GEOLOGY OF PART OF THE CENTRAL DAMARA BELT
AROUND THE TUMAS RIVER, SOUTH WEST AFRICA

by

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Thesis submitted for the degree
of Master of Science at Rhodes
University, Grahamstown,
South Africa.

DECLARATION

All work in this thesis is the original work of the author except where specific acknowledgement is made to the work of others.

Signed

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ABSTRACT

The investigation covering 1500 square kilometres within the central granite zone of the Damara belt, South West Africa, revealed Pre-Damara (Abbabis) basement unconformably overlain by metasediments of the Damara Supergroup. The term Leeukop member is proposed for basal metaconglomerates of the Nosib Group that immediately overlie the basement augen-gneisses. Augen-gneiss clasts are present within the Leeukop metaconglomerates.

The Damara orogeny has only partly affected the Abbabis rocks of the Tumas River Inlier but further to the west the Husab suite of red granites and granite-gneisses, as field and geochemical evidence suggest, were derived syntectonically during the Damara orogeny by reactivation of the Pre-Damara basement. Rössing alaskitic granites represent late stage melts, that were also derived from Pre-Damara basement rocks during orogenesis, which accumulated post-tectonically in structural traps at the base of the Khomas Subgroup. Salem granitoids are present in synclinal structures associated with metasediments of the Khomas Subgroup and syntectonic derivation by anatexis during the Damara orogeny is suggested. In the east the differentiated Gawib granitoid stock was emplaced post-tectonically through basement rocks into the Damara metasediments. A deep seated origin is indicated by high crystallisation temperatures (>850°C) obtained from quaternary Qz-An-Ab-Or-H₂O plots.

The metamorphic grade increases westwards from medium grade to high grade. In the east, the metapelites contain andalusite, and coexisting muscovite and quartz. This indicates that temperatures of 600°C at 3.5-4 kbar pressure were attained. In the west, coexisting wollastonite and anorthite in the Khan gneisses indicate pressure-temperature conditions of 720°C at 4.5-5 kbar. Two tectonic events were responsible for the regional structure. An early F₁ episode produced east-west oriented overturned folds and was followed by an intense F₂ episode of isoclinal folding which is responsible for the dominant northeast-southwest regional fabric. The interference of these folds in the proximity of the underlying basement produced the complex dome and basin structures seen in the central and western parts of the area. An F₃ episode of minor importance was also recognized.

The presence of continental basement rocks in this central part of the Damara belt is evidence for formation of the orogen by in-situ deformation rather than continental collision.
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**MAP SYMBOLS**

The symbols used by Jacob (1974) for denoting the different rock suites on the maps were adopted during this investigation.

The Nosib group rocks are prefixed by the symbol NS and the Swakop group rocks by D. (D = Damara group in the terminology adopted by Jacob (op.cit.)).
The symbols C, Q, Gn, S and R denote the predominant lithology in each stratigraphic unit, which are calcareous, quartzitic, gneissic, schistose and rudaceous respectively.

The granitoids and granite-gneisses are denoted by the symbols G and Gn respectively, with the Abbabis complex granite-gneisses being distinguished by the symbol AGn₁.

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Thanks are due to Professor H.V. Eales, Dr. J.S. Marsh and Dr. P.A. Snowden of the Rhodes University Geology Department for the many valuable discussions and advice given. Also to the field staff of Anglo American Prospecting Services, for their cooperation whilst the field work was being undertaken, and to the management for their support and interest.

The writer gratefully acknowledges the financial assistance provided by Anglo American Corporation of South Africa, and the CSIR, without which the completion of this project would not have been possible.

Finally to my wife Jill, I would like to express my appreciation for all her help throughout the project.
1. **INTRODUCTION**

1.1 **GENERAL**

The area under investigation is situated in the vicinity of the Tumas River within the Namib Desert Park, South West Africa. The western boundary lies approximately 25 kilometers inland from the Atlantic coastline, and the area is bounded by latitudes 22° 42' and 23° 02' and longitudes 14° 46' and 15° 20' (Fig. 1). The area, covering approximately 1500 square kilometers, is centrally placed in the Damara orogen in a zone where intense metamorphism, granitisation and tectonism has occurred.

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**Fig. 1**

Map showing the location and extent of the area investigated (♀♂♀♂♀)
1.2 PREVIOUS INVESTIGATIONS

Early investigations into Damara geology, which incorporated descriptions of this area, were undertaken by Gevers (1931) and Bethlehem Steel Corporation in 1954. These were reconnaissance studies done without topographic control or aerial photographs but provided a valuable background for more detailed work. Maps of Gevers' work covering latitudes 22° to 23° South and longitudes 15° to 16° East on a scale of 1:100 000 were published by Gevers (1934), but the work of Bethlehem Steel Corporation, is unpublished.

In a review of the Precambrian geology of South West Africa up to the early 1960's, Martin (1965) included an overall interpretation of the Damara Orogen.

Recent descriptions of separate areas within the Damara orogen immediately to the north of the area covered by the present investigation were made by Smith (1965) and Jacob (1974). Their field and petrogenetic interpretations have provided a basis for all current interpretations in the central part of the orogen.

More recently, research workers from the University of Göttingen, Germany, have undertaken a series of studies throughout the Damara belt to the north and north-east, separately covering aspects of metamorphism, tectonism, granitisation and associated detailed regional studies. To date, a large proportion of their results is unpublished. Where available, however, reference to this work is mentioned in the text.

The relative locations of all recent investigations covering aspects of Damaran geology within the central part of the belt are shown in Fig. 2.
1.3 PRESENT INVESTIGATION

The current study was begun whilst the writer was employed on exploration in the area by Anglo American Corporation during the period November 1972 to November 1975 inclusive. Additional field work was undertaken in June - July 1976, and laboratory investigations and compilation of the data were performed at Rhodes University Grahamstown during 1976 - 1977.

In compiling the geological map (Map 1), field mapping was first completed by traversing the area at 1km intervals on foot. Subsequent follow-up work was then undertaken on areas requiring more detailed investigation.
The field information was recorded on 1:25 000 field sheets, the data from these then being incorporated onto a 1:50 000 geological map along with data from aerial photograph interpretations done in conjunction with the field mapping. Laboratory studies at Rhodes University involved petrography, geochemistry and structural interpretations and were undertaken using techniques listed in the Appendix.

It is the aim of the study to present a detailed geological account of the area with emphasis on the basement (Abbabis)/cover (Damara) relationships. This is considered under topics such as field relationships, petrography, metamorphism, geochemistry and structure.

1.4 PHYSIOGRAPHY

A description of the physiography of the overall region has been provided by Gevers (1931) and for the area to the north by Smith (1965). A detailed physiographic and geomorphological investigation within the current area is at present being undertaken by Wilkinson (pers. comm) so it will suffice to give only a brief outline here.

The area forms part of the arid Namib Desert and occurs on the Namib plain which is characterised by a gradually rising surface from the coast to an elevation of 700m above sea level in the extreme east of the area.

The local topography forms a broad undulating plain and a wide depression occurs in the central part of the area which is drained by the Tumas River. There is some confusion over the use of the term Tumas as the name for this river, as some local people refer to it as the Tubas river. This drainage system has, however been labelled on the 1:25 000 Government topographic sheets as the Tumas river, and this is the term adopted for this study.

The Tumas river drains the Gawib and Tinkas Flats to
the north-east and east respectively, and the river course passes through the central part of the area from east to west. The Swakop river occurs 3km to the north of the area, and in contrast to the Tumas river, has caused deep dissection of the Namib plain and is flanked by rugged 'badland' terrain. The extreme northern part of the area drains into the Swakop river, and the terrain here is of the 'badland' type.

Throughout the central and western part of the area, numerous 'inselbergs' occur on the otherwise undulating scree-covered Namib plain (Plate 1). These inselbergs generally rise to elevations of about 50m above the surface of the plain and are almost exclusively composed of metasedimentary rocks. Marble and dolomite (Karibib Formation) most commonly form the inselbergs, e.g. Witpoortberg, Tubasberg, Leeukop, while quartzite (Etusis Formation) forms the Rabenrücken and GLÜCKS hills and banded gneisses (Khan Formation) build the prominent Zebraberge. The predominance of marble forming the elevated topographic features in the desert region is in contrast to the inland areas in the vicinity of Usakos, where quartzites form the prominent hills (Chuos Mountains, Otjipateraberge etc.) and the marble formations are topographically subordinate.

In the eastern part of the area schists (Kuiseb Formation) form low irregular hills, elongated parallel to the dominant regional foliation, e.g. Lucasberge.

Throughout the area the granitoid rocks have been preferentially weathered to the level of the Namib plain with the exception of two 'bornhardt'-type features composed of rocks of the Salem granitoid suite e.g. Klein Tubasberg.

The Tinkas and Gawib flats are composed of calcrete and gypsum deposits (1 - 50m in thickness) and these extend westwards into the depression occupied by the Tumas river,
where a series of terraces has been developed. Vegetation throughout the area is extremely sparse, consisting of low shrubs and is generally confined to drainage areas. Subsurface water only occurs, and is confined to the main drainage channels, the largest occurrence being in the Tumas river 8 - 14km west of 15° 00'.

1.5 REGIONAL GEOLOGY

The Damara orogenic belt is one of the younger (late Precambrian) Pan-African mobile belts that transect the African continent. Their origin has been discussed as either intracontinental (ensialic orogeny), Shackleton (1973, 1976) or intercontinental (continental collision) Burke and Dewey (1973).

The Damara orogen extends south-westwards into South West Africa from Botswana, in the north-east, and swings into a NNW-SSE trend which runs parallel to the Atlantic coastline (Fig. 2).

The rock types include geosynclinal sediments which have been subjected to high-grade metamorphism and tectonism and are now represented as quartzites, meta-conglomerates, marbles, calc-granofelses, a variety of micaceous schists and amphibolites. The Damaran metasediments are underlain by granites and gneisses belonging to the Abbabis Complex, which near Usakos, have been dated at 2040 ± 40my (Jacob, Burger and Kröner 1977). Pre-Damaran basement rocks are exposed at the northern and southern limits of the orogen and also in a central zone where they have in part, been subjected to reactivation during the Damaran orogenic period. The metasediments of the Damara group have been loosely subdivided into a marginal miogeosynclinal facies ('Outjo Facies' – Martin 1965) and a central eugeosynclinal facies ('Swakop Facies' – Martin 1965). However, the central part of the eugeosynclinal facies, in the vicinity of the pre-Damara granite-gneisses is
composed largely of metamorphosed miogeosynclinal-type lithologies (conglomerates, quartzites, marbles and calc-granofelses) rendering the eugeosynclinal-miogeosynclinal concept an oversimplification.

The area under investigation is situated on the southern margin of the centrally occurring Abbabis and Husab granite-gneisses and overlying miogeosynclinal-type metasediments (Fig. 2).

1.6 CLASSIFICATION AND NOMENCLATURE

Proposals for the stratigraphic classification and nomenclature of the rocks of the Damara orogen have been recently introduced by one of the South African Committees for stratigraphy (in prep.) after initial suggestions by Kroner (Ed.) (1974). These proposals were deemed necessary due to the variation and complexity of the rock associations causing unnecessary confusion in the terminology being adopted. The terminologies used by Smith (1965), Jacob (1974) and in the present study are listed in Table 1. The origin of the Formation and Group names used in the classification of the units occurring in the Damara belt to date has been discussed by Jacob (1974, p.3) and need not be duplicated here. The new terminology adopted in this study, where different from that used by previous workers, is discussed below.

1.6.1 DAMARAN METASEDIMENT

A large occurrence of metaconglomerate occurs at the base of the Etusis Formation and above the pre-Damara (Abbabis Complex) augen-gneisses around the margin of the Tumas River Inlier (see Map 1). This unit has lithologic characteristics differing from the typical Etusis feldspathic quartzites (see section 2.1.2.1) and as it occurs in the vicinity of Leeukop Mountain it is classified as the LEEUKOP MEMBER of the Etusis Formation.

The sequence of the interbedded marbles, quartzites
## Table 1

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### Notes
- **SMITH (1965)**: Upper Damara System
- **JACOB (1974)**: Lower Damara System
- **PRESENT INVESTIGATION**: Not observed

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**Stratigraphic Terminology adopted for the Damara Supergroup**

**Group Rocks**: The classification of rocks into groups, subgroups, and formations within the Damara Supergroup follows a hierarchical system. Each level of classification is designed to provide a clear and distinct boundary between rock units, facilitating easier identification and study of geological features. The terminology used in this context is essential for accurate stratigraphic correlation and understanding the temporal and spatial relationships of geological layers.
and schists (Dome-Gorge Formation), formerly assigned to the Lower stage of the Hakos series by Smith (1965) and the Rössing Formation by Jacob (1974), which occur below the tillite (Chuos Formation), have not been recognised in the present area. The Dome Gorge Formation and the Chuos, Karibib and Kuiseb Formations have now been classified into the Ugab and Khomas Subgroups respectively, which constitute the Swakop Group. The Swakop Group is equivalent to the former eugeosynclinal Swakop facies of the Damara System (Martin, 1965).

The terms KARIBIB FORMATION and KUISEB FORMATION suggested by the S.A.C.S. Working Group are adopted for the calcareous and pelitic metasediments similar to those occurring at the type localities. The terms Upper stage of the Hakos Series and Khomas Series (Smith 1965), Husab Formation, Tinkas Formation and Witpoort Formation (Jacob 1974) were previously used for these metasediments. The terms Tinkas Formation and Witpoort Formation were used by Jacob to define different facies of the Khomas Subgroup. The term Kuiseb Formation is now used for the metasediments of the former Khomas Subgroup and the terms TINKAS MEMBER and WITPOORT MEMBER are used, to reflect the differences of the facies. The interbedded sequence of marbles and biotite schists at the eastern boundary of the GLÜCKS dome immediately overlying the Etusis quartzites are considered to be the chronostratigraphic equivalent of the Karibib Formation. For purposes of the lithostratigraphic classification however, these rocks are grouped as interbedded Karibib (marbles) and Kuiseb (metapelites and calcgranofelses) Formations. The term granofels is used after the definition by Goldsmith (1959) to define metamorphic rocks that have a massive rather than foliated appearance and are
A Summary of the Lithologies Present in the Rocks Found in the Area.
composed of an equigranular or granoblastic mosaic of minerals. The term calc-granofels is used to distinguish rocks containing calcium-bearing minerals and is used in preference to the term calc-silicate rock.

The different lithologies of the metasedimentary units occurring in this area are summarised in Table 2.

1.6.2. GRANITOID ROCKS

In the area under investigation it was possible to distinguish between granites and gneisses that had been in existence prior to Damaran sedimentation and those that appear to have been generated during the Damara orogenic period.

This former group, composed largely of augen gneisses with minor metasedimentary relicts, is confined largely to an area east of 15°00' longitude, which is referred to as the Tumas River Inlier. The name ABBABIS COMPLEX is adopted for these rocks, after the Abbabis Formation, consisting of similar rock types, described by Gevers (1931) and Smith (1965) which occur in the Karibib area.

The latter group of Damaran-generated rocks, consisting of migmatites, granites, granite-gneisses, pegmatites and sheared augen gneisses, occur largely to the west of 15°00' longitude. These rocks are grouped under the name HUSAB GRANITE-GNEISS SUITE as they occur in the large antiform on the western side of the Husabberge-Witpoortberge and in the vicinity of the Husab fluorspar mine and Husab Gorge. The name Arandis granite, suggested by the S.A.C.S. Working Group for these rocks, is perhaps inappropriate in that the majority of the rocks in the vicinity of Arandis are augen gneisses that can be correlated with the pre-Damaran Complex (Jacob, pers comm.)

The S.A.C.S. Working Group proposed the term Waldau
granite to distinguish those red granites that are intrusive into Nosib and Swakop groups, from the Arandis granites. For the purposes of this study these rocks have been grouped into the Husab granite-gneiss suite.

The term SALEM GRANITOID SUITE is adopted for the rocks of essentially granodioritic composition, formerly grouped under the name 'Salem Granite' by earlier workers.

'Granitoid' is a sack term suggested by Streckeisen (1973, p. 30) for alkali granites, granites, granodiorites, diorites and tonalites.

The late-kinematic to post-kinematic pegmatitic granites, and dykes formerly termed 'Alaskitic Pegmatitic Granite' by Jacob (1974) and 'Pegmatitic granite' by Smith (1965) that occur in this area are called the RÖSSING ALASKITIC GRANITES after similar rocks occurring at Rössing Mine. The name Rössing Mine Granites is suggested by the S.A.C.S. Working Group, but is not preferred in this study as they have a rather more widespread distribution than is suggested by the name.

The association of the different granitoid rocks, relative to one another, is shown in a schematic manner in Fig. 3b.
2. **METASEDIMENTARY ROCKS**

2.1 **LITHOLOGIES AND FIELD RELATIONSHIPS**

The distribution and lithologic characteristics of the various rock units are discussed and comparisons drawn with their occurrences outside the area mapped.

2.2.1 **ABBABIS COMPLEX (AGn₁)**

The rocks which constitute the Abbabis Complex are of pre-Damaran age and occur in anticlinal and domal structures where they are overlain unconformably by Damara metasediments (refer to structure, Chapter 5).

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**Fig 3a**

Schematic representation of Damara geosynclinal sediments (Fig 3a) and Granitoid rocks (Fig 3b) as found in the Tumas River Area.
The predominant rock types of the Complex are augen gneisses with subordinate occurrences of other granitic and gneissic rocks, which are discussed later in section 4.1.1. Metasedimentary rocks are represented by amphibolite (possibly orthoamphibolite), calc-granofelses and biotite schists, which are present as xenoliths and remnants within the gneissic rocks. These occur as discontinuous bands, generally less than 5 metres in width, which lie parallel to the foliation in the enclosing gneisses (Plate 2) and are found intermittently throughout the outcrops of the Abbabis rocks. The metasedimentary xenoliths are conspicuous due to their dark grey-green colouration and siliceous nature and form protruberences above the level of the surrounding preferentially weathered granitic rocks.

Amphibolite and calc-granofels remnants occurring within the Husab granite-gneiss suite west of 15°00' longitude are considered to be of pre-Damaran age although the enclosing rocks have been remobilised during the Damara orogenic period. The Abbabis metasediment relics are best preserved in the Tumas River Inlier where Damara tectonism and ultrametamorphism has not been marked. To the south of the Witberg, a narrow E-W trending syncline occurs within the Abbabis rocks. The rock types occurring in this syncline are marbles, calc-granofelses and quartzites that are similar in lithology to Damara metasediments. This is considered therefore to represent a tight infolding of Damara rocks during the F1 phase of deformation (refer chapter 5).

2.1.2 THE NOSIB GROUP OF THE DAMARA SUPERGROUP

2.1.2.1 ETUSIS FORMATION (NS1,Gn, NS1,R, NS1,Q)

The principal rock types are psammitic and psamphitic in nature and include meta-conglomerates, pelitic schists and pink feldspathic quartzites. These rocks
represent the earliest metasediments in the Damara sedimentation sequence. The Etusis rocks occupy anticlinal and domal structures i.e. Glucks dome, or are present as remnants in the cores and on the limbs of the larger domes. They are distributed throughout the area although in the central part they are absent or poorly developed. Where they directly overlie the pre-Damara Abbabis Complex the relationship is unconformable, although immediately adjacent to the contact, the foliation in the underlying basement rocks has been transposed parallel to the contact (refer Structure, Chapter 5). In the west, where they overlie the granites and gneisses of the Husab suite, the bedding in the quartzites and the foliation in the granite-gneisses is paraconformable.

The Rabenrücken occurrence in the extreme east of this area and the Langer Heinrich and Horebis occurrences described by Jacob (1974) represent the easternmost exposures of Etusis metasediments in the central part of the Damara belt. Further to the east the overlying Swakop group metasediments occur.

On the flanks of the Tumas River Inlier which contains Abbabis Complex augen gneisses, a considerable development of conglomerate occurs at the base of the Etusis Formation below feldspathic quartzites. This conglomerate has a grey-green appearance in outcrop and contains an assortment of pebbles and boulders, mainly of granitic character (Plate 3). Augen-gneiss pebbles identical to the rocks of the underlying Abbabis Complex are common as clasts (Plate 4), in addition to leucogranite, pegmatite, foliated granite and occasionally schist and quartzite. Small alkali-feldspar relics from basement augen gneisses and pegmatites, often form small grit sized clasts. The matrix is migmatitic and semi-pelitic in nature, containing the minerals quartz, biotite, diopside, hornblende and epidote. These rocks constitute an easily recognisable field unit and the term LEBUKOP MEMBER
(Ns₁R) is used to distinguish this local basal rudaceous sequence from the overlying arenaceous rocks of the Etusis Formation. The largest occurrence of the Leeukop conglomerate lies immediately to the northeast of Leeukop Mountain. The thickness of the unit there is approximately 500m and on the western flanks of the Tumas River Inlier 250 - 350m of conglomerate is present. It is not developed in the north-eastern part of the Inlier, where feldspathic quartzites directly overlie the basement rocks.

Elsewhere, the lower part of the Etusis Formation is characterised by metasediments of a more pelitic nature (biotite schists, migmatites and gneisses – NS₁Gn). This is particularly apparent in the core of Glück's dome and in the anticline immediately east of the Klein Tubasberg. In this latter locality the metapelites and migmatites are intimately associated with granite-gneisses of the Husab suite. The migmatitic gneisses developed, exhibit classical migmatitic features such as those documented by Mehnert (1968) and can be classified as metatexites in which biotite rich restites remain as identifiable portions of the original rock (Mehnert, op.cit, p. 253). In areas where there is a complex association of the Etusis migmatites and the Husab granite-gneisses, separation of the two units was not possible and they are shown as stippled areas on the geological map.

The pink feldspathic quartzites which predominate in the upper part of the Formation are similar in appearance to those occurring in the Otjipateraberge and the Chuos Mountains, described by Smith (1965) and Jacob (1974). Their lithology changes from white, sugary fine-grained (< 1.5mm) quartzites and glassy grey magnetite bearing quartzites in the west to coarser grained quartzites and grits interbedded with conglomerate bands (< 4m wide) in the central and eastern part of the area (Plate 5). They constitute the most common rock type of the Etusis Formation in this area.
and are commonly finely bedded, with cross-bedding features apparent in some of the exposures (Plate 6). The thickness of the quartzites is extremely variable and they are absent in the central part of the area adjacent to the Witpoortberge as at this locality the Karibib marbles directly overlie the Abbabis rocks. The largest development of feldspathic quartzite, with minor biotite gneiss and conglomerate is in the Glucks Dome, where the sequence is approximately 1200 metres in thickness. At the Rabenrücken in the east, and west of the Welleberge in the extreme western part of the area the thickness of the formation is approximately 500 – 600 metres.

2.1.2.2. KHAN FORMATION (Ns₂Gn)

These rocks form the uppermost stratigraphic unit of the Nosib Group and are found conformably overlying the Etusis quartzites and paraconformably underlying the marbles of the Karibib Formation (Swakop Group). Intercalation of the Khan banded-gneisses and the Etusis quartzites is observed at the base of the Khan Formation immediately to the north of the Welleberge and this feature was also noted by Jacob (1974) for the Khan/Etusis contact in the area to the north. This formation is confined to the north-western part of the area where it forms small elevated topographic features such as the Zebraberge, Welleberge and End-Klippe.

The formation contains banded-gneisses, biotite schists, quartzites and amphibolites. The banded gneisses constitute 90 per cent of the unit and form an easily identifiable field unit due to their grey-green colouration and their migmatitic appearance. The migmatitic structures are predominantly stromatic (banded) where mafic layers of amphibole and pyroxene have segregated from felsic bands of quartz and feldspar (Plate 7). Occasional spotted or fleck (stictolithic) structures
(Mehnert 1968 p37) are also observed where diopside and hornblende 'spots' are surrounded by leucocratic halos.

The upper 10 - 30 metres of the formation, on the eastern side of the Zebraberge consists of biotite schists which contain pyrite, pyrrhotite and small amounts of chalcopyrite. At Endklippe there is a large oval-shaped occurrence of ortho-amphibolite within the upper part of the formation.

Narrow quartzite and conglomerate bands are occasionally found within the sequence of banded gneisses at the southern closure of the Ida Dome and in the Welleberge.

The rocks of this formation, occurring to the north of the area under review which are described by Smith (1965) as calc-granulites; by Nash (1971) as basic gneisses, and by Jacob (1974) as banded gneisses are similar in nature to those found in this area. The occurrences of Khan metasediments are noticeably confined to the western parts of the area studied by Smith and Jacob respectively, and this is also apparent in this area. This formation is thus of restricted geographical extent.

2.1.3 THE SWAKOP GROUP OF THE DAMARA SUPERGROUP

This group is characterised by calcareous and pelitic rocks and paraconformably overlies the Nosib Group. In this area the Nosib and Swakop groups were found to be conformable although a local unconformity occurs to the south of the Tumas River, at 15°00' longitude, where the Karibib marbles have a 20-30° angular relationship to the underlying Etusis (Leeukop) conglomerate. A paraconformable relationship was also found between the two groups by Smith (1965) and Jacob (1974). Separation of the metasediments into two groups is based primarily on the different nature of the rock types i.e. arenaceous/rudaceous Nosib group and calcareous/argillaceous Swakop group, and also on the
apparent observance by earlier workers of an unconformity present in the northern part of the Damara belt.

2.1.3.1 CHUOS FORMATION \( (D_2G) \)

The rocks of this formation are metamorphosed diamictites, although they are commonly referred to as tillites of glacio-marine origin. The extent of the formation is widespread throughout the Damara belt and has been discussed by Gevers (1931), Smith (1965), Martin (1965), de Waal (1966), Hälöch (1970), Miller (1972) and Jacob (1974). It forms an important marker horizon when present as it lies between the Dome-Gorge Formation and the Karibib Formation enabling these two calcareous suites to be distinguished.

In this area the continuity and extent of the Chuos Formation outcrops is very limited, and it is almost entirely absent from the central part. In the extreme west it occurs locally above the Etusis and Khan Formation metasediments. In the south-western part of the area, quartz biotite schists containing scattered pebbles and boulders are associated with haematite-magnetite-ore accumulation. This association of iron ore and tillite is well known particularly in the Kaokoveld and is documented by Martin (1965). On the western flanks of the Witpoortberge a boulder-bearing amphibolite/metapelite occurs in very localised outcrops which directly overlie gneisses of the Abbabis Complex. This occurrence at the Witpoortberge is of extremely limited extent and it remains uncertain as to whether it is a basal conglomeratic zone of the Nosib Group or the Chuos diamictite. The rocks at this locality contain large (0.5m) clasts of augen gneiss and granite enclosed in a biotite-hornblende matrix.

The largest occurrence of the tillite is in the extreme east, on the western side of the Rabenrücken where it overlies the quartzites of the Etusis Formation and
directly underlies the Kuiseb Formation metapelites. In outcrop, the unit is characterised by a grey quartz-biotite granofels/schist matrix which encloses boulders and pebbles of varying sizes (0.02 - 1.0m). Commonly the clasts do not form a very large percentage of the rock (< 20%) and they consist of red granite, leucogranite, biotite gneiss, augen gneiss (Plate 8), quartzite and vein quartz.

In general, the occasional angularity of clasts, the abundant matrix and the poorly sorted and packed arrangement of the clasts in the majority of the exposures supports a glacio-marine origin for these rocks.

2.1.3.2. KARIBIB FORMATION (D₃C)

This formation characteristically consists of a thick marble sequence with subordinate interbedded biotite schist and calc-silicate rocks. There is gradual change in lithology apparent across the area:

i) In the west, the sequence is composed predominantly of marble, with minor amounts of calc-silicate and biotite schist.

ii) In the central part, narrow marble bands interbedded with schist are developed locally at the base of massive marble bands.

iii) Immediately west of Glück's Dome the sequence consists of thin marble and schist bands at the base, followed by a thick massive marble unit with interbedded thin marble bands and biotite schist zones at the top of the sequence. These latter schist zones are regarded as part of the overlying Kuiseb Formation.

iv) Immediately east of Glück's Dome the formation is represented by narrow (1 - 20m) (relative to the massive marbles in the west) marble bands only, interbedded with biotite schist of the Kuiseb Formation.
This latter occurrence represents the easternmost appearance of the Karibib Formation as further to the east, in the Rabenrücken anticline, the Karibib Formation is absent. The Chuos and Kuiseb Formations at the Rabenrücken are in direct conformable contact.

The lithology of the Karibib Formation is similar to that described by Smith (1965) for the Upper stage of the Hakos Series, and Jacob (1974) for the Husab Formation, in the area to the north.

Throughout this area, the Karibib Formation generally overlies the metasediments of the Nosib Group para-conformably (the Chuos Formation only occurs locally). In the central part of the area, however the Karibib Formation marbles directly overlie the granites and gneisses of the Husab and Abbabis suites. It appears that in this central part, the marbles were deposited directly onto the pre-Damaran basement. Tight infolding of the marbles with the granite gneisses occurs throughout the central part of the area, and several of these units may prove to be of pre-Damaran age.

The Formation, in appearance, is characterised by white/cream fine to coarsely crystalline marble. Impure zones are distinguished by the development of calc-silicate bands containing diopside, brown calcite, wollastonite and garnet (Plate 9).

This unit generally occurs steeply dipping, on the limbs of the anticlines and synclines, and is present in thicknesses of up to 600 metres.

2.1.3.3. **KUISEB FORMATION (D₄S)**

The Kuiseb Formation is represented in this area by a sequence of pelitic rocks that can be separated into two different units, each having a different lithology. Erosion, and ultrametamorphism to the Salem granitoids in the west, has reduced the thickness of the Kuiseb Formation in this area to approximately 2 000 metres.
In the west the WITPOORT MEMBER is developed, and is characterised by a homogeneous sequence of biotite schists, occasionally containing cordierite and garnet. These rocks are migmatitic in parts where in situ pegmatitic segregations have formed. The rocks are found in syncline and basin structures overlying the marbles of the Karibib Formation.

In the eastern part of the area, rocks of the TINKAS MEMBER constitute the predominant rock type, and occur in both anticlinal and synclinal structures. They are regarded as the stratigraphic equivalent of the Karibib Formation and the Witpoort member of the Kuiseb Formation that occurs in the west as they occupy equivalent stratigraphic levels and they directly overlie the Chuos Formation at the Rabenrückchen. The Witpoort and Tinkas members are not observed in contact in the field.

The Tinkas metasediments are a sequence of calc-granofels bands (generally 0.05 - 2m thick) interbedded with the pelitic schists and quartz-biotite granofelses (generally 0.1 - 10m thick), see Plate 10. The calc-granofels bands are more resistant to weathering than the intervening schists and granofels and an extremely irregular 'topography' results on the scale of the outcrop. There is only very minor development of pegmatite in this sequence and this occurs in the lower part where an origin from depth and not in-situ development is considered. The Tinkas Member of the Kuiseb Formation in this area is very similar to the Tinkas Formation described by Jacob (1974), however marble bands are not observed in the sequence in this area. The marble bands occurring at the base have been classified in this study as part of the Karibib Formation. The considerable development of interbedded calc-granofels/schists continues further to the southeast out of the area mapped, with the amount of calc-granofels in the sequence decreasing towards the biotite schists of the Khomas trough.
2.1.4 MISCELLANEOUS ROCKS

2.1.4.1 BASIC INTRUSIVES

Several small gabbroic intrusives occur in the vicinity of 23°00' latitude at 8-10km west of 15°00' longitude, associated with the red granites and gneisses of the Husab suite. These bodies are small (less than 50m in diameter) and appear to occur in a cluster. They are extremely dark in appearance and form weather-resistant hummocks strewn with in-situ developed boulders. The intrusive bodies appear to be metamorphosed and foliated at the margins, implying that emplacement occurred prior to or during the Damara orogenic period.

The rocks are characterised by the mineral assemblage plagioclase, olivine, orthopyroxene, hornblende, biotite, ± calcite ± clinopyroxene ± ore minerals. The olivine grains, which in part are now altered to serpentine, are surrounded by successive rims of orthopyroxene, hornblende and biotite (Plate 11), indicative of prolonged reaction of the early formed crystals with the melt. A basic dyke that occurs within the Husab suite adjacent to 15°00' longitude, north of the Swakop/Windhoek road may also be related to this group of rocks. Basic intrusives were reported by Smith (1965, p14) occurring within the Abbabis Complex and the rocks in this area may be related.

2.1.4.2 DOLERITE

Dolerite occurs throughout the area as dykes and sheets which are conspicuous as they form elevated black ridges that traverse considerable distances. The dykes are generally less than 3 metres wide, near vertical in attitude and trend predominantly NNW to NNE or ENE. There are little or no contact metamorphic effects associated with the intrusives which are probably of Karoo age.
2.1.4.3 SUPERFICIAL DEPOSITS

Considerable accumulations of calcrete and gypsum occur under the Gawib flats and along and adjacent to the Tumas River. These deposits are of Cainozoic age and form very extensive flat areas, except where dissected by recent drainage such as along the lower part of the Tumas River where a series of terraces occurs. The lithology of these accumulations is cemented grits, sands and conglomerate beds which are occasionally in excess of 50m in thickness and are capped by a resistant gypsiferous layer. Unconsolidated and consolidated red sands are found within the deposits along the Tumas river channel. Higher level accumulations of calcrete, gypsum and conglomeratic beds are found to the south and west of the Zebraberge. In the extreme east of the area the Tinkas flats are underlain by another high level calcrete deposit.

The development of gypsum appears to be restricted to the western and central parts of the area, as in the east (Tinkas Flats) calcrete alone is developed. Gypsum-rich deposits in the Namib desert appear to be confined to a coastal zone occurring up to approximately 70km inland from the Atlantic Ocean. The development of gypsum in this zone has been attributed to the presence of the sulphur laden mist that develops due to the cold Benguela current interacting with the warmer air from the continental mass (Martin, 1965). To the east of this zone the presence of the mist is less common and its effects are not reflected in the formation of gypsum.

2.1.5 DISCUSSION

It is apparent that the stratigraphy and lithology of the different metasedimentary units in the central part of the Damara belt, established from investigations in the area to the north, are essentially similar to those occurring in this area. The present knowledge of the nature and character of the metasedi-
ments throughout the major part of this central zone allows an attempt to be made to reconstruct the paleoenvironment in existence at the time of geosynclinal infilling and sedimentation.

The conclusions of Jacob (1974) regarding the presence of a ge-anticlinal ridge strongly influencing the nature and extent of the sedimentation of the Damara sediments, are reinforced by the results obtained in the present study.

The existence of the Tumas River Inlier, flanked by the coarse Leeukop conglomerates and coarse-grained Etusis quartzites is evidence for a high energy environment existing close to a pre-existing landmass, and is not in agreement with derivation due to deep-water sedimentation as should be expected if this centrally occurring zone occupied the trough of the Damara geosyncline. The extent of this ge-anticlinal landmass was probably considerable, as the nature, thickness and extent of the sediments indicates.

Occurrences of exposed pre-Damaran basement rocks flanked by psephitic sediments extend from the Abbabis Inlier near Karibib, through the Tumas River Inlier (near 15°00' longitude and 23°00' latitude) to additional occurrences 15 - 30km south of 23°00'. This represents a strike length of at least 200km and a minimum width of approximately 70km. (Rabenrücken to Arandis). The existence of a ge-anticlinal ridge had a pronounced effect on the geographical extent of the various formations. The basal Leeukop conglomerate occurs as a distinct unit over a fairly restricted area (+160 sq kms) and is similar to an occurrence reported by Gevers (1931) and Smith (1965) on the flanks of the Abbabis Inlier. The quartzites of the Etusis Formation are extremely varied in lithology, ranging from fine-grained cross-bedded sediments to quartzites with interbedded coarse conglomeratic bands. This is indicative of separate basins with variable energy environments, or uneven subsidence of the geosynclinal floor. The restriction of the metasediments
of the Khan Formation to an area west of where the Marmor Pförte-Husaberg-Witpoortberge now lie, indicates that the ge-anticlinal ridge existed at this time in this vicinity, and conditions of deposition were such that the semi-calcareous sediments of the Khan and the calcareous and pelitic sediments of the Dome-Gorge Formation were deposited only in this area. During deposition of the Chuos diamictite and the calcareous beds of the Karibib Formation the landmass in existence east of the Khan sediments may have subsided or been eroded as the Karibib marbles directly lie on the pre-Damaran rocks at the Witpoortberge. A major landmass further to the northwest appears to have existed as the thick marble bands gradually thin out in the northwest and become interbedded with deeper water argillaceous sediments towards the east, disappearing altogether in the vicinity of the Rabenrücken and Langer Heinrich areas. Subsidence of the ge-anticlinal ridge must have continued until eventually the central part of the area was receiving deeper water sediments (Kuiseb Formation) from the edges of the Damara geosyncline, and a limited amount of volcanic material represented by the Matchless/Gorob orthoamphibolite belt (Finnemore, 1975).

A schematic interpretation of the stratigraphic units during pre-orogenic times is shown in Fig. 3a.

2.2 PETROGRAPHY

The petrography of the metasediments is described with regard to characterising the mineral assemblages that are representative of the different lithologic and stratigraphic units. The assemblages presented account for all the minerals commonly found in the rocks. The stable mineral assemblages in the sense of Winkler (1976) are later considered in relation to their significance in establishing metamorphic grade, the position of isograds and isoreaction-grads, metamorphic-tectonic relationships and pressure-temperature estimates (Chapter 3).
The modal-percentage estimates for the metasedimentary rocks were determined by standard petrographic techniques and visual comparison with the abundance diagrams of Terry and Chilinger (1955).

The terminology used in this study to define the various lithologic units is a compromise between broadly used field terms and the more restrictive classification used in petrographic studies. The term 'quartzite' is used for rocks containing greater than 80% quartz (Winkler 1976, p327), the majority of psammitic rocks however, in this area, contain major amounts of feldspar or minor amounts of mica, and are termed feldspathic and micaceous quartzites respectively. The pelitic or schistose rocks in the area are all characterised by the presence of biotite. However, it is obvious from the subsequent tables that the biotite content in all the samples is lower than 50% which is required to warrant the term 'biotite schist' (Winkler 1976, p326). These rocks which are invariably termed biotite-schists from their field appearance are more correctly described as quartz-biotite schists. Calc-granofels and granofels are used in the context described in section 1.6.1. The term amphibolite is used for rocks containing modal amounts of plagioclase and hornblende together in excess of 60% (95% Winkler 1976, p166), and hornblende gneiss or schist applies to rocks containing these minerals in lesser amounts.

2.2.1 ABBABIS COMPLEX

The petrography of the calc-granofels and amphibolite/hornblende-gneiss lenses occurring within the granites and granite-gneisses of the complex are considered here. The petrography of the granitic rocks is dealt with in a later chapter. The characteristic mineral assemblages are listed in Table 3.
The amphibolites/hornblende gneisses exhibit a foliation due to the sub-parallel alignment of hornblende and biotite grains. These rocks contain quartz, plagioclase, hornblende, biotite, epidote and ore minerals with accessory amounts of retrograde clinozoisite. Clinopyroxene is noticeably absent. The grain size in these rocks is 0.03 - 6mm with hornblende generally forming large dark blue-green to green poikiloblastic subidiomorphic crystals. Plagioclase is also observed as poikiloblastic grains, but is most commonly polygonal in form. Biotite, when present, is in the form of brown, subidioblastic grains, commonly associated with hornblende. Sphene has been observed as small subidioblastic grains occurring as inclusions aligned parallel to the cleavage in hornblende grains. They also occur scattered throughout the sections. It is noticeable that sphene and biotite do not occur in the same rocks, indicating competition for titanium ions between these minerals. Ore minerals are generally xenoblastic in form and are distributed throughout the sections.

The texture of the calc-granofelsen is granoblastic polygonal. These rocks are somewhat inequigranular and grain sizes vary between 0.03mm and 10mm. The general lack of foliation in calc-silicate rock types, although they occur in high grade metamorphic terrains, has been attributed to the low ionic mobility promoted by the presence of calcite (Spry, 1969).

The rocks are characterised by the presence of diopside, epidote, clinozoisite (retrograde), calcite, sphene and occasionally scapolite, garnet and hornblende. Biotite and ore minerals are noticeably absent. Of the feldspars, plagioclase is developed almost exclusively, with alkali feldspar occurring in very few of the assemblages. Quartz and plagioclase occur as xenoblastic or polygonal grains although plagioclase grains are occasionally subidioblastic and are commonly saussuritised.
Diopside is commonly present as colourless or pale green xenoblastic to subidioblastic grains. It is often associated with epidote and hornblende, where this latter mineral is present. These aggregates give a faint gneissic banding to the rocks. Clinozoisite occurs as small xenoblastic ragged grains, between the mineral grains that constitute the fabric. Sphene is present in all the samples investigated, as small idioblastic to subidioblastic grains (larger grains are often embayed) and appears to have concentrated in zones.
2.2.2 THE NOSIB GROUP OF THE DAMARA SUPERGROUP

2.2.2.1 ETUSIS FORMATION

The mineral assemblages found in the rocks of this formation are listed in tables 4, 5 and 6.

The metaconglomerates (Leeukop Member NS2 R) owe their grey-green appearance to the presence of diopside, hornblende, epidote, garnet, biotite and also ore minerals in these rocks. Quartz, plagioclase or alkali feldspar, scapolite, calcite, sphene, allanite, apatite, zircon and chlorite are also present.

The grain size of the minerals in these rocks (excluding clasts) is extremely variable (1 - 10 mm) and the texture is heteroblastic. A gneissic texture is apparent due to the aggregation of the calc-silicate minerals into zones separate from quartz and feldspar.

Quartz occurs as large (4 - 10 mm) irregular xenoblastic grains, although it is also occasionally polygonal. Plagioclase feldspar is invariably subidioblastic in form and usually smaller in size than quartz. Sericitisation is apparent in some of the rocks. Alkali feldspar, when present, occurs in a similar habit to quartz, and is characterised by typical cross-hatch twinning (microcline). Scapolite occurs as subidioblastic crystals and is generally found in association with diopside, epidote and hornblende. Clinoziosite is found as retrograde ragged masses occurring interstitially between diopside and scapolite grains. Epidote also occurs as individual xenoblastic crystals forming part of the fabric and thus possibly represents prograde crystallisation. Diopside is present as colourless xenoblastic grains with rounded grain boundaries. Green pleochroic hornblende occurs as a replacement mineral, replacing diopside along cleavages and fractures. In rocks with higher proportions of hornblende, the grains are large (~ 5 mm) and poikiloblastic, often with inclusions of diopside, and they are arranged in a sub-parallel alignment.

Biotite, where present, is subidioblastic in form and
### TABLE 4

**MODAL ESTIMATES (VOL %) OF CONGLOMERATE (MATRIX)-(NS, R) AND QUARTZ-BIOTITE SCHIST -(NS, GN) FROM THE ETUSIS FORMATION (a=accessory amounts < 1%)**

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<th>BIOTITE</th>
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<th>HORNBLENDE</th>
<th>EPIDOTE</th>
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**META-FELITES (QUARTZ-BIOTITE-SCHIST/GNEISS NS, GN)**

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aligned in a similar subparallel manner. Chlorite has formed as a retrograde phase from the biotite in most sections examined. Sphene is scattered throughout the rocks, though it tends to form in aggregates of small idioblastic to xenoblastic crystals. Ore minerals are xenoblastic, often with very irregular and embayed boundaries, and they tend to be distributed throughout. The clasts present are largely granitic and may constitute 60 - 70% of the rock (Plate 3).

The mineral assemblages present in the matrix of the Leeukop conglomerates are similar to those in the amphibole-pyroxene gneisses of the Khan Formation and can be grouped into two types:

i) Quartz-plagioclase-diopside-hornblende-epidote + scapolite, calcite, sphene and ore.

ii) Quartz-alkali feldspar-biotite-hornblende + epidote, sphene, ore.

The metapelites (quartz-biotite schist/gneiss, migmatite) - Ns$_1$Gn, occurring near the base of the Etusis Formation have a simple mineralogy.

The most common minerals are quartz, plagioclase, alkali-feldspar, biotite, muscovite, chlorite and zircon. The texture of the rocks is lepidoblastic for the varieties with abundant biotite, but granoblastic or gneissic in others where feldspar and quartz form distinct leucosome layers. The grain size of the schist is small (maximum 2mm), with coarser crystals occurring in the gneisses (maximum 4mm).

Quartz occurs as xenoblastic grains with curved and embayed boundaries. The crystals generally show some degree of fracturing and strain. Plagioclase feldspar, when present, occurs as multiply-twinned subidioblastic to xenoblastic grains, and often contains products of sericitisation along cleavage and composition planes. Alkali-feldspar, predominantly microcline, exhibiting
### TABLE 5

**MODAL ESTIMATES (VOL%) OF METAPSAMMITES (QUARTZITE, FELDSPATHIC QUARTZITE) FROM THE ETUSIS FORMATION EAST OF THE TUMAS RIVER INLIER**

(a = accessory amounts < 1%)

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crosshatched twinning, occurs as xenoblastic grains. Perthitic intergrowths are often developed. Biotite is present as subparallel subidioblastic flakes in places flattened around large grains of quartz and feldspar. Chlorite is developed as an alteration product of the biotite. Muscovite coexists with biotite and usually occurs as ragged unoriented flakes. Xenoblastic zircon grains have distinctive pleochroic halos when the grains are present as inclusions in biotite flakes. Sillimanite (present as fine needlelike inclusions in quartz grains), diopside, sphene, hornblende, calcite, apatite and ore minerals are present in small quantities in some of the rocks.

The metapsammites (quartzite, feldspathic quartzite, micaceous quartzite)-Ns_{1,0} of the Etusis Formation have a simple mineralogy and consist predominantly of quartz, feldspar, biotite and muscovite. The quartz content is extremely variable (30 - 90%). The texture of the rocks is granoblastic-polygonal, except where a faint fabric is apparent due to flattening of the grains.

The characteristic mineral assemblages are listed in Tables 5 & 6. The rocks have arbitrarily been subdivided into two groups (east and west of the Tumas River Inlier) as there are differences apparent in their mineral contents.

The quartz grains are usually equant, rounded to polygonal in shape, with curved and scalloped boundaries and occasionally exhibiting triple point intersections. The grains are often fractured, flattened and strained and contain inclusions of other minerals. These rocks are generally feldspathic and micaceous in nature, with alkali-feldspar and plagioclase occurring in most of the samples investigated. The plagioclase feldspar present is commonly untwinned albite that has undergone sericitisation. Quartz and feldspar are often
observed as recrystallised relic clasts. Muscovite is present in all of the quartzites east of the Tumas River Inlier, with the exception of the hornblende-rich

**TABLE 6**

Modal estimates (vol %) of metapsammites (quartzites, feldspathic quartzites, micaceous quartzites) from the Etohis formation west of the Tumas River Inlier (a=accessory amounts <1%)

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varieties at the Rabenrücken. It usually occurs as subidioblastic/xenoblastic ragged flakes distributed throughout the rocks in an unorientated (retrograde) manner. It also occurs as prograde flakes present in a subparallel-parallel orientation, in some of the samples. Biotite occurs as brown subidioblastic/idioblastic flakes, usually arranged in subparallel orientation, and often shows retrograde alteration to chlorite. Sillimanite is present as large (maximum 8mm) fibrous aggregates that can be seen in thin section (Plate 12) and outcrop (Plate 13). Small needles also occur as inclusions in quartz and muscovite. Hornblende is a major constituent of the impure quartzites at the Rabenrücken and occurs as pale green xenoblastic grains scattered throughout the sections. Additional minerals, present in minor amounts in the quartzites are zircon, epidote, calcite, tourmaline and apatite.

The quartzites west of the Tumás River Inlier have a noticeably simpler mineralogy than those to the east. They are composed predominantly of quartz and feldspar with significantly low contents of biotite and muscovite; sillimanite is absent from the assemblages. The proportion of ore minerals in the rocks is conspicuously higher than that of the quartzites to the east.

2.2.2.2 KHAN FORMATION

The predominant rock types in this suite are banded amphibole-pyroxene gneisses, quartzites, quartz-biotite schists/gneisses and amphibolites.

Mineral assemblages are listed in Table 7. The most common minerals occurring in this suite are quartz, plagioclase, K-feldspar, diopside, hornblende, epidote, calcite, sphene and ore. Wollastonite scapolite, garnet and biotite are present in some of the specimens.
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The texture of these rocks is generally granoblastic polygonal although a gneissic banding is obvious from the separation of the calc-silicate minerals (diopside, hornblende, epidote, scapolite and wollastonite) and quartz and feldspar into alternate zones.

The rocks have a range in grain size from 0.2 mm to about 2 mm, the larger grains being wollastonite and hornblende. Wollastonite has been observed in thin sections as large poikiloblastic grains up to 20 mm in size. Quartz occurs as rounded to polygonal equidimensional grains, often with embayed boundaries and occasional triple point intersections. The rocks tend to have either a predominance of plagioclase or of alkali feldspar, which have a texture similar to quartz. Colourless, pale green diopside occurs as subidioblastic/xenoblastic grains. Scapolite is present as xenoblastic to polygonal grains, commonly associated with diopside. Green, strongly pleochroic hornblende occurs associated with diopside as a retrograde product forming along the cleavages of the diopside grains and often forms poikiloblastic grains enclosing diopside. Epidote occurs as yellow pleochroic grains, with subidioblastic prograde form or as colourless clinozoisite in ragged xenoblastic retrograde grains occurring interstitially between diopside, hornblende and scapolite. Biotite occurs as brown subidioblastic flakes aligned subparallel in zones and commonly altered to chlorite. Garnet is present as large poikiloblastic grains. Wollastonite is present in some samples as elongate xenoblastic twinned grains and calcite occurs as twinned xenoblastic crystals. Sphene has been observed as xenoblastic to idioblastic grains scattered throughout the rocks or occurring in aggregates. It is present in aggregates developed along cleavage planes in hornblende in the amphibolite developed at the top of the formation in the west. Here the hornblende forms large poikiloblastic grains enclosing
inclusions of diopside. Grains of opaque ore are deeply embayed and poikiloblastic and are often associated with sphene.

In the biotite schists at the top of the sequence on the eastern side of the Zebraberge, the mineralogy is simple, consisting mainly of quartz and biotite with abundant ore grains.

2.2.3. THE SWAKOP GROUP OF THE DAMARA SUPERGROUP

2.2.3.1 CHUOS FORMATION

The mineral assemblages are listed in Table 8. The matrix of the Chuos tillite is largely pelitic and is characterised by a very fine grained (generally less than 1mm) assemblage of quartz, biotite and muscovite. Feldspars are present in low proportions.

