular and coalescent. Furthermore, as is to be expected, the uppermost portions of the flows exhibit a more pronounced amgydaloidal character than the lower portions. Leaching of amygdale minerals on weathered surfaces frequently leaves the rock with a pitted and 'vesicular' appearance. The colour is also variable, the most common being greenish-grey. Dark red and black rocks may also be prominent, as for example on southern Guperas-Wittmanshaar where a dark maroon lava is the dominant variety.

A relatively rare rock type quite distinct from those described above is a true felspar-phyric lava, which is indistinguishable in hand-specimen from certain basic dykes and intrusive bodies of the Guperas Formation (Sections 3.1.6.2 and 3.1.6.3). The only occurrence of this lava is immediately northeast of Fächerberg trigonometrical beacon where it is interbedded with the more common finer-grained, sparsely porphyritic and highly amygdaloidal lava.

Petrography

Under the microscope the coarser-grained lavas have typical intergranular textures, subhedral plagioclase laths of maximum lengths of 20 mm and random orientations, with interstices occupied by ferromagnesian minerals (BW 483, BW 1506). As the grain size decreases the amount of interstitial glass increases, resulting in an intersertal texture (BW 1489) which may further grade into a hyalopilitic texture where the plagioclase crystals and microlites 'float' in a glassy mesostasis (BW 487). In rare instances a vague flow orientation of plagioclase microlites results in a pilotaxitic modification of otherwise hyalopilitic textured varieties (BW 1513). Most of the lavas are sparsely porphyritic with phenocrysts of plagioclase felspar and/or clinopyroxene.

Without exception, the lavas are amygdaloidal, but often only very sparsely so, and the amygdales may be extremely small (< 1 mm). The shape of the amygdales is not constant although most commonly they are spherical or ovoid. The higher the degree of vesicularity the more irregular are the cavities.

Secondary or deuteric minerals filling the cavities are calcite, quartz, chlorite, zeolites and epidote, in order of decreasing importance. Frequently, silica minerals line cavities which are otherwise filled by any one or more of the other above minerals.

Phenocrysts of plagioclase felspar constitute a maximum of 8 per cent of the total rock volume and are subhedral to euhedral in shape, some having a resorbed outline. Their size ranges from 0,4 to 2,0 mm and glomeroporphyritic aggregates frequently occur. The composition of phenocrystic plagioclase could not be determined due to extensive alteration. Plagioclase of the groundmass generation ranges in size from 0,03 mm (virtually cryptocrystalline) to about 0,8 mm and mainly has a subhedral to euhedral development. Crystals and microlites are distinctly prismatic and almost needle-like in shape. The composition is remarkably constant at An_{30}- An_{33} throughout the area and succession. Normal type zoning of the plagioclase is common, but not strongly developed.

Ferromagnesian minerals are the clinopyroxenes augite and pigeonite and possible minor constituents of orthopyroxene, collectively making up a maximum of 21 per cent by volume. Pigeonite is also somewhat rare in comparison to augite. Both are pale-brown to neutral in colour and usually display severe or complete alteration to chlorite and pale-green fibrous uralite amphibole. The presence of a second ferromagnesian mineral (orthopyroxene ?), as evidenced by serpentinised and chloritised grains with a red stain due to iron exsolution, is suspected in a few lavas. These crystals display an earlier-formed habit than the clinopyroxene, but when present are volumetrically subordinate to clinopyroxene.

Pyroxene occasionally forms crystals of phenocryst dimensions (maximum size 2,0 mm) and may also occur in glomeroporphyritic aggregates. 'Groundmass' pyroxene is also intergranular. The crystal form is anhedral and twinning is present but not ubiquitous.

Interstitial glass, crowded with very fine granules of opaque ore minerals, is a very common constituent of these lavas (up to 42 per cent by volume). Alteration and the presence of ore inclusions have resulted in a characteristic dark brown colour and a murky appearance.

The rare felspar-phyric lava (e.g. BW 1456) has such a distinctly atypical texture that its petrography is discussed separately below:

The rock is composed of 18 per cent plagioclase phenocrysts set in a fine-grained intergranular groundmass of predominantly plagioclase and clinopyroxene, or their alteration products. Small spherical quartz-filled amygdales are sparse but always present. Plagioclase phenocrysts may be up to 1,3 cm in size and are of subhedral to euhedral nature. Their composition could not be determined due to extensive saussuritisation. Groundmass plagioclase crystals are subhedral, range in size from 0,05 mm to 0,5 mm and have a composition in the mid-andesine range. Zoning and incipient to advanced alteration does not allow more accurate determination to be made. Clinopyroxene consists of pale-brown augite which is invariably extensively altered to chlorite and uralite. A cryptocrystalline mesostasis is present, crowded and darkened by opaque ore granules and alteration.

3.1.6.2 Basic Intrusives (excluding dykes)

Intrusions of basic material relating to the Guperas cycle occur mainly within the area of outcrop of Guperas sedimentary rocks and lava in the eastern half of the area. Although undoubtedly of similar age, these intrusives vary considerably as regards texture and mineralogy and will, therefore, be described individually.

(i) On northern Aruab an elongate intrusion aligned approximately east-west, with an outcrop area of roughly 2 km^2, crosscuts Guperas sandstone and conglomerate and is in turn intruded by rhyolitic material of the same Formation. The rock has a very distinctive porphyritic character with well-formed plagioclase insets within a fine-grained greenish-grey coloured groundmass. The weathered colour is a pale brown. In hand specimen (and in thin-section) this rock is indistinguishable from the basic porphyry dykes discussed in Section 3.1.6.3.

Petrography: Relatively large (up to 1,2 cm in length) and randomly oriented subhedral to euhedral plagioclase phenocrysts, making up 14 per cent of the total rock volume, are set in a fine-grained intergranular groundmass (e.g. BW 1551). The insets are unzoned, simply and polysynthetically twinned, only moderately saussuritised, and have a composition of An_{54}.

The groundmass is composed of subhedral laths of plagioclase felspar (0,05 to 0,8 mm in length), partly interstitial clinopyroxene and a very cloudy interstitial mesostasis of hypocrystalline material with minor quartz. Plagioclase of this generation is mainly normally zoned, but very faint oscillatory zoning was also observed. The composition ranges between about $An_{49}$and An_{44} from core to margin respectively. Pyroxene is pale-brown anhedral augite which is always partially altered to chlorite. Opaque ore granules are always present.

(ii) A basic body with a total outcrop area of about 1 km^2 is intrusive into the Guperas sandstones,quartzites and conglomerates in the eastern part of Naudaus. The occurrence is rather poorly exposed due to scree cover and river gravels but the intrusive relationship can be clearly seen on the banks of a southerly-trending tributary of the Naudus River. The rock is a dark greyish-green coloured pyroxene porphyry of fairly homogeneous character, which weathers to a strong rusty red-brown colour.

Petrography: Subhedral to euhedral and frequently glomeroporphyritic clinopyroxene phenocrysts (up to 23 per cent by volume) are set in an intergranular groundmass which is composed of plagioclase laths, clinopyroxene and opaque ore minerals (e.g. BW 1390). Alteration of all components is extensive, and in the case of the groundmass constituents is complete. Saussuritic pseudomorphs mark the presence of plagioclase, and the pyroxene has been reduced to pale-green fibrous amphibole (actinolite) and chlorite. Pyroxene phenocrysts are only partly and marginally altered, and unaltered portions indicate pale-brown augite.

A few plagioclase crystals (pseudomorphs) exceed the average groundmass grain size (0,3 mm) by a sufficient order of magnitude to be termed phenocrysts (maximum size of 2,0 mm) but the rock remains dominantly a pyroxene porphyry.

Many small opaque ore mineral crystals of both primary and secondary (alteration of ferromagnesian minerals) origin are present in the groundmass.

Small quartz-filled amygdales can occasionally be observed.

(iii) A poorly exposed outcrop of a fine-grained and dense basic rock appears on central Naudaus, forming two low scree-covered hills. Contacts with the Guperas quartzite and sandstone are not well exposed, but from limited field evidence it can be inferred that the relationship is an intrusive one. Total outcrop area is in the order of 4 km^2. The body is cut by a number of quartz-porphyry dykes and one basic dyke, of the Guperas Formation.

Petrography: The rock is highly altered and its texture, which was probably fine-grained intergranular, has been largely obliterated by the growth of secondary or late-deuteric fibrous uralite after the ferromagnesian minerals, presumably pyroxene. Alteration products of plagioclase felspar, are plentiful, and the presence of interstitial alkali felspar is also suggested by cloudy remnants (e.g. BW 1491).

(iv) A small (0,07 km^2) oval-shaped intrusive of medium-grained quartz-hypersthene gabbro, occurs on Persia and displays, in outcrop, a very faint and roughly horizontal preferred orientation of plagioclase crystals. An intrusive relationship to the flaggy sandstones and conglomerate of the Guperas Formation is evident. The rock is of medium-grey colour on fresh surface, weathering to a light reddish-brown.

Petrography: Subhedral labradoritic plagioclase laths (0,2 to 4,5 mm in length) with a well-pronounced preferred orientation (igneous lamination) are found together with mainly interstitial ferromagnesian minerals hypersthene and pigeonite (total of 17 per cent by volume), and lesser amounts of interstitial micropegmatite and quartz. The hypersthene is not markedly pleochroic and is partly serpentinised (e.g. BW 1380).

(v) The youngest basic intrusive body belonging to this group (on the basis of field evidence at presently exposed erosion levels) and in fact the youngest known such intrusive of the entire Sinclair Group, is a gabbro occurring on northern Naudaus. The country-rocks are basal Guperas quartz-felspar prophyry lava and underlying conglomerates. The shape of the intrusion is elongate (about 2 km long, and a maximum of 0,5 km wide) in an approximately north-south direction, tapering quickly northward into a dyke of variable width, but averaging about 15 m. The rock is a pale greyish-green colour and weathers to a rust red-brown. The presence of sparse patches of dark green altered glomeroporphyritic pyroxene imparts a spotted character to the rock.

Petrography: The texture is intergranular with porphyritic tendencies. Pyroxene forms infrequent glomeroporphyritic aggregates and occasionally plagioclase may also be phenocrystic. The grain size distribution is of a rather seriate nature (0,1 to 5,0 mm), so that the porphyritic textural element is not very obvious. Furthermore, pyroxene is not always strictly intergranular or phenocrystic and often tends to be slightly ophitic.

The plagioclase felspar, making up 57 per cent of the rock, is subhedral to rarely euhedral, and is extensively saussuritised, the latter property inhibiting a determination of the composition. Augite and pigeonite are present (16 per cent in total) and both are pale brown in colour, anhedral in outline and frequently twinned. The pyroxene is always chloritised to some extent and the larger glomeroporphyritic crystals are usually completely altered. Occsionally, pyroxene crystals are thinly rimmed by brown pleochroic amphibole. Also present is a very cloudy interstitial mesostasis of alkali felspar (?) and clear quartz, the latter filling the central portions of the interstices and constituting 27 per cent of the rock volume. Opaque ore is always present in accessory amounts.

(vi) The occurrence of a laminated uralitised gabbro body in the southern tip of Losberg on Nubib West is rather problematical. It is the only basic intrusive of post-Nubib granite age that occurs outside the main area of Guperas Formation emplacement (as seen today). Whether it can be directly related to the set of intrusions discussed above is not certain. The gabbro intrudes and, therefore, post-dates the Nubib granite, and 3 km to the southeast a small dyke-like body very similar to the gabbro cross-cuts basic lava of the Barby Formation. Furthermore, xenoliths of basic amygdaloidal lava within the main gabbro undoubtedly belong to the Barby Formation. Aplite veins also cross-cut the gabbro and may possibly be derived from rheomorphism of Nubib granite. The lamination dips to the south-southwest at an angle of approximately 75° and is considered to represent primary crystal settling reoriented to a steep angle after consolidation.

Petrography: The gabbro consists of roughly equal amounts of saussuritised plagioclase laths and mainly interstitial clinopyroxene that is almost completely uralitised and chloritised. Minor quartz is interstitial, and accessory opaque ore minerals are both primary and secondary as a result of alteration of ferromagnesian minerals. Other alteration products and accessory minerals are epidote and sphene.

3.1.6.3 Basic Dykes

Distribution and Field Character

Basic dykes relating to the Guperas cycle of igneous activity are fairly common throughout the area although they are subordinate to the quartz-porphyry dykes of the same Formation (Section 3.1.6.5). In the extreme southwest, on Vergenoeg and adjoining farms Kumbis, Sonop and to a lesser extent on Spes Bona, a very dense swarm is present. The strike of these dykes varies between 35° and 45° and their attitude is vertical or steeply dipping to the northwest. The width varies from 1 m or slightly less up to a maximum of 25 m, and individual dykes can be traced along strike for distances of up to 6 km. However, since their terminations are frequently sand covered, this is a minimum figure.

Morphologically, the dykes may form either positive or negative features depending on the nature of the country-rock, e.g. positive ridge-like features in Kumbis adamellite and slightly negative features when bordered by Barby lava.

In this southwestern area basic dykes cross-cut, and therefore post-date, the following units: Kumbis Formation, Kunjas Formation, Barby Formation, Spes Bona syenite and Nubib granite. Intersections between quartz porphyry and basic dykes indicate that the former post-date the latter.

From the accompanying geological map it would appear that basic dykes cutting the Barby Formation are totally absent, but only a representative proportion of the dykes has been shown in order to avoid obliteration of other geological features, and a large number of these basic dykes are in fact present cutting the lavas. The greater resistance to intrusion offered by the Barby lavas is considered to have contributed significantly to this decrease in intrusion density.

The second major occurrence of basic dykes of Guperas age takes the form of a small but very dense swarm on the northern part of Blutpütz West and the adjoining part of Naudaus. Within the swarm, however, basic dykes are once again subordinate to the Guperas quartz-porphyry dykes. Field characteristics here are essentially similar to those of the southwestern swarm, with the exception that the dykes cut Guperas basal sandstone, conglomerate, quartzite and Nubib granite, and strike remarkably constantly in a direction 355° - 360°, with a vertical attitude. Very few basic dykes have been found cutting the Guperas rhyolites, and this is considered significant since it is felt that the dykes are related to the intrusion of bodies of basic material into the Guperas sequence and, the extrusion of varying thicknesses of basic lava intercalated with the Guperas sediments and underlying the felsic lavas. In at least one instance, on northern Naudaus, one of these basic dykes swells along strike and passes into a basic 'body'.

Petrography

Two different, essentially holocrystalline, textural varieties of these dykes are apparent in thin-section (and also in hand-specimen):

(i) Porphyritic with phenocrysts of plagioclase felspar and/or, more rarely, clinopyroxene set in a fine-grained intergranular groundmass of predominantly plagioclase and clinopyroxene, or their alteration products.

(ii) Fine- to medium-grained intergranular ± interstitial micropegmatite.

Sparse small and well-rounded quartz-filled amygdales are frequently present. In hand-specimen both types have a greenish-grey colour on fresh surface, weathering to a reddish-brown. The porphyritic type is a very distinctive rock, particularly when weathered, with the cream-coloured plagioclase insets contrasting strongly with the red-brown groundmass.

Plagioclase felspar of the first above-mentioned textural variety is of two generations, i.e. relatively large (up to 1,5 cm in length) phenocrysts forming subhedral to euhedral laths (25 per cent by volume) set in a finer-grained (average length of 0,3 mm) groundmass generation of mainly subhedral laths, which bear an intergranular relationship to other groundmass components. In the intergranular non-porphyritic rocks the plagioclase makes up an average of 56 per cent of the rock volume and occurs as a loosely-packed mosaic of lath-shaped crystals (0,1 to 1,5 mm in length). Very extensive saussuritisation has affected the plagioclase and frequently only pseudomorphic shapes remain. Relatively unaltered crystals, or parts of crystals, always exhibit typical polysynthetic twinning. Zoning is ubiquitous and normal and is most severe in the marginal parts of the crystals. The extensive alteration and zoning inhibit, to a large extent, the accurate determination of composition but a limited number of analyses from non-porphyritic and groundmass plagioclase indicate compositions ranging from An_{50-56} to An_{40-44} in crystal cores and margins respectively. Minor, but in some cases significant, amounts of alkali felspar occur interstitially, and it is less altered than the plagioclase felspar (e.g. BW 130).

Clinopyroxene constituting a maximum of 47 per cent of the rock volume, is the major ferromagnesian mineral and is a pale purplish-brown augite which is usually largely interstitial but in some cases (e.g. BW 1411) it may also be phenocrystic (maximum size of 2,0 mm). Ophitic tendencies were noted in one dyke (BW 468). The crystal outline is anhedral. Slight zoning is almost universal, and often a distinct core zone, as emphasised by the extinction and higher state of alteration, is present. Alteration of pyroxene to pale-green fibrous uralite and chlorite (penninite) is extensive.

Quartz is nearly always present in small amounts, once again interstitially, and is very often intergrown with felspar and the form of micropegmatite. Some chloritised biotite was observed in one specimen (BW 130). Opaque ore and apatite are accessories. Alteration products have already been mentioned, with the exception of calcite which is invariably present in minor quantities.

3.1.6.4 Rhyolites

The term rhyolite is used here in a general sense to include all extrusive or closely related sub-volcanic rocks of highly siliceous nature and a more precise chemical and mineralogical classification is applied in a later Section (Section 3.3.3). Reasons for employing the term 'rhyolite' or 'rhyolitic' are:

(i) to distinguish between extrusive, and related, but solely intrusive and texturally similar, features of the same igneous episode (e.g. the term quartz-porphyry is applied to the late felsic dykes of the Guperas Formation),

(ii) to avoid the usage of a complex terminalogy in the following decriptions, and

(iii) to conform to modern general usage of the term as applied to felsic extrusives in orogenic belts (e.g. Hyndman, 1972; McBirney & Weill, 1966; Pichler & Zeil, 1972).

Rhyolitic volcanic activity during the Guperas cycle is manifested by:

(i) Lava flows

(ii) Domes and plugs

(iii) Volcaniclastic rocks (Fisher, 1966, p.289)

(a) Ash-flow tuff (ignimbrites)

(b) Tuff, lapilli-tuff and tuffite beds.

On the accompanying geological map the lavas and volcaniclastics have not been differentiated and appear under the common label of 'quartz-porphyry lava and agglomerate'. Where it is considered necessary, text figures are included in this thesis to indicate greater lithological detail. The stratigraphy of the Guperas Formation and the general geographic and stratigraphic distribution of the volcanic units within the total succession have already been discussed (Section 2.2.7).

Present exposures of Guperas rhyolites are confined to three areas or 'fields':

(i) Kronenberg-Auramberg mountain range on western Ginas,

(ii) Blutpütz West and immediately adjacent areas to the south and east, and

(iii) Rooiberg mountain.

It can only be speculated as to what extent these three areas represent main eruptive centres or parts thereof. The Kronenberg-Auramberg and Blutpütz West occurrences both show marked wedging out in a southeasterly and southerly direction respectively, and this indicates a maximum development of volcanic activity in a direction coinciding with the present outcrop areas. Since these two areas are now separated by a younger cover of the Auborus Formation within a north-south trending graben structure, it is impossible to prove whether they originally formed a single, considerably larger, volcanic field or not, but such a possibility would seem likely.

Due to younger cover and erosion, the southern and southeastern limits of the Rooiberg occurrence cannot be determined, but an original surface connection with the Blutpütz West field is considered likely, despite the almost complete lack of subvolcanic rhyolite intruding the Guperas sediments, Barby basic lavas and Nubib granite in the intervening area on Naudaus.

A brief visit to the Awasib Mountains in the Sperrgebiet to the west of the area studied has revealed the presence of an appreciable thickness of Guperas lithic sandstone and conglomerate which can be correlated with the succession of the Naudaus trigonometrical beacon area. There are no volcanics overlying or intercalated within these beds, indicating that rhyolitic extrusive activity did not extend very far west of the Kronenberg-Auramberg Mountains.

(i) Rhyolitic lava flows

Distribution and Field Character

Rhyolitic lava is volumetrically by far the most dominant extrusive rock of the Guperas sequence. A thickness of 1000 m is present in the Kronenberg-Auramberg mountains, where the flows building up this pile dip at low angles to the northeast and east. The preserved thickness in the Blutpütz West mountains is approximately 650 m, and the orientation of the lavas is predominantly horizontal, with dips as high as 25^{o} locally. The Rooiberg occurrence has the form of a very shallow north-south trending synclinal structure and here the lava pile has an exposed development of 800 m. All the thicknesses mentioned here must be considered as minimum figures since the upper surface of the lava piles is nowhere preserved or exposed.

Due to the generally thick nature of the flows their stratiform development is not always obvious, and textural and compositional variations are usually not sufficient to produce any marked differences in weathering characteristics which might accentuate the layer-pile nature. Distal terminations are always abrupt where observed, without the tapering character of a less viscous lava flow.

The extremely resistant nature of the lava ensures that exposures are always prominent and of considerable relief. Rugged mountains of towering proportions with steep slopes and vertical cliff faces, often many hundreds of metres in height, are characteristic topographic features associated with the rhyolite lava outcrops.

The relationship between the lava and other rhyolite units of the Guperas Formation is discussed in the following subsections, but it is appropriate to mention here that the lava sequence is, with only minor exceptions, devoid of intercalations of clastic or pyroclastic horizons. The exceptions are found about 6 km east of the Fächerberg trigonometrical beacon where 60 m of ash-flow tuff and three bands of typical Guperas lithic sandstone and conglomerate, 30 to 150 m thick, occur as intercalations within the rhyolite lavas.

The lava is always porphyritic, with phenocrysts of quartz and/or alkali felspar and/or plagioclase felspar set in an aphanitic groundmass of typical red, reddish-brown, maroon or lilac colour. Grey to green-coloured varieties are common in the Kronenberg-Auramberg mountains but are still subordinate to reddish-coloured lavas. On weathering, the rocks take on a red-brown to yellow hue, which is not always markedly different from the fresh surface colours.

Flow-banding is well developed in the vast majority of flows, and ranges from highly contorted fluidal structures to roughly linear laminae paralleling the base of the flow unit. In the absence of flow-banding the groundmass is massive and structureless.

A vesicular texture is not abundant in the lavas but frequently occurs in flows of the Blutpütz West field, particularly in those near Fächerberg.

The cavities may be highly irregular or spherical and randomly oriented, but more commonly they are severely stretched within the flow-banding. Vesicles may either be completely or only partly filled (lined) by quartz.

Lithophysal structures were noted in a few flows in the vicinity of Fächerberg where the spherical forms are small, ranging in diameter from about 2 mm to 3 cm, and are not very tightly packed.

Petrography

A porphyritic character is typical of these lavas, with phenocrysts of quartz and/or plagioclase and/or alkali felspar constituting up to 18 per cent of the total rock volume. Slightly more than 50 per cent of all extrusive rhyolites examined in thin-section display some sort of flow structure. A true fluidal texture is confined to the groundmass but other features such as lenticular concentrations of glomeroporphyritic quartz phenocrysts (e.g. BW 103), preferentially oriented plagioclase laths (BW 103) and drawn-out (or flattened) vesicles, also contribute a directional element to the texture of some of the flows, and are the result of a viscous flow mechanism.

Quartz forms the dominant and most persistent phenocryst mineral, and is a characteristic inset of all flows with the exception of the basal units at the southern end of the Rooiberg field on Naudaus and Rooiberg Süd, where it is completely absent as a phenocryst mineral, as is alkali felspar. Quartz phenocrysts constitute an average of 5 per cent of the total rock volume with a maximum of 10 per cent. The shape of quartz phenocrysts varies to a great extent depending on the state of the ubiquitous corrosion or resorption. All grains show at least some degree of 'rounding' and the more advanced state of resorption has resulted in perfectly rounded, irregular or deeply embayed crystal outlines. Rounded, embayed, bipyramidal or hexagonal high-temperature forms are common. Generally, it would appear that most crystals were euhedral, or nearly so, prior to resorption. The quartz has been more severely affected by corrosion than any of the other phenocryst minerals. Undulatory extinction is not a common feature and the quartz is clear and free of inclusions. Individual crystals range in size from less than 0,1 to 3,2 mm in diameter, and glomeroporphyritic aggregates are not uncommon.

Basal flows occurring 3 km east of Fächerberg display all the characteristics of a lava flow in outcrop and in hand-specimen. In thin-section, however, it is apparent that much of the phenocryst quartz has been highly fragmented, prior to emplacment. The sharply angular and chip-like nature and wide size-range of the quartz cannot be accounted for by resorption as in other samples where the angularity has a 'smoothed' outline. It seems possible that this flow actually represents a highly welded ash-flow deposit and sparse lithic inclusions visible in outcrop support this postulate. Complete devitrification of the groundmass, however, renders the recognition of shards impossible. It may thus be possible that many of the severely devitrified flows, which do not display highly fluidal textures typical of lava flows, but which have a less well-defined flow structure, may actually be welded and compacted ash-flow tuffs. Unfortunately the poor state of preservation of fine original textures does not allow for a definite decision to be made concerning the mechanism of emplacement.

Plagioclase felspar phenocrysts constitute up to 13 per cent of the rock, with an average of 3,1 per cent and are, therefore, subordinate to quartz phenocrysts in amount. The crystals are euhedral to subhedral and usually do not show well-developed resorption effects; only very occasionally do they appear as highly corroded 'skeletal' remnants. Individual crystals range from less than 0,2 to 4 mm in length and are rather stumpy in outline. A glomeroporphyritic habit is well-developed, and aggregates measuring up to 6 mm and comprising more than 20 individuals were noted. Twinning is universal and according to the albite, carlsbad and pericline laws. A reddish-brown cloudiness is ubiquitous and alteration to sericite and lesser amounts of epidote, although not extensive, is present in most samples.

The composition ranges from An_0 to An_{10} and a predominance of values indicating pure albite possibly suggests secondary albitisation processes. This is supported by the extensive albitisation of the alkali felspar which is discussed below.

A second generation of plagioclase of pure albite composition (An_0-An_5) is clear in comparison to the primary phenocrystic plagioclase and displays a characteristic 'chess board' twin pattern (e.g. BW 1293 and BW 99). Most alkali felspar phenocrysts exhibit some degree of replacement by 'chess-board' albite, regardless of whether primary plagioclase phenocrysts are present in the rock or not. Replacement occurs as irregular patches within the crystals and is frequently complete.

The chess-board twin pattern appears to be characteristic of secondary albite replacement of potassic felspar, and according to Callegari and de Pieri (1967) the twinning is the result of internal stresses that develop due to marked differences in the cell volumes involved.

Alkali felspar is the least prominent phenocryst mineral and makes up an average of 3 per cent of the total rock volume, with a maximum of 10 per cent. These figures probably do not represent the original total alkali felspar content because of the extensive albitisation which has affected the crystals as described above.

The size of phenocrysts ranges from less than 0,2 mm to 0,8 cm and the crystal form is variable from anhedral to euhedral. Resorption effects are not as prominent as in the quartz phenocrysts but certainly more pronounced than the plagioclase. Resorption results in generally rounded outlines and in an advanced state it may reduce the crystal to a skeletal framework. Perthitic exsolution is common and the crystals may exhibit carlsbad twinning.

A few samples (e.g. BW 1501) contain chlorite, opaque ore minerals and calcite, all pseudomorphic after ferromagnesian mineral phenocrysts. These crystals are never large (maximum diameter 1,2 mm), and their outlines are invariably rounded. Generally, phenocrysts of ferromagnesian minerals are not common and on the average constitute trace amounts only.

Occasionally, small (maximum diameter 0,8 mm), well-rounded and resorbed phenocrysts of opaque iron ore and 'micropegmatite phenocrysts' (Williams, Turner and Gilbert, 1958, p. 126), in which alkali felspar and quartz are graphically intergrown, may also occur. Sample BW 1334 shows an unusually abundant development of ore phenocrysts.

The groundmass texture can be broadly classified into two types: fluidal and massive.

Massive varieties consist of micro- to cryptocrystalline, granular, poikilitic, and spherulitic or plumose growths of quartzo-felspathic material all resulting from devitrification of an original glassy matrix.

Radiating and spherulitic structures composed of fibrous felspar and cristobalite vary greatly in size and development. Small spherulites up to about 0,7 mm in diameter may develop sporadically throughout a cryptocrystalline granular groundmass (Plate 21), or the groundmass may consist entirely of such spherulitic growths. Also common are large (up to 4,5 mm in diameter) spherulites which frequently possess phenocryst nuclei. The 'interstices' are composed of cryptocrystalline granular material. When these larger, better-defined spherulites develop in tightly-packed fashion, growth interference is apparent and polygonal forms have resulted. The tendency for the growth of large devitrification spherulites to proceed from phenocryst nuclei is not consistent, and often radiating structures have developed completely independently of any phenocrysts.

Fluidal textured varieties range from rhyolites, with barely discernible flow structure, to rocks displaying incredibly contorted banding that could only have arisen in a fluid, but presumably very viscous, medium. Fluidal banding is recognised or accentuated by a number of textural and mineralogical properties which, in fact, characterise the rhyolites of this type:

(a) Alternating bands of variable width in which reddish-brown iron oxide staining varies in intensity due to varying concentrations of the oxides, or alternatively, varying states of oxidation caused by "streaming of volatiles into minute tension cracks diagonal to the fluidal lamination" (Williams, Turner and Gilbert, 1958, p. 122).

(b) Concentration of opaque ore granules within certain bands.

(c) Development of 'pegmatite' zones, due to vapour phase crystallisation (Ross and Smith, 1961, p. 26) within tension spaces perpendicular to the flow lamination. Felspar-rich marginal zones and quartz-rich cores are typical.

(d) Devitrification has resulted in (i) axiolitic structures developing along flow laminae and (ii) varying grain size between adjacent laminae or bands. This latter case refers particularly to granular devitrification products, but in some instances concentrations of spherulites within bands bordered or separated by granular or axiolitic material are also present (Plate 21). On the other hand, large spherulitic growths often tend to obscure the original fluidal texture, and were it not for pegmatite zones flow structure in thin-section would be obliterated.

(e) Parallel alignment of small, slender, opaque ore needles.

The groundmass of most of the rhyolites is characterised by a reddish-brown colour due to ferric oxide. A few lavas have a pale greenish-brown coloured groundmass with relatively high proportions of very fine-grained alteration products (mainly chlorite, ore, epidote) of pre-existing ferromagnesian mineral granules, resulting in the overall greenish-grey colour of the rock in hand-specimen.

Minute opaque ore mineral grains are always an important accessory constituent of the groundmass, and frequently devitrification has brought about a redistribution of these grains. Growth of radiating structures without inclusion of the ore has concentrated them at the outer margins or in the interstices of the spherulites.

Perlitic cracks, now largely obscured by extensive devitrification are apparent in some flows.

Slender needles of cristobalite measuring up to 0,3 mm in length characterise the groundmass of some rhyolites (e.g. BW 101), and those in contact with quartz phenocrysts are in optical continuity.

Chlorite, epidote, sericite and opaque ore are alteration products, and accessory amounts of zircon, apatite, sphene and secondary calcite and hematite were noted.

Vesicular rhyolites are not very common in the Guperas volcanic sequence and are mainly restricted in occurrence to the Fächerberg trigonometrical beacon area. Large and irregular, randomly-oriented vesicles, completely or only partly filled with quartz, are subordinate to lenticular quartz-filled amygdales, which are flattened and streaked out within the flow laminae.

(ii) Domes and Plugs

Distribution and Field Character

Numerous rhyolitic intrusive bodies are associated with all three of the Guperas rhyolitic volcanic fields. Twenty-eight such intrusions were located within the Kronenberg-Auramberg field, fifteen in the Blutpütz West field and four are associated with the Rooiberg occurrence. Many of these, the larger in particular, are undoubtedly of a composite nature, and a more detailed field study would certainly reveal a greater number of individual intrusions.

The distribution of the bodies indicates their close association with the extrusives of similar composition, appearance and age (Fig. 7). All but one of the intrusives of the Kronenberg-Auramberg occurrence have been intruded into the rhyolite lava pile, and the exception intrudes Barby basic lava a mere 1 km from the closest outcrop of acid lava.

Volumetrically, the vast majority of rhyolite intrusives of the Blutpütz West field occur within an area corresponding very closely to that covered by the rhyolite extrusives. A few small dike- and sill-like bodies intrude Guperas sedimentary strata to the southeast of the main volcanic field, in the vicinity of, and roughly paralleling, major north-northwesterly striking post-Auborus faulting. A deeper level of erosion and the more highly dissected nature of the Blutpütz West field as compared with the Auramberg-Kronenberg Field results in a relatively high percentage of intrusive material being exposed cutting the Guperas sediments rather than the overlying rhyolite lava. By comparison with the above two fields the Rooiberg occurrence has remarkably few associated acid intrusives, and those present are volumetrically insignificant.

Basic inclusions, usually not exceeding 5,0 cm in size, are present in all the intrusive rhyolites but are not abundant, except in the case of a fairly large body on northwestern Blutpütz West, in which basic inclusions are present to the extent that large hand-specimen-sized samples always contain at least one exposed inclusion.

In general, the massive or fluidal textures of the rhyolites indicate fluid magma on intrusion. However, a significantly large portion of the Blutpütz Ost body is of fragmental nature as seen in hand-specimen, particularly on weathered surface, and in thin-section. Occasionally well-rounded pebbles have been included, which were derived from the conglomerate country-rock.

Petrography

The petrographic properties of the intrusive rhyolites are essentially similar to those of the rhyolite lavas. Phenocrysts of quartz and plagioclase and/or alkali felspar make up an average of 11 per cent (range 5 to 19 per cent) of the total rock volume, and are set in a groundmass composed of products of devitrification and vapour phase crystallisation of a micro- to cryptocrystalline nature. Flow-banding is weakly developed and appears in approximately 25 per cent of the occurrences.

Quartz phenocrysts, which are well-rounded and embayed due to resorption, and occasionally exhibit bipyramidal or hexagonal outlines, constitute up to 12 per cent of these rocks, with an average content of 6 per cent. It is clear that most crystals were euhedral prior to resorption which was apparently more pronounced here than in the quartz phenocrysts of the lavas discussed earlier. Glomeroporphyritic aggregates do occur but are not particularly common. Maximum size of the quartz phenocrysts is about 2 mm in diameter. Strain features were not observed.

Plagioclase phenocrysts have a compositional range of An_0-An_8, are usually subhedral to less commonly euhedral in outline, and never constitute more than 5 per cent by volume of the rock. Resorption effects are notable, but are generally not as advanced as those affecting the quartz, although very highly corroded crystals do occasionally occur. Twinning is ubiquitous and according to the albite, carlsbad and pericline laws and glomeroporphyritic aggregates are common. Crystals are rather stumpy in character and measure up to 3,0 mm in length. Sericitisation is incipient and a reddish-brown cloudiness is characteristic.

Alkali felspar phenocrysts make up a maximum of 12 per cent of the rock and is a cloudy orthoclase perthite. Resorption effects are evident in rounded outlines and replacement by clear 'chess-board' albite is well advanced in some specimens. Granular quartz as inclusions within the phenocrysts is secondary and evidently accompanies the albitisation. Crystals measure between 0,2 and 2,8 mm in diameter.

Rare aggregates of chlorite and opaque ore minerals represent the altered and redistributed remains of small ferromagnesian phenocrysts. Even more rare are small (maximum 0,3 mm in diameter), well-rounded and resorbed phenocrysts of an opaque ore mineral.

Crystallisation of the groundmass has resulted in a generally coarser-grained character than that found in the lavas. Plumose and spherular growths of quartzo-felspathic material measuring about 0,3 mm in diameter predominate. Less common are larger and better-defined spherulites up to 3,5 mm in diameter. Granular devitrification products are also present, but are less frequent than the fibrous varieties. Very fine granules of opaque ore minerals are a ubiquitous groundmass constituent.

The development of 'pegmatite' patches in the groundmass of these rocks, due to vapour phase crystallisation, is a frequent characteristic. Specimen BW 578 displays a remarkable development of such pegmatite patches, each consisting of a relatively coarse-grained marginal zone of predominantly alkali felspar, and a thicker 'core' zone of quartz. In the case of the lavas, the shape and distribution of these pegmatitic aggregates is controlled by flow, and they tend to form streaks, bands or lenticles, whereas in the intrusive rhyolites, where flow structure is absent or only weakly developed, the pegmatites have a random and patchy distribution.

Basic rock fragments are always present as inclusions but in extremely variable, and usually low, concentrations.

The above description is applicable to all but three of the intrusive rhyolite occurrences examined in the field, and in thin-section. The exceptions are:

(a) A small plug on northern Aruab (e.g. BW 1552). This body consists of a greenish-grey porphyritic rhyolite which is comparatively rich in phenocrysts (total 29 per cent) composed of 17 per cent quartz and 12 per cent plagioclase of composition An_8. Maximum size of both phenocryst minerals is about 2,5 mm. Alkali felspar is absent as a phenocryst mineral. The groundmass is crowded with epidote granules and chlorite. Some of the quartz phenocrysts appear alien, in that they consist of angular aggregates of highly strained crystals, whereas most of the quartz phenocrysts are typically rounded by corrosion and totally unstrained.

The above properties and the chemical composition (Section 3.3.3) suggest that this rock may have been derived by gross contamination of a basic magma by siliceous rock material or magma.

(b) The major parts of the Blutpütz Ost intrusion, and the smaller plug in the vicinity of the Naudaus trigonometrical beacon, consist of fragmented rhyolitic material. Although not always immediately apparent in hand-specimen, the fragmental texture is unmistakeable in thin-section. Phenocrysts of quartz, plagioclase (An_0 to An_{10}), perthitic alkali felspar and highly angular fragments of rhyolite appear in a matrix of very fine-grained granular and fibrous devitrification products. The rock fragments which make up as much as about 75 per cent of the rock and range in size from < 0,1 mm to about 1,0 cm, are themselves devitrified, the resulting quartzo-felspathic intergrowth having a fabric direction peculiar to each fragment, a property which greatly facilitates recognition of the fragments as such. The fragments are invariably a darker reddish-brown colour than the groundmass, and they also contain a relatively high proportion of ore grains.

Alkali felspar phenocrysts measure up to 3,0 mm in diameter and are smoothly corroded remnants of once euhedral crystals (e.g. BW 1509). Replacement by 'chess-board' albite and minor quartz is common.

Primary plagioclase phenocrysts are totally absent from certain parts of the Blutpütz Ost body (e.g. BW 1509),whilst in other parts (e.g. BW 1507) they form the dominant phenocrystic felspar. The maximum size is 1,2 mm and the nature is obvious from the outlines of the grains.

Quartz phenocrysts vary considerably in their mode of occurrence. In specimen BW 1509 from Blutpütz Ost they appear as well-formed originally euhedral, resorbed and embayed insets, with only minimal evidence of fragmentation. The size range in this case is from less than 0,1 to 2,0 mm. In samples BW 1507 from Blutpütz Ost and BW 139 from the Naudaus mountain plug, however, the quartz occurs as highly angular chips with only very occasional evidence of corrosion, and the crystals fragments are relatively small ($\sim$ 0,3 mm average diameter).

(iii) Volcaniclastic Rocks

(a) Ash-flow Tuff (Ignimbrite)

Distribution and Field Character

The terminology applied in the following descriptions is in accordance with that suggested by Ross and Smith (1961), and Smith (1960) in his classic paper on ash-flows. Ross and Smith (1961, p.3) state that "the consolidated deposits of volcanic ash resulting from an ash-flow are called ash-flow tuff. Ash-flow is here used as an adjective to indicate the mechanism of dispersal, and tuff indicates the state and size of the material. Ash-flow tuff is an inclusive, general term for consolidated ash-flow beds that may or may not be either completely or partly welded". With regard to size of material, Smith (1960, p.80) defines the basic unit, the ash-flow, as consisting of "50 or more weight percent of ash and fine ash exclusive of foreign inclusions. Ash ($\frac{1}{4}$ to 4 mm) and fine ash (less than $\frac{1}{4}$ mm) refer to the Wentworth and Williams (1932, p.45) size scale".

Welded ash-flow tuff (Ross and Smith, 1961, p.800) or ignimbrite (Marshall, 1935, p. 4-10) occurring within the Guperas rhyolite succession are volumetrically most prominent in the eastern and southeastern part of Blutpütz West field.

In this locality, three successive 'sheets' (Smith, 1960, p.800) occur at the base of the volcanic sequence, conformably overlying the Guperas sandstone and conglomerate.

The distribution is such that the sheets comprise a distinct wedge along the Aruab-Guperas boundary, thinning to the south where they are overlain by Guperas sedimentary rocks. Moderate dips of 20° - 25° in the southern part of the exposure allows the wedge-like character to be readily observed in the field. When traced northward the ash-flow sheets thicken noticeably, a near horizontal attitude is adopted and the overlying clastic beds are replaced by rhyolitic lava flows. Where exposed beneath the lava on Blutpütz Ost, the ash-flow sheets are seen to terminate against a prominent northwest-striking fault.

On the basis of field and petrographic characteristics the three successive sheets can be regarded as separate cooling units (Smith, 1960, p.801), possibly of a multiple-flow nature, and each probably represents a distinct phase of emplacement. For convenience of description the units have been numbered 1 to 3 from the base upwards (Fig. 8).

Contact relationships are such that the lower contact of flow 1, with the sandstones and conglomerates, is sharp, whereas the upper contact with the base of flow 2 is transitional over a thickness of 1-2 metres. No appreciable time lapse occurred between the two eruptions, and the transitional nature of the contact is attributed to 'mixing' of the adjacent non-welded zones of the two ash-flows. A discontinuous band of sedimentary material is interbedded between flow units 2 and 3 and may reach a thickness of 30 m. The nature of the sediment also varies considerably due to variable proportions of normal clastic and pyroclastic material. The upper and presumably non-welded parts of flow unit 2 were reworked and stratified with the incorporation of normal clastic debris. This clastic material increases with increasing thickness of the unit, producing tuffaceous sandstones and finally, but rarely, normal Guperas lithic sandstones. Coarse-grained beds are developed to a lesser extent, and range from agglomerates to normal petromict conglomerates. The former were possibly formed as a coarse residue by the winnowing of non-welded ash.

Flow unit 3 lies with a sharp contact on unit 2 and the sedimentary lenses described above, but the upper contact of unit 3 is transitional through stratified tuff and tuffaceous sandstone into normal Guperas lithic sandstone and conglomerate. The transition can once again be attributed to reworking, winnowing, and stratification of non-welded tuff, with the incorporation of increasingly larger amounts of epiclastic material upward in the sequence.

The maximum thickness attained by these units varies between approximately 300 m for unit 3 and 350 m for units 1 and 2. The flows can be traced southwards along strike for 6 km in the case of unit 1, and for 12 km in the case of flow units 2 and 3. In a westerly direction only unit 1 can be traced at the base of the rhyolite sequence, thinning rapidly from the northern Guperas area and pinching out within a distance of 5 km. On Blutpütz Ost the ash-flow units are terminated by a prominent fault, north of which the ash-flow tuffs do not appear at the base of the rhyolite succession.

The changes in thickness and distribution of the flows point to a source area in the vicinity of central Blutpütz Ost, i.e. immediately northeast of the terminating fault. In this area a large elongate, and partly fragmental, rhyolitic intrusive, is present (Fig. 8) which may have been the main source of the ash-flows (see Section 3.2).

Calculations based on recorded thicknesses and present distribution indicate a total volume in the order of 75 km^3 for the ash-flows, but since the easterly extent is completely unknown this figure must be a minimum. Assuming that the intrusive body on Blutpütz Ost represents a source of the flows, and that the flows are distributed symmetrically along its exposed length, the above figure can be increased to approximately 580 km^3.

However, considering the highly mobile nature of ash-flows (Smith, 1960), it is probable that much of the southward and westward thinning observed in these flow units is due to the effect of an irregular topography prevalent at the time of emplacement. Certainly it would appear that flow unit 1 represents the infilling of a pre-existing topographic low, since its terminations are more abrupt than would normally be expected, and the 'thickness to distance-travelled' ratio is greater than in the case of the overlying units. The latter must have flowed over a more nearly horizontal and less irregular surface, and would consequently have travelled further. According to Cook (1966, p.164) the "variation in thickness of an ignimbrite is controlled more by irregularities of the surface on which it was deposited than by any depositional thinning, away from the source of eruption".

Welded ash-flow tuff of the Guperas Formation can, in all instances, be distinguished from true rhyolitic lava flows of the same sequence by virtue of its fragmental and obviously pyroclastic nature. Rock fragments are a ubiquitous constituent and may be of considerable size (maximum 5 cm). Collapsed pumice fragments are also always present, but never in excessive amounts, and appear as dark-grey compressed lenticles parallel to the base of the units, resulting in a very poorly developed eutaxitic texture. Lithic fragments are never deformed. The tuff is rich in crystal fragments, which consist almost exclusively of quartz and felspar.

The three flow units as discussed in the preceding paragraphs may be recognised in the field using the above criteria, but distinct differences in some gross characteristics are obvious and also allow differentiation between the units to be made in the field.

At its thick northern limit, flow unit 1 is not particularly rich in rock fragments, and extreme welding has resulted in a rock that appears very much like the true rhyolitic lava further north. A conspicuous foliation is considered to be due to compaction rather than flow, and constitutes a fairly well-defined 'eutaxitic' texture which is further enhanced by light red and relatively coarse-grained lenticular streaks in the dark greyish-brown groundmass (Plate 23). These streaks probably formed as a result of vapour-phase crystallisation and this is supported by the fact that central cavities frequently appear within these more coarsely crystalline streaks. On fresh surface the rock is virtually indistinguishable from some of the typical Guperas rhyolite lavas, and were it not for the presence of rock fragments and lack of true flow texture the original nature of the deposit might be in dispute.

It is for this reason that the possibility of some of the rhyolite lava flows being highly-welded and recrystallised ash-flow tuffs is not entirely

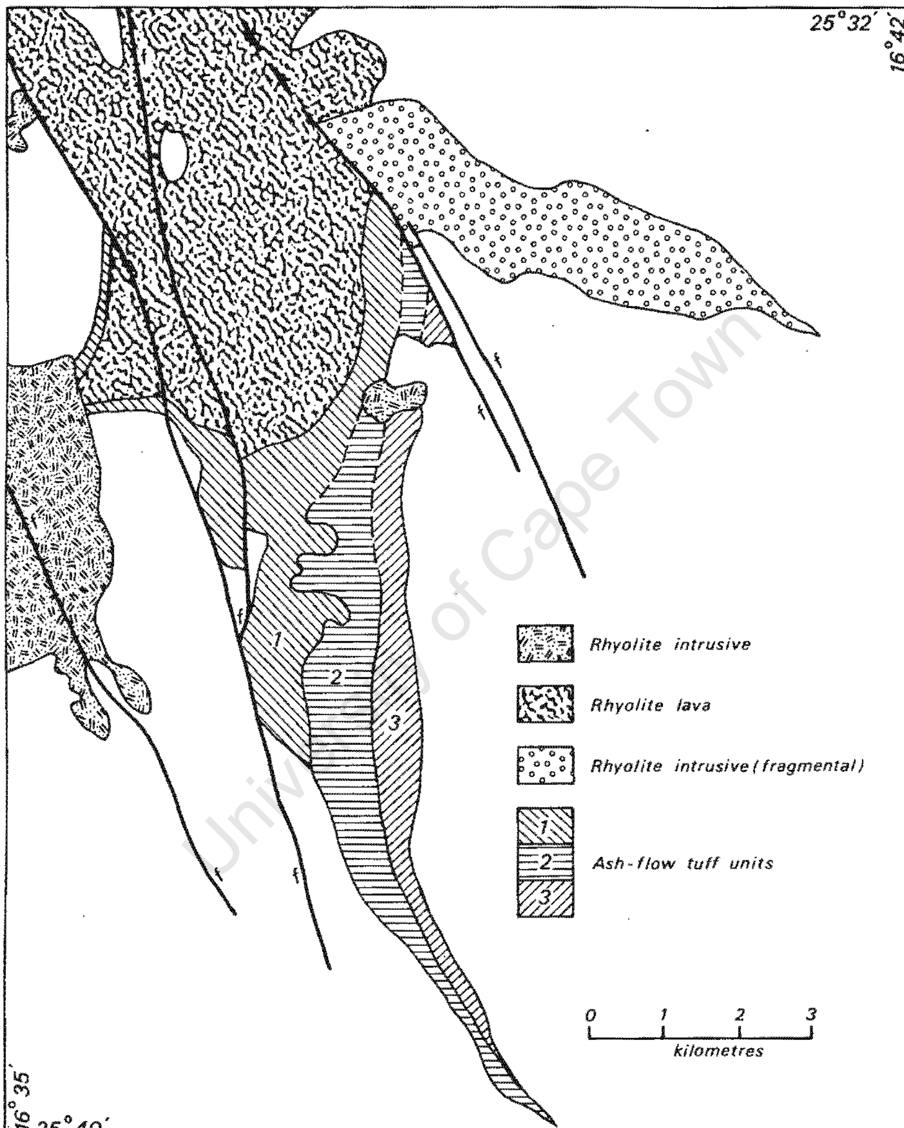


Fig. 8. Geological map showing the distribution of the major ash-flow tuff units of the Guperas Formation and their possible source vent

ruled out. On weathered surface, which is a brilliant yellow streaked with red (the coarser-grained areas), the fragmental character of the rock can be recognised and lithic fragments, which invariably constitute red rhyolitic material, become conspicuous. The latter are noticeably well-rounded (Plate 23). The coarser-grained streaks are more resistant to weathering than the rest of the rock, and frequently stand out with considerable relief on weathered surface. Traced toward its extremities, especially on northeastern Aruab, the fragmental nature becomes more easily identifiable, lithic fragments are more angular, increasing slightly in amount, and the rock is unmistakeable as an ash-flow tuff.

Unit 2 contains a relatively high proportion of lithic fragments, consisting mainly of felsic volcanic material, but basic volcanic and clastic fragments are also frequently included. This unit is remarkably uniform in appearance and consists of a dark grey to black rock, with variously-coloured felsic volcanic inclusions (red, green and grey mainly). Welding has produced a very hard compact character. The weathered rock is light brown in colour. Southward, toward its distal margins welding is less well developed and the grey variety grades into a more easily weathered reddish-grey coloured rock.

Flow unit 3 is a pale greenish-grey coloured rock. The lithic fragment content falls between the two extremes of units 1 and 2,and the grain size of fragments is generally much finer in the other units. Dark-grey lenticles of collapsed pumice, although not very abundant, are easily recognisable in the lighter-coloured matrix. As with unit 2, the rock takes on a reddish hue toward its apparently less highly-welded distal margins.

With the exception of the upper reworked parts of flow units 2 and 3, no sorting and/or bedding was found within any particular vertical section. In flow unit 2 there is a barely discernible reduction in rock fragment content and size away from the presumed source toward the distal end. Such a mild sorting is attributed to the natural ability of finer more easily buoyed particles to be transported a further distance than the larger ones. In flow unit 1 the reverse situation is present, i.e. the lithic fragment content increases slightly toward the distal zones. Since the more lithic part of the flow must represent the initial outburst, it is perhaps not unreasonable to assume that it owes its higher rock fragment content to the 'clearing' or fragmentation of chilled vent or fissure precursor rhyolite, followed by a magma less contaminated by foreign particles.

Other occurrences of ash-flow tuff in the Guperas rhyolite succession are:

(i) At the base of the lava sequence below the Auramberg trigonometrical beacon, and further to the southeast underlying the thick locally developed basic lava member (Section 2.2.7). The two exposures occupy similar stratigraphic positions and are almost certainly part of an extensive flow with a minimum lateral distribution of 12 km. The thickness is fairly constant at 50 m. The rock colour ranges from a deep red near Auramberg to a predominantly paler red and lilac east of Hahnenkamm. Lithic fragments and crystals are abundant, and in sections east of Hahnenkamm beacon a definite layered character is notable. The basal 1-2 metres of the flow are crowded with rock fragments, many of which consist of basic lava derived from the underlying Barby Formation. Upward in the flow the rock fragment content disappears completely, and there is a corresponding increase in the degree of welding. This layering is essentially a sorting effect, the larger and heavier fragments concentrating at the base of the flow, but the effect of foreign particles picked up from the underlying surface during flow was probably also not insignificant. Compaction has resulted in the development of a laminar parting.

(ii) Intercalated within the rhyolite lava flows 4 km east of the Fächerberg beacon is a band of welded tuff which is exceptionally rich in rock fragments of felsic composition. Lack of exposure and faulting prevented the determination of the lateral extent of this bed but its thickness is in the order of 60 m. Lithic fragments are red in colour and are set in a dark-maroon groundmass. Fragment/groundmass boundaries are usually sharp in the case of the larger fragments, which may measure up to 5 cm in size, but in the case of the smaller fragments they are frequently of a diffuse nature.

Petrography

The pyroclastic nature of these rocks is very well displayed in thin-section. Abundant crystals, crystal and lithic fragments and, less commonly, pumice fragments are set in a groundmass of welded, compacted glass shards, fine-grained crystal fragments and dust (Plate 24 and 25).

Glass shards display typical cusp-, lune-, tricupidate and Y-shaped outlines, which have generally not been destroyed by welding or devitrification (e.g. BW 1523). Crystallisation of the glass has taken place on an almost submicroscopic scale, preserving the outline of individual shards. As a result of this fine-grained and fibrous devitrification, axiolitic structures are commonly developed within the confines of individual shards (Plate 25) (e.g. BW 1522). The size of the shards varies considerably, ranging from less than 0,05 mm to 0,7 mm, and their colour is yellowish- to reddish-brown.

Welding and compaction have been moderate to severe resulting in a coherent structure, with the shards displaying marked distortion and stretching, particularly in the vicinity of phenocrysts (Plate 27) and inclusions and this simulates to some extent, the flow structure of rhyolite lava (e.g. BW 1523 and BW 1549). Evidence of plasticity of the glass during welding and distortion by compaction is well illustrated by the manner in which deep resorption embayments in quartz phenocrysts have been penetrated by glassy material (e.g. BW 1549).

Extreme welding and granular devitrification in the northern parts of flow unit 1 (BW 1542) have resulted in almost total eradication of the vitroclastic character (Plate 28). The development of bands of granular devitrification products with varying grain size and pegmatitic streaks due to vapour-phase crystallisation in these rocks, has produced a 'eutaxitic' fabric not unlike the flow texture observed in some of the lavas, but which is caused by compaction rather than flow. The eutaxitic character is further enhanced by varying concentrations of iron oxides within the compaction/crystallisation banding. Further evidence for a vapour phase in these eutaxitic varieties can be found in small (<1,5 cm) highly irregular vesicles.

Flow unit 3 also displays a rather ill-defined fabric, most apparent in hand-specimen and due to dark streaks representative of flattened pumice fragments. The writer prefers, however, to restrict the application of the term 'eutaxitic' to the northern parts of flow unit 1 where the texture best fits the original definition of the term. Fritsch and Reiss (1868, p.414) proposed the name eutaxite for a volcanic rock composed of ejected fragments of different colours, and with a texture as follows: "The different fractions in general lie beside one another as streaks, bands and lenses in seemingly well-ordered distribution".

The textures observed in the eutaxitic welded tuffs of unit 1 are comparable to those of the eutaxitic ignimbrites of the older Nagatis Formation, described by von Brunn (1967) and occurring in the adjacent area to the south. As with the Guperas eutaxitic volcaniclastics, the Nagatis eutaxites are subordinate to normal ignimbrites (von Brunn, 1967). With the exception of the eutaxitic variety, welded tuffs of the Guperas Formation are notably rich in phenocrysts and crystal fragments.

Quartz is the dominant phenocryst mineral, constituting between 4 and 26 per cent of the total rock volume. The majority of the quartz fraction is comprised of highly fragmented and angular grains from less than 0,1 to 3,0 mm. The larger phenocrysts may be fairly sharply euhedral, exhibiting a high temperature hexagonal symmetry, but usually they have some degree of rounding and embayment. Many of the smaller fragmented crystals also have 'smoothed' resorption features. The fragmented nature of quartz (Plate 25) appears to be an important diagnostic feature of flows with pyroclastic affinities within the Guperas sequence, and is in direct contrast to the generally more rounded or euhedral phenocrysts of the lava flows. Furthermore, true glomeroporphyritic aggregates of quartz phenocrysts were not observed in the welded tuffs, whereas such textural phenomena are common in the lavas.

Plagioclase felspar phenocrysts make up a maximum of 12 per cent of the total rock volume and plagioclase is the dominant felspar present as an inset mineral. Unzoned, and generally not as highly fragmented as the quartz, the crystals range in size from about 0,1 to 2,5 mm and are subhedral to rarely euhedral in form. A fairly advanced degree of resorption is evident in most specimens, a feature that was not prominent in the lavas. The composition is in the range An_0-An_{12} and faintly developed twinning is according to the albite, carlsbad and pericline laws. Mild alteration to sericite is ubiquitous.

Alkali felspar is a relatively rare phenocryst constituent, and usually forms rounded and resorbed crystals of non-perthitic and untwinned character. Incipient albitisation is also evident and the possibility that much of the plagioclase was originally alkali felspar, is not entirely ruled out, although the absence of 'chess-board' albite, places some doubt on this interpretation.

No trace of ferromagnesian minerals or their alteration products was found in the tuffs, but small and sparse opaque ore grains, occasionally reaching phenocryst dimensions, are always present.

