THE STRATIGRAPHY, SEDIMENTOLOGY AND STRUCTURE OF THE CANGO GROUP NORTH OF OUDTSHOORN, C.P.

by

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A B S T R A C T

This investigation aims at determining the paleogeographic and tectonic setting of a Precambrian suite of low grade metamorphic rocks north of Oudtshoorn, Cape Province, by means of a stratigraphical, sedimentological and structural analysis.

The Cango Formation is raised to Group status, while formal names are introduced in a new, more detailed stratigraphic subdivision.

The lowest unit, the Matjies River Formation, was deposited in an environment varying from that of an unstable shelf (Nooitgedagt Member) to an isolated platform (Cave Member). The main source area lay to the west. After submergence the turbidites of the Groenefontein Formation were deposited in deeper water. Transport was in a westerly direction along the trough axis, with the main supply from the flanks to the north and south. During subsequent uplift the basin was locally eroded and the Vaartwell conglomerates (partly cannibalized from the Groenefontein Upper Member) were deposited in alluvial fans (polymictic member) and on beaches (monomictic member). Transport was roughly in an easterly direction. Westward transgression of the sea resulted in upward grading of the Vaartwell to infralittoral (Uitvlugt Formation) and circalittoral (Gezwinds Kraal Formation) deposits. The latter grade laterally into the fluvio-deltaic sediments of the Schoongezigt Formation, derived from the east. An overall flysch- molasse - flysch cyclicity is in evidence.

The regional setting is visualized as an embayment situated at the southern edge of a continent and connected in the southeast to the open sea. The source areas were probably of a non-volcanic nature.

The Cango rocks were subsequently deformed by horizontal, northward-directed compression as shown by the style of folding and thrust faulting. The regional strike is arcuate, convex to the north and conforms to the outcrop pattern of the overlying Cape Fold Belt in this area. This is either a

post-Cape phenomenon affecting both the Cango and Table Mountain Groups, or it indicates that the pre-Cape trend controlled the deformation of the latter.

Metamorphism of the Cango rocks did not exceed the lower green=schists facies.

The Schoemans Poort Formation was deposited unconformably in isolated depressions (mainly as fluvial sediments) along the margin of elevated highlands after uplift and erosion of the Cango Group. It represents a molasse deposit.

A suite of basic rocks subsequently intruded both the Cango Group and Schoemans Poort Formation.

The distinctive paleogeography of the Cango Group and the Schoemans Poort Formation precludes any correlation with the possibly time-equivalent Malmesbury Group or Klipheuwel Formations respectively.

Seen in perspective the depositional and deformational history of the Cango sediments in all important aspects resembles the marginal basin association of the miogeosynclinal zone. As such they may contribute to a better understanding of the fundamental processes responsible for diastrophic cycles.

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1. INTRODUCTION

The main objective of the International Geodynamics Project is to gather information concerning movement of the litho= sphere and its influence upon geologic features. The ultimate aim: to establish the origin of the deep seated processes, can only be reached through compilation of the information obtained from all areas and fields of study in the Project. The purpose of the present study was to determine whether the geologic history of the Cango followed a pattern similar to the sequences of events which in younger areas are connected to plate tectonics. If the answer to this proved to be positive, it would be necessary to find out where the Cango Group fitted into the model and what interrelationship existed between the early sedimentary tectonics, processes of sedimentation and later deformation. With these objects in mind, a study of the stratigraphy, sedimentology and structure of the Cango Group commenced in March 1975.

TABLE 1

DEVELOPMENT OF THE STRATIGRAPHIC COLUMN

Dunn (1877)	Corstorphine and Rogers (1896)	Rogers and Schwarz (1900)	McIntyre (1932)	Stocken (1954)	Mulder (1954)		Proposed New Subdivision.	
			Quartzite and sandstone with conglomerate bands	The Unconformable Feldspathic Grit Formation			Schoemans Poort Formation	
			Cango conglomerate	The Upper Greywacke Formation	Upper Fine-grained Series		Gezwinds Kraal Formation	Schoongezigt Formation
		Cango Conglomerate	Quartzite and sandstone, shaly at base with limestone		Cango conglomerate Series	Feldspathic Stage	Uitvlugt Formation	Danzers Kloof Member
Namaqualand Schists	Slates, dolomites, grits and conglomerate		Main limestone formation	The Crossbedded Grit Formation		Conglomeratic Stage		Rooiberg Member
		Slates and limestone	Shales and limestone				Vaartwell Formation	Monomictic Member
			Quartz grits and shales			Fine-grained Stage		Polymictic Member
			Arkoses, grits and conglomerate	The Lower Greywacke Formation	Lower Fine-grained Series		Groenefontein Formation	Upper Member
								Middle Member
								Lower Member
			Dolomite, granite, gneiss and foliated sediments.	The Limestone-Shale Formation	The Cango Valley Series		Matjies River Formation	Cave Member
								Nooitgedagt Member

2. STRATIGRAPHY

2.1. PREVIOUS WORK

The pre-Cape rocks south of the Swartberg appeared for the first time in 1887 on the "Geological Sketch Map of South Africa" of E.J. Dunn, who regarded these "Namaqualand Schists" to be older than the Malmesbury Beds.

In 1897 Corstorphine and Rogers described the sediments in the "First Annual Report of the Geological Commission."

A year later Rogers and Schwarz gave a description of the Cango Beds in the vicinity of Ladismith, which they followed eastwards into the Oudtshoorn District in 1899. Although they believed the "Cango Conglomerate" to be younger than the slates and dolomitic limestones, no other indications of age relationships were given. A tentative correlation of the slates and limestones with the Malmesbury Series was made, but the conglomerate was separated from this on the belief that its granite pebbles had been derived from the younger Cape intrusives.

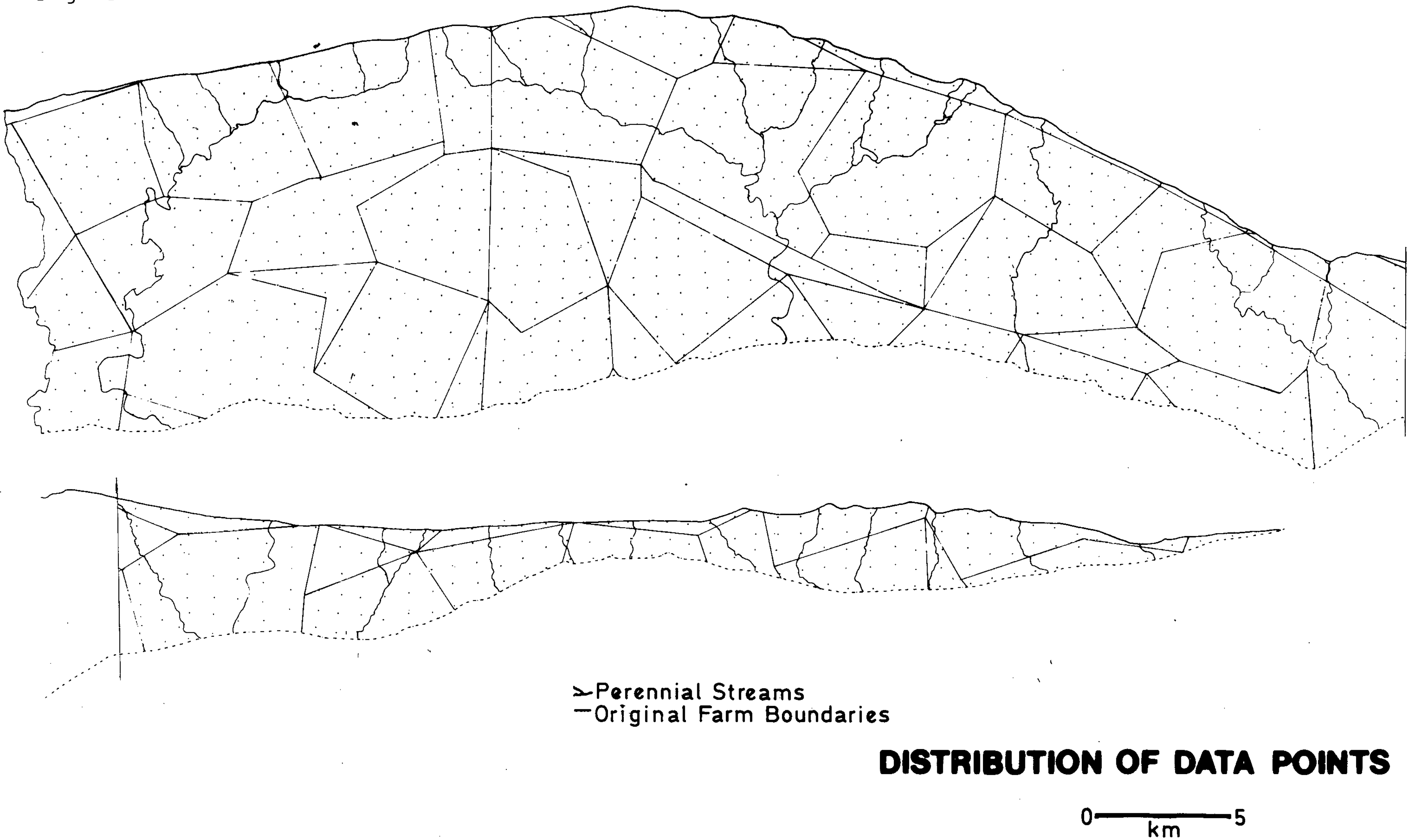
The economical possibilities of the limestone occurrences were studied by Wybergh in 1920.

In 1932 McIntyre distinguished rock formations of two different ages, viz. a "pre-Cango" and younger "Cango Series." His structural, lithological and stratigraphical description was accompanied by a sketch map through Schoemans Poort.

The first detailed regional mapping of the Central Cango was undertaken in 1951 and 1952 by C.G. Stocken, who proposed a new and entirely different stratigraphic sequence for the area between 22°00′ and 22°20′E. He confirmed the existence of several faults (which were largely responsible for previous misinterpretation of the strati= graphy), and tentatively correlated the Cango Beds with the pre-Cape rocks of the Gamtoos and George areas.

M.P. Mulder mapped the Western Cango and part of the eastern

Fig. 1.



DISTRIBUTION OF DATA POINTS

remainder during 1952 and 1953. Except for the use of different formation names, his work basically confirmed the results of Stocken.

2.2. PRESENT INVESTIGATION

The Cango Belt east of Potgieters Poort (henceforth referred to as "the study area") was remapped in detail using 1:40000 aerial photographs, while the remaining westerly part was reconnoitred during January 1976. In the study area traverses were run on foot along north-south lines approximate=ly 1 km apart, while dip and strike measurements of foliation planes were taken at similar intervals. (Fig.1.) Contact mapping was used to fill in important details. Out=side the study area investigations were confined to roads and readily accessible outcrops.

Although the stratigraphic sequence of Stocken and Mulder was found to be essentially correct, its interpretation in the western area, possibly because of an unrecognized thrust sheet, needed revision. The present subdivision (Table 1) is based on rock-stratigraphic units, for which type sections are designated in accordance with the South African Code of Stratigraphic Terminology and Nomenclature (1971). Formal names derived from these type sections, are used throughout.

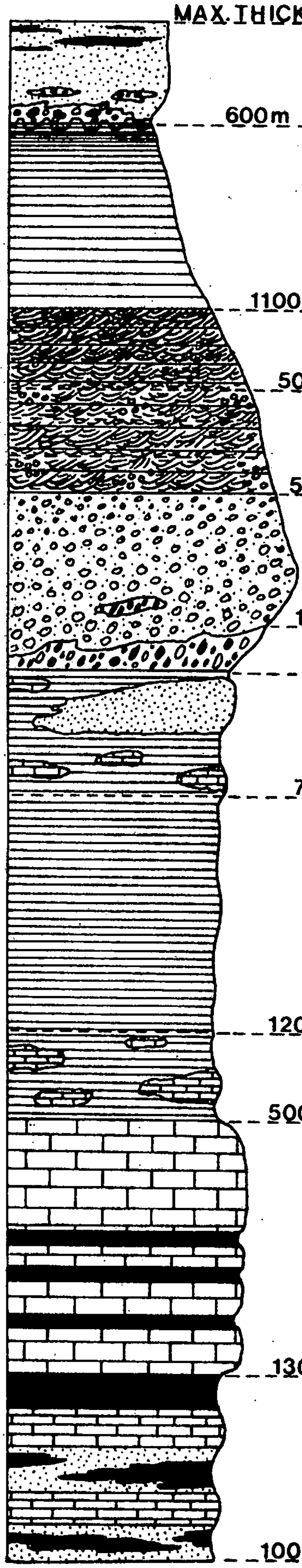
The "Cango Conglomerate Series" of Mulder, which corresponds to the "Crossbedded Grit formation" of Stocken, has been subdivided into two formations. Further subdivision into members is suggested for all but the Gezwinds Kraal and Schoongezigt Formations.

The "Unconformable Feldspathic Grit formation" of Stocken becomes the "Schoemans Poort Formation," which is excluded from the Cango Group.

2.3. GENERAL DESCRIPTION

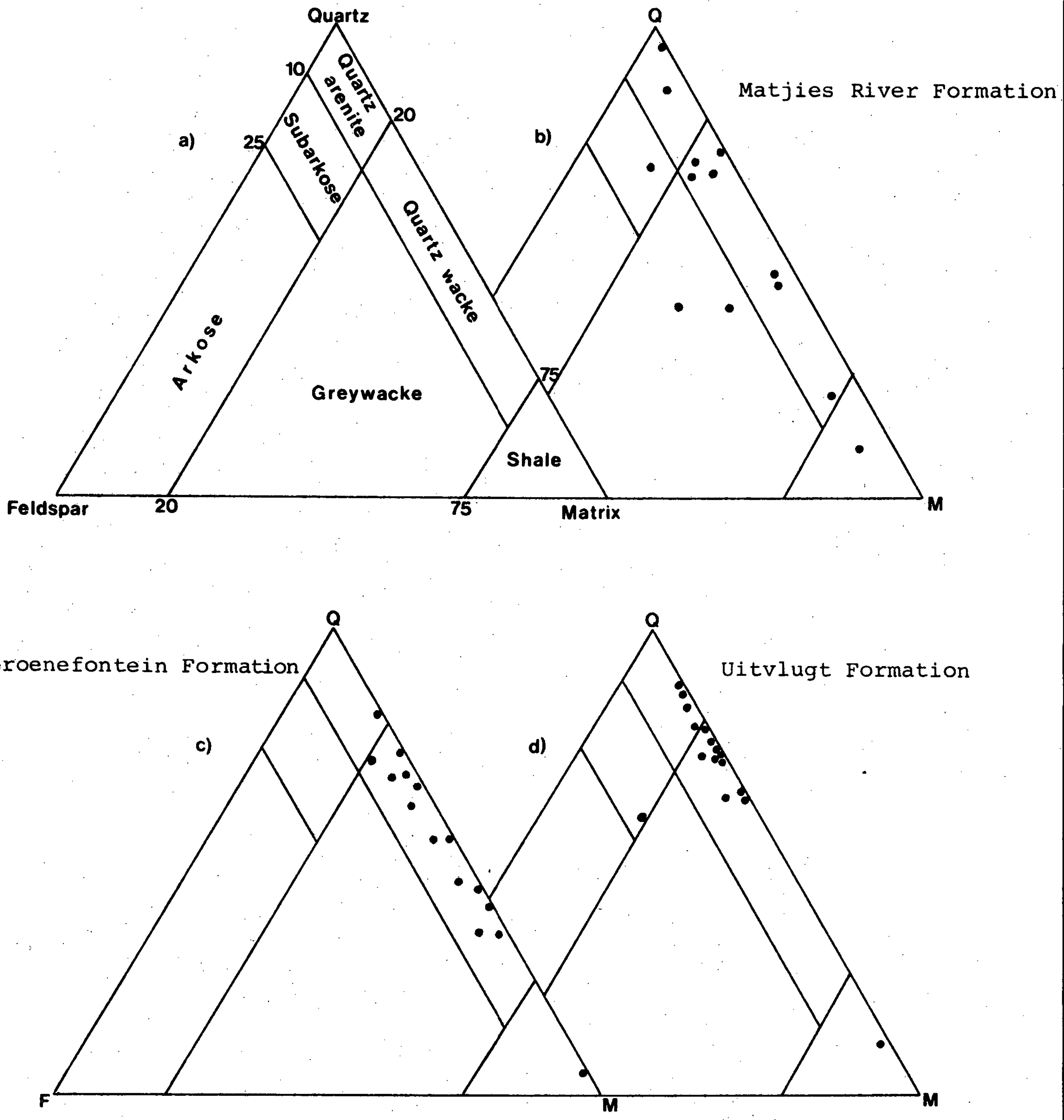
The following takes the form of a short field description to provide the reader with a background of the geology of the Cango while introducing the new stratigraphic subdivi=

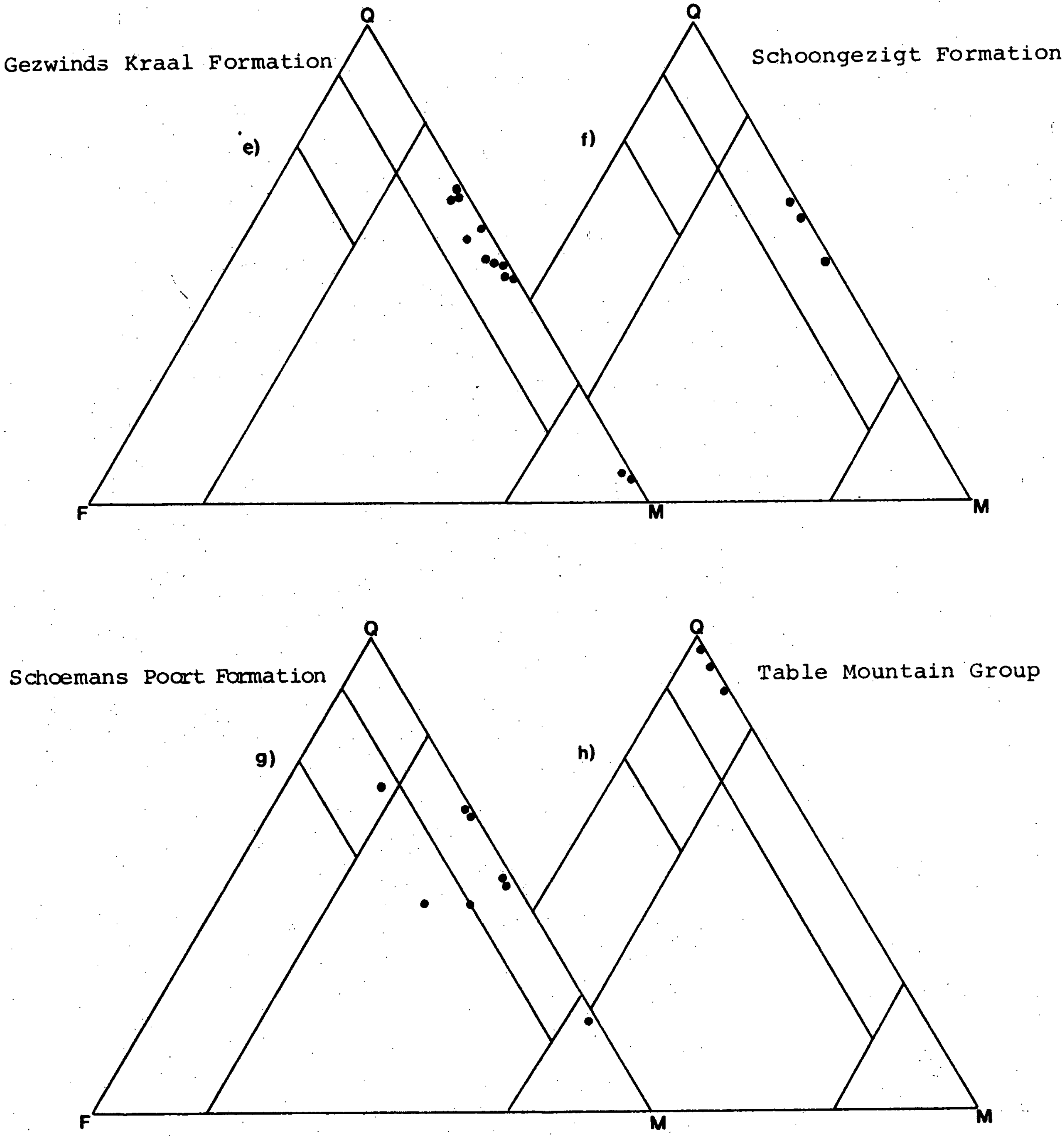
Fig. 2. The stratigraphic column.



MAX. THICKNESS	MEMBER	FORMATION
600m		Schoemans Poort
1100⁺m		Gezwinds Kraal
500m	Rooiberg	Uitvlugt
500m	Danzers Kloof	Uitvlugt
1000m	monomictic	Vaartwell
	polymictic	Vaartwell
700m	Upper	Groenefontein
1200m	Middle	Groenefontein
500m	Lower	Groenefontein
1300m	Cave	Matjies River
1000⁺m	Nooitgedacht	Matjies River

Fig. 3. Classification and mineral composition (sandstones).




Gezwinds Kraal Formation
e)
Q
F
M
Schoongezigt Formation
f)
Q
M
Schoemans Poort Formation
g)
Q
F
M
Table Mountain Group
h)
Q
M

sion. Relevant petrographic and sedimentological aspects are discussed in more detail in the next chapter, while information on topographic expression or the distribution of formations (which is available from the accompanying maps and diagrams) is omitted from the description.

The sandstone classification scheme (Fig. 3.(a)) is based on an unpublished chart of Pettijohn (1944)(see Krumbein and Sloss, 1963, p. 157), but terms such as "subgreywacke" and "quartzose sandstone" are abandoned in favour of "quartz wacke" and "quartz arenite" in accordance with current usage (Pettijohn, 1972, pp. 155-159.).

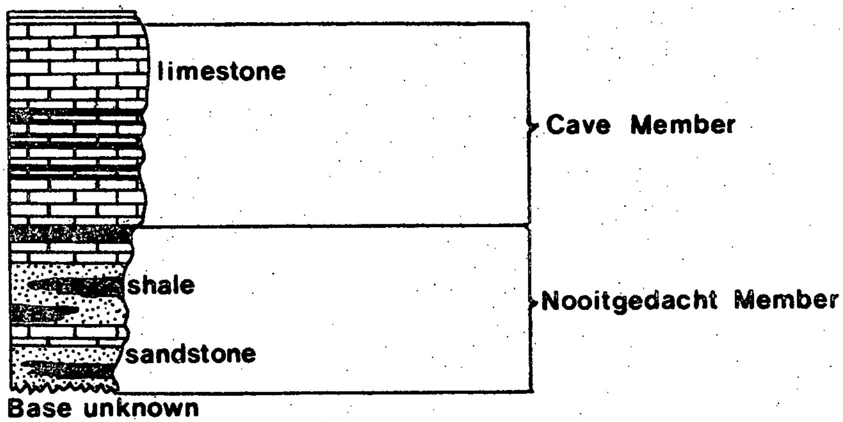
2.3.1. THE CANGO GROUP

A proposal to raise the status of the Cango Series or Cango Formation to that of a Group, has been submitted to S.A.C.S., because the sequence can be divided into several formations (Fig. 2.)

2.3.1.1. MATJIES RIVER FORMATION

TYPE LOCALITY Nooitgedagt ($33^{o}23'$S.; $22^{o}10'$E)

SECTION (Fig.4.) Matjies River area



The Matjies River Formation has an unconformable upper contact with the Table Mountain Group in the north and elsewhere grades conformably into the Groenefontein Formation. It consists of limestone, shale and sandstone, of which the limestones are the most conspicuous and give the formation its distinctive character. Two members are recognized.

NOOITGEDAGT MEMBER

Five units of the member are exposed in the core of the Cango Valley Anticline. The lowermost consists of sandstones with subordinate lenses of shale which are followed upwards by a thin limestone-shale band, another sandstone-shale unit and finally a second limestone-shale horizon topped by a rather consistent shale band.

Sandstones dominate in the area east of 22°07'E, while shales become more abundant westwards. In general the two rock types either grade into and form lenses within each other or occur as alternating zones. Sharp contacts are occasionally observed. The sandstones display great variability in composition and texture and, as shown by Fig. 3 (b), range from quartz arenite to subarkose, greywacke and quartz wacke. Some lenses of calcareous wacke are occasionally found in close proximity to limestone horizons.

Although bedding is mostly massive and obscure, lenses of sandstone may show alternating layers of pale and dark grey material varying from a few cm to more than 1 m in thickness. Internal lamination consists of dark and light as well as coarse and finer intercalations on a mm scale, which may have either sharp or grading contacts.

The wackes are generally coarse grained, poorly sorted and may contain up to 25% orthoclase and microcline. Plagio= clase is present in minor amounts (less than 5%.) These feldspathic types consist of subrounded to angular quartz grains with large (up to 2 cm) fragments of K - feldspar in a quartz-sericite matrix. Colours range from dark grey or black to green and cream, with red, brown, orange and yellow produced by weathering of the feldspar. Dark colours are usually the result of blue or black quartz grains.

The better sorted quartz arenites consist of subrounded quartz grains which may constitute more than 90% of the rock. These display a mosaic intergrowth of grains with sutured contacts. Dark colours are the rule.

Occasional conglomeratic lenses have inclusions of granite, gneiss and subrounded quartz up to 4 cm in diameter. Intraformational conglomerates indicating penecontemporaneous erosion were observed on the farms Kombuys (22°13′E), Welgevonden (21°57′E) and elsewhere. These are described in the next chapter.

The shales of the Nooitgedagt Member are predominantly olive-green in colour, with bluish, yellowish and dark green, red to brown and even purple varieties also present in places. Dark grey tints are typical of calcareous types where these grade into limestone, while an abundance of carbonaceous matter is indicated by a dark to black appearance. Oxidation produces reddish brown to black colours.

The shales are generally well bedded, with light and darker intercalations some mm to a few cm in width. These are separated by sharp contacts. Cleavage is often strongly developed and where at an angle to the bedding produces "pencil" weathering.

In composition the shales range from calcareous types, usually associated with limestone, to silty, slaty and quartzose varieties. They may grade completely into sandstone and limestone.

The limestone units of the Matjies River Formation have a characteristic cyclicity which is diagrammatically illustrated in Fig. 5. The megacycles form fairly continuous bands in the east, but give way to lenses west of 22°07′E. They consist of alternating bands, 15-20 cm or more in width, of limestone and shale. The latter may be partly produced by the leaching of limestone layers originally containing large amounts of silt. Dolomitic limestone, calcitic dolomite and black chert are present as rarer intercalations.

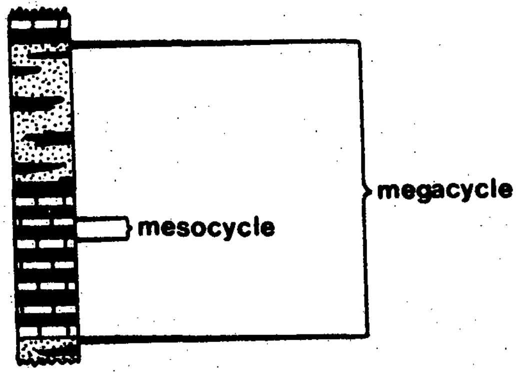


Fig. 5. Cyclicity of Cango limestones.

The limestone is bluish black on fresh surfaces (the colour fading as the Mg-content increases), and weathers with a dark grey or brown ferruginous, wrinkled crust. It is frequently replaced by limonite along fault zones. Red clay, which is the ultimate product of weathering, favours the growth of aloes. These help in finding smaller lenses in the field.

Intraformational conglomerate with elongated, rounded fragments of limestone differing slightly in colour from the host, indicates some transport during penecontemporaneous erosion. Rare inclusions of grit and granite were also encountered.

On Kombuys a lens of fine-grained (2-10µ) dolomitic limestone, 10-25 m wide, is pale blue or bluish grey in colour and weathers to a dark brown crust. When broken a foetid odour is emitted.

CAVE MEMBER

Like the other limestone units the Cave Member is composed of limestone, dolomitic limestone and shale or siltstone. West of $22^{\circ}07'$ the shales frequently constitute more than 50% of the horizon with limestone forming lenses therein, but the latter becomes the dominant constituent eastwards and develops excellent layering. The shales vary in colour from pale brown, yellow and bluish to olive green.

On Kombuys the Cave Member can be divided into 7 units, farther west at De Hoek (22°10'E) into 5 and beyond 22°07' no subdivisions are possible. The lower limestone zones on Kombuys and De Hoek contain interbedded shales and phyllites with occasional sandy material, which is absent from the upper parts. Type sections on the two farms are given below. (Fig. 6.)

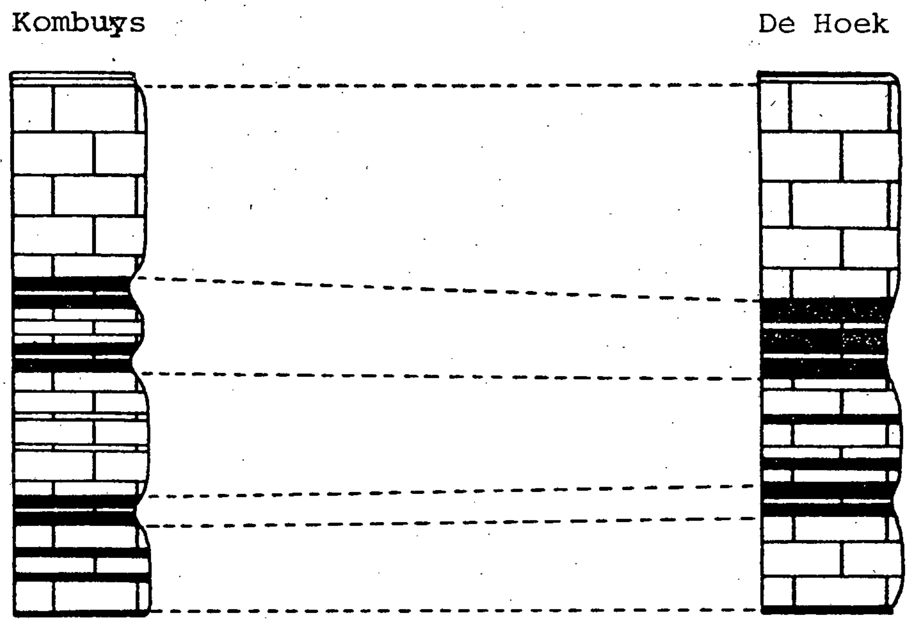


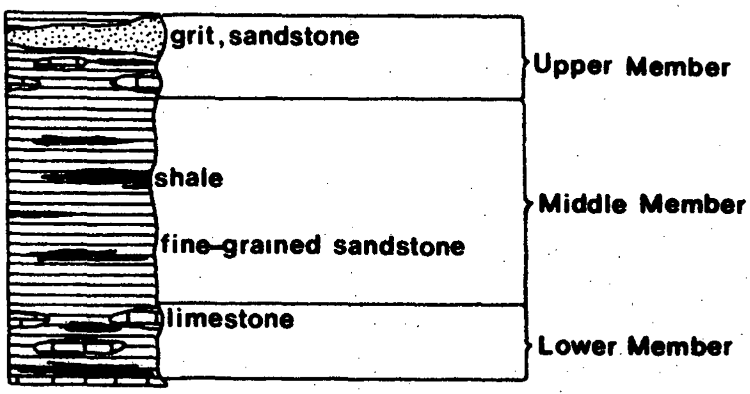
Fig. 6. Cave Member: Type sections at Kombuys and De Hoek.

The limestone is similar to that of the Nooitgedagt Member. In thin section it is seen to consist of a matrix of rounded, even-textured grains in the calcilutite or calcisiltite range, throughout which are scattered oölites and quartz grains. Some recrystallized calcite grains are up to 600μ in diameter and may be concentrated in zones along the cleavage. The latter is sometimes visible to the naked eye as dark zones a few mm wide. Small calcite, chert and quartz veins criss-cross the rock at all angles, but mostly subparallel to the foliation.

2.3.1.2. GROENEFONTEIN FORMATION

TYPE LOCALITY Groenefontein(33°24'S.; 22°15'E)

SECTION (Fig. 7.)



A rather monotonous succession of fine-grained, silty quartz wackes and shale with occasional limestone and coarser sand=stone lenses constitutes the Groenefontein Formation. Although previously believed to be irregularly disposed, more detailed mapping revealed these lenses to occur at fixed stratigraphic horizons in the central and eastern parts of the Cango. This allows the further subdivision of the forma=tion into three members: The Lower Member has limestone but no sandstone lenses, the Middle Member contains neither, and the Upper Member is characterized by both. Outside the study area this distribution could not be verified because of the difficulty in determining the stratigraphic position, and neither limestone nor sandstone lenses were encountered west of 21°43'E.

LOWER MEMBER

The contact with the Matjies River Formation is of the mixed gradational type, i.e. while individual contacts between the silty wackes of the Lower Member and the Cave Limestone appear sharp in the field, the limestone occurring as lenses at the base of the formation is similar to that of the Cave Member. Except for the large lens on Rust En Vriede

(22°20′E), they occur mostly within 800 m from the lower contact. In length they range from only a few meters to over 2 km, as for example on Nooitgedagt. The lenses may consist of either massive limestone or else are composed of intercalated shale and limestone bands as in the Matjies River Formation. They grade into the surrounding quartz wackes and shale.

West of 22°07′E limestone lenses are less frequently encountered.

MIDDLE MEMBER

Characterized by its uniform lithology and complete absence of marker beds, the Middle Member is composed of fine-grained, greenish to bluish grey quartz wackes that weather to silky, buff-coloured outcrops. Shales and phyllites complete the succession.

Although previously classified as "greywacke" or subgreywacke," the sandstones have less than 5% feldspar (mainly plagioclase) and few recognizable rock fragments, so that the term quartz wacke is more appropriate (Fig. 3 (c)). They consist of 45-70% quartz grains, which are angular to subangular, moderately sorted and often polycrystalline as a result of deformation. The matrix, composed of silt, sericite and chlorite, forms the remaining 30-55% of the rock and contains fragments of calcite, cubes of pyrite (limonite) magnetite octahedrons and larger, rounded inclusions of hornblende or pyroxene. Mica flakes (concentrated along the cleavage) are conspicuous in hand specimens.

Graded bedding, sole structures (Fig. 8 (a)) and a single ripple mark (Plate 3 (b)) were observed in the Huis River Pass (21°33′E), while false sole markings are sometimes produced by the weathering of crosscutting foliation planes (Fig. 8 (b)). Convolute lamination was found on the farm Buffelsvalley (22°39′E).

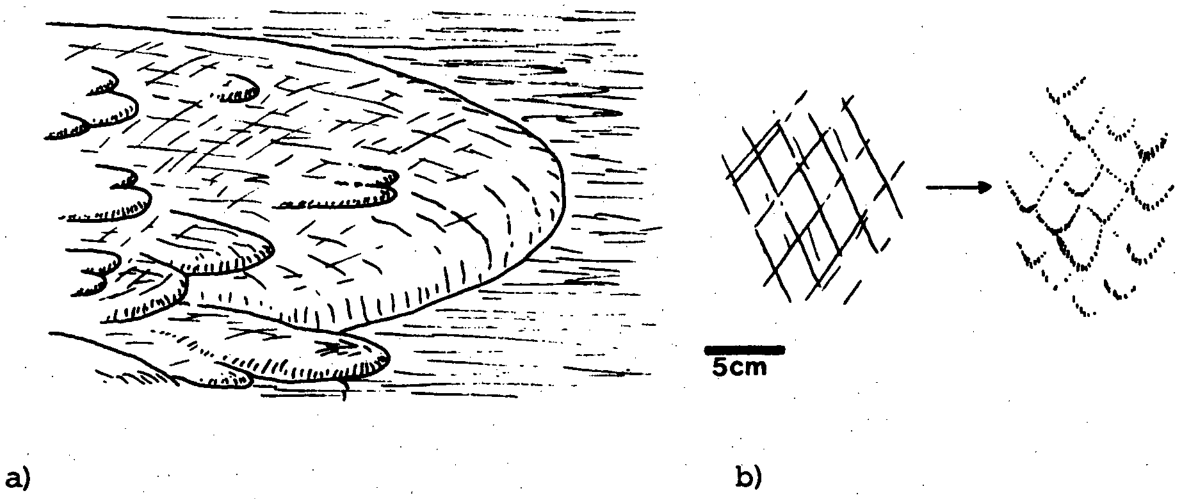


Fig. 8. (a) Flute casts, Huis River Pass.
(b) False sole structures produced by crosscutting foliation.

The shales occur as small bands, a few to 100 cm wide, which frequently show green to yellow laminations. They are commonly arenaceous and grade into the associated quartz wackes. West of Coetzees Poort shales become more abundant and form large lenses consisting of green and dark grey bands a few cm wide. Broad zones of shale and subordinate quartz wacke are separated by more resistant zones of almost exclusively quartz wacke. In the extreme west beyond Coetzees Poort, shale, with colours ranging from green to purple, is the dominant constituent.

Traced eastwards, the rocks of the Middle Member show a slight increase in grain size.

UPPER MEMBER

Distinguished from the Lower Member by bands and lenses of coarser grained sandstone, the Upper Member is also marked by the renewed appearance of limestone lenses. These differ from the limestone of the Lower Member in many cases by their paler, greyish blue to white colour and coarser crystalline texture. They are also less frequently interbedded with shale. Followed westwards beyond the study area the lenses become more abundant and larger, the limestones being bluish grey, frequently arenaceous and grading into quartz wacke with which they seem to occur more often than shale. Bedding in these lenses is up to 1 m wide and may exhibit cm scale laminations. In the westernmost part of the Cango the lenses are absent, while those in the eastern outcrop of the Upper Member are much smaller, but more consistent and some= times metamorphosed to marble. One occurrence on the farm De Oude Muragie (22°29E) is 4 m wide, 500 m long and inter= bedded with shale.

Sandstone lenses cap many of the higher hilltops in the north; in the south prominent ridges are formed by more continuous bands. Generally white, grey and reddish in colour, they are mostly massive and medium to coarse grained. Their composition varies from quartzose to rare arkosic types which are poorly sorted, but most contain more than 20% matrix and should be classified as quartz wackes.

On the farm Buffels Kraal (22°21′E), massive, bluish grey quartzitic sandstone has well developed beds up to 1,5 m in thickness (Plate 2 (c)). These are very similar to the wackes of the Uitvlugt Formation and have at places been mistaken as such by previous workers, although crossbedding is absent.

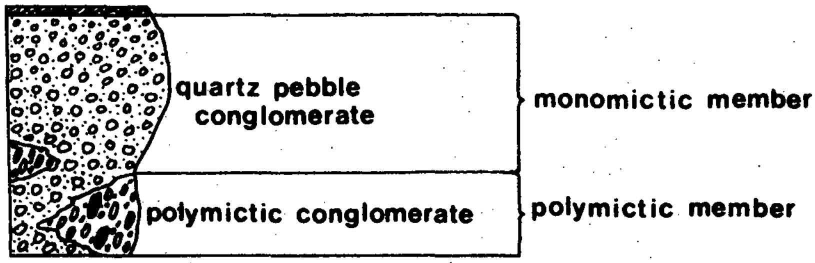
Thin sections from two samples taken some 25 km apart (Buffels Kraal; Welbedagt 22°06′E) show a remarkably similar and distinctive texture. Large, angular quartz grains with smaller, better rounded quartzitic fragments scattered inbetween, are set in a matrix of chlorite and sericite, which makes up some 15 to 30% of the rock.

Conglomeratic sandstone containing quartzite pebbles occurs as a large lens on the farm De Oude Muragie, immediately underlying the Vaartwell Formation.

2.3.1.3. VAARTWELL FORMATION

TYPE LOCALITY Vaartwell ($33^{\circ}27'S$,; $21^{\circ}33'E$)

SECTION (Fig. 9.)



Stocken and Mulder included the Vaartwell with the overlying crossbedded sandstones as one formation. However, the two units differ from each other as much as the Uitvlugt and Gezwinds Kraal Formations, and although these workers distinguish between the two rock types, their maps show large areas as "crossbedded grits," which in fact are under= lain by conglomerate. Outliers of conglomerate overlying crossbedded sandstone were possibly regarded as lenses within the latter, but as Stocken and Mulder do not mention the size and extent of their "lenses", it is difficult to say whether they refer to those (a few meters in length) that occur at the base of the Uitvlugt Formation, or to the larger conglo= merate sheets which occupy several square km in the south. Whatever the case may be, these outliers have a tectonic origin and are not part of the original stratigraphy.

West of Potgieters Poort an inverted sequence is shown in the north were conglomerate overlies or intertongues with cross= bedded sandstones. At Coetzees Poort, where the hanging wall of the overthrust is composed of highly sheared polymictic conglomerate, this is shown as crossbedded grit, whereas the footwall (which is crossbedded sandstone with a few scattered quartz pebbles) is mapped as conglomerate. This contact is described as "an example of a pseudo= unconformity." It is thus comprehensible why, with large volumes of conglomerate apparently appearing high up in the crossbedded grit sequence, these workers chose to include the two types of sediment in one formation.

In the present investigation the distinction between Vaartwell Conglomerate and the Uitvlugt Formation was made as follows: if pebbles or conglomeratic lenses become sufficiently abundant to warrant the use of the term conglomerate, the unit is regarded as part of the Vaartwell Formation; if the pebbles or lenses are so widely scattered that the general description "sandstone with conglomeratic lenses" is more appropriate, it is regarded as Uitvlugt Formation. It must be stressed however, that the contact between the two formations is a continuously grading one and the boundary is therefore inevitably somewhat arbitrary. A decrease in pebble size, more conspicuous in the upper parts, completes the transition.

The contact with the underlying Groenefontein Formation can be sharp or perfectly gradational. In the latter case small pebbles make their appearance in the quartz wacke, becoming larger or forming small lenses increasing in size. In the Gamka River the actual contact is sharp, but the quartz pebbles are smaller and more widely scattered at the base. They exceed polymictic types in number in a 5 m transitional zone at the Wynands River (Wildehonde Kloof, 22°00′E). Shearing at contacts between the polymictic member and Groenefontein Formation is a common feature, as for example on Opzoek (22°31′E) and in Schoemans Poort (21°15′E). Although the contact is apparently conformable in the study area, the presence of pebbles derived from the underlying formation indicates local unconformable relationships.

The polymictic and monomictic members of the Vaartwell Formation grade into each other both vertically and laterally, thus making it impractical to map them as separate units on a regional scale. The term "member" is therefore used informally and the two types of conglomerate are indicated on the geological map by the letters "p" and "m".

THE POLYMICTIC MEMBER

Polymictic conglomerate is usually found at the base of the formation, although it may comprise its entire thickness.

The member is completely unstratified and owes its existence mainly to erosion of the underlying rocks. Inclusions of fine-grained quartz wacke, shale, limestone and coarser sandstone were obviously derived from the Upper Member of the Groenefontein Formation, while quartz, leucocratic granite, gneiss and chert were transported from the provenance area. The granite, fine to medium grained and pinkish in colour, is composed of K-feldspar, quartz, muscovite, epidote and a little biotite. The pebbles are poorly sorted and set in a green, argillaceous, strongly sheared matrix, which characteristically weathers with brown and reddish streaks.

While the less competent pebbles are sheared and flattened into an a-lineation, the more resistant quartz and granite inclusions show only occasional weak orientation which may be partly due to primary imbrication. The matrix folia appear to "flow" around these pebbles.

The average longest diameter of about a thousand pebbles measured in the study area was found to be 5,47 cm, but elongated boulders of the less competent rocks may reach lengths of up to 150 cm.

THE MONOMICTIC MEMBER

This conglomerate reaches a thickness of 1000 m or more on the Central Plateau and is the most competent unit in the Cango Group. Pebbles of vein quartz and quartzite, averaging 3 cm but up to 20 cm in diameter, constitute 80 to 95% of the inclusions. They are well rounded in the study area but often subangular in the west. Black chert is ubiquitous but relatively rare, while silty quartzwacke, shale and coarse grained sandstone may also be present in varying amounts.

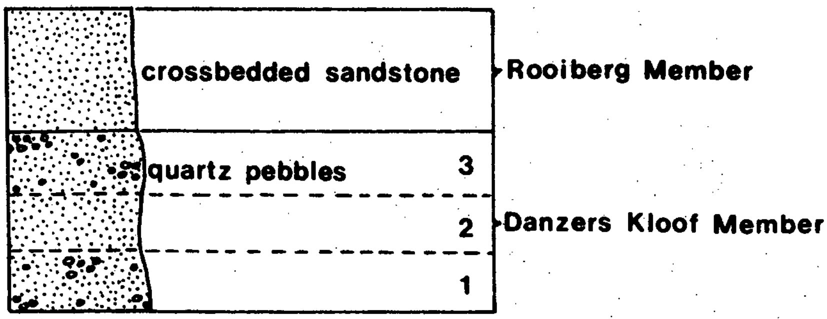
A conspicuous feature is that the conglomerate usually consists of numerous lenses with scattered pebbles inbetween. These lenses may be a few to 50 m or more in length, with poorly sorted sandstone or grit filling the gaps. The pebbles are set in a massive, arenaceous matrix and are usually spaced a few cm apart, although closer packing may occur.

Intercalations of conglomeratic and more gritty horizons produce massive bedding which can often be recognized only at a distance. Upwards in the succession the arenaceous lenses become more abundant and finally grade into the Uitvlugt Formation.

2.3.1.4. UITVLUGT FORMATION

TYPE LOCALITY Uitvlugt (23°26'S ; 22°07'E)

SECTION (Fig. 10.)



Building most of the rugged Central Plateau, the sandstones of the Uitvlugt Formation reach a thickness of 600 to 800 m on an overturned flank in the north. Traced eastwards a rapid thinning is encountered on the farm Voorzorg (22°24'E) and the formation finally disappears at Doorn Kraal (22°37'E), where a facies change brings the Groenefontein and Schoongezigt Formations in direct contact. In the north the formation appears beneath the Swartberg on the farm Tafel Berg (22°18'E), providing valuable evidence that the fine-grained quartz wackes north of the Boomplaas Fault can be correlated with those occupying most of the Cango outcrop area in the west and east.

DANZERS KLOOF MEMBER

At its type locality on Uitvlugt three zones can be distinguished in the Danzers Kloof Member. At the base a complete transition with the Vaartwell Formation comprises zone 1, with widely scattered and sporadic conglomeratic lenses and abundant crossbedding, which serve to distinguish this horizon from the Vaartwell conglomerate. Upwards there is a gradation to zone 2 which is relatively free of quartz pebbles. This gives way to another conglomeratic horizon, zone 3, which grades into the Rooiberg Member. Though not definable as a conglomerate at its type locality on Uitvlugt, zone 3 can be mapped as such on the farms Rykdom ($22^{o}09'$E) and De Olykraal ($22^{o}13'$E). The quartz pebbles are well rounded and average 1-2 cm in diameter.

ROOIBERG MEMBER

Quartz pebbles are rare in the Rooiberg Member and never sufficiently concentrated to form conglomeratic lenses. The member is composed of crossbedded, fine to coarse grained sandstones (Fig. 3 (d)) which are greenish grey in fresh outcrops and weather reddish brown, yellow and white. Most common are quartz wackes with a matrix of sericite and chlorite forming 20 to 40% of the rock, in which are set angular to subrounded quartz grains with minor feldspar. Sorting is generally poor. Second in order of abundance are quartz arenites (10 to 20% matrix) which are poorly to moderately sorted and often display sutured grain boundaries. Orthoclase and microcline can be present in amounts of up to 20%, forming conspicuous subarkosic horizons with a pinkish mottled texture. Plagioclase occurs in minor amounts (less than 2%).

The sandstones are massively bedded near the base and single horizons can attain thicknesses of up to 4 m. They are often separated by green to bluish green shale bands averaging 30 cm in width and spaced up to 10 m apart. Sandstones in the eastern section have 30 to 50 cm thick beds which show laminations a few mm to 3 cm wide.

Crossbedding is mainly of the tabular type (Plate 3 (d)), but festoon or trough cross-stratification is also abundant. (Plate 3 (c)). Inclination angles vary between 10^{o} and 40^{o} with an average of about 15^{o}, but the larger angles of repose are probably a result of tectonic deformation. Single foresets can be up to 4 m in length.

Ripple marks were observed in Schoemans Poort (Plate 3 (f)) and on the farm Koetzers Kraal (22°02'E). The crests are slightly asymmetric, wave lengths in the first case being 2,5 to 5 cm and the amplitude 2 to 3 mm. The ripples on Koetzers Kraal seem to have undergone considerable deformation.

Sandstone-shale contacts show many interesting features such as load casts (Fig. 11 (a)), small scale erosion scour (Fig. 11 (b)) and shale injections parallel to cleavage in the sandstone. Rip-up clasts of shale are common near such contacts.

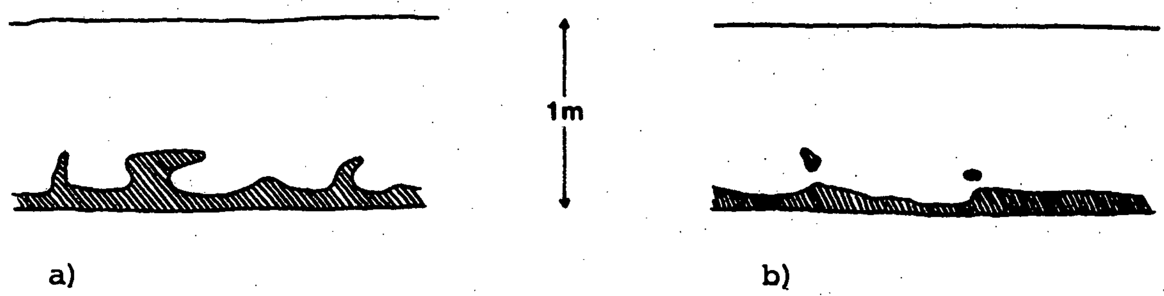
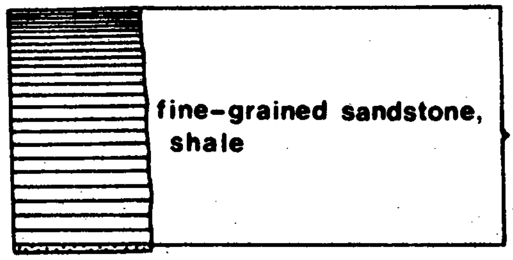


Fig. 11. (a) Load casts of sandstone in shale, Potgieters Poort.
(b) Erosion scour with rip-up clasts of shale, Koetzers Kraal.

2.3.1.5. GEZWINDS KRAAL FORMATION

TYPE LOCALITY Gezwinds Kraal (33°26'S.; 22°04'E).

SECTION (Fig. 12.)



A complete, continuous gradation with the crossbedded sandstones of the Uitvlugt Formation defines an arbitrary lower contact for the Gezwinds Kraal Formation. The broad difference in lithology is however reflected in the type and colour of weathering, the Uitvlugt sandstones being white or cream at a distance as opposed to the reddish brown of the Gezwinds Kraal Formation. The latter is also characterized by a smoother topography.

In fresh outcrops the quartz wackes are green to greyish blue with intercalated olive green or blue shale bands. Compositional and textural variations are minimal, the Gezwinds Kraal sediments as a group being the most homogeneous in the Cango. The wackes are composed of very fine-grained, angular to subangular quartz grains (45-70%) with minor plagioclase ($<4\%$), alkali feldspar ($<1\%$) and calcite fragments. The matrix (30-55%) is chlorite and sericite (Fig. 3 (e)).

Although the typical massive bedding of the Uitvlugt Formation is absent, minor crossbedding sporadically occurs in the basal part of the Gezwinds Kraal Formation. Both tabular and trough varieties are present, with inclination angles generally higher than those of the Uitvlugt. Bedding thickness varies between 30 and 100 cm for the quartz wacke and 5 to 60 cm for shale. Although these wackes and shales are rhythmically intercalated in the study area, the alter= nation is less regular in the west.

Upwards in the succession shale becomes more prominent while the wackes become increasingly finer grained. This is accompanied by a marked thinning of beds so that the two rock types alternate on a cm scale in the uppermost parts. Grading was not recognized in the wackes and the contacts with shale are always sharp.

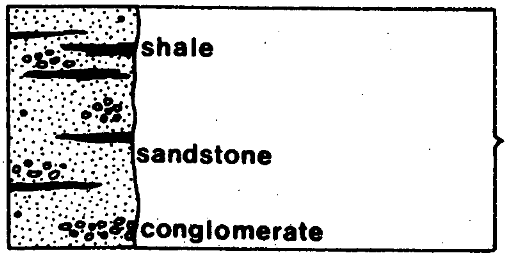
Chloritic phyllite can cover large areas, as for example on Buffels Bosch Rivier ($22^{\circ}01'$E) and in Meirings Poort ($22^{\circ}33'$E).

A gradual change in facies is observed eastwards. Sandy and occasional grit bands with rare crossbedding make their appearance, while still farther east well rounded quartz pebbles necessitate the establishment of a lateral facies boundary.

2.3.1.6. SCHOONGEZIGT FORMATION

TYPE LOCALITY Schoongezigt (33°28'S.; 22°48'E)

SECTION (Fig. 13.).



The Schoongezigt Formation is separated from the Gezwinds Kraal by a lateral facies boundary, which is defined by the first appearance of quartz pebbles. Composed of fine, medium and coarse grained quartz wackes with scattered pebbles and conglomeratic lenses, the formation is also more massively bedded.

Fresh samples are bluish green, greenish grey or grey, which weather first to cream and then buff to reddish brown tints. Where Tertiary terraces have only recently been removed by erosion the underlying wackes show a marked difference in colour, often creating the impression of unconformities at a distance.

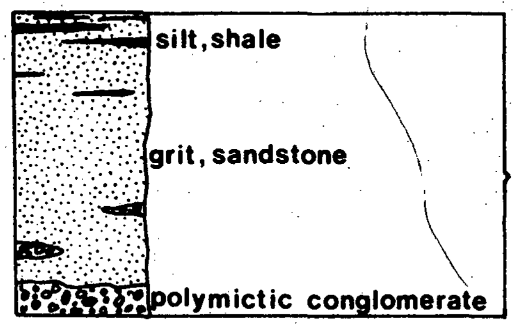
There is an increase eastwards in number, type and size of pebbles, but in the extreme eastern part of the outcrop inclusions again seem to be absent. The full range of pebble types includes quartz, black and brown chert, quartz wacke, grit, sandstone and shale. The majority are well rounded and vary in size from 1 to 15 cm.

In thin section, apart from an increase in grain size, the wackes are similar in texture to those of the Gezwinds Kraal Formation, while the compositional plot of three samples on the Q F M diagram fall in the same field (Fig. 3 (f)).

2.3.2. THE SCHOEMANS POORT FORMATION

TYPE LOCALITY Schoemans Poort (33°29'S.; 22°15'E)

SECTION (Fig. 14.)



The "Unconformable Feldspathic Grit Formation" was tentatively correlated on lithological grounds with the Klipheuwel Formation in the Western Cape. Although Stocken bases this correlation partly on the observation that it is cut by "dolerite dykes similar to the Western Province dolerites," thus distinguishing it from the Nama Group, he neither described them nor did he indicate their position on his map. As the present writer could find no such intrusives in the formation, it must be assumed that the dykes referred to by Stocken probably belong to the same suite cutting the Cango Group, as it seems very unlikely for younger dykes to be restricted to the limited outcrop of this formation. There is no evidence that the Cango diabase is related to the dolerites of the Cape and it therefore should not be used for correlation purposes. (See p.29). Furthermore, according to Hartnaday et al (1974) the Klipheuwel Formation is distinguished from the Malmesbury rocks partly by a "characteristic reddish colouration", which is in contrast with the pale grey to white appearance of the Schoemans Poort Formation. It should therefore suffice to say that the latter is of post- Cango, pre-Cape age, without attempting to correlate it with distant counterparts.

It is clear from its unconformable relationship with the

Uitvlugt Formation on the farm Rykdom (22°10′E), that the Schoemans Poort Formation was deposited during the waning stages of or after the pre-Cape deformation, following the subsequent uplift and erosion of the Cango Group. Farther to the east it is separated from the Groenefontein Formation by a tectonic contact.

Because of extensive shearing the bedding is often completely destroyed near the upper contact with the Table Mountain Sandstone in the south, so that an unconformable relationship could not be confirmed in this case. Both formations dip south at about 30°.

The sediments consist of greenish white sandstones and grits weathering to pale grey or white, with bands and lenses of conglomerate in the basal and irregular shale and silt lenses in the uppermost parts.

The conglomerate is similar to that of the Vaartwell polymictic member, but weathers to lighter colours and does not display the characteristic a-lineation of pebbles in the latter. Quartz, sandstone and shale pebbles in the Schoemans Poort Formation are subrounded and can be 30 cm or more in diameter. As Stocken points out, the cobbles may well be derived from reworking of the Vaartwell conglomerate, but it is also possible that they were eroded directly from the formations of the Cango uplifted elsewhere.

The grits and sandstone are massively bedded and consist of medium to coarse, subrounded or angular quartz with up to about 20% orthoclase, (Fig. 3 (g)), set in a fine quartz-sericite matrix. Rock fragments and flakes of muscovite are present in the coarser grit and conglomeratic layers.

As is the case with most of the Cango sediments, the Schoemans Poort rocks have a large percentage of matrix (up to 50%) of which a considerable amount is the result of secondary alteration of quartz and feldspar to sericite. This aspect is more fully discussed under "Diagenesis and Metamorphism". (Chapter 4).

2.3.3. TABLE MOUNTAIN GROUP

A thin shell of Table Mountain Sandstone occurs as a remnant on the northern side of the Cango Fault, where it unconformably overlies different formations of the Cango Group. The quartz arenites (Fig. 3 (h)) are generally medium to coarse grained, massively bedded and contain many well rounded pebbles of white and blue quartz, quartzite and sandstone, which are up to 15 cm in diameter. These form small conglomeratic bands and lenses which are sporadically developed in the sand= stone over its entire outcrop area in the south.

2.3.4. TERTIARY TERRACES

During the Tertiary an extensive erosion bevel developed over the Cango which today is evident in many high level gravel remnants associated with a shoulder at 1000 to 1500 m along the southern flank of the Swartberg. (Plate 1 (a)). Debris from the upper slopes of the mountain accumulated on a flat plain along the northern fringe of the Cango. A projection of the remnants of this Tertiary surface to the south discloses that part of the Central Plateau east of Coetzees Poort must still have been above base level, which is confirmed by a similar shoulder appearing on the northern side of the Central Divide (Plate 1 (b)). To the west the peneplain also reveals itself as a gently sloping surface in the higher hilltops (Plate 1 (c)). That larger rivers such as the Gamka drained the plain towards the south at this time, is shown by the sloping on both sides towards the river of what is still clearly the original erosion surface.

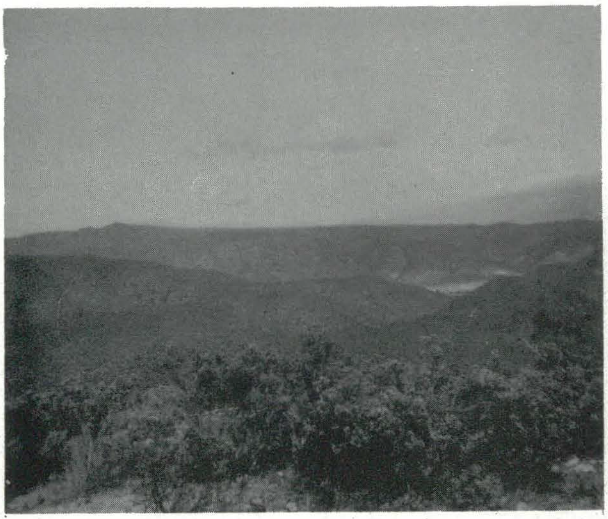
While in total 41 terraces unconformably overlie the Cango sediments, another 25 are situated on the erosion shoulder of the Swartberg. Most occur east of 21°20′, which could indicate a slower rate of erosion and possibly a retarted post-Tertiary epeirogenic uplift in this area. The remnants generally lie within 3 km from the northern Table Mountain Sandstone contact at heights of 730 to 980 m. The average slope is about 5° southwards.

Lithologically the terraces consist of loose or consolidated boulders and cobbles which have been derived from the Table Mountain Sandstone. The boulders can be more than 5 m in

PLATE 1.

(a) Tertiary erosion shoulder on southern slopes of Swartberg (in background).

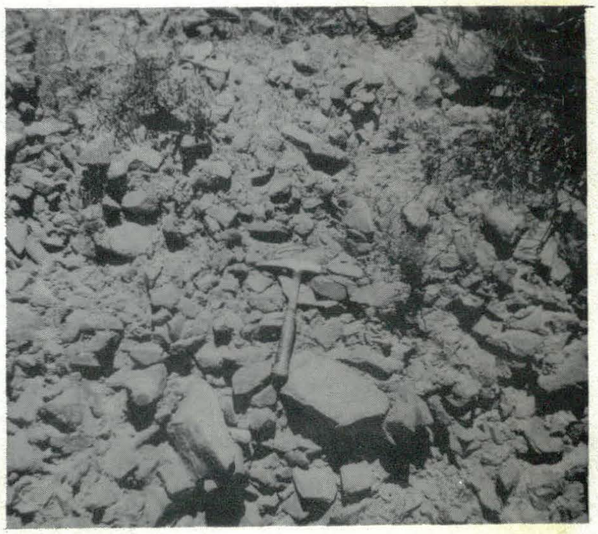
(b) Similar shoulder on northern slope of Central Divide (in background).



(c) Tertiary erosion surface in Cango beyond Coetzees Poort. Looking west.

(d) Paleoterrace of the Gamka River. Looking south.

(e) Calcrete.



(f) Talus. Note angular cobbles as opposed to rounded river gravels in (d).

diameter and are relatively well rounded. The cement usually consists of secondary silica.

In the eastern section one terrace has three distinct layers and an estimated thickness of 15-20 m.

On the farm Schoongezigt a terrace straddles the Cango Fault, proving that virtually no movements has occurred on the latter since the Tertiary. Another terrace overlies the Boomplaas Fault on Rust En Vriede.

2.3.5. RECENT DEPOSITS.

2.3.5.1. CALCRETE

Superficial calcareous deposits are usually found on flat surfaces such as the larger river valleys and in the vicinity of limestone or diabase outcrops. The calcrete is white to greyish blue with angular to subrounded inclusions of lime= stone, quartz and shale, which are a few mm to 100 cm in diameter. (Plate 1 (e)). Although only 30 to 60 cm thick, up to three layers of calcrete are superposed on one another.

On the farm Kombuys a calcareous grit, which is thought to be of post-Tertiary origin, is described by Stocken.

2.3.5.2. FERRICRETE

On the farm Voorzorg (22°24′E) a patch of ferricrete occurs on the Wildehonde Kloof Fault line. It is 20 cm or more in thickness and contains subrounded and rounded inclusions of the surrounding rock.

2.3.5.3. GRAVEL

Two types of unconsolidated conglomerate are present. Angular, flat pebbles and boulders a few cm to more than 2 m in length are found on steep slopes (Plate 1 (f)), while well rounded, large boulders occur in old river beds. The latter type forms conspicuous terraces on the flanks of the Gamka River at heights of about 20 and 80 m above the present river-bed (Plate 1 (d)). Most of the boulders were derived from the Swartberg Table Mountain Sandstone.

2.3.6. IGNEOUS ROCKS

2.3.6.1. FIELD DESCRIPTION

Although McIntyre (1932) reported occurrences of granite and gneiss near the southern entrance of Schoemans Poort, no acid intrusions were found by subsequent workers and during the present investigations. Basic intrusions in the form of dykes and sills are however abundant in the study area.

Two generations of dykes were recognized. Those of the older suite are rarely observed and so much sheared and altered that they resemble shaly horizons in the country rock. One example occurs at the Gamka River outside the study area. Although no thin sections were made of these, they probably represent true greenschists as they were evidently deformed and metamor= phosed with the Cango Group.

The second generation of dykes intruded the Cango rocks after their deformation and are presumably of the same age as the sills. The green to bluish rock is much weathered on the surface (typically producing spheroidal crusts), but fresh samples were obtained by breaking down large boulders. The dykes commonly vary from 1-3 m, but seldom exceed 30 m in width, and range from 50 m to more than 4 km in length. They usually show up as dongas due to a more rapid rate of erosion. One such weathered dyke is responsible for the rock fall that produced the famous "Devil's Workshop" in the Cango Caves.

Of the 11 known sills, 10 occur outside the study area in the vicinity of the Gamka River, but a large intrusion near the southern entrance of Potgieters Poort (the Drooge Kraal Sill) has been investigated in more detail. This intrusion is over 14 km long, 500 m wide and on estimation 500 m in thickness. It has intruded the country rock parallel to the bedding, showing a dark chill zone not more than 3 m wide.

Although previous workers alotted a pre-Cape age to this intrusion on the grounds that no contact metamorphism is visible in the Table Mountain Sandstone directly overlying it in this area, this assumption should not be made without further consideration. It has been observed (See p. 99) that the rocks of the Cango Group show little evidence of contact metamorphism except for some induration, which would be very difficult to recognize in the Table Mountain Sandstone. That the sill is older however than the post-Cape orogeny, is shown by a 1 - 1,5m wide zone along the contact, in which the diabase is sheared and intruded by quartz veins. A weak foliation occuring away from this contact may also represent cleavage resulting from the post-Cape deformation (Plate 7 (b)).

2.3.6.2. PETROGRAPHY

A rather detailed petrographic description of the western intrusions is given by Mulder (1954).

In summary: the intrusions consist of highly altered diabase or epidiorite with subophitic or poikilitic textures in which no orientation of crystals was detected. The dykes are usually fine-grained and now consist mainly of the alteration products of plagioclase, pyroxene (actinolite) and ilmenite. Some quartz (probably derived from the country rock) is always present. The sills are mostly medium to coarse grained (crystals 1 - 5 mm) and originally consisted of more than 90% plagioclase and pyroxene in about equal amounts. Plagioclase occurs as idiomorphic or hypidiomorphic crystals with polysynthetic twinning and an An - content of 50-70%. Much of the feldspar is now altered to kaolinite, saussurite and chlorite, while myrmekitic quartz is present as an exsolution product. The pyroxenes constitute 35-55% by volume of the rock, with clinopyroxene (pigeonite - augite) predominant and orthopyroxene (hypersthene) present in minor amounts. Hornblende is the main alteration product and in many cases forms a major part of the rock, while chlorite represents a still later stage in the alteration. Among the accessory minerals ilmenite is predominant as large irregular grains with kelyphitic rims of leucoxene. A minor amount of clinozoisite and zoisite is present, while apatite, magnetite, titanite, zircon, pyrite, biotite and serpentine (an alteration product of olivine) constitutes the bulk of the remainder. Interstitial quartz is present in the contact zones.

Olivine was observed in thin sections of samples taken from the base of the sill. These rounded grains must have settled through fractional crystallization from an originally more simatic magma.

2.3.6.3. PETROCHEMISTRY

Chemical analyses of 3 samples were made, two of these being from the Drooge Kraal (JPRD) and Gamka River (JPRC) sills respectively, and the other (JPRB) from a dyke in the vicinity of the Wynands River (Table 1 (a)).

TABLE 1 (a) CANGO DIABASE

SAMPLE	SiO_2	TiO_2	Al_2O_3	Total Fe	MnO	MgO	CaO	Na_2O	K_2O	P_2O_5	LOI H_2O	Total
JPRB	48,69	2,83	12,67	14,34	0,24	5.36	10,48	1,87	0,12	0,31	2,33	99,24
JPRC	47,54	1,36	15,59	12,82	0,19	7,18	8,38	1,68	1,09	0,29	3,01	99,13
JPRD	49,30	1,21	9,75	19,79	0,27	13,13	6,35	1,23	0,64	0,28	2,01	103,96

Analysis by D.H. Cornell (1977).

TABLE 1 (b) AVERAGE COMPOSITION

SAMPLES	SiO_2	TiO_2	Al_2O_3	Total Fe	MnO	MgO	CaO	Na_2O	K_2O	P_2O_5
CANGO (3)	48,51	1,80	12,67	15,65	0,23	8,56	8,40	1,59	0,62	0,29
W.P. (8)	48,66	2,04	15,34	12,31	0,22	6,21	9,06	2,64	1,22	0,37
GEORGE (17)	46,28	1,83	15,02	11,93	-	6,64	9,76	2,15	0,54	0,45

Data from Potgieter (1949), Nell and Brink (1944).

With the exception of SiO_2, the chemical composition of the 3 samples analised is rather variable, and for comparison with the Karoo, Western Province and George dolerites two diagrams were selected. On the total alkalis: silica diagram (See Gresse, 1976) the composition of the Cango samples plots in the tholeiite and high-alumina basalt fields, which distinguishes them from the Western Province dolerites (Fig. 15). The average Karoo dolerite plots at a higher SiO_2 content, while the George metadolerites plot to the left of the Cango samples. Two of the latter also plot at a lower alkali content than the Karoo, Western Province or George dolerites.

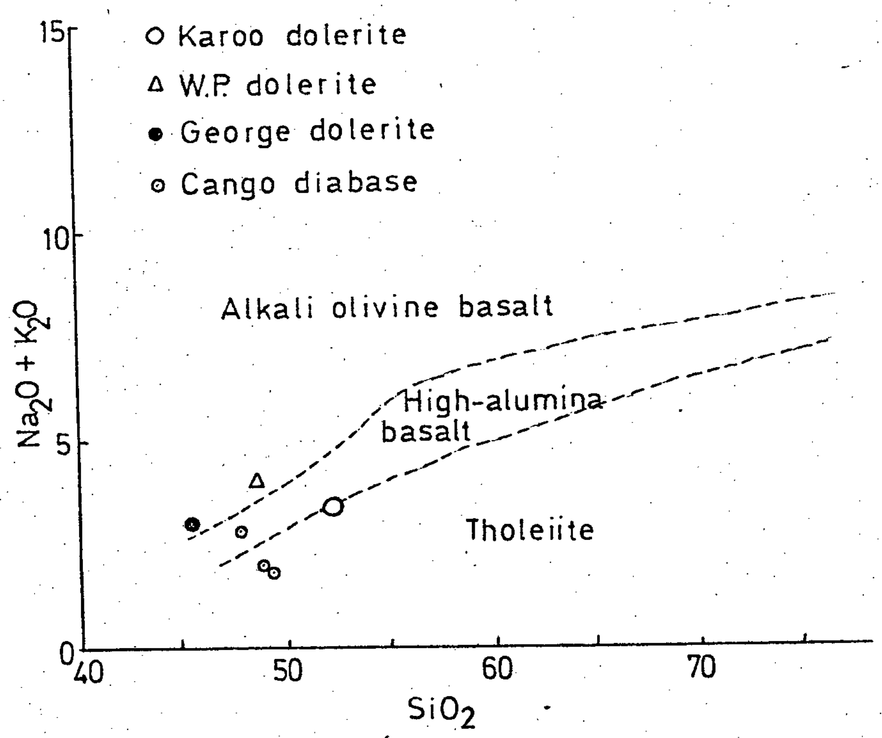


Fig. 15. Alkali - silica diagram.

On the molecular ratio diagram (See Nell and Brink, 1944) the Cango diabase is clearly distinguished from the Karoo dolerites by plotting in the Western Province dolerite field (Fig. 16).

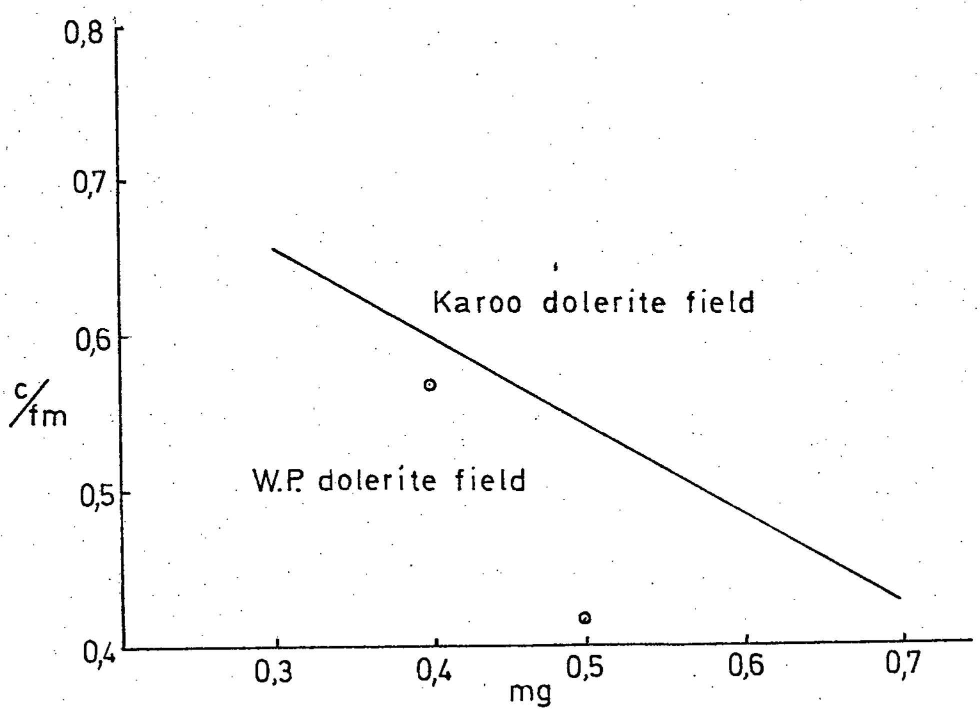


Fig. 16. Molecular ratio diagram (After Nell and Brink, 1944).

Average compositions of the Cango, Western Province and George dolerites (Table 1 (b)) reveal no clear chemical relationships and there is therefore no evidence that any of these intrusions belong to the same suite.

3. SEDIMENTOLOGY

3.1. INTRODUCTION

The sediments of the Cango Group have been affected both in composition and texture by the processes of diagenesis, metamorphism and weathering. It is therefore of prime importance to evaluate first these obscuring features before attempting to read petrographically the origin and environments of deposition of the rocks. For a description of the textural and compositional changes induced by these processes the reader is referred to the next chapter.

Although only those grains showing little evidence of modifi= cation were used for determining graphic parameters such as sorting, the results are principally intended to serve as comparative data among the rocks of the Cango Group them= selves. Comparison with younger sediments are based primarily on larger external characteristics such as sedimentary structures.

3.2. DEPOSITIONAL ENVIRONMENTS

3.2.1. MATJIES RIVER FORMATION

Although Stocken regarded the limestone-shale horizon as an unstable shelf deposit, (p.52), he did not provide any evi= dence beyond very broad considerations. In the following discussion the general principles of carbonate deposition are adhered to in an attempt to reconstruct the probable conditions of deposition. Field evidence, petrographical observations and geochemical data are used in conjunction with theoretical aspects.

3.2.1.1. CHEMICAL CONSIDERATIONS

(Rodgers, 1968; Newell and Rigby, 1957; Ginsburg, 1957; Fairbridge, 1957; Friedman, 1964; Milliman, 1976.)

Before the end of the Precambrian, algae were the only organisms capable of precipitating Ca-and Mg-carbonate, so that chemical deposition predominated during this period. From the early Cambrian onwards more advanced organisms appeared which were capable of extracting the excess $CaCO_3$ in sea water to build hard parts. Organic limestones however only surpassed chemical limestones in the Middle Ordovician, when $CaCO_3$ was in general used up fast enough to prevent chemical precipitation. It can therefore be assumed that the Cango limestones, Precambrian in age, were either deposited chemically (inorganically) or precipitated by obscure, lime-secreting algae. The presence of organic carbon seen in thin sections as black specks, veinlets and rims around limestone intraclasts or quartz grains may suggest an organic control in this case.

It is generally agreed that sea water saturated or supersatu= rated with $CaCO_3$ is most favourable for large scale carbonate deposition. As the solubility of $CaCO_3$ increases with colder temperatures and water depths, it is in the warm, shallow seas of the low latitudes that production of marine limestone is greatest.

CO_2, which is transferred from the atmosphere to the oceans according to the reactions:

$$H_2O + CO_2 \rightleftharpoons H^+ + HCO_3^-$$

$$HCO_3^- \rightarrow H^+ + CO_3^{=}$$

plays an important part in carbonate deposition. The partial pressure of CO_2 varies with changes in salinity, temperature, pressure and biologic activity. Increasing water turbulence also causes increased uptake of CO_2 from the atmosphere, so that an influx of CO_2 occurs during storms and alkalinities are lowered. Inversely, removal of CO_2 raises the pH and favours precipitation of $CaCO_3$. During photosynthesis algae can reduce the partial pressure of CO_2 while respiration produces the opposite effect. It is thus clear that the CO_2 system is easily affected by external factors, and these variations may be responsible for the rhythmic intercalation of shale and limestone so typical of the Cango carbonates.

3.2.1.2. EVIDENCE FOR MARINE ORIGIN

STABLE ISOTOPES

According to Urey (1947), the isotope composition of a solid precipitated from solution is related to the isotopic composition of the precipitating medium. Milliman (1976) has plotted ^{13}C of modern carbonate sediments against ^{18}O and concluded that fresh water limestones have considerably lower ^{18}O and ^{13}C values than marine carbonates. ^{13}C and ^{18}O values of 7 samples of Cango limestone and dolomite determined by Schidlowski, et al (1974) give an average value of +0,74 for ^{13}C and -8,68 for ^{18}O as P.D.B. * As the ^{13}C ratio appears to be almost constant from the early Precambrian up to recent times, these values indicate that the limestones of the Cango are probably of marine origin (Fig. 19, Milliman). However, the steady increase of photosynthetic oxygen throughout the sedimentary record (Fig. 12, Schidlowski et al,) probably accounts for the ^{18}O values of the Matjies River limestones being lower than the average for marine carbonates. Exposure to fresh water may also produce lower (negative) values for ^{18}O (Bathurst, 1971)

TRACE ELEMENTS

Friedman (1969) has shown that the variables in the chemistry of the precipitating medium are reflected in the geochemical composition of carbonate sediments. The concentration of the trace elements Ba, Fe and Mn is commonly greater in fresh water than in sea water environments. Three carbonate samples from the Cango Group were analysed (Table 2). The results indicate a low concentration of Fe and Ba, which is less than the general values for river and lagoonal sediments. (Fe mostly $>$ 500 ppm, Ba commonly $>$ 40 ppm.) Mn, however, greatly exceeds the average values for marine carbonates, which are usually less than 30 ppm.

* S M O W values for ^{18}O given by these authors were recalculated to P.D.B. using the formula $^{18}O = (^{18}O \times 1,03) + 29,5$:- Milliman, 1976).

Table 2.

SAMPLE	Fe	Mn	Ba (ppm)
1 (Matjies River Formation)	6 ± 5	500 ± 50	12 ± 10
2 (Matjies River Formation)	56 ± 10	170 ± 20	17 ± 10
3 (Groenefontein Formation)	210 ± 20	1980 ± 150	17 ± 10

Analysis by D.H. Cornell.

3.2.1.3. EVIDENCE FOR SHALLOW WATER DEPOSITION

FIELD EVIDENCE

As the deposition of deep sea carbonates only commenced with the evolution of coccolithopods and foraminifera during the early and mid-Mesozoic, most sedimentation prior to this time probably occurred in shallow water areas or semi-enclosed basins (Milliman, 1976). Deeper water carbonates therefore, if they occurred during the Precambrian, must have been derived from shallow water carbonate areas by slumps, turbidity currents or sliding of large blocks down the steep slopes bordering such areas. The total absence of turbidity features such as sole markings and graded bedding in the Cango limestone makes a similar origin very unlikely. Moreover, olistoliths of limestone were not observed, while the absence of large scale convolutions also suggests that these carbonates were not deposited on a steep slope. The typical, regular alternation of shale and fine-grained limestone of the order of a few decimeters per cycle, is a common feature of many shallow water marine successions on the other hand. (Duff, et al, 1967).

Evidence for penecontemporaneous erosion is abundant in the form of intraclast limestone in the carbonates and intra= formational conglomerate in the sandstones of the formation. The allochems of the former consist of elongated and subrounded fragments of limestone differing slightly in colour from the host. They are up to 1 cm in diameter and in thin section are seen to consist of subhedral and rounded grains of calcite considerably coarser (up to 700μ) than the surrounding matrix of the host (30-180μ). The boundaries are irregular and cut across individual grains in the intraclasts. Oölites which occur in these intraclasts are of roughly the same size as the other grains and show much evidence of recrys= tallization, suggesting that the original limestone was probably oölitic and lost its character after metamorphism. The presence of intraclasts in limestone is generally indicative of the proximity of semi-consolidated outcrops, either subaerial or subaqueous, which during storms provide broken fragments of limestone that are shifted and incorporated in later lime muds accumulating on the depositional surface. The amount of lithification necessary can be the result of contact with fresh water during exposure above tide level (Bathurst, 1971).

Intraformational conglomerates are found in the sandstones of the Nooitgedagt Member. On the farm Kombuys (22°13′E) for example, a lens of brownweathering wacke has a poly= mictic paraconglomerate at the base which contains unsorted pebbles of limestone, quartz, shale and sandstone. The limestone inclusions, being the largest, obviously did not suffer considerable transport, in contrast to the other types which are also better rounded. This suggests erosion from the limestone lens immediately underlying the conglo= merate in this area. On Welgevonden (21°57′E) limonite has replaced most of the grit matrix of a conglomerate with subrounded and rounded inclusions of quartz, grit, chert and limestone, the latter up to 30 cm in length. Pebbles of gneiss, granite and dolomitic limestone are occasionally encountered.

PETROGRAPHICAL EVIDENCE

The sandstones of the Nooitgedagt Member display a diversity of types that range from quartz arenite, subarkose and quartz wacke to typical greywacke. This probably indicates an area with considerable local variation in depositional conditions, which would not be expected in a deep sea environment. Together with the lenticular, massive nature of the sandstone the variation can best be explained by a supra - to sublittoral zone where local topographic features could provide different milieus for winnowing, sorting and rounding of the terrigenous detrital matter delivered into the site by rivers.

According to Chilingar et al (1967), widespread accumulations of pure to relatively pure limestones, and in particular fine-textured varieties, represent moderately shallow to very shallow marine realms. This is supported in the Matjies River limestones by the occurrence of oölites and rare dolomitic lenses, the former providing possibly the strongest evidence for shallow water conditions. Modern oölites seem to be restricted to tropical and subtropical environments where waters are generally less than 2 m deep, although they can form at depths up to 15 m (Milliman, 1976). Most investigators agree that oölites form in environments of agitation where waves and tidal currents are strong enough to keep the ooids in suspension in a turbulent solution. Diffusion rates in water are probably too slow and will hamper the growth of oölites in a stagnant environment (Bathurst, 1971). In a strongly agitated environment the ooids are kept in suspension for longer periods, a high level of supersaturation can thus be maintained at the ooid surface and several lamellae can form around the nucleus. The neces= sary ions are provided by new supplies of supersaturated waters brought over the edge of platforms during tides. Oölites formed in non-agitated environments tend to have discontinuous layers around the nuclei and rough surface textures.

Thin section studies of the Cango oölites show them to have diameters of 100-750μ with an average of 250μ. Two or three, even up to five lamellae surround large nuclei of dirty brown calcite, the individual coatings ranging from 15 to 85μ in thickness. Occasionally these laminae offlap to leave the nucleus exposed. The oölites are scattered through a microgranular, recrystallized matrix, and were probably formed in a moderately agitated environment.

Oölites in the intraclasts show better rounding and polish= ing by abrasion, have diameters of up to 600μ and 3 to 4 laminae which are fairly thick (20-140μ). The nuclei are usually very small, but where large recrystallized nuclei occur they consist of coarse, clear calcite. Recrystallization proceeds from the centres outward.

These oölites seem to have formed in a more violently turbulent environment - probably on the shelf margin - as suggested by their smoother outlines, smaller nuclei, a large number of thicker lamellae on the average, and their occurrence in intraclasts.

The presence of thin, intercalated beds of calcitic dolomite in the limestones is another indication that deep water was not involved. Although there is still some controversy as to the origin of dolomite, it seems as if modern occur= rences are mostly of shallow water origin. In the Bahamas dolomite is formed at or near the surface on exposed supratidal flats, where flooding by spring high tides and storms is followed by long periods of subaerial exposure. (Shinn et al, 1965).

The typical features of late stage diagenetic metasomatism like coarse recrystallization and the absence of fine inter= bedded limestone (Fairbridge, 1957) are not observed in the Cango dolomites. An early diagenetic origin is therefore most likely. According to Usdowski (1968) dolomitization during early diagenesis can be caused by sea water concentra= ted by evaporation of H_2O at the sediment surface or in lagoons. Reaction of the pre-existing $CaCO_3$ with the more concentrated solutions can produce dolomite at low temperatures:

$$CaCO_3 + Mg^{++} + CO_3^{=} \rightarrow Ca\,Mg\,(CO_3)_2$$

The $CO_3^{=}$ can be supplied during weathering of silicates which releases OH^- ions to react with the CO_2 in the air:

$2OH^- + CO_2 \rightarrow CO_3^{=} + H_2O$ (Lippman, 1968). Shelf brines according to Deffreys et al (1965), are the most effective agents of dolomitization, because loss of Ca by precipitation causes their Mg/Ca ratios to exceed that of normal marine waters.

The limited extent of the Matjies River dolomitic lenses suggests that supratidal conditions were reached only at favourable localities, which could have been analogous to the modern palm hummocks of the Bahamas.

Further evidence for subaerial exposure, and therefore shallow water, comes from the dolomitic lens on Kombuys (22°14′E). Coarse "dogtooth" crystals of calcite (up to 500 µ) line the walls of microgeodes in a fine (2-10 µ) dolomitic matrix. The crystals are obviously void fillings and become larger from the walls of the druses towards the centres. According to Friedman (1964) a drusy calcite mosaic occurring in marine limestones indicate that they were subjected to subaerial diagenetic changes or to fresh waters in the subsurface. A modern example of drusy mosaic described by Shinn et al (1965) occurs at Andros Island, where dolomite crops out in crustlike layers above the normal high tide level.

Considering the evidence outlined above, it seems to be fairly well established that most of the Matjies River limestones were deposited in waters probably less than 10 m in depth, with local islands or sand banks providing supratidal environments for the formation of dolomite.

3.2.1.4. PALEOGEOGRAPHY AND PALEOSLOPE

NOOITGEDAGT MEMBER

As mentioned earlier, the lenticular beds and variability of the Nooitgedagt sandstones indicate a complex environment of deposition. In spite of the rare penecontemporaneous erosion and scour, the absence of crossbedding or well-defined ripple marks suggests that current flow was generally weak and variable. Although some characteristics of alluvial deposits are thus displayed by the sandstones, they lack the abundant, well-orientated crossbedding and channel-fill structures of this type of environment. The poor sorting and roundness, relatively high feldspar content and general immaturity of most of the sediments could indicate that these were poured into a body of quiet, standing water such as a shallow restricted bay or shelf, where tidal and longshore currents were too weak to be effective agents of winnowing. The many small topographic irregularities characteristic of most shelves, such as sand swells, terraces and shallow channels, can cause some parts to be subjected to wave turbulence while others are quiet. This variability should be sufficient to account for most textural, structural and compositional differences among the sandstones under discussion, but deposits from marginal marine areas, deltas and fluvial tracts may all be included in the succession.

Plots of Phi Median Diameter (MdØ), Phi Deviation Measure (σØ) and Phi Skewness Measure (αØ) of 8 Nooitgedagt samples compared with the grain-size parameters of various modern sediments (Fig. 17 (a) and (b)) show that river channels, bays and lagoons can all be considered as possible depositional environments.

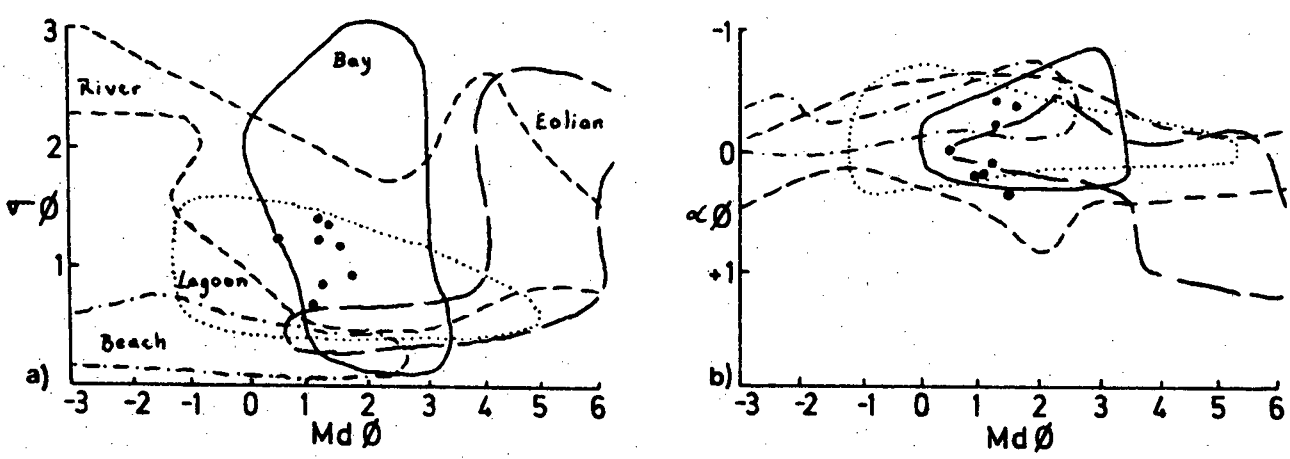


Fig. 17 (a) Nooitgedagt Member: Sorting vs grain size.
(b) Nooitgedagt Member: Skewness vs grain size.
(Data compiled from Kukal, 1971).

The presence of at least two horizons of marine limestone means that, if not wholly a marine deposit, two or more cycles of transgression must have intervened during deposition of the Nooitgedagt. The two carbonate horizons are probably part of a larger, unexposed cycle of clastic/chemical interbedding, of which the alternative origins are outlined below.

In the first case, a relatively stable depositional site with little variation in sea level is envisaged, and the megacycles are considered to be the result of external influences on the sedimentation. These in turn can be traced back to:

i) Climatic changes involving prolonged dry periods during which influx of terrigenous matter was small, or

ii) Rhythmic tectonic movements in the source area, including periods when the continental interior was low and yielded no important volume of clastic material.

The presence of calcareous sandstones, and the fact that the carbonate mesocycles are composed of limestone and argillaceous material, suggest that shifts in the CO_2 system because of climatic variations could have had very little influence on the larger cycles. While the mesocycles may have formed as a result of rhythmically interrupted carbonate precipitation against a constant "background" of clay deposition, the megacycles are characterised by coarse clastics. During periods of rapid terrigenous influx the large-scale precipitation of limestones can be prevented by "dilution" of the carbonate grains or by burial of carbonate producing algae. The abundance of unstable feldspar in the sandstones supports the hypothesis of rapid burial, but it is also an indication that the source area was not too far away. If this was the case, the effect of up=lift in the provenance would have been felt in the area of deposition, causing regression of the sea level and deposition of coarse clastics into fluvio - deltaic or alluvial environments along the coast. Evidence for a stable shore=line site of deposition would thus lean heavily on climatic control over erosion and transport of detrital matter from the source area.

In an unstable environment of deposition, uplift of the interior could cause regression of the sea level, produce higher rainfall and increased erosion on the seaward side of newly emerged mountain ranges and thus simultaneously change the depositional milieu and amount of terrigenous material reaching the coast.

Assuming that changes in sea level are the main cause for the limestone megacyclicity, it remains to be solved whether the sandstones of the Nooitgedagt are marine deposits, or whether they represent terrigenous sediments periodically flooded by transgressing seas. While there is no evidence in favour of the latter against a nearshore marine environ= ment, and considering that the well sorted, mature blanket arenites of advancing beaches are not represented in the succession, the first alternative is regarded the most likely. (Fig. 18 (a) and (b)).

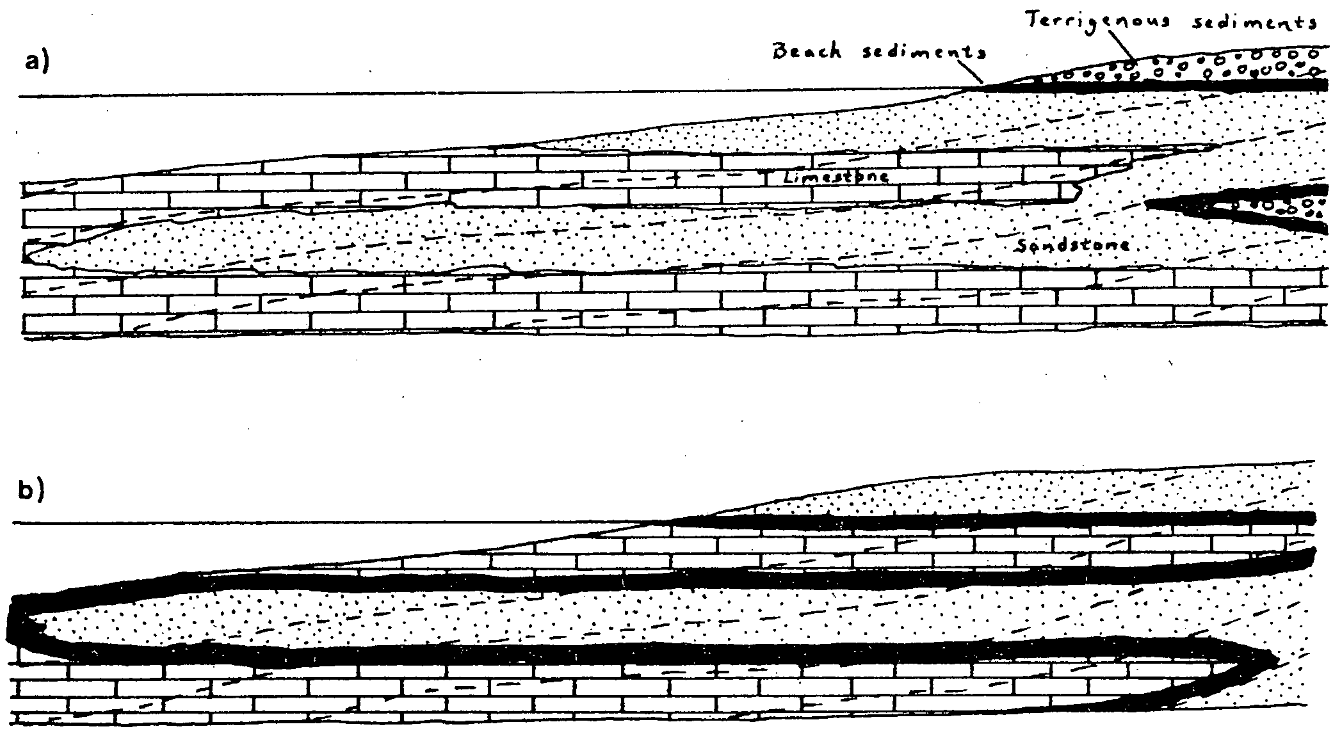


Fig. 18. Stratigraphic cycles in: a) wholly marine deposits.
b) marine/terrigenous deposits.

The Nooitgedagt depositional environment is therefore envisaged as a shallow, unstable shelf or bay bordering an uplifted interior with short, swift streams carrying detritals to the coast. The shelf has an irregular bottom topography with quiet waters near the shore, in which sand and clay accumulate. Wave and current action is restricted to shoal areas. On the shelf edge inflowing water is warmed and saturated with $CaCO_3$.

resulting in a broad belt of limestone parallelling the margin. Wave turbulence is greater on this shallower part of the shelf, which is favourable for the growth of oölites. At the margins of small islands or hummocks protruding above high tide level and occasionally flooded during storms, crusts of dolomite develop.

A modern example of such an environment can be found on the continental shelf off the southeastern United States south of Cape Hatteras, where a nearshore zone of quartz sand 30 km wide lies next to an offshore zone of pure carbonate sands. Any shift in the boundary between the two mileus would produce discreet interbeds of sandstone and limestone (Gorsline, 1963). Another example described by Gould and Stewart (1955) lies off the northern Florida panhandle and Alabama.

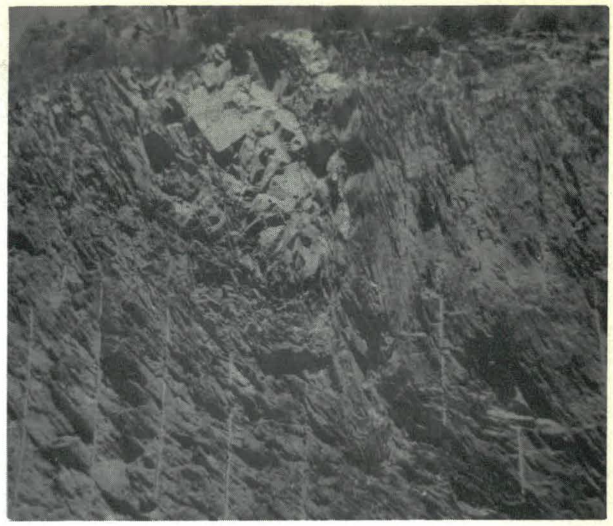
In an attempt to reconstruct the paleoslope of the shelf in Nooitgedagt times some factors were taken into consideration which will be discussed separately. Owing to the limited north-south extent of the outcrops, variations in this direction could not be determined.

BEDDING CHARACTERISTICS

Traced from west to east, there is a marked tendency in the sediments of the Nooitgedagt to develop better layering. Although the limestones are still somewhat lenticular in the east, the outcrops are more continuous with fewer breaks inbetween. The mesocyclic bedding shows a similar change, the western area featuring mostly isolated lenses of lime= stone in shale within megacycles, while the eastern limestones usually display intercalations of shale and limestone on a decimeter scale. The sandstones also form isolated, massive lenses in the west and develop internal bedding towards the east. (Plate 2 (a)).

PLATE 2.

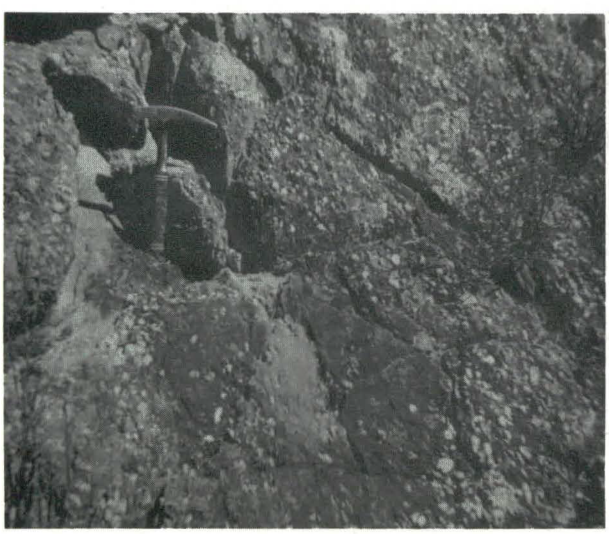
(a) Rhythmic bedding in Nooitge= dagt sandstone, De Cango. Looking east.



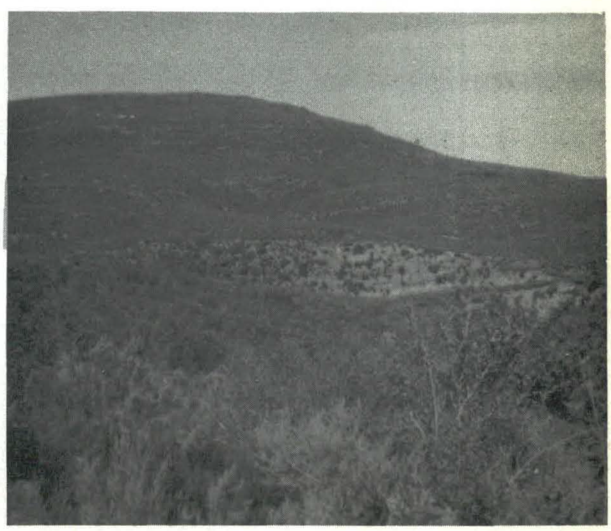
(b) Lenticular bedding in Nooit= gedagt sandstone, De Cango. Looking east.

(c) Medium-grained sandstone in Upper Member, Buffels Kraal. Looking west.

(d) Polymictic conglomerate, Vaartwell Formation. Note lineation of pebbles paral= lel to pencil.



(e) Quartz pebble conglomerate, Vaartwell Formation.



(f) Horizontal bedding in Uit= vlugt Formation, Central Plateau. Looking east.

SAND-SHALE RATIO

Fig. 19 shows that the sand-shale ratio increases towards the east. Determinations were made at 5 km intervals.

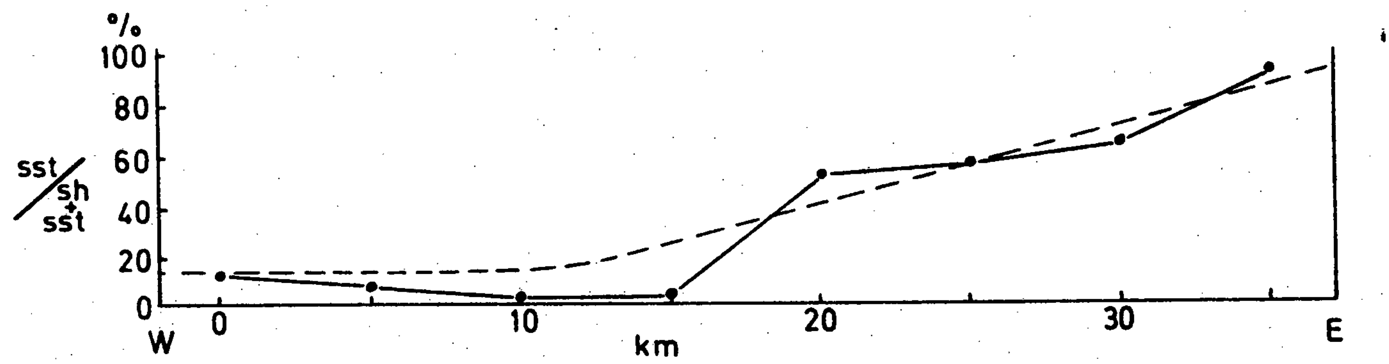


Fig. 19 . Nooitgedagt Member: Sandstone-shale ratio vs distance.

GRAIN SIZE

Thin sections of 8 samples representative of the sandstones in their specific environments were studied. The longest axes of 100-200 unaltered grains were measured using an auto= matic point counter for random selection of points. Cumu= lative curves were drawn on a log scale and the median grain sizes read from the 50% line. A decrease in grain size from west to east is shown in Fig 20.

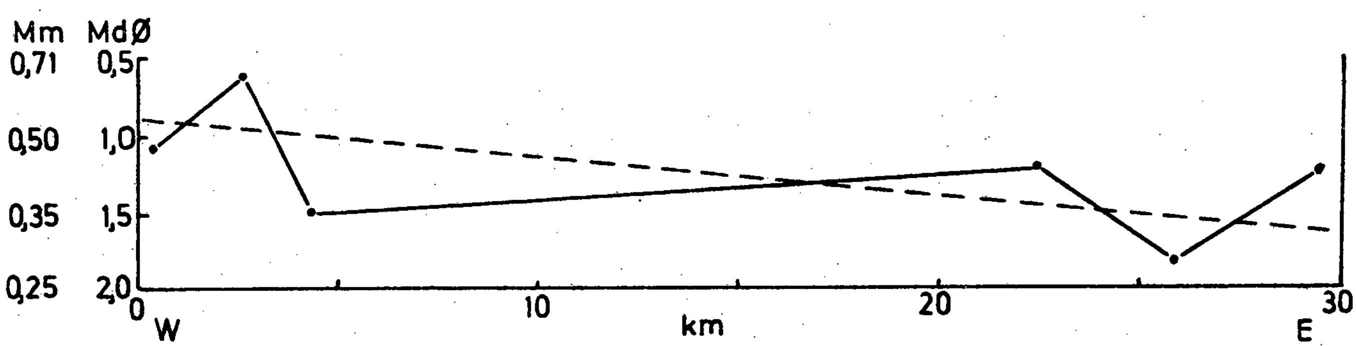


Fig. 20 .Nooitgedagt Member: Grain size vs distance.

SKEWNESS:

Using the formula $\left[\frac{\phi 84+\phi 16-2\phi 50}{2(\phi 84-\phi 16)} + \frac{\phi 95+\phi 5-2\phi 50}{2(\phi 95-\phi 5)} \right]$

phi skewness was determined for the same samples used above. The values decrease from west to east. (Fig. 21).

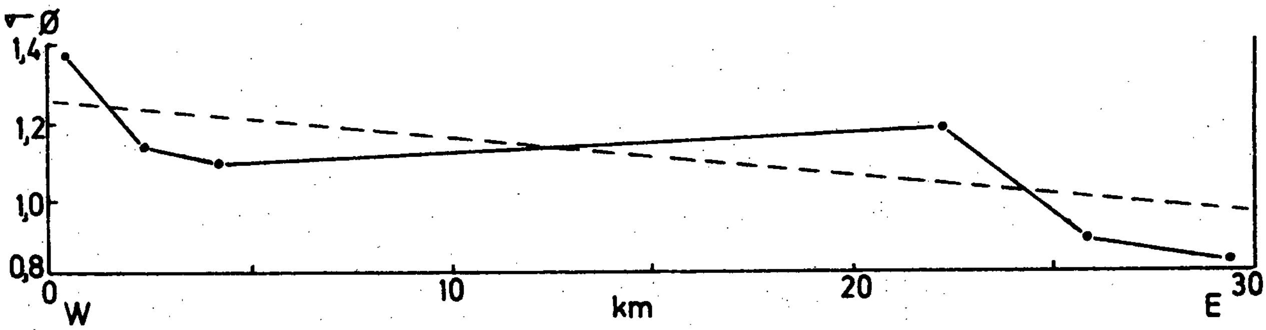


Fig. 21 . Nooitgedagt Member: Skewness vs distance

SORTING

The coefficient of sorting $\left[\frac{\phi 84-\phi 16}{4} + \frac{\phi 95-\phi 5}{6{,}6} \right]$

decreases eastwards, thus indicating improved sorting in that direction. (Fig. 22).

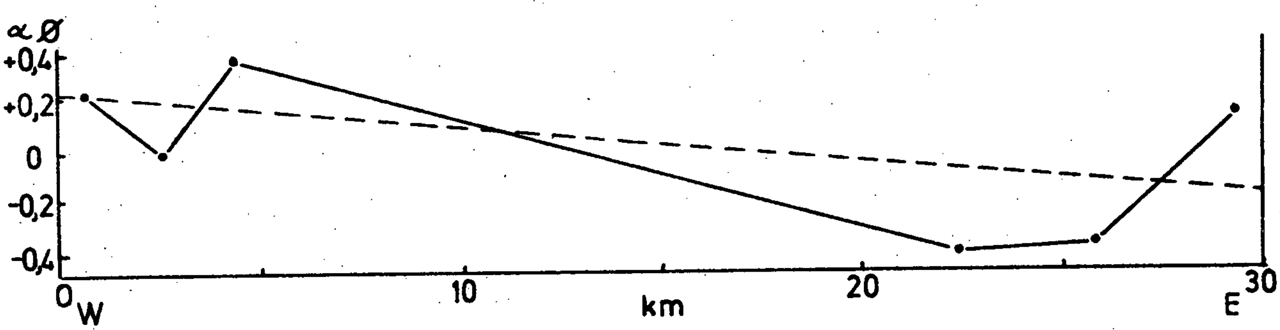


Fig.22 . Nooitgedagt Member: Sorting vs distance

MATRIX

Although the amount of matrix undoubtedly increased since deposition, a large volume of the secondary matrix is directly dependent on the abundance of argillaceous rock fragments in the original sediment (p.97). Since both the original matrix and amount of rock fragments will tend to decrease with winnowing and sorting, it can be used as an index for distance of transport. Fig. 23 indicates a possible decrease of the matrix percentage towards the east.

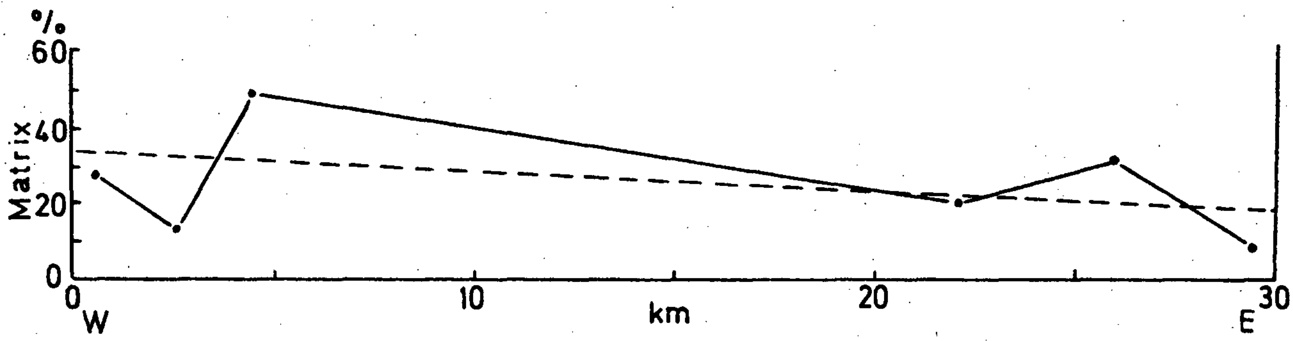


Fig.23 . Nooitgedagt Member: Matrix content vs distance

The grain-size parameters discussed above serve only to confirm features readily observed in the field, as the number of samples analysed is too small to give a true indication of their variation.

When applying these indices to determine the paleo=slope direction, it is necessary to consider the conditions in modern carbonate shelves. Tidal currents and waves are usually more active on the margins of sheltered shelves while less vigorous in protected places near the shore. Thus, there is a tendency for fine sediments to accumulate on the inner shelf, while greater water turbulence on the outer shelf generally results in good sorting of the sands. According to Milliman (1976) the carbonate content of shelf sediments also increases offshore, because the inner shelf is exposed to greater sedimentary influx and sil=tation from neighbouring rivers. The increase in the sand-shale ratio and carbonate content towards the east, and simultaneous decreases in grain size, sorting coefficient and matrix percentage of the Nooitgedagt sandstones, are all indications that the shore line was located somewhere to the west of the present outcrop boundary.

CAVE MEMBER

With the onset of the Cave deposition, some event occurred which eliminated coarse clastics from the depositional cycle. From the base of the member upwards the limestones display a marked increase in purity. While the lower parts have numerous intercalations of phyllite, shale and silty sandstone, the uppermost 600 m consists of pure, massive limestone. Simultaneously there is an upward elimination of the thin layers and lenses of dolomitic limestone which occur in the lower parts. Analyses by Iskor (1920) of samples from the lower zone show values of up to 15% for Si and 14,5% Mg, while the higher limestones have an average of less than 0,3% Si and 0,8% Mg. This is probably an indication of a steady increase in water depth with deposition, which could also mean a greater distance from the shore during another cycle of transgression.

The abrupt cessation in the deposition of coarse clastics may however have another explanation, which can be illustra= ted best in view of a modern analogue in the Great Bahama Bank (described by Newell and Rigby, 1957.) Many of the distinctive characteristics of the Bank sediments can be recognized in the upper parts of the Cave Member, among which the high purity, monomictic and non-porous nature of the limestones and the absence of crossbedding are prominent. (See E.W.Beales, 1958).

The fact that limestone, virtually free from terrigenous material is being formed at present on the submerged platforms of the Bahamas, is probably a result of the inter= vening trenches or channels between the mainland and the platforms, which present a barrier to sediments arriving from the continent. The origin of these troughs have been attributed by Hess (1933) to gradual subsidence of a structure controlled topography, accompanied by rapid deposition of calcareous deposits which maintained the higher prominences near sea level. Newell and Rigby stress the fact that the Bahamian troughs are flat-floored rather than V-shaped, suggesting that they are not simply erosional forms produced by turbidity currents as proposed by Ericson et al (1952). They attribute the existence of the troughs to sedimentary bypassing and deposition over the platforms.

The present writer is of the opinion that the trenches may have had their origin in block faulting of the graben type, because of frictional drag and local extension of the crust along the trailing edge of the drifting North American continent. The cessation of coarse detrital influx into the Matjies River shelf can thus be a result of similar faulting which isolated the shelf from the mainland. Fine silt would still be able to cross the newly formed troughs in suspension during the initial stages, but as the separation proceeded chemical precipitation would have become the main agent of deposition on the Cave platform. An example of an intermediate stage can be found in the Campeche Bank connected to the mainland of Mexico, which is only semi-isolated and has limestones with an admixture of terri= genous sediments. (Kornicker and Boyd, 1962).

During the initial stages the rate of sedimentation on the Cave platform was great enough to balance the rate of subsidence, but as the submergence proceeded deeper water conditions prevailed which started the deposition of the Groenefontein Formation.

As regards the paleoslope, the same variations in bedding characteristics can be observed in the Cave Member as described for the Nooitgedagt limestones, viz. a better developed bedding and an increase in the carbonate content from west to east.

3.2.2. GROENEFONTEIN FORMATION

3.2.2.1. THE "GREYWACKE" PROBLEM

The silty sandstones of the Groenefontein Formation have the typical dark, fine-grained appearance of greywackes and were regarded as such by Stocken and Mulder. Micro=

scopically they also display textures similar to the greywackes described in literature, viz. angular, often polycrystalline quartz grains and feldspar with hazy margins in a fine-grained matrix consisting of an intimate intergrowth of chlorite, sericite and silt-sized particles of quartz. However, the modal and chemical composition of the Groene= fontein wackes are somewhat different from the "classical" greywackes. Little feldspar was found under the microscope, in spite of chemical staining to expose untwinned varieties.

Geochemically there are also important deviations from the composition of true greywackes. According to Pettijohn (1972), the bulk composition of the latter is surprisingly homogeneous and much alike the world over, being rich in Al_2O_3, FeO + Fe_2O_3, MgO and Na_2O. They differ from arkoses in the dominance of Na_2O over K_2O and (usually) MgO over CaO. Figs. 24 (a) and (b) show that the Groenefontein wackes in this respect more resemble arkoses than grey= wackes, although they are similar to the latter in SiO_2, TiO_2, Al_2O_3, MnO, total Fe and total H_2O. (Table 5.) However, the chemical composition of the rocks is in the first place a reflection of the provenance, and indirectly also of the tectonic setting of the latter. Moreover, the albitic nature of many greywacke plagioclases may not be a primary feature, but could be the result of albitization in situ (Pettijohn, 1972), while the absence of K-feldspar could be due to post-depositional solution (Gluskoter, 1964). The chemical inconsistency of the Groenefontein wackes with classical greywackes does therefore not necessarily mean that a different type of sediment or depositional setting should be involved. In fact, the lithology and sedimentary structures of the Groenefontein Formation are very characteristic of greywacke suites all over the world, regardless of geologic age. Although they may not be greywackes sensu stricto, these quartz wackes almost certainly belong to the same clan and should be considered in this context.

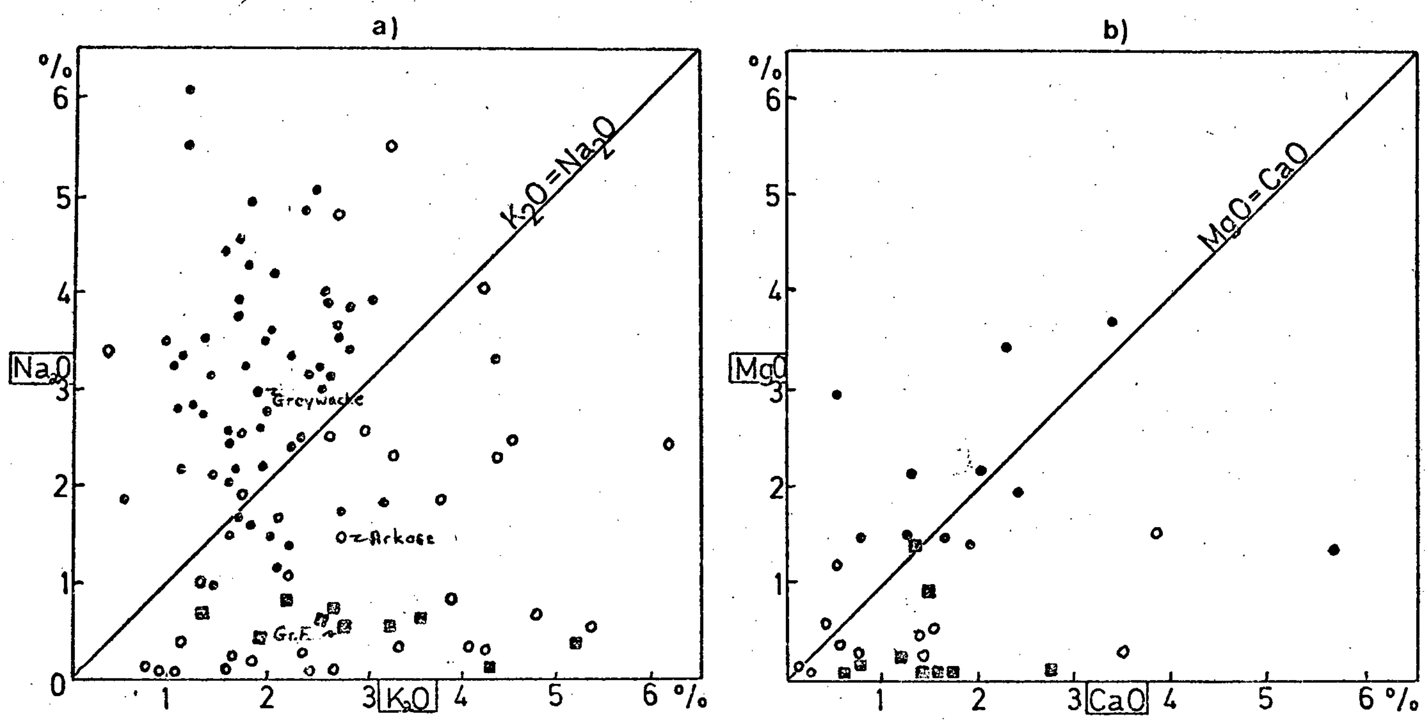


Fig. 24. (a) Groenefontein Formation: Na2: K_2O ratio
(b) Groenefontein Formation: MgO: CaO ratio

TABLE 5

Average composition of represen tative	SiO_2	TiO_2	Al_2O_3	Tot Fe	MnO	MgO	CaO	Na_2O	K_2O	P_2O_5	Tot H_2O
Quartz Arenite	95.4	0.2	1.1	0.6	-	0.1	1.6	0.1	0.2	-	0.3
Lithic Arenite	66.1	0.3.	8.1	5.2	0.1	2.4	6.2	0.9	1.3	0.1	4.3
Arkose	77.1	0.3	8.7	2.2	0.2	0.5	2.7	1.5	2.8	0.1	0.9
Grey wacke	66.7	0.6	13.5	5.1	0.1	2.1	2.5	2.9	2.0	0.2	3.0
Groene- fontein F.	71.63	0.65	12.96	5.06	0.11	0.28	1.09	0.59	2.79	0.07	2.67

(Data from Pettijohn 1963, p. 15).

3.2.2.2. EVIDENCE FOR A DEEP WATER, TURBIDITE ORIGIN

Most greywackes of the geologic record are turbidites representing a special depositional environment, and the Groenefontein wackes display several features which are typical of this type of environment.

LITHOLOGY

Of the characteristic features of deep sea turbidites listed by Kuenen (1964, pp. 9-13,16), the uniform and rhythmic bedding with lack of markers is most conspicuous in the Middle Member. Except for the lenses of grit and medium to coarse grained sandstone in the Upper Member, most beds in the Groenefontein Formation contain nothing beyond the size of fine silt. Sorting is moderate to moderately well (See table XI, Appendix) and improves with decrease in grain size. Flakes of mica are very conspicuous in hand specimens, quartz grains are mostly angular and there is a large percentage of matrix. Pyrite of early diagenetic origin (now pseudomorphs of limonite) are common, especially in the Upper Member where cubes up to 2 cm in diameter are found. These usually indicate strongly reducing conditions.

Deep water is not only suggested by mm scale laminations in the shales, but also by the total absence of all elements indicative of shallow water. There are no winnowed sands, large-scale crossbedding, wedging, lenticular beds or scour-and-fill channels. Plots of the grain size parameters of 8 representative samples (Figs. 25 (b) and (c)), indicate that the sandstones can be either pelagic sediments or turbidite deposits, although the skewness values of the former are usually somewhat higher. In Fig. 25 (a) the plots lie somewhere between those for deposition from tranquil water, under wave action or by unidirectional currents.

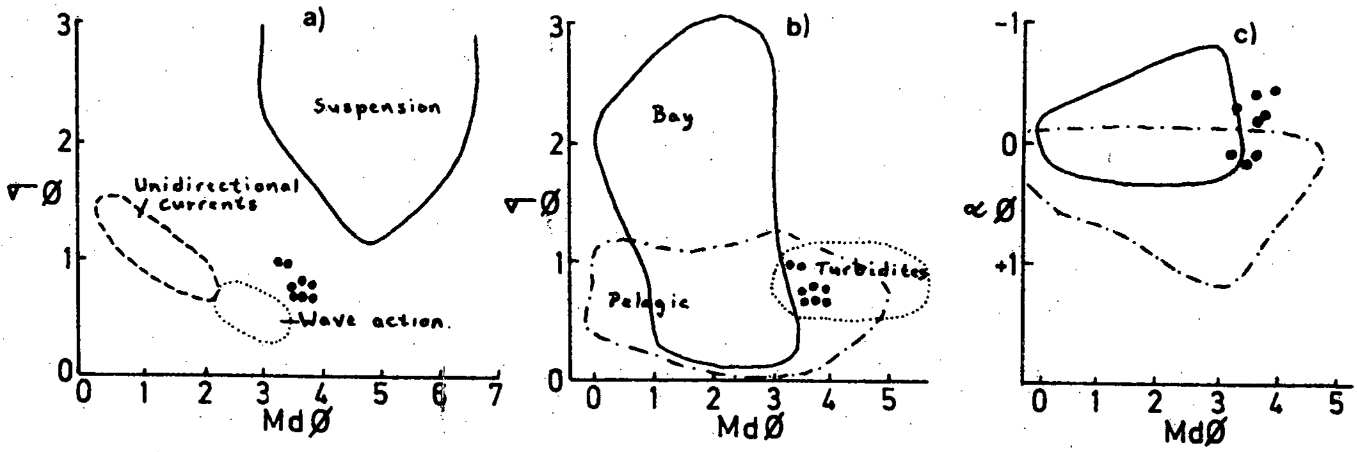


Fig.25. (a) Groenefontein Formation: Sorting vs grain size (After Stewart, 1958).
(b) Groenefontein Formation: Sorting vs grain size (Data from Kukal, 1971).
(c) Groenefontein Formation: Skewness vs grain size (Data from Kukal, 1971).

SEDIMENTARY STRUCTURES

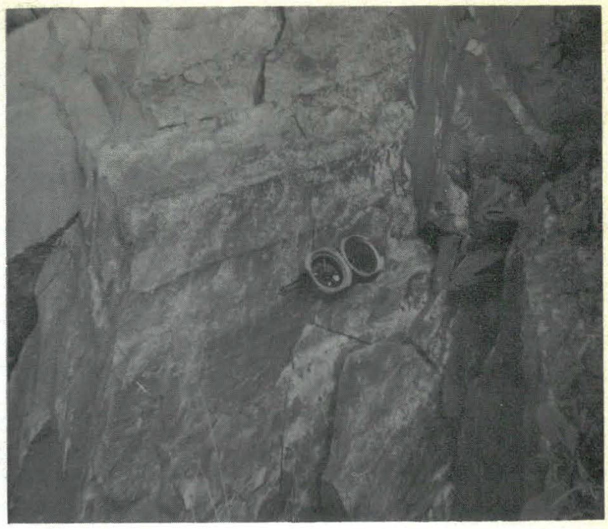
Probably as a result of the intense deformation suffered by this incompetent formation, bedding surfaces are seldom preserved or sufficiently exposed for studies of primary sedimentary structures. Most observations were limited to the fresh road cuttings made during the construction of the Huis River Pass (outside the study area).

Flute casts exposed in these new outcrops are of two types: medium sized linguoid or bulbous varieties, and giant deltoid casts. The former have dimensions of a few to 60 cm in length and are up to 25 cm in width. (Fig. 8 (a)). They display good bilateral symmetry in most cases. The deltoid casts are up to 1 m long and 50 cm wide. (Plate 3 (a)). Other types of sole markings encountered are groove casts some 3 m in length, and very rare, asymmetric ripple marks. (Plate 3 (b)). These have amplitudes of 2,5 cm and wave lengths of 10 cm. Cm-scale crosslaminations in their crests distinguish them from tectonic ripples.

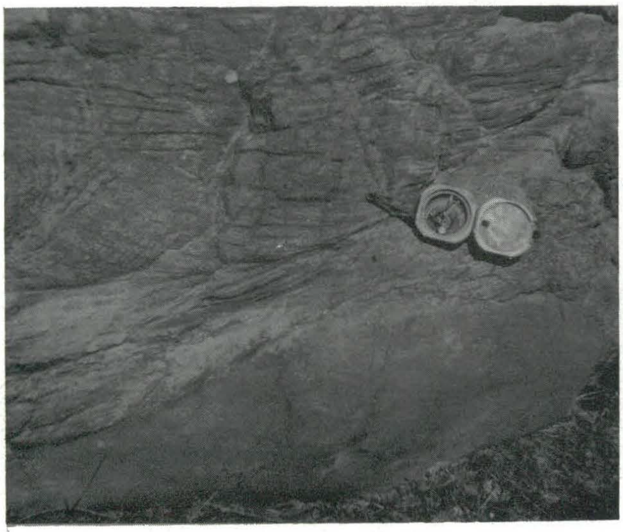
Graded bedding is not often conspicuous in the wackes because

PLATE 3.

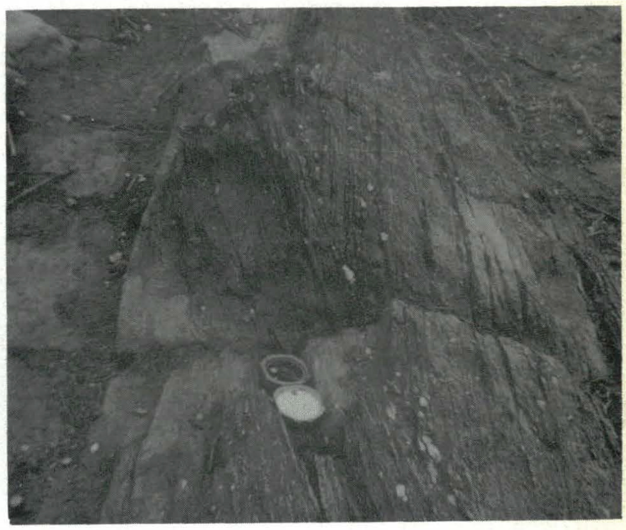
(a) Giant flute casts in Groene= fontein Formation, Huis River Pass.



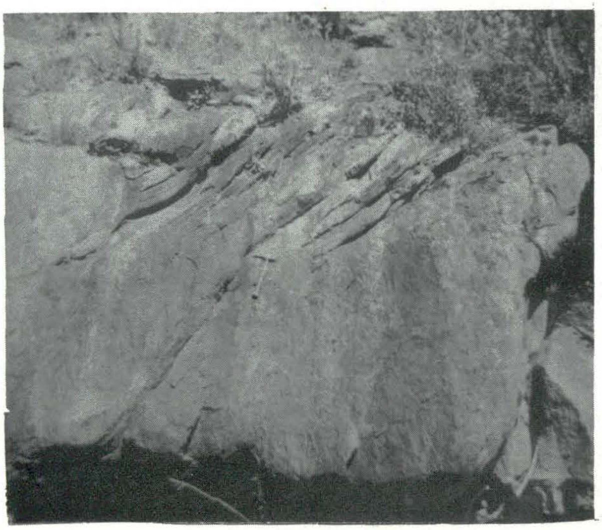
(b) Ripple marks in Groenefon= tein Formation, Huis River Pass.



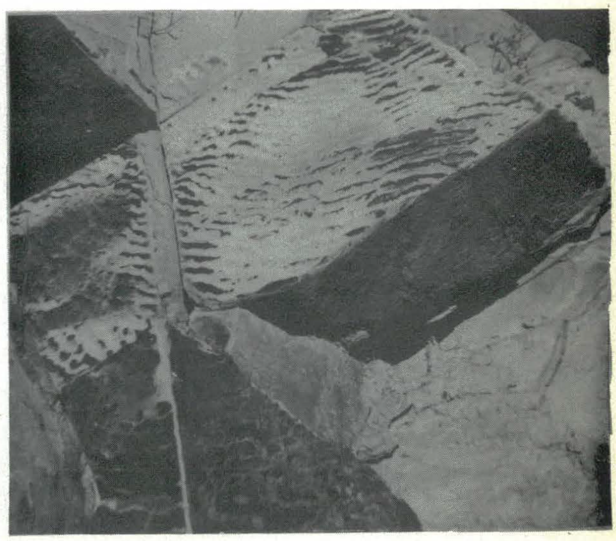
(c) Festoon crossbedding, Uit= vlugt Formation.



(d) Tabular crossbedding, Uit= vlugt Formation.



(e) Part of large festoon-trough, Uitvlugt Formation. Looking west.



(f) Ripple marks in Uitvlugt Formation, Schoemans Poort.

of their fine grain size, but its presence is revealed by a slight change in colour from grey at the base to dark green at the top of every cycle. A complete transition from silty sandstone to shale occurs, the former usually constituting 80% of every unit and grading to shale only in the uppermost 1 to 20 cm. Each unit has sharp contacts with its neighbours. Although rhythmically interbedded, the cycles have no fixed width; a succession of thick units (1 to 5 m) can for example be followed by sandstone - shale intercalations of 3 to 30 cm.

Convolute lamination (Fig. 26) was found in loose boulders on Buffelsvalley in the extreme east (22°39′E), but nowhere else. According to Kuenen (1964) and Pettijohn (1972) it is relatively common in turbidite sequences, possibly being a result of current movement deforming transverse ripple marks. (Kuenen, 1953). Holland (1959) however suggested that convolute deformation arises in response to irregular distribution of pressure and suction in the turbidity current, a view which this writer prefers.

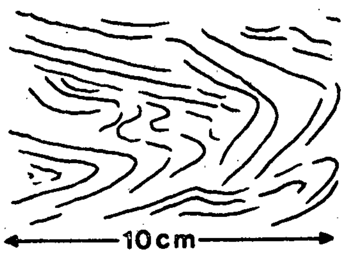


Fig. 26. Convolute lamination in Groenefontein quartz wacke, Buffelsvalley.

3.2.2.3. PALEOGEOGRAPHY

LOWER MEMBER

The occurrence of limestone lenses in a transitional zone at the base of turbidite deposits can be explained in the following way: Continued submergence of the Cave platform eventually surpassed the rate of carbonate build-up and resulted in transgression across the shelf area. The deepening water is not only reflected in the general absence of oölites in the limestone lenses at the base of the formation, but also by the lack of dolomitic intercalations. Fine silt subsequently poured into the subsiding basin which either "drowned" algae responsible for limestone deposition, or "diluted" carbonate precipitating inorganically. Locally conditions still allowed limestone lenses to form, but as the amount of influx increased and the water deepened, limestone disappeared from the sedimentary record. Part of the calcite probably went into solution because of increasing pressure and colder water temperatures away from the platform.

The Lower Member with its associated limestone lenses is therefore regarded as a "downslope" or intermediate deposit between the platform and basin facies. The rhythmic limestone=shale intercalations in some lenses and the absence of grading, is in contrast with limestone turbidites, which are more commonly found on relatively steep slopes or in the more distal parts of basins. The absence of slump or scour structures and olistoliths of limestone also suggest that slopes in this case were very gentle, such as are typical of the aprons peripheral to many carbonate shelves.

Lithologically the limestones display some of the character=istics of offshelf deposits mentioned by Wilson (1969), among which the dark, evenly planar beds of limestone with occasional mm-scale lamination is very prominent, but small scale grading or cross-lamination and major discontinuities were not observed. However, as Sloss (1969) pointed out, many of the features commonly attributed to specific environments of limestone deposition are not necessarily depth dependent, and interpretations should be in harmony with the pattern of the depositional history. The association of the Lower Member limestones with sediments of the greywacke suite above and platform facies below, is clearly an indication of a transitional environment between the two. Dzulynski and others (1959) have referred to this type of sediment as "fluxoturbidites".

MIDDLE MEMBER

The turbidites of the Middle Member were probably poured into a marginal cratonic basin of bathial depth. Tectonic slope steepening, storms or earthquakes could have acted to trigger slides down submarine canyons cutting into the basin slope, shelf or other source areas. The presence of carbonate flakes in some samples might suggest that some of the turbidity currents originated on platform areas where carbonates were present. However, calcite which replaces both matrix and detrital grains in some samples could be a post-depositional product, possibly even a result of the albitization of feldspar.

Although each turbidite cycle is topped by a shaly part, the absence of fossils makes their recognition as pelagic strata very difficult. These clayey horizons may be part of the "tails" of turbidity currents, or could have accumulated during the quiet periods following the rapid deposition of silt by the currents. It is also possible that true pelagic strata are not present at all, due to a too short time interval between currents, or as a result of having been swept away by subsequent turbidity slides.

UPPER MEMBER

The renewed appearance of limestone lenses in the Upper Member, together with grit and coarser sandstone bands, suggests a return to the conditions which prevailed during deposition of the Lower Member. The influx of the shallow water Vaartwell conglomerates immediately hereafter marks another major cycle of uplift and regression, the initial stage of which is probably represented by the Upper Member. The coarse sandstone lenses are possibly related to the faults and earth-shocks accompanying the initial uplift, as a preliminary event to the major tectonism later on. Microscopically they can be classified as microbreccias, while their grain size parameters (two representative samples) plot in the fields for bay or river sediments. (Fig. 27 (a) and (b)). It is therefore suggested that these lenses are relics of submarine river and traction channels or slump deposits, originated by tectonic movement which triggered the flow of rapidly accumulating sediments away from shore into shallow depressions or "traps".

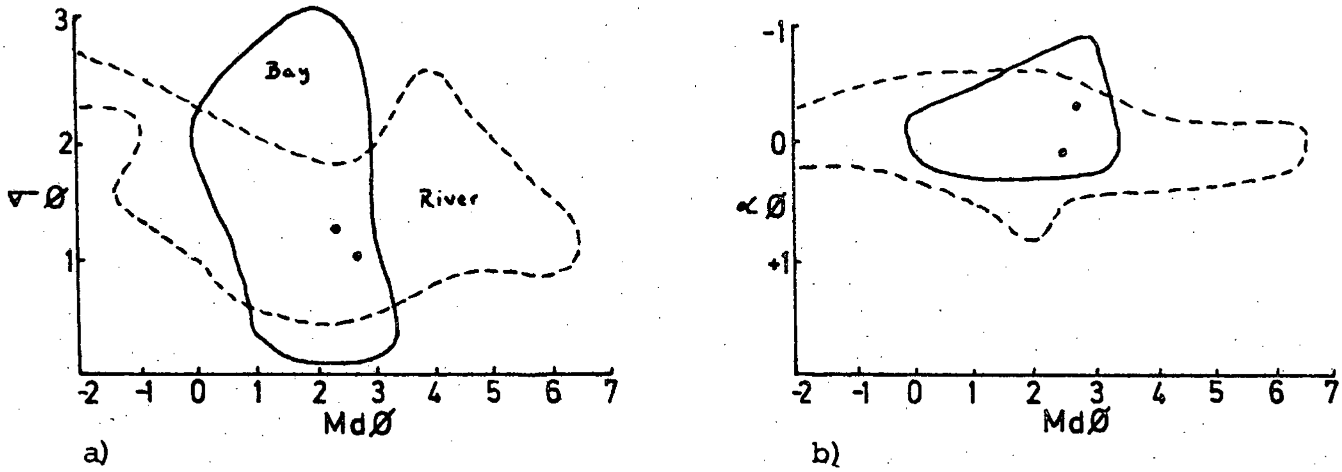


Fig. 27 Upper Member: (a) Sorting vs grain size
(b) Skewness vs grain size

3.2.2.4. PALEOSLOPE

LITHOFACIES

The distribution of limestone and grit lenses in the Lower and Upper Members is possibly related to water depth and distance from the source area, but the true nature of this interdependence can only be guessed at. In the Lower Member, limestone lenses are more abundant east of 22°7′ and also smaller west of this longitude, thus seeming to conform to the earlier depositional pattern of the Matjies River limestones. The reappearance of limestones in the Upper Member together with coarse sandstone lenses shows a similar distribution. Abundant in the southeastern and central parts of the Cango, they disappear towards the west beyond the Calitzdorp dam area. Laminations on a cm scale which is typical of deeper water, together with very rare grading, was observed more often in the western lenses, while those in the east appear more massive and mostly lack internal struc= tures. To the north beneath the overturned Uitvlugt and Vaartwell Formations the limestone is also less common than in the south. Grit and sandstone bands are closely correlated with the distribution of limestone lenses, being more common in the east and southeast than in the west and north. They also possess better bedding in the southeast, especially on the farm Buffels Kraal and east through Meirings Poort.

The resemblance between the distribution of limestones in the Lower Member and those of the Matjies River Formation can similarly be explained by proximity to the source area in the west, with accordingly greater dominance of terri= genous matter, but the lithofacies of the Upper Member presents serious problems. Coarse detrital lenses would certainly be expected to occur closer to the provenance, which in this case may thus have been situated in the east or southeast. If so, the greater abundance of limestone lenses in this area must reflect shallower water conditions. Another possibility is that the Upper Member limestones and grits have been removed by erosion because of uplift above sea level in the west and north. As previously mentioned, contact relationships between the Groenefontein and Vaartwell Formations are mostly obscure and moreover not fully investigated by this writer outside the study area. Also, the complex folding in the Groenefontein Formation makes any determination of positions in the stratigraphic sequence very difficult at any distance from the upper or lower contacts.

In the Middle Member there is a marked increase in the abundance of shale and arenaceous shale towards the west, although this proved impractical to map. It could however indicate deep water in this area, representing "shaly flysch" basinward of the "normal flysch" of Vassoevic (1957) (Fig. 28).

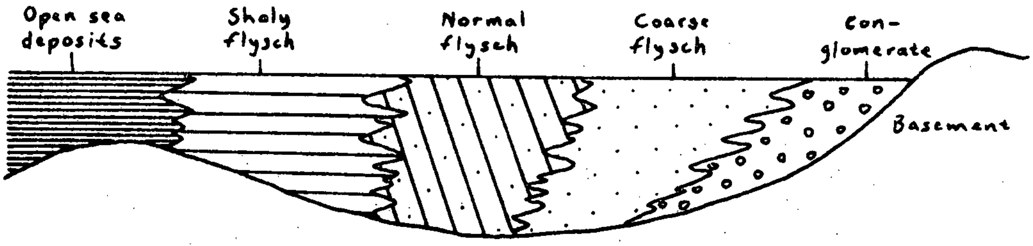


Fig 28. Facies distribution in flysch basins (After Vassoevic, 1957)

Gradation and sole markings are also more obvious in the Huis River Pass than eastwards, although this may be due to poorer exposures elsewhere. Kuenen (1964) reports thinner beds and improved gradation away from the source area, but according to Pettijohn (1972) thick, graded beds with large sole markings can be expected near the provenance, which grade distally into thin, weakly graded siltstones with few sole markings.

SOLE MARKINGS

The direction of transport as determined by stereogram are 200° for the giant deltoid casts and 337° for the lingoid flute casts. The ripple marks occurring near the latter give a direction towards 322°. Turbidity currents at these localities thus could have flowed in a north-westerly to south-south-westerly direction, but neither the attitudes of fold axes nor the effect of plastic deformation were taken into account during reconstruction of the original bedding-orientation.

GRAIN SIZE

Turbidites usually become finer grained and better sorted away from the source area (Kuenen 1964). From east to west the impression is gained in the Middle Member that the overall grain size decreases, which is borne out by MdØ values of 8 representative samples of quartz wacke shown in Fig. 29. (See also Pettijohn 1972, p. 79).

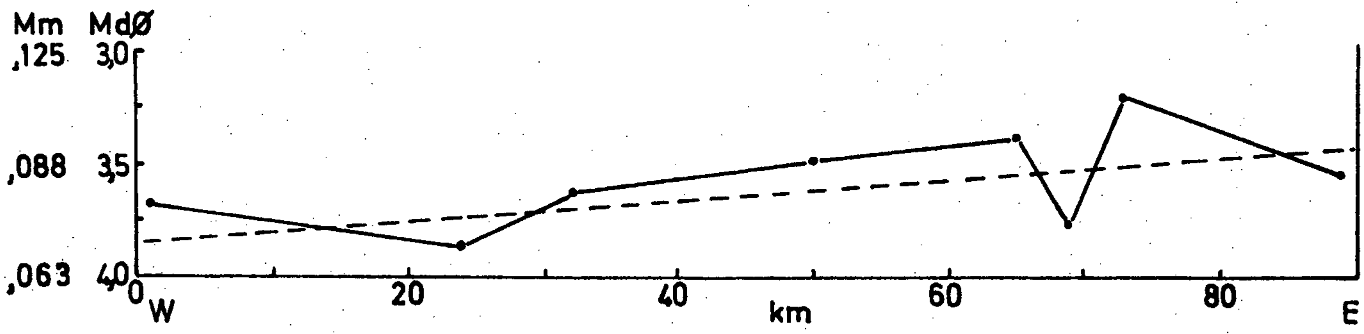


Fig. 29. Groenefontein Middle Member: Grain size vs distance.

SORTING

The degree of sorting varies inversely with the median grain size but does not show a clear trend, although two higher σ-Ø values are recorded in the east. (Fig. 30).

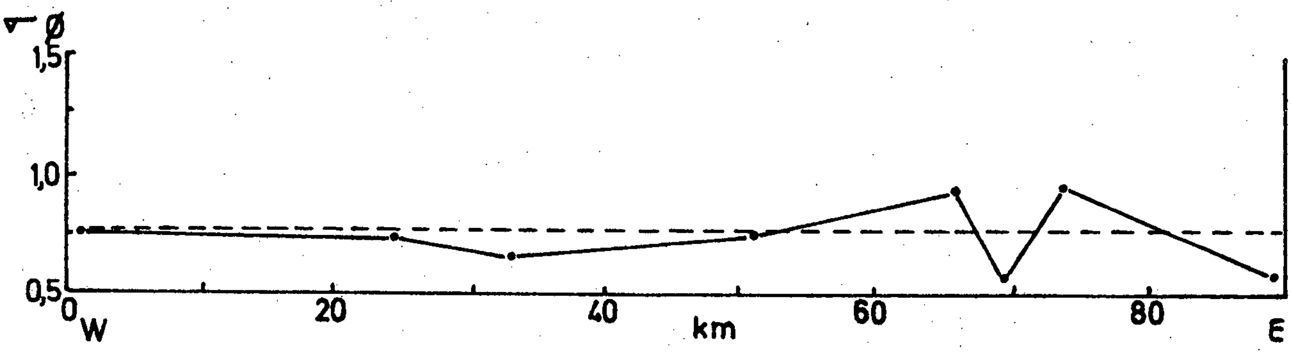


Fig. 30. Groenefontein Middle Member: Sorting vs distance.

SKEWNESS

The values of αØ vary from negative in the west to positive in the east, which means an increase in the fine fraction of every sample in that direction. (Fig. 31.).

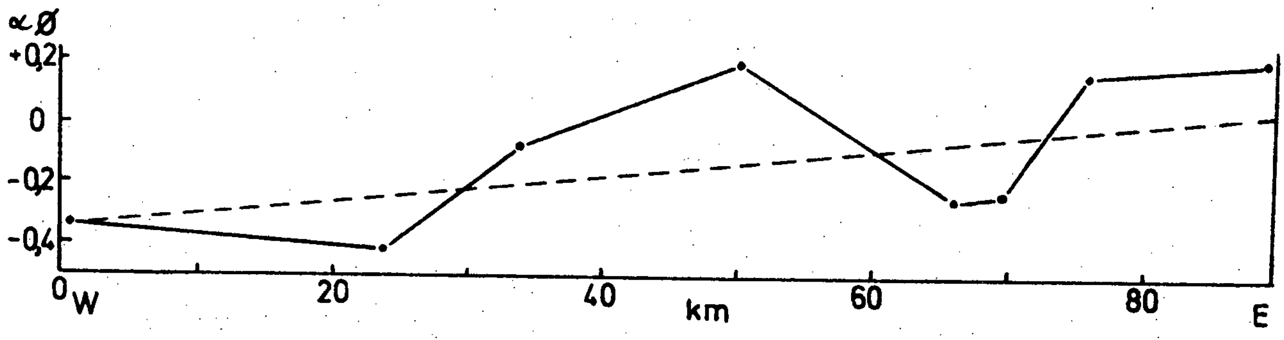


Fig. 31. Groenefontein Middle Member: Skewness vs distance.

In river sediments ∝Ø usually decreases with distance of transport, but whether this applies to turbidites also, is uncertain. It is interesting to note that the value of ∝Ø varies inversely with the amount of matrix (Fig. 32.), which is in agreement with data on greywackes published by Kelling (1962) and Mizutani (1957). (See Dzulynski and Walton,1965, Fig. 2 c)

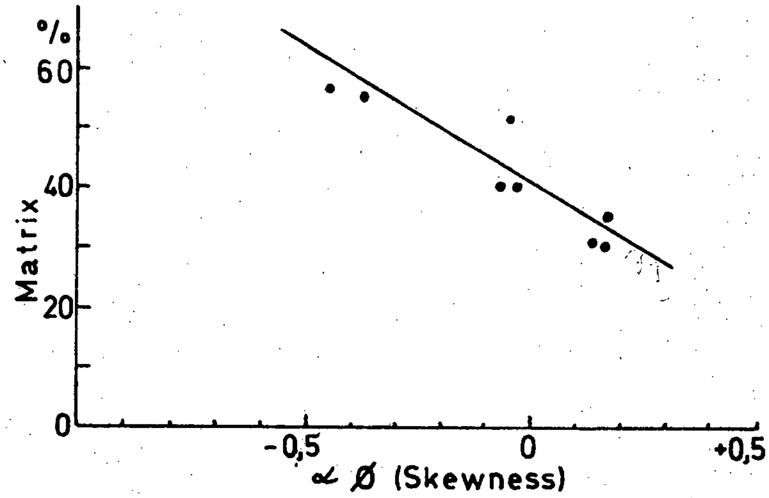


Fig. 32. Groenefontein Middle Member: Skewness vs matrix content.

MATRIX (Fig. 33).

Contrary to river sediments, where fractional settling from suspension is less important, one would expect the very fine detritals and flat mica flakes to stay waterborne for longer distances in a turbidity current, thus increasing the percentage of primary matrix away from the provenance. Unfortunately the original amount of matrix is not known in this case, but assuming that diagenesis had more or less the same effect on all samples, the increase in matrix percentage towards the west could indicate longer distances from the source area. As abrasion is of little importance in turbidity currents, the amount of argillaceous rock fragments should not vary appreciably with distance.

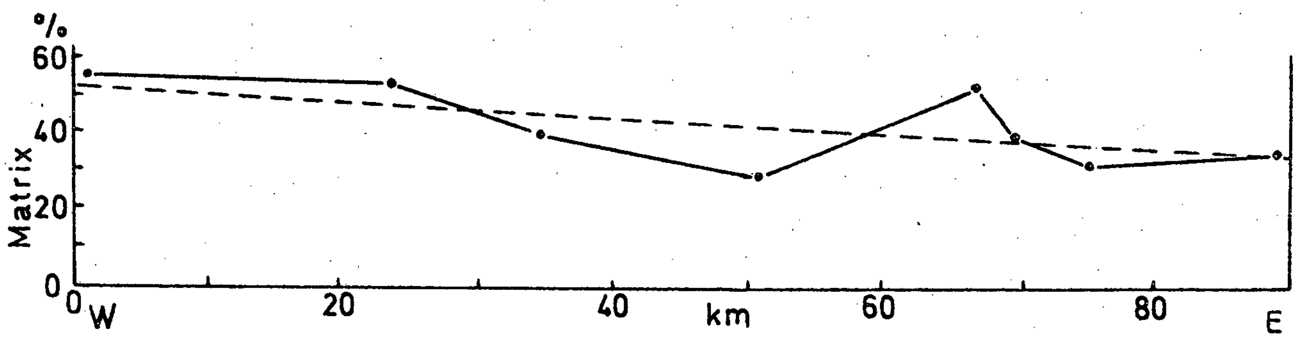


Fig. 33. Groenefontein Middle Member: Matrix content vs distance.

P_2O_5 CONTENT

According to Kukal (1971), phosphorus usually increases towards the marine environment. The westward decrease in P_2O_5 shown by Fig. 34 could be the result of an increase in the distance from the connection to the open sea (Fig. 64), but this is doubtful. The trend obtained may in fact reflect the abundance of the heavy mineral apatite, which would decrease away from its source of granitic rocks because of settling from suspension.

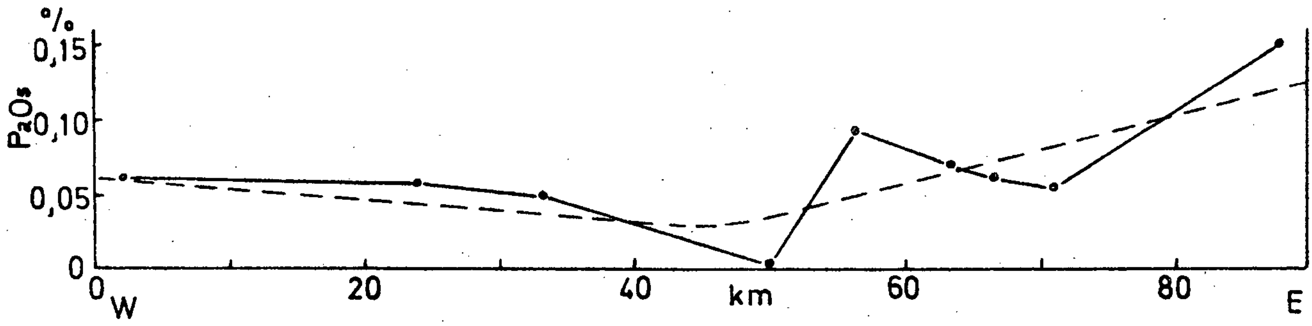


Fig. 34. Groenefontein Middle Member: P_2O_5 content vs distance.

In conclusion it may be said that, although the evidence outlined above is not clear-cut and certainly controversial, the presence of entry points or " sources" towards the east or southeast should not be discarded entirely. In fact, there is very little indication of east-moving currents at all, except maybe for the Lower Member which could have been derived from the subsiding shelf roughly to the west. Although longitudinal transport probably played a dominant role, it is likely that the main contributions came from the flanks of the basin to the north or south. It is therefore suggested that the turbidites were derived from these slopes and flowed west along the trough axis in the manner illustrated by Fig.35. (after Cummins, 1957).

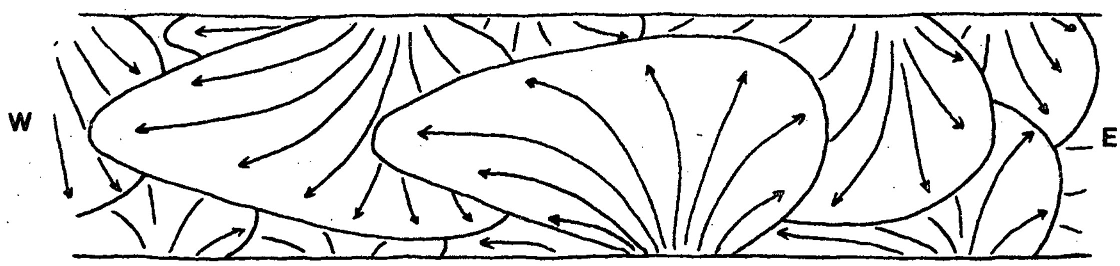


Fig. 35. Turbidity flow down trough axis (After Cummins, 1957).

3.2.3. VAARTWELL FORMATION

Because of the coarseness of the sediments, thin sections or geochemical studies of the Vaartwell conglomerates are of limited value in environmental analyses. The latter was therefore based primarily on broad lithological characteristics. The more relevant aspects of these will consequently be described in more detail, but should be considered in conjunc=tion with other features already outlined in the previous chapter.

3.2.3.1. PALEOGEOGRAPHY

THE POLYMICTIC MEMBER

The polymictic conglomerate, a mixture of heterogeneous pebble types set in an argillaceous or sandy matrix which commonly constitutes more than 50% of the rock, is a poorly stratified horizon widely varying in thickness. Although it may comprise the entire width of the Vaartwell at places, this conglomerate is usually confined to its base or has local, restricted distributions within the formation. This irregular distribution, together with the lenticular cross sections and poor stratification of the member, is strongly reminiscent of fluviatile conditions. Strong currents, such as would not be expected in marine environments, must have been responsible for the transportation of large boulders (some with diameters of 1000 to 1500 mm) which are not infrequently encountered. According to data furnished by Hjulström (1935) and Nevin (1946), the transportation of such boulders would require current velocities of at least 1000 cm/sec. Tidal currents in nearshore marine environments have velocities usually in the range of 150-160 cm/sec., which would make them capable of transporting only grains up to about 120 mm in diameter. Deposition therefore probably occurred near the upper courses of the transporting rivers.

The chief constituents of the conglomerate are metastable rocks representing all the lithologic types crossed by the deliver=ing streams. These inclusions were derived from the Upper Member of the Groenefontein Formation and are moderately rounded, poorly sorted (average σ_{ϕ} 1,17) and, as mentioned, often very large in spite of their low resistance to abrasion. The more stable exotic pebbles of quartz, chert and granite are well rounded in contrast and never very large. The average median size of all pebble types (excluding the "matrix") is 10 cm and the range 0,7 to 60 cm, but up to 7Ø size classes can be distinguished on the Wentworth scale (Figs. 36, 37 (c)). Many of the histograms show a bimodal distribution of size ranges (Fig 37 (a) and (b)).

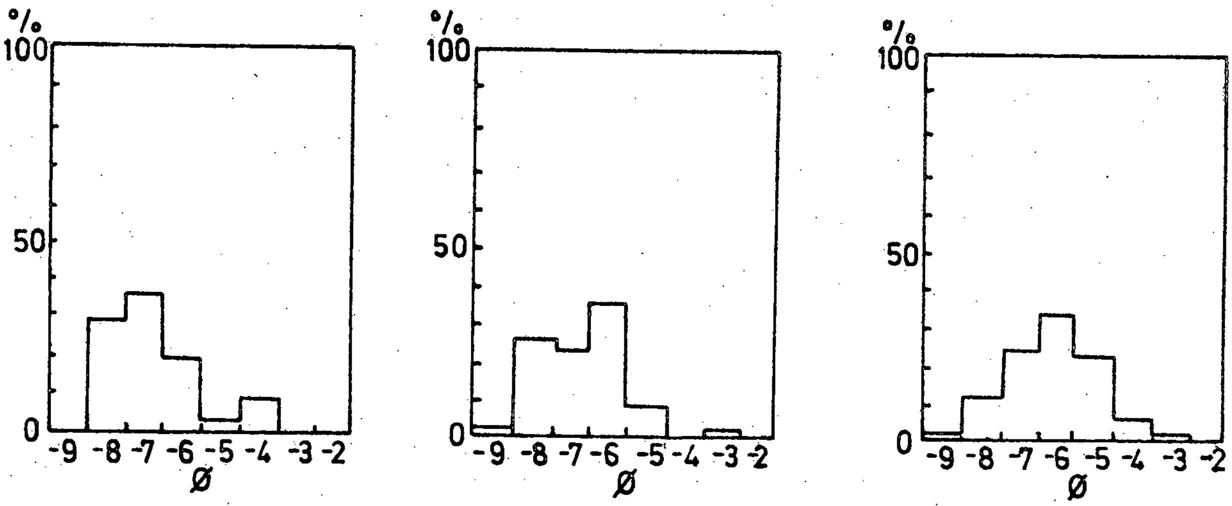


Fig. 37. Polymictic conglomerate: Histograms of pebble size. (a) and (b): selected stations; (c) synoptic (all stations).

These compositional and textural features are very similar to those of the "greywacke conglomerate" of Krumbein and Sloss (1963), which usually occurs in thick, wedgelike deposits and is typical of rapidly subsiding depositional areas. These con= glomerates appear to be mainly fluvial and represent a pouring of coarse clastics into geosynclines from rapidly elevated, adjacent highlands. Apart from stream channels, deposition also occurs in alluvial fans or piedmonts at the margins of basins where the high gradients of mountain streams are abruptly checked. As mentioned previously, the presence of derivatives from the underlying formation indicates that the conformable, gradational contacts observed in the study area transgress laterally into unconformable relationships. This in turn suggests that uplift in the source area was extensive enough to raise at least part of the Groenefontein sediments above sea level so that erosion set in. A high relief, rigorous climate and rapid rate of erosion are also suggested by the presence of limestone and granite cobbles in the member, which, if not for these special conditions, would have disintegrated or dissolved very quickly. (See Pettijohn 1957, p. 252).

The polymictic member is therefore regarded as a fluviatile or fanglomerate deposit having accumulated in stream channels and alluvial fans along the apron of the elevated source area. It thus marks the transgression from marine to continental environments in the Cango Group.

THE MONOMICTIC MEMBER

This conglomerate consists of 90 to 95% pebbles of well rounded vein quartz together with minor amounts of quartzite, black chert, silty sandstone and shale. The pebbles have a median diameter of 6 cm in a range of 2-20 cm, and moderate sorting (0,83). Only 4 size classes can be distinguished, excluding the matrix which has possibly been introduced after deposition. The histograms show unimodal size distributions (Fig. 38).

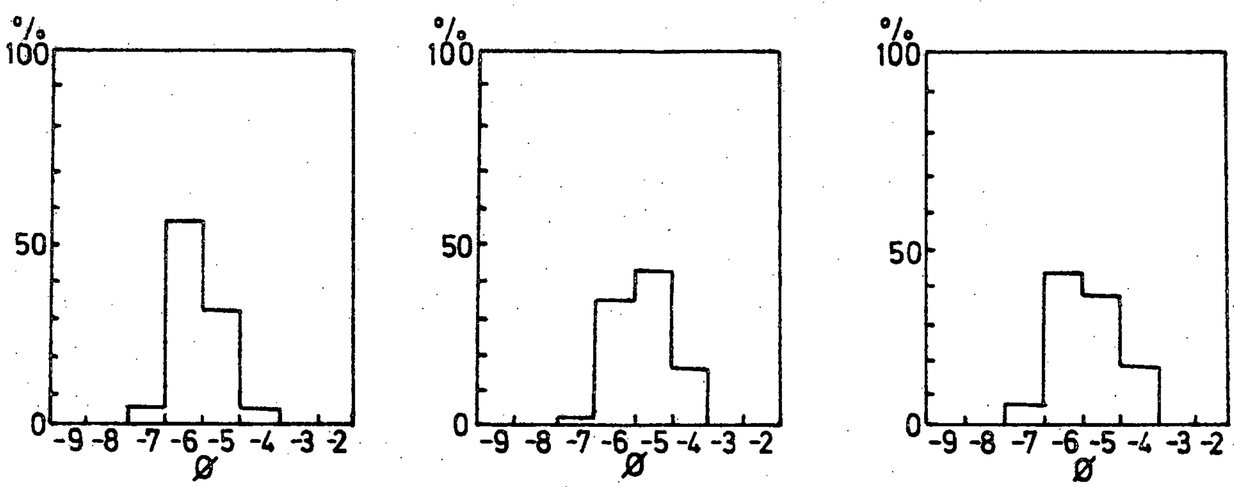


Fig. 38. Quartz pebble conglomerate: Histograms of pebble size. (a) and (b): selected stations; (c) synoptic (all stations)

The matrix consists of grit and sandstone composed predominantly of quartz with minor amounts of feldspar and mica. Cross=bedding is found higher up in the succession where the lenses, merging into one another at the base, become increasingly scattered and grade into the Uitvlugt Formation. Stratifi=cation is coarse and often only recognizable at a distance as a result of differences in weathering between alternating pebble - rich and pebble-poor beds.

The outcrop which forms the Central Divide north of the Wildehonde Kloof Fault is typically a thick, blanket-type deposit, but elsewhere, notably towards the east and north, it is often very thin or even absent.

According to Krumbein and Sloss (1963) and Pettijohn (1957) this type of conglomerate is commonly representative of the basal beach deposits of marine transgression over a surface of low relief. However, according to Kukal (1971), beach gravels usually have a thickness restricted to several metres, except in cases where the beach represents the extension of alluvial cones washed down from the nearby mountains directly to the sea. Such deposits are more poorly sorted and can extend to depths of a few hundred metres. This picture fits the environment visualized for the polymictic member, and could explain the lateral grading of the two types of conglomerate over short distances into each other. The lack of unstable pebbles in the monomictic member can thus be attributed to wave action which winnowed and sorted the mixed assortment brought into the area by streams, thereby eliminating the softer rock types.

The upward gradation from polymictic to quartz pebble conglomerate can be traced back to two causes. In the first place, if the latter represents beach deposits, its wide distribution must be the result of slow transgression over areas previously occupied by polymictic conglomerate, implying a significant lowering or downcarving of the previously elevated highlands.

The second cause can be attributed to streams, initially arising on the uplifted slopes of the Groenefontein Formation near the depositional area, transporting at first only derivatives from the latter and later, with progressive headward erosion, cutting into the metamorphic terrain farther back and acquiring a load of stable vein quartz pebbles which became rounded and sorted with the increased distance of transport.

In, summary, the Vaartwell depositional environment is visualized as a narrow coastline peripheral to an elevated, hilly country from which short streams and alluvial fans emerged, the latter reaching down to the coast at places. Down-carving of the highlands resulted in transgression of the ocean with quartz pebble conglomerate deposited along the prograding shore.

3.2.3.3. PALEOSLOPE

ISOPACH VARIATIONS

Thickness determinations of the Vaartwell are limited to the overturned limbs of anticlines, these being the only places where both basal and upper contacts are exposed. In the area west of Coetzees Poort the formation probably reaches its greatest width, values of 700 to 1300 m being common, but only a few km to the east the conglomerate, except for scattered pebbles in the Uitvlugt Formation, is absent. This highly erratic distribution can be the result of either, as mentioned above, the accumulation of conglomerates in alluvial fans or deltas from several entry points, or a very uneven rate of subsidence with deposition concentrated in depressions or troughs. On the Central Divide a minimum thickness of 800 to more than 1000 m is reached in the monomictic member, the basal contact in this case being a normal, high-angled fault. In the south between Schoemans Poort and Koetzers Kraal the thickness varies between 60 and 300 m. East of Schoemans Poort there is another increase to 1200 m which wedges out at a rate of about 100 m/km to 150 m at the Brakke River (22°24'E). The area to the north and east of Nels River is characterized by intermittently developed conglomerate bands 50 to 250 m thick. Generally speaking therefore, there is a decrease in thickness from west to east with a simultaneous decrease in lateral persistence. As fanglomerates are characterized by rapid down-dip thinning, the uplands were probably located roughly to the west, although the local thickening in the central and eastern sections could indicate other entry points from the flanks of the trough.

DISTRIBUTION OF PEBBLE TYPES

Although lateral grading commonly occurs between the conglomerate types, the distribution of pebble types in general seems to follow a clearly recognizable pattern. In the southern outcrops, which are mainly polymictic, there is an increase eastwards of stable quartz and black chert with a simultaneous decrease in silty wackes and shale derived from the Groene=fontein Formation. Coarser sandstone and "quartzite" maintain a steady pattern. This outcrop when traced eastwards swings towards the north in the vicinity of Schoemans Poort, so that Fig. 39 in fact represents an easterly to north=easterly trend.

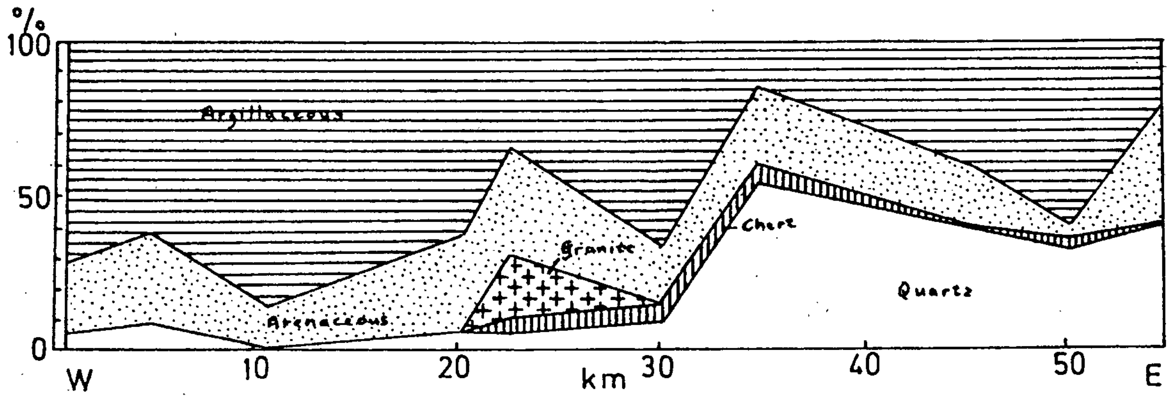


Fig. 39. Southern outcrops: Distribution of pebble types.

On the Central Divide, although the base of the Vaartwell Formation is not exposed, quartz pebbles dominate over polymictic types, suggesting a possible increase of stable pebbles towards the north. These outcrops show a pronounced decrease in unstable pebbles from west to east (Fig. 40).

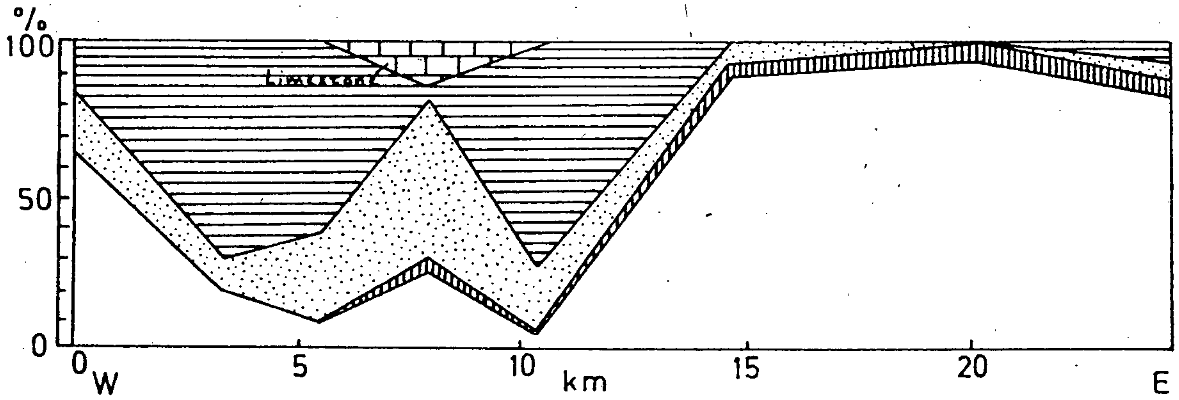


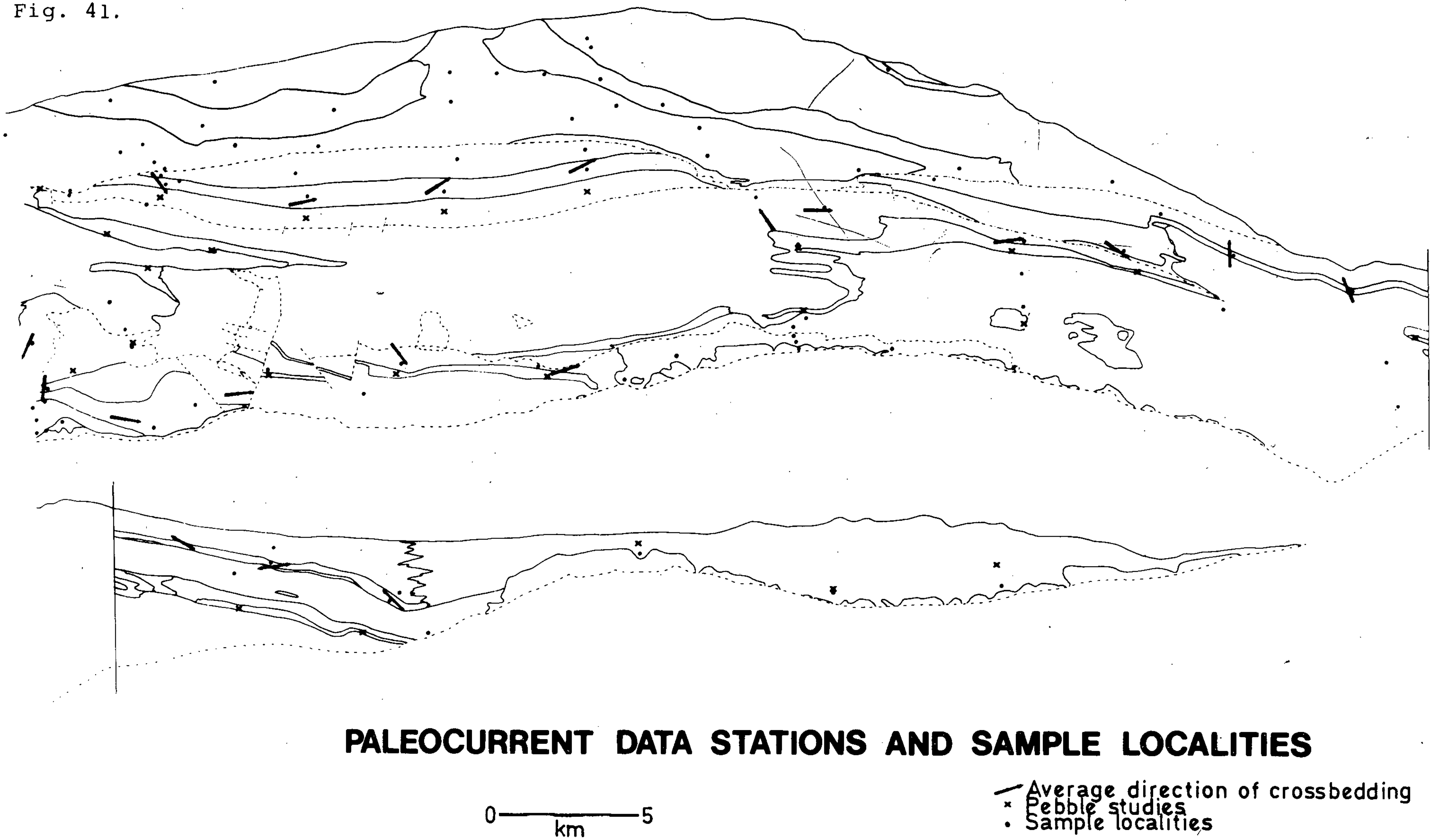
Fig. 40. Central Divide: Distribution of pebble types.

Detailed pebble studies were not carried out in the area west of Potgieters Poort, but an analysis given by Mulder on Vaartwell (21°32'E) suggests quartz as the major constituent, quartzite second in abundance, greywacke fairly common, and granite and "banded ironstone" present in minor amounts. The outliers of the Potgieters Poort Nappe have polymictic conglomerate at the base and grade upward into the monomictic type. Since the thrust sheet was derived from the south, it is obvious that the relationships mentioned above might not hold for other areas, and quartz pebbles may well become the dominant constituent outside the Cango outcrop to the west or south. Inside the study area however, the increase in stable pebbles and simultaneous decrease of less resistant types towards the north and east, is indicative of a northeasterly direction of transport.

PEBBLE SIZE AND SORTING

Pebble size measurements were obtained from two sets of stations arranged from west to east (Fig. 41), one following the southern outcrop of the Vaartwell from Koetzers Kraal to the Brakke River and commencing again in the syncline on De Oude Muragie (22°30'E), while the other measurements were taken on the Central Divide. The longest axes of 100 pebbles were determined at each station and plotted on cumulative curves, from which the median pebble size and

Fig. 41.



PALEOCURRENT DATA STATIONS AND SAMPLE LOCALITIES sorting were calculated. As a check on these figures the five largest pebbles at each locality were also measured and their average expressed in Ø. A close correlation was found between the two parameters.

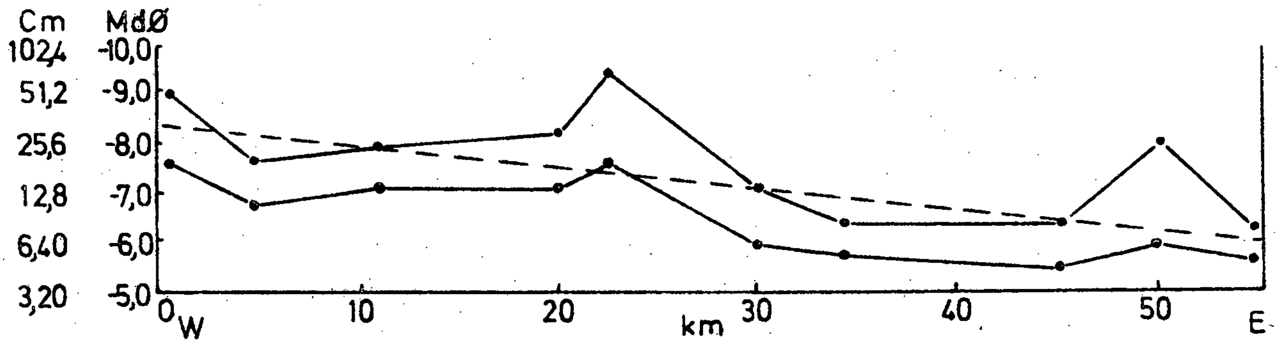


Fig. 42. Southern outcrops: Pebble size vs distance.

As illustrated in Figs. 42 and 43, the Md Ø increases in value from west to east, thus showing a decrease in pebble size in that direction.

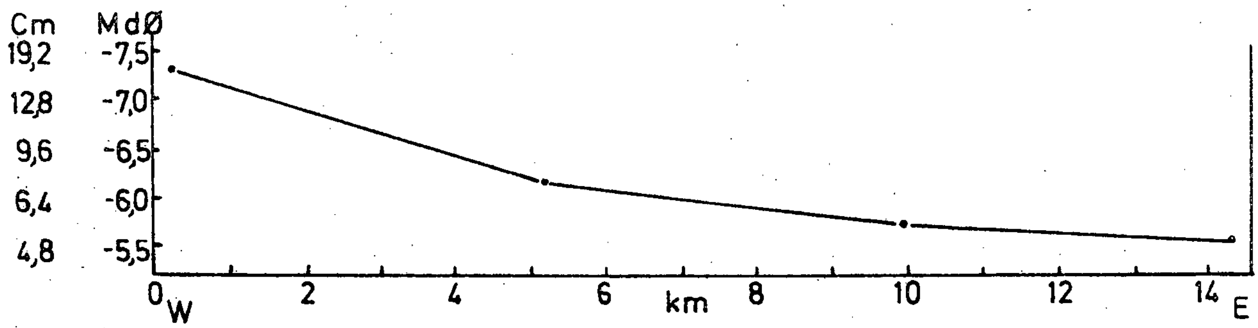


Fig. 43. Central Divide: Pebble size vs distance.

Sorting is somewhat variable but also seems to improve eastwards (Fig. 44). These data therefore confirm the picture of a west-to-east direction of transport.

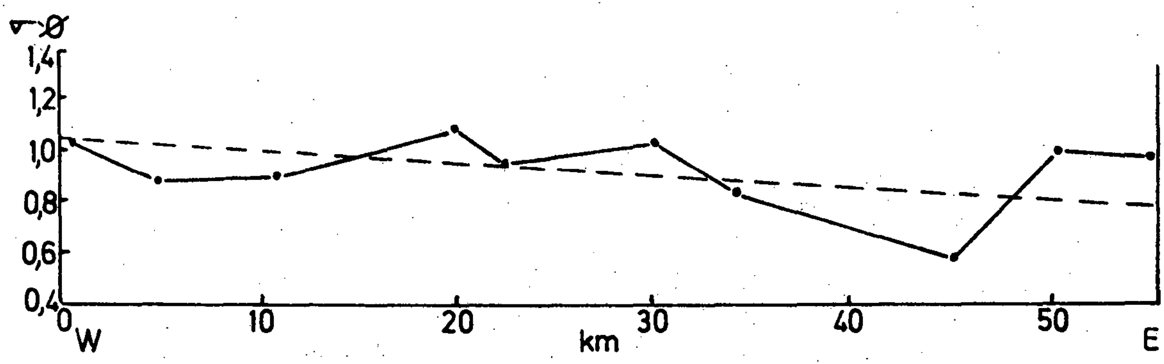


Fig. 44. Vaartwell Formation: Sorting vs distance.