The texture is granoblastic with subidioblastic to idioblastic biotite and muscovite grains, usually poorly aligned. In sample X150 there is an alignment of biotite and muscovite and the biotite flakes are deformed around clasts in the rock. Muscovite occurs in a similar habit to biotite in the rocks. However it is also present as large subidioblastic unorientated flakes, representing retrograde post-deformational growth. Quartz is present as xenoblastic rounded to irregular grains that are occasionally flattened parallel to biotite (sample X150).

The clasts present in the thin sections studied are aggregates of quartz, sericitised plagioclase, and hornblende. Zircon and apatite are common accessory minerals in these rocks.

2.2.3.2 KARIBIB FORMATION

The mineral assemblages in the marbles are listed in Table 8. The dominant minerals in these rocks are calcite and/or dolomite which constitute between 60 and 90% (Volume percent) of the thin sections studied. No attempt was made to distinguish between calcite
TABLE 8

MODAL ESTIMATES (VOL %) OF METASEDIGENTS FROM SWAKOP GROUP

(a = accessory amounts <1%)

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and dolomite as Jacob (1974) describes mineral assemblages in the marbles occurring immediately to the north of this area and reports that the relative concentration of these minerals is extremely variable with compositions ranging from pure dolomite to pure calcite rich marbles.

The textures are granoblastic to granoblastic-polygonal with the grains usually showing some degree of flattening and in some places deformation of twin lamellae is obvious. The grain size in the thin sections investigated varies from less than 0.3mm to 10mm.

Olivine (forsterite), altered to serpentine, phlogopite, quartz, ore minerals and apatite occur in the marbles, occasionally in significant amounts.

2.2.3.3. KUISEB FORMATION

The rocks of this suite are essentially metapelites with interbedded calc-granofels bands particularly in the east.

The WITPOORT MEMBER rocks are quartz-biotite schists and migmatites, and modal estimates are presented in Table 8.

The schists of the Witpoort Member are composed almost exclusively of quartz, plagioclase, alkali feldspar and biotite with accessory zircon.

These rocks are inequigranular and generally 0.5mm - 2.5mm in grain size. Where porphyroblasts of garnet, cordierite and andalusite are present, these often exceed 8mm in size. The texture is granoblastic to lepidoblastic, depending on the proportion and alignment of biotite in the rocks. The biotite flakes are subidioblastic to idioblastic and range from unorientated to strongly aligned. The biotite flakes are altered to chlorite in some of the specimens. Muscovite
is not common in these rocks.

Garnet is present in two of the specimens investigated and it occurs as rounded to highly irregular xenoblastic, poikiloblastic grains. In specimen 165 it is associated with cordierite, and biotite flakes are conspicuously flattened around the garnet, implying pre-tectonic growth of the garnet. In specimen X142, collected from the opposite limb of the syncline, the garnets are of extremely irregular shape and contain inclusions of biotite that are strongly aligned parallel to the biotite flakes in the matrix, and this appears to indicate syn-tectonic growth of the garnets.

Quartz and feldspar occur as rounded to irregular xenoblastic grains. Plagioclase is generally poly-synthetically twinned and sericitised. Alkali-feldspar grains commonly contain inclusions of quartz, plagioclase (perthite) and biotite. In several specimens, aligned biotite inclusions in the alkali feldspar grains have an orientation that is invariably not parallel to biotite flakes in the remainder of the rock (Plate 14). This suggests growth of the alkali feldspar grains after an initial deformation event, with a further foliation imprinted in the rocks post-dating the alkali-feldspar growth.

Apatite, allanite and zircon are common accessory minerals in most of the rocks, with zircon surrounded by pleochroic haloes when present as inclusions in biotite flakes.

The TINKAS MEMBER calc-granofelses are characterised by the mineral assemblages quartz, plagioclase, hornblende, calcite, epidote and diopside. Alkali-feldspar, biotite, garnet and sphene also occur in significant amounts in some of the specimens, see Table 9. The grain size of the calc-granofelses is usually less than 0.3mm. However, porphyroblasts up to approximately
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10mm of either garnet, hornblende, diopside or epidote do occur. The texture of the rocks is granoblastic to porphyroblastic. Quartz is present as rounded to polygonal grains and is conspicuously more abundant than the feldspar minerals.

The porphyroblasts of garnet, hornblende, diopside and epidote are usually very ragged in form and deeply embayed. Hornblende is commonly poikiloblastic and is commonly concentrated at the contact of the calc-granofels adjacent to the interbedded metapelite bands.

Unoriented idioblastic grains of hornblende representing post-deformational growth are present in specimen X125 (Plate 15).

The quartz-biotite schists and granofelses constituting the remainder of the suite are composed predominantly of quartz, plagioclase feldspar and biotite. Cordierite, andalusite, muscovite, chlorite, ore minerals and zircon are also present in minor amounts.

The texture of these rocks is granoblastic where biotite is poorly aligned or lepidoblastic where subparallel subidioblastic/idioblastic grains of biotite are abundant. In specimen 19211, transposition of biotite grains formed during an earlier deformation has occurred (Plate 16). Muscovite exhibits varying degrees of alignment, occurring both parallel to biotite flakes and also as larger ragged unoriented flakes.

The grain size is fine grained (maximum 1mm) particularly in comparison to the metapelites of the Witpoort Member. Quartz and plagioclase occur as rounded, occasionally polygonal, xenoblastic grains. Alkali feldspar is very scarce in these rocks and in several of the rocks investigated, feldspar minerals are entirely absent, or present in minor amounts only.
Cordierite and andalusite are present as conspicuous porphyroblasts, commonly up to 5mm in size. The andalusite porphyroblasts are extremely embayed and are replaced largely by quartz and muscovite, leaving a skeletal andalusite grain. The porphyroblasts have been observed elongated parallel to the foliation which is indicative of syntectonic growth (Plate 17).

Cordierite occurs as oval-shaped porphyroblasts, often altered at the margins to pinite and commonly containing abundant inclusions of quartz and biotite (Plate 18). The biotite inclusions are generally poorly orientated and smaller in size in comparison to the matrix biotites around the porphyroblasts (Plate 19), indicating syntectonic or late-tectonic growth of cordierite (parallel to $F_2$ axial planar foliation).
3. METAMORPHISM

3.1 GENERAL

The stable mineral assemblages of the metamorphic rocks in this area are interpreted according to the concepts discussed by Winkler (1976), in order to establish metamorphic grade, position of isograds and isoreactiongrads, and the physical conditions (pressure and temperature) of metamorphism. Winkler subdivides levels of metamorphism into the very broad grouping of very low-, low-, medium-, and high-grade (Fig 4a), which are broadly similar to the major facies groups, Turner (1968), see Fig 4b.

Fig 4a
Grades of metamorphism
(after Winkler 1976)

Fig 4b
Major facies groups
(after Turner 1968)
The metamorphic facies and sub-facies concepts (Turner and Verhoogen (1960), Miyashiro (1961), Winkler (1967) and Turner (1968)) that have been used extensively in the evaluation of metamorphic belts have become rather unwieldy, as discussed by Winkler (1976). The use of the sub-facies subdivision is no longer preferred by Turner (1968) and Den Tex (1971), owing to the variety and number of different sub-facies possible.

The emphasis in defining metamorphic grade in the sense of Winkler (1976), is on the changes in mineral assemblages and not on the characteristic mineral parageneses that occur at different levels of metamorphism, as in the case with the facies concept. The number of variations of mineral assemblages possible in rocks which is dependent on different rock compositions is almost infinite. Within each metamorphic grade, particular mineral reactions and their resultant parageneses characterise certain metamorphic zones, from which accurate PT estimates can be made. Only certain rock types contain assemblages of petrogenetic importance.

The mineral reactions that produce the changes in the mineral parageneses can be delineated in the field and are termed 'isoreactiongrads', Winkler (op.cit.). The term 'isograd', introduced by Tilley (1924) is still used to define the appearance or disappearance of a particular index mineral in the field.

The majority of rock units in the field exhibit simple mineralogies and therefore specific rock types containing significant indicator minerals and mineral assemblages were located. The most sensitive and useful rocks for establishing PT conditions are the metapelites. Mineral parageneses containing cordierite, andalusite, sillimanite, muscovite, garnet and quartz in the metapelites and metapsammites in this area enable conclusions to be made as to the type and grade of metamorphism.
Prior to any investigation into the significance of different mineral parageneses in the interpretation of metamorphic grades, consideration must be made as to whether equilibrium conditions existed or not, as only equilibrium mineral assemblages are significant. In the Khan/Swakop area to the north, Jacob (1974) analysed coexisting cordierite and biotite for FeO, MgO, Na₂O and MnO, and found when plotting the concentration of Fe/Fe + Mg, Na₂O and Mn in these minerals regular trends resulted. These trends, resulting from partitioning of the elements into coexisting minerals, indicate that equilibrium existed in the metapelites (Saxena 1968, Gorbatschev 1968), particularly in regard to cordierite and biotite and thus it has been assumed to have been attained in the rocks as a whole.

The composition of the fluid phase present during metamorphism has significant effects on the stability fields and equilibrium temperatures of metamorphic reactions. For graphite-free pelitic rocks the condition $P_T = P_f = P_{H_2O}$ is assumed, as water bound up in hydrous minerals and in pores in the rock is the only significant constituent of the fluid phase.

The condition $P_T > P_f = P_{H_2O}$ which can be expected if rocks have been subjected to more than one phase of metamorphism, has the effect of lowering equilibrium temperatures of dehydration reactions. There is little experimental data, however, enabling quantitative conclusions to be made.

With carbonate-, iron-, and graphite-bearing mineral assemblages however, CH₄, HCl, CO₂, O₂, H₂ etc., released during metamorphism contribute to the fluid phase and these have varying effects on the equilibrium temperature of reactions.

Increasing mole fractions of CO₂ ($X_{CO_2}$) have been shown to increase equilibrium temperatures in carbonate-bearing rocks, (Greenwood, 1967; Winkler, 1976) and these rocks due to their bivariant nature ($H_2O - CO_2$) are not particularly good for PT estimates.
Using information from other reactions in pelitic rocks however, approximate $X\text{CO}_2$ or $X\text{H}_2\text{O}$ values can be established and PT conditions can be estimated from carbonate bearing assemblages.

### 3.2 PELITIC AND PSAMMITIC ROCKS

The metamorphic grade increases from the east to the west in this area for, in the extreme east, there is no trace of anatexis having occurred in the rocks, but migmatites and granites are common in the central and western part. The metapelites in the east contain biotite, andalusite and cordierite which have significance in PT estimates.

The presence of staurolite, although occurring in the area to the northeast (Jacob 1974), and also to the southeast, was not detected in this area. This is possibly due either to unfavourable bulk compositions of the rocks or to the disappearance of staurolite from the assemblages further to the east of the area studied. According to Winkler (1976, p218), the presence of an earlier Mg-rich chlorite in the rocks forms cordierite instead of staurolite due to reaction 1.

$$\text{chlorite} + \text{muscovite} + \text{quartz} = \text{cordierite} + \text{biotite} + \text{Al}_2\text{SiO}_5 + \text{H}_2\text{O}$$

The latter possibility whereby staurolite has already disappeared from the assemblages may be correct, in that staurolite was only found as inclusions in garnet and cordierite grains in the area to the northeast (Jacob 1974) and that Göttingen workers place the 'staurolite-out' isograd considerably further to the east than this part of the orogen, Porada (1976).

Andalusite occurs sporadically, usually coexisting with cordierite and its disappearance from mineral assemblages coincides with the easternmost appearance of sillimanite, in the metapsammites and metapelites of Glück's Dome. Andalusite may have been formed at
an early stage from staurolite by reaction (2).

\[
\text{staurolite + muscovite + quartz} = \text{biotite + almandine + andalusite + } H_2O
\]  

\[\text{-(2)}\]

The disappearance of andalusite from the mineral assemblages appears to be due to either inversion (reaction 3) or to reaction 4.

\[
\text{andalusite} = \text{sillimanite}
\]  

\[\text{-(3)}\]

\[
\text{biotite + andalusite + quartz} = \text{Cordierite + K-felds + Ca-rich plagioclase + garnet + } H_2O
\]  

\[\text{-(4)}\]

The coexistence of andalusite with cordierite implies that reaction (4) may have been involved, as relict inclusions of andalusite appear to be present in some of the cordierite porphyroblasts (Plate 20). Other products of this reaction, K-feldspar and garnet are either present in accessory amounts or absent in the sections studied.

Experimental derivation of the phase boundaries between the \(\text{Al}_2\text{SiO}_5\) polymorphs (Althaus 1967, 1969; Richardson et al. 1968, 1969) enables these minerals to be used as PT indicators, provided that there is no metastable coexistence outside their normal stability fields (Chinner, 1966; Winkler 1976, p.93). Andalusite and sillimanite were not observed coexisting in the samples investigated in this study and there was no overlap in their respective field occurrences. This suggests that there has been no metastable behaviour in this area although it was found to happen to the north (Jacob, 1974), and is evidence for reaction 3 having occurred.

An andalusite-out/sillimanite-in isograd can be drawn along the western boundary of the Glucks Dome (Fig 5). The position of this isograd coincides with the westernmost occurrence of andalusite in the area to the north (Jacob 1974, p130). The andalusite/sillimanite isoreaction—
Fig. 5

- isograd/isoreaction-grad
- Salem granitoid suite
- Husab and Abbabis granite-gneisses
- Gawib granitoid suite
- Swakop group metasediments
- Nosib group metasediments

The position of isograds and isoreaction-grads in the Tumas River Area.
grad in the area to the north was placed 12km further to the east Jacob (op. cit.), due to the metastable occurrence of andalusite in that area.

The disappearance of andalusite from the mineral assemblages and the presence of sillimanite prior to the start of anatexis in the field, is more in agreement with the experimentally derived position of the andalusite/sillimanite phase boundary by Althaus (1967, 1969) and Weill (1966) than that of Richardson et al., (1968, 1969). The reaction curve of Althaus (1969) is shown in fig 6.

Cordierite is found in the metapelites in the east (Tinkas Member) occurring in narrow bands. The presence of cordierite or cordierite + almandine garnet in the rocks is significant in that cordierite + almandine garnet is only stable within a restricted PT range (Winkler 1976, p230), see Fig 6.

The presence of garnet, however, is also controlled by the whole rock chemistry and Wynne-Edwards and Hay (1963) report that the rock composition is as important a factor as pressure in determining whether cordierite and/or garnet will form in pelitic metasediments. The plots of Jacob (1974, p. 105, 107) show that some of the rocks from the Khan/Swakop area do have suitable whole rock chemistry for the formation of both cordierite and garnet. Therefore assuming favourable rock compositions, the assemblages in the east, that contain cordierite but no garnet, are possibly significant in that they appear to have formed at pressures less than 5kb, at a temperature of approximately 600°C (below curve iv and anatexis boundary in Fig. 6)

Sillimanite is present, as mentioned previously, in the metapelite and metapsammite assemblages at Glucks Dome and it also occurs in similar lithotypes on the western flanks of the Chameleon Hills (Plates 12 & 13). Muscovite and quartz are stable in these assemblages, with prograde muscovite obvious from its strongly aligned manner, distinct from the retrograde, ragged
PT diagram compiled from experimentally determined equilibrium curves for reactions referred to in the text.

Curve  
(i) Kya/And, Kya/Sill determined from an average of data of Richardson et al. (1969) and Althaus (1969).
(iii) Upper limit of stability of cord+gnt (Almandine) in rocks with FeO/(MgO+FeO) ratios of 0.4, Currie (1971).
(iv) Lower limit of stability of cord+gnt (almandine) in rocks with FeO/(MgO+FeO) ratios of 0.8, Currie (1971).
(viii) Bio+sill+Qz = cord+K.Feld+H2O (present investigation)
(x) Gnt-cord stability field, Hirschberg and Winkler (1968).

See page 60.
unorientated flakes.

The transition from medium-grade to high grade metamorphism is characterised by the breakdown of muscovite in the presence of quartz (reaction 5) and the appearance of migmatites at pressures greater than 3kb, i.e. the beginning of anatexis (Winkler 1976, p82).

\[
\text{muscovite + quartz = K-feldspar + sillimanite + H}_2\text{O}
\]

The position of this 'anatexis-in-gneiss' boundary in the field is well defined and occurs along the eastern edge of the Tumas River Inlier (Fig 5). On the high temperature side of the boundary the migmatitic Leeukop metaconglomerates are present on the western flanks of the Chameleon Hills. The boundary passes to the southeast of Leeukop where the easternmost outcrops of the Salem granitoid suite are located.

The majority of the metapelites west of the 'anatexis-in-gneiss' boundary in the high grade metamorphic region exhibit migmatisation effects. The Kuiseb Formation metapelites have largely been transformed into granite-gneisses of the Salem granitoid suite. Biotite and quartz-biotite schists, preserved on the limbs of the synclines contain the assemblage cordierite-garnet-K-feldspar-plagioclase-quartz. This coexistence of garnet (almandine) and cordierite as mentioned earlier is stable over a restricted PT range (Hensen and Green, 1971, Currie 1971 and Hirschberg & Winkler 1968) See Fig 6. The coexistence of the minerals has been attributed to the following two reactions (6 & 7) Hensen & Green (1971) and Currie (1971).

\[
\text{cordierite} = \text{almandine + sillimanite + quartz} - (6)
\]

\[
\text{cordierite + biotite} = \text{almandine + K-feldspar + H}_2\text{O} - (7)
\]

As the minerals cordierite and garnet are present (samples 165 + 905) in the metapelites beyond the
anatexis boundary in this area, minimum temperature and pressure values, using data of Currie (1971) and assuming a FeO/(FeO+MgO) ratio in the rocks to be 0.8, can be placed at 650°C @ 4.5kb to 800°C @ 3.5kb (see fig 6). No analyses of metapelitic rocks were undertaken during this study but data from Jacob (1974) indicates that the FeO/(FeO+MgO) ratios of the metapelitic rocks in the area to the north were approximately 0.6, thus if metapelitic rock compositions are similar in this area, PT conditions that existed could be expected to be higher than those given above.

The coexistence of quartz, biotite, sillimanite, cordierite and K-feldspar has been attributed to reaction 8; Hess (1969), Schreyer and Seifert (1969), Hoffer (1976).

\[
\text{biotite} + \text{sillimanite} + \text{quartz} = \text{cordierite} + \text{K-feldspar} + \text{H}_2\text{O} \quad (8)
\]

The mineral parageneses present in the metapelites beyond the anatexis boundary (Samples 165 + 905) (Plate 21) contain biotite and quartz as inclusions in cordierite and K-feldspar porphyroblasts which appear to have been reactants in the above reaction (8). Sillimanite was not observed but this is probably due to its having been used up in the reaction.

The position of this reaction curve within the PT field is still inconclusively known. Schreyer and Seifert (1968) in studying the pure Mg-system place it in the position as shown in Fig 6, curve vi).

Hess (1969) considers that with increasing Fe content this reaction will occur at lower pressures, although exact values are not known. Experimental work by Hoffer (1976) places this reaction curve at considerably lower temperatures (curve vii fig 6) but it is difficult to reconcile the data of Hoffer (op. cit.) with evidence from mineral parageneses present in this area.
for the following reasons:-

i) The assemblage biotite + quartz + cordierite + K-feldspar (+ sillimanite?) is present in this area 8 km inside the anatexis boundary.

ii) Garnet and cordierite coexist in the above assemblages suggesting that the reaction should lie within the garnet-cordierite stability fields delineated by Hensen and Green (1971), Currie (1971) or Hirschberg and Winkler (1968) and these lie above the position of the reaction curve defined by Hoffer (1976).

It is necessary therefore to disregard the data of Hoffer (1976) and assume the reaction curve lies in a similar position as placed by Jacob (1974), below the reaction curve of Schreyer and Seifert (1969), (viii in Fig 6).

The mineral assemblages of the metapelites and meta-psammites from the Nosib metasediments west of the anatexis boundary do not contain minerals of any significance for metamorphic evaluation. Migmatitic development in the metapelites and banded gneisses (Plate 7), is significant however in indicating that pressures and temperatures exceeded values required for anatexis to begin.

Adjacent to the Leeukop metaconglomerates, immediately to the northeast of Leeukop, sillimanite is developed in the granite-gneisses as fibrolitic masses. These rocks may have been partly dehydrated owing to the pre-existence of granitic and gneissic rocks, developed during an earlier metamorphic episode, (Abbas Complex) and this sillimanite possibly formed either during this early phase or during the Damara metamorphic period.

3.3 CALC-GRANOFELSES AND AMPHIBOLITES

The mineral assemblages of the calc-granofels and amphibolites of the Abbas Complex appear similar to those in corresponding rock types in the Damara metasediments. These rocks apparently have been
little effected by the Damara metamorphism as it could have been expected that they would contain high temperature mineral assemblages, due to the dehydration and consequent lowering of equilibrium temperatures by the earlier Abbabis metamorphism. The metasedimentary remnants in the Abbabis Complex were not exhaustively sampled and thus the presence of high-temperature mineral assemblages may have been overlooked.

The carbonate-bearing assemblages of the Tinkas Member calc-granofelses contain quartz, plagioclase garnet (grandite), epidote and calcite. Jacob (1974, p.69) determined the refractive indices and cell edge lengths of garnets and epidotes from the south-east of the Khan/Swakop area and established that the compositions of garnets in the Tinkas Member were that of grandite \((Gr_{80}And_{20})\) and that the epidote was \(Ps_{10-17}\).

Evidence for the stability of grossularite, calcite and quartz over a very restricted PT range is given by Storre and Nitsch (1972) for pressures of 2kb. The presence of the assemblages is also indicative of low \(XCO_2\).

The stability of grossularite, anorthite, calcite and quartz extend over a wide temperature range \((400 - 600^\circ C\) at 2kb) and low \((< 0.2) XCO_2\) values. These minerals can coexist due to reaction 9 (Winkler 1976),

\[
grossularite + CO_2 = anorthite + calcite + quartz
- (9)
\]

Epidote, although present, does not constitute a significant proportion of the rocks, which may be due to reactions 10 and 11, having proceeded to the right. These reactions were considered by Jacob (1974) to have formed grossularite in the calc-granofelses in the area to the north.

\[
zoisite + CO_2 = anorthite + calcite + H_2O
- (10)
\]

\[
zoisite + quartz = anorthite + grossularite + H_2O
- (11).
\]
The coexistence of grandite, plagioclase and quartz may be significant in that Winkler (1976, p.144) states that the paragenesis grossularite + anorthite + quartz is stable at 500° - 600°C (2kb) and at very low XCO₂ (0.02 - 0.2) at 2kb. The stability of this paragenesis is unknown for higher pressures (4 - 5kb).

Diopside, which is occasionally present in the mineral assemblages of the calc-granofelses in the east is reported to form at temperatures below 600°C if XCO₂ < 0.1, by reaction 12, Metz (1970).

1 tremolite + 3 calcite + 2 quartz = 5 diopside + 3CO₂ + 1H₂O
- (12)

Wollastonite, although not observed in the eastern mineral assemblages has been reported by Jacob (1974) to occur intermittently in the area to the north. Wollastonite forms at temperatures above 600°C at 2kb only at very low XCO₂ (< 0.1). This is further evidence for very low XCO₂ conditions, (probably due to high water contents derived from the adjacent inter-bedded metapelites.)

The banded gneisses of the Khan Formation in the west, in general provide little information on PT conditions. Prograde epidote, coexisting with quartz, is indicative of high Fo₂ conditions existing in the banded gneisses. Liou (1973) has shown that epidote is stable up to 750°C at Pf = 5kb under conditions of high Fo₂.

Mineral assemblages in the banded gneisses at the Welleberge and north of Klein Tubasberg contain wollastonite, plagioclase, grandite, calcite and quartz (Plate 22). These minerals represent an invariant isobaric paragenesis. Wollastonite can form by reaction 13 (Winkler 1976), or reaction 14 (Boettcher 1970).

grossularite + CO₂ = anorthite + calcite + wollastonite
- (13)
1 grossularite +1 quartz = anorthite +2 wollastonite 
- (14)

Reaction 14 is possibly responsible for the formation of the wollastonite and anorthite intergrowths in the banded gneisses, where $X_{CO_2}$ is probably very low. The PT limits of this reaction (Boettcher, 1970) from the system \( \text{CaO - A}_2\text{O}_3 - \text{SiO}_2 - \text{H}_2\text{O} \), are shown in Fig 6, curve ix.

The Karibib Formation marbles were not extensively sampled but the assemblages west of the anatexis boundary contain olivine (forsterite) whereas the samples from east of the anatexis boundary contain quartz and no forsterite. The development of forsterite has been attributed to reactions 15 and 16, after Metz (1970).

\[
1 \text{ tremolite} + 1 \text{ dolomite} = 8 \text{ forsterite} + 13 \text{ calcite} + 9 \text{ CO}_2 + 1 \text{ H}_2\text{O} 
- (15)
\]

\[
1 \text{ diopside} + 3 \text{ dolomite} = 2 \text{ forsterite} + 4 \text{ calcite} + 2 \text{ CO}_2 
- (16)
\]

The equilibrium temperatures for both of these reactions are very similar for a given fluid pressure. As neither tremolite nor diopside was observed in the forsterite-bearing assemblages both reactions appear to have proceeded to the right. The occurrence of these minerals together is restricted to an univariant equilibrium curve in \( P - T - X_{CO_2} \) diagrams (Puhan, 1977).

3.4 DISCUSSION

Pressure and temperature conditions that existed during metamorphism can be estimated by the appearance or disappearance of particular minerals or mineral assemblages throughout the area. These PT estimates are made by considering the position of the experimentally determined stability fields for mineral reactions within Fig. 6.
In the metapelites in the east, the absence of staurolite, the presence of andalusite without sillimanite and the coexistence of the stable mineral pair, muscovite and quartz, is indicative of minimum temperatures of 600°C and pressures of 3.5 - 4 kb (see A in fig 6). The assemblages grossularite, quartz, anorthite and calcite with only minor epidote, in the calc-granofelites and the apparent lack of forsterite in the marbles east of Gücks Dome is in agreement with this conclusion. In the metapelites in the syncline on the western side of the Tumas River Inlier, garnet (almandine) and cordierite coexist, and forsterite is present in marbles of the Karibib Formation. Considering that wollastonite and anorthite are present in the banded gneisses considerably further west in a higher temperature zone apparent from the abundance of anatectic granite-gneisses (Husab granite gneiss suite), the above mineral assemblages are indicative of a minimum temperature of 675°C at pressures of 4 - 4.5 kb, adjacent to the western boundary of the Tumas River Inlier. (See B fig 6). The assemblage biotite, cordierite, K-feldspar, garnet (+ sillimanite) is also present in this area but it is difficult to reconcile the data of Hoffer (1976), reaction vii, fig 6, with this and the inferred position (reaction viii, fig 6) would be in agreement with data from other reactions in this area. Adjacent to the Welleberge and north of the Tubasberg, the banded gneisses contain anorthite and wollastonite which suggests that a temperature of 720°C and pressures of 4.5 - 5 kb must have been exceeded in this part of the area (see C fig 6). The absence of wollastonite in the banded gneisses of the Zebra-berge enables a 'wollastonite-in' isograd to be tentatively placed in the position shown in fig 5.

The data from the mineral assemblages in this area therefore indicates increasing pressure and temperature conditions during the peak of metamorphism from a
minimum if approximately 600°C at 3,5 - 4kb, in the east to approximately 720°C at 4,5 - 5kb in the west. These values are similar to those obtained by Jacob (1974) for the adjacent Khan/Swakop area, but with lower pressures in the east apparent from the data presented in this study. This is due to the acceptance of the position of the andalusite/sillimanite reaction curve of Althaus (1969) as suggested by Winkler (1976, p92), rather than that of Richardson et al., (1969) as used by Jacob (1974) and also the restriction of the PT field by the absence of staurolite in the assemblages of this area. It is difficult to reconcile the pressures of 6 - 8kb as suggested by Nash (1971) for the SJ area with the assemblages present in the western part of the area, as exceptionally high temperatures would be required at these pressures for the wollastonite and anorthite to coexist.

Temperature and pressure estimates by Göttingen workers are very low in relation to the current results. Temperature determinations by Puhan (1976) using the dolomite-calcite solvus geothermometer for unmixed magnesian calcite crystals on samples collected from the marbles of the Karibib Formation yielded a temperature of 620°C. This value considered with regard to the experimentally determined stability field for the assemblage forsterite, diopside, tremolite, dolomite, calcite gives a pressure of 2,8 ± 0,3kb (Puhan 1977). The position of this sample in the field lies 30km inside the anatexis-in-gneiss boundary defined by Jacob (1974) for the Khan/Swakop area, and as the PT values of Puhan (1976) lie below the anatexis field, they would appear to be incorrect.
4. GRANITOIDS

4.1 FIELD RELATIONSHIPS AND CHARACTERISTICS

On the basis of field characteristics and relationships, five different suites of granitoid rocks can be distinguished. These are presented schematically in Fig 3b, and comprise:

PRE-DAMARA (BASEMENT) GRANITE-GNEISSES
  ABBABIS COMPLEX (AGn1)

DAMARA SYN-TECTONIC GRANITE-GNEISSES AND GRANITOIDS
  HUSAB GRANITE-GNEISS SUITE (Gn1)
  SALEM GRANITOID SUITE (Gn2)

DAMARA LATE (POST)-TECTONIC GRANITOIDS
  GAWIB GRANITOID SUITE (G3)
  ROSSING ALASKITIC GRANITE (G4)

4.1.1 ABBABIS COMPLEX (AGn1)

These rocks outcrop extensively immediately east of 15°00' longitude in the core of a large anticlinal/domal structure, the Tumas River Inlier, and also on the eastern flanks of the Rabenrücke Hills in the extreme east of the area. They are overlain unconformably by metasediments of the Nosib Group.

The principal rock types are augen-gneisses (Plate 23) which constitute approximately eighty percent of the suite. Similar augen-gneisses occurring within the Abbabis Inlier (Type area) have been reported by Gevers (1931), and Smith (1965). The augen-gneisses are red to grey in appearance and contain abundant conspicuous feldspar augen, 1 to 4cm in size. The rocks are strongly foliated, defined by alignment of the augen, and biotite rich layers. This foliation (As1) has been modified in places by later Damara tectonic events (refer to Structure section 5.4.1.)

In addition to the augen-gneisses, sillimanite-rich red biotite granite-gneisses occur in the south-eastern part of the Tumas River Inlier. These rocks
are unconformably overlain by the Etusis basal conglomerates (Leeukop member) and therefore are also pre-Damaran in age.

The Leeukop conglomerates flank the Tumas River Inlier and contain clasts of granites and gneisses (Plate 4) which form the underlying basement rocks. This is conclusive evidence that the adjacent granite-gneisses of the Tumas River Inlier are of pre-Damaran age.

There is a small occurrence of augen-gneisses present on the western flanks of the Witpoortberge, unconformably overlain by marble of the Karibib Formation. These rocks have been grouped into the Abbabis Complex but, further to the west, rejuvenation during the Damara orogeny has largely obliterated the original character of the rocks, which are therefore grouped into the Husab granite-gneiss suite.

Metasedimentary xenoliths of probable pre-Damaran age occur within the granite-gneisses parallel with the As₁ fabric. These rocks are described in section 2.1.1.

4.1.2 HUSAB GRANITE-GNEISS SUITE (Gn1)

This suite of rocks is the most heterogeneous, and is composed of red granites and gneisses, foliated granites, leucogranites, migmatites and pegmatites with a complex association of metasedimentary remnants. These rocks outcrop widely west of 15°00' longitude and west of the Witpoortberg Hills, where a major proportion of the rocks exposed belong to this suite. As mentioned in the previous section, subordinate amounts of these rocks are associated with the Abbabis suite of augen gneisses in the Tumas River Inlier.

Similarly, augen-gneisses of the Abbabis Complex occur within the Husab suite of rocks in the west. Separate representation of these units on the geological map
however, is not possible owing to the scale and complexity of their relationships. Sheared and deformed augen-gneisses can be recognised in the Husaberg anticlinorium but are subordinate to the medium-grained red biotite rich granite-gneisses with which they are associated.

The Husab granite-gneiss suite is characterised by a marked red colour which is particularly obvious in the biotite-rich granite-gneisses in the Husabberg anticlinorium. In the west (Welwitschia anticlinorium) the rocks have a low biotite content and leucogranites and pale pegmatitic granites are common. Leucogranites and pegmatites associated with the red granite-gneisses are commonly contorted and folded which indicates that they were syn-tectonically emplaced. The foliated rocks of this suite occur below the Nosib and Swakop group metasediments.

Mobilisates of red and grey homogeneous granite are found associated with the Damaran metasediments in several localities. Granite dykes can be seen cross-cutting the Nosib metasediments in the vicinity of Hollands anticline. Khomas Subgroup metasediments and the Salem granitoid suite are cut by several dykes of red granite on the western side of the Tumas River Inlier approximately 2.5km north of the intersection of 23°00' latitude and 15°00' longitude. There are no contact metamorphic affects present in the host rocks. Dykes of this granite are also observed as contorted remnants in the late to post tectonic Rössing alaskitic granite (Plate 24). Throughout the area, red granite, gneiss and leucogranite have been complexly interfolded with basal Nosib metasediments and migmatites. These areas are defined on the geological map as the stippled part (~) of the Husab granite-gneiss suite.

It is apparent that the granite-gneisses and granites of the Husab suite were emplaced syn- and late-tectonically
during the Damaran orogenic period and were generated by remobilisation of the Abbabis granite-gneisses. A small amount may have formed by contamination with, and melting of, overlying Damaran metasedimentary rocks.

4.1.3 SALEM GRANITOID SUITE (Gn2)

Grey porphyritic biotite granite/granodiorite, non-porphyritic granodiorite and porphyritic leucogranite are the major varieties of rocks comprising this suite.

The porphyritic biotite granite/granodiorite, which is the most abundant, constitutes approximately 85% of the rocks, and characteristically contains coarse ( \( < 4 \mathrm{~cm} \) ) phenocrysts of K-feldspar that are commonly strongly aligned and often anhedral in shape (Plate 25). The phenocrysts lie in a well-foliated biotite rich matrix, the alignment of which appears to run parallel to the limbs of the major structures and therefore is of \( F_2 \) age.

The non-porphyritic granite is also strongly foliated but K-feldspar phenocrysts are generally absent. Interrelated zones of non-porphyritic and porphyritic granodiorite/granite give the suite a stromatic, migmatitic appearance (Plate 26). The non-porphyritic granodiorite occurs intermittently throughout the occurrences of the Salem rocks.

The porphyritic leucogranite occurs as elongate and contorted bodies associated with the porphyritic and non-porphyritic granite/granodiorite. The porphyritic leucogranite differs from the porphyritic biotite granite/granodiorite in that it contains less biotite in the matrix and is generally non-foliated. The porphyritic leucogranite cuts across the foliated rocks of the suite with sharp contacts and is clearly a post-tectonic phase (Plate 26).

Within the eastern part of the Salem granitoid occurrence in the Tubasberg syncline, a medium-grained, grey-garnet-bearing granite is present. It is possible
that this granite may relate to other red-grey granites of the Husab suite that intrude the Salem granitoid rocks. Occasionally the Salem rocks have a strong red colouration which may represent some degree of admixture with red granites and gneisses of the Husab suite.

The Salem granitoids occur to the west of the Tumas River Inlier in this area and stratigraphically lie above the marbles of the Karibib Formation. They are found in synclinal and basin structures bounded by the marbles and remnants of the Witpoort biotite schists (Kuiseb Formation). An exception occurs in the south-western part of the area immediately east of the two small basin structures containing Salem granitoids. The Salem rocks here are in direct contact with rocks of the Husab suite. There is evidence that the contact is a shear zone as intermittent marble bands occur along with vein quartz and breccia, along the contact.

4.1.4 RÖSSING ALASKITIC GRANITE (G4)

This suite comprises pegmatites, pegmatitic granites and leucogranites containing very low proportions of mafic minerals. They are present as dykes and large irregular bodies (Plate 24), intruding granitgneisses of the Husab suite, and the Damara supergroup metasediments.

Pegmatites and leucogranites occurring within the Husab suite of rocks, that are complexly deformed and diffusely related to the surrounding granites and gneisses with gradational contacts are classified as part of the Husab suite. The term Rössing alaskitic granite is used to define leucogranites and pegmatitic granites that show definite late to post tectonic features with obvious intrusive relationships with the adjacent rocks.

The Rössing alaskitic granites most frequently occur at, or near the base of the Karibib Formation (marbles)
and within the banded gneiss sequence of the Khan Formation. The granites are extremely variable in grain size and large parts are pegmatitic. Zoning, as often found in pegmatites, is not observed in these rocks. However, local clusters of smoky quartz, biotite and iron oxides give the rocks a patchy appearance in some places. The rocks are generally non-foliated and have been intruded both concordantly along foliation planes in the host rocks, and discordantly across lithologic and structural boundaries. Contact metamorphic features are not prominent where the granites have intruded gneisses or quartzites and contacts are either sharp or diffuse where digestion and minor recrystallization of the host rock has occurred. Where intruded into the marbles (Karibib Formation), skarn zones are generally present at granite/marble contacts. The most widely developed of these skarn zones is seen immediately to the southeast of the Zebraberge. At this locality, extensive but discontinuous skarn bodies are developed at the contact with the marbles over a strike length of approximately 3km and width of 2 to 10 metres, extending further north beyond the area.

These skarns, which are dark brown in colour and therefore conspicuous in relation to the surrounding felsic and carbonate rocks, are composed predominantly of garnet with subordinate amounts of diopside, scapolite, wollastonite, epidote, vesuvianite and tremolite. Elsewhere, narrow (1-10m) dykes of the granite intrude into the marble, and wollastonite is developed intermittently along the contacts.

Occurrences of the Rössing alaskitic granites are confined to the western part of the area where their association with the Khan Formation metasediments is strikingly apparent. On the eastern side of Glück Dome and the Tumas River Inlier, there is abundant development of pegmatites within the Karibib and Kuiseb metasediments. These pegmatites are concordantly
emplaced within the host rock foliations and are often zoned. These pegmatites are not classified with the Rössing alaskitic granites as they probably have developed syn-tectonically from metasediments at depth.

4.1.5 GAWIB GRANITOID SUITE (G3)

The Gawib granitoids occur in the form of a stock in the extreme eastern part of the area. The stock is oval in shape and varies from 7.5km in width in the northern part of the mapped area, to 2km in the south. The body is approximately 16km long (N-S axis) and has been emplaced discordantly and across the foliations and lithological boundaries of the enclosing metasediments. The enclosing country rocks comprise Kuiseb and Etusis metasediments, tillite (Chuos Formation) and augen gneisses (Abbabis Complex). Contact metamorphic effects associated with the intrusion of the stock are restricted to a faint discoloration of the country rocks, particularly the quartzites (Etusis Formation), immediately adjacent (1-5 metres) to the contact.

The stock is composed of several different, but related rock types;

i) A foliated granodiorite, characterised by a high content of mafic minerals (hornblende and biotite) which has a foliation developed parallel to the contacts of the stock. These rocks occur at the margin of the stock, forming an envelope around the interior. Their greatest development is in the north-eastern part of the body, but occur intermittently around the perimeter.

ii) A medium-grained non-porphyritic granite containing scattered phenocrysts of K-feldspar. This variety is widely developed in the core of the stock in the northern and southern extremities.
(iii) A porphyritic granite/granodiorite, similar in character to (ii) but characterised by abundant development of K-feldspar phenocrysts, occurs in the central part of the stock.

(iv) Pale and pink aplastic dykes. These occur within the stock and at the contact with the country rocks.

The different members of the Gawib stock have both sharp intrusive contacts and gradational relationships. From field relationships the stock appears to be a differentiated body composed of a central core of late-phase porphyritic granite/granodiorite surrounded by the non-porphyritic granite. These two members form the bulk of the stock, and are enclosed in a narrow envelope of early-phase foliated granodiorite.

4.1.6 DISCUSSION

The augen gneisses present in the Tumas Inlier are undoubtedly pre-Damaran in age as a consequence of their structural discordance with, and their incorporation as clasts into, the rudaceous beds of the Damara Supergroup. The Husab suite of granite-gneisses therefore have been formed primarily by reactivation of the pre-Damaran basement rocks as is evident by the relicts of pre-Damaran gneisses present throughout the Husab suite occurrences. Husab granites derived by anatexis of Nosib metasediments appear to be of comparatively minor extent, although migmatites have been developed from these rocks.

The Salem granitoid suite occurs exclusively in synclinal and basin structures which contain remnants of Kuiseb metasediments. This supports earlier suggestions by Smith (1965), Miller (1973) and Jacob (1974) that it has developed by anatexis of the Kuiseb metasediments. It is found only on the high temperature side of the anatexis-in-gneiss boundary.
located in the field (see fig 5).

The intrusive features, contact metamorphic affects and pegmatitic nature of the Rössing alaskitic granites suggest that this suite has originated from deeper sources rather than from in-situ segregation, and at a late phase in the tectonism and granitisation.

The well differentiated nature of the Gawib stock and its emplacement across Damaran structures and the underlying pre-Damara (Basement) rocks implies a post tectonic origin from depth.

The granitoids in this area exhibit characteristics and relationships similar to their equivalents in the Damara belt further to the north formerly described by Gevers (1931), Smith (1965), Nash (1971), Miller (1973), Jacob (1974), Faupel (1974), Nießerding (1976) and Hoffman (1976). It is apparent that the granitoid suites occurring throughout the central part of the Damara belt retain a constancy in field characteristics and associations.

4.2 PETROGRAPHY

4.2.1 ABBABIS COMPLEX (AGn1)

Augen-gneisses, the most common of the rocks in this suite are characterised by well developed K-feldspar or K-feldspar/quartz augen (1-4cm in length). The augen are separated by biotite and quartz-feldspar layers which wrap around the augen (Plate 23). The texture of the rocks is medium to coarse grained and gneissose.

Modal analyses (Table 10) show the major phases to be microcline, quartz, plagioclase (An$_{26-35}$), and biotite. Accessory minerals include: muscovite (+ sericite), zircon, Fe-ore, and apatite. Microcline exhibits typical cross-hatched twinning and is occasionally poikilitic and perthitic. Quartz is present in the form of large strained anhedral grains
and plagioclase crystals are often zoned, with albite-rich rims. Subhedral, subparallel, often deformed grains of biotite occur in distinct bands. A pale red foliated granite, associated with the augen-gneisses on the eastern flanks of the Raben-rücken is composed almost exclusively of felsic minerals with accessory amounts of Fe-ore and muscovite. The rock is medium grained and contains narrow zones of fine-grained quartz and K-feldspar. Later fractures contain iron oxide.

4.2.2 HUSAB GRANITE-GNEISS SUITE (Gn1)

The rocks types that occur within this suite are red granite, red granite-gneiss, leucogranite, pegmatite, migmatite, paragneiss and amphibolite which are all intimately associated. Despite the variety of rock types obvious in the field, the rocks have very similar modal compositions (Table 10).

The granitoids all contain as major constituents: quartz, microcline, plagioclase, and biotite. Accessory amounts of Fe-ore, muscovite, zircon, apatite, sphene, epidote and calcite also occur. Microcline, the predominant feldspar is typically anhedral, exhibits cross-hatched twinning and is occasionally poikilitic and also perthitic. Myrmekite is often developed at K-feldspar/plagioclase boundaries. Plagioclase is invariably multiply twinned and has sericitic alteration along the cleavage planes. Zoning is often observed in the plagioclase grains and distinct albite rims occur on some. The composition of plagioclase varies between An\textsubscript{26} - An\textsubscript{30}, with the exception of the plagioclase rich rocks (GB 228, GB 229) and (GB 210), where the composition is predominantly sodic (An\textsubscript{7} - An\textsubscript{12}). Antiperthite has been observed in some samples. Quartz, occurring as discrete grains or aggregates, commonly exhibits strained extinction. Biotite occurs as euhedral to subhedral grains, usually in distinct zones and exhibiting green-brown pleochroism.
In many of the rocks, chlorite forms as a retrograde phase of the biotite. Zircon grains are often observed as inclusions in the biotite, and are surrounded by pleochroic haloes. Calcite occurs as a secondary mineral in fractures within the rocks.

TABLE 10

<table>
<thead>
<tr>
<th>SAMPLE NO</th>
<th>ROCK TYPE</th>
<th>QUARTZ</th>
<th>MICROCLINE</th>
<th>Biotite</th>
<th>ACCESORIES</th>
<th>AVERAGE FELICIAS COMPOSITION</th>
<th>LOCALITY</th>
</tr>
</thead>
<tbody>
<tr>
<td>GB200</td>
<td>augen-gneiss</td>
<td>35</td>
<td>36</td>
<td>14</td>
<td>12</td>
<td>musc, zir, ore, epid</td>
<td>An35</td>
</tr>
<tr>
<td>GB205</td>
<td>&quot;</td>
<td>31</td>
<td>47</td>
<td>12</td>
<td>8</td>
<td>&quot; &quot; &quot;</td>
<td>An31</td>
</tr>
<tr>
<td>GB209</td>
<td>&quot;</td>
<td>24</td>
<td>45</td>
<td>22</td>
<td>7</td>
<td>musc, ore</td>
<td>An35</td>
</tr>
<tr>
<td>GB213</td>
<td>&quot;</td>
<td>28</td>
<td>27</td>
<td>20</td>
<td>23</td>
<td>musc, zir, ore</td>
<td>An26</td>
</tr>
<tr>
<td>GB226</td>
<td>pink foliated granite</td>
<td>32</td>
<td>48</td>
<td>19</td>
<td>1</td>
<td>musc, ore</td>
<td>An26</td>
</tr>
<tr>
<td>GB207</td>
<td>red foliated granite-gneiss</td>
<td>30</td>
<td>47</td>
<td>17</td>
<td>5</td>
<td>musc, zir, ap, ore</td>
<td>An20</td>
</tr>
<tr>
<td>GB208</td>
<td>red feldspar granite</td>
<td>35</td>
<td>38</td>
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<td>5</td>
<td>&quot; &quot; &quot;</td>
<td>An21</td>
</tr>
<tr>
<td>GB211</td>
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<td>26</td>
<td>12</td>
<td>&quot; &quot; &quot;</td>
<td>An27</td>
</tr>
<tr>
<td>GB217</td>
<td>red granite</td>
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<td>27</td>
<td>3</td>
<td>&quot; &quot; &quot;</td>
<td>An24</td>
</tr>
<tr>
<td>GB227</td>
<td>red foliated granite-gneiss</td>
<td>34</td>
<td>42</td>
<td>19</td>
<td>4</td>
<td>&quot; &quot; &quot;</td>
<td>An25</td>
</tr>
<tr>
<td>GB221</td>
<td>&quot;</td>
<td>31</td>
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<td>4</td>
<td>&quot; &quot; &quot;</td>
<td>An23</td>
</tr>
<tr>
<td>GB223</td>
<td>red granite</td>
<td>30</td>
<td>32</td>
<td>34</td>
<td>2</td>
<td>&quot; &quot; &quot;</td>
<td>An27</td>
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<tr>
<td>GB219</td>
<td>pale-grey granite</td>
<td>24</td>
<td>35</td>
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<td>An24</td>
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<tr>
<td>GB225</td>
<td>pale-red granite</td>
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<td>29</td>
<td>26</td>
<td>5</td>
<td>&quot; &quot; &quot;</td>
<td>An26</td>
</tr>
<tr>
<td>GB226</td>
<td>pale granite-gneiss</td>
<td>39</td>
<td>21</td>
<td>39</td>
<td>1</td>
<td>&quot; &quot; &quot;</td>
<td>An25</td>
</tr>
<tr>
<td>GB229</td>
<td>leuco-granite-gneiss</td>
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<td>33</td>
<td>40</td>
<td>1</td>
<td>&quot; &quot; &quot;</td>
<td>An12</td>
</tr>
<tr>
<td>GB210</td>
<td>&quot;</td>
<td>35</td>
<td>37</td>
<td>26</td>
<td>2</td>
<td>&quot; &quot; &quot;</td>
<td>An7</td>
</tr>
</tbody>
</table>
4.2.3 ROSSING ALASKITIC GRANITES (G4)

This leucogranitic-pegmatitic suite of rocks has an extremely variable grain size. The minerals are generally 0.5mm - 4mm in size, but occasional crystals of K-feldspar up to 30cm in size are observable in outcrop. The texture is hypidiomorphic granular, and the main mineral phases present are quartz, microcline and plagioclase, with minor to accessory amounts of biotite, ore, muscovite sphene, zircon and calcite. Microcline is the predominant constituent with the exception of KH 256, KH 472/3 and KH 496, which contain predominantly plagioclase (Table 11).

The plagioclase composition of the granite varies between An_{12} to An_{19} (soda rich oligoclase) and the crystals are commonly zoned.

Microcline often contains perthitic intergrowths of plagioclase and occasionally K-feldspar crystals are enclosed in larger plagioclase crystals (antiperthite), particularly in the plagioclase rich rocks. (KH 256, KH 472/3 and KH 496).

Quartz is occasionally strained and often smoky grey in appearance, particularly where associated with aggregates of biotite and ore.

Microcline exhibits cross-hatch twinning and plagioclase crystals are generally multiply twinned and slightly sericitized along composition planes. Myrmekitic intergrowths are present along contacts of K-feldspar and plagioclase.

4.2.4 SALEM GRANITOID SUITE (Gn2)

This suite of rocks (Table 12) includes porphyritic granite/granodiorite, porphyritic leucogranite and fine-grained granodiorite.

The porphyritic biotite granite/granodiorite and porphyritic leucogranite are both coarse-grained rocks, but the former contains abundant biotite and has a gneissic texture. The fine-grained granodiorite, unlike the former two, has a very minor phenocryst development but is rich in biotite.
TABLE 11

<table>
<thead>
<tr>
<th>SAMPLE NO.</th>
<th>ROCK TYPE</th>
<th>QUARTZ (vol%)</th>
<th>MICROLINE (vol%)</th>
<th>PLAGIOCLASE (vol%)</th>
<th>Biotite (vol%)</th>
<th>Accessories</th>
<th>Average Plagioclase Composition</th>
<th>Locality</th>
</tr>
</thead>
<tbody>
<tr>
<td>GB236</td>
<td>Fine grained pale granite</td>
<td>27</td>
<td>47</td>
<td>25</td>
<td>tr</td>
<td>1</td>
<td>muscovite, ore, sph</td>
<td>An15</td>
</tr>
<tr>
<td>KH106/7</td>
<td>Med.</td>
<td>21</td>
<td>56</td>
<td>23</td>
<td>tr</td>
<td>tr</td>
<td>muscovite, cc</td>
<td>An12</td>
</tr>
<tr>
<td>KH132/3</td>
<td>&quot;-&quot; coarse pink granite</td>
<td>12</td>
<td>63</td>
<td>23</td>
<td>1</td>
<td>1</td>
<td>ore</td>
<td>An17</td>
</tr>
<tr>
<td>KH255</td>
<td>&quot;-&quot; pale granite/pegmatite</td>
<td>32</td>
<td>42</td>
<td>25</td>
<td>1</td>
<td>tr</td>
<td>ore</td>
<td>An18</td>
</tr>
<tr>
<td>KH256</td>
<td>Med-fine grain granite</td>
<td>29</td>
<td>7</td>
<td>64</td>
<td>tr</td>
<td>tr</td>
<td>muscovite, sph</td>
<td>An17</td>
</tr>
<tr>
<td>KH472/3</td>
<td>Med-fine gray granite</td>
<td>33</td>
<td>4</td>
<td>58</td>
<td>5</td>
<td>tr</td>
<td>muscovite, sphe, zir, cc</td>
<td>An19</td>
</tr>
<tr>
<td>KH496</td>
<td>Med-coarse gray granite</td>
<td>15</td>
<td>7</td>
<td>75</td>
<td>3</td>
<td>tr</td>
<td>ore</td>
<td>An18</td>
</tr>
<tr>
<td>KH497</td>
<td>&quot;-&quot; pale granite/pegmatite</td>
<td>29</td>
<td>43</td>
<td>28</td>
<td></td>
<td>-</td>
<td></td>
<td>An14</td>
</tr>
</tbody>
</table>

Microcline is the phenocryst phase in the porphyritic varieties and the matrix feldspar is predominantly plagioclase. Microcline is perthitic, and myrmekite is developed between K-feldspar and plagioclase crystals. Hornblende exhibits strong green/yellow-brown pleochroism and is commonly associated with biotite. Muscovite is present as distinct subhedral crystals often occurring in aggregates. Biotite, in places has suffered retrograde alteration to chlorite.
<table>
<thead>
<tr>
<th>SAMPLE NO</th>
<th>ROCK TYPE</th>
<th>QUARTZ</th>
<th>MUSCOVITE</th>
<th>MICROCLINE</th>
<th>PLagioclase</th>
<th>Biotite</th>
<th>Hornblende</th>
<th>Accessories</th>
<th>Average Plagioclase Composition</th>
<th>Locality</th>
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<tbody>
<tr>
<td>GB216</td>
<td>porphyritic foliated granodiorite</td>
<td>22</td>
<td>24</td>
<td>31</td>
<td>17</td>
<td>2</td>
<td>2</td>
<td>4, Ap, sph, ore, zir</td>
<td>An25</td>
<td>P10</td>
</tr>
<tr>
<td>GB220</td>
<td>&quot;</td>
<td>23</td>
<td>9</td>
<td>50</td>
<td>13 tr</td>
<td>5</td>
<td>5</td>
<td>Ap, musc, &quot;&quot;, &quot;&quot;, cc</td>
<td>An41</td>
<td>Q6</td>
</tr>
<tr>
<td>GB237</td>
<td>fine grained granodiorite</td>
<td>19</td>
<td>11</td>
<td>43</td>
<td>15</td>
<td>8</td>
<td>2</td>
<td>Ap, zir, cc, ore, musc</td>
<td>An25</td>
<td>R13</td>
</tr>
<tr>
<td>GB254</td>
<td>dark foliated granodiorite</td>
<td>18</td>
<td>16</td>
<td>41</td>
<td>15</td>
<td>8</td>
<td>2</td>
<td>epid, sph, all, ore, ap</td>
<td>An35</td>
<td>L28</td>
</tr>
<tr>
<td>GB221</td>
<td>&quot;</td>
<td>22</td>
<td>10</td>
<td>39</td>
<td>18</td>
<td>5</td>
<td>6</td>
<td>epid, sph, all, cc, ore, ap</td>
<td>An31</td>
<td>L28</td>
</tr>
<tr>
<td>GB203</td>
<td>slightly porphyritic</td>
<td>21</td>
<td>28</td>
<td>35</td>
<td>9</td>
<td>1</td>
<td>6</td>
<td>epid, cc, sph, ore, all, zir</td>
<td>An29</td>
<td>L27</td>
</tr>
<tr>
<td>GB201</td>
<td>non-porphyritic granite</td>
<td>33</td>
<td>16</td>
<td>43</td>
<td>6</td>
<td>tr</td>
<td>2</td>
<td>epid, cc, sph, ore, all, zir</td>
<td>An27</td>
<td>M25</td>
</tr>
<tr>
<td>GB202</td>
<td>&quot;</td>
<td>25</td>
<td>17</td>
<td>49</td>
<td>7</td>
<td>tr</td>
<td>2</td>
<td>epid, all, musc, cc, sph, ore</td>
<td>An28</td>
<td>L26</td>
</tr>
<tr>
<td>GB223</td>
<td>porphyritic granite</td>
<td>23</td>
<td>27</td>
<td>40</td>
<td>7</td>
<td>tr</td>
<td>3</td>
<td>epid, ap, sph, ore, zir</td>
<td>An29</td>
<td>N26</td>
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<td>GB224</td>
<td>&quot;</td>
<td>27</td>
<td>21</td>
<td>43</td>
<td>7</td>
<td>tr</td>
<td>2</td>
<td>epid, ap, sph, ore, zir</td>
<td>An29</td>
<td>N25</td>
</tr>
</tbody>
</table>

### 4.2.5 Gawib Granitoid Suite (G3)

The different members of the Gawib stock are:

i) foliated granodiorite

ii) porphyritic granite/granodiorite

iii) non-porphyritic granite

iv) aplite
Varieties i) to iii) show increasing quartz and microcline, and decreasing hornblende and biotite contents from the margin to the centre of the stock (Table 12). The texture of the rocks is hypidiomorphic granular and a faint foliation and distortion of biotite crystals is obvious in samples GB 204, GB 221 and GB 203 from the margin of the stock. These also contain abundant accessory minerals which include epidote, sphene, ore, allanite, apatite, muscovite and calcite.

Biotite, hornblende and the most common accessory mineral in the marginal members, epidote, generally occur together, the biotite being largely altered to chlorite in some samples.

Microcline has typically cross-hatched twinning and is often poikilitic, containing blebs of plagioclase and quartz. Myrmekitic intergrowths occur between the two feldspars, and plagioclase which generally is multiply twinned, is severely sericitised along composition planes.

Plagioclase grains are commonly zoned, with the rims, richer in albite. Average plagioclase compositions vary from An$_{35}$ to An$_{27}$ with those from foliated granodiorites being slightly higher in the An component. Subhedral crystals of muscovite present in some rocks as inclusions within plagioclase probably represent the products of sericitisation.

Quartz is present as anhedral strained crystals particularly in the foliated granodiorite. Calcite generally occurs interstitially and also in fractures. Zircons with radioactive haloes are observed in biotite grains.

4.2.6 DISCUSSION

Modal quartz, alkali feldspar and plagioclase components of the rocks are plotted on triangular diagrams after the methods proposed by Streckeisen (1973), see
Fig 7. The modal classifications are generally in agreement with currently used field terminology. Modal analyses are plotted in Figs 8, 9 and 10 with fields which represent modal data for similar rock suites elsewhere in the Damara belt studied by previous workers, shown for comparison.

1a Quartzolite (silexite)  
1b Quartz-rich granitoids  
2 Alkali-feldspar  
3 Granite  
4 Granodiorite  
5 Tonalite  
6a Alkali-feldspar quartz syenite  
7a Quartz syenite  
8a Quartz monzonite  
9a Quartz monzodiorite/quartz monzogabbro  
10a Quartz diorite/quartz gabbro/quartz anorthosite  
6b Alkali-feldspar syenite  
7b Syenite  
8b Monzonite  
9b Monzodiorite/monzogabbro  
10b Diorite/gabbro/anorthosite

Fig. 7
General classification and nomenclature of granitoids as proposed by Streckeisen (1973)

Rocks of the HUSAB GRANITE-GNEISS suite from this area plot in the centre of the granite field (Fig 8), with slight extension into the granodiorite field (Jacob, 1974)
and the alkali feldspar granite and quartz-syenite fields (Smith, 1965). The augen gneisses of the ABBABIS Complex fall within the centre of the granite field. The crystallisation sequence of the minerals in these rocks is not clear from petrographic interpretation.

Fig. 8

Red Granites, Gneisses, leucogranites
Augen gneisses
Red granite-gneiss (Smith 1965)  
" " " (Jacob 1974)  
" " " (Hoffmann 1976)

Streckeisen Plot of Modal Compositions
For Husab and Abbabis Granite-gneiss Suites.
The RÖSSING alaskitic granites, from field characteristics would be expected to have an alkali-feldspar granite composition but with varying proportions of modal plagioclase (An 12 - 17) not sodic enough to be incorporated in the alkali-feldspar component (A of the triangular plots), these rocks range in composition from tonalites to quartz-syenites according to the Streckeisen plot. They generally have low quartz contents compared to the other granites. Inhomogeneous sampling has undoubtedly led to the spurious results from these plots.

The SALEM GRANITOID suite of rocks from the central part of the belt compositionally fall in the monzogranitic and granodiorite fields (Fig 9) with several rocks of tonalitic composition noted by Jacob (1974) and Hoffman (1976). The Salem suite from an area 150km north of the central zone of the Damara belt at Omgambo and Otjousondjou, investigated by Miller (1973) shows a considerable range in composition (see Fig 9). His samples of the dioritic envelope at Otjousondjou (Miller 1973, Fig 3) are of quartz-diorite/quartz-monzodiorite/tonalite/granodiorite composition, whilst those further to the west in the larger body at Omgambo are of granodioritic and granitic composition.

In comparison to the Salem granitoid suite from the central part of the Damara belt the dioritic/monzodioritic varieties of the Salem suite from the north are extremely rich in hornblende, with this mineral occurring in similar or greater proportions than biotite. These rocks are also much lower in modal quartz, which generally constitutes less than 25 volume percent of the rocks.

The GAWIB GRANITOIDs are predominantly granodioritic but also partly granitic in composition (Fig 10). The Gawib, Achas and other granodiorites occurring near the Donkerhoek granite (Jacob 1974; Nieberding 1976) have similar compositions. Further to the north
Fig. 9

Porphyritic biotite Gr. ○
Non-porphyritic Gr/Gn ●

Salem Granite (Smith 1965) /\  | Salem granite (Ongambo) Miller (1973)
" " (Jacob 1974) \   | Salem granite - (Otjosondjou) Miller (1973)
" " (Hoffmann 1976) \  |-

Streckeisen plot of modal compositions for Salem Granitoid Suite.

Smith (1965) mapped similar bodies which he termed quartz-diorites and diorite-gneisses. These rocks grade into what he called 'Salem-type' granite (Smith, 1965, p54) which are possibly equivalent to the porphyritic granite from the interior of the
Fig. 10

Foliated Granodiorite
Non-porphyritic Granite
Porphyritic Granite
Quartz-diorite, Diorite-gneiss (Smith 1965)
Gawib, Achas Granites (Jacob 1974)
Achas Granite, Granodiorite (Nieberding 1976)

Streckeisen plot of modal compositions for the Granodiorite Suite incorporating (Gawib, Achas, Quartz-diorite, Diorite-gneiss.)

Gawib stock, and are therefore similarly differentiated. Modally these rocks are quartz monzodiorite, tonalite and granodiorite in composition (see Fig 10).
The Gawib stock therefore appears to belong to a granodioritic suite of rocks of similar character which intrude Kuiseb Formation metasediments throughout the central Damara belt. Their relationship to the Salem granitoid suite is uncertain from field evidence but they may be allochthonous equivalents. The sequence of crystallisation from petrographic evidence is unclear, but the abundance of zoned plagioclase grains suggests prolonged crystallisation of this mineral.

4.3 GEOCHEMISTRY

4.3.1 SCOPE OF THE INVESTIGATION

The geochemistry of the granitic rocks was investigated in order to establish the geochemical character of the different suites and to examine petrochemically the evidence of their origin and relationships that are apparent from field and petrographic observations.

The field relationships (see section 4.1) of the Abbabis granite-gneisses, the Husab granite-gneisses and the Rössing Alaskitic granites indicate that these rock suites are related. The evidence suggests that the Abbabis rocks represent pre-Damaran basement material which, during the Damaran orogeny, was reactivated in part by partial and/or complete melting processes, to form the syn-kinematic Husab granite-gneisses and the late-kinematic Rössing alaskitic granites. Earlier conclusions by Smith (1965) Jacob (1974) and Hoffmann (1976) imply that the suite of granite-gneisses now termed the Husab suite, formed largely by partial melting (anatexis) of Nosib metasediments. Consequently, the Rössing alaskitic granites were considered to represent late-phase mobilisates similarly derived by anatexis of Nosib metasediments. An extensive geochemical investigation of both the Damara metasediments and the granitic rocks was beyond the scope of this study. Attention was therefore focused on the granitic rocks in order to establish, geochemically, the validity of the conclusions on their origin made from field observations in this study.
The syn-kinematic Salem and the post-kinematic Gawib granitoid suites have similarities in petrography and field setting (see sections 4.1 and 4.2) as they are both composed largely of porphyritic granite and grano­diorite and they occur largely within the same stratigraphic level above the Karibib Formation (see Fig 3). This has led to conclusions that the Gawib stock possibly originated from the same magma that produced the Salem suite (Jacob 1974). The Gawib stock in this area, however, is a differentiated body that cross-cuts pre-Damaran basement rocks and an allochthonous deep-seated origin appears probable rather than an autochthonous anatectic origin, as is implied for the Salem suite.

A geochemical study was therefore undertaken, to reveal if differences exist between the geochemistry of these rocks suites and to also assist in establishing their origin. At present the geochemical characteristics of the Salem granitoid suite from the type area are not known. Rocks of the Salem granitoid suite used for comparison with those of the Gawib granitoid stock are the limited number of samples collected in this area, and those described by Miller (1973) for the Salem granitoid suite occurring to the north at Omangambo and Otjosondjou.

**4.3.2 PRESENTATION OF DATA**

The rocks analysed, their sample numbers and sample localities are presented in tables 10,11 and 12 and Map (1) respectively. The rocks were analysed by X-Ray fluorescence techniques (XRF) on a Phillips PW1410 spectrometer. Sample collection, preparation and analytical methods, along with data compilation techniques and values for standards are given in the Appendix.

The major and trace element data are listed in tables 13, 14 and 15 with trace element ratios, cation percentages and mesonorm values. Relevant data are also presented in variation diagrams. These diagrams provide a simple visual means of interpreting and
presenting the data whereby comparison within and between different rock suites can be made.

There is little agreement as to the correct variable for use as abscissa in the variation diagrams, but the majority of workers favour one of the following:

\[
\text{SiO}_2 \quad \text{Harker diagram}
\]

\[
\left( 1/3 \text{ Si + K} \right) - (\text{Ca + Mg}) \quad \text{Nockolds and Allen (1953) modified Larsen index}
\]

\[
\text{Normative } \text{Qz+Ab+Or+Ne+Kp+Lc} \quad \text{Thornton and Tuttle (1960) Diff Index}
\]

\[
\text{MgO} \quad \text{Wright (1974)}
\]

In the present investigation, after consideration of the above indices, MgO was chosen as the abscissa for the variation diagrams. It presents the best spread and representation of all the rock groups in relation to observed field and petrographic associations and is considered a good index of differentiation as MgO generally decreases continuously during fractional crystallisation irrespective of starting composition and pressure (Wright 1974), although in rare instances under oxidizing conditions it has been observed to increase with differentiation (Marsh, pers. comm.).

In the presentation and use of normative rock compositions, mesonorm rather than CIPW norm values were used. The mesonorm values have close similarities to the modal compositions, incorporate hydrous minerals, and have a wider applicability in comparing the data with that available from experimental systems determined from laboratory studies.

Laboratory investigations determining phase relations and equilibria by melting-experiments on natural and synthetic assemblages have provided reliable data for the granite system (Qz - Ab - Or - An - H₂O) (See Fig 11) since the early work of Tuttle and Bowen (1958).
<table>
<thead>
<tr>
<th>Table 13</th>
</tr>
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<tbody>
<tr>
<td>Major element, trace element, trace element ratio, cation percentage and mesonorm values for rocks analysed from the Aabab and Husab suites.</td>
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Major element, trace element, trace element ratio, cation percentage and mesonorm values for rocks analysed from the Salem and Gamib Granitoid suites.
The results from various workers including Luth et al. (1964), Piwinski and Wyllie (1968, 1970) Robertson and Wyllie (1971), Piwinski (1973), and Winkler (1967, 1974) show the stability of different minerals and mineral assemblages at different temperatures, pressures and degrees of water saturation, from $H_2O$-deficient to the $H_2O$ saturated melts. More recently, Winkler et al. (1975) have shown the validity of plotting normative or carefully determined modal compositions within the system Qz - Ab - Or - An - H$_2$O in order to determine crystallisation sequences and temperatures of formation for given pressures ($PT = PH_2O$). It is then possible to infer origins (partial melting (anatexis) or fractional crystallisation from depth) for natural granitic, granodioritic and tonalitic rocks, by comparing the natural data with results gained from controlled melting experiments.

Mesonorm values for the granitoids from this area have been plotted on the Qz - Ab - Or and An - Ab - Or projections of Winkler et al., (1975), in an attempt to establish crystallization sequences and temperatures. The methods of Winkler et al., (op. cit.) using central projections were used in this study although Hoffmann (1974), considers that distorted plots result from the use of this method of projection and he prefers parallelly projected points.

4.3.3 NORMATIVE CHEMISTRY RELATED TO EXPERIMENTAL SYSTEMS

4.3.3.1 ABBABIS-, HUSAB-, GRANITE-GNEISSES AND ROSSING ALASKITIC GRANITES

The data from table 16 is presented on Qz - Ab - Or and An - Ab - Or projections (Figs 12 a and b) of the Qz - Ab - An - Or - H$_2$O tetrahedron (Fig 11). The phases not represented at the apices of the projections (An and Qz respectively) are shown by figures (weight percent) beside each plotted rock position.
**TABLE 16**

Ab-An-Or and Qz-Ab-Or coordinates used in triangular plots

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Fig. 11

Tetrahedron representing system Qz-Ab-Or-An-H$_2$O (water saturated) at 5kb pressure; perspective view (after Winkler et al., 1975).

The ABBABIS augen-gneisses (AGn$_1$) lie within both the quartz and alkali feldspar fields of Fig 12 and at maximum distances of approximately 5% An and 10% An below the Qz+plag+L+V cotectic surface and plag+K-feldspar+L+V cotectic surface respectively (see Fig 12a), but only 2% (excluding GB 200) away from the Qz-alk feld+L+V cotectic surface (see Fig 12b).
Qz-Ab-Or projection (Fig 12a) and An-Ab-Or projection (Fig 12b) showing the position of rocks from the Abbabis (AGn1), Husab (Gn1) and Rössing suites.
This implies that during crystallisation from a melt the crystallisation sequence for these rocks begins with either quartz or alkali feldspar

i.e. i) quartz → quartz + alkali feldspar → quartz + alkali feldspar + plagioclase

ii) alkali feldspar → alkali feldspar + quartz → alkali feldspar + quartz + plagioclase

As the samples all plot relatively close to cotectic surfaces, the crystallisation of all the phases, by analogy with the results of Winkler et al., (1975), occurred over a small temperature interval. By tracing the paths of crystallisation from Fig 12, it can be deduced that the first minerals to crystallise (quartz or alkali feldspar) appeared at approximately 680-710°C, with the first appearance of plagioclase occurring at approximately 670°C.

The HUSAB granite-gneisses all plot close to the cotectic line P-E, with the majority lying within the alkali feldspar field, and the remainder occurring within the quartz field. This implies that as the samples lie less than 5% An and 5% Qu away from the cotectic planes, plag + alk. felds + L + V/Quz + plag + L + V and Quz + alk. felds + L + V respectively, crystallisation of the three phases occurred over a very small temperature interval. The crystallisation sequence is similar to that concluded for the Abbabis augen gneisses. In the majority of the rocks, alkali-feldspar crystallised first from the melt, then, after a small temperature decrease, quartz and alkali feldspar crystallise together followed by plagioclase. The temperature interval from the first appearance of alkali feldspar to the first appearance of plagioclase appears to be in the range 700-670°C which is similar to temperatures deduced from the Abbabis plots.

The ROSSING alaskitic granite samples exhibit a considerable scatter in their plotted positions. The
samples with 10% An component lie within the plagioclase field at approximately 10% An away from the cotectic surfaces Qz+plag+L+V and plag+alk.-feld+L+V. Winkler et al., (1975, p262) found that for samples occurring 10% An away from a cotectic surface, the beginning of crystallisation could have occurred at approximately 100°C higher than the temperatures implied from the adjacent isotherms. The beginning crystallisation for these An rich samples, by analogy with the results of Winkler et al.,(op. cit.) may have been 770-790°C.

The majority of the other samples lie within the alkali feldspar field with several of the low An component samples lying at considerable distances from the cotectic surfaces plag+alk.feld+L+V and Qz+alk.-feld+L+V. At present, An component data for the higher temperature isotherms is not known but it is possible that for the samples containing 3% and 4% An lying furtherest from the eutectic P-E (see Fig 12 a), alkali feldspar may have begun crystallising at temperatures greater than 750°C. For these latter samples, firstly alkali feldspar and then alkali feldspar and plagioclase crystallise over a considerable temperature interval prior to the appearance of quartz at approximately 660°C.

4.3.3.2 GAWBH AND SALEM GRANITOID SUITES

The normative data from Table 16 is plotted in the Qz-Ab-Or and An-Ab-Or projections (Fig 13 a and b), of the Qz-Ab-Or-An-H₂O system (Fig 11).

The GAWBH granitoid rocks plot within the plagioclase field above the cotectic surface quartz+plagioclase+melt+vapour. One of the samples (18% An) lies above the cotectic surface plagioclase+alkalifeldspar+melt+vapour. The gneissic-granodioritic members of this suite from the margin of the stock, occur the greatest distance away and the porphyritic granite members from the core of the stock, lie closest to the cotectic surface. As mentioned in the previous section, by
An Content (Wt %)

\[ P_{H_2O} = 5 \text{ kb} \]

Fig 13a

Qz-Ab-Or projection (Fig 13a) and An-Ab-Or projection (Fig 13b) showing the position of rocks from the Salem (Gn2) and Gawib (G3) granitoid suites.
analogy with Winkler et al., (1975, p262), rocks plotting greater than 10% An away from the cotectic surfaces Qz+plag+L+V or plag+alk.+feld+L+V imply that temperatures at the beginning of crystallisation are approximately 100°C higher than those indicated by the isotherms on the underlying cotectic surfaces. The rocks of the Gawib suite plot between the 650°C and 685°C isotherms (see Fig 13 a) but lie at distances of 9 to 24% above the cotectic surfaces. The gneissic-granodioritic members of the Gawib stock therefore appear to have begun crystallising at considerably higher temperatures than the granodiorite studied by Winkler et al., (op. cit.). Data are not available at present for rocks plotting such large distances away from cotectic surface to enable crystallising temperature to be predicted with confidence. However, temperatures greater than 850°C for the first appearance of plagioclase from the melt are possible.

From the position of the plots in Fig 13 the crystallisation sequence for the gneissic granodioritic rocks from the periphery of the Gawib stock can be interpreted at:

i) +850°C - crystallisation of plagioclase from the melt

ii) 675 - 680°C - crystallisation of plagioclase and quartz from melt

iii) 660°C - crystallisation of plagioclase, quartz and alkali feldspar from the melt.

The remaining rocks that occupy the core of the Gawib stock (porphyritic and non-porphyritic granite-granodiorite) lie at distances of approximately 10-12% An above the cotectic surfaces Qz+plag+L+V and plag+alk.+feld+L+V, within Fig 13, implying crystallisation of plagioclase at temperatures of about 760°C. The crystallisation sequence for these rocks is the same as that envisaged for the gneissic granodiorites but the
plagioclase crystallised over a narrower temperature interval, with quartz crystallising with the plagioclase at a temperature of approximately 660°C and then a small temperature decrease to approximately 655°C prior to the first appearance of alkali feldspar.

The SALEM granitoid suite rocks also plot within the plagioclase field (Fig 13) above the cotectic plane quartz+plagioclase+L+V but lie at a distance of 7–9% An above the cotectic plane. By analogy with the results of Winkler et al., (1975), temperatures at the beginning of crystallisation of plagioclase could be expected to be approximately 90°C above those indicated by the isotherms on the cotectic surface below the respective samples, giving a temperature of 740–770°C. The sequence of crystallisation for the Salem rocks is plagioclase, followed by quartz and plagioclase and then alkali feldspar appeared with the other two minerals at approximately 655–660°C.

4.3.4 MAJOR ELEMENTS

4.3.4.1 ABBABIS-HUSAB-GRANITE-GNEISSES AND RÖSSING ALASKITIC GRANITES

The data presented in variation diagrams (Fig 14), suggest there is a relationship between these rocks as the major element proportions of the suites and the variation between the suites is of a similar order. Although there is a scatter in some of the plots, the trends are significant considering the broad distribution of the sample localities and the heterogeneous nature of the rocks.

The ABBABIS augen gneisses have a similar major element chemistry suggesting homogeneity throughout the complete occurrence at both the Tumas River Inlier and the Rabenrücken. One of the samples (GB 200) from the Rabenrücken has significantly lower Al₂O₃ and a higher CaO content than the other augen gneisses and this may be due to alteration and the presence of epidote.
Fig. 14

Variation of oxide percentages plotted against MgO in rocks of the Abbabis (+), Husab (·), and Rössing (o) suites.
The HUSAB granites and granite-gneisses similarly have a very uniform major element chemistry. The scatter obvious in the CaO and Na₂O contents is probably due to variation in plagioclase composition and content as the high CaO samples are low in Na₂O and vice versa. Apart from the grouping of the rocks, obvious from the diagrams, at MgO 0.25 - 0.50% there are two samples of low MgO content (GB 228 & 229) and one of a high MgO content (GB 211).

GB 211 is a sample of pink foliated granite associated with the augen gneisses of the Abbabis suite. This rock has major element proportions similar to the augen gneisses and may represent a less differentiated derivative of these rocks, in comparison to the other rocks of the suite.

GB 228 and GB 229 are samples of a pale red foliated granite and a leucogranite respectively, that occur in the western part of the area, closely associated with feldspathic quartzites of the Nosib group. Their relatively low proportions of MgO, total Fe and K₂O, and high proportion of SiO₂ and Na₂O in relation to the remainder of the Husab suite may imply that assimilation of the Nosib quartzites with the Husab granitic melts has occurred.

The ROSSING alaskitic granites and pegmatites exhibit a comparatively wide variation in major element proportions. SiO₂, Al₂O₃ and K₂O show the widest scatter which is due to the coarse pegmatitic nature of the rocks, resulting from the samples being non-homogeneous.

The majority of the samples have a low MgO content. However, KH 472/3 and KH 496 are relatively enriched in MgO. These rocks show significant differences in the majority of the major element contents, relative to the other rocks of the suite. KH 256 (MgO 0.18%) has a similar 'abnormal' major element chemistry, although the MgO content is the same as the bulk of the samples. Their differences can be attributed to the high modal amounts of plagioclase and biotite.
with very low microcline in comparison to the other rocks. This may represent some degree of assimilation of the adjacent basic gneisses of the Khan Formation by the granites.

In general, major element variation in the Abbabis, Husab and Rössing suites reflects the difference in their respective mineral contents. The trends exhibited in the variation diagrams, however, do not contradict the field evidence implying that they represent a genetically associated group of rock suites.

4.3.4.2 GAWIB AND SALEM GRANITOID SUITES

The Gawib granitoids exhibit a linear variation in major-element composition (see Fig 15). The major rock types in the Gawib stock are each represented by two samples and there is very good agreement in the values obtained.

The porphyritic gneissic-granodiorite (GB 203) that occurs spatially between the extreme varieties of the Gawib rocks has intermediate major element abundances and serves as a link with the exception of relatively high K₂O and low Al₂O₃ proportions.

The changes in the major element proportions reflect the different mineral compositions but are also in agreement with a differentiation sequence. With increasing differentiation (decreasing MgO), SiO₂ and Na₂O increase and TiO₂, total Fe, CaO and P₂O₅ decrease. K₂O and Al₂O₃ remain relatively unchanged except for the 'anomalous' amounts present in the intermediate member, GB 203.

The Salem granodiorite-gneisses exhibit a relatively varied major element content (see Fig 15). One of the samples of porphyritic biotite 'granite', GB 220 is very similar to a sample of gneissic-granodiorite (GB 221) from the periphery of the Gawib stock, differing only in CaO and Na₂O content.
Fig. 15

Variation of oxide percentages plotted against MgO in rocks of the Salem and Gawib granitoid suites.

Gawib Rocks:  
- porphyritic granite
- non-porphyritic granite
- porphyritic gneissic-granodiorite
- foliated granodiorite

Salem rocks:  
+ Salem granitoids  
(present investigation)  
× Salem granitoids (Miller 1973)
The other Salem granite samples show considerable variation and differ from the trend exhibited by the Gawib granite-granodiorite suite in SiO₂, Al₂O₃ (lower) and TiO₂, total Fe and P₂O₅ (higher).

Rocks of the Salem granitoid suite from Omangambo and Otjosondjou (Miller 1973) containing similar amounts of MgO to the Salem and Gawib rocks in this area are plotted on the variation diagrams (Fig 15), for comparison. The majority of the diorite and monzonite rocks of the Salem suite at Otjosondjou are extremely enriched in MgO, and CaO and depleted in SiO₂, Na₂O and K₂O in relation to the rocks in this area and these were not plotted for comparison. The Salem rocks (calcic and potassic adamellite) of Miller (1973) exhibit very similar major element contents to the Gawib granitoid rocks rather than to the Salem rocks from this area (see Fig 15). Notable differences between these suites occur in the CaO, Na₂O and K₂O contents where CaO is depleted and K₂O enriched in the Salem rocks (Miller 1973) relative to the Gawib rocks from this area. NaO decreases and K₂O increases with increasing MgO content in the Salem rocks from Omangambo and Otjosondjou whereas the Gawib rocks exhibited an increase in Na₂O and anomalous variation in K₂O with increasing MgO content.

4.3.5 TRACE ELEMENTS

The distribution of trace elements in rocks has significance in that genetic associations between rock suites, mode of origin (anatexis, fractionation etc.) and fractionation controls may be established. Trace element distribution and substitution are considered using the following criteria:

i) Theoretical principles of behaviour based on ionic size, bond strength and electronegativity values and ionization potential as well as crystal field theory consideration.

These principals have application to distribution be-
tween solid phases and have been reviewed by Taylor (1965), after formulation by Goldschmidt (1937).

ii) Partition coefficient values for specific minerals and trace elements, based on trace element distribution preferences between solid and liquid phases of a magma.

The partition coefficient values have been determined from natural and synthetic systems in the laboratory by a number of workers, for different mineral and trace element combinations and the partition coefficient \( D \) is defined as:

\[
D = \frac{\text{concentration of the trace element in the crystal}}{\text{concentration of the trace element in the liquid}}
\]

4.3.5.1 ABBABIS-, HUSAB-GRANITE-GNEISSES AND ROSSING ALASKITIC GRANITES

The trace element abundances and variation diagrams (trace elements vs MgO) are presented in Tables 13 and 14 and Fig 16 respectively.

The ABBABIS augen gneisses have uniform trace element abundances reflecting the homogeneity of the suite that was apparent from the major element data. The minor variation in the trace element abundances can be attributed to corresponding variation in the major elements that commonly form a geochemical association with these elements i.e. \( K^+ \) and \( Rb^+ \); \( Ca^{2+} \) and \( Sr^{2+} \), (Taylor 1965) (see Fig 17). The anomalously low Sr in the Ca/Sr plot is probably due to additional Ca present from extraneous epidote.

The HUSAB granites and granite-gneisses also show a uniform trace element distribution. Although there is a considerable scatter in some of the variation diagrams, a general homogeneity is apparent.

The Husab rocks, containing low MgO (GB 228 and GB 229) differ from the remainder of the suite in that they are depleted in rubidium, niobium, zirconium and thorium. This may be attributed to major element
Fig 16

Variation in trace element concentration plotted against MgO in rocks of the Abbabis (+), Husab (.) and Rossing (o) suites.
Fig. 17

Rb vs $K_2O$ and Sr vs CaO concentrations in rocks of the Abbabis suite.

differences from the remainder of the suite, particularly for Rb as the $K_2O$ content in these rocks is correspondingly low.

The Rössing alaskitic granites exhibit a variation in trace element content which can be related to the wide range in the major element chemistry exhibited. This variation is undoubtedly indirectly due to the pegmatitic nature of the rocks and the samples, which are from boreholes and are consequently inhomogenous.
The high MgO samples (KH 472/3 and KH 496) are relatively depleted in barium and strontium and enriched in niobium, yttrium, uranium, thorium and lead, in comparison to the remainder of the samples of the suite. The remainder of the suite which have similarly low MgO contents, exhibit a wide variation in strontium, rubidium and uranium contents, but otherwise have uniformly low concentrations of the remaining trace elements analysed.

\[ \text{Fig. 18} \]

\[ \text{U/Th and Pb/Sr trace element ratios in rocks of the Abbabis (+), Husab (•) and Rössing (○) suites. (All U/Th ratios in the Rössing alaskitic granites are greater than 0.38)} \]
The concentration of barium shows progressive decrease in the Abbabis, Husab and Rössing suites, and the U/Th ratio shows a marked increase from the Abbabis to the Husab suite with the Rössing alaskitic granites having the highest values, Fig 18. This increase in the U/Th is in agreement with increases predicted during advanced fractionation, due to the weaker U-O bond than the Th-O bond (Taylor 1965) and from partition coefficient data, which for all rock forming minerals D is considerably lower than 1 (Dostal and Capedri 1975).

The K/Rb ratio is very similar as are the other trace element ratios for the different suites (see Tables 13 and 14), although marginal changes with decreasing MgO do occur.

The trace element concentrations and ratios for the majority of the Husab granites and granite-gneisses are similar to those of the Abbabis granite-gneisses, and derivation of the bulk of the Husab suite from the pre-Damara (Abbabis) granite-gneisses is a distinct possibility. Complete melting of the Abbabis suite may explain the similarity in the trace element contents as partial melting would have resulted in selective incorporation of trace elements into the melt and thus, a different trace element chemistry for the two suites could be expected. The difference in the barium content and the U/Th ratio with decreasing MgO through the suites suggests that a small amount of fractionation may have occurred.

4.3.5.2 GAWIB AND SALEM GRANITOID SUITES

Trace element abundances are listed in Table 15 and are plotted against MgO in variation diagrams, (see Fig 19). The individual members of the GAWIB granitoid suite have similar trace element contents, but variation between these members forms well-defined trends across the stock. The porphyritic and non-porphyritic granites forming the bulk of the stock have very similar trace
Fig. 19

Variation in trace element concentration plotted against MgO in rocks of the Salem and Gawib granitoid suites.

Gawib rocks:  
- porphyritic granite
- non-porphyritic granite
- porphyritic gneissic-granodiorite
- foliated granodiorite

Salem rocks:  
+ Salem granitoids (present investigation)
× Salem granitoids (Miller 1973)
element abundances as was found with the major elements. This indicates a high order of homogeneity throughout the major part of the stock.

Significant trace element variations from the high MgO, foliated granodiorite envelope to the low MgO, granite-granodiorite core across the stock are:

i) A marked increase in the strontium and lead contents.

ii) A marginal increase in the barium, rubidium and niobium content.

iii) Increasing Pb/Sr and U/Th ratios (see Fig 20)

iv) Decreasing K/Rb, Ba/Sr and Tl/Nb ratios (see Fig 20)

The contents of the uranium and zirconium are relatively constant for all members of the suite, whilst the yttrium content is slightly lower in the porphyritic and non-porphyritic granite-granodiorite core. Thorium values are also similar for the majority of the rocks with GB 203 (MgO, 1.59%) containing an anomalously high amount.

Relative bond strengths and ionic size differences between the trace elements in the ratios above result in changes with increasing fractionation (Taylor, 1965). The above variations that are exhibited between the core rocks relative to those at the margin of the Gawib stock imply that the core granite and granodiorite are more fractionated.

Evidence of the controls of fractionation during evolution of the stock can be seen from the variation in the barium, strontium and rubidium contents of the rocks, (Fig 19).

Table 17 lists partition coefficient data for the major mineral constituents that occur in the Gawib suite of rocks.
Fig. 20
Variation in trace element ratios plotted against MgO in rocks of the Gawib granitoid suite.
Gawib rocks: ⨁ porphyritic granite
○ non-porphyritic granite
● porphyritic gneissic-granodiorite
● foliated granodiorite
The foliated granodioritic rocks from the margin of the stock have the most primitive geochemistry (Fig 19). If it is assumed that the porphyritic and non-porphyritic granite of the core represent a residual melt that remained after crystallisation of the granodiorite now found at the margin, the early crystallisation of this latter member influenced the trace element distribution now found in the rocks. The dominant mineral phase in the foliated granodiorites is plagioclase with hornblende and alkali feldspar occurring in smaller but similar amounts (see Table 12). The partition-coefficient data indicates that rubidium would be retained in the melt fraction unless biotite ($D_{Rb} = 1-4$) fractionation is significant, as fractionating plagioclase, alkali feldspar and hornblende to a lesser extent, reject rubidium. The increase in the barium content in the later formed rocks however, suggests that fractionation of alkali feldspar ($D_{Ba} = 4-7$) and biotite ($D_{Ba} = 1.09-15$) has not been dominant. Fractionation therefore must have been dominated by plagioclase and/or hornblende.

Evidence from the experimental plots (section 4.3.3.2) and from petrographic studies (section 4.2.5) indicate that early fractionation of plagioclase occurred but the strontium increase in the later formed members of the stock cannot be explained by plagioclase dominated fractionation owing to the partition-coefficient values of $D_{Sr}^{plag} = 1-3$. It appears therefore that although plagioclase fractionation occurred, fractionation of hornblende, possibly at an earlier stage controlled the evolution of the stock. This is supported

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<td>0.08-0.67</td>
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<tr>
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<td>0.05-0.59</td>
<td>1-3</td>
<td>0.025-0.19</td>
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(From Phillipotts and Schnetzler (1970))
by the J-trend exhibited in the CaO plot, (Fig 21). Lambert and Holland (1974) recognise J- and L-type fractionation trends relative to a standard (calc-alkaline) trend when considering CaO vs Y contents, and conclude that for granodioritic rocks, the J-trend is due to hornblende, sphene and apatite fractionation, whereas the L-trend is due to plagioclase and/or biotite and other Ca-poor minerals fractionating.

Fig. 21
CaO/Y plot with standard (calc-alkaline) trend from Lambert and Holland (1974)
(see Fig. 19 for legend to symbols)

There is considerable variation in the trace element contents of the rocks of the SALEM granitoid suite
and there are insufficient samples to characterise typical Salem rocks.

The porphyritic biotite granite samples (GB 216 and GB 220) show significant differences in the levels of niobium, yttrium and lead present, which is an indirect reflection of their different mineralogy and major element chemistry.

The trace element content and variation in the Salem granitoid rocks from Omangambo and Otjosondjou (Miller, 1973) are shown on the variation diagrams for comparison with the Salem and Gawib suites from this area.

4.3.6 DISCUSSION

Extrapolation from the water saturated Qz-Ab-An-Or-H₂O system at 5kb indicates the Abbab and Husab granites to have a similar crystallisation sequence in that alkali feldspar (or quartz) crystallised from the melt first, with plagioclase crystallising last. The crystallisation interval (alkali feldspar-plagioclase) is small (670°-710°).

Major element, trace element and crystallisation path similarities that exist between the Abbabis and Husab granite gneisses support conclusions from field relationships that the Husab suite is derived by melting of the Abbabis suite during the Damara orogeny. This evidence appears more conclusive than that cited by Hoffmann (1976, p960), who inferred from the Qz-Ab-An-Or-H₂O system that Red Granites (Husab granite-gneisses) formed by fractional melting of Nosib metasediments.

Results from the Rössing alaskitic granite plots are possibly spurious owing to the inhomogeneity of the samples, but higher temperatures of 750°-790°C are indicated for the 1st crystallisation of the minerals. The large crystallisation interval implied from the plots for the alkali feldspar is supported by the presence of large "phenocrysts" (up to 30cm) of this
mineral in the rocks.

In the Salem and the Gawib rock suites plagioclase crystallised first from the melt over a long crystallisation interval (Gawib rocks), followed by quartz and then finally alkali feldspar. The parent melts of the Gawib granodiorite from the margin of the stock appear to have begun crystallising (plagioclase) at very high temperatures (850°C) with the granitic and granodioritic melts that crystallised in the core of the stock crystallising plagioclase at lower temperatures.

The presence of hornblende and biotite, which make up as much as 23% of the marginal granodiorites (see Table 12) may have affected the crystallisation paths of some of the rocks. For the majority, however, the constituents Qz, Ab, An and or form more than 80% (vol %) of the rock and their representation in the quaternary plots is valid.

The crystallisation sequences derived from the experimental systems are supported by petrographic evidence for the respective suites.

Estimates of pressure from metamorphic assemblages were made during this investigation (see section 3) and pressures were concluded to lie in the range 4.5-5 kb for the area of granitic occurrences. This is in agreement with results of 4 - 5 kb by Jacob (1974) for the area to the north, and comparison within the experimental system of Winkler et al., (1975) for 5 kb is therefore valid.

It is unlikely however, that magmatic processes always occur in the presence of excess water (Pivinskii and Wyllie, 1968, 1970; Luth, 1969) and therefore estimates of crystallisation temperatures from the quaternary plots, which assume water saturation are probably minimum estimates of the actual temperatures that existed at the time of granite emplacement.

Robertson and Wyllie (1971) show that considerable changes occur in the liquidus position and the liquidus-solidus interval for different degrees of water
Fig. 22

Diagram showing variation in liquidus and liquidus/solidus interval for different degrees of water saturation (after Robertson and Wyllie, 1971).

saturation in comparison to water-saturated melts, (see Fig 22). It is reasonable to equate the pre-Damaran Abbabis granite-gneisses, their Damara derivatives, Husab granite-gneisses and Rossing alaskitic granites with the water deficient and vapour absent type (Type 11) of Robertson and Wyllie (1971). Dehydration would have occurred during the pre-Damara metamorphism and granitisation of the Abbabis rocks and therefore water is likely to be present only in their hydrous minerals (biotite, muscovite and hornblende)
The Salem granitoid suite of rocks appears to have formed by anatexis of the Kuiseb metasediments as field relationships in this area and evidence elsewhere in the Damara belt (Smith, 1965; Miller, 1973 and Jacob, 1974) indicate. The metasediments from which the Salem suite originated therefore contained a small amount of water in addition to that present in the hydrous minerals and the Salem suite can be equated with the water deficient and vapour present type (Type 111) of Robertson and Wyllie (1971).

The temperatures obtained from the quaternary plots (water saturated) of 740° - 770°C for the beginning of crystallisation for the Salem rocks probably approximate better the actual temperatures achieved in this part of the Damara belt, than those indicated (670 - 710°C) from the plots of the Abbabis and Husab suites of rocks. These latter suites appear to have been considerably more dehydrated than the Salem rocks and the results from the quaternary plots which assume water saturation will, as a result, be less indicative of the temperatures achieved.

As the degree of undersaturation of the Salem rocks could not be established, the actual temperature achieved above those indicated (740° - 770°C) cannot be determined. Results from investigation of the stable metamorphic mineral assemblages indicate that for the western part of the area where the bulk of the anatectic granitoids occur, minimum temperatures of 720°C were attained (refer section 3.4). The results therefore from the two different sources show a measure of correspondence.

The minimum temperature estimate of 770°C at 5kb in the central part of the Damara belt is indicative of an exceptionally high geothermal gradient (see fig 23) in excess of 30°C/km and perhaps in the region of 40°C/km.
The Gawib granitoid stock in the eastern part of the area, from field evidence, appears to have been emplaced from depth.

With data available on crystallisation paths from plotting the mesonorm values of Qz, Ab, An and Or in the experimental system, it is concluded (with the aid of trace element data) that the Gawib stock originated by plagioclase and early hornblende controlled fractionation.

Temperatures obtained from the experimental plots for the first appearance of plagioclase from the melt are approximately 850°C, although this temperature is speculative owing to incomplete experimental data from the quaternary plots for rocks plotting away from the cotectic surfaces. The temperature indicated from the stable metamorphic mineral assemblages in the metasediments in this eastern part of the area is approximately 600°C at 3.5 - 4kb, which
is considerably lower than the minimum temperature (Gawib granitoid is probably undersaturated) indicated from the experimental plots. This is evidence in support of the field relationships which suggest the Gawib body is an allochthonous intrusive.

Comparison of the geochemistry of the Gawib and Salem suites from this area is hampered by the variable chemistry and inadequate number of samples of the Salem rocks. Differences are apparent in major and trace element contents (particularly strontium, niobium and zirconium, see Fig 19), but crystallisation paths are similar. The major and trace element chemistry of the Salem granitoid suite at Omangambo, and Otjosondjou (from Miller 1973) is very similar to that of the Gawib suite, (see Figs 15 and 19). The differentiated nature of the Salem rocks described by Miller (op.cit.) appear similar to that of the Gawib rocks. Considering the implication that the Gawib stock has probably been emplaced from depth through the pre-Damara basement, the rocks from Omangambo and Otjosondjou may have originated similarly. Isotopic studies and a geochemical investigation into the Salem granitoid suite from the type area may clarify the relationship of these rock suites.
5. **STRUCTURE**

5.1 **GENERAL**

The structural investigation was undertaken to:

i) characterise the type and style of deformation in the area with respect to the different lithologies and stratigraphic units.

ii) give an account of the symmetry and orientation of the deformation.

iii) establish the tectonic history.

Conclusions are then compared with results from adjacent structural studies undertaken in the Damara belt. Structural investigations in the Damara belt have been made in a number of disconnected areas. In the central part of the belt, adjacent to the current study area Roering (1961), Smith (1965), Nash (1971) and Jacob (1974) have defined the structural features of their respective study areas. Frets (1969), Guj (1970) and Miller (1972) have investigated separate areas several hundred kilometres to the north of the previously mentioned investigations, and adjacent to the southern boundary of the Damara belt investigations have been undertaken by De Waal (1966), Hälbic (1970), Von Groote-Bidlingmaier (1973) and Porada and Wittig (1975).

5.2 **METHODS, TERMINOLOGY AND NOTATION**

In defining the styles and phases of deformation the following procedure was adopted:

i) The macrostructure was defined by interpretation of the outcrop patterns exhibited by the different lithologies (Maps 1 and 2).

ii) Sub-areas were selected throughout the area where the deformation episodes were represented and outcrop exposure enabled quantitative measurements to be taken.
iii) Within the sub-areas (see Map 2) all planar structures (foliations) and linear structures (lineations) were measured and deformation styles were recorded. Planar structures measured include bedding surfaces, bedding foliation, fracture cleavage, slaty cleavage, jointing and schistosity. Linear structures measured were elongate transposition features, boudins, mullions, intersections of early foliations with axial plane foliations, elongate minerals and pebbles, and minor parasitic fold axes.

iv) The structural measurements were plotted on stereographic projections and synoptic plots were constructed in order to define the orientation and style of the deformation.

The notation adopted for this study is:

Bedding or bedding foliation (S₀), tectonic foliation (S₀₁, S₁, S₂ etc.), deformation (fold) phase (F₁, F₂ etc), fold axes (B₁, B₂ etc), lineations (l₁, l₂ etc).

The chronologic order of the deformation is denoted by numerals i.e. F₁ being the earliest fold phase and S₁, B₁, l₁ relate to this phase. Deformation episodes in the pre-Damara Abbadis rocks not recognised in the Damaran rocks are given the prefix A, i.e. AF₁ etc.

The term 'tectonic foliation' (S₁) is used to describe planar surfaces of metamorphic origin. These are predominantly axial plane foliations and surfaces that may have developed due to direct or indirect influence from the original sedimentary layering. e.g. where flexural slip movement parallel to bedding planes is considered to have developed on the limbs of folds. This latter foliation is termed S₀₁ as previously denoted by Jacob (1974). See table 18 for a summary.

Regional structural elements and features that are apparent from the geological mapping (i.e. macroscopic structures - Turner and Weiss, 1963) are qualitatively introduced under 'Macrostructure'. Quantitative studies undertaken at selected sub-areas are discussed under 'Mesostructures'.
TABLE 18
A SUMMARY OF THE TECTONIC FEATURES DEVELOPED IN THE DIFFERENT LITHOLOGIES AND STRATIGRAPHIC UNITS

<table>
<thead>
<tr>
<th>LITHOLOGY</th>
<th>VIRTRI BEDDING FEATURES</th>
<th>F1</th>
<th>F2</th>
<th>F3</th>
</tr>
</thead>
<tbody>
<tr>
<td>KUISEB FORMATION</td>
<td>calc-granofels/metapelite inter-layering (Tinkas) (S0)</td>
<td>fabric indirectly controlled by S0 calc-granofels/metapelite (Tinkas) migmatite (Upper) (S0) axial plane foliation S1 and minor folds S2</td>
<td>transposition foliation, axial plane slaty cleavage foliation in metapelite.</td>
<td>Large scale folding no penetrative in metapelite.</td>
</tr>
<tr>
<td>KARIBIB FORMATION</td>
<td>Impure calc-silicate recrystallisation zones and bedding (S0)</td>
<td>Alignment and stretching of clasts</td>
<td>Small and large scale folding (incl. flow folding) S2</td>
<td>Large scale folding in matrix</td>
</tr>
<tr>
<td>CHUDS FORMATION</td>
<td>Not recognised</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>EKAN FORMATION</td>
<td>Quartzite, conglomeratic layers (S0)</td>
<td>Migmatitic banding subparallel to bedding S0 (S0)</td>
<td>Polding of migmatitic layers, occasional transposition foliation (S2)</td>
<td>Large scale folding.</td>
</tr>
<tr>
<td>EKUSIS FORMATION</td>
<td>Grit and conglomeratic layers, cross bedding (S0)</td>
<td>Migmatitic development in Ns, Gn, Ns, R, (S0)</td>
<td>Fracture cleavage in Large scale folding quartzites, folding and axial plane foliation. Slaty cleavage in metapelite (S2)</td>
<td>Large scale folding.</td>
</tr>
</tbody>
</table>

| ABRABIS COMPLEX | gneissosity developed in granites augen gneisses | Folding, crenulation and refoliation of AP, migmatite and granite-gneiss development | | | |

AP4 (PRE-DAMARA) | F2 (DAMARA) |
5.3 MACROSTRUCTURES

On a macroscale the area can be divided into two distinct structural domains.

5.3.1 EASTERN STRUCTURAL DOMAIN (LUCASBERGE TO RABENRÜCKEN)

On the eastern side of the Tumas River Inlier and Glück's dome (refer Maps 1 and 2 and geological sections), the structural elements are regular and continuous, owing to the folds being of the similar type (see structural form lines on maps 1 and 2) and the folding, tight to isoclinal, (Hobbs et al., 1976, p171), with fold axes all plunging to the southwest. There is little evidence on the macroscale of interference folding having occurred in this part of the area and the structural form lines on the maps correspond to the So1 foliation having been folded during the F2 fold phase, producing the dominant NE-SW regional fabric.

Two shear zones of regional significance occur in this part of the area. On the eastern side of the tight-isoclinal folds SE of Glück's dome, stratigraphic evidence (see geological sections) and the presence of a well developed foliation in the metapelites suggests a shear zone exists here. This shear zone passes up the eastern side of Glück's dome and the northern part of the Tumas River Inlier and effectively subdivides the two regional structural domains. The eastern side has suffered downthrow relative to the western side. A smaller shear-zone with similar relative movement to the one previously described occurs on the eastern side of the Rabenrücken hills. At this locality the eastern limb of the Rabenrücken anticline has been sheared out.

A post-tectonic granitoid stock (Gawib) intrudes across the structures in the east and deflects the regional fabric slightly in the northern part. The regularity of the structures in this domain may be explained by the predominance of one rock type, the metasediments of the Kuiseb formation.
5.3.2 WESTERN STRUCTURAL DOMAIN (ENDKLIPPE TO GLÜCKS HILLS)

The structural features of this domain are extremely complex with the orientation, scale, and style of the structures showing considerable variation. The most conspicuous features (see Maps 1 and 2) are the steeply dipping NE-SW trending Damara metasediments that occur on the limbs of broad anticlines and narrow synclines. These anticlinal structures contain granites and granite-gneisses of the Abbabis and Husab suites in their cores whilst the synclines contain Salem granitoids and remnant Kuiseb metasediments. The anticlines and synclines are doubly plunging, both to the NE and SW which has resulted in closures and the development of complex dome and basin features. The structures developed are asymmetrical and the folding is non-cylindrical. Basin structures are present in the south-western part of the area but generally the broad (10-12km wide) anticlinal and domal features are separated by narrow (2-6km wide) synclines. The orientation of the intervening synclines is variable but NE-SW and E-W trending synclines predominate.

The Klein Tubasberg, Tubasberg and Saltbush synclines are major NE-SW trending synclines, whilst the Welleberge, Arcadia and Scorpion synclines represent E-W trending structures. The folds are assymetrical with the limbs of the NE-SW trending structures usually overturned to the west and those of the E-W structures overturned to the south which is undoubtedly due to non-cylindrical folding.

Within the Husaberg anticlinorium numerous narrow bands of marble occur and these possibly represent tight infolding of the Karibib metasediments. It is difficult to reconcile the outcrop pattern in this part of the area with the interference patterns of Ramsay (1967).
owing to the irregularity of the structures. The southern part of the Tumas River Inlier has similarities however, with Type 2 interference patterns of Ramsay (op. cit.) where the orientation of the respective fold phases is at right angles. (Fig 24 a & b).

Fig 24 a
Outcrop pattern in the southern part of the Tumas river area.

Fig 24 b
Ramsay (1967), Type 2 interference pattern

The dominance and continuity of the NE-SW trending structures over the E-W trending structures implies that if two periods of deformation were responsible for the structures present in the area the E-W structures were developed during an early (F₁) phase, followed by
the NE-SW \((F_2)\) phase. The attitude of the strata in the E-W synclines and the minor folding on the eastern limb of the Klein Tubasberg syncline near the Eisenhügel (refer Map 2) implies that the \(B_1\) folds were overturned to the south. The lack of continuity or regularity of these features throughout the area suggests that the intensity of the folding during this \(F_1\) phase was extremely variable.

Broad changes in the orientation of the regional fabric are apparent (Maps 1 and 2) and this may be due to a post-\(F_2\) phase of open buckling.

The complexity of the structure in the western domain appears to be due to the presence of the underlying Abbabis basement Complex. The dome structures can be described as 'mantled gneiss domes', originally described by Eskola (1949). The development of such structures can be attributed to either diapiric uprise of granitic rocks or to interference folding. The broad anticlines and domes with narrow inter-spersed synclines are similar to structures in the French Alps and in other African orogenic belts as reported by Ramsay (1967, p383). Their formation is attributed by Ramsay (op. cit.) to rocks of contrasting viscosities having been subjected to compressive stress acting from all directions or in a series of superimposed deformations. In this area the underlying basement granites and gneisses represent high viscosity material whilst the overlying metasediments would have relatively low viscosity.

5.4 MESOSTRUCTURES

Observations and measurements from outcrops within the sub-areas (see map 2) are discussed in this section, and the structural styles and orientations are characterised and documented.

The orientations of the structural elements referred to in the text are quoted as an azimuth from true North. The direction of dip is indicated in abbreviated form after the dip angle.
5.4.1 ABBABIS COMPLEX

The Abbabis Complex pre-dates the deposition of the Damara metasediments, and pre-Damara and Damara imposed structural features can be recognised.

5.4.1.1 PRE-DAMARA STRUCTURES

The predominant rocks of the Abbabis Complex, the augen gneisses exhibit a strong gneissic foliation, defined by the alignment of augen encased in biotite laminae (Plate 23). This foliation undoubtedly pre-dates the deposition of the Damara Supergroup sediments as clasts of augen gneiss are present in the Leeukop basal conglomerate (Etusis Formation), and in the Chuos Formation, (Plates 4 and 8). The orientation of this foliation throughout the Tumas River Inlier trends 040° with a variable dip, (fig 25). The structural form lines (Maps 1 and 2) within the Inlier correspond to this foliation and they conspicuously lie at an angle to the Damara trends defined by the Damara metasediments on the limbs.

It is apparent however, that this pre-Damara foliation has been affected locally in the Tumas River Inlier by Damaran folding as the foliation in the augen gneisses has been folded about an axis trending 038° and plunging 38°N (fig 25). This Damaran refoliation is evident predominantly at the margins of the Inlier adjacent to the Damara metasediments, where the foliation in the augen gneisses can often be observed to lie parallel to the contact.

Transposition of the pre-Damara foliation also occurs on the western flanks of the Witpoortberg where the Abbabis gneisses are in contact with the overlying Karibib marbles (fig 27).

The Tumas River Inlier affords the best examples of pre-Damaran structures as it appears the least affected of the basement exposures by Damaran tectonism. Damaran effects on the Abbabis rocks are discussed in more detail in the next section.
Fig. 25

94 poles to foliation (As₁) in the augen-gneisses, contours 1-3-5-8-11% per 1%.

F₂ axial plane (position inferred

* position of B₂ fold axis at northern closure of Inlier.

Structural data from the Tumas river Inlier.

Fig. 26

101 poles to foliation (As₁) in the augen gneisses, contours 1-3-9-15-35% per 1% area.

x A₁ lineations (elongate augen)

Average attitude of As₁ foliation.

Structural data from the northern part of subarea 1, adjacent to the Rabenrücken hills.
5.4.1.2 DAMARAN STRUCTURES

Pre-Damaran rocks are preserved in three separate domains within the area, namely: The Tumas River Inlier, the Rabenrücken anticline, and within the anticlinoria west of the Witpoortberge. Damaran tectonism and granitisation has affected these rocks to varying extents.

The Tumas River Inlier, as mentioned in the previous section, contains Abbabis rocks that have been least affected by the Damara tectonism in this area. Damara modification of the pre-Damara fabric are transposed foliations which occur adjacent to the pre-Damara/Damara contact, localised chevron folding in the augen gneisses (Plate 27) and anatexis/melting causing rotation of isolated augen gneiss skialiths (Plate 28).

The orientation of the chevron folding where observed in the Tumas River Inlier is variable but commonly trends NNW-NNE, and shearing has been observed in zones parallel to the axial planes of these folds. U-Pb dating on zircons from one of the homogeneous granite bodies (Gn1) within the Tumas River Inlier
(Jacob pers. comm) has given a minimum age of 572 m.y. The deformation which has modified the surrounding augen gneiss fabric by folding, shearing, boudinage of calc-silicate/amphibolite remnants, rotation of augen gneiss blocks in anatetic mobilisates and has affected this granite is therefore certainly Damara in age.

On the eastern side of the Rabenrücken hills, Abbabis augen gneisses exposed in an anticlinal core of a B2 fold have a well-developed AS1 foliation. The eastern limb of this fold is absent and has apparently been sheared out. The structural data from this subarea (1) are plotted in figs 26 & 28, and are similar in orientation to data obtained from measurements on the overlying Nosib group rocks (Jacob, 1974, Fig 15a, b, c).

The AS1 fabric in the augen gneisses appears to be folded in the southern part of this subarea and the A1 lineations also show evidence of rotation (Fig 28). This folding is probably due to the Damara F2 phase as the data from the overlying Nosib metasediments is very similar. The coaxial nature of the fabric in the Abbabis and the Damara rocks however, does not allow for simple interpretation and the possibility exists that the folding of the AS1 fabric in this sub-area is of pre-Damaran age.

The Abbabis rocks that occurred in the anticlinoria to the west of the Witpoortberg have now largely been refoliated or melted during the Damara orogenesis and the resultant granites and gneisses are now classified as the Husab granite-gneiss suite. Augen gneisses occur in localised areas, (Witpoortberg as mentioned previously) where their original AS1 fabric can still be recognised. Other remnant Abbabis rocks that can be recognised are strongly sheared augen gneisses where the AS1 fabric no longer predominates, although augen can still be recognised disrupted by a Damara gneissic foliation (S1). The
Damara biotite-rich granite-gneisses, leucogranites and red homogeneous granites derived from the Abbabis rocks have an $S_1$ fabric that has been folded during the $F_2$ phase of deformation, as is evident on a macroscale from the structural form lines (see maps 1 and 2.)

Fig. 28

49 poles to foliation ($A_{S_1}$) in augen gneisses, contours 1-3-5-8% per 1% area

$\times$ $A_{L_1}$ lineations (elongate augen)

Structural data from the southern part of subarea 2 adjacent to the Rabenrücken hills.
5.4.2 DAMARA METASEDIMENTS

Structures discussed in this section occur in the Damara metasediments and were developed during the Damara orogenesis. Bedding features that have been preserved are also discussed.

5.4.2.1 BEDDING STRUCTURES

Bedding features that have not been modified by a tectonic foliation are restricted to the cross-bedded quartzites (Etusis Formation) occurring adjacent to the Chameleon hills (Plate 6). Elsewhere bedding surfaces (So) are recognisable by lithologic changes such as:

i) conglomerate/quartzite - Etusis Formation

ii) quartzite/banded gneiss - Khan Formation

iii) marble/impure marble - Karibib Formation

iv) calc-granofels/pelite - Kuiseb Formation

These however have largely been modified by a tectonic foliation (So1) which occurs parallel to the bedding. This foliation is considered to have been caused by flexural-slip movement along the bedding planes during the first period of deformation F1. This lithologic-controlled banding in the Swakop/Khan area is considered by Jacob (1974) to represent possible axial planar foliation (S1) in places where folding was isoclinal.

The preservation of original bedding structures is possibly due to the buffering effect of the adjacent underlying basement rocks and the competent nature of the quartzitic and arkosic metasediments. The F2 folding in these rocks consequently is of the open buckle (concentric) type. Ripple-mark-like structures occur at the contact of the Tinkas calc-granofelses. These are not considered to be of sedimentary nature but to represent minor folding and buckling, as the intensity of metamorphism and folding in the adjacent beds would have disrupted any primary sedimentary structures.
5.4.2.2  \( F_1 \) STRUCTURES

The earliest tectonic surfaces developed in the area are the banding (\( S_{01} \)) in both the Khan and Kuiseb (Tinkas Member) formations. (The development of migmatites and the fabric in the Husab and Salem suites is also considered to have developed during this phase.) As discussed previously, the banding in the metasediments is considered to be caused by Flexural-slip movements subparallel to bedding during the \( F_1 \) deformation, and in many places cannot be distinguished from bedding. Minor folds formed in the metasediments during this phase of deformation have only been recognised in the Karibib marbles (Plate 9) and in the Tinkas calc-granofels bands where minor \( B_1 \) folds can be recognised occurring in the \( B_2 \) fold closures (Plate 29). Recognition of these features on the limbs of \( B_2 \) folds is difficult owing to distortion of the early structures resulting in a coaxial attitude. In sub-area 2 at Bonfire Gorge, small scale recumbent \( B_1 \) folds and folded axial planar foliations (\( S_1 \)) were recognised in the interbedded sequence of calc-granofelses and meta-pelites (Kuiseb Formation, Plates 30 and 31). These \( F_1 \) features are preserved in the hinges of minor \( B_2 \) folds and measurements of \( B_1 \) axial planes and lineations (\( B_1 \) fold axes) indicate that these folds are coaxial but have flat lying folded axial planar surfaces perpendicular to those of the later \( B_2 \) folds (fig 29).

The coaxial nature of this folding is possibly due to the extension of the \( B_1 \) folds parallel to the \( B_2 \) fold axes during the \( F_2 \) deformation phase, as evidence on a macroscale indicates an EW orientation for the development of \( B_1 \) folds. Conversely, the orientation of the \( F_1 \) phase of deformation in the east may have a different orientation to that found in the remainder of the area.

On the limbs of the Tubasberg syncline occupied by
Salem granitoids (adjacent to 15°00' longitude) minor folding within the Karibib marble bands appears to represent B_1 folds as the folds here plunge to the north, whereas the Tubasberg syncline (F_2) plunges to the south. At sub-area 3 measurements of bedding (So) and bedding foliation (S_o_1) of minor folds in the marbles (Plate 9), indicate fold axes plunging 47° in a north-easterly direction (Fig. 30). With unfolding of the B_2 fold stereographically (Fig. 31) the orientation of the B_1 fold would lie in an ESE-WNW direction.

Fig. 29

35 poles to axial planar surfaces of minor B_1 folds in metapelites (Kuiseb Formation), contours 3-6-9-14% per 1% area.
x 1 lineations (minor B_1 fold axes).
Structural data from subarea 2 (Bonfire Gorge).
Fig. 30

36 poles to bedding (So) and bedding foliation (So₁) in Karibib marbles, contours 3-6-8-11% per 1%.

× 1 lineations (minor B₁ fold axes)
/ inferred position of B₁ axial plane
Structural data from subarea 3.

Fig. 31
Stereographic construction utilizing data from subarea 3 and the Tubasberg syncline.
Within the limbs of the V-shaped marble occurrence (von Stryk's anticline) fold closures appear to be of F₁ age and lineations defined by the minor B₁ fold axes plunge at ± 20° to the south.

5.4.2.3 F₂ STRUCTURES

The dominant NE-SW trend of the structures throughout the area is due to the F₂ deformation episode. The folds produced during this deformation are of the open-concentric type in the competent quartzites and arkoses of the Etusis Formation, and of the similar-isoclinal type in the Tinkas metapelite/Calc-granofels sequence. In the Etusis quartzites the bedding/foliation (So/S₀₁) is disrupted by an S₂ fracture cleavage (Plate 32). This fracture cleavage, minor B₂ folds and associated lineations (l₂) are the main mesoscopic F₂ features developed in these rocks.

In zones of intense deformation the banded gneisses of the Khan Formation generally exhibit transposition of the S₀₁ fabric in the B₂ folds parallel to B₂ axial planes (Plate 33), but this S₀₁ fabric (banding) is generally folded in a flexural-slip manner (Plate 34). Fracture cleavage (S₂) is also apparent in these rocks.

The Swakop Group metasediments exhibit varied response to the F₂ deformation. In sub-area 4 the three predominant rock types; marble, calc-granofels and meta-pelite can be observed together (fig 32). The marble bands have resorted to flexural-slip/flow parallel to the bedding surfaces and in these bands the S₂ fabric is only developed at the nose of folds where minor transposition parallel to the S₂ axial planes occurs. The metapelite bands have a prominent S₂ fabric developed parallel to the axial planes of the B₂ folds and transposition of the earlier foliations is complete. The calc-granofels bands in response to deformation are transposed parallel to S₂ axial planes on the B₂ fold limbs (Plate 35), but often only an S₂ axial planar fracture cleavage is developed in the B₂ fold hinges (Plate 29). Transposition is marked where the B₂
folds are isoclinallly folded, adjacent to the south-eastern closure of Glücks dome. Further to the east (Lucasberg, Rabenrücken) transposition parallel to $S_2$ is weakly developed and is often only found in the metapelite bands. A lineation ($l_2$) can often be observed developed at the contact of calc-granofels bands and the metapelites due to the intersection of the $S_2$ axial planar foliation in the metapelites and the $S_{01}$ surface of the calc-granofels. Minor $B_2$ folds on the limbs of larger structures are often developed in the calc-granofels, along with related mullion and boudin structures ($l_2$), (Plates 36 and 37).

![Schematic diagram showing varied response to deformation in Swakop group metasediments at subarea 4.]

In the areas of more intense deformation, shearing along fold limbs parallel to the axial planes has occurred. This is recognisable in the field by a well developed slatey cleavage in the metapelite.
and calc-grano-fels bands where the original \( S_{01} \) fabric is completely obliterated, e.g. on the eastern side of the Glücks dome. On a macroscale this is evident from stratigraphic and structural discontinuities, i.e. Rabenrücken, Orlog and in the SW corner of the area (see geological sections). The orientation of the structures developed during the \( F_2 \) deformation was determined by quantitative measurements taken at the different sub-areas (see map 2). In the east (sub-area 2) adjacent to Bonfire Gorge, the \( B_2 \) folds in the Kuiseb metasediments plunge to the SSW (200\(^\circ\)), Fig 33. The spread of the \( S_{01} \) poles (fig 33) indicates that the folding here was not tightly isoclinal but more open (Plate 37). The axial plane of the \( B_2 \) folds dips 80\(^\circ\)W.

![Diagram](image)

Fig. 33
- 104 poles to bedding foliation (\( S_{01} \)) and axial plane foliation (\( S_1 \)) in Kuiseb metasediments.
- 30 poles to axial planes of minor \( B_2 \) folds and axial planar foliation (\( S_2 \)).
- 1\(_2\) lineations (mullions, minor \( B_2 \) fold axes intersection of \( S_1 \) and \( S_2 \) surfaces.

Structural measurements defining the \( B_2 \) folds at subarea 2 (Bonfire Gorge).
Measurements from the Kuiseb formation metasediments at the Lucasberge (sub-area 5) indicate B2 folds plunging 40° to 231° with an axial plane dipping 80°W (fig 34). The distribution of So1 poles in the plot indicates open type folding, although the structural form lines on a macroscale indicate similar type folding (see maps 1 and 2).

The tightly folded Karibib marbles and Kuiseb metapelites and calc-jranofelses (sub-area 4) on the eastern side of Glücks dome occur in a tightly folded syncline plunging 50° towards 220° (fig 35). The axial plane dips steeply at 85°W and the distribution of the So1 poles in the plot is indicative of tight isoclinal folding. The quartzites exposed in Glücks dome were investigated in two sub-areas, one in the south and one in the north. At the southern closure (sub-area 6) the quartzites have a flat lying attitude and they are folded about a B2 axis plunging 22° towards 210°, (fig 36). The plunge of the folds decreases northwards. The axial plane (fig 36) dips steeply to the east (82°) and the distribution of the So1/So poles in the plot indicates concentric/buckle type folding. The quartzites at the northern closure of Glücks dome (sub-area 7) are folded about a B2 axis plunging 25° to 065°, (fig 37) with an axial plane dipping to the west. The trend of the axial surface (S2) at the two closures of Glücks dome differs. This is possibly due to either a later F3 deformation event trending WNW-ENE or to the proximity of the underlying basement.

In the Zebraerbeuge, at the southern closure of the Ida dome (sub-area 8) Khan Formation gneisses have been folded about a B2 axis, plunging 28° towards 232° (fig 38). The axial plane of these folds dips 80° to the NW (fig 38).

In the Welleberge (sub-area 9), banded gneisses of the Khan Formation (Ns2Gn) occur in an overturned
Fig. 34

- 100 poles to bedding (So), bedding foliation (So₁) in Kuiseb metasediments.
- 27 poles to axial plane cleavage/foliation (S₂)
- 12 lineations (So₁/S₂ intersection, mullions, minor fold axes)

Structural measurements defining B₂ folds at subarea 5 (Lucasberge).

Fig. 35

73 poles to bedding (So)/bedding foliation (So₁) in Kuiseb metasediments and Karibib marbles, contours 1-3-7-10% per 1% area.
- 35 poles to axial plane foliation (cleavage (S₂)
- 12 lineations (mullions, minor fold axes, So₁/S₂ intersections).

Structural measurements from subarea 4
Fig. 36

97 poles to bedding ($S_0$) in Etusis quartzites and metapelites, contours 1-3-6-9-13% per 1% area.
- 32 poles to axial plane fracture cleavage ($S_2$)
- $l_2$ lineations (minor $B_2$ fold axes, $S_0/S_2$ intersections.)
Structural measurements from the southern closure of Glück's dome (subarea 6)

Fig. 37

41 poles to bedding ($S_0$) in Etusis quartzites, contours 1-2-5-8% per 1% area.
- 9 poles to axial plane fracture cleavage ($S_2$).
- $l_2$ lineations (minor $B_2$ fold axes, $S_0/S_2$ intersections.)
Structural data from subarea 7 (northern closure of Glück's dome).
syncline. The \( F_2 \) folding here is of very open buckle type and the regional foliation (\( S_0_1 \)) trends E-W. Lineations (\( L_1 \)) defined by hornblende growth on \( S_0_1 / S_1 \) banding/foliation indicate an early phase of deformation orientated E-W, which has subsequently been modified by a broad buckling during \( F_2 \) deformation. The plunge of the \( F_2 \) fold is to the NE and is therefore overturned which is undoubtedly due to the orientation and dip of the early axial surface.

Fig. 38

- 72 poles to banding (\( S_0_1, S_1 \)) in Khan gneisses
- 35 poles to axial planar fracture cleavage (\( S_2 \)) and to axial planes of minor \( B_2 \) folds.
- 12 lineations (elongate minerals, \( S_0_1 / S_2 \) foliation intersection, minor \( B_2 \) fold axes.)

Structural data from subarea 8 (Zebraberge).
30 poles to bedding (So) foliation (So₁) in Karibib marbles and Kuiseb metasediments.
• 13 poles to axial planar cleavage (S₂) (slatey cleavage in metapelites and fracture cleavage in calc-granofels).
× 1₂ lineations (mullions and minor B₂ fold axes).

Structural data from subarea 10.

South of the Endklippe hills in the western part of the area (sub-area 10) B₂ folds measured in interbedded Karibib marbles and Kuiseb metapelites/calc-granofels plunge steeply at 68° towards 200° (fig 39). The axial plane dips steeply at 85° to the east.

A synoptic plot of the Π points and lineations (l₂) measured in the respective sub-areas and elsewhere in the area is shown in fig. 40. From this plot it is apparent that the orientation of the B₂ folds is
Fig. 40

average axial plane (S₂) attitude for each respective subarea
× l₂ lineations (minor B₂ fold axes, mullions, boudins, elongate minerals, S₀₁/S₂ intersections).
○ △ points from the different subareas.
Synoptic plot of structural data from throughout the area.

Fig. 41

31 poles to joints in Etusis quartzites.
Probable B₃ fold axial direction.
Measurements of jointing at subareas 6 and 7.
NE-SW with near vertical axial planes and the majority of the folds measured plunge to the southwest.

5.4.2.4 F₃ STRUCTURES

A variety of non-penetrative post-F₂ structures have been observed. These rarely exceed several metres in scale and their orientation appears to coincide with the broad buckling (F₃) apparent on a macroscale. Adjacent to the northern closure of Glücks dome in the metapelites of the Kuiseb Formation (sub-area 11) localised disruption of S₂ axial planar slaty cleavage was observed and minor folding (Plate 38) occurs about an axis trending between 115° and 125°. This disruption of the S₂ cleavage is in the form of vertical kinking and jointing within the metapelites. In the quartzites of Glücks dome (sub-areas 6 & 7) minor buckling of the bedding/bedding foliation (S₀/S₀₁) was observed without a penetrative axial planar foliation developed. The approximate axis of this buckling was orientated at 120°. Joints developed in these rocks may, however represent fracture parallel to the axial planar direction in this phase of deformation (see below).

5.4.2.5 JOINTING

In the quartzites of Glücks dome (sub-areas 6 & 7) joints are present throughout which have a fairly constant orientation of 120° (fig 41). These joints appear to represent an axial plane fracturing in response to broad F₃ phase buckling. Horizontal joints are developed in the Khan Formation gneisses at the south western extension of the Ida Dome. These may have developed at a much later period due to compensation from unloading during erosion.

5.4.2.6 FAULTS

Fault zones trending NS occur in the central part of the area and are recognised by the presence of red quartz-vein breccia and chert. They can be traced
for up to 10km and in the central part of the area at the Witberg a vertical displacement of approximately 50 metres can be recognised. The faults post-date the Damara folding but pre-date the intrusion of the Karoo dolerites, which can be observed cutting the fault zones.

5.4.3 DAMARA GRANITOIDS

The granitoids occurring throughout the area can be classified into syn and late to post-tectonic suites. The Rössing alaskitic granites and the Gawib granitoid suite represent late to post tectonic granitoids and the Husab and Salem suites are syn-tectonic granitoids.

5.4.3.1 SYN-TECTONIC GRANITOIDS

The syn-tectonic granites are those of the Husab granite-gneiss suite and the Salem granitoid suite as discussed previously. The rocks of the Husab suite exhibit a complex structure and the foliation developed in these rocks has been folded during the \( F_2 \) deformation phase as is evident from the structural form lines on maps 1 and 2. This implies that these granite gneisses and the associated migmatites of the Lower Nosib group were developed during the \( F_1 \) phase of deformation.

Varieties of the Husab suite occurring as narrow (± 2m) dykes show distinct late-tectonic features. Dykes of red homogeneous granite are observed as contorted remnants in the Rössing Alaskitic granites (Plate 24) and elsewhere red and grey dykes cross-cut both the Husab, Abbabis and Salem suites.

The Salem granitoid suite of rocks also invariably exhibits a strong foliation (Plates 25 & 26) whenever seen in this area, although weakly to non-foliated
varieties also occur. In the Tubasberg syncline a penetrative foliation was observed that was not confined to the margins but which appeared to be conformable with the lithologic contacts at the limbs of the syncline. The predominant rock type in this suite is the porphyritic biotite granite-gneiss and the K-feldspar phenocrysts in these rocks were commonly observed to have rounded and sheared margins. This suggests that the foliation is tectonic and not due to flow. The presence of the strong tectonic fabric, which is conformable with the margins of the syncline (particularly the Tubasberg syncline) suggests that this granitoid suite developed at an early stage (F₁) and the foliation (S₁) was further deformed during the F₂ phase. The porphyritic leucogranite member of this suite commonly has crosscutting features and is not foliated which suggests emplacement postdated the F₂ deformation.

5.4.3.2 LATE- TO POST-TECTONIC GRANITOIDS

The GAWIB granitoid body occurs as a large stock which has been intruded across the regional F₂ fabric and is non-foliated with the exception of a narrow envelope surrounding the central core. This foliation within the granodiorite envelope is parallel to the margin of the stock and is considered to be related to the intrusion rather than to a tectonic foliation. The core of the stock is characterised by large unoriented K-feldspar phenocrysts. The ROSSING alaskitic granites occur in the vicinity of the Zebraberge as coarse pegmatitic bodies associated with the metasedimentary rocks of the Khan and Karibib Formations. Narrow granite/pegmatitic veins or large (+ 1 km) ovoid granitic/pegmatitic/masses often crosscut the F₂ structural fabric or occasionally develop along F₂ axial planes (Plate 34). Leucogranitic and pegmatitic bodies that exhibit a foliation and which are folded also occur, but these rocks are generally associated with the granites and
gneisses of the Husab suite into which they have been grouped.

5.5 DISCUSSION

Evidence suggests that two main deformational events, $F_1$ and $F_2$ occurred in this area, producing the NE-SW fabric with minor development of E-W trending structures. Post $F_2$ deformation appears to have been broad open buckling of the early structures with the axis of folding oriented in an ESE-WNW direction.

In order to develop the dome and basin features in the central part of the Damara belt previous writers are divided in their interpretation as to the orientation of the deformation phases involved (fig 42).

![Diagram]


There is general agreement on the orientation of the $F_2$ deformation phase as this was the most intense and is responsible for the strongly developed NE-SW fabric. Smith (1965) considers that 1st folds were oriented NW-SE and thus interference between this phase and the following $F_2$ phase produced the irregular pattern of dome and basin structures.

Jacob (1974) stated that the early plunging $F_2$ folds become more pronounced during the later part of this
deformational period and where anatexis occurred, migmatitic uprise and differential flattening produced the dome structures, which were locally accentuated during the F$_4$ deformation phase. The F$_1$ structures in the area studied by Jacob (op.cit.) are coaxial with F$_2$ and thus dome and basin structures are considered unlikely to have formed from interference between these deformation phases.

Evidence for an E-W trend of the structures produced during F$_1$ is as follows:

i) E-W trending Scorpion, Welleberge and Arcadia synclines.

ii) Complex E-W oriented reclined folding on the eastern limbs of the F$_2$ Klein Tubasberg-syncline, south of the Eisenhügel.

iii) Tight infolding in the marbles on the limbs of major Tubasberg syncline.

It is apparent from the Welleberge, Arcadia and Scorpion synclines that the B$_1$ folds were overturned to the south. Interference between the F$_1$ developed structures and the later B$_2$ north-easterly oriented folds appears to have formed the complex dome and basin features. The complexity of the structures in the west may also have been influenced by the presence of the underlying basement granites and gneisses primarily due to the viscosity difference between these rocks and the overlying metasediments and due to diapiric uprise which may also have occurred.

In the east, the basement rocks are present only at depth, and there is a greater regularity in the orientation and plunge of the folds. The lack of any major post-F$_2$ deformation phase precludes an interference of F$_2$ and post-F$_2$ structures accentuating the previously developed domes and basin as suggested by Jacob (1974).
Smith (1965) describes the B₂ folds as predominantly plunging NE, whereas in the area studied by Jacob (1974) the B₂ fold axes were found to plunge in both a NE and SW direction. In this area, the majority of the B₂ folds, particularly in the east, plunge in a SW direction (Fig 44). This implies that between the area to the north (Smith, 1965) and the present area, either large scale buckling may have occurred post-dating the F₂ deformation phase, or the F₁ structures in the Khan/Swakop area may have been overturned to the north, with interference between the F₁ and F₂ fold phases also producing a variation in the direction of plunge of the F₂ structures throughout this part of the orogen.
6. SUMMARY AND CONCLUSIONS

The Damara geosynclinal sediments present in this area are shallow water sediments that were deposited in a high energy environment and in isolated basins.

The sedimentation sequence began with psephitic material (Leeukop conglomerate) deposited in restricted basins and continued with deposition of the psammitic sediments of the Etusis Formation. These sediments are invariably crossbedded and often interbedded with conglomeratic zones.

In the west, deposition of semi-calcareous sediments (Khan Formation) followed. A period of erosion probably terminated the deposition of the Khan sediments and glacial conditions ensued during which sporadic glacio-marine deposits of diamictite (Chuots Formation) accumulated.

A sequence of carbonate sediments (Karibib Formation were subsequently deposited which, towards the east gradually became interlayered with pelites (Kuiseb Formation). In the extreme east of the area deeper water conditions apparently existed as there is no accumulation of carbonates and pelitic to semi-calcareous sediments (Tinkas member-Kuiseb Formation) are present.

Deeper water conditions ensued throughout the area after the carbonate sedimentation and pelitic sediments (Witpoort member-Kuiseb Formation) accumulated over a very large area.

The pre-existing Abbabis Complex in the form of a prominent landmass, influenced strongly the deposition of the early Damara sediments. The extent and thickness of the Nosib Group metasediments is extremely variable and the direct deposition of carbonate sediments (Karibib Formation) onto the Abbabis rocks as seen at the Witpoortberge, and the incorporation of clasts of Abbabis rocks into
the Chuos diamictites implies that a ridge existed nearby, up until the beginning of deposition of the Karibib sequence.

The Damara sediments in this area are similar to those reported to the north by Smith (1965), Nash (1971) and Jacob (1974). Psephitic sediments however, appear to be better developed in this area.

The Damara geosyncline was later subjected to intense deformation with accompanying metamorphism and granitization. Two main deformation events appear to be responsible for the structures observed today. The first phase of folding was orientated WNW-ESE and flexural slip movements during recumbent and overturned folding occurred. The axial planes of the folds dip to the north. The second phase of folding produced the dominant NE-SW regional fabric and in the east isoclinal and open type folds were developed. In the central and western parts of the area, interference between the \( F_1 \) and \( F_2 \) fold phases have formed the complex dome and basin structures. The asymmetrical structures of broad anticlines with narrow interspersed synclines imply that the proximity of, and possible coincidental diapiric uprise of the underlying basement (Abbabis) rocks may have contributed to the complexity of the structures as observed today.

The Abbabis rocks have not been significantly affected by the Damara orogenesis except where remelting has occurred. The original pre-Damara fabric \( (AS_1) \) that is preserved in clasts of Abbabis augen-gneiss and granites present in the Damara metasediments, has been modified locally by shearing, kink folding and transposition, particularly at the margins of the Abbabis occurrences where Damara effects have been greatest. The Tumas River Inlier occurrence of pre-Damara rocks comprising predominantly augen-gneisses, has been least affected by Damara tectonism and granitization in this part of the Damara belt and can be equated with the Abbabis Inlier near Karibib, which has now been dated at 1925 ± 300my by Jacob et al, (In press).
The metamorphic events appear to have reached a peak during the $F_1$ and $F_2$ deformation phases and amphibolite facies mineral assemblages were developed.

The metamorphic grade increases from medium-grade assemblages in the east to high-grade assemblages in the west. The anatexis-in-gneiss boundary occurs on the eastern margin of the Tumas river Inlier and to the west of this anatetic melts were developed.

The mineral assemblages present in the metasedimentary rocks throughout the area enable temperature and pressure estimates to be made by making comparisons with experimentally derived phase equilibria. In the eastern part of the area, the absence of staurolite, the presence of andalusite and the coexistence of muscovite and quartz in the metapelites is indicative of temperatures of approximately 600°C at 3.5-4 kb. Grossularite, quartz, anorthite, and calcite, with minor epidote are found in the calc-granofelses and, together with the absence of forsterite in the marbles, stability fields of these mineral associations agree with the data obtained from the metapelites.

Almandine garnet and cordierite coexist in the mineral assemblages of the metapelites and forsterite is present in the marbles on the eastern limb of the Tubasberg syncline. These minerals are significant in that minimum temperatures of 675°C at 4-4.5 kb must have been exceeded. The highest temperatures and pressures indicated by the mineral assemblages in the metasediments of this area are 720°C at 4.5-5 kb. These figures are implied by wollastonite and anorthite coexisting in the Khan gneisses in the west.

In the comparison with experimentally derived phase equilibria use was made of the position of the andalusite-sillimanite phase boundary as determined by Althaus (1969).
Andalusite was replaced by sillimanite in the mineral assemblages in this area before the anatexis-in-gneiss boundary was reached and this supports the results of Althaus (op.cit.) rather than those of Richardson et al. (1969).

The mineral assemblages in the metapelites, calc-granofelses and marbles in this area are essentially similar to those established by Jacob (1974) in the area to the north.

The lower PT estimates for the eastern part of this area in comparison to the eastern part of the area studied by Jacob (op.cit) are due to the absence of staurolite in this area and the acceptance of the equilibrium curve of Althaus (1969) for andalusite-sillimanite inversion as opposed to the equilibrium curve of Richardson et al (1969), used by Jacob (op.cit.).

The Abbabis augen-gneisses in the Tumas River Inlier and also relict zones that occur further to the west were undoubtedly in existence in their present form prior to the deposition of the Damara sediments. The Husab granite-gneisses, however, underlie the Nosib metasediments, and derivation from the Abbabis rocks by partial and complete remelting is apparent from field relationships. Derivation of the Husab and Salem suites appears to have occurred at an early period of the Damara orogeny as the gneissic fabric in these suites has been deformed during the F₂ phase.

Late stage mobilisates however, form part of both suites in the form of leucogranites. The Rossing alaskitic granite constitutes a large mass of late-phase mobilisate derived from the Husab and Abbabis melting. The age of these granites has been placed at approximately 470 my. (Kröner and Hawkesworth, 1977).
Geochemical data on melting paths and crystallisation temperatures were obtained by plotting normative values on quarternary plots in the system Qz-Ab-An-Or-H₂O after the method of Winkler et al. (1975). The temperatures obtained for the beginning of crystallisation for the Salem and Husab suites are 740⁰-770⁰C and 670⁰-710⁰C respectively.

The Salem granitoid suite derived from geosynclinal sediments probably more closely approximate the water-saturated conditions assumed in the quarternary plots and values of 740-770⁰C are quite realistic for the beginning of crystallisation of this suite. The Husab suite in comparison was probably undersaturated, having undergone a metamorphic episode during Abbabis times and the values from the quarternary plots are likely to be lower than those in nature.

Major-element, trace-element and crystallisation-path similarities existing between the Husab and Abbabis suites support field evidence that the former suite was derived predominantly by melting from the latter and not from anatexis of Nosib metasediments.

The Gawib granitoid suite which occurs in the form of a stock in the eastern part of the area, is a differentiated body apparently emplaced from depth. Evolution of the stock appears to be controlled by plagioclase and hornblende fractionation with crystallisation beginning at temperatures greater than 850⁰C.

There are similarities in field appearance between the Gawib and Salem granitoid suites but the attempt to establish geochemically any links or differences was inconclusive.

Differences are apparent between the major- and trace-element chemistries of the two suites from this area, but crystallisation paths are similar.
A study of the geochemical character of the Salem granitoid suite from the type area would be a useful research project in that meaningful comparisons could then be made between the granodiorite bodies occurring throughout the Damara belt.

The Salem granitoid rocks from Omangambo and Otjosondjou (Miller, 1973), have a similar major and trace-element chemistry to the Gawib rocks from this area and in field setting are of a similar differentiated nature. This implies that either the Salem and Gawib suites are both autochthonous or the Salem suite at Omangambo may be allochthonous and related to the Gawib granitoid suite.

The presence of a large landmass of pre-Damara rocks supplying clastic sediments from the centre of the Damara geosyncline and the occurrence of high-temperature metamorphism and subsequent granitization in this zone at a later period, indicates that the central part of the orogenic belt underwent rapid subsidence during the latter phase of Damara sedimentation. The occurrence of pre-Damara (Abbabis) basement rocks underlying the Damara metasediments in the central part of the geosyncline implies that the remainder of the geosyncline is also underlain by continental rocks. The existence of underlying continental crust and the presence of shallow water sediments deposited at the beginning of the Damara sedimentation cycle in the centre of the Damara belt is more in accordance with continental destruction and in-situ deformation (Shackleton, 1973, 1976), Krüner (1977), than with intercontinental development due to continental collision (Burke and Dewey, 1972).
APPENDIX

A MODAL ANALYSES
Modal estimates for the majority of the thin sections investigated were determined by direct comparison with the charts of Terry and Chillinger (1955).

For the specimens that were analysed by X-Ray Fluorescence (Granitoids) the modal compositions were determined using a Swift automatic point counter. For these specimens 800 to 1000 counts per section were carried out. In the case of porphyritic varieties, rock slabs were stained using Sodium cobaltinitrite after the method of Lyons (1971), and the modal compositions were determined on duplicate slabs under the binocular microscope using the manual Cawood counter developed by Mr D. Cawood of the Rhodes University Geology Dept.

B WHOLE-ROCK CHEMICAL ANALYSES
1. SAMPLING AND PREPARATION
Most of the samples were collected from outcrops in the field and the remainder are quartered core samples from exploratory boreholes.

The samples were cleaned with distilled water and then reduced to 2cm chips by means of a rock splitter and then further reduced in size by passing through a jaw crusher with hardened steel jaws. The samples were then quartered to 500g-1kg and passed through the swing-disc mill (Cr steel) for 25 secs where the chips were reduced to a fine powder. Sufficient powder for producing glass fusion discs, whole rock pellets and for FeO determination was then taken and ground by hand with an agate and mortar to -300 mesh.
Whole rock pellets were made using approximately 5gm of the powder and it was found necessary to use a few drops of binding solution in each pellet to prevent shearing, which is common in powders with a high SiO₂ content. 12 ton p.s.i. was found to be adequate for formation of the whole rock pellets.

The preparation and analytical methods used are those advocated by Norrish and Hutton (1969), and currently adopted at the Rhodes University Geology Department.

2. ANALYTICAL PROCEDURE

The major elements and the trace elements Rb, Sr, Y, Nb, Zr, Th and Pb were determined by the writer using X-Ray Fluorescence techniques on a Phillips PW 1410 spectrometer.

The trace elements Ba and U were determined by the Anglo American Research Laboratories, Johannesburg. Ba was determined by atomic absorption analysis and U was determined by X-Ray Fluorescence analysis after the method of Feather and Willis (1976).

2.1. MAJOR AND TRACE ELEMENTS

The elements Si, Al, Ti, Ca, Mg, K, Mn, P and Total Fe were determined on the glass fusion discs whilst Na and all the trace elements were determined on the whole-rock powder pellets.

The analytical conditions used and the values of the various standards employed are set out in the following Tables (I, II and III).
Table (1)

Instrumental conditions for analysis of major elements.

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<th>Sample Preparation</th>
<th>Fusion Disc</th>
<th>Whole rock pellet</th>
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<td><strong>Element</strong></td>
<td>Si  Ti  Al  Fe  Mn  Mg  Ca  K  P  Na</td>
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<td><strong>KV Generator MA</strong></td>
<td>50  50  50  50  50  50  50  50  50  50</td>
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<tr>
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<td><strong>Standards</strong></td>
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(Pulse height discrimination applied)
Table II
Instrumental conditions for analysis of trace elements

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<th>SAMPLE PREPARATION</th>
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<td>220</td>
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<tr>
<td>TIME</td>
<td>100 sec on background, 200 sec on peak</td>
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<tr>
<td>POSITION</td>
<td>Kα, Kα, Kα, Kα, Kα, ThLα, PbLα</td>
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<td>BLANKS</td>
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</tr>
<tr>
<td>STANDARDS</td>
<td>USGS international standards, NIMROC standards and UCT in house standards (See Table III for standard values used)</td>
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Pulse height discrimination applied. Full corrections for background, live interference and mass absorption applied.

Table III
Standard values employed in the determination of trace elements (values in ppm).

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<thead>
<tr>
<th>ELEMENT</th>
<th>Nb</th>
<th>Zr</th>
<th>Y</th>
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<th>Rb</th>
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<td></td>
<td>G-2</td>
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<td>324</td>
<td>10.7</td>
<td>492</td>
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<td>UCT in</td>
<td>OK272</td>
<td>74.5</td>
<td>164</td>
<td>29.0</td>
<td>113</td>
<td>88.0</td>
<td>-</td>
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<tr>
<td>house Stds</td>
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<td>199</td>
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<td>212</td>
<td>-</td>
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<td>NIMROC Standards</td>
<td>NIM-G</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>50</td>
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</tbody>
</table>
FeO was determined after the method of Shapiro and Brannock (1962) using 0.5gm of powder which was digested in a mixture of $\text{H}_2\text{SO}_4$ and HF and this solution was titrated against a standard potassium dichromate solution. Each sample was analysed in duplicate.

3. PROCESSING OF DATA

The data from the X-Ray spectrometer printout was incorporated into several computer programmes adopted for use at the University Computer Centre by Dr. J.S. Marsh of the Geology Dept., whereby the resultant major trace element values, mass absorption coefficients and mesonorm values were obtained.

C DETERMINATIVE MINERALOGY

Plagioclase compositions for the different rocks and rock suites were determined using the Universal stage and the method of Rittmann (1929). Curves given by Tröger (p.129, 1971) were used. In the case of zoned crystals an average value between the core and the rim was taken.

D STRUCTURAL INVESTIGATION

Measurements of the structural data were made using a Brunton compass either directly on linear and planar surfaces where the character of the outcrop allowed, or on a small 15cm by 15cm hardboard square wedged or placed onto the respective surfaces being measured.

Structural data from the field were plotted to Lambert equal area stereonets and contoured where necessary using specially prepared plots.
REFERENCES CITED


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Additional References


Plate 1
Typical Namib landscape of broad plains with inselbergs.

Plate 2
Contorted metasedimentary remnants of Abbabis age lie within highly deformed augen-gneisses (AGn₁) in the Tumas River Inlier.

Plate 3
Basal conglomerate (Leeukop Member) of the Nosib Group with large clasts of red and grey granite, leucogranite, quartzite and occasional biotite schist.
Plate 4
Augen-gneiss, quartzite, leucogranite and amphibolite clasts occurring within the Leeukop metaconglomerate (Nosib Group).

Plate 5
Interbedded coarse conglomeratic beds and fine grained feldspathic quartzites at the eastern side of the Rabenrücken.

Plate 6
Cross-bedded Nosib feldspathic quartzites on the eastern flank of the Tumas River Inlier adjacent to the Chameleon Hills.
Plate 7
Typical migmatitic banded gneiss of the Khan Formation showing a boudinaged garnet-diopside-hornblende band adjacent to quartz-feldspar and pelitic layers.

Plate 8
Chuos tillite west of the Rabenrücken showing an augen-gneiss clast associated with smaller granitic and quartzitic clasts in a pelitic matrix.

Plate 9
Pale Karibib marble with impure zones (dark) composed of calc-silicate minerals, which probably reflect original bedding layers. The folding of the unit as seen here is considered to have occurred during the $P_1$ tectonic episode.
Plate 10

Typical Tinkas member (Kuiseb Formation) sequence of interbedded pelitic schist (dark) and calc-granofels bands (pale).

Plate 11

Photomicrograph of intrusive gabbro showing serpentinized olivine (ol) surrounded by orthopyroxene (opx), hornblende (hb) and plagioclase (plag) (Specimen No X74).

Plate 12

Needles of sillimanite and quartz occur as inclusions within alkali feldspar in Etusis quartzites adjacent to the Chameleon Hills, (Specimen No.144).
Plate 13
Fibrous aggregates of sillimanite present in feldspatic quartzites (Etusis Formation) west of the Chameleon Hills.

Plate 14
Alkali feldspar grains containing numerous small biotite inclusions. The alignment of the inclusions in the feldspar grains is not parallel to the dominant fabric in the rocks, shown in this section by the alignment of the larger matrix biotite grains. (Witpoort meta-pelite, Specimen No. 905).

Plate 15
Large unorientated porphyroblasts of hornblende in a calc-granofels band, (Tinkas Member/Kuiseb Formation) indicating post-deformational growth. (Specimen No. 7611).
Plate 16
Transposition of biotite flakes formed during a previous deformation. (Tinkas metapelite, Specimen No.19211).

Plate 17
Light coloured elongate syntectonic andalusite (and) present in a well foliated biotite rich matrix. (Metapelite - Tinkas Member/Kuiseb Formation). (Specimen No.X123).

Plate 18
Pinotized cordierite grains (cord) coexisting with alkali-feldspar (K.feld) in biotite schist. (Specimen No.X61).
Plate 19

A large cordierite porphyroblast, containing unorientated biotite inclusions, occurring in a fine grained well foliated biotite rich matrix. There is no distortion of the matrix biotites adjacent to the porphyroblast. (cordierite schist—Tinkas Member/Kuiseb Formation).
(Specimen No. 1731).

Plate 20  (bottom plate)

Coexisting andalusite (and) and cordierite porphyroblasts in a metapelite (Tinkas Member/Kuiseb Formation)
(Specimen No. 1731).

Plate 21  (centre plate)

K feldspar coexisting with cordierite (partly pinotized) biotite and quartz. (Specimen No. X61).
Plate 22
Banded gneiss of the Khan Formation, containing intergrowths of plagioclase (plag) and wollastonite (wo).
(Specimen No. 17001).

Plate 23
Typical augen-gneiss of the Abbabis Complex, showing strongly aligned feldspar augen in a biotite matrix.

Plate 24
Contorted red homogeneous granite dyke (G1) situated within a large mass of Rössing alaskitic pegmatitic granite.
Plate 25

Strongly aligned feldspar phenocrysts typical of the grey porphyritic Salem granitoids.

Plate 26

Dark non-porphyritic granodiorite/gneiss associated with porphyritic Salem granite/granodiorite. A narrow dyke of pale leucogranite can be seen crosscutting the other granitoids in the background.

Plate 27

Chevron folding (Damara) of the Pre-Damara fabric in augen-gneisses in the Tumas River Inlier.
Plate 28
Blocks of augen-gneiss occurring within a homogeneous granite and showing evidence of reorientation (Tumas River Inlier).

Plate 29
Damara (S2) axial planar cleavage developed in a calc-granofels band (fracture cleavage) and the underlying pelitic schist (slatey cleavage). Minor B1 folds can be recognised within the calc-granofels band.

Plate 30
Minor B1 folds occurring in the hinge of a B2 fold with Tinkas metasediments at Bonfire Gorge. The B1 folds have been deformed along the axial planes of the B2 fold.
Plate 31
B2 fold closure in interbedded calc granofels/metapelite bands (Tinkas Member/Kuiseb Formation). S1 axial plane foliation can be observed in the metapelitic band which has been folded during the F2 fold phase.

Plate 32
S2 axial planar fracture cleavage developed in Etusis quartzites east of Witkop at the closure of a B2 fold.

Plate 33
Transposition of S01 banding in Khan gneiss occurring parallel to the S2 axial planar direction of the B2 folds. This transposition is commonly seen in zones of more intense deformation.
Plate 34

Folding of S01 banding, commonly seen in the Khan gneisses in Hollands anticline, where the flexural slip folding has occurred without transposition to the S2 axial plane direction. A narrow alaskitic granite veinlet (G4) has developed parallel to the S2 axial plane direction and quartz has accumulated in low pressure zones at the hinges of the B2 folds.

Plate 35

Transposition of a calc-granofels band of Tinkas Member (Kuiseb Formation) occurring parallel to S2 axial planar foliation in the surrounding metapelite on the limb of a B2 fold.

Plate 36

Minor B2 folds developed in a calc-granofels band of the Tinkas metasediments on the limb of a larger B2 fold east of the Lucasberg Mullion structures (12) are also developed.
Plate 37  (bottom plate)
An open B2 fold showing mullions (12) developed in a calc-granofels band within the Tinkas metasediments at Bonfire Gorge. S2 axial planar cleavage can be seen in the metapelite layer in the core of the fold.

Plate 38  (top plate)
A minor B2 fold developed in Tinkas metapelites east of Glück's Hills. Gentle buckling appears to have occurred at right angles to the B2 axis and was probably developed during the F3 phase.