Pumice fragments are ubiquitous but generally sparse. Flow unit 3 has the highest concentration of pumiceous material (maximum of 9 per cent by volume). In hand-specimen pumice appears as dark grey streaks flattened in the plane of compaction and measuring up to 2,5 cm in 'length' and less than 3 mm in thickness.

In thin-section, pumice has a reddish-brown colour and a pronounced collapse structure (Plate 24). Commonly, the fragments have a fibrous appearance parallel to the plane of compaction (Plate 29) due to the flattening and stretching of originally spherical vesicles. Fragment ends typically have a ragged or frayed appearance (Plate 24). It is also evident that much of the pumice was porphyritic (Plate 29). Compression has caused the fibrous fabric of the pumice to wrap around the relatively rigid phenocrysts, which are always quartz, and in some cases has resulted in crystals projecting beyond the general outline of the fragment. Invariably, however, a thin layer of streaky pumice still remains enveloping the phenocryst in such cases, thus providing evidence that it is an original inset of the fragment and not of the groundmass in general. Phenocrysts of this mode of occurrence are unfragmented and bounded by smoothly resorbed subhedral to euhedral outlines. Pumice fragments are frequently compressed against, and bent about, normal quartz phenocrysts (as well as other phenocrysts and rock fragments) (Plate 24), but these can be distinguished from the above by the lack of a thin enveloping pumice layer and a generally more angular nature.

In general, the welded tuffs are rich in lithic fragments which, in flow units 2 and 3, constitute up to 44 per cent by volume. Flow unit 1 has a low but variable rock fragment content, ranging from less than 1 per cent in the north to about 3 per cent in the southern and western exposed distal portions.

The welded tuff occurring 4 km east of Fächerberg is exceptionally rich in lithic material which may constitute 50 per cent of the whole rock. The deposit occurring on Ganaams, 2 km east of Hahnenkamm trigonometrical beacon, has a rock fragment content ranging from about 15 per cent at the base to virtually nil in the upper parts of the unit.

Compositionally, the fragments comprise the following main types:

(i) Devitrified felsic volcanic material.

(ii) Divitrified welded tuff.

(iii) Sedimentary rock, usually quartzite.

(iv) Basic material, frequently porphyritic and presumably of volcanic or subvolcanic nature.

The felsic volcanic rock fragments are by far the most common and, together with the sedimentary fragments, usually constitute the largest, measuring up to 5 cm in diameter.

Alien fragments are invariably angular but reasonably equidimensional. The felsic rock fragments of the highly welded eutaxitic tuff (of flow unit 1) have a distinctly rounded character, and appear to have suffered a certain amount of magmatic corrosion, probably prior to vesiculation and extrusion.

The compositional characteristics of the various fragment types requires that they were derived from two major sources:

(i) Rocks through which the tuff-forming magma intruded on its way to the surface.

(ii) Material picked up from the surface by the overriding flow during emplacement.

As mentioned earlier in this Section, the upper, and presumably non-welded, parts of flow units 2 and 3 have been reworked by normal sedimentary processes and grade upward into lithic sandstones and conglomerates of the Guperas Formation. In thin-section the transition is characterised by the following properties, increasing in importance from (i) to (iii) with the degree of reworking:

(i) Lack of welding and compaction structure, but otherwise compositionally similar to the underlying welded zone. Foreign crystal and rock fragments not present in the welded zones, but common in Guperas sedimentary rocks, e.g. microcline, granophyre, and quartzite, and quartz displaying strongly strained extinction, are incorporated.

(ii) Marked reduction in the proportion of shards, although the presence of fine-grained material of low to absent birefringence in the groundmass possibly represents mechanically reduced glassy substance. Accompanying the reduction in shard content is a relative decrease in the matrix content, and a corresponding increase in the coarser, predominantly lithic fraction.

(iii) Fine-grained material that can be termed matrix eventually becomes virtually absent, and no shard structure is recognisable in the upper parts of these beds. Normal clastic material may constitute more than 30 per cent of the total volume and at this stage the rock is best described by the term volcanic sandstone (Pettijohn, Potter and Siever, 1972, p. 172), rather than tuff.

The above transition takes place over a stratigraphic thickness of a few metres, and essentially indicates the reworking of the upper non-welded zones of the tuffs, with the incorporation of normal clastic material and the progressive winnowing out of the finer glassy matrix upward in the sequence toward the lithic sandstones.

(b) Tuff, Lapilli-tuff and Tuffite

Distribution and Field Character

In the following description the classification of 'volcaniclastic' rocks of Fisher (1961 and 1966) has been employed. Grade size limits of this scheme are <2 mm for tuffs, >2mm <64 mm for lapillistone and >64 mm for agglomerate or pyroclastic breccia. A mixture term for grades less than 64 mm is lapilli-tuff (Fisher, 1966, p. 292).

Pyroclastic rocks of the non-flow or non-ignimbritic type are represented only scantily within the Guperas rhyolite volcanic sequence. The greatest development of volcaniclastics of this type occurs at the base of the rhyolite lava succession on the eastern flanks of the Rooiberg Mountain in the vicinity of the trigonometrical beacon, and 8 km further to the west in an identical stratigraphic position.

In these outcrops Guperas sandstone and conglomerate grade upward into lapilli-tuff (Fisher, 1966). The pure pyroclastic rocks consist entirely of angular fragments of red felsitic-porphyritic material and quartz and felspar crystal fragments. The pyroclastics are generally completely unsorted, bedding is virtually non-existent and only in rare instances are vague and inconsistent laminae of finer or coarser material present. Flat or elongate grains are randomly oriented.

Petrographic characteristics of, and differences between, the lapilli-tuffs of the western and eastern Rooiberg occurrences are presented in greater detail in the following subsection, but it can be mentioned that in the field the most obvious differences are the generally larger and more highly angular fragments and strong iron staining in a tuffaceous matrix of the western outcrop, as compared with that to the east.

Lithic fragments form the coarser and dominant component and impart a deep red colour to the rock, which weathers to a yellow-brown shade. Due to the remarkably uniform character of the rock fragments, the fragmentary nature of the rock is not always immediately obvious on fresh surface. On weathered surface, however, the clastic nature is unmistakeable.

Volcaniclastic rocks transitional between the pure pyroclastic and normal clastic end members have very variable modal compositions. Characteristically, the incorporation of angular fragmentary felsic volcanic material (lithic and crystal fragments) increases progressively upward until normal clastic material is totally excluded. A correspondingly poor development of bedding accompanies the transition, which takes place over a thickness of about 10-30 m. The total thickness of the lapilli-tuff, excluding the transitional beds, is approximately 80 m in the eastern outcrop and possibly as much as 150 m to the west of Rooiberg.

Very thin sporadic lapilli-tuff beds with a maximum thickness of 2 m, occur at the base of the ash-flow tuff sheet east of Hahnenkamm trigonometrical beacon. In most cases these lapilli-tuffs appear as local developments gradational downwards into Guperas conglomerate which has an epiclastic volcanic sandstone matrix. The volcaniclastics differ from the conglomerates by possessing virtually no clastic pebble fraction and having higher and dominant proportions of angular felsic volcanic rock and crystal fragments. A small proportion of basic volcanic lithic fragments, displaying some degree of rounding, persists from the underlying conglomerates, and it would seem that the lapilli-tuff is composed of a mixture of pyroclastic and epiclastic volcanic (mainly basic lava) material, the former being dominant. No signs of stratification are present.

On south-central Guperas, approximately 2 km southeast of the southern termination of the ash-flow tuff units, a highly ferruginous well-stratified tuffite horizon of 30 m maximum thickness is developed. Its lateral extent is about 1,7 km, but it wedges out toward the west and terminates against a large fault in the east. Stratigraphically, this tuff is interbedded with typical Guperas sedimentary strata, and is situated about 100 m below a level equivalent to the base of the lava/ash-flow tuff pile lying to the northwest. According to Fisher (1966, p. 296) the term tuffite can be applied to rocks composed of admixtures of pyroclastic fragments in epiclastic volcanic rocks. The definition as given by Gary *et al.* (1972) states that a tuffite is "a tuff containing both pyroclastic and detrital material, but predominantly pyroclastic", can be more suitably applied to the Guperas tuffites, since the normal clastic fraction is probably not 'epiclastic' in the sense of the term as used by Fisher (1966).

Petrography

Volcaniclastic beds of the Rooiberg mountain occurrences consist of highly angular fragments of devitrified and frequently porphyritic rhyolite, and fragmented quartz and felspar crystals set in a sparse matrix of finer-grained crystal chips, mainly quartz, and minor amounts of devitrified material (Plate 30).

Quartz and felspar crystals occurring as phenocrysts within the lithic fragments have rounded and embayed outlines, whereas those deposited as discrete grains are highly angular. The size of lithic fragments ranges from less than 0,01 mm to about 1,0 cm and that of the crystals from less than 0,1 to 2 mm, placing the rocks in the lapilli-tuff class of Fisher (1966). The eastern deposits are better sorted than those of the western occurrence and it is further evident that, although in both occurrences the grains are highly angular, those of the eastern beds have a considerably higher sphericity (e.g. BW 1593 and BW 1594) than those of the west (e.g. BW 1446). Alkali felspar is also more common in the latter. In addition to alkali felspar (perthitic), plagioclase of albitic composition is also an important crystal constituent. Opaque ore mineral grains are always present, and rare chlorite patches mark the presence of altered ferromagnesian minerals. Equally rare epidote grains were also noted. Hematite staining of the matrix or finer fraction may be conspicuous, particularly in the western outcrop.

The above description applies to the pure pyroclastic beds, and in view of the modal character and nature of the fragments the assignation of the term 'rhyolitic lapilli-tuff' is considered justifiable.

The thin volcaniclastic beds occurring immediately east of the Hahnenkamm trigonometrical beacon do not differ significantly from the Rooiberg 'rhyolitic lapilli-tuff' and, therefore, the above descriptions suffice, with the exceptions that the Hahnenkamm volcaniclastics have a small epiclastic fraction consisting of basic volcanic rock fragments and the maximum grain size is larger, attaining about 4 cm.

The ferruginous tuffite horizon on south-central Guperas is composed of angular fragments of quartz, altered felspar and basic felsic rock material set in a matrix of very fine-grained quartz and possibly devitrified glass. The maximum grain size is 1,0 mm. Shard-like outlines in the matrix are sparse and vaguely recognisable, since they are obscured to a large extent by a very strong ferruginous stain permeating the entire rock. Chloritic pseudomorphs after ferromagnesian minerals were frequently noted, as were detrital grains of ore. Calcite is a prominent secondary mineral. Bedding is present and emphasized by varying concentrations of ferruginous material.

These pyroclastic horizons described above have been included under the general heading of 'Guperas rhyolites' since they represent the initial manifestations of acid volcanic activity in this sequence, and because, with the exception of the tuffite horizon, they are composed predominantly (in some cases exclusively) of rhyolitic volcanic fragments and crystals.

3.1.6.5 Quartz-porphyry Dykes

Distribution and Field Character

All felsic dykes occurring in the Sinclair area can be related to the Guperas cycle of (felsic) igneous activity, and the general term 'quartz-porphyry' defined by Gary *et al.* (1972) as "a porphyritic extrusive or hyperbyssal rock containing phenocrysts of quartz and alkali felspar (usually orthoclase) in a microcrystalline or cryptocrystalline groundmass" can be applied here. These dyke rocks contain, almost without exception, phenocrysts of quartz and alkali felspar, and otherwise meet the textural requirements of the definition.

The dykes appear throughout the area mapped, with the exception of the Nubib mountain and Losberg mountain massifs. Isolated examples do occur in places, but generally the dykes constitute swarms often of very high intrusion density and only a representative proportion has been indicated on the accompanying geological map to prevent obliteration of other geological detail.

Four main areas of intrusion can be defined:

(i) Vergenoeg-Kumbis-Sonop area. Abundant quartz-porphyry dykes trend 35°- 45°, and constitute a very dense swarm. Their attitude is vertical or steeply dipping to the northwest. Individual dykes can be traced for up to 12 km along strike, and the rock units invaded belong to the Kumbis, Kunjas and Barby Formations and the Spes Bona Syenite. This swarm is composite in that the felsic dykes have been intruded parallel to basic dykes discussed in Section 3.1.6.3.

(ii) The area in the vicinity of the Tafelberg and Blutpütz-Naudaus trigonometrical beacons. This is an extremely dense swarm in which the quartz-porphyry dykes, together with basic dykes as discussed in Section 3.1.6.3, were intruded with such intensity that the country-rock has, in some localities, been totally obliterated. Dykes situated in the central part of the swarm have invaded Guperas clastic rocks, whereas in the outlying areas the country-rock is basic lava of the Barby Formation, Nubib granite and all the various units of the Guperas Formation rhyolite member.

(iii) On southern Blutpütz West, Aruab, Guperas and Wittmanshaar, quartz-porphyry dykes have intruded all members of the Guperas Formation, Barby lava and intrusives, and Nubib granite. This swarm, which could probably be regarded as the southward extension of the previous one, has a low intrusion density and dykes are distributed in small groups or as individuals throughout this area. The most significant difference between these dykes and those of the Tafelberg-Blutpütz-Naudaus swarm, other than density of intrusion, is the more variable trend and a generally lower degree of linearity. In the northern parts of this area the trend varies from 320° through 0° to 15° with a mean at about 345°. South of latitude 25°41'S the trend swings to 0°-50° with a mean at about 35°.

(iv) On the farm Vrede, and to a lesser extent on Betta and Gorab, quartz-porphyry dykes trending 350°- 15° have intruded Kumbis, Kunjas and Barby rocks and Nubib granite. Some of these dykes show a tendency to swell and coalesce, resulting in small intrusive bodies.

The width of the dykes is variable from less than 1 metre to about 60 m, and single dykes can be traced for up to 12 km along strike. The extremely hard and compact nature of the quartz-porphyry ensures that the dykes are resistant to weathering, and hence form conspicuous morphological features. Chill margins may stand out as parallel ridges.

The dyke material consists of phenocrysts of quartz, pink to red alkali felspar and white plagioclase, set in an aphanitic groundmass of strong red colour. In the coarser varieties small irregular green-coloured aggregates mark the presence of altered ferromagnesian minerals.

Basic inclusions are fairly common, particularly in the Vergenoeg-Kumbis-Sonop dykes, and these have caused varying degrees of hybridisation resulting in a greenish-grey coloured rock.

Von Brunn (1967, p. 26-28) distinguished three varieties of 'late acid' dykes; "quartz-porphyry", "hybrid granophyre" and "granite-porphyry" the last-mentioned frequently being composite with a basic component. The granite-porphyry of von Brunn (1967) differs from his quartz-porphyry by virtue of being generally coarser grained and possessing a higher modal proportion of phenocrysts. Furthermore, the two varieties are located in geographically distinct areas.

In the area under discussion, dykes fitting the description of von Brunn's (1967) hybrid granophyre and granite-porphyry are only poorly represented. They are not confined to any particular locality, but appear either as isolated occurrences or as minor constituents of the swarms described above, in which they are much subordinate to the more typical quartz-porphyry dykes. A strict distinction between the three types of dykes is made even more difficult by the presence of transitional varieties. For these reasons all felsic dykes have been included under the general heading of 'quartz-porphyry' and where appropriate, differences will be noted in the following petrographic descriptions.

Petrography

Characteristically, these rocks are porphyritic with phenocrysts of quartz, alkali felspar and plagioclase felspar set in a crypto- to microcrystalline groundmass (Plate 31). The total phenocryst material present in the rock increases with grain size.

Flow-banding was not observed either in thin-section or hand-specimen, and in many cases the quartz-porphyry is indistinguishable in hand-specimen from much of the non-flow-banded rhyolite lava of the Guperas Formation. In thin-section, however, the two can usually, but not always, be distinguished by the coarser crystallinity of the dyke rock groundmass.

Basic inclusions are a common constituent and range in size from 0,6 mm to 8 cm.

Quartz forms well-rounded and embayed phenocrysts measuring up to 4,5 mm in diameter and constituting a maximum of 11 per cent of the rock. Most crystals show evidence of relict euhedral shapes, with high-temperature bipyramidal hexagonal symmetry. Glomeroporphyritic aggregates measuring up to 5,0 mm in diameter are also present. In the coarser-grained varieties quartz is always subordinate in amount to, and smaller in size than, felspar as a phenocryst mineral, whereas in the relatively finer-grained types this property is variable. Extinction may be undulose, but this is never a well-developed feature.

Plagioclase felspar is the dominant phenocryst mineral in the majority of these dykes, making up a maximum of 21 per cent of the total rock volume. The compostion is in the range An_7 to An_{15}. Three modes of occurrence have been noted:

(i) Primary insets which measure up to 1,0 cm in length, and generally have a stumpy and slightly rounded subhedral form. Glomeroporphyritic aggregates are very common, and slight resorption of the clusters as units indicates an early magmatic aggregation. Twinning is according to the albite, carlsbad and, to a lesser extent, the pericline laws. Alteration to sericite and epidote minerals varies from slight to extreme, and often inhibits determination of the composition. In the coarser-grained porphyries, the plagioclase crystal clusters often include small and completely altered phenocrysts of ferromagnesian minerals. Rarely, small subhedral crystals of plagioclase appear as inclusions in alkali felspar phenocrysts. The plagioclase always appears cloudy in thin-section and is distinctively white in hand-specimen.

(ii) Perthitic exsolution lamellae within the alkali felspar phenocrysts. These have the form of small patches of irregular shape and distribution throughout the host crystal, and when sufficiently large they display polysynthetic twinning.

(iii) Replacement features in alkali felspar phenocrysts (Plates 31 and 32). These replacement or 'albitisation' phenomena were also noted in many of the rhyolites, but are particularly well developed and displayed in the dyke rocks, especially the coarser varieties, where a rapakivi-like character has resulted due to mantling of perthite crystals by albitic plagioclase. The mantles do not always have a symmetrical distribution about the perthite phenocrysts, and have irregular inner margins. Glomeroporphyritic clusters of alkali felspar crystals frequently display mantling by albite, which has an optical continuity independent of the various crystal boundaries of the aggregate components. Indications are, therefore, that the albite is replacing the alkali felspar, transforming it into a 'pseudo-' plagioclase phenocryst. The replacement process is invariably accompanied by the introduction of small amounts of granular quartz as inclusions within the crystal and this is another factor pointing to a late- or post-magmatic origin for these replacement phenomena. All stages of transformation from virtually non-existent to complete can be observed in these rocks, and even in any one particular specimen the degree of replacement varies considerably.

The process generally proceeds inward from the margins of a perthite host crystal (Plate 32) but replacement cores have also been noted. These may be ovoid-shaped or may form highly irregular patches apparently developed from the extended growth and coalescence of exsolution lamellae. The albite cores discussed here are a distinctly different feature to the albite crystal inclusions mentioned in (i) above. Replacement mantles or patches usually form optically continuous units but may also consist of numerous independent segments. Characteristically, the replacement albite has polysynthetic twinning, developed in two directions and resulting in a typical 'chess-board' pattern.

Alkali felspar phenocrysts (0 to 12 per cent by volume) are orthoclase perthite which, as discussed above, displays varying degrees of replacement by albitic plagioclase. The crystals are large with respect to quartz and plagioclase, and attain dimensions of up to 2 cm length in the coarsest rock types, i.e. "granite-prophyry" of von Brunn (1967, pp. 36-38). Original euhedral to subhedral outlines have been smoothed by resorption in many instances. Glomeroporphyritic aggregates do occur but are not as common as those involving plagioclase phenocrysts. Carlsbad twinning is occasionally developed.

Alkali felspar phenocrysts (up to 1,6 cm in length) are pale to deep red in colour, contributing to the overall red hue of most of the dyke rocks. White albite rims contrast strongly with the red-coloured perthite host crystals and this is particularly well displayed in hand-specimens of the coarser-grained quartz porphyries (e.g. BW 935 and BW 516) (Plate 31).

Evidence of relatively small ferromagnesian phenocrysts can occasionally be found in partly redistributed patches of chlorite and opaque iron oxides. Very rarely, opaque grains of ore minerals form small rounded phenocrysts.

Quartz and potash felspar predominate in the groundmass and textural relationships can be described as follows:

(i) Fine-grained intergranular in which small, cloudy and anhedral potash felspar crystals with maximum dimensions of 0,4 mm form a framework, with clear quartz filling the interstices.

(ii) Microcrystalline granular or felsitic texture is a common variety, which may constitute the entire groundmass or, more frequently, may occupy interspherulite spaces.

(iii) Radiating or plumose fibrous intergrowths forming spherulitic structures often large enough to be discernible with the naked eye. Their size ranges from <0,2 to 4,0 mm, and the larger structures frequently have phenocryst nuclei. If the nucleus crystal is larger than the intergrowth, a corona-like structure results (e.g. BW 516).

The nature of the spherulitic structures ranges from small ill-defined and patchy forms, through well-formed fibrous and 'feathery' growths of variable size, to the relatively coarse-grained granophyric intergrowths discussed below.

Except in the case of the smaller and poorly-defined types, spherulitic structures rarely constitute the entire groundmass and granular or fine granophyric-textured material usually makes up a fair proportion of the groundmass.

(iv) Granophyric intergrowths between quartz and potash felspar, which may be gradational from the coarser spherulitic varieties but retaining a somewhat radiating character. These intergrowths are most commonly confined to the interspherulitic areas.

All of the above textures may be slightly modified by redistribution of quartz. Alkali felspar of the groundmass is always cloudy and reddish-brown in colour, hence the overall red colour of the rock.

Opaque ore mineral granules and trichites are ubiquitous accessory constituents of the groundmass. Irregularly-shaped grains or masses of opaque minerals represent alteration products of pre-existing ferromagnesian minerals, and are usually associated with chloritic aggregates.

The coarser grained quartz-porphyries may contain flakes or shreds of green chloritised biotite, and green hornblende was also observed as a groundmass constituent.

Accessory minerals include zircon, apatite and leucoxene. Secondary minerals resulting from the alteration of plagioclase have already been mentioned. Calcite, and particularly epidote, may also be important secondary minerals. The great abundance of epidote in some of these rocks is attrituted to contamination by basic material.

3.1.7 ROOIBERG GRANITE

Distribution and Field Character

Three small plutons of this granite with a total outcrop area of only 2 km^2 occur on the farms Rooiberg and Rooiberg Süd.

Field relations to other units are as follows:

(i) In the southern outcrops, at the Rooiberg—Rooiberg-Süd boundary, the granite has intruded the uppermost clastic and volcaniclastic beds of the Guperas Formation, and is overlain by sedimentary beds of the Nama Group in the same outcrops.

(ii) In the northern occurrences, on Rooiberg, the granite has intruded fluidal-textured rhyolitic lava of the Guperas Formation, and is also overlain by both the Auborus Formation and Nama Group.

The Rooiberg granite is a red or grey, fine- to medium-grained, often mildly porphyritic rock, which is very similar to certain textural varieties of the Nubib granite. Its weathered colour is a pale reddish- or yellowish-brown. In the absence of the contact relationships discussed above it would, in most instances, be impossible to distinguish with any certainty, between the Rooiberg and Nubib granites. A vague foliation striking approximately northwest is present in the northern outcrops, and possibly indicates repeated younger movement along the Nam Shear Belt.

With the possible exception of some of the Guperas quartz-porphyry dykes, this granite constitutes the youngest igneous episode of the Sinclair magmatic event.

Petrography

Microscopic examination of samples from the various outcrops of this granite reveals variations in texture and mineralogy which are not obvious in the field. Two main 'primary' textures are present:

(i) Fine- to medium-grained hypidiomorphic granular.

(ii) Fine- to medium-grained porphyritic-granophyric. Most of the crystals have a somewhat resorbed outline, particularly those of quartz. Occasional large grains of alkali felspar and quartz lend a porphyritic character to the rock. Cuneiform and plumose-type granophyric intergrowth of alkali felspar and quartz is interstitial, and frequently reaches such proportions that the early-formed crystals appear to 'float' within the groundmass.

A protomylonitic texture has been impressed on the granite occurring in the northern outcrops on Rooiberg. Originally the rock probably had a hypidiomorphic granular character but, in its present state, it consists of relatively large, and highly strained crystals, or slightly lenticular crystal aggregates, separated by strained and granulated material. The directional element of this texture is not well developed and is only faintly apparent in hand-specimen.

The general mode of occurrence of the major mineral constituents of the Rooiberg granite is as follows:

Quartz - In the granitic textured varieties this mineral forms an interlocking mosaic with alkali felspar and plagioclase and the crystal form is anhedral. Quartz of the porphyritic-granophyric granites is of two generations; early-formed subhedral crystals which are usually resorbed and embayed to varying degrees, and a 'second generation' comprising part of the groundmass granophyric intergrowth. The size of the early-formed quartz crystals is very variable ranging from 0,2 to 4,5 mm, and it is the larger examples of this type that define the porphyritic nature of the rock. Frequently these crystals also act as nuclei for the later granophyre development. Quartz makes up an average of 35 per cent of the total rock volume.

Alkali felspar - This may be either orthoclase perthite or microcline perthite in the case of the undeformed occurrences, whereas in the sheared bodies microcline is prevalent. Perthite is of the patch type, and in some instances exsolution of the sodic component has been extensive, giving rise to large patches of albitic plagioclase (similar to those formed in the Nubib granite), often constituting up to 40 per cent of the total volume of the mineral. Incomplete albite rims are also present, and have the appearance of replacement rather than exsolution features. The replacement plagioclase, which has a composition of An_0-An_5, commonly displays a chess-board twin pattern, and is invariably in optical continuity with the perthitic plagioclase of the host alkali felspar.

The crystal form is generally subhedral and occasional twinning is according to the carlsbad law. Alkali felspar is the dominant phenocrystic mineral in the porphyritic type. The alkali felspars of the undeformed granite have a typical reddish-brown cloudiness which imparts a reddish-brown hue to the rock. Alkali felspar of the sheared granite is noticeably less cloudy and, as a result of this, the overall colour of the rock tends to be somewhat grey.

Alkali felspar of the porphyritic granite belongs to two generations, i.e. early-formed, relatively large (up to 6,5 mm) and often resorbed, subhedral to euhedral crystals of porphyritic or glomeroporphyritic nature, and later-formed intergrowths with quartz. Alkali felspar constitutes an average of 42 per cent of the rock.

Plagioclase felspar (20 per cent by volume) nearly always displays good crystal form and crystal size ranges from 0,5 to 3,0 mm. This mineral is also present as the patchy exsolution and/or replacement features discussed above. Commonly a plagioclase crystal may act as a nucleus for granophyric growth. The composition of plagioclase ranges from An_8 to An_{12}. Sericitisation is only slight.

Ferromagnesian and accessory minerals - The presence of original dark-minerals in the Rooiberg granite is apparent from rather scattered and minor aggregates of secondary minerals, principally chlorite, opaque ore and epidote. In rare instances the remnant outline and basal cleavage of original biotite, which was probably the only ferromagnesian mineral, can be recognised. Certainly, ferromagnesian minerals, or rather the alteration products thereof, form a very minor constituent of the rock (<3 per cent). All the opaque ore minerals are apparently secondary and derived from alteration of biotite. Zircon and apatite are accessories.

3.2 MODE OF EMPLACEMENT

3.2.1 GRANITES

The granites of the Sinclair Group display all the characteristics of typical high-level granites emplaced into the epizone as defined by Buddington (1959).

Following the suggestions of Buddington (1959), the characteristics of the Sinclair granites which allow them to be classified as high-level granites intrusive into the epizone, may be listed as follows:

(a) Wholly discordant nature in relation to the country-rock.

(b) Lack of any lineation or foliation with the exception of those granites involved in the Nam shear belt.

(c) Close genetic relationship to volcanic rocks. Although a direct connection between the granites and felsic intrusives of the Sinclair cannot be demonstrated in the field, the close temporal association between the two, repeated occurrences of both igneous phases throughout the Sinclair and the identical chemical characteristics (Section 3.3.3) are highly suggestive of a genetic relationship between the two (Section 3.4.4).

(d) Sharp contacts with country-rock and lacking or very limited associated contact metamorphism.

(e) Porphyritic texture.

(f) Occasional miarolitic texture.

(g) Granophyric texture.

(h) Absence of pegmatites. The pegmatitic segregations within the Nam Shear Belt (Section 3.5) are not primary features of the Nubib granite and cannot be considered as evidence against an epizonal emplacement.

(i) Minor aplitic phases.

(j) Minor amounts of contact breccia.

(k) Flow-banding, developed in a few of the quartz-felspar-porphyry intrusions that have been included in the general grouping of 'Nubib granite' (Section 3.1.4).

In discussing the mode of emplacement of Sinclair granites west of Helmeringhausen, von Brunn (1967) presents evidence for both forcible and passive magmatic intrusion. North of Sinclair there is very little evidence for forcible emplacement and only rarely, where the granite intrudes the sedimentary beds of the Kunjas Formation, is minor contorting of the country-rock in evidence. However, since the country-rock invaded by the granites is predominantly basic lava of a highly competent nature, deformation by an intruding granitic magma might not have been very easily achieved.

Field evidence rather points to a fracturing and piece-meal stoping process having been dominant during emplacement. The fracturing of country-rock and incorporation of blocks of all sizes into the granite magma is in evidence at nearly all of the granite/basic lava contacts and is particularly well displayed in parts of the Nubib mountains. Very large xenoliths of basic lava occur ubiquitously throughout the major outcrops of granite and many of these could well represent roof pendants.

In the Nubib mountains, where Nubib granite invades a sequence of basaltic and rhyolitic extrusives of the Barby Formation, only basic xenoliths are present in the granite. It is probable, therefore, that the total absence of rhyolitic xenoliths can be attributed to the remelting and incorporation as a liquid, of the felsic lithotypes, into the intruding granitic magma. The incorporation of this material would not be expected to significantly alter the bulk composition of the granite due to the very similar chemical compositions of the Barby rhyolites and Nubib granite (Section 3.3.3). It can be postulated that the small occurrence of grey granophyre on the eastern flanks of Losberg represents a 'pocket' of remelted or reconstituted dark grey to black felsic extrusive of the Barby and the apparently transitional contact relationships between this grey granophyre and the normal red-coloured Nubib granite could then be suitably accounted for.

A similar reconstitution of felsic volcanics has been noted by von Brunn (1967) in the Tumuab mountains with regard to the Nagatis rhyolitic extrusives and the intrusive Tumuab granite.

It would seem, therefore, that emplacement of the Sinclair granites, the Nubib granite in particular, took place mainly by a passive stoping mechanism in which large and small blocks of country-rock sank through the granitic magma, thereby creating space for the intruding magma. Country-rock with a relatively low melting point, i.e. rhyolite, could be remelted on incorporation into the granitic magma whilst the more refractory blocks of basic lava remained generally unaffected. Only in rare cases, such as in the area below the Eckberg beacon, is there any evidence of reaction between the invading granitic magma and basic xenoliths (Section 3.1.4). It is suggested that these xenoliths were incorporated in the melt at greater depths than was generally the case elsewhere, possibly in or approaching the mesozone (Buddington, 1959), where the temperature of the magma might have been sufficiently high to allow some assimilation to take place.

On a regional scale the intrusion of Nubib granite displays a strong structural control. As pointed out in Section 3.1.4, the major outcrops of this granite occur in a broad northwest-trending belt traversing the region between the Persia/Osis and Nubib/Nubib West areas, and paralleling the strong northwesterly grain of the Nam Shear Belt. This parallelism of trends and the development of the shear belt toward the end of the period of Nubib granite intrusion (Section 3.1.5) suggests that the shear belt is the surface or high-level expression of an ancient zone of fundamental crustal weakness that was utilised by the granitic magmas for ascent to epizonal levels in the crust. The presence of large-scale vertical faulting, trending northwestward, bordering the main zone of granite intrusion to the southwest (on the farms Vrede, Kameelhof, Betta and Naudaus), and the presence of probable pre-Sinclair basement rocks in the northern part of the Nam Shear Belt on Steinfeld (Section 3.1.5 and Fig. 6) is also suggestive of an *en bloc* subsidence of this zone of plutons. The cause of the subsidence was probably a gross mass deficiency at depth resulting from the transfer of large quantities of granitic magma from lower levels in the crust to the epizone.

The above discussion refers particularly to the emplacement of the Nubib granite. Outcrops of the younger Rooiberg granite are extremely limited but textural similarities with the Nubib granite (i.e. granophyric and porphyritic), its completely discordant contacts and intrusion into a pile of felsic volcanics also suggest emplacement into a high crustal level.

The Haremub granite also provides evidence for high-level magmatic intrusion; exposed contacts with country rock are sharp and basic xenoliths, invariably with sharp margins, are common.

3.2.2 ASH-FLOW TUFFS OR IGNIMBRITES

According to Cook (1966, p. 159), it is generally believed that ignimbrites have been formed " ... by deposition from a hot, rapidly expanding, turbulent, highly mobile, magmatic gas 'cloud' (*nuée ardente* or pyroclastic flow) which carries with it intratelluric crystals, liquid droplets of the exploding magma (and the resulting glass shards), as well as rock fragments torn from the walls of the vent or picked up from the ground surface ... ". The lower incandescent part of the flow constitutes that portion which may become welded.

The textural and distribution characteristics of the various ash-flow tuff units of the Sinclair (as described in Sections 3.1.2.1 and 3.1.6.4) suggest that they were emplaced by just such a mechanism as described by Cook (1966) and quoted in part above.

An extremely high degree of mobility for the ash-flow tuffs of the Sinclair is readily apparent from the distribution characteristics. The ash-flow tuff of the basal volcaniclastic member of the Barby Formation is relatively thin (<120m) but is remarkably persistent throughout the area (Section 3.1.2.1). A felsic unit of such a thickness and wide distribution requires extreme mobility for its constituent flows at the time of emplacement and this could only have been achieved by a fluidised solid-in-gas system. It has also been pointed out in Section 3.1.6.4 that the main ash-flow tuff units of the Guperas Formation extend far beyond the geographical limits of the rhyolite lavas and thin towards their extremities far less rapidly than do the lavas.

It can only be speculated as to the nature of the source vent that gave rise to the ash-flow tuffs of the Barby Formation. Although difficult to estimate accurately, the total volume of these deposits certainly exceeds 10 km^3 and could not, therefore, have erupted from volcanic domes or open craters as *nuées ardentes* (Smith, 1960). *Nuées ardentes* or Pelean-type deposits generally have volumes of less than 1 km^3 with a maximum of 10 km^3 (Smith, 1960) and cannot compare with the volumes that must have been erupted during various stages of development of the Barby Formation. In the absence of any positive evidence for large-scale collapse structures it can be suggested that the ash-flow tuffs of the Barby were erupted from fissures or, more likely, from a series of vents situated along deep fractures.

As discussed in Section 3.1.6.4 and shown in Fig. 8 the major ash-flow tuff units of the Guperas Formation, occurring along the Aruab-Guperas boundary, display a distinct thinning toward the south and west, away from a large intrusive rhyolite body separated from the tuff units by a north-northwesterly striking fault. The eastern limits of the ash-flow units are not known due to younger cover and faulting. The absence of any ash-flow tuff at the base of the felsic volcanic succession of the Guperas to the northwest of the above rhyolite intrusive and the thickening of the ash-flow units toward it, strongly suggest that the intrusive represented a source vent or series of source vents for the extrusion of the gas-charged pyroclastic material. Thin-section examination of samples from the rhyolite intrusive indicates that it is largely fragmental, not unlike much of the ash-flow tuff, thereby strongly supporting the above suggestion. The composite nature of the intrusive as well as an apparently elongate form further suggest that the source area consisted of a line of volcanic vents situated along an approximately northwest trending fissure zone. Estimated volumes for the ash-flow tuffs of the Guperas (Section 3.1.6.4) are consistent with their derivation from a multiple vent source (Smith, 1960).

3.2.3 RHYOLITE LAVA, DOMES, PLUGS AND QUARTZ-PORPHYRY DYKES

Outcrops of rhyolite lava units of the Barby Formation are generally not extensive and their more limited lateral extent is apparent when compared with associated basic lava units. The restricted distribution and highly flow-banded nature of the lavas of the 'Rhyolite Member' certainly indicate flow of an extremely viscous medium. A slightly more even but nevertheless restricted distribution of the high-Ca rhyolites of the Barby, together with a finer more regular flow-banding aligned parallel to 'bedding', possibly indicate a somewhat more mobile and less viscous magma. No possible source areas or specific vents could be located in the area under investigation.

A low degree of mobility for the rhyolite lavas of the Guperas Formation is very well displayed by the extremely limited lateral extent of individual flows, the generally high degree of syn-depositional flow folding and the abrupt terminations. The lava apparently flowed only for very short distances. The very close association of rhyolite plugs with rhyolite intrusives (Fig. 7) suggests that these intrusions, rather than the quartz-porphyry dykes, acted as feeders to the lava. The swarms of quartz-porphyry dykes do not coincide with the present-day outcrop of extrusive rhyolite, whereas the plugs are restricted almost exclusively to the outcrop areas of rhyolite extrusives. Since the present-day outcrop of rhyolite lava is not considered to be extensively different from the original distribution (Section 3.1.6.4) it is suggested that the rhyolite plugs represent fillings of the vents through which felsic magma was carried to the surface to be erupted as Guperas rhyolite lava. It can also be speculated that the large dome-like intrusion underlying the Ganaams beacon area represents a partially extrusive or near extrusive body since it intrudes the stratigraphically highest exposed lava flows. However, since the top of the lava sequence is not preserved it is impossible to determine whether this dome had an extrusive component or not.

There exists no direct obvious relationship between the various rhyolite occurrences of the Guperas as discussed in Section 3.1.6.4 and local structural features, with the exception of the following:

(i) Block subsidence of the lava pile along an arcuate normal fault system has occurred in the vicinity of the Fächerberg trigonometrical beacon. Scanty outcrops of older Barby and Kunjas rocks on Spes Bona allow an inferred extension of this fault system to be made in a northwest and northerly direction, paralleling the present trend of the Kronenberg-Auramberg mountains.

(ii) Another arcuate fault on Rooiberg Süd and Naudaus could be associated with the *en bloc* subsidence of the Rooiberg rhyolite field.

(iii) Normal faulting, following the north to north-northwest regional structural trend is best developed in the Blutpütz West and adjacent areas, coinciding with the present outcrops of rhyolite in this area. This may indicate the development of unstable conditions due to emplacement of large quantities of magma into high crustal and extrusive positions.

The intrusion of quartz-porphyry dykes of the Guperas can generally be related to major crustal tensional features. The emplacement of the extremely dense swarm on northern Blutpütz West and the adjoining part of Naudaus appears to have been controlled by a prominent north to north-northwesterly horst-and-graben type fault zone that dominated during Guperas and post-Guperas times. The swarm in the vicinity of Vergenoeg, Kumbis and Sonop also appears to have intruded into a tensional zone developed in the outer parts of a large, shallow, and roughly northeasterly-plunging anticlinal structure which is apparent from the outcrop pattern of the Kunjas and Barby Formations in this area. The dykes have been injected approximately parallel to the axial plane of this anticline which dips steeply to the northwest. No obvious relation exists between the north-northeasterly trend of the quartz-porphyry dykes in the Vrede area and any major structures but it is most likely that intrusion has also been influenced by the stress field that dominated during Guperas and post-Guperas times.

3.2.4 SPES BONA SYENITE AND ASSOCIATED DIORITE AND MONZONITE

The lack of any evidence for a composite character of the Spes Bona Syenite, such as textural and compositional variability, strongly suggests that the body was emplaced in a single magmatic phase. The generally smooth outline of the intrusion (Fig. 5) supports this and together with the conspicuous lack of chill zones and contact metamorphism, is possibly also indicative of a partly crystalline or 'mushy' state during emplacement.

It is probable that emplacement of the syenite body into its present position took place to a large extent by vertical fracturing and upward displacement of overlying country-rocks (Fig. 9). Such a mechanism could also account for the lack of xenoliths which would be expected in abundance if a stoping intrusive mechanism has operated, and for the lack of contortion of the country-rock. The only deformation noted in the latter and associated with the emplacement of the syenite is a vertical shear developed within the adjacent Barby lavas on Sonop. Shearing of this nature is consistent with an emplacement of the syenite involving *en bloc* vertical displacement of the roof rocks.

Fracturing and uplift of the overlying and immediately adjacent country-rocks prior to the rise of the syenite might have provided potential channels for the transport of the more mobile and genetically related dioritic and monzonitic magmas into the crust (Fig. 9). Since such fractures would be expected to delineate roughly the border areas of the rising syenite mass, it follows that these more basic intrusives would be injected into, and preserved within, positions peripheral to the syenite, as indeed they are (Section 3.1.3 and Fig. 5).

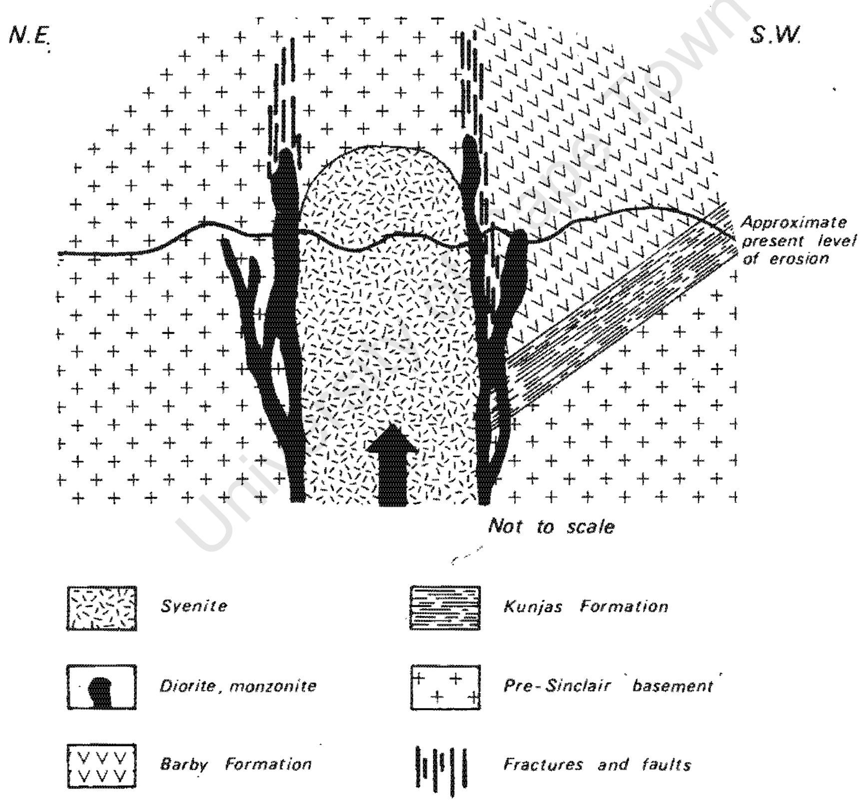


Fig. 9. Sketch section showing the probable mode of emplacement of the Spes Bona syenite and associated dioritic and monzonitic intrusives

A conspicuous feature of the Spes Bona syenite is its pronounced elongation along a northwesterly-trending axis (Fig. 5). This elongation is parallel to the trend of the main Nubib granite belt, the Nam shear belt and the major faulting in the Betta-Vrede area as discussed earlier in this section with regard to the emplacement of Nubib granite.

It would appear, therefore, that up to Guperas times at least, the emplacement of the major igneous intrusive units was strongly influenced by a northwesterly-trending structural grain, which may reflect the late stages of brittle deformation in the Namaqua Mobile Belt to the south and southeast (Jackson, 1974; Blignault *et al.*, in press).

3.2.5 BASIC LAVA AND DYKES

The extrusion of basic lava during development of the Sinclair Group took place under subaerial conditions. This is indicated by the complete lack of pillow structure, as well as the close association of the lavas with coarse-grained clastic rocks and ash-flow tuffs. The presence of a pre-existing basement, built up largely of granitic material and felsic and basic intrusives, further suggests a continental and, therefore, probably subaerial environment. The paucity of primary basic pyroclastic deposits is indicative of relatively quiet extrusion interrupted only periodically by violent explosive activity.

An areally extensive distribution of generally uniform Barby basalt types occurring north of latitude 25°30'S is indicative of a high mobility of flow and it is most likely that extrusion took place from extensive fissures. Since there is, however, no field evidence for feeder dykes in the area, this suggestion is tentative.

South of latitude 25°30'S, lava types are somewhat more variable in vertical section and distinct lateral variations in gross texture and colour can also be recognised. From the lateral and vertical variations in gross characteristics as discussed in Section 3.1.2.1, it is apparent that a number of distinct centres of eruption existed in the area extending from Sonop to Aruab. Unfortunately, the exact nature and positions of these centres cannot be determined due to a lack of detailed lithological data on the distribution of the various members of the Barby lava succession. Once again, no feeders are exposed and only two thin dykes of undoubted Barby age were located. The small size of the latter certainly rules them out as primary feeder channels. The inferred presence of various volcanic 'centres' and the lateral interfingering of lava types might suggest the presence of a number of large single volcanoes (central volcanoes?). Outpourings of lava probably followed one another in rapid succession, building up the tremendous thickness of the Barby lava sequence without appreciable deposition of clastic or volcaniclastic material between flows.

The limited outcrop of Guperas basic lava precludes a lengthy discussion on the conditions prevailing during, or controlling, their extrusion. However, the close association with coarse clastic beds, ash-flow tuff horizons and non-bedded tuff horizons of non-flow origin, is strong evidence for a subaerial emplacement. The extremely local development of basic lava on

Ganaams does not necessarily imply limited mobility of the magma and it is suggested that the rapid and comtemporaneous extrusion of rhyolitic lava in immediately adjacent areas (Fig. 1) developed barriers beyond which the basic lava could not flow. The infilling of a relatively small and rapidly deepening 'basin' forming simultaneously with the extrusion of the basic lava could have resulted in the local and uniquely thick pile of basic lava.

Basic dykes of Guperas age occur in dense composition swarms together with the quartz-porphyry dykes mentioned earlier and intrusion has apparently been controlled by the same tensional features as discussed for the emplacement of the felsic dykes. Whether these basic dykes represent feeders to the limited amount of basic lava occurring in the Guperas sequence is not certain, due to lack of field evidence. However, chemical data from basic dykes of probable Guperas age in the area west of Helmeringhausen (von Brunn, 1967), indicate that the dykes have similar compositions to the lavas and the age relationships within the Guperas sequence (Section 3.1.6.3) are also in agreement with the possibility of the dykes representing feeders to the lavas.

3.3 CHEMICAL COMPOSITION AND CLASSIFICATION

3.3.1 BASIC ROCK UNITS

Whole-rock chemical data for the basic igneous rock units of the Sinclair Group are shown in Tables 3, 6, 8, 10 and 12. The data represent twenty-nine analyses of lava and intrusives from the area north of the Sinclair Mine and seven analyses of Barby lava and Guperas basic dykes from the area west of Helmeringhausen (von Brunn, 1967) (Fig. 33). The former are new analyses carried out by the writer during 1972 (for analytical details see Appendix).

3.3.1.1 Barby Formation

(i) Extrusives

In the classification of these basic extrusives, a primary subdivision based mainly on SiO_2 content has been employed (in part after Taylor *et al.*, 1969). In this scheme, *basalts* have a SiO_2 content lower than 53 per cent[1] and *andesites* range from 53 to 62 per cent with a subgroup of low-Si (basaltic) andesite with 53 to 57 per cent SiO_2. Since the highest silica content of any Sinclair Group basic lava is 56,1 per cent, all 'andesites' considered here are of the low-Si or basaltic andesite type, although the term 'andesite' is frequently used for the sake of simplicity (e.g. Section 3.1.2.1).

In Section 3.1.2.1 an informal subdivision into 'members' has been applied to the extrusive sequence of the Barby Formation. This subdivision is based on field and petrographic characteristics, and wherever possible, on chemistry. The term 'basaltic' is therefore applied to members composed predominantly of lava with a SiO_2 content of less than 53 per cent (e.g. the Basalt member). Use of the term 'andesite' or 'basaltic andesite' in a member name implies a dominance of lavas with a SiO_2 content within the range 53-57 per cent. Those lavas which contain anomalously high contents of K_2O have been prefixed by the term 'trachy-' (e.g. trachybasalt or trachyandesite). In the case of these potassic lavas, gross petrographic characteristics such as the dominant phenocryst mineral present and the relative size of the inset crystals allow further distinction to be made between members, resulting in most of the member names used in Section 3.1.2.1 (e.g. Large-felspar trachyandesite, Pyroxene trachyandesite, etc.).

Calc-Alkaline Lavas

One of the most striking and characteristic geochemical features of much of the Barby basic lava sequence is the relatively high potassium content and the steep slope of the correlation on the K_2O vs. SiO_2 variation diagram (Fig. 10).

[1]All oxide concentrations given in the following Sections are weight percentages

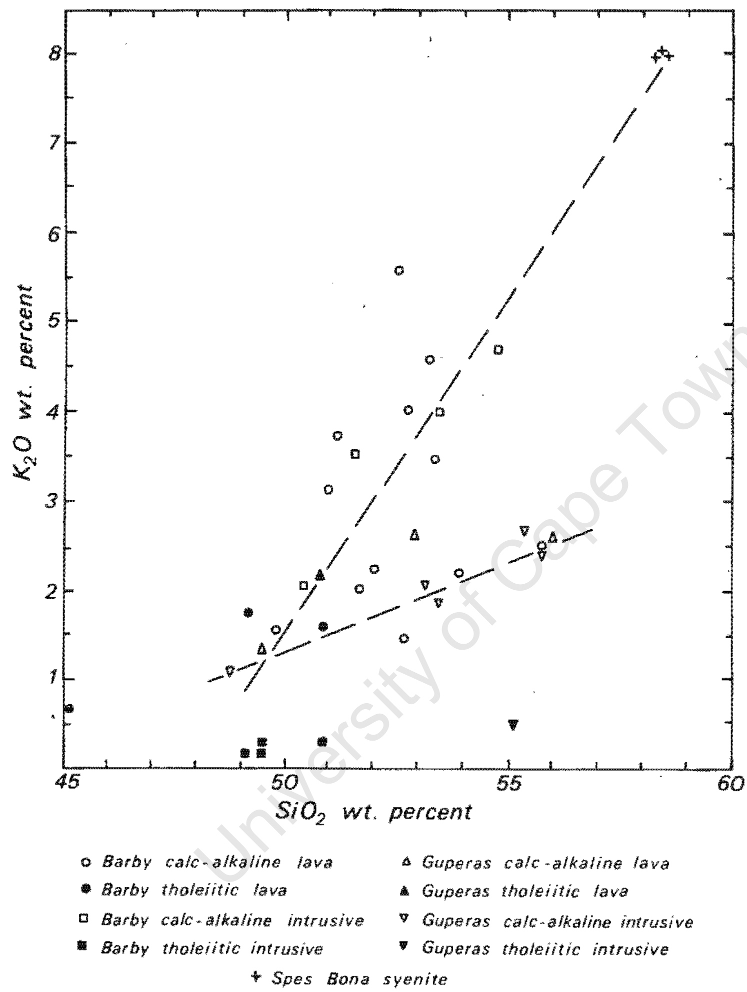


Fig. 10. K_2O vs. SiO_2 variation diagram for the basic and syenitic igneous rocks of the Sinclair Group

The strongly alkaline and generally low-silica nature of many of the Barby lavas (Fig. 11) might suggest that they belong to the alkali olivine basalt series, but a number of factors rule out such a classification:

(a) On an AFM diagram $((Na_2O+K_2O):(FeO+0,9Fe_2O_3):MgO)$ it is apparent that the lavas plot in the area of marked iron enrichment (relative to magnesium with increasing alkali content) typical of the tholeiitic and alkali olivine

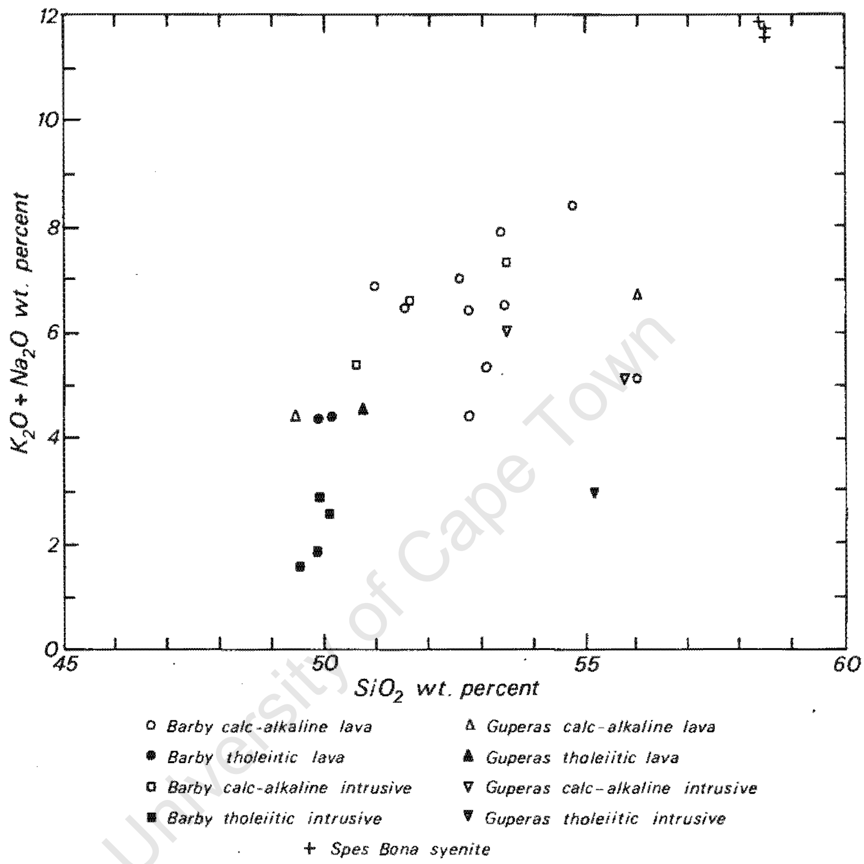


Fig. 11. Plot of total alkalis vs. SiO_2 for the basic and syenitic igneous rocks of the Sinclair Group

basalt series, as well as in the area of low or virtually no iron enrichment as typified by the calc-alkaline series (Fig. 12).

All of those samples that plot within the field of low iron enrichment are those of relatively high potassium content (Table 3). *The potassic lavas of the Barby, therefore, display calc-alkaline affinities and are not typical alkali olivine basalts.*

Figure 12 indicates, therefore, that a range of basic rock types exists within the Sinclair Group, ranging from tholeiites (of variable but generally low K_2O content) to rocks of distinctly calc-alkaline character and generally high to very high potassium content.

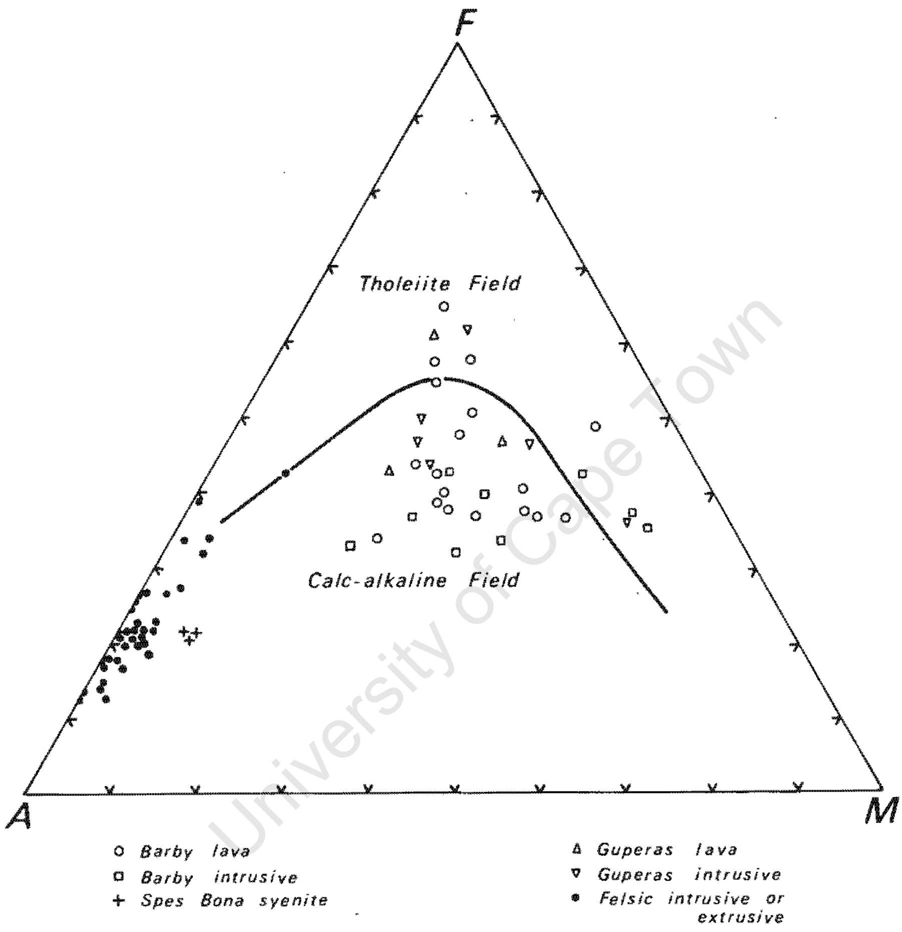


Fig. 12. AFM diagram for basic, syenitic and felsic igneous rock units of the Sinclair Group, including the data of von Brunn (1967) for basic rocks. The boundary curve between tholeiitic and calc-alkaline fields is that proposed by Irvine and Baragar (1971)

A plot of MgO/Al_2O_3 vs. $(Na_2O+K_2O)/(\text{Total } FeO+TiO_2)$ can also be used to discriminate between tholeiitic and calc-alkaline trends (Green, 1973). When plotted on such a diagram (Fig. 13) the distribution of data between the calc-alkaline and tholeiite fields is identical to that in the AFM plot.

