TR 82-32

THE STRATIGRAPHY AND STRUCTURE
OF THE
KOMMADAGGA SUBGROUP AND CONTIGUOUS ROCKS

by

ROGER SWART B.Sc.(Hons)

Thesis presented in fulfilment of the requirements
for the degree of Master of Science
in the
Department of Geology,
Rhodes University,Grahamstown.

January 1982

ABSTRACT

The Lake Mentz and Kommadagga Subgroups were deposited in a marine environment and are characterised by a heterogeneous sequence of sediments, which range in grain size from clays to grits. During the first phase of deposition the Kweekvlei Shale and Floriskraal Formations were deposited in a prograding shoreline environment, whereas the succeeding Waaipoort Shale Formation is interpreted as represnting a reworked shoreline. The final phase of deposition of the Cape Supergroup was a regressive one in which the Kommadagga Subgroup was formed. The coarsening upward cycle of this subgroup represents a deltaic deposit. A significant time gap appears to exist before the deposition of the glacial-marine Dwyka Tillite Formation.

Structurally, the area was subjected to deformation by buckle folding at about 250 Ma into a series of folds with southward dipping axial planes. Only one phase of deformation is recognised in the study area.

A decrease in pore space, mineral overgrowths,formation of silica and calcite cements and development of authigenic minerals such as opal, stilpnomelane, analcite, prehnite, muscovite and various clay minerals are the characteristic diagenetic features of the sediments.The mineralogical evidence suggests that the maximum temperature and pressure of burial was 150 C and 4 to 5 Kbar respectively.

CONTENTS

Chapter One : INTRODUCTION

1 1	Regional Setting	1
1.2	Regional Geological Setting	1
1.3	Previous work	3
1.4	Methods of Study	8
1.5	Acknowledgements	8

Chapter Two : STRATIGRAPHY AND SEDIMENTOLOGY

2.1	Introduction	10
2 2	Lithological Description	10
2.2.1	Kweekvlei Shale Formation	10
2.2 2	Floriskraal Formation	11
2.2.3	Waaipoort Shale Formation	12
2.2.4	Miller Diamictite Formation	20
2.2.5	Swartwaterspoort Sandstone Formation	26
2.2 6	Soutkloof Shale Formation	26
2.2.7	Dirkskraal Sandstone Formation	30
2.2.8	Dwyka Tillite Formation	34
2.3	Palaeoenvironments	36
2.3 1	Previous Interpretations	36
2.3.2	This Study	37
2.4	Provenance	43

Chapter Three : STRUCTURE

3.1	Introduction	45
3.2	Major Structures	46
3.2.1	Folds	46

3 2.2 Faults 48

3 3 Minor Structures 49

3.3.1 Tension Gashes 49

3.3.2 Axial Planar Cleavage 52

3.3 3 Lineations 62

3.3.4 Slickensides 62

3.3.5 Boudins 62

3.4 Microstructures 63

Chapter Four : DIAGENESIS

4.1 Introduction 67

4.2 Methods of Study 70

4.3 Cements 70

4.4 Authigenic Minerals 78

4.4.1 Feldspars 78

4.4.2 Prehnite 85

4.4.3 Stilpnomelane 88

4.4.4 Chlorite 90

4.4.5 White Mica (Muscovite) 90

4.4.6 Clay Minerals 91

4.4.7 Zeolites 93

4.5 Pressure and Temperature of Burial 95

Chapter Five : SYNTHESIS 100

REFERENCES: 104

FIGURES

1: Regional setting of the study area.

2: Topo-cadastral features of the study area.

3: Ridges consisting of Floriskraal Formation arenites.

4: Schistose rock fragment in a thin section of the Floriskraal Formation.

5: Geology of the Kommadagga area.

6: Sedimentary logs of the Waaipoort Formation.

7: Explanation of the symbols used in the logs.

8: Quartz arenite with ripple marks.

9: Slump structures in the Waaipoort Formation.

10: Rhythmites in the Waaipoort Formation.

11: Micro-bedding in a quartz arenite.

12: A euhedral zircon crystal with zoning.

13: Crawling trails in shale.

14: Features which resemble fossil fungae.

15: Possible fungal spores in the Waaipoort Formation.

16: Acritarchs in the Waaipoort Formation.

17: Numbers of different Acritarch species through time.

18: Exposure of the Miller Diamictite in a cliff face.

19: Sedimentary log of the Miller Diamictite.

20: Quartz overgrowths in the Swartwaterspoort Sandstone Formation.

21: Sedimentary log of the Soutkloof Shale Formation.

22: Field exposure of the Soutkloof Shale Formation.

23: Cone-in-cone structures in a carbonate band.

24: Sedimentary log of the Dirkskraal Sandstone Formation.

25: Mottling in the Dirkskraal Formation.

26: A view of the main Dwyka Tillite Formation.

27: Crude bedding near the base of the tillite.

28: A mixed tidal flat deposit overlain by storm deposits.

30: A north-south structural section through the study area.

31: Stereograms showing axial planar cleavage in the northern domain.

32: Brecciation in a fold hinge.

33: En echelon tension gashes in a quartz arenite.

34: Model explaining the origin of tension gashes.

35: Stereogram showing relationship of cleavage and tension gashes.

36: Stereograms of cleavage orientation from various parts of the study area.

37: Crenulation of bedding.

38: Examples of rough cleavage.

39: Classification of cleavage in psammitic rocks.

40: Diagrammatic sketches of rough cleavage fabrics.

41: Dwyka Tillite with rough cleavage.

42: Orientations of lineations in the study area.

43: Boudinage of carbonate units.

44: Pressure shadow formed behind a haematite grain.

45: Deformation of plagioclase feldspar.

46: Submicroscopic kinking of platy minerals.

47: Extreme deformation of quartz in the Dwyka Tillite Formation.

48: Quartz overgrowths.

49: Pressure solution features.

50: Scanning electron micrograph of opal.

51: Close up of ridges.

52: Ridge like forms on quartz overgrowths.

53: Quartz crystal with ridges.

54: Calcite cementing an arenite.

55: Relationship of pH and solubilities of quartz, calcite and amorphous silica.

56: Authigenic feldspar crystal.

57: Plagioclase overgrowths.

58: Stability relationships of feldspars at low temperatures.

59: The Kastner and Siever model for the origin of authigenic feldspars

60: Activity diagram for the system $Na_2O-Al_2O_3-SiO_2-H_2O$.

61: Prehnite in the Dirkskraal Formation.

62: Stilpnomelane and muscovite.

63: Halloysite showing its tubular structure.

64: Discrete grains of kaolinite.

65: Scanning electron micrograph of smectite.

66: Zeolites from the Ecca Group.

67: Zeolites from the Waaipoort Formation.

68: Temperature-pressure relationships of stilpnomelane and muscovite.

69: Schematic composite log of the Lake Mentz and Kommadagga Subgroups.

70: The Witteberg Basin.

CHAPTER ONE

INTRODUCTION

1.1. Regional Setting :

The study area is situated 60km west-north-west of Grahamstown (see fig.1). The topography is undulating with a maximum altitude of 737m (2433 feet) at Kommadaggakop (fig 2; see back folder), and a minimum altitude of 379m (1250 feet) in the valley of the Bushmans River. The Little Fish River flows east-west across the north of the area, while the Bushmans River flows in the same direction in the south (fig.2.). The drainage over most of the area, which receives 278mm of rain annually (Onesta and Verhoef, 1976), consists mainly of ephemeral streams.

The main community in the area is the hamlet of Kommadagga which is a railway siding on the main Port Elizabeth to De Aar route. Cuttings along this railway line and the National Road, both of which run north-south through the area, give fine exposures of the geology.

1.2. Regional geological setting :

The regional geological setting of the area is shown in fig.1. Structurally the area lies on the northern margin of the Cape Fold Belt, which is an east-west striking orogenic zone characterised by brittle deformation, very low grade metamorphism and a lack of igneous intrusions (Newton, 1973; Lock, 1980). The basin in which the shelf sediments of this belt were deposited started forming by the early

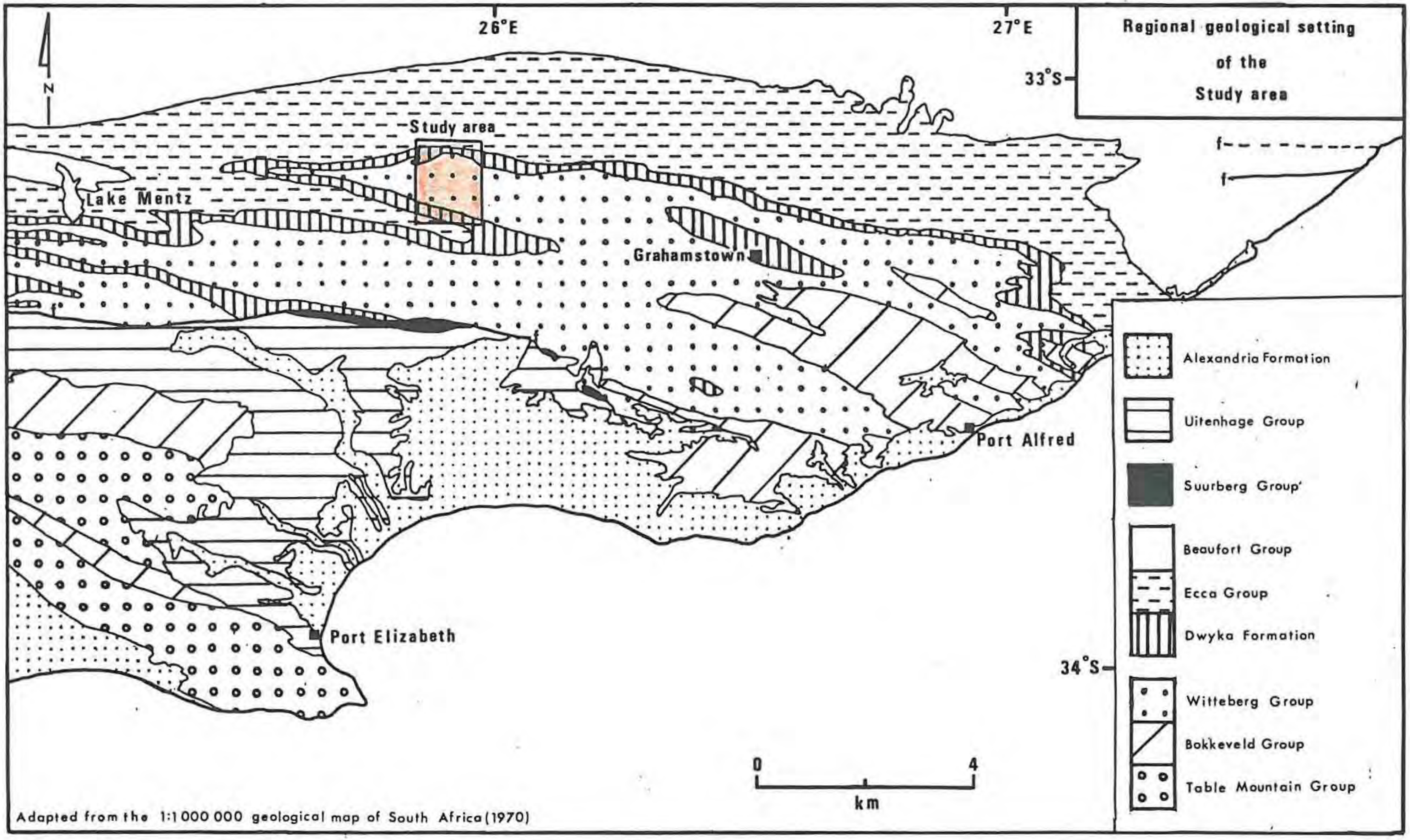

N
26°E
27°E
Regional geological setting
of the
Study area
33°S
Study area
Lake Mentz
Grahamstown
Port Alfred
Port Elizabeth
f
f
Alexandria Formation
Uitenhage Group
Suurberg Group
Beaufort Group
Ecca Group
Dwyka Formation
34°S
Witteberg Group
Bokkeveld Group
Table Mountain Group
0
4
km
Adapted from the 1:1 000 000 geological map of South Africa (1970)

Fig.1

Ordovician (Lock, 1973; 1978; Rust, 1973, p.274), and underwent many phases of transgression and regression before culminating in a deformational event which began some 277Ma (Halbich, 1981).

Formations found in the study area are from the Cape and Karoo Sequences. The stratigraphy of these sediments has recently been reviewed by Johnson (1976) and the bulk of this work has been incorporated into the report of the South African Committee for Stratigraphy (SACS, 1980). This stratigraphy is summarised in table 1 along with the work of Du Toit (1926) and Loock (1967).

The age of the upper Witteberg and Dwyka sediments is uncertain because of the lack of palaeontological evidence. Stapleton (1977), using palynological evidence, dated the Waaipoort Formation as being middle to late Devonian, but Gardiner (1969) has dated fish from the same formation as being early Carboniferous in age. Stapleton (1977) has dated the Dwyka as being early Permian. No age dates have been reported for the Kommadagga Subgroup, but it is probably early Carboniferous in age.

1.3. Previous work :

Prior to the 1970's little was known about the Cape Fold Belt, and it was not until very recently that detailed work was published on the belt (Johnson,1966;1976; Rust,1973; Newton,1973; Lock,1978;1980; De Swardt and Rowsell,1974). The first geological map to include the study area was published in 1928 at a scale of 1 inch : 3,75 English miles and was based on the work of Rogers, Schwarz and Haughton. In 1944 De Villiers published a review of what was known then about the Cape Orogeny, while in 1945 Edgar D. Mountain, in a presidential

DU TOIT (1926)

		Upper
	Ecca Series	Middle
		Lower
		Upper Shales
Karoo System	Dwyka Series	Boulder Beds
		Lower Shales
Cape System	Witteberg Series	

Table 1-a

LOOCK(1967)

	Ecca Group		
		Dwyka Tillite Formation	
Karoo Supergroup	Dwyka Group		
			Dirkskraal Siltstone Member
		Kommadagga Formation	
			Dirkskraal Shale Member
			Swartwaterspoort Sandstone Member
Cape Supergroup	Witteberg Group	Lake Mentz Formation	
			Bergplaas Shale

Table 1-b

JOHNSON (1976); SACS (1980)

	Ecca Group		
Karoo Sequence			Dwyka Tillite Formation
			Dirkskraal Sandstone Formation
		Kommadagga Subgroup	Soutkloof Shale Formation
			Swartwaterspoort Sandstone Formation
Cape Supergroup	Witteberg Group		Miller Diamictite Formation
		Lake Mentz Subgroup	Waaipoort Shale Formation
			Floriskraal Formation
			Kweekvlei Shale Formation

Table 1-c

address to the Geological Society of South Africa, described the formations of the Cape System as being "relatively uninteresting" (Mountain,1945, pxxii).

Meyer (1965) described the geology of an area to the south-east of the current study area, while Loock (1967), in a much broader study of the Witteberg-Dwyka contact included the south-eastern portion of the present study area in his thesis. Johnson (1966;1976) attempted to rationalise the stratigraphy of the Cape and Karoo Sequences of the Eastern Cape, while Marais (1963) was the first to subdivide the Upper Witteberg sediments.

Theron (1962-a) published an analysis of the Cape folding in the Willowmore area, but it was not until a paper by Newton (1973) appeared proposing a gravity folding origin for the belt that any detailed work was done. This paper was followed by the work of De Swardt and others who examined the origin of foliation in the belt, and also the relationship between diagenesis and deformation (De Swardt and Rowsell,1974; De Swardt, Fletcher and Toschek, 1974). The onset of the National Geodynamics Project in 1974 prompted the most detailed study of the belt yet undertaken. Prof. I.W. Halbich and students of his at the University of Stellenbosch undertook a study of the belt between longitudes 22 E and 23 E. The main summary of this work is still to be published (Sohnge and Halbich, in press), but important contributions have already appeared (Halbich, 1977; 1981). Plate tectonic models for the origin of the belt have also been proposed by a number of workers (Newton,1974;1980; De Beer et al., 1974; Lock,1980), while Rust (1973) and Lock(1978) have examined the tectonic evolution of the basin. More recently Hiller and Snowden (1981) have interpreted the structural history of an area of the Baviaanskloof which lies west of the current study area.

1.4. Methods of study :

A total of two and a half months were spent in the field mapping the area in detail. As no detailed geological maps of the area have been published mapping was done on a scale of 1:12 500 using enlargements of the 1:50 000 topographic map in conjunction with aerial photographs. In addition, detailed sections of all sedimentary units were studied and logs of these units were drawn up.

Samples were collected from all units and examined using the conventional petrographic microscope and the scanning electron microscope. The electron microprobe was used for chemical analysis of plagioclase and prehnite grains. Modal and grain size analyses were not done on the samples because of the alteration of the mineralogy and grain size by diagenetic processes. Visual estimation of the original grain size, shape and sorting of the sediments was done with the aid of charts published in Griffiths (1967).

The structural data was recorded on the map, and then plotted on a Lambert equal area projection. This data was contoured on the basis of number of per cent per one per cent (%/1%) of the area whenever it was felt necessary to do so.

1.5 Acknowledgements :

This thesis would not have been possible without the help of numerous people. Firstly Dr. Norton Hiller is thanked for his invaluable advice and constant encouragement and stimulation. Dr. Phil Snowden provided constructive criticisms on the structural portion of this study. Prof. Hugh Eales was responsible for organising the finances behind this study and for this I am most grateful. The financial support of the C.S.I.R. was invaluable.

Drs. Ivan Reynolds and Roger Jacob were pestered many times for help in mineral identification and were always willing to assist. Robin Cross of the Rhodes University Electron Microscope Unit took the scanning electron micrographs and Dennis Gouws helped with the microprobe analyses.

Mike Kokkinn and Llewellyn Parker provided invaluable assistance and constant entertainment. Caroline Halpin and Tammy Griffin typed rough drafts of this thesis.

A special thank you to Goonie and Janet Marsh and Daz Dunn for helping me get through these last two years.

CHAPTER TWO

STRATIGRAPHY AND SEDIMENTOLOGY

2.1. Introduction :

A total of about 1450m of sediments is found in the study area, and the sediments range from shallow marine sandstones and shales (approximately 600m) to extensive diamictites of glacial origin (850m). These sediments have not previously been documented with only Loock (1967) and Johnson (1966;1976) having written any accounts of the units in this area. However neither used detailed sedimentary logs of the units in their interpretation, nor was there any attempt to model the facies. The object of this part of the study was therefore to attempt to fill the gap in our knowledge.

2.2. Lithological Description :

2.2.1. Kweekvlei Shale Formation :

In the study area this formation is only exposed in one small outcrop in the core of an anticline at Wittepoort. Exposure is poor as the shales are highly weathered and as a result little could be seen in the way of sedimentary structures. The shales appear dark coloured as reported by Loock (1967) and horizontal lamination was the dominant sedimentary structure observed. The full thickness of the shale is not exposed in the study area, but Loock (1967) reported a thickness of 80m for this unit.

2.2.2. Floriskraal Formation :

Two main outcrops of this formation are found in the study area. These form linear topographic highs in the central and southern parts of the area (see fig.3). This topographic expression of this unit is distinctive, as is its clean white-weathering nature. Petrographically it is a medium to coarse grained quartz arenite, cemented by silica and with less than 5% matrix. Grain contacts are often sutured and overgrowths are common (see chapter 4). Heavy minerals are rare with a few isolated grains of tourmaline, sphene, zircon and anatase(?) being the only species observed in thin section. Rock fragments are rarely seen and those that are found are of schistose rocks (see fig.4). Some grey weathering shales are interbedded with the Floriskraal sandstones, but these are very thin (up to 0.5cm thick) and are highly weathered.

The sandstone units are for the most part massively bedded although some low angled cross stratification was found. It is possible that the massive beds originally contained more sedimentary structures but have since been homogenised by bioturbation.

The total thickness of the Floriskraal Formation in the study area is about 30m and, as SACS (1980) has demonstrated, thins toward the east.

No fossils were found in this unit.

2.2.3. Waaipoort Shale Formation :

This unit has the most extensive outcrop in the study area and its distribution is shown in fig.5 (see back folder). It is a heterogeneous sequence with fine grained shales, arkoses, minor quartz

Fig.3: Linear ridges formed by resistant quartz arenites of the Floriskraal Formation.

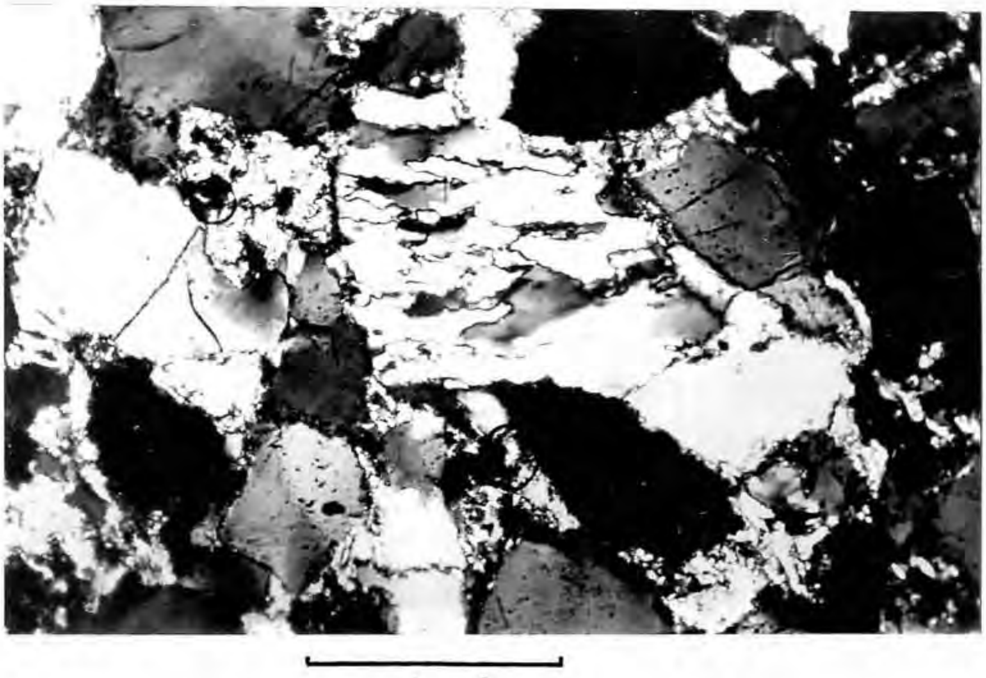


Fig.4: Photomicrograph showing a schistose rock fragment in a Floriskraal Formation quartz arenite.

(10*; crossed nicols; sample K-95)

arenites and local pebble gravels. Detailed sedimentary logs of various parts of this sequence are shown in fig.6 (see back folder). It is characterised in the field by a greenish appearance, the presence of rhythmites and a highly sericitic nature.

The medium grained units of the Waaipoort have dominantly flat to gently inclined bedding. Ripple marks are found, and these may occasionally have trace fossils on the top surface. The ripples are both assymmetrical and symmetrical in form, with assymetrical ripples which have wavelengths up to 10cm, and amplitudes of 2cm being more common. Interference ripples are also found in the study area. Occasional clean quartz arenite lenses are dominated by a rippling with the ripples being defined by heavy mineral bands (see fig.8).

Fig.8: A coarse grained quartz arenite, possibly representing storm deposits, from the Waaipoort Formation.

Slumping is found in the fine to medium grained sediments and is easily recognised by its convolute form (see fig.9). These structures indicate a certain degree of slope instability during deposition of the Waaipoort sediments, and, according to Reineck and Singh (1973), they are associated with rapid sedimentation. Rhythmites are a common sedimentary structure in the study area (fig 10), comprising alternating coarse and fine grained material with different compositions and colours associated with each.

Nodules are found fairly frequently in the finer to medium grained sediments. They are generally calcareous in nature, and may be elongated parallel to bedding. Elsewhere fish fossils have been found in nodules from the Waaipoort Formation (Gardiner,1969; Marais,1963; Theron,1962-b), but no fish remains were found in nodules from the study area. These nodules may be rimmed by pyrite, the precipitation of which is favoured by the reducing conditions which might have existed at the time causing the demise of the fish (Theron, 1962-b).

The dominant bedding forms in the arenites are parallel, horizontal and gently inclined laminations, while small scale trough cross beds are also present. The units with inclined laminations may be topped by a scour surface, or, more commonly, by horizontal lamination.

Coarser grained lenses with well rounded pebbles up to two centimeters in diameter occur very rarely. They are found in scour channels and probably represent a gravel lag. These lenses are not very big, always being less than 10cm thick, 10cm to 20cm wide, and cannot be traced across the width of a road cutting.

The coarse grained arenites of this formation are characterised by a nearly monomineralic composition, with 98% quartz and scattered heavy minerals. Rock fragments and feldspar grains are absent, while there is only an occasional mica flake. Triple point and sutured grain

Fig.9: Slump structures from a medium grained unit of the Waaipoort Formation

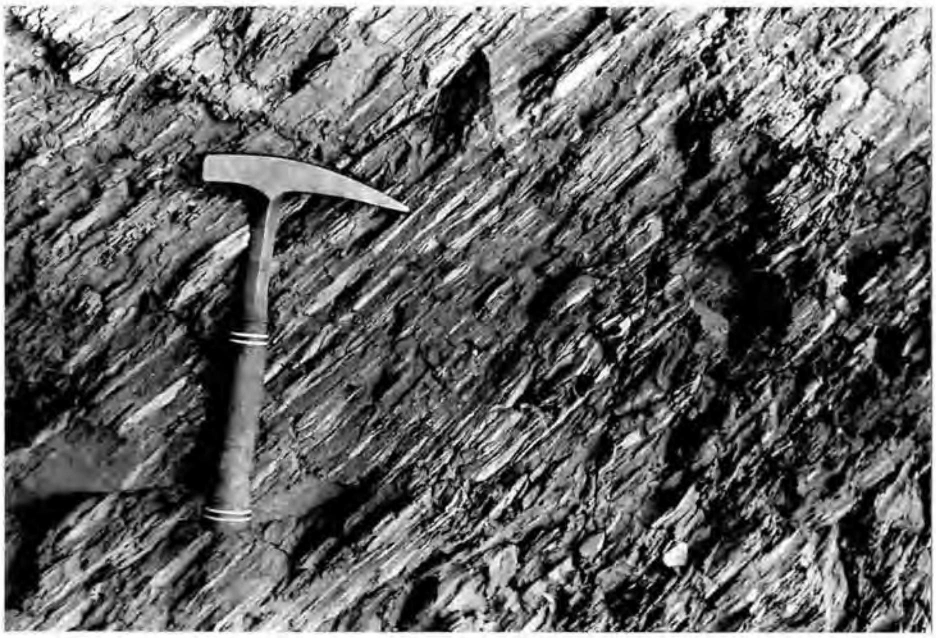


Fig.10: Tidal rhythmites from the Waaipoort Formation.

boundaries are common and are possibly caused by the interlocking of grain overgrowths.

The medium grained rocks have a more heterogeneous composition with an important proportion of feldspar and matrix. In general grains have a roundness between 0.3 and 0.5 (estimated visually from charts in Griffiths, 1967) and are moderately sorted. However, locally, sorting may be good, and fig.11 shows an arenite bedded on a microscale with good sorting within each individual layer.

Quartz is the dominant mineral in these sediments and is generally colourless with no alteration features. Many of the quartz grains poikolitically enclose acicular crystals of rutile which indicates a granitic source for these sediments (Folk, 1965; Blatt et al., 1980, p292).

Feldspar grains are common in some of the arenites, and are characterised by polysynthetic twinning and varying degrees of alteration which gives the grains a brown coating in plane polarised light. This brown alteration helps in the identification of untwinned feldspar grains. It is possible that much of the original feldspar content has been completely altered to clay minerals and that the matrix which we see today is secondary and not primary (Galloway, 1974; Wilson and Pittman, 1977; Brenchley, 1969).

Rock fragments representing a wide range of lithologies are common in these sediments. Many fragments are difficult to identify and the only ones identified with certainty were metamorphic quartz grains, cherts, quartzites and shale fragments. The shale fragments have now been compacted so that they now have very irregular outlines,

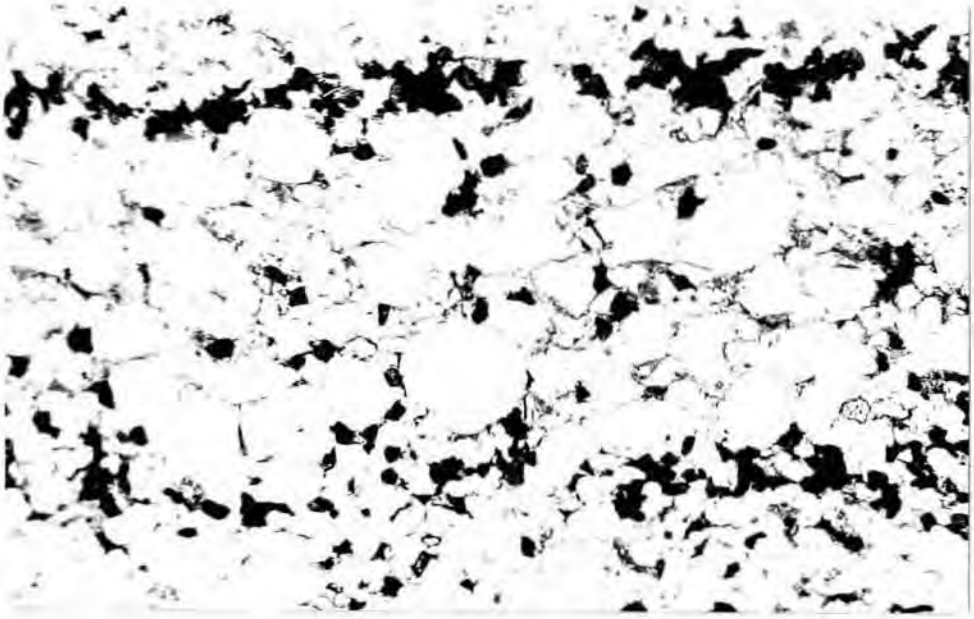


Fig.11(a)

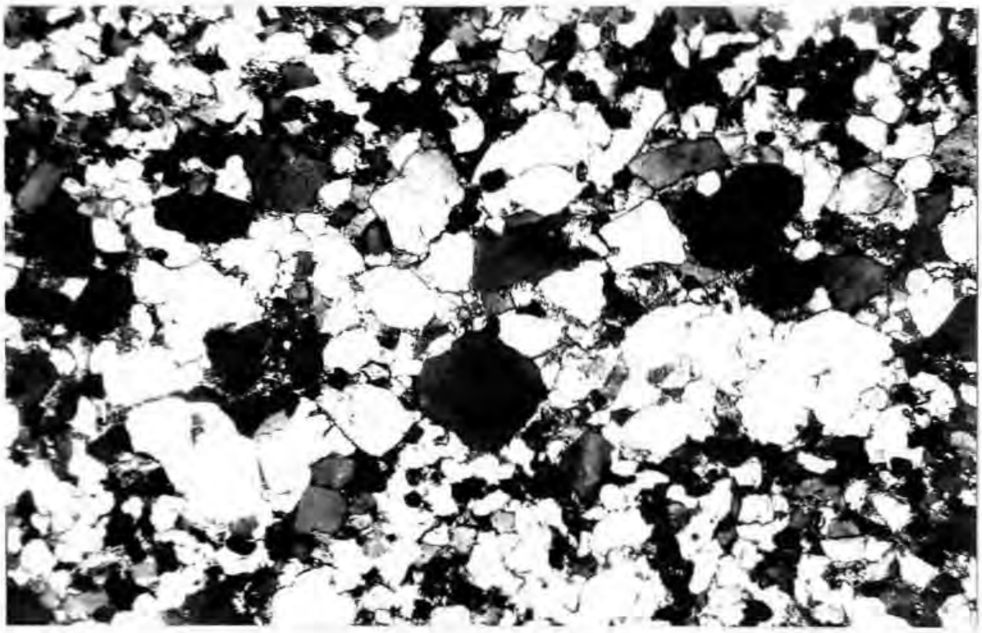


Fig.11(b)

Fig.11(a)&(b): Photomicrographs of a thin section showing well developed bedding and sorting on a microscale.
(2.5*; a-plane polarised light; b-crossed nicols; sample K-72)

conformable with the edges of the surrounding grains.

Tourmaline, zircon and rutile are the most abundant heavy minerals, being found in varying proportions in most units. Of the tourmaline group both schorl and dravite are found, with schorl being the more abundant. Schorl indicates either a metamorphic or igneous source, while dravite is indicative of a metamorphic source area (Blatt et al., 1980, p.314). These grains have very irregular to well rounded outlines and are larger than any other member of the heavy mineral suite.

Zircon grains are colourless and well rounded, although some grains may show their original euhedral crystal form (fig.12). Euhedral zircon grains suggest an igneous source as zircons from metamorphic and sedimentary rocks are generally well rounded (Blatt et al., 1980 p.313). Rutile grains are very common, being well rounded and ranging in size from a few microns to 0.6mm in diameter. The abundance of rutile suggests a metamorphic source area (Force, 1976; 1980).

In addition to the above heavy minerals, garnet is found in the Waaipoort Formation.The grains often appear platy, possibly as a result of a parting parallel to the (110) crystal face, indicating that the grains are almandine (Heinrich, 1965, p.61). Apatite also occurs as a trace component in most units although Loock (1967) reported that there was no apatite in these sediments. Sphene is found in most samples in trace amounts, and opaque oxide minerals are common.

Plant specimens are found, but are too fragmentary to identify postively. Previously the following genera of plants have been

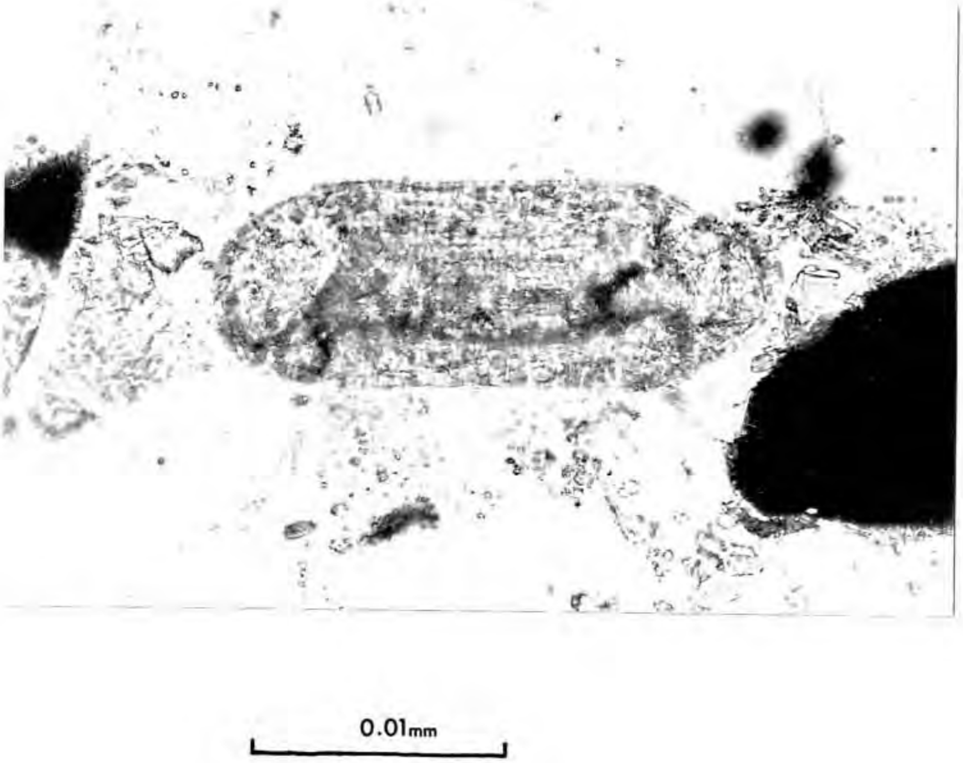


Fig.12: A zoned euhedral zircon crystal from the Waaipoort Formation. (60*; plane polarised light; sample K-45)

Fig.13: Shale bed with a number of crawling trails.

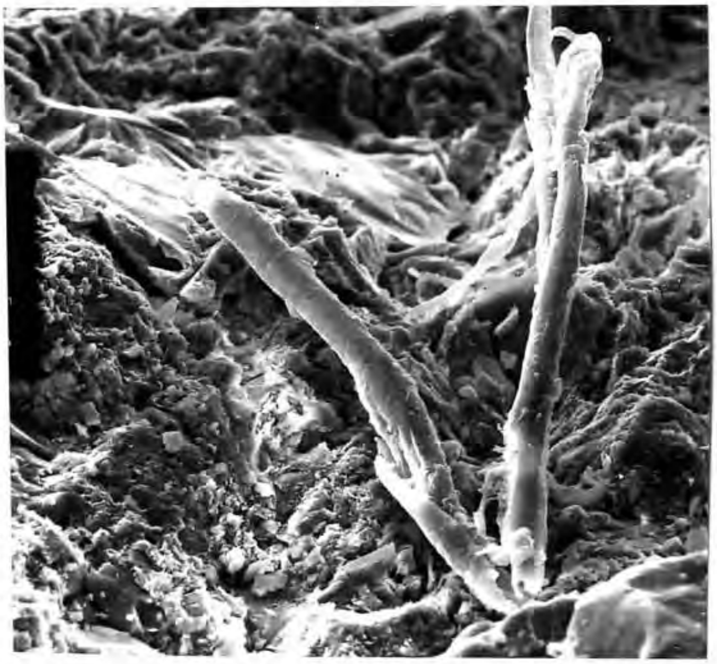
reported from the Witteberg Group : Dutoitia, Protolepidodendron, Haplostigma, Archaeosigillaria, Leptophloem and Platyphytum (Haughton, 1969; Plumstead, 1967). No invertebrate fossils have been found, nor were any fish fossils discovered although they have been reported from elsewhere in the Waaipoort Formation (Marais, 1963; Theron, 1962-b). Some evidence for biological activity is found, mainly in the form of crawling trails (fig.13), but is not common.

During the course of routine SEM examination of these rocks a number of unusual features were observed which have been interpreted as being microfossils. The first type resembles the fossil fungae described by Hallbauer (1975) from the Precambrian Witwaterstrand Supergroup (see fig.14) and the second, possibly related feature shown in fig.15, has been interpreted as fossil fungal spores.

A third problematical feature is shown in fig.16. These features have been allocated to the group Acritarchs, an informal group to which organic walled unicelluar creatures are assigned until their true affinities are known. These forms were very common in the Devonian (see fig.17).

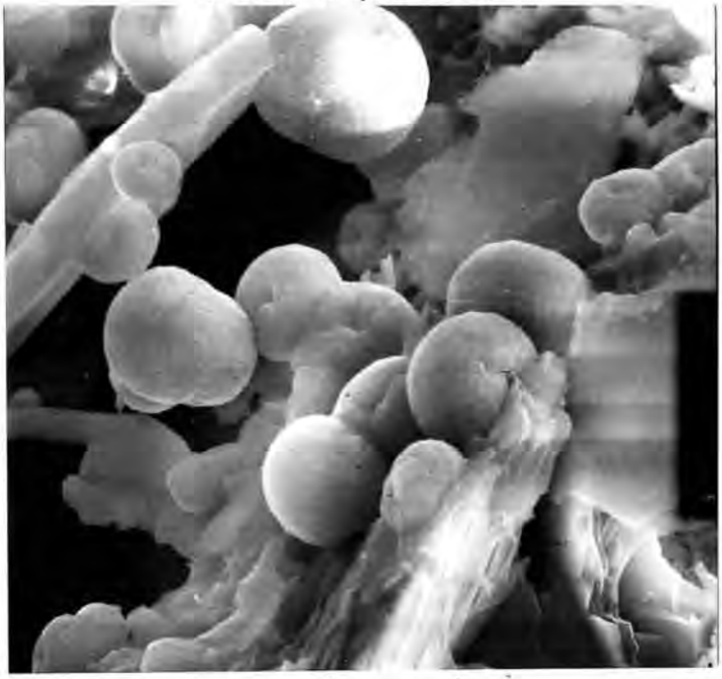
2.2.4 Miller Diamictite Formation:

The Miller Diamictite outcrops discontinuously along strike, being at its thickest of 4m in a cliff face adjacent to a stream north of Saltaire station (fig.18). A sedimentary log of this formation is shown in fig.20. Johnson (1976) reported that the diamictite had a maximum thickness of 6m in the same general area. At Bergplaas the diamictite can be followed along strike and can be seen to be pinching out. It thins from being 0.5m thick to nothing over a distance of 5m



40μ

Fig 14: A problematical feature which is similar in form to fossil fungae described from the Witwatersrand Group by Hallbauer (1975) (340*; sample K-86)



4μ

Fig.15: A feature found in close assocciation with those in fig.14 and are possibly fossil fungal spores.

(7000*; samole K-14)

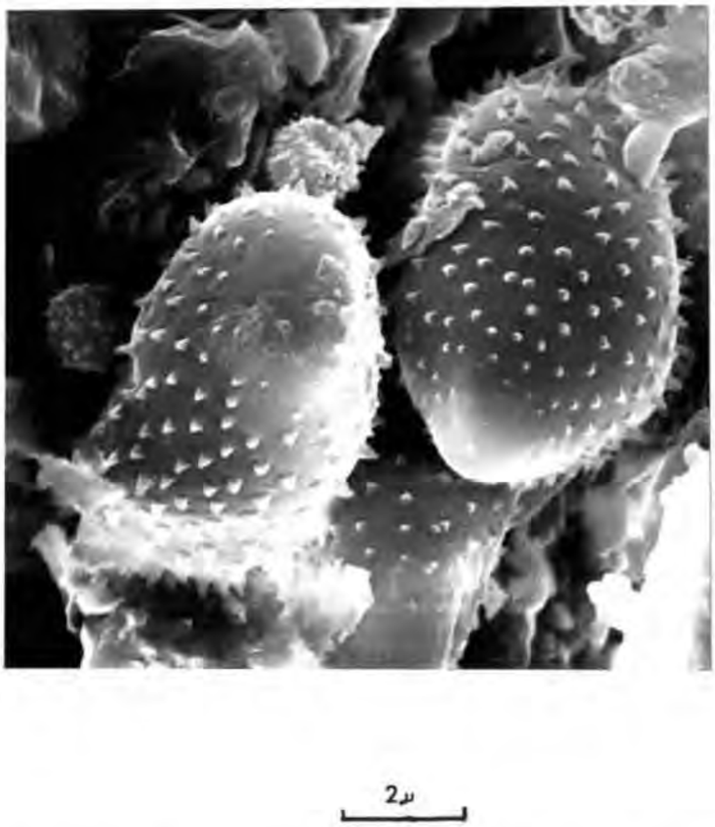


Fig.16: An unidentified feature observed under the SEM. These features have tentatively been assigned to the group Acritarchs. (6000*; sample K-13)

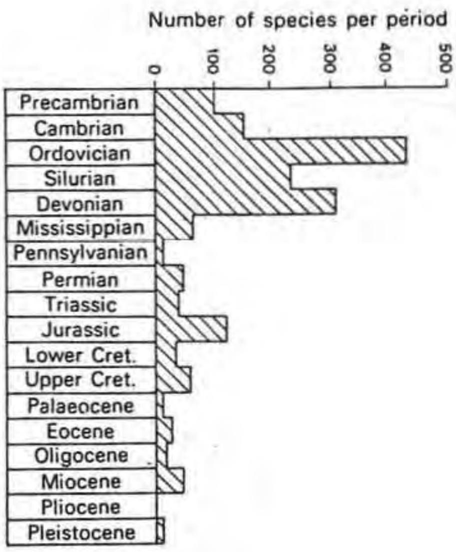


Fig.17: Distibution of Acritarchs through time. (after Brasier, 1980, fig.5.2, p.33)

Fig.18: Exposure of the Miller Diamictite in a cliff face near Saltaire station.

to 6m,but no evidence was found for inter-tonguing with the overlying Swartwaterspoort Sandstone which Johnson (1976) described.

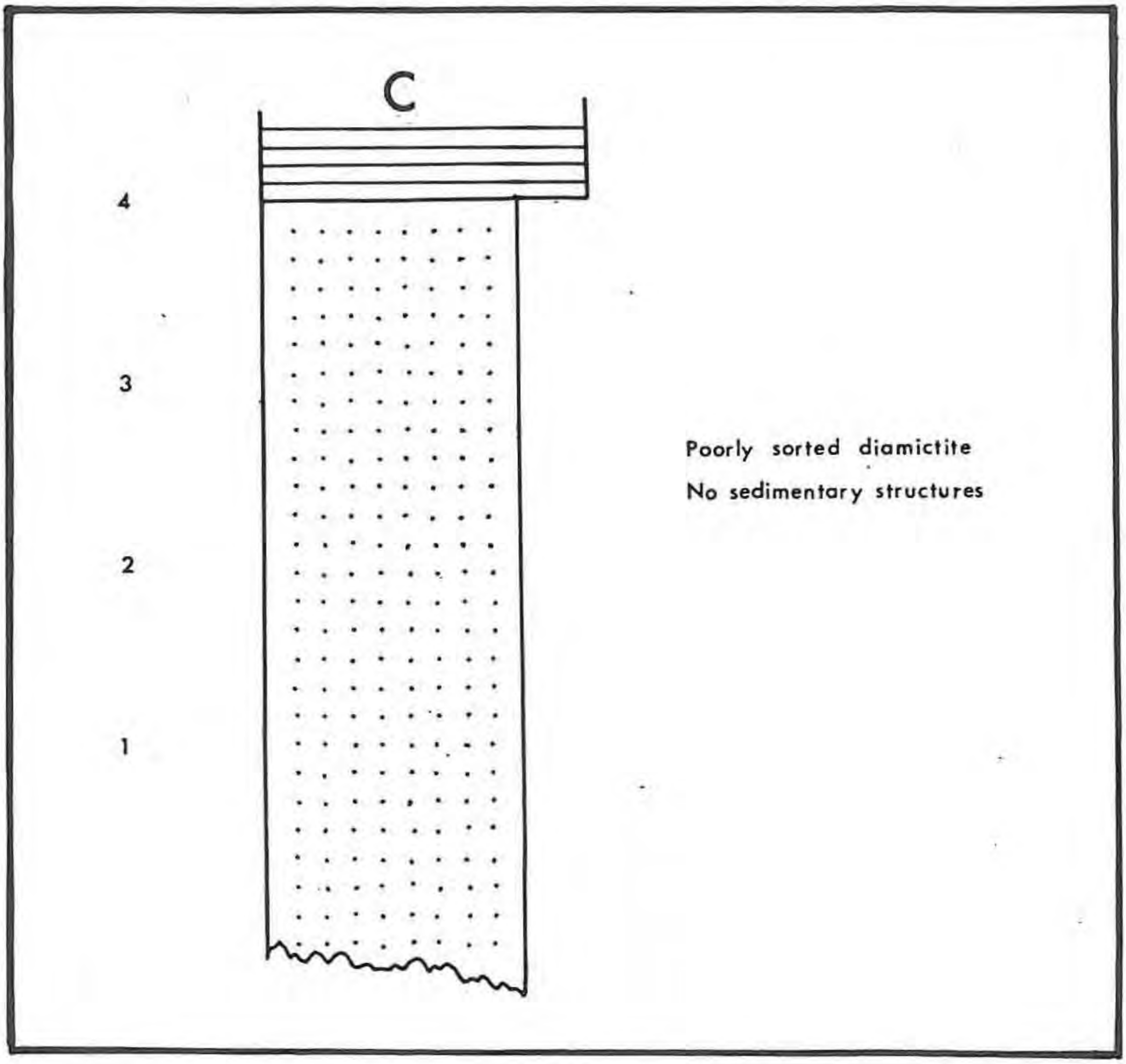


Fig.19: A detailed log of the outcrop of the Miller Diamictite near Saltaire station. See fig.5 for locality of the section and fig.7 for an explanation of the symbols.

In hand specimens the diamictite is fine to medium grained with occasional pebbles up to 2cm in long dimension. Fresh specimens are dark blue but it weathers to a dull brown colour.

No sedimentary structures were found despite careful examination of fresh and weathered outcrops. It is possible that the dark colour of the rock obscures any sedimentary structures that are present, and X-ray photography may reveal the presence of such structures.

Thin sections show that this formation is characterised by a wide range of grain sizes. Individual grains are not in contact with each other, but are rather "floating" in the matrix. Quartz grains are generally free of inclusions, but some may enclose rutile whereas others enclose zircon crystals, both types indicating a granitic source.

Feldspar grains are rare and are generally altered. Similarly rock fragments are uncommon and rutile was the most abundant heavy mineral observed, whereas only isolated grains of zircon and tourmaline are found.

The sorting of the diamictite is very poor, the maximum grain size being 2cm and the minimum less than 0.1mm, with a continuous gradation in size between these two extremes. Sphericity and rounding of grains is high, and this suggests a certain degree of textural maturity, although the poor sorting contradicts this.

No evidence of bioturbation was found in the diamictite, nor were there any body fossils apparent except for a single plant fragment, possibly of Archaeosigillaria, found at Kommadagakop.

2.2.5 Swartwaterspoort Sandstone Formation :

Outcrops of this formation are discontinuous in the study area where it has a maximum observed thickness of 5m. In the field it is easily recognised by its coarse grain size, clean white weathering nature, presence of iron oxide staining and slump structures. It is resistant to weathering and often forms the capping for koppies.

A 0.1cm to 3cm thick lamination, defined by iron oxide staining, is common in the Swartwaterspoort. Bell (1981) feels that these are laterally discontinuous and can rarely be traced for greater than 0.5 meters. Soft sediment deformation structures are the only sedimentary structures commonly observed.

Quartz is the most common type of clast found in this formation with other grains rarely comprising more than 2% of the rock. Overgrowths are common on quartz grains, and are easily recognised as the original grain shape is outlined by a clay layer (see fig.20). The clasts found in this unit are of the same composition as those in the Miller Diamictite

No fossils have been found in this unit.

2.2.6 Soutkloof Shale Formation :

The light coloured shales of the Soutkloof Shale Formation are poorly exposed throughout the study area because of their susceptibility to weathering and erosion. In the field this unit generally forms valleys, and the best exposures are in road and railway cuttings. The maximum measured thickness of this unit in the study area is 127m, consisting essentially of finely laminated shales. A sedimentary log of this formation is shown in fig.21 (see back folder).

0.25mm

Fig.20: Photomicrographs of quartz overgrowths in the Swartwaterspoort Sandstone Formation.
(10*; crossed nicols; sample K-17)

Fig.22: Exposure of the Soutkloof Shale in a roadcutting.

Very thin horizontal lamination is the only sedimentary structure seen in the shale often being less than 1mm in width (see fig.22). It is possible that other sedimentary structures such as bioturbation do exist, but have been obscured by weathering.

Fig.23: Cone-in-cone structures in a carbonate band in the Soutkloof Shale Formation.

Intermittent carbonate layers with a maximum thickness of 6cm occur interbedded with the shale. These carbonate layers consist dominantly of calcite with some minor clays. In hand specimen the characteristic feature of these carbonate bands is their brown-grey colour and the presence of cone-in-cone structures (fig.23), the surface of which is covered by a clay layer. The origin of these structures is unknown, but is probably related to pressure. The c-axes of the calcite grains have grown or recrystallised perpendicular to bedding and sections cut parallel to this plane show a texture comparable to a dove's feather.

The shale is very fine grained and because of this little information was gathered from thin section work. Quartz grains form less than 5% of the specimens examined, and other minerals present are either clays or micas, none of which could be identified with certainty. The micaceous minerals often show a preferred orientation parallel to bedding.

No fossils have been found in this unit.

2.2.7 Dirkskraal Sandstone Formation :

The Dirkskraal Sandstone Formation, the uppermost unit of the Witteberg Group, is present everywhere in the study area and is never more than 100m thick. It is characterised by a fine grained nature, the presence of mottling and a ripple lamination. Sedimentary logs of this formation are shown in fig.24 (see back folder). The basal contact with the Soutkloof Shale is poorly exposed because of the highly weathered nature of the shale, but sections from outside the study area show a gradational contact. The upper contact with the Dwyka Formation is sharp, and shows no sign of glacial deformation.

The ripple lamination is evident throughout the sequence. These layers are defined by layers of dark organic(?) material, but are laterally discontinuous, nor does there appear to be a preferred orientation to these structures. Horizontal lamination is also found, being only 1mm-2mm thick. Primary current lineations are associated with these horizontal laminations.

The sandstones may occasionally appear massive but when they are followed along strike units which initially appear massive actually have a well developed ripple or horizontal lamination.

A distinguishing feature of the formation is the widespread mottling found within it (fig.25). Mottling in sedimentary rocks is often caused by bioturbation (Pettijohn, 1975), but this is unlikely in this case as:

(a) the mottles are too fine, rarely being more than 0.5cm in diameter and

(b) the mottling does not disturb the lamination at all (fig.25).

Some mottles may be flattened parallel to the lamination (fig.25). The fact that the lamination is not disturbed by the mottling would suggest a syn-depositional origin for the mottles, but no textural differences can be seen between mottles and host rock. However diagenetic prehnite is closely associated with these mottles, and their origin is therefore probably a compositional one (see chapter four).

Scour surfaces asociated with channelling are common and these may be found with mud flake conglomerates. These channels may be quite large, cutting down 1m-1,5m and may be 3m-4m wide. Trough cross bedding is occasionally present, and load casting is found where sands overlie finer grained material. Large concretions with a maximum

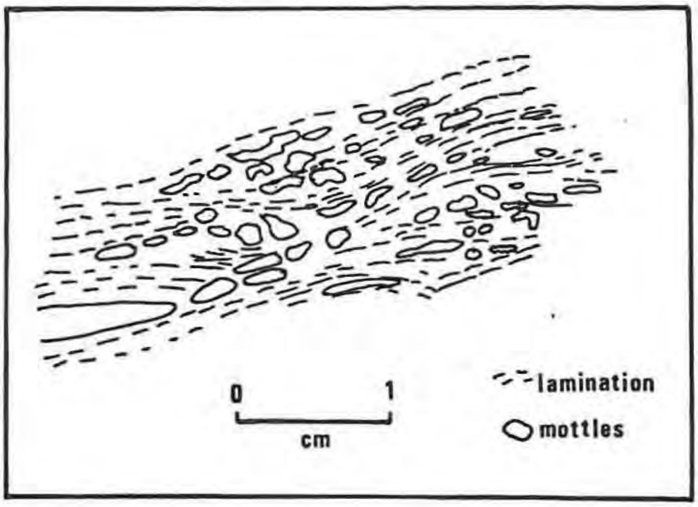


Fig.25: Mottling in the Dirkskraal Formation.

Drawn from a handspecimen

diameter of 20cm are found but are not common. No fossil material was found within these concretions.

In places the silts show numerous thin horizontal feeding trails on their top surfaces, but no vertical burrowing was observed. The sediments of this formