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

by

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of Master of Science at Rhodes
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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 .....[signature].....

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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 ($> 850^{\circ}$C) obtained from quarternary Qz-An-Ab-Or-H_2O 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 kb 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 kb.

Two tectonic events were responsible for the regional structure. An early F_1 episode produced east-west oriented overturned folds and was followed by an intense F_2 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 F_3 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.

CONTENTS

PLATES

CONTENTS OF POCKET: Geological Map (Map 1)
                    Structural Map (Map 2)
                    Geological Sections

ABBREVIATIONS

(used in text and diagrams)

Ab	– albite	Qu,Qz,Qtz	– quartz
All	– allanite	Scap	– scapolite
An	– anorthite	Sill	– sillimanite
And	– andalusite	Sph	– sphene
Ap	– apatite	Stt	– staurolite
Cc	– calcite	Tm	– tourmaline
Czt,Clz	– clinozoisite	Tr	– tremolite
Chl	– chlorite	Wo	– wollastonite
Cord	– cordierite	Zeol	– zeolite
Diop	– diopside	Zir	– zircon
Do	– dolomite	Zo	– zoisite
Epid	– epidote	L	– liquid
Fo	– forsterite	V	– vapour
Gnt	– garnet	km	– kilometre(s)
Gr	– grossularite	m	– metre(s)
Hb	– hornblende	cm	– centimetre(s)
K.feld	– alkali feldspar	mm	– millimetre(s)
Kya	– kyanite	kb	– kilobar(s)
Law	– lawsonite	T	– temperature
Musc	– muscovite	P_t	– pressure(total)
Opx	– orthopyroxene	P_f	– pressure(fluid)
Or	– orthoclase	P_1	– pressure(load)
Plag	– plagioclase	D	– distribution coefficient
Preh	– prehnite	my	– million years
Pump	– pumpellyite	p.s.i	– pounds per square inch
		secs	– seconds

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 AGn_1.

ACKNOWLEDGEMENTS

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 $22^{\circ} 42'$ and $23^{\circ}02'$ and longitudes $14^{\circ}46'$ and $15^{\circ}20'$ (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.
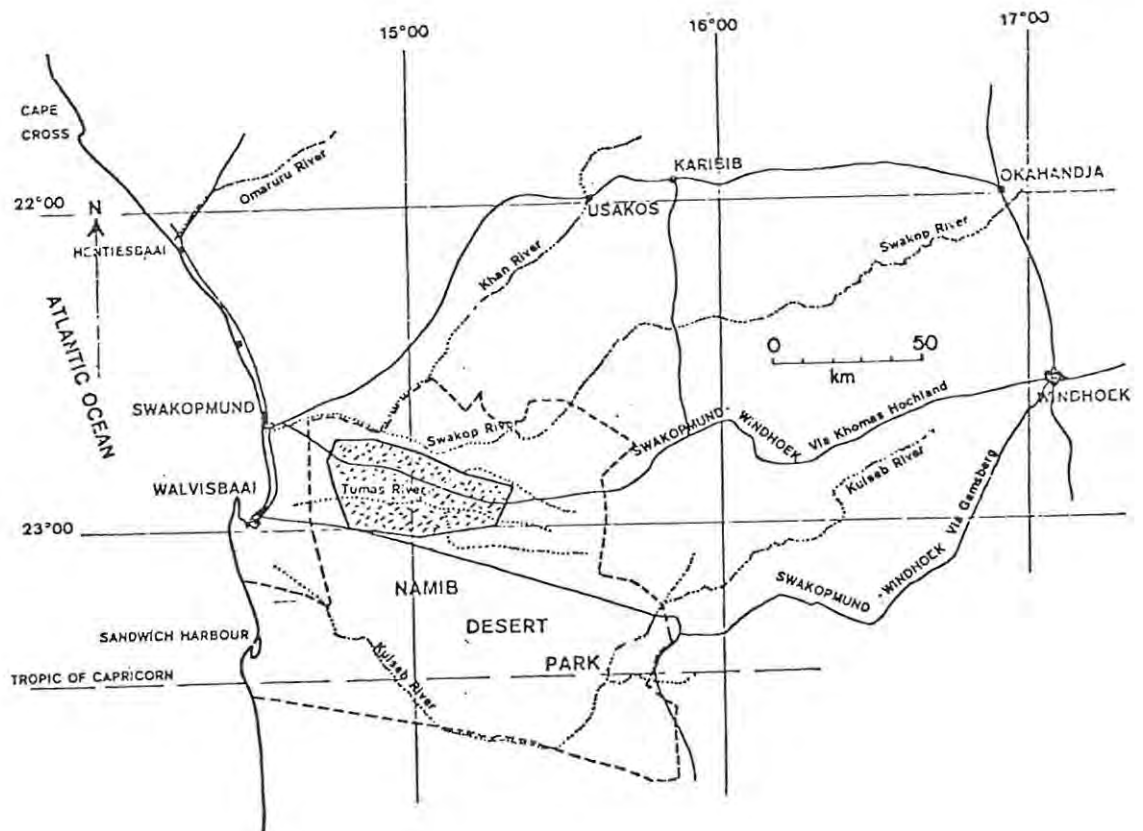
Fig. 1

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, seperately 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.

I Smith (1965)

2 Jacob (1974)

3 Faupel (1974)

4 Nieberding (1976)

5 Present Investigation

☐ Post Damara

Central Granite
Zone
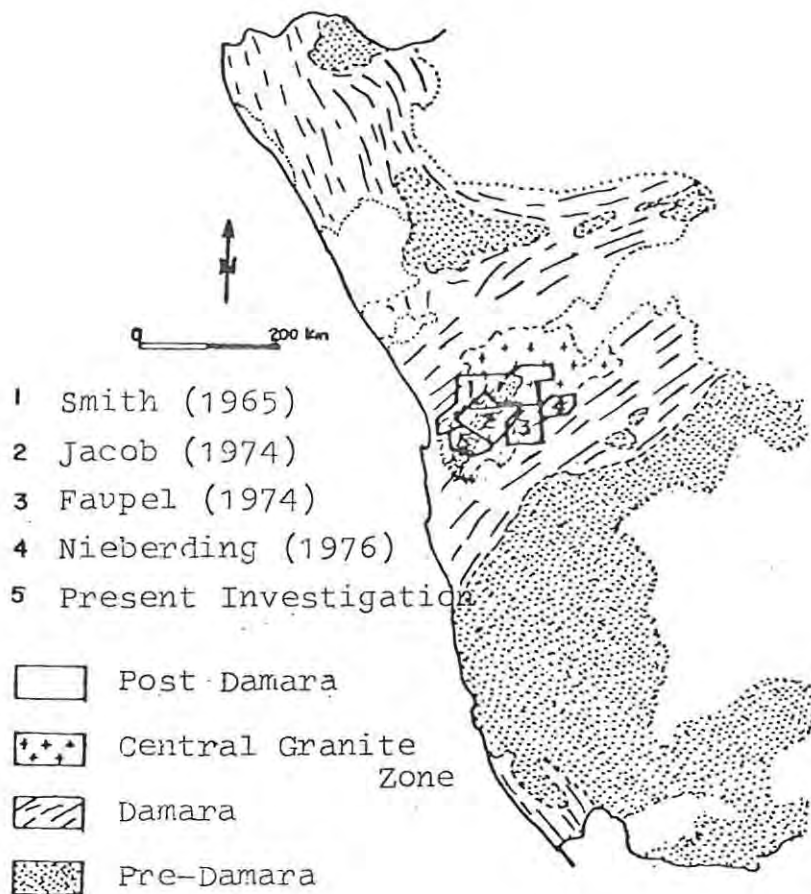
Damara

Pre-Damara

Fig. 2
Map showing the relative location of recent investigations
undertaken in the Central Granite Zone of the Damara
belt.

1.3   PRESENT INVESTIGATION

The current study was begun whilst the writer was em-
ployed 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 per-
formed 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 inter-
vals on foot. Subsequent follow-up work was then under-
taken on areas requiring more detailed investigation.

The field information was recorded on 1 : 25 000
field sheets, the data from these then being in-
corporated onto a 1 : 50 000 geological map along
with data from aerial photograph interpretations
done in conjunction with the field mapping. Labo-
ratory studies at Rhodes University involved petro-
graphy, geochemistry and structural interpretations
and were undertaken using techniques listed in the
Appendix.

It is the aim of the study to present a detailed geo-
logical account of the area with emphasis on the
basement (Abbabis)/cover (Damara) relationships.
This is considered under topics such as field rela-
tionships, 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 gradu-
ally 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 adop-
ted 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 undu-
lating 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 dolo-
mite (Karibib Formation) most commonly form the in-
selbergs, 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 vicini-
ty 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 For-
mation) 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 collison) 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-miogeo-
synclinal 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 com-
plexity of the rock associations causing unnecessary
confusion in the terminology being adopted. The ter-
minologies 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 classifi-
cation 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 METASEDIMENTS

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

Stratigraphic terminology adopted for the Damara Super Group Rocks.

TABLE 1

8

SMITH (1965)			JACOB (1974)			S.A.C.S. (1976)			PRESENT INVESTIGATION		
KHAN/SWAKOP AREA			KHAN/SWAKOP AREA			CENTRAL DAMARA BELT			TUMAS RIVER AREA		
SERIES	STAGE		SUBGROUP	FORMATION		SUB-GROUP	FORMATION		SUB-GROUP	FORMATION	
KHOMAS	(D_2)	DAMARA SYSTEM	DAMARA GROUP	KHOMAS	DAMARA SUPER-GROUP	SWAKOP GROUP	KHOMAS	DAMARA SUPER-GROUP	SWAKOP GROUP	KUISEB	(Tinkas member-E) KUISEB (Witpoort member-W (D_4S)
				TINKAS $(D_4S$ (east) WITPOORT $(D_4S$ (west) TINKAS $(D_4S$ (east)							
HAKOS	UPPER (D_1U)			HUSAB $(D_3C$ (west)			KARIBIB			KARIBIB (D_3C)	
	CHUOS (D_1M)			CHUOS (D_2G)			CHUOS			CHUOS (D_2G)	
	Trans. O/lap			Unconformity (D_1C)			Discordance			NOT OBSERVED	
	LOWER (D_1L)			RÖSSING (D_1Gn)		UGAB	DOME GORGE				
Conformable Transit.			Para-conform./unconform			Unconf. or conf/Trans			Paraconformity		
NOSIB FORMATION	UPPER (NS_1Q)		NOSIB GROUP	KHAN (NS_2Gn)		NOSIB GROUP	KHAN		NOSIB GROUP	KHAN (NS_2Gn)	
	LOWER (NS_1Q)			ETUSIS (NS_1Q/NS_1Gn)			ETUSIS			ETUSIS (NS_1Q) (Leeukop member $NS_1R)$	
MAJOR UNCONFORMITY											
ABBABIS FORMATION						ABBABIS COMPLEX			ABBABIS COMPLEX (AGn_1)		

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

TABLE 2

			FORMATION	MEMBER	LITHOLOGY
DAMARA SUPER-GROUP	SWAKOP GROUP	KHOMAS SUBGROUP	KUISEB	WITPOORT	biotite schist/gneiss, migmatite ± cordierite, garnet, sillimanite
				TINKAS	biotite schist ± cordierite, andalusite, sillimanite, quartz-biotite granofels
			KARIBIB		dolomitic and calcitic marble, calc-granofels, biotite schist
			CHUOS		tillite (diamictite) -schistose matrix with granitic, gneissic and metasedimentary clasts
		UGAB SUB-GROUP	DOME GORGE		not recognised in the area under investigation
			PARACONFORMITY		
	NOSIB GROUP		KHAN		banded amphibole-pyroxene gneiss partly migmatitic;±hornblende, diopside,scapolite,wollastonite pyritic biotite schist,quartzite conglomerate
			ETUSIS		pink feldspathic quartzites, biotite schist, gneiss,conglomerate ±sillimanite
				LEEUKOP	conglomerate,migmatite,impure quartzites ±hornblende,diopside, scapolite
			MAJOR UNCONFORMITY		
			ABBABIS COMPLEX		augen-gneiss,granite,calc-granofels,amphibolite,biotite schist and gneiss

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 lar-
gely 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, consis-
ting of migmatites, granites, granite-gneisses, peg-
matites 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. <u>METASEDIMENTARY ROCKS</u>

2.1 <u>LITHOLOGIES AND FIELD RELATIONSHIPS</u>

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 (AGn_1)

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
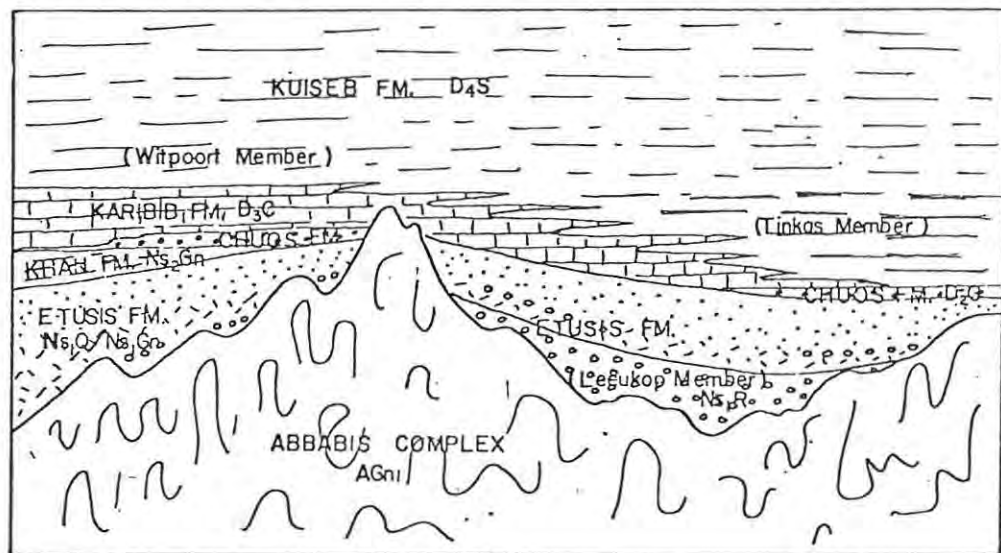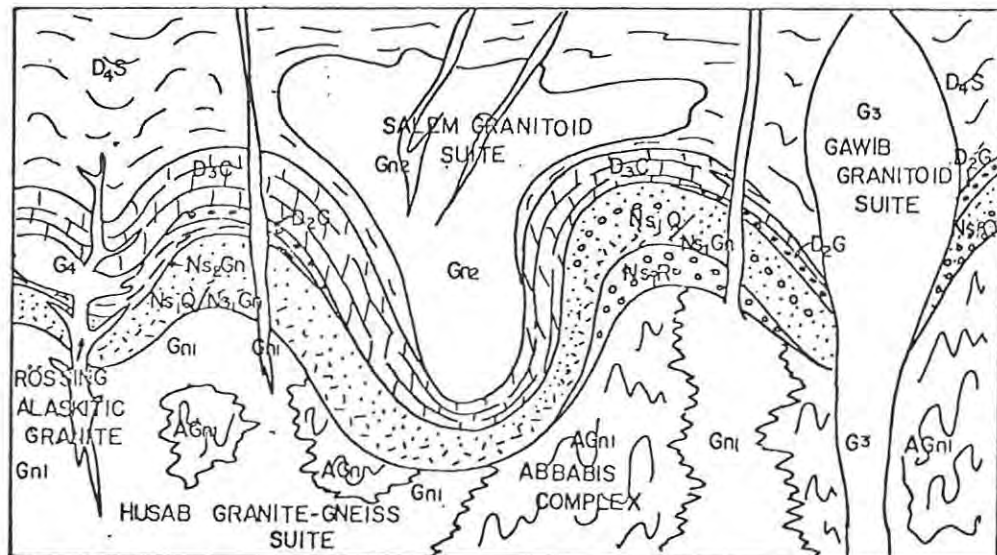Damara metasediments (refer to structure, Chapter 5).

Fig 3a

Fig 3b

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 $15^{\circ}00'$ 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 F_1 phase of deformation (refer chapter 5).

2.1.2 THE NOSIB GROUP OF THE DAMARA SUPERGROUP

2.1.2.1 ETUSIS FORMATION (Ns_1Gn, NS_1R, NS_1Q)

The principal rock types are psammitic and psephitic 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. Glücks 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 relation-
ship is unconformable, although immediately adjacent
to the contact, the foliation in the underlying base-
ment 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 deve-
lopment 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 relicts from basement augen gneisses and
pegmatites, often form small grit sized clasts. The
matrix is migmatitic and semi-pelitic in nature, con-
taining the minerals quartz, biotite, diopside, horn-
blende and epidote. These rocks constitute an easily
recognisable field unit and the term LEEUKOP MEMBER

(NS_1R) is used to distinguish this local basal ruda-
ceous 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 600m and on the wes-
tern 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 -
NS_1Gn). This is particulary apparent in the core
of Glücks 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 mig-
matitic gneisses developed, exhibit classical migma-
titic 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 conglo-
merate 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
Glücks 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 approxi-
mately 500 - 600 metres.

2.1.2.2.   KHAN FORMATION   (Ns$_2$Gn)

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 con-
stitute 90 per cent of the unit and form an easily
identifiable field unit due to their grey-green coloura-
tion 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 diop-
side and hornblende 'spots' are surrounded by leu-
cocratic 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 occa-
sionally 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 re-
stricted 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 $15^{\circ}00'$ longitude,
where the Karibib marbles have a 20-30° angular re-
lationship to the underlying Etusis (Leeukop) conglo-
merate. 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 cal-
careous/argillaceous Swakop group, and also on the

apparent observance by earlier workers of an uncon-
formity present in the northern part of the Damara
belt.

2.1.3.1 CHUOS FORMATION (D$_2$G)

The rocks of this formation are metamorphosed dia-
mictites, 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), Hälbich (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 out-
crops 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 boul-
ders 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 ex-
posures supports a glacio-marine origin for these
rocks.

2.1.3.2. KARIBIB FORMATION (D_3C)

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 inter-
bedded with schist are developed locally at the base
of massive marble bands.

iii) Immediately west of Glücks Dome the sequence
consists of thin marble and schist bands at the base,
followed by a thick massive marble unit with inter-
bedded 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 Forma-
tion.

iv) Immediately east of Glücks Dome the formation
is represented by narrow (1 - 20m) (relative to the
massive marbles in the west) marble bands only, inter-
bedded 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 (D_4S)

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 Rabenrücken. 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 occuring 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 $23^{\circ}00'$ latitude at 8-10km west of $15^{\circ}00'$ 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 boul-
ders. 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, bio-
tite, $\pm$ calcite $\pm$ clinopyroxene $\pm$ ore minerals. The
olivine grains, which in part are now altered to ser-
pentine, are surrounded by successive rims of ortho-
pyroxene, hornblende and biotite (Plate 11), indica-
tive of prolonged reaction of the early formed crystals
with the melt. A basic dyke that occurs within the
Husab suite adjacent to $15^{\circ}00'$ 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 pro-
bably 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 Caino-
zoic 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 ter-
races 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. Un-
consolidated 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 approxi-
mately 70km inland from the Atlantic Ocean. The de-
velopment of gypsum in this zone has been attributed
to the presence of the sulphur laden mist that deve-
lops 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 investiga-
tions in the area to the north, are essentially simi-
lar 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 paleoen-
vironment 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 na-
ture 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 en-
vironment 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 addi-
tional 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 conglome-
rate occurs as a distinct unit over a fairly restricted
area ($\pm$ 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 geo-
synclinal floor. The restriction of the metasediments

of the Khan Formation to an area west of where the
Marmor Pforte-Husabberg-Witpoortberge now lie, indi-
cates 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 sub-
sided 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 Raben-
rücken and Langer Heinrich areas. Subsidence of the
ge-anticlinal ridge must have continued until even-
tually 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 sig-
nificance in establishing metamorphic grade, the posi-
tion 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 tech-
niques and visual comparison with the abundance dia-
grams 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 folia-
tion due to the sub-parallel alignment of hornblende
and biotite grains. These rocks contain quartz, pla-
gioclase, hornblende, biotite, epidote and ore mine-
rals with accessory amounts of retrograde clinozoi-
site. Clinopyroxene is noticeably absent. The grain
size in these rocks is 0,03 - 6mm with hornblende
generally forming large dark blue-green to green poiki-
loblastic 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, com-
monly associated with hornblende. Sphene has been
observed as small subidioblastic grains occurring as
inclusions aligned parallel to the cleavage in horn-
blende grains. They also occur scattered throughout
the sections. It is noticeable that sphene and bio-
tite do not occur in the same rocks, indicating com-
petition for titanium ions between these minerals.
Ore minerals are generally xenoblastic in form and
are distributed throughout the sections.

The texture of the calc-granofelses 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 diop-
side, 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 exclu-
sively, 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 com-
monly saussuritised.

TABLE 3

MODAL ESTIMATES (Vol %) OF METASEDIMENTS FROM THE ABBABIS COMPLEX.   (a = accessory amounts<1%)

SAMPLE NO	QUARTZ	PLAGIOCLASE	HORNBLENDE	DIOPSIDE	BIOTITE	EPIDOTE	MUSCOVITE	SPHENE	CALCITE	ORE	APATITE	OTHERS	LOCALITY
AMPHIBOLITES/HORNBLENDE-GNEISSES													
83	30	45	3		12	2		a			2	Zir(a),All(a) Czt(a)	L17
X53	10	40	25	10	a					7	3	Chl(2)	H16
X135	15	15	50		a	3		9		6		Chl(2)	L16
X136	10	35	30	20	a					5	a		L16
X138	30	40	15					3		5	3	Zir(a)	L16
+X132	50	15			30			a		2	a		M28
CALC-GRANOFELS/GNEISSES												All(a)	
84	40	35	4	5	3			3	2		a	Chl(a),Zir(a)	L17
132	70	5	a		1		2	2	5	1		Gnt(2),Scap(3) Chl(a),Czt(3)	N16
153	20	35		25				7	2			K.Feld(10)	O16
X39	35	35	5	3	3			3	2		a	Chl(2),Zir(a)	O16
X133	55			2	17	2		a			2	Chl(3),Zir(a)	M28

+ porphyroblastic biotite gneiss

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. Clino-zoisite 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 NS_1R) 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 - 10mm) 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 - 10mm) 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 ($\pm$ 5mm) 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$_1$R) AND QUARTZ—BIOTITE SCHIST —(NS$_1$GN) FROM THE ETUSIS FORMATION (a=accessory amounts <1%)

SAMPLE NO	QUARTZ	PLAGIOCLASE	K.FELDSPAR	BIOTITE	DIOPSIDE	HORNBLENDE	EPIDOTE	SCAPOLITE	CALCITE	SPHENE	ORE	OTHERS	LOCALITY
META CONGLOMERATES (LEEUKOP MEMBER) (NS$_1$R)													
Q1	35	7		20	6	3		15	4	4	3	All(a)	M13
Q2	35	20		10	3	7			4	4	3	garnet(10), All(a)	N13
94	45	5		5		7	10	10	7	2		Ap(a)	L16
148	30	20		15	1	10		15	3	3	a	garnet(11),Wo(a)	O13
X22	35		45	a		10					4	Chl(a)	K16
X36	25		35	20		10	2			a	5	All(a),Chl(a), Zir(a)	R16
X37	12		35	20	7	7	3		a	2	6	Musc(a),Chl(1), All(a)	Q17
X41	45	5	10	10		10	12		3	3	a		L14
META-PELITES (QUARTZ-BIOTITE-SCHIST/GNEISS NS$_1$Gn)													
7001	40		45	10							2	Zir(a)	O5
7002	45	5	40	5			a				1	Musc(a),Chl(a), Ap(1)	O4
8005	40	5	15	30	5				3			Zir(a)	N4
19001	30	25		15		8	2		10	2	3	Musc(a),Chl(a) Ap(1)	I5
X29	60	15	5	10					a	a	a	Musc(a),Chl(4)	M28
1361	50	5	20	15								Musc(4),Chl(1) Sill(1)	N21
1541	60	5	15	6							1	Musc(8),Chl(2)	O20
1602	25	25	30	7			a				3	Musc(3),Chl(3), Zir(a)	J7
1803	55	10	20	12								Musc(a),Chl(1), Zir(a)	I8
2004	50		40	7					1				H8
2204	50	25	5	15								Musc(1),Chl(2) Zir(a)	G9
2401	40	35	5	12							7	Zir(a)	F6

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 irregu-
lar and embayed boundaries, and they tend to be dis-
tributed 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, migma-
tite)-Ns$_1$Gn, occurring near the base of the Etusis
Formation have a simple mineralogy.

The most common minerals are quartz, plagioclase, al-
kali-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 dis-
tinct leucosome layers. The grain size of the schist
is small (maximum 2mm), with coarser crystals occur-
ring 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%)

SAMPLE NO	QUARTZ	PLAGIOCLASE	ALKALI FELDSPAR	BIOTITE	MUSCOVITE	EPIDOTE	ZIRCON	SILLIMANITE	CHLORITE	ORE	OTHERS	LOCALITY
96	70	5	20	a	3					a		L15
144	80	a	5	5	3			6		1	Ap(a)	O15
147	50		40	5	1					3	Ap(a)	O14
661	60	a	35		2		a			2		K22
1151	65	a	30	3	1			a		a		M19
1221	80	1	5	5	5		a			1	Ap(a),Tm(a)	N22
1351	80	3	4	5	7				a	1		N21
1531	85	7	a	3	3			a				O20
11311	65	1	10			5	2			2	hb(12)	M28
12111	50	20	10			a	a			a	hb(17)	N27
X20	55	5	35	3	a		a			2		K16
X28	75	10	5		2				4	1	cc(3)	M28
X34	50	20	15	6	3		a		2	a	cc(a)	S17
X35	70	5	a	5	6	3	a		2	a	cc(2)	R16
X38	75		10	a	7			6		a	hb(a)	Q17
X40	45		45	5	1					a		O16
X82	75	2	3		10				3		cc(3)	Q20
X83	80	1	4		3	5				a	cc(5)	P20
X84	70	10	5	a	9		a		3			Q20
X85	60	15	5	3	7	a			4	a	cc(2)	P19
X86	70	5	5	5	7			a	3		cc(a)	P19
X87	80		a		10				4	a	cc(4)	P19
X88	75	5	10	3	3			a	2			P19
X90	60	15	10	2	6				4	a		P18
X91	75	10	2		10					1	cc(2)	Q15

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 in-
clusions in biotite flakes. Sillimanite (present as
fine needlelike inclusions in quartz grains), diop-
side, 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_1Q 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 sub-
divided 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 poly-
gonal 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 under-
gone sericitisation. Quartz and feldspar are often

TABLE 6

MODAL ESTIMATES (VOL %) OF METAPSAMMITES (QUARTZITES, FELDSPATHIC QUARTZITES, MICACEOUS QUARTZITES) FROM THE ETUSIS FORMATION WEST OF THE TUMAS RIVER INLIER (a=accessory amounts < 1%)

SAMPLE NO	QUARTZ	PLAGIOCLASE	ALKALI FELDSPAR	BIOTITE	MUSCOVITE	CHLORITE	ZIRCON	ORE	OTHERS	LOCALITY
1201	50	5	40		a			2		L3
2005	80				1	1		12	Ep(a),cc(a),hb(2), Diop(2)	H9
2202	70	15	5	a	a		a	5		G8
5003	60		25	7	3			a		P4
6001	55		30	7	1		a	2		O5
6003	65		25	2	2		1	3		O4
10001	90		1	6	a		a			M5
10005	55	35	a	a	1	a	a	3		M3
12001	40	30	15	1	3	a	1	4		L4
18001	30	30	20	3			1	7		I4
19004	45	10	40		a			5		I3
X10	80		10	3	1	a			cc(2)	F1

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

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 (maxi-
mum 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 musco-
vite. Hornblende is a major constituent of the im-
pure 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, tour-
maline and apatite.

The quartzites west of the Tumas River Inlier have a
noticeably simpler mineralogy than those to the east.
They are composed predominantly of quartz and feld-
spar with significantly low contents of biotite and
muscovite; sillimanite is absent from the assemblages.
The proportion of ore minerals in the rocks is con-
spicuously 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, epi-
dote, calcite, sphene and ore. Wollastonite scapo-
lite, garnet and biotite are present in some of the
specimens.

TABLE 7

MODAL ESTIMATES (VOL %) OF METASEDIMENTS FROM THE KHAN FORMATION
(a = accessory amounts < 1%)

SAMPLE NO	QUARTZ	PLAGIOCLASE	ALKALI FELDSPAR	DIOPSIDE	HORNBLENDE	EPIDOTE	BIOTITE	GARNET	CALCITE	SCAPOLITE	SPHENE	ORE	OTHERS	LOCALITY
F4	40	20	5	20	a		a		5		2	4		B2
F7	20	5	30	20		3			3	7	3	5		B2
1801	25		55		2		8					5	Musc(2),Ap(2)	I6
1805	7	40		a	12	a	25					5	Chl(1),	I10
2006	40	30	5		10		7		2			3	Chl(1),Zir(a)	H9
2403	25	20	a	15	10	10		5	2	5	3		Zir(1)	F6
2406	20	25	a	20	10		20		a		1		Musc(1)	F11
2603	20	a	25	20	25						3	2	Zir(a)	E10
2803	40	25		20		3			a		2	5		D11
2804	60	30	a								a	1	Musc(1)	D11
2805	25	5	30	10	20	2	3				4	a		D12
9001	25	35	a	15	8	6			5		3	5		N5
10006	30	25	5	10	7	7			4		5	1	Wo(a),Scap(a)	M4
12002	35	15	20	5	3	2			5		4	1	Sericite(10)	L3
13004	35	a	3	30	10	2			5	10	3	1		L3
17001	a	a		30				5	2	35	2		Wo(25)	J6
19005	10		20	7				15	4	30	3		Wo(10)	I2
22001	25	30	2	20	15	3			3	1	4		Zir(1)	G6
23001	30		55				5					6	Zir(a)	G6
23002	20	5	35	15	5	3		5	3		3	2		G5
29001	25	35	5	15	5	2		5	3	a	2		Zir(a)	D2
30002	35	20	3	20	5			7	2		5	1	Chl(a)	C2
X146	30	7	20	7				5	5		3		Wo(20)	H5
KH254	60	5	2				25					5	Zir(a),Ap(1)	D13
GB225	1	40		7	40				a		7	1		B2
GB226	3	40		10	35	2			a		5	5		B2
32002	45	3	40						a		a	5	Musc(a)	B4
33002	20	20	5	30	12				2		2		Chl(1),Ap(a)	B4

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 0,2mm to about 2mm, the larger grains being wollastonite and hornblende. Wollastonite has been observed in thin sections as large poikiloblastic grains up to 20mm 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 occuring 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 METASEDIMENTS FROM SWAKOP GROUP
(a = accessory amounts < 1%)

SAMPLE NO	QUARTZ	PLAGIOCLASE	ALKALI FELDSPAR	BIOTITE	MUSCOVITE	CHLORITE	HORNBLENDE	ANDALUSITE	CORDIERITE	GARNET	CALCITE	OLIVINE/SERP.	SPHENE	ORE	OTHERS	LOCALITY
TILLITE (CHUOS FORMATION – D_2G) – MATRIX																
X127	65	2	5	10	15	a								a	Zir(2)	G25
X147	70		20	7						a				2	Zir(a),Ap(a)	F1
X149	65	a	a	10	10	1								a	Zir(a),Ap(a) Clast(10)	L28
X150	55	5		15	5	2								a	Zir(a),Ap(a) Clast(15)	L28
MARBLE (KARIBIB FORMATION – D_3C)																
154											70	20		2	Ap(a)	O14
X60											65	30			Ap(2)	M11
X77			a								85	2			Ap(4)	J12
X81	7	2	a		2						85		1	2		Q20
X122	3		a	1							90			1		P21
META–PELITES (KUISEB FORMATION – WITPOORT MEMBER – D_4S)																
905	35	5	30	25	a									a	Zir(2),All(a)	N11
1005	30	10	30	25		a	a								Zir(1)	M11
1006	40	20	20	15	1									a	Zir(2)	M11
2007	30	30	5	25		3								2	Epid(3),Ap(a)	H11
X32	35	35	a	20	a	2					1		a	a	Zir(a)	S14
X33	60	15		20	1						a		a	a	Ap(a)	S17
X61	30	5	35	20	a	3			5						Zir(1),All(a)	O10
X63	30	5	35	30	a						a			a	Zir(1)	O10
165	50	15		25					3	5				a	Ap(1),All(a), Zir(1)	P13
X142	35	5	15	25						15				a	Zir(2)	N12
X165	15	25		25				25						5	Ap(5),Zir(a)	R7

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 speciments investi-
gated 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 speci-
men X142, collected from the opposite limb of the syn-
cline, 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, plagio-
clase (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 al-
kali-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, horn-
blende, calcite, epidote and diopside. Alkali-feld-
spar, biotite, garnet and sphene also occur in sig-
nificant 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

TABLE 9

MODAL ESTIMATES (VOL %) OF METASEDIMENTS FROM KUISEB FORMATION (TINKAS MEMBER D_4S)   (a = accessory amoumts < 1%)

SAMPLE NO	QUARTZ	PLAGIOCLASE	ALKALI FELDSPAR	BIOTITE	MUSCOVITE	CHLORITE	HORNBLENDE	ANDALUSITE	CORDIERITE	GARNET	CALCITE	SPHENE	ORE	OTHERS	LOCALITY
CALC-GRANOFELSES														
1931	45	5	20				5				5	2		Diop(15)	Q24
19311	70	3	2				15				3			Epid(2),Zir(1	Q25
20111	60	10					25			5	1			Epid(1)	R25
X120	65	2		3	1		15			1	2	3	a	Epid(2)Zir(a) Diop(3)	Q21
X125	35			20	1		15				2	2	a	Epid(2)Diop(20)	N24
META PELITES														
161	50	20	5	20	a								a	Sill(a)	P17
651	55	10		15	5	3					5	1			K23
831	15		2	35					50						L23
1021	55	15		25		3	a				1				M23
1111	30		15	25			25					1	a	Zir(1.)	M24
1121	50	20		20	3						1	a			M23
1421	35		5	15				a	40		1				O23
1731	40		2	30				2	25				a		P22
1811	50	30		12	2	2							a	Zir(1)	Q23
1961	50	30	5	15		2								Zir(a)	Q23
8111	65	10		20									a	Zir(1)	L28
15311	75	2		20							1				O27
19211	70	7		20							1			Zir(a)	Q26
X80	45	25	5	15	5	1							a	a Zir(1)	M23
X121	65	5		20	5	a							a	Ap(1),Zir(a)	Q21
X123	55	a		25	10	1	3						1	Ap(1),Zir(1)	Q21
X124	60	10		15	2	3	a						a	Clasts(5)	R21
X201	30	a	1	30	15			2	20				1	Ap(a)	L29

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. <u>METAMORPHISM</u>

3.1 <u>GENERAL</u>

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.
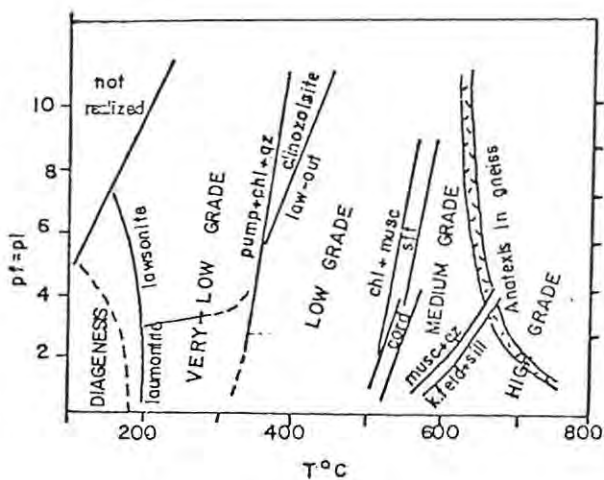
Fig 4a

Grades of metamorphism
(after Winkler 1976)
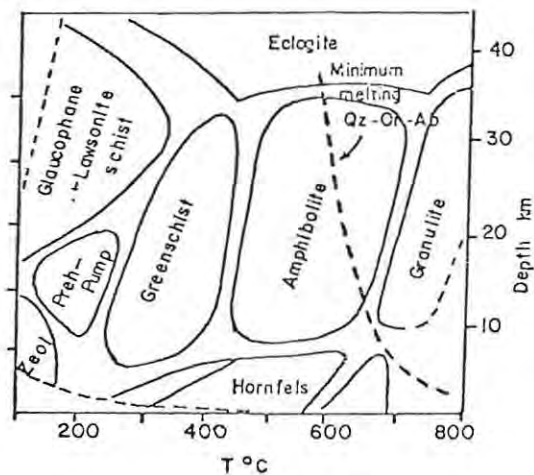
Fig 4b

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, Na_2O and MnO, and found when plotting the concen-
tration of Fe/Fe + Mg, Na_2O 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, Gorbatshev 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 condi-
tion $P_T = P_f = PH_2O$ 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 $P_T > P_f = PH_2O$ which can be expected if
rocks have been subjected to more than one phase of
metamorphism, has the effect of lowering equilibruim
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, CH_4, HCl, CO_2, O_2, H_2 etc.,
released during metamorphism contribute to the fluid
phase and these have varying effects on the equilibrium
temperature of reactions.

Increasing mole fractions of CO_2 (XCO_2) have been
shown to increase equilibrium temperatures in carbo-
nate 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 XCO_2 or XH_2O values
can be established and PT conditions can be esti-
mated 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 signi-
ficance 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 pre-
sence of an earlier Mg-rich chlorite in the rocks
forms cordierite instead of staurolite due to reaction 1.

chlorite + muscovite + quartz = cordierite + biotite +
$$Al_2SiO_5 + H_2O$$
$$-(1)$$
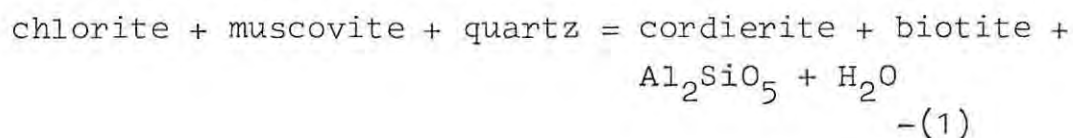
The latter possibility whereby staurolite has already
disappeared from the assemblages may be correct, in
that staurolite was only found as inclusions in garnet
and cordierite grains in the area to the northeast
(Jacob 1974) and that Göttingen workers place the
'staurolite-out' isograd considerably further to the
east than this part of the orogen, Porada (1976).

Andalusite occurs sporadically, usually coexisting
with cordierite and its disapperance from mineral
assemblages coincides with the easternmost appearance
of sillimanite, in the metapsammites and metapelites
of Glücks Dome. Andalusite may have been formed at

an early stage from staurolite by reaction (2).

staurolite + muscovite + quartz = biotite + almandine +

andalusite + H_2O

$-(2)$

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

anadalusite   = sillimanite

$- (3)$

biotite + andalusite + quartz = Cordierite + K-felds +
+ Na plagioclase   Ca-rich plagioclase $\pm$
                                     garnet + H_2O

$-(4)$

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 Al_2SiO_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 Glücks 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
- 卌 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 temparature 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
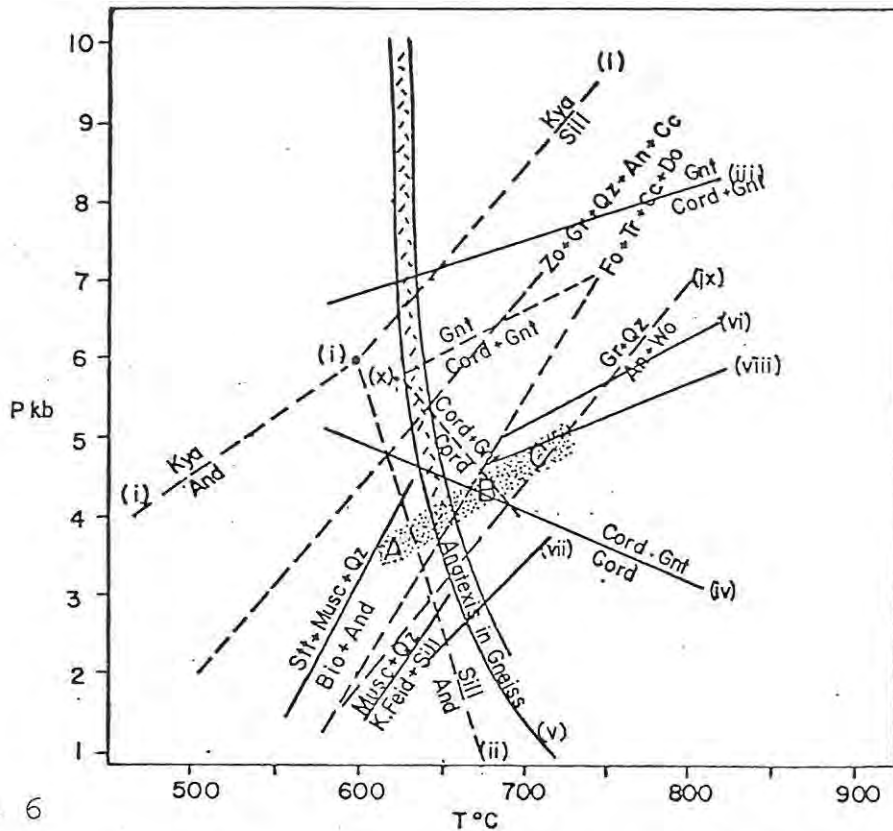
Fig. 6

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).

"   (ii) Andalusite=sillimanite, 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).

"   (v) Anatexis-in-gneiss boundary, Winkler (1976).

"   (vi) bio+sill+Qz = cord+K.Feld+H_2O, Schreyer
         and Seifert (1968).

"   (vii) bio+sill+Qz = cord+K.Feld+H_2O, Hoffer (1976).

"   (viii) bio+sill+Qz = cord+K.Feld+H_2O (present
         investigation)

"   (ix) Gr+Qz = An+Wo, Boettcher, (1970).

"   (x) gnt-cord stability field, Hirschberg and
         Winkler (1968).

ABC   See page 60.

unorientated flakes.

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

muscovite + quartz = K-feldspar + sillimanite + H_2O
$$- (5)$$
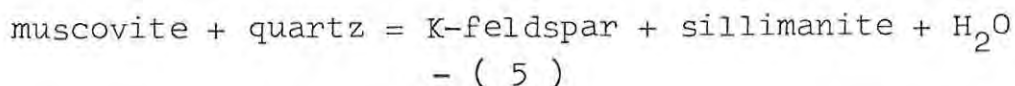
The position of this 'anatexis-in-gneiss' boundary
in the field is well defined and occurs along the
eastern edge of the Tumas River Inlier (Fig 5). On
the high temperature side of the boundary the migma-
titic Leeukop metaconglomerates are present on the
western flanks of the Chameleon Hills. The boundary
passes to the southeast of Leeukop where the eastern-
most 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 attri-
buted to the following two reactions (6 & 7) Hensen &
Green (1971) and Currie (1971).

cordierite = almandine + sillimanite + quartz
$$- (6)$$

cordierite + biotite = almandine + K-feldspar + H_2O
$$- (7)$$
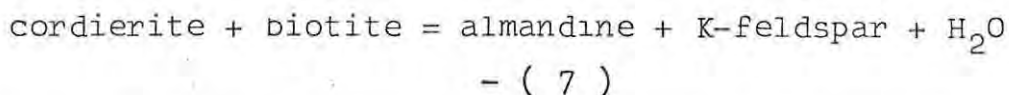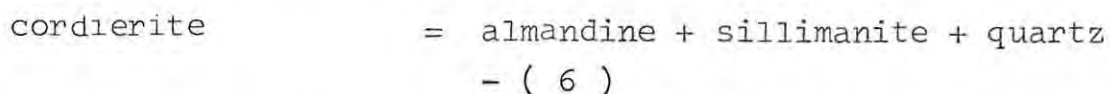
As the minerals cordierite and garnet are present
(samples 165 + 905) in the metapelites beyond the

anatexis boundary in this area, minimum temperature
and pressure values, using data of Currie
(1971) and assuming a FeÓ/(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 compo-
sitions 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 reac-
tion 8; Hess (1969), Schreyer and Seifert (1969),
Hoffer (1976).

biotite + sillimanite + quartz = cordierite + K-feldspar +
$$H_2O \qquad (8)$$
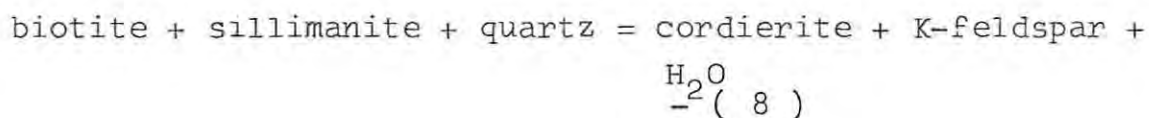
The mineral parageneses present in the metapelites
beyond the anatexis boundary (Samples 165 + 905)
(Plate 21) contain biotite and quartz as inclusions
in cordierite and K-feldspar porphyroblasts which
appear to have been reactants in the above reaction (8).
Sillimanite was not observed but this is probably 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 Hof-
fer (1976) places this reaction curve at considerably
lower temperatures (curve vii fig 6) but it is diffi-
cult 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 as-
semblages 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 sig-
nificance for metamorphic evaluation. Migmatitic de-
velopment 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, immediate-
ly to the northeast of Leeukop, sillimanite is deve-
loped in the granite-gneisses as fibrolitic masses.
These rocks may have been partly dehydrated owing to
the pre-existence of granitic and gneissic rocks, de-
veloped during an earlier metamorphic episode, (Abba-
bis 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 Abbabis 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 dehy-
dration 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 indi-
ces 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 temparature 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

$$- (9)$$
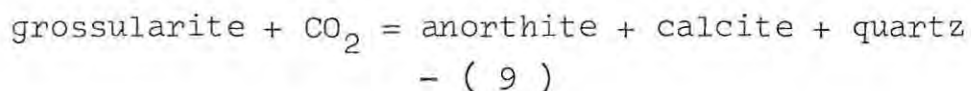
Epidote, although present, does not constitute a sig-
nificant proportion of the rocks, which may be due
to reactions 10 and 11, having proceeded to the right.
These reactions were considered by Jacob (1974) to
have formed grossularite in the calc-granofelses in
the area to the north.

zoisite + CO_2 = anorthite + calcite + H_2O

$$- (10)$$

zoisite + quartz = anorthite + grossularite + H_2O

$$- (11).$$
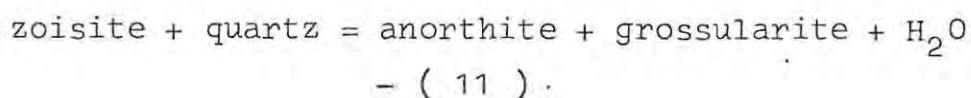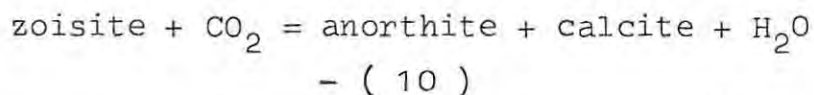
The coexistence of grandite, plagioclase and quartz may be significant in that Winkler (1976, p.144) states that the paragenesis grossularite + anorthite + quartz is stable at $500^{\circ} - 600^{\circ}C$ (2kb) and at very low XCO_2 (0,02 - 0,2) at 2kb. The stability of this paragenesis is unknown for 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 $600^{\circ}C$ if XCO_2 < 0,1, by reaction 12, Metz (1970).

$$1 \text{ tremolite} + 3 \text{ calcite} + 2 \text{ quartz} = 5 \text{ diopside} + 3CO_2 + 1H_2O$$

$$- (12)$$
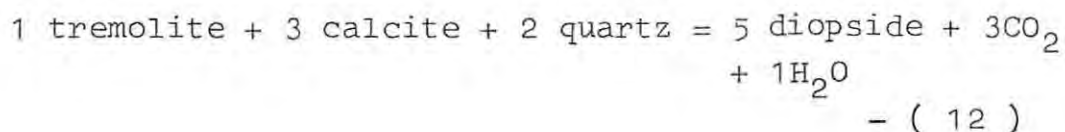
Wollastonite, although not observed in the eastern mineral assemblages has been reported by Jacob (1974) to occur intermittently in the area to the north. Wollastonite forms at temperatures above $600^{\circ}C$ at 2kb only at very low XCO_2 (< 0,1). This is further evidence for very low XCO_2 conditions, (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 fo_2 conditions existing in the banded gneisses. Liou (1973) has shown that epidote is stable up to $750^{\circ}C$ at Pf = 5kb under conditions of high fo_2.

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).

$$\text{grossularite} + CO_2 = \text{anorthite} + \text{calcite} + \text{wollastonite}$$

$$- (13)$$

1 grossularite +1 quartz = 1 anorthite +2 wollastonite

$$- (14)$$
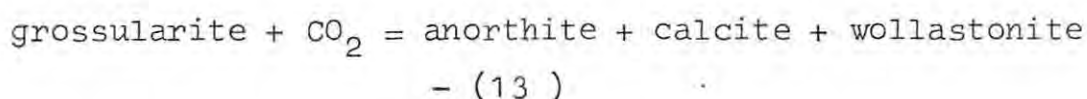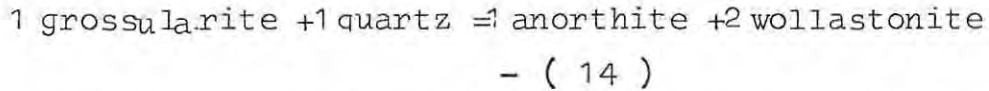
Reaction 14 is possibly responsible for the formation of the wollastonite and anorthite intergrowths in the banded gneisses, where XCO_2 is probably very low. The PT limits of this reaction (Boettcher,1970) from the system $CaO - Al_2O_3 - SiO_2 - H_2O$, are shown in Fig 6,curve ix.

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 tremolite + 1 dolomite = 8 forsterite + 13 calcite
$$+ 9 CO_2 + 1 H_2O$$

$$- (15)$$

1 diopside + 3 dolomite = 2 forsterite + 4 calcite
$$2 CO_2$$
$$- (16)$$
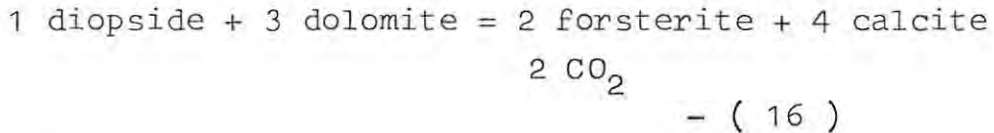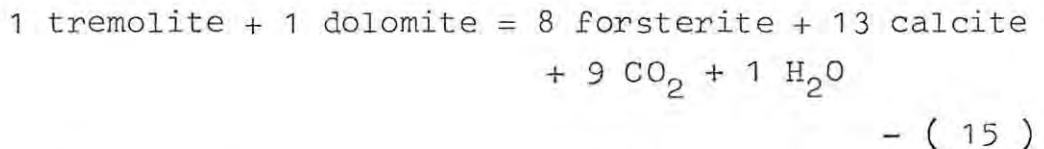
The equilibrium temperatures for both of these reactions are very similar for a given fluid pressure. As neither tremolite nor diopside was observed in the forsterite-bearing assemblages both reactions appear to have proceeded to the right. The occurrence of these minerals together is restricted to an univariant equilibrium curve in $P - T - XCO_2$ 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 stau-
rolite, the presence of andalusite without sillimanite
and the coexistence of the stable mineral pair, mus-
covite and quartz, is indicative of minimum temperatures
of $600^{\circ}C$ and pressures of 3,5 - 4kb (see A in fig 6).
The assemblages grossularite, quartz, anorthite and
calcite with only minor epidote, in the calc-granofel-
ses and the apparent lack of forsterite in the marbles
east of Glücks Dome is in agreement with this conclu-
sion. In the metapelites in the syncline on the
western side of the Tumas River Inlier, garnet (alman-
dine) 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 ana-
tectic granite-gneisses (Husab granite gneiss suite),
the above mineral assemblages are indicative of a
minimum temperature of $675^{\circ}C$ at pressures of 4 - 4,5kb,
adjacent to the western boundary of the Tumas River
Inlier. (See B fig 6). The assemblage biotite, cor-
dierite, K.feldspar, garnet ($\pm$ 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
$720^{\circ}C$ and pressures of 4,5 - 5kb 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 ten-
tatively 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 $600^{\circ}C$ at 3,5 - 4kb, in the
east to approximately $720^{\circ}C$ 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 accep-
tance 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 stau-
rolite in the assemblages of this area. It is dif-
ficult to reconcile the pressures of 6 - 8kb as sug-
gested 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. Tem-
perature determinations by Puhan (1976) using the
dolomite-calcite solvus geothermometer for unmixed
magnesi n calcite crystals on samples collected from
the marbles of the Karibib Formation yielded a tem-
perature of $620^{\circ}C$. This value considered with regard
to the experimentallly determined stability field for
the assemblage forsterite, diopside, tremolite, dolo-
mite, calcite gives a pressure δf 2,8 $\pm$ 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 relation-
ships, five different suites of granitoid rocks
can be distinguished. These are presented schema-
tically 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)
     RÖSSING ALASKITIC GRANITE (G4)

4.1.1   ABBABIS COMPLEX (AGn1)

These rocks outcrop extensively immediately east of
$15^{\circ}00'$ longitude in the core of a large anticlinal/
domal structure, the Tumas River Inlier, and also
on the eastern flanks of the Rabenrücken 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
(As_1) 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 pre-
sent 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
As_1 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 $15^\circ 00'$ 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
respresentation 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 anticlino-
rium) 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 intersec-
tion of $23^{\circ}00'$ latitude and $15^{\circ}00'$ 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 metasedi-
ments 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 gene-
rated by remobilisation of the Abbabis granite-
gneisses. A small amount may have formed by con-
tamination 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 compr sing this suite.

The porphyritic biotite granite/granodiorite, which
is the most abundant, constitutes approximately 85%
of the rocks, and characteristically contains coarse
($\ll$ 4cm) 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. In-
terrelated zones of non-porphyritic and porphyritic
granodiorite/granite give the suite a stromatic, mig-
matitic appearance (Plate 26). The non-porphyritic
granodiorite occurs intermittantly 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 por-
phyritic 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 occur-
rence 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 in-
termittent 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 granite-
gneisses 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 smokey 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 Glücks 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 en-
closing metasediments. The enclosing country rocks
comprise Kuiseb and Etusis metasediments, tillite
(Chuos Formation) and augen gneisses (Abbabis Complex).

Contact metamorphic effects associated with the in-
trusion of the stock are restricted to a faint disco-
louration 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 con-
tacts 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 aplitic 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-Damara 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 syn-clinal 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), Nieberding (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 musco-vite. 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. Acces-sory amounts of Fe-ore, muscovite, zircon, apatite, sphene, epidote and calcite also occur. Microcline, the predominant feldspar is typically anhedral, ex-hibits cross-hatched twinning and is occasionally poikilitic and also perthitic. Myrmekite is often developed at K-feldspar/plagioclase boundaries. Pla-gioclase is invariably multiply twinned and has seri-citic alteration along the cleavage planes. Zoning is often observed in the plagioclase grains and dis-tinct albite rims occur on some. The composition of plagioclase varies between An_{26} – An_{30}, with the exception of the plagioclase rich rocks (GB 228, GB 229) and (GB 210), where the composition is predominantly sodic (An_7 – An_{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

SAMPLE NO	ROCK TYPE	MODAL PERCENTAGES (VOL%) (tr = 0-0.5%)					ACCESSORIES PRESENT	AVERAGE PLAGIOCLASE COMPOSITION	LOCALITY
		QUARTZ	MICROCLINE	PLAGIOCLASE	BIOTITE	ACCESSORIES			
	ABBABIS COMPLEX								
GB200	augen-gneiss	35	36	14	12	3	musc, zir, ore, epid	An_{35}	M28
GB205	" "	31	47	12	8	2	" " " "	An_{31}	M28
GB209	" "	24	45	22	7	2	musc, ore	An_{35}	L18
GB213	" "	28	27	20	23	2	musc, zir, ore	An_{26}	Q14
GB206	pink foliated granite	32	48	19	–	1	musc, ore	An_{30}	M29
	HUSAB GRANITE-GNEISS SUITE								
GB207	red foliated granite-gneiss	30	47	17	5	1	musc, zir, ap, ore	An_{20}	G15
GB208	red fgnd. granite	35	38	18	4	5	" " " ", cc	An_{21}	M18
GB211	pink granite	32	28	26	12	2	" " " "	An_{27}	O15
GB217	red granite	29	39	27	3	2	" , ore, ap	An_{24}	Q8
GB227	red foliated granite-gneiss	34	42	19	4	1	" " ", sph, epid	An_{25}	C6
GB231	" " " "	31	39	24	4	2	" " ", zir	An_{23}	R8
GB235	red granite	30	32	34	2	2	" " "	An_{27}	E13
GB212	pale-grey granite	34	35	25	5	1	", ap, cc	An_{24}	O15
GB215	pale-red granite	40	29	26	5	tr	", zir, cc	An_{26}	O10
GB228	pale granite-gneiss	39	21	39	–	1	", ore, sph, cc	An_{12}	I5
GB229	leuco-granite-gneiss	26	33	40	1	tr	", ap	An_{10}	I5
GB210	" " "	35	37	26	tr	2	", ore, zir	An_{7}	L17

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

SAMPLE NO	ROCK TYPE	MODAL PERCENTAGES (VOL %) (tr = 0-0.5%)					ACCESSORIES PRESENT	AVERAGE PLAGIOCLASE COMPOSITION	LOCALITY
		QUARTZ	MICROCLINE	PLAGIOCLASE	BIOTITE	ACCESSORIES			
ROSSING ALASKITIC GRANITE (G$_4$)									
GB236	fine grained pale granite	27	47	25	tr	1	musc,ore,sph	An15	D12
KH106/7	med. " " "	21	56	23	tr	tr	" , cc	An12	D13
KH132/3	"-coarse grnd pink granite	12	63	23	1	1	" ,ore	An17	D13
KH255	"- " pale granite/pegmatite	32	42	25	1	tr	ore	An16	D13
KH256	med-fine grnd pale granite	29	7	64	tr	tr	musc,sph	An17	D13
KH472/3	med pale grey granite	33	4	58	5	tr	" ",zir,cc	An19	D13
KH496	med-coarse grey granite/pegmatite	15	7	75	3	tr	ore	An18	D13
KH497	" - " pale granite/pegmatite	29	43	28	-	-		An14	D13

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 placioclase 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 12

SAMPLE NO	ROCK TYPE	MODAL PERCENTAGES (VOL %) (tr = 0-0.5)							AVERAGE PLAGIOCLASE COMPOSITION	LOCALITY
		QUARTZ	MICROCLINE	PLAGIOCLASE	BIOTITE	HORNBLENDE	ACCESSORIES	ACCESSORIES PRESENT		
	SALEM GRANITOID SUITE									
GB216	porphyritic foliated granodiorite	22	24	31	17	2	4	Ap,sph,ore,zir, all	An25	P10
GB220	" "	23	9	50	13	tr	5	Ap,musc,",",",cc	An41	Q6
GB237	fine grained granodiorite	19	11	43	15	8	2	Ap,zir,cc,ore, musc	An25	R13
	GAWIB GRANITOID SUITE									
GB204	dark foliated granodiorite	18	16	41	15	8	2	epid,sph,all, ore,ap	An35	L28
GB221	" " "	22	10	39	18	5	6	epid,sph,all,cc, ore,ap	An31	L28
GB203	slightly porphyritic	21	28	35	9	1	6	ep,cc,sph,ore, all,zir	An29	L27
GB201	non porphyritic granite	33	16	43	6	tr	2	epid,cc,sph,ore all,zir,musc	An27	M25
GB202	" " "	25	17	49	7	-	2	epid,all,musc, cc,sph,ore	An28	L26
GB223	porphyritic granite	23	27	40	7	tr	3	epid,ap,sph,ore zir	An29	N26
GB224	" "	27	21	43	7	tr	2	epid,ap,sph,ore zir	An29	N25

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 hypi-
diomorphic granular and a faint foliation and dis-
tortion 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 mine-
rals 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 multiplytwinned, 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 compo-
nent. 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 par-
ticularly in the foliated granodiorite. Calcite
generally occurs interstitially and also in fractures.
Zircons with radioactive haloes are observed in bio-
tite grains.

4.2.6 DISCUSSION

Modal quartz, alkali feldspar and plagioclase compo-
nents 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.
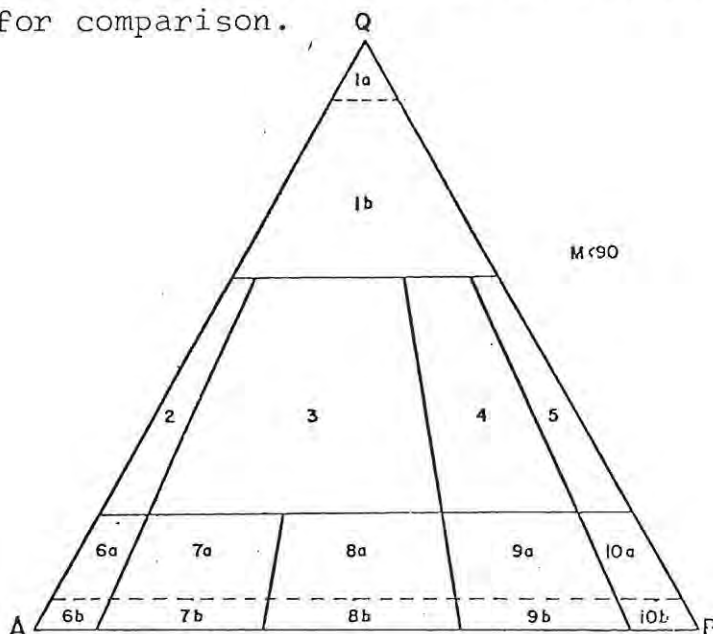
1a Quartzolite (silexite)
1b Quartz-rich granitoids
2 Alkali-feldspar
3 Granite
4 Granodiorite
5 Tonalite
6a Alkali-feldspar quartz syenite
7a Quartz syenite
8a Quartz monzonite

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 inter-
pretation.
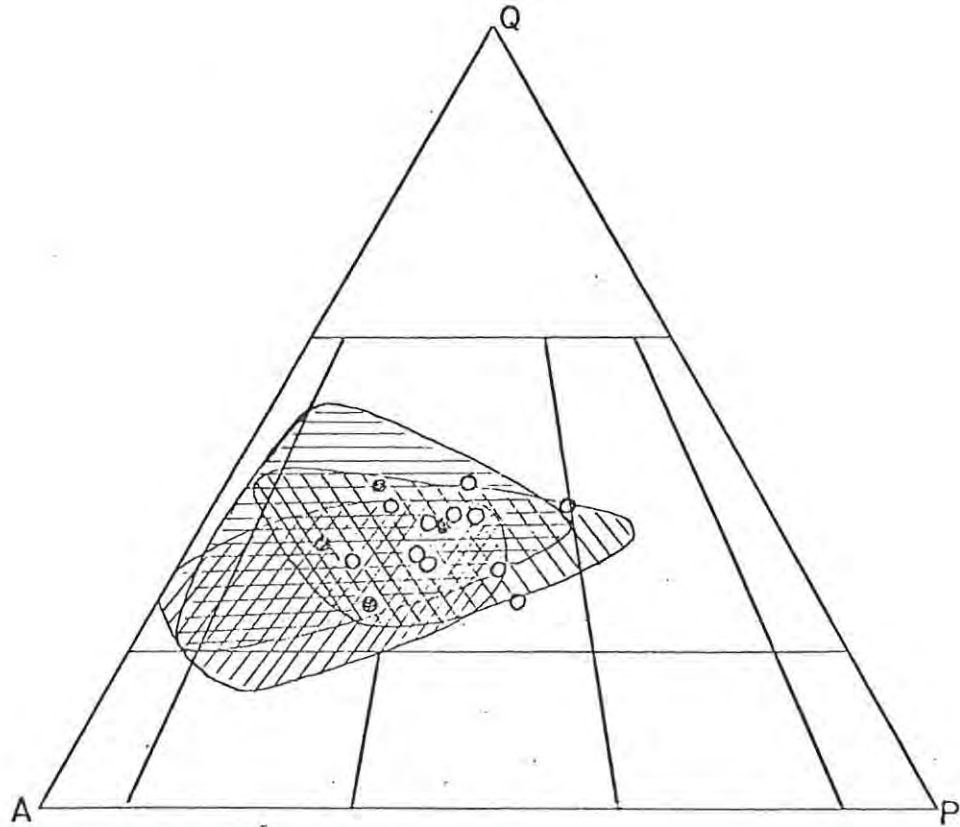
Fig. 8

Red Granites, Gneisses, leucogranites   o

   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 monzo-granitic 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 Omangambo 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 Omangambo are of granodioritic and granitic composition.

In comparison to the Salem granitoid suite from the central part of the Damara belt the dioritic/monzo-dioritic 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
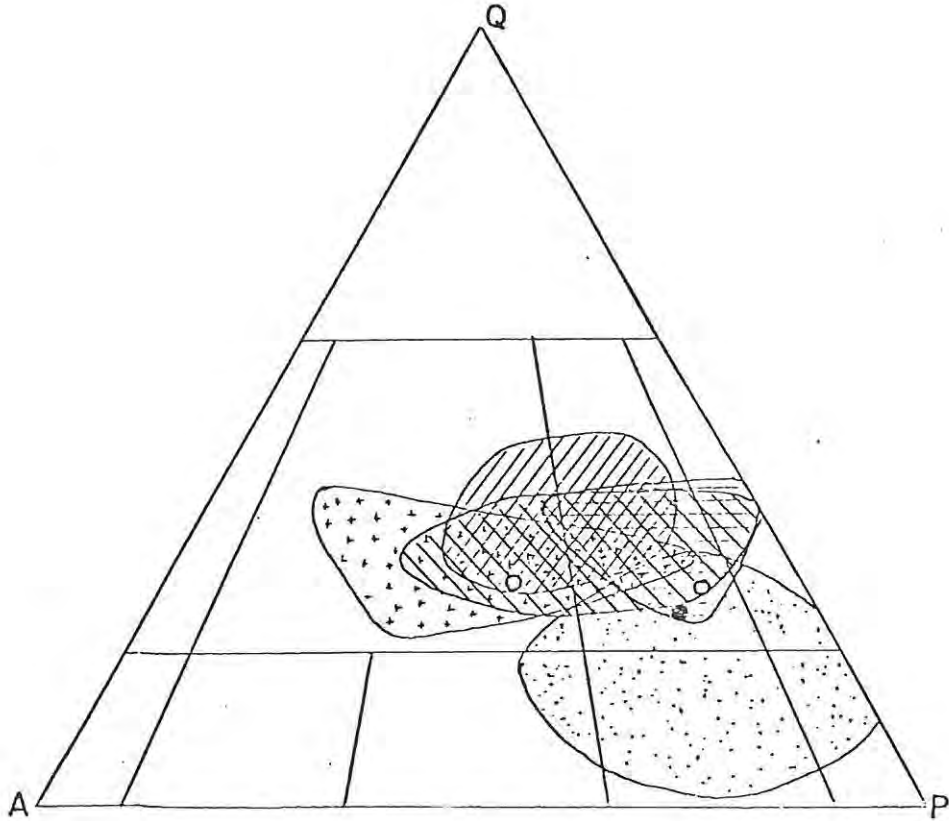
Fig. 9

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

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

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
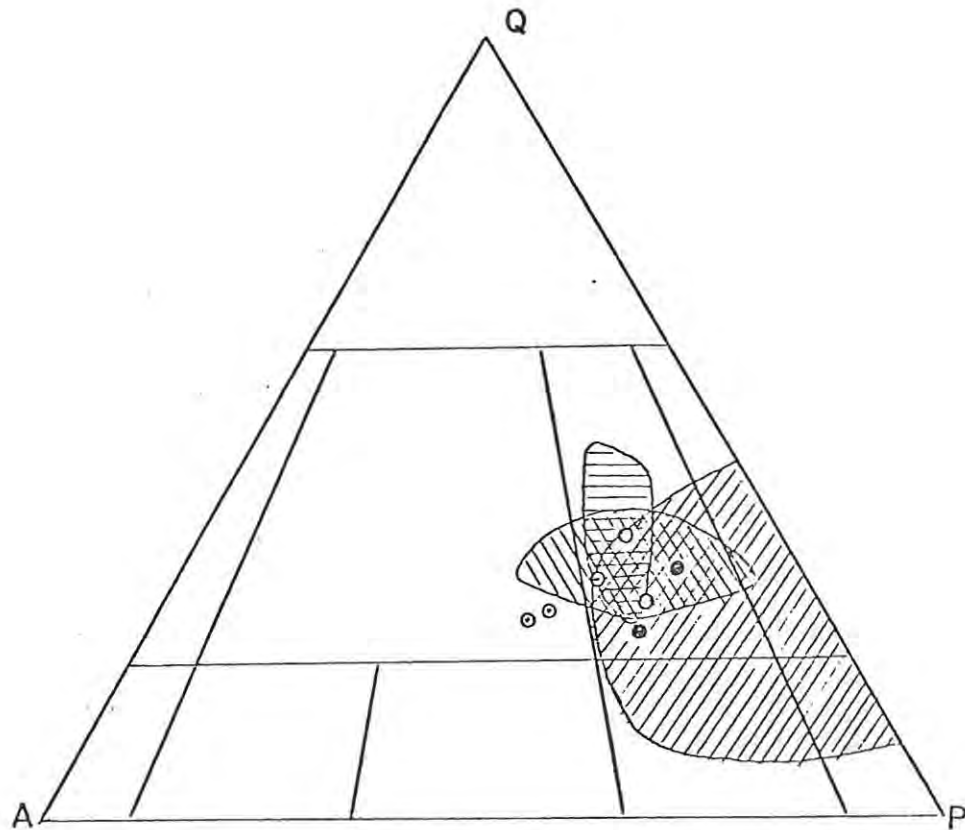
Fig. 10

Foliated Granodiorite   ✽

Non-porphyritic Granite   O

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 differen-
tiated. 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 through-
out 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 plagio-
clase 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 sug-
gests 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 alas-
kitic 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 meta-
sediments. Consequently, the Rössing alaskitic gra-
nites were considered to represent late-phase mobili-
sates similarly derived by anatexis of Nosib meta-
sediments. An extensive geochemical investigation
of both the Damara metasediments and the granitic rocks
was beyond the scope of this study. Attention was there-
fore 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 grano-diorite 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 autochtonous 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:

SiO_2 Harker diagram

$(^1/_3 Si + K) - (Ca + Mg)$ Nockolds and Allen (1953)
modified Larsen index

Normative Qz+Ab+Or+Ne+Kp+Lc   Thornton and Tuttle (1960)
Diff Index

MgO   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 compo-
sitions, mesonorm rather than CIPW norm values were
used. The mesonorm values have close similarities
to the modal compositions, incorporate hydrous mine-
rals, 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 $(Qz - Ab - Or - An - H_2O)$ (See
Fig 11) since the early work of Tuttle and Bowen (1958).

Table 13

Major element , trace element , trace element ratio: ,cation .
percentage and mesonorm values for rocks analysed from the
Abbabis and Husab suites.

| | ABBABIS GRANITE-GNEISS SUITE | | | | HUSAB GRANITE-GNEISS SUITE | | | | | | | | | | |
	GB200	GB205	GB209	GB213	GB207	GB208	GB211	GB212	GB215	GB217	GB227	GB231	GB235	GB228	GB229
SIO2	71.21	70.65	71.11	69.66	73.13	71.74	69.55	73.55	73.71	72.78	72.11	73.38	74.13	76.18	77.65
TIO2	0.68	0.47	0.54	0.64	0.34	0.58	0.50	0.25	0.21	0.26	0.40	0.30	0.27	0.17	0.03
AL2O3	12.14	13.69	13.82	14.12	13.49	13.77	14.43	13.81	13.72	13.81	14.03	14.09	13.77	13.67	13.57
FE2O3	2.26	1.10	2.49	0.82	1.57	1.76	1.02	0.16	0.31	1.09	1.24	1.33	1.42	1.27	0.12
FEO	2.18	2.10	1.47	3.67	0.67	0.72	2.52	1.63	1.44	0.84	1.18	0.91	0.43	0.24	0.12
MNO	0.09	0.06	0.05	0.07	0.03	0.05	0.06	0.04	0.02	0.03	0.02	0.02	0.02	0.02	0.02
MGO	1.21	0.92	0.87	1.10	0.38	0.40	0.75	0.41	0.30	0.42	0.51	0.51	0.27	0.04	0.06
CAO	2.27	1.11	1.34	1.69	0.58	1.56	1.87	1.30	1.36	1.25	1.27	1.31	0.85	1.08	0.48
NA2O	2.34	1.97	2.68	1.97	2.57	2.58	2.64	2.41	2.89	2.93	2.66	3.10	3.74	4.98	5.02
K2O	4.39	7.42	5.35	5.98	7.04	6.02	4.96	5.92	5.17	5.73	6.31	5.20	5.14	2.56	3.76
P2O5	0.19	0.13	0.13	0.17	0.07	0.06	0.15	0.06	0.05	0.08	0.15	0.05	0.08	0.05	0.02
H2O+	1.06	1.02	0.43	0.87	0.48	1.60	1.00	0.92	0.76	0.76	0.79	0.66	0.67	0.37	0.22
H2O-	0.05	0.11	0.07	0.04	0.06	0.06	0.04	0.04	0.05	0.06	0.07	0.11	0.09	0.08	0.05
TOTAL	100.07	100.73	100.35	100.80	100.41	100.50	99.49	100.50	99.97	100.04	100.74	100.97	100.88	100.71	101.12
TRACE ELEMENTS (PPM)															
BA	N.D.	1500	1500	1500	900	1500	1100	1200	900	900	1000	1100	800	1000	900
RB	243	353	295	283	322	276	242	232	255	319	319	243	260	72	90
SR	100	92	115	145	104	154	113	145	96	115	85	110	49	59	68
NB	15.4	9.2	13.9	13.0	9.0	9.4	14.9	12.5	14	13.9	19.3	10.2	15.2	7.1	0.7
ZR	267	204	254	261	237	384	381	186	200	223	318	248	227	155	56
Y	30.1	29.6	27.4	29.2	47	17.7	23.6	20.6	31.2	24.9	36.1	36.3	38.2	25.4	6.0
U	N.D.	3	5	5	19	10	8	10	11	12	9	14	22	8	2
TH	39	38	49	50	148	68	68	65	79	82	126	93	82	53	16
PB	37.7	58	46	68	68	96	68	81	58	63	24.5	53	41	51	60
TRACE ELEMENT RATIOS															
K/RB	149	174	150	175	181	180	170	211	168	149	164	177	164	291	347
CA/SR	162	86	83	83	40	73	118	64	101	77	107	85	124	131	50
RB/SR	2.44	3.85	2.57	1.96	3.11	1.80	2.14	1.61	2.67	2.78	3.76	2.22	5.33	1.24	1.32
K/BA	N.D.	47	30	33	65	33	37	41	48	53	52	39	53	21	35
BA/SR	N.D.	14.1	13.0	10.3	8.7	9.7	9.7	8.3	9.4	7.8	11.8	10.0	16.3	16.9	13.2
U/TH	N.D.	0.08	0.10	0.10	0.13	0.15	0.12	0.15	0.14	0.15	0.07	0.15	0.27	0.15	0.13
PB/SR	0.38	0.63	0.40	0.47	0.65	0.62	0.60	0.56	0.60	0.55	0.29	0.48	0.84	0.86	0.88
TI/NB	265	307	231	295	228	241	202	121	90	110	123	173	106	145	291
CATION PERCENT															
SI	68.37	66.82	67.18	65.98	68.82	68.42	66.50	69.50	69.78	68.77	67.81	68.69	69.15	70.69	71.21
TI	0.49	0.34	0.38	0.46	0.24	0.27	0.36	0.18	0.15	0.18	0.28	0.21	0.19	0.12	0.02
AL	13.74	15.27	15.40	15.77	14.96	15.49	16.27	15.39	15.32	15.38	15.55	15.55	15.14	14.95	14.68
FE+++	1.64	0.78	1.77	0.59	1.11	1.26	0.74	0.11	0.22	0.77	0.88	0.94	0.99	0.88	0.08
FE++	1.75	1.66	1.17	2.90	0.53	0.57	2.01	1.29	1.14	0.66	0.93	0.71	0.34	0.18	0.09
MN	0.07	0.05	0.04	0.05	0.02	0.04	0.05	0.03	0.02	0.03	0.02	0.01	0.02	0.01	0.01
MG	1.73	1.29	1.23	1.56	0.53	0.57	1.07	0.58	0.42	0.59	0.72	0.71	0.38	0.05	0.09
CA	2.34	1.15	1.36	1.71	0.58	1.60	1.92	1.32	1.38	1.26	1.28	1.31	0.85	1.07	0.47
NA	4.36	3.61	4.91	3.61	4.69	4.41	4.90	4.41	5.31	5.37	4.85	5.62	6.77	8.97	8.93
K	5.37	8.96	6.45	7.22	8.45	7.32	6.05	7.14	6.24	6.90	7.57	6.20	6.12	3.03	4.40
P	0.15	0.10	0.11	0.14	0.06	0.05	0.12	0.05	0.02	0.07	0.12	0.04	0.06	0.06	0.02
MESONORM															
Q	35.49	27.51	31.10	30.97	28.66	30.48	30.58	32.55	32.57	29.82	28.68	30.93	29.16	32.81	30.36
C	0.82	1.44	2.43	2.89	1.33	1.28	2.62	1.71	1.37	1.17	1.53	1.66	1.12	1.18	0.51
OR	22.50	40.46	29.68	29.07	41.35	35.58	25.65	32.62	28.77	33.02	35.79	29.41	29.94	15.03	21.75
AB	21.80	18.07	24.57	18.07	23.45	22.04	24.52	22.04	26.57	26.87	24.25	28.10	33.86	44.83	44.64
AN	7.96	3.12	4.00	5.13	1.23	6.22	6.73	5.31	5.98	4.85	4.01	5.18	2.83	4.45	2.10
BI	7.29	6.95	4.14	11.27	1.47	1.63	7.38	4.93	3.91	2.40	3.27	2.59	1.05	0.18	0.41
SPH	1.48	1.01	1.14	1.37	0.73	0.81	1.09	0.54	0.45	0.55	0.84	0.62	0.56	0.36	0.07
MT	2.45	1.17	2.66	0.88	1.58	1.72	1.10	0.17	0.33	1.16	1.31	1.41	1.01	0.55	0.12
HE	0.00	0.00	0.00	0.00	0.06	0.12	0.00	0.00	0.00	0.00	0.00	0.00	0.32	0.52	0.00
AP	0.40	0.27	0.28	0.37	0.15	0.13	0.33	0.13	0.06	0.18	0.32	0.10	0.16	0.10	0.04

Table 14

	GB236	KH106/7	KH152/3	KH255	KH256	KH472/3	KH496	KH497
SIO2	75.48	71.38	73.10	77.46	74.65	68.28	67.59	75.43
TIO2	0.08	0.02	0.02	0.03	0.03	0.25	0.32	0.01
AL2O3	14.18	16.02	16.00	13.54	14.62	16.42	16.97	13.39
FE2O3	0.07	0.13	0.04	0.13	0.12	1.18	1.28	0.17
FEO	0.09	0.13	0.08	0.18	0.18	0.49	0.60	0.11
MNO	0.01	0.01	0.00	0.01	0.01	0.01	0.02	0.01
MGO	0.07	0.10	0.02	0.13	0.18	0.91	0.92	0.04
CAO	0.68	1.08	0.76	1.11	2.14	1.56	2.07	0.88
NA2O	3.23	3.41	3.79	4.05	6.09	4.54	5.97	3.90
K2O	6.59	8.08	7.08	3.74	0.96	5.62	3.00	5.04
P2O5	0.03	0.28	0.05	0.07	0.13	0.02	0.02	0.02
H2O+	0.21	0.40	0.20	0.24	0.29	0.52	0.44	0.14
H2O-	0.05	0.06	0.05	0.10	0.11	0.08	0.04	0.06
TOTAL	100.77	101.00	101.19	100.79	99.51	99.68	99.24	99.20
TRACE ELEMENTS [PPM]								
BA	900	800	900	800	1200	600	600	700
RB	310	373	342	169	29.8	288	164	223
SR	129	118	72	123	169	58	63	41.2
NB	31	4.8	1.3	5.2	7.6	29.8	33.4	*
ZR	52	11.5	15.6	25.5	31.0	63	77	67
Y	15.8	26.2	6.5	15.6	21.7	54	33.0	15.4
U	10	11	5	28	12	1013	338	44
TH	23	13	13	20	16	130	62	17
PB	14.3	20.1	61.1	38.3	23.8	89.4	46	19.7
TRACE ELEMENT RATIOS								
K/RB	176	179	171	184	266	162	152	188
CA/SR	38	65	75	64	91	167	235	153
RB/SR	2.41	3.17	4.76	1.37	0.18	4.97	2.60	5.41
K/BA	61	84	65	39	265	78	41	60
BA/SR	7.0	6.8	12.5	6.5	7.1	10.3	9.5	17.0
U/TH	0.43	0.85	0.38	1.40	0.75	7.79	5.45	2.59
PB/SR	0.11	0.17	0.85	0.31	0.14	1.54	0.73	0.48
TI/NB	14.5	25.0	73.8	34.6	25.2	50.9	57.6	72.0
CATION PERCENT								
SI	69.93	65.51	66.86	71.88	69.22	63.27	62.39	70.85
TI	0.05	0.01	0.01	0.02	0.02	0.18	0.22	0.01
AL	15.49	17.34	17.25	14.82	15.98	17.94	18.47	14.83
FE+++	0.05	0.09	0.03	0.09	0.08	0.82	0.89	0.12
FE++	0.07	0.10	0.06	0.14	0.14	0.38	0.46	0.09
MN	0.00	0.01	0.00	0.00	0.01	0.01	0.01	0.01
MG	0.10	0.13	0.03	0.18	0.24	1.25	1.27	0.06
CA	0.68	1.06	0.74	1.10	2.13	1.55	2.05	0.88
NA	5.81	6.08	6.72	7.29	10.94	8.16	10.69	7.10
K	7.79	9.46	8.26	4.43	1.13	6.64	3.53	6.04
P	0.03	0.21	0.04	0.05	0.10	0.02	0.02	0.02
MESONORM								
Q	27.93	17.51	20.57	34.74	29.09	16.41	15.91	29.69
C	0.73	0.45	0.93	1.12	0.02	0.86	0.65	0.00
OR	38.67	46.98	41.19	21.66	5.07	31.09	15.48	30.09
AB	29.04	30.38	33.58	36.44	54.71	40.80	53.46	35.51
AN	2.92	3.43	3.36	4.95	9.72	5.70	9.00	4.20
BI	0.41	0.52	0.20	0.74	0.94	3.36	3.47	0.20
SPH	0.16	0.04	0.03	0.06	0.07	0.53	0.67	0.03
MT	0.07	0.14	0.04	0.14	0.13	1.13	1.33	0.18
HE	0.00	0.00	0.00	0.00	0.00	0.07	0.00	0.00
AP	0.07	0.57	0.10	0.14	0.26	0.05	0.04	0.04

Major element, trace element, trace element ratio, cation percentage and mesonorm values for rocks analysed from the Rössing Alaskitic granites.

Table 15

	SALEM GRANITE				GAWIB GRANITE						
	GB215	GB220	GB237		GR201	GB202	GB203	GB204	GB221	GB223	GB224
SIO2	65.36	60.64	59.55		70.31	68.40	64.80	60.69	59.57	69.62	70.00
TIO2	0.91	1.02	1.62		0.30	0.36	0.61	0.92	0.90	0.33	0.33
AL2O3	14.68	15.90	14.35		15.46	15.85	15.13	16.77	15.85	15.24	15.58
FE2O3	1.26	2.66	1.94		1.25	1.26	2.21	2.39	2.40	1.48	1.40
FEO	4.00	3.38	6.54		1.50	1.82	2.42	3.93	4.36	1.30	1.31
MNO	0.11	0.09	0.16		0.10	0.10	0.12	0.16	0.14	0.12	0.12
MGO	1.17	2.57	1.88		0.64	0.76	1.59	2.25	2.59	0.82	0.79
CAO	2.87	3.84	4.19		2.73	3.16	3.80	5.40	5.44	2.88	2.88
NA2O	3.03	2.54	2.83		4.07	3.92	3.19	2.70	2.58	3.08	3.95
K2O	4.59	3.53	3.89		3.42	3.29	4.06	3.31	3.42	3.62	3.79
P2O5	0.29	0.27	0.61		0.09	0.12	0.21	0.25	0.23	0.09	0.09
H2O+	0.98	2.68	1.29		0.53	0.85	1.42	1.18	1.86	0.59	0.67
H2O-	0.02	0.05	0.01		0.04	0.05	0.05	0.02	0.04	0.03	0.02
TOTAL	99.28	98.97	98.86		100.44	99.92	99.61	99.96	99.33	100.00	100.93
TRACE ELEMENT (PPM)											
BA	1200	1200	1800		1300	1500	1200	1200	1300	1500	1400
RB	220	232	135		136	121	174	117	121	153	171
SR	157	239	261		705	819	661	483	518	681	695
NB	37.1	19.8	50.6		18.1	19.0	23.5	12.7	12.2	16.2	16.3
ZR	373	308	480		181	199	238	213	199	165	165
Y	67	29.6	61		14.9	16.2	26.6	30.3	28.5	18.1	18.3
U	2	5	4		8	7	10	3	7	8	10
TH	34	27	18		24	24	43	25	23	23	25
PB	27.8	8.1	24.0		39.1	31.4	25.3	16.7	9.6	31.4	37.1
TRACE ELEMENT RATIOS											
K/RB	173	126	239		209	225	193	234	234	196	184
CA/SR	131	115	115		28	28	41	80	75	30	30
RB/SR	1.40	0.97	0.52		0.19	0.15	0.26	0.24	0.23	0.22	0.25
K/BA	32	25	18		22	18	28	23	22	25	22
BA/SR	7.6	5.4	6.9		1.8	1.8	1.8	2.5	2.5	1.9	2.0
CA/Y	306	926	490		1309	1596	1021	1274	1363	1137	1123
U/TH	0.06	0.19	0.22		0.33	0.29	0.23	0.12	0.30	0.34	0.43
PB/SR	0.18	0.03	0.09		0.055	0.056	0.038	0.035	0.018	0.046	0.053
TI/NB	148	310	318		99	113	155	434	441	122	121
CATION PERCENT											
SI	62.53	59.29	57.81		65.55	64.34	61.80	57.70	57.43	65.27	65.00
TI	0.66	0.75	1.16		0.21	0.25	0.44	0.66	0.65	0.23	0.23
AL	16.56	18.53	16.43		17.00	17.58	17.01	18.80	18.01	16.85	17.06
FE+++	0.91	1.96	1.42		0.88	0.89	1.59	1.71	1.74	1.04	0.98
FE++	3.20	2.76	5.31		1.17	1.45	1.93	3.13	3.51	1.02	1.01
MN	0.09	0.07	0.13		0.08	0.08	0.10	0.13	0.12	0.10	0.10
MG	1.66	3.75	2.72		0.89	1.06	2.27	3.19	3.72	1.14	1.09
CA	2.94	4.02	4.35		2.73	3.19	3.88	5.50	5.61	2.89	2.86
NA	5.62	4.44	5.32		7.37	7.15	5.89	4.98	4.81	7.05	7.11
K	5.60	4.40	4.82		4.07	3.94	4.93	4.01	4.20	4.32	4.49
P	0.24	0.22	0.50		0.07	0.09	0.17	0.20	0.19	0.07	0.07
MESONORM											
Q	24.42	26.25	21.54		26.24	25.26	22.37	21.06	20.12	25.86	24.94
C	1.56	3.71	1.62		0.76	0.93	0.00	0.78	0.00	0.41	0.43
OR	20.53	12.66	11.70		17.51	16.16	19.18	10.73	10.85	18.72	19.60
AB	28.10	22.18	26.59		36.83	35.75	29.47	24.89	24.07	35.23	35.55
AN	9.45	14.47	11.67		12.00	13.90	15.45	22.56	22.49	12.68	12.58
BI	11.99	14.96	19.66		4.52	5.68	8.79	14.91	16.26	4.64	4.56
ACT	0.00	0.00	0.00		0.00	0.00	0.61	0.00	1.14	0.00	0.00
SPH	1.97	2.26	3.55		0.63	0.76	1.31	1.97	1.95	0.70	0.69
MT	1.36	2.93	2.13		1.32	1.33	2.38	2.56	2.61	1.57	1.47
AP	0.63	0.60	1.54		0.19	0.25	0.44	0.53	0.50	0.20	0.19

Major element, trace element, trace element ratio, cation percentage and mesonorm values for rocks analysed from the Salem and Gawib Granitoid suites.

The results from various workers including Luth
et al. (1964), Piwinskii and Wyllie (1968, 1970)
Robertson and Wyllie (1971), Piwinskii (1973), and
Winkler (1967, 1974) show the stability of different
minerals and mineral assemblages at different tem-
peratures, 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 deter-
mined modal compositions within the system Qz – Ab –
Or – An – H_2O 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 crys-
tallisation from depth) for natural granitic, grano-
dioritic 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 pro-
jections 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 paral-
lelly projected points.

4.3.3   NORMATIVE CHEMISTRY RELATED TO EXPERIMENTAL SYSTEMS

4.3.3.1   ABBABIS–, HUSAB–, GRANITE–GNEISSES AND RÖSSING
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_2O tetrahedron (Fig 11).
The phases not represented at the apices of the pro-
jections (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

SAMPLE NO	Qz	Ab	An	Or	Or	An	Ab	Qz	Ab	Or
GB 200	41	25	9	26	43	15	42	45	27	28
205	31	20	4	45	66	5	29	32	21	47
209	35	28	5	33	51	7	42	36	29	35
213	37	22	6	35	56	10	35	40	23	37
GB 207	30	25	1	44	63	2	36	31	25	44
208	32	23	7	38	56	10	35	35	25	40
211	35	28	8	29	45	12	43	38	30	32
212	35	24	6	35	54	9	37	37	25	37
215	35	28	6	31	47	10	43	37	30	33
217	32	28	5	35	51	7	42	33	30	37
227	31	26	4	39	56	6	38	32	27	40
228	34	46	5	16	23	7	70	35	48	16
229	31	45	2	22	32	3	65	31	46	22
231	33	30	6	31	47	8	45	35	32	33
235	30	35	3	31	45	4	51	31	36	32
GB236	28	30	3	39	55	4	41	29	30	40
KH 106/7	18	31	4	48	58	4	38	18	32	50
KH 132/3	21	34	3	42	53	4	43	22	35	43
255	36	37	5	22	34	8	58	37	39	23
256	30	56	10	5	7	14	79	33	62	6
472/3	18	43	6	33	40	7	53	19	46	35
496	17	57	10	17	20	12	69	19	63	18
497	30	36	4	30	43	6	51	31	37	32
GB 216	30	34	12	25	35	16	48	33	38	28
220	35	29	19	17	26	29	45	43	36	21
237	30	37	16	16	23	23	53	36	44	20
GB.:201	28	40	13	19	26	18	56	33	46	22
202	28	39	15	18	25	21	54	33	46	21
203	26	34	18	22	30	24	46	31	42	27
204	27	31	29	14	18	39	43	37	44	19
221	26	31	29	14	19	39	42	37	44	20
223	28	38	14	20	28	19	53	32	44	23
224	27	38	14	21	29	19	52	31	44	24
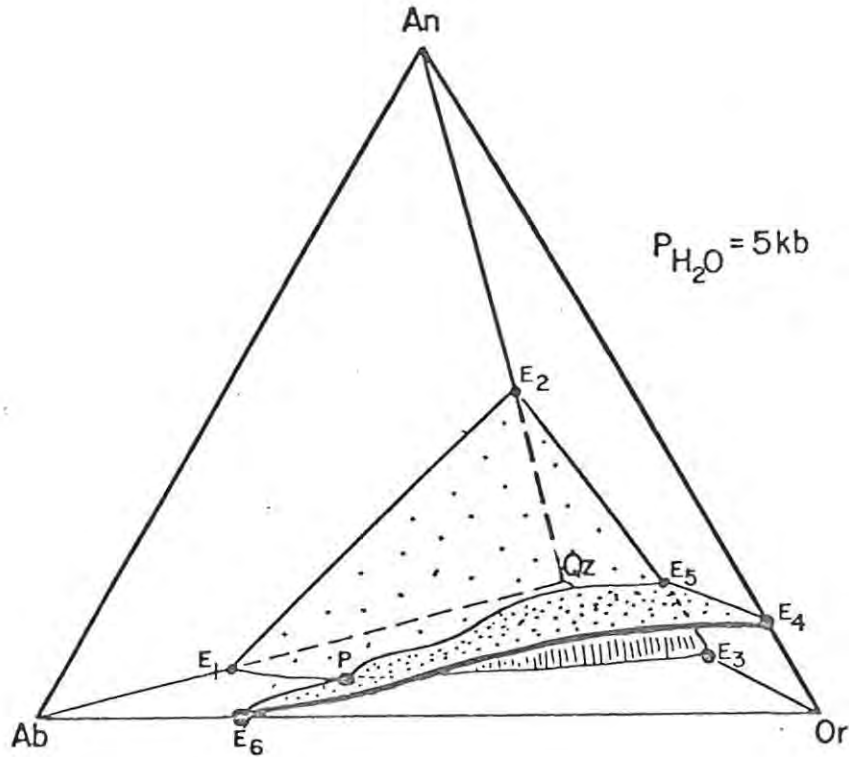
Fig. 11

Tetrahedron representing system Qz-Ab-Or-An-H_2O (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).
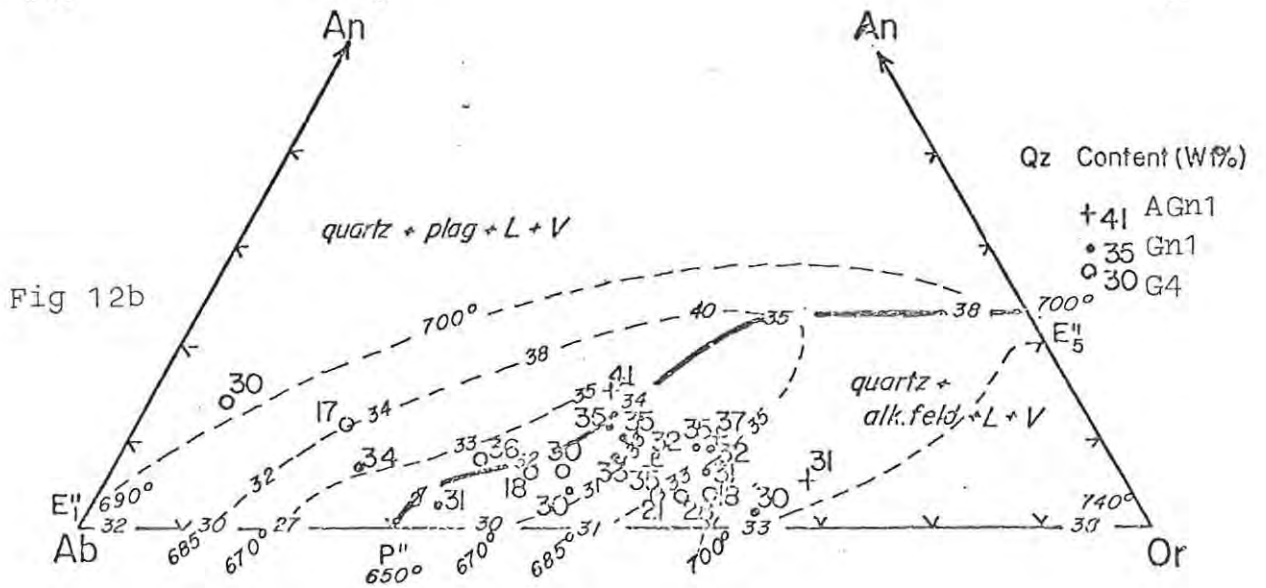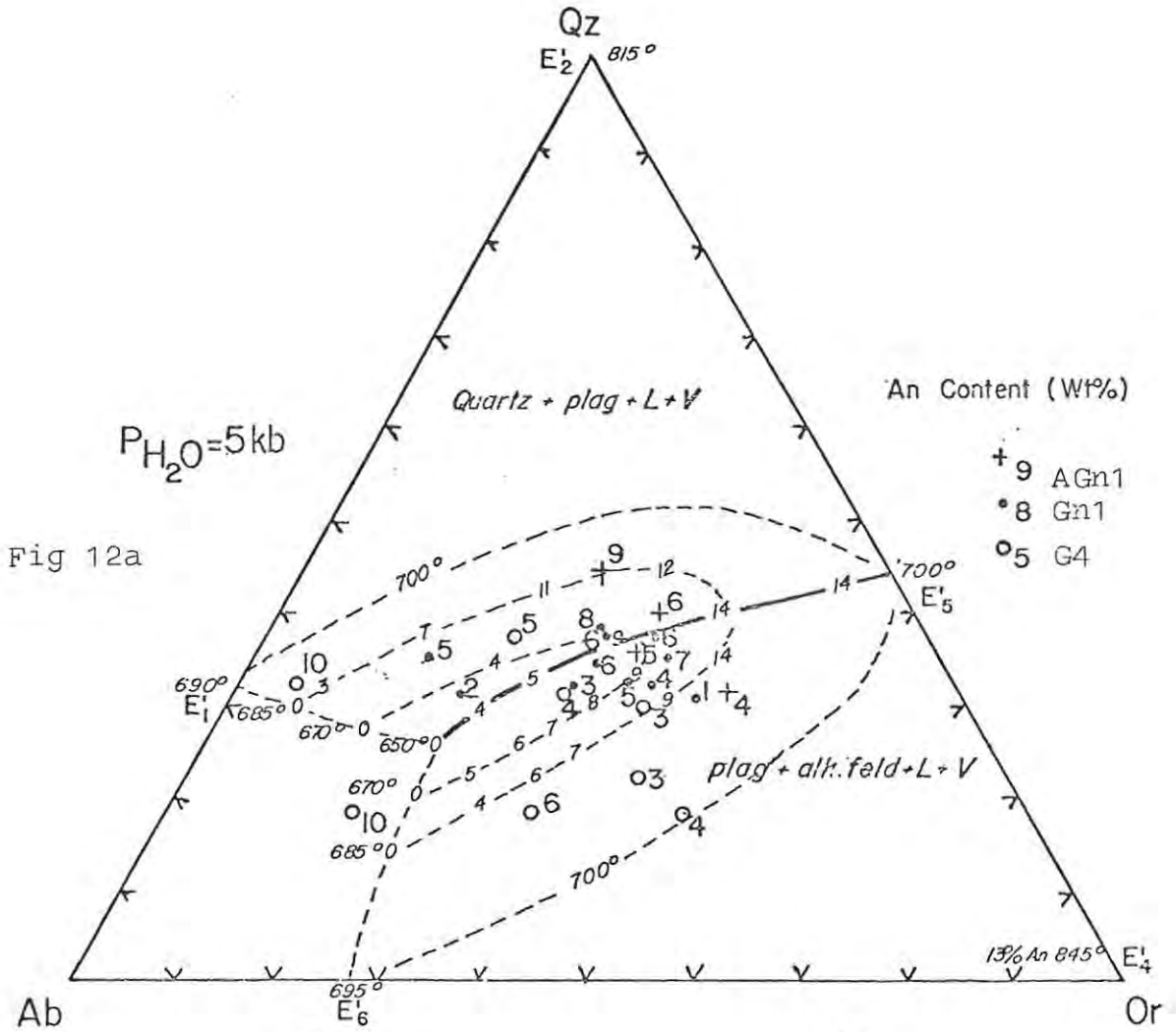
Qz-Ab-Or projection (Fig 12a) and An-Ab-Or projection (Fig 12b) showing the position of rocks from the Abbabis (AGn1) Husab (Gn1) and Rössing suites.

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 cotec-
tic line P-E$_5$, with the majority lying within the al-
kali feldspar field, and the remainder occurring with-
in 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/Qz+plag+L+V and Qz+alk.
felds+L+V respectively, crystallisation of the three
phases occurred over a very small temperature inter-
val. The crystallisation sequence is similar to that
concluded for the Abbabis augen gneisses. In the ma-
jority 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 appear-
ance of plagioclase appears to be in the range 700–670°C
which is similar to temperatures deduced from the
Abbabis plots.

The RÖSSING alaskitic granite samples exhibit a con-
siderable scatter in their plotted positions. The

samples with 10% An component lie within the pla-
gioclase 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 sur-
face, the beginning of crystallisation could have
occurred at approximately 100°C higher than the tem-
peratures implied from the adjacent isotherms. The
beginning crystallisation for these An rich samples,
by anology 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 com-
ponent 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$_5$ (see Fig 12 a), al-
kali feldspar may have begun crystallising at tempe-
ratures 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 approxi-
mately 660°C.

4.3.3.2 GAWIB 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-H$_2$O system (Fig 11).

The GAWIB 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
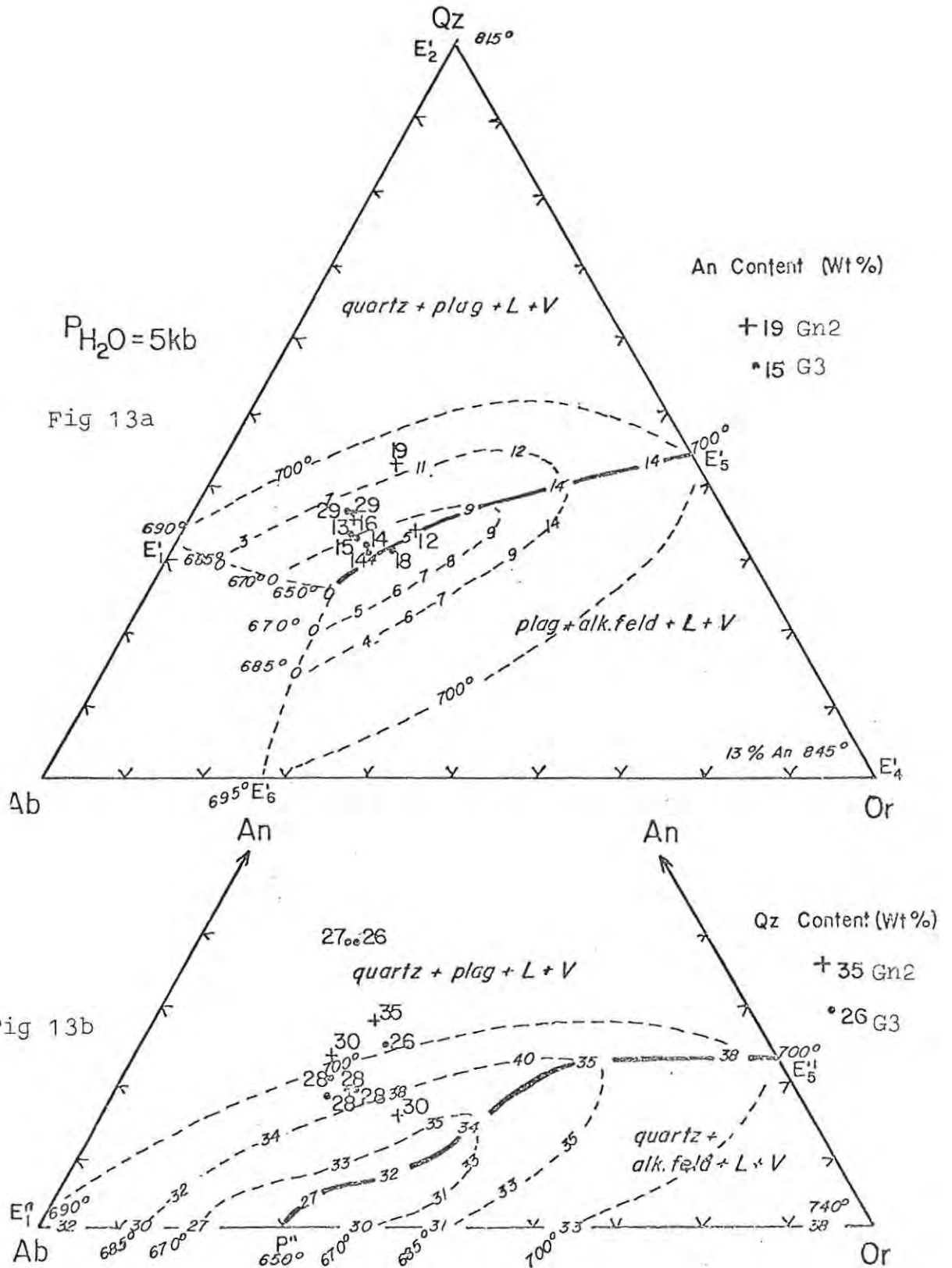
Qz–Ab–Or projection (Fig 13a) and An–Ab–Or projection
(Fig 13b) showing the position of rocks from the Salem
(Gn2) and Gawib (G3) granitoid suites.

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) $+850^{\circ}$C - crystallisation of plagioclase from the melt

ii) $675 - 680^{\circ}$C - crystallisation of plagioclase

n 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 Al_2O_3 and a higher CaO content than the other augen gneisses and this may be due to alteration and the presence of epidote.
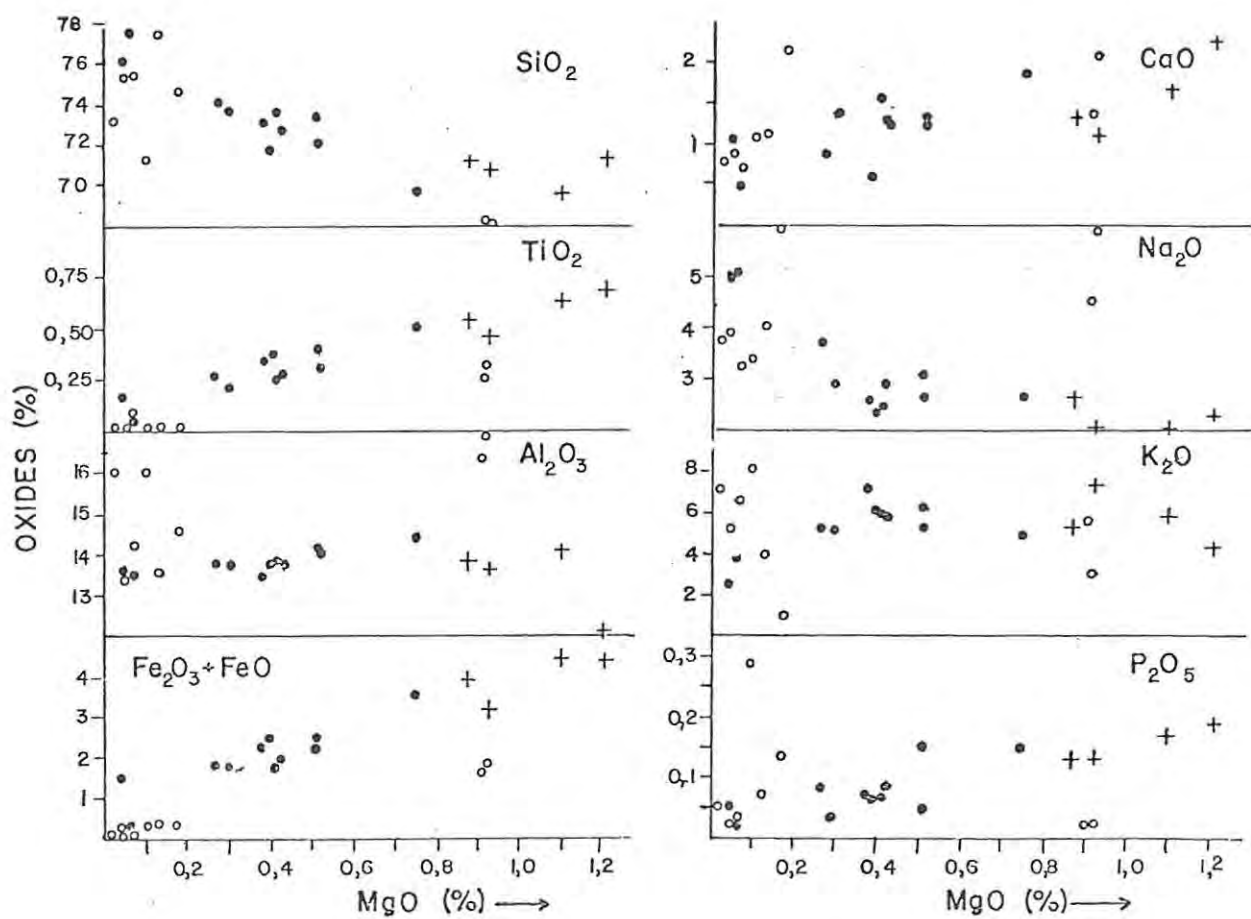
Fig. 14

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 Na_2O contents is
probably due to variation in plagioclase composition
and content as the high CaO samples are low in Na_2O
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 K_2O,
and high proportion of SiO_2 and Na_2O in relation to
the remainder of the Husab suite may imply that assi-
milation of the Nosib quartzites with the Husab granitic
melts has occurred.

The ROSSING alaskitic granites and pegmatites exhibit
a comparitively wide variation in major element propor-
tions. SiO_2, Al_2O_3 and K_2O 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 exhibi-
ted in the variation diagrams, however, do   not con-
tradict the field evidence   implying that they re-
present 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 K_2O and low Al_2O_3 proportions.

The changes in the major element proportions reflect
the different mineral compositions but are also in
agreement with a differentiation sequence. With in-
creasing differentiation(decreasing MgO), SiO_2 and
Na_2O increase and TiO_2, total Fe, CaO and P_2O_5 decrease.
K_2O and Al_2O_3 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, dif-
fering only in CaO and Na_2O 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 SiO_2, Al_2O_3 (lower) and TiO_2, total Fe and P_2O_5 (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 SiO_2, Na_2O and K_2O 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, Na_2O and K_2O contents where CaO is depleted and K_2O enriched in the Salem rocks (Miller 1973) relative to the Gawib rocks from this area. NaO decreases and K_2O increases with increasing MgO content in the Salem rocks from Omangambo and Otjosondjou whereas the Gawib rocks exhibited an increase in Na_2O and anomalous variation in K_2O 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 RÖSSING 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 <u>ABBABIS augen gneisses</u> 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 <u>HUSAB granites and granite-gneisses</u> 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
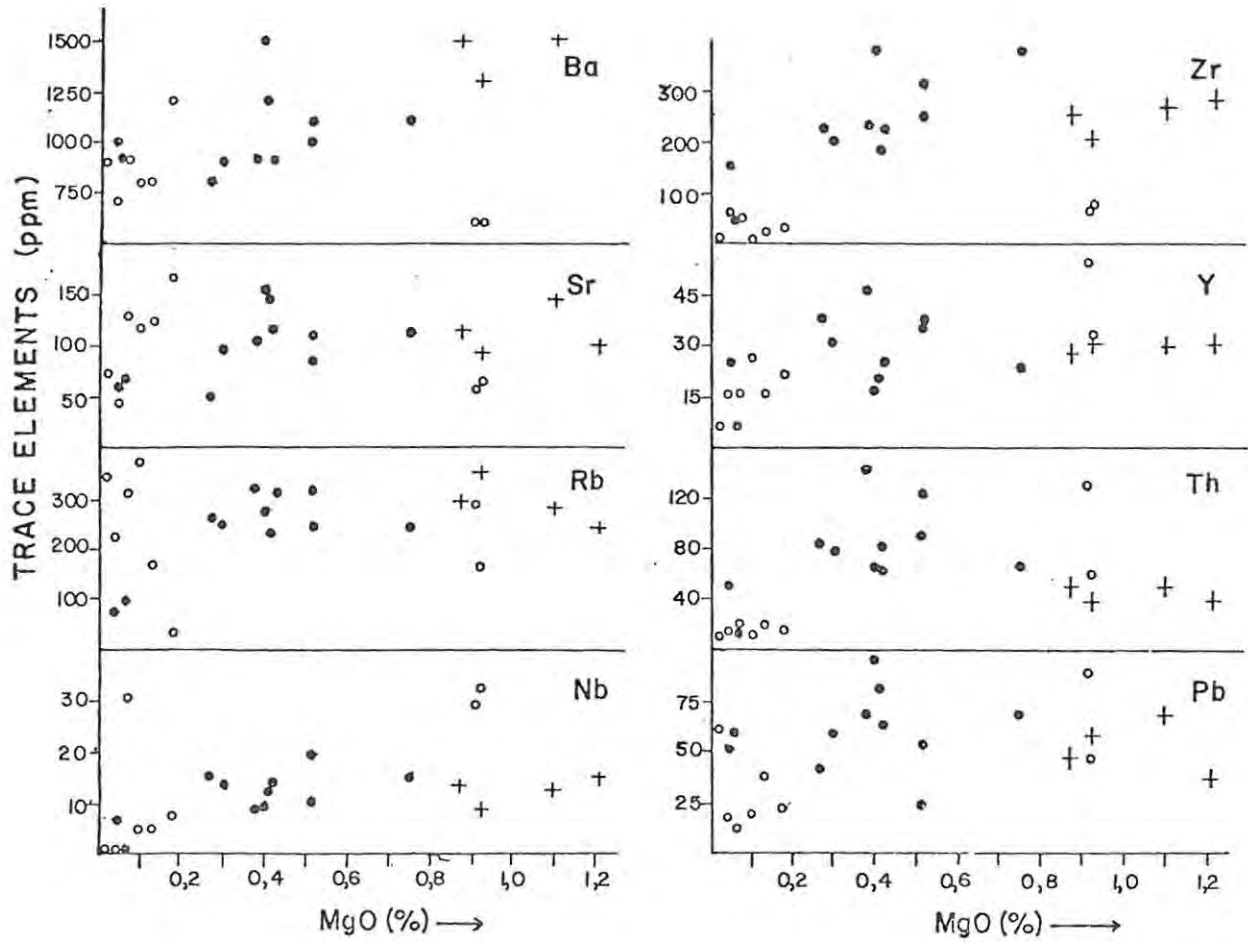
Fig 16

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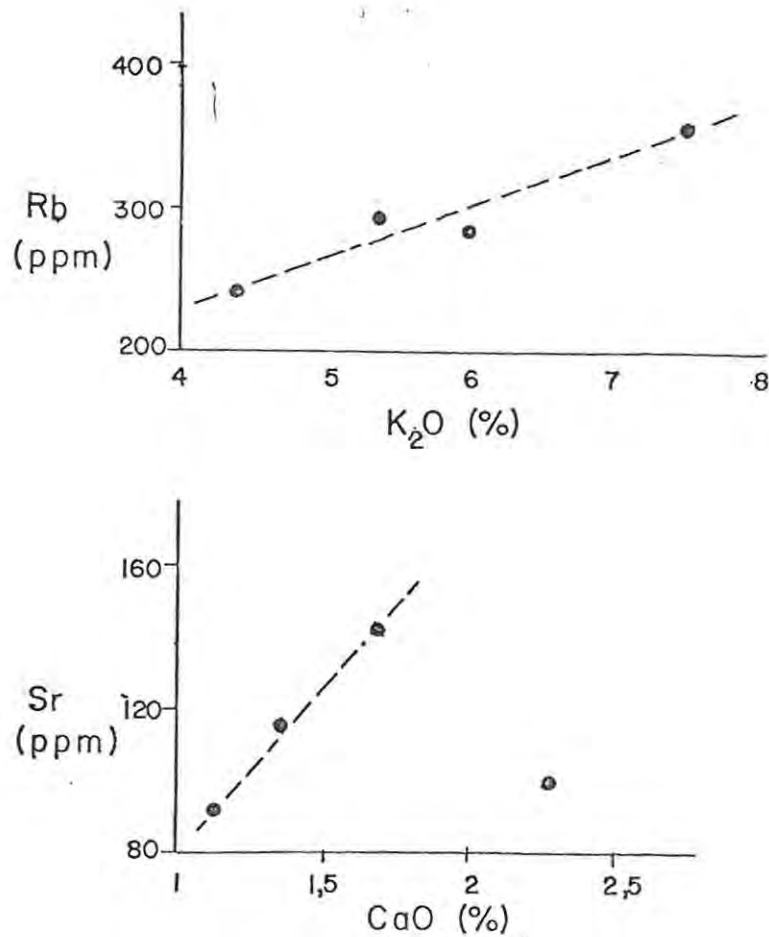
Fig. 17

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

differences from the remainder of the suite, parti-
cularly for Rb as the K_2O content in these rocks is
correspondingly low.

The R̈ossing 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 re-
latively 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 simi-
larly low MgO contents, exhibit a wide variation in
strontium, rubidium and uranium contents, but other-
wise have uniformly low concentrations of the remaining
trace elements analysed.
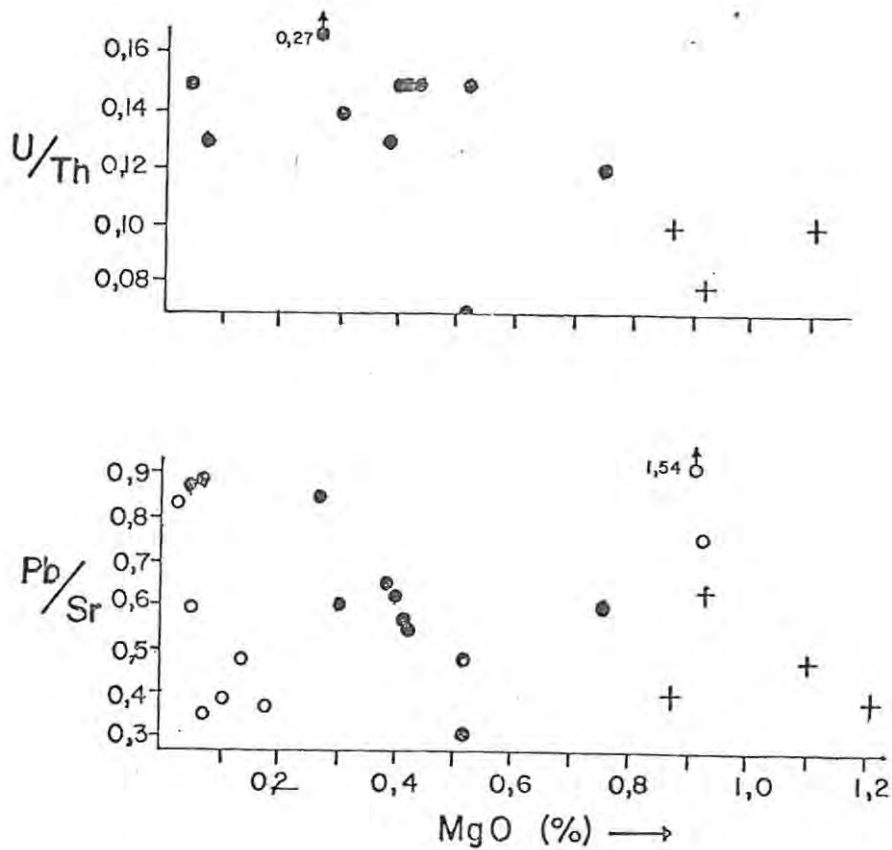
Fig. 18

U/Th and Pb/Sr trace element ratios in rocks of
the Abbabis (+), Husab (.) and Rössing (o) suites.
(All U/Th ratios in the Rössing alaskitic granites
are greater than 0.38)

The concentration of barium shows progressive de-
crease 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 con-
tents as partial melting would have resulted in selec-
tive 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
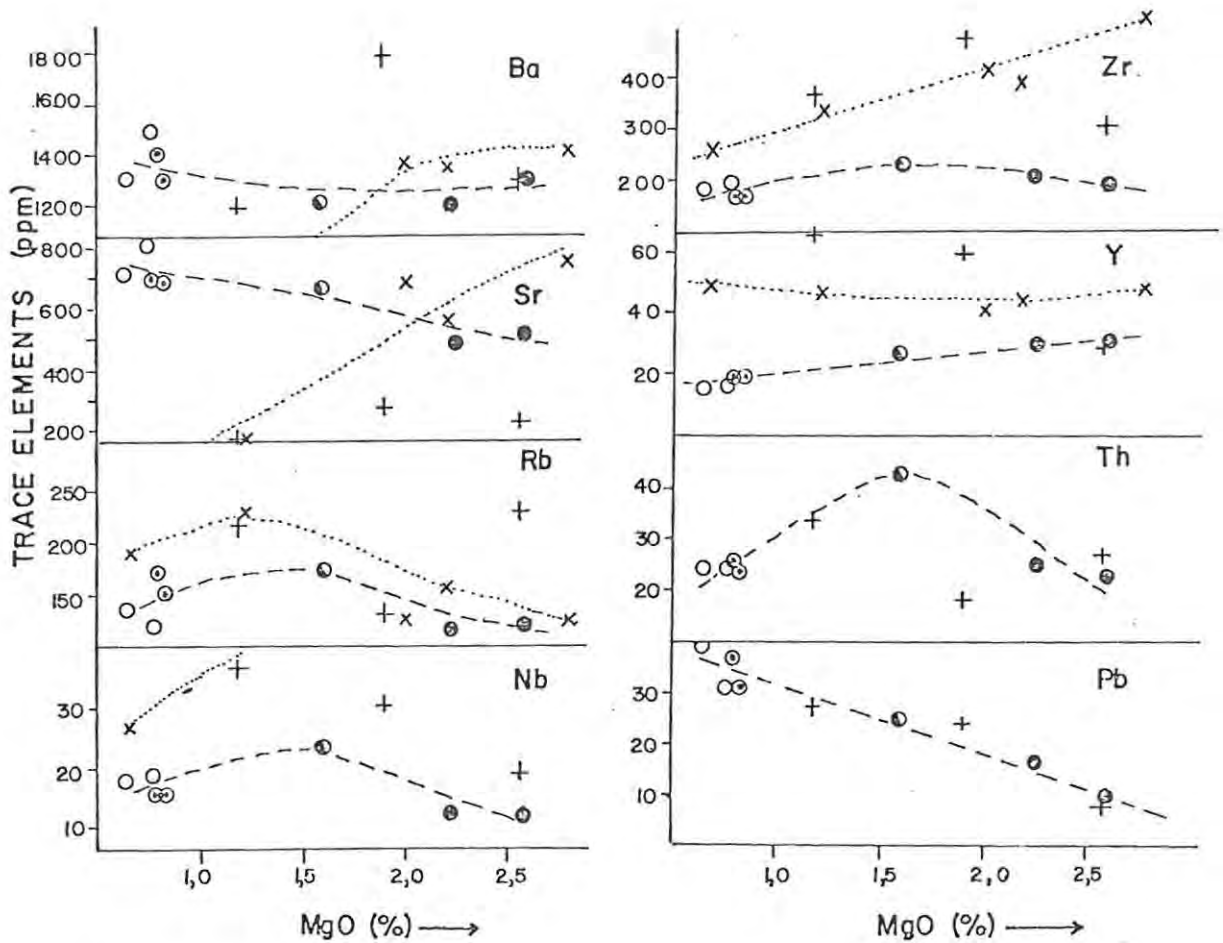
Fig. 19

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 Ti/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.
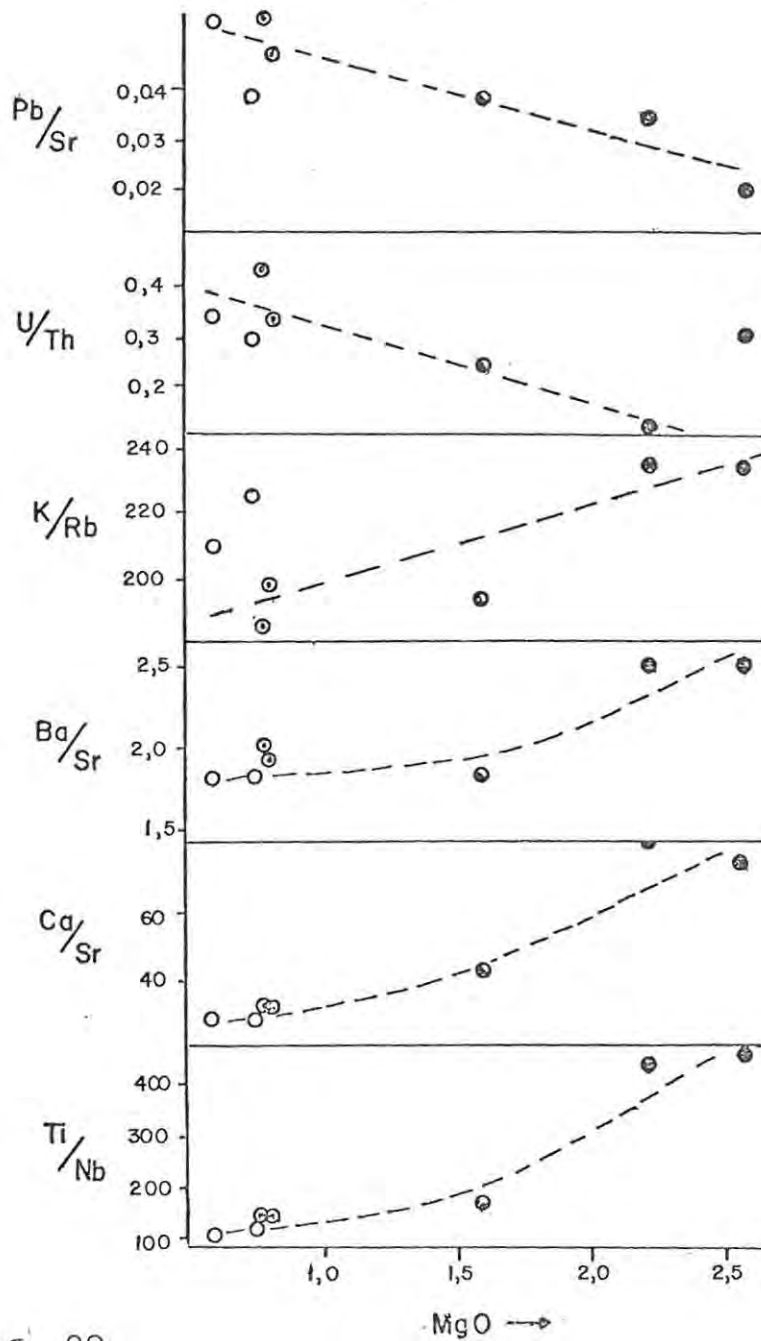
Fig. 20

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

Table 17

	Ba	Sr	Rb
alkali feldspar	4-7	3-5	0.3-0.6
biotite	1.09-15	0.08-0.67	1-4
hornblende	0.04-0.05	0.02-0.1	1.0
plagioclase	0.05-0.59	1-3	0.025-0.19

(From Phillpotts and Schnetzler (1970))

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-porphy-
ritic 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 granodio-
rites is plagioclase with hornblende and alkali feld-
spar occurring in   smaller but similar amounts
(see Table 12). The partition-coefficient data indi-
cates that rubidium would be retained in the melt
fraction unless biotite (D_{Rb} = 1-4) fractonation 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) indi-
cate that early fractionation of plagioclase occurred
but the strontium increase in the later formed members
of the stock cannot be explained by plagioclase domi-
nated fractionation owing to the partition-coefficient
values of D_{Sr}^{plag} = 1-3. It appears therefore that
although plagioclase fractionation occurred, fractiona-
tion of hornblende, possibly at an earlier stage con-
trolled the evolution of the stock. This is supported

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.
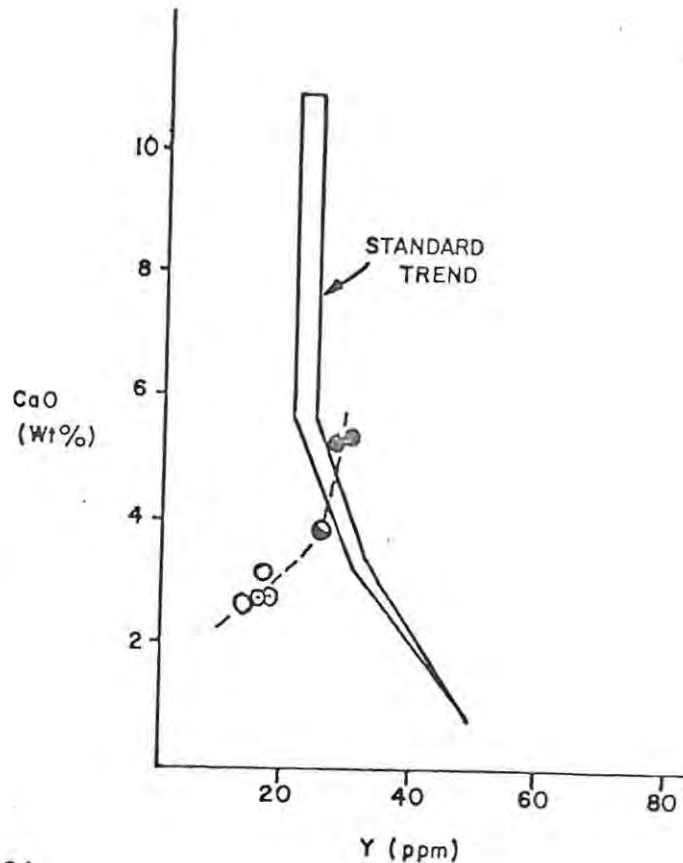
Fig. 21

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 $Qz-Ab-An-Or-H_2O$
system at 5kb indicates the Abbabis 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 crys-
tallisation 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 $Qz-Ab-An-Or-H_2O$ 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^OC 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 quarternary 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 4,5-5kb for the area of granitic occurrences. This is in agreement with results of 4 - 5kb by Jacob (1974) for the area to the north, and comparison within the experimental system of Winkler et al., (1975) for 5kb is therefore valid.

It is unlikely however, that magmatic processes always occur in the presence of excess water (Piwinskii and Wyllie, 1968, 1970; Luth, 1969) and therefore estimates of crystallisation temperatures from the quarternary 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
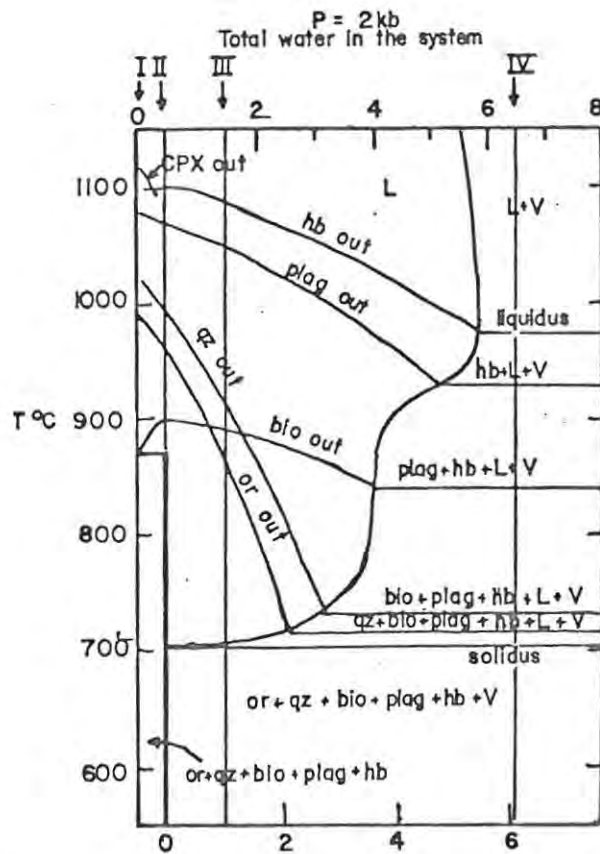
Fig. 22

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 deriva-
tives, Husab granite-gneisses and Rössing 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 else-where 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 quarternary plots. (water saturated) of $740^{\circ} - 770^{\circ}C$ 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 $(670 - 710^{\circ}C)$ 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 quarternary 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 $(740^{\circ} - 770^{\circ}C)$ 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 $720^{\circ}C$ were attained (refer section 3.4). The results therefore from the two different sources show a measure of correspondence.

The minimum temperature estimate of $770^{\circ}C$ at 5kb in the central part of the Damara belt is indicative of an exceptionally high geothermal gradient (see fig 23) in excess of $30^{\circ}C/km$ and perhaps in the region of $40^{\circ}C/km$.
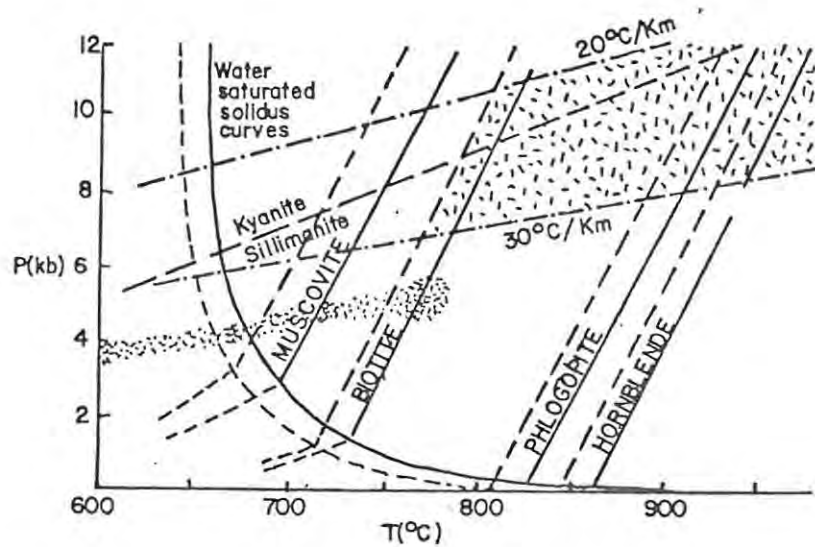
Fig. 23

PT diagram after Brown and Fyfe (1970) showing
PT field of Tumas River area (░░░░).

The Gawib granitoid stock in the eastern part of the
area, from field evidence, appears to have been em-
placed 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 con-
trolled fractionation.

Temperatures obtained from the experimental plots
for the first appearance of plagioclase from the
melt are approximately 850°C, although this tempera-
ture 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 assem-
blages 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) indi-
cated 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 crystallisa-
tion 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 Otjosond-
jou 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), Hälbich (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 orien-
tation and style of the deformation.

The notation adopted for this study is:
Bedding or bedding foliation (So), tectonic foliation
(So_1, S_1, S_2 etc.), deformation (fold) phase (F_1, F_2 etc),
fold axes (B_1, B_2 etc), lineations (l_1, l_2 etc).
The chronologic order of the deformation is denoted
by numerals i.e. F_1 being the earliest fold phase and
S_1, B_1, l_1 relate to this phase. Deformation episodes
in the pre-Damara Abbabis rocks not recognised in the
Damaran rocks are given the prefix A, i.e. AF_1 etc.

The term 'tectonic foliation' (S_1) is used to describe
planar surfaces of metamorphic origin. These are pre-
dominantly 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 So_1 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. macros-
copic structures - Turner and Weiss, 1963) are quali-
tatively introduced under 'Macrostructure ', Quanti-
tative 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

	BEDDING FEATURES	F_1	F_2	F_3
KUISEB FORMATION	calc-granofels/ metapelite inter-layering (Tinkas) (So)	Fabric indirectly controlled by So calc-granofels/ metapelite (Tinkas) migmatite (Witpoort) (So_1)-axial plane foliation S_1 and minor folds B_1	Transposition foliation, axial plane slatey cleavage in metapelite. Folding of calc-granofelses (S_2)	Large scale folding no penetrative foliation. Disruption of early foliation
KARIBIB FORMATION	Impure calc-silicate zones and banding (So)	Recrystallisation parallel to bedding So (So_1)	Small and large scale folding (incl. flow folding) S_2	Large scale folding.
CHUOS FORMATION	Not recognised	Alignment and stretching of clasts	foliation developed in matrix	Not recognised
KHAN FORMATION	Quartzite, conglomeratic layers (So)	Migmatitic banding subparallel to bedding So (So_1)	Folding of migmatitic banding, occasional transposition foliation (S_2)	Large scale folding. No penetrative foliation
ETUSIS FORMATION	Grit and conglomeratic layers, cross bedding (So)	Migmatitic development in Ns_1Gn, Ns_1R, (So_1)	Fracture cleavage in quartzites, folding and axial plane foliation. Slatey cleavage in metapelites (S_2)	Large scale folding No penetrative foliation

	AF_1 (PRE-DAMARA)	F_2 (DAMARA)
ABRABIS COMPLEX	gneissosity developed in granites augen gneisses	Folding, crenulation and refoliation of AF_1, migmatite and granite-gneiss development

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
Glücks 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 south-
west. 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 So_1 foliation having been folded
during the F_2 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 Glücks 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 Glücks 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 re-
lative movement to the one previously described occurs
on the eastern side of the Rabenrücken hills. At this
locality the eastern limb of the Rabenrücken anti-
cline 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 similari-
ties 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).
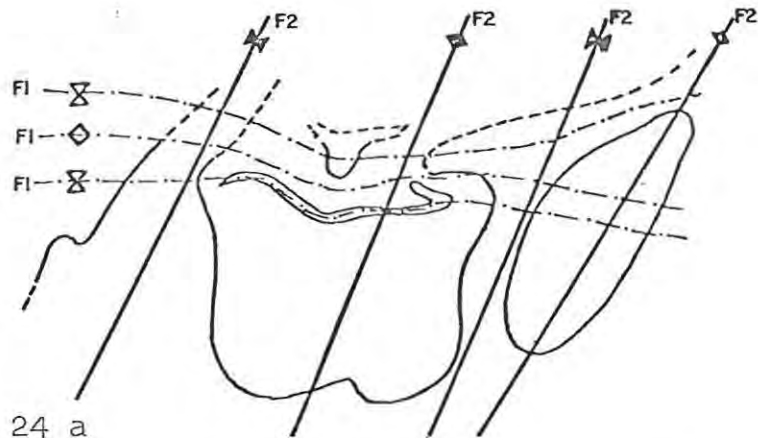
Fig 24 a

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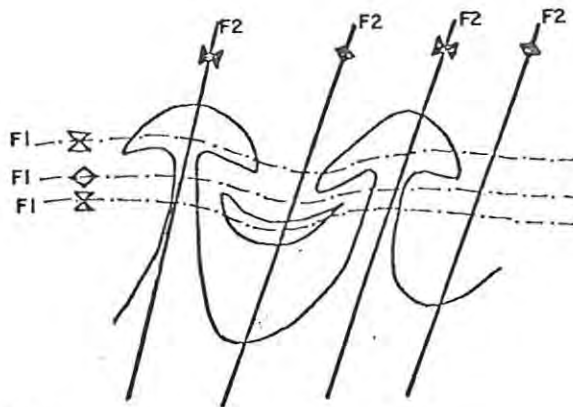
Fig 24 b

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 (F_1) 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 Eisen-hügel (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 Fench Alps and in other African orogenic belts as reported by Ramsay (1967, p383). Their formation is attributed by Ramsay (op. cit.) to rocks of con-trasting viscosities having been subjected to com-pressive stress acting from all directions or in a series of superimposed deformations. In this area the underlying basement granites and gneisses repre-sent 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 sedi-
ments 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 con-
spicuously 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 $38^{O}N$ (fig 25). This Damaran refoliation
is evident predominantly at the margins of the Inlier
adjacent to the Damara metasediments, where the folia-
tion 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 affec-
ted of the basement exposures by Damaran tectonism.
Damaran effects on the Abbabis rocks are discussed
in more detail in the next section.
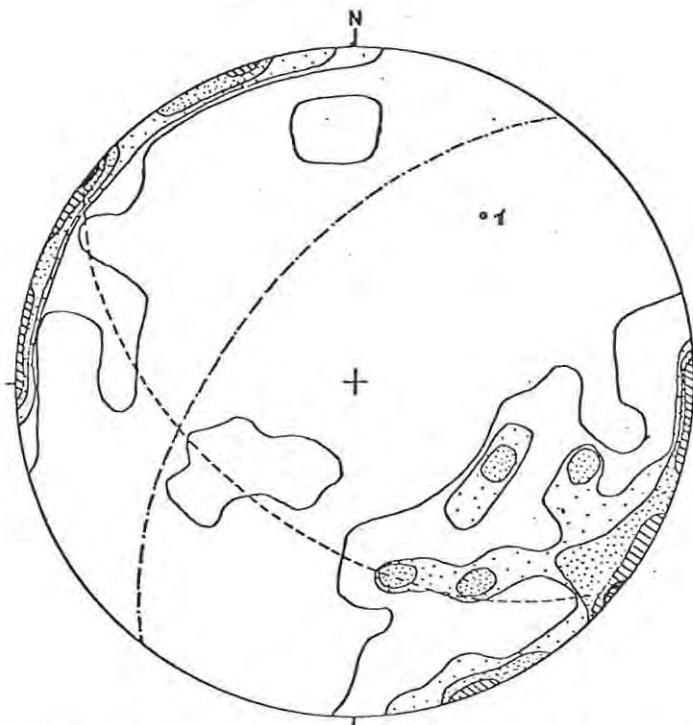
Fig. 25

    94 poles to foliation (As$_1$) in the augen-gneisses, contours 1-3-5-8-11% per 1%.

    ⌒⁓ F$_2$ axial plane (position inferred

    °₁ position of B$_2$ fold axis at northern closure of Inlier. Structural data from the Tumas river Inlier.
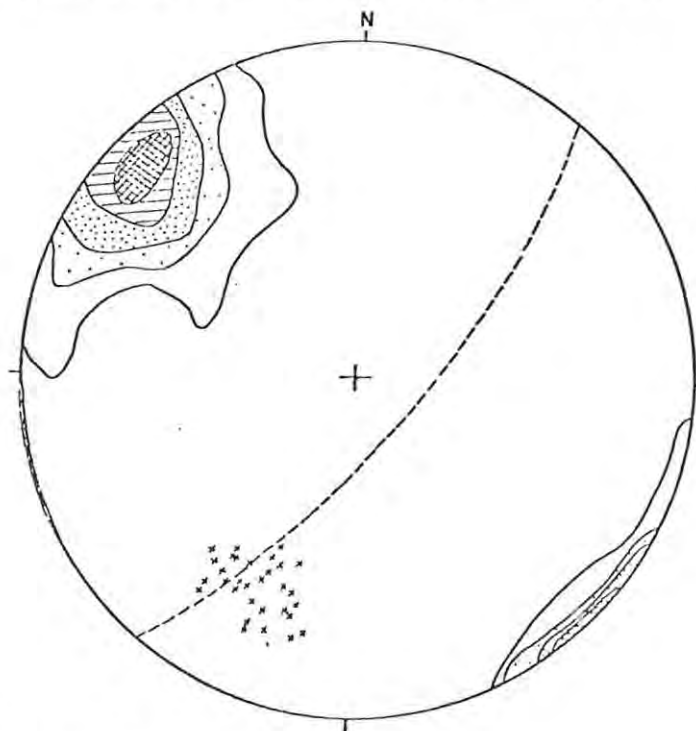
Fig. 26

    101 poles to foliation (As$_1$) in the augen gneisses, contours 1-3-9-15-35% per 1% area.

    × Al$_1$ lineations (elongate augen)

    ⌒⁓ Average attitude of As$_1$ foliation.

Structural data from the northern part of subarea 1, adjacent to the Rabenrücken hills.
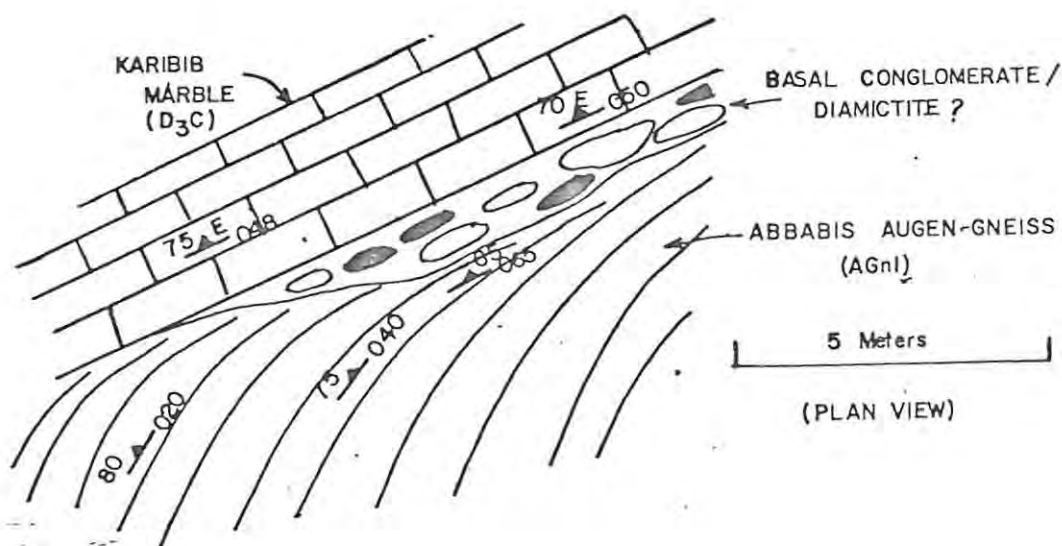
Fig. 27

Schematic diagram showing the basement (Abbabis)/
cover (Damara) relationship on the western flanks
of the Witpoortberge.

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. Dama-
ra 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 572m.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 anatectic 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
B_2 fold have a well-developed AS_1 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 simi-
lar in orientation to data obtained from measurements
on the overlying Nosib group rocks (Jacob, 1974,
Fig 15a, b, c.).

The AS_1 fabric in the augen gneisses appears to be
folded in the southern part of this subarea and the
Al lineations also show evidence of rotation (Fig 28).
This folding is probably due to the Damara F_2 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 AS_1 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, (Witpoortberge
as mentioned previously) where their original AS_1
fabric can still be recognised. Other remnant Abbabis
rocks that can be recognised are strongly sheared
augen gneisses where the AS_1 fabric no longer pre-
dominates, although augen can still be recognised
disrupted by a Damara gneissic foliation (S_1). 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.)
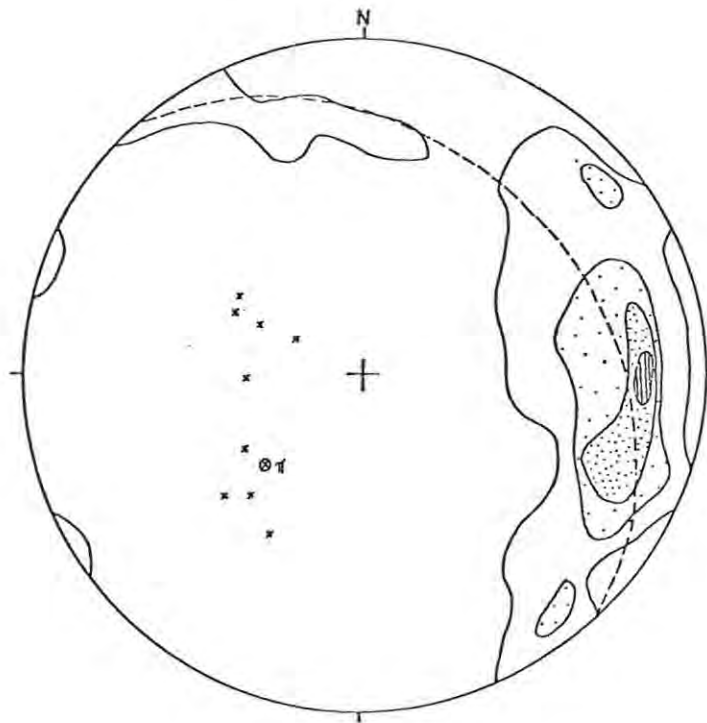
Fig. 28

49 poles to foliation (As_1) in augen gneisses,
contours 1-3-5-8% per 1% area
  × Al_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 (So_1) 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 F_1. This
lithologic-controlled banding in the Swakop/Khan
area is considered by Jacob (1974) to represent
possible axial planar foliation (S_1) 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
F_2 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 (So_1) 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 l_1 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 $15^{\circ}00'$ 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 (So_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.
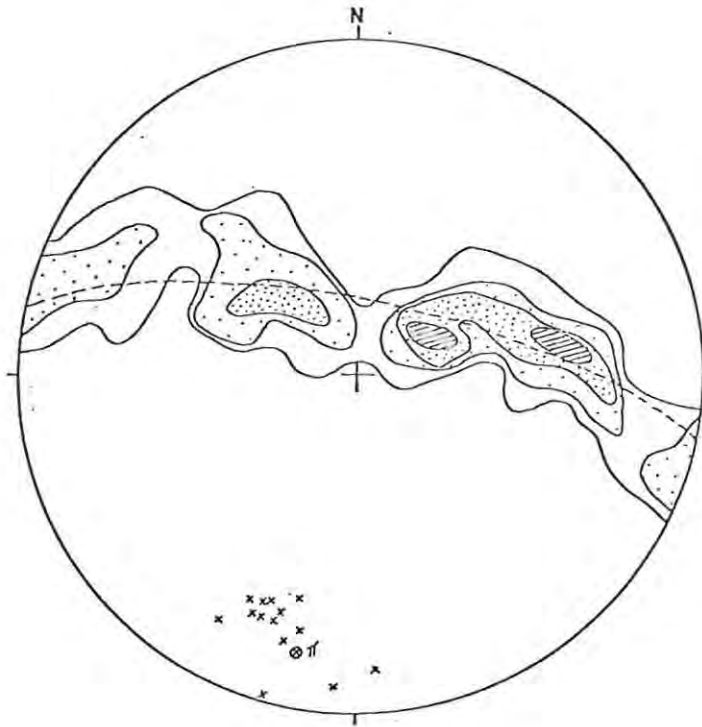
Fig. 29

35 poles to axial planar surfaces of minor B_1 folds in metapelites (Kuiseb Formation), contours 3-6-9-14% per 1% area.
x l_1 lineations (minor B_1 fold axes.
Structural data from subarea 2 (Bonfire Gorge).
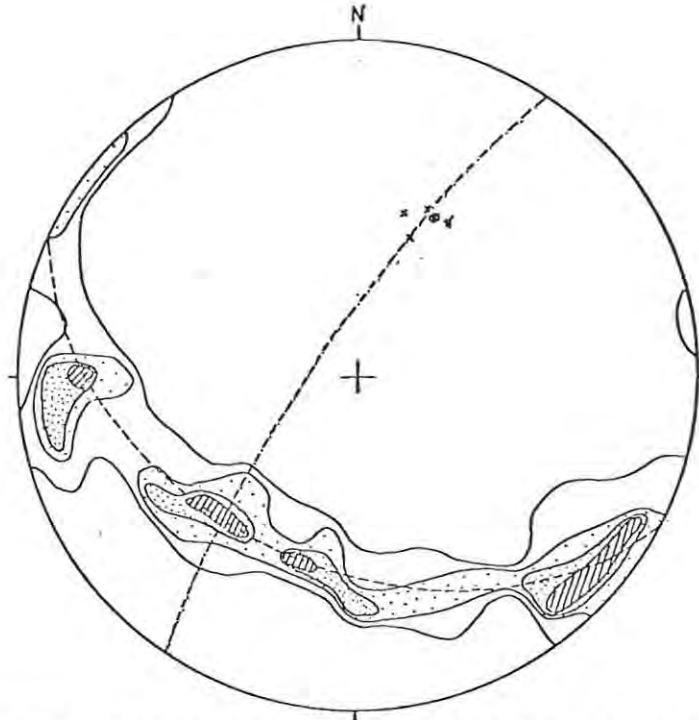
Fig. 30

36 poles to bedding (So) and bedding foliation (So$_1$) in Karibib marbles, contours 3-6-8-11% per 1%.

× l$_1$ lineations (minor B$_1$ fold axes)

⟋ inferred position of B$_1$ axial plane

Structural data from subarea 3.
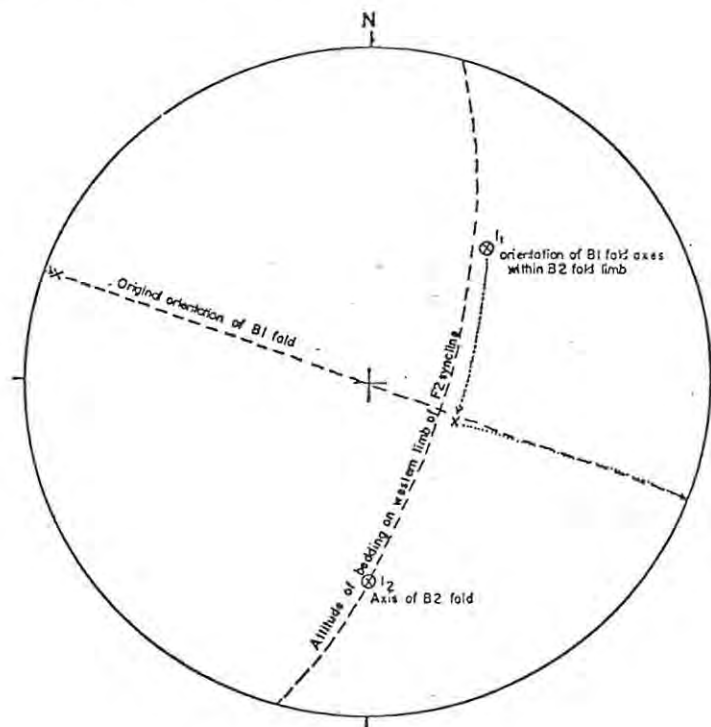
Fig. 31

Stereographic construction utilizing data from subarea 3 and the Tubasberg syncline.

Within the limbs of the V-shaped marble occurence
(von Stryk's anticline) fold closures appear to be
of F_1 age and l_1 lineations defined by the minor B_1
fold axes plunge at $\pm\ 20^{\circ}$ to the south.

5.4.2.3 F_2 STRUCTURES

The dominant NE-SW trend of the structures throughout
the area is due to the F_2 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/bed-
ding foliation (So/So_1) is disrupted by an S_2 fracture
cleavage (Plate 32). This fracture cleavage, minor
B_2 folds and associated lineations (l_2) are the main
mesoscopic F_2 features developed in these rocks.
In zones of intense deformation the banded gneisses
of the Khan Formation generally exhibit transposition
of the So_1 fabric in the B_2 folds parallel to B_2 axial
planes (Plate 33), but this So_1 fabric (banding) is
generally folded in a flexural-slip manner (Plate 34).
Fracture cleavage (S_2) is also apparant in these rocks.

The Swakop Group metasediments exhibit varied response
to the F_2 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 flexual-slip/flow
parallel to the bedding surfaces and in these bands
the S_2 fabric is only developed at the nose of folds
where minor transposition parallel to the S_2 axial
planes occurs. The metapelite bands have a prominent
S_2 fabric developed parallel to the axial planes of
the B_2 folds and transposition of the earlier folia-
tions is complete. The calc-granofels bands in
response to deformation are transposed parallel to
S_2 axial planes on the B_2 fold limbs (Plate 35), but
often only an S_2 axial planar fracture cleavage is
developed in the B_2 fold hinges (Plate 29).
Transposition is marked where the B_2

folds are isoclinally 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 So_1 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).

TINKAS META-PELITES
(D_4S)
axial planar slatey
cleavage (S_2)

KARIBIB MARBLES (D_3C)
bedding foliation (So_1)

TINKAS CALC-GRANOFELSES
(D_4S)
transposed parallel to S_2 on limbs
fracture cleavage (S_2) in hinge zone

~20 Meters
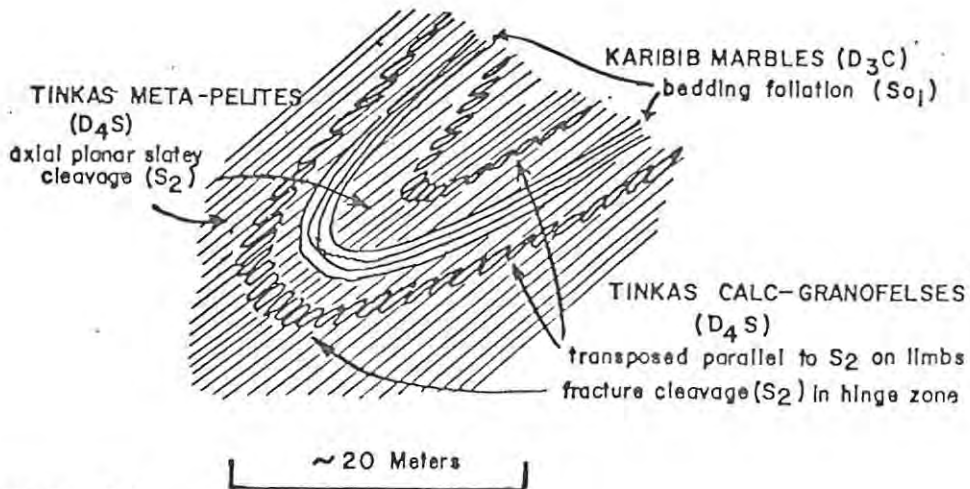
Fig. 32

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–granofels bands where the original So_1
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, ie. 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 meta-
sediments plunge to the SSW (200°), Fig 33. The
spread of the So_1 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° W.
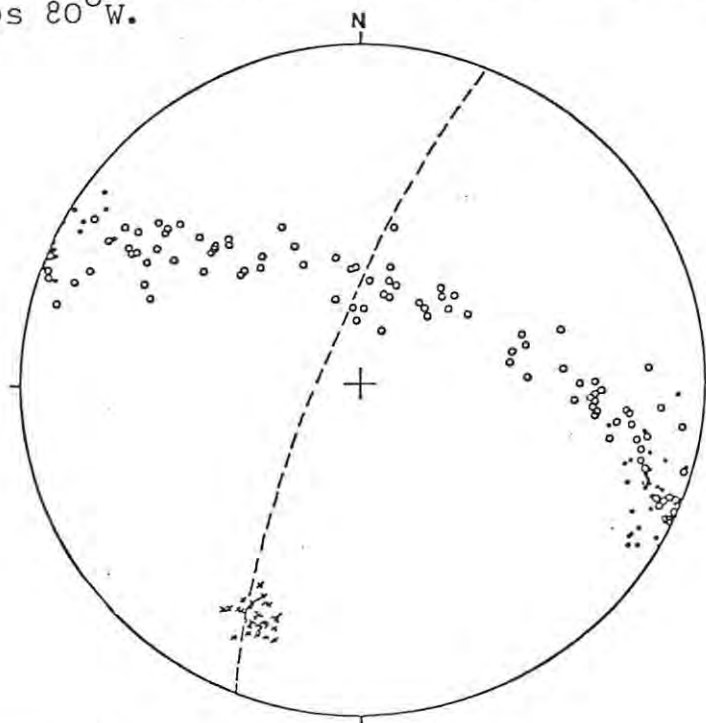
Fig, 33
o   104 poles to bedding foliation (So_1) and
axial plane foliation (S_1) in Kuiseb metasediments.
   •   30 poles to axial planes of minor B_2 folds and
axial planar foliation (S_2).
   x l_2 lineations (mullions, minor B_2 fold axes
intersection of S_1 and S_2 surfaces.
Structural measurements defining the B_2 folds at
subarea 2 (Bonfire Gorge).

Measurements from the Kuiseb formation metasediments
at the Lucasberge (sub-area 5) indicate B_2 folds plun-
ging 40° to 231° with an axial plane dipping 80^OW
(fig 34). The distribution of So_1 poles in the
plot indicates open type folding, although the
structural form lines on a macroscale indicate simi-
lar type folding (see maps 1 and 2).

The tightly folded Karibib marbles and Kuiseb meta-
pelites and calc-granofelses (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^OW and the distribution
of the So_1 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 B_2 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
So_1/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 B_2
axis plunging 25° to 065°, (fig 37) with an axial
plane dipping to the west. The trend of the axial
surface (S_2) at the two closures of Glücks dome differs
This is possibly due to either a later F_3 deformation
event trending WNW-ENE or to the proximity of the
underlying basement.

In the Zebraberge, at the southern closure of the Ida
dome (sub-area 8) Khan Formation gneisses have been
folded about a B_2 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 (Ns_2Gn) occur in an overturned
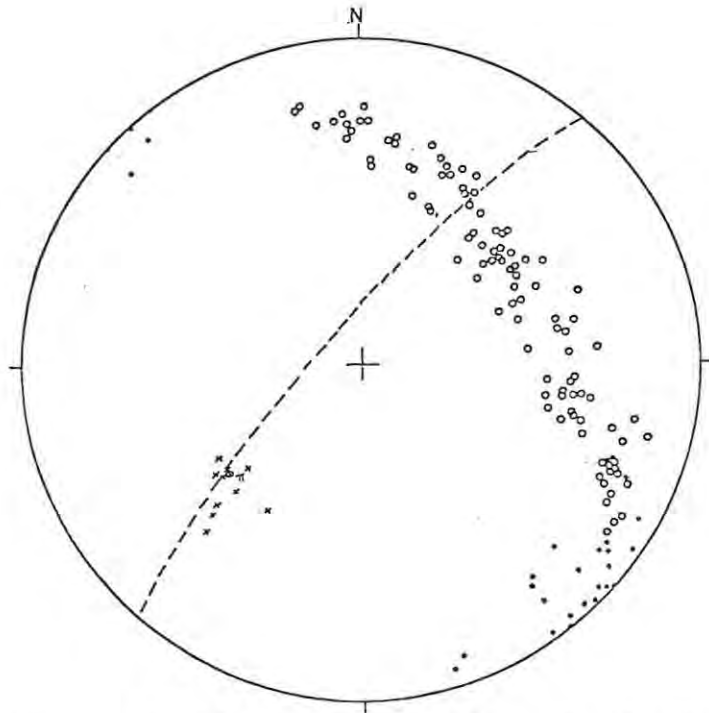
Fig. 34

  o 100 poles to bedding (So),bedding foliation (So$_1$) in Kuiseb metasediments.

     • 27 poles to axial plane cleavage/.foliation (S$_2$)

     ✗ l$_2$ lineations (So$_1$/S$_2$ intersection,mullions,minor fold axes)
     Structural measurements defining B$_2$ folds at subarea 5
     (Lucasberge).
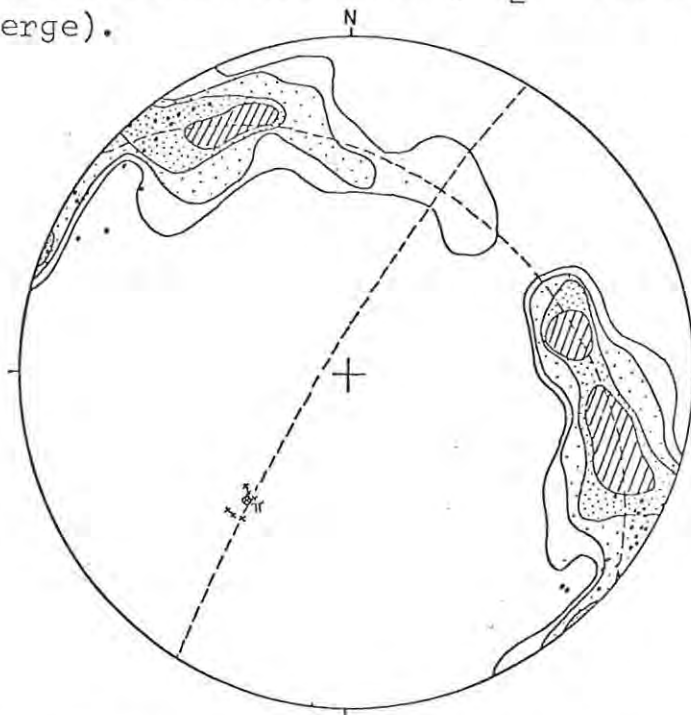
Fig. 35

    73 poles to bedding (So)/bedding foliation (So$_1$) in Kuiseb metasediments and Karibib marbles,contours 1-3-7-10% per 1% area.

     • 35 poles to axial plane foliation (cleavage (S$_2$)

     ✗ l$_2$ lineations (mullions,minor fold axes,So$_1$/S$_2$ intersections).

Structural measurements from subarea 4
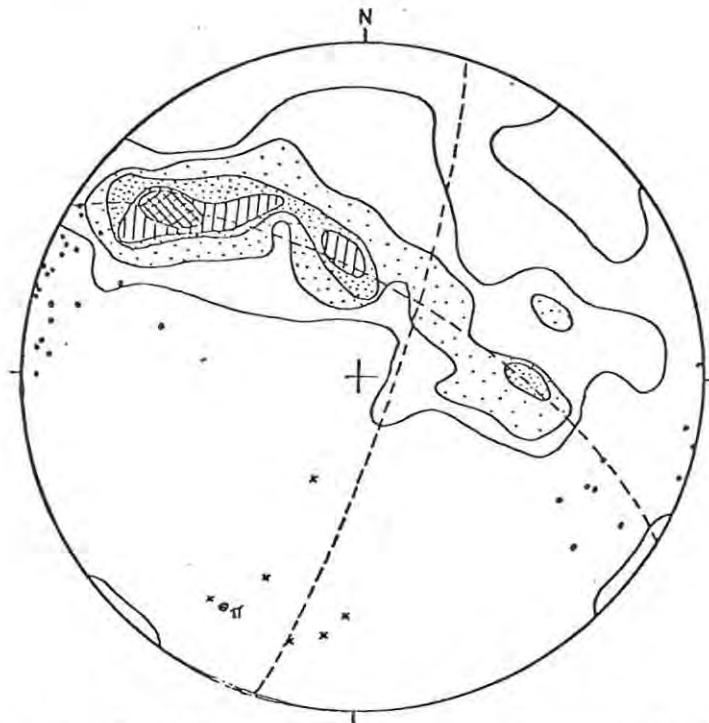
Fig. 36

97 poles to bedding (So) 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, So/S_2 intersections.) Structural measurements from the southern closure of Glücks dome (subarea 6)
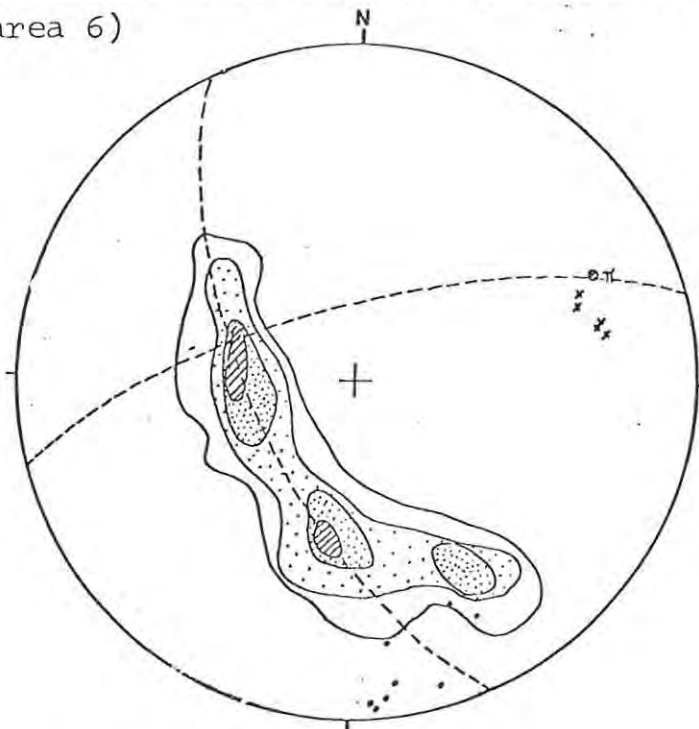
Fig 37

41 poles to bedding (So) 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, So/S_2 intersections.) Structural data from subarea 7 (northern closure of Glücks dome).

syncline. The F_2 folding here is of very open buckle type and the regional foliation (So_1) trends E-W. Lineations (l_1) defined by hornblende growth on So_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.
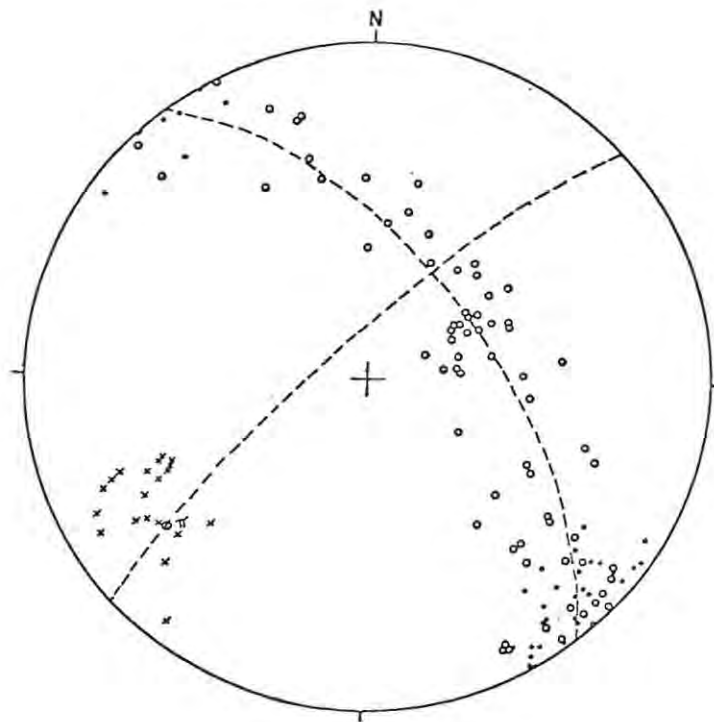
Fig. 38

o   72 poles to banding (So_1, Si_1) in Khan gneisses
•   35 poles to axial planar fracture cleavage (S_2) and to axial planes of minor B_2 folds.
x l_2 lineations (elongate minerals, So_1/S_2 foliation intersection, minor B_2 fold axes .)
Structural data from subarea 8 (Zebraberge).
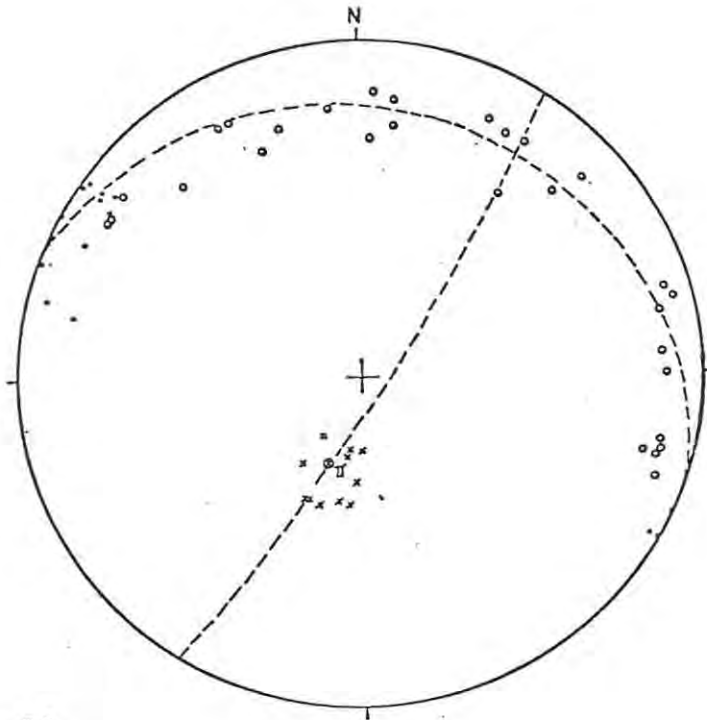
Fig 39

o   30 poles to bedding (So) foliation (So$_1$) in
Karibib marbles and Kuiseb metasediments.

•   13 poles to axial planar cleavage (S$_2$) (slatey
cleavage in metapelites and fracture cleavage in calc-
granofels).

×   l$_2$ lineations (mullions and minor B$_2$ fold axes).
Structural data from subarea 10.

South of the Endklippe hills in the western part of
the area (sub-area 10) B$_2$ folds measured in inter-
bedded 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 (l$_2$)
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 B$_2$ folds is
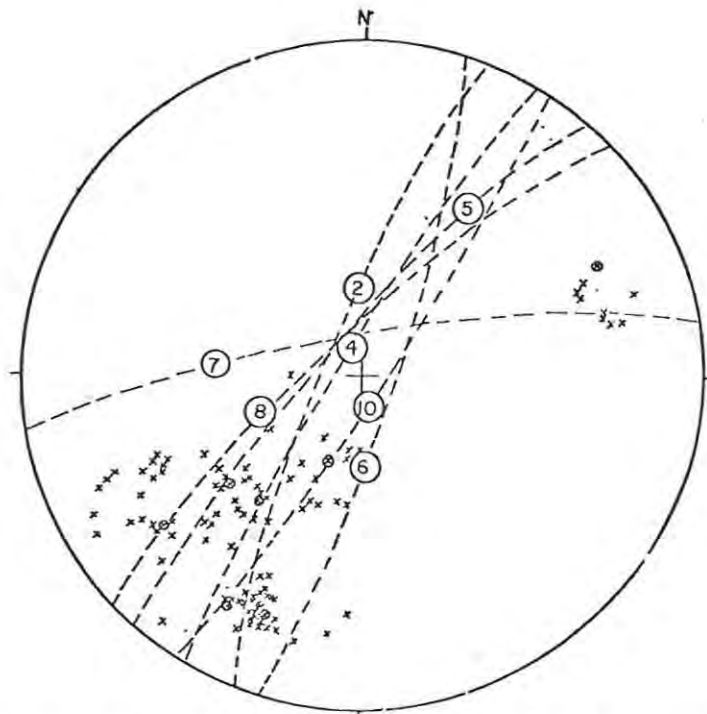
Fig. 40

/ average axial plane (S_2) attitude for each respective subarea

× l_2 lineations (minor B_2 fold axes, mullions, boudins, elongate minerals, So_1/S_2 intersections).

⊛ π points from the different subareas.

Synoptic plot of structural data from throughout the area.
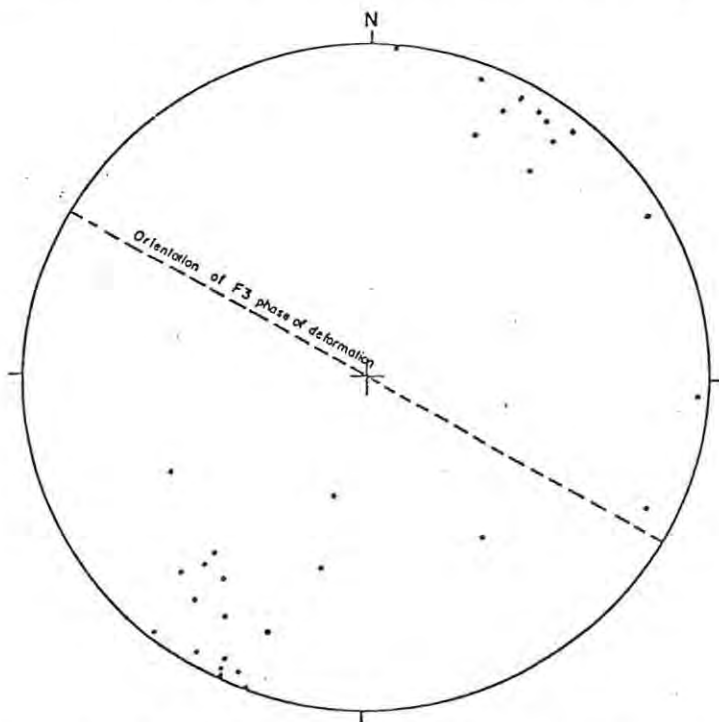
Fig. 41

· 31 poles to joints in Etusis quartzites.

/ Probable B_3 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 F_3 STRUCTURES

A variety of non-penetrative post-F_2 structures have been observed. These rarely exceed several metres in scale and their orientation appears to coincide with the broad buckling (F_3) 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 S_2 axial planar slaty cleavage was observed and minor folding (Plate 38) occurs about an axis trending between 115° and 125°. This disruption of the S_2 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 (So/So_1) 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 F_3 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 dispacement 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 grani-
toid suite represent late to post tectonic granitoids
and the Husab and Salem suites are syn-tectonic grani-
toids.

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 deve-
loped 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
($\pm$ 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 phenocyrsts in these rocks were com-
monly 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) sug-
gests that this granitoid suite developed at an early
stage (F_1) and the foliation (S_1) was further deformed
during the F_2 phase. The porphyritic leucogranite
member of this suite commonly has crosscutting features
and is not foliated which suggests emplacement post-
dated the F_2 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 F_2 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 unorien-
tated K-feldspar phenocrysts.
The ROSSING alaskitic granites occur in the vicinity
of the Zebraberge as coarse pegmatitic bodies asso-
ciated with the metasedimentary rocks of the Khan
and Karibib Formations. Narrow granite/pegmatitic
veins or large ($\pm$ 1km) ovoid granitic/pegmatitic/
masses often crosscut the F_2 structural fabric or
occasionally develop along F_2 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 struc-
tures. 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 orien-
tation of the deformation phases involved (fig 42).

Roering (1961)   Smith (1965)   Jacob (1974)   Present
                                                Investigation
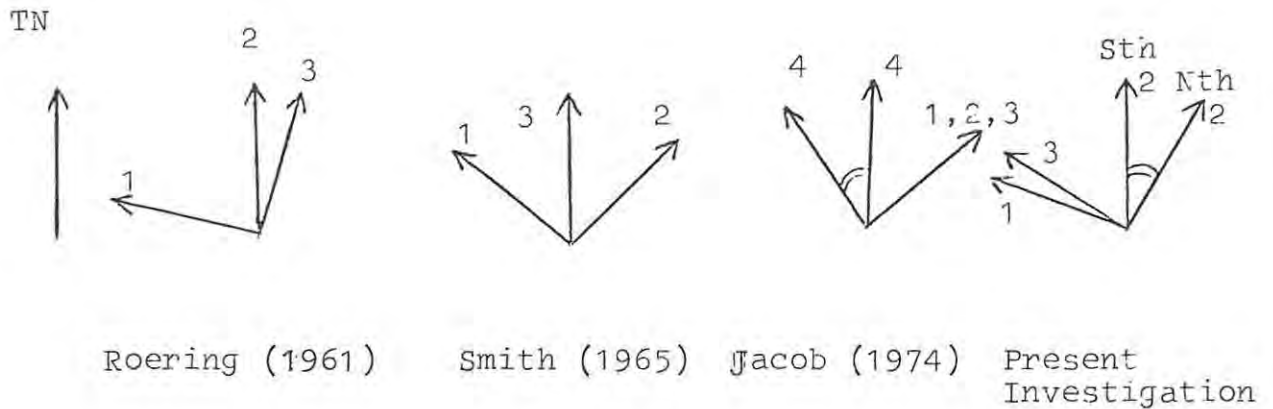
Fig 42

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 $\dot{B}_1$ 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 B_2 folds as predominantly plunging NE, whereas in the area studied by Jacob (1974) the B_2 fold axes were found to plunge in both a NE and SW direction. In this area, the majority of the B_2 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 F_2 deformation phase, or the F_1 structures in the Khan/Swakop area may have been overturned to the north, with interference between the F_1 and F_2 fold phases also producing a variation in the direction of plunge of the F_2 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 (Chuos 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 granit-
ization. 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 predominently 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 $\pm$ 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 anatectic 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 3,5-4kb. 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 4,5-5kb. 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 F_2 phase.

Late stage mobilisates however, form part of both
suites in the form of leucogranites. The Rössing
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 470my. (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-H_2O after the method of Winkler et al, (1975). The temperatures obtained for the beginning of crystallisation for the Salem and Husab suites are 740°-770°C and 670°-710°C 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 740-770°C 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 850°C.

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), Kröner (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 SiO_2 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 absorbtion 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 (I)

Instrumental conditions for analysis of major elements.

SAMPLE PREPARATION	Fusion Disc									Whole rock pellet
ELEMENT	Si	Ti	Al	Fe	Mn	Mg	Ca	K	P	Na
GENERATOR KV	50	50	50	50	50	50	50	50	50	50
MA	40	40	40	40	40	40	40	40	40	40
TUBE	Cr	Cr	Cr	Cr	Cr	Cr	Cr	Cr	Cr	Cr
COUNTER	Flow	Flow	Flow	Flow	Flow	Flow	Flow	Flow	Flow	Flow
COLLIMATOR	Coarse	Fine	Coarse	Fine	Coarse	Fine	Fine	Fine	Coarse	Coarse
CRYSTAL	PET	LiF 200	PET	LiF 200	LiF 200	LiF 200	TlAP	LiF 200	Ge	TlAP
TIME	40	10	40	20	40	200	10	10	20	200
POSITION	$K\alpha$ lines →									
Al-FILTER	Out	Out	Out	Out	In	Out	Out	Out	Out	–
BLANKS	Fe/Ca	SiO_2	SiO_2	SiO_2	SiO_2	SiO_2	SiO_2	SiO_2	SiO_2	SiO_2
STANDARDS	USGS international standards, NIMROC standards and UCT.in house Stds (Values for USGS Int.Stds,as listed by Flanagan (1973))									

(Pulse height discrimination applied)

Table II

Instrumental conditions for analysis of trace elements

SAMPLE PREPARATION	Whole Rock Pellet						
ELEMENT	Nb	Zr	Y	Sr	Rb	Th	Pb
GENERATOR KV	55	55	55	55	55	55	55
MA	40	40	40	40	40	40	40
TUBE	W	W	W	W	W	W	W
COUNTER	Scintillation ————————————————→						
COLLIMATOR	Fine	Fine	Fine	Fine	Fine	Fine	Fine
CRYSTAL	LiF 220	LiF 220	LiF 220	LiF 220	LiF 220	LiF 220	LiF 220
TIME	100 sec on background, 200 sec on peak						
POSITION	$K\alpha$	$K\alpha$	$K\alpha$	$K\alpha$	$K\alpha$	$ThL\alpha$	$PbL\beta$
BLANKS	SiO_2	SiO_2	SiO_2	SiO_2	SiO_2	SiO_2	SiO_2
STANDARDS	USGS international standards, NIMROC standards and UCT in house standards (See Table III for standard values used)						

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).

ELEMENT		Nb	Zr	Y	Sr	Rb	Th	Pb
USGS	AGV	11,0	237	20,3	670	65,0	7	35
STANDARDS	GSP	24,5	535	25,0	238	255	108	55
	BCR	10,5	195	34,5	340	46,5	–	–
	G–2	9,0	324	10,7	492	170	25	29
UCT in	OK272	74,5	164	29,0	113	88,0	–	–
house Stds	M–38	18,5	199	40,5	80,0	212	–	–
NIMROC Standards	NIM-G	–	–	–	–	–	50	48

FeO was determined after the method of Shapiro and Brannock (1962) using 0.5gm of powder which was digested in a mixture of H_2SO_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 out-crop 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.

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PLATES

Plate 1

Typical Namib landscape of broad plains with inselbergs.

Plate 2

Contorted metasedimentary remnants of Abbabis age lie within highly deformed augen-gneisses (AGn_1) 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 1

Plate 2

Plate 3

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 4

Plate 5

Plate 6

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 F_1 tectonic episode.
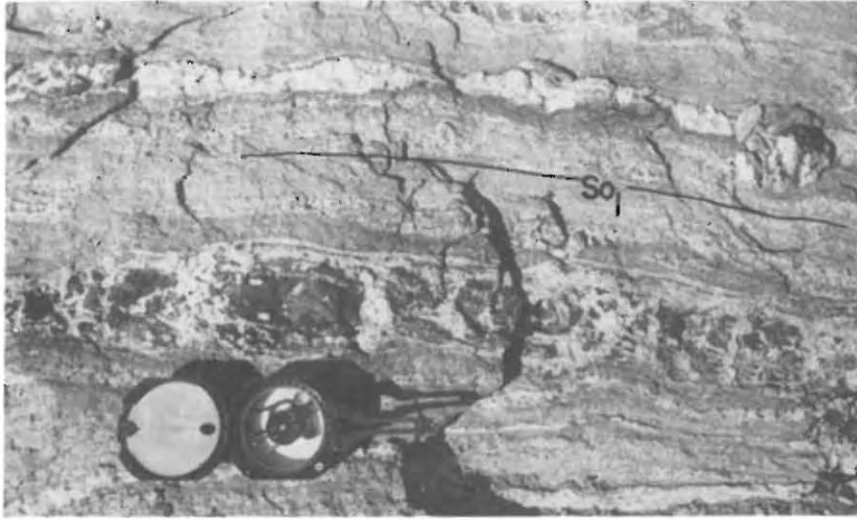
Plate 7

Plate 8

Plate 9

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 10
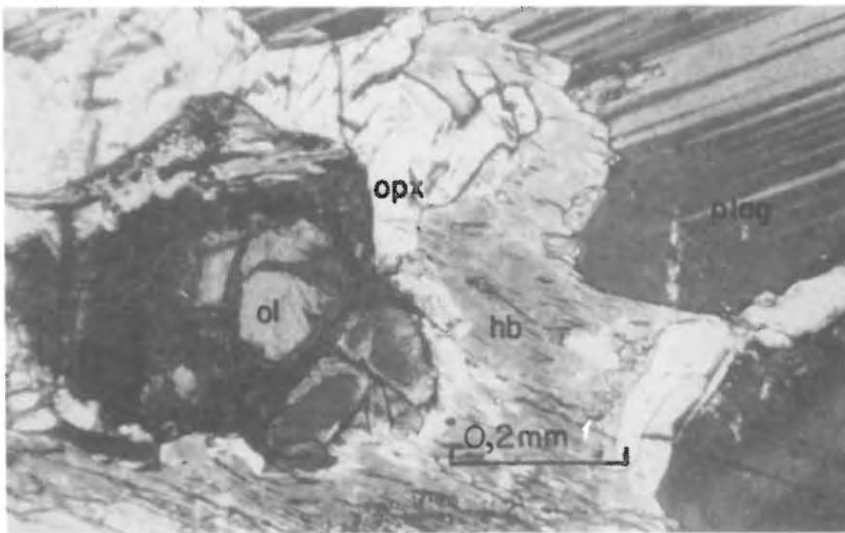
Plate 11
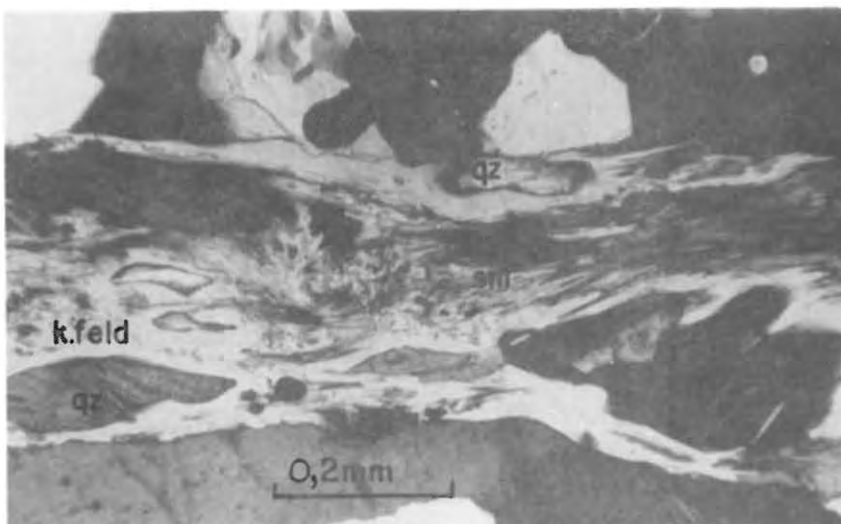
Plate 12

Plate 13

Fibrous aggregates of sillimanite present in feldspathic
quartzites (Etusis Formation) west of the Qhameleon 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).
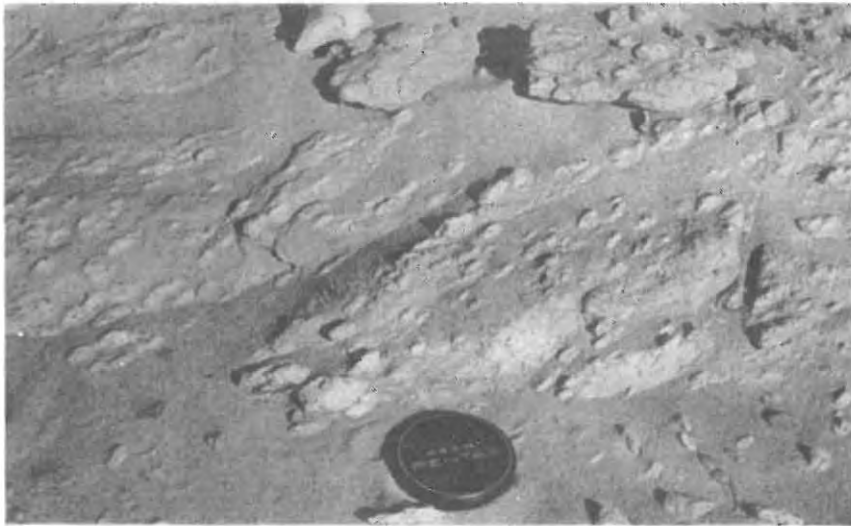
Plate 13
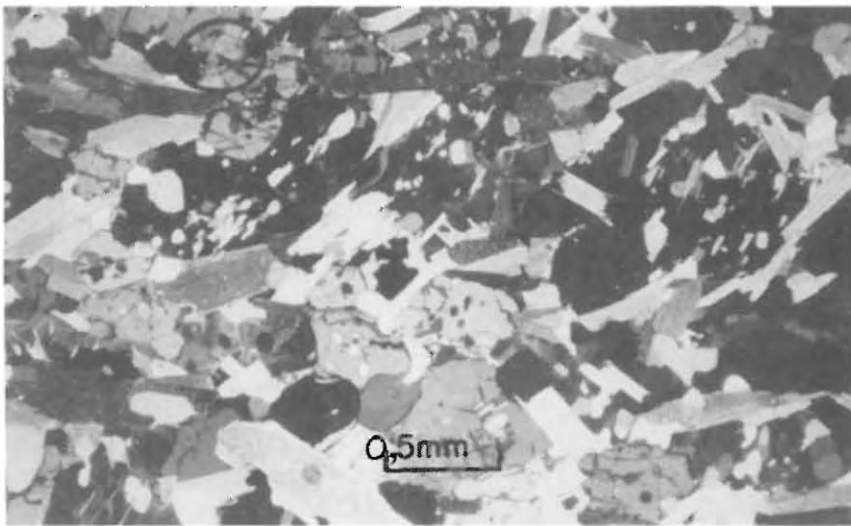
Plate 14
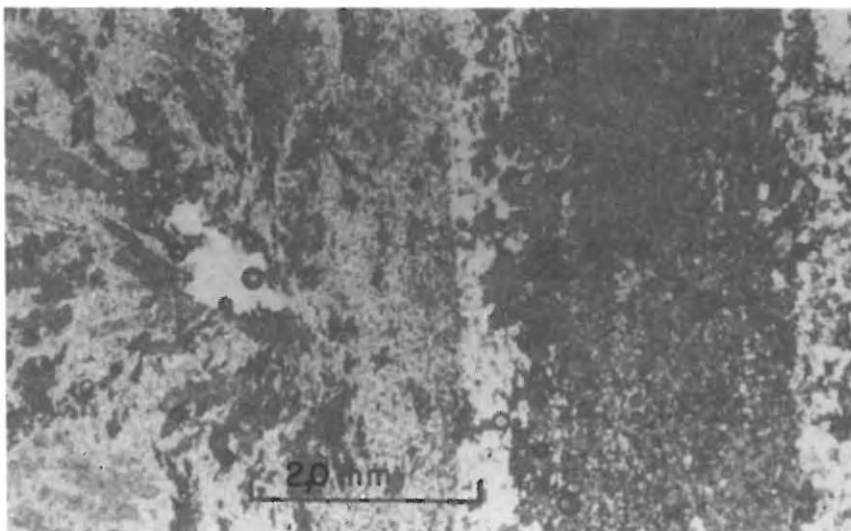
Plate 15

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).
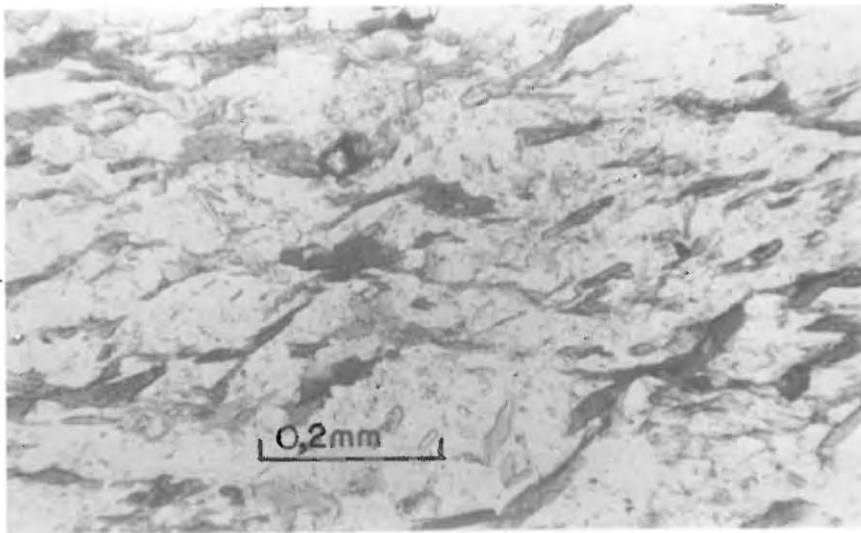
Plate 16
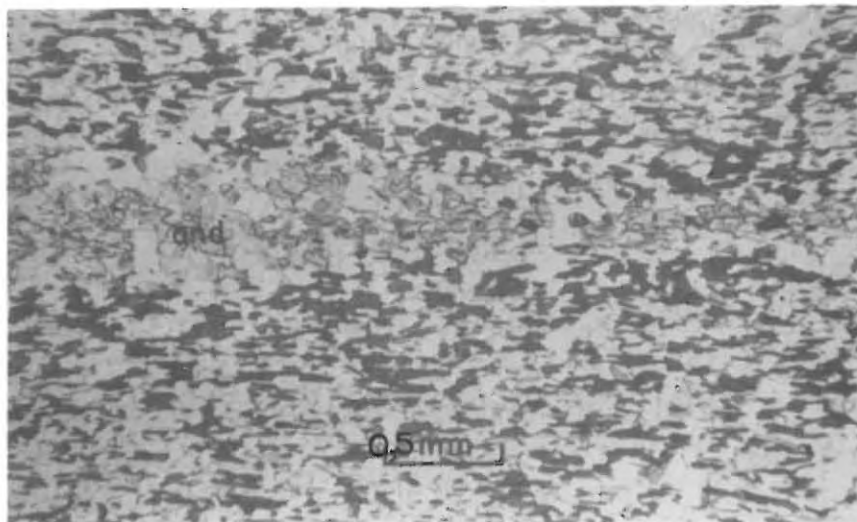
Plate 17
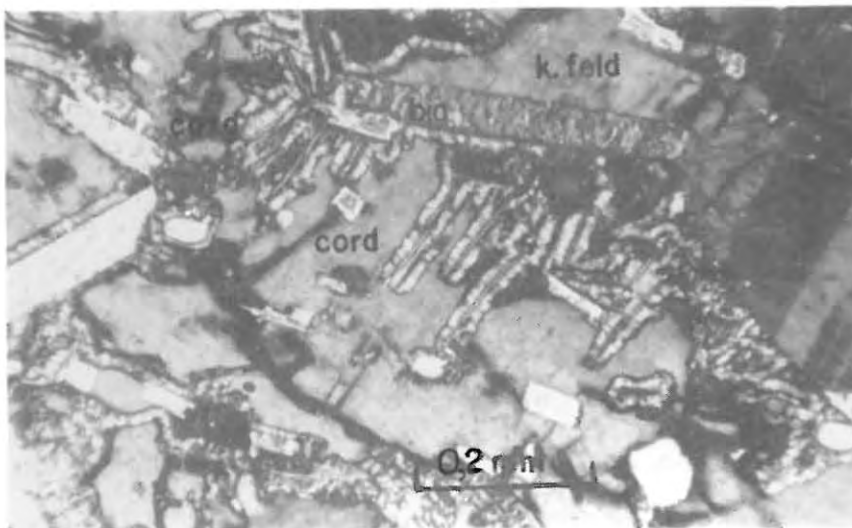
Plate 18

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).
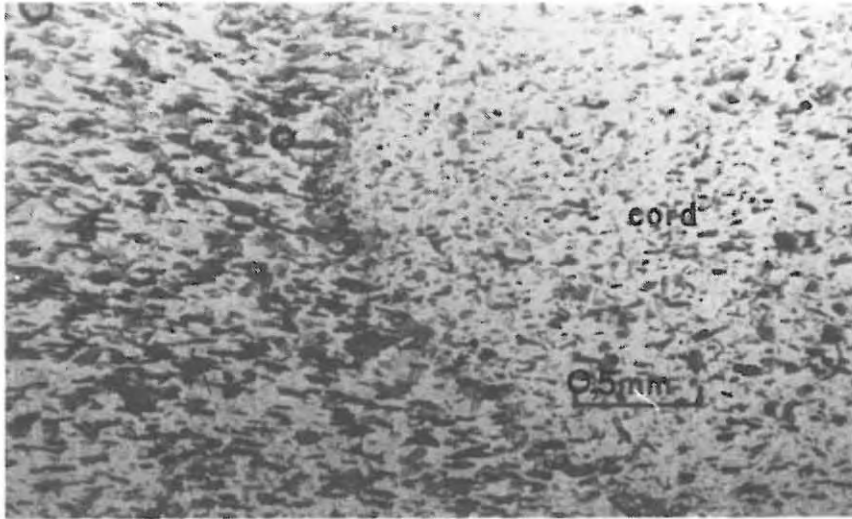
Plate 19
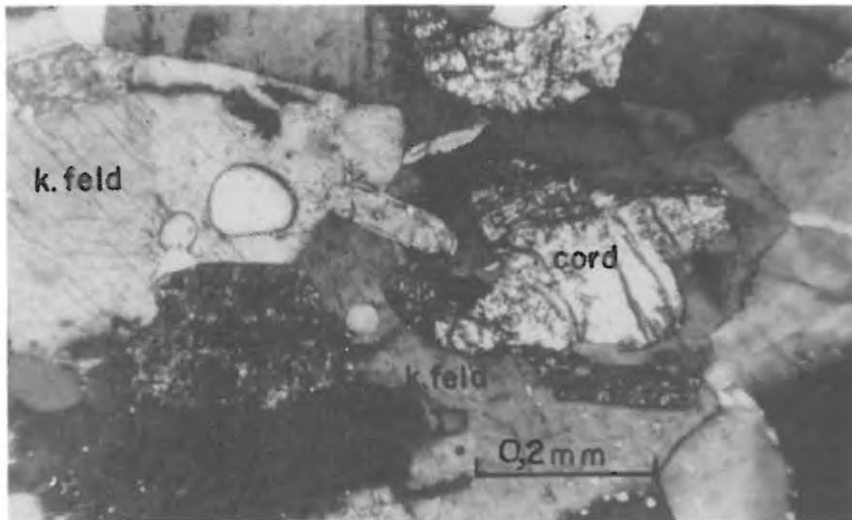
Plate 21
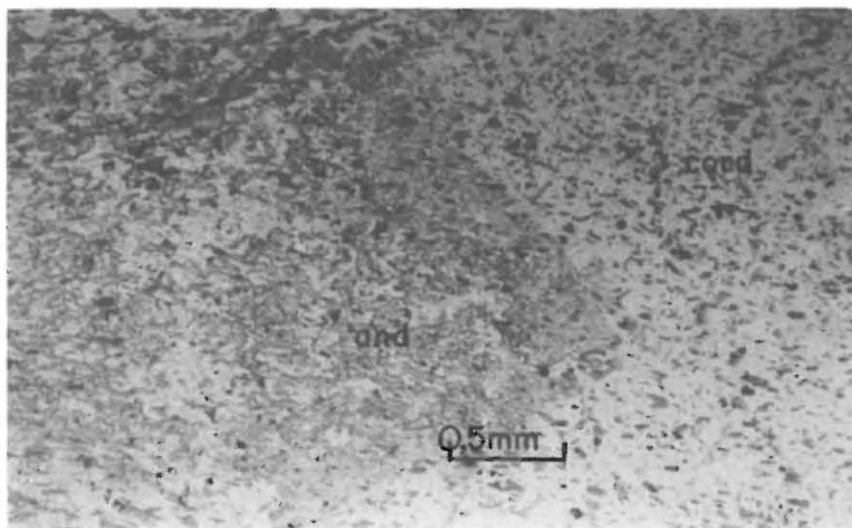
Plate 20

Plate 22

Banded gneiss of the Khan Formation, containing inter-
growths 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 .
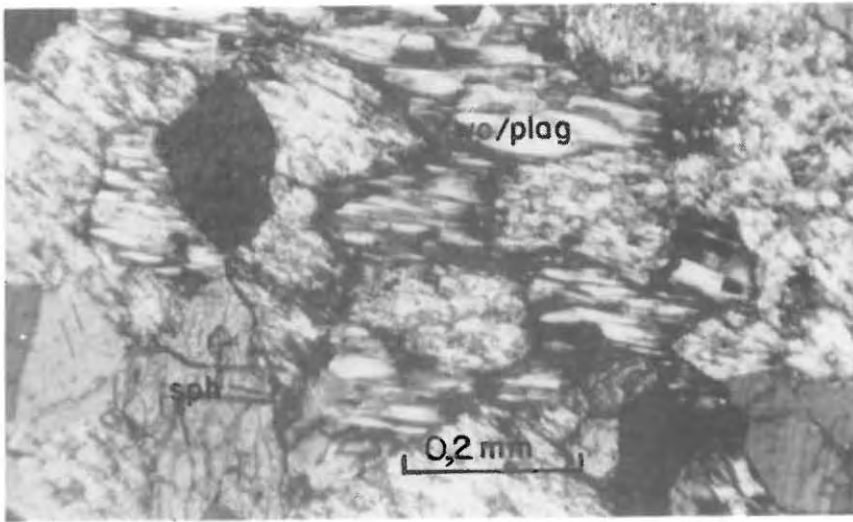
Plate 22

Plate 23

Plate 24

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.
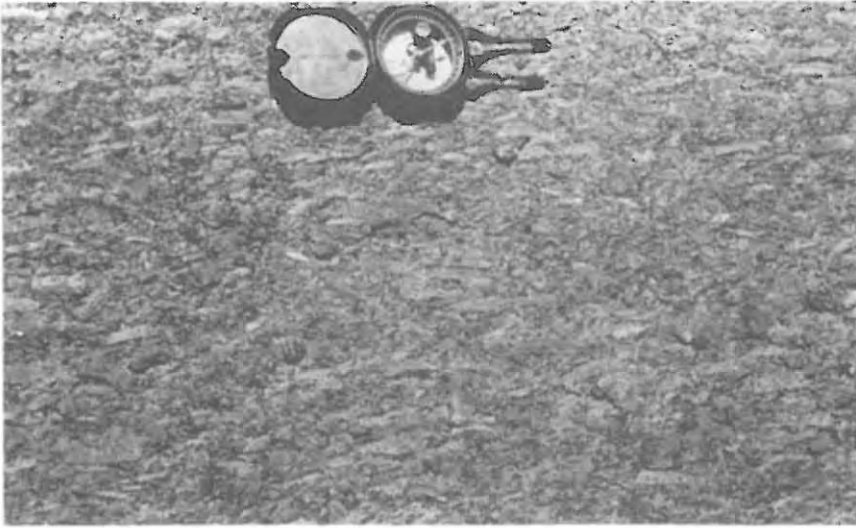
Plate 25

Plate 26

Plate 27

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.
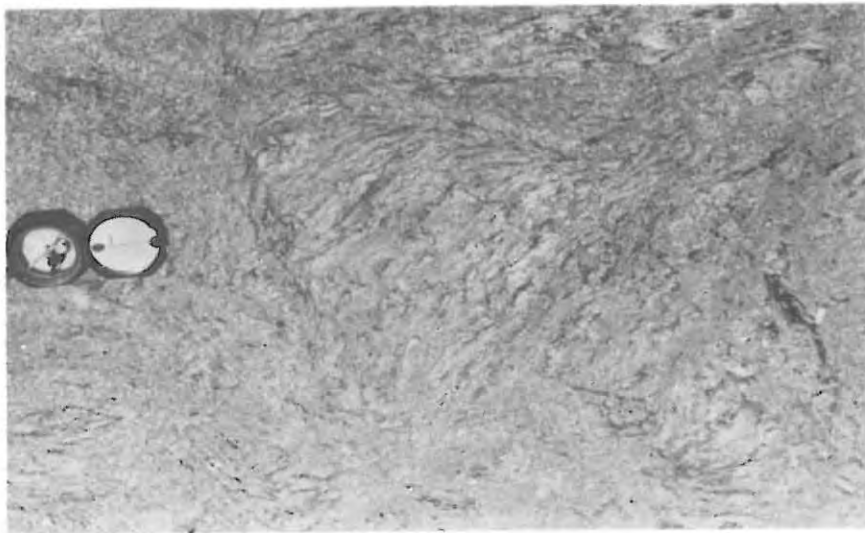
Plate 28

Plate 29
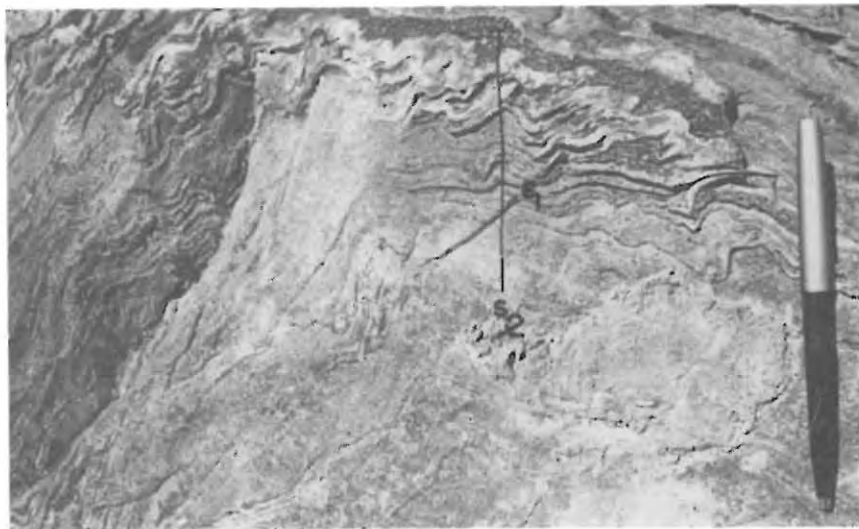
Plate 30

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 So1 banding in Khan gneiss occurring
parallel to the S2 axial planar direction of the B2
folds. This transposition is commonly seen in zones
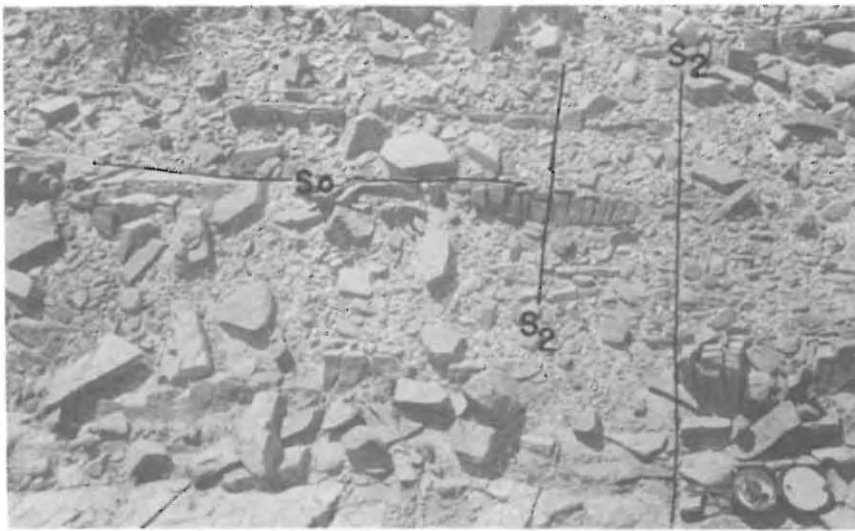of more intense deformation.
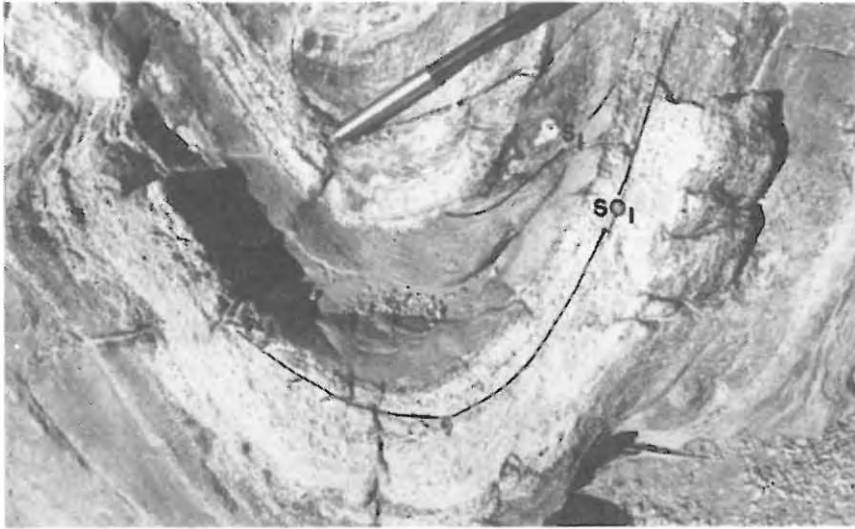
Plate 31

Plate 32

Plate 33

Plate 34

Folding of So1 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 Lucasberge. Mullion structures (12) are
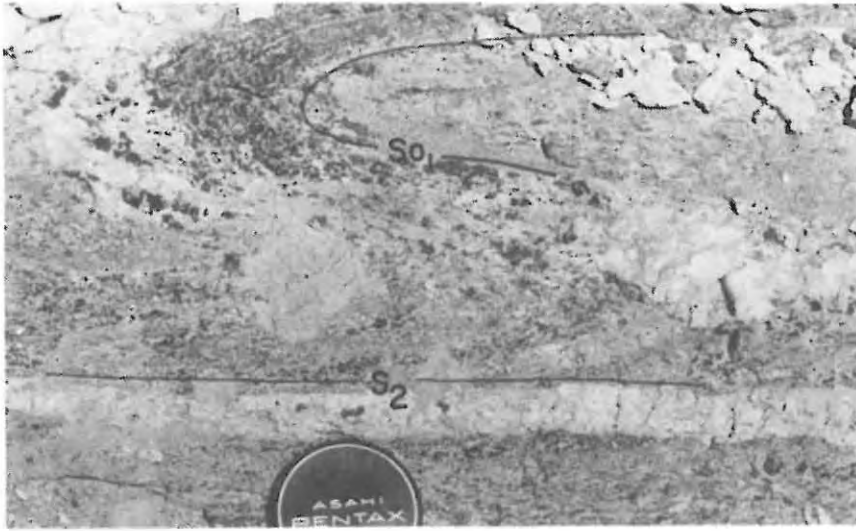also developed.

Plate 34

Plate 35

Plate 36

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 Glücks Hills. Gentle buckling appears to have
occurred at right angles to the B2 axis and was probably
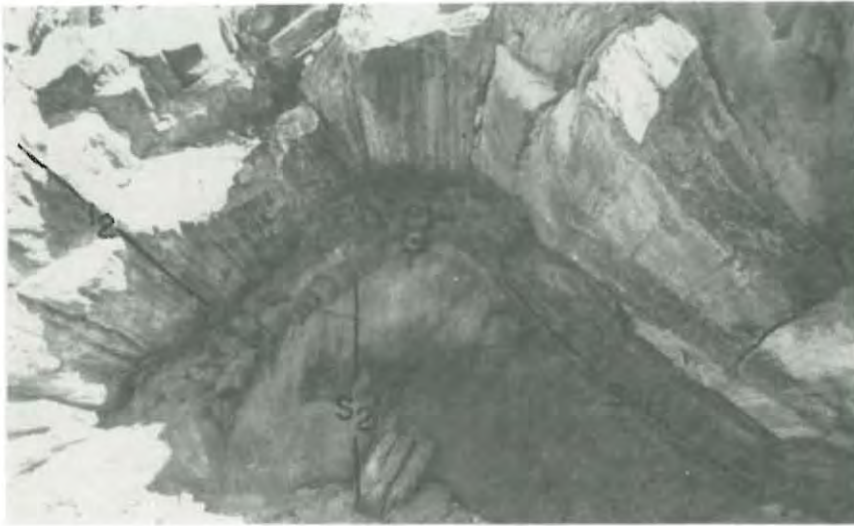developed during the F3 phase.

Plate 38

Plate 37