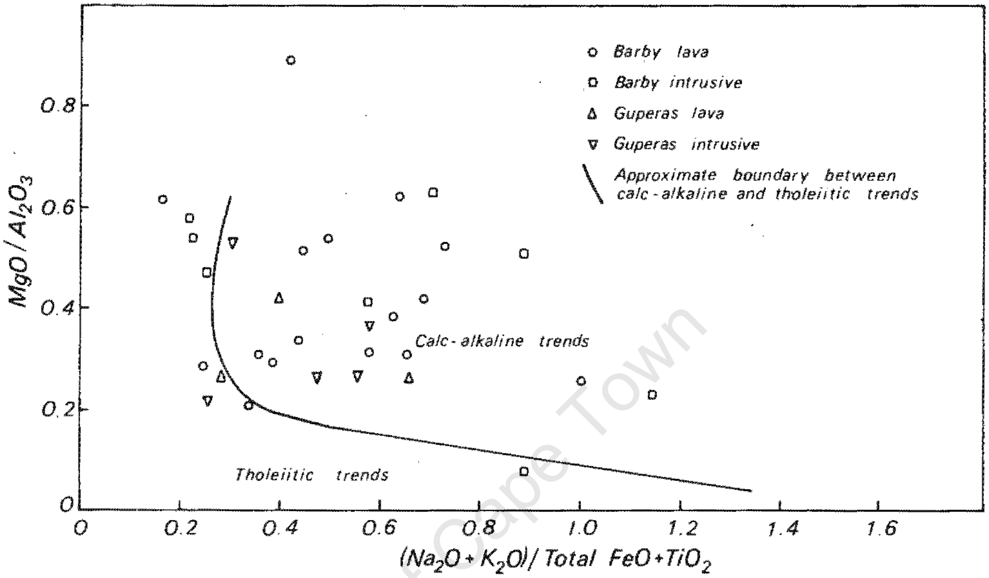


Fig. 13. Plot of MgO/Al_2O_3 vs. $(Na_2O+K_2O)/(Total\ FeO+TiO_2)$ for the Sinclair Group basic rocks (including those of von Brunn, 1967). The boundary line between the tholeiitic and calc-alkaline fields follows the data of Green (1973)

(b) The K_2O/Na_2O ratios of the Barby potassic extrusives are typically in the order of one or greater (Table 3, Fig. 14) whereas in the case of the alkali olivine basalt series this ratio is 0,5 or less (Dickinson *et al.*, 1968; Manson, 1967).

(c) By definition, an alkali olivine basalt must possess both olivine and nepheline in the norm (Yoder & Tilley, 1962; Green & Ringwood, 1967). The Barby calc-alkaline potassic lavas contain hypersthene in the norm and no nepheline, and many are quartz- rather than olivine-normative (Table 3, Fig. 15).

The geochemical and petrographic (Section 3.1.2.1) properties, as displayed by the calc-alkaline potassic basaltic-andesite lavas of the Barby Formation, strongly suggests that these rocks belong to the "shoshonite rock association", a name proposed by Joplin (1964, 1965 and 1968) for a group of potassium-rich basaltic and andesitic rocks. Joplin *et al.* (1972) have shown that the shoshonites have affinities with the calc-alkaline series and are not related to the alkali basalts. Other recent studies of high potassium

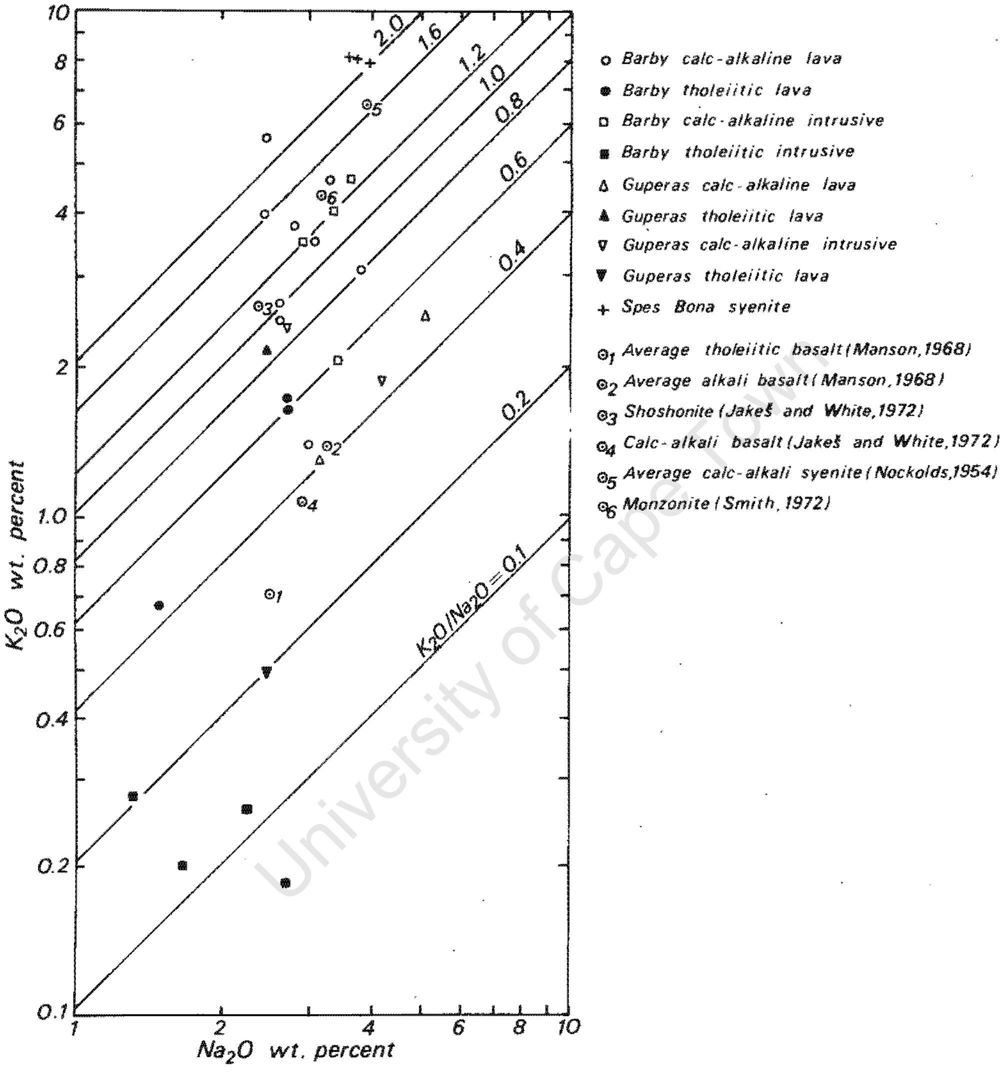


Fig. 14. Logarithmic plot of K_2O vs. Na_2O for the basic rocks and syenite of the Sinclair Group compared with typical average values

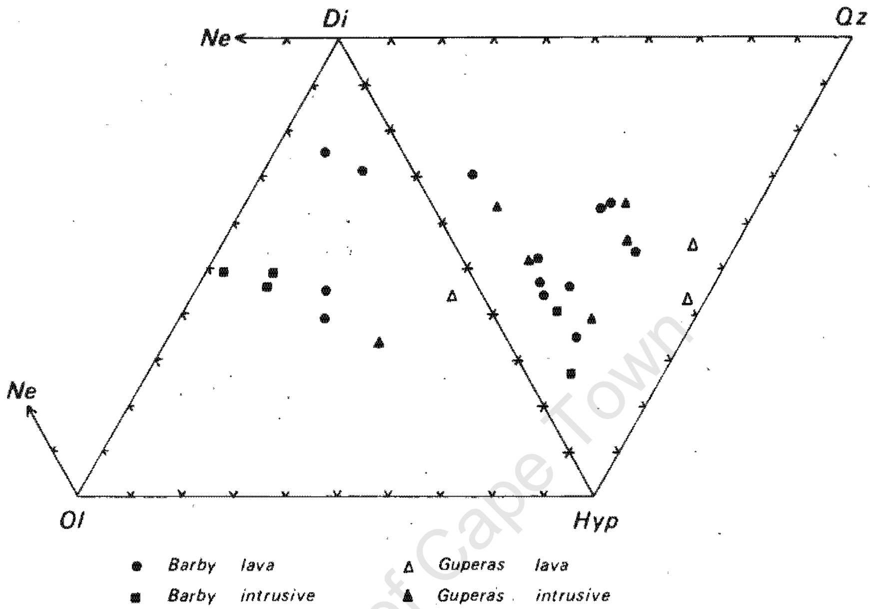


Fig. 15. Projection into the basalt tetrahedron (from plagioclase) of the CIPW norms of analysed basic rocks of the Sinclair Group (including the samples of von Brunn, 1967), following the procedure of Green (1973)

basalt-andesite suites have unequivocally shown the close relationship between these rocks and the calc-alkaline association (Jakes & White, 1971; Gill, 1970; Jakes & Smith, 1970; Jakes & White, 1972).

Analyses of typical shoshonites with a similar range in SiO_2 content as that of the Barby potassic lavas, have been included in Table 3 for comparison.

High K_2O contents, ranging from 1,41 to 5,57 per cent characterise the Barby calc-alkaline lavas and hence they compare favourably with typical shoshonites. Many modern calc-alkaline suites including rocks of the shoshonite association, display distinctly flatter trends on the K_2O vs. SiO_2 diagram than that displayed by the Barby calc-alkaline lavas (Jakes & White, 1972; Joplin, 1968) but shoshonitic associations characterised by steep rather than flat K_2O vs. SiO_2 relationships have been described (Gill, 1970; Ninkovich & Hays, 1972).

The Al_2O_3 contents of the Barby calc-alkaline lavas vary widely from 12,54 to 17,24 per cent, with an average of 14,24 per cent, the fluctuation being independent of the SiO_2 content, with a consequent wide scatter of points and lack of correlation in the Al_2O_3/SiO_2 variation diagram (Fig. 16).

It is apparent from the average major element data (Table 18) that the felsic igneous rocks of the Sinclair can be divided into two fairly distinct groups: a low-Ca group (average CaO contents of 0,33-0,83 per cent) and a high-Ca group (average CaO contents of 1,52-1,77 per cent). Falling within the latter group is the Haremub granite and those rhyolites which occur in the lower part of the Barby succession south of latitude $25^{\circ}30$'S. The former group is made up of the Nubib granite, rhyolites of the Barby occurring north of latitude $25^{\circ}30$'S (the 'Rhyolite member'), and the rhyolite lava, plugs and domes, and quartz-porphyry dykes of the Guperas Formation. The felsic effusives of the Nagatis formation (average CaO content of 0,83 per cent) and the Rooikam and Tumuab granites (average CaO contents of 0,79 and 0,74 per cent respectively) of von Brunn (1967) also fall within this low-Ca grouping.

It is emphasised that these groupings have been made on the basis of average chemical compositions only and, although holding true for most of the individual rock analyses, a few relatively high-Ca specimens (BW 935, BW 1403, BW 581) do occur in the low-Ca grouping and *vice versa* (BW 1566). Nevertheless the felsic magmas produced during certain periods of igneous activity were of a predominantly relatively high-Ca or low-Ca character.

3.3.3.1 Low-Ca Units

The units belonging to this group constitute the bulk of felsic rocks of the Sinclair Group and generally have compositions falling within the range of the average calc-alkali granite (and rhyolite) and average alkali granite (and rhyolite) of Nockolds (1954). Figures 21 and 22 show the positions of the Sinclair felsic rocks relative to these average compositions of Nockolds (1954) in terms of the major normative constituents Qz, Or, Ab and An. The highly siliceous nature of these felsic rocks (74,46 to 76,87 per cent SiO_2) as compared with the slightly lower silica values in Nockolds' average granites and rhyolites results in the Sinclair rocks plotting generally closer to the Qz apex of the Qz-Ab-Or diagram (Fig. 22). The only other notable differences in major element chemistry of the low-Ca felsic units when compared with Nockolds' (1954) average calc-alkaline and alkali compositions are a higher state of iron oxidation ($Fe_2O_3/FeO>1$) and a slightly, but consistently, lower Al_2O_3 content. The very low normative anorthite content of most of the Sinclair felsic rocks can be attributed to the low alumina content in combination with the low concentrations of calcium.

From the major element compositions presented in Tables 13-18, and the clustering of points in the diagrams incorporating the major normative constitutents An, Ab, Or and Qz, Ab, Or (Figs. 21 and 22 respectively), it is apparent that the low-Ca felsic rock units of the Sinclair have very uniform chemical characteristics, suggesting that these granites, rhyolites and quartz-porphyries belong to a single, extended but episodic, magmatic event during which similar conditions in the source region produced magmas of only limited compositional variability. A corresponding uniformity in the average chemical compositions of felsic rock units of the Sinclair Group occurring to the south of the present study area, has been noted by von Brunn (1967). This is to be expected since most of the felsic units from the latter area can be directly correlated with those north of the Sinclair mine.

cent).

The syenite is characterised by a very high K_2O content (about 8 per cent) and a high K_2O/Na_2O ratio in excess of 2 (Fig. 14). On the AFM diagram of Fig. 12 the syenite falls within the calc-alkali field, and the major element composition compares very favourably with the average calc-alkali syenite of Nockolds (1954) (Table 9). Small but notable deviations in compositions from this average syenite are a higher state of iron oxidation with consequent high Fe_2O_3/FeO ratio, and lower CaO and MgO contents in the case of the Spes Bona syenite. On the K_2O vs. SiO_2 diagram (Fig. 10) the syenite samples lie on a continuation of the steep trend defined by the calc-alkaline lavas and intrusives of the Barby. The chemical properties and the close association in the field and in time (Section 3.1.3) between the syenite and the diorites, monzonites, trachybasalts and trachyandesites of the Barby, suggests that the syenite belongs to the same high-potassium calc-alkaline (shoshonitic) suite as the Barby rocks.

In Table 9 the analyses of two syenites from a shoshonitic intrusive series ranging from gabbros through monzonites into syenites (Smith, 1972), are presented for comparison with the Spes Bona syenite. Although the latter does not compare well with either one of the shoshonitic syenites, it does correspond remarkably well with a rock intermediate in composition between the two. The K_2O/Na_2O ratios are high in these shoshonitic syenites as are the Fe_2O_3/FeO ratios, and both compare well with those of the Spes Bona syenite. On the basis of major element data alone, therefore, it would seem that the syenite does belong to a high-potassium, calc-alkaline or shoshonite suite.

The contents of Rb, Sr and Ba are exceptionally high (Table 9), exceeding even the elevated values observed in the highly potassic lavas and intrusives of the Barby (Table 8). Such high concentrations of large trace element cations are characteristic of the shoshonite association as already discussed, and the extremely high contents of Sr (1072-1300 ppm) and Ba (2418-3200 ppm) as present in the Spes Bona syenite are comparable with similar high concentrations in the shoshonitic syenites described by Smith (1972) (Table 9). K/Rb ratios (Fig. 19) are low (194-213) and fall within the range covered by the calc-alkaline potassic lavas and intrusives of the Barby.

3.3.3 FELSIC ROCK UNITS

Thirty-seven samples from the major felsic rock units of the Sinclair Group (as occurring north of the Sinclair Mine) were analysed for all major elements and six trace elements and the results appear in Tables 13-17. A further 18 analyses of felsic rocks from the Sinclair Group have been presented by von Brunn (1967) and these include data from four effusives of the Nagatis Formation (see Sections 4 and 5) which does not occur in the present study area.

In Table 18 the average chemical compositions of the various felsic units of the Sinclair Group (from the present study area) are compared with average granite compositions from Nockolds (1954) and Turekian and Wedepohl (1961).

Trace element contents are significantly different from those of the calc-alkaline intrusives and further serve to distinguish between the two groups of intrusives (Table 11).

	Barby tholeiitic intrusives		Barby calc-alkaline intrusives	
	Average	Range	Average	Range
Rb	5,2	(3,1-8,6)	159	(87-200)
Sr	224	(107-349)	847	(729-1018)
Ba	108	(76-178)	1214	(1099-1317)
Zr	17,1	(10,9-27,0)	124	(75-188)
Nb		<0,53[1]	3,3	(1,2[1]-5,2)
Y	7,0	(4,1-9,7)	16,3	(12,8-19,7)

[1]below detection limit

Table 11. Comparison between the trace element contents (in parts per million) of the tholeiitic and calc-alkaline intrusive rocks of the Barby Formation

Rb, Sr and Ba, which have high concentrations in the calc-alkaline intrusives are relatively low in their tholeiitic counterparts, and even Zr, Nb and Y, which were not anomalously high in the calc-alkaline rocks, are comparatively low in the tholeiites.

K/Rb ratios are variable (267-573) (Fig. 19), but are nevertheless appreciably higher than those of the calc-alkaline intrusives (161-204).

3.3.1.2 Guperas Formation

Chemical analyses of three samples of lava and three of intrusives from the Guperas Formation are available from the area north of Sinclair (Table 12). Von Brunn (1967) has also presented three analyses of basic dykes which belong to the Guperas Formation.

Limited chemical data for the Guperas Formation, therefore, precludes a lengthy discussion of the chemical characteristics of this basic igneous suite.

It is evident, however, that the compositional trends present in these rocks are very similar to those in the Barby lavas and intrusives. Both tholeiitic and calc-alkaline varieties are present (Figs. 12 and 13) and the

Table 9. Comparison of the chemical compositions and normative mineralogy of the calc-alkaline intrusives of the Barby Formation and the Spes Bona syenite with that of high-potassium calc - alkaline intrusives

	BW 1606	BW 1256	BW 1613	BW 1565	1	2	3	4
SiO_2	50,65	51,62	53,52	54,82	50,5	54,9	55,0	52,6
TiO_2	0,80	0,79	0,82	0,90	0,8	0,73	0,81	0,58
Al_2O_3	16,75	14,23	15,05	17,11	17,5	14,8	15,3	17,9
Fe_2O_3	3,55	3,41	3,03	4,10	4,75	4,05	4,5	4,4
FeO	5,12	5,04	4,42	2,24	4,25	3,95	3,95	3,5
MnO	0,14	0,14	0,12	0,13	0,15	0,14	0,14	0,15
MgO	6,94	9,01	7,79	3,88	8,4	7,2	3,95	2,85
CaO	8,80	7,41	6,55	4,96	3,95	5,35	7,2	8,15
Na_2O	3,42	3,03	3,35	3,62	3,85	3,25	3,15	3,3
K_2O	2,06	3,52	3,99	4,73	3,3	3,5	4,31	4,05
P_2O_5	0,58	0,54	0,49	0,51	0,6	0,52	0,34	0,3
l.o.i.	0,28	0,23	0,24	2,26	1,38	1,05	1,15	1,79
H_2O^-	0,02	0,06	0,04	0,19	0,2	0,27	0,21	0,17
	99,11	99,03	99,41	99,45	99,63	99,71	100,01	99,74
Rb	87	143	206	200				
Sr	1018	780	729	859	950	700	820	1000
Ba	1099	1317	1147	1293	1000	1250	1200	1800
Zr	75	83	151	188	90	270	115	<100
Nb	<1,6	1,8	5,2	4,9	<40	<40	<40	<40
Y	13	16	17	20				
K/Rb	197	204	161	196				
K_2O/Na_2O	0,60	1,16	1,19	1,31	0,86	1,08	1,37	1,23
Fe_2O_3/FeO	0,69	0,68	0,68	1,83	1,12	1,02	1,14	1,26
K_2O+Na_2O	5,48	6,55	7,34	8,35	7,15	6,75	7,46	7,35
CIPW NORM (Wt. per cent)								
Q				1,73		2,17	2,71	
Or	12,17	20,80	23,58	27,95	19,50	20,68	25,47	23,93
Ab	28,94	25,64	28,34	30,63	32,57	27,50	26,65	27,92
Ne								
An	24,27	14,84	14,25	16,47	15,68	15,46	14,88	22,07
C					1,85			
Di Wo	6,51	7,68	6,28	2,01		3,21	7,77	6,85
Di En	4,55	5,62	4,62	1,73		2,46	5,63	4,82
Di Fs	1,42	1,34	1,06			0,41	1,43	1,45
Hy En	2,54	0,64	2,62	7,93	5,43	15,47	4,20	1,22
Hy Fs	0,79	0,15	0,60		0,74	2,58	1,06	0,37
Ol Fo	7,15	11,33	8,51		10,85			0,74
Ol Fa	2,46	2,97	2,15		1,64			0,25
Mt	5,15	4,94	4,39	5,05	6,89	5,87	6,52	6,38
Il	1,50	1,48	1,54	1,69	1,50	1,37	1,52	1,09
Hm				0,61				
Ap	1,37	1,28	1,16	1,21	1,42	1,23	0,81	0,71
	98,82	98,71	99,10	97,01	98,07	98,41	98,65	97,80

	5	6	BW 1617	BW 1598	BW 1599	7	8	9
SiO_2	51,00	51,74	58,45	58,40	58,49	60,41	59,7	59,41
TiO_2	0,13	0,14	0,60	0,67	0,67	0,52	0,28	0,83
Al_2O_3	17,21	4,76	18,82	18,73	18,69	15,22	19,7	17,21
Fe_2O_3	4,23	1,86	1,96	2,06	2,19	4,88 †	1,24	2,19
FeO	2,41	6,96	1,60	1,60	1,64		0,78	2,83
MnO		0,19	0,05	0,06	0,06	0,09	0,06	0,08
MgO	6,19	7,33	1,45	1,36	1,49	3,38	0,29	2,02
CaO	9,15	8,42	3,26	3,14	3,22	4,93	2,2	4,06
Na_2O	2,88	2,68	3,63	3,90	3,68	4,25	2,5	3,92
K_2O	4,93	3,93	8,03	7,98	8,05	5,76	10,9	6,53
P_2O_5		0,47	0,50	0,48	0,50	0,61	0,04	0,38
l.o.i.	0,63 (l.o.i. + H_2O^-)	0,52	0,76	0,74	0,77		1,37	0,63
H_2O^-			0,06	0,07	0,01		0,13	
	98,76	100,00	99,17	99,19	99,46	100,05	99,19	100,00
Rb			312	341	329			
Sr			1300	1072	1180	>1000	>1000	
Ba			3200	2418	2850	>2000	1450	
Zr			69	90	85	<100	190	
Nb			<1,5	2,0	2,2			
Y			9	12	12	40	<40	
K/Rb			213	194				
K_2O/Na_2O			2,21	2,05	2,19	1,36	4,36	1,67
Fe_2O_3/FeO	1,75	0,27	1,22	1,29	1,33		1,59	0,77
K_2O+Na_2O	7,81	6,61	11,66	11,88	11,73	10,01	13,4	10,45
CIPW NORM (Wt. per cent)								
Q								1,67
Or	29,13	23,22	47,45	47,16	47,57		64,71	38,59
Ab	10,83	20,69	30,71	30,79	30,79		18,20	33,17
Ne	7,33	1,07		1,19	0,19		1,60	
An	19,48	16,64	11,35	10,04	10,71		10,35	9,84
C								
Di Wo	10,82	9,21	0,65	1,00	0,83		0,13	3,27
Di En	9,03	5,66	0,51	0,82	0,69		0,10	2,12
Di Fs	0,42	3,03	0,06	0,06	0,04		0,01	0,93
Hy En			0,06					2,91
Hy Fs			0,01					1,27
Ol Fo	4,47	8,83	2,13	1,80	2,12		0,43	
Ol Fa	0,23	5,20	0,29	0,15	0,15		0,05	
Mt	6,13	2,70	2,84	2,99	3,18		1,80	3,18
Il	0,24	2,14	1,13	1,26	1,26		0,53	1,56
Hm								
Ap		1,11	1,18	1,14	1,18		0,09	0,90
	98,11	99,50	98,37	98,40	98,71		97,70	99,41

BW 1606 - Diorite, Barby Formation
BW 1256 - Olivine monzonite, Barby Formation
BW 1613 - Olivine monzonite, Barby Formation
BW 1565 - Monzonite, Barby Formation
1. Gabbro (Smith, 1972)
2. Monzonite (Smith, 1972)
3. Monzonite (Smith, 1972)
4. Gabbro (Smith, 1972)
5. Monzonite (shoshonite) (Joplin, 1968)
6. Augite - biotite - olivine monzonite (Nockolds, 1954)
BW 1617 - Calc-alkaline syenite, Spes Bona syenite
BW 1598 - Calc-alkaline syenite, Spes Bona syenite
BW 1599 - Calc-alkaline syenite, Spes Bona syenite
7. Syenite (Smith, 1972)
8. Syenite (Smith, 1972)
9. Average calc-alkaline syenite (Nockolds, 1954)

† Total Fe

Table 10. Comparison of the chemical compositions and normative mineralogy of the tholeiitic intrusives of the Barby Formation with that of the average gabbro, norite and localised anorthosite of Nockolds (1954)

	BW 334	BW 293	BW 323	BW 312	1	2	3
SiO_2	49,48	49,15	50,89	49,41	48,36	50,28	49,98
TiO_2	0,16	0,17	0,34	0,15	1,32	0,89	0,14
Al_2O_3	18,61	17,50	17,03	27,30	16,84	17,67	28,94
Fe_2O_3	1,71	1,85	2,51	1,11	2,55	1,30	0,80
FeO	5,09	5,83	5,68	2,01	7,92	7,46	1,43
MnO	0,12	0,14	0,14	0,05	0,18	0,14	0,07
MgO	9,98	9,90	8,09	2,11	8,06	9,27	0,84
CaO	12,22	13,19	12,40	14,25	11,07	9,72	14,01
Na_2O	1,33	1,67	2,27	2,72	2,26	1,96	2,75
K_2O	0,27	0,20	0,26	0,19	0,55	0,63	0,42
P_2O_5	0,01	0,02	0,05	0,02	0,24	0,21	0,09
l.o.i	0,56	0,07	0,18	0,20	0,64	0,47	0,55
H_2O^-	0,02	0,05	0,04	0,04			
	99,56	99,74	99,88	99,56	99,99	100,00	100,02
Rb	8,6	5,5	3,8	3,1			
Sr	187	107	349	253			
Ba	114	76	178	64			
Zr	17	14	27	11			
Nb	<1,6	<1,6	<1,6	<1,5			
Y	6,6	7,7	9,7	4,1			
K/Rb	267	302	573	494			
K_2O/Na_2O	0,20	0,12	0,11	0,07	0,24	0,32	0,15
Fe_2O_3/FeO	0,34	0,32	0,44	0,55	0,32	0,17	0,56
K_2O+Na_2O	1,60	1,87	2,53	2,91	2,81	2,59	3,17
CIPW NORM (Wt. per cent)							
Q			0,09				1,27
Or	1,60	1,18	1,54	1,12	3,25	3,72	2,48
Ab	11,25	14,13	19,21	23,01	19,12	16,58	23,27
An	44,01	39,67	35,51	61,73	34,18	37,56	65,39
Di Wo	6,91	10,71	10,72	3,69	8,00	3,88	1,47
Di En	4,82	7,24	7,11	2,32	4,93	2,43	0,76
Di Fs	1,51	2,65	2,83	1,15	2,61	1,21	0,68
Hy En	19,26	9,94	13,04	2,84	8,82	18,73	1,33
Hy Fs	6,02	3,63	5,19	1,41	4,68	9,29	1,19
Ol Fo	0,54	5,24		0,07	4,43	1,35	
Ol Fa	0,19	2,11		0,04	2,59	0,74	
Mt	2,48	2,68	3,64	1,61	3,70	1,88	1,16
Il	0,30	0,32	0,64	0,28	2,48	1,67	0,26
Hm							
Ap	0,02	0,05	0,12	0,05	0,57	0,50	0,21
	98,91	99,55	99,64	99,32	99,36	99,54	99,47

BW 334 - Norite, Barby Formation
BW 293 - Gabbro, Barby Formation
BW 323 - Norite, Barby Formation
BW 312 - Anorthosite, Barby Formation
1. Average Gabbro (Nockolds, 1954)
2. Average Norite (Nockolds, 1954)
3. Anorthosite (localised) (Nockolds, 1954)

SiO_2, ranging from 2,06 per cent in the diorite (BW 1606) through 3,52 and 3,99 in the olivine monzonites (BW 1256, 1613) to 4,73 per cent in the monzonite (BW 1565). This trend of concomitantly increasing K_2O and SiO_2 coincides with, and reinforces, the same steeply positive trend observed for the calc-alkaline lavas (Fig. 10). The CaO content shows a regular decrease from 8,80 per cent in the diorite to 4,96 per cent in the monzonite. MgO is extremely variable (3,88 to 9,01 per cent). Similar highly variable MgO contents characterise the calc-alkaline lavas of the Barby (Table 3) and the shoshonite intrusives (Smith, 1972) presented in Table 9 for comparison purposes. Al_2O_3 contents are highly variable (14,23 to 17,11 per cent) and show no correlation with SiO_2 (Fig. 16). As with the calc-alkaline lavas, TiO_2 is always low (0,79-0,90 per cent), the K_2O/Na_2O ratios are high (0,6 to 1,3), increasing with rising K_2O content, and the P_2O_5 content is generally moderately high. K/Rb ratios are consistently low (161-204) (Fig. 19).

From Table 8 (p. 127) it can be seen that there is a very close similarity between the trace element contents of the Barby calc-alkaline extrusives and intrusives.

Tholeiitic Intrusives

Tholeiitic intrusive rocks of the Barby Formation are gabbroic and noritic in composition and are predominantly undersaturated with respect to silica in possessing olivine in the norm. Only one of the four samples analysed contains normative quartz. Sample BW 312 is exceptional in that it plots within the calc-alkaline field of Fig. 12 in spite of the fact that it is closely associated with the tholeiitic rocks. However, this anomaly is almost certainly due to the sample being collected from an anorthositic lense of cumulative origin with consequent lack of ferromagnesian constituents and, therefore, lack of iron enrichment.

In Table 10 the chemical compositions of the gabbro, norites and anorthosite are presented and compared with the average gabbro, average norite and "localised anorthosite" of Nockolds (1954). The only significant differences are lower TiO_2 (0,16 to 0,34 per cent) and higher CaO (12,22 to 13,19 per cent) in the case of the Barby gabbro and norites. The relatively high CaO content is reflected in a higher normative plagioclase content in the Barby rocks as compared with the averages of Nockolds (1954). Since the Barby rocks show evidence of magmatic mineral segregation and layering due to crystal settling and are, to varying degrees, cumulates, a direct comparison of chemical composition with that of average values is possibly not valid for most elements. The chemistry does, however, allow these tholeiitic intrusives to be clearly distinguished from the calc-alkaline intrusives of the Barby (Tables 9 and 10). In the tholeiites, K_2O contents are very low, ranging from 0,20 to 0,27 per cent, as opposed to a range of 2.06 to 4,73 per cent in the calc-alkaline rocks, for similar ranges in SiO_2 content. CaO contents are considerably higher in the tholeiites, 12,22 to 13,19 per cent, as compared with 4,96 to 8,80 per cent in the calc-alkaline rocks.

Further, but less marked, differences in major element chemistry are the higher Al_2O_3 and lower Na_2O in the tholeiitic intrusives as compared with the calc-alkaline intrusives.

	Barby calc-alkaline lavas		Barby tholeiitic lavas		Tholeiitic basalts (Prinz, 1967)	
	Average	Range	Average	Range	Average	Range
Rb	139	(36,2-294)	62,8	(31,4-91,7)	30	(2-125)
Sr	721	(446-1003)	253	(245-267)	450	(57-2000)
Ba	1363	(623-2386)	299	(189-461)	244	(10-1200)
Zr	151	(92-288)	131	(27,0-255)	108	(20-335)
Nb	4,4	(2,7-11,6)	5,0	(0,17[1]-11,6)		
Y	22,0	(17,3-34,0)	28,2	(12,6-47,7)		(12-90)

[1]below detection limit

Table 7. Comparison between the trace element contents (in parts per million) of the Barby tholeiitic lavas, the calc-alkaline lavas of the Barby and average tholeiites of Prinz (1967)

	Barby calc-alkaline lavas		Barby calc-alkaline intrusives		Spes Bona syenite	
	Average	Range	Average	Range	Average	Range
Rb	139	(36,2-294)	159	(87-200)	246	(312-341)
Sr	721	(446-1003)	847	(729-1018)	1184	(1072-1300)
Ba	1363	(623-2386)	1214	(1099-1317)	2822	(2418-3200)
Zr	151	(92-288)	124	(75-186)	81	(69-90)
Nb	4,4	(2,7-11,6)	3,3	(1,2[1]-5,2)	1,8	(1,3[1]-2,2)
Y	22,0	(17,3-34,0)	16,3	(12,8-19,7)	10,8	(9,1-11,8)

[1]below detection limit

Table 8. Comparison between the trace element contents (in parts per million) of the Barby calc-alkaline lavas, intrusives and Spes Bona syenite

Table 6. Comparison of the chemical compositions and normative mineralogy of tholeiitic basalts of the Barby Formation with that of average basalts and basaltic andesites

	BW 1415	BW 811	BW 1351	1	2	3	4	5	6
SiO_2	45,13	49,34	50,94	50,83	47,90	51,5	51,57	59,64	57,40
TiO_2	0,61	1,41	2,22	2,03	1,65	1,2	0,80	0,76	1,25
Al_2O_3	15,86	17,94	12,74	14,07	11,84	16,3	15,91	17,38	15,60
Fe_2O_3	3,87	4,67	7,28	2,88	2,32	2,8	2,74	2,54	3,48
FeO	8,14	6,78	8,34	9,00	9,80	7,9	7,04	2,72	5,01
MnO	0,20	0,16	0,20	0,18	0,15	0,17	0,17	0,09	
MgO	9,88	3,70	3,66	6,34	14,07	5,9	6,73	3,95	3,38
CaO	10,85	8,51	6,17	10,42	9,29	9,8	11,74	5,92	6,14
Na_2O	1,51	2,69	2,74	2,23	1,66	2,5	2,41	4,40	4,20
K_2O	0,66	1,73	1,64	0,82	0,54	0,86	0,44	2,04	0,43
P_2O_5	0,10	0,15	0,33	0,23	0,19	0,21	0,11	0,28	0,44
l.o.i.	2,11	1,91	3,46	0,91	0,59	0,81	0,45	1,08	
H_2O^-	0,12	0,11	0,09						
	99,04	99,10	99,81	99,94	100,00	99,95	100,11	100,80	97,33
Rb	31	92	65						
Sr	246	267	245						
Ba	189	246	461						
Zr	27	112	255						
Nb	<1,7	3,3	11,6						
Y	13	24	48						
K/Rb	175	156	209						
K_2O/Na_2O	0,44	0,64	0,60	0,37	0,33	0,34	0,18	0,46	0,10
Fe_2O_3/FeO	0,47	0,69	0,87	0,32	0,24	0,35	0,39	0,93	0,69
K_2O+Na_2O	2,17	4,42	4,38	3,05	2,2	3,36	2,85	6,44	4,63
CIPW NORM (Wt. per cent)									
Q		2,78	10,55	3,72		3,22	2,24	8,89	13,52
Or	3,90	10,22	9,69	4,85	3,19	5,08	2,60	12,05	2,54
Ab	12,78	22,76	23,18	18,87	14,05	21,15	20,39	37,23	35,54
An	34,55	31,77	17,62	25,96	23,26	30,72	31,30	21,65	22,45
Di – Wo	7,78	3,95	4,52	10,12	9,01	6,90	10,95	2,46	2,15
Di – En	5,01	2,21	2,59	5,68	6,00	3,85	6,57	1,86	1,33
Di – Fs	2,25	1,59	1,73	4,03	2,35	2,77	3,80	0,35	0,68
Hy – En	7,42	7,00	6,52	10,11	14,25	10,84	10,19	7,98	7,08
Hy – Fs	3,34	5,02	4,36	7,18	5,57	7,80	5,89	1,49	3,63
Ol – Fo	8,53				10,36				
Ol – Fa	4,23				4,46				
Mt	5,61	6,77	10,56	4,18	3,36	4,06	3,97	3,68	5,05
Il	1,15	2,65	4,17	3,81	3,10	2,25	1,50	1,43	2,35
Hm									
Ap	0,24	0,36	0,78	0,54	0,45	0,50	0,26	0,66	1,04
	96,79	97,08	96,27	99,05	99,41	99,14	99,66	99,73	97,36

BW 1415 – Tholeiitic basalt, Barby Formation
BW 811 – K – rich tholeiitic basalt, Barby Formation
BW 1351 – K – rich tholeiitic basalt, Barby Formation
1. Normal tholeiitic basalt (Nockolds, 1954)
2. Tholeiitic olivine basalt (Nockolds, 1964)
3. Average continental tholeiitie (Manson, 1967)
4. Island arc tholeiite (Jakes and White, 1972)
5. Calc – alkaline andesite (Jakes and White, 1972)
6. Tholeiitic andesite (Jakes and White, 1972)

Al_2O_3 content of 17, 94 per cent is also high by normal tholeiitic standards and compares more favourably with alumina contents as found in calc-alkaline suites (cf. data of Taylor & White, 1966; Jakes & White, 1972). The MgO content (3,70 per cent) is very low and is comparable with that of tholeiitic and calc-alkaline andesites (Table 6).

Sample BW 1351 is also slightly enriched in K_2O and depleted in MgO when compared with average tholeiitic compositions. CaO (6,17 per cent) is also low for a tholeiite (Table 6). The constituent minerals of this sample are fairly altered so that the abnormally low calcium content might well be due to the decomposition of the constituent plagioclase and subsequent leaching. A very high Fe_2O_3 content and a significant TiO_2 content reflect an abundance of ilmenite in the rock.

Trace element data allow a fairly clear distinction to be made between the Barby calc-alkaline and tholeiitic rocks. The three elements Rb, Sr and Ba, high concentrations of which served to characterise the calc-alkaline rocks, have a markedly lower content range in the tholeiites. These low concentrations are typical of tholeiitic basalts in general (Table 7). Y and Zr also have contents coinciding with the normal range found in tholeiitic basalts and do not differ significantly from the contents found in the calc-alkaline lavas of the Barby (Table 7).

(ii) Intrusives

Basic intrusive rocks of the Barby Formation display chemical properties very similar to those of the extrusive members. Chemical analyses and CIPW norms of specimens from these intrusive units are presented in Table 9. All the samples analysed are hypersthene normative and, in addition, range from moderately undersaturated with up to 14,30 per cent olivine (BW 1256) to mildly oversaturated with 1,73 per cent quartz (Fig. 15 and Table 9). As in the case of the lavas, the intrusives fall within both the tholeiitic and calc-alkaline fields when plotted on an AFM diagram (Fig. 12). Characteristically, the calc-alkaline intrusives are high in potassium content whereas the tholeiites are relatively low in this element.

Calc-Alkaline Intrusives

The calc-alkaline intrusives display identical chemical characteristics to those of the potassium-rich lavas of the Barby and are, therefore, also comparable with rock-types of the shoshonitic association. In Table 9 the chemical compositions of the calc-alkaline intrusives are compared with those of similar potassium-rich calc-alkaline or shoshonitic intrusives from south-eastern Papua (Smith, 1972), with a monzonite (shoshonite) from Joplin (1968), and with the augite-biotite-olivine monzonite of Nockolds (1954).

On the basis of these comparisons it can be seen that the Barby calc-alkaline intrusives have a chemistry closely similar to that described by Smith (1972) for the Papuan shoshonitic intrusives.

The SiO_2 content (50,65 to 54,82 per cent) is fairly limited and is almost exactly the same as the range observed for the calc-alkaline lavas of the Barby (51,0 to 55,98 per cent). The K_2O content increases regularly with increasing

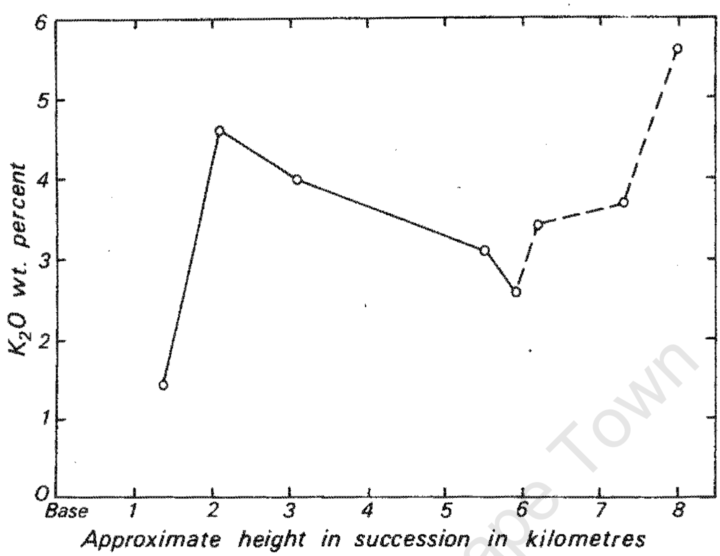


Fig. 20. Variation of K_2O content with height in the succession for the Barby calc-alkaline lavas of the Klein Haremub-Aubures-Aruab area

Tholeiitic lavas

Three samples of the Barby basic lava (BW 1415, 811, 1351) plot within the tholeiitic field of the AFM diagram (Fig. 12). All three are hypersthene-normative and, in addition, BW 1415 contains olivine in the norm whilst BW 811 and BW 1351 are quartz-normative. These samples do, therefore, fit the definition of a tholeiite as defined by Green & Ringwood (1967). The chemical compositions and normative mineralogy of the tholeiitic lavas are presented in Table 6 together with average tholeiitic compositions.

The olivine tholeiite (BW 1415) has a 'normal' major element composition intermediate between the 'normal tholeiitic basalt' and 'tholeiitic olivine basalt' of Nockolds (1954). It also compares favourably with the 'island arc tholeiite' of Jakes & White (1972), with the exception of possessing slightly higher MgO and lower silica contents than the latter.

Sample BW 811 plots very close to the calc-alkaline field in Figs. 12 and 13 and, consequently, although classified here as a tholeiite in view of the moderately high degree of iron enrichment and normative mineral constituents, it also displays certain chemical characteristics which mark it as a type transitional between a true tholeiite and calc-alkaline basalts of the shoshonite association already discussed. The potash content of 1,73 per cent in sample BW 811 is high for tholeiites since Manson (1967) records a maximum of 1,30 per cent of K_2O in continental tholeiites with an average of 0,86 per cent. The

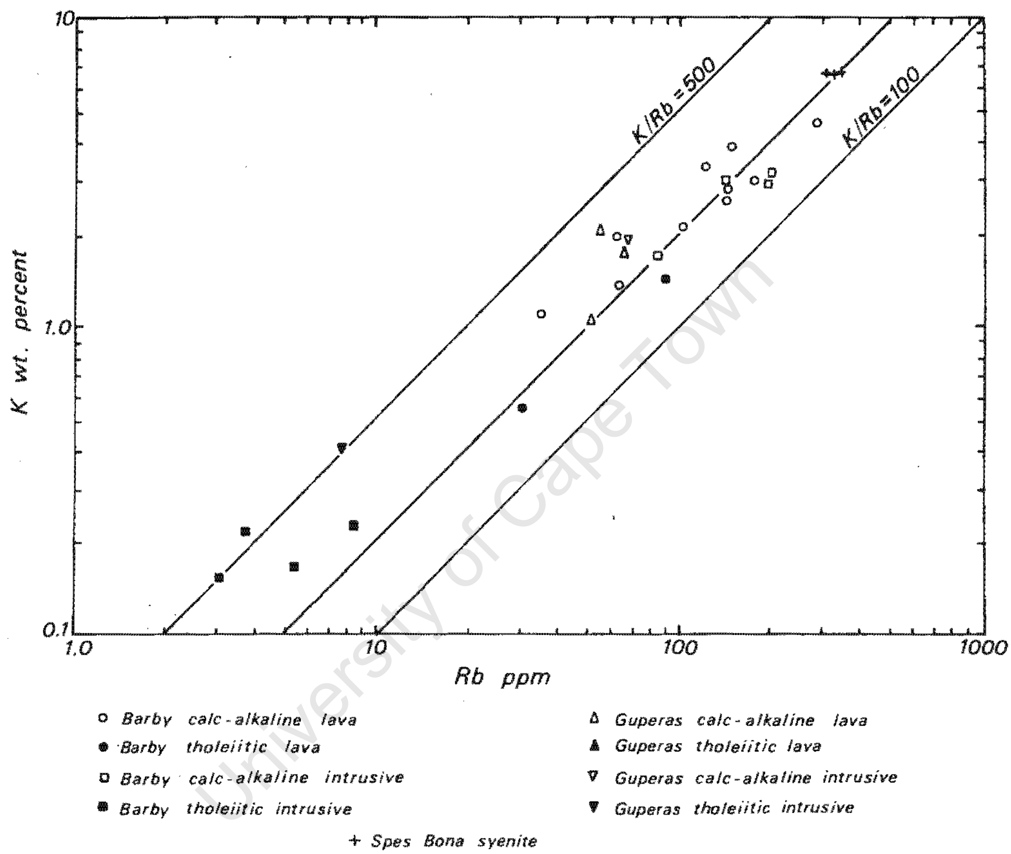


Fig. 19. Logarithmic plot of K vs. Rb for volcanic and intrusive basic rocks of the Barby and Guperas Formations, and the Spes Bona syenite

From brief visits to the areas west and north of the region discussed here, the impression was gained that the correlates of the Barby basic lavas in these adjacent areas are not of the excessively potassium-rich type but more closely resemble the tholeiitic basalts (north of latitude 25°30'S) and the slightly K-rich calc-alkaline andesites occurring near the base of the succession south of latitude 25°30'S.

The range of Sr values in the Barby rocks (446-903 ppm) is also high when compared with average values for tholeiitic (Prinz, 1967) (Table 4) and calc-alkaline basaltic suites (Taylor & White, 1966; Jakes & White, 1972), but is indistinguishable from that of many alkali olivine basalts which frequently have Sr contents in excess of 1000 ppm (Prinz, 1967).

Ba displays a very pronounced enrichment in the Barby calc-alkaline lavas, ranging from 759 to 2386 ppm. The extremely high concentrations of this element are evident when compared with the maximum value of 1200 ppm for all basaltic rocks as recorded by Prinz (1967). Normal calc-alkaline basaltic rocks (low-K) typically have average Ba contents in the order of 340 ppm according to Jakes & White (1972).

Enrichments in the trace elements Rb, Sr and Ba, as discussed above, are characteristic of the high-potassium calc-alkaline suite or shoshonitic association. Complete trace element data for rocks of this type are relatively few, but Table 5 shows a comparison between the trace element concentrations of the Barby calc-alkaline lavas and those of a few typical shoshonite suites. The data serve to illustrate the close chemical similarity between the Barby lavas and typical shoshonites. Rubidium, however, appears even more enriched in the Barby rocks than the shoshonites presented in Table 5. All other trace elements determined (Y, Zr and Nb) display normal distribution characteristics when compared with basalts in general (Prinz, 1967) and shoshonites in particular (Table 5).

K/Rb ratios of the Barby calc-alkaline lavas are low (158-328) (Fig. 19) and are comparable over most of the range of ratios with those of other high-potassium calc-alkaline suites which typically have K/Rb ratios of 200-320 (Jakes & White, 1970). The exceptionally low K/Rb ratios of some of the Barby lavas can be attributed to the very high Rb contents.

Within the area mapped calc-alkaline basaltic and basaltic-andesite lavas only seem to occur south of latitude $25^{\circ}30'$S, whereas north of this latitude outcrops are apparently almost entirely composed of tholeiitic basalt. Considering the limited chemical data available such a generalisation might well be unfounded, however, but it is evident from field work that the very highly potassic lavas with their characteristic red-brown colour are present in the south only.

In the southern outcrops of calc-alkaline lava flows, there is an erratic but overall increase in potassium content upward in the succession which is most apparent in the Klein Haremub-Aubures-Aruab area. In Fig. 20 an attempt has been made to represent this trend graphically.

It is also evident from the great thickness of the pyroxene trachyandesite member in the Aruab area (Section 3.1.2.1) that there is a maximum development of highly potassium-rich lava here. The three samples analysed from this member (BW 1557, BW 1558 and BW 1622) have very high K_2O contents, ranging from 3,45 to 5,57 per cent. Away from this area, toward the west and south, the K_2O content of the flows building up the sequence diminishes, but still remains abnormally high (e.g. BW 1615 from Vergenoeg and the samples of basic lava analysed by von Brunn, 1967).

	Barby Formation calc-alkaline lavas		Tholeiitic basalts (Prinz, 1967)		Alkali basalts (Prinz, 1967)		Basaltic calc-alkaline, Japan (Taylor & White, 1966)		Basaltic calc-alkaline, Island arcs (Jakes & White, 1972)	
	Average	Range	Average	Range	Average	Range	Average	Range	Average	Range
Rb	139	(36,2-294)	30	(2-125)	51	(1-150)	13	(9-17)		(10-30)
Sr	721	(446-1003)	450	(57-2000)	774	(18-2000)	360	(326-400)	330	
Ba	1363	(623-2386)	244	(10-1200)	444	(20-1000)	199	(170-220)	340	
Zr	151	(92-288)	108	(20-335)	138	(15-320)	98	(46-120)	100	
Nb	4,4	(2,7-11,6)					3,5	(2,1-5)		
Y	22	(17,3-34,0)	32	(12-90)	30	(10-100)	22	(13-26)	20	

Table 4. Comparison between the trace element contents (in parts per million) of the Barby calc-alkaline lavas and other basaltic rock types

	Barby calc-alkaline lavas	Guperas calc-alkaline lavas	1	2	3	4	5
Rb	36,2-294	53,4-57,0	50-80	30-187		35-180	75
Sr	446-1003	460-541	450-950	645-2010	∿1500	485-1115	700
Ba	623-2386	450-1021	400-1500	456-992	1000-1500	1360-2140	1000
Zr	92-288	90-304	130-220		∿300	110-145	100
Nb	2,7-11,6	2,0-10,8			10-15		
Y	17,3-34,0	18,1-43,8			∿30		

1. High-potassium calc-alkaline basaltic andesite lavas, Papua (Jakes & Smith, 1970)
2. High-potassium basaltic lavas, Viti Levu, Fiji (Gill, 1970)
3. Potassic basaltic-andesitic lavas, Nevada (Robinson, 1972)
4. Shoshonitic lavas, Wyoming (Nicholls & Carmichael, 1969)
5. Shoshonite, New Guinea (Jakes & White, 1972)

Table 5. Comparison between the trace element contents (in parts per million) of calc-alkaline lavas of the Barby and Guperas Formations and those of typical high-potassium calc-alkaline (shoshonite) suites

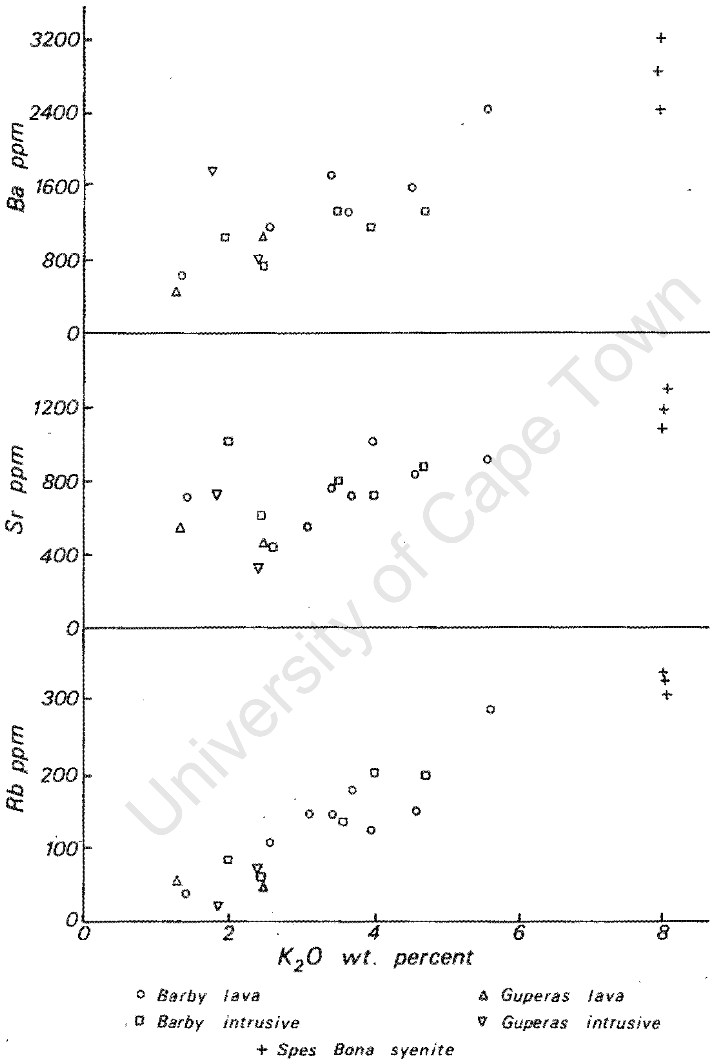


Fig. 18. Plots of Rb, Sr and Ba vs. K_2O showing the positive correlation between these trace elements and the potassium content in the calc-alkaline rocks of the Barby and Guperas Formation and Spes Bona Syenite

TiO_2 is low (0,67 to 1,52 per cent) and Fe_2O_3/FeO ratios vary widely, although there may be a correlation between the Fe_2O_3/FeO ratio and K_2O content possibly indicating an increase in the state of oxidation with increasing potassium content. High P_2O_5 is also a characteristic of the Barby calc-alkaline lavas.

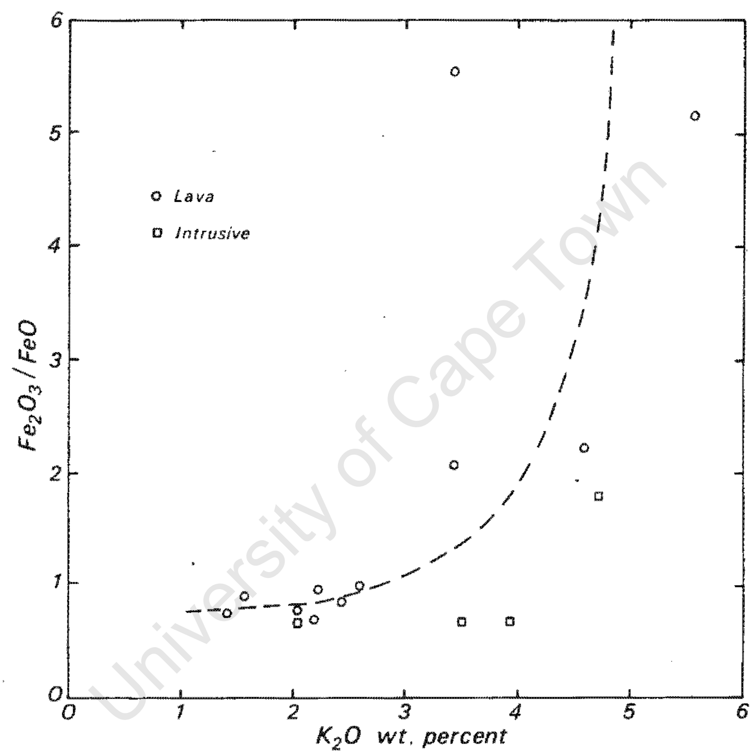


Fig. 17. Plot of the Fe_2O_3/FeO ratio vs. K_2O content for calc-alkaline rocks of the Barby Formation

Trace element chemistry of the Barby calc-alkaline lavas is very distinctive and characterised by very high contents of rubidium, strontium and barium[1]. All three of these elements show a marked correlation with K_2O, their respective contents increasing with increasing K_2O content (Fig. 18).

Rb ranges from 36,2 to 294 ppm and is, therefore, highly enriched when compared with average values for other major basaltic and basaltic-andesite suites (Table 4). Some of the analysed Barby lavas have Rb contents exceeding the maximum value of 160 ppm for all basaltic rocks (Prinz, 1967).

[1]All trace element concentrations given in the following Sections are in parts per million

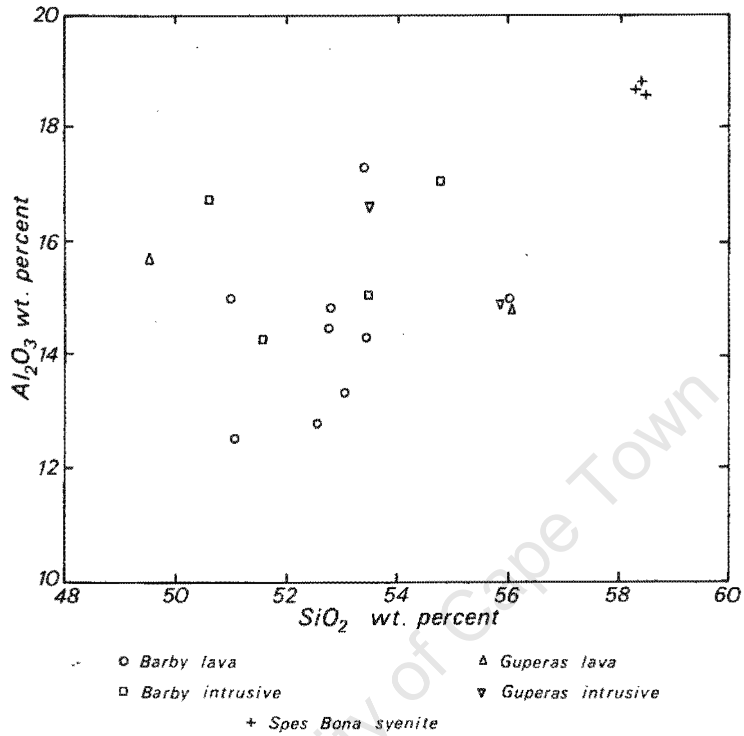


Fig. 16. Plot of Al_2O_3 vs. SiO_2 for the basic calc-alkaline and syenitic rocks of the Sinclair Group, showing a complete lack of correlation between the two oxides

The large variation in, and low average of, Al_2O_3 content present in the Barby rocks is in agreement with similar tendencies in the shoshonite association in general and, according to the data of Joplin (1968), Nicholls & Carmichael (1969), Jakes & White (1969 and 1972), and Jakes & Smith (1970), fluctuations of more than 5 per cent are common. Such wide fluctuations and low average values are in direct contrast to the majority of calc-alkaline assemblages where the average Al_2O_3 content is high (17,2 per cent, after Taylor & White, 1966) with fluctuations in the order of 2 per cent or less (Taylor, 1969).

The Na_2O content of the Barby calc-alkaline lavas has a fairly narrow range of 2,42 to 3,26 per cent and compares favourably with typical shoshonites and low silica calc-alkaline rocks in general (Jakes & White, 1972; Taylor & White, 1966). CaO varies from 5,07 to 8,22 per cent, a range which is appreciably lower than that for tholeiitic and alkali basalts (cf. data of Nockolds, 1954; Manson, 1967) and high-Al, low-Si calc-alkaline basalts (Jakes & White, 1972), but is entirely in accord with typical shoshonitic values (Table 3).

lavas of the Barby Formation with that of some shoshonite

BW 406	1	2	3	4	5	6	7	8
51,00	53,74	53,07	56,14	55,30	52,65	48,5	48,6	51,6
0,88	1,05	0,73	0,57	0,28	0,72	2,2	1,7	1,6
15,00	15,84	17,71	18,34	15,63	17,36	16,3	15,5	16,4
6,81	3,25	4,70	2,90	2,12	7,00	3,1	2,6	3,2
3,31	4,85	2,87	3,67	4,03	2,02	8,0	8,7	7,4
0,15	0,11				0,17	0,17	0,17	0,17
5,87	6,36	3,78	3,17	7,27	3,29	6,6	8,4	5,6
7,62	7,90	6,56	7,36	7,04	7,70	9,9	10,3	9,8
3,78	2,38	3,39	3,22	2,15	3,43	3,0	2,3	2,5
3,13	2,57	3,79	3,13	5,13	3,90	1,0	0,6	0,8
0,43	0,54				0,56	0,36	0,23	0,21
2,04		2,67	0,93	1,05	1,96	0,9	0,9	0,7
0,11								
100,13	98,59	99,27	99,43	100,00	100,76	100,03	100,00	99,98
147								
555								
95								
<1,7								
21								
0,83	1,08	1,11	1,02	2,38	1,14	0,33	0,26	0,32
2,06	0,67	1,64	0,79	0,53	3,46	0,39	0,30	0,43
6,91	4,95	7,18	6,35	7,28	7,33	4,0	2,9	3,3
er cent)								
	5,34	1,36	5,57		0,99			4,76
18,50	15,19	22,40	18,50	30,31	23,05	5,91	3,55	4,73
31,40	20,14	28,68	27,24	18,19	29,02	25,38	19,46	21,15
0,31								
14,72	24,95	21,92	26,35	17,85	20,46	28,06	30,20	31,17
8,46	4,47	4,44	4,24	7,13	5,88	7,81	8,10	6,71
7,32	3,15	3,77	2,76	5,06	5,08	4,78	4,95	3,94
	0,94	0,08	1,20	1,45		2,59	2,70	2,45
	12,69	5,64	5,14	9,10	3,11	4,97	11,20	10,01
	3,79	0,13	2,23	2,61		2,69	6,11	6,23
5,12				2,76		4,69	2,35	
				0,87		2,80	2,01	
8,62	4,71	6,81	4,20	3,07	5,03	4,49	3,77	4,64
1,65	1,97	1,37	1,07	0,53	1,35	4,13	3,19	3,00
0,86					3,53			
1,02	1,28				1,33	0,85	0,54	0,50
97,98	98,62	96,60	98,50	98,93	98,83	99,15	98,13	99,29

5. Shoshonite, Viti Levu, (Gill, 1970)
6. Average of 247 olivine alkaline basalts (Manson, 1967)
7. Average of 182 olivine tholeiitic basalts (Manson, 1967)
8. Average of 715 quartz basalts (Manson, 1967)

Table 3. Comparison of the chemical compositions and normative mineralogy of the calc-alkaline lavas of the Barby Formation with that of some shoshonite lavas and average basalts

	BW 1615	BW 381	BW 1558	BW 1557	BW 1622	BW 253	BW 409	BW 390	BW 406	1	2	3	4	5	6	7	8
SiO_2	55,98	53,45	53,45	51,15	52,62	52,79	53,11	52,79	51,00	53,74	53,07	56,14	55,30	52,65	48,5	48,6	51,6
TiO_2	0,88	0,90	0,73	0,83	0,67	0,82	1,52	0,83	0,88	1,05	0,73	0,57	0,28	0,72	2,2	1,7	1,6
Al_2O_3	14,96	17,24	14,30	12,54	12,80	14,53	13,33	14,85	15,00	15,84	17,71	18,34	15,63	17,36	16,3	15,5	16,4
Fe_2O_3	3,63	4,77	5,84	7,89	7,44	3,80	6,29	5,48	6,81	3,25	4,70	2,90	2,12	7,00	3,1	2,6	3,2
FeO	4,30	2,12	2,82	1,41	1,43	5,06	6,69	3,43	3,31	4,85	2,87	3,67	4,03	2,02	8,0	8,7	7,4
MnO	0,14	0,09	0,14	0,13	0,13	0,15	0,16	0,13	0,15	0,11				0,17	0,17	0,17	0,17
MgO	4,91	4,54	6,02	7,82	6,83	7,63	4,34	4,60	5,87	6,36	3,78	3,17	7,27	3,29	6,6	8,4	5,6
CaO	7,16	5,07	7,44	8,22	7,20	7,61	5,40	8,16	7,62	7,90	6,56	7,36	7,04	7,70	9,9	10,3	9,8
Na_2O	2,60	3,26	3,06	2,76	2,42	2,98	2,61	2,41	3,78	2,38	3,39	3,22	2,15	3,43	3,0	2,3	2,5
K_2O	2,46	4,63	3,45	3,69	5,57	1,41	2,62	3,99	3,13	2,57	3,79	3,13	5,13	3,90	1,0	0,6	0,8
P_2O_5	0,29	0,49	0,32	0,43	0,49	0,30	0,52	0,48	0,43	0,54				0,56	0,36	0,23	0,21
l.o.i.	1,67	2,25	1,52	1,96	1,48	2,11	2,35	2,29	2,04		2,67	0,93	1,05	1,96	0,9	0,9	0,7
H_2O^-	0,08	0,05	0,05	0,14	0,09	0,09	0,23	0,17	0,11								
	99,06	98,86	99,14	98,97	99,17	99,28	99,17	99,61	100,13	98,59	99,27	99,43	100,00	100,76	100,03	100,00	99,98
Rb	62	152	146	180	294	36	106	126	147								
Sr	609	835	754	665	904	719	446	1003	555								
Ba	759	1592	1716	1328	2386	623	1135										
Zr	179	159	110	111	92	114	288	122	95								
Nb	3,3	3,9	2,7	3,3	3,8	2,3	11,6	2,1	<1,7								
Y	25	19	20	20	17	19	34	23	21								
K/Rb	328	253	196	170	158	323	205										
K_2O/Na_2O	0,95	1,42	1,13	1,34	2,30	0,47	1,00	1,65	0,83	1,08	1,11	1,02	2,38	1,14	0,33	0,26	0,32
Fe_2O_3/FeO	0,84	2,25	2,07	5,59	5,20	0,75	0,94	1,60	2,06	0,67	1,64	0,79	0,53	3,46	0,39	0,30	0,43
$K_2O + Na_2O$	5,06	7,89	6,51	6,45	7,99	4,39	5,23	6,40	6,91	4,95	7,18	6,35	7,28	7,33	4,0	2,9	3,3
CIPW NORM (Wt. per cent)																	
Q	10,44	1,22	2,70			3,88	10,48	4,43		5,34	1,36	5,57		0,99			4,76
Or	14,54	27,36	20,39	21,81	32,91	8,33	15,48	23,58	18,50	15,19	22,40	18,50	30,31	23,05	5,91	3,55	4,73
Ab	22,00	27,58	25,89	23,35	20,48	25,21	22,08	20,39	31,40	20,14	28,68	27,24	18,19	29,02	25,38	19,46	21,15
Ne									0,31								
An	21,89	18,74	15,10	10,93	7,62	22,11	16,92	17,92	14,72	24,95	21,92	26,35	17,85	20,46	28,06	30,20	31,17
Di-Wo	4,90	1,34	8,23	11,29	10,40	5,71	2,70	8,11	8,46	4,47	4,44	4,24	7,13	5,88	7,81	8,10	6,71
Di-En	3,44	1,16	7,12	9,76	8,99	4,11	1,73	6,72	7,32	3,15	3,77	2,76	5,06	5,08	4,78	4,95	3,94
Di-Fs	1,05					1,09	0,79	0,39		0,94	0,08	1,20	1,45		2,59	2,70	2,45
Hy-En	8,79	10,15	7,87	5,57	2,61	14,89	9,08	4,71		12,69	5,64	5,14	9,10	3,11	4,97	11,20	10,01
Hy-Fs	2,67					3,96	4,15	0,27		3,79	0,13	2,23	2,61		2,69	6,11	6,23
Ol-Fo				2,91	3,79				5,12				2,76		4,69	2,35	
Ol-Fa													0,87		2,80	2,01	
Mt	5,26	4,52	7,44	2,52	3,05	5,51	9,12	7,95	8,62	4,71	6,81	4,20	3,07	5,03	4,49	3,77	4,64
Il	1,65	1,69	1,37	1,56	1,26	1,54	2,85	1,56	1,65	1,97	1,37	1,07	0,53	1,35	4,13	3,19	3,00
Hm		1,65	0,71	6,16	5,34				0,86					3,53			
Ap	0,69	1,16	0,76	1,02	1,16	0,71	1,23	1,14	1,02	1,28				1,33	0,85	0,54	0,50
	97,32	96,57	97,58	96,88	97,61	97,05	96,61	97,17	97,98	98,62	96,60	98,50	98,93	98,83	99,15	98,13	99,29

1. Shoshonite, New Guinea (Jakes and White, 1969)
2. Average of 7 shoshonites, Yellowstone Park, U.S.A. (Joplin, 1968)
3. Average of 3 shoshonites, Lesser Sunda Islands, Indonesia (Joplin, 1968)
4. Average of 3 shoshonites, Italy (Joplin, 1968)

5. Shoshonite, Viti Levu, (Gill, 1970)
6. Average of 247 olivine alkaline basalts (Manson, 1967)
7. Average of 182 olivine tholeiitic basalts (Manson, 1967.)
8. Average of 715 quartz basalts (Manson, 1967)

Table 13. Chemical composition and normative mineralogy of the Nubib granite and Gneiss (mylonitised granite)

	BW 1589	BW 1450	BW 1595	BW 1596	BW 1340	BW 1180	BW 1175	BW 682	BW 1448	BW 899
SiO_2	75,10	75,36	73,95	75,93	78,53	75,52	74,75	74,48	76,37	74,10
TiO_2	0,19	0,19	0,31	0,19	0,21	0,20	0,26	0,19	0,15	0,28
Al_2O_3	11,37	11,79	12,27	11,83	10,82	11,87	12,23	11,88	11,77	12,01
Fe_2O_3	1,24	1,40	1,59	0,72	0,37	1,42	1,51	0,91	1,24	2,13
FeO	1,13	0,87	0,91	0,68	0,87	0,87	0,39	0,64	0,70	1,04
MnO	0,05	0,03	0,06	0,04	0,02	0,05	0,03	0,03	0,01	0,06
MgO	0,11	0,25	0,29	0,27	0,23	0,16	0,27	0,19	0,05	0,16
CaO	0,93	0,51	0,83	0,58	0,14	0,58	0,74	0,53	0,22	0,91
Na_2O	3,16	3,67	3,57	3,45	3,47	3,94	3,34	3,58	3,88	3,52
K_2O	5,22	4,67	5,21	4,96	4,33	4,57	5,60	5,03	5,12	4,95
P_2O_5	0,02	0,02	0,05	0,02	0,03	0,02	0,03	0,02	0,01	0,04
l.o.i.	1,09	0,41	0,48	0,56	0,33	0,37	0,40	0,57	0,12	0,22
H_2O^-	0,07	0,08	0,04	0,07	0,02	0,06	0,08	0,09	0,04	0,04
	99,68	99,25	99,56	99,30	99,37	99,63	99,63	99,14	99,68	99,46
Rb	191	164	253	212	171	175	272	205	214	207
Sr	33	71	47	44	35	44	71	33	23	52
Ba	632	636	482	162	420	624	477	505	160	622
Zr	437	390	371	159	265	396	279	222	306	544
Nb	22,1	22,7	23,7	32,0	22,5	21,2	20,5	20,1	27,4	25,3
Y	115	103	114	65	93	105	93	106	142	135
K/Rb	227	237	171	194	211	217	171	204	198	199
CIPW NORM (Wt. per cent)										
Q	34,93	34,75	31,55	35,23	40,77	33,94	32,34	34,18	33,99	33,02
Or	30,85	27,60	30,79	29,31	25,59	27,01	33,09	29,72	30,26	29,25
Ab	26,74	31,05	30,21	29,19	29,36	33,34	28,26	30,29	32,04	29,78
An	1,43	1,91	2,07	2,15	0,50	1,21	1,84	1,50		2,36
C					0,24					
Ac									0,69	
Wo	0,23					0,02			0,08	0,33
Di Wo	1,05	0,20	0,72	0,25		0,63	0,68	0,42	0,34	0,46
Di En	0,27	0,14	0,62	0,15		0,40	0,59	0,28	0,12	0,40
Di Fs	0,83	0,04		0,09		0,19		0,10	0,23	
Hy En		0,48	0,10	0,53	0,57		0,08	0,19		
Hy Fs		0,15		0,33	0,99			0,07		
Mt	1,80	2,03	2,25	1,04	0,54	2,06	0,60	1,32	1,45	2,76
Il	0,36	0,36	0,58	0,36	0,39	0,38	0,49	0,36	0,28	0,53
Hm			0,04				1,09			0,23
Ap	0,05	0,05	0,12	0,05	0,07	0,05	0,07	0,05	0,02	0,09
	98,54	99,12	99,05	98,68	99,02	99,23	99,13	98,48	99,50	99,21

Table 14. Chemical composition and normative mineralogy of high - Ca rhyolites (BW 235, BW 1228, BW 1566) and low - Ca rhyolites (BW 1597, BW 733, BW 756) of the Barby Formation, granite (BW 1592) and Haremub granite (BW 1624)

	BW 235	BW 1228	BW 1566	BW 1597	BW 773	BW 756	BW 1592	BW 1624
SiO_2	66,93	67,49	69,36	75,24	74,81	74,40	74,44	71,97
TiO_2	0,81	0,79	0,63	0,26	0,19	0,27	0,24	0,26
Al_2O_3	13,68	13,88	13,73	11,78	12,02	12,14	12,25	13,75
Fe_2O_3	2,66	2,68	2,35	1,75	2,20	2,00	1,41	1,41
FeO	2,45	1,95	1,47	1,06	0,93	0,91	0,96	0,94
MnO	0,06	0,07	0,09	0,06	0,04	0,03	0,04	0,04
MgO	0,69	0,75	0,46	0,07	0,04	0,04	0,41	0,59
CaO	1,75	1,70	1,12	0,36	0,63	0,44	0,66	1,77
Na_2O	3,44	4,50	4,18	3,85	4,14	3,83	3,49	4,06
K_2O	5,60	4,53	5,75	4,66	4,39	4,78	4,77	3,81
P_2O_5	0,20	0,20	0,11	0,01	0,01	0,02	0,03	0,07
l.o.i.	0,98	0,88	0,27	0,25	0,30	0,47	0,60	0,73
H_2O^-	0,05	0,02	0,04	0,05	0,00	0,01	0,05	0,05
	99,30	99,44	99,56	99,40	99,70	99,34	99,35	99,45
Rb	169	141	181	132	134	187	167	69
Sr	238	195	155	50	86	43	46	380
Ba	1810	1351	1427	708	804	674	469	925
Zr	433	400	399	641	449	536	333	127
Nb	10,3	9,5	9,1	28,1	13,9	27,1	23,8	5,4
Y	48	44	44	86	88	121	119	11,1
K/Rb	275	267	264	292	272	213	238	456
CIPW NORM (Wt. per cent)								
Q	21,21	20,29	20,94	34,21	32,94	32,94	33,82	29,26
Or	33,09	26,77	33,98	27,54	25,94	28,25	28,19	22,51
Ab	29,11	38,07	35,37	32,57	35,03	32,41	29,53	34,35
An	5,35	4,30	1,72	1,10	1,25	1,82	3,08	8,05
C							0,22	
Ac								
Wo					0,64			
Di Wo	0,84	1,18	1,30	0,26	0,12	0,10		0,12
Di En	0,49	0,93	1,12	0,12	0,10	0,08		0,09
Di Fs	0,31	0,11		0,26	0,12	0,10		0,01
Hy En	1,23	0,93	0,02	0,05		0,02	1,02	1,38
Hy Fs	0,79	0,11		0,06			0,29	0,20
Mt	3,86	3,89	3,25	2,54	2,59	2,27	2,04	2,04
Il	1,52	1,48	1,18	0,49	0,36	0,51	0,45	0,49
Hm			0,11		0,41	0,40		
Ap	0,47	0,47	0,26	0,02	0,02	0,05	0,07	0,17
	98,27	98,53	99,25	99,22	99,52	98,95	98,71	98,67

Table 15. Chemical compositions and normative mineralogy of the rhyolite lava of the Guperas Formation

	BW 1403	BW 1333	BW 581	BW 1501	BW 1284	BW 1435	BW 1293	BW 1478
SiO_2	68,86	77,49	72,24	75,63	73,51	75,50	76,49	75,94
TiO_2	0,62	0,15	0,39	0,22	0,36	0,20	0,14	0,16
Al_2O_3	12,85	10,47	11,27	11,23	11,23	11,27	11,50	11,36
Fe_2O_3	2,32	0,75	1,85	1,26	2,44	1,26	1,51	1,66
FeO	2,04	1,13	2,98	1,48	1,75	1,11	0,44	0,70
MnO	0,09	0,03	0,10	0,05	0,07	0,05	0,01	0,04
MgO	0,71	0,14	0,21	0,06	0,24	0,20	0,10	0,11
CaO	1,60	0,35	1,36	0,89	0,85	0,89	0,13	0,59
Na_2O	3,27	3,62	3,01	3,57	2,18	3,44	2,34	2,98
K_2O	4,69	3,72	4,46	4,24	5,70	4,15	6,44	5,18
P_2O_5	0,18	0,01	0,09	0,01	0,07	0,02	0,01	0,02
l.o.i.	1,65	1,10	0,84	0,57	1,01	1,17	0,55	0,74
H_2O^-	0,06	0,11	0,07	0,00	0,10	0,07	0,09	0,08
	98,94	99,07	98,87	99,21	99,51	99,33	99,75	99,56
Rb	168	115	122	125	164	124	271	140
Sr	183	90	109	57	58	39	72	88
Ba	980	357	2087	1137	2208	367	238	1382
Zr	422	381	693	474	652	447	290	300
Nb	14,2	18,9	21,7	21,7	22,1	19,4	26,0	19,8
Y	68	91	99	96	97	76	149	74
K/Rb	231	269	304	281	289	277	198	307
CIPW NORM (Wt. per cent)								
Q	27,50	40,76	33,55	36,51	36,69	37,38	37,84	37,51
Or	27,71	21,98	26,36	25,06	33,68	24,52	38,06	30,61
Ab	27,67	30,63	25,47	30,21	18,44	29,11	19,80	25,21
An	6,56	1,34	4,07	2,10	3,76	3,06	0,58	2,33
C					0,10		0,47	
Ac								
Wo								
Di Wo	0,08	0,14	0,87	0,94		0,51		0,20
Di En	0,05	0,03	0,12	0,10		0,20		0,17
Di Fs	0,03	0,12	0,83	0,94		0,31		
Hy En	1,72	0,32	0,40	0,05	0,60	0,29	0,25	0,10
Hy Fs	0,97	1,15	2,67	0,48	0,75	0,45		
Mt	3,36	1,09	2,68	1,83	3,54	1,83	1,04	1,93
Il	1,16	0,28	0,73	0,41	0,68	0,38	0,26	0,30
Hm							0,79	0,33
Ap	0,43	0,02	0,21	0,02	0,17	0,05	0,02	0,05
	97,24	97,86	97,96	98,65	98,41	98,09	99,11	98,74

Table 16. Chemical composition and normative mineralogy of rhyolite intrusives (plugs and domes) of the Guperas Formation.

	BW 578	BW 1281	BW 1470	BW 1502	BW 1504	BW 1552
SiO_2	76,36	78,15	77,25	76,35	76,25	70,78
TiO_2	0,10	0,11	0,16	0,11	0,14	0,62
Al_2O_3	11,63	10,15	11,33	11,86	11,81	12,58
Fe_2O_3	1,76	1,57	1,81	1,75	0,88	2,05
FeO	0,18	0,13	0,20	0,29	1,44	2,71
MnO	0,01	0,03	0,03	0,02	0,05	0,12
MgO	0,03	0,07	0,15	0,08	0,05	1,02
CaO	0,40	0,12	0,40	0,37	0,38	2,64
Na_2O	4,17	2,33	2,37	2,81	3,78	3,97
K_2O	4,56	5,67	4,62	5,53	4,48	1,43
P_2O_5	0,02	0,01	0,02	0,00	0,01	0,18
l.o.i	0,46	0,49	0,81	0,68	0,59	1,30
H_2O^-	0,11	0,11	0,11	0,08	0,11	0,10
	99,79	98,94	99,26	99,93	99,97	99,50
Rb	164	169	134	173	111	44
Sr	17	40	79	21	20	235
Ba	86	491	1367	343	725	641
Zr	291	291	258	322	387	404
Nb	23,3	20,7	17,5	23,3	21,3	10,8
Y	92	131	73	126	106	62
K/Rb	231	279	286	265	334	268
CIPW NORM (Wt. per cent)						
Q	34,40	42,57	44,76	37,93	35,45	34,41
Or	26,95	33,51	27,30	32,68	26,47	8,45
Ab	34,44	19,71	20,05	23,78	31,98	33,59
An		0,49	1,85	1,84	1,82	11,92
C			1,75	0,58	0,08	0,13
Ac	0,74					
Wo	0,69					
Di Wo	0,09	0,01				
Di En	0,07	0,01				
Di Fs						
Hy En		0,16	0,37	0,20	0,12	2,54
Hy Fs					1,78	2,51
Mt	0,33	0,21	0,29	0,69	1,28	2,97
Il	0,19	0,21	0,30	0,21	0,26	1,16
Hm	1,28	1,43	1,61	1,27		
Ap	0,05	0,02	0,05		0,02	0,43
	99,23	98,33	98,33	99,18	99,26	98,11

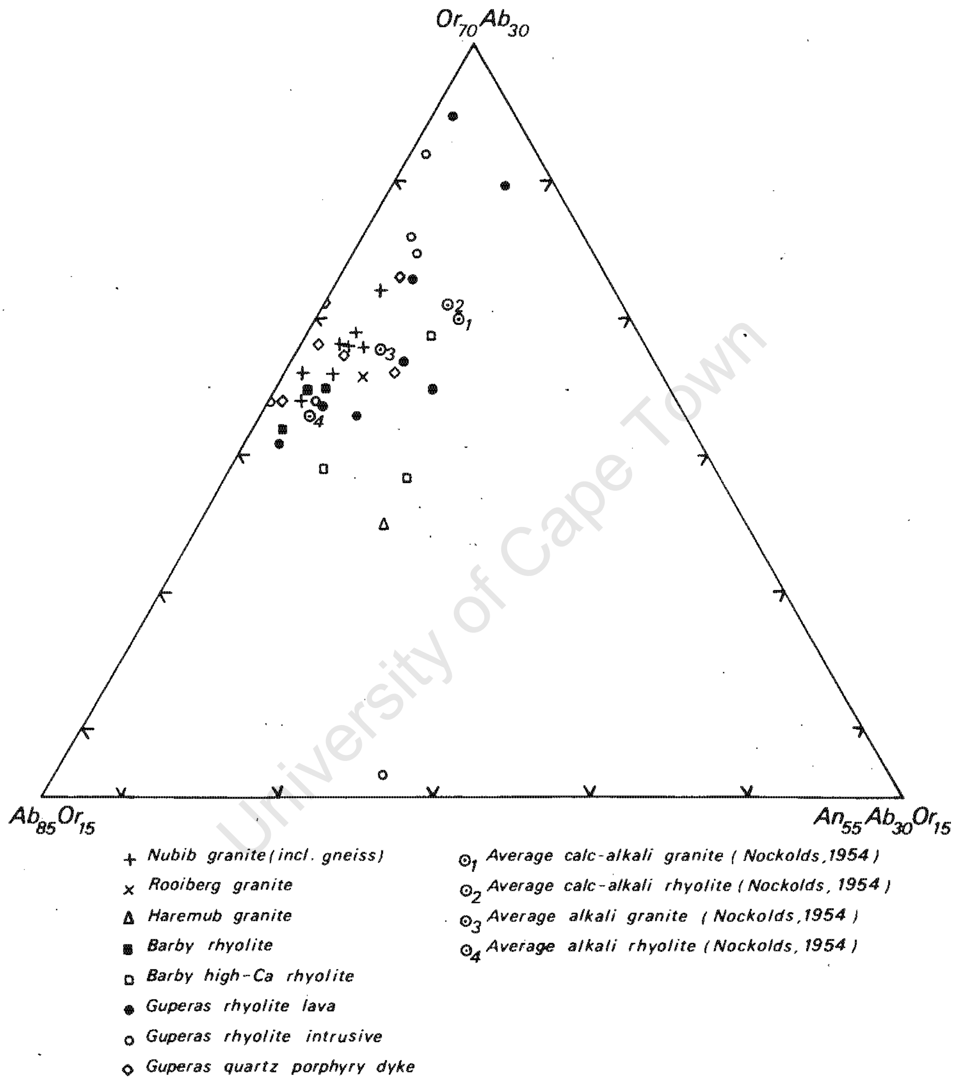


Fig. 21. Comparison of the felsic rocks of the Sinclair Group with those of average granites and rhyolites of Nockolds (1954) in terms of the normative mineral constituents Or, Ab, An

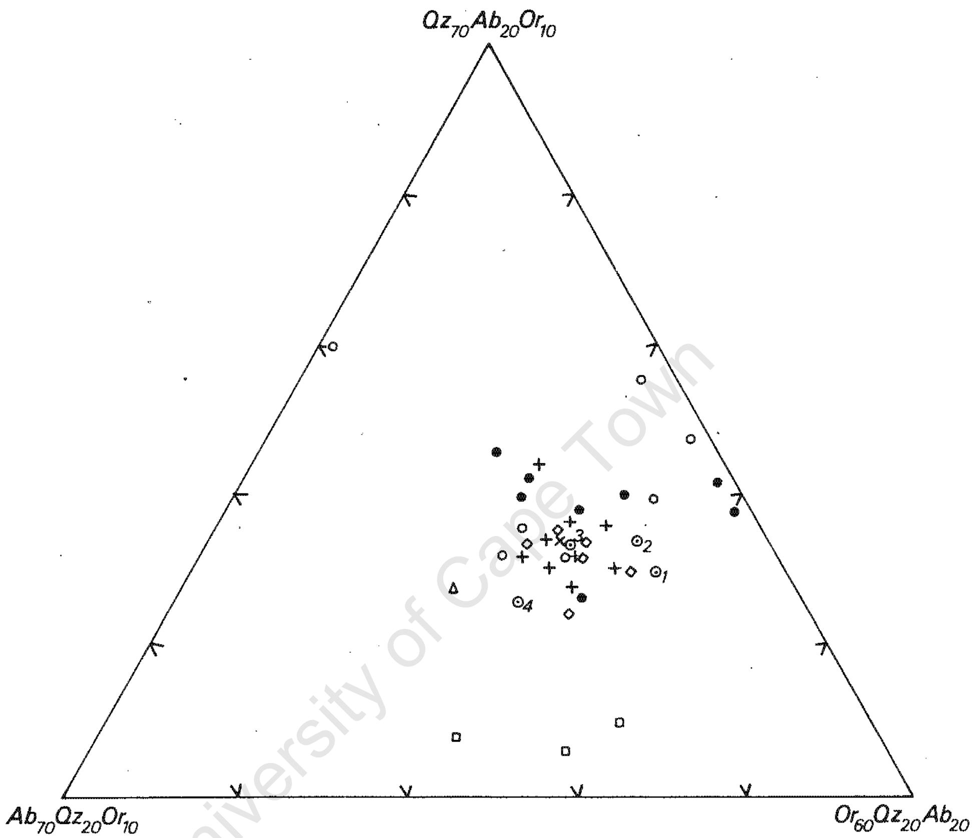


Fig. 22. Comparison of the felsic rocks of the Sinclair Group with those of average granites and rhyolites of Nockolds (1954) in terms of the normative mineral constituents Qz, Ab, Or. Symbols representing individual samples and average rocks have the same connotation as in Figure 21

The close compositional similarities noted above between the various low-Ca units are further substantiated by the trace element contents.

Trace element data from the low-Ca felsic units are compared with those of the average low-Ca granite of Turekian and Wedepohl (1961) and the average granite of Taylor (1964) (Table 18). On the basis of these comparisons, the following points can be noted with regard to the Sinclair rocks:

(i) Rb contents are 'normal' with the exception of the Guperas quartz-porphyry dykes which are slightly enriched.

(ii) Sr is very low.

(iii) Nb contents are 'normal' with the exception of the Guperas quartz-porphyry dykes which have slightly enriched concentrations of this element.

(iv) Zr and Y are markedly enriched.

(v) Ba shows a fairly erratic distribution but in general compares well with the average values of the literature.

Possible explanations accounting for these trace element anomalies will be presented in Section 3.4.5 which deals with the petrogenesis of these rocks. However, the low Sr content of the low-Ca units requires some comment at this stage. When the Sr values are plotted against the Ca contents of these rocks (Fig. 23) it is evident that they all fall either within or below the 'main granite sequence' of Turekian and Kulp (1956), thus indicating 'normal' to enriched values of Sr for the corresponding Ca concentrations. Turekian and Kulp (1956) found that the Sr content of granitic rocks, and to a somewhat lesser extent of rhyolites, is dependent on the Ca content, thus defining a positive relationship on a log Sr vs. log Ca diagram. On the basis of 170 analyses of granitic rocks they established a "main sequence" for this relationship, into which the vast majority of granitic rocks fall, due to the strong coherence between Ca and Sr at the low concentration levels of Ca found in granites.

K/Rb ratios for the Sinclair low-Ca felsic rocks range from 203 to 279 and, therefore, compare favourably with the values of Turekian and Wedepohl (1961) and Taylor (1964).

A single analysis of Nubib mylonitic gneiss (Tables 13 and 18), suggests that this rock is essentially similar in composition to the undeformed Nubib granite, thereby supporting the evidence presented in Section 3.1.5 for the development of a shear belt affecting various units of the Sinclair Group. A single analysis can hardly be regarded as representative of the gneisses as a whole, but in view of the fact that the compositional range of the nine analysed samples of Nubib granite is small, it is perhaps reasonable to assume that the gneisses will display a similar lack of variation, provided that the transformation from granite to gneiss was isochemical. This is likely when the low grade metamorphic and mechanical nature of the transformation (mylonitisation) is considered (Engel & Engel, 1958).

All the chemical features discussed above, that characterise the felsic units of the low-Ca grouping, and the Nubib granite in particular, are well displayed by the gneiss. The only element that differs appreciably between gneiss and granite is Zr, which is higher in the former by a factor of 1,7. However, due to the poor sampling statistics concerning the gneiss, it is impossible to decide whether this difference is real or not. The CIPW norm, calculated from the gneiss analysis is presented in Tables 13 and 18, together with norms for the Nubib granites. Once again, no significant differences between the two are apparent.

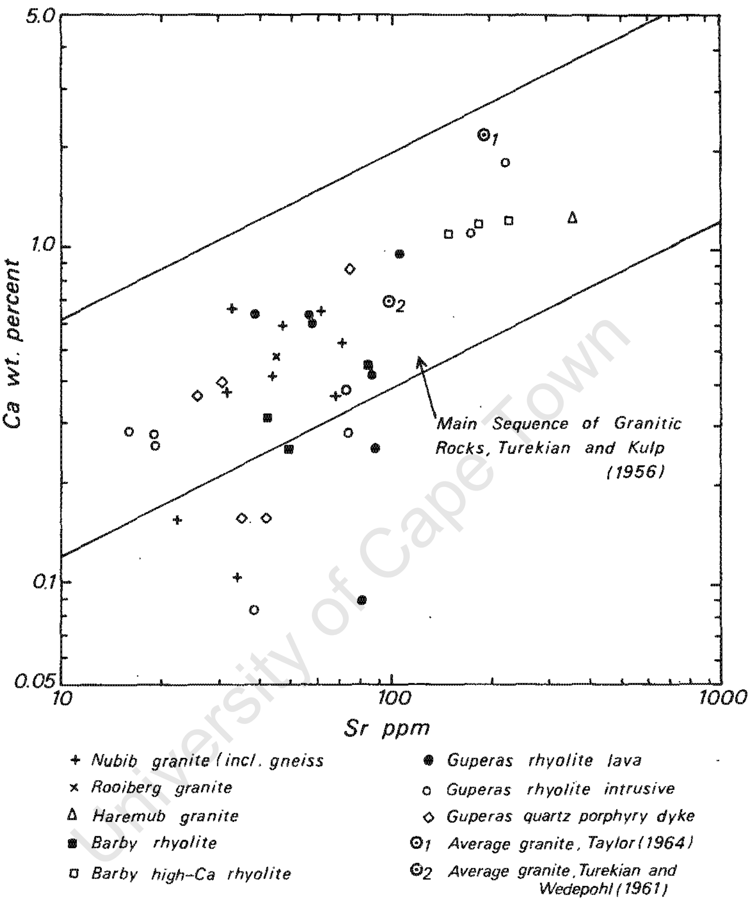


Fig. 23. Logarithmic plot of Sr vs. Ca for the felsic rock units of the Sinclair Group

3.3.3.2 High-Ca Units

The high-Ca rhyolites of the Barby, belonging to this grouping, compare well with the average alkali granite and rhyolite of Nockolds (1954) (Table 18). The CaO content of the Barby rocks are, however, notably higher, as is the Fe_2O_3 content. SiO_2 contents are relatively low.

The Haremub granite, which is also included in this grouping, displays granodioritic tendencies, having an excess of Na_2O over K_2O, an enriched CaO content when compared with average calc-alkali and alkali granites, and a fairly low SiO_2 content. This is in agreement with the observation in the field that the Haremub granite is heterogeneous, grading into granodioritic rock in a few localities.

Although somewhat more variable than those of the low-Ca units, trace element contents are fairly distinctive and further serve to separate these felsic rocks from those of the low-Ca group. Compared with the trace element contents of the average granites of Turekian and Wedepohl (1961) and Taylor (1964), the high-Ca felsic units are characterised by the following:

(i) 'Normal' to low Rb concentrations (Barby high-Ca rhyolites and Haremub granite respectively).

(ii) Due to the strong coherence between Sr and Ca at low concentrations of Ca (Turekian & Kulp, 1956), the high-Ca rocks still plot within the 'main sequence of granitic rocks' (Turekian & Kulp, 1956) on the log Sr vs. Ca diagram (Fig. 23). The Sr contents can, therefore, be regarded as 'normal'.

(iii) Considerably depleted Nb contents.

(iv) Enriched Zr contents.

(v) Y contents are 'normal' to low.

(vi) Ba contents are variable but slightly high.

K/Rb ratios for the high-Ca rhyolites average at 269 and are normal to slightly high when compared with average granite compositions. The Haremub granite has a very high K/Rb ratio, probably due to the depleted Rb content.

As for the low-Ca units, possible explanations for any anomalies in the trace element contents of the high-Ca units as well as for any chemical differences from the low-Ca units, will be presented in Section 3.4.5 which deals with the petrogenesis of these rocks.

Since only one sample of Haremub granite was analysed, it might be considered that the chemical characteristics of this granite as discussed above are possibly not typical of the granite as a whole. The Haremub granite is almost certainly a correlate of the Kotzerus granite to the south (Section 4) of which three chemical analyses are available (von Brunn, 1967). These analyses show compositions that characterise the geochemistry of the Haremub granite and thus provide a basis for the separation from the low-Ca units. It can be concluded, therefore, that the single analysis of Haremub granite, while possibly not being fully representative of the composition of the entire granite mass, does at least display compositional tendencies which might be considered characteristic of this intrusive phase.

The relative contents (without reference to average granitic and rhyolitic compositions) of the various oxides and elements that serve to distinguish between the low-Ca and high-Ca felsic units of the Sinclair may be summarised as follows:

Low-Ca units	High-Ca units
Low CaO	High CaO
High SiO_2	Low SiO_2
Low Al_2O_3	High Al_2O_3
Low MgO	High MgO
Low Sr	High Sr
High Nb	Low Nb
High Y	Low Y

Sample BW 1552, a quartz-felspar porphyry from a small intrusive body of the Guperas Formation, has been excluded from the averages in Table 18 and from the discussion above. On Figures 21 and 22 this sample plots away from any of the other felsic rocks of the Sinclair and does not appear to fit into either of the groups discussed above. The high-Ca character, together with relatively high MgO, Sr and Zr contents and low Rb, Nb, and Y contents, might suggest that the rock belongs to the high-Ca group, but the K_2O content is very low and not consistent with the high SiO_2 content. This low K_2O also rules out a favourable comparison with average granodiorite or rhyodacitic compositions. As discussed in Section 3.4.5, it is probable that this rock is the result of extreme contamination, possibly combined with fractionation, of a basic magma, and cannot, therefore, be included in either of the two groups discussed in this Section.

3.4 PETROGENESIS AND GEOTECTONIC SETTING

In the following Section various processes that could have given rise to the basic, syenitic and felsic magmas of the Sinclair Group are discussed in the light of available field, petrographic and chemical characteristics.

The concept of plate tectonics is employed to a considerable degree since many magma generating processes are thought to be controlled by geotectonic processes and the separation of the pure petrogenetic aspects from those of geotectonic processes is not considered reasonable, especially in view of the presence of the highly unique shoshonitic rock-types in the Sinclair. The presence of shoshonites and, therefore, the generation of potassium-rich magmas, is regarded as being indicative of a particular geotectonic environment as will be discussed presently.

The validity of applying plate tectonic theory to the evolution of late-Precambrian rock associations such as the Sinclair, might be considered questionable but in view of certain evidence (other than the presence of shoshonites) to be discussed in Section 5.2, it is considered reasonable to do so in the case of the Sinclair. The use of global tectonic theory has, furthermore, been applied by many workers in formulating plausible models for ancient tectonic and magmatic events, frequently dating back into the Precambrian (e.g. Burke and Dewey, 1971; Thorpe, 1972; Badham, 1973)

3.4.1 CALC-ALKALINE BASIC UNITS

Any process formulated for the production of calc-alkaline magmas such as those giving rise to the Sinclair basic rock units must also satisfactorily explain the highly enriched potassium contents of these rocks. Processes capable of enriching a basic magma (or rock) in potassium may be listed as follows:

(i) *In situ* metasomatism.

(ii) Wall-rock reaction at high pressures (upper mantle).

(iii) Wall-rock reaction at low pressures (crustal regions).

(iv) Crystal-liquid fractionation.

(v) Generation of a primary potassic magma.

Each of these processes will be discussed in turn in order to determine whether any one or more can be applied to the Sinclair rocks or not.

(i) *In Situ* metasomatism

Such a process cannot seriously be considered as being capable of enriching the Sinclair calc-alkaline lavas in potassium and associated elements since the high-K lavas are confined to mappable units (Section 3.1.2) and the chemical compositions of the basic potassic units are similar but distinctly different from other basic units of the region, e.g. tholeiitic basalts and gabbros.

(ii) Wall-rock reaction at high pressures

The process of wall-rock reaction at high pressures has been discussed by Green and Ringwood (1967) and involves the incorporation of the low melting fraction of mantle wall-rock into a basaltic magma, accompanied by the precipitation of the liquidus phase of the basalt. Such a low melting fraction of mantle peridotite is likely to involve the melting of potassium-rich accessory mineral phases such as phlogopite (Kushiro *et al.*, 1967) and perhaps amphibole (Lambert and Wyllie, 1968), apatite and zircon (Appleton, 1972), and will be highly enriched in the 'incompatible elements' (Green and Ringwood, 1967). The incompatible elements (including K, Ti, P, U, Th, Ba, Rb, Sr, Cs, Zr, Hf and the rare-earth elements) are those characterised by their inability to substitute to a great extent for elements forming the major mineral phases of the upper mantle (olivine, aluminous pyroxenes).

It could possibly be envisaged that for the Sinclair lavas a parent calc-alkaline basaltic magma of initially relatively low potassium content became enriched in potassium and 'potassium-like' elements by wall-rock reaction during slow ascent to the surface. Varying potassium contents would then be the result of varying speed of ascent, initial temperature of the different magma batches, and the thickness of mantle through which the magma travelled.

However, several objections can be raised against such a hypothesis. At high pressures, ($<\sim$35 km depth) plagioclase is not a stable phase in the mantle wall-rock mineralogy and strontium behaves as an incompatible element since it does not readily enter the pyroxene structure. At low pressures, plagioclase is a stable phase and strontium, because of its substitution for calcium in plagioclase, will behave as a compatible element. According to Green and Ringwood (1967), this difference in partition behaviour of strontium at low and high pressures makes it possible to use ratios such as K/Sr and Rb/Sr to distinguish between the effects of deep-level wall-rock reaction and a shallow-level process, i.e. rocks enriched in incompatible elements at low pressure should be characterised by rapid increases in such ratios as K/Sr and Rb/Sr through quite large variations in overall incompatible element abundance (Green and Ringwood, 1967, p. 179).

Applying this principle to the Sinclair potassic rocks (Table 19), it is apparent that in general there are wide variations in the K/Sr and Rb/Sr ratios as a result of very large variations in the abundance of the most prominent incompatible element, potassium. Therefore, if wall-rock reaction is to be postulated, the data would favour a low pressure rather than a high pressure process. It should be noted, however, that despite these indications for low pressure reaction processes in a régime where strontium behaves as a compatible element, it does show some degree of enrichment in parallel with the potassium content (Section 3.3.1, Fig. 18). This aspect is discussed in more detail later in this Section.

The process of wall-rock reaction is accompanied by the complementary process of precipitation of the liquidus phase or phases. This infers, therefore, that the basaltic liquid will not only change in bulk composition by assimilation of the low melting fraction of the wall-rock, but also by the

Sample No.	K (wt. per cent)	K/Sr	Rb/Sr
BW 1615	2,04	34	0,10
BW 381	3,85	46	0,18
BW 1558	2,87	38	0,19
BW 1557	3,06	46	0,27
BW 1622	4,62	51	0,33
BW 253	1,17	16	0,05
BW 409	2,18	49	0,24
BW 390	3,12	31	0,12
BW 406	2,60	47	0,27
BW 1565	3,93	46	0,23
BW 1606	1,71	17	0,09
BW 1613	3,31	45	0,28
BW 1256	2,92	37	0,18
BW 1541	2,11	46	0,12
BW 483	1,08	20	0,10
BW 1439	1,97	62	0,22
BW 1390	1,54	21	0,03

Table 19. Potassium contents, K/Sr and Rb/Sr ratios for the calc-alkaline basic rock-types of the Sinclair Group.

process of fractional crystallisation.

Fractional crystallisation of a parent calc-alkaline magma would result in silica enrichment (Green and Ringwood, 1968) and although the Sinclair calc-alkaline rocks show a range of compositions from olivine-normative to quartz-normative (SiO_2 contents from 49,54 to 56,14 per cent), the degree of fractional crystallisation required to produce this amount of silica enrichment is relatively small when compared with that envisaged by Green and Ringwood as accompanying an extensive wall-rock reaction process. Furthermore, the element Ti, which should be concentrated by virtue of its incompatible behaviour during high-pressure upper mantle wall-rock reaction processes, is characteristically low in the Sinclair potassic rocks.

Gast (1968) has concluded that wall-rock reaction would not be capable of producing an alkali basalt from an olivine tholeiitic parent magma. Since the potassium contents of the Sinclair calc-alkaline rocks greatly exceeds that of average alkali basalts (Table 3 and Fig. 14), it is even less likely that these potassium-rich rocks could have formed by reaction with a peridotitic wall-rock.

The arguments presented above do not rule out high pressure wall-rock reaction as an accessory and contributing process in the formation of the Sinclair potassic magmas, but such a process was not the prime cause of the very high enrichment of the major incompatible elements in these magmas.

(iii) Wall-rock reaction at low pressures

According to Green and Ringwood (1967) it is possible that an environment suitable for extensive wall-rock reaction may occur under low pressure conditions. Such might be the case if repeated basic magmatic injections caused wall-rock temperatures in the lower crust to approach the magma temperature.

The erratic but general increase in potassium and associated elements in the Barby lavas upward in the succession might be considered evidence for such a high-level enrichment process, i.e. progressive heating of the lower crust with continued injection of basic magma could have led to an increased ability for the magma to assimilate crustal material. The presence of small, highly resorbed phenocrysts of olivine in quartz-normative potassic lavas such as BW 381, could be considered as additional evidence in favour of a low pressure assimilative process. For a number of reasons, however, a low-pressure wall-rock reaction process cannot satisfactorily account for the observed chemical characteristics of the Sinclair potassic calc-alkaline lavas.

The partial melting of lower crustal material of presumed average intermediate composition (Green and Ringwood, 1967; den Tex, 1965) would first produce liquids of quartz-normative character, possessing an appreciable quantity of normative albite in addition to normative orthoclase. The assimilation or hybridisation of liquids of such compositions by a basaltic magma would result in concomitant increases in silica and sodium as well as potassium, and such has certainly not been the case with the Sinclair basic magmas. The Na_2O content remains relatively stable when compared with the K_2O content (Fig. 14) and increases in K_2O are not necessarily accompanied by increases in SiO_2 (as illustrated by the very steep trend in the K_2O vs. SiO_2 diagram of Fig. 10). It is, therefore, difficult to see how the assimilation of 'granite' liquids could enrich a basaltic magma in potassium without appreciably affecting the bulk sodium and silica contents.

It has already been pointed out that at low pressures plagioclase is a major stable phase in the wall-rock mineralogy, and that Sr behaves as a 'compatible' element under these conditions. Since reaction with crustal wall-rock material would involve mainly the 'granitic components' whereas plagioclase would remain largely unaffected, as a refractory residue, the reaction process would not enrich the basaltic magma in strontium. This is contrary to the observation that the potassic rocks of the Sinclair are somewhat enriched in strontium despite the possible indications of a low-pressure wall-rock reaction process as indicated by the K/Sr and Rb/Sr ratios.

(iv) Crystal-liquid fractionation

The experimental work of Green and Ringwood (1968) has shown that during fractional crystallisation of calc-alkaline magma, the composition of the residual liquid will follow a typical calc-alkaline trend. Garnet and clinopyroxene are important co-existing liquidus mineral phases in magmas of calc-alkaline basaltic composition at high pressures and because of their sub-silicic character the residual liquid will be enriched in silica and alkalis and generally depleted in iron and magnesium. As shown by the very steep trend defined by the calc-alkaline rocks on the K_2O vs. SiO_2 plot (Fig. 10), the increase in potassium content (and, therefore, total alkali content) is not accompanied by as large a corresponding increase in silica as would be expected from the experimental results of Green and Ringwood (1968).

For example, Green and Ringwood (1968) found that, at 27 kb, an increase of approximately 2,8 per cent of SiO_2 is accompanied by a corresponding increase of about 2,3 per cent of K_2O. Since the Na_2O content of these rocks does not vary appreciably (Fig. 14) this increase in K_2O will closely approximate the increase in total alkalis for the Sinclair rocks, which is considerably higher (more than three times) than that which would be expected if a normal fractionation trend toward silica and alkali enrichment had been followed.

Furthermore, although the Sinclair potassic rocks fall within the calc-alkaline field by virtue of their low degree of iron enrichment relative to MgO with increasing total alkali content, the MgO contents do not decrease regularly with increasing alkali content as would be expected if a fractionation process was the cause of the high potassium contents (and, therefore, high total alkali content). For example, if samples BW 1557 and BW 253 are compared it is notable that not only are the alkali contents of these two rocks considerably different for approximately equal MgO contents, but sample BW 1557, which has the higher MgO content, is actually more highly enriched in K_2O, quite contrary to what would be expected in a normal calc-alkaline trend.

	MgO (wt. per cent)	$K_2O + Na_2O$ = Total alkalis (wt. per cent)
BW 1557	7,82	3,69 + 2,76 = 6,45
BW 253	7,63	1,41 + 2,98 = 4,39

It has already been mentioned that the Sinclair calc-alkaline basic rocks range from normatively undersaturated (olivine-normative) to oversaturated (quartz-normative) with respect to silica, so although normal calc-alkaline differentiation was not the prime cause of the high potassium contents, it seems likely that at least some small degree of fractionation following such a trend did take place.

It has been suggested by O'Hara and Yoder (1967) that at high pressures, eclogite fractionation of a primary hypersthene-normative picrite magma, formed by partial melting of a garnet-peridotite mantle at about 30-40 kb, will result in residual magmas characterised by high concentrations of K, Na, Sr, Ba, Rb, Ti, P, Mg plus high K/Na ratios. Such a residual magma, they claim, might well have the composition of a potassic mafic lava on eruption. The precipitation of garnet and clinopyroxene at high pressure from a liquid that is already silica-poor with respect to garnet-clinopyroxene mixtures is expected to decrease the proportion of SiO_2 in relation to CaO, Al_2O_3, MgO, FeO and alkalis (+ associated minor elements) in the residual liquid. These residual liquids are also likely, therefore, to contain abundant normative olivine, nepheline, 'dicalcium silicate', leucite and kalsilite.

If such a process is postulated for the origin of the Sinclair potassic magmas, however, numerous problems are encountered:

(a) The Sinclair potassic basaltic rocks are not nepheline normative but range from olivine-normative to quartz-normative (see Section 3.3.1). As mentioned above, the trend of eclogite fractionation as proposed by O'Hara and Yoder (1967) indicates that nepheline will appear in the norm as fractionation proceeds and potassium will be enriched whereas silica will be depleted. Furthermore, as shown in Fig. 10, the trend of the Sinclair rocks is rather towards absolute silica enrichment with increasing K_2O rather than relative depletion of SiO_2 with increasing K_2O as occurring in an eclogite fractionation process.

(b) MgO and, to a lesser extent, CaO contents of the Sinclair rocks are variable, but generally low, and do not show any sign of the marked enrichment that should accompany high pressure fractionation. It is possible, however, that medium pressure fractionation involving olivine and pyroxene during rise of the magma from the high pressure zone could have led to extensive depletion of Mg.

(c) TiO_2 values for the Sinclair rocks are characteristically low and remain stable over a wide range of K_2O contents. They certainly do not show the enrichment features following an increase in K_2O as predicted by O'Hara and Yoder (1967) for eclogite fractionation processes. In an attempt to account for similarly anomalously low TiO_2 contents in potassic lava, Appleton (1972) has proposed a possible depletion of Ti by low pressure fractionation of ulvospinel.

(d) It is not likely that high pressure fractionation of this type would result in magmas of a calc-alkaline nature due to the enrichment of both Mg and Fe accompanying the increase in alkalis. O'Hara and Yoder (1967) point out, however, that the increase in Fe/Mg ratio is nowhere as great as in the case of low pressure fractionation of basaltic liquids.

It can be concluded from the foregoing discussion that neither wall-rock reaction processes nor crystal-liquid fractionation processes could have produced the highly potassic Sinclair calc-alkaline rocks. This conclusion does not, however, eliminate at least high pressure (mantle) wall-rock reaction and calc-alkaline fractionation as having been contributing processes.

It would seem that the potassic nature of these rocks is either of primary origin or is the result of a fractionation process that has enriched the magma in potassium and associated minor elements, but not appreciably in silica and sodium. Since such a fractionation trend is not supported by experimental evidence it is probable that the highly potassic nature of the Sinclair rocks is a primary magmatic feature.

(v) Generation of a primary potassic magma

In the preceding Section the great similarities between the chemical and mineralogical characteristics of the potassic calc-alkaline rocks of the Sinclair and those of the 'shoshonite association' (Joplin, 1968) were pointed out. It was further suggested that the Sinclair calc-alkaline rocks probably belong to such an association.

Data concerning the nature and origin of the high-potassium calc-alkaline rocks or shoshonites are limited and direct experimental studies involving these rocks have not, to the writer's knowledge, been carried out. However, the recognition of these rocks as an essential part of the complete orogenic volcano-plutonic cycle has placed considerable significance on their presence, and a pattern in their occurrence is beginning to emerge.

Joplin (1968) noted that rocks of the shoshonite association seem to be particularly characteristic of newly stabilised or just consolidated orogenic regions. This is in agreement with more recent studies on shoshonitic occurrences (Jakes & White, 1969; Gill, 1970; Jolly, 1971) in which it has been established that high potassium basaltic rocks constitute the final stages in the magmatic evolution of island and continental margin arc structures, and invariably occupy the zone most remote from the oceanic and, therefore, unstable side of the arc region. The importance of potassium-rich igneous rocks associated with island and continental margin magmatic arcs has recently been demonstrated in many parts of the circum-Pacific orogenic belt and numerous examples can be cited:

(a) The Melanesian island arcs (Jakes & White, 1969; Jakes & Smith, 1970; Smith, 1972).

(b) Viti Levu, Fiji, island arc (Dickinson *et al.*, 1968; Gill, 1970).

(c) Andean continental margin magmatic arc (Lefèvre, 1973; Pichler & Weyl, 1973).

(d) Continental margin magmatic arc, western U.S.A. (Robinson, 1972).

(e) Puerto Rico, island arc (Jolly, 1971).

Island arc and continental margin magmatism appears to be the main process by which sialic crust is formed (Taylor, 1967; Jakes & White, 1971). Sialic continental plates have, therefore, developed through time by the progressive accretion of successive volcano-plutonic orogens of arc-type marginal to early shield nucleii (Dickinson, 1972). It follows that the evidence for ancient shoshonitic magmatic activity such as that which gave rise to the Sinclair shoshonites will be found in, (i) island arcs or arc remnants that have become accreted to, or incorporated within, continental crustal plates, and (ii) volcano-plutonic arcs developed at the margins, or 'paleo' margins of sialic crustal plates.

The compositional structure of island arcs has recently been reviewed by Jakes and White (1971, 1972) and, briefly, it appears that arcs are built up of three rock associations each having distinctive chemical characteristics. The earliest rocks are tholeiitic and have very low K, Rb, Ba and Sr contents with moderate degrees of iron enrichment. In the later stages of arc evolution, calc-alkaline rocks are erupted together with tholeiites and in the latest stages shoshonites appear together with tholeiites and calc-alkaline rocks. Throughout this development there is a progressive increase in K_2O content for any given SiO_2 content and K_2O/Na_2O ratio, and decrease in the amount of iron-enrichment and K/Rb ratio. The three associations are spatially arranged with the tholeiites closest to the trench region and the shoshonites farthest from the trench. The start of eruption of one successive association does not imply the cessation of the preceding one and the eruption of shoshonitic magmas during the late stages of arc development will, for example, be accompanied by tholeiitic and calc-alkaline activity. Relationships between the three associations are, therefore, expected to be entirely transitional.

It is not proposed to review modern plate tectonic theory here but seismic data (Sykes, 1966; Isacks *et al.*, 1968), morphology, gravity and heat-flow measurements indicate that large volumes of oceanic crust are being consumed along Benioff zones by the process of subduction beneath island arcs and 'active' continental margins. It is now widely accepted that the igneous activity producing island arcs and continental margin magmatic arcs is the result of this geotectonic process of subduction. Opinion differs, however, as to the exact nature of the role played by the subducted slab of oceanic crust in the generation of new magma.

Oceanic crust generally consists of the tholeiitic basaltic material with a thin veneer of ocean-floor sediment and possibly some alkali volcanoes (Jakes & White, 1970; Dickinson, 1972). On subduction the tholeiite undergoes a dehydration process and, through progressive metamorphism, will be transformed to amphibolite. At still deeper levels the amphibolite will convert to a garnet-pyroxene-quartz (eclogite) assemblage (Fitton, 1971; Boettcher, 1973). The experimental work of Allen *et al.* (1972) indicates that amphibolite will remain the stable hydrous mineral phase to depths of up to about 75 km, and at levels deeper than this and up to 175 km, phlogopite is the stable hydrous phase. Below about 175 km the mica is no longer stable.

Geotherms calculated for the upper parts of a subducted oceanic slab (Oxburgh & Turcotte, 1971, Toksöz *et al.*, 1971) indicate that partial melting of the upper surface of the descending slab may occur.

At shallow depths the partial melts produced will be in equilibrium with amphibole and, according to Fitton (1971), they will give rise to a magma suite showing a high degree of iron enrichment and corresponding to the 'island arc tholeiite series' (Jakes & White, 1970, 1972). The trace element chemistry of the melt is likely to reflect, in part, that of amphibole, and the K/Rb ratios will be low. Nicholls and Ringwood (1973) claim that partial melting of the subducted oceanic crust will not occur at these relatively shallow depths and propose that the water from dehydration of the descending slab will rise into the overlying mantle, thus increasing mobility, triggering off gravitational instability and causing diapirs of wet peridotite to rise and undergo partial melting in the wedge of mantle overlying the subduction zone. The olivine tholeiite magma produced in this way would then undergo olivine fractionation during ascent and erupt as representatives of the island arc tholeiite series.

Partial melting at greater depths, where the amphibole has been replaced by an eclogite assemblage, will be controlled by equilibrium with garnet and pyroxene, and the increasing importance of iron-rich garnet in the residuum with depth will result in the low degree of iron enrichment characteristic of the calc-alkaline series. Furthermore, the decreasing ratio of melt to residuum with increasing depth (Jakes & White, 1970) together with the inability of potassium to enter garnet and pyroxene will lead to an increased potassium content of the calc-alkaline melt.

It has also been proposed that the production of calc-alkaline magmas may not be the direct result of partial melting of the transformed oceanic crust (McBirney, 1969; Hamilton, 1969; Nicholls & Ringwood, 1973). Water from the dehydration reactions may rise into the overlying mantle where partial melting of wet peridotite or fractionation of basaltic magma under hydrous conditions at depths of about 30-40 km will produce a magma of typical calc-alkaline character (Green & Ringwood, 1968). Nicholls and Ringwood (1973) consider it likely that temperatures will exceed the solidus of eclogite under high water pressures within the depth interval of 100-150 km, and, in view of the experimental results of Green and Ringwood (1968), they feel that the liquids produced under these conditions will be calc-alkaline but of a highly siliceous (>65 per cent SiO_2) nature. These acidic liquids, which will also have a high K/Na ratio, could react with the mantle overlying the Benioff zone causing diapirs of wet pyroxenite to rise from depths of 100-150 km and undergo partial melting during ascent to produce typical calc-alkaline magmas. This scheme, a three-stage process for the derivation of the calc-alkaline suite, is also supported by Gill (1974).

The presence of shoshonite occurrences at distances furthest removed from the trench region of an island or continental margin magmatic arc, would infer that the production of potassic calc-alkaline melts in such regimes is associated with the partial melting of transformed oceanic crust (eclogite) at the deepest and most highly dehydrated levels of the subduction zone. This is in agreement with experimental evidence (Green & Ringwood, 1968) and accords well with the general trends in the chemical composition of the tholeiitic and calc-alkaline magmas discussed above, i.e. general increase in K_2O content, decreasing K/Rb ratio and decreasing amount of iron enrichment, with increasing depth to the subduction zone.

In view of the remarkable similarities between the potassic calc-alkaline rocks of the Sinclair Group and those of the shoshonite association, and considering the association of shoshonites with the orogenic calc-alkaline and tholeiitic rock series, it is proposed that the Sinclair calc-alkaline magmas represent primary melts resulting from processes possibly related to the subduction of oceanic crust in Late Precambrian times.

It is probable that much of the water contained within a subducted slab of oceanic lithosphere will have been lost by depths of about 150 km or possibly even less, either by dehydration or by melting reactions followed by the uprise of magma as discussed above for the generation of the calc-alkaline and arc tholeiite suites (Wyllie, 1973). The only water present at these deeper levels is likely to be contained in the stable,hydrous phase, phlogopite (Allen *et al.*, 1972). Any partial melting that takes place at depths of about 150 km or greater will, therefore, take place under relatively 'dry' conditions. According to the experimental studies of Green and Ringwood (1968), the 'dry' partial melting of quartz eclogite (transformed tholeiite) under conditions of high pressure will produce either andesite or basaltic andesite liquids depending on the degree of partial melting. Melting of 40-50 per cent will result in andesite with a 'normal' SiO_2 and alkali content whereas at the deeper levels of a subduction zone, where increasingly lower degrees of partial melting are to be expected (Jakes & White, 1970), liquids progressively poorer in silica and richer in alkalis will be produced (Green & Ringwood, 1968; Jakes & Smith, 1970). Such liquids then have compositions as typified by the shoshonites, including the Sinclair potassic calc-alkaline rocks.

As mentioned above, the stable hydrous mineral phase within quartz eclogite at the depths under consideration (>150 km), will be phlogopite which being highly potassic and of a low melting character, will enter the partial melts thus enriching them in potassium and increasing the K/Na ratio. The proportion of phlogopite in the total amount of liquid produced will, therefore, affect the K/Na ratio of the melt since in phlogopite this ratio is extremely high. It is probable that small degrees of partial melting (<20 per cent) will produce liquids with K/Na ratios greater than one (Jolly, 1971), as is the case with the Sinclair potassic basic rocks and shoshonites in general.

The retention of iron-rich garnet in the residuum of partially melted quartz-eclogite, as mentioned before, will result in a strongly calc-alkaline character for these potassic melts and the low K/Rb ratios of the potassic phase (phlogopite) will be reflected in low K/Rb ratios of the partial melts. Garnet and pyroxene, major constituents of the eclogite, also have low K/Rb ratios and, if entering the liquid phase, would also contribute toward a general lowering of the K/Rb ratios. Low K/Rb ratios are an important characteristic of the Sinclair shoshonites. It follows that trace elements that commonly substitute for potassium at high pressures (Rb, Ba and, to a lesser extent, Sr) will be highly concentrated in phlogopite and will, therefore, enter the melt together with potassium, resulting in the high concentration of these elements observed in the Sinclair calc-alkaline rocks (Section 3.3.1).

Using the models of Toksöz *et al.* (1971) for the thermal regime of a downgoing (subducted) slab, it appears that sufficiently high temperatures will occur in the upper parts of the slab at depths of about 150-175 km, to allow for low degrees of partial melting to take place. The liquidus temperature of dry calc-alkaline basaltic andesite at these pressures ($\sim$50 kb) can be derived by extrapolation of the data of Green and Ringwood (1968) and is about 1600°C. The presence of small amounts of water from the decomposition of phlogopite will lower this temperature considerably, thus increasing the chances for partial melting to take place.

Magma generation in a dehydrated environment such as that discussed above will result in magmas which, on reaching the surface, will erupt in a predominantly non-explosive manner, due to the low content of gaseous phases. This is certainly the case with the Sinclair calc-alkaline lavas since basic pyroclastics of primary origin are not at all common; most of the volcaniclastic deposits associated with these lavas are probably of very locally derived epiclastic origin. A certain amount of water contained within these magmas is, however, indicated by the frequent presence of amygdales. Amygdaloidal texture is only sparsely developed in the less potassium-rich flows and is generally prominent in the highly potassic red-coloured lavas of the pyroxene trachyandesite member as occurring on the farms Aubures and Aruab. This concomitant increase in vesicularity with potassium content may well be explained by the relative increase in the proportion of the hydrous phase phlogopite entering the melt. The amount of water provided by this mica will be relatively small but, in view of the low degrees of partial melting envisaged, it might well be sufficient to cause the observed vesicularity.

The higher Fe_2O_3/FeO ratios accompanying the higher potassium contents as noted in the Sinclair rocks could possibly be attributed to a relatively high P_{O_2}, the result of a slightly increasing water content accompanying the increase in the proportion of phlogopite incorporated into the melt. This accords well with the observation by Jakes and White (1972) that, in general, the proportion of more highly oxidised rocks increases away from the trench in a magmatic arc environment.

The general increase in the potassium content of the calc-alkaline lavas upward in the Barby succession, as discussed in Section 3.9.1, is in agreement with the observation that shoshonites are generated during late stages of magmatic arc development. It does not explain, however, why lava of a generally lower potassium content should be erupted during a later period - that of the Guperas Formation. Unfortunately, insufficient chemical data for the Guperas basic lavas are available to be certain of the true nature of the magmas that were produced during this younger period and it may be that the calc-alkaline lavas of the Guperas merely represent magmatic products generated by the continuation of the same subduction process that operated during development of the Barby. The great similarities in chemical composition between the basic lava types of the Barby and Guperas, and the coincidence in the distribution of the very highly potassic lavas of the two formations, suggests that the magma-generating processes were the same and were similarly distributed during the two periods. For example, the only known basic lava of the Guperas with a distinctly red colour occurs on the farm Guperas, i.e. in the same area in which the red-coloured and highly potassic Barby lavas attain their maximum development. Chemical analysis of the Guperas lava mentioned above (BW 1541) confirms that its potassium content is the highest of all Guperas basic rocks analysed and that it is calc-alkaline in character.

The general increase in the potassium content of the Barby basic lava pile between the Vergenoeg and Aubures/Aruab areas could suggest that the depth of magma generation, i.e. depth to the subduction zone, increases in a southeasterly direction. In the light of other evidence (Section 5) it is shown that this is a reasonable assumption, but it should also be noted that much of the Barby lava occurring in the area west of Helmeringhausen (von Brunn, 1967), to the south of the Aubures/Aruab area, has a lower potassium content than the highly potassic lavas of the latter area. However, it is probable that the lavas occurring west of Helmeringhausen are representative of the lower part of the Barby volcanic succession and, therefore, their observed relatively low potassium contents might be explained in the light of the vertical increase in the potassium content discussed above.

The above deductions concerning the geographic and stratigraphic distributions of the potassic basic lava types must, however, be regarded as highly tentative and far more chemical data than are presently available are needed to obtain a clear impression of the temporal and spatial relationships between the relatively potash-poor and potash-rich basic lava types.

It has been suggested (Dickinson & Hatherton, 1967; Dickinson, 1968; Hatherton & Dickinson, 1968, 1969; Dickinson, 1970) that the relationship between potash content of calc-alkaline suites and distance from trench zone may be used to determine the depth of magma generation in the region of a subduction zone and, therefore, to reconstruct the position and attitude of ancient seismic zones. A recent statistical study by Nielson and Stoiber (1973) of the relationship of potassium content of basaltic andesite and andesitic lavas to silica content and depth to the inclined seismic zone in magmatic arc systems has demonstrated that the K-h plots do not provide a reliable method for determining the depth of magma generation. They conclude that "if the potassium content is determined by depth at which melting occurs, then some other factor or group of factors sufficiently alters the potassium content to render it of doubtful use as a quantitative indicator of the depth of origin of the lava" (Neilson and Stoiber, 1973, p.6887).

It is, in any event, doubtful whether such a technique can be applied to the Sinclair calc-alkaline rocks due to the very high potassium content and the very steep slope on the K_2O vs. SiO_2 variation diagram. It is evident as a consequence of this very steep slope that unreasonable depths for magma generation are obtained when the Sinclair data are applied to the 'K-h' plots of Dickinson (1970). For example the K_{55} value for the Sinclair shoshonites (i.e. K_2O percentage at a value of 55 per cent SiO_2) is approximately 5,3. Using the 'K-h' plots (where h = depth to seismic zone) of Dickinson (1970), depths of magma generation are indicated as being in the order of 400-500 km. These depths are certainly grossly excessive and do not agree with the conclusions arrived at earlier for the generation of the Sinclair potassic magma in the depth region of about 150-175 km where phlogopite would be a stable phase in a 'dry' environment.

The normative olivine- to quartz-bearing character of the Sinclair calc-alkaline rocks could be entirely the result of varying degrees of partial melting were it not for the fact that, invariably, those rocks which have the highest silica contents are also those with the highest potassium contents. In terms of partial melting, as pointed out above, this is inconsistent with the expected chemical trends.

If, however, the Sinclair primary melts, probably of a dominantly olivine-normative character, underwent fractionation during ascent to the surface, the observed general concomitant increase in K_2O with SiO_2 can be accounted for. Fractionation would have followed a calc-alkaline trend of increasing SiO_2 and alkalis by the precipitation of garnet and clinopyroxene at high pressures together with plagiocase at lower pressures (Green & Ringwood, 1968). In many of the Sinclair rocks there is good evidence for the precipitation of olivine prior to eruption or intrusion and it is, therefore, probable that the crystallisation of olivine accompanied that of the other liquidus minerals at some stage during the ascent of the magma. The crystallisation of olivine might be expected to raise the Fe/Mg ratio of the melt and cause the fractionating liquid to follow a trend of high iron enrichment rather than a typical calc-alkaline trend of low iron enrichment. However, it is most likely that

Table 17. Chemical composition and normative mineralogy of quartz-porphyry dykes of the Guperas Formation

	BW 1548	BW 1404	BW 935	BW 1365	BW 1497	BW 466
SiO_2	74,69	76,81	71,98	76,63	76,45	75,48
TiO_2	0,17	0,06	0,30	0,07	0,16	0,16
Al_2O_3	12,26	11,77	13,11	11,84	11,38	12,07
Fe_2O_3	0,91	1,07	1,02	1,17	1,86	0,86
FeO	1,28	0,20	1,72	0,18	0,14	1,06
MnO	0,04	0,01	0,05	0,01	0,02	0,05
MgO	0,10	0,04	0,40	0,10	0,05	0,18
CaO	0,56	0,22	1,24	0,51	0,22	0,53
Na_2O	3,16	3,65	3,57	3,98	3 ,53	3,59
K_2O	5,68	4,89	4,99	4,61	5,25	5,05
P_2O_5	0,01	0,01	0,06	0,01	0,01	0,01
l.o.i.	0,58	0,49	0,91	0,46	0,25	0,57
H_2O^-	0,11	0,10	0,09	0,11	0,13	0,10
	99,55	99,32	99,44	99,68	99,45	99,71
Rb	215	296	215	354	131	182
Sr	33	38	77	27	54	76
Ba	601	199	760	85	958	889
Zr	353	166	339	172	257	228
Nb	24,3	57,3	18,3	55,9	23,8	15,7
Y	81	80	72	129	83	57
K/Rb	220	137	193	108	334	230
	CIPW NORM (Wt. per cent)					
Q	32,66	36,36	28,32	34,98	35,66	33,55
Or	33,56	28,90	29,49	27,24	31,02	29,84
Ab	26,74	30,88	30,21	33,67	29,31	30,37
An	2 ,50	1,03	5,01	0,83		1,91
C		0,10				
Ac					0,49	
Wo				0,39	0,28	
Di - Wo	0,09		0,31	0,29	0,14	0,27
Di - En	0,01		0,11	0,25	0,12	0,08
Di - Fs	0,08		0,21			0,20
Hy - En	0,23	0,10	0,89			0,36
Hy - Fs	1,32		1,72			0,87
Mt	1,32	0,51	1,48	0 ,41	0 ,06	1,25
Il	0,32	0,11	0,56	0,13	0,30	0,30
Hm		0,72		0,88	1,65	
Ap	0,02	0,02	0,14	0,02	0,02	0,02
	98,85	98,73	98,45	99,09	99,05	99,02

Table 18. Average chemical composition and normative mineralogy of the felsic rock units of the Sinclair Group compared with average granite and rhyolite compositions

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15
SiO_2	75,67	67,93	74,82	74,46	76,87	75,34	71,97	74,10	74,44	72,08	73,66	73,86	74,57	74,23	69,10
TiO_2	0,21	0,74	0,24	0,28	0,13	0,15	0,26	0,28	0,24	0,37	0,22	0,20	0,17	0,20	0,38
Al_2O_3	11,76	13,76	11,98	11,40	11,35	12,07	13,75	12,01	12,25	13,86	13,45	13,75	12,58	13,60	14,55
Fe_2O_3	1,16	2,56	1,97	1,63	1,56	1,15	1,41	2,13	1,41	0,86	1,25	0,78	1,30		
FeO	0,78	1,96	0,97	1,45	0,45	0,76	0,94	1,04	0,96	1,67	0,75	1,13	1,02	1,83 †	3,47 †
MnO	0,04	0,08	0,04	0,06	0,03	0,03	0,04	0,06	0,04	0,06	0,03	0,05	0,05	0,05	0,04
MgO	0,20	0,63	0,05	0,22	0,08	0,14	0,59	0,16	0,41	0,52	0,32	0,26	0,11	0,27	0,26
CaO	0,56	1,52	0,48	0,83	0,33	0,55	1,77	0,91	0,66	1,33	1,13	0,72	0,61	0,71	2,21
Na_2O	3,56	4,04	3,94	3,05	3,09	3,58	4,06	3,52	3,49	3,08	2,99	3,51	4,13	3,48	3,73
K_2O	4,97	5,29	4,61	4,82	4,97	5,08	3,81	4,95	4,77	5,46	5,35	5,13	4,73	5,06	4,02
P_2O_5	0,02	0,17	0,02	0,05	0,01	0,02	0,07	0,04	0,03	0,18	0,07	0,14	0,07	0,14	0,07
l.o.i	0,48	0,71	0,34	0,95	0,61	0,54	0,73	0,22	0,60	0,53	0,78	0,47	0,66		
H_2O	0,06	0,03	0,02	0,07	0,11	0,11	0,05	0,04	0,05						
	99,47	99,42	99,48	99,27	99,59	99,52	99,45	99,46	99,35	100,00	100,00	100,00	100,00	99,57	97,83
Rb	206	164	151	154	150	232	69	207	167					170	150
Sr	45	196	59	87	35	51	380	52	46					100	285
Ba	434	1529	729	1094	602	582	925	622	469					840	600
Zr	314	411	542	457	310	252	127	544	333					175	180
Nb	23,6	9,6	23,1	20,5	21,2	32,6	5,4	25,3	23,8					21	20
Y	104	45	98	94	109	84	11,1	135	119					40	40
K/Rb	203	269	259	270	279	204	456	199	238					247	223
Fe_2O_3/FeO	1,49	1,31	2,03	1,12	3,47	1,51	1,50	2,05	1,47	0,52	1,67	0,69	1,27		
	CIPW NORM (Wt. per cent)														
Q	34,67	20,84	33,40	36,00	39,08	33,79	29,26	33,02	33,82	31,45	33,07	34,03	31,02		
Or	29,37	31,26	27,24	28,48	29,37	30,02	22,51	29,25	28,19	32,62	31,61	29,61	27,95		
Ab	30,12	34,18	33,34	25,81	26,14	30,29	34,35	29,78	29,53	25,04	25,30	29,44	34,94		
An	1,43	3,79	1,39	3,18	1,57	1,87	8,05	2,36	3,08	4,89	5,15	2,06	1,82		
C					0,31				0,22	0,70	0,86	1,26			
Ac															
Wo			0,21					0,33							
Di - Wo	0,51	1,10	0,14	0,25		0,31	0,12	0,46					0,31		
Di - En	0,33	0,78	0,12	0,09		0,17	0,09	0,40					0,09		
Di - Fs	0,14	0,22		0,17		0,13	0,01						0,23		
Hy - En	0,17	0,78		0,45	0,20	0,18	1,38		1,02	1,25	0,80	0,60	0,18		
Hy - Fs	0,07	0,22		0,81		0,13	0,20		0,29	1,39	0,05	0,70	0,47		
Mt	1,68	3,71	2,58	2,36	1,18	1,67	2,04	2,76	2,04	1,26	1,81	1,36	1,88		
Il	0,39	1,39	0,45	0,53	0,24	0,28	0,49	0,53	0,45	0,56	0,41	0,32	0,23		
Hm			0,19		0,75			0,23							
Ap	0,05	0,40	0,05	0,12	0,02	0,05	0,17	0,09	0,07	0,33	0,17	0,26	0,17		
	98,93	98,67	99,11	98,25	98,86	98,89	99,17	99,21	98,71	99,46	99,23	99,64	99,29		

† Total Fe

1. Average of 9 Nubib granites
2. Average of 3 high-Ca rhyolites, Barby Formation
3. Average of 3 low-Ca rhyolites, Barby Formation
4. Average of 8 rhyolites lavas, Guperas Formation
5. Average of 5 rhyolites intrusives, Guperas Formation
6. Average of 6 quartz-porphyry dykes, Guperas Formation
7. Haremub granite (BW 1624)
8. Mylonitised Nubib granite (BW 899)
9. Rooiberg granite (BW 1592)
10. Average calc-alkali granite (Nockolds, 1954)
11. Average calc-alkali rhyolite (Nockolds, 1954)
12. Average alkali granite (Nockolds, 1954)
13. Average alkali rhyolite (Nockolds, 1954)
14. Average low-Ca granite (Turekian and Wedepohl, 1961)
15. Average granite (Taylor, 1964)

Table 13. Chemical composition and normative mineralogy of the Nubib granite and Gneiss (mylonitised granite)

	BW 1589	BW 1450	BW 1595	BW 1596	BW 1340	BW 1180	BW 1175	BW 682	BW 1448	BW 899
SiO_2	75,10	75,36	73,95	75,93	78,53	75,52	74,75	74,48	76,37	74,10
TiO_2	0,19	0,19	0,31	0,19	0,21	0,20	0,26	0,19	0,15	0,28
Al_2O_3	11,37	11,79	12,27	11,83	10,82	11,87	12,23	11,88	11,77	12,01
Fe_2O_3	1,24	1,40	1,59	0,72	0,37	1,42	1,51	0,91	1,24	2,13
FeO	1,13	0,87	0,91	0,68	0,87	0,87	0,39	0,64	0,70	1,04
MnO	0,05	0,03	0,06	0,04	0,02	0,05	0,03	0,03	0,01	0,06
MgO	0,11	0,25	0,29	0,27	0,23	0,16	0,27	0,19	0,05	0,16
CaO	0,93	0,51	0,83	0,58	0,14	0,58	0,74	0,53	0,22	0,91
Na_2O	3,16	3,67	3,57	3,45	3,47	3,94	3,34	3,58	3,88	3,52
K_2O	5,22	4,67	5,21	4,96	4,33	4,57	5,60	5,03	5,12	4,95
P_2O_5	0,02	0,02	0,05	0,02	0,03	0,02	0,03	0,02	0,01	0,04
l.o.i.	1,09	0,41	0,48	0,56	0,33	0,37	0,40	0,57	0,12	0,22
H_2O^-	0,07	0,08	0,04	0,07	0,02	0,06	0,08	0,09	0,04	0,04
	99,68	99,25	99,56	99,30	99,37	99,63	99,63	99,14	99,68	99,46
Rb	191	164	253	212	171	175	272	205	214	207
Sr	33	71	47	44	35	44	71	33	23	52
Ba	632	636	482	162	420	624	477	505	160	622
Zr	437	390	371	159	265	396	279	222	306	544
Nb	22,1	22,7	23,7	32,0	22,5	21,2	20,5	20,1	27,4	25,3
Y	115	103	114	65	93	105	93	106	142	135
K/Rb	227	237	171	194	211	217	171	204	198	199
					CIPW NORM (Wt. per cent)					
Q	34,93	34,75	31,55	35,23	40,77	33,94	32,34	34,18	33,99	33,02
Or	30,85	27,60	30,79	29,31	25,59	27,01	33,09	29,72	30,26	29,25
Ab	26,74	31,05	30,21	29,19	29,36	33,34	28,26	30,29	32,04	29,78
An	1,43	1,91	2,07	2,15	0,50	1,21	1,84	1,50		2,36
C					0,24					
Ac									0,69	
Wo	0,23					0,02			0,08	0,33
Di Wo	1,05	0,20	0,72	0,25		0,63	0,68	0,42	0,34	0,46
Di En	0,27	0,14	0,62	0,15		0,40	0,59	0,28	0,12	0,40
Di Fs	0,83	0,04		0,09		0,19		0,10	0,23	
Hy En		0,48	0,10	0,53	0,57		0,08	0,19		
Hy Fs		0,15		0,33	0,99			0,07		
Mt	1,80	2,03	2,25	1,04	0,54	2,06	0,60	1,32	1,45	2,76
Il	0,36	0,36	0,58	0,36	0,39	0,38	0,49	0,36	0,28	0,53
Hm			0,04				1,09			0,23
Ap	0,05	0,05	0,12	0,05	0,07	0,05	0,07	0,05	0,02	0,09
	98,54	99,12	99,05	98,68	99,02	99,23	99,13	98,48	99,50	99,21

Table 14. Chemical composition and normative mineralogy of high - Ca rhyolites (BW 235, BW 1228, BW 1566) and low - Ca rhyolites (BW 1597, BW 733, BW 756) of the Barby Formation, granite (BW 1592) and Haremub granite (BW 1624)

	BW 235	BW 1228	BW 1566	BW 1597	BW 773	BW 756	BW 1592	BW 1624
SiO_2	66,93	67,49	69,36	75,24	74,81	74,40	74,44	71,97
TiO_2	0,81	0,79	0,63	0,26	0,19	0,27	0,24	0,26
Al_2O_3	13,68	13,88	13,73	11,78	12,02	12,14	12,25	13,75
Fe_2O_3	2,66	2,68	2,35	1,75	2,20	2,00	1,41	1,41
FeO	2,45	1,95	1,47	1,06	0,93	0,91	0,96	0,94
MnO	0,06	0,07	0,09	0,06	0,04	0,03	0,04	0,04
MgO	0,69	0,75	0,46	0,07	0,04	0,04	0,41	0,59
CaO	1,75	1,70	1,12	0,36	0,63	0,44	0,66	1,77
Na_2O	3,44	4,50	4,18	3,85	4,14	3,83	3,49	4,06
K_2O	5,60	4,53	5,75	4,66	4,39	4,78	4,77	3,81
P_2O_5	0,20	0,20	0,11	0,01	0,01	0,02	0,03	0,07
l.o.i.	0,98	0,88	0,27	0,25	0,30	0,47	0,60	0,73
H_2O^-	0,05	0,02	0,04	0,05	0,00	0,01	0,05	0,05
	99,30	99,44	99,56	99,40	99,70	99,34	99,35	99,45
Rb	169	141	181	132	134	187	167	69
Sr	238	195	155	50	86	43	46	380
Ba	1810	1351	1427	708	804	674	469	925
Zr	433	400	399	641	449	536	333	127
Nb	10,3	9,5	9,1	28,1	13,9	27,1	23,8	5,4
Y	48	44	44	86	88	121	119	11,1
K/Rb	275	267	264	292	272	213	238	456
				CIPW NORM (Wt. per cent)				
Q	21,21	20,29	20,94	34,21	32,94	32,94	33,82	29,26
Or	33,09	26,77	33,98	27,54	25,94	28,25	28,19	22,51
Ab	29,11	38,07	35,37	32,57	35,03	32,41	29,53	34,35
An	5,35	4,30	1,72	1,10	1,25	1,82	3,08	8,05
C							0,22	
Ac								
Wo					0,64			
Di Wo	0,84	1,18	1,30	0,26	0,12	0,10		0,12
Di En	0,49	0,93	1,12	0,12	0,10	0,08		0,09
Di Fs	0,31	0,11		0,26	0,12	0,10		0,01
Hy En	1,23	0,93	0,02	0,05		0,02	1,02	1,38
Hy Fs	0,79	0,11		0,06			0,29	0,20
Mt	3,86	3,89	3,25	2,54	2,59	2,27	2,04	2,04
Il	1,52	1,48	1,18	0,49	0,36	0,51	0,45	0,49
Hm			0,11		0,41	0,40		
Ap	0,47	0,47	0,26	0,02	0,02	0,05	0,07	0,17
	98,27	98,53	99,25	99,22	99,52	98,95	98,71	98,67

Table 17. Chemical composition and normative mineralogy of quartz-porphyry dykes of the Guperas Formation

	BW 1548	BW 1404	BW 935	BW 1365	BW 1497	BW 466
SiO_2	74,69	76,81	71,98	76,63	76,45	75,48
TiO_2	0,17	0,06	0,30	0,07	0,16	0,16
Al_2O_3	12,26	11,77	13,11	11,84	11,38	12,07
Fe_2O_3	0,91	1,07	1,02	1,17	1,86	0,86
FeO	1,28	0,20	1,72	0,18	0,14	1,06
MnO	0,04	0,01	0,05	0,01	0,02	0,05
MgO	0,10	0,04	0,40	0,10	0,05	0,18
CaO	0,56	0,22	1,24	0,51	0,22	0,53
Na_2O	3,16	3,65	3,57	3,98	3,53	3,59
K_2O	5,68	4,89	4,99	4,61	5,25	5,05
P_2O_5	0,01	0,01	0,06	0,01	0,01	0,01
l.o.i.	0,58	0,49	0,91	0,46	0,25	0,57
H_2O^-	0,11	0,10	0,09	0,11	0,13	0,10
	99,55	99,32	99,44	99,68	99,45	99,71
Rb	215	296	215	354	131	182
Sr	33	38	77	27	54	76
Ba	601	199	760	85	958	889
Zr	353	166	339	172	257	228
Nb	24,3	57,3	18,3	55,9	23,8	15,7
Y	81	80	72	129	83	57
K/Rb	220	137	193	108	334	230
CIPW NORM (Wt. per cent)						
Q	32,66	36,36	28,32	34,98	35,66	33,55
Or	33,56	28,90	29,49	27,24	31,02	29,84
Ab	26,74	30,88	30,21	33,67	29,31	30,37
An	2,50	1,03	5,01	0,83		1,91
C		0,10				
Ac					0,49	
Wo				0,39	0,28	
Di - Wo	0,09		0,31	0,29	0,14	0,27
Di - En	0,01		0,11	0,25	0,12	0,08
Di - Fs	0,08		0,21			0,20
Hy - En	0,23	0,10	0,89			0,36
Hy - Fs	1,32		1,72			0,87
Mt	1,32	0,51	1,48	0,41	0,06	1,25
Il	0,32	0,11	0,56	0,13	0,30	0,30
Hm		0,72		0,88	1,65	
Ap	0,02	0,02	0,14	0,02	0,02	0,02
	98,85	98,73	98,45	99,09	99,05	99,02

Table 18. Average chemical composition and normative mineralogy of the felsic rock units of the Sinclair Group compared with average granite and rhyolite compositions

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15
SiO_2	75,67	67,93	74,82	74,46	76,87	75,34	71,97	74,10	74,44	72,08	73,66	73,86	74,57	74,23	69,10
TiO_2	0,21	0,74	0,24	0,28	0,13	0,15	0,26	0,28	0,24	0,37	0,22	0,20	0,17	0,20	0,38
Al_2O_3	11,76	13,76	11,98	11,40	11,35	12,07	13,75	12,01	12,25	13,86	13,45	13,75	12,58	13,60	14,55
Fe_2O_3	1,16	2,56	1,97	1,63	1,56	1,15	1,41	2,13	1,41	0,86	1,25	0,78	1,30		
FeO	0,78	1,96	0,97	1,45	0,45	0,76	0,94	1,04	0,96	1,67	0,75	1,13	1,02	1,83 †	3,47 †
MnO	0,04	0,08	0,04	0,06	0,03	0,03	0,04	0,06	0,04	0,06	0,03	0,05	0,05	0,05	0,04
MgO	0,20	0,63	0,05	0,22	0,08	0,14	0,59	0,16	0,41	0,52	0,32	0,26	0,11	0,27	0,26
CaO	0,56	1,52	0,48	0,83	0,33	0,55	1,77	0,91	0,66	1,33	1,13	0,72	0,61	0,71	2,21
Na_2O	3,56	4,04	3,94	3,05	3,09	3,58	4,06	3,52	3,49	3,08	2,99	3,51	4,13	3,48	3,73
K_2O	4,97	5,29	4,61	4,82	4,97	5,08	3,81	4,95	4,77	5,46	5,35	5,13	4,73	5,06	4,02
P_2O_5	0,02	0,17	0,02	0,05	0,01	0,02	0,07	0,04	0,03	0,18	0,07	0,14	0,07	0,14	0,07
l.o.i	0,48	0,71	0,34	0,95	0,61	0,54	0,73	0,22	0,60	0,53	0,78	0,47	0,66		
H_2O	0,06	0,03	0,02	0,07	0,11	0,11	0,05	0,04	0,05						
	99,47	99,42	99,48	99,27	99,59	99,52	99,45	99,46	99,35	100,00	100,00	100,00	100,00	99,57	97,83
Rb	206	164	151	154	150	232	69	207	167					170	150
Sr	45	196	59	87	35	51	380	52	46					100	285
Ba	434	1529	729	1094	602	582	925	622	469					840	600
Zr	314	411	542	457	310	252	127	544	333					175	180
Nb	23,6	9,6	23,1	20,5	21,2	32,6	5,4	25,3	23,8					21	20
Y	104	45	98	94	109	84	11,1	135	119					40	40
K/Rb	203	269	259	270	279	204	456	199	238					247	223
Fe_2O_3/FeO	1,49	1,31	2,03	1,12	3,47	1,51	1,50	2,05	1,47	0,52	1,67	0,69	1,27		
CIPW NORM (Wt. per cent)															
Q	34,67	20,84	33,40	36,00	39,08	33,79	29,26	33,02	33,82	31,45	33,07	34,03	31,02		
Or	29,37	31,26	27,24	28,48	29,37	30,02	22,51	29,25	28,19	32,62	31,61	29,61	27,95		
Ab	30,12	34,18	33,34	25,81	26,14	30,29	34,35	29,78	29,53	25,04	25,30	29,44	34,94		
An	1,43	3,79	1,39	3,18	1,57	1,87	8,05	2,36	3,08	4,89	5,15	2,06	1,82		
C					0,31				0,22	0,70	0,86	1,26			
Ac															
Wo			0,21					0,33							
Di - Wo	0,51	1,10	0,14	0,25		0,31	0,12	0,46					0,31		
Di - En	0,33	0,78	0,12	0,09		0,17	0,09	0,40					0,09		
Di - Fs	0,14	0,22		0,17		0,13	0,01						0,23		
Hy - En	0,17	0,78		0,45	0,20	0,18	1,38		1,02	1,25	0,80	0,60	0,18		
Hy - Fs	0,07	0,22		0,81		0,13	0,20		0,29	1,39	0,05	0,70	0,47		
Mt	1,68	3,71	2,58	2,36	1,18	1,67	2,04	2,76	2,04	1,26	1,81	1,36	1,88		
Il	0,39	1,39	0,45	0,53	0,24	0,28	0,49	0,53	0,45	0,56	0,41	0,32	0,23		
Hm			0,19		0,75			0,23							
Ap	0,05	0,40	0,05	0,12	0,02	0,05	0,17	0,09	0,07	0,33	0,17	0,26	0,17		
	98,93	98,67	99,11	98,25	98,86	98,89	99,17	99,21	98,71	99,46	99,23	99,64	99,29		

† Total Fe

1. Average of 9 Nubib granites
2. Average of 3 high-Ca rhyolites, Barby Formation
3. Average of 3 low-Ca rhyolites, Barby Formation
4. Average of 8 rhyolites lavas, Guperas Formation
5. Average of 5 rhyolites intrusives, Guperas Formation
6. Average of 6 quartz-porphyry dykes, Guperas Formation
7. Haremub granite (BW 1624)
8. Mylonitised Nubib granite (BW 899)
9. Rooiberg granite (BW 1592)
10. Average calc-alkali granite (Nockolds, 1954)
11. Average calc-alkali rhyolite (Nockolds, 1954)
12. Average alkali granite (Nockolds, 1954)
13. Average alkali rhyolite (Nockolds, 1954)
14. Average low-Ca granite (Turekian and Wedepohl, 1961)
15. Average granite (Taylor, 1964)

Table 15. Chemical compositions and normative mineralogy of the rhyolite lava of the Guperas Formation

	BW 1403	BW 1333	BW 581	BW 1501	BW 1284	BW 1435	BW 1293	BW 1478
SiO_2	68,86	77,49	72,24	75,63	73,51	75,50	76,49	75,94
TiO_2	0,62	0,15	0,39	0,22	0,36	0,20	0,14	0,16
Al_2O_3	12,85	10,47	11,27	11,23	11,23	11,27	11,50	11,36
Fe_2O_3	2,32	0,75	1,85	1,26	2,44	1,26	1,51	1,66
FeO	2,04	1,13	2,98	1,48	1,75	1,11	0,44	0,70
MnO	0,09	0,03	0,10	0,05	0,07	0,05	0,01	0,04
MgO	0,71	0,14	0,21	0,06	0,24	0,20	0,10	0,11
CaO	1,60	0,35	1,36	0,89	0,85	0,89	0,13	0,59
Na_2O	3,27	3,62	3,01	3,57	2,18	3,44	2,34	2,98
K_2O	4,69	3,72	4,46	4,24	5,70	4,15	6,44	5,18
P_2O_5	0,18	0,01	0,09	0,01	0,07	0,02	0,01	0,02
l.o.i.	1,65	1,10	0,84	0,57	1,01	1,17	0,55	0,74
H_2O^-	0,06	0,11	0,07	0,00	0,10	0,07	0,09	0,08
	98,94	99,07	98,87	99,21	99,51	99,33	99,75	99,56
Rb	168	115	122	125	164	124	271	140
Sr	183	90	109	57	58	39	72	88
Ba	980	357	2087	1137	2208	367	238	1382
Zr	422	381	693	474	652	447	290	300
Nb	14,2	18,9	21,7	21,7	22,1	19,4	26,0	19,8
Y	68	91	99	96	97	76	149	74
K/Rb	231	269	304	281	289	277	198	307
	CIPW NORM (Wt. per cent)							
Q	27,50	40,76	33,55	36,51	36,69	37,38	37,84	37,51
Or	27,71	21,98	26,36	25,06	33,68	24,52	38,06	30,61
Ab	27,67	30,63	25,47	30,21	18,44	29,11	19,80	25,21
An	6,56	1,34	4,07	2,10	3,76	3,06	0,58	2,33
C					0,10		0,47	
Ac								
Wo								
Di - Wo	0,08	0,14	0,87	0,94		0,51		0,20
Di - En	0,05	0,03	0,12	0,10		0,20		0,17
Di - Fs	0,03	0,12	0,83	0,94		0,31		
Hy - En	1,72	0,32	0,40	0,05	0,60	0,29	0,25	0,10
Hy - Fs	0,97	1,15	2,67	0,48	0,75	0,45		
Mt	3,36	1,09	2,68	1,83	3,54	1,83	1,04	1,93
Il	1,16	0,28	0,73	0,41	0,68	0,38	0,26	0,30
Hm							0,79	0,33
Ap	0,43	0,02	0,21	0,02	0,17	0,05	0,02	0,05
	97,24	97,86	97,96	98,65	98,41	98,09	99,11	98,74

Table 16. Chemical composition and normative mineralogy of rhyolite intrusives (plugs and domes) of the Guperas Formation.

	BW 578	BW 1281	BW 1470	BW 1502	BW 1504	BW 1552
SiO_2	76,36	78,15	77,25	76,35	76,25	70,78
TiO_2	0,10	0,11	0,16	0,11	0,14	0,62
Al_2O_3	11,63	10,15	11,33	11,86	11,81	12,58
Fe_2O_3	1,76	1,57	1,81	1,75	0,88	2,05
FeO	0,18	0,13	0,20	0,29	1,44	2,71
MnO	0,01	0,03	0,03	0,02	0,05	0,12
MgO	0,03	0,07	0,15	0,08	0,05	1,02
CaO	0,40	0,12	0,40	0,37	0,38	2,64
Na_2O	4,17	2,33	2,37	2,81	3,78	3,97
K_2O	4,56	5,67	4,62	5,53	4,48	1,43
P_2O_5	0,02	0,01	0,02	0,00	0,01	0,18
l.o.i	0,46	0,49	0,81	0,68	0,59	1,30
H_2O^-	0,11	0,11	0,11	0,08	0,11	0,10
	99,79	98,94	99,26	99,93	99,97	99,50
Rb	164	169	134	173	111	44
Sr	17	40	79	21	20	235
Ba	86	491	1367	343	725	641
Zr	291	291	258	322	387	404
Nb	23,3	20,7	17,5	23,3	21,3	10,8
Y	92	131	73	126	106	62
K/Rb	231	279	286	265	334	268
	CIPW NORM (Wt. per cent)					
Q	34,40	42,57	44,76	37,93	35,45	34,41
Or	26,95	33,51	27,30	32,68	26,47	8,45
Ab	34,44	19,71	20,05	23,78	31,98	33,59
An		0,49	1,85	1,84	1,82	11,92
C			1,75	0,58	0,08	0,13
Ac	0,74					
Wo	0,69					
Di - Wo	0,09	0,01				
Di - En	0,07	0,01				
Di - Fs						
Hy - En		0,16	0,37	0,20	0,12	2,54
Hy - Fs					1,78	2,51
Mt	0,33	0,21	0,29	0,69	1,28	2,97
Il	0,19	0,21	0,30	0,21	0,26	1,16
Hm	1,28	1,43	1,61	1,27		
Ap	0,05	0,02	0,05		0,02	0,43
	99,23	98,33	98,33	99,18	99,26	98,11

the appearance of olivine as a liquidus phase would be delayed until relatively shallow levels (<35 km) are reached (Green & Ringwood, 1967), and it would not be capable of appreciably affecting the strong calc-alkaline character inherited by the partial melting of eclogite at depth and by subsequent fractionation leading to silica and alkali enrichment. The rare presence of small, highly resorbed olivine crystals in quartz-normative lavas (e.g. BW 381) might be the result of the incomplete reaction of olivine phenocrysts with a liquid that was rapidly enriched and oversaturated in silica by the appearance of plagioclase as a major liquidus phase at low pressures. Notably, BW 381 contains an abundance of large plagioclase phenocrysts.

The generally lower MgO and CaO contents of those Sinclair potassic rocks that contain normative quartz or only minor amounts of normative olivine (e.g. BW 381, BW 1565, BW 409) as compared with the generally higher MgO and CaO contents of the more highly olivine-normative bearing rocks (e.g. BW 1256, BW 1622), is consistent with the early removal of clinopyroxene from the melt (cf. experimental data of Green & Ringwood, 1968, p.133-141), and does not require extensive olivine fractionation.

Summary of Conclusions:

1. Processes such as *in situ* metasomatism, wall-rock reaction at high and/or low pressures and crystal liquid fractionation cannot adequately account for the very high potassium contents of the Sinclair calc-alkaline magmas.

2. The calc-alkaline magmas represent primary potassic melts resulting from processes possibly related to the subduction of oceanic crust in Late Precambrian times.

3. The primary magma was derived by low degrees of partial melting of eclogite (transformed oceanic crust) in the depth interval $\sim$150 to 175 km, i.e. in a 'dry' environment but still within the depth range where phlogopite is stable.

4. The low melting character of the phlogopite would have caused it to enter the partial melts, thus enriching them in potassium and associated minor elements such as Rb and Ba.

5. The primary potassic magmas would have been of a predominantly silica-undersaturated nature and relatively minor amounts of fractionation, of the ascending magma, following a typical calc-alkaline trend would have further enriched the melt in silica and alkalis.

3.4.2 SPES BONA SYENITE

The close temporal and spatial relationships between the diorites, monzonites and the Spes Bona syenite, together with chemical characteristics very similar to the Barby calc-alkaline potassic intrusives and extrusives as discussed in Section 3.3.1, are highly suggestive of a genetic link between these various units and the syenite. This can either be considered as entirely fortuitous or the association can be interpreted as co-magmatic. The latter is considered most likely especially in view of the similar chemical characteristics exhibited by the syenite and the other potassium-rich units.

On chemical variation diagrams (Figs. 10,11,12,14 and 18) the syenite plots within and extends all the trends exhibited by the calc-alkaline rocks. The position of the syenite compositions in a trend of increasing SiO_2 contents with increasing K_2O contents (see Fig. 10) suggests that this rock might have been derived from a normal calc-alkaline differentiation toward alkali and silica enrichment with decreasing MgO, FeO and CaO by the precipitation of mainly clinopyroxene and garnet. The syenite has enriched SiO_2 and K_2O contents when compared with the calc-alkaline lavas and intrusives. The trace elements associated with potassium also show enriched concentrations. The high concentration of strontium may also reflect a dominantly high pressure fractionation since at lower pressures this element would enter plagioclase and decrease in concentration in the residual liquid fraction.

However, it is apparent from the normative compositions of the syenite that fractionation was not the only process in the production of the syenite magma. Normal calc-alkaline differentiation will lead to quartz-normative products and it is notable that all three of the analysed syenites contain olivine in the norm and two have small amounts of normative nepheline. This indicates that even though the SiO_2 content of the syenite is higher than that of any of the calc-alkaline lavas and intrusives, large amounts of silica have been used by the potash to form potassium felspar, resulting in an undersaturated character in the norm (and mode).

The very high K_2O/SiO_2 ratio of the syenite cannot be the result of normal fractionation and it can only be concluded that either,

(i) the fractionation followed a trend not of the calc-alkaline type, and with potassium increasing at such a rate relative to silica so as to retain a normatively undersaturated character, or

(ii) the original magma was of a highly potassium-rich nature and was even more undersaturated than the present syenite. Subsequent fractionation following a calc-alkaline trend would thus have increased the silica and alkali content, reducing the K_2O/SiO_2 ratio. The fractionation apparently did not proceed far enough for the final syenite magma to attain an oversaturated quartz-normative composition.

Alternative (ii) is preferred as a reasonable explanation of the observed chemical properties of the syenite since the present calc-alkaline character of the rock does not support any fractionation trend other than a calc-alkaline one. A process such as that outlined in (ii) would also fit the scheme presented earlier in this Section for the derivation of the potassic calc-alkaline basic magmas of the Sinclair.

Partial melting of eclogite (transformed tholeiite) within a subduction zone, involving a low melting potassic mineral phase such as phlogopite, could well give rise to liquids of syenitic character (Green & Ringwood, 1968) if the degree of partial melting was low enough. Degrees of melting of 5 per cent or less would probably be necessary to produce such highly potassic liquids of silica undersaturated character (Jakes & White, 1972). Whether or not separation of the liquid phase from the crystalline 'residue' is possible with such low degrees of partial melting, is uncertain and some authors (Green,

Green & Ringwood, 1967; Green & Ringwood, 1967 and 1968) consider 20 per cent as a reasonable minimum. However, Jakes and White (1972) infer that 5 per cent or less of partial melting will produce liquids capable of separation, accumulation and movement toward the surface and Gast (1968), in discussing the origin of alkali basalts, considers that liquids from partial melting may segregate into small conduits and channels when the proportion of liquid reaches only 3-6 per cent of the total mass of crystal residuum plus liquid. It is further possible that in the upper parts of a subducting slab, where the shoshonitic and syenitic magmas are thought to have originated, shearing would have been well developed and could have provided additional channels and hydrostatic conditions favourable to the coalescence and accumulation of liquids from small degrees of partial melting.

If the origin of the Spes Bona syenite can be regarded in terms of low degrees of partial melting of eclogite and subsequent fractional crystallisation, then it might be possible to provide an explanation for the variation in trend on the K_2O vs. SiO_2 diagram observed in different shoshonitic rock suites. It was noted earlier in this Section that most shoshonite suites define a fairly shallow trend on the K_2O vs. SiO_2 diagram (Fig. 10) but not uncommonly, as in the case of the Sinclair shoshonites, a steep trend is characteristic (Gill, 1970; Jakes & Smith, 1970; Joplin, 1968).

Partial melting of eclogite (under dry conditions) to degrees of about 40-50 per cent wil produce liquids of andesitic composition (Green & Ringwood, 1968) with relatively high SiO_2 contents and 'normal' K_2O contents. With lower degrees of partial melting the K_2O content of the liquid will increase while the SiO_2 content will decrease (Green & Ringwood, 1968; Jakes & Smith, 1970; Jolly, 1971). Therefore, the shoshonitic magmatic products generated by low and variable degrees of partial melting will define a steep *negative* trend on the K_2O vs. SiO_2 diagram. If the effects of normal calc-alkaline fractionation are superimposed on the magma compositions, as is envisaged for the Sinclair calc-alkaline rocks, the trend will move through the vertical to steeply positive depending on the degree of fractionation (Fig. 24). The reason for this change in trend is the tendency for calc-alkaline differentiation to produce quartz-normative end products (i.e. the K_2O/SiO_2 ratio decreases). Under these conditions the occurrence of shoshonites may be characterised by syenite as well as the more typical basaltic andesites. Supporting this is the fact that those shoshonite suites possessing steep trends on the K_2O vs. SiO_2 diagram (Gill, 1970; Jakes & Smith, 1970; Joplin, 1968), including the Sinclair shoshonite suite, are characterised by an abundance of rocks of undersaturated character (normative olivine and nepheline) and those members that are normatively oversaturated with respect to silica generally contain only very minor amounts of quartz in the norm.

If, on the other hand, the shoshonites were produced by fairly constant degrees of partial melting with initially constant K_2O/SiO_2 ratios and then underwent normal calc-alkaline differentiation, a relatively flat trend would be followed, roughly parallel to that of the more typical low potassium calc-alkaline suites.

Joplin (1968) considered two different differentiation trends to be present in the shoshonite association, one toward quartz-bearing rocks and the

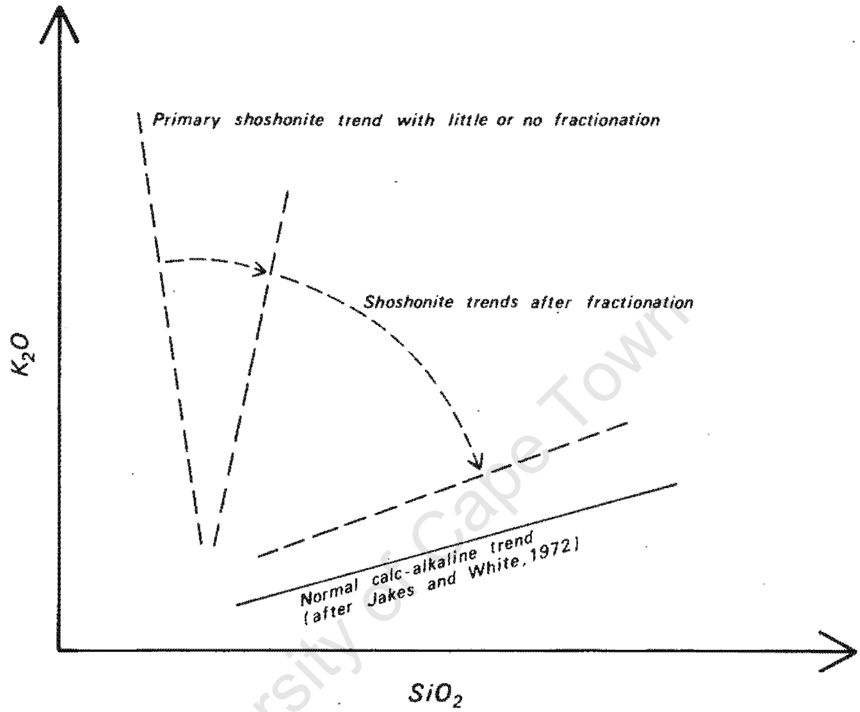


Fig. 24. Schematic change in the K_2O vs. SiO_2 trend of shoshonitic magmatic products when normal calc-alkaline fractionation is superimposed on primary magmas

other toward potassium enrichment. In view of the reasoning presented above it is proposed here that the potassium enrichment trend (with decreasing SiO_2 contents) of Joplin (1968) is a result of the degree of primary partial melting, whereas the trend toward quartz-normative or quartz-bearing potassic rocks reflects fractional crystallisation along a typical calc-alkaline trend.

A combination of the two processes may provide a wide range in compositions with respect to silica and potassium in particular. The slope of the trend on the K_2O vs. SiO_2 diagram may, therefore, provide a rough guide as to which process has been dominant in establishing the chemical characteristics of a particular shoshonite occurrence.

A process of alkali diffusion or 'potash enrichment' has been proposed for the origin of the Deboullie calc-alkali syenite by Boone (1962). Differentiation by fractional crystallisation of a dioritic magma and selective upward transfer of H_2O and potash in the melt is considered to have produced a magma rich in potash (syenite) at the top of the magma column, leaving a granodioritic 'residue' at deeper levels which then intruded the earlier-emplaced syenite.

Although such a process could well be considered for the genesis of the Spes Bona syenite, the eclogite melting hypothesis, as proposed earlier, is preferred for the following reasons:

(i) The concentration of H_2O in the magma would result in a very high volatile content and, consequently, extensive soaking of the country-rock and miarolitic texture in the syenite might be expected. There is no evidence for these in the case of the Spes Bona syenite.

(ii) There is also no field evidence for a complementary granodioritic residual fraction, although this might exist but not be exposed at the present level of erosion.

(iii) Due to the proposed genetic link between the shoshonitic rocks of the Barby and the Spes Bona syenite, if potash enrichment is proposed for the genesis of the syenite, then it must also be considered for the entire succession of Barby calc-alkaline lavas.

Since many of the Barby calc-alkaline lavas contain nearly as much potassium as the Deboullie syenite (cf. data of Boone, 1962, p.1460) enormous quantities of complementary residual granodioritic magma would have been produced and there is no field evidence in support of this.

(iv) The extremely high K_2O content of the Spes Bona syenite (~8 per cent) when compared with the Deboullie syenite (5,87 per cent) requires unrealistic amounts of potassium to be transferred by volatile enrichment.

3.4.3 THOLEIITIC BASIC UNITS

Tholeiites of the Sinclair Group range from olivine-normative to quartz-normative in composition and, therefore, display the effects of low pressure fractionation. According to Green *et al.* (1967) basaltic liquids fractionating from olivine tholeiite can only cross the plane of silica saturation (basalt tetrahedron - Yoder and Tilley, 1962) into the quartz tholeiite field at depths corresponding to pressures of less than 4,5 kb (~15 km). As pointed out by O'Hara (1965, p.35), "no intrusive tholeiitic magma body displayed among crustal rocks, without tectonic contacts, displays any fractionation trend other than that passing through the condition olivine + two pyroxenes + plagioclase + liquid towards silica enriched compositions".

The tholeiitic gabbros and norites occurring on farms Kumbis,166 (Springbokvlakte), and Saffier, provide evidence for such high-level fractionation, both *in situ* and at some slightly deeper level, but still within the low pressure regime of the crust (upper crust). Fractionation *in situ* has resulted in the layered character typical of many of the intrusive bodies.

The Kumbis intrusives, in particular, contain layers of olivine-gabbro, quartz-gabbro and anorthosite due to the settling out of olivine and plagioclase crystals. Quartz-bearing layers formed in this manner are of relatively minor importance and it would seem that the original magma intruded into the site presently occupied by the Kumbis body was of olivine tholeiitic character.

The presence of 'non-layered' and relatively homogeneous basic bodies ranging from picrite through olivine-rich gabbro to quartz-gabbro is indicative of the tapping at different times and/or at various levels of a large fractionating magma reservoir (or reservoirs) below the present level of exposure. Once again, the trend of differentiation is typically of the low pressure type resulting in quartz tholeiitic end products. That olivine was an early liquidus phase is apparent from its textural relationships to other minerals in these rocks. The picrite sill intruding the gabbro and norite body on Kumbis must have been the result of accumulation of olivine by early precipitation and crystal settling and intrusion as an olivine crystal-liquid mush, followed by final crystallisation of the liquid *in situ*, resulting in interstitial plagioclase, minor amounts of biotite and poikilitic clinopyroxene.

Extrusive tholeiitic rocks of the Barby and Guperas Formations also range from olivine-normative to quartz-normative in composition and, therefore, also show the effects of a low pressure fractionation process involving the precipitation of olivine, pyroxene and plagioclase.

It is thus apparent that fractionation of olivine tholeiitic magma in the low pressure regime (<15 km) of the crust would have been capable of producing intrusive and extrusive rocks as represented by the Sinclair tholeiites. Any deductions concerning the source of the primary magmas at depth must, however, remain speculative.

It is generally agreed that tholeiitic basalt magmas, with the possible exception of the island arc tholeiites (Jakes & White, 1970; Jakes & Gill, 1970; Fitton, 1971), are derived by partial melting of upper mantle material (O'Hara, 1965; Green & Ringwood, 1967; Green, Green & Ringwood, 1967, Nicholls & Ringwood, 1973). The study of liquid-crystal relationships on natural and synthetic rock materials at varying pressures and temperatures in the system $CaO-MgO-Al_2O_3-SiO_2$ has provided much information which allows speculation as to the origin of basaltic magmas. The composition of the liquid formed by partial melting of mantle peridotite will depend on a number of factors:

(i) Pressure/depth at which magma segregation from residual crystal occurs.

(ii) Degree of partial melting (temperature and pressure dependent).

(iii) Partial pressures of volatiles.

(iv) Crystallisation differentiation during ascent of the liquids to the surface.
This fractionation process will be continuous and, in addition, extensive crystallisation during interruption of ascent at some specific level or levels may take place, thereby further changing the composition of the liquid.

Taking these factors into account it is possible, on the basis of experimental evidence (Yoder & Tilley, 1962; O'Hara, 1965; Green & Ringwood, 1967; Green, 1969), to deduce the following processes for the derivation of magmas of tholeiitic composition:

(i) Continuous fractionation of only olivine from magma generated at depths of up to about 100 km (O'Hara, 1968). According to Wyllie (1971, p.198) "a magma batch formed and fractionated at various levels is taken from a particular pressure and temperature at such speed that its composition remains near the boundary of, but always just within, the olivine primary phase volume; hence it fractionates olivine only during its ascent. This is possible because of the increase in normative olivine content of liquids with pressure."

(ii) Continuous fractionation, while the magma ascended toward the surface, of olivine, two pyroxenes and the appropriate alumina-rich phase from either a hypersthene-normative picrite or a nepheline-normative picrite generated by partial melting at depths of about 75-100 km and 30-75 km respectively (O'Hara, 1965). In this scheme the end product at the surface will be a silica-saturated tholeiitic magma.

(iii) Fractionation with decreasing pressure of

(a) hypersthene-normative picrite formed by 35-40 per cent melting in the depth interval 70-100 km, or

(b) low-Al_2O_3 olivine tholeiite formed by 25-30 per cent melting in the depth interval 35-70 km, or

(c) high-Al_2O_3 olivine tholeiite, formed by 20-25 per cent melting at about 30 km depth,

will give rise to high-alumina basalts and with increased precipitation of plagioclase as well as olivine and two pyroxenes in the low pressure regime (<15 km) will further give rise to olivine tholeiites and finally quartz tholeiites. The precipitation of alumina-rich enstatite in the high pressure regime will initially deplete the liquid in alumina but at intermediate pressures/depths a subaluminous enstatite is the liquidus pyroxene and the liquid will become relatively enriched in Al_2O_3, thereby producing Al_2O_3 basalts at a high level. With further fractionation involving plagioclase, olivine and pyroxene, the liquid will become poorer in Al_2O_3 and trend toward a tholeiitic composition (Green & Ringwood, 1967; Green, 1969).

(iv) Fractionation with decreasing pressure of alkali olivine basalt formed either by 10-20 per cent melting of mantle peridotite or by fractionation of low-Al_2O_3 olivine tholeiite at a constant depth, i.e. 35-70 km (Green & Ringwood, 1967; Green, 1969).

(v) Fractionation of high-alumina 'basalt-like' liquid in the intermediate part of the low pressure regime to produce a high-alumina basaltic magma which, when fractionating toward the surface, can give rise to high-alumina hypersthene-normative basalts (O'Hara, 1965). The parent high-Al liquid could have originated by fractionation of hypersthene-normative picrite or nepheline-normative picrite as outlined in (ii).

(vi) The generation of tholeiitic magmas in an island arc type environment, which may be directly or indirectly related to partial melting of the upper mantle, has been discussed earlier in this Section.

It is suggested in Section 3.4.5 that the thickness of the continental crust in the Sinclair area, during development of the Sinclair Group, was in the order of 36 km. Such a thickness precludes, therefore, the formation of basaltic magmas by partial melting of the mantle at depths shallower than about 36 km. The generation of high-Al_2O_3 olivine tholeiite at about 30 km depth as discussed in (iii c) above is consequently ruled out but the low pressure fractionation scheme (outlined in (iii c)) could still have been operative in this pressure regime, affecting the compositions of liquids produced at deeper levels. The complete lack of alkali olivine basalt in the Sinclair area probably also rules out the derivation of the Sinclair tholeiites by fractionation of a low-Al_2O_3 alkali olivine basalt, generated at depths of 35-70 km as discussed in (iv) above.

Continuous (as opposed to interrupted) fractionation, up to the low pressure regime (<15 km) of liquids formed at depths greater than about 36 km is, therefore, considered to be the most likely process for derivation of the primary tholeiitic magmas of the Sinclair Group. In detail, any one or more of the schemes outlined in (i), (ii & v), (iii a & b) and (vi) above would be applicable.

The complete range in basalt types from typically tholeiitic to typically calc-alkaline within the Sinclair, as shown by the AFM plot of Fig. 12, the presence of transitional tholeiitic basalts with an abnormally high potassium content (BW 811 and BW 1351) and contemporaneity between both lava types, is highly suggestive of a transition and possible relationship between the processes giving rise to the calc-alkaline potassic magmas and those giving rise to the tholeiites.

If the generation of calc-alkaline magmas of the Sinclair is related to subduction processes (see p.147), then it is possible that a model similar to that of Nicholls and Ringwood (1973) for the production of the island arc tholeiites can be applied here for the generation of the Sinclair tholeiitic magmas.

The combination of increasing pressure and temperature with depth, affecting the upper surface of a subducted slab of oceanic lithosphere, will lead to its progressive dehydration. Water will be lost from the slab over a broad depth interval, but the breakdown of amphibole at depths between 70-100 km (Allen & Boetcher, 1971; Allen *et al.*, 1972) will release a considerable amount of water that would rise up into the overlying mantle causing a reduction in viscosity, the introduction of gravitational instability and conditions suitable for the initiation of partial melting of rising mantle peridotite (Nicholls & Ringwood, 1973). Partial melting at depths greater than 70 km will produce picrite magmas which could then, by fractionation toward the surface, as outlined in schemes (i) or (iii) above, produce basaltic liquids of compositions similar to those of the Sinclair tholeiites.

The transitional tholeiites with high potassium contents might be fractionation products of those magmas produced by partial melting of upper mantle peridotite in proximity to the subduction zone, where interaction between the picritic melts and minor potassic partial melts resulting from the breakdown of amphibole and/or phlogopite, might more easily take place. The tholeiites of relatively low potassium content could then have an origin more remotely associated with the subduction zone as in the model discussed above. It is possible that these tholeiites represent 'island arc tholeiites' (Jakes & Gill, 1970). Minor chemical differences do occur between the low-potassium tholeiites of the Sinclair and typical arc tholeiites (Table 6, Section 3.3.1), but the data are not sufficient at present to establish whether these differences are significant or not.

The high Al_2O_3 content of one Sinclair tholeiite, BW 811, might also reflect a transitional character toward the calc-alkaline rocks, since typical high-alumina tholeiites, produced by fractionation of basalt magma at depths of 15-35 km (or partial melting of mantle peridotite at depths of 15-35 km), will be olivine-normative (O'Hara, 1965; Green, Green & Ringwood, 1967; Green & Ringwood, 1967). The tholeiite in question (BW 811) is quartz-normative and could not, therefore, have undergone extensive low pressure fractionation since this would have reduced the alumina content (by the precipitation of plagioclase) whilst producing a quartz-normative composition. It is concluded, therefore, that the high alumina content of this rock, together with a quartz-normative character, could indicate a close relationship to the calc-alkaline suite.

Summary of Conclusions:

1. Continuous (as opposed to interrupted) crystal fractionation up to the low pressure regime (<15 km) of primary liquids formed by partial melting of mantle peridotite at depths of >∿36 km could have given rise to the primary tholeiite magmas.

2. High-level fractionation (<15 km depth) during both the continuous and interrupted rise of the primary magma produced normatively oversaturated (with respect to silica) compositions.

3. Partial melting within the mantle could have been the result of unstable conditions developed in the mantle 'wedge' above a subduction zone; the unstable conditions arising from the release of water into the mantle from the underlying oceanic plate.

4. The high-potassium nature of some of the Sinclair tholeiites might reflect the interaction between the primary tholeiitic melts (picritic) of the mantle and calc-alkaline melts resulting from partial melting in the subduction zone.

5. The high-potassium and high-alumina contents of some of the tholeiites suggests a close and probably transitional link with the high-potassium calc-alkaline rocks.

3.4.4 GEOTECTONIC SETTING DEDUCED FROM TRACE ELEMENT DATA

The model proposed for the generation of the basic and syenite magmas of the Sinclair Group, the calc-alkaline magmas in particular, implies that the Sinclair Group constitutes part of a magmatic arc, developed at the margin of an ancient continental plate during active convergence with, and consumption of, an oceanic plate. This aspect of the regional tectonic setting will be discussed further in Section 5.

In a recent publication, Pearce and Cann (1973) have emphasized that certain types of basic volcanic rocks are characteristic of particular geotectonic settings. Their studies also indicate that inter-element relationships between the trace elements Ti, Zr, Y, Nb and Sr are particularly useful for discriminating between the different types of basic volcanic rocks typically found in particular tectonic settings. It should, therefore, be possible to discriminate between various tectonic environments by employing the relationships between the trace elements in basic volcanic rocks mentioned above. By making use of modern examples of established tectonic setting, Pearce and Cann (1973) have been able to determine fairly distinct fields on variation diagrams in which basic volcanic rocks will plot according to their tectonic setting.

The basic volcanic rocks of the Sinclair have been plotted on these variation diagrams in Figs. 25, 26 and 27. On the Ti vs. Zr and Ti-Zr-Sr plots there is a fairly wide scatter of points, but a majority of compositions fall within or very close to the calc-alkali basalt fields. On the Ti-Zr-Y plot, however, there exists a very close clustering of compositions within the calc-alkali basalt fields. Sample BW 1415 plots consistently in the low-potassium tholeiite fields of all these diagrams, which is consistent with its classification based on major element chemistry (Section 3.3.1). The presence of a sample plotting in the 'ocean-floor basalt' fields of Figs. 25 and 26 could be considered problematical since there is no field evidence that might suggest the presence of ocean-floor basalts in the Sinclair area. However, as pointed out by Pearce and Cann (1973) the fields on their diagrams are not based on a statistically 100 per cent distinction of magma types so that a certain amount of overlap of fields, not indicated on the diagrams, is to be expected. Furthermore, the determination of the fields was based on a limited number of examples only and although they are regarded as providing a good indication of the magma-type, they are probably still not well enough defined to allow a perfect distinction of magma types.

A number of samples in Fig. 26 and one in Fig. 27 do not plot within any of the fields and this can probably also be attributed to the statistical error in defining exact field boundaries as discussed above. Certainly in the case of Fig. 26 points plotting outside of the fields form part of a close clustering of samples in the calc-alkali field.

It would appear, therefore, that the relationships between the elements Ti, Zr, Y, Nb and Sr provide a reasonably good indication that the basic volcanic rocks of the Sinclair Group belong to the "volcanic arc" basalt type (Pearce & Cann, 1973, p.291) developed at the margins of converging crustal plates. In the proposed classification scheme for basic volcanic rocks based on tectonic setting as given by Pearce and Cann, calc-alkaline basalts and

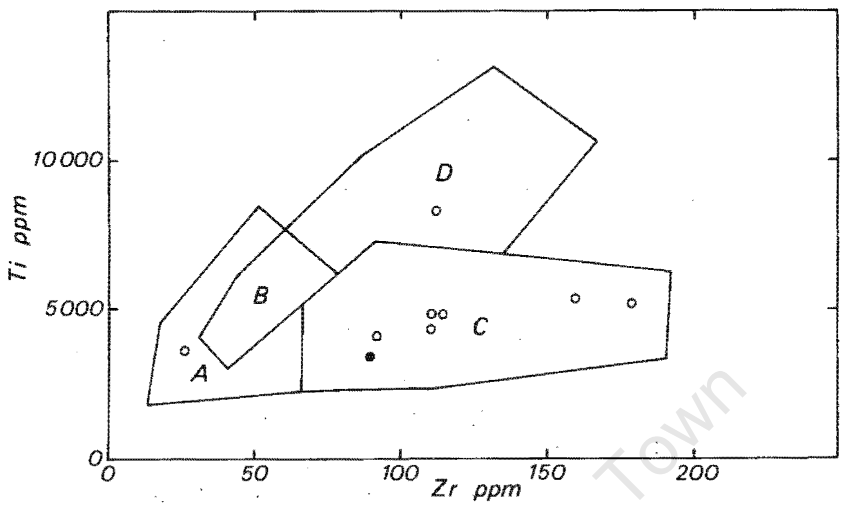


Fig. 25. Discrimination diagram for the basic lavas of the Barby Formation (open circles) and the Guperas Formation (full circles) using Ti and Zr. Ocean-floor basalts plot in fields D and B, low-potassium tholeiites in fields A and B, and calc-alkali basalts in fields C and B (*after* Pearce and Cann, 1973)

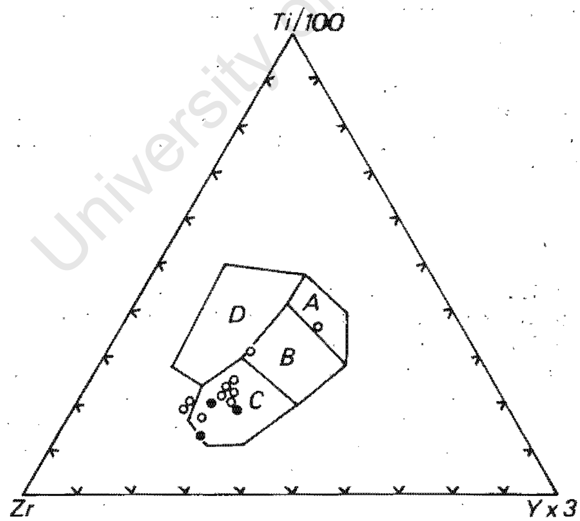


Fig. 26. Discrimination diagram for the basic lavas of the Barby Formation (open circles) and the Guperas Formation (full circles) using Ti, Zr and Y. Oceanic island or continental basalts plot in field D, ocean-floor basalts in field B, low-potassium tholeiites in fields A and B, and calc-alkali basalts in fields C and B (*after* Pearce and Cann, 1973)

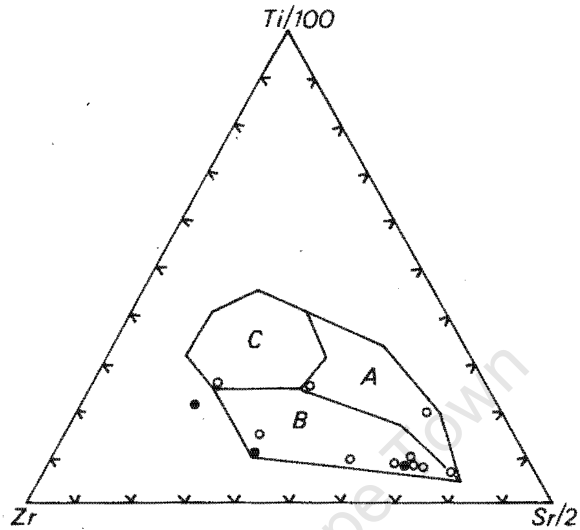


Fig. 27. Discrimination diagram for the basic lavas of the Barby Formation (open circles) and the Guperas Formation (full circles) using Ti, Zr and Sr. Ocean-floor basalts plot in field C, low-potassium tholeiites in fields A, and calc-alkali basalts in field B (*after* Pearce and Cann, 1973)

and low-potassium tholeiites are considered characteristic of this particular geotectonic environment. These conclusions support those arrived at earlier in this Section as regards the processes involved in the generation of the calc-alkaline magmas.

3.4.5 FELSIC UNITS

It has been pointed out in Section 3.3.3 that the felsic rock units of the Sinclair Group are comprised of two fairly distinct groups. The distinction is made primarily on chemical grounds, and on trace element contents in particular. The bulk of the felsic rocks comprise a highly siliceous granitic group, i.e. the Barby rhyolites, Nubib granite (≡Rooikam and Tumuab granites of von Brunn, 1967), rhyolite extrusives, plugs, domes and quartz-porphyry dykes of the Guperas Formation, Rooiberg granite and Nagatis rhyolite extrusives (von Brunn, 1967). The second and volumetrically less significant group consists of the high-Ca rhyolites of the Barby Formation and the Haremub granite (≡Kotzêrus granite of von Brunn, 1967) which is also relatively Ca-rich and tends toward a granodioritic composition.

In an attempt to formulate an acceptable mechanism (or mechanisms) for the generation of the granites, rhyolites and other felsic rock units of the Sinclair Group, a possible origin for these rocks will be discussed below in the light of current ideas on the genesis of granitic magmas.

Processes capable of producing magmas of granitic composition may be listed as follows:

(i) Partial fusion of crustal rocks (Eskola, 1932; Brown and Fyfe, 1970; Fyfe, 1973; Brown, 1973).

(ii) Fractional crystallisation of basic magma (Bowen, 1928).

(iii) Partial melting of oceanic basaltic crust or upper mantle along a subduction zone inclined under a continental margin (Dickinson, 1968, 1970; Hamilton, 1969). A corollary to this process is the melting of ocean-floor sediments dragged down into a subduction zone (Gilluly, 1971; Huang and Wyllie, 1973).

The composition of many felsic rock-types can be represented within the system An-Ab-Or-Qz-H_2O (Tuttle and Bown, 1958; von Platen, 1965; Kleeman, 1965; Winkler, 1967; Presnall and Bateman, 1973).

Before attempting to refer the Sinclair rocks to this so-called "granodiorite" system, it must first be established to what extent the normative constituents of the system are representative of the bulk normative constituents of the magmas which crystallised to form the Sinclair felsic units, i.e. what proportion of the rock compositions is represented by the normative system An-Ab-Or-Qz. All of the 38 analyses of Sinclair felsic rocks (Tables 13-17) contain more than 88 per cent of these normative constituents and the average value is 94 per cent. The system An-Ab-Or-Qz-H_2O is, therefore, considered to be a very good approximation of the bulk composition of the Sinclair felsic rocks, and almost certainly of the magmas from which they crystallised, in view of the lack of post-consolidation metamorphism.

In Fig. 28 the normative compositions of the Sinclair felsic rock-types (including 17 analyses from von Brunn, 1967) have been plotted on the Ab-Or-Qz face ('granite' system of Tuttle and Bowen, 1958) of the granodiorite tetrahedron. From this diagram it is apparent that the rock compositions form a prominent cluster within the low temperature trough of the system. The only Sinclair felsic rock-types not lying in proximity to the low-temperature region of this diagram are the three Ca-rich Barby rhyolites and a rhyolite intrusive (BW 1552) of the Guperas Formation. The former fall well within the orthoclase field in spite of the fact that only plagioclase phenocrysts are present in the rock. The Guperas rhyolite plots well within the Qz field, and lies close to the Qz-Ab side-line.

It has, however, been pointed out by Kleeman (1965) that even where rocks contain as little as 2 per cent normative anorthite, it is misleading to interpret them in the Ab-Or-Qz system without consideration of the effects of the An content. Von Platen (1965) has also demonstrated that in the process of partial fusion of rocks within the An-Ab-Or-Qz system the decreasing Ab/An ratio of the source rock results in the position of the cotectic line being displaced progressively toward the Qz apex of the tetrahedron when viewed on the Ab-Or-Qz face. Such a displacement has the effect of changing the position of the ternary minimum and, therefore, the composition of the first products of anatexis or "minimum melt" (Winkler, 1967, p.201).

Since the Sinclair granites and rhyolites have a range in normative An from 0 to about 12 per cent, with an average of 2,5 per cent, it is necessary

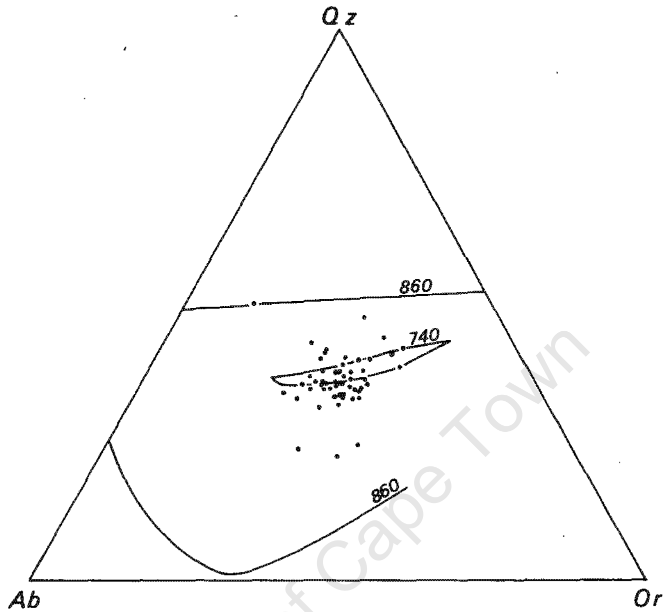


Fig. 28. Distribution of the Sinclair felsic rock-units within the system Ab-Or-Qz, with respect to the experimentally determined liquidus boundaries at 740°C and 840°C at 1 kb water pressure (*after* Tuttle and Bowen, 1958)

to refer them to the An-Ab-Or-Qz system in order to gain a possibly more realistic impression of their position in relation to the low-temperature trough. Fig. 29 shows the normative compositions of the Sinclair felsic rock-types (including 17 analyses from von Brunn, 1967) plotted on the An-Ab-Or face of the An-Ab-Or-Qz system, or 'granodiorite' system.

Once again it is apparent that there is an extremely close clustering within the low-temperature region of the system. Notably, those rhyolites of the Barby Formation which fall within the felspar field on the Ab-Or-Qz diagram (Fig. 28) now also plot in the low-temperature trough, illustrating the possibility of error in referring rocks to the Ab-Or-Qz system alone, without consideration of the anorthite content. The Guperas rhyolite (BW 1552) still plots well away from the minimum, and a possible reason for this will be discussed below.

The fact that the Sinclair felsic rocks, almost without exception, plot in the low-temperature trough of the An-Ab-Or-Qz system is in support of a magmatic origin for these rocks. Further evidence for a magmatic state can be found in the glassy or fine-grained and intrusive characteristics of the granites.

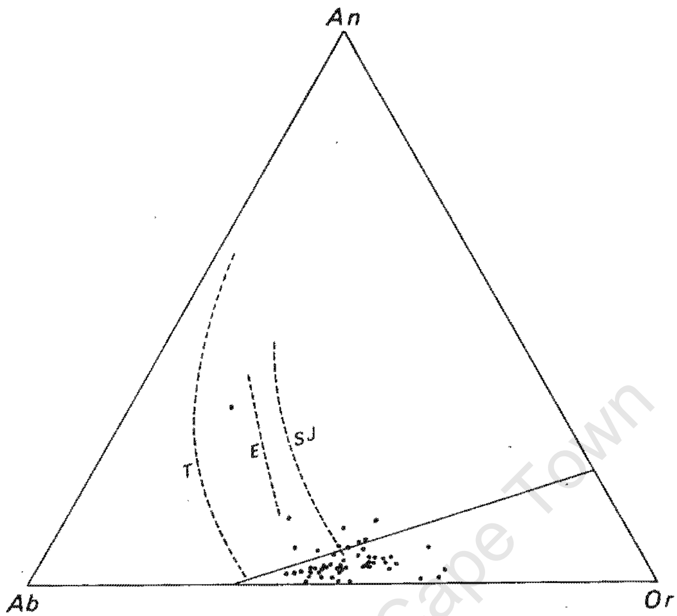


Fig. 29. Distribution of the Sinclair felsic rock-units within the system An-Ab-Or, with respect to the low-temperature divide (*after* Presnell and Bateman, 1973). Differentiation trends (T = Thingmuli, E = Eskadalemuir, SJ = San Juan) are *after* Kleeman (1965)

Felsic magmatic products can be derived by the fractional crystallisation of a parent basic magma and it might be considered that the Sinclair felsic magmas were derived by such a process.

It has been demonstrated for the Skaergaard mafic layered intrusion (Wager & Deer, 1939) for example, that differentiation of basic magma would certainly result in an end product of granitic composition falling within the low-temperature trough of the Ab-An-Or-Qz system. This felsic end product would, however, constitute only a fraction of a per cent of the original volume of basic magma (Hyndman, 1972).

A preponderance of basic rocks could be expected, therefore, in any particular environment where granitic rocks occur as a result of fractional crystallisation of a basic magma. Furthermore, the process should give rise to magmas ranging in composition from the most basic members through to the most felsic or granitic end members, with possible evidence of an ultrabasic accumulative fraction.

In the Sinclair-Helmeringhausen area quite the reverse situation is found. In terms of outcrop area, the ratio of felsic to basic rocks is in the order of 3:2, north of the Sinclair Mine, and even though enormous volumes of basic eruptives and intrusives are present, they are volumetrically dominated by granites and rhyolites. If the situation west of Helmeringhausen (von Brunn, 1967) is also taken into account, the above ratio of felsic to basic rock-types outcropping at the surface, is not changed significantly. The almost complete lack of ultrabasic rocks either as intrusives or xenoliths in the basic lava is also significant. The small picrite sill associated with the Barby gabbros (Section 3.1.2.2) on farm Kumbis is the only evidence of a possible ultrabasic accumulative fraction, but is, in any case, volumetrically insignificant if large-scale fractionation processes are to be envisaged. The lack of rock-types intermediate in composition between the Sinclair basalts and andesites on the one hand, and the granites and rhyolites on the other hand, which is very well displayed by a distinct hiatus in major element composition in the AFM diagram of Fig. 12, is also inexplicable in terms of a fractional crystallisation origin for the felsic magmas.

A 'bimodal' compositional character for part of the Sinclair Group west of Helmeringhausen has already been convincingly demonstrated by von Brunn (1967). Furthermore, elements such as K and Rb, which would be expected to be significantly concentrated during crystal fractionation processes (Ewart and Stipp, 1968), are not markedly enriched in the Sinclair rocks.

If a differentiating basic magma is able to assimilate relatively large amounts of sialic material, then the siliceous residual liquid could amount to as much as 15 per cent of the total mass of basic magma plus sialic material (Kleeman, 1965). Even such a process, however, cannot be reconciled with the observed volume relationships in the Sinclair-Helmeringhausen area and the problem concerning the lack of a differentiation 'series' still exists. Kleeman (1965, p. 45) states that "the differentiation of a contaminated basalt will provide a range of liquids with compositions lying in the plagioclase field ... of the An-Ab-Or-Qz system ... near the low-temperature trough but none can lie in the orthoclase field. This is the range of compositions represented by the acid rocks of the British Tertiary Province and of the San Juan and Thingmuli Provinces." These compositional trends of these provinces are shown in Fig. 29 and it is immediately apparent that not only do the vast majority of Sinclair felsic rock compositions plot in the orthoclase field, but there is also no evidence of a compositional trend as displayed by the above-mentioned examples.

Granites and rhyolites derived in this fashion from contaminated basic magmas are generally of a subaluminous or metaluminous character "because the basic magma has enough calcium to use up any excess alumina" (Kleeman, 1965, p. 50) in the ingested sialic rocks and hornblende would be expected as a typical mafic mineral. However, the Sinclair granites and rhyolites are typically peraluminous and hornblende is a fairly uncommon mafic constituent of the granite, biotite being rather more characteristic.

The position of sample BW 1552 in the system An-Ab-Or-Qz (coinciding roughly with the differentiation trends of Fig. 29) is suggestive of an origin by fractional crystallisation of a basic magma which has been contaminated by sialic material as suggested by Kleeman (1965). The body of which sample BW 1552 is representative is very small (0,3 km^2) and can be satisfactorily explained by this mechanism. The high Qz content of the rock relative to the low Or content is probably the result of assimilation of a quartz-rich basement rock. Field evidence for this can be found in the occasional presence of metaquartzite micro-xenoliths in this rock.

In view of the arguments presented above, it is concluded that fractional crystallisation of basic magma, whether contaminated by sialic material or not, could not provide a suitable mechanism for the production of the vast majority of granitic magmas of the composition and volumes observed in the Sinclair Group felsic units.

It would seem, therefore, that an origin for the felsic rocks of the Sinclair by partial melting would explain more reasonably the tendency for the compositional clustering in the low-temperature trough of the An-Ab-Or-Qz or 'granodiorite' system. The nature of the source rock that underwent partial melting remains to be discussed.

The new concept of global tectonics has led to the suggestion that many of the world's granitic batholiths have been fed by partial melts derived from subducted oceanic crust (Dickinson, 1970; Matsumoto, 1968; Hamilton, 1969). These granitic batholiths are, therefore, envisaged as having a similar origin to the basalt, andesite and rhyolite of continental masses, e.g. the coastal batholiths of western South and North America.

In view of the fact that the Sinclair Group may have developed on the margin of an old continental mass, the so-called Kalahari Craton, at a time when subduction of oceanic crustal material was possibly taking place along this margin (Sections 3.4.1 and 5), it is pertinent to review modern opinion on a direct subduction origin for large granitic masses.

As Presnell and Bateman (1973) point out, the frequent association of granite batholiths with subduction zones marginal to continents does not necessarily imply that the magma has been derived by partial fusion of subducted oceanic crust. In their recent study on the genesis of the Sierra Nevada batholith, they show convincingly that the complete range of 'granitic' rocks constituting the batholith could be produced by a combination of equilibrium and fractional fusion of the lower crust.

It is also notable that granitic batholiths are restricted exclusively to areas underlain by continental crust. Granites associated with purely oceanic crustal settings are very minor when compared with the voluminous batholiths and rhyolites of continental environments. As an example of this relationship, Gilluly (1971) pointed out that in the western two-thirds of the Aleutian arc, where subducted oceanic crust underlies oceanic crust, the magmas produced are dominantly andesitic with only minor albite granite and rhyolite, whereas in the eastern part of the arc where sialic crust has been underthrust by the oceanic plate, stocks of 'granitic' rocks are abundant.

Brown (1973) has calculated that estimated continental crustal accretion rates can be satisfied at present solely by the introduction of 'andesite' magmas into plate margin magmatic belts. To account for the large volumes of felsic magma by the same subduction mechanism, rates of plate consumption would have to exceed by three times the present global estimate of oceanic crust consumption of 3,4 km^3yr^{-1} (Wright, 1971).

The probability also exists that considerable volumes of ocean-floor sediment are dragged down along subduction zones as the upper part of the consumed oceanic crust (Gilluly, 1971). Taking the North American plate as an example, Gilluly (1971) has concluded on the basis of volume relationships, that most of the ocean-floor sediments carried toward the continental margin during Mesozoic and Cenozoic times were transported down the subduction zone and not scraped off the diving slab to remain as a mēlange.

The possibility that these sediments would melt and the resulting magmas be added to the magmatic arc at the surface was suggested by Gilluly (1971) and Oxburgh and Turcotte (1970). Recent experimental work by Huang and Wyllie (1973) has shown that oceanic sediments, while being carried down in a subduction zone, will undergo two episodes of melting to produce granitic liquids, each episode beginning at different depths but possibly overlapping to some extent. The first liquid dissolves all pore fluid in the source rock, and the second liquid results from the dissociation of muscovite in a drier environment. These liquids will be produced at depths of 60 or 17 km and 92 or 21 km respectively, according to the thermal models for subducted slabs of Toksöz *et al.* (1971) and Oxburgh and Turcotte (1970). On migrating upwards the magma is emplaced into the overlying mantle, where the temperature would exceed that of the granite liquid by possibly as much as 200° - 300° C. Its chances of reaching high levels in the upper mantle and crust before crossing the granite solidus curve and solidifying are thereby considerably enhanced, providing that the rise of magma is slow, and that thermal equilibrium can be reached with the surrounding mantle.

Assuming that a subduction zone existed beneath the continental margin on which the Sinclair Group evolved (Sections 3.4.1 and 5), it is possible that the production of granite liquids from oceanic sediments took place in the manner outlined above. It is, however, doubtful whether such a process could account for the Sinclair felsic magmas. In the first place, magmas produced at the depths postulated by Huang and Wyllie (1973) would intrude the crust at distances of up to about 100 km from the continental margin and would, therefore, not coincide with the main axis of Sinclair felsic magmatic activity, which would have been at least 150-200 km away from the presumed margin of the Kalahari Craton during Sinclair times (Section 5). Furthermore, it is unlikely that large volumes of granite liquids of high water content, could ever reach high crustal levels or be extrusive (Cann, 1970), even considering the possibility of heating during their passage through the mantle.

The close association between granite batholiths (and related extrusives) and continental crust is now well established, and one is forced to the conclusion that felsic magmas, when produced on a large scale, have originated by partial fusion of sialic crust. The process of basic and intermediate magma production in the region of a subduction zone underlying the margin of a continent, and the production of felsic magma on a grand scale by fusion of sialic crust are not, however, unrelated processes. It is perhaps reasonable to suppose that the 'basic' magma on rising through the upper mantle and crust, provided the heat necessary for partial fusion.

In view of the arguments presented above in favour of a crustal origin for granite magmas in general, and other more specific evidence to be presented below, an origin by partial fusion of sialic crustal material is considered most probable for the Sinclair Group felsic magmas.

The Sinclair felsic rocks are all highly siliceous, ranging in SiO_2 content from 66,9 to 78,5 per cent, with an average value of 74,5 per cent (a few of the highest values may be due to secondary silicification). It is not likely that such silica-rich magmas could be produced in large volumes by partial melting of oceanic crustal material, especially in view of the absence in the field of the more basic members of the 'granite' family. For example, the majority of 'granites' occurring in the western parts of North and South America, proposed by Dickinson (1968, 1970) and Hamilton (1969) as having been produced in the region of a subduction zone, range from quartz-diorites and quartz-monzonites through granodiorites to true granites.

The strong association between granitic provinces and sialic crust, which was noted above, also exists in the Sinclair-Helmeringhausen area, thus further suggesting an origin by crustal fusion. Pre-Sinclair rocks, or 'basement', consist of the Kumbis Formation (Section 2.1) and the intrusive 'Grey granodiorite' (von Brunn, 1967) both of which are part of the sialic plate of the Kalahari Craton (Section 4). The peraluminous character of the Sinclair felsic rocks and the presence of biotite as the dominant mafic constituent also point to a crustal origin. The lack of calcium in the first products of partial fusion of sialic crustal material would result in an excess of alumina, as opposed to the fractionation of a basic magma discussed earlier, where the calcium is present in sufficient quantity to use up excess alumina, resulting in a sub- or metaluminous magma.

Major element data and the normative mineral constituents derived therefrom indicate an origin by either partial fusion or fractional crystallisation of a basic magma but, with the exception of the relative alumina content, do not provide the evidence necessary for distinguishing between the two processes. A study by Taylor *et al.* (1968) on leucogranites and rhyolites has shown that it may be possible to distinguish between the two processes by means of trace elements and, to a certain extent, interelement ratios. They found that partial fusion products are relatively depleted in Rb and enriched in Ba and Sr and, consequently, the Ba/Rb and K/Rb ratios are significantly greater for rocks of partial fusion origin than for those of fractional crystallisation origin. In Table 20 average values of the trace elements Rb, Sr, Ba, Zr and the ratios Ba/Rb and K/Rb for the Sinclair felsic rocks are compared with those of rhyolites (of probable partial fusion origin) and those of leucogranites (of probable fractional crystallisation origin) as given by Taylor *et al.* (1968). Trace element contents and interelement ratios of the average low-Ca granite of Turekian and Wedepohl (1961) are included as a 'standard' reference. Due to the fact that we are dealing here with particular, and possibly geologically restricted cases, it is not possible to make direct comparisons between the Sinclair felsic rocks and the rhyolites and leucogranites of Taylor *et al.* (1968). Nevertheless, using the principles involved in their study, it is apparent that, on the basis of trace element chemistry, the Sinclair felsic units compare more favourably with the rhyolites supposedly derived by partial fusion processes.

Rubidium values, although higher than those of the rhyolites of Taylor *et al.* (1968) are comparable with those of the average low-Ca granite of Turekian

	Sinclair Group felsic rock units		Rhyolites from Taylor *et al.* (1968)		Leucogranites from Taylor *et al.* (1968)		Average low-Ca granite (Turekian & Wedepohl 1961)
	Average	Range	Average	Range	Average	Range	
Rb	175	(44-354)	108	(93-160)	390	(220-720)	170
Sr	88	(17-379)	125	(41-170)	42	(7-100)	100
Ba	778	(85-2208)	870	(630-1080)	270	(22-670)	840
Zr	363	(127-693)	160	(80-220)	88	(52-140)	175
Ba/Rb	5,3		8,1		0,71		4,9
K/Rb	246		250		100		247

Table 20. Comparison of the trace element contents (in parts per million) of the felsic rock units of the Sinclair Group with those of rhyolites and leucogranites of Taylor *et al.* (1968) and the average low-Ca granite of Turekian and Wedepohl (1961)

and Wedepohl (1961), and generally do not show signs of enrichment. Barium values are comparable with those of the rhyolites of Taylor *et al.* (1968) and are possibly slightly enriched when compared with the average granite values of Turekian and Wedepohl (1961).

Zirconium is certainly enriched in the Sinclair rocks and lends further support to a crustal fusion origin (El-Hinnawi, Pichler and Zeil, 1969). Ba/Rb and K/Rb ratios are normal when compared with those of the rhyolites of Taylor *et al.* (1968) and certainly do not reflect the relative enrichment of Rb typical of rocks of fractional crystallisation origin.

From Table 20 it seems that the low Sr values of the Sinclair felsic rocks would favour an origin by fractional crystallisation, but as pointed out in Section 3.3.3, when the Sr values are plotted against the Ca contents of these rocks (Fig. 23) it is evident that they all plot either within or below, the main granite sequence of Turekian and Kulp (1956), thus indicating 'normal' to enriched values of Sr. Samples plotting below the "main sequence" of the log vs. log Ca plot are, according to Turekian and Kulp (1956), best explained by magmatic differentiation.

The spread of Sinclair felsic rocks in Fig. 23, ranging from approximately midway in the "main sequence" with Sr/Ca ratios of about 100 to positions below the "main sequence" with Sr/Ca ratios up to 1000, therefore indicates some degree of magmatic fractionation which probably took place at a fairly high level in the crust as the granite liquids rose to the surface. The porphyritic nature of the rhyolites and much of the granite further indicates the likelihood of such differentiation processes having operated at high levels.

Heat Source for Fusion of the Crust

A source capable of providing sufficient heat to cause fusion within the crust during the genesis of the Sinclair Group can be found in the repeated injections of basic magma. The cyclic development of the Sinclair Group will be discussed in greater detail in Section 5 but it is important at this stage, to note that although there is a certain amount of overlap in events, each major phase of felsic activity is preceded by a phase of basic activity, the intensity of which is variable and to a great extent determines the intensity of the succeeding felsic event, i.e.

(i) The youngest and least intense phase of felsic igneous activity, that of the Guperas Formation and Rooiberg granite, follows on the relatively mild basic intrusive and extrusive activity of the same formation.

(ii) The most intense phase of felsic igneous activity, that of the Nubib granite (≡ Rooikam and Tumuab granites of Helmeringhausen area, von Brunn, 1967) and the rhyolites of the Barby Formation, was preceded by, and was partly contemporaneous with, the massive outpourings of basic lava and lesser basic intrusives constituting the bulk of the Barby Formation.

(iii) There is little evidence for a phase of intense basic igneous activity preceding the Nagatis Formation 'rhyolites' which must have covered a considerable area at the time of their emplacement (von Brunn, 1967). Only thin local flows of basic lava are evident in the field. However, it is quite possible that the first phase of basic igneous activity, initiating the development of the Sinclair Group, on intruding into a relatively 'cool' crust, would have solidified more readily and at a lower crustal level than the basic magma of subsequent cycles, and might never have reached the surface in appreciable quantities. Furthermore, intrusion, extrusion and sedimentary deposits of the younger cycles could very easily have obliterated all evidence of such basic magmatic activity at the present level of exposure.

The most reasonable explanation for the observations presented above (i.e., basic magmatic activity followed by felsic magmatic activity in roughly proportional quantities) is that the basic magma, on passing through the crust, raised the crustal temperature gradient to a level where partial fusion of the sialic rocks could take place.

Von Brunn (1967, p.156), in considering the Sinclair granites and rhyolites west of Helmeringhausen has suggested that a concentration of radioactivity in the sialic crust, together with deep-seated crustal deformation, might have provided the heat necessary to cause fusion and production of felsic magmas. Although these might have been contributing processes, particularly since the rate of radiogenic heat production is considered to have been much higher in Precambrian than in Recent times (Dickinson and Luth, 1971), it seems unnecessary to invoke them as principal heat sources in view of the basic-felsic succession of events as discussed above, and the inevitable transfer of heat from mantle regions to the crust by means of repeated basic magma injections.

If the conclusions of Presnell and Bateman (1973) for the generation of the Sierra Nevada granite magmas are applied to the Sinclair magmas, a semiquantitative impression can be gained of the possibility of fusion of pre-Sinclair crust through heat supplied by basic magma.

Assuming that the Sinclair potassic calc-alkaline magmas were derived from a subduction zone and regions within the upper mantle above such a zone (Sections 3.4.1 and 5), their depth of origin would be in the region of 150 - 175 km. Taking 160 km ($\sim$50 kb) as an average and extrapolating the data of Green and Ringwood (1968), the temperature required to produce a liquid of basaltic andesite composition (the most common 'basic' lava type of the Sinclair Group) at that depth would be in the order of 1600°C under anhydrous conditions.

It is, however, unlikely that the magma would be completely anhydrous due to the incorporation of low-melting hydrous mineral phases such as phlogopite into the basic melts, and Wyllie (1973) considers 2 per cent as a reasonable maximum water content to allow melting in the depth interval of 100 - 200 km along the upper parts of a subducted slab.

The presence of 1 per cent water would lower the liquidus temperature to about 1500°C (Presnell and Bateman, 1973; Wyllie, 1973), a temperature compatible with those calculated for the upper parts of a subduction zone at a depth of 160 km, according to the models of Toksöz *et al.* (1971) and Oxburgh and Turcotte (1970). Higher water contents would lower this melting temperature considerably, but such a water-saturated magma could not rise very far before the water vapour pressure would exceed the confining pressure and the liquid would be forced to crystallise. Wyllie (1973) has further argued that subducted oceanic crust would, in any event, probably have lost most of its pore fluid at depths of 150 km.

If the basaltic andesite magma rose fairly rapidly it would reach the base of the crust ($\sim$36 km or $\sim$10 kb) at a temperature of about 1400°C (Presnell and Bateman, 1973), and would be in a superheated condition since the liquidus temperature for basaltic andesite with 1 per cent water would, at this pressure, be in the region of 1150-1200°C (Wyllie, 1973).

The presence of plagioclase and clinopyroxene phenocrysts in most of the Sinclair basaltic andesites indicates that the magma was erupted at temperatures at or slightly below the liquidus curve, further suggesting that the superheat was lost to the crustal rocks, raising the geothermal gradient and initiating partial fusion.

Even if the rise of basaltic andesite magma was slow and it reached the base of the crust approximately at its liquidus temperature, it would still have the capacity to raise the crustal temperature gradient. The melting temperature vs. pressure relationship is approximately positively linear (Green and Ringwood, 1968; Green, 1972) so that, as the magma rose through the crust and the confining pressure decreased, the liquidus temperature would also decrease. Therefore, in order for the magma to reach the surface at or slightly below the liquidus, heat must be lost to the crust.

Basic magma which cooled sufficiently to solidify within the crust before actually reaching a high-level, as might have been the case during the initial stages of the Nagatis eruptive phase, would serve as a heat reservoir. The solidus temperature of basaltic andesite (plus 1 per cent H_2O) under the pressures found within the crust would be about 1000-1050°C), still considerably above that required to cause partial fusion of sialic crustal rocks.

Crustal thickness

Throughout this discussion it has been assumed that the thickness of the continental crust in the Sinclair-Helmeringhausen area during the development of the Sinclair Group was in the order of 36 km. Some authors (e.g. Windley, 1973; Shackleton, 1973) have argued for a thinner crust during the Middle Precambrian, but recently Condie (1973) has demonstrated that crustal thickness in stable areas has remained essentially constant at about 35-38 km from the Archean (> 2,500 Ma) to the present. Condie's (1973) estimates of stable crustal thickness throughout geological time are based on estimated burial

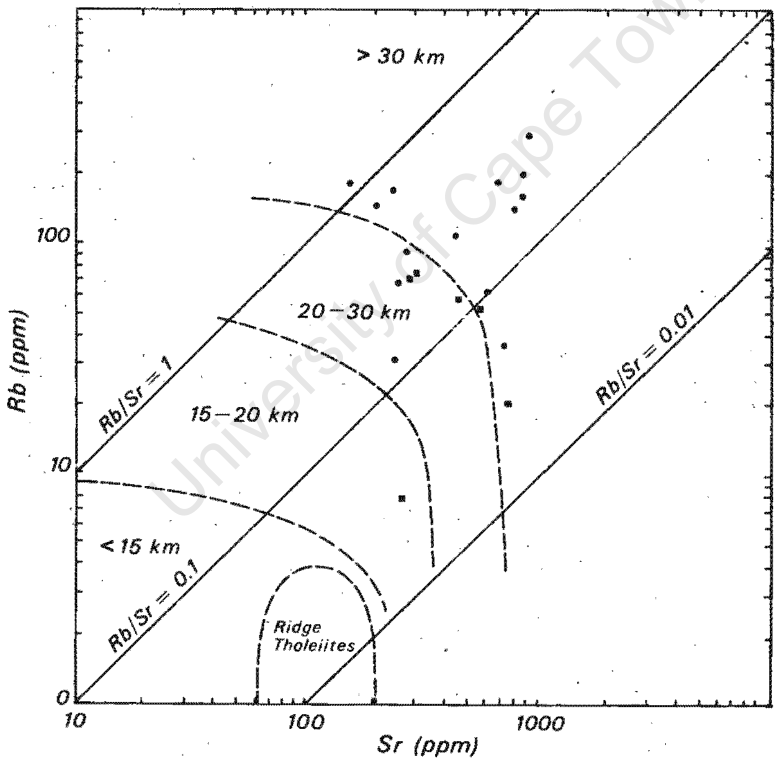


Fig. 30: Rb-Sr distribution in the calc-alkaline (solid circles) and tholeiitic (solid squares) basic volcanic rocks of the Sinclair Group, with inferred crustal thicknesses *after* Condie (1973)

depths of high P-T mineral assemblages exposed in Archean granulite facies terraines, exposed stratigraphic sections in Archean greenstone belts and certain geochemical indices. He found that a relationship exists between crustal thickness and the K_2O, Rb and Sr values of volcanic rocks of the arc-tholeiite, calc-alkaline and shoshonite series.

Plotting the Sinclair basic volcanic rocks on the Rb-Sr crustal thickness grid of Condie *et al.* (1972) in Fig 30, the indications are that the crust was thicker than 30 km during the evolution of the Sinclair Group. The stability of the crust in the Sinclair area (part of the large Kalahari Craton), is evidenced by the virtually complete lack of large-scale deformation since Sinclair times. No data are available on the present thickness of the Kalahari Craton, but the extreme stability would suggest thicknesses comparable to those of other modern stable crustal areas, i.e. in the order of ∿35-40 km. Since the Craton has remained essentially unaffected by large-scale tectonic events, since Sinclair times, it is regarded as having had approximately the same thickness throughout this period, comparable with thicknesses of continental crust in similar modern environments. The effects of possible thickening by under- or intra-plating by younger magmas (Pakiser & Ziets, 1965) in post-Sinclair times cannot be fully evaluated, but in view of the evidence presented by Condie (1973) for a constant continental crustal thickness since the Archean, these effects are probably not significant. A total lack of post-Sinclair magmatic activity is further evidence against thickening by magmatic injection.

It is almost certain that, during the evolution of the Sinclair Group, the crust underlying the Sinclair area increased in thickness due to the considerable basic extrusive activity; so it is possible that a thickness of ∿36 km can be considered a minimum especially for the younger units of the Group. In any event, a crustal thickness of > ∿36 km would not invalidate the arguments presented in this Section for a crustal fusion origin of the felsic units of the Sinclair Group. On the contrary, a thicker crust would increase the probability of crustal anatexis.

Geothermal gradient

It is not possible at this stage to determine the geothermal gradient for the crustal segment underlying the Sinclair-Helmeringhausen area during development of the Sinclair Group but, by analogy with modern examples, reasonable estimates can be made. Thermal gradients observed in stable continental areas are in the order of 10^{o}-$15^{o}C\ km^{-1}$ (Hyndman, 1972; Winkler, 1967). Such gradients are not capable of causing fusion within the crust, but much higher heat flow values characterise the marginal zones of continental masses during the development of magmatic belts such as the 'Andean' or 'Cordilleran' type. Uyeda and Watanabe (1970) have determined that, for the South American continent, the geothermal gradient attains a peak of about $40^{o}C\ km^{-1}$ over the Andean mountain chain, which has been a zone of extensive magmatism from the Mesozoic to the Present. Normal gradients for this continent are less than $20^{o}C\ km^{-1}$. Uyeda and Watanabe (1970) also recorded a few very high values in the order of 40-$60^{o}C\ km^{-1}$, but these are probably suspect since they are associated with local geothermal activity. It can be speculated, therefore, that the geothermal gradient prevailing in the Sinclair-Helmeringhausen area during the development of the Sinclair Group would have fallen into the limits set by the $20^{o}C\ km^{-1}$ and $40^{o}C\ km^{-1}$ gradients.

It has been suggested (Saggerson and Owen, 1969; Ray 1970) that geothermal gradients were higher in Precambrian times than today, due to the thinner crust and possibly also due to a higher rate of radiogenic heat production (Dickinson and Luth, 1971) since the heat producing radioactive elements were distributed throughout the crust rather than concentrated in the uppermost levels as they are today (Heier, 1973; Lambert, 1971). However, as pointed out earlier, a crust of 'normal' thickness ($\sim$36 km) is advocated for during the evolution of the Sinclair Group so it is considered acceptable to use gradients observed in modern continental crust and in similar geological environments. Furthermore, the effects of radiogenic heat production might not have been very different from those of today.

Fusion Relationships

The temperature of beginning of melting in the crust may, under water-saturated conditions, be below 700°C and possibly as low as 640°C (Tuttle & Bowen, 1958; von Platen, 1965; Winkler, 1967), the minimum temperature for the granite system. It is, however, not likely that high-level felsic intrusives and extrusives, such as those constituting a large proportion of the Sinclair Group, were formed under conditions of water-saturation and at such low temperatures. A study of the phases present in glassy rocks (Carmichael, 1967) further indicates that the temperatures of eruptive granitic liquids may be in the order of 900°C. Granite melts formed at depth and having high water contents would not be able to rise very far in the crust before the solidus curve was intersected and the magma was forced to crystallise (Fig. 31). (Harris *et al.*, 1970; Cann, 1970). Heat of crystallisation may increase the magma temperature but the cooling effects of exsolution of the dissolved water and its adiabatic expansion might be sufficient to counteract most of the heat of crystallisation (Harris *et al.*, 1970).

The high-level and extrusive nature of the Sinclair felsic rocks indicate that the melts had low water contents and were, therefore, able to exist at shallow depths within the crust without crystallising to the extent where mobility was appreciably restricted. A complete lack of contact metamorphic aureoles and pegmatite phases further rules out the possibility of a high water content. On the other hand, the presence of at least some water in the Sinclair felsic magmas is indicated by the presence of ash-flow beds, biotite as a mafic mineral in the granites and sparsely distributed miarolitic cavities in some of the granites.

Considering the extrusive and high level emplacement of the Sinclair felsic magmas and resultant low pressure crystallisation conditions ($< \sim 1$ kb), reference to Fig. 31 suggests maximum water contents for these magmas in the order of 4 per cent, with an average of about 2 per cent. It is generally agreed that felsic extrusives and high-level granites have had water contents in the region of 1-2 per cent (Fyfe, 1970; Tuttle and Bowen, 1958). Lower crustal regions are considered to be in a dessicated state (Ringwood and Green, 1966; Fyfe, 1970 and 1973; Presnall and Bateman, 1973; Heier, 1973) and Wyllie (1971) regarded 1 per cent pore fluid as a generous estimate for lower crustal rocks. It is, therefore, likely that fairly extensive melting of the lower crust will produce granite magmas of low water content (1-2 per cent), provided a high enough temperature is attained.

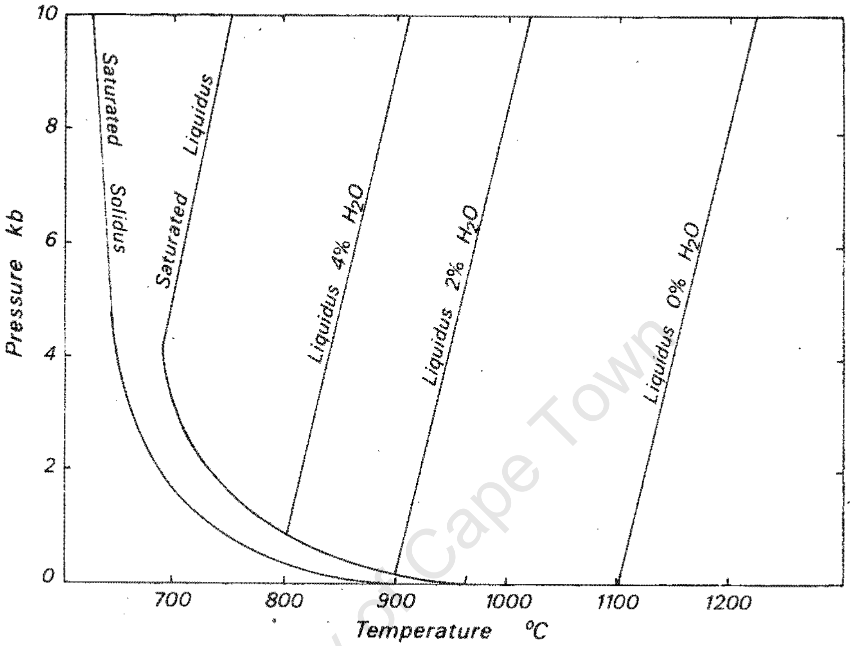


Fig. 31. Anhydrous and water-saturated liquidus and water-saturated solidus curves for granite. The anhydrous solidus at atmospheric pressure is denoted by a short hatch mark between 900 and 1,000°C. (*after* Harris, Kennedy and Scarfe, 1970)

According to Brown and Fyfe (1970) the low water contents of felsic magmas that are extrusive or that intrude high levels in the crust can be provided solely by the melting of hydrated phases such as biotite and hornblende. In such cases primary melting of the quartzo-felspathic crustal host rock is delayed until the hydrous phase becomes unstable, often at considerably elevated temperatures. On the basis of experimental work, Brown and Fyfe (1970) have determined the pressure - temperature relationships for the beginning of melting of dry granitic material when the only water available is that provided by the decomposition of the various hydrated mineral phases (Fig. 32).

Considering the extrusive and high-level nature of the Sinclair felsic units and the mafic mineralogy of the granites (mainly biotite with lesser amounts of hornblende and rare muscovite), it is probable that the magmas formed as a result of fusion of crustal material within the range of temperatures and pressures in which the melting of biotite and hornblende is dominant. This area of Sinclair felsic magma production has been indicated in Figure 32 by shading, and is bounded at the high pressure end by the lower boundary of the crust which in this instance is taken as about 36 km. The minimum depth at which granite liquids could be produced would have been limited by the maximum geothermal gradient operative during Sinclair times, which, as already discussed, is taken as 40°C km^{-1}.

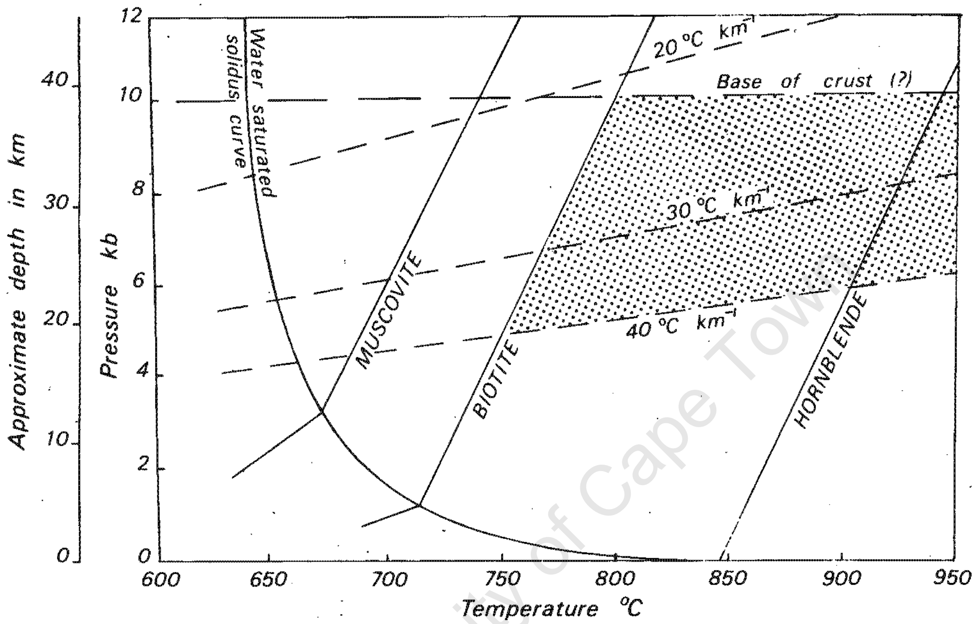


Fig. 32. The beginning of melting recognised within dry granitic material resulting from the dissociation of the various hydrous phases (*after* Brown and Fyfe, 1970). Also indicated is the water-saturated solidus curve for granite and possible geothermal gradients operative during development of the Sinclair Group. Major production of Sinclair granitic liquids most probably took place within the shaded area

Granitic liquids formed by the melting of muscovite would possible contain sufficient water to restrict the magma's ability to rise to high levels in the crust, since at a depth of about 12 km the solidus curve would be intersected and the liquid would crystallise. Furthermore, the melting of a quartzo-felspathic host and muscovite only is not likely to produce biotite in the final rock due to the lack of ferro-magnesian elements in the resulting granitic liquid.

It is thus proposed that the majority of the Sinclair felsic magmas were produced by partial fusion of the crust between depths of 18 km and 36 km, that is, at all levels throughout the lower crust and possibly also the lower parts of the upper crust, and at temperatures between about 750°C and 950°C.

From Figures 31 and 32 it is further apparent that, during a particular geothermal cycle in which the temperature gradient increases to a peak, the earliest granitic melts will be the wettest, since with increasing temperature the water contents of the magma would decrease as the hydrated phase changes by undergoing melting. It is then perhaps significant that the vast majority of explosively-eruptive felsic rocks of the Sinclair, the ash-flow tuffs or ignimbrites, are among the first representatives of felsic igneous activity within all three cycles of the Sinclair Group.

Vesiculation of a granitic liquid giving rise to a fluidised solid-in-gas system (Smith, 1960; Boyd, 1961), to be erupted as ash-flow tuff, must take place in the P-T region bounded by the liquidus and solidus curves in Figure 31, i.e. vesiculation will not occur until the water saturated liquidus curve is reached, when $P_{H_2O} > P_{Total}$. Once the solidus is intersected by the cooling curve, no vesiculation can take place due to a completely crystalline state. It follows that the more vitric ash-flow tuffs of the Sinclair like some of those constituting parts of the Basal volcaniclastic and Rhyolite members of the Barby Formation, would have vesiculated at P-T conditions very close to the liquidus and hence do not carry many phenocrysts. The crystal-rich ash-flow tuffs of the Guperas Formation, on the other hand, owe their characteristic strongly phenocrystic nature to vesiculation at P-T conditions well within the area bounded by the liquidus and solidus curve.

The exact P-T conditions under which a granitic liquid will vesiculate depend on the initial H_2O content. Boyd (1961) has demonstrated that for ash-flow tuff to display any degree of welding, the initial water content cannot exceed 4 per cent. For higher water contents the cooling factor, due to the evaporation and expansion of gas, is too high to allow welding to take place. Since the ash-flow tuffs of the Sinclair Group invariably display a welded character it is assumed that their initial water contents were at or below 4 per cent. From Figure 31 it can be deduced, therefore, that vesiculation of the Sinclair felsic magmas (which formed ash-flow tuffs) took place between the temperatures of about 800°C and 950°C, and at pressures of between 0 and 1 kb (0-4 km depth). The upper temperature limit is fixed by the intersection of the saturated liquidus with atmospheric pressure conditions. Above this temperature the ascending magma will never intersect the saturated liquidus curve, P_{H_2O} will always remain less than P_{Total} and extrusions will take place without vesiculation. Limits of ~1,5 to ~4 per cent can, therefore, be placed on the initial water contents of those granitic magmas that gave rise to ash-flow tuff within the Sinclair Group. Since the fluidised system would have resulted in very rapid transport to the surface, it is probable that eruptive temperatures were not much below the temperatures at which vesiculation took place (i.e. 800-950°C).

As pointed out by Brown and Fyfe (1970), the normative orthoclase contents of felsic magmas produced by partial fusion are dependent on the hydrated mineral associated with fusion. High temperature melts from hornblende tend to be granodioritic and lower temperature biotite and muscovite melts are granitic. The presence of biotite and amphibole as the dominant mafic minerals in the Sinclair granites to a great extent precludes muscovite melting only, for the reasons already discussed. Furthermore, the Ab/Or ratio of the melt will increase as a function of increasing pressure, complimenting the above trend for the production of granodioritic liquids at higher temperatures and granitic magmas at lower temperatures. The increase in the Ab/Or ratio is due to the fact that under conditions of increasing pressure, the composition of the granite ternary minimum becomes progressively enriched in the albite component (Tuttle and Bowen, 1958; Luth, Jahns and Tuttle, 1964; Kleeman, 1965).

Applying these principles to the Sinclair Group, it can be postulated that those felsic units that show granodioritic and rhyodacitic tendencies, namely the Haremub-Kotzērus granite and the high-Ca Barby rhyolite would be generated at the highest temperatures and pressures, i.e. at the deepest crustal levels involving the melting of amphibole. The presence of biotite as a mafic constituent of the Haremub-Kotzērus granite can possibly be accounted for by the early removal of Ca from the melt during the crystallisation of plagioclase. The residue of Mg and Fe in the melt resulting from such a process would, therefore, be used to form the late-crystallising biotite observed in these rocks.

Further evidence for the involvement of amphibole in the formation of the Haremub-Kotzērus granite and high-Ca Barby rhyolite melts can be found in the very low Nb contents of these rocks as compared with the other felsic rock-types of the Sinclair Group. Nb most commonly substitutes for Ti and since hornblende is appreciably poorer in Ti than is biotite, it is reasonable to suppose that melts involving amphibole as a hydrous phase will reflect this in having relatively low concentrations of Nb. Analyses of hornblende and biotite from dioritic, monzonitic and adamellitic rocks by Miller (1972) for example, indicate that the Ti content of hornblende is usually less than half of that in biotite. It follows that the Nb content of hornblende is very much lower than that of biotite.

With the exception of the Haremub-Kotzērus granite and the high-Ca Barby rhyolites, variations in bulk chemical composition of the Sinclair felsic rocks are relatively minor. This is to be expected if the bulk composition of the crustal rocks undergoing partial melting is not vastly different from that of the granodioritic system cotectic composition. The lower crust is considered to be of intermediate composition (Ringwood and Green, 1966; den Tex, 1965), and "... of the common rock-types andesite would have the most mafic composition capable of satisfying typical lower crustal densities and seismic velocities under relatively dry conditions" (Presnall and Bateman, 1973; p. 3187).

The average lower crust is, therefore, likely to be more felsic than andesite and, although progressive melting will move the resulting liquid toward the bulk composition of such lower crustal source rocks, there will be a large melting interval during which appreciable volumes of liquid approximating the cotectic composition will be produced.

Despite this lack of variation in the quartzo-felspathic normative constituents of most of the Sinclair felsic rocks, there is a moderate variation in the trace element contents of these rocks which is not always consistent with variations in major element chemistry (Section 3.3.3), e.g. variations in Sr content are not always consistent with variations in Ca content. Partial melting of crustal material, involving the decomposition of biotite or hornblende, will result in a melt with trace element contents that might in part reflect those of the most dominant of these two hydrous minerals.

Variations in trace element chemistry, such as those observed in the Sinclair granites, rhyolites and quartz-porphyries could possibly be explained, therefore, in terms of the melting model discussed above involving the hydrated minerals biotite and hornblende. Concentrations of the trace elements Rb, Sr, Ba and Nb in biotite and hornblende differ very markedly in the two minerals and since melting at slightly varying depths and temperatures would produce liquids in which the proportion of biotite to hornblende varies considerably, it is possible that the differing trace element contents of these two minerals would be combined in varying proportions to produce the variations in trace element chemistry existing in the Sinclair felsic rocks.

Summary of Conclusions:

1. The normative compositions of the Sinclair felsic rocks can be plotted in the An-Ab-Or-Qz system and all except one plot within the low temperature trough of the system, indicating a magmatic origin for the vast majority of these rocks.

2. The felsic magmas were derived by partial fusion of crustal material.

3. A heat source for fusion within the crust can be found in the repeated injection of basic magma.

4. The provision of heat for crustal fusion by basic magma injections has resulted in a basic, followed by felsic, cyclic development of the Sinclair Group.

5. Crustal thickness during development of the Sinclair can be estimated at about 35-38 km.

6. The geothermal gradient operative during development of the Sinclair Group is estimated as having reached a peak of about $40^{o}C\ km^{-1}$.

7. Water contents of the felsic magmas, giving rise to the rocks exposed today, were probably in the region of 1-2 per cent and could not have exceeded about 4 per cent.

8. The dissociation of biotite and hornblende, together with the quartzo-felspathic crustal host rock, could have provided the water necessary for melting.

9. The Sinclair magma originated from melting of the crust between depths of about 18 km and 36 km and between temperatures of about $750^{o}C$ and $950^{o}C$.

10. The high-Ca Barby rhyolites and the Haremub-Kotzérus granite are the result of melting at the deepest levels mainly involving the decomposition of hornblende.

11. Ash-flow tuffs of the Sinclair are the result of vesiculation of felsic magma at depths of between 0 and 4 km, at temperatures of between $800^{o}C$ and $950^{o}C$ and with water contents restricted to a range of about 1,5 to 4 per cent.

12. Variations in trace element chemistry in the Sinclair felsic rocks can possibly be attributed to the melting of varying proportions of biotite and amphibole.

4 CORRELATION AND AGE

4.1 CORRELATION WITH ADJACENT AREAS

As a result of the present study (Area A in Fig. 33) and the investigation of von Brunn (1967) (Area B in Fig. 33), it has been possible to determine more fully the geographical extent of the various units comprising the Sinclair Group and it is also probable that this Group can now be defined in its entirety.

The respective geological successions in the areas north of Sinclair and west of Helmeringhausen (von Brunn, 1967) are presented in Table 22 for comparison. The latter succession has been slightly modified by the writer from that originally proposed by von Brunn (1967) but the only changes introduced are those of interpretation involving units whose stratigraphic position was considered uncertain by this author. The correlations suggested in Table 22 and discussed below have been used in this generalised map.

The oldest unit belonging to the Sinclair Group is the *Nagatis Formation* (von Brunn, 1967, 1969) which is restricted to areas south of latitude 25°45'S and does not, therefore, occur in the present study area.

On the basis of stratigraphic position, field characteristics, and petrographic and chemical properties (Table 23), a correlation of the *Haremub granite* with the *Kotzērus granite* of von Brunn (1967) is most probable.

The *Kunjas Formation* is present in both areas although as discussed in Section 2.2.3, the upper limit of this formation has been redefined.

Outcrops of the *Barby Formation* are also continuous from one area to the other so that the extrusives of this unit are well represented between latitudes 25°00'S and 26°00'S. Changes in facies do occur, however, and these have been discussed in Section 3.1.2. Basic intrusives which are abundantly exposed in the southwestern part of the Sinclair area are only sparsely, if at all, represented further south. A few small basic bodies of post-Kunjas age are present about 15-20 km west-southwest of Helmeringhausen, but since these are very poorly exposed and were not discussed in any detail by von Brunn (1967), a correlation with the basic intrusives north of Sinclair would be highly speculative.

No large syenitic intrusions such as the *Spes Bona syenite* occur in the area west of Helmeringhausen although von Brunn (1967) noted the presence of a localised syenitic rock associated with the large mass of Tumuab granite. Despite similarities in major element chemistry between this rock and the Spes Bona syenite, it can be concluded from the trace element data, Rb and Sr in particular (Table 23), and the description of von Brunn (1967, p. 107-110) that this syenite represents a local felspar-rich phase of the Tumuab granite as originally suggested by von Brunn and is not related to the older potassic basic magmatic event of the Barby Formation.

On the basis of field relationships and great similarities in petrology and chemistry (Table 23), the *Nubib granite* as occurring north of the Sinclair Mine is the direct correlate of the *Rooikam granite* of von Brunn (1967). The relationship of the *Tumuab granite*, occurring west of Helmeringhausen (von Brunn, 1967) to other units of the Sinclair Group is not clear due to the lack of suitably exposed contacts. An intrusive relationship to the Nagatis Formation and the Kotzērus granite makes it younger than these units and on the basis of 1020±80 Ma and 1290±80 Ma ages obtained from the Tumuab granite (von Brunn and Dodson, 1967), von Brunn (1967) assumed two main phases of intrusion, i.e. the first being a pre-Kunjas and post-Kotzērus phase and the second being of post-Barby age and synchronous with the Rooikam phase. The Tumuab granite is also cut by quartz-porphyry and granite-porphyry dykes of probable Guperas age, so it would appear to pre-date this Formation. The above-mentioned absolute ages were derived from isotopic Rb and Sr determination of whole-rock material and the isochrons were based on only two samples in each case. These ages are, therefore, not considered very reliable. The age of 1290±80 Ma seems to correspond fairly well with the U/Pb zircon age of 1360±50 Ma for the Rooikam granite in the immediate vicinity of Helmeringhausen (Burger and Clifford *in* Burger and Coertze, 1973). In view of the above ages and the close petrographic and chemical similarities between the Tumuab and Rooikam granites (von Brunn, 1967), it is considered likely that the two can be correlated and together with the Nubib granite, represent a single episode of intense felsic magmatic activity.

An age of 940±40 Ma was also obtained by Burger and Clifford (*in* Burger and Coertze, 1973) for a 'granophyric granite', on the farm Dabis near Helmeringhausen (probably the Rooikam granite). In view of the other ages available for the granites of the Sinclair Group, this age is regarded as being far too low for the intrusive event.

The *Nam Shear Belt* and the mylonitised rocks within this belt might well represent the effects of tectonics that gave rise to a similar northwesterly-trending shear zone in the southern part of the Helmeringhausen area in which the *Tiras gneiss* of von Brunn (1967) is the prominent rock-type. Muscovites from small secretion pegmatites in the central parts of the Nam Shear Belt have yielded K/Ar ages of 1127±45 Ma, 1129±45 Ma and 1143±46 Ma (D.C.Rex, pers. comm.). These ages must, however, be regarded as minimum ages in view of the probability of argon loss. The localities of these samples appear in Fig. 36 in the Appendix. The trend and minimum age of the above shear belts (1127±45 Ma) is suggestive of a relationship to the youngest and brittle phases of deformation of the Namaqua Mobile Belt lying to the south and southeast (Joubert, 1971).

The *Guperas Formation* is very well developed in the area north of Sinclair whereas, south of latitude 25°45'S the formation is only represented by minor occurrences of conglomerate, sandstone and flagstone, and the 'granite-porphyry' and 'late acid' dykes of von Brunn (1967) which are marginal to the dense felsic dyke swarms north of latitude 25°45'S.

The *Rooiberg granite* represents the youngest phase of igneous activity within the Sinclair Group. No equivalent granite of post-Guperas age has been reported from the area west of Helmeringhausen. A minimum age of 1270 Ma has been obtained for the Rooiberg granite (U/Pb method; A.J.Burger, pers. comm.).

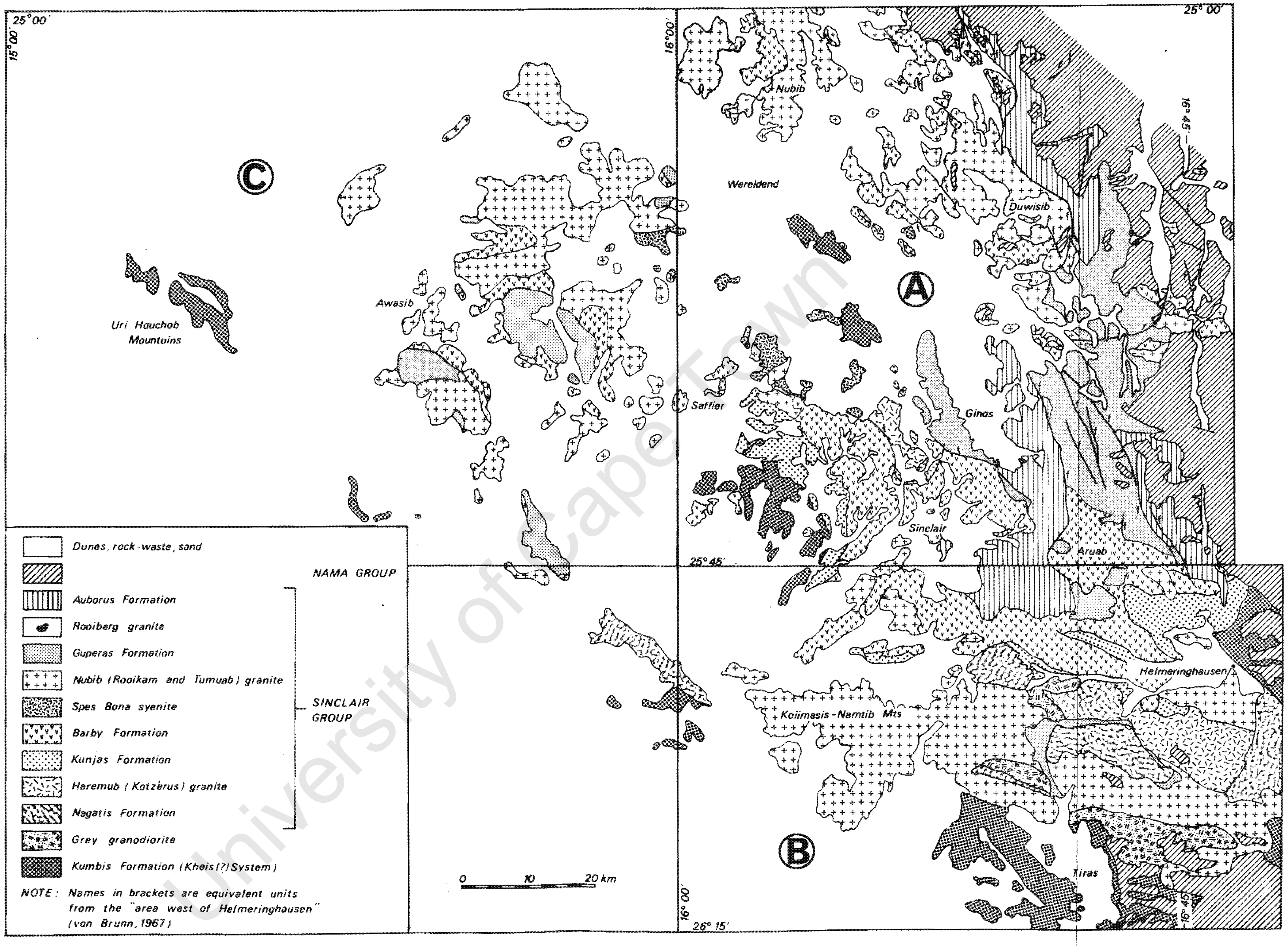


FIGURE 33. GENERALISED GEOLOGICAL MAP OF THE SINCLAIR (A), HELMERINGHAUSEN (B) AND AWASIB (C) AREAS

TABLE 22. TENTATIVE CORRELATION OF THE LATE PRECAMBRIAN FORMATIONS COMPRISING THE REHOBOTH MAGMATIC ARC

	HELMERINGHAUSEN AREA (modified *after* von Brunn, 1969)	SINCLAIR AREA (Watters, 1972, 1974)	AWASIB MOUNTAINS AREA (in part *after* unpublished map, C.D.M.)	NAUCHAS AREA (de Waal, 1966)	REHOBOTH AREA	BOTSWANA (Boocock, 1968; Thomas, 1973)
SINCLAIR GROUP in the Helmeringhausen, Sinclair and Awasib areas	<u>Auborus Formation</u> Red felspathic sandstone, conglomerate	<u>Auborus Formation</u> Red felspathic sandstone, conglomerate, shale			<u>Klein Aub Formation</u> Conglomerate, shale, quartzite (Schalk, 1973)	<u>Ghanzi Formation</u> Felspathic sandstone, shale, conglomerate
		<u>Rooiberg granite</u> (1270 Ma)				
	<u>Guperas Formation</u> Basal facies only - lithic sandstone, conglomerate, felsic porphyry dykes (Okarus granite porphyry intrusion) ?	<u>Guperas Formation</u> Lithic sandstone, conglomerate, rhyolite lava, ash-flow tuff, tuff, rhyolite plugs, quartz-porphyry dykes, basic lava, intrusives and dykes	<u>Guperas Formation</u> Non-volcanic facies - conglomerate, grit, sandstone, quartzite, shale, felsic porphyry intrusives	"<u>Guperas member</u> of the Sinclair Formation" (de Waal, 1966, p.107) Rhyolite lava, pyroclastic basic lava, sedimentary beds, quartz porphyry and granite porphyry dykes.	<u>Doornpoort Formation</u> (incl. Skumok porphyry, 1132±75 Ma) Red quartzites, conglomerate, slate, felsic volcanics, basic lava (Schalk, 1973) <u>Grauwater Formation</u> (incl. Opdam Formation) Younger Abbabis Formation. Quartzite basic lava, felsic volcanics and pyroclastics (Schalk, 1973)	<u>Kgwebe Formation</u> Sandstone, tuffaceous sandstone, diabase, felsic porphyry volcanics and intrusives (950±37, 1020±50 Ma)
	<u>Rooikam and Tumuab</u> granites (1020±80, 1290±80, 1360±50 Ma)	<u>Nubib granite</u>	<u>Younger Awasib granite</u>	<u>Nauchas granite suite</u> (including <u>Gamsberg granite</u>)	<u>Gamsberg granite</u> and Biesiespoort granite (1356±51 Ma) ?	
	<u>Barby Formation</u> Basic to intermediate lava	<u>Spes Bona syenite</u> <u>Barby Formation</u> Porphyritic trachyandesite and trachyandesite basalt, basalt, rhyolite, ash-flow tuff, tuff, volcanic conglomerate, quartzite, gabbroic, dioritic and monzonitic intrusives	<u>Barby Formation</u> Basic to intermediate lava, basic intrusives	<u>Alberta basic complex</u>		
	<u>Kunjas Formation</u> Basal conglomerate, arkose, shale	<u>Kunjas Formation</u> Basal conglomerate, arkose, shale		Gaub Valley schists and conglomerates?	<u>Gaub Valley member</u> of the Older Abbabis Formation ? (Hälbich, 1970)	
	<u>Kotzërus granite</u>	<u>Haremub granite</u>	<u>Older Awasib granite</u>			
	<u>Nagatis Formation</u> Ash-flow tuff, felsic lava, basic lava, agglomerate, arkose, shale, grit					

Table 23. Comparison between the average chemical compositions of some Late Precambrian rock units from southern and central South West Africa

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15
SiO_2	71,97	72,19	58,45	58,84	75,67	73,93	74,60	73,00	49,48	49,15	50,89	51,76	48,06	46,01	46,64
TiO_2	0,26	0,36	0,65	0,23	0,21	0,26	0,23	0,38	0,16	0,17	0,34	0,42	0,42	0,39	0,21
Al_2O_3	13,75	12,52	18,74	16,94	11,76	13,06	13,08	13,96	18,61	17,50	17,03	18,79	14,74	20,43	14,54
Fe_2O_3	1,41	2,34	2,07	0,42	1,16	1,00	0,73	0,76	1,71	1,85	2,51	2,19	1,93	2,75	2,29
FeO	0,94	2,23	1,61	0,90	0,78	1,35	1,03	1,26	5,09	5,83	5,68	5,93	8,80	5,87	4,93
MnO	0,04	0,11	0,06	0,09	0,04	0,04	0,03	0,07	0,12	0,14	0,14	0,15	0,21	0,13	0,14
MgO	0,59	0,33	1,43	<0,20	0,20	<0,20	<0,20	0,54	9,98	9,90	8,09	6,25	12,84	7,66	13,62
CaO	1,77	1,84	3,21	5,45	0,56	0,79	0,74	1,46	12,22	13,19	12,40	9,86	11,88	13,91	14,92
Na_2O	4,06	4,68	3,73	4,54	3,56	3,08	3,32	2,85	1,33	1,67	2,27	2,72	0,84	1,50	1,20
K_2O	3,81	2,80	8,02	9,13	4,97	5,63	5,47	4,68	0,27	0,20	0,26	0,35	0,08	0,24	0,12
P_2O_5	0,07	0,07	0,49		0,02	0,05	0,02	0,09	0,01	0,02	0,05	0,10			0,01
l.o.i.	0,73	0,56	0,76	3,88	0,48	0,66	0,43		0,56	0,07	0,18				
H_2O^-	0,05	0,13	0,05	0,20	0,06	0,27	0,18		0,02	0,05	0,04				
	99,45	100,16	99,27	100,82	99,47	100,32	100,06	99,05	99,56	99,74	99,88	98,52	99,80	98,89	98,62
Rb	69	76	327	270	206	263	260		8,6	5,5	3,8				
Sr	380	156	1184	234	45	67	54		187	107	349				
Ba	925		2822		434				114	76	178				
Zr	127		81		314				17,0	13,6	27,0				
Nb	5,4		1,8		23,6				<1,6	<1,6	<1,6				
Y	11,1		11		104				6,6	7,7	9,7				
K/Rb	456	306	203	281	203	179	176		267	302	573				

1. Haremub granite
2. Average of 3 Kotzerus granites (von Brunn, 1967)
3. Average of 3 Spes Bona syenites
4. Syenite (von Brunn, 1967)
5. Average of 9 Nubib granites
6. Average of 3 Rooikam granites (von Brunn, 1967)
7. Average of 4 Tumuab granites (von Brunn, 1967)
8. Gamsberg granite (de Waal, 1966)
9. Norite, Barby Formation
10. Gabbro, Barby Formation
10. Gabbro, Barby Formation
11. Norite, Barby Formation
12. Gneissic amphibolite, Alberta Complex (de Waal, 1966)
13. Meta hyperite, Alberta Complex (de Waal, 1966)
14. Amphibolite, Alberta Complex (de Waal, 1966)
15. Amphibolite, Alberta Complex (de Waal, 1966)

The discordant nature of the U/Pb determination for this sample (Table 28 in Appendix) makes it probable that the actual age of this rock is slightly higher than 1270 Ma. This age is, however, acceptably younger, by a reasonable margin, than that of 1360±50 Ma obtained for the Rooikam granite as discussed above. These two ages probably represent the two most reliable and useful ages thus far obtained for rocks of the Sinclair Group.

The *Auborus Formation* is best developed between latitudes 25°00'S and 25°45'S and does not extend farther south than the northern parts of the area mapped by von Brunn (1967).

From the study of unpublished geological reconnaissance maps of Consolidated Diamond Mines of South West Africa and aerial photographs of the Awasib Mountain Land in Diamond Area 2, as well as a brief field excursion to this area, it is apparent that the Sinclair Group extends, in part, to the west of longitude 16°00'E. A tentative geological succession for this area with regard to the Sinclair Group is presented in Table 22, and the geographical distribution of the various units is indicated in Fig. 33 (area C). With the possible exception of the basic lavas of the Barby Formation, volcanic activity does not appear to have been as prominent in the Awasib mountains as it was to the east. The absence of volcanic beds within the Guperas equivalents of this area has already been discussed in Section 2.2.7. On the other hand, granitic and to a lesser extent, rhyolitic, intrusive activity is well developed. At least two ages of granite are present; the older pre-dating basic lavas (Barby equivalents) and the younger invading the lavas but overlain by sedimentary beds correlated with the Guperas Formation. The latter are intruded by quartz-porphyry and felsitic bodies and dykes which represent the only evidence for post-Guperas magmatic activity. Small basic intrusives, probably emplaced at the same time as the gabbros of the Sinclair area, are numerous.

On the basis of the correlation discussed above, a reasonably accurate geological succession can be drawn up for what is proposed as the Sinclair Group (Table 22).

4.2 REGIONAL CORRELATION

The distribution of all rock units within the age range ±900-1350 Ma and occurring in western Botswana and in southern and central South West Africa north of the Namaqua Mobile Belt (i.e. broadly speaking, pre-Damara, pre-Nama and post-'basement') is shown on the generalised geological map of Fig. 34. The data have been obtained from published sources (Geological Map of South West Africa, 1963; Martin, 1965; de Waal, 1966; von Brunn, 1967; Boocock, 1968; Hälbich,1970; Schalk, 1973). From this map it is apparent that:

(i) The distribution pattern of these units has the form of a prominent and extensive curvilinear feature or 'arc'.

(ii) The Sinclair, Helmeringhausen and Awasib areas form a prominent part of this arc.

(iii) The arc separates the undeformed Late Precambrian deposits (Nama Group) of the stable platform area (Kalahari Craton) from the highly deformed Late Precambrian deposits of the Damaran Orogenic Belt, suggesting that the arc might occupy a marginal cratonic position.

(iv) The trend of the arc and of the arc-structures, such as major faulting, parallels that of the Damaran structural trends suggesting a possible relationship between the geotectonic development of the arc and the Damaran orogeny.

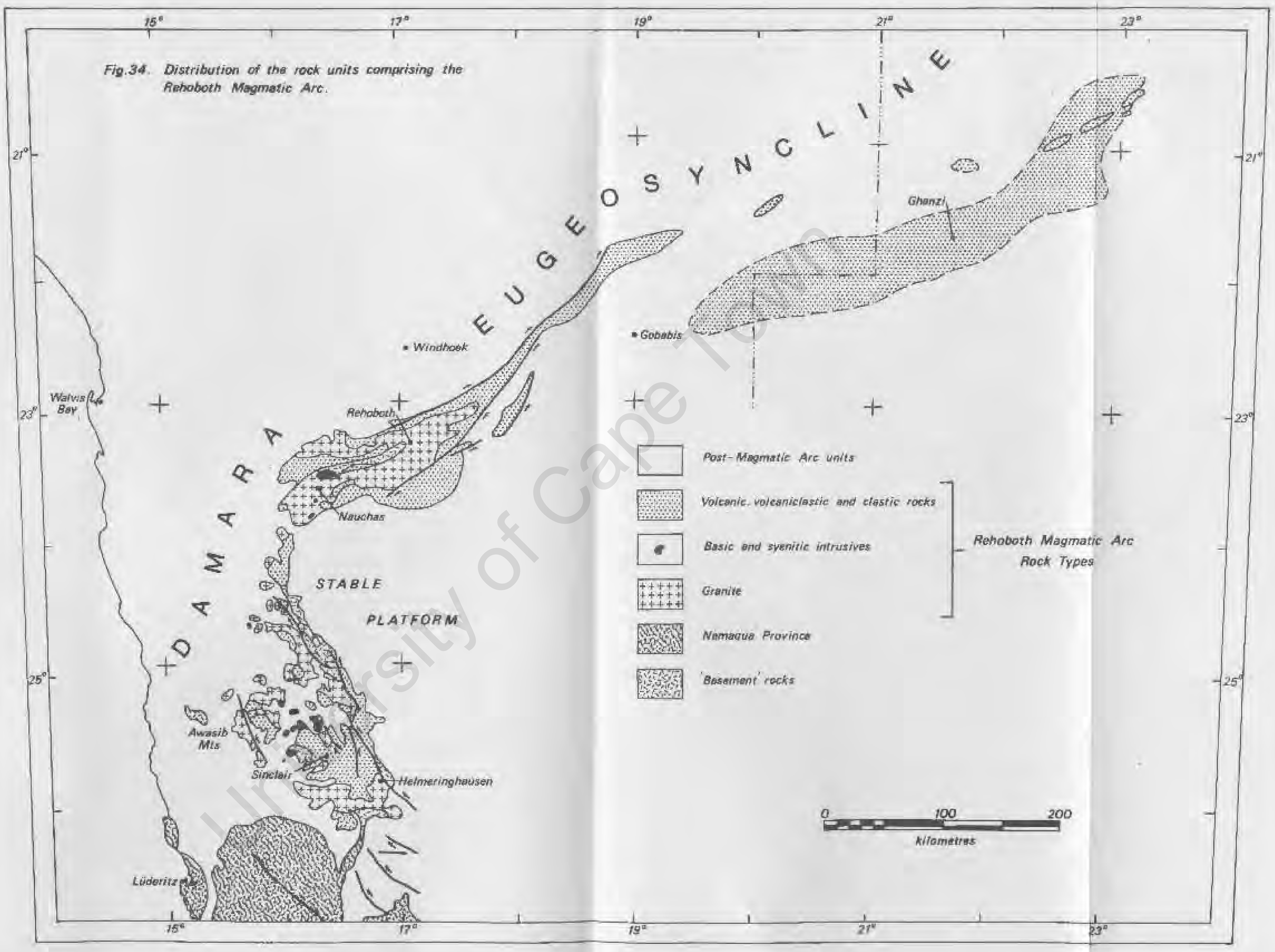
The geology along this arc outside the Sinclair-Helmeringhausen-Awasib area also consists essentially of various sequences of basic to felsic lava, volcaniclastic and various sedimentary units frequently of obvious local provenance, deposited on an older sialic basement of unknown age, and intruded by large amounts of co-magmatic high-level granite, rhyolite and basic material. The name "Rehoboth Magmatic Arc" (Watters, 1974) has been proposed for this volcano-plutonic belt.

With the exception of the Nauchas area (de Waal, 1966) and a small portion of the area south of Windhoek (Hälbich, 1970), detailed information concerning the geology of the post-basement, pre-Damara-Nama formations along the Rehoboth Magmatic Arc, has not been published. Schalk (1973) has, however, presented a generalised map showing the distribution of the Late Precambrian Formations between the Nauchas area and the South West African-Botswana border, so that sufficient data are available to gain a general impression of the geology and to make very broad and tentative correlations with the sequences of the Sinclair and Helmeringhausen areas. Direct correlation between lithological types is probably not possible at this stage due to the typical local development and consequent extreme lateral variation of most of the units, e.g. the Guperas and Nagatis Formations.

It has been suggested by von Brunn (1967) that the *Gamsberg Granite* (de Kock, 1934; de Waal, 1966), which forms part of the Nauchas granite suite in the Nauchas area (de Waal, 1966) (Fig. 34), could be a correlate of the Rooikam Granite. The great similarities between the average Nubib granite composition and that of the Gamsberg granite (Table 23) support this suggestion. It must be realised, however, that such a correlation, based on major element chemical data only, is far from conclusive. It might also be speculated that the other members of the Nauchas granite suite (*Korabis* and *Koepel granite* and *Piksteel granodiorite*) and the *Weener Quartz Diorite* (de Waal, 1966) are also time equivalents of the Nubib granite and Barby Formation respectively. The probability of a correlation between the Gamsberg granite and the *Biesiespoort granite* (named after the farm Biesiespoort 175 on which the samples for radiometric work were collected), to the east of the Nauchas area, also supports a Gamsberg≡Rooikam≡Nubib granite correlation since the Biesiespoort granite has been dated at 1356±51 Ma (K/Ar on biotite - unpublished data forwarded to the Geological Survey Office, Windhoek by M. Dirac, 1967, then of the University of Cambridge), an age that corresponds very well with that of 1360±50 Ma obtained for the Rooikam granite in the vicinity of Helmeringhausen (Burger and Clifford *in* Burger and Coertze, 1973)

The *Alberta basic complex*, described by de Waal (1966), predates the Gamsberg granite in the Nauchas area and displays a very similar character and intrusive sequence to the main tholeiitic gabbro-norite body on Kumbis in the Sinclair area. The Alberta complex consists of a main 'Layered Sequence' which is intruded by a very coarse-grained 'gabbro pegmatoid' considered by

Fig.34. Distribution of the rock units comprising the Rehoboth Magmatic Arc.



2. *Gamsberg Granite*

1. *Gaub Valley Member* of the *Older Abbabis Formation* (Hälbich, 1970) - schists and conglomerates.

If a correlation of these rocks with the type Sinclair sequence is to be attempted (Table 22), it can be tentatively suggested that the Grauwater Formation (and the Doornpoort Formation ?) might be the time equivalent of the Guperas Formation, in view of the fact that the Grauwater is of post-Gamsberg granite age (Schalk, pers. comm.), and also considering the probable Gamsberg-Nubib granite correlation. The 1132±75 Ma age obtained for the Skumok porphyry (van Niekerk, 1968) may not support such a correlation but since this age can probably be regarded as a minimum age, any discrepancies in age between the Doornpoort and upper Guperas (> 1270 Ma) may not constitute a problem. The *Klein Aub Formation* may represent a local basin infilling similar to the Auborus Formation to the south.

Pre-Gamsberg units in this area are included in the *Older Abbabis Formation* of Hälbich (1970). The youngest of these is the *Gaub Valley Member*, equivalent to the Gaub Valley schists and conglomerates of de Waal (1966) and, therefore, possibly equivalent to the pre-Nubib granite formations of the type Sinclair Group. The oldest member of the Older Abbabis Formation is the Elim Member, which, according to Schalk (pers. comm.) constitutes 'basement' to Late-Precambrian formations in this area.

In western Botswana along strike with the Nauchas-Rehoboth-Dordabis sector of the Arc, the sandstones, volcaniclastic and felsic extrusives and intrusives of the *Kgwebe Formation*, dated at 950±37 Ma and 1020±50 Ma (Burger and Coertze, 1973), are unconformably overlain by the *Ghanzi Formation* consisting of felspathic sandstone, shale and conglomerate. Since the ages obtained from the Kgwebe porphyries are probably minimum ages (Rb-Sr method), the inclusion of these formations into the Rehoboth Magmatic Arc is considered reasonable. Their position in the succession relative to other parts of the Arc is not known, however, and the inference that the Kgwebe and Ghanzi Formations relate to post-Gamsberg/Nubib granite units in other parts of the Arc, (Table 22) must at this stage remain highly speculative.

It has been shown in this section that the Sinclair Group as exposed in the Sinclair and Helmeringhausen areas does not constitute an isolated occurrence of intense magmatic expression, but rather forms part of an extensive volcano-plutonic province. The form of this province is a curvilinear arc, named the Rehoboth Magmatic Arc.

5. EVOLUTIONARY MODEL AND GENERAL CONCLUSIONS

The general conclusions arising from this study and from a model for the evolution of the Sinclair Group and its correlates are presented below. The writer is well aware that several alternative models could be formulated, but the model presented here is considered the most plausible in the light of the data provided in the present study and other works dealing with the correlates of the Sinclair Group.

The evolution of the 'type' Sinclair Group, in the Sinclair and Helmeringhausen areas is first discussed in a restricted or local sense before a model is formulated which concerns the role of the Sinclair Group in a regional setting, i.e. within the Rehoboth Magmatic Arc.

5.1 LOCAL SEQUENCE OF EVENTS

The evolution of the Sinclair Group proceeded within three major cycles which are broadly delineated as follows:

3rd Cycle	Auborus Formation	Felspathic sandstone and conglomerate
	Rooiberg granite	
	Guperas Formation	Rhyolitic intrusives and extrusives
		Basic lava and intrusives
2nd Cycle	Guperas Formation	Lithic sandstone and conglomerate
	Nubib/Rooikam/Tumuab granite	
	Spes Bona syenite	
	Barby Formation	Basic lava and intrusives, rhyolite extrusives
1st Cycle	Kunjas Formation	Arkose, grit, shale
	Haremub/Kotzérus granites	
	Nagatis Formation	Rhyolite extrusives and minor basic lava, arkose, grit, shale
	(? Unexposed basic intrusives and extrusives ?)	

The basis for this cyclic subdivision is seen in a set of common factors or rather in a repeated general pattern. This pattern of development which is complete in all cycles with the possible exception of the first, can be regarded as follows:

Generation and emplacement of basic magma into and onto the continental crust, followed by the generation and emplacement of felsic magma into high crustal levels, followed by vertical tectonics and production of local fault-trough basins, followed by erosion of the basic and felsic rock-units and deposition of immature clastic debris into local basins, followed finally by relative stabilisation with respect to faulting and tilting.

Without exception, the type of sediment deposited toward the close of each cycle directly reflects the composition (i.e. felsic or basic) of the dominant igneous phase of the same cycle.

1ST CYCLE

The first cycle is incomplete when compared with the later cycles in that it lacks an initial basic magmatic phase. The Nagatis Formation apparently represents the initial phase of this cycle. It is, however, possible that the representatives of the basic phase were entirely intrusive and never reached the surface due to emplacement into a relatively cool crust and consequent rapid and early chilling. Alternatively, since exposure of the Nagatis Formation is relatively poor due to the emplacement and deposition of younger units of the Sinclair, considerable volumes of basic magma might have reached the surface but are now obliterated.

According to von Brunn (1967), both ignimbrite and rhyolite were extruded intermittently during the first period of Nagatis volcanism, probably from multiple vents concentrated along fissures. The rhyolite lavas flowed only over short distances before congealing into large pillow-like masses, whereas the ignimbrites, having far greater mobility, spread over most of the existing topography. Violent explosive volcanicity toward the end of the Nagatis times led to the formation of agglomerates. Relatively small amounts of basic lava were also extruded but these are nowhere comparable in quantity to the volumes brought to the surface during the second and third cycles. These small volumes of basic magma could not have successfully raised the regional geotherms to allow extensive crustal fusion to take place unless, as discussed above, they were accompanied by a large intrusive and predominantly pre-Nagatis felsic phase. During and immediately following the extrusion of basic lava, erosion and subsequent deposition resulted in beds of arkose, grit, shale and volcanic conglomerate being formed in local basins.

Renewed felsic volcanism then gave rise to initial ignimbritic eruptions becoming intimately associated with porphyritic rhyolite lava. Eruption during this phase was concentrated in the eastern part of the Helmeringhausen area and the extrusion of large volumes of felsic material resulted in the collapse of the main volcanic centre.

The plutonic felsic phase following the Nagatis resulted in the emplacement of the Haremub and Kotzërus granites. The generation of felsic magma for these plutons took place at a deep level in the crust where it was possible for amphibole to become involved in the partial fusion process. Fusion probably took place on a fairly limited scale and to varying degrees, resulting in small volumes of magma ranging in composition from granodiorite to granite. Contemporaneous rise and partial mixing of these magmas resulted in a hybrid magma which, on emplacement into a high crustal level by means of a stoping mechanism, incorporated xenoliths of basic country-rock without appreciable assimilation. The emplacement of this granite could well have occurred contemporaneously with the late stages of Nagatis volcanism, the granitic magmas representing the deepest crustal derivatives of partial melting, whereas the

Nagatis felsic magmas were derived from a slightly higher crustal level, but still within the lower crust, where the decomposition of biotite rather than amphibole was dominant.

Intrusion of the Haremub and Kotzērus granites did not take place on a large scale and it is possible that only two main masses were intruded, one a few kilometres south and southwest of Helmeringhausen and the other centred on the farm Haremub in the Sinclair area. The granite invaded rocks of the pre-Sinclair basement (Kumbis Formation) and Nagatis effusives.

Following the igneous activity of the first cycle, considerable erosion of the felsic volcanic piles of the exposed basement (mainly granitic) and of Haremub/Kotzērus granite took place, leading to deposition of the predominantly arkosic sediments of the Kunjas Formation. The Haremub granite pluton must have formed a prominent topographic high at this time, onto which only a thin succession of Kunjas sedimentary beds was deposited. Von Brunn (1967) suggested that deposition of the Kunjas took place in northwest-trending fault-troughs which developed as a result of crustal subsidence following on the period of extensive magmatic activity. Outcrops of the Kunjas Formation are, however, too limited to allow for accurate interpretation of the regional pre- or syndepositional tectonics. The nature and extent of the Kunjas beds indicate moderate relief in closely adjacent source areas and large depositional basins. The frequent occurrence of coarse beds of conglomerate and grit at various levels in the sequence are indicative of frequent periods of rejuvenation of relief. The numerous grit beds in the vicinity of the Haremub granite pluton could be due to the active rise and erosion of this granite mass during deposition of the Kunjas.

The increasing frequency of orthoquartzite beds toward the top of the Kunjas succession is indicative of the filling-up of the depository with the deposition of mature clastic debris and quiet tectonic conditions.

2ND CYCLE

Renewed igneous activity beginning in Barby times marks the start of the second cycle. The extrusion of minor quantities of highly gas-charged felsic magma from a few widely-spaced vents resulted in the deposition of thin, but widespread ash-flow tuff beds. The highly vitric pyroclastic material had a very high degree of mobility, and flooded the entire Sinclair-Helmeringhausen area. The extrusion of hot ash was followed in some areas, probably adjacent to the major vents, by the explosive eruption and deposition of cooler ash into local shallow-water basins, by airfall processes, thus producing the fine laminations, occasional cross-bedding and crude graded bedding, so typical of the stratified tuffites of the Basal volcaniclastic member of the Barby Formation. The incorporation of foreign clastic grains indicates a mild influx of normal sedimentary clastic material. Periodic congealing of magma in the source vents resulted in a build-up of pressure with eventual release by violent explosive activity. This is shown by coarse angular felsic rock fragments which form beds of volcanic conglomerate in areas adjacent to vents.

The extrusion of very large quantities of basic magma followed the deposition of these volumetrically insignificant felsic volcanic deposits. Mainly calc-alkaline lava was extruded in the southern Sinclair and Helmeringhausen areas, whilst to the north and northwest, tholeiitic lava types were the dominant products of basic volcanism.

It was concluded in Section 3.4 that the basic magmas are the result of partial melting along the upper surface of a southeasterly-dipping subducting slab of oceanic crust, and also of partial melting in the upper mantle overlying the slab. The calc-alkaline magmas were most probably derived by direct partial melting of eclogite (transformed oceanic crust), within the subduction zone, in a relatively 'dry' environment at depths of about 150-175 km. The prominence of the potassium-rich mica, phlogopite, as the stable hydrous phase at these depths and its incorporation into the low-melting fraction of the eclogite, on entering a suitable pressure-temperature regime, may have led to the production of the characteristically potassium-rich melts emplaced during this cycle.

The general increase in potassium content of the calc-alkaline lavas in a roughly southwesterly direction is regarded as indicative of increasing depth of magma generation, and hence increasing depth to the subduction zone in this direction.

This 'polarity' is further emphasized by the presence of tholeiitic basic lavas occurring to the north and northwest of the calc-alkaline occurrences. These rocks are considered to have originated in the upper mantle above a relatively shallow part of the subduction zone, where the release of H_2O and possibly minor amounts of low-melting products from the sinking amphibolite/eclogite (transformed oceanic basalt) slab could have produced unstable conditions in the overlying mantle, favouring the production of tholeiitic liquids by partial melting of peridotite.

The calc-alkaline magmas underwent varying degrees of fractionation *en route* to the surface and, consequently, were extruded in a partly crystalline state, resulting in strongly porphyritic rocks. Eruption took place under subaerial conditions and was generally quiet and non-explosive, although many of the flows emplaced toward the end of this basic phase are highly amygdaloidal and must have flowed in a nearly frothy state. Numerous volcanic centres must have been present in the south of the Sinclair area and these were probably of the central type.

In the southern portion of the Sinclair area the extrusion of calc-alkaline lava was interrupted by the outflow of rhyolitic lava which, in the vicinity of Aruab, was preceded by the explosive brecciation of both the already-deposited basic lava and of congealed rhyolitic vent-filling material. Thus angular volcanic fragments were deposited over a restricted area to form beds of agglomerate. The compositional characteristics of this felsic lava indicates an origin by partial fusion of deep crustal material. Renewed basic calc-alkaline volcanism following this short interlude of felsic magmatic activity was broken by brief periods of quiescence, during which local erosion of the lava flows took place with subsequent deposition of the debris in very small local basins.

On cessation of the basic volcanic activity of this phase, at least in the south of the Sinclair area, similar local erosion of the lavas and deposition of volcanic debris took place once again, but in this case appreciable amounts of normal clastic material were brought in from more distant source areas leading to the incorporation of well-rounded and mature grains and pebbles of quartz into the sediments.

In the northern part of the Sinclair area, the extrusion of tholeiitic lava was broken at some unknown stage by a period of diminished igneous activity during which influx of clastic debris occurred, depositing beds of arkose and quartzite. The prominence of these beds diminishes in a southeasterly and southerly direction, possibly indicating a source area to the north and/or northwest. Numerous thin basic lava flows intercalated with these sedimentary beds provide evidence that basic volcanic activity had merely lessened in intensity but had not ceased altogether.

The deposition of sedimentary beds was followed by a period of extensive rhyolitic volcanism during which the extrusion of basic lava continued at a diminished rate. Both ash-flow tuff and rhyolite lava were extruded initially, the more mobile ash-flows spreading over an area as wide as that covered by the earlier sedimentation, i.e. as far southeast as Duwisib.

The rhyolite lava was extruded as thick viscous units flowing over short distances and developing a tortuously flow-banded character. Intermittent violent explosive volcanicity, possibly accompanied by rapid erosion of the locally rugged volcanic landscape, provided volcanic debris of both basic and felsic nature and with an extremely wide range in grain or clast size. Deposition in local basins took place with a minimum amount of transport and/or reworking allowing rounding of only the largest clasts to take place.

Alternating extrusions of basic and felsic flows and the deposition of volcanic conglomerates were gradually replaced by the emplacement of thick flows of rhyolite which in turn, eventually gave way to renewed and extensive tholeiitic basaltic volcanic activity.

Toward the end of this basic volcanic phase of the Barby it became increasingly more difficult for the tholeiitic magmas to reach the surface and much of the material accumulated in both large and small chambers. This occurred at levels generally below the lava pile where, as the magma cooled slowly, crystal settling took place. These chambers were apparently fed by deeper and larger reservoirs in which similar crystal settling and fractionation processes may have taken place, but on a larger scale. Repeated injections from these main reservoirs, tapped at different levels, resulted in compositional variations between successive emplacements to higher crustal levels.

At the close of the Barby extrusive phase, vertical block movement resulted in tilting of the beds to moderate angles. The continued rise of the Haremub granite pluton at this time produced a dome structure and the rise of the basement complex in the Vergenoeg area produced a plunging synclinal structure in the overlying Kunjas and Barby strata. At the same time, a northwest-trending fracture system developed in the Spes Bona-Aandster area, marking the rise of the Spes Bona syenite body.

The syenite magma was probably derived by low degrees of partial melting of eclogite within a subduction zone, as also postulated for the calc-alkaline lavas of the Barby. The syenite represents the final 'basic' magmatic products of this cycle, which, on rising toward high crustal levels, underwent considerable fractionation. It is likely that the magma fed into a large reservoir at fairly high levels in the crust where it crystallised almost completely. The syenite mass continued to rise *en bloc*, pushing up the overlying crustal material. Potassic calc-alkaline magma, still present at depth following the main extrusive phase, rose up along the fractures developing in zones marginal to the rising syenite and overlying crust to produce the monzonite and diorite bodies that occur around the syenite body. The progressive increase in potassium content, with time, of the calc-alkaline magmas of the Barby, culminating in the extremely high potash content of the calc-alkali syenite, indicates a general progressive decrease in the degree of partial melting of 'dry' eclogite in the underlying subduction zone.

At approximately the same time as the development of the northwest-trending zone of weakness marking the site of syenite emplacement, deep crustal fractures developed in the northern part of the Sinclair area. These are represented by prominent northwesterly-trending normal faults, northeastward-tilted blocks, and a prominent zone of mylonitization, possibly marking the marginal site of a rising block of pre-Sinclair basement to the north and northeast. This mylonite zone (the Nam Shear Belt) is a site of fundamental crustal weakness and repeated shear-deformation, and it is most probably related to the late-stage deformation in the Namaqua Mobile Belt to the south. It allowed the tapping of considerable volumes of felsic magma generated in the lower crust by large-scale partial fusion processes. The partial fusion took place in response to a rise in geotherms resulting from the voluminous injection of Barby basic magma into the crust.

The emplacement of this felsic magma to form the Nubib granites was, therefore, facilitated by movement in the mylonite belt, thereby concentrating the plutons in a prominent northwest-trending zone. To the south small amounts of felsic magma rose up along zones of weakness marginal to the Spes Bona syenite body. In the Helmeringhausen area intrusion of granitic magma during this phase also took place along large-scale zones of weakness trending roughly northwest (von Brunn, 1967).

Movement along the Nam Shear Belt continued throughout most of the intrusion of the Nubib granite, reaching a peak during its emplacement, but diminishing rapidly toward the end of this felsic phase. Cataclastically deformed Nubib granite, Barby basic lava, and possibly basement rock in the extreme north of the area have, therefore, resulted from this shearing.

Ascent of the granite magma was accomplished by a stoping mechanism, with the magma incorporating xenoliths of basic country-rock without much reaction taking place. Xenoliths of felsic country-rock in the Nubib mountains were remelted and incorporated into the invading felsic magma.

The intrusion of the Nubib (≡ Rooikam ≡ Tumuab) granite was followed by a period of normal faulting during which a deep north-south trending trough developed along the eastern part of the Sinclair area. There is thus a marked change in the dominant structural trend at this time, probably reflecting a tectonic consolidation of the Namaqua Mobile Belt to the south and a consequent reduction in the influence that the final movements of the Namaqua Episode could exert on the stable crustal block underlying the Sinclair and Helmeringhausen areas.

The trough formed a depositional basin which received large amounts of immature clastic debris from local provenance areas, to form the lower parts of the Guperas Formation. Coarse conglomerate and lithic sandstone were deposited and the abundance of basic volcanic fragments and pebbles in these rocks reflects the erosion of adjacent areas of exposed Barby basic lavas, and rapid deposition of the resultant debris. In areas where outcrops of granitic rock were prominent, quartzo-felspathic grains dominate in the sediments. The fairly good sorting of the sandstone constituents is further indication of a high energy sediment transport system, probably reflecting fluviatile conditions.

Deposition of sedimentary detritus during early Guperas times was widespread and not restricted to the main Guperas trough and scattered local basins occurred as far west as the Awasib Mountains. The prominent north-south trough, extending from Aruab in the south to Rooiberg (at least) in the north, eventually became the main depository of the Guperas, receiving large quantities of clastic material derived from adjacent areas of extreme positive relief.

3RD CYCLE

The onset of basic igneous activity during the Guperas phase marks the start of the third and last cycle in the development of the Sinclair Group. As was the case between the first and second cycles, the transition from the second to third cycle is not marked by an unconformity or hiatus but rather by the renewal of igneous activity. Whereas the Barby volcanism was extensive and rapid in its build-up, thus preventing the further development of the sedimentary Kunjas Formation, the basic volcanicity of the Guperas that initiated the third cycle, tended to be either very local or sporadic, resulting in flows intercalated with beds of lithic sandstone and conglomerate. The lavas were of both calc-alkaline and tholeiitic types and are considered to be the result of similar generative processes as proposed for the production of the Barby basic lavas. It is likely that the lava flows were fed by magma erupting along north-south striking fissures which, on solidification, formed a multitude of basic dykes.

The injection of basic magma into the sialic crust once again raised the geotherms to a level where partial melting could take place, thus producing great volumes of felsic magma. The geothermal gradient was probably still at a fairly high level from the previous cycle and, consequently, only small volumes of basic magma were necessary to raise the temperature sufficiently for partial fusion to take place.

Felsic magma rose to the surface *via* major tension features and hence was extruded within, or in areas adjacent to, the main sedimentary trough of the Guperas. Eruption took place from a series of vents which were frequently interconnected. The first felsic magma was rich in volatiles and on reaching the low-pressure near-surface regime of the crust it vesiculated and was erupted in three major pulses as thick ash-flow tuff sheets from a series of vents in the Blutpütz Ost area. The ash was in a highly mobile state and flowed almost exclusively southward into what was then the most rapidly subsiding part of the Guperas trough. Minor ash-flows were also erupted in the Ganaams area from an unknown source. At about this time, highly gas-charged but violently explosive eruptions took place in the Rooiberg area, depositing angular and unsorted felsic volcanic debris over this portion of the trough.

Subsequent extrusions of felsic magma took the form of highly viscous rhyolite flows that spread out laterally, in a very limited manner, from the source vents. These lavas were emplaced as large pillow-like masses, building up thick piles in the northern and central parts and on the south-western flank areas of the main Guperas trough.

Basic volcanicity continued in the Ganaams area during the extrusion of rhyolite, filling up a local 'basin' which developed between steep-sided piles of rhyolite. South of the main Blutpütz West rhyolite pile, on Aruab and Guperas, the deposition of immature clastic debris from the erosion of nearby exposed pre-Guperas units continued whilst the extrusion of rhyolite was taking place to the north and northwest. The high mobility of the ash-flows, when compared with the overlying rhyolites resulted in the former being intercalated within the otherwise purely sedimentary succession occurring to the south of the rhyolite lava pile on Blutpütz West.

Mild basic igneous activity continued during emplacement of the rhyolites as evidenced by basic flows intercalated within the sedimentary succession to the south of the Blutpütz West rhyolite field, and by basic intrusions within the sedimentary beds underlying the felsic lavas.

The strong east-west crustal tension prevailing at this time also facilitated the intrusion of felsic magma in very dense dyke swarms. However these dykes probably did not provide feeder channels to the rhyolite lavas.

As the felsic lava piles thickened, it became difficult for the magma to reach the surface and some material remained in small resevoirs within or below the rhyolites to crystallise as plutons of Rooiberg granite.

The emplacement of large volumes of felsic magma during Guperas times resulted in the collapse of the three main volcanic centres. This subsidence took place along arcuate faults in the case of the Kronenberg-Auramberg and Rooiberg fields, whereas subsidence of the Blutpütz West lava pile took place mainly along north and north-northwest trending normal faults within or bounding the Guperas trough.

Further trough-faulting along the prominent post-Barby north-south trend followed and caused a rejuvenation of relief. Subsequent sedimentation took place within locally developed basins elongated in a generally north-south direction, building up the sequence of the Auborus Formation.

Clastic debris was once again derived mainly from adjacent exposed old members of the Sinclair Group, and the predominance of felsic igneous activity in post-Barby times contributed to the arkosic nature of the Auborus beds. The deposition of very fine-grained material took place as the basins filled up and the adjacent topography became more mature.

The complete absence of igneous activity during or following the deposition of the Auborus sequence implies that this Formation represents the closing phase in the evolution of the Sinclair Group. The inclusion of the Auborus Formation into the Sinclair Group is a logical consequence since this unit completes the final cycle and represents the deposition, into entirely local basins, of the planation products of a mountainous volcanic and granite terrain.

Normal faulting along the north-south trend continued during post-Auborus times, and preserved the Auborus sedimentary strata in pre-Nama graben structures.

5.2 REGIONAL EVOLUTION

The theory of sea-floor spreading and plate tectonics has been successfully applied in explaining and unifying modern global tectonic features, and has recently been applied to older tectonic and magmatic events (Burke and Dewey, 1970; White *et al.*, 1971; Thorpe, 1972; Badman, 1973; Dewey and Burke; 1973). It seems reasonable, therefore, to examine the evolution of the Sinclair Group and its correlates in the light of the new global tectonics although the validity of applying plate tectonic models to Precambrian events has been questioned (e.g. Shackleton, 1969, 1973) and many writers suggest that it should be applied with reservation (e.g. Clifford, 1972; Windley, 1973; Rutland, 1973). In the case of the Sinclair Group and its correlates, however, there is some justification in proposing a plate tectonics evolutionary model.

Firstly, there seems little doubt that large rigid continental plates or cratons did exist during the Late Precambrian (Clifford, 1972). The Kalahari Craton, for example, along which the Sinclair Group is situated, has remained essentially undeformed since Sinclair times (< ∿ 1400 Ma). Such a stable and consolidated nature infers that the craton had already attained its present thickness in pre-Sinclair times.

Secondly, the presence of highly characteristic high-potassium calc-alkaline basic rocks or shoshonites, occurring prominantly within the Sinclair Group does, by analogy with modern examples, strongly suggest development of the Group over an ancient active zone of subduction of oceanic crust. It would appear that the partial melting of 'dry' eclogite within a subduction or 'Benioff' zone offers the most satisfactory explanation for the genesis of the shoshonite magmas.

Thirdly, the distribution of the Sinclair Group and its correlates, as discussed in Section 4 and illustrated in Figure 34, describes an extensive curvilinear belt, called the Rehoboth Magmatic Arc. Throughout its length, the magmatic arc has as its base the stable continental plate of the Kalahari Craton. It is further apparent that this magmatic arc roughly separates the undeformed Late Precambrian and Cambrian (?) deposits (Nama Group) of the stable platform area from the highly deformed Late Precambrian deposits of the Damaran orogenic belt, suggesting that the arc occupies a marginal cratonic position.

In terms of plate tectonics, curvilinear volcano-plutonic arcs, such as the Rehoboth Magmatic Arc, develop along the margins of stable continental plates when an oceanic plate is being actively consumed (subducted) beneath its leading edge (James, 1971; Dickinson, 1972).

It is conceivable, therefore, that the Rehoboth Magmatic Arc developed along the western and northwestern margin of the Kalahari Craton as a result of the active consumption of ancient oceanic crust, lying to the northwest and west of the craton, along an approximately southeasterly dipping subduction zone.

Many of the other characteristic properties of the Rehoboth Magmatic Arc, or parts thereof, are closely analagous to typical features of young continental margin volcano-plutonic arcs such as the Cordilleran or Andean-type belts. These characteristics are discussed below:

(i) The polarity in the distribution of lava-types in the Sinclair area, i.e. a progressive change from tholeiitic through calc-alkaline into shoshonitic basaltic-andesite rock-types in a generally southeasterly direction, is consistent with the compositional trends observed in younger volcano-plutonic belts (Dickinson, 1970; Jakes and White, 1972; Lefèvre, 1973) and supports the proposal for a southeasterly-dipping subduction zone beneath the margin of the Kalahari Plate during the evolution of the Sinclair Group.

(ii) Voluminous 'granite' batholiths characterise continental-margin magmatic belts (Hamilton, 1969; Dickinson, 1970) and these range from diorites, monzonites and tonalites through granodiorites to true granites.

It was shown earlier that diorites and monzonites are present in the Sinclair area as the intrusive, co-magmatic, representatives of an extensive period of volcanism (Section 3.4), and true granitic rocks are also exceptionally voluminous. The intermediate varieties, such as granodiorite, are not well represented in the Sinclair area, but occur abundantly in other parts of the Rehoboth Magmatic Arc. De Waal (1966) has described an extensive suite of plutonic rock types ranging from quartz diorite through granodiorite into granite (see Section 4.2). Members of this suite have also been reported from the adjacent area to the east (Hälbich, 1970).

(iii) Rhyolitic extrusives, including extensive ash-flow deposits, invariably related to the granites, are common in active continental margin environments (James, 1971; Hyndman, 1972).

In the Sinclair area, a co-genetic relationship between the rhyolites and granites seems almost certain and both rhyolites and ash-flow tuffs were emplaced at various stages throughout the evolution of the Sinclair Group. Rhyolitic extrusives also appear ubiquitously throughout the entire Rehoboth Magmatic Arc and the apparent lack of ash-flow tuff in sequences other than the Sinclair succession may be due to the lack of detailed petrographic investigations and, therefore, a case of non-recognition rather than absence.

(iv) A characteristic 'pulsing' of igneous activity in magmatic arcs has been noted by Dickinson (1972).

An episodic or cyclic development for the Sinclair Group has already been discussed earlier in this Section and from the limited data available it is apparent that an episodic character is present in all parts of the Rehoboth Magmatic Arc (Section 4.2), but does not appear to be as well developed as in the Sinclair area.

(v) Crustal dilations, resulting in vertical tectonics and extensive block-faulting paralleling the trend of magmatic arcs is dominant in modern examples (James, 1971).

Many large-scale normal faults or tension features are present or can be inferred paralleling the general trend of the Rehoboth Magmatic Arc. The dominance of approximately north-south-striking tension features in post-Barby times in the Sinclair area has been emphasized earlier in this Section. In pre-Guperas stages of evolution the tectonic influence of the Namaqua Mobile Belt to the south was dominant.

In the Nauchas-Gobabis sector of the Rehoboth Magmatic Arc most tensional features have been obliterated by the younger Damaran tectonism, but from the strike of pre-Damara dyke swarms and the occurrence of thick local sequences it can be inferred that tension and consequent vertical block-faulting also played an important role in the development of this sector of the arc.

(vi) The presence of abundant pyroclastics and sedimentary suites composed of first-cycle derivatives of older volcano-plutonic units, deposited in parallel subsiding troughs, is particularly characteristic of the continental margin arc tectonic environment under consideration (Dickinson, 1970).

Pyroclastic beds directly associated with volcanicity occur ubiquitously throughout the Rehoboth Magmatic Arc and, as discussed earlier in this Section, the sedimentary sequences in the Sinclair area are built up predominantly of material derived from local denudation and rapid deposition in nearby fault-troughs. In the sector of the Rehoboth Magmatic Arc northeast of the Nauchas area, coarse immature and locally-derived clastic material was apparently available to a lesser extent than in the Sinclair area since mature fine-grained and quartzitic beds are frequently encountered there (Schalk, 1973). A possible explanation for a difference in the mode of sedimentation between this sector of the Arc and the Helmeringhausen-Nauchas sector can be found in the greater intensity of igneous activity, extrusive in particular, that prevailed in the latter. The enormous outpourings of lava, intrusion of plutonic masses, and the consequent crustal readjustment produced an igneous terrain of almost constant extreme topography. Northeast of Nauchas however, less intense volcanic activity was accompanied by intense plutonism, and this led to a topography of only moderate relief, allowing sedimentary detritus to be brought in from more distant source areas (possibly basement terrains). Although trough-faulting probably did take place, the consequent erosion of older exposed sequences of the Rehoboth Magmatic Arc, which would have contained a high proportion of sedimentary beds, probably added multi-cycle clastic material to the basins.

In view of the evidence discussed above, it would seem therefore, that a good case can be made for the application of active continental margin plate tectonic processes to the evolution of the Rehoboth Magmatic Arc.

It is suggested, as a consequence, that the presence of the Rehoboth Magmatic Arc can be used to define a paleo-continental margin, i.e. the margin of a certain sector of the Kalahari Craton during the period $\sim$1400 Ma to $\sim$1250 Ma. According to Dickinson (1973) the distance between the continental margin and the axis of a volcano-plutonic 'orogen', developed above a subduction zone at such a margin, is in the order of 225 ± 50 km. To the west of the Sinclair area which is regarded as approximating the axis of the Rehoboth Magmatic Arc in that sector, younger Damara rocks are encountered 180 km west of the study area, along the present coast of South West Africa. It is probable therefore, that the appearance of these younger rocks delineates fairly accurately the position of the ancient continental slope regions of the Kalahari Craton.

In the Nauchas-Gobabis area, however, the picture is somewhat different,

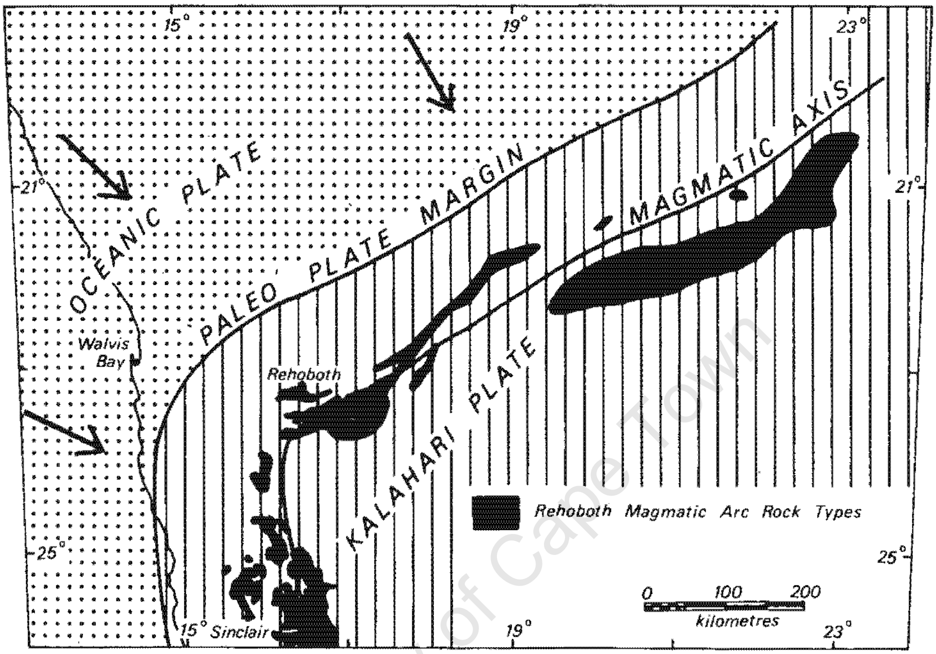


Fig. 35. Tentative reconstruction of the geotectonic setting that may have prevailed during the evolution of the Rehoboth Magmatic Arc, superimposed on a sketch map showing the present-day distribution of the Arc units. Arrows indicate the approximate directions of plate movements.

since the younger Damara units have been thrust southeastward over the Rehoboth Magmatic Arc units, therefore making any estimate of the original position of the continental margin extremely speculative. This becomes even more apparent when it is considered that the position of the axis of the volcano-plutonic belt in this area is not known, although it can probably be regarded as coinciding with the prominent occurrences of granitic intrusions passing through the Nauchas-Rehoboth area (Fig. 35).

The reconstruction of the position of the Kalahari Plate margin during the evolution of the Rehoboth Magmatic Arc, as shown in Figure 35, must, therefore, be regarded as highly speculative.

The model proposed above for the evolution of the Sinclair Group and its correlates may have far-reaching implications for the subsequent development of the Damaran Belt, since it implies that this orogen is not entirely of ensialic development.

It is suggested, therefore, that the possible existence of oceanic crust, during post-Arc or post-Sinclair times, in the area now underlain by the Damaran Belt, should be considered in future research on the evolution of this geosyncline.

APPENDIX

1 ANALYTICAL TECHNIQUES

1.1 SAMPLING

Sample sites were selected to satisfy the following requisites as far as possible:

(i) Samples representative of the body, succession or particular part of the succession.

(ii) Wide geographical spread.

(iii) The size of the sample taken was in direct proportion to the average grain size (i.e. according to Smales and Wager, 1960) although, as in the case of felsic lavas, inhomogeneities due to flow-banding were also taken into account.

1.2 SAMPLE PREPARATION

After an initial breaking up with a large hammer the samples were washed and reduced to blocks of 3 cm and less by means of a splitter with carbon steel 'knife edges'.

The blocks were then individually scrubbed under running water with a nylon bristle brush, rinsed in distilled water and dried in an oven at 120°C.

The material was then passed through a jaw-crusher with carbon steel jaws (no Ni, Cr and Co contamination) which reduced it to chips of 0,5 cm and less. At this stage the material was quartered, the larger fraction stored and the smaller fraction crushed to -120 mesh in a Siebtechnik mechanical agate mill.

From this stock sample, 6 g were ground for one hour in an automatic agate mortar until >95 per cent of the sample was -300 mesh.

The procedure as outlined above ensures that contamination is kept to a minimum, the only contaminating elements being Fe and Si.

Of the finely-ground material, 4 g were then moistened with a mowiol solution, to increase the cohesive properties of the oversaturated rocks, and briquetted at a pressure of 15 tons according to the method of Baird (1961).

Approximately 1,5 g of the remaining powder was heated at 120°C for 8 hours in preheated vitreosil crucibles and then at 1050°C for a further 8 hours. The weight losses at these temperatures determined hygroscopic water (H_2O^-) and loss on ignition (l.o.i.) respectively.

Two 0,28 g portions of the ashed sample were used to make fusion discs according to the method of Norrish and Chappell (1967) for major element analysis. The flux used was Johnston-Matthey Spectorflux 105(Li_2CO_3, $Li_2B_4O_7$ and La).

1.3 ELEMENT AND OXIDE ANALYSES

Major elements and all trace elements except barium were determined by X-ray fluorescence on a Philips PW 1220 semi-automatic spectrometer with a 2 kW generator. Barium was determined by X-ray fluorescence on a Philips PW 1540 spectrometer with a 1 kW generator.

All major elements were determined in duplicate in two consecutive runs, and trace elements were determined once in each of two consecutive runs, in both cases in a fixed sequence in order to check reproduceability. Sets of standards were determined at least 4-6 times during any particular run.

In order to obtain a quantitative check on precision in sample preparation and instrument operation, eleven fusion discs of sample BW 1598 were prepared and analysed. The mean, standard deviation and coefficient of variation for each major element (except for sodium) are presented in Table 24. The results are very satisfactory, and even MgO, which normally has notably poor counting statistics, has a coefficient of variation of less than 2,3 per cent.

Operational conditions for the analyses are presented in Tables 25 and 26.

Ferrous iron was determined titrimetrically by a method modified after Shapiro and Brannock (1956). 0,500 g of powdered rock sample (-120 mesh) was digested in HF and H_2SO_4, this solution being titrated against a standard solution of $K_2Cr_2O_7$ with diphenylamine sulfonic acid as an indicator.

Table 24. Mean, standard deviation and coefficient of variation of major elements in eleven analyses of sample BW 1598 (syenite)

Oxide	Mean of 11 analyses	Standard deviation	Coefficient of variation (%)
SiO_2	58,403	0,276	0,472
TiO_2	0,672	0,004	0,569
Al_2O_3	18,728	0,089	0,475
Fe_2O_3	3,663	0,033	0,903
MnO	0,058	0,001	1,960
MgO	1,358	0,031	2,294
CaO	3,144	0,021	0,654
K_2O	7,981	0,022	0,274
P_2O_5	0,447	0,007	1,588

Table 25. Operational conditions for major element analyses

ELEMENT	Si	Ti	Al	Fe	Cr	Mn	Mg	Ca	Na	K	P
Spectrometer	PW 1220	PW 1220	PW 1220	PW 1220	PW 1220	PW 1220	PW 1220	PW 1220	PW 1220	PW 1220	PW 1220
Target Tube	Cr	Cr	Cr	Au	Au	Au	Cr	Cr	Cr	Cr	Cr
Generator kV	50	50	50	55	55	55	50	50	60	50	50
Generator mA	32	20	32	36	36	36	32	20	32	20	32
Collimator	Coarse	Coarse	Coarse	Fine	Fine	Fine	Coarse	Fine	Coarse	Coarse	Coarse
Crystal	Pet	LiF (200)	Pet	LiF (220)	LiF (220)	LiF (220)	ADP	LiF (200)	RAP	LiF (200)	Ge
Counter	Flow	Flow	Flow	Flow	Flow	Flow	Flow	Flow	Flow	Flow	Flow
Counting Time (sec)	20	10	60	10	200	100	240	10	200	20	120
2θ	109,28	86,15	145,03	85,82	107,23	95,30	137,82	113,21	54,23	136,70	140,90
Sample	F.D.	F.D.	F.D.	F.D.	F.D.	F.D.	F.D.	F.D.	W.R.P.	F.D.	F.D.

F.D. = Fusion Disc W.R.P. = Whole-rock pellet

Table 26. Operational conditions for trace element analyses

ELEMENT		Nb	Zr	Y	Sr	Rb	Ba	S
Spectrometer		PW 1220	PW 1220	PW 1220	PW 1220	PW 1220	PW 1540	PW 1220
Target Tube		W	W	W	W	W	Cr	Cr
Generator	kV	70	70	70	70	70	50	50
	mA	28	28	28	28	28	20	32
Collimator		Fine	Fine	Fine	Fine	Fine	Fine	Coarse
Crystal		LiF (220)	LiF (220)	LiF (220)	LiF (220)	LiF (220)	LiF (220)	Ge
Counter		Scint	Scint	Scint	Scint	Scint	Flow	Flow
Counting Time (sec)		240	120	120	120	120	320	240
2Θ		30,44	32,11	33,90	35,865	38,00	124,24	110,60
Sample		W.R.P.	W.R.P.	W.R.P.	W.R.P.	W.R.P.	W.R.P.	W.R.P.

W.R.P. = Whole-rock pellet

2 ANALYTICAL DATA FOR AGE DETERMINATIONS

2.1 POTASSIUM-ARGON AGES

Analytical data for three K-Ar ages, determined by D.C. Rex of Leeds University on muscovite from secretion pegmatites of the Nam Shear Belt, are presented in Table 27 below.

Table 27. Analytical data for K-Ar age determinations

Sample	Number	K%	Radiogenic ^{40}Ar (s.c.c./g.x10^{-4})	Radiogenic ^{40}Ar %	Age Ma
Muscovite	BW 910	8,70	5,355	98,9	1127±45
Muscovite	BW 1155	8,72	5,414	98,9	1129±45
Muscovite	BW 1156	8,60	5,369	97,7	1143±46

$$\lambda_\beta = 4,72 \times 10^{-10}yr^{-1}$$
$$\lambda_e = 0,584 \times 10^{-10}y^{-1}$$
$$^{40}K/K = 0,0119 \text{ at.\%}$$

2.2 URANIUM-LEAD AGES

Analytical data for a single U-Pb age, determined by A.J. Burger of the National Physical Research Laboratory of the South African Council for Scientific and Industrial Research are presented in Table 28.

Table 28. Analytical data for U-Pb (zircon) age determination of sample BW 1592

Concentration (wt. %)		Pb Isotope Ratios		
Pb	U	Pb^{206}/Pb^{204}	Pb^{207}/Pb^{204}	Pb^{208}/Pb^{204}
0,009956	0,07051	924,21	90,19	144,27

Calculated ages (Ma)			N^{207}/N^{235}	N^{206}/N^{238}
Pb^{207}/Pb^{206}	Pb^{207}/U^{235}	Pb^{206}/U^{238}		
1270	922	786	1,4524	0,1284

3 SAMPLE LOCALITIES

The localities of all analysed samples are given (to the nearest degree) in Table 29, and also on the generalised geological map of Figure 36.

Table 29. Localities of analysed samples

Sample No.	Rock-type, member, Formation	Longitude (to nearest 1')	Latitude (to nearest 1')	Farm
578	Rhyolite intrusive, Guperas Formation	16°23'	25°34'	Klein Haremub
1281	" " " "	16°22'	25°28'	Kronenhof Wes
1470	" " " "	16°34'	25°33'	Blutpütz West
1502	" " " "	16°35'	25°34'	Blutpütz West
1504	" " " "	16°34'	25°34'	Blutpütz West
1552	" " " "	16°36'	25°37'	Aruab
1548	Quartz-porphyry dyke, Guperas Formation	16°37'	25°37'	Aruab
1404	" " " " "	16°37'	25°24'	Naudaus
935	" " " " "	16°16'	25°06'	Nubib
1365	" " " " "	16°38'	25°28'	Persia
1497	" " " " "	16°35'	25°32'	Blutpütz West
466	" " " " "	16°16'	25°29'	Spes Bona
1403	Rhyolite lava, Guperas Formation	16°38'	25°24'	Naudaus
1333	" " " "	16°37'	25°13'	Rooiberg
581	" " " "	16°24'	25°34'	Ginas
1501	" " " "	16°37'	25°33'	Blutpütz West
1284	" " " "	16°22'	25°28'	Kronenhof Wes
1435	" " " "	16°36'	25°22'	Naudaus
1293	" " " "	16°37'	25°19'	Duwisib
1478	" " " "	16°33'	25°30'	Blutpütz West
1589	Nubib granite	16°44'	25°26'	Osis 73
1450	" "	16°37'	25°26'	Naudaus
1595	" "	16°29'	25°16'	Duwisib
1596	" "	16°03'	25°04'	Nubib West
1340	" "	16°45'	25°25'	Osis 74
1180	" "	16°22'	25°15'	Betta
1175	" "	16°24'	25°13'	Gorab
682	" "	16°11'	25°08'	Vrede
1448	" "	16°33'	25°25'	Naudaus
899	Nubib mylonitic gneiss	16°16'	25°01'	Steinfeld
1592	Rooiberg granite	16°40'	25°18'	Rooiberg

Table 29. Continued

Sample No.	Rock-type, member, Formation	Longitude (to nearest 1')	Latitude (to nearest 1')	Farm
1624	Haremub granite	$16^\circ21'$	$25^\circ37'$	Klein Haremub
1597	Rhyolite lava, Rhyolite member, Barby Formation	$16^\circ06'$	$25^\circ05'$	Nubib West
773	" " " " " "	$16^\circ08'$	$25^\circ02'$	Nubib West
756	" " " " " "	$16^\circ09'$	$25^\circ02'$	Nubib West
235	Rhyolite lava, High-Ca rhyolite member, Barby Formation	$16^\circ14'$	$25^\circ36'$	Vergenoeg
1228	" " " " " " "	$16^\circ11'$	$25^\circ33'$	Sonop
1566	" " " " " " "	$16^\circ34'$	$25^\circ40'$	Aruab
334	Norite, Barby Formation	$16^\circ01'$	$25^\circ41'$	State Land
293	Gabbro, Barby Formation	$16^\circ02'$	$25^\circ40'$	Kumbis
323	Norite, Barby Formation	$16^\circ01'$	$25^\circ41'$	State Land
312	Anorthosite, Barby Formation	$16^\circ01'$	$25^\circ40'$	State Land
811	Tholeiitic basalt, Basalt member, Barby Formation	$16^\circ05'$	$25^\circ03'$	Nubib West
1351	" " " " " "	$16^\circ48'$	$25^\circ26'$	Osis 73
1415	" " " " " "	$16^\circ34'$	$25^\circ27'$	Naudaus
1606	Diorite, Barby Formation	$16^\circ00'$	$25^\circ22'$	Aandster
1613	Olivine monzonite, Barby Formation	$16^\circ00'$	$25^\circ31'$	166
1256	" " " "	$16^\circ05'$	$25^\circ29'$	Saffier
1565	Monzonite, Barby Formation	$16^\circ34'$	$25^\circ40'$	Aruab
1598	Spes Bona syenite	$16^\circ16'$	$25^\circ29'$	Spes Bona
1599	" " "	$16^\circ05'$	$25^\circ22'$	Aandster
1617	" " "	$16^\circ06'$	$25^\circ26'$	Saffier
1615	Trachyandesite, Small-felspar trachyandesite member, Barby Formation	$16^\circ15'$	$25^\circ35'$	Vergenoeg
381	Trachyandesite, Large-felspar trachyandesite member, Barby Formation	$16^\circ34'$	$25^\circ37'$	Ginas
1558	Trachyandesite, Pyroxene trachyandesite member, Barby Formation	$16^\circ38'$	$25^\circ42'$	Aruab
1557	Trachybasalt, Pyroxene trachyandesite member, Barby Formation	$16^\circ35'$	$25^\circ43'$	Aruab

Table 29. Continued

Sample No.	Rock-type, member, Formation	Longitude (to nearest 1')	Latitude (to nearest 1')	Farm
1622	Trachybasalt, Pyroxene trachyandesite member, Barby Formation	16°31'	25°40'	Aubures
253	Basaltic-andesite, Small-felspar trachyandesite member, Barby Formation	16°15'	25°33'	Spes Bona
409	Trachyandesite, Pyroxene trachybasalt member, Barby Formation	16°29'	25°39'	Aubures
390	Trachybasalt, Pyroxene trachybasalt member, Barby Formation	16°26'	25°37'	Ginas
406	Trachybasalt, Pyroxene trachybasalt member, Barby Formation	16°29'	25°39'	Aubures
1541	Trachyandesite lava, Guperas Formation	16°43'	25°43'	Guperas
1591	Basalt, Guperas Formation	16°40'	25°18'	Rooiberg
483	" " "	16°29'	25°38'	Ganaams
1551	Andesite intrusive, Guperas Formation	16°35'	25°37'	Aruab
1439	Trachyandesite intrusive, Guperas Formation	16°36'	25°22'	Naudaus
1390	" " " "	16°40'	25°26'	Naudaus
Samples analysed for K-Ar age determinations:				
910	Muscovite mica, Nam Shear Belt	16°18'	25°00'	Steinfeld
1155	" " " " "	16°28'	25°10'	Eldorado
1154	" " " " "	16°28'	25°10'	Eldorado

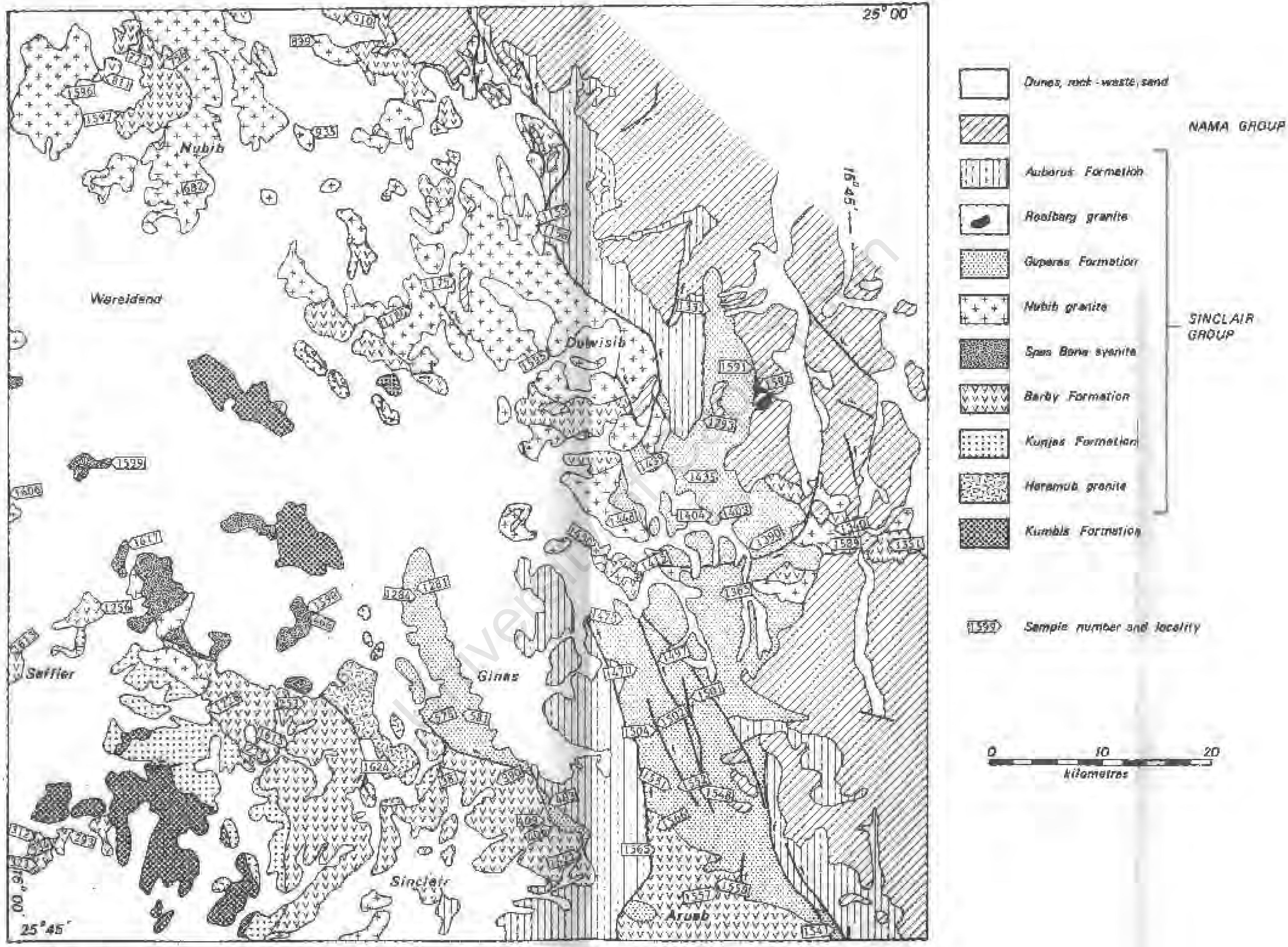


FIGURE 36. GENERALISED GEOLOGICAL MAP SHOWING THE LOCALITIES OF ANALYSED SAMPLES

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PLATES 1-32

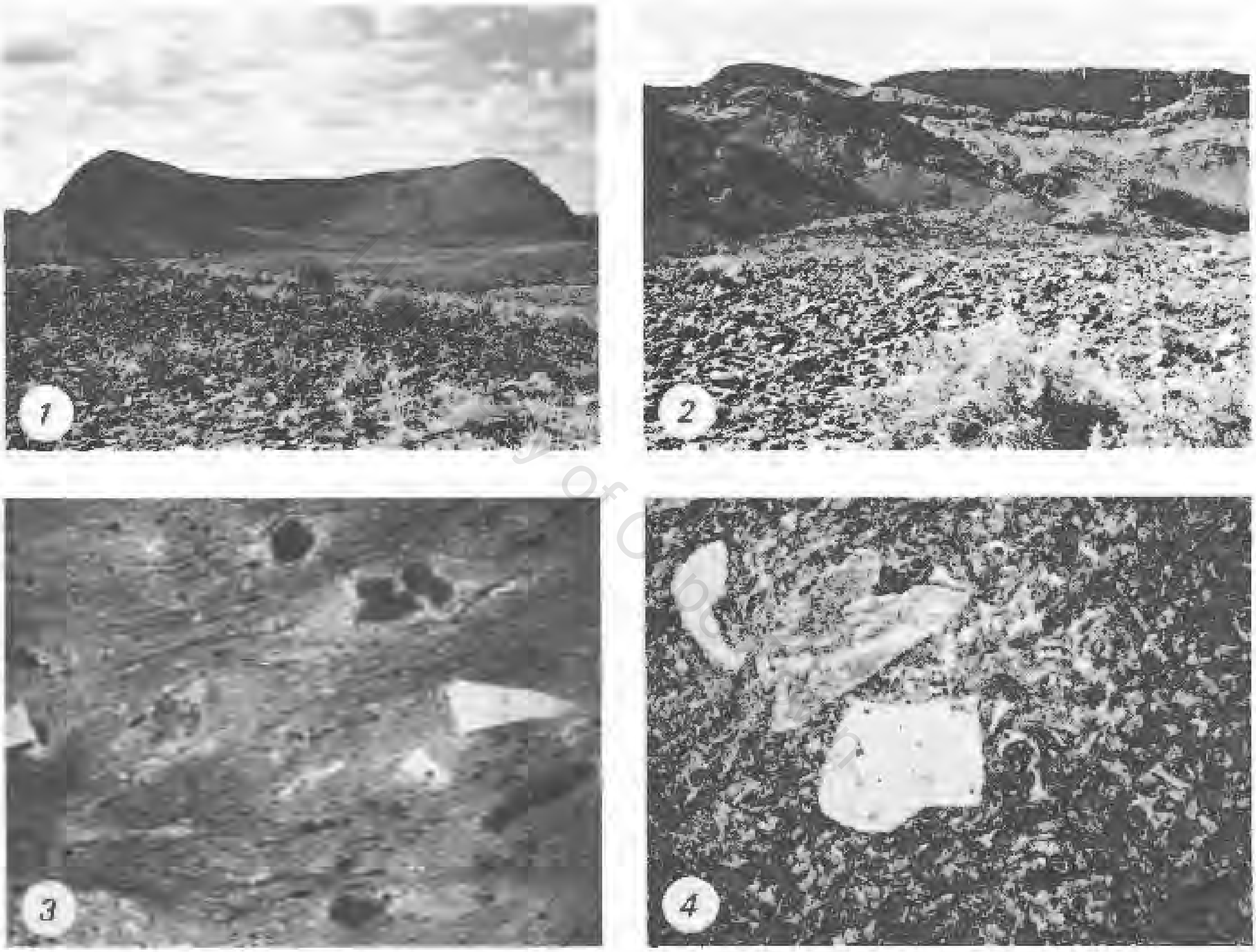
Explanation of Plates 1-4

Plate 1. Easterly-plunging and fault-truncated syncline of unique basal orthoquartzite of the Guperas, overlying Barby basic lavas on the farm Aruab. The more typical and easily-eroded Guperas conglomerate and sandstone overlie the orthoquartzite and occupy the centre of the synclinal structure

Plate 2. Calc-alkaline basic lava flows of the Barby Formation, dipping at moderate angles to the southeast and overlain unconformably by Auborus strata. The flows display varying resistances to erosion depending on the grain-size and degree of vesicularity

Plate 3. Photomicrograph of a welded ash-flow tuff from the Basal volcaniclastic member of the Barby Formation, showing angular chip-like grains of quartz and felspar set in a groundmass of highly compressed glass shards. Devitrification and compression has all but destroyed the original vitroclastic texture (x27)

Plate 4. Photomicrograph of a welded ash-flow tuff from the Basal volcaniclastic member of the Barby Formation, showing angular phenocryst mineral fragments set in a devitrified groundmass possessing a well-preserved shard structure (x40)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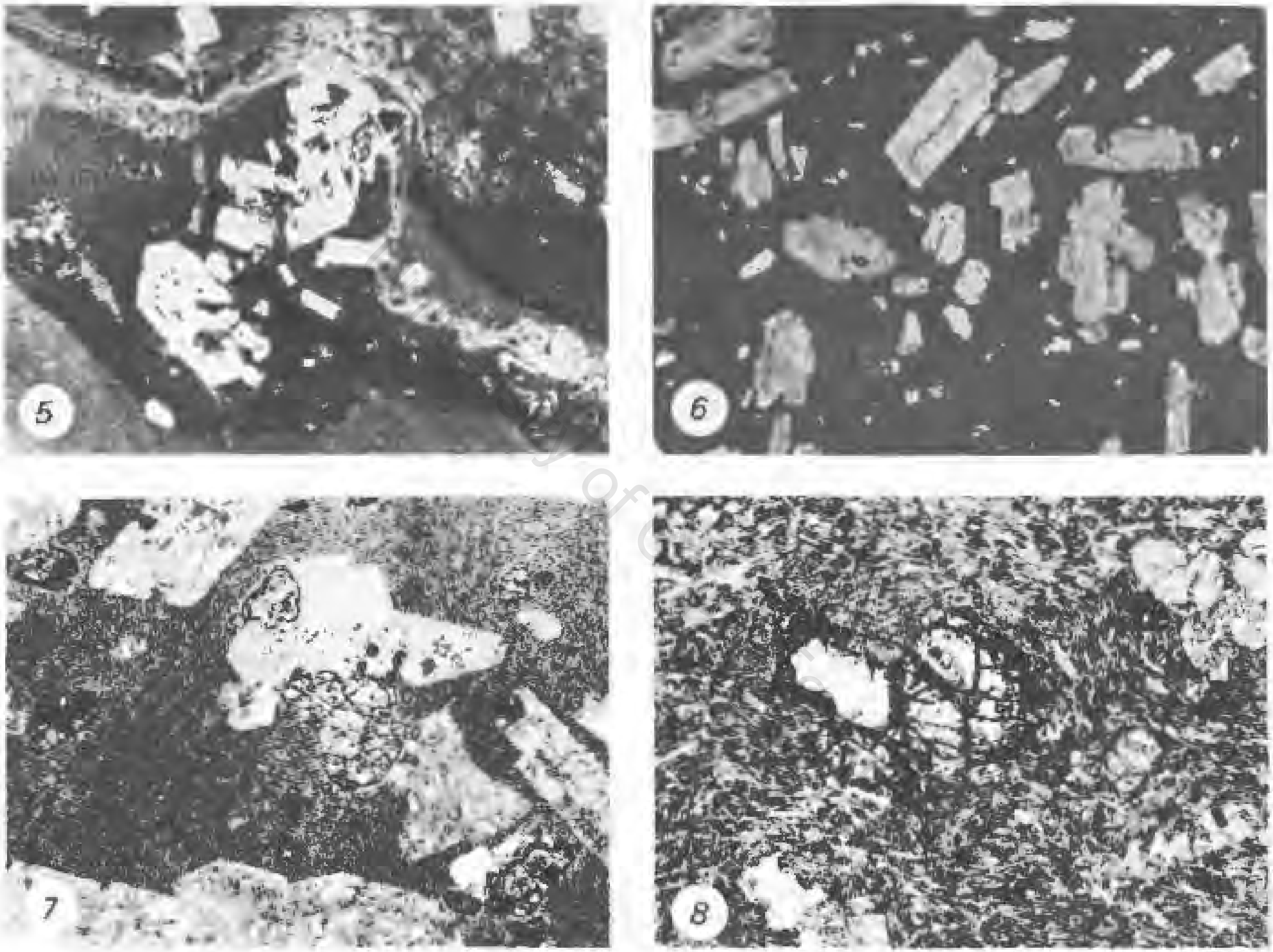
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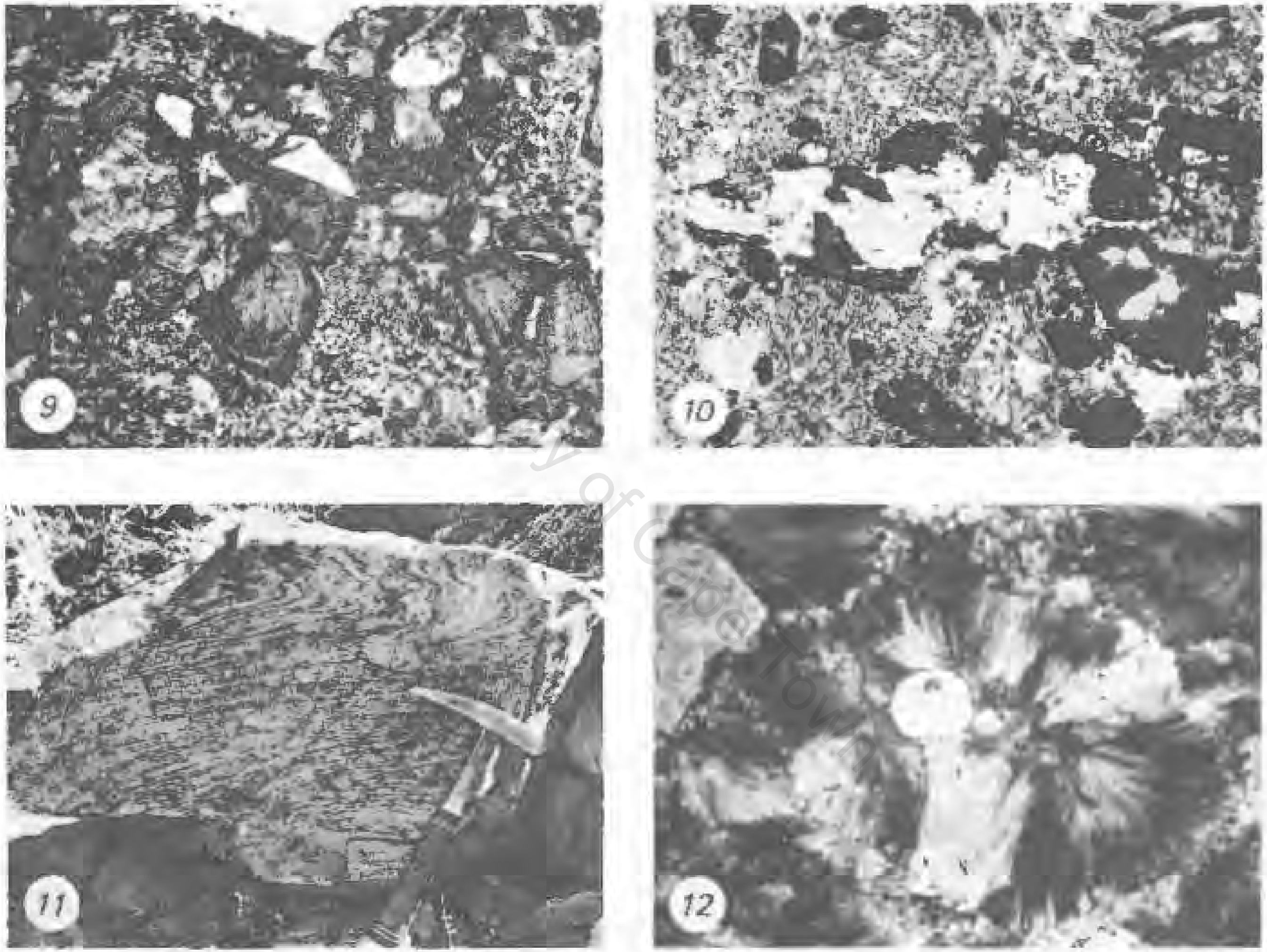
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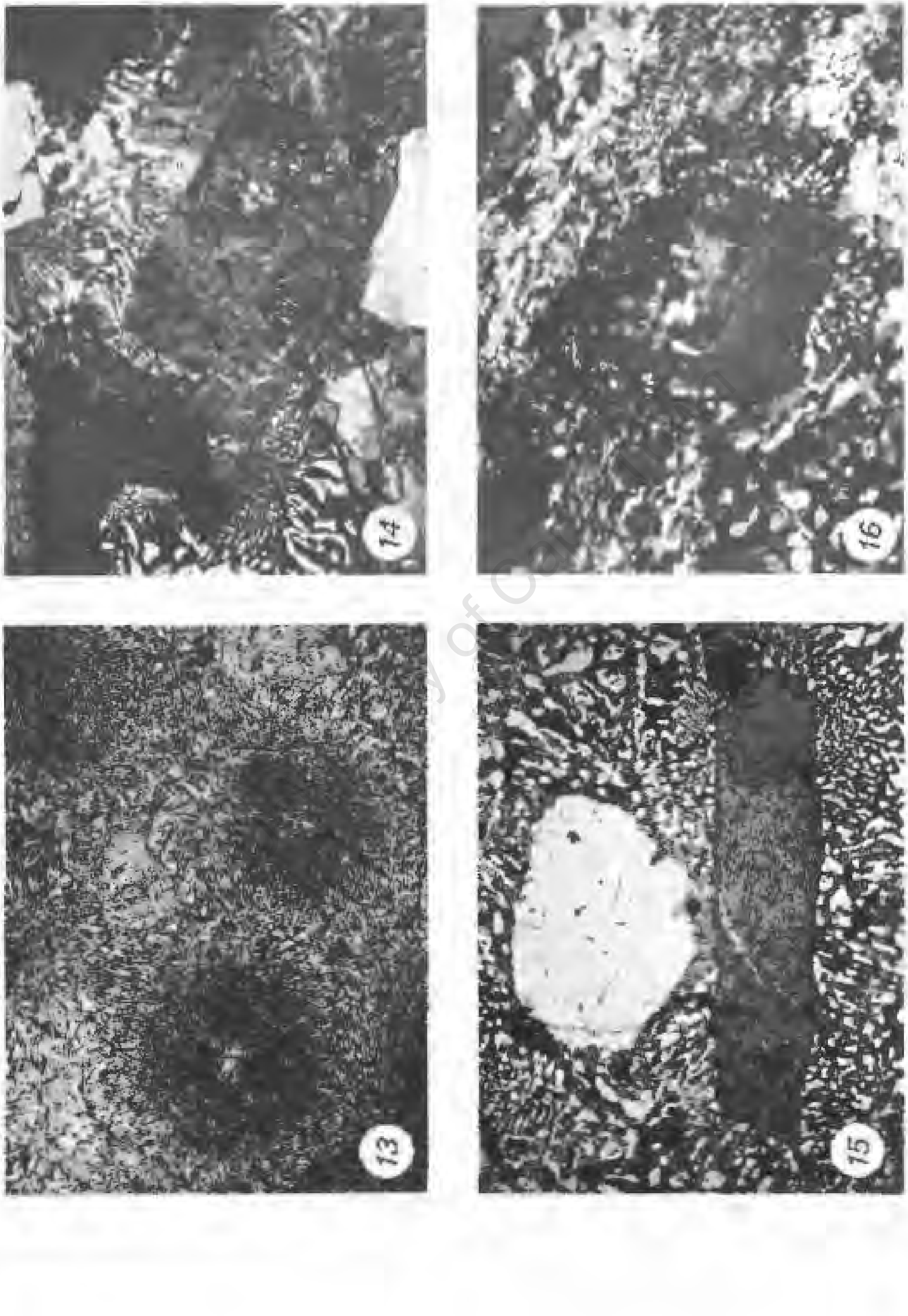
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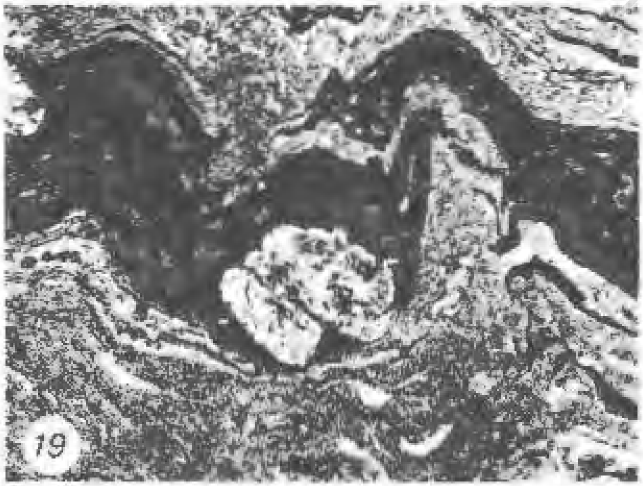
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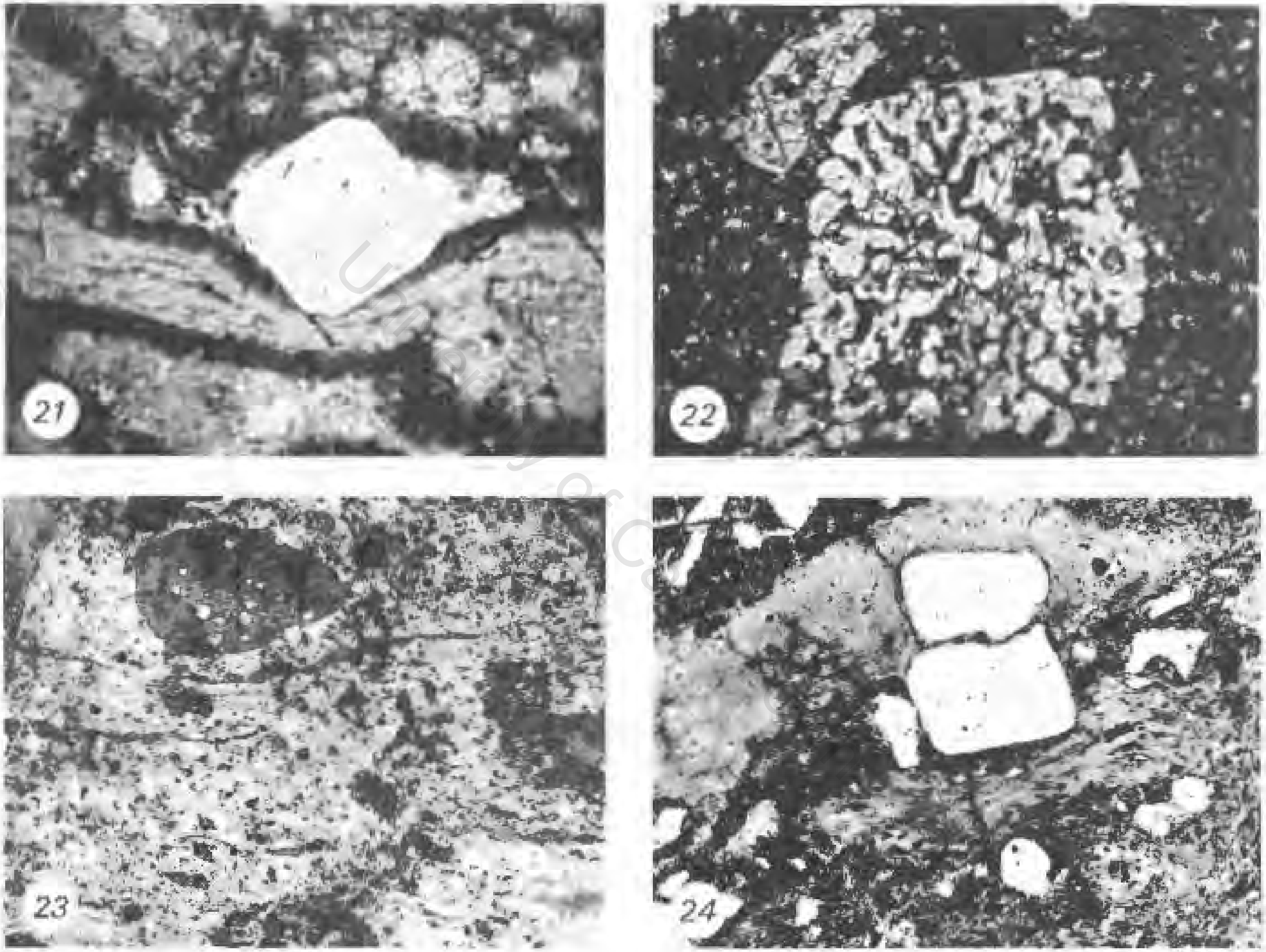
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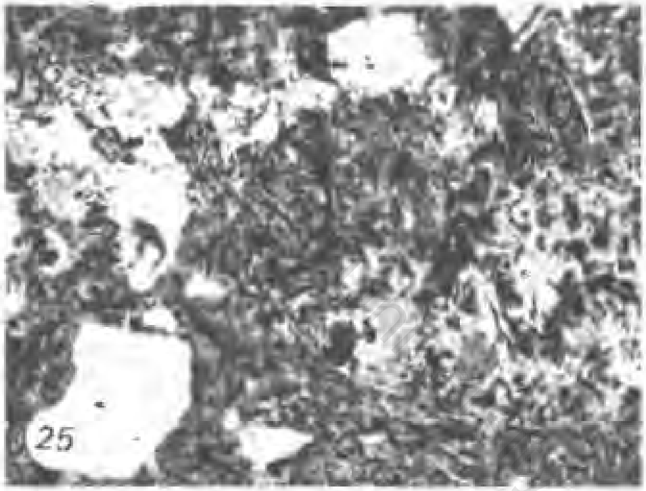

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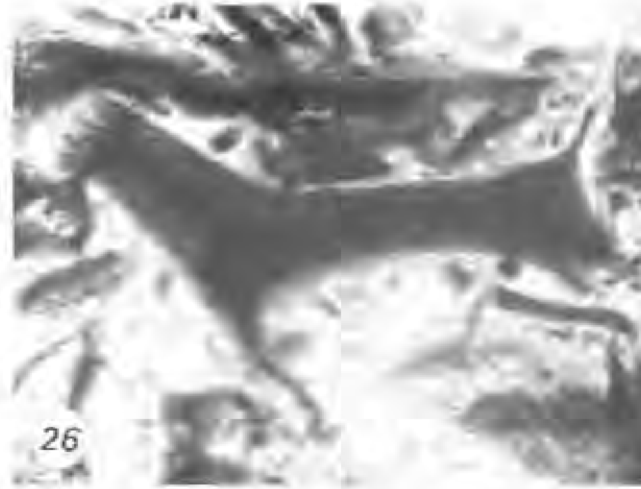

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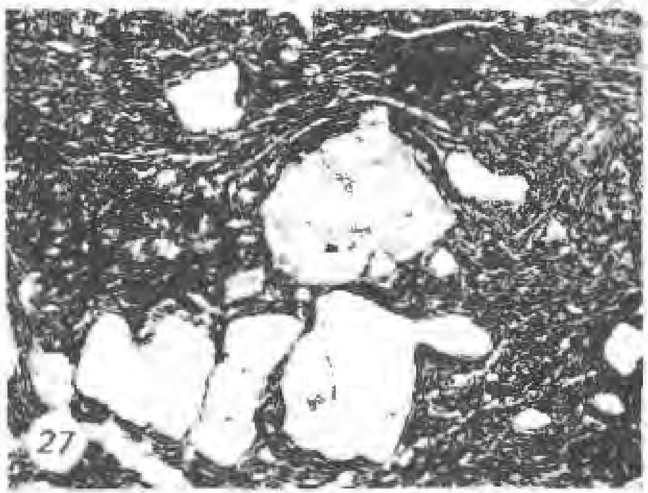

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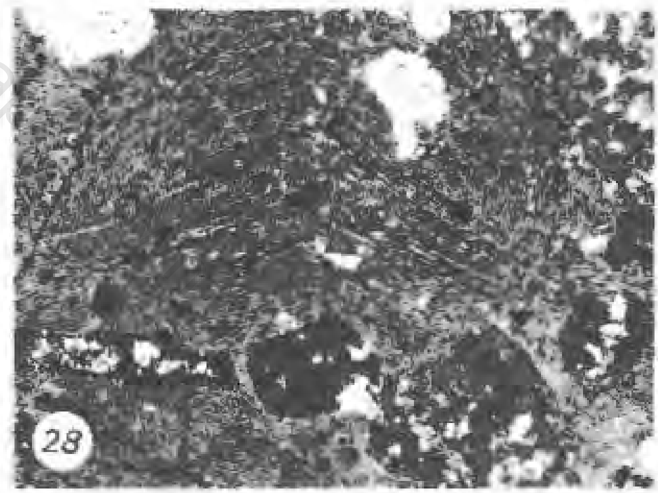

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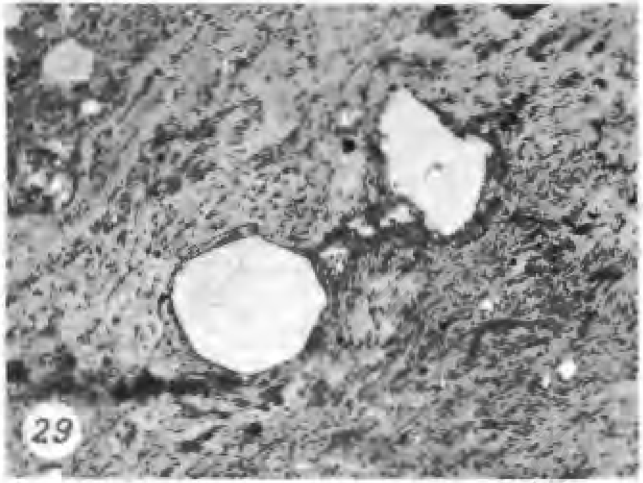
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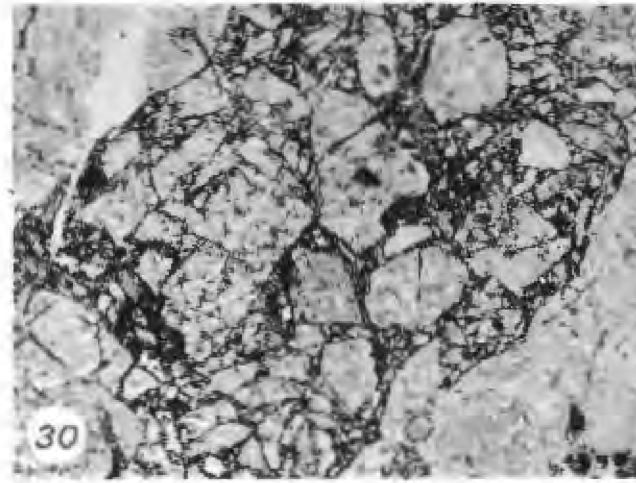
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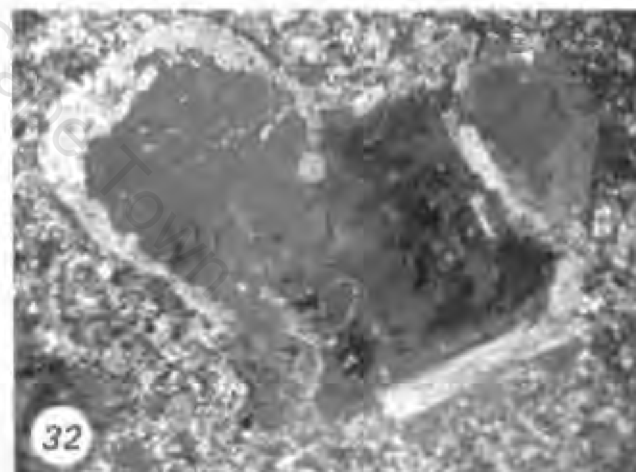
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