IMBRICATION

Stocken (pp 33.) reported a primary orientation of the longest axes of resistant pebbles in an east-west direction. In order to verify this, quartz pebble orientations were measured at nine of the stations used in determining size distributions. It must be emphasized however, that the alignment of pebbles could have been influenced by tectonism in some cases and may not necessarily give a true indication of current directions.

According to many workers (e.g. Johnston, 1922; Krumbein, 1940), the longest axes of pebbles are usually parallel to the current, while the inclination is generally upstream in fluviatile deposits (Twenhofel, 1932). Roller pebbles on beaches tend to align themselves with their longest axes parallel to the shore line (Krumbein and Sloss, 1963) or in the case of disk shaped pebbles, show a preferred orientation with their long axes perpendicular to the shore line (Kukal, 1971). In such cases there is a small inclination seaward, although inverse imbrication has also been observed.

Rectified stereogram plots of quartz pebbles in the polymictic conglomerate indicate that streams flowed primarily in an easterly direction. Diskshaped quartz pebbles on the Central Divide have a north-south alignment and a rectified northerly dip, which could indicate a shore-line towards the south in beach deposits. Occasional crossbedding in the "matrix" between conglomeratic lenses suggests that transport was from the west, thus making a combination of "roller" and "disk" orientation mechanisms possible. The stippled arrows in Fig. 45 are perpendicular to the longest intercepts of quartz pebbles in the monomictic member and point away from the probable source area.

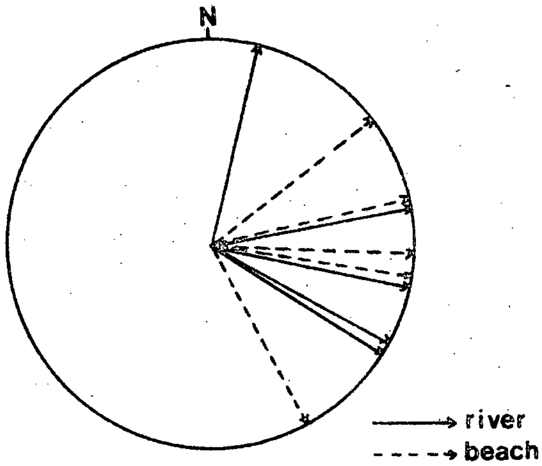


Fig. 45. Paleocurrent directions derived from imbrication of quartz pebbles.

In conclusion, most features investigated in the Vaartwell Formation are indicative of transport in a north-easterly direction, which places the provenance roughly to the southwest.

3.2.4. UITVLUGT FORMATION

3.2.4.1. PALEOGEOGRAPHY

Stocken, by including the conglomerates of the Vaartwell Formation as a basal part of the "Crossbedded Grit Formation", regarded both to be fluviatile origin. The upward gradation from conglomerate to sandstone and ultimately the siltstones of the present Gezwinds Kraal Formation he explained by: "a rejuvenated river eroding and smoothing down its bed, thus carrying to the coast a load continuously decreasing in grain size", or alternatively; "delta or alluvial fan formation which by their growth appreciably lessen the slope of the stream thus reducing its competency."

Although these factors undoubtedly contributed to the upward change in lithology, the proposed mechanisms do not take into account the possible effects of changes in the environment of deposition on the sedimentology. It has been shown that, although the polymictic conglomerates are probably fluvial deposits, the monomictic member may represent beach gravels and could indicate a transgression of the sea towards the lowering source area. If this was the case, an upward fining of sediments would be expected in response to successively deeper water cycles passing any one locality. The problem that emerges from these two hypotheses can therefore be summarized in the following question: Are the Uitvlugt and Gezwinds Kraal Formations representative of fluviatile and therefore continental deposits formed in a tectonically stable environment, or are they marine sediments indicating another cycle of subsidence and transgression across part of the former provenance? In the following discussion an attempt is made to distinguish between the two alternatives.

COMPOSITION AND TEXTURE

Plots of grain size parameters, except that they suggest deposition from unidirectional currents (Fig. 46 (a)), afford no evidence in favour of either one of the two possibilities, as they fall in the overlapping area of the respective fields for river and bay sediments (Figs 46 (b) and (c).

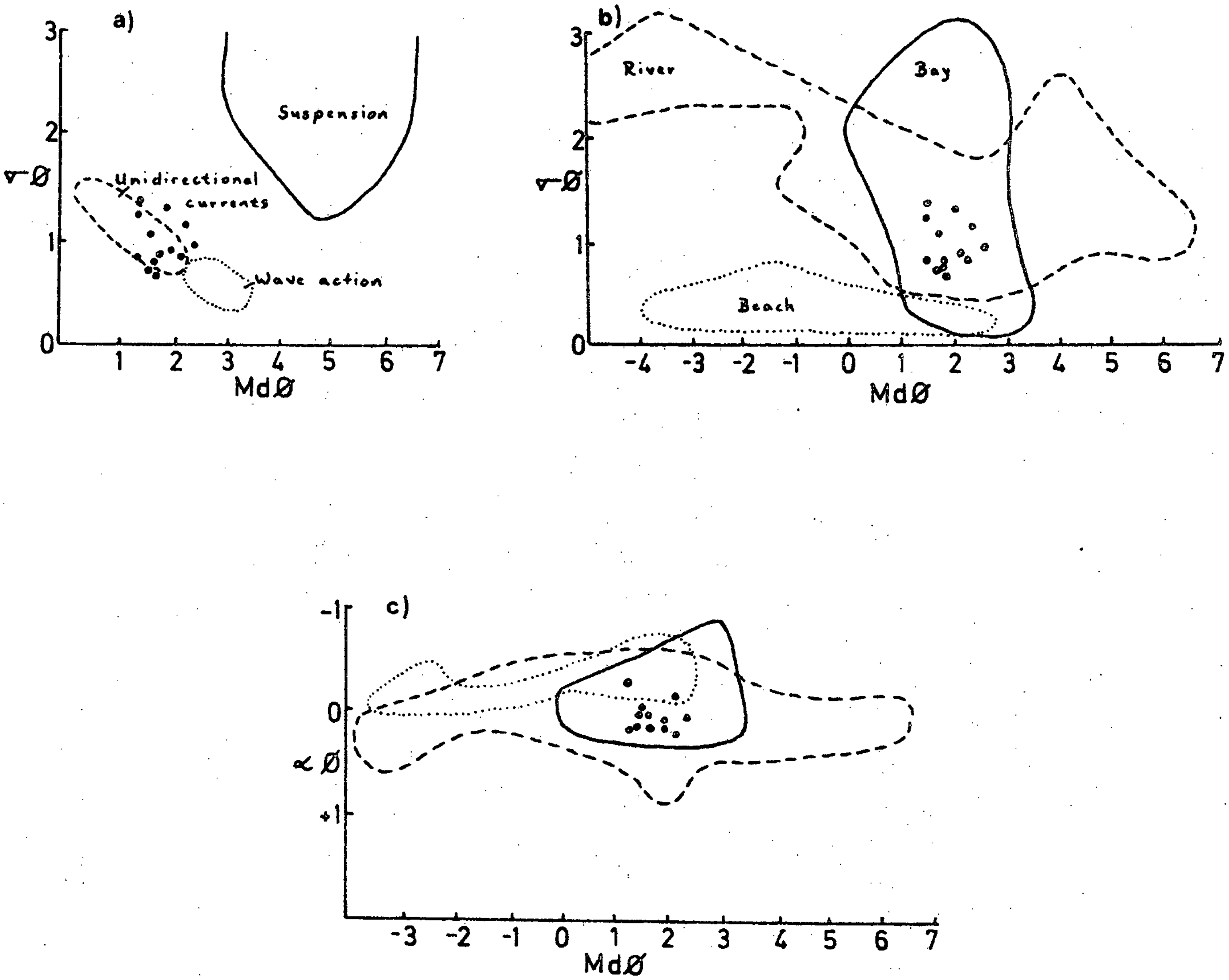


Fig. 46. Uitvlugt Formation: (a) Sorting vs grain size. (Stewart, op cit)
(b) Sorting vs grain size. (Kukal, op cit)
(c) Skewness vs grain size.

In composition the sandstones of the Uitvlugt Formation are rich in quartz and poor in feldspar, an exception being the conspicuous bands of subarkose in the basal part of the succession near Potgieters Poort. Carbonaceous flakes and fragments of shale are present in some samples. The matrix of fine quartz and clay minerals forms up to 33% of the rock. Although a few samples plot in the quartz arenite field (Fig. 3 (d)) these are probably not true orthoquartzites but more likely the products of local winnowing of quartz wacke sands, as suggested by the angular to subrounded grains and presence of unstable heavy minerals. The average particle size varies from 0,2 to 0,35 mm, while sorting is moderate to poor ($\sigma\phi$ 0,7 - 1,3). Quartz wackes of this type generally represent conditions of moderate subsidence in unstable depositional areas, where the rate of burial is rapid enough to prevent thorough winnowing during transportation (Krumbein and Sloss, 1963).

SEDIMENTARY STRUCTURES

Although some of the primary structures in the Uitvlugt sandstones, such as conglomeratic lenses (Danzers Kloof Member) and abundant cross-stratification, can be regarded as characteristic of river sediments, there is a lack of other features which would be expected in such an environment. Stream channels and larger cut-and-fill structures were not observed in the study area; in contrast the rhythmic, uniform and often thickly bedded units can be followed over relatively long distances and do not resemble elongated lenses as in stream deposits. There is very little textural variation between beds and abrupt lateral changes in particle size have not been observed. Giant ripples are common in most large rivers, but absent from the Uitvlugt, while small current ripples were encountered only on one or two occasions.

That strong currents nevertheless transported and deposited the Uitvlugt sandstones, is clearly shown by the abundant planar and occasional festoon-shaped crossbedding, the latter reaching up to 7 metres in cross sections (Plate 3 (e)). Small-scale scour and rip-up-clasts are also quite common, usually occurring in the vicinity of festoon-crossbedded "lenses". The average inclination (13°) of crossbeds in the Uitvlugt Formation is closer to the average for marine (18°) than for river deposits ($19,3^{\circ}$) (Pettijohn, 1962), but whereas marine crossbedding has a histogram peak between 18° and 24°, the Uitvlugt peak (Fig. 47) lies between 12° and 18° as in river sediments.

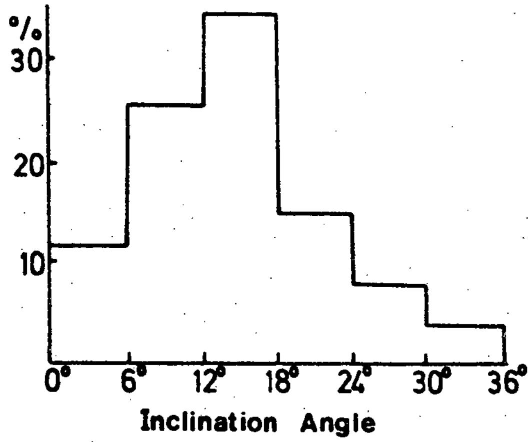


Fig. 47. Uitvlugt Formation: Histogram of crossbedding inclination angles.

It must be assumed, however, that compaction and deformation could have affected the original angles of inclination. A study of the azimuths of the Uitvlugt cross-stratification seems to provide more positive evidence for a marine environment. Although described by Stocken as "invariably from the west," actual measurements show a variable pattern both locally and regionally. Fig. 48 portrays an example of opposite current directions as defined by crossbedding and ripple marks (Schoemans Poort).

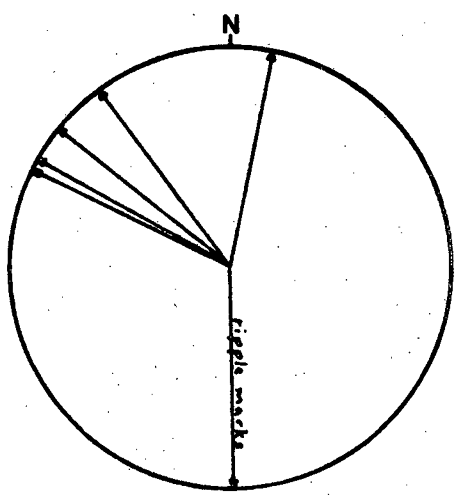


Fig. 48. Current directions in Uitvlugt Formation, Schoemans Poort.

Bipolar crossbedding was encountered in about 20% of the individual outcrops studied, which is characteristic of the infralittoral zone of the neritic environment. This zone extends from low tide to a depth of about 50 m, forming a broad band parallel to the coast where clastic material derived from the landward side by streams is spread over the bottom in response to wave and current action. These deposits are usually lenticular near the shore with a greater range in particle dimensions. The variation in wave and current trend is also great. In deeper water the units are better defined and more regular (Twenhofel,1932). It is therefore suggested that the Danzers Kloof Member, with its small conglomeratic lenses and festoon - crossbedding, was deposited near the shore, while the deposition of the Rooiberg Member took place in slightly deeper water. A depth of 20 m or more is indicated by the general absence of ripple marks.

In contrast to the environment responsible for the Nooitgedagt deposition, which was probably a restricted shelf with very variable currents and an irregular topography, the Uitvlugt depositional environment is visualized as a shallow, open shelf under the influence of strong, permanent ocean currents. That transgression may have proceeded in cycles is suggested by the stratigraphy of the Danzers Kloof Member, where a zone of offshore sediments containing few pebbles is overlain by a more conglomeratic horizon, presumably of nearshore origin, which finally grades into the Rooiberg Member.

3.2.4.2. PALEOSLOPE

ISOPACH VARIATIONS

The Uitvlugt Formation maintains a steady thickness of 700-800 m on the Central Divide, but on the farm Voorzorg in the eastern section there is a rapid decrease to about 200 m. While the Danzers Kloof and Rooiberg Members are of equal thickness in the central area, the latter becomes the dominant unit eastwards where it attains up to 6 times the width of the sporadically developed Danzers Kloof Member. Bedding thickness is also greatest in the central and southwestern parts, where it can be up to 4 m in places, but decreases to a steady

15-30 cm in the east. The extreme eastern tip of the formation finally wedges out between the deep-water sediments of the Groenefontein Formation and the shallow water equivalent of the Gezwinds Kraal, the Schoongezigt Formation. A decrease in both formation and bedding thickness can be expected away from the source area as the distance of transport increases, but it may also be the result of variations in the slope of deposition or rate of submergence across the depositional environment.

CROSSLAMINATION

Five measurements were made at each of 19 stations, the foresets plotted together with the orientation of the bedding and the latter rotated to the horizontal by use of stereographic methods. The mean direction at each station was determined and plotted on Fig. 41. The synoptic plot of the measured azimuths (Fig. 49) indicates a general easterly flow with a unipolar maximum due east and a bipolar submaximum displaying a southeast to northwesterly direction.

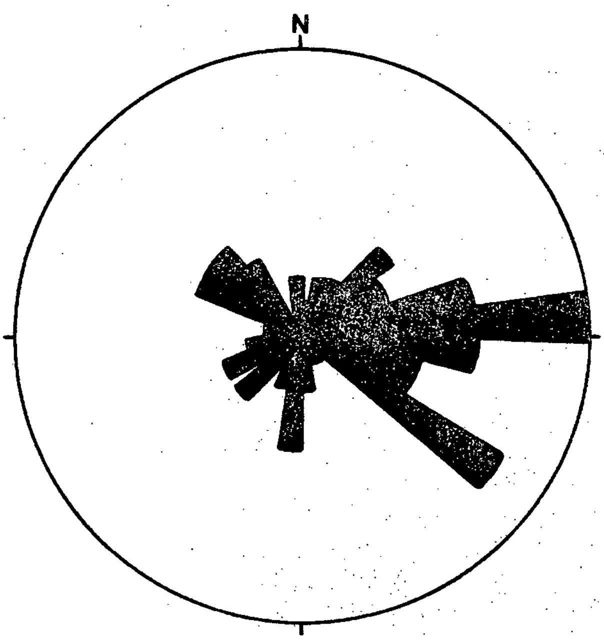


Fig. 49. Synoptic plot of crossbedding directions (95 measure= ments), Uitvlugt Formation.

It is possible that the main maximum is the result of a strong longshore or offshore current, while the submaximum represents ebb- and-flow tides. As longshore currents are often accom=

panied by waves moving obliquely towards the shore line, the sub-direction could indicate a beach to the north parallel to the transporting current. (Fig. 50) In most marine sandstones however, the net crossbedding is commonly down=slope (Pettijohn 1962).

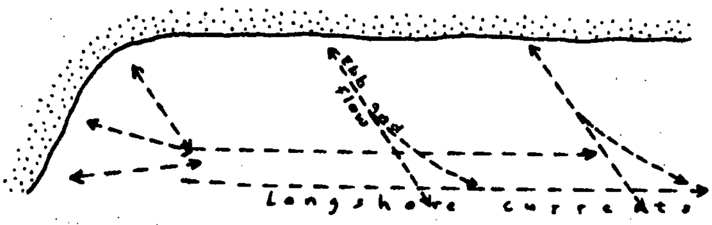


Fig. 50. Interaction of tidal and longshore (offshore) currents.

In the central area cross-stratification is well developed and shows a fairly constant direction of inclination, but in the extreme eastern part it is less well defined with more variable directions, changing from an easterly to northerly direction of transport.

GRAIN SIZE

Although there is no clear trend from west to east of median grain sizes, greater variability seems to occur in the west (Fig. 51). This may be attributed to a slightly more heterogeneous, nearshore environment proximal to the source area.

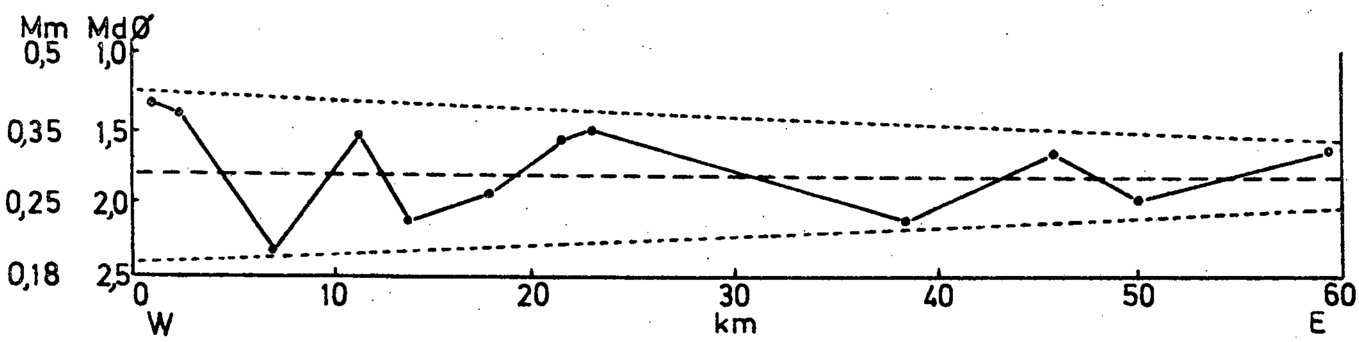


Fig. 51. Uitvlugt Formation: Grain size vs distance.

SORTING

The value of $\sigma\phi$ sharply decreases during the first 20 km and then stays fairly constant in the eastern samples, indicating improved sorting in that direction. (Fig. 52).

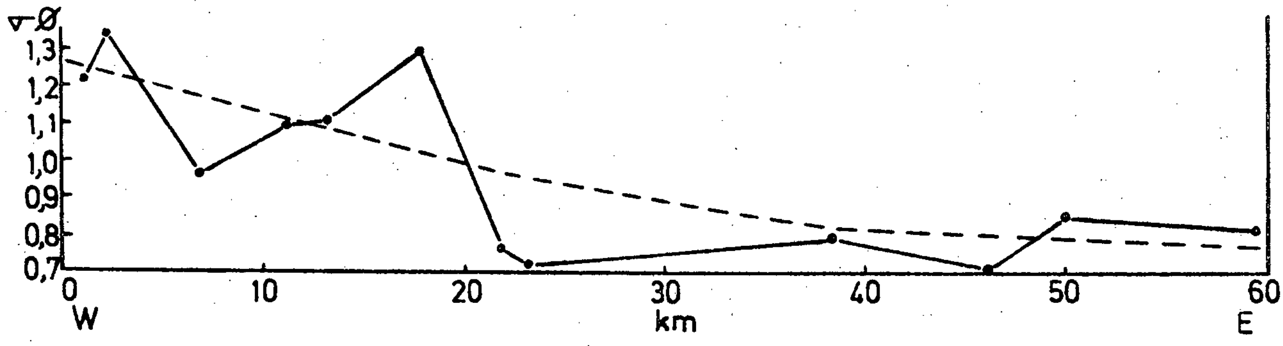


Fig. 52. Uitvlugt Formation: Sorting vs distance.

SKEWNESS

A "tail" in the coarse grain fractions increasing towards the east is shown by the values of $\alpha\phi$ (Fig. 53). This is in accordance with observations in some river sediments, where $\alpha\phi$ drops from positive to zero and even negative values in the direction of flow (e.g. Naidu and Borreswara, 1965).

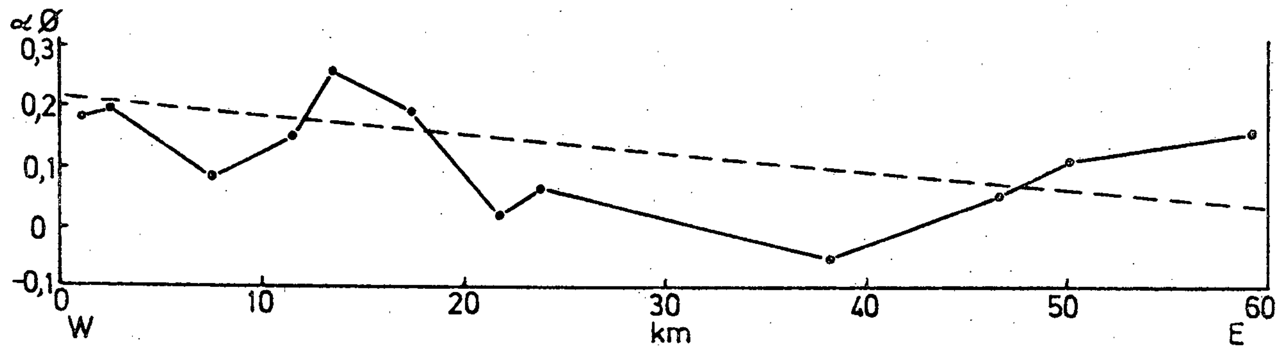


Fig. 53. Uitvlugt Formation: Skewness vs distance.

MATRIX

The percentage of matrix steadily decreases towards the east, suggesting "cleaner" sand, less argillaceous rock fragments and greater winnowing, which is probably a function of the distance of transport. (Fig. 54).

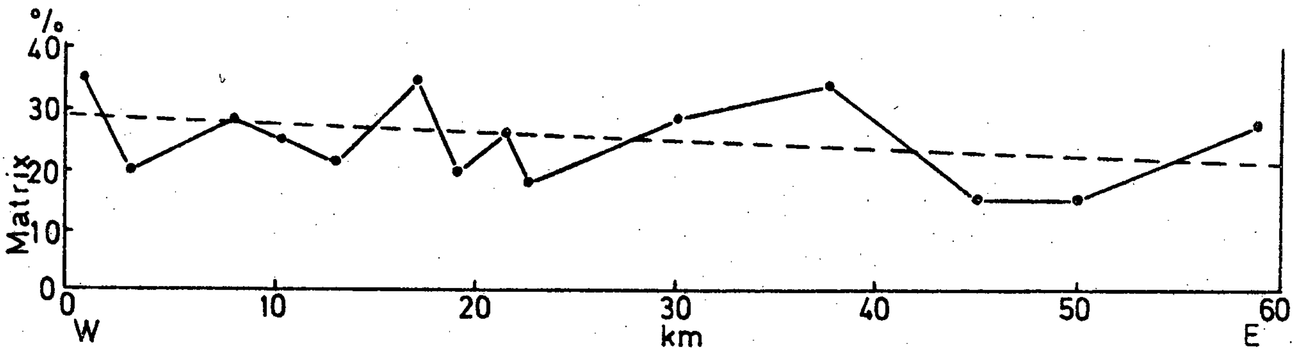


Fig. 54. Uitvlugt Formation: Mātrix content vs distance.

The decrease in K_2O content (Fig. 55) may be partly related to the clay minerals of the matrix (see Fig. 65) and partly to a decrease in the amount of akali-feldspar.

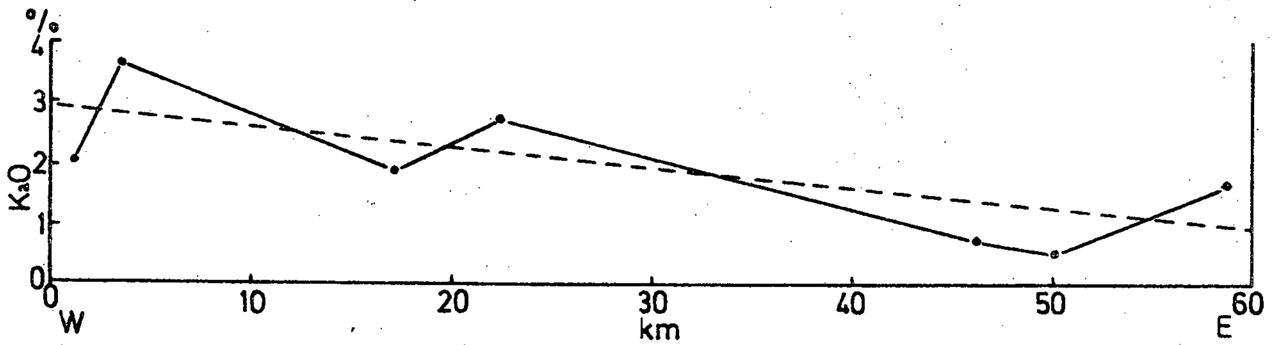


Fig. 55. Uitvlugt Formation: K_2O content vs distance.

Considering the evidence outlined above, there is no doubt that the bulk of the Uitvlugt sandstones were transported in an easterly direction over most of their outcrop area, except in the east where currents apparently turned north=south.

3.2.5. GEZWINDS KRAAL AND SCHOONGEZIGT FORMATIONS

These two formations occupy the same stratigraphic position in the Cango Group, and for the purpose of paleoslope discussion are therefore considered as one unit. Their environments of deposition however, if conclusions reached during the present investigation are correct, were very different.

3.2.5.1. PALEOGEOGRAPHY

GEZWINDS KRAAL FORMATION

When considering the depositional environment of the Gezwinds Kraal Formation, it must be kept in mind that its deposition was part of a continuous process which started with the Vaartwell sedimentation. In the foregoing pages it has been proposed that the upward fining of sediments is mainly the result of a major transgressive cycle gradually increasing the water depth and distance from the shore. This was presumably accompanied by an approach towards an equilibrium profile because of a lack of further diastrophic movement in the source area, which progressively reduced the transporting capacity of the contributing streams. Although Stocken accepted the latter as the only mechanism responsible for the change in lithology, and regarded the sediments as fluviatile deposits, there are many arguments against such an origin. Strong textural variations, scour-and-fill structures and shallow water features such as ripple marks are entirely absent in the formation, while its most conspicuous characteristic, viz. rhythmic, laterally homogeneous bedding, is somewhat unusual for river sediments. Many of the graphic plots of parameters also fall outside the field for fluviatile deposits, but Figs. 56 (a) and (b) show that a slight shift in the field for bay sediments towards finer grain sizes would incorporate the samples as regards both their sorting and skewness measure, which may indicate a similar, marine environment at a greater distance from the shore.

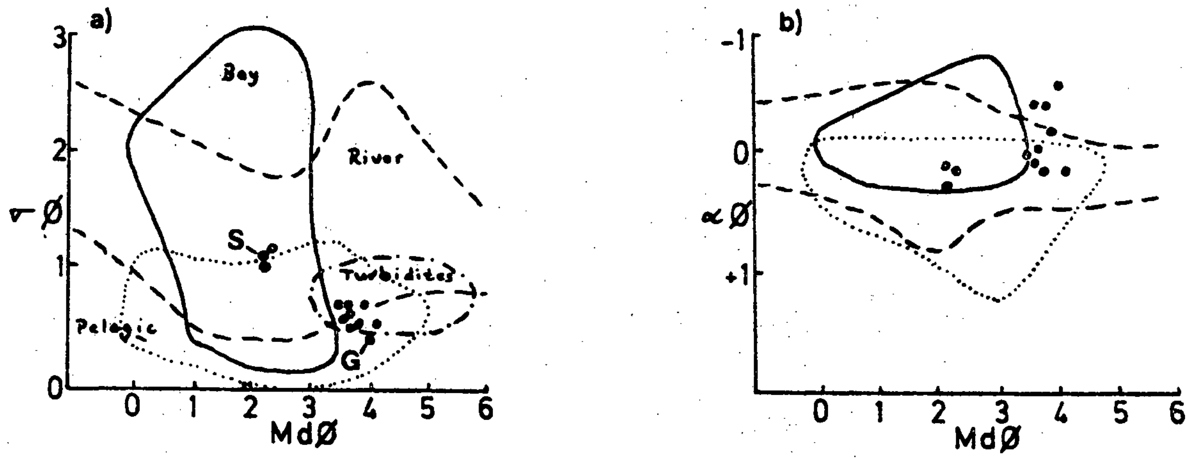


Fig. 56. Gezwinds Kraal (G) and Schoongezigt (S) Formations:
(a) Grain size vs sorting (Kukal, op cit).
(b) Grain size vs skewness (Kukal, op cit).

Unfortunately data for this type of environment, transitional between shallow water and pelagic deposits, are not available.

The cyclicity of sandstone and shale has been attributed by Stocken to seasonal flooding of the transporting river, which, in the alternative case, could equally well apply to a rhythmic influx of coarser material into the depositional basin during periods of increased erosion. The bipartite structure however may be related to longer-period climatic cycles and is not necessarily a result of yearly variations.

Vertically the rhythmic alternation of fine-grained sandstone and shale is somewhat similar to the Groenefontein Formation, although shale is more prominent in every cycle and becomes the dominant constituent higher up in the succession. This is partly a result of the sandstones becoming increasingly finer grained and thus passing into the silt or shale category.
The respective shale and sandstone beds are sharply demarcated and display no evidence of grading, which distinguish them from the Groenefontein cycles. Although some of the graphic parameters plot inside the field for turbidites, such an origin for the formation as a whole is very doubtful. While bedding thickness in the Groenefontein wackes is very irregular in a vertical sense, the progressive decrease in thickness of the

Gezwinds Kraal beds towards the upper parts of the formation would require trigger mechanisms regularly fixed in time, which is extremely unlikely. An exception occurs in the eastern part of the Gezwinds Kraal outcrop (near the facies boundary) where bedding is less regular and sole markings were observed on loose blocks in the formation. These could indicate turbidity currents, probably of a very local nature, flowing from the shallow water environment of the Schoongezigt downslope into the deeper Gezwinds Kraal basin. (Fig. 57).

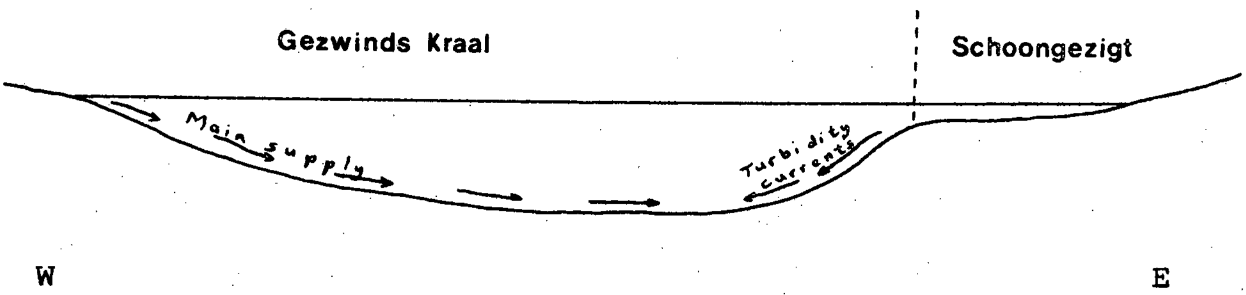


Fig. 57. Cross section through Gezwinds Kraal - Schoongezigt basin.

Although the absence of typical shallow water features may indicate deep water, the lithological characteristics of these deposits are probably more dependent on the distance from the land mass than the absolute depth of sedimentation. With tur= bidity currents ruled out as a transport mechanism, bottom ocean currents must have been responsible for the distribution of terrigenous material delivered from the source area. Both distance and depth will become limiting factors in this type of traction transport, but according to Kukal (1971, p. 226) these currents are operative to a depth of at least 200 m, which means that the actual depth of sedimentation could have been several hundred metres. Sedimentary structures typical of much deeper water, such as gradation or lamination, however, were never observed. It is therefore suggested that the infralittoral environment of the Uitvlugt gradually gave way to circalittoral conditions, but probably did not pass beyond the neritic zone.

SCHOONGEZIGT FORMATION

Although in the central area there is hardly any recognizable variation laterally in the Gezwinds Kraal Formation, a gradual change can be observed from the vicinity of Meirings Poort eastwards. Bedding becomes less regular and more massive, the average grain size increases and pebbles of vein quartz, grit, sandstone and shale make their appearance. While the general lithology of the Schoongezigt Formation rather resembles the Danzers Kloof Member, cross-stratification is absent and the bedding more variable both vertically and laterally. The distribution of conglomeratic lenses is also erratic and not comparable to the upward gradation of the Danzers Kloof Member.

Pebble types in the Schoongezigt Formation are similar to those of the Vaartwell polymictic member, although limestone and granite inclusions were not encountered. This suggests a common or at least similar source area, which may include reworked Groenefontein material mixed with "exotic" vein quartz and black chert. As the Uitvlugt sandstones wedge out with the appearance of pebbles in the Schoongezigt Formation, the direct exposure of the Groenefontein sediments above sea level affords no problem. This uplift, which could also explain the thinning and ultimate disappearance of the Uitvlugt Formation in the east, would have initiated fluvial sedimentation in this area, of which the Schoongezigt Formation may be the result.

There is little doubt that these sediments are indeed fluviatile or shallow water deposits. The occurrence of pebbles up to 15 cm in diameter obviously demands strong transporting currents (at least 200 cm/sec), which usually are found in the lower courses of rivers (Hjülstrom,1935; Nevin,1946) near the marine environment. The lenticular bedding, high textural variability and poor to moderate sorting is typical of the alluvial environment, although the absence of crossbedding, scour or other shallow water features weakens this interpretation. Grain size parameters plot in the overlapping area of bay and river sediments (Figs. 56 (a) and (b)). One possible solution is that the Schoongezigt Formation represents nearshore stream deposits occasionally brought under the influence of wave action during storms, which disturbed current features formed previously. In this sense it may be regarded as a transitional, fluvio-deltaic deposit between the continental and marine realms. This type of environment can be found in "tectonic delta" complexes, which form as thick sequences marginal to alluvial fans and are composed of conglomerates, coarse sandstones and shale. An example is the Catskill Formation of New York State, where a similar facies gives way over a narrow lateral transitional zone to a marine sequence dominated by fine clastics, which are regarded as slope turbidites (Allen and Friend, 1968; Walker and Harms, 1970). The resemblance to the Gezwinds Kraal Formation is clear.

3.2.5.2. PALEOSLOPE

As in the previous formations, directional studies in the Gezwinds Kraal and Schoongezigt Formations are severely hampered by their very restricted north - south distributions. Furthermore, the upward decrease in grain size of the former makes it essential to obtain samples from the same strati= graphic positions, which is not always possible. Although similar distances from the lower contact with the Uitvlugt Formation were used, the latter itself is very difficult to place with accuracy. In the Schoongezigt Formation the irregular distribution of sediment types presented similar problems. The variable values of the grain size parameters are therefore possibly a reflection of these inevitable inaccuracies and should not be leaned upon too heavily.

GRAIN SIZE (Fig. 58)

The median grain size stays fairly constant from west to east for the first 30 km, decreases during the next 20 km, and increases again near the lateral facies boundary of the Schoongezigt Formation.

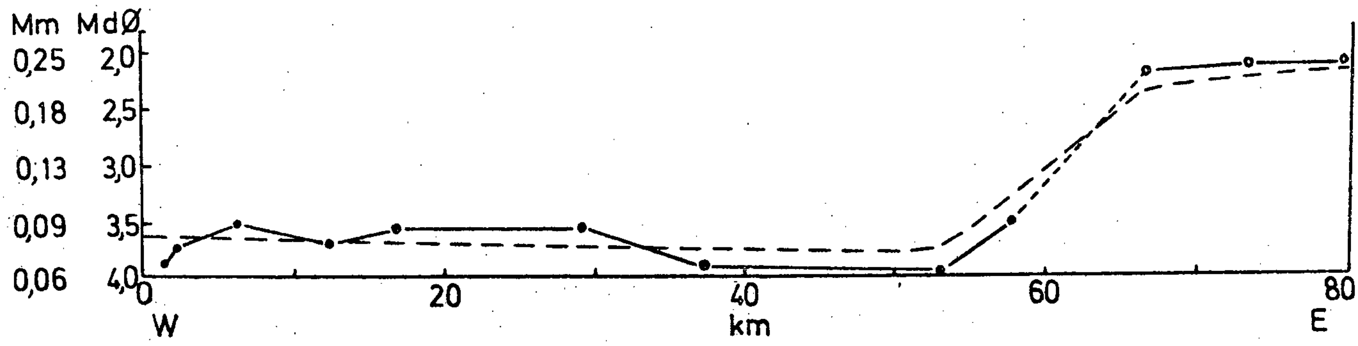


Fig. 58. Gezwinds Kraal and Schoongezigt Formations: Grain size vs distance.

SORTING (Fig. 59)

A similar pattern is followed by the coefficient of sorting, which improves after 30 km and then deteriorates approaching the Schoongezigt Formation.

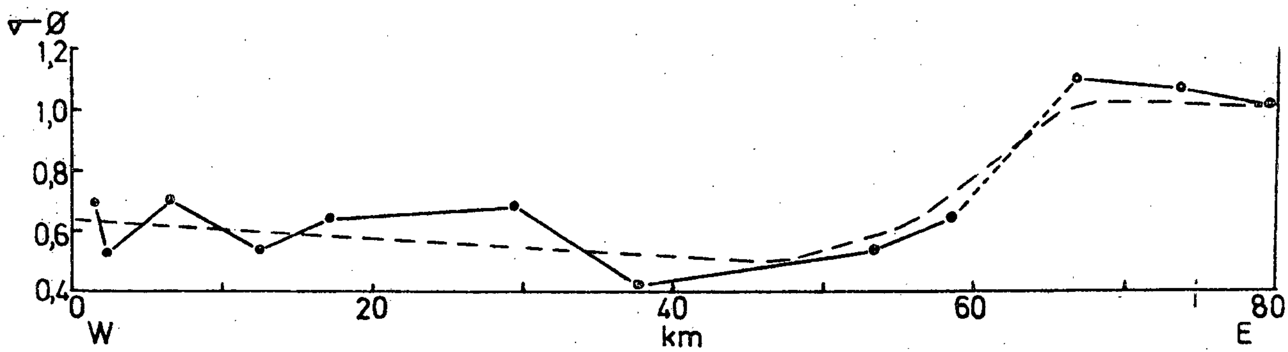


Fig. 59. Gezwinds Kraal and Schoongezigt Formations: Sorting vs distance.

SKEWNESS (Fig. 60)

Predominantly negative values in the west give way to positive values in the Schoongezigt Formation and vicinity.

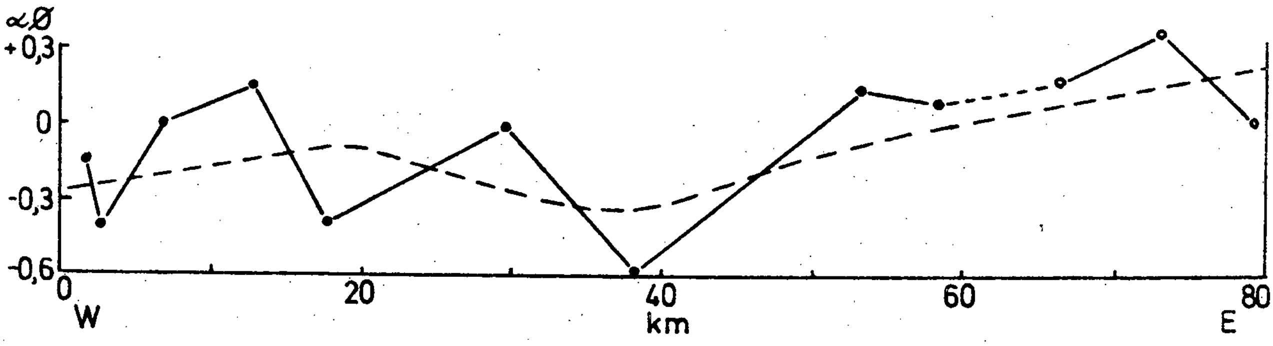


Fig. 60. Gezwinds Kraal and Schoongezigt Formations: Skewness vs distance.

MATRIX (Fig. 61)

The percentage of matrix decreases towards the east in the Gezwinds Kraal Formation as a whole, and slightly increases in that direction in the Schoongezigt Formation.

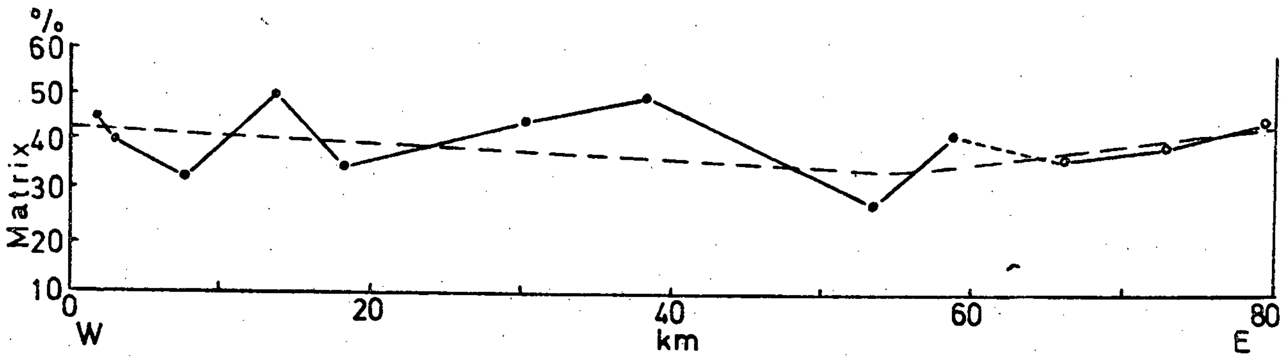


Fig. 61. Gezwinds Kraal and Schoongezigt Formations: Matrix content vs distance.

The K_2O content, as in the Uitvlugt Formation, follows the same trend. (Fig. 62)

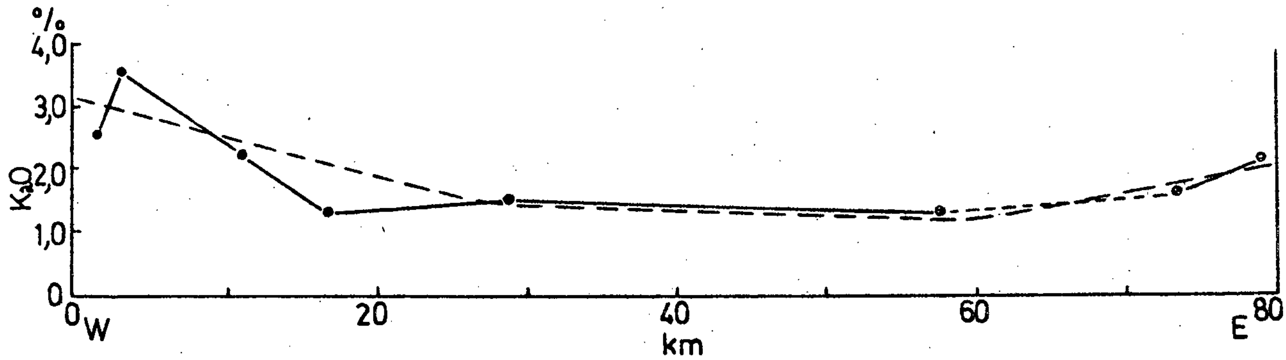


Fig. 62. Gezwinds Kraal and Schoongezigt Formations: K_2O content vs distance.

Although scant, the evidence outlined above confirms the regional paleogeographic picture of two separate source areas. The Gezwinds Kraal Formation, deposited as a continuation of the Uitvlugt sedimentation, was presumably derived mainly from the source area in the west, while the Schoongezigt Formation represents a separate, but essentially time= equivalent process of sedimentation. Its provenance was probably located in the east at the other end of the Cango basin, as shown in Fig. 64. Some of the "overflow" from these shallow water sediments probably contributed to the Gezwinds Kraal basin, possibly by local turbidity currents down these steeper slopes. (Fig. 57).

3.2.6. SCHOEMANS POORT FORMATION

As no special environmental studies were undertaken in the Schoemans Poort Formation, only very broad inferences can be made. The lenses of polymictic conglomerate at the base of the formation were probably derived from the Groenefontein Formation or may represent reworked material from the polymictic member of the Vaartwell Formation. The large size (30 cm or more) of some of the pebbles demands current velocities of more than 300 cm/sec, which can be reached in the middle or upper courses of rivers (Nevin, 1946). Graphic parameters of the sandstones also plot in the field for fluviatile sediments. (Fig. 63).

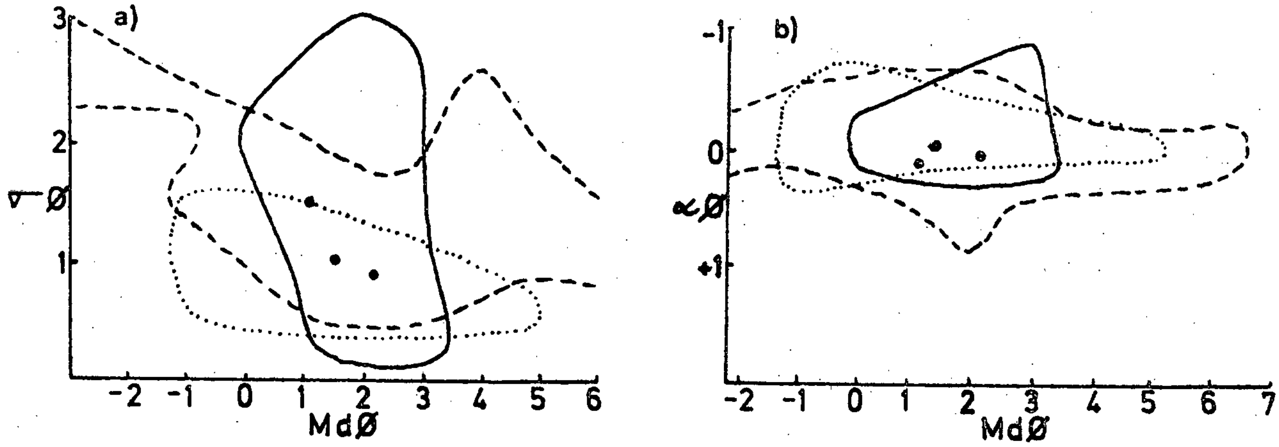


Fig. 63. Schoemans Poort Formation:
(a) Grain size vs sorting (Kukal, op cit).
(b) Grain size vs skewness.

That deposition occurred near the source area is indicated by the poor sorting of the formation as a whole, large textural variability and high feldspar content of the sandstones. Together with the absence of semi-permanent current features such as crossbedding, this could be the result of sedimentation during sudden, violent rainstorms followed by rapid run-off, which left behind a rubble of pebbles, coarse sand and silt in depressions along the base of steep slopes or mountain fronts. That these may have had very restricted distributions is suggested by the fact that no similar sediments appear any= where else beneath the Table Mountain Group in the Cango out= crop area. Correlation with the Klipheuwel Formation in the Western Province therefore seems to be rather presumptuous and the writer would suggest that this view be abandoned.

No directional studies were undertaken in this formation, but the restricted occurrence of the rocks suggests an origin probably towards the south, in the area now overlain by Cretaceous sediments.

3.3. THE SOURCE AREA

A paleogeographic reconstruction not taking into account the nature and character of the source area would be incomplete. Furthermore, because source areas are positive or tectonically active regions, the matter of provenance is directly coupled with tectonics. An attempt is made therefore to decipher the location, distance, relief and rock types of the provenance.

3.3.1. LOCATION AND REGIONAL PALEOGEOGRAPHY

From the sedimentological data outlined in the preceding pages the existence of possibly two separate source areas was postulated. Sediments during Matjies River, Vaartwell and Uitvlugt times were derived from a provenance roughly to the west, a more exact determination by petrographic methods made impossible by the restricted north-south exposure of the pre-Cape rocks. The Groenefontein and Schoongezigt sediments on the other hand, were apparently transported in a westerly direction, with the distribution of facies possibly resulting from source areas in the east or south-east.

The regional structural trends in the Cango Group, with overfolding and thrustfaulting in a northerly direction, give a fairly reliable indication that the craton was situated in the north. As the east-west trending fold axes are probably parallel to the axis of the sedimentary trough, longitudinal transport must have been the major factor during sedimentation, while the direction of current flow was determined by the slope along the trough axis. Transport directions as a key to the location of the source areas may therefore be somewhat misleading and other factors must be considered in order to complete the regional picture.

As cratons are characterized by their tectonically stable nature, it must be assumed that the strong uplift and rapid influx of sediments associated with the Vaartwell and Uitvlugt Formations was the result of tectonism in a land mass to the south or southwest. An east-northeastward direction of transport and a possible coast line in the south was postulated on account of pebble studies in the Vaartwell Formation (p.69), which seems to confirm this possibility.

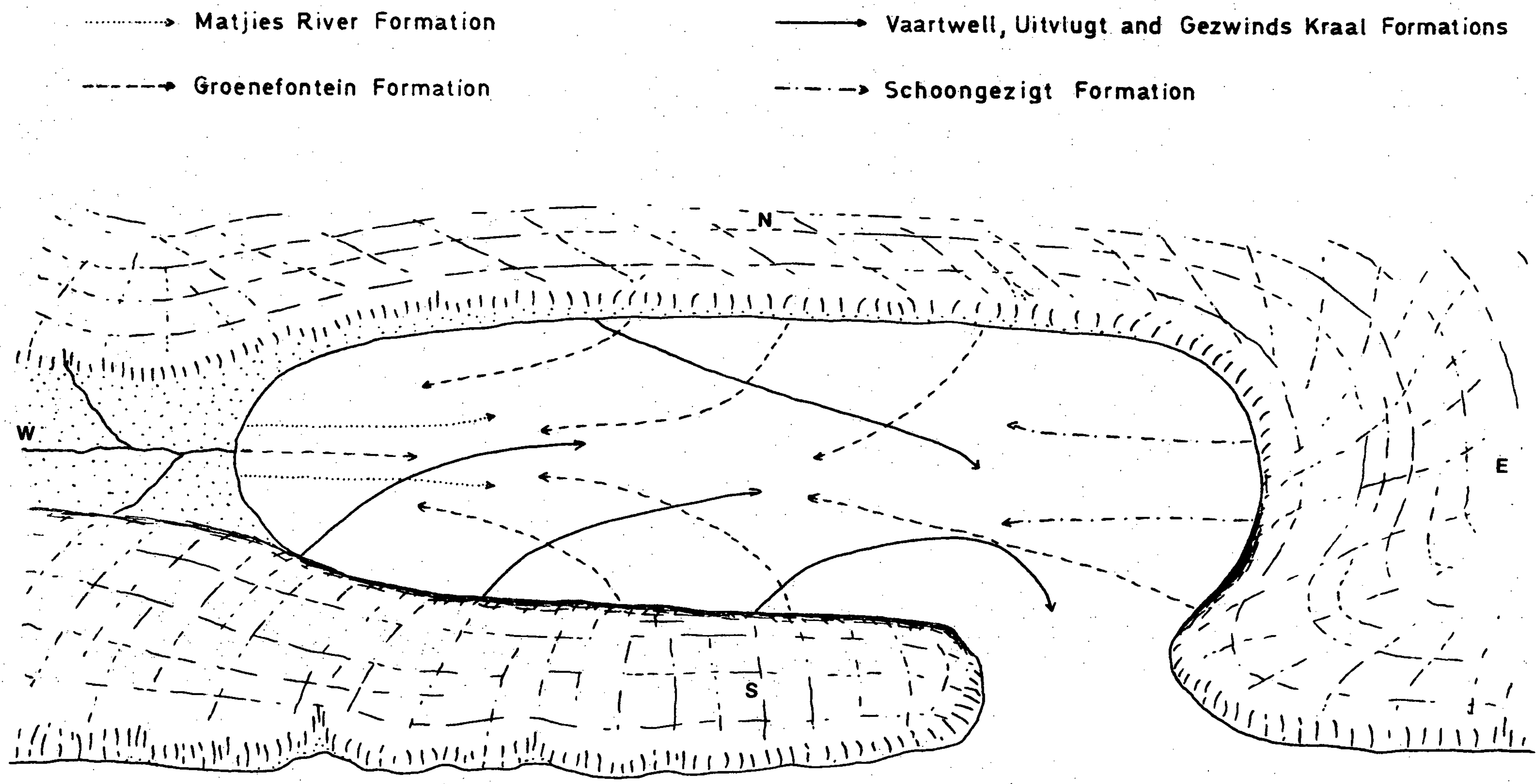


Fig. 64. Paleogeography and transport directions in Cango basin.

Similarly, the lateral distribution of coarse sandstone lenses in the Upper Member of the Groenefontein Formation suggests a source southeast of the present outcrop, which may represent a preliminary uplift in this part of the southern land mass before the major elevation in the southwest exposed part of the Groenefontein sediments above sea level.

Apart from a westerly direction of flow, there is no evidence that the turbidites of the Middle Member were derived from either the craton or the southern coast, and both must be considered possible source areas. This also applies to the upper parts of the Gezwinds Kraal Formation.

Facies patterns in the Matjies River Formation, which are more dependent on environmental influences than actual current directions, suggest an influx from the west which cannot be attributed as easily to longitudinal transport. A source to the west should therefore be kept in mind in attempting to reconstruct the regional picture.

Finally, the sediments of the Schoongezigt Formation denote shallow water and a tectonically active provenance in the east.

Putting the pieces together, a picture emerges showing a craton to the north, a tectonically active land mass in the south and coast lines to the west and east. The paleogeography shown in Fig. 64 depicts a long, narrow embayment such as the Gulf of California or the Red Sea, which is intended only to convey the general idea rather than an attempt to picture a particular stage in the development of the Cango paleogeography. This model can explain most features in the Cango Group, as for example the change in current direction towards the north in the eastern part of the Uitvlugt Formation, which is attributed to strong inflowing tides in the area opposite the bay entrance.

3.3.2. DISTANCE

There are several indications that large distances of transport were not involved in the sedimentation of the Cango Group. The abundance of potash feldspar as angular grains erratically distributed throughout the Nooitgedagt and Upper Members and also near the base of the Uitvlugt Formation, is a sure indication of the proximity of their source, while the presence of unstable heavy minerals such as diopside, hypersthene and epidote leads to the same conclusion. In addition, the preservation of granite, limestone and other unstable cobbles of the polymictic member, together with their large size, can only signify a relatively short distance of transport. The only sediments likely to have travelled, some distance are those of the Groenefontein and Gezwinds Kraal Formations.

3.3.3. RELIEF AND CLIMATE

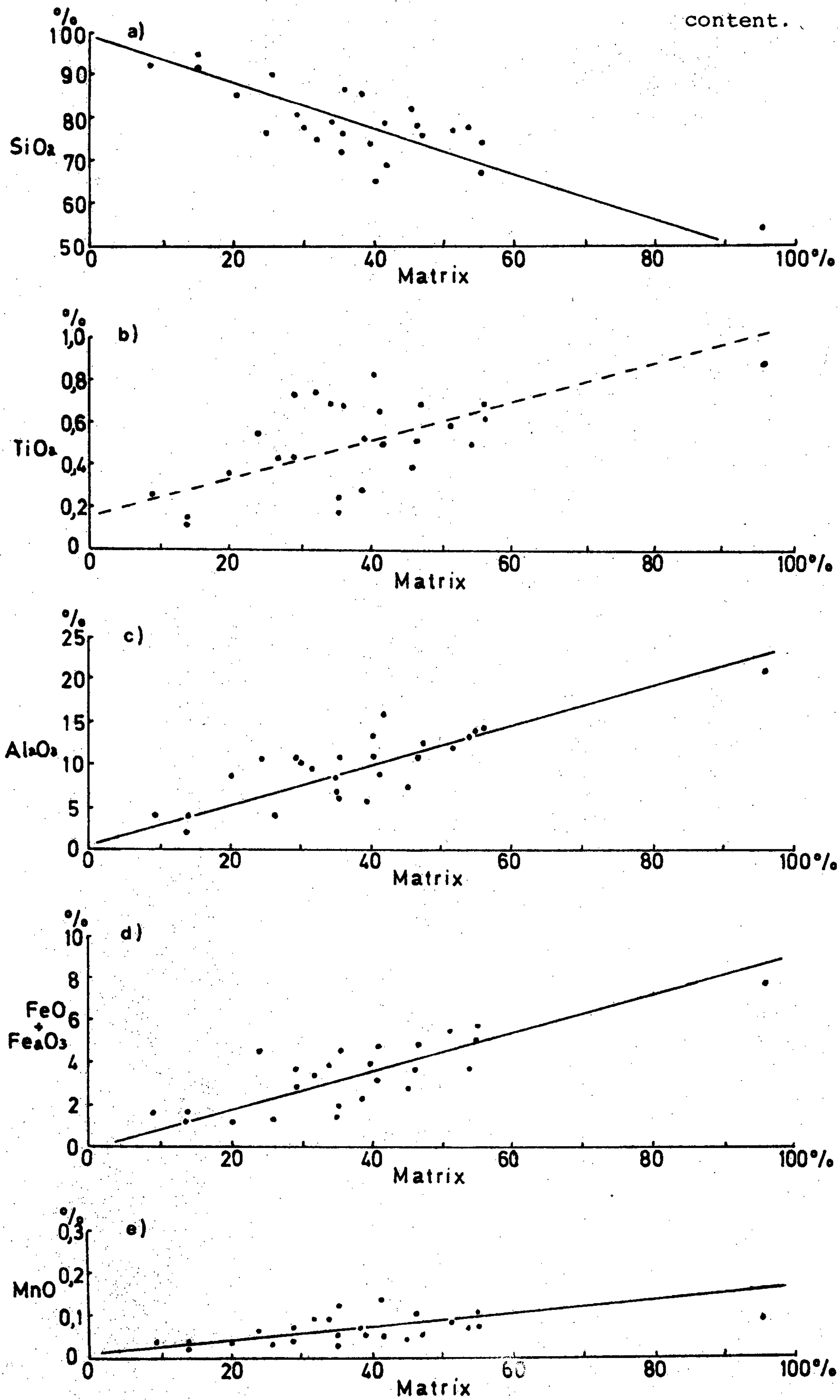
High reliefs and erosion rates were evidently prevalent in the source areas of the Vaartwell, Uitvlugt and Schoongezigt Formations, while a more subdued topography is probably responsible in part for the fine texture of the Groenefontein and Gezwinds Kraal sediments. Although the relief of the Matjies River provenance may have been mountainous, the recurrence of marine transgressions during its sedimentation (p. 40) is not in favour of a high topography. The alter= native is a large river, fed by short tributaries running down nearby slopes, draining an extension of the gulf area to the west. A tropical or subtropical climate would have been favourable for limestone precipitation.

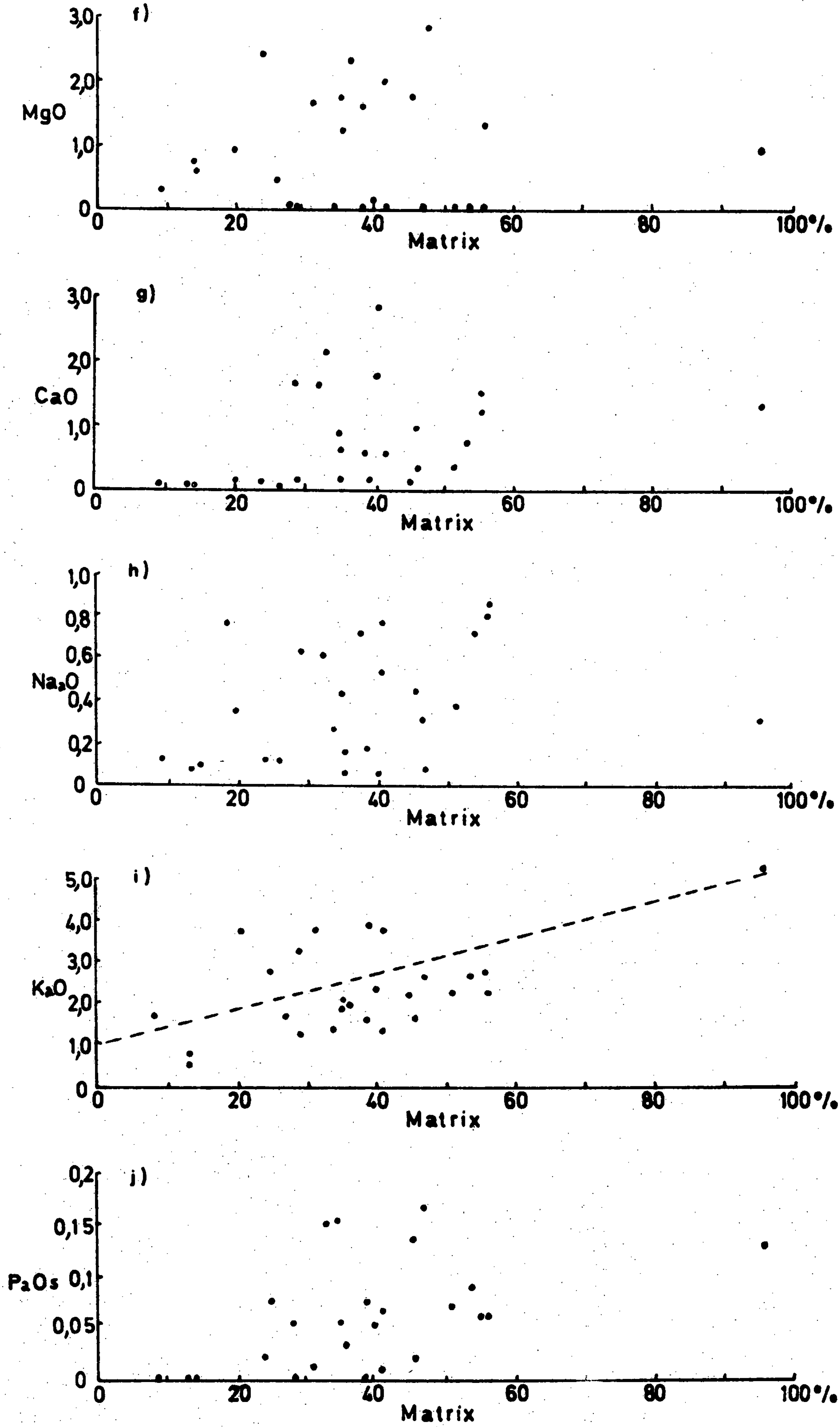
3.3.4. THE SOURCE ROCKS

If the paleogeographic reconstruction shown in Fig. 64 is basically correct, a large variation in the type of detritus supplied from both sides of the basin would not necessarily be expected, especially as the new trough could have developed in a region previously underlain by similar rock types. A difference may occur however, if one of the flanks of the trough was tectonically more active, thus being uplifted and eroded more frequently and so gradually unroofing older sediments or deep seated igneous rocks.

Lithologically the difference between the sediments presumably derived from the craton and those of the southern, more active land mass, lies in their grain size and feldspar content. Geochemically there is also a marked variation (Table 6), but how much of this is due to principal differences in the source rocks is uncertain. It seems as if the matrix definitely plays a major role in the geochemistry of these rocks, as shown by Fig. 65. The SiO_2 content varies

Fig. 65. Chemical composition of Cango sediments as a function of matrix content.




f)
3,0
2,0
MgO
1,0
0
0
20
40
60
80
100%
Matrix
g)
3,0
2,0
CaO
1,0
0
0
20
40
60
80
100%
Matrix
h)
1,0
0,8
0,6
Na₂O
0,4
0,2
0
0
20
40
60
80
100%
Matrix
i)
5,0
4,0
3,0
K₂O
2,0
1,0
0
0
20
40
60
80
100%
Matrix
j)
0,2
0,15
P₂O₅
0,1
0,05
0
0
20
40
60
80
100%
Matrix

inversely with the percentage of matrix, while TiO_2, Al_2O_3, total Fe, MnO and K_2O show positive correlations. MgO, CaO, Na_2O and P_2O_5 are evidently not dependent on the amount of matrix.

TABLE 6. GEOCHEMISTRY

FORMATION	Number of Samples	SiO_2	TiO_2	Al_2O_3	Total Fe	MnO	MgO	CaO	Na_2O	K_2O	P_2O_5
MATJIES RIVER	2	83.69	0.49	6.74	1.72	0.06	0.95	0.76	0.37	2.64	0.01
UITVLUGT	7	84.03	0.29	6.01	1.85	0.03	1.13	0.15	0.20	1.91	0.24
SCHOONGEZIGT	2	83.63	0.33	6.41	2.48	0.05	1.59	0.37	0.31	1.84	0.01
GEZWINDS KRAAL	6	76.26	0.60	11.20	4.24	0.03	0.80	1.05	0.40	2.17	0.10
GROENEFONTEIN	9	71.63	0.65	12.96	5.06	0.11	0.28	1.09	0.59	2.79	0.07

The heavy mineral suite, with the exception of pyroxene, anatase, ilmenite and pyrope, is similar in all the formations (Table 7). Point counting of species in specially mounted crops revealed a very erratic distribution both laterally and vertically in the Cango Group, thus eliminating the possible existence of heavy mineral provinces.

TABLE 7. HEAVY MINERALS

FORMATION	Number of Samples	Mgt	Ilm	Py	Zr	Ap	Ep	Tour	Rut	Opx	Cpx	Anat	Alm	Pyr
MATJIES RIVER	3	X	X	X	X	X	X	X	X				X	
UITVLUGT	13	X	X	X	X	X	X	X	X	X	X	X	X	X
SCHOONGEZIGT	3	X	X	X	X	X	X	X	X			X	X	X
GEZWINDS KRAAL	7	X		X	X	X	X	X	X	X		X	X	X
GROENEFONTEIN	9	X		X	X	X	X	X	X	X	X	X	X	X
SCHOEMANS POORT	1	X			X	X	X	X		X				

Some of the most useful information on the nature of the source rocks is provided by pebbles in the Vaartwell and Schoongezigt Formations. Both contain metamorphic rock inclusions such as slate, quartzite and vein quartz, the latter indicative of numerous tension joints, faults and segregation veins in the source area. Sediments are represented by shale and quartz wacke, coarser sandstone and grits, subarkose and black chert probably derived from the lower Cango formations or similar sediments in the original provenance.

Well rounded cobbles of granite in the Vaartwell Formation indicate a gradual uncovering of plutonic stocks and bosses, a deduction which can also be made from the shapes and types of heavy minerals in the Cango Group. Whereas most of the zircons (pinkish in colour) are well rounded and suggestive of reworked sedimentary cycles, those of the Uitvlugt Formation in particular, include bipyramidal or prismatic idiomorphs probably derived directly from sodium-rich granite or granodiorite plutons. The presence of acid igneous rocks is furthermore denoted by angular fragments or irregular grains of apatite and rutile, and by the abundance of the potash feldspars orthoclase and microcline.

Short, tabular crystals of tourmaline and angular diopside grains may have originated in the contact aureoles of the igneous intrusions, the tourmaline possibly representing pneumatolytic veins and the diopside a progressive thermal metamorphism of calcium-rich sediments.

The distribution of angular grains of augite, enstatite, ilmenite and magnetite throughout the succession is suggestive of basic igneous intrusions in the source area, while rare specimens of pyrope may have been derived from ultrabasic rocks. (A microprobe analysis of these red, well rounded grains reveals the presence of Mg, Al, Si and Fe).

Regional dynamothermal metamorphism of the source area is indicated by the presence of slaty and schistose rock particles in some thin sections, angular grains of epidote, and possibly almandine (dodecahedron faces, subconchoidal fracture, dark to black colour and microprobe analysis showing the presence of Fe, Al, and Si). Pebbles of gneiss also occur in the polymictic member.

Although polycrystalline quartz can be used as a criterion for distinguishing between metamorphic and sedimentary source areas (e.g. Blatt and Christie, 1967), the abundance of this textural type in the Cango rocks can be attributed mainly to the post=Cango deformation.

The tectonic setting of the source area is partly revealed by the geochemistry of the Cango sediments. On p. 47 the origin of the high K^+ and Ca^{++} content of these rocks has been considered somewhat problematical, and as the answer evidently lies in the clay minerals of the matrix, this aspect justifies more consideration.

The presence of a high percentage of matrix can be traced back to the following sources:-

i) Deposition as interstitial material with the original detrital grains.

ii) Deformation of argillaceous rock fragments.

iii) Alteration of feldspars by diagenesis, metamorphism or weathering.

No doubt these processes all played a role in producing the ultimate matrix, but the introduction of K^+ and Ca^{++} cations is most likely to have occurred during deposition or early diagenesis.

If the clay minerals of the matrix were transported by rivers from the source area, upon encountering sea water a chemical reaction would have taken place in order to reach equilibrium with the new environment. This halmyrolitic reaction in some cases can result in the uptake of K^+ and Na^+, as experiments by Potts (quoted in Keller, 1963) suggest. Weaver (1958) has shown that montmorillonite derived from the weathering of muscovite readily takes up K^+ in sea water and becomes illitic, whereas montmorillonite formed by the devitrification of volcanic glass does not respond similarly. This could explain the low Na^+ / K^+ ratio of the Cango quartz wackes as opposed to the high ratio of the "classical" greywackes (Fig. 24), and may therefore indicate a different, non-volcanic tectonic setting. As a decrease in the amount of fixed Mg^{++} and an increase in Ca^{++} also accompany the reactions in muscovite - derived montmorillonite, such a solution seems very likely, especially since granite and possibly pegmatite occurred in the provenance of the Cango Group.

During later diagenesis a partial conversion of montmorillonite to illite may occur because of a decrease in the expandability of montmorillonite at greater depths of burial (See Burst, 1959). During the higher temperatures of low-grade metamorphism mixed-layer montmorillonite - illite however may convert back to well-crystallized mica and chlorite, as they seem to have a stability field confined to the lower temperature ranges (See Perry and Hower, 1970). This may explain the present mineralogical composition of the rocks.

4. DIAGENESIS AND METAMORPHISM

4.1. INTRODUCTION

According to Pettijohn's usage (1972) diagenesis includes all those processes, chemical and physical, which affect the sediment after deposition up to the greenschist facies of metamorphism. In practice however, it is very difficult to distinguish between textures resulting from diagenesis or metamorphism and those which are the products of later weathering, and although every effort was made to obtain fresh samples, the effects of weathering cannot be overlooked entirely. Recent road cuttings such as in the Huis River Pass show that the relatively impermeable Groenefontein wackes, for example, are discoloured to a depth of at least 5 meters. It must be emphasized therefore, that some of the textures described do not necessarily have a purely diagenetic or metamorphic origin.

4.2. DIAGENESIS

The effects of diagenesis are very obscure in thin sections, largely because the dynamothermal metamorphism left an overprint on the textures which makes the evaluation of earlier changes very difficult. Furthermore, sedimentary structures indicate that the rocks may not have been fully consolidated at the onset of deformation, which means that all pore fluids did not escape during the earlier stages of compaction. Diagenesis and metamorphism in this case is therefore regarded as a continuous process only differing in its degree of intensity.

Of the chemical and compositional changes which affected the Cango rocks, only that of dolomitization can be considered a truly diagenetic process.

4.3. REGIONAL DYNAMOTHERMAL METAMORPHISM

4.3.1. TEXTURAL AND COMPOSITIONAL CHANGES

The most striking feature in the petrography of the Cango sandstones is the large amount of clay matrix, which averages

33% in about 60 thin sections. Undoubtedly this is not entirely of detrital origin. A closer examination of textures reveals that the original matrix has been increased by at least two processes: Although very few rock fragments are now recognizable as such, the diagenetic modification and squashing of argillaceous rock particles during deformation must have been responsible for at least part of the matrix. The other process is that of chemical alteration and replacement of detrital quartz and feldspar by sericite, chlorite and other clay minerals. Evidence for such reactions is abundant and includes relic or "phantom" structures of potash feldspar replaced by clay minerals along its cleavage and the etching or forming of embayments along the boundaries of detrital quartz grains.

An important aspect is the preference which the matrix shows in replacing quartz that exhibits strain features such as undulatory extinction, deformation lamellae and polygonization (Fig.66 (a)). Although the growth of these minerals generally proceeds along grain boundaries or cracks in quartz grains, rounded xenoblastic nuclei often start growing from the inside and along the contacts of strained and unstrained portions of single grains (Fig. 66 (b). Polygonization can be observed as proceeding from the grain boun= daries inwards until the whole fragment has a polycrystalline texture, which may lead to its being mistaken for a quartzitic rock particle. Ultimately the grains are shattered and replaced by muscovite and chlorite porphyroblasts along the regional cleavage. These features strongly suggest that alteration took place during or after deformation, but was of no great importance before.

Further effects of the deformation can be seen in segregation bands of mica and quartz, recrystallization along cleavage zones and elongated or stretched grains of quartz held intact by stylolite-like strands of mica. (Fig. 66(c)).

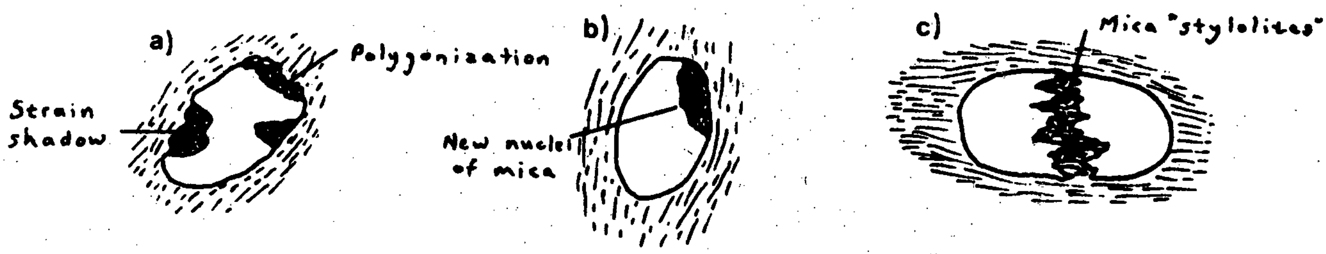


Fig. 66. Effects of deformation on grain replacements.

A continuation of the deformation process is shown by flakes and stringers of mica bent and twisted along post-crystalline slip surfaces or wrapped around detrital grains, and in segregation bands or deformation lamellae which have subsequently been folded.

Later generations of quartz may be observed to grow from partly replaced feldspar grains, veins of calcite and inside micro= geodes in limestone. One or two samples display an interlocking crystalline mosaic of quartz with sutured contacts, which results from secondary enlargement of the original grains by chemical precipitation.

The limestones of the Cango, though now composed of calcite and minor dolomite, were probably precipitated as aragonite. Modern oölites are exclusively aragonitic in composition, but the mineral being stable only at high pressures and inverting under normal conditions to its calcite polymorph. These sediments reacted to deformation largely by recrystallization. Under high magnification a weak foliation induced by the parallel elongation of irregularly bounded lensoid grains of calcite is seen, while larger, subhedral crystals in other cases concentrate along the cleavage planes. Some of the oölites show signs of plastic deformation in their stretched shapes and ends fading out into the matrix.

4.3.2. METAMORPHIC FACIES

Mineral assemblages in the Cango rocks show little variation. In the pelitic sediments of the Matjies River, Groenefontein and Gezwinds Kraal Formations the lower greenschists facies is represented by the assemblage: muscovite - chlorite - quartz - albite, although the occasional appearance of brown biotite in the Groenefontein rocks indicates that temperatures have locally risen above this subfacies. The mineral assemblage in these cases is: biotite - muscovite - quartz - (albite). Calcite may appear as a subordinate member.

The quartzo-feldspathic rocks of the Matjies River and Uitvlugt Formations contain the same mineral assemblage but muscovite and chlorite are less abundant, while calcite may take the place of albite.

Occasional limestone lenses in the Groenefontein Formation have been metamorphosed to schistose marble, recrystallization producing a coarser texture than normally encountered. Quartz may be an associated mineral, and its occurrence with dolomitic limestones in the Nooitgedagt Member indicates either low temperatures or high CO_2 pressures, for otherwise tremolite, diopside or grossularite would have formed.

During dynamothermal metamorphism the confining load is supplemented by directed pressure, and the Pt therefore can vary between 2 and 10 Kb. Pressures in the greenschist facies are usually in the range of 2 to 6 Kb which corresponds to a depth of 6 to 20 km, while temperatures are around 400° C (Winkler, 1976).

4.4. CONTACT METAMORPHISM

The influence of the basic intrusions on the country rocks was minimal, being restricted to induration and recrystallization into finer grained textures up to about 3 m from the contact. With the exception of occasional biotite no change in mineral composition was observed in thin sections.

5. STRUCTURE

5.1. FAULTS (Structural Map).

5.1.1. THE CANGO FAULT

This large normal fault, previously referred to as the Great Fault, Swartberg or Oudtshoorn Fault, forms the southern boundary of the Cango outcrop. It brings the Enon Formation and rarely Bokkeveld Shale in contact with Table Mountain Sandstone, the Schoemans Poort sediments and various formations of the Cango Group.

On aerial photographs the Cango Fault emerges as a thin, dark line in the easternmost part of the area, which in the field is seen to consist of dongas 10 to 20 m wide and overgrown with larger shrubs. Many of the small streams arising in the Swartberg vanish when crossing the line, thus indicating a porous, brecciated fault zone. The rather straight trend of the fault in this area also suggests a high dip angle.

Traced westwards, an occasional line of slightly larger shrubs is usually the only betrayal of its presence beneath the surface, the true contact mostly being obscured by talus from the poorly consolidated Enon Conglomerate. The latter dips gently towards the north and frequently displays angular, brecciated boulders near the contact.

Breccia (sensu stricto) is only rarely encountered in spite of the large displacement. This builds massive, quartzitic ridges 10-20 m wide. They are confined to the competent Table Mountain Sandstone and apparently form where the steepening fault plane leaves bedding surfaces in the latter and breaks through the formation.

In Schoemans Poort the Table Mountain Sandstone is criss-crossed by numerous joints and quartz veins over a distance of nearly 100 m from the fault line (Plate 5 (a)). Small low angle reverse faults and steeper normal faults are ubiquitous, commonly favouring thin shaly horizons in the massive sandstone.

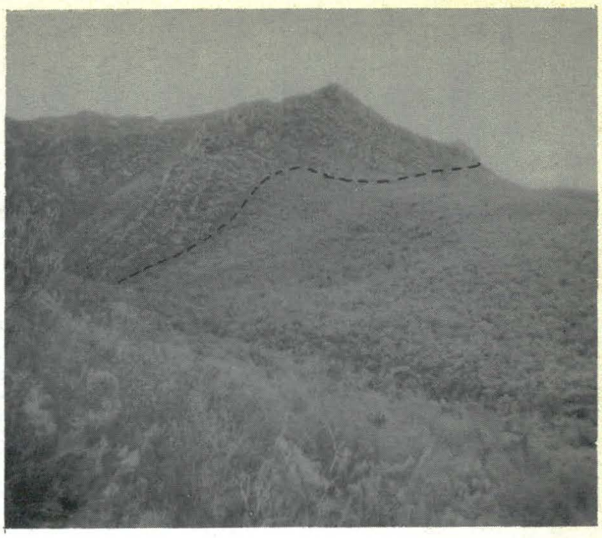
In the Huis River Pass west of the study area, the contact between the Table Mountain and Cango Groups dips at an extremely low angle, and according to previous workers this represents the fault surface. Although the Groenefontein wackes are brecciated along the contact, this is a common feature in other areas and can be attributed to differential movement between the competent Table Mountain Sandstone and underlying, less competent Cango Beds. This contact therefore does not necessarily represent the Cango Fault, and further work may be necessary in the Huis River Mountains to the south where, according to the Prince Albert 1:125000 map, breccia occurs in the Table Mountain Sandstone south of the contact.

To a large extent the position and orientation of the fault appears to be related to pre-existing structures in the Table Mountain Group. Like most of the other normal faults in the Cape Fold Belt it is situated on the southern limb of a northwards overturned anticline. This might be explained by the fact that the overturned flanks of anticlines in general are more complexly folded than the straight, drawn-out back limbs. Large scale slip should therefore occur more easily on these normal limbs where bedding surfaces are available as slip-planes. The course of the Cango Fault parallels the bedding in many places, as for example east of Potgieters Poort, where the Enon rests directly against bedding surfaces of the Table Mountain Sand= stone. Bedding-controlled slip could therefore explain the low angles exhibited by some Table Mountain Sandstone against Enon contacts, for example 35° east of Coetzees Poort (Plate 4 (a)) and 45° on the Waterkloof (22°42′E).

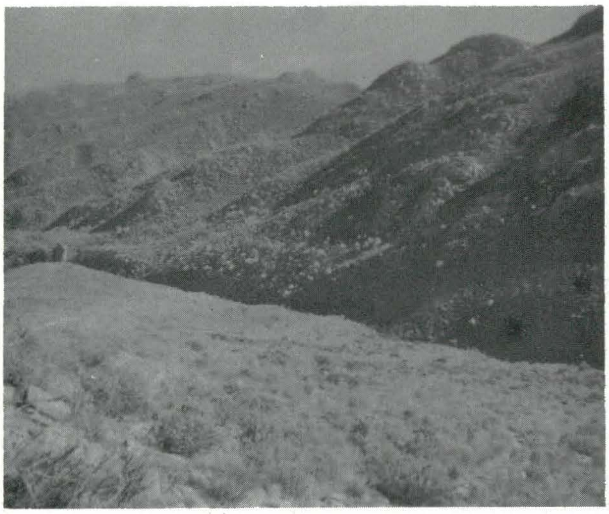
It has also been suggested that the present contacts between these formations may not represent the original Cango Fault, which probably had its main displacement during the Mid-Cretaceous. If the Enon was continually reworked during further growth of the fault, it may have in some cases transgressed onto the fault erosion scarp, thus concealing the fault plane completely (Fig 67 (b)). This view is substantiated by data from a bore-hole just south of the contact on Buffels Kloof (21° 53′E), which encountered Enon, Table Mountain Sandstone and finally Cango rocks without passing through any breccia or other indication of a fault.

PLATE 4.

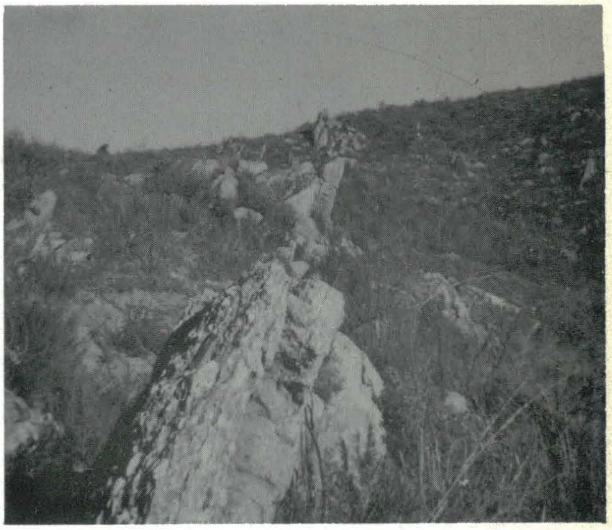
(a) Cango Fault east of Coetzees Poort. Enon Formation on right (south).



(b) Boomplaas Fault. Vinkenest= rivier. Looking west.

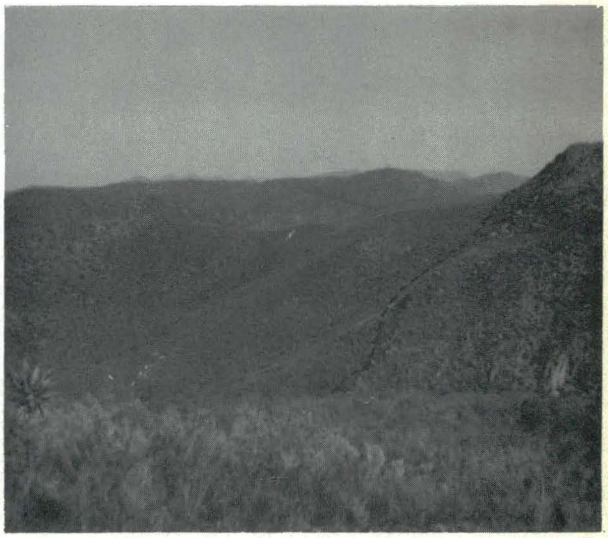


(c) South-facing erosion scarp of Wildehonde Kloof Fault, Central Plateau.



(d) Induration ridge of Wilde= honde Kloof Fault, Central Plateau. Looking west.

(e) Welbedagt Fault, Koetzers Kraal. Groenefontein Forma= tion on left (north).



(f) Stratigraphic effect of Welbedagt Fault, Koetzers Kraal. From left (north) to right: Vaartwel, Groenefon= tein and Uitvlugt Formations.

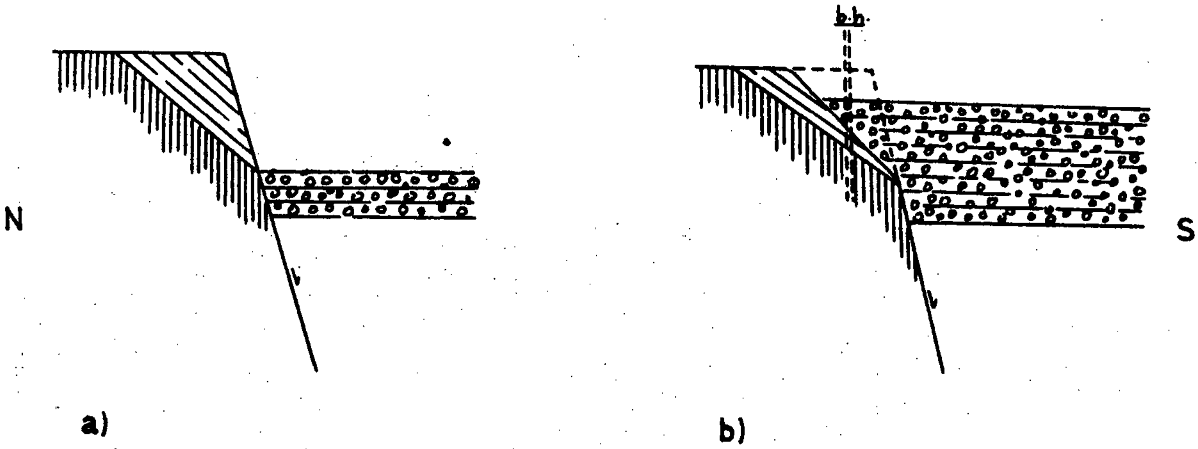


Fig. 67. (a and b) Possible development of Cango Fault at Buffels Kloof.

Minor displacements probably occurred for some time after the original rupture. Although a Tertiary terrace straddling the fault on Schoongezigt has suffered no measurable displace= ment, a very faint line continuing across the terrace can be seen on aerial photographs, suggesting that at least shattering occurred along the Cango Fault since the Tertiary.

The stratigraphic displacement of the fault in this area may be between 2000 and 4000 m, taking into account the estimated thickness of the Table Mountain Group and assuming that the rarely exposed shale just south of the fault (e.g. on Schoon= gezigt) belongs to the Bokkeveld Group.

5.1.2. THE BOOMPLAAS FAULT

Although conspicuous by its effect on the stratigraphy, this fault is very often obscure in the field. On some aerial photographs a faint line can be seen along the contact, which in most cases brings silty wackes or shale of the Gezwinds Kraal Formation adjacent to shales, sandstones or limestone of the Matjies River Formation. Where shales of both formations are brought together it is often impossible to locate the position of the fault correctly within 5 to 10 m. This becomes even more difficult in the east, where the similar silty wackes of the Groenefontein and Gezwinds Kraal Formations are in contact.

The only direct, knife sharp contact observed, is on the farm

De Cango ($22^{\circ}19'$E), where limestone of the Cave Member lies adjacent to Gezwinds Kraal quartz wacke. No breccia was encountered here.

That the contact is indeed a tectonic one, is however proved not only by the stratigraphic sequence, but also by the occurrence of local tectonic features near the fault line. On Vinkenestrivier (Plate 4 (b)) sandstone of the Matjies River Formation is cracked and cut by numerous quartz veins, while small iron-manganese gossan lenses with quartz veins occur on the fault south of the Cango Caves ($22^{\circ}13'$E) and elsewhere.

Tectonic conglomerate was encountered at the Wynands River and on Voorbedagt ($22^{\circ}05'$E). In the first case the Gezwinds Kraal wackes contain rounded fragments of limestone and quartz wacke which are orientated in the cleavage to form an a-lineation. The limestone "pebbles" are up to 60 cm in length and have pro= bably been derived from a lens in the underlying Matjies River Formation. They have a whiter colour because of recrystallization.

On Voorbedagt, shale of the Gezwinds Kraal Formation is considerably harder at the fault with small, flattened inclusions of shale again oriented in the cleavage. A narrow, sharp ridge has formed as a result of the induration.

On Nooitgedagt, thin limestone lenses occur near the fault in the Gezwinds Kraal and high up in the Uitvlugt Formation. These exhibit intense flowage and shear phenomena. Shale injections occur throughout the lenses and are also concentrated in shear zones some 30 to 60 cm wide. These zones are more resistant to weathering and form sharp, dark ridges 3 to 10 mm high. White, recrystallized limestone fragments are spindle shaped and cleavage-orientated, while darker, angular inclusions of up to 100 cm are fringed by shale injections. Fragments of shale and quartz veins complete the brecciated structure. Small outcrops of limestone parallel to the cleavage are also abundant in the faulted area. These float in a phyllitic matrix and are similar to the tectonic conglomerate described above, although the inclusions are much larger.

Contrary to Stocken, who interpreted the lenses to be later post-Cape upthrusts along shear cleavage planes, the present writer is inclined to regard their emplacement as simultaneous with the Boomplaas Fault. The overridden Cave Member probably provided the limestone.

Although previously assumed to be a thrust fault, there is no convincing evidence to support this view. A normal fault with a low southward dip could in fact have produced the same effects on the stratigraphy. However, the fact that it is situated on the northern, overturned limb of an anticline makes the first alternative more likely. Both solutions to the problem are given below (Fig. 68).

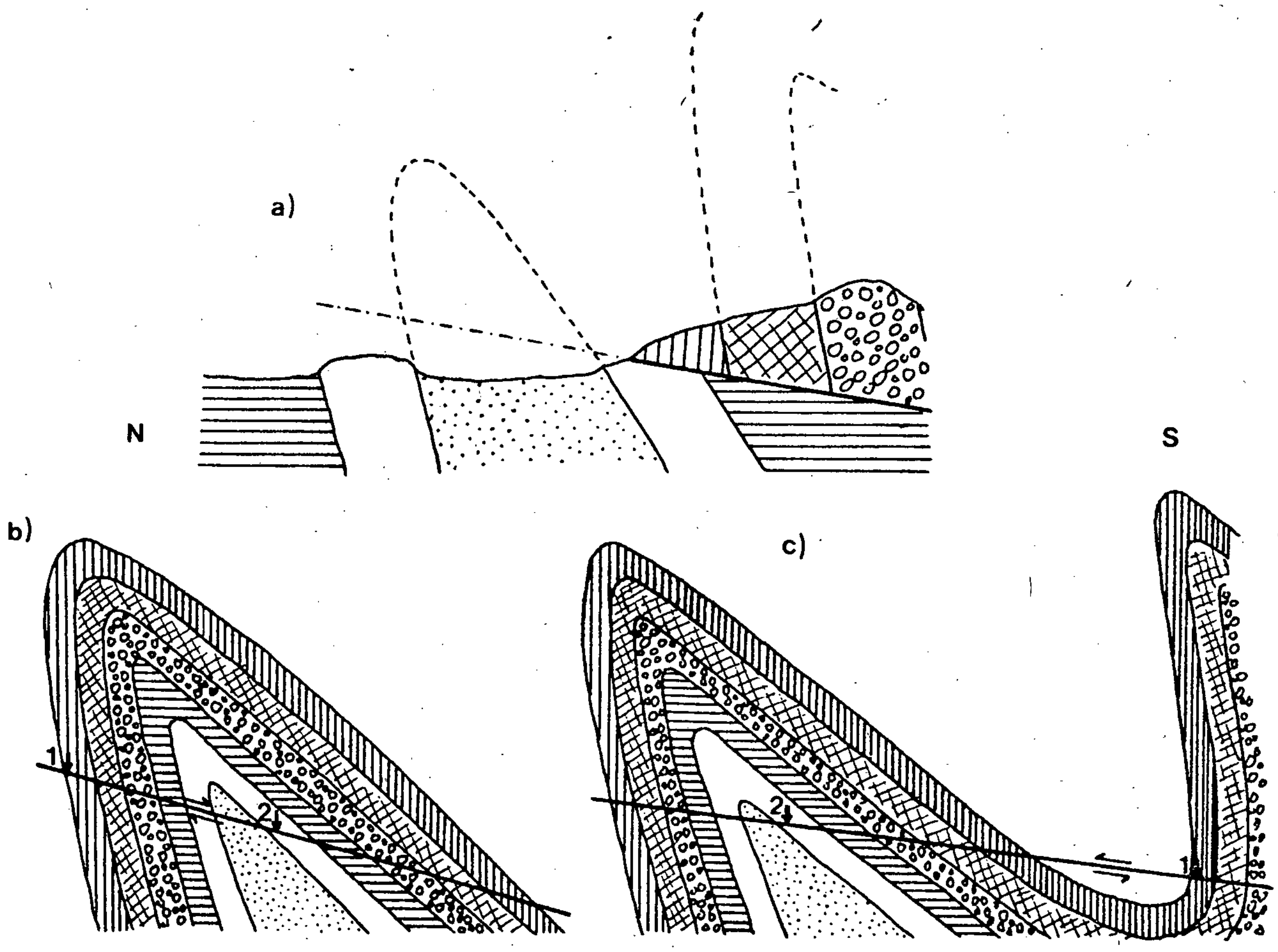


Fig. 68. Boomplaas Fault (a) as a normal (b) and thrust fault (c).

Taking into account structural trends in the area and the estimated thicknesses of the formations involved, a minimum displacement along slip of 3400 m for a normal fault or 3000m for a thrust is possible.

5.1.3. THE WILDEHONDE KLOOF FAULT

This pre-Tertiary normal fault extends some 4 km farther west than shown on previous maps. It is parallel or sub-parallel to bedding on both sides for most of its course, with overturned wackes of the Uitvlugt Formation in the south brought into contact with the Vaartwell Conglomerate on the northern footwall. At $22^{o}15'$E it cuts the Boomplaas fault, causing the Uitvlugt to lie directly against the Matjies River Formation. Farther east crossbedded sandstones appear on both sides of the contact, but with the Danzers Kloof Member on the upthrow against the Rooiberg Member on the downthrow side. Beyond $22^{o}25'$E the fault possibly continues beneath the alluvium of the Nels River.

The greater resistance of the Vaartwell Formation to weathering produces a small south-facing fault-line scarp on the Central Plateau (Plate 4 (c)), while induration on both sides of the contact is also responsible for a narrow intermittent ridge at places (Plate 4 (d)).

Although the actual contact is invariably sharp and breccia or other phenomena usually associated with a fault are absent, field evidence is not entirely lacking. At the Wynands River the crossbedded sandstones are sericitized on the contact, the streaks of mica forming an a - lineation which is parallel to the 80^{o} dip of the fault. Small quartz veins are common.

In Schoemans Poort, just south of the fault, a zone of 30 m in the Uitvlugt Formation shows intense cracking and numerous quartz veins (Fig. 69). Shearing and small displacements are abundant, the latter generally low angled and mostly restricted to shaly horizons. These features are strongly reminiscent of the effects of the Cango Fault in the Table Mountain Sandstone farther to the south (Plate 5 (d)).

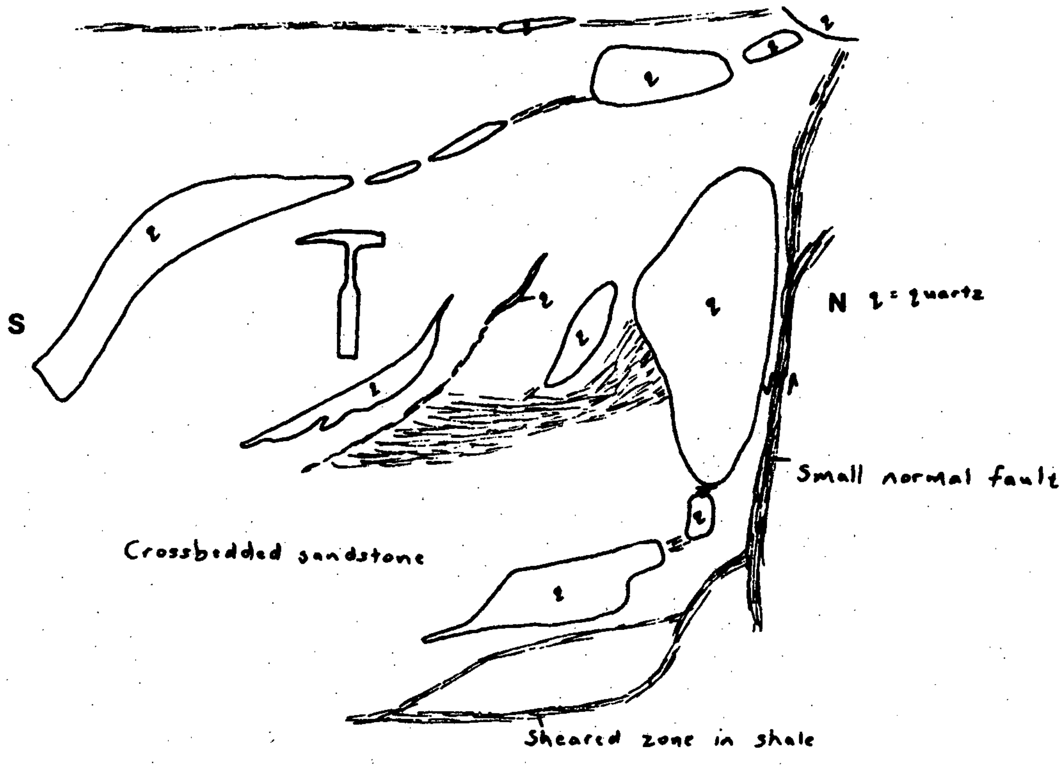


Fig. 69. Shatter zone of Wildehonde Kloof fault, Schoemans Poort.

Although the Wildehonde Kloof Fault is undoubtedly a normal one, the mechanism responsible for the juxtaposition of sequences is somewhat controversial. According to Stocken, the repetition of overturned sequences necessitates the presence of a thrust fault allowing an anticline to ride forward over the earlier thrust sheet from the south (Fig. 70 (a)) but this is not necessarily the case.

Smaller faults in the Cango are often observed to parallel bedding planes for some distance, and after breaking across the foliation, finally return to parallelism with the layers (Plate 5 (e)).

As illustrated in Fig. 70 (b), this could also be the case with the Wildehonde Kloof Fault, and as there is no evidence for the existence of a thrust sheet, the alternative, more simple solution is preferred. A third explanation involves two normal faults, the first being a low angle fault cut by a later, high angle fault parallel to and controlled by the bedding in both formations. (Refer to Fig. 70 (a), reversing the direction of stage (1).

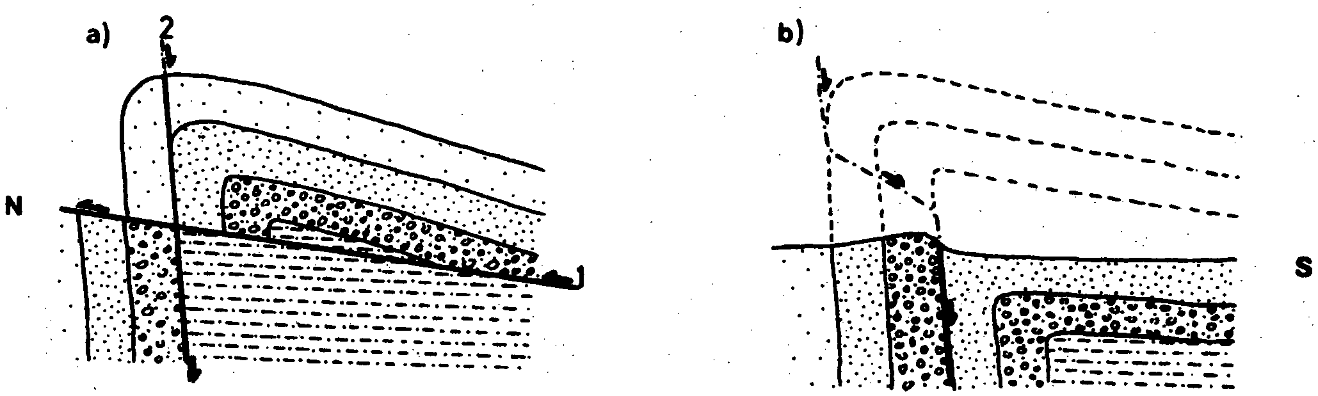


Fig. 70. Wildehonde Kloof Fault:
(a) Solution proposed by Stocken, 1954.
(b) Alternative, normal fault solution.

5.1.4. THE WELBEDAGT FAULT

Apparent in most cases only by its displacement of the strati= graphy, this fault is extremely difficult to trace where similar rocks appear on both sides. This may account for previous incorrect interpretations of its course on Koetzers Kraal in the west, where the actual continuation of the fault lies farther to the north than indicated on existing maps. The breccia along the Droë River in the south probably belongs to a tear or hinge fault similar to the one crossing the farm Abrahams River about 3 km to the east, as the tectonic contact of Uitvlugt against Gezwinds Kraal sandstones exhibits the same strike.

In the east grits of the Schoemans Poort Formation are in contact with the silty Groenefontein wackes, and the marked difference in lithology and topography makes it an easy task to follow the fault trace, even though the actual contact is nearly always obscured by talus.

On the farms Rykdom and Koetzers Kraal the Welbedagt Fault causes a repetition of formations (Plate 4 (f), Fig. 71). The best outcrops, proving without doubt the tectonic nature of the contact, were found on these farms. Breccia and kink folds in the Groenefontein wackes on Rykdom give the fault a dip of 35 to 50° south.

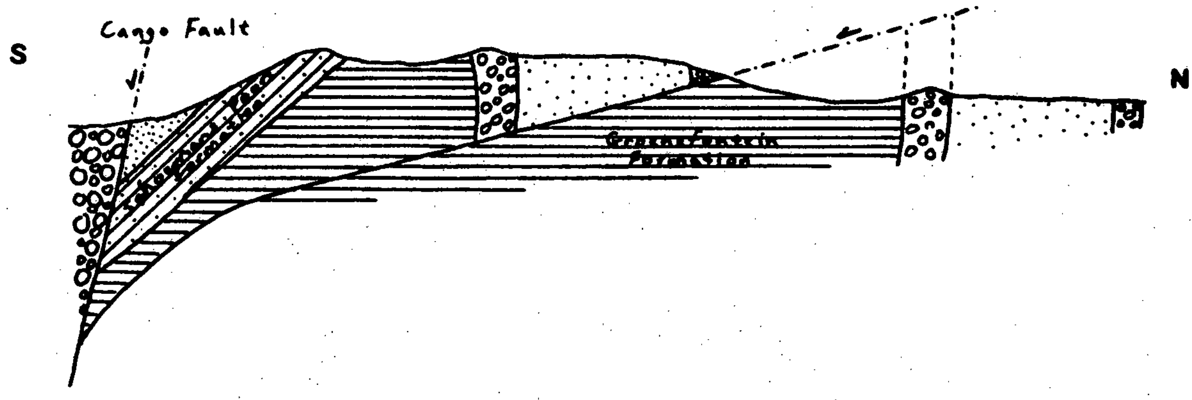


Fig. 71. The Welbedagt as a normal fault on Rykdom.

At Koetzers Kraal the breccia is up to 3 m wide in places, while small quartz veins are observed in the fault zone. An excellent outcrop on this farm shows a very sharp contact between Groenefontein quartz wacke and sandstone of the Uitvlugt Formation. (Plate 4 (e)). Only a small lens of breccia is developed. The fault dips 60°S and is parallel to bedding in the Groenefontein Formation, but makes a sharp angle with its cleavage. The crossbedded sandstones have a horizontal to slightly northerly dip, but within 25 m of the fault a sudden steepening to 54°S is observed, the bedding becoming almost parallel to the fault plane. Although not necessarily so, this may result from downward shear of the hanging wall.

Another well exposed contact is found on the eastern side of the Matjies River valley (22°02'E), where massive beds of the Uitvlugt Formation on both sides of the fault are cut off against its surface. The latter shows evidence of shearing, slickensides and the ever-present quartz veins. Higher up on the valley wall the fault runs into the conglomerate of the Koetzers Kraal Klippe, which is displaced slightly downwards in the south.

Although previously regarded as a thrust fault, the present writer is in favour of a normal fault solution on the following grounds:

i) The numerous drag features which according to Stocken indicate a clear northward thrust are very doubtful evidence. "Drag" of the type described was readily found, but can in all cases be attributed to northward creep of the incompetent wackes on steep inclines. This is a very common feature in the Cango rocks and can be found on many similar slopes.

ii) The Schoemans Poort Formation is also affected by this fault, which proves it to be considerably younger than the main deformational phase in the Cango.

iii) Two diabase dykes are offset by the fault at the Wynands River. The displacement, which is relatively small in this area, can hardly be reconciled with thrusting but indicates down-faulting to the south.

iv) Although the dip appears to vary considerably, steep dips such as the 60° measured on Koetzers Kraal is unusual for the type of thrust sheet proposed.

v) Both the Koetzers Kraal and Welbedagt Klippes are cut by the fault and apparently displaced downwards in the south, although the actual amount of movement is very small in these areas.

As a result of the hinge action of the Droë River Fault, the rocks south of the Welbedagt Fault to the west of the former are part of the Potgieters Poort Nappe, while those east of the hinge crop out below the thrust plane of the sheet. In this area two estimations of the amount of dip displacement (Koetzers Kraal and Rykdom) give a similar figure of 1100 m in spite of the differences in dip (60° and 35°). It is suspected however, that the amount of slip decreases from east to west, the large displacement on Koetzers Kraal being the result of an independent, additional slip along the northwest striking tear fault crossing Abrahams River.

West of the Koetzers Kraal Fault on the farm Drooge Kraal ($22^{\circ}00'$E), the effect of the Welbedagt Fault dies out and the tectonic contact passes into a normal, gradational one.

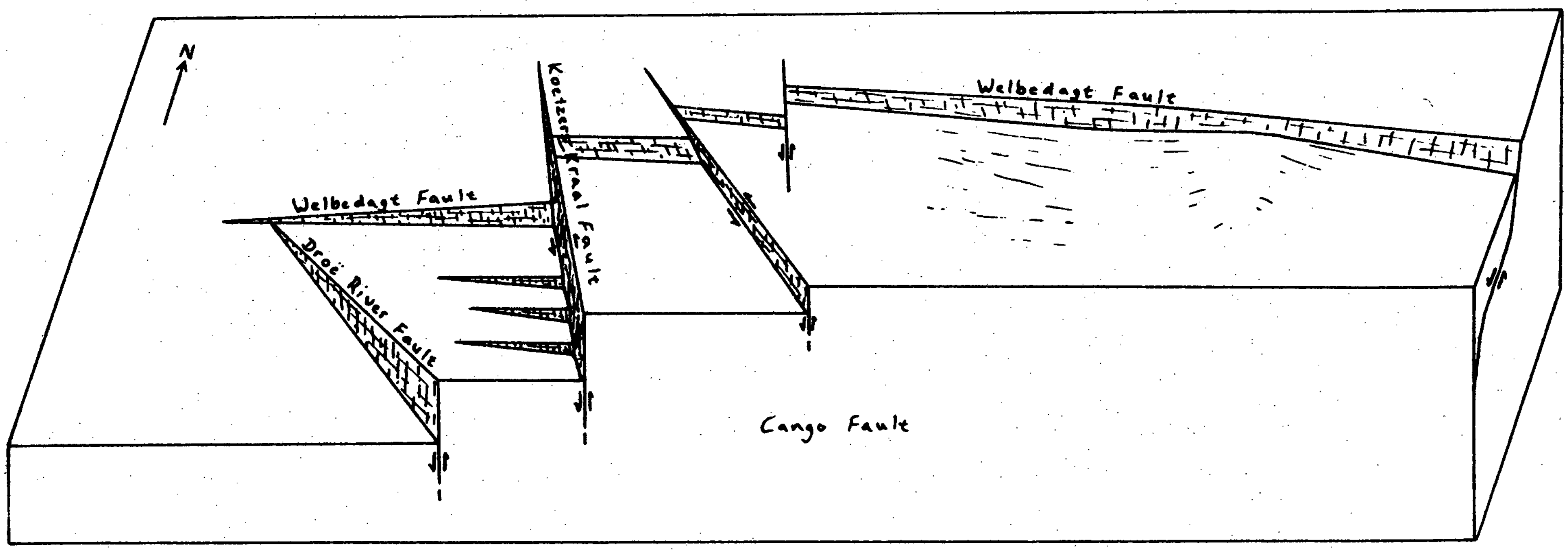


Fig. 72. Diagrammatic interpretation of the Welbedagt and transverse faults in the south.

5.1.5. THE TRANSVERSE FAULTS IN THE SOUTH. (Fig. 72).

Two sets of faults, striking 020° and 320° respectively, exert considerable influence on the stratigraphy between longitudes 22°00′and 22°06′E.

5.1.5.1. THE KOETZERS KRAAL FAULT

Although this feature has the appearance of a tear fault in plan and was previously regarded as such, the stratigraphic dis= tribution on both sides of the fault is probably the result of a scissor action combined with oblique slip towards the south. A wrench fault possessing the large offset shown by the Koetzers Kraal in the south would have displaced at least to some degree the contacts in the north. The fact that older rocks crop out in the east, however, is indicative of a descending western wall with the axis of rotation located to the north. Fig. 73 shows why a small amount of oblique slip is necessary to explain the southward shift of the Welbedagt Fault trace.

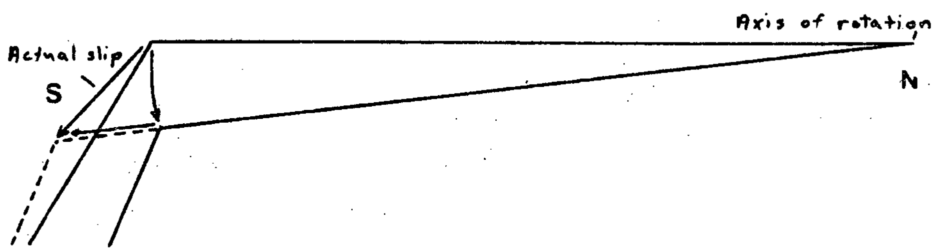


Fig. 73. Actual displacement on the Koetzers Kraal Fault.

5.1.5.2. THE DROË RIVER FAULT

As mentioned on p. 107 , breccia was encountered along the Droë River west of the Koetzers Kraal Fault. This zone of shattered sandstone, much wider than is characteristic of most faults in the Cango, strikes north-west and brings the Gezwinds Kraal Formation in contact with Uitvlugt sandstones. Structural trends in this area preclude the possibility of a normal, sheared contact,while a glance at the geological map will show why, with the area to the west of the contact belonging to the root zone of the Potgieters Poort Nappe, and the area east thereof being beneath the thrust plane, a scissor action similar to that of the Koetzers Kraal Fault is postulated.

5.1.5.3. AGE-RELATIONSHIPS

Although the Welbedagt and Koetzers Kraal Faults are considered to be contemporaneous with the development of the Cango Fault, the Droë River Fault is evidently much older, as it does not affect the strip of Table Mountain Sandstone in the south. The effects of the other transverse faults cannot be traced farther south and whether they run into the Cango Fault is uncertain. However, they displace the Welbedagt Fault and therefore probably result from further movements on the Cango Fault. The similar strike directions of pre- and post-Cape faults suggest that pre-existing planes of weakness or faulting had been re-activated during the later stress phases.

5.1.6. THE POTGIETERS POORT NAPPE

Detailed mapping on the Central Plateau led to the discovery of several remnants of a large overthrust sheet which had not previously been recognized. The outliers occur between $21^{o}50'$ and $22^{o}11'$E over a distance of 33 km. In all cases conglomerates of the Vaartwell Formation have been thrust over either the Groenefontein or Uitvlugt Formations.

5.1.6.1. THE COETZEES POORT REMNANT

This is by far the largest of the remnants, extending from Kweek Kraal ($21^{o}50'$E) over a distance of more than 15 km to Koetzers Kraal in the east and crossing Coetzees and Potgieters Poorts on the way.

In Potgieters Poort the sheet has polymictic conglomerate at the base, with elongated pebbles of quartz wacke, shale and quartz. No inclusions resembling the underlying crossbedded sandstones were observed. Upwards the polymictic conglomerate gives way to the monomictic quartz pebble type. From north to south there is a complete transition from conglomerate to crossbedded sandstones and the interbedded sandstone and shale of the Gezwinds Kraal Formation.

In the poort, near the cattle gate, the Uitvlugt Formation underneath the nappe dips south at an angle which varies from 35°S to nearly horizontal. Crossbedding proves the beds to be right side up. The contact is almost without exception knife sharp, with breccia or gouge only very rarely encountered. In a few places the overlying sheet appears to erode the sandstones, but in general the fault surface is parallel to bedding in the latter. A thin shale horizon has in some places acted as a slip-surface. Drag on the very faint, but still recog= nizable regional cleavage shows the sheet to have been thrust northwards (Fig. 74).

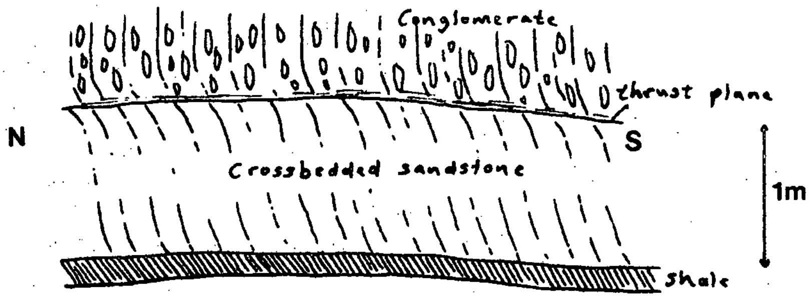


Fig. 74. Drag on shear cleavage at base of Coetzees Poort Remnant, Potgieters Poort.

On Vogelfontein (21°58′E) a quartz vein, 3-5 m wide, separates the crossbedded sandstones from conglomerate along the thrust. While the attitude of cleavage in the sandstones is 100°/30°S, an orientation of 112°/70°S was measured in the conglomerates. About two km east of this contact the nappe rests directly on the Groenefontein Formation. In the vicinity of the Wynands River quartz wacke of this formation is cracked and crumpled near the contact, while small quartz veins are ubiquitous.

West of Potgieters Poort a reconnaissance survey revealed that, because of the topography, the remnant splits into two parts, the northernmost of which is illustrated in (Fig. 75). From a distance the unconformable relationship in this part of the nappe is clearly demonstrated.

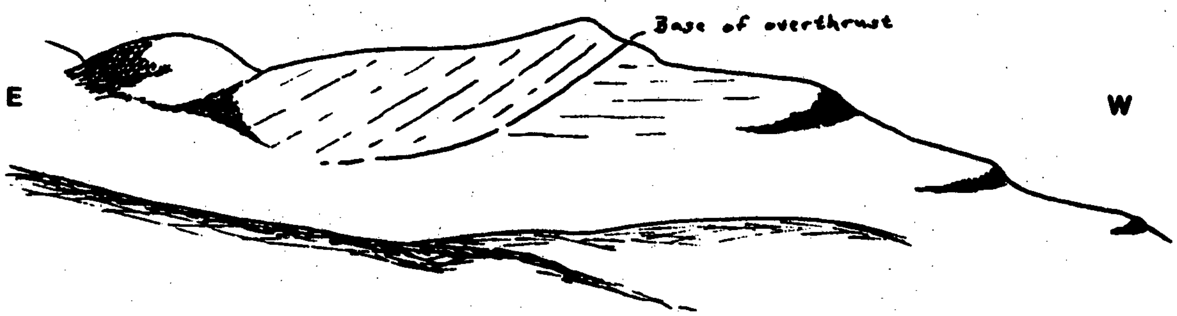


Fig. 75. Coetzees Poort Remnant west of Potgieters Poort, viewed from a distance.

In the south the nappe is cut by the Drooge Kraal diabase sill. The fault plane in this area steepens markedly and a "mélange" contact with the underlying Groenefontein Formation was observed.

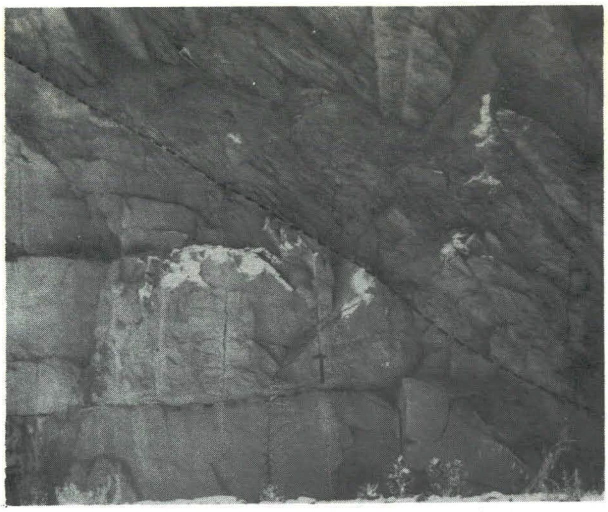
An excellent contact in Coetzees Poort displays massive bedding of the Uitvlugt Formation cut off by the thrust and overlain by highly sheared polymictic conglomerate (Plate 5 (a)). The dip of the fault plane changes over a short distance from 32° S to 10° N and again to 20° S.

5.1.6.2. THE KOETZERS KRAAL KLIPPE

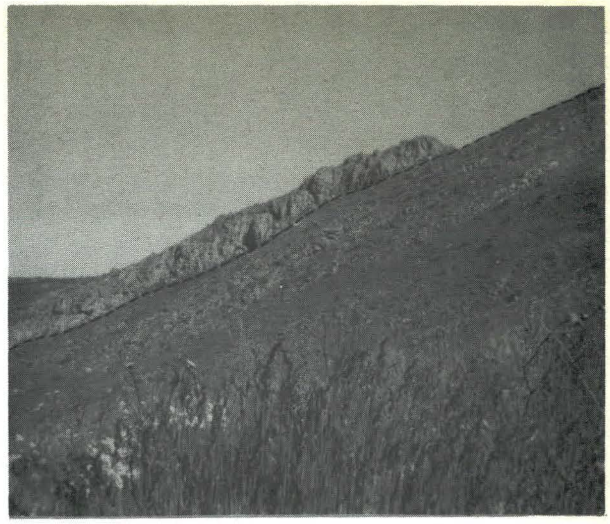
This outlier is terminated in the east by the Koetzers Kraal Fault and extends for 4 km in a north-south direction. The thrust plane, which usually dips south at a few degrees, steepens to 40° in the north. The best proof of tectonic relationships can be found here.

Seen from the west and north, the massive bedding in the conglomerate (which represents the northern flank of an anticline) is sharply terminated against the fault surface, while the gently south-east dipping beds of the underlying sandstones are also cut off by the fault. (Fig. 76).

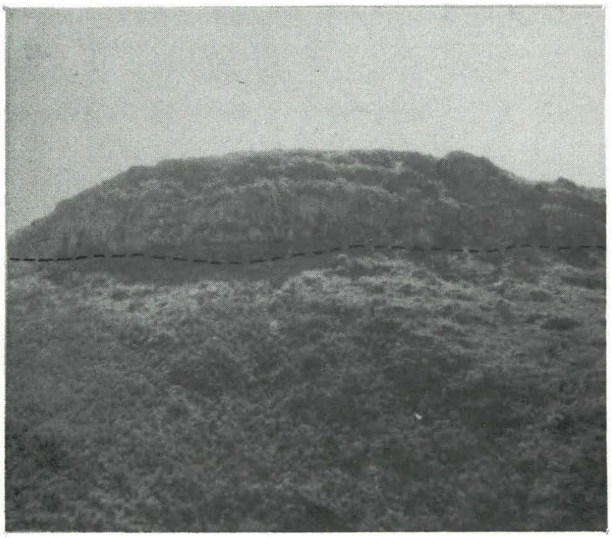
PLATE 5



(a) Potgieters Poort thrust, Coetzees Poort. Looking east.

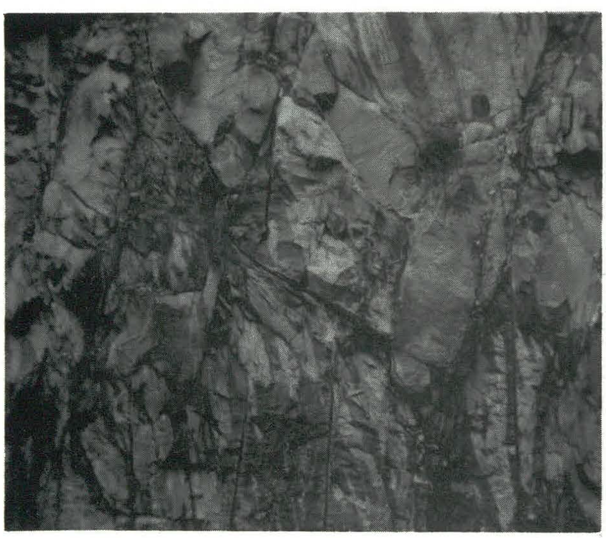


(b) Northernmost part of Koet=zers Kraal Klippe, Abrahams River. Looking west.



(c) Welbedagt Klippe. Looking east.

(d) Shatter zone in Table Moun=tain Sandstone north of Cango Fault, Schoemans Poort. Looking east.



(e) Small normal fault changing dip from parallel to bedding across bedding and parallel again. Schoemans Poort, looking east.

(f) Brecciated zone (white) in Nooitgedagt sandstone, Rau=benheimer dam. Looking east.

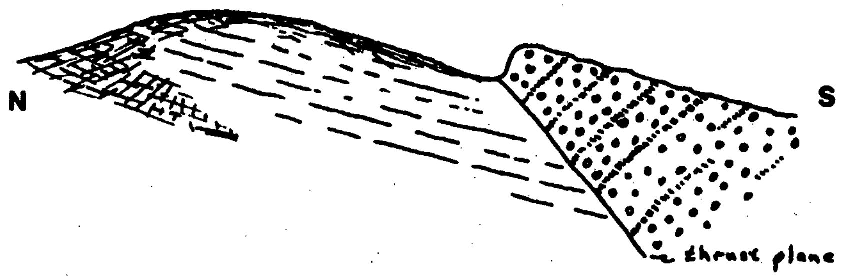


Fig 76. Eastward view on Koetzers Kraal Klippe, Wildehonde Kloof.

An east-west striking hinge fault complicates the interpretation of structures in this part of the klippe. Viewed from the north, the remnant appears to wedge out, with crossbedded sandstones both on top of and underlying the nappe, thus creating the impression of a huge conglomeratic lens in the sandstone. (Fig. 77).

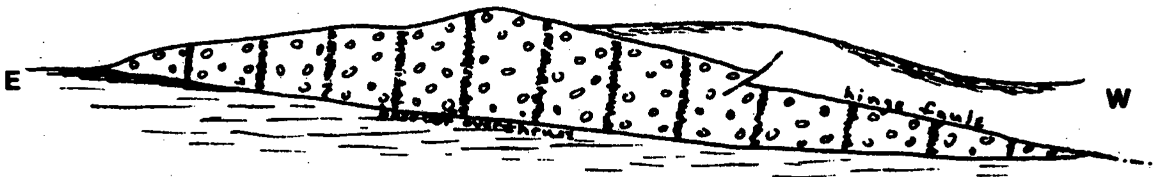


Fig. 77. Southward view of klippe on Gezwinds Kraal,

It appears that after overthrusting of the sheet the northern extremity was hinged down (Fig. 78), which effectively protected this part of the nappe against further erosion, while the southern parts (with exception of the eastern remnant similarly protected by the Koetzers Kraal Fault) were stripped away.

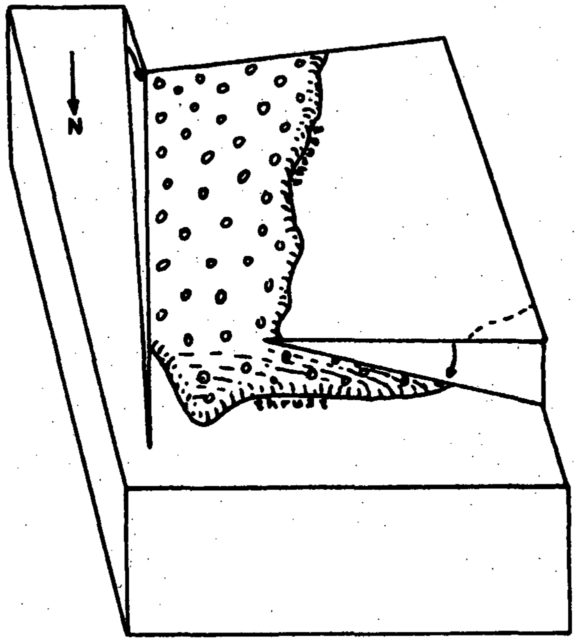


Fig. 78. Diagrammatic interpretation of faults in northern part of Koetzers Kraal Klippe.

On closer inspection the hinge fault is very sharp and mostly parallel to the bedding (which is nearly vertical at this point) in both formations. One of the small overfolds common on the Central Plateau possibly provided a plane of weakness in the underlying crossbedded sandstones along which bedding-slip could occur.

In the southern part of the Koetzers Kraal Klippe a few parallel east-west trending normal faults also complicate the outcrop pattern. An analysis of the stratigraphy here shows that these displacements are arranged in a step-like manner, each causing a northward and eastward ascending of the stratigraphy. Small, steep-sided stream channels forming tributaries of the Matjies River mark the courses of these faults, which probably terminate against the Koetzers Kraal Fault in the east. From this it seems likely that the hinge action of the latter triggered similar mechanisms on the descending wall, which caused displacements of the type illustrated in Fig. 72.

The thrust plane of the nappe is always sharp and signs of movement are rare. However, an excellent exposure just east of the Matjies River provides direct evidence for a tectonic contact. This outcrop can be reached by following the Welbedagt Fault up the eastern side of the river. Underneath a large overhang of conglomerate there are small drag folds, cuspate structures and injection dykes of sandstone in shale (Fig. 79), quartz filled tension gashes indicating the orientation of the strain ellipse, and a sheared slip-surface in shale changing laterally into a "sedimentary" type of contact. In spite of the massive sheet of conglomerate sliding towards the north, the sheared zone is only a few to 15 cm wide.

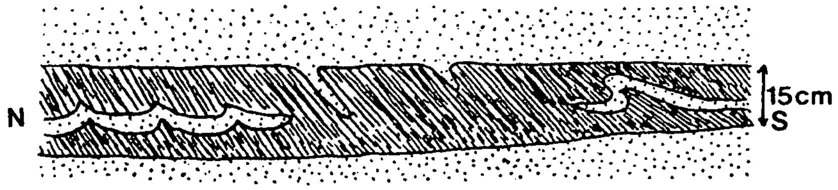


Fig. 79. Cuspate structures, drag folds and injection dykes at base of Koetzers Kraal Klippe, east of Matjies River.

An interesting, but potentially controversial feature, is that a bed devoid of pebbles and closely resembling the Uitvlugt sandstones, occurs at the base of the conglomerate above the main slip-surface (Fig. 80). This bed is cut off by the fault in the southern part of the outcrop and its upper contact closely resembles some of the other contacts found everywhere along the nappe structures. Again this is knife sharp, with small lenses similar in appearance to the lower slip-surface. Although this bed may be an elongated lens in the conglomerate and thus part of the original sheet, it may also have originated in the following way: The northward sliding nappe, encountering a large amount of friction along its lower contact with the Uitvlugt sandstones, passed an area in the latter where a shale band was located about a meter below the fault plane. Because of less resistance in this horizon the fault broke through the upper sandstone bed and, incorporating it into the lower part of the sheet, proceeded along the shale band.

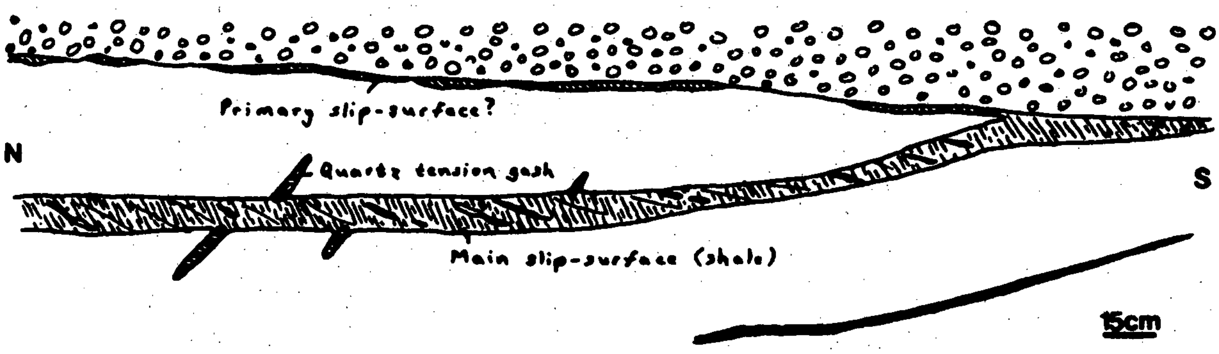


Fig. 80. Two slip-surfaces at base of klippe at Matjies River.

5.1.6.3. THE WELBEDAGT AND OTHER KLIPPES

Covering a little more than 1 km^2, the Welbedagt Klippe (Plate 5 (c)) exhibits the same basic features displayed by the others. Among those is a flat erosion terrace above the fault plane where the overlying sheet had been eroded away. In this case it is more than 100 m wide, suggesting some induration of the rocks adjacent to the slip-surface. Bedding is nearly horizontal and parallel to the fault.

Two other small remnants cap the higher hilltops farther east. In the larger one on Oorsprong van Kansa (22°08′E) the fault plane has a dip of 25° S, while a very small remnant on the eastern border of the farm shows numerous quartz veins on the contact.

5.1.6.4. DYNAMICS OF THE POTGIETERS POORT OVERTHRUST

That large displacement of several km along a gently inclined thrust plane is possible, has been shown by Hubbert and Rubey (1959). The normal component of effective stress is sufficiently reduced during periods of orogeny to lessen the friction which makes such movement impossible under normal conditions. This

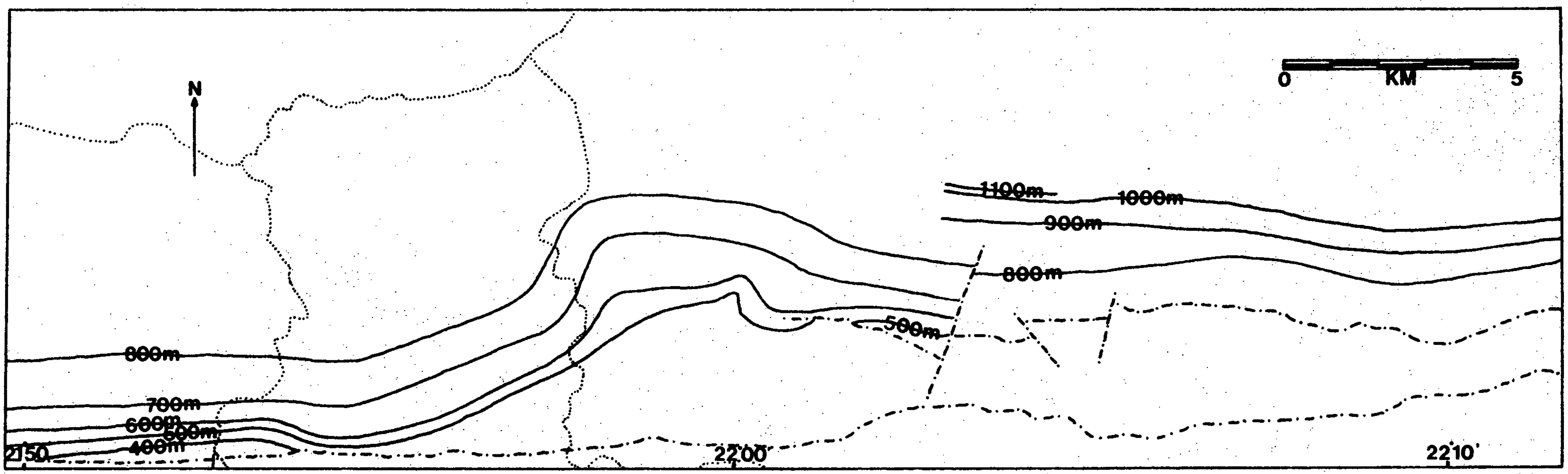


Fig. 81. Structure Contours Drawn On Base Of Potgieters Poort Nappe

is accomplished by a build-up of abnormal pressures in horizons possessing low permeability (e.g. shale), where the fluid pressure cannot be rapidly dissipated under compression. The critical value of shear stress required to produce sliding is accordingly reduced under these conditions and low angle thrusts can consequently occur under stresses of only moderate magnitude.

The actual displacement can be attributed to either compressive thrust or sliding down an inclined plane. Structural features cannot be regarded as decisive (De Sitter, 1964) and evidence must therefore be of a circumstantial nature. For sliding a forward slope must be demonstrated in the field, while a downward plunge of the fault plane at the rear can be regarded as fair evidence of a compressive thrust mechanism.

A structure contour map (Fig. 81) of the sheet base has been compiled on 1:50000 topographic maps. The result shows a scoop-shaped structure interrupting the general east-west strike in the centre of the nappe. West of the Koetzers Kraal Fault the contours show a southward steepening of the thrust plane, a feature which can also be observed directly on Drooge Kraal and in Coetzees Poort. It is possible that this plunge represents the root of the nappe, and could therefore be regarded as evidence for "thick-skin" tectonics.

5.1.7. OTHER FAULTS

Small reverse and normal faults are very abundant in the Cango rocks, but are only conspicuous in road cuttings or in rare instances as siliceous breccia, iron-enriched ridges or small displacements on the surface. Transverse faults are usually of the left lateral type and orientated parallel to the shear joint system.

Breccia is very rare, except in some instances where poorly bedded, massive sandstones of the Matjies River Formation are involved. At the Raubenheimer dam for example, a normal fault has produced a brecciated, powdered zone 1 m in width (Plate 5 (f)).

Faults in the Uitvlugt and Vaartwell Formation are without exception very sharp, the presence of quartz veins being their only betrayal on the surface. The direction of movement can often be inferred from tension cracks in these veins, as for instance on Buffels Kraal (22°21'E) (Fig. 82).

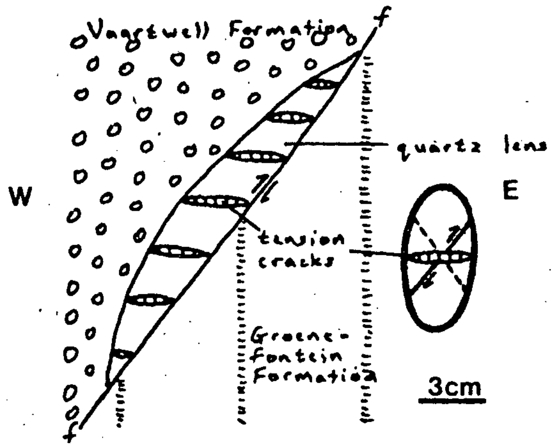


Fig. 82. Direction of shear inferred from tension cracks in quartz veins, Buffels Kraal.

5.2. JOINTS

The rose diagram (Fig. 83) shows no clear resolution into different sets of joints.

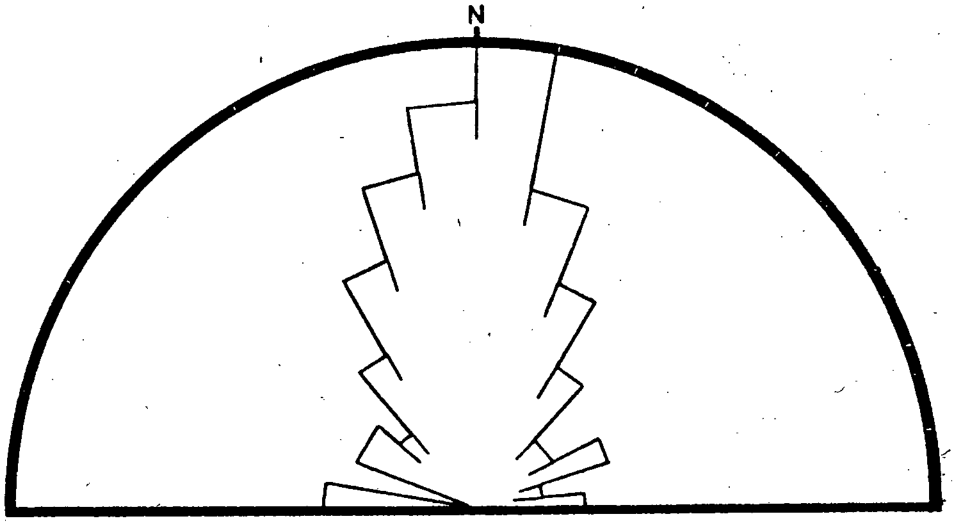


Fig. 83. Rose diagram of joint strike - frequency.

Field observations however support the existence of at least two generic types, viz. shear and tension joints. The latter type, as experiments by Hubbert and Willis (1957) suggested, develop at right angles to the axis of least principal stress (σ_3). The two sets of tension joints in the Cango are at right angles to each other, being orientated north-south (ac) and east-west (ab) respectively. The ac-system shows up on aerial photographs as straight lines frequently followed by streams and is nearly vertical in dip, while quartz veins often characterize the ab joints. These dip south at angles varying from 50° to 90° and are often parallel to the regional cleavage and bedding planes, especially on the overturned limbs of anticlines. The assumption that this set represents longitudinal extension fractures on fold crests does not apply therefore, an alternative explanation being needed for the paradox of rectangular tension features.

Shear joints in the Cango are orientated at angles of less than 45° (commonly $10-30^{\circ}$) with the axis of greatest principal stress (σ_1). Their nature is clearly demonstrated by the fact that some actually display shear. These shear joints are evidently younger not only than the Table Mountain sediments, into which they pass with a slight deflection in strike, but also cross or even displace the possibly Mid-Cretaceous (Stocken, 1954) Wildehonde Kloof Fault. While these joints and associated oblique faults are evidently related in orientation to the stress fields of pre- and post-Cape deformation, some developed after the youngest compressive phase and even post-date the tensional forces that influenced this subcontinent.

N.J. Price (1959), proposes that joints in general develop during post-tectonic uplift because of lateral expansion releasing the residual stresses built up during the later phases of tectonic compression. (Fig. 84).

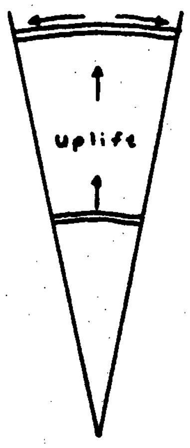


Fig. 84. Lateral expansion as a result of tectonic uplift. (After Price, 1959).

This is accompanied by an inversion in the directions of principal stress, so that conditions for the development of shear and tension joints are successively satisfied as uplift proceeds.

The orientation and frequency of joints are related to the residual stresses in the rocks, which in turn depend on the original state of strain induced by tectonism. The tectonic axes can therefore be inferred from the orientation of joints even if these are actually younger than the deformation. Considering these principles, the progressive evolution of joints in the Cango may be explained in the following way. (Fig. 85).

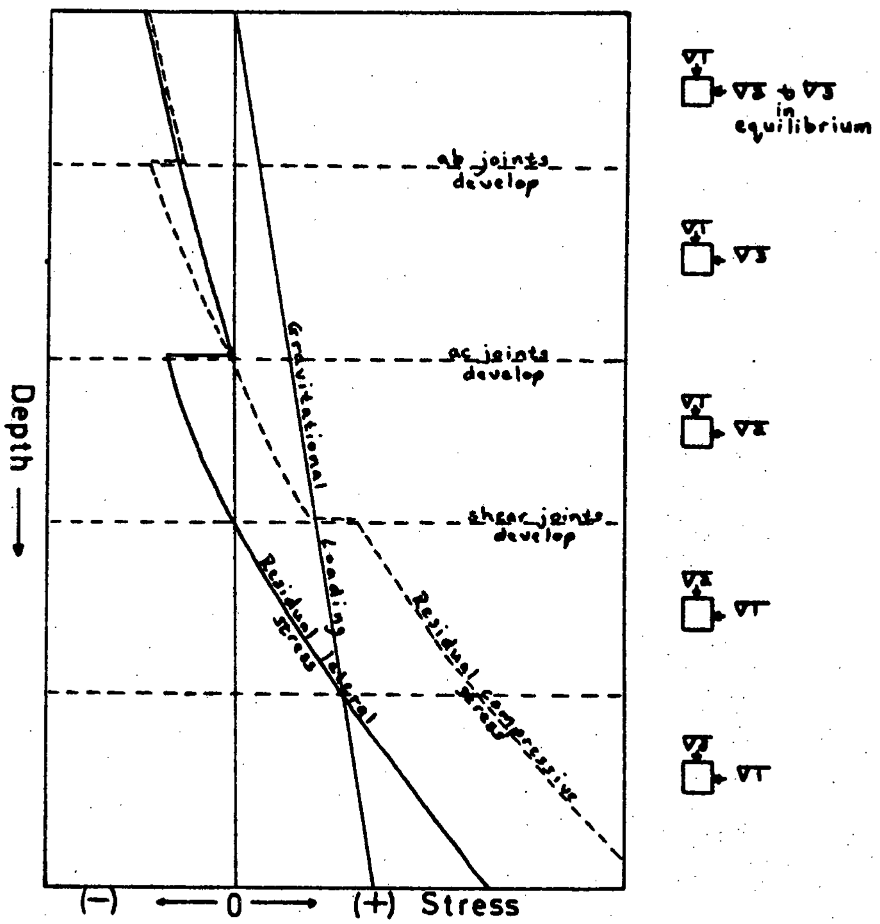


Fig. 85. Development of joint systems as a result of uplift. (After Price, 1959).

With σ_1 originally acting in a north-south direction, σ_2 orientated east-west and σ_3 vertical, initial uplift caused σ_2 and σ_3 to change position, σ_2 becoming vertical and thus satisfying one of Mohr's (1914) conditions for the development of wrench shear planes. Futher uplift produced joints of this type, leading to the partial dissipation of residual stresses while gravitation caused σ_1 to assume the vertical position. Still further uplift favoured the development of ac tension joints, σ_2 and σ_3 again changing positions and ab joints forming as a result, thus explaining their orientation at right angles to each other.

5.3. CLEAVAGE

5.3.1. GENERAL DESCRIPTION

Two secondary foliations, slaty and fracture cleavage, are present in the Cango Group. Although the fracture cleavage, which effected non-penetrative, planar discontinuities, is usually subparallel to the slaty cleavage, it is in some thin sections seen to cross and displace the older foliation. Part of this cleavage is probably the result of post-Cape deformation, but very often it only reflects differences in competence between beds.

Slaty cleavage produces a planar fabric that is penetrative throughout the rock material. On the overturned limbs of folds it is commonly inclined at angles of less than 10° to the bedding. In thin section a preferred orientation of mineral grains is conspicuous in most samples, which is probably the result of recrystallization combined with the mechanical reorientation of crystal particles.

A conspicuous feature of both types of cleavage is a regionally consistent dip which is apparently independent of the attitude of bedding. Because it is also subparallel to the overturned limbs of mega-anticlines, it is very likely that the regional cleavage developed simultaneously with the later phases of folding or even afterwards.

The occurrence of small shale injections parallel to the cleavage in sandstone and vice versa (Figs. 80 and 86) is possibly the result of water driven off perpendicular to the direction of maximum principal stress during metamorphism (Maxwell, 1962) causing the rotation of mineral particles to form slaty cleavage. If this is true, it means that the sediments at this late stage were not yet fully consolidated. The escape of pore water during compaction could have been retarded by the thin, impermeable shale horizons in the Uitvlugt Formation, resulting in prolonged high pore pressures at depth. On the other hand, it is also possible that these "injections" in fact originated as load or flame structures while the sand=stone was still waterladen, and were later rotated into and flattened parallel to the developing cleavage.

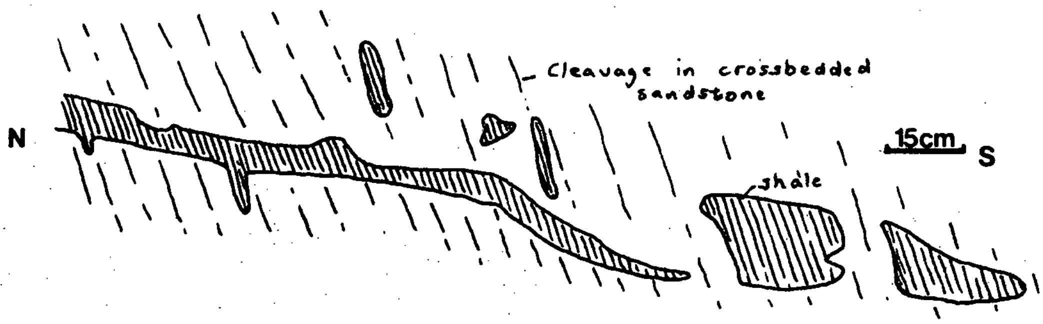


Fig. 86. Shale injections parallel to cleavage in sandstone, Uitvlugt Formation, Potgieters Poort.

According to Hills (1972) the expulsion of water from the original constituents and voids results in the formation of sericite and chlorite by recrystallization of the clay minerals, while quartz and sedimentary micas are in general only mechanically deformed. Chlorite and sericite porphyroblasts are common in most of the Cango thin-sections and are generally orientated parallel to the cleavage. Quartz however also displays signs of recrystallization parallel to the foliation, the new grains showing a polycrystalline texture.

In the limestones of the Matjies River Formation cleavage forms by recrystallization and plastic deformation, as shown by zones of coarsely recrystallized calcite and deformed oölites.

5.3.2. REGIONAL CLEAVAGE TRENDS (Data points: Fig.1; Subareas: Fig. 87).

The strike of all cleavage in the Cango ranges between 060° and 160° with a maximum at 110° (Fig.88 (f)). The variation in strike (60°) is less in the Uitvlugt Formation (Fig.88 (c)) than in the other units, in which it may vary as much as 100°. This may be attributed partly to the competence of the formation, which presumably would be affected less by rotation along shear faults or by "einengungstektonik (see p.126), and partly to its narrow exposure in the east, where less readings were taken as a result. The most incompetent rocks on the other hand, those

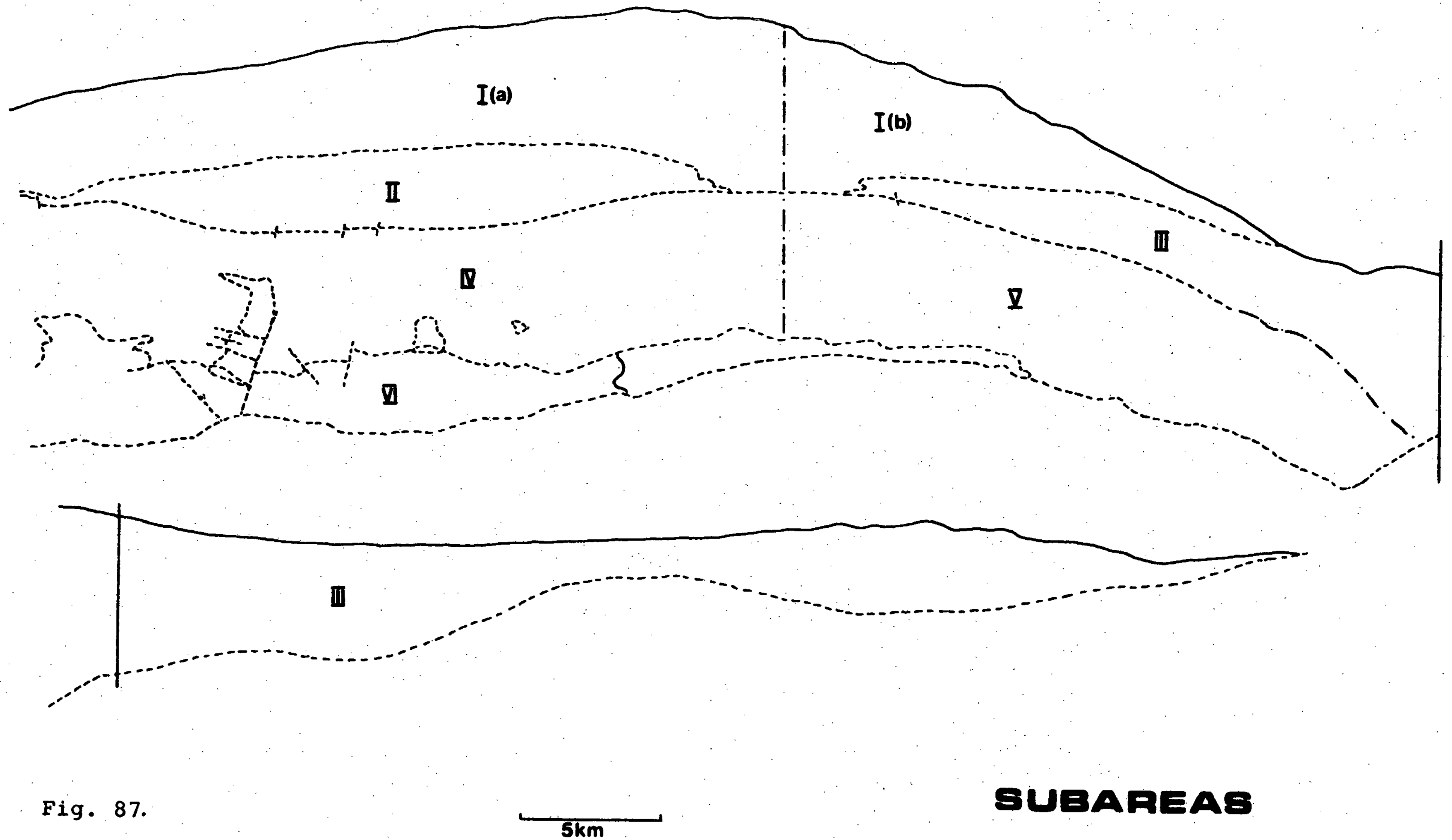


Fig. 87.

SUBAREAS

Fig. 88. Histograms of cleavage strike.

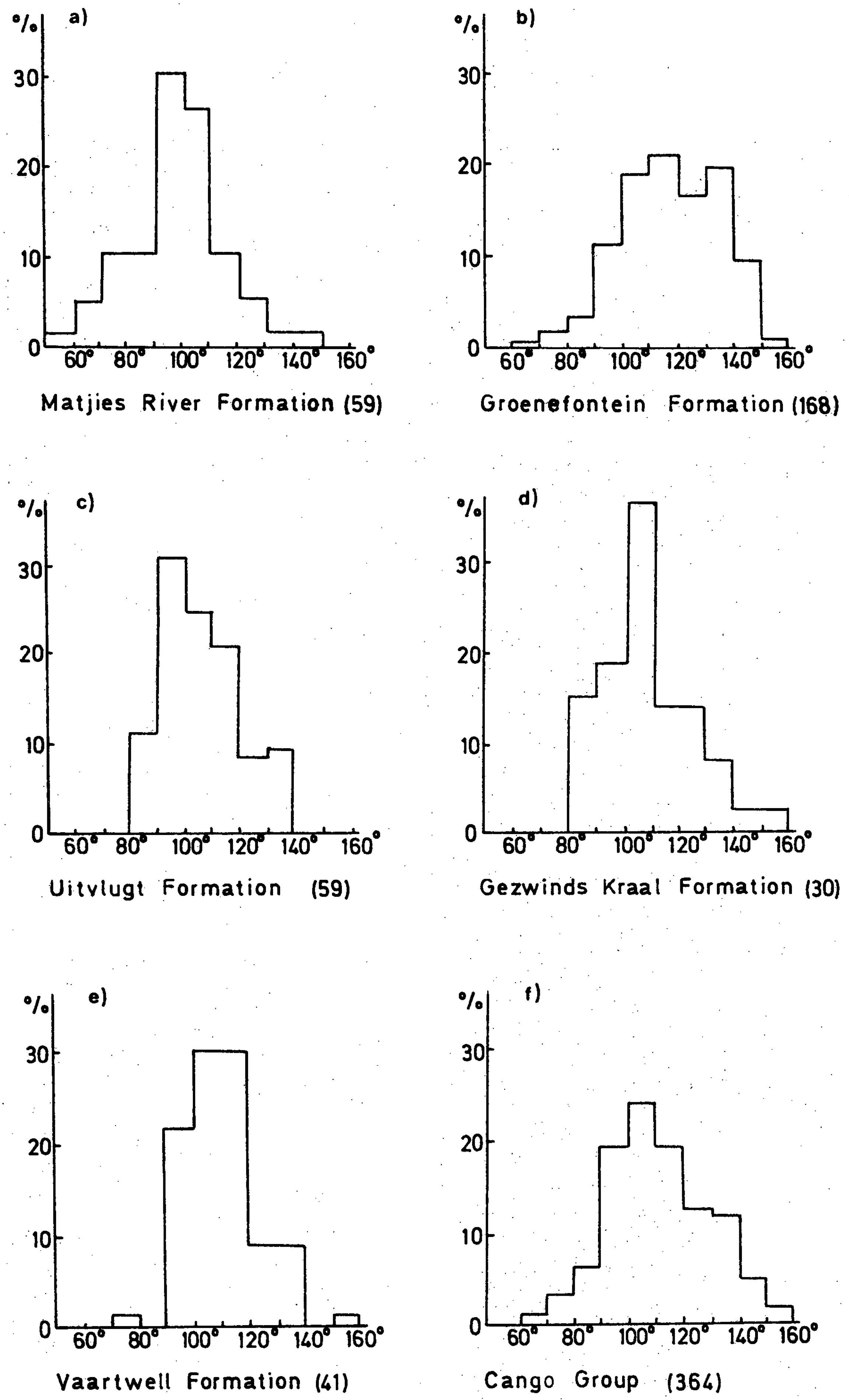


Fig. 89. Histograms of cleavage dip.

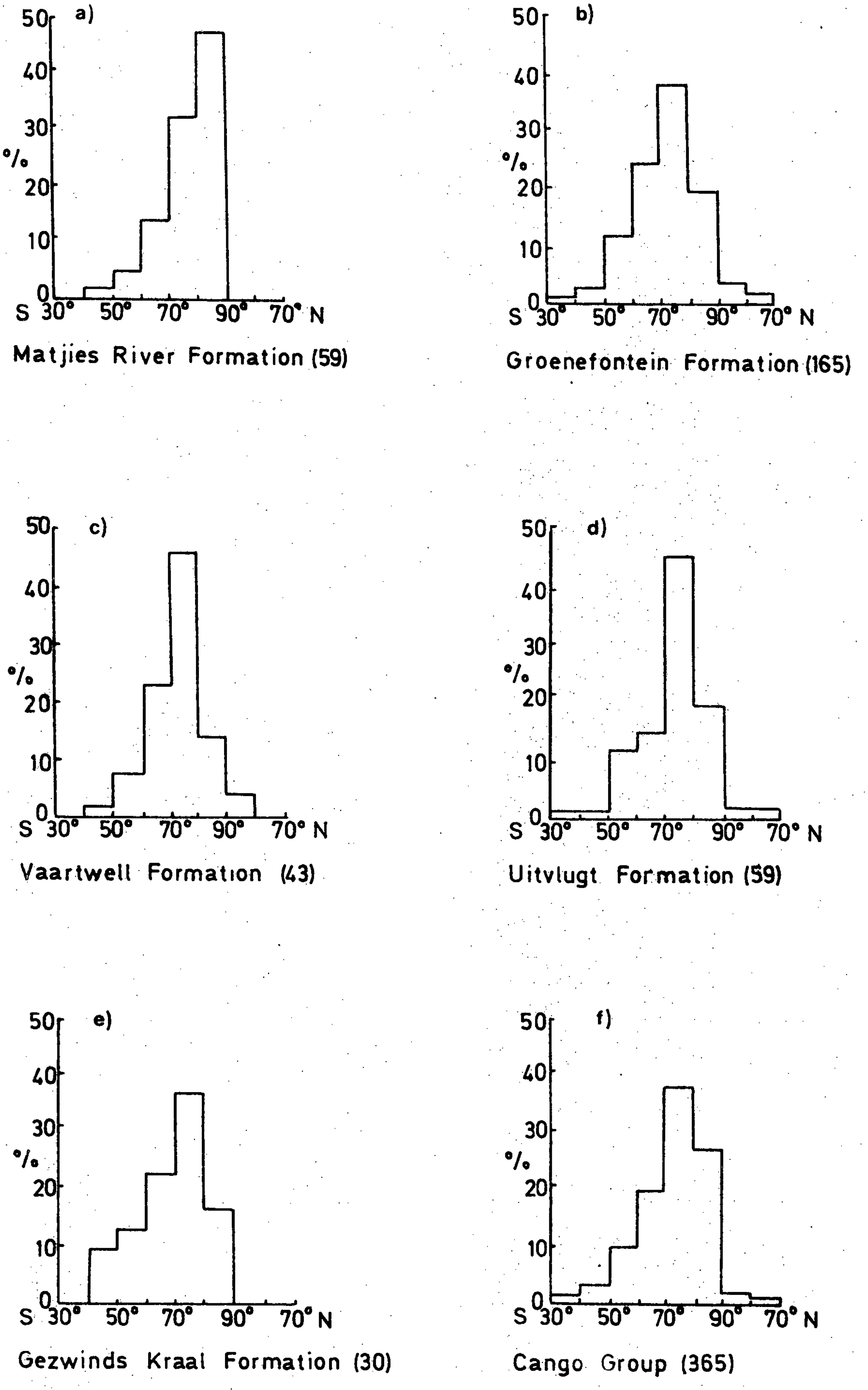
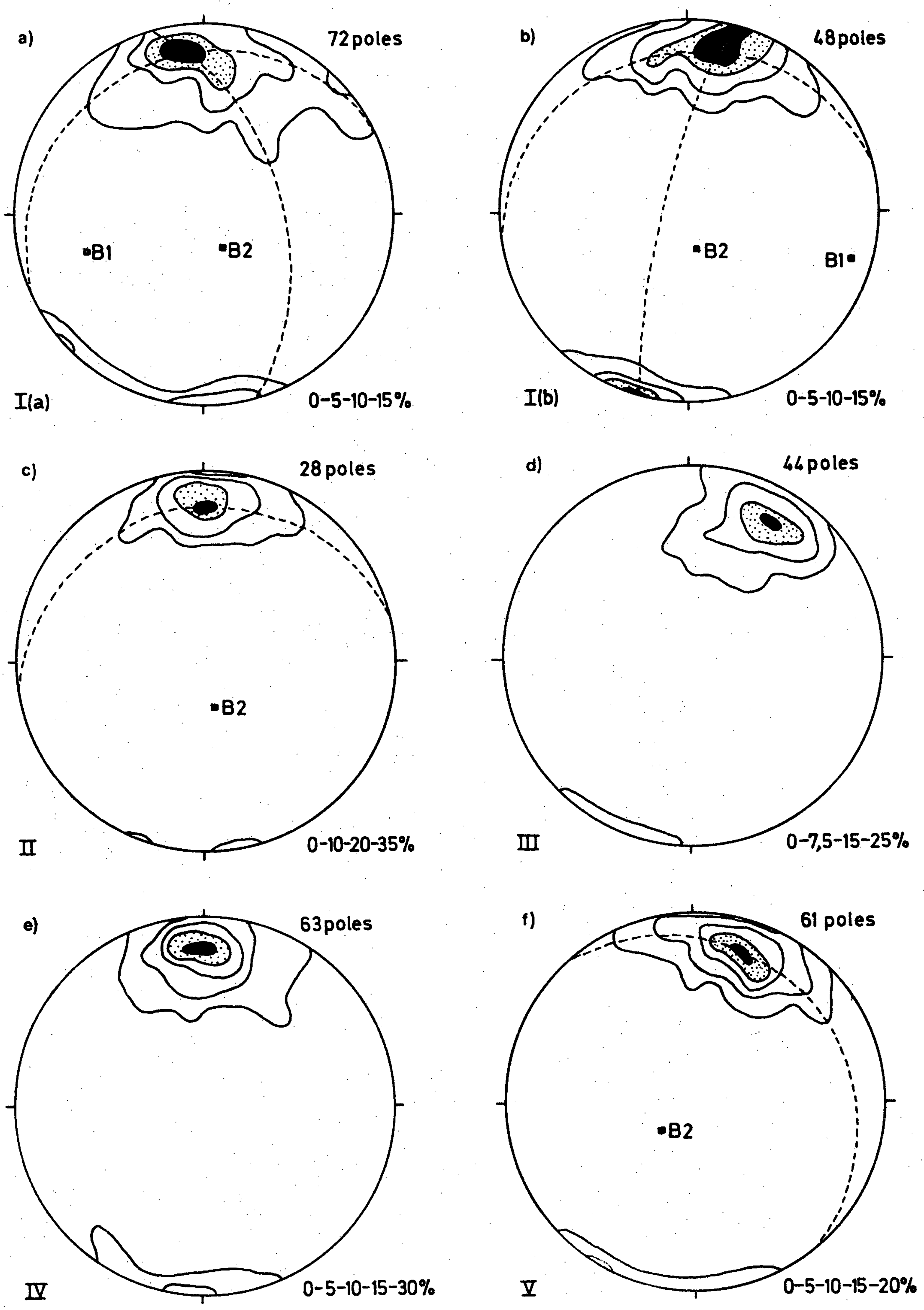
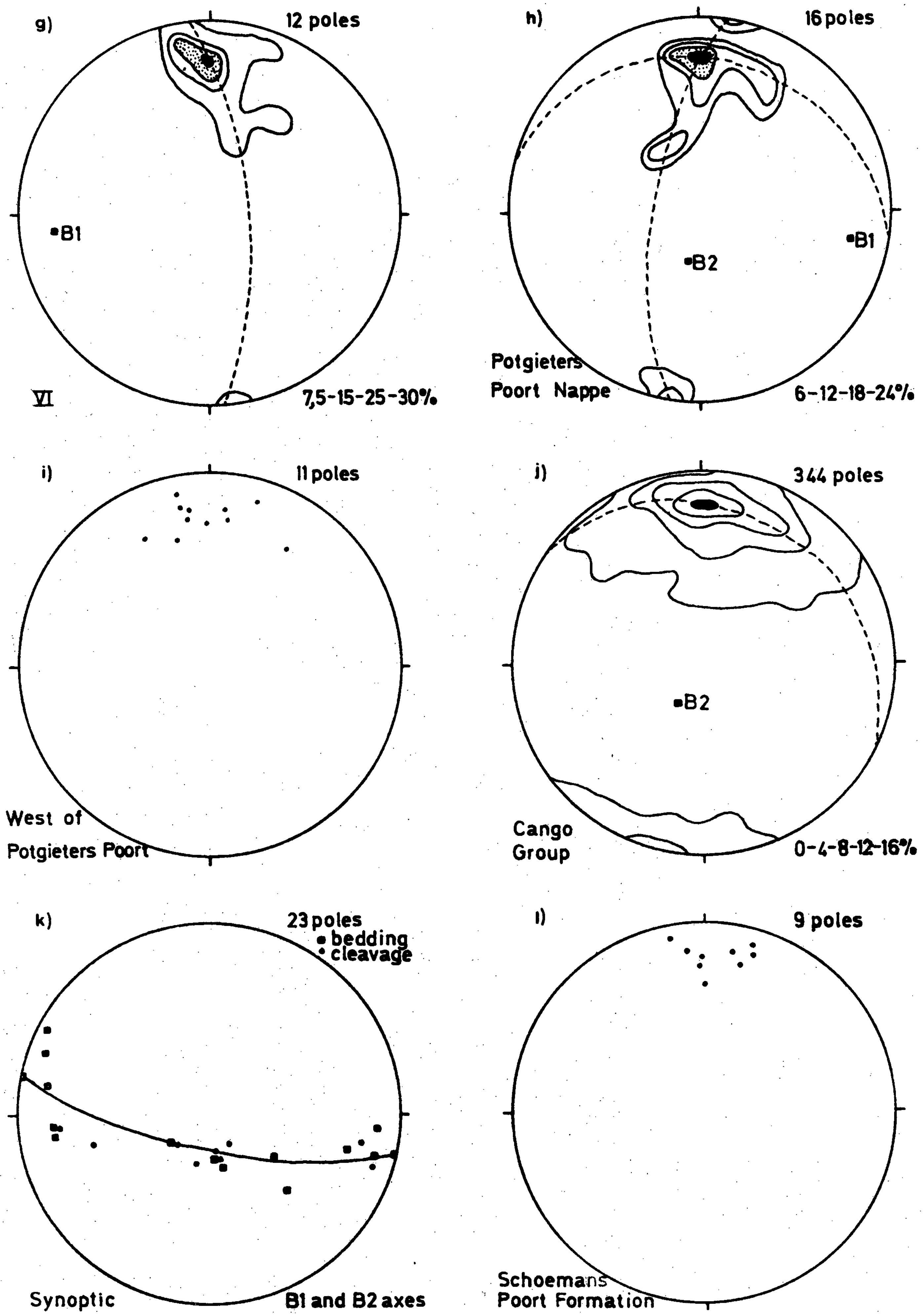


Fig. 90. Cleavage : Π-diagrams.




g)
12 poles
B1
VI
7,5-15-25-30%
h)
16 poles
B1
B2
Potgieters
Poort Nappe
6-12-18-24%
i)
11 poles
West of
Potgieters Poort
j)
344 poles
B2
Cango
Group
0-4-8-12-16%
k)
23 poles
bedding
cleavage
Synoptic
B1 and B2 axes
l)
9 poles
Schoemans
Poort Formation

of the Groenefontein Formation (Fig. 88 (b)) show the greatest deviation and most subdued maximum, although this may also be traced back to their larger outcrop area.

A unimodal distribution is shown by the histograms of dip angle, with a maximum between 70° and 80° in all but the Matjies River sediments. (Fig. 89 (a) and (f)). The regional cleavage strike east of 22° 15′ swings from due east to south of east (Fig. 90 (a) - (g): (i)), the implications of which are discussed later on. Cleavage plots of the Potgieters Poort Nappe define a strike and dip which conform to the regional trend of the area. (Fig. 90 (h)). This result also favours a thrust origin for the nappe.

The few readings taken in the Schoemans Poort Formation (Fig. 90 (1)) confirm that the post-Cape compression was exerted in the same general direction, i.e. from south to north.

5.4. FOLDS

5.4.1. DISCUSSION OF SUBAREAS (Fig. 87).

5.4.1.1. SUBAREA I

THE MATJIES RIVER FORMATION

Detailed mapping has made possible the correlation of the Cave Member with a partly exposed limestone band in the south, which has hitherto been regarded as a lower stratigraphic horizon. The megastructure within the Matjies River Formation is thereby revealed as a single overturned anticline (the Cango Valley Anticline), with both limbs dipping steeply towards the south. The fold axis plunges eastwards.

The strike of the axial trace as inferred from outcrop patterns, bedding and cleavage strike and other field observations, changes between 21° 05′E and 21° 09′E, where the Cave limestone is facially replaced by shale. A possible explanation is that, during the final stages of folding, the competent Cave Member on both sides of the Bassons Rust gap acted as rigid walls between which the less competent shale was sheared towards the north. This "Einengungstektonik" (Sander, 1948) led to a local redistribution of stress fields and the rotation of the rigid sides away from the area of compression (Fig. 91 (a)). As revealed by the structural map (folder), the limestones on both sides of the gap are more intensely overfolded than elsewhere, which could be a direct result of the combined forces acting in the area.

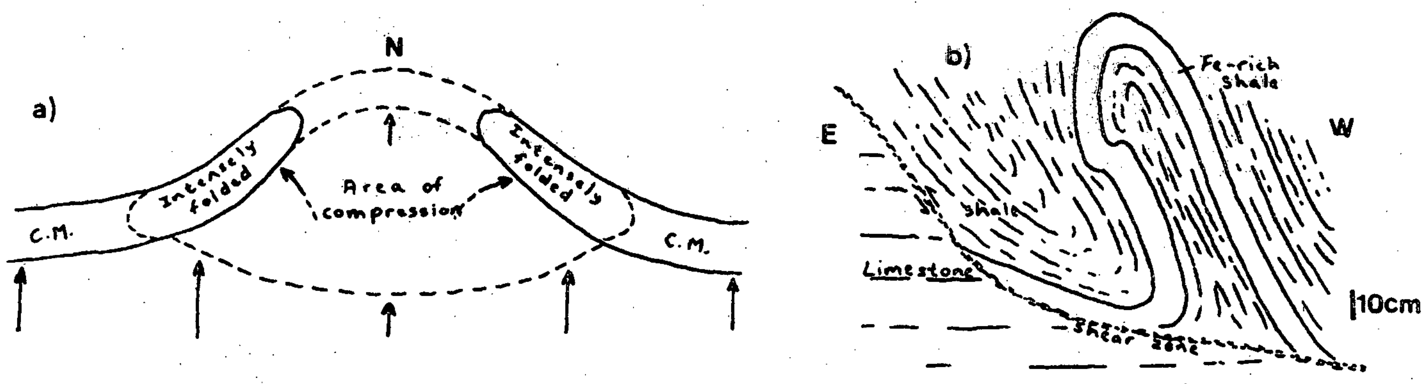


Fig. 91. (a) "Einengungstektonik" in the Bassons Rust gap.
(b) Sheared anticline in same area.

The existence of a complex stress pattern in the vicinity of the gap is confirmed by an analysis of small scale structures. The Cave Member shows evidence of first order folding which is not observed elsewhere and which is probably responsible for the wider outcrop on Nooitgedagt (Plate 6 (b)). Fig. 91 (b) illustrates an isoclinal fold in shales of the Nooitgedagt Member, in which the overturned limb has been sheared towards the east. Intensely folded-calcite veins, plunging boudins and many displacement features are also tell-tales of a more complex deformation than is apparent at first glance.

Small scale folds elsewhere in this formation occur mainly in shale. Isoclinal and similar fold varieties display the typical axial plane cleavage, although proof for flexural-slip mechanisms also exists.

THE GROENEFONTEIN FORMATION

Stratigraphic subdivison of the formation and readings taken at regular intervals provide a clearer picture of its megastructures than has previously been possible. The southern flank of the large anticline within the subarea (structural map) has a dip of 30° to 60° over most of its extent, but becomes steeper east of 22° 17′ where the axial planes are spaced more closely. From here eastwards all folds are isoclinal.

Smaller isoclinal folds are fairly common, while some of the larger limestone lenses in the Lower Member also exhibit these features.

π - DIAGRAMS (Fig. 92 (a,b))

The subarea is divided into two parts because of its great lateral extent. The western part (a) is characterized by a complex pattern and a number of π- circles can be drawn. π-1 defines the main direction of compression with a fold axis plunging at 098°/15 E, while the other possible circles can be attributed to the redistribution of stress in the Bassons Rüst gap.

5.4.1.2. SUBAREA II (Fig. 92 (c))

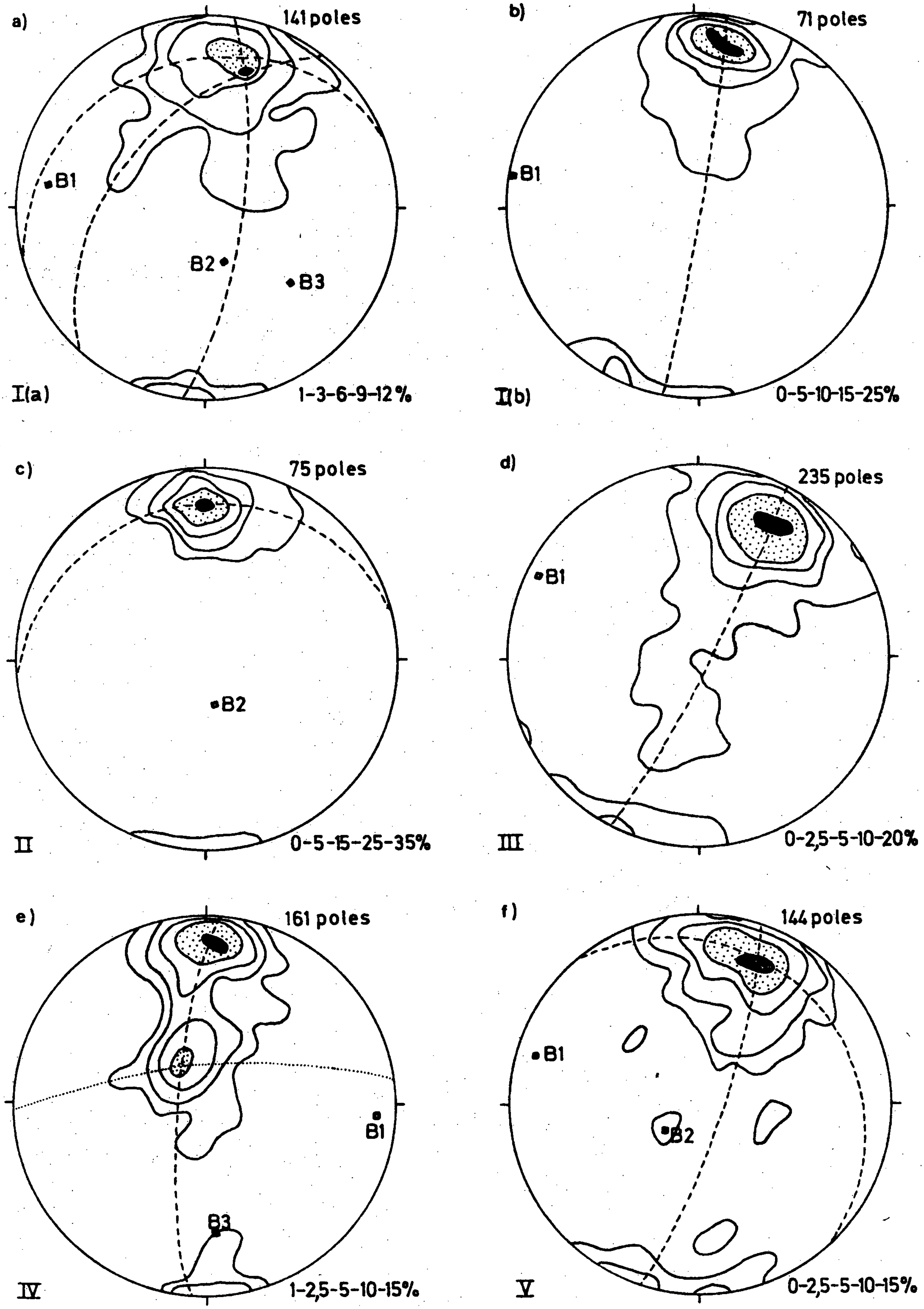
The Vaartwell, Uitvlugt and Gezwinds Kraal Formations are overturned to the north. Only the reversed limb of the anticline, which has a southerly dip of about 70°, is exposed. Higher order folds were not observed.

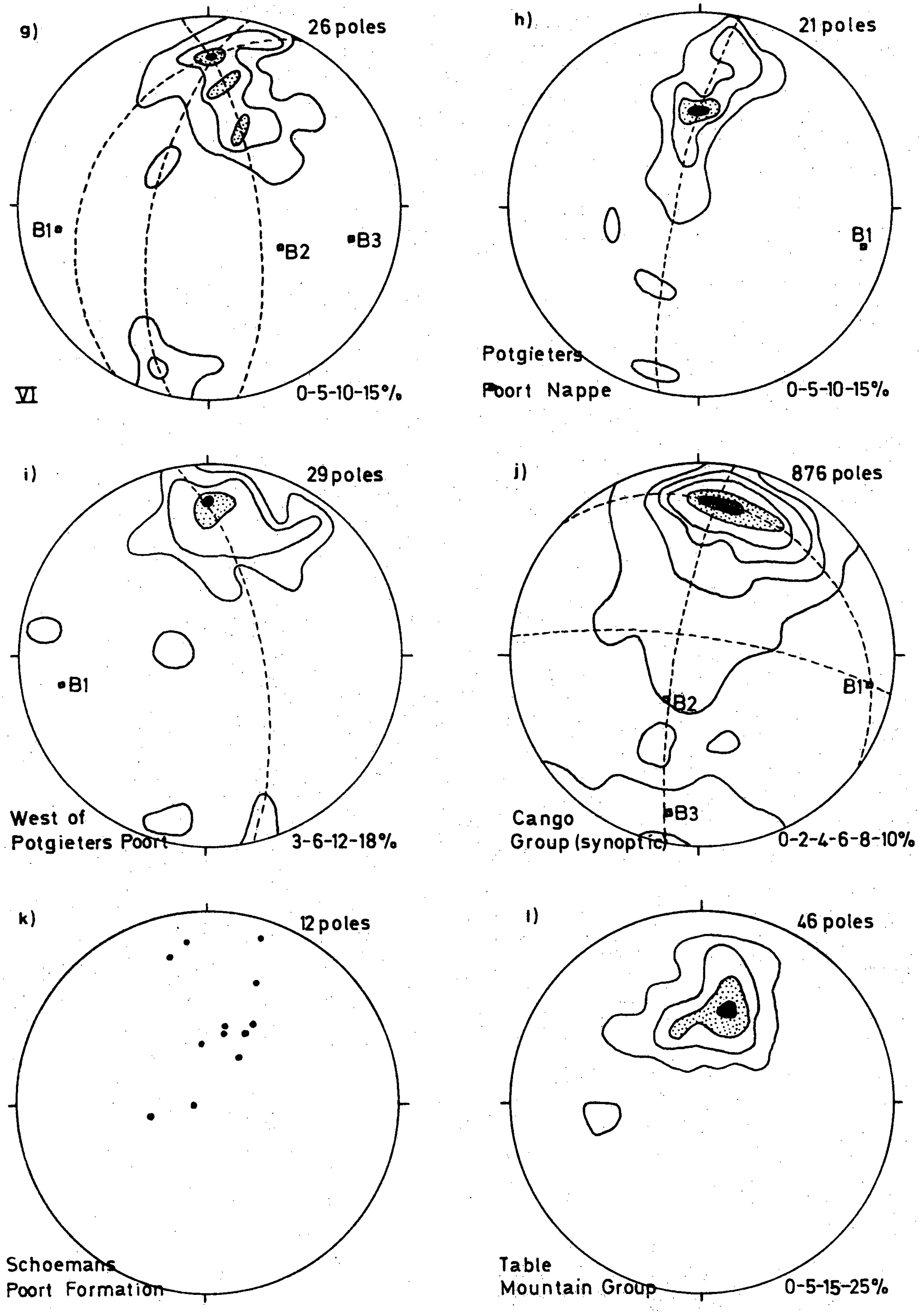
5.4.1.3. SUBAREA III (Fig. 92 (d))

In spite of its great lateral extent, bedding pole plots of this subarea define a single great circle. The axes of the two major folds crossing the area from west to east are nearly horizontal with a strike of 118°, which implies a deviation of 20° or more from the trend in other subareas.

Steep dips dominate in the strip south of the northernmost syncline, with tight, isoclinal folding being characteristic of the structural trend in this area. North of this syncline a more open type of folding is revealed where dips average 30 to 60° over large areas, steepening however to more than

Fig. 92. Bedding : Π-diagrams.




g)
26 poles
B1
B2
B3
VI
0-5-10-15%
h)
21 poles
B1
Potgieters
Poort Nappe
0-5-10-15%
i)
29 poles
B1
West of
Potgieters Poort
3-6-12-18%
j)
876 poles
B1
B2
B3
Cango
Group (synoptic)
0-2-4-6-8-10%
k)
12 poles
Schoemans
Poort Formation
l)
46 poles
Table
Mountain Group
0-5-15-25%

60° east of Meirings Poort.

The variation in the attitude of bedding on opposite sides of the syncline is probably a result of the gradual change in the Gezwinds Kraal sediments, which pass into the coarser and more competent Schoongezigt Formation beyond 22°37'E.

The Uitvlugt and Vaartwell Formations are considerably reduced in thickness in this subarea and play a minor role in determining structural trends. An exception may be the syncline crossing Meirings Poort at 33°28'S, which exposes these two units in its core. The Vaartwell here shows a considerable increase in thickness and, combined with the massive sandstone lenses in the Upper Member, could have determined the position of the trough zone.

Rarely observed first order folds in the sediments of the Groenefontein Formation are again of the similar, isoclinal type (Fig. 93).

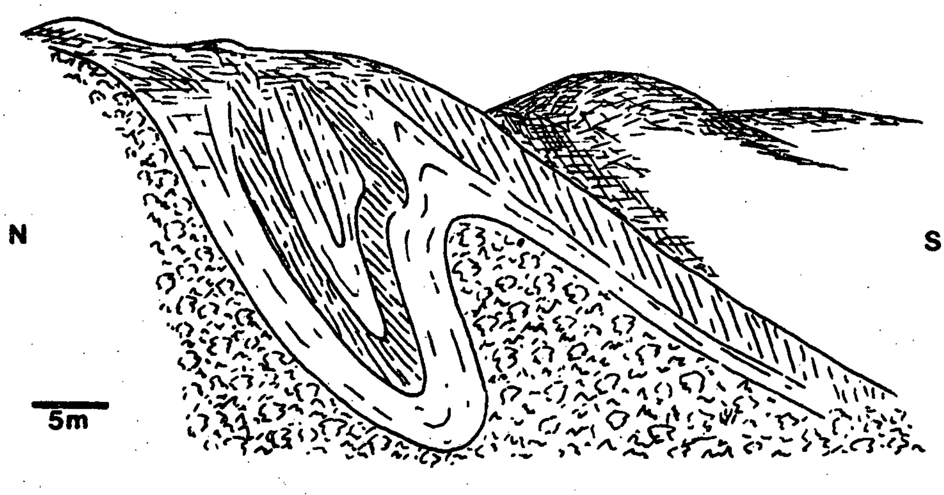
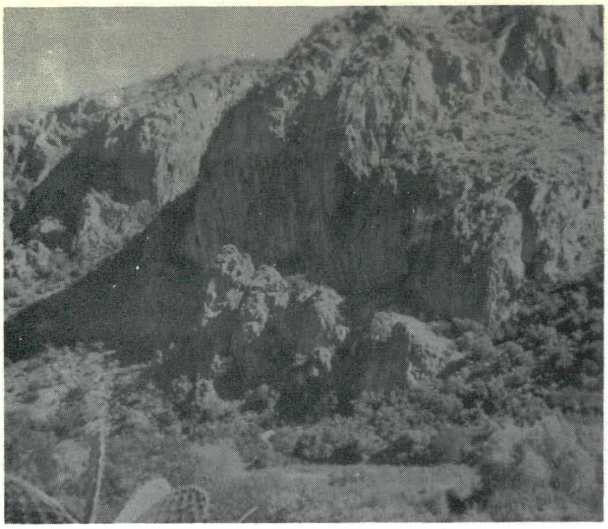


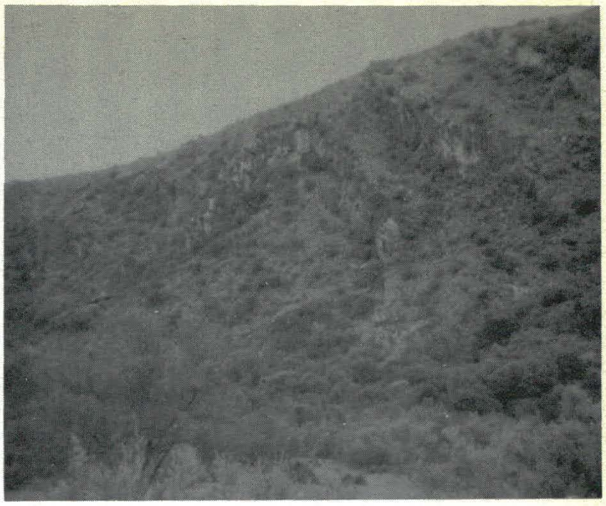
Fig. 93. Isoclinal fold in Groenefontein Formation, De Oude Muragie.

5.4.1.4. SUBAREA IV (Fig. 92 (e))

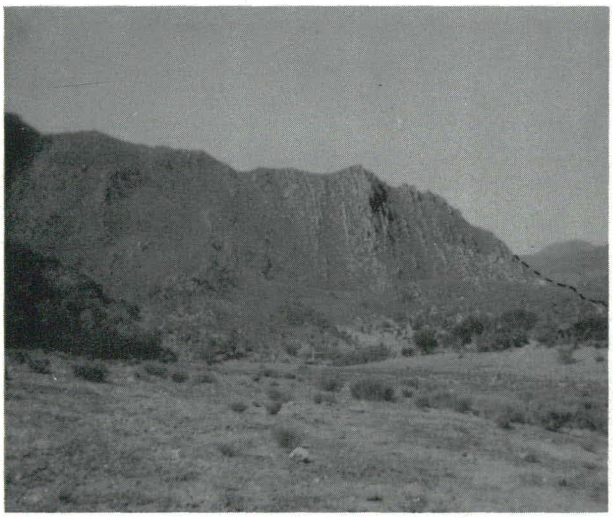
Dominated on the Central Plateau by the Uitvlugt Formation, this area is extremely suitable for structural studies. Good



(a) Uppermost part of Cave Mem= ber on overturned limb of Cango Valley Anticline, De Hoek. Looking west.



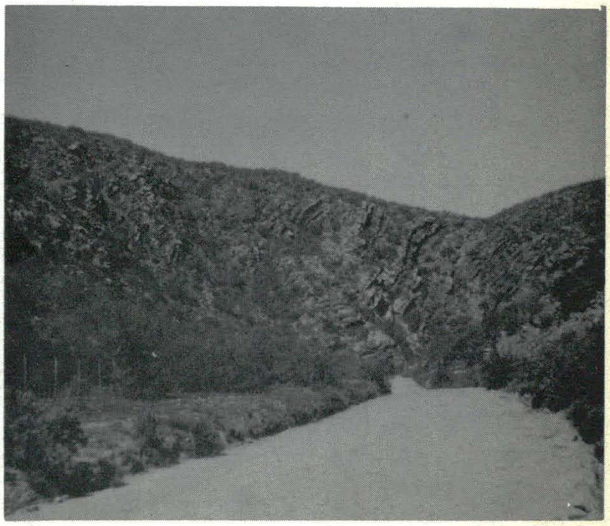
(b) First order folds in Cave Member. Looking east from Bassons Rust gap.



(c) Overturned Uitvlugt Forma= tion (left) grading into Vaartwell Formation north of Wildehonde Kloof Fault, Central Plateau. Looking east.

(d) Overturned Uitvlugt Forma= tion south of Wildehonde Kloof Fault, Central Pla= teau. Looking west.

(e) Uitvlugt sandstones dipping southeast on Central Pla= teau. Looking east.



(f) Monoclinal fold in Uitvlugt Formation , Potgieters Poort. Looking southeast.

outcrops and an abundance of crossbedding are of great help in direct observations, one drawback being the ruggedness of the plateau which makes it only accessible on foot. The megastructure as shown by the structural map and the π- diagram is an overturned anticline, the normal limb of which steepens markedly before plunging down into the overturned flank. This effect is even more conspicuous in the Groenefontein Formation to the west, where dips of 60° to 90° S occur over a wide area. The hinge zone of this anticline is also much narrower than those to the north.

A major part seems to have been played by the Vaartwell Formation in controlling the position of the mega-anticline. The geo= logical map shows that, while the formation is absent south of the axial zone, it is well developed in the overturned limb (Fig. 94).

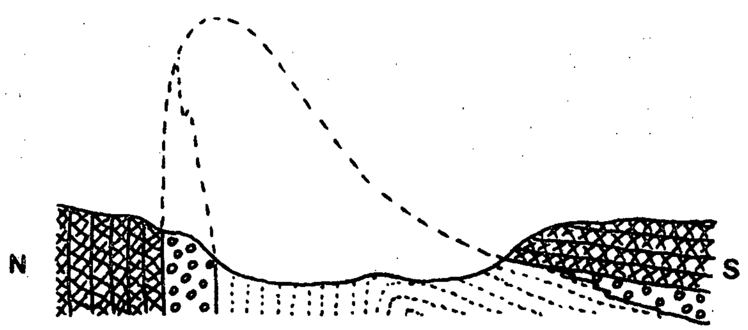


Fig. 94. Anticline controlled by absence of Vaartwell Formation, Wildehonde Kloof.

During folding a weak zone due to the absence of this competent horizon could thus have given rise to excessive buckling which developed into a major anticline. The proposed mechanism is explained by Fig. 95: Where lenses of a competent nature separated by weaker zones are subjected to folding, the competent lenses will deform into synclines while the structural weaknesses will form anticlines. When pressure is exerted from one direction overfolding results.

Fig. 95. Folding as a result of stratigraphic control.

In deformation of this type narrow anticlinal hinge zones as well as a steepening of dips adjacent to the crests of folds can be expected. In contrast to the tight folding of the other formations, the Uitvlugt on the normal limb of the anticline displays gentle, open folds of the first order. (Fig. 96 (a)).

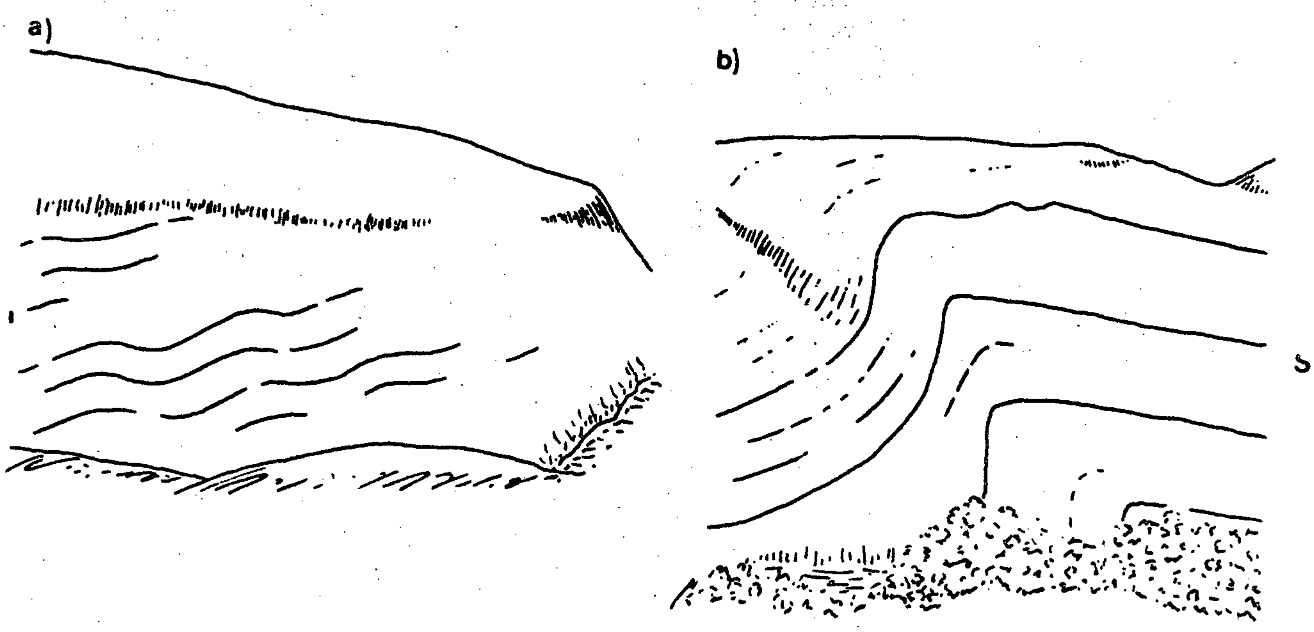


Fig. 96. (a) Open, first order folds in Uitvlugt Formation, Abrahams River.
(b) Overturned fold in Uitvlugt Formation, Welbedagt.

These folds are of the concentric type and were produced by buckling with slip along bedding and in shaly horizons, as a more detailed investigation of cleavage in the latter suggests.

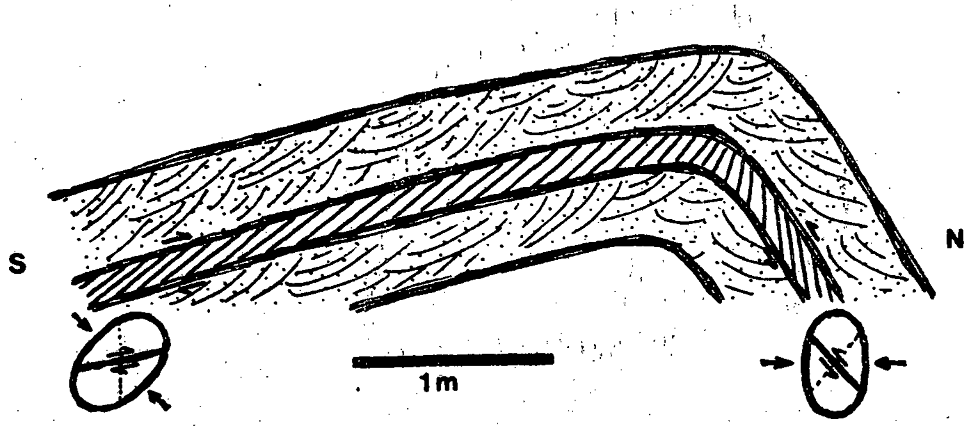


Fig. 97. Cleavage suggesting buckling and flexural slip in shale between sandstone beds, Uitvlugt Formation, Koetzers Kraal.

Another feature encountered on the normal limb of the mega-anticline is the occurrence of smaller overfolds which zonally interrupt the open fold structure. (Fig. 96 (b)).

5.4.1.5. SUBAREA V (Fig. 92 (f))

The mega-structure of the previous subarea continues towards the east, but in this part the Groenefontein Formation is exposed in the core of the anticline. As shown by the structural map, the Groenefontein Formation tends to deform into tight, isoclinal folds as opposed to the open, concentric folds of the Uitvlugt Formation. This implies disharmonic folding between the two formations, the crossbedded sandstones deforming mainly by buckling and flexural slip between layers, while the fine-grained Groenefontein wackes react by slip, shear or plastic flow perpendicular to the compression. A single anticline in the Uitvlugt can thus split into two or more in the underlying formation (Fig. 98).

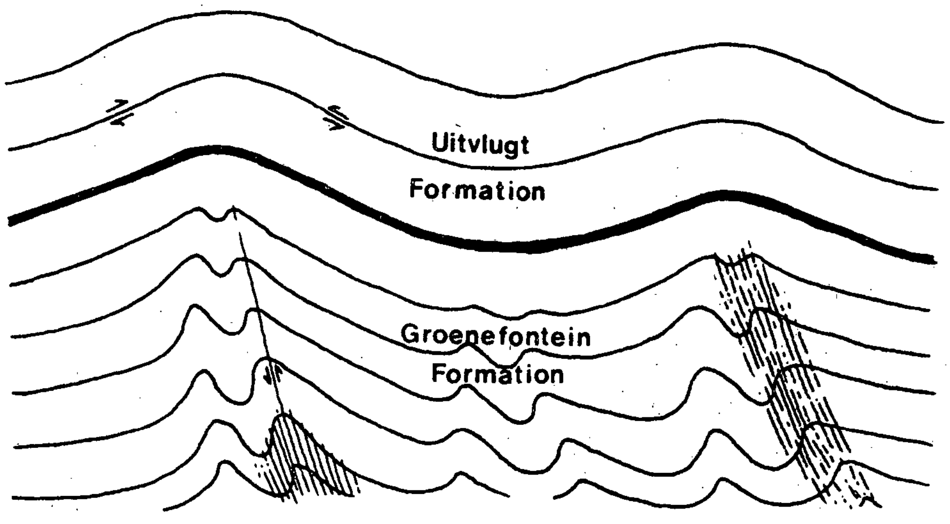


Fig. 98. Disharmonic folding between Uitvlugt and Groenefontein Formations.

5.4.1.6. SUBAREA VI (Fig. 92 (g))

The area east of the Koetzers Kraal Fault is dominated by a single mega-anticline with partly overturned northern limb and intermediate dips on the normal flank. West of the fault a traverse along the Matjies River reveals a succession of first order folds with gently dipping or horizontal back limbs and vertical or steeply dipping north flanks. The result is an ascent in the stratigraphy towards the north (Fig. 99 (a)).

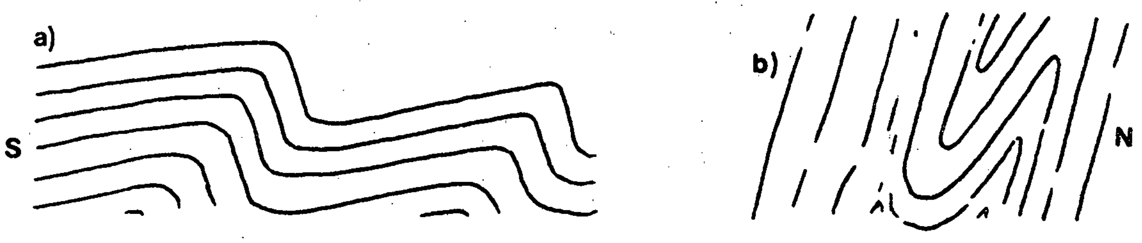


Fig. 99. (a) Monoclinal folding in Uitvlugt Formation, Subarea VI.
(b) Tight folding in Uitvlugt Formation, west of Matjies River.

Very tight varieties of these folds are sometimes encountered, (Fig. 99 (b)).

Sliding along bedding surfaces is revealed in the many slicken= sides, sheared quartz veins and cut-off features excellently exposed in the steep-sided river walls.

Bedding pole plots of this subarea (Fig. 92 (g)) have a somewhat more complex pattern which can be attributed to:

i) The relatively small amount of readings taken, and
ii) Effects of the Welbedagt and transverse faults in the vicinity.

5.4.1.7. THE POTGIETERS POORT NAPPE (Fig. 92 (h)).

Although bedding in many parts of the nappe is horizontal and parallel to the fault plane, the northernmost part in the Koetzers Kraal Klippe dips steeply towards the north. This probably represents the sheared-off limb of an anticline thrust in from the south. In the southern part of this outlier the bedding is once more inclined northwards. (Fig. 100).

Fig. 100. Folds and faults in Koetzers Kraal Klippe, seen from the east.

In other parts of the nappe, e.g. on Vogelfontein, the bedding at a distance can be observed dipping east at about 20°, while that in the Welbedagt Klippe is almost horizontal. The root zone west of the Droë River Fault is characterized by gently southward dipping beds.

5.4.1.8. WEST OF POTGIETERS POORT (Fig.92 (i))

Outcrop patterns (Prince Albert 1:125000 map) indicate a single mega-anticline which apparently conforms to the structure of the Table Mountain Group. The average dip is about 60°S.

In the Huis River Pass concentric folds with amplitudes of 50 m or more were observed in the Groenefontein Formation, some with fold axes striking more or less in a northerly direction.

5.4.1.9. THE SCHOEMANS POORT FORMATION (Fig.92 (k))

The largest concentration of bedding plots defines an average dip of about 30° S. Some plots indicate dips of 50 - 80° S, but bedding is extremely difficult to recognize within this formation and these could in fact represent secondary foliations.

5.4.1.10 THE TABLE MOUNTAIN SANDSTONE (Fig. 92 (l))

The π- diagram represents data from the remnant of the normal, southern limb of the anticlinorium, which has an average dip of 40° S. The fold axis direction however cannot be deduced, as no readings were taken on the northern, overturned limb of the an= ticline. Some of the contact relationships in the north are illustrated in Fig. 101 (a) and (b))

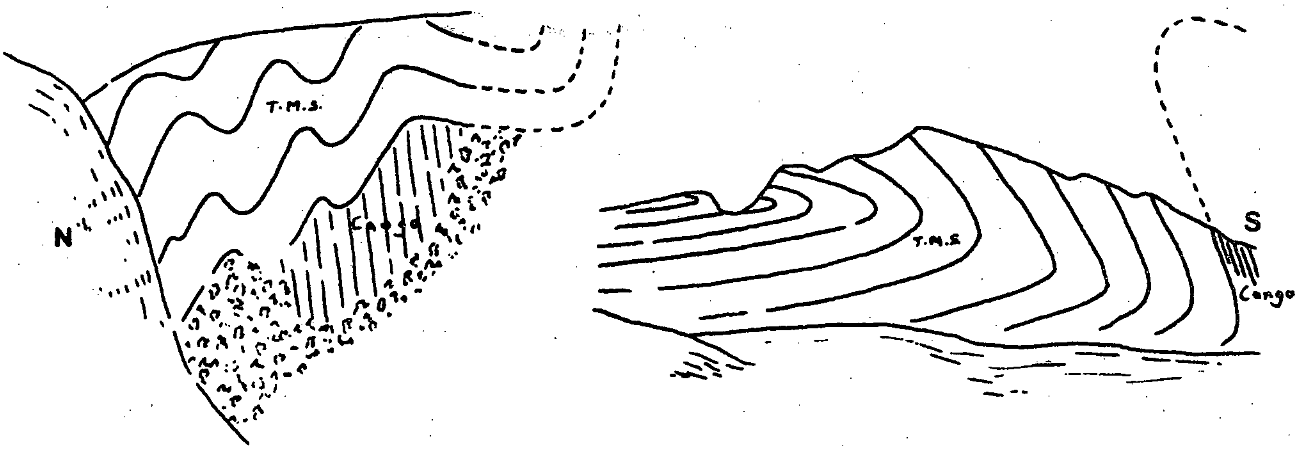


Fig. 101. Contact relationships between Table Mountain and Cango Groups on (a) Tafel Berg and (b) Spitzkop.

There is considerable evidence of differential shear between the Table Mountain Sandstone and the Cango rocks in the south. On the farm Koetzers Kraal for example, crumpled bedding occurs in chloritic phyllite of the Gezwinds Kraal Formation directly below the contact. In the north unfortunately, the contact is nearly always obscured by scree.

5.4.2. REGIONAL BEDDING TRENDS

The synoptic plot of bedding planes in the Cango Group east of Potgieters Poort reveals a large spread in the direction of strike, which is directly related to the arcuate structure discussed later on.

As a result two B-axes (B1 and B2) at right angles to each other are defined in many of the subarea π- diagrams, and also in the synoptic plot of bedding in the Cango Group as a whole (Fig. 92 (j)). The B2-axis has a steep southerly dip and represents the northern, overturned limbs of anticlines in general, as these are far more often exposed than the normal limbs. Subarea IV (the Central Plateau) is the only region where adequate readings were taken on the latter. If this limb is similarly folded around a north-south axis a third Π-circle would be expected in the Π- diagram, corresponding to a B3 - axis dipping south at about 30°. The exposure of the Groenefontein Formation both east and west of the Central Plateau, which is occupied by Uitvlugt sandstones (Fig. 102), confirms the existence of such a north-south trending axis on the normal limb. A third B - axis is therefore postulated in Fig. 92 (c), normal to the poles of the southern limb and defining a Π-circle passing through a weak westward bulge in the contours. A similar Π-circle on the synoptic diagram passes through the B1 axis, thus proving that the two directions of folding are at right angles to each other.

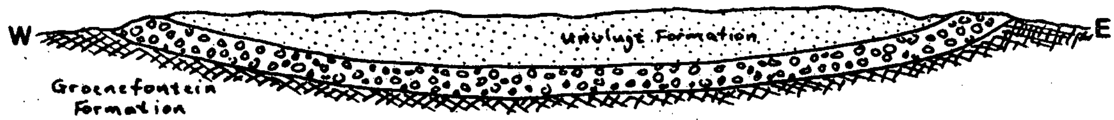


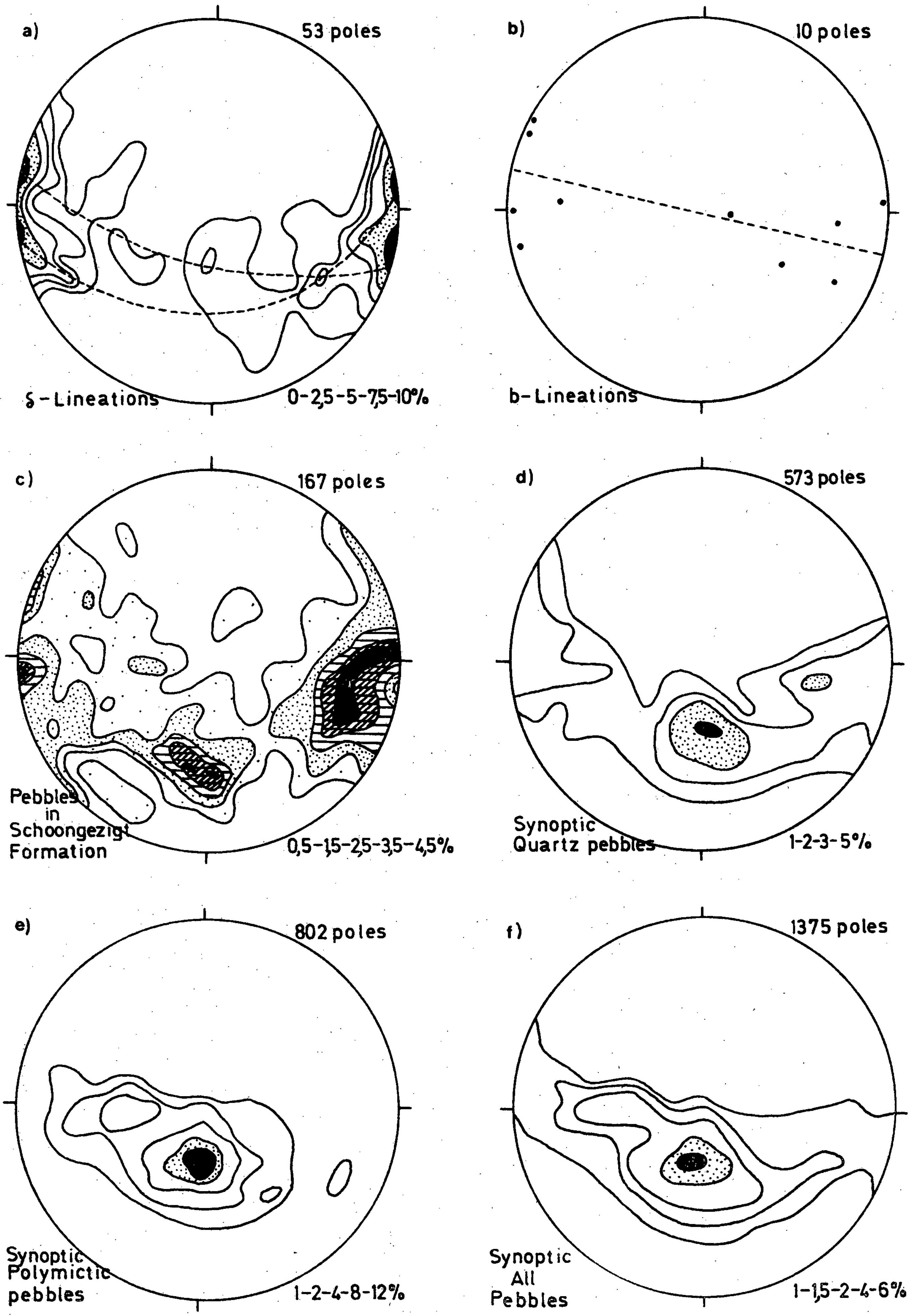
Fig. 102. Gentle warping around north-south axis, Central Plateau.

5.5. LINEATION

5.5.1. δ -LINEATION (intersection of bedding and cleavage)

For this analysis the Cango was treated as a single domain. The locus of the δ- poles (Fig. 103 (a)) defines two great circles of which one approximately coincides with the mean orientation of cleavage. As cleavage and bedding dips on the overturned limbs of anticlines are nearly parallel, this

Fig. 103. Lineations : Synoptic diagrams.



π-1 circle is taken to represent δ- lineations on these limbs, while π-2 represents the normal flanks. Steep lineation plunges on these great circles indicate noncylindrical folding on a macroscopic scale, which is in accordance with the arcuate style of folding postulated above. A similar lineation plot will result if the cleavage was superimposed on strata already folded or partly folded in this manner.

Although some of the cleavage planes measured possibly belong to the post-Cape deformation, it is likely that most of the cleavage was formed during the later stages of the post-Cango deformation, when the original strike directions of fold axes had been modified slightly. The synoptic bedding and cleavage maxima are in fact separated by about 5°.

5.5.2. b - LINEATION (Fig. 103 (b))

A few axes of small folds, kink bands and quartz rods were measured. Their distribution is similar to that of B1 - axes.

Quartz rods are more abundant in shale bands between the massive sandstones of the Uitvlugt Formation than elsewhere. This can be related directly to the style of deformation of this unit, in which segregational quartz veins were ruptured as a result of shear between beds, so that their longest dimensions are parallel to the fold axes. (Fig. 104)

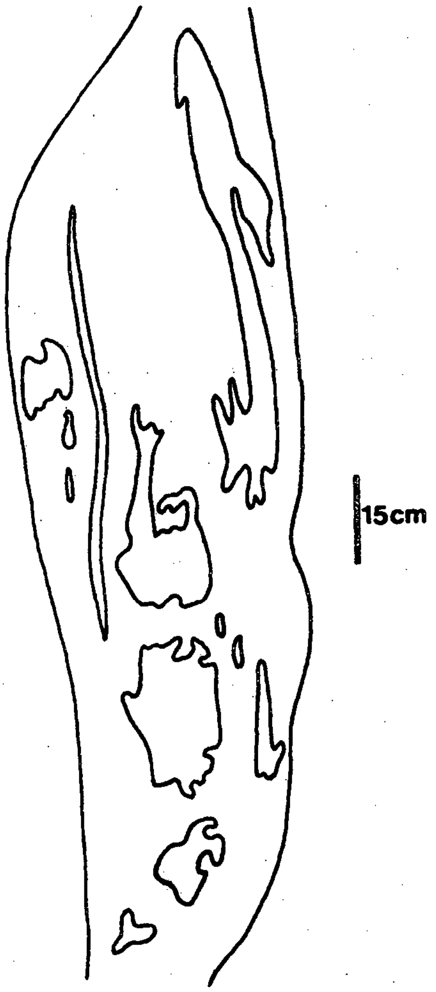


Fig. 104.
Ruptured quartz veins forming b-lineations, Schoemans Poort.

5.5.3. a-LINEATION (Fig. 103 (c), (d) and (3))

The longest axes of deformed pebbles in the Cango have a maximum orientation parallel to cleavage and perpendicular to the fold axes. This a-lineation is more pronounced in the polymictic (Fig. 103 (e)) than in the quartz pebble conglomerate (Fig. 103 (d)) While the latter have Y/Z ratios seldom exceeding 2:1, the incompetent varieties show values ranging from 6:1 to 12:1 for fine grained quartz wacke and up to 18:1 for shale and limestone.

Because these pebbles usually occur in a matrix of approximately similar competence, their deformation proceeded by stretching parallel to the direction of tectonic transport and plastic flow. The quartz pebbles in polymictic conglomerate however, are considerably more competent than the argillaceous matrix of the latter, and if oriented, did so by rotation. Rhombic outlines displayed by some are reminiscent of shearing, carried to the extreme where fractured pebbles are offset in an en echelon manner. (Fig. 105).

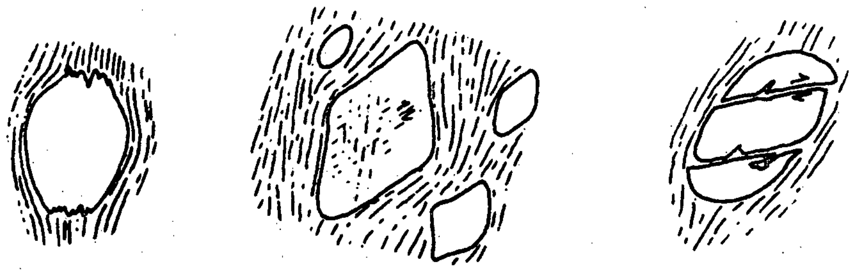


Fig. 105. Shapes of quartz pebble altered by deformation.

The majority of quartz pebbles in monomictic conglomerate (sandy matrix) are not visibly deformed or rotated and can therefore be used in imbrication studies of stream directions.

A closer look at the plots of long-axes shows that, while pebbles in the Vaartwell Formation are in general parallel to the bedding, in many instances the latter is more steeply inclined. This is to be expected on the overturned limbs of anticlines if the alignment is parallel to an axial plane cleavage. Whereas Whitten (1966, p. 269) states that elongated components normal to B possibly develop during the initial deformation of incompetent rocks, while lineations parallel to B develop during flexural-slip folding of more competent, thoroughly lithified units, both types of lineation according to Ramsay (1967) can occur in different parts of the same fold. The inverted limbs of anticlines normally show an a-lineation, while the hinge zone may have pebbles arranged parallel to the fold axis.

The orientation of pebbles at 19 data stations (not shown) in the Vaartwell Formation varies in no fixed manner along the bedding. Not one of the individual diagrams shows a clearly delineated b-direction, which may result from the fact that hinge zones are seldom exposed in the Vaartwell Formation. At 5 stations there is a clearly defined a-lineation.

Pebbles in the Schoongezigt Formation lie approximately in the bedding. The synoptic diagram (Fig. 103 (c)) shows both a - and b - lineations, with the latter somewhat better defined. Quartz pebbles are either randomly orientated in the bedding or slightly aligned in the b- direction, while the polymictic types posses both a - and b- lineations and show all transitions between the two. This may be attributed to the exposure of both the hinges and limbs of folds in this formation.

The synoptic plot of all pebble orientations in the Cango (Fig. 103 (f)) define a great circle striking approximately 098° and with a maximum concentration of a-lineation poles at 65° due south.

5.6 SUMMARY AND CONCLUSIONS

The structural analysis of the Cango Group was intended primarily to determine:

i) The style and intensity of deformation;
ii) The orientation of finite strain axes;
iii) The nature of the responsible forces.

The style of deformation obviously depends on the competence of the affected formations and the width of structural lithic units (Curry et al, 1962). Whereas the Groenefontein sediments show evidence of plastic deformation into isoclinal, similar type folds, the Uitvlugt Formation reacted by buckling and flexural slip to form concentric fold types. The Gezwinds Kraal Formation apparently followed the structural style of the latter and although considered incompetent, is not deformed into the isoclinal structures of its lithologic counterpart, the Groenefontein Formation. This may possibly relate to its higher position in the stratigraphy and consequently shallower depth of deformation, and also to its gradational contact with the competent Uitvlugt Formation.

The exposed part of the Matjies River Formation is deformed into a single, isoclinal anticline with steeply dipping limbs.

While the Vaartwell Formation apparently helped in determining the position of some of the fold axes, its two members show different styles of deformation. The incompetent pebbles of the polymictic member are elongated in a well-defined a-lineation on the overturned limbs of folds, indicating a large amount of plastic flow and stretching parallel to the clea= vage. Selective thinning of these north limbs could result in an anisopach style of folding for this member. On the other hand, the matrix of the quartz pebble conglomerate is similar, if somewhat coarser than the competent Uitvlugt sandstones, while bedding is even more massive and less well defined. Flexural slip or plastic flow would therefore be reduced to a minimum in this member. As its competence differs sharply from that of the underlying sediments, and while grading contacts are the rule between the other formations, it is not surprising that the thrust plane of the Potgieters Poort Nappe more or less follows the base of this formation.

The style of deformation of the Cango Group as a whole can be inferred from a geometrical analysis of the structural elements. The π- diagrams of cleavage and bedding poles show two B-axes approximately at right angles to each other. Whereas B_1 is obviously the result of north-south compression, B_2 apparently represents cross-folding in an east-west direction.

The B_1 and B_2 axes of all π- circles (Fig. 90 (1)) define a great circle which coincides with that of δ- lineation (π-1), the synoptic diagram of pebble orientations and the cleavage maximum plane. The steep dip (75°) of this great circle results from:

i) The isoclinal folding of the Groenefontein and Matjies River Formations, and

ii) The exposure of the Vaartwell, Uitvlugt and Gezwinds Kraal sediments mostly on the overturned limbs of anticlines.

The locus of the B-axes can be explained by Fig. 106, which depicts a warping of the axial surfaces in an arcuate structure towards the north. Whether this is a result of superimposed folding or simply a case of heterogeneous, noncylindrical B_1 folding reflecting trends in the underlying basement, is uncertain. The orientation of axes at 90° to each other is not an uncommon feature in metamorphic terrains elsewhere - it might even result from an inversion of the process responsible for rectangular joint systems, i.e. lateral compression because of burial. Another possibility which should be kept in mind, is the possible effect of the Cedarberg folding in the Western

Cape, which is also perpendicular to the north-south compression responsible for the Cape Fold Belt.

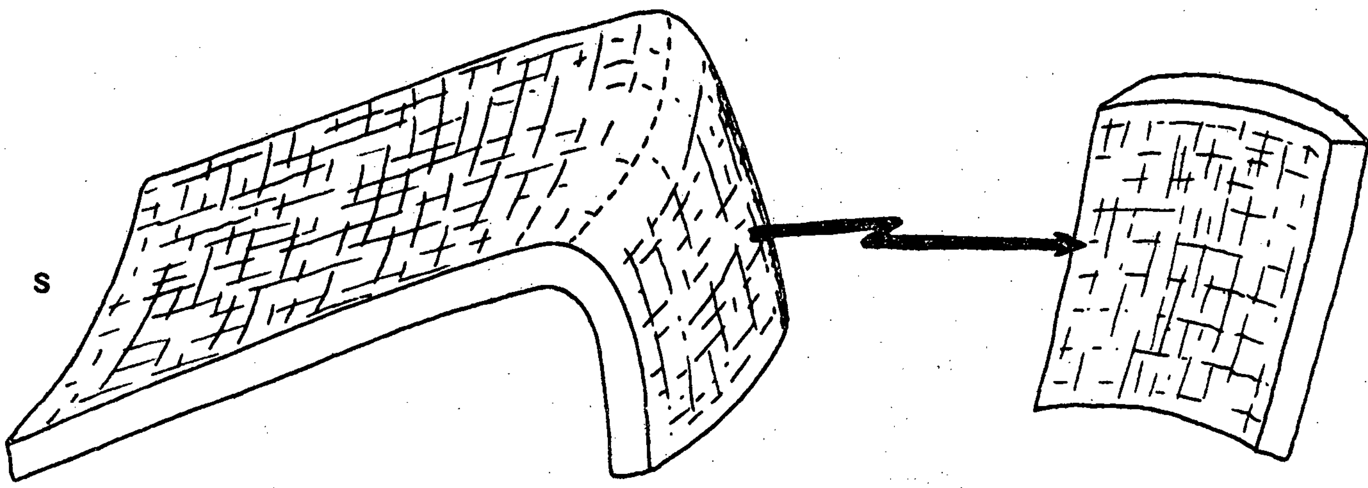


Fig. 106. Diagrammatic representation of Cango megastructure.

Although some units are affected by thrust faulting and isoclinal folding the overall intensity of deformation in the study area cannot be regarded as very high. Large parts of the Central Plateau are characterized by gentle, open folds or simply flat south-east dipping beds. Recumbent folds are completely absent, their place being taken by a brittle phase of deformation. This may be attributed to the expulsion of pore fluids during the formation of cleavage. Finally, only one phase of pre-Cape deformation is apparent in the study area, as opposed to the recognition of at least two stages in the pre-Cape rocks of the George area. (Gresse, 1976).

Table 8 summarizes the main orientation of the tectonic axes as inferred from synoptic diagrams of the various structural elements.

TABLE 8

	Bedding	Cleavage	Joints	δ-lineation	b-lineation	a-lineation.
σ_1	010°	015°	005°	012°	012°	008°
σ_2	100°	105°	095°	102°	102°	098°
σ_3	68°S	70°S	(See Fig. 85).	72°S	Vertical	65°S

The nature of the deformational forces should be sought in the style of deformation. According to De Sitter (1964) folds originating under lateral compression are characterized by stretched, thinned or ruptured overturned limbs, while those forming as a result of gravitational sliding show none of these effects.

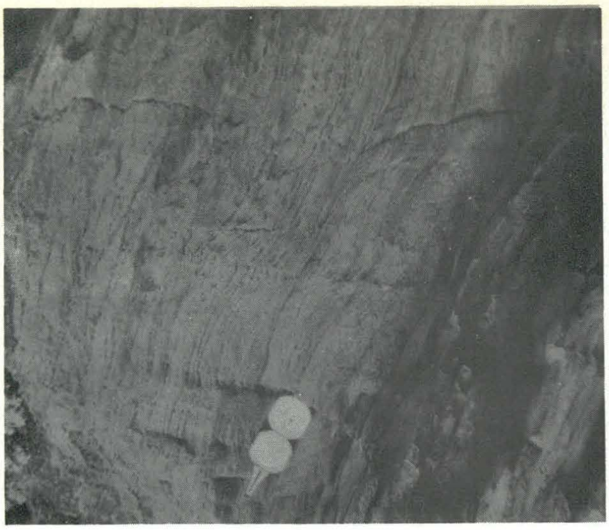
As considerable evidence of stretching and rupturing of inverted flanks in the Cango exists (e.g. Boomplaats Fault, the a-lineation of incompetent pebbles), while none of the features typical of gravitation folding is present, the forces responsible for the Cango orogeny are attributed to tangential compression from the south.

The structure of the Cango basement rocks also casts more light on the origin of the Cape Fold Belt. Of the two hypotheses proposed to explain this orogeny, i.e. tangential compression (De Villiers, 1944; Le Roux, 1974) and gravitational sliding (Newton, 1973), the latter has merit only if the basement as a whole did not participate in this event.

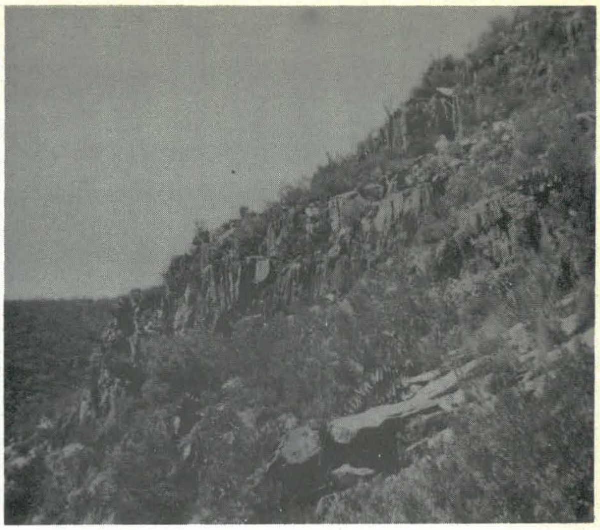
At the contacts between the Cape rocks and the Cango Group immediately north of the Cango Fault, shearing and distortion of the pre-Cape sediments is a common feature. This effect decreases away from the contact and is therefore not of pre-Cape origin, but results from differential shear between the two units. Similar shearing occurs between the Table Mountain Sandstone and Schoemans Poort Formation and between the former and the Drooge Kraal diabase in Potgieters Poort. Because the above-mentioned phenomena pertain to the contact, they could still be explained by thin-skinned gravitational sliding, but their combination with the following features shifts the balance in favour of tangential compression: Flow cleavage in the Cango rocks as revealed in thin sections is commonly displaced by fracture cleavage which suggests a more brittle reaction to what could have been the Cape orogeny. This fracture cleavage in turn has been deformed by kink bands which are even younger. A faint foliation, (possibly fracture cleavage) was also observed in places in the Drooge Kraal Sill (Plate 7 (b)), which intruded the Cango rocks after their deformation.

A regional feature previously mentioned, is the arcuate outcrop of the Table Mountain Sandstone in the Swartberg, which conforms roughly to the structure of the Cango as revealed in the present analysis. There may be two reasons for this: The post-Cape structural trend was either controlled by existing lineaments in the Cango basement, or the major pre-Cape folds were modified to conform to the Cape Fold pattern. A combination of both mechanisms is also likely, a small adaptation on either side then being sufficient to accommodate the difference, if any, in the original direction of maximum compression. Whatever the case may be, this precludes the possibility of independent sliding of Cape rocks over the Cango basement.

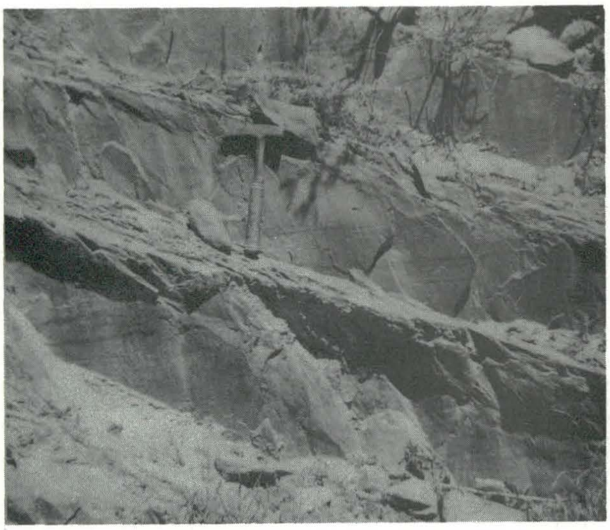
PLATE 7.



(a) Fracture cleavage deformed by kink bands in Uitvlugt Formation. Looking west.

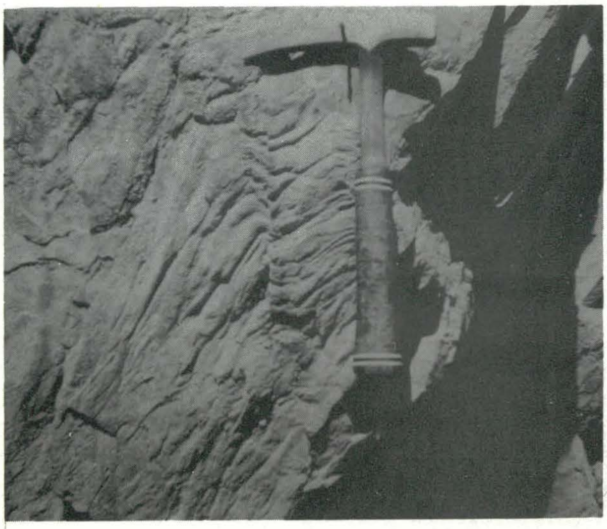


(b) Cleavage in Drooge Kraal Diabase sill. Looking east.



(c) Kink bands in Groenefontein Formation. Looking south.

(d) Folded cleavage in Gezwinds Kraal Formation, De Cango. Looking east.



(e) False sole structures dis= placed along joint, Groene= fontein.

(f) Quartz rod in Uitvlugt For= mation, De Kruys. Looking east.

6. SYNTHESIS

The geologic history of the Cango Group must now be examined within its larger tectonic framework, in order to establish a model which integrates the geological features most satisfactorily. The sequence of events as inferred from the present study can be summarized as follows:-

The regional depositional setting seems to have been a long, narrow embayment (Fig. (64.) subject to periodic flooding of its shores. Tectonic uplift was restricted mainly to the southern land arm.

The sediments of the Nooitgedagt Member (Matjies River Forma= tion) resemble shelf deposits, where repeated transgressions of the sea towards the west resulted in inner shelf clastics interbedded with outer shelf limestones. An abrupt cessation of coarse detrital matter marked the onset of the Cave Limestone deposition, possibly as a result of isolation from the source area to the west. A tropical or subtropical climate would have been favourable for this large scale carbonate deposition.

Continued submergence of the isolated platforms surpassed the rate of carbonate build-up and deeper water environments prevailed. Slope facies are represented by the transitional Lower Member of the Groenefontein Formation, while turbidity currents supplied the fine-grained quartz wackes of the Middle Member. These sediments may have been derived from the flanks of the trough and were apparently transported in a westerly direction along its axis. The coarser grained sandstone and limestone lenses of the Upper Member suggest a return to shallower water conditions.

This uplift rapidly accelerated in the southwest and part of the Groenefontein basin was exposed above sea level. Polymictic conglomerates, partly derived from the Upper Member, were deposited in alluvial fans along the margin of the newly elevated highlands. Where these piedmonts extended down to the coast quartz pebble conglomerate was deposited in beach environments, both through elimination of softer pebble types by wave action and by the extension of the supplying rivers deeper inland. A more rigorous climate prevailed during Vaartwell times.

Another major cycle of transgression is reflected in an upward transition to the nearshore deposits of the Danzers Kloof Member (Uitvlugt Formation) which grade upwards into the Rooiberg Member. The latter is interpreted as offshore, infralittoral deposits under the influence of powerful ocean currents travelling in an easterly direction.

As the sea invaded the western part of the embayment deeper water environments developed in which the sedimentation of the Gezwinds Kraal Formation took place. In the east, however, the facial-equivalent Schoongezigt Formation was deposited in shallow water surroundings.

The Cango rocks were subsequently deformed as a result of lateral tangential compression as shown by the style of folding and thrust faulting in the area. The arcuate megastructure conforms to the outcrop pattern of the overlying Cape sediments. This indicates that the Cango basement either controlled the trend of the Cape Fold Belt in this area or took part in the post-Cape orogeny, which strongly reduces the possibility of a gravitational - sliding mechanism for the latter.

Metamorphism reached the lower greenschist facies.

Uplift subsequently exposed the whole Cango basin,with resulting erosion and deposition of the unconformable Schoemans Poort sediments along the southern part of the present Cango outcrop.

From the preceding discussion it is clear that there is a marked relation between tectonism and concomitant sedimentation in the Cango Group. Strong uplift in the source area was accompanied by rapid filling of the subsiding basin by coarse clastics, while subsidence exceeding the rate of accumulation resulted in greater depths of deposition. These diastrophic movements - only slightly disturbing the attitude of the sediments - were largely epeirogenic in the depositional basin but probably had a more orogenic nature in the source areas to the south.

Placing the Cango Group in its correct tectonic or geosyn=clinal setting requires a closer look at the lithologic associations. (See Krumbein and Sloss, 1963). The Matjies River sediments display many of the characteristic attributes of unstable shelf deposits, among which may be mentioned the cyclical repetition of beds as a result of oscillatory movements, poor to moderately sorted sandstones occurring in sheets or lenticular masses, and the abundance of calcareous, silty shales and fine-grained limestones containing a considerable clay admixture. The relatively great thickness and the absence of disconformities in these deposits however, combined with their stratigraphic relation=ships, shift the balance towards the marginal basin association. During times of relative inactivity in the adjacent eugeo=syncline, sediments deposited in marginal basins are similar in many respects to shelf deposits, especially if access to the open sea permits normal circulation conditions. However, the more rapid rate of burial and increased thick=ness of this association reflect a greater rate of subsidence than normally occurs on shelves, while less frequent uplift above depositional base level also ensures more continuous sedimentation.

Assuming that the Matjies River Formation represents a marginal basin association, the remaining Cango formations must be investigated in this framework.

Marginal basins represent the transitional realms between eugeosynclines and cratons and in location thus resemble the original "miogeosynclines" of Stille (1936). These were interpreted as continuous, linear belts separating cratons from the eugeosynclines and differing from the latter only in their less intense deformation and absence of volcanics. This original concept is however not fully supported by more recent evidence (e.g.Sloss, 1956), which suggests the existence of localized tectonic centres such as a series of marginal basins in a zone paralleling the eugeosyncline. "It is almost as though the craton deteriorated irregularly along the edges, in a transitional zone that... may contain both positive and negative tectonic features; and that may at times receive sedimentary fillings characteristic either of the craton or of the eugeosyncline, depending on the degree of tectonic activity in the geosynclinal belt." (Krumbein and Sloss, 1963).

Sediments may be derived directly from the eugeosynclines during times of marked orogenic activity or may result from erosion of local marginal uplifts along the basins, these uplifts usually displaying structural trends parallel to the orogenic trends in the eugeosyncline.

The fine-grained quartz wackes of the Groenefontein Formation are typical of the flysch deposits common in eugeosynclines, but differ in the absence of associated volcanics, in their low metamorphic grade and in their geochemistry. It has been suggested that the high K_2O/ Na_2O and CaO/MgO ratios of these rocks possibly reflect the non-volcanic nature of the source areas (p. 94), which can therefore be assumed to have been marginal uplift areas rather than orogenic belts in the eugeosyncline. The Groenefontein Formation thus represents a "greywacke" association of intermediate or shallow burial similar to the "type" Alpine flysch, which was deposited in marginal basins on either flank of the Alpine orogenic trend.

The Vaartwell and Uitvlugt Formations may be regarded as clastic wedge associations. Comparison with the Frontier Formation of Wyoming (Goodell, 1957) reveals a striking resemblance in most important attributes. These characteristics include the derivation from a source area dominated by sedimentary rocks (including carbonates, quartzite, chert and argillite) which were deposited in both marine and non-marine environments, the vector properties such as crossbedding exhibited by the sandstones, the wedge-shaped cross sections which are thickest near the provenance, and the presence of very coarse conglomerate indicating high stream gradients "scarcely typical of known delta systems." Accumulation of the Frontier was on a broad piedmont plain lying between the sea and the elevated mountains and rock fragments were cannibalized from older sediments. The tectonic setting of this clastic wedge is interpreted as a broad shelf area interrupted by marginal basins, and may thus be regarded as a miogeosynclinal zone. Similar sediments of the Alpine foreland have been termed molasse, and this is adopted for the Vaartwell and Uitvlugt Formations in the present study.

The combined Gezwinds Kraal and Schoongezigt Formations seem to represent a rather special case and do not fit into the classical scheme of lithologic associations of Krumbein and Sloss. (1963) A return to deeper water, "flysch" environ= ments is however suggested by the sedimentology of the Gezwinds Kraal.

It is evident from the above that the marginal basin asso= ciation within the miogeosynclinal zone, represents the most satisfactory model in explaining the development of the Cango Group. The embayment visualized from the environmental studies hopefully contributes to a better understanding of the paleogeography of the "miogeosyncline", which at present is still in a state of flux as more and more know= ledge accumulates.

Generalizations about continental drift as the ultimate source of sedimentary tectonics in the Precambrian will be premature at this stage, as much more evidence is needed to verify such an hypothesis. However, the writer would like to draw attention to the many similarities between the development of geosynclines throughout the geologic record. The time sequence of events in the Cango Group is matched by many older, time- equivalent and younger geosynclinal areas, and this certainly points to a more fundamental process in the geodynamic history of the earth.

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A P P E N D I X

TABLE X - 1 GRAIN SIZE PARAMETERS

SAMPLE	Mdφ	σφ	αφ
MATJIES RIVER FORMATION			
M 1^1	1,07	0,63	+0,165
M 1^2	1,24	0,81	+0,095
T 76	1,20	1,19	-0,431
M 2	1,81	0,88	-0,400
M 2^1	1,25	1,48	-0,270
W 23	1,46	1,10	+0,365
W 1	1,07	1,37	+0,230
GROENEFONTEIN FORMATION			
G 7	2,38	1,17	+0,126
G 6	3,37	0,93	-0,247
E	3,69	0,74	-0,365
G 11	2,71	1,03	-0,254
G 1	3,55	0,57	+0,190
G 10	3,66	0,65	-0,060
B	3,87	0,72	-0,415
G 4	3,18	0,93	+0,154
G 5	3,79	0,57	-0,239
G 8	3,47	0,72	+0,190
UITVLUGT FORMATION			
U 13	1,40	1,32	+0,184
U 10	1,98	1,28	+0,196
W 51	3,09	0,96	-0,170
U 9	2,14	1,10	+0,258
U 12	1,35	1,21	+0,174
U 6	1,29	0,75	-0,128
U 11	1,51	0,71	+0,053

SAMPLE	Mdφ	σφ	αφ
U 2	1,67	0,81	+0,156
U 16	1,59	0,75	+0,017
U 7	2,12	0,79	-0,053
U 8	2,34	0,95	+0,073
U 14	1,55	1,08	+0,154
U 3	1,97	0,84	+0,115
U 4	1,69	0,68	+0,047
GEZWINDS KRAAL FORMATION			
L 3	3,17	0,77	+0,292
G 23	4,13	0,45	-0,493
Gz 8	3,62	0,64	-0,401
Gz 9	3,77	0,53	-0,414
Gz 3	3,98	0,42	-0,626
Gz 4	3,63	0,68	-0,015
Gz 5	3,87	0,69	-0,547
Gz 1	3,58	0,63	+0,063
Gz 6	3,52	0,71	-0,005
Gz 7	3,75	0,55	+0,123
Gz m	4,03	0,53	+0,109
SCHOONGEZIGT FORMATION			
D 1	2,17	1,01	+0,063
D 2	2,18	1,08	+0,352
D 3	2,24	1,11	+0,151
SCHOEMANS POORT FORMATION			
S 2	1,25	1,50	+0,105
S 3	1,41	0,97	-0,014
S 69 -1	2,22	0,92	+0,033

TABLE X - 2 CHEMICAL COMPOSITION (XRF)

SAMPLE	SiO_2	TiO_2	Al_2O_3	Total Fe	MnO	MgO	CaO	Na_2O	K_2O	P_2O_5	H_2O+ LOI	Total
M 2	75.00	0.73	9.25	3.28	0.09	1.58	1.52	0.61	3.67	0.01	2.92	98.66
M 1[2]	92.38	0.24	4.22	0.15	0.03	0.32	0.00	0.12	1.61	0.00	0.83	99.90
E	73.78	0.62	13.55	4.92	0.07	0.00	1.37	0.87	2.22	0.06	2.66	100.12
B	67.10	0.68	13.87	5.54	0.10	1.32	1.23	0.80	2.70	0.06	2.77	96.17
G 10	65.92	0.31	10.75	8.82	0.52	0.06	2.76	0.06	2.37	0.05	4.42	96.54
G 8	80.25	0.42	10.36	2.89	0.04	0.00	0.16	0.62	3.24	0.00	1.99	100.00
S 70	54.75	0.87	20.96	7.89	0.09	0.90	1.33	0.32	5.17	0.13	3.84	96.25
G 6	76.72	0.49	12.68	3.53	0.07	0.00	0.63	0.74	2.58	0.09	2.13	99.66
G 5	72.45	0.52	13.39	3.97	0.05	0.00	0.13	0.72	3.59	0.07	2.64	97.53
G 4	77.70	0.72	10.42	3.56	0.07	0.00	1.53	0.75	1.29	0.05	1.50	97.59
G 1	76.00	0.69	10.67	4.36	0.11	0.23	0.68	0.41	1.98	0.15	2.05	97.33
U 12	70.92	0.18	6.51	1.98	0.02	1.70	0.80	0.15	2.00	0.05	15.65	99.96
U 13	84.68	0.36	8.14	1.10	0.03	0.84	0.10	0.33	3.74	0.00	1.07	100.39
U 15	86.61	0.21	6.46	1.46	0.04	1.20	0.14	0.50	1.93	0.03	1.18	99.76
U 16	76.98	0.54	10.46	4.29	0.06	2.42	0.02	0.11	2.74	0.02	1.97	99.61
U 4	91.50	0.14	3.85	1.57	0.04	0.70	0.00	0.09	0.73	0.00	1.99	100.61
U 3	93.93	0.16	2.44	1.31	0.02	0.58	0.00	0.09	0.57	0.00	0.86	99.96
U 2	89.89	0.42	4.22	1.27	0.03	0.50	0.00	0.12	1.68	0.07	0.86	99.06
Gz 5	75.60	0.69	12.35	4.89	0.06	0.00	1.01	0.08	2.63	0.16	2.47	99.94
Gz 9	68.86	0.65	15.68	4.74	0.05	0.00	0.60	0.79	3.72	0.01	2.34	97.44
Gz 7	76.71	0.58	11.88	5.28	0.08	0.00	0.38	0.39	2.25	0.07	2.24	99.86
Gz 8	79.00	0.70	8.30	3.80	0.08	0.00	2.15	0.26	1.39	0.15	2.64	98.47
Gz 4	78.27	0.51	10.37	3.74	0.10	2.80	0.38	0.32	1.68	0.13	1.81	100.11
Gz 1	79.11	0.48	8.32	3.01	0.13	2.00	1.76	0.53	1.34	0.06	2.35	99.09
D 2	85.91	0.27	5.36	2.08	0.06	1.52	0.54	0.19	1.62	0.00	1.52	99.07
D 1	81.35	0.39	7.46	2.88	0.04	1.66	0.19	0.43	2.05	0.02	1.41	97.88
S 3	78.82	0.21	10.21	1.85	0.04	1.50	0.38	0.01	3.50	0.09	1.70	98.31
S 2	78.07	0.23	10.54	2.08	0.05	2.13	0.08	0.08	4.45	0.08	1.72	99.51

M.R. = Matjies River Formation; Gr = Groenefontein Formation;
Ui = Uitvlugt Formation; Gz = Gezwinds Kraal Formation;
Sc = Schoongezigt Formation; Sp = Schoemans Poort Formation;


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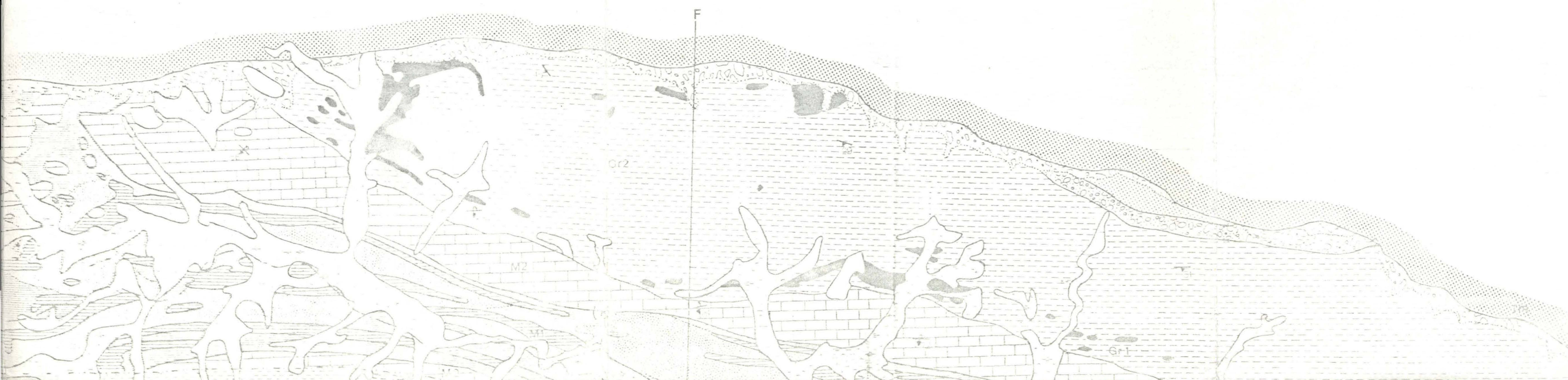
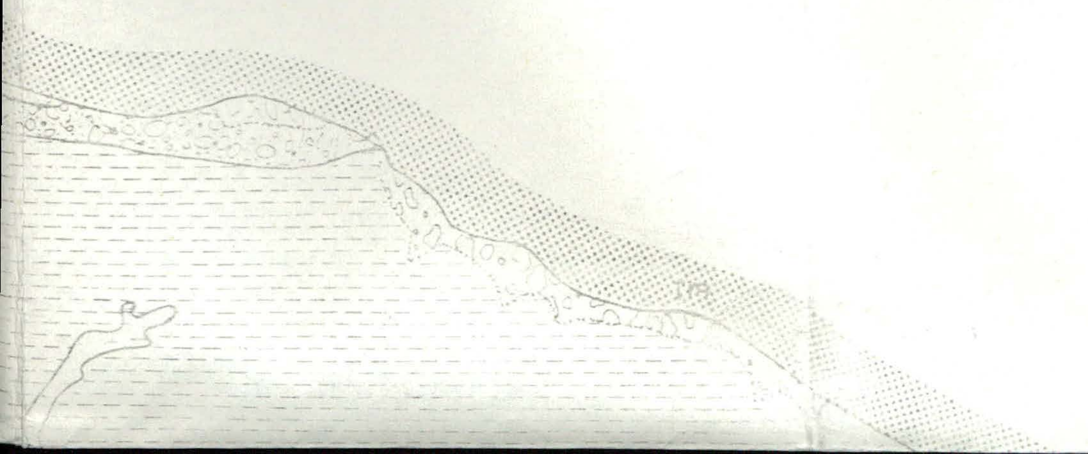


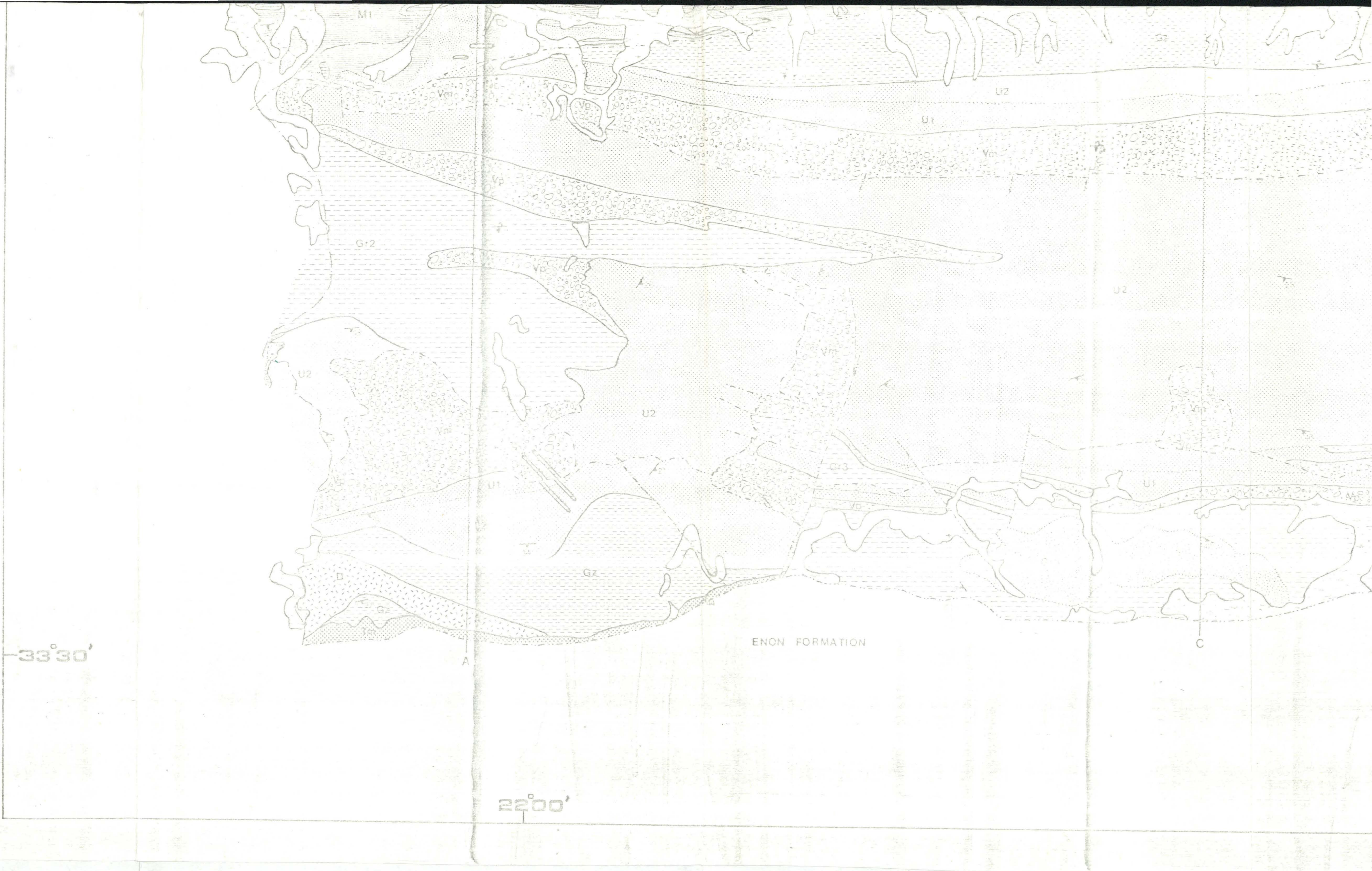
ROUP EAST OF 21°57'

LEGEND

	…uvium			
	Tertiary Terraces			
Tm	Table Mountain Sandstone			
D	Diabase			
Sp	Grits, wackes, subarkose		SCHOEMANS POORT FORMATION	
Sc	Conglomerate, wackes, shale		SCHOONGEZIGT FORMATION	CANGO GROU
Gz	Fine-grained wackes, shale		GEZWINDS KRAAL FORMATION	
U2	Crossbedded wackes, arenites, subarkose	ROOIBERG MEMBER	UITVLUGT FORMATION	
U1	Crossbedded wackes, quartz pebbles	DANZERS KLOOF MEMBER		
Vm	Quartz pebble conglomerate	MONOMICTIC MEMBER	VAARTWELL FORMATION	
Vp	Polymictic conglomerate	POLYMICTIC MEMBER		
Gr3	Grits, arenites, limestone lenses	UPPER MEMBER	GROENEFONTEIN FORMATION	
Gr2	Fine-grained wackes, shale	MIDDLE MEMBER		
Gr1	Fine-grained wackes, limestone lenses	LOWER MEMBER		
M2	Limestone, siltstone, shale	CAVE MEMBER	MATJIES RIVER FORMATION	
	Limestone-shale	NOOITGEDAGT MEMBER		
M1	Shale			
	Wackes, arenites, subarkose			

Main contacts
Other contacts
Faults
Main roads
Dip and strike of bedding

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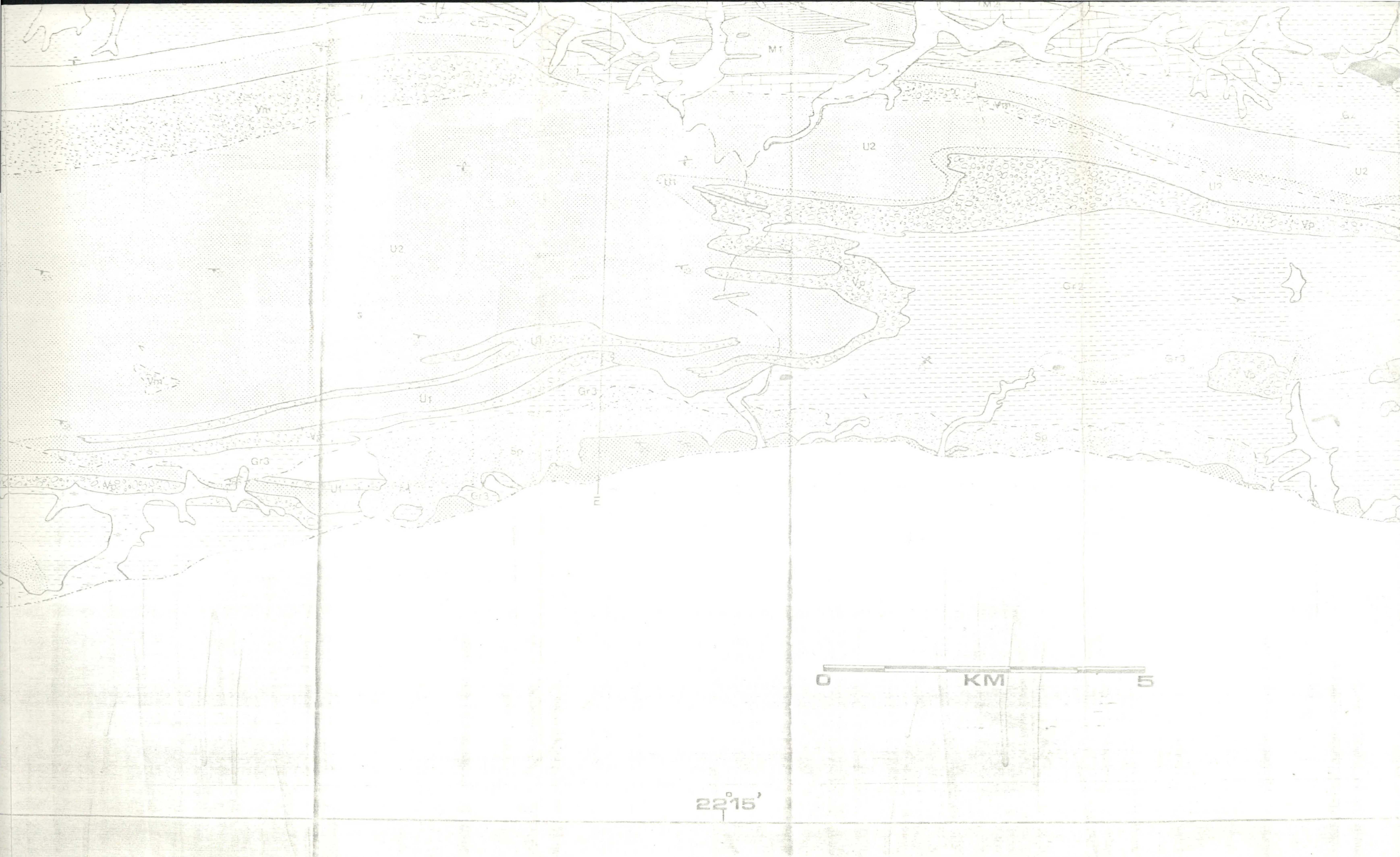
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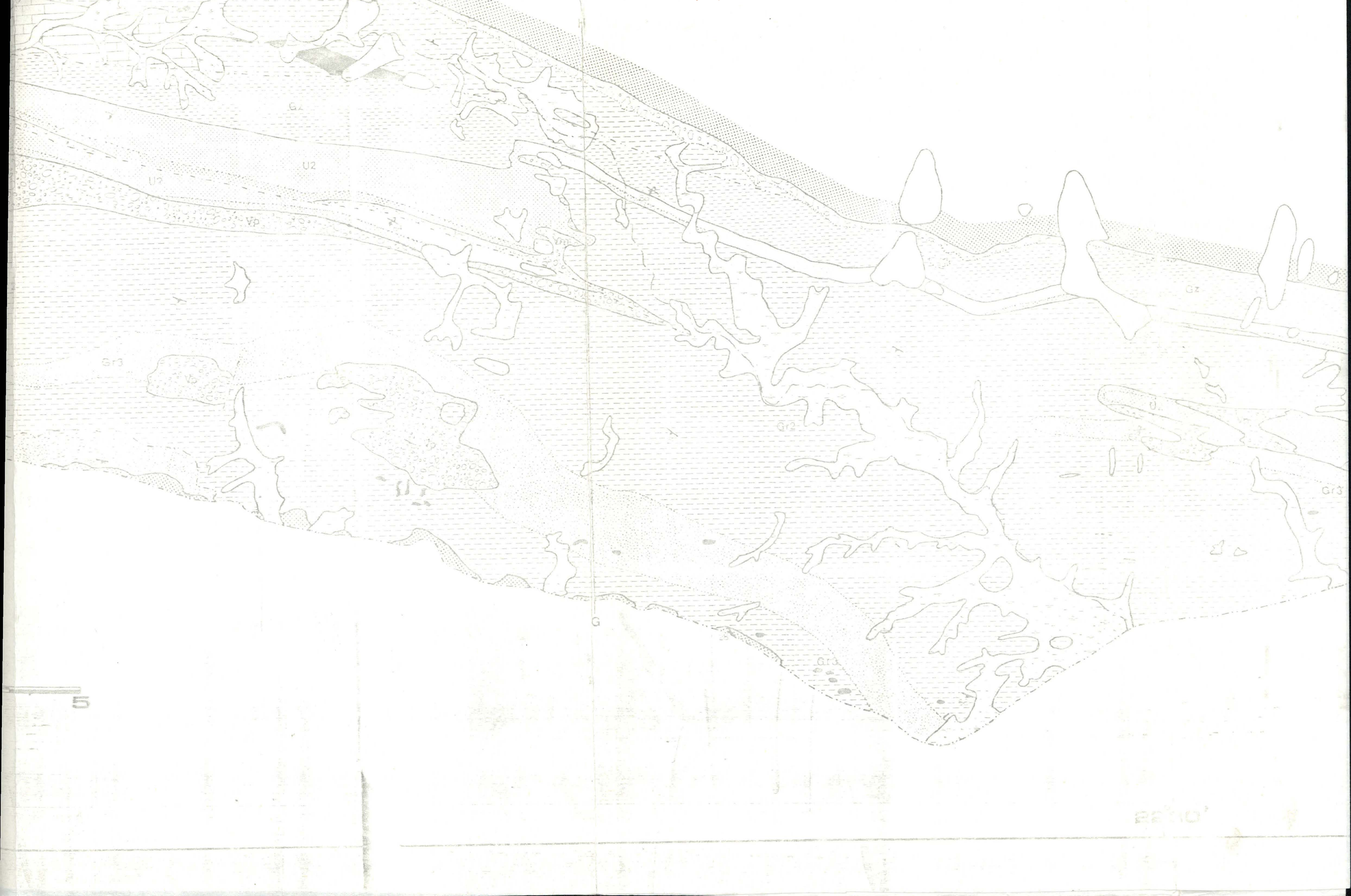
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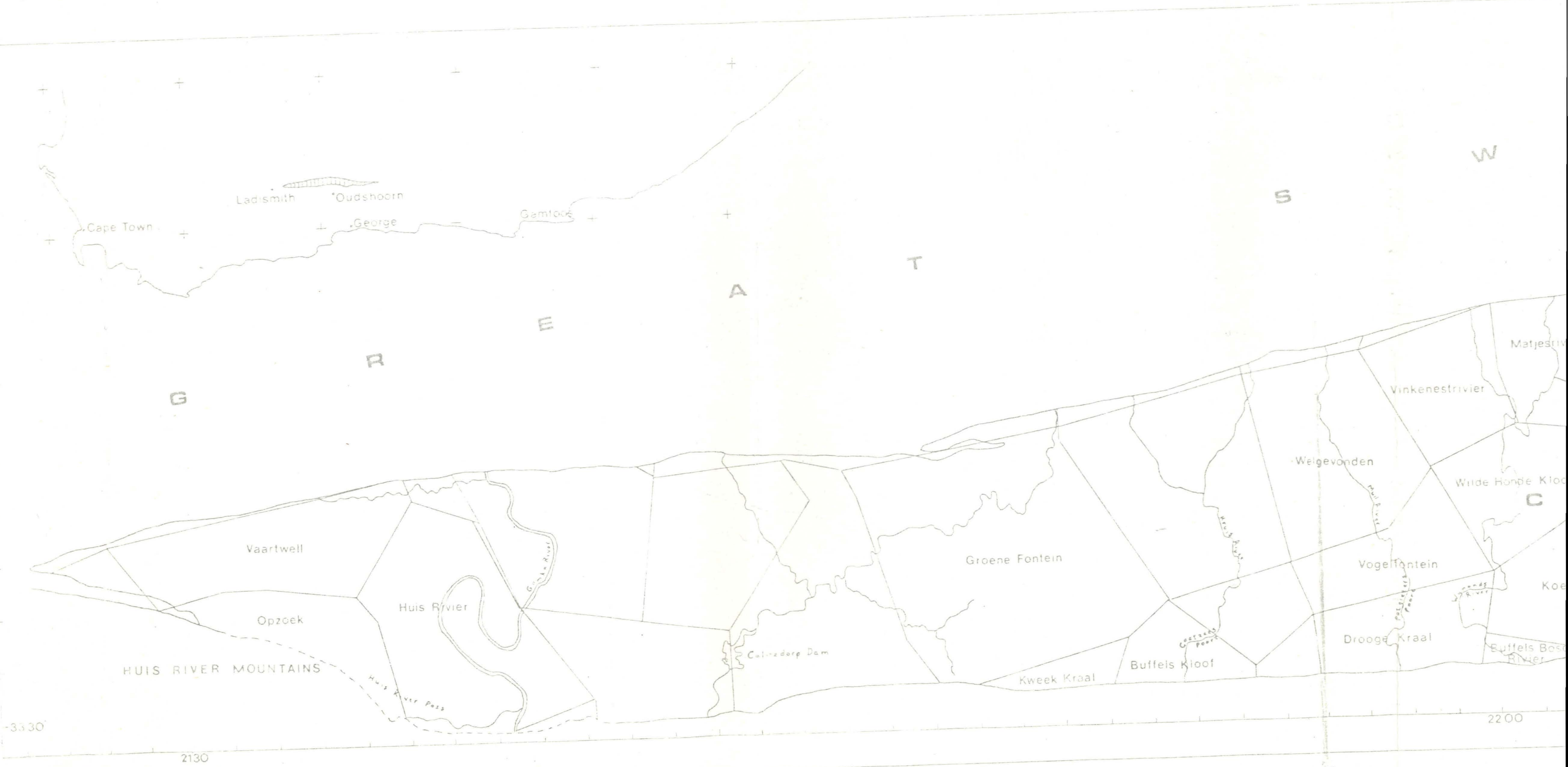
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Groene Fontein
Vogelfontein
Opzoek
Huis Rivier
Drooge Kraal
Buffels Kloof
Kweek Kraal
Calitzdorp Dam
HUIS RIVER MOUNTAINS
Huis River Pass
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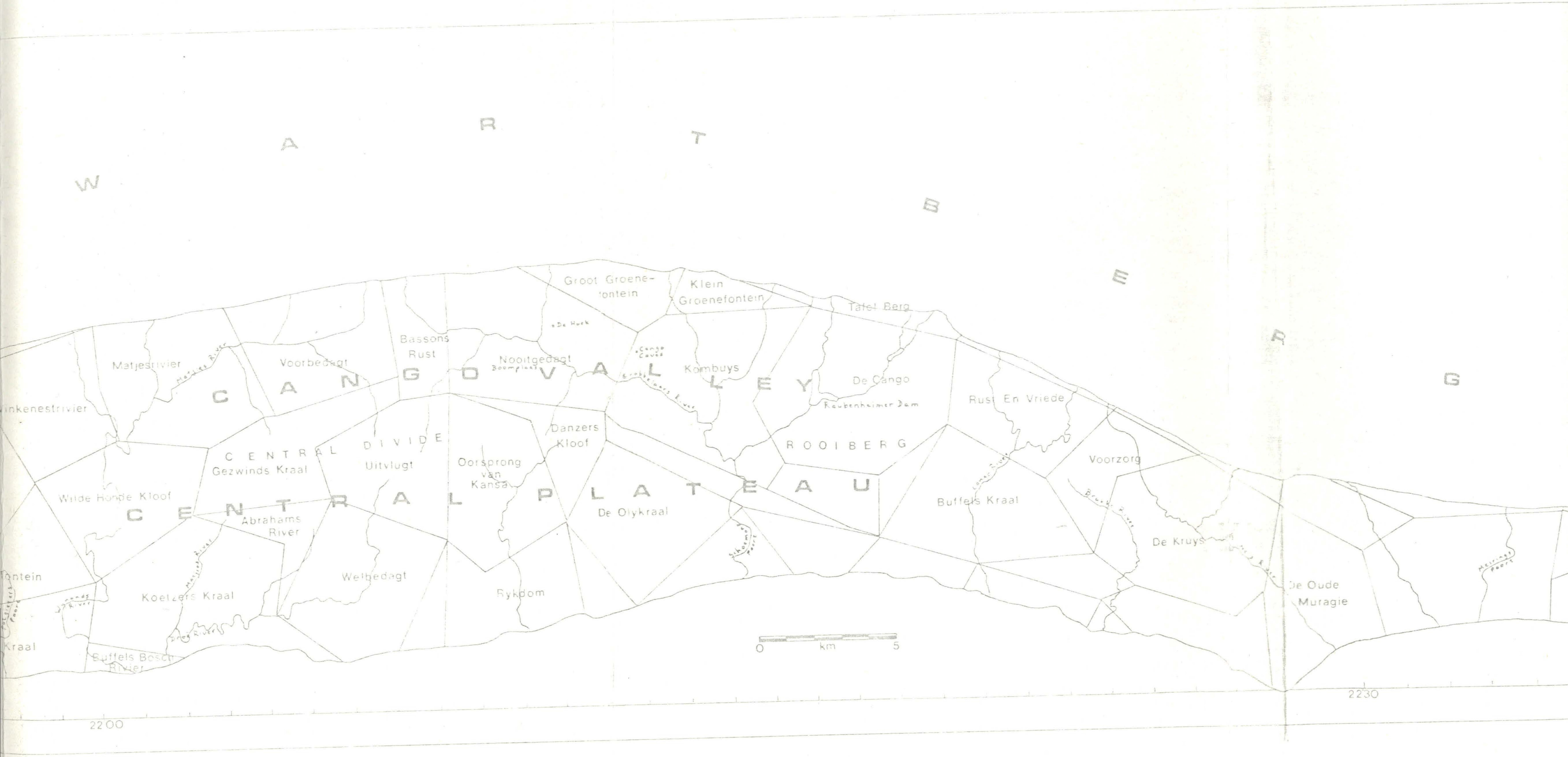
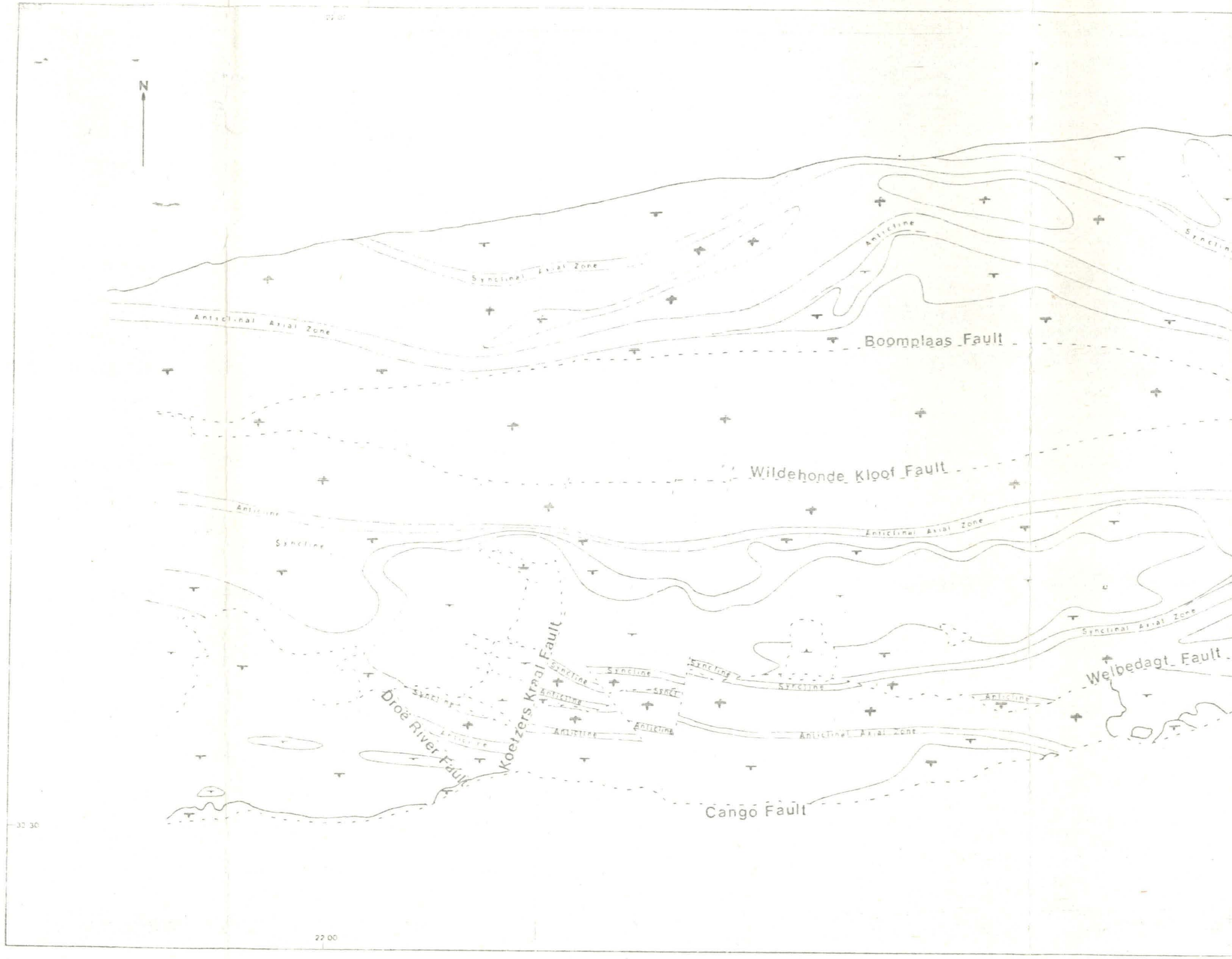
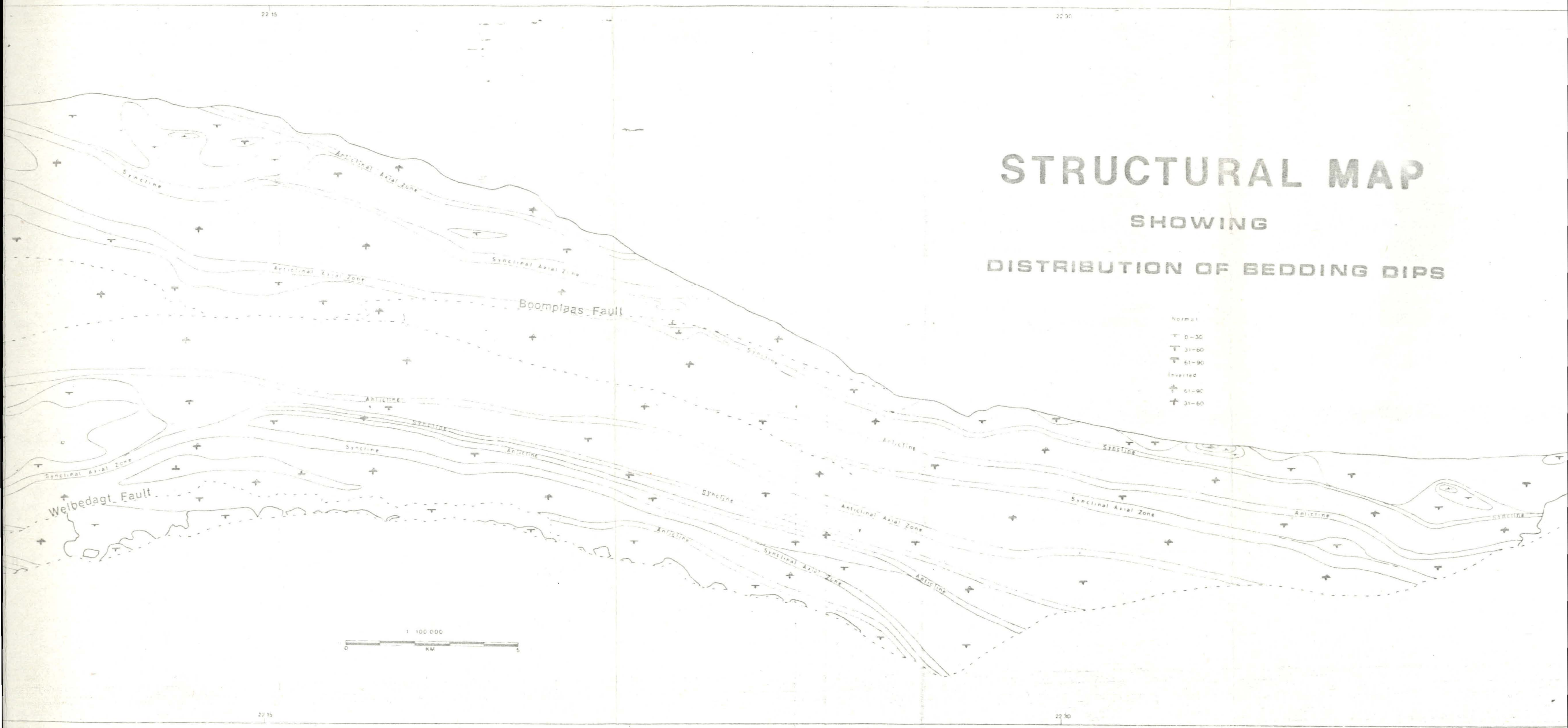
2130
2200


W A R T B E R G
C A N G O V A L L E Y
C E N T R A L P L A T E A U
C E N T R A L D I V I D E
R O O I B E R G
Groot Groene-
fontein
Klein
Groenefontein
Tafel Berg
Bassons
Rust
Nooitgedagt
Kombuys
De Cango
Rust En Vriede
Matjesrivier
Voorbedagt
Danzers
Kloof
Voorzorg
Gezwinds Kraal
Uitvlugt
Oorsprong
van
Kansa
Wilde Honde Kloof
Abrahams
River
De Olykraal
Buffels Kraal
De Kruys
Koelzers Kraal
Welbedagt
Rykdom
De Oude
Muragie
Buffels Bosch
Rivier
Kraal
0 km 5
2200
2230

LOCALITY MAP


N
Synclinal Axial Zone
Anticlinal Axial Zone
Boomplaas Fault
Wildehonde Kloof Fault
Anticlinal Axial Zone
Synclinal Axial Zone
Welbedagt Fault
Koetzers Kraal Fault
Droë River Fault
Cango Fault
22.00


STRUCTURAL MAP
SHOWING
DISTRIBUTION OF BEDDING DIPS
Normal
0-30
31-60
61-90
Inverted
61-90
31-60
22 15
22 30
Boomplaas Fault
Welbedagt Fault
Syncline
Anticline
Anticlinal Axial Zone
Synclinal Axial Zone
1 : 100 000
0
KM
5

URAL MAP

WING

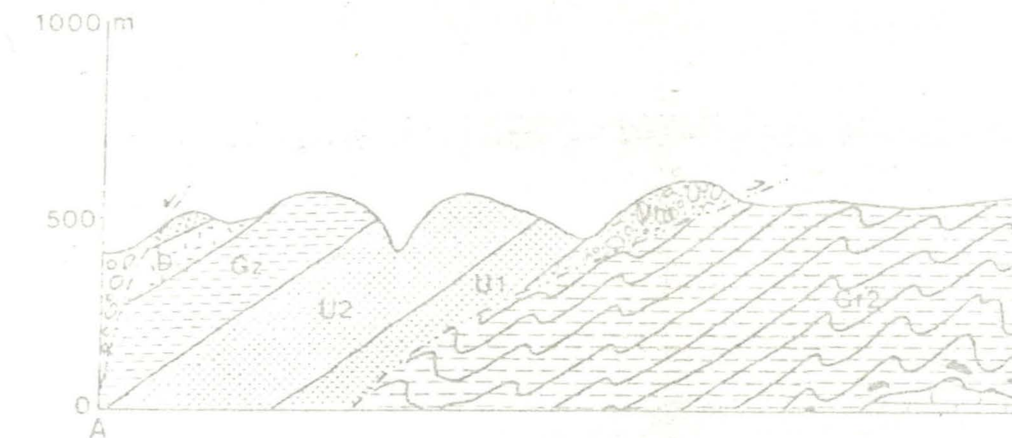
OF BEDDING DIPS

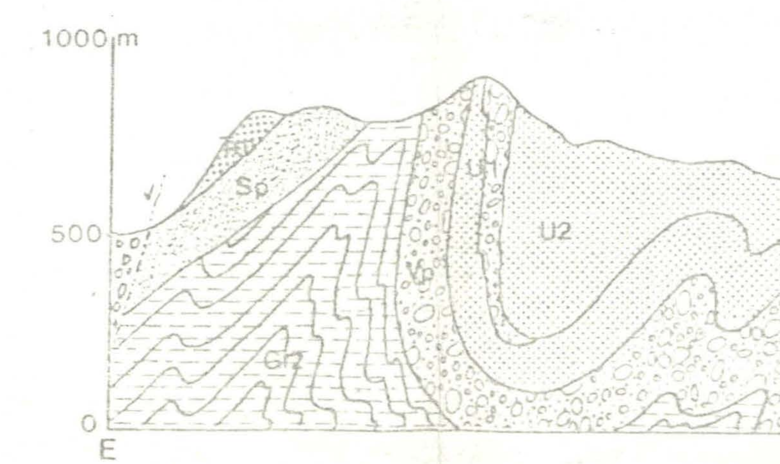
T OF 21°57'

LEGEND

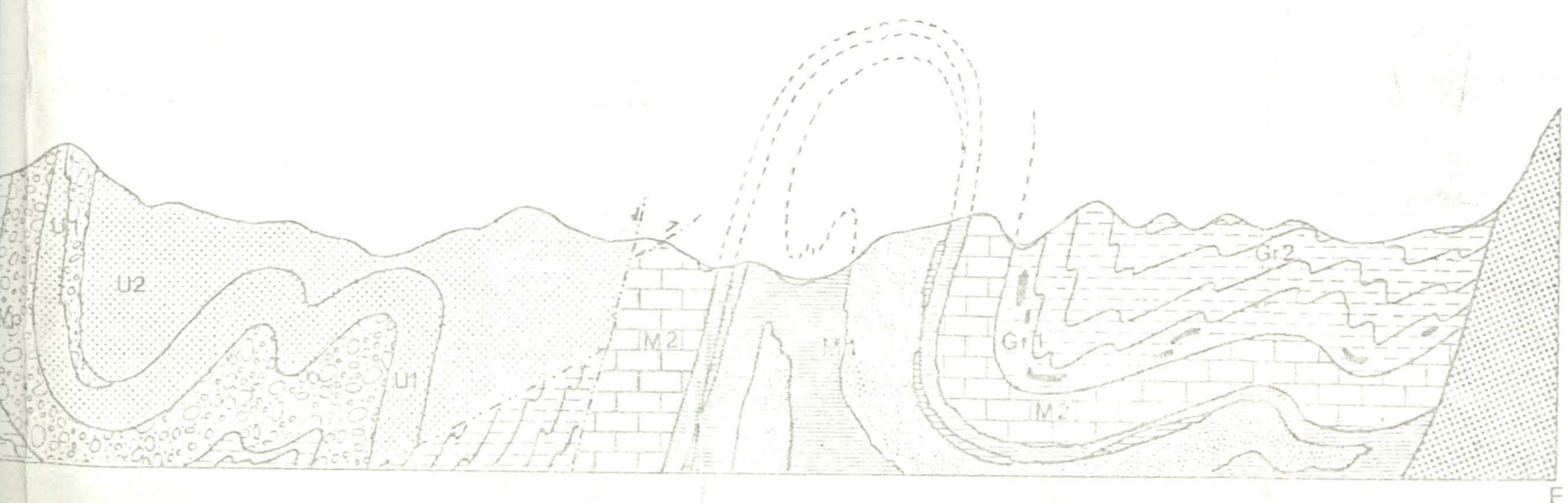
	Talus			
	Alluvium			
	Tertiary Terraces			
Tm	Table Mountain Sandstone			
D	Diabase			
Sp	Grits, wackes, subarkose		SCHOEMANS POORT FORMATION	
Sc	Conglomerate, wackes, shale		SCHOONGEZIGT FORMATION	CANGO GROUP
Gz	Fine-grained wackes, shale		GEZWINDS KRAAL FORMATION	
U2	Crossbedded wackes, arenites, subarkose	ROOIBERG MEMBER	UITVLUGT FORMATION	
U1	Crossbedded wackes, quartz pebbles	DANZERS KLOOF MEMBER		
Vm	Quartz pebble conglomerate	MONOMICTIC MEMBER	VAARTWELL FORMATION	
Vp	Polymictic conglomerate	POLYMICTIC MEMBER		
Gr3	Grits, arenites, limestone lenses	UPPER MEMBER	GROENEFONTEIN FORMATION	
Gr2	Fine-grained wackes, shale	MIDDLE MEMBER		
Gr1	Fine-grained wackes, limestone lenses	LOWER MEMBER		
M2	Limestone, siltstone, shale	CAVE MEMBER	MATJIES RIVER FORMATION	
M1	Limestone-shale Shale Wackes, arenites, subarkose	NOOITGEDAGT MEMBER		

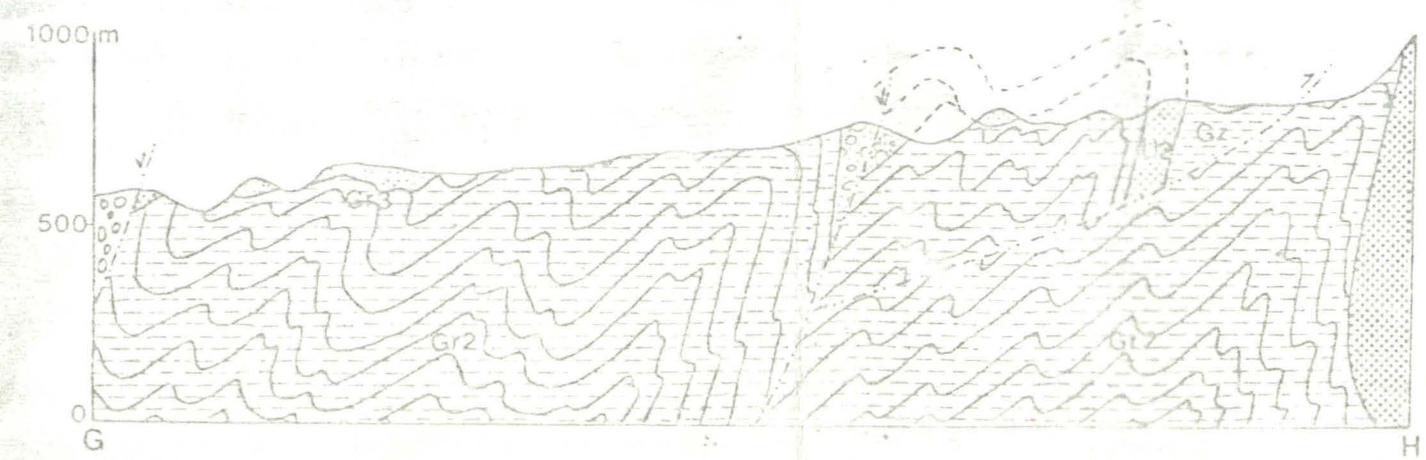
- Main contacts
- Other contacts
- Faults
- Main roads
- Dip and strike of bedding

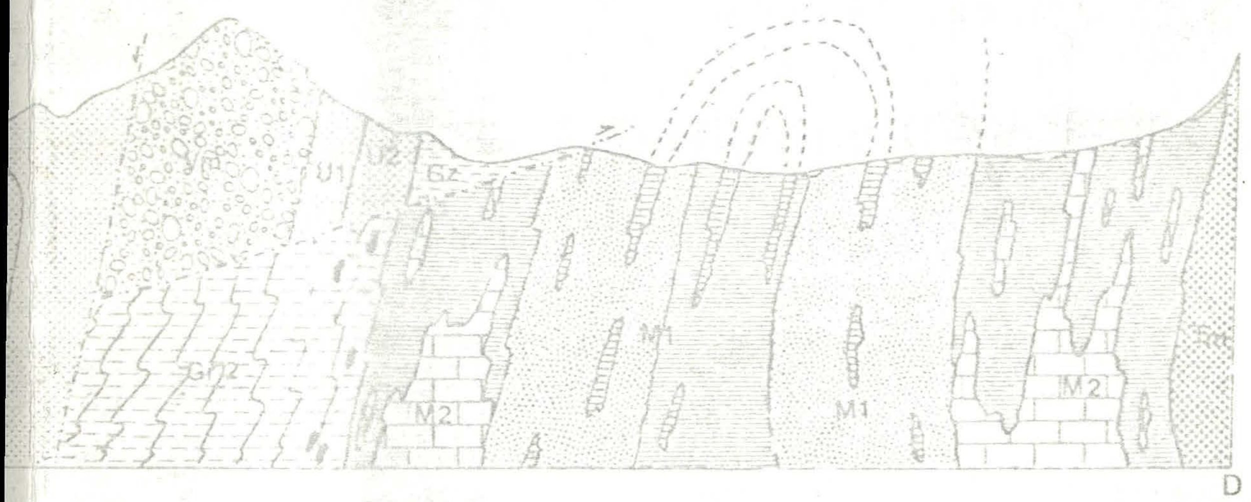


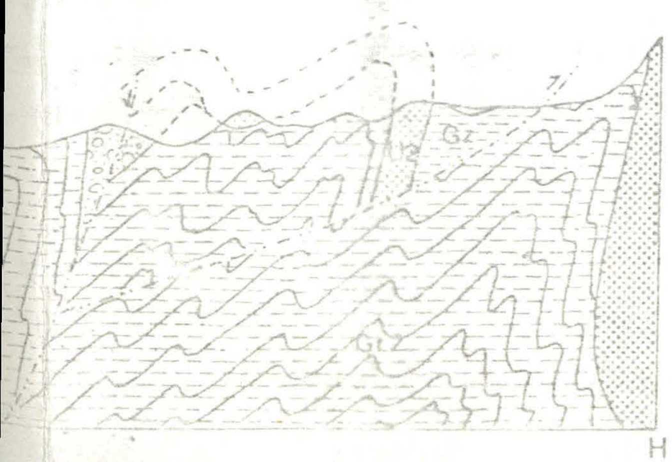


GEOLOGICAL SECTIONS

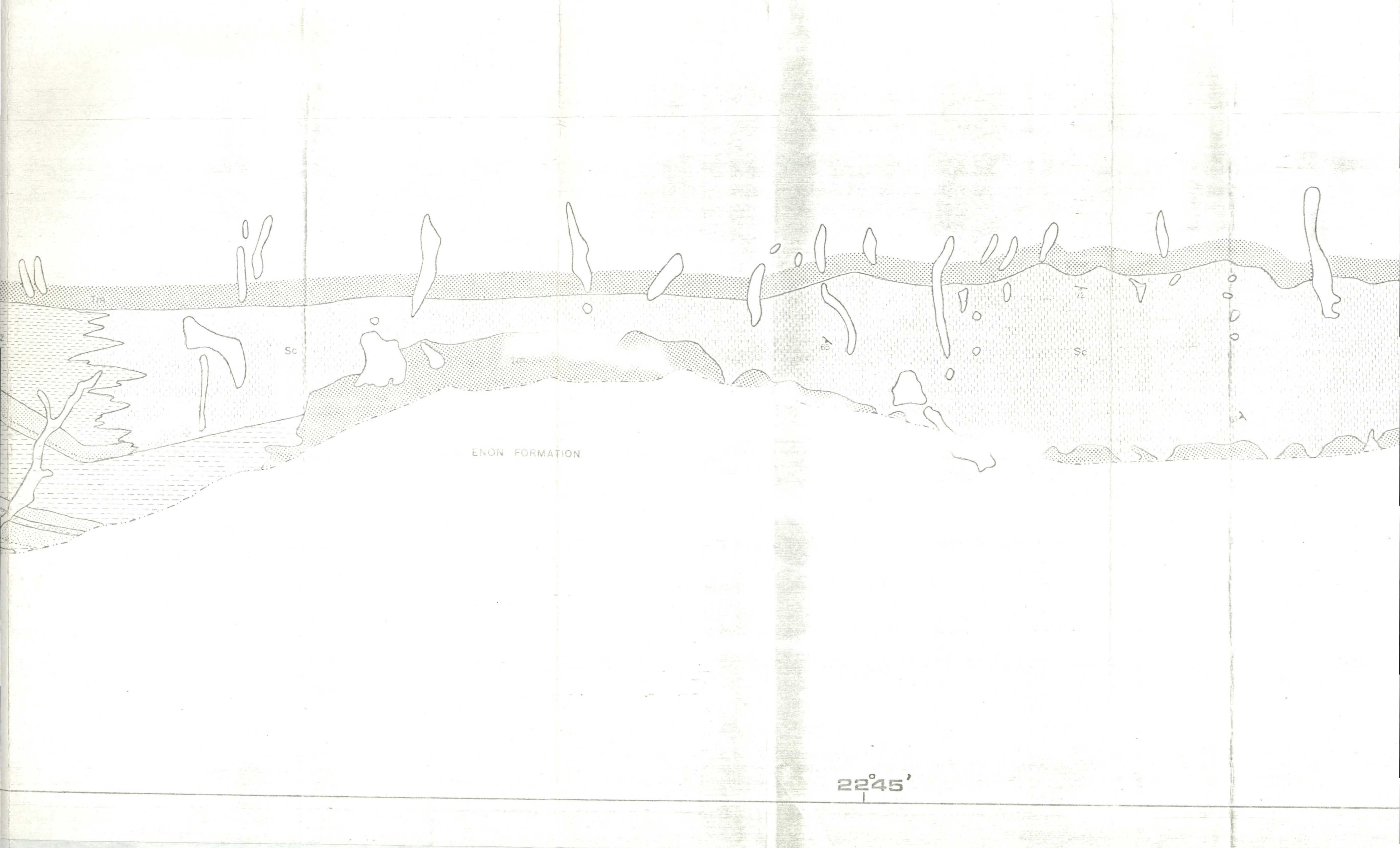


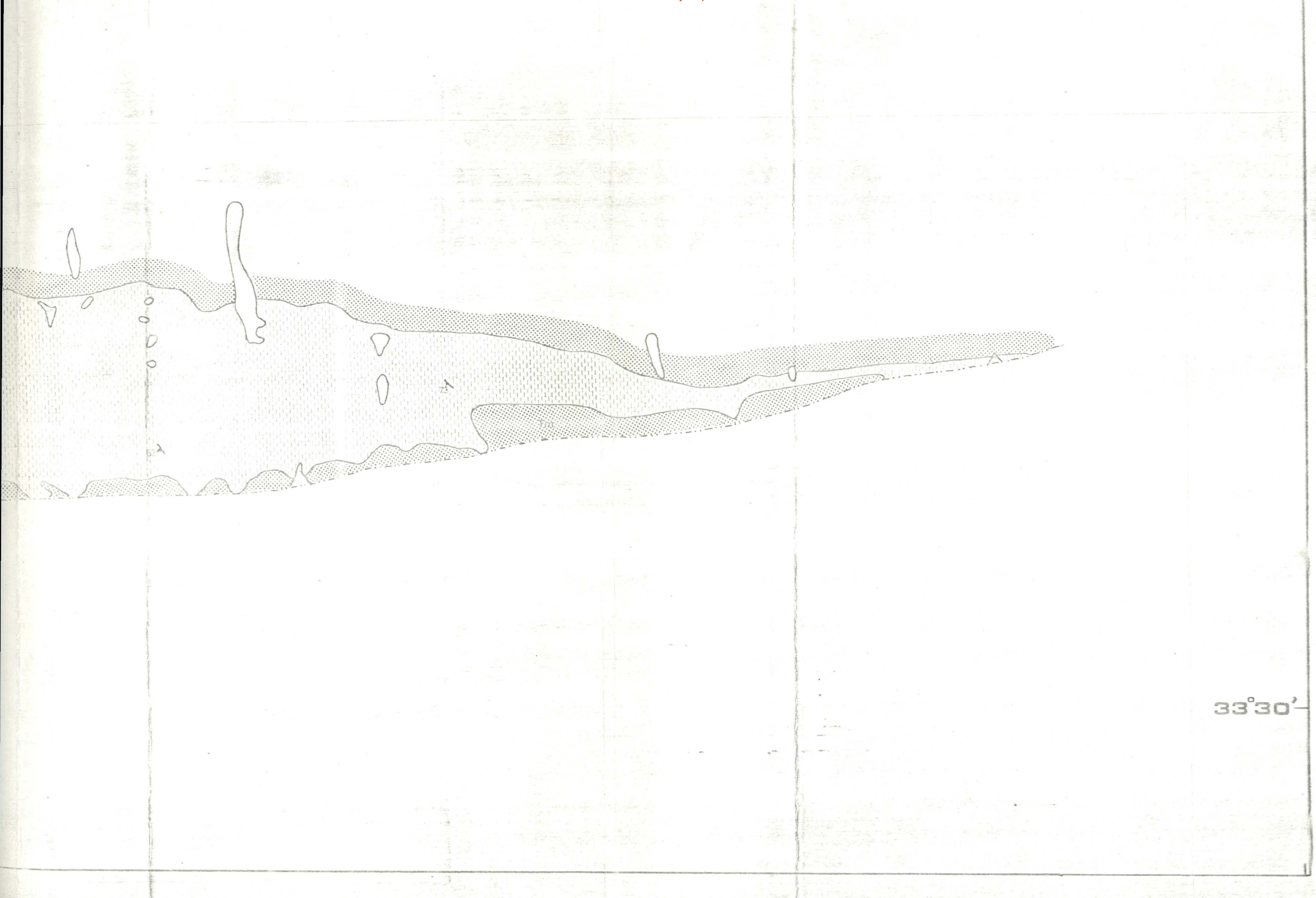



U1
U2
Gz
M1
M1
M2
M2
M2
D


Gz
H


Tm
Gz
U2
Vp
Gr2
Gr3
G
22°30'


Sc
ENON FORMATION
Sc
22°45'


33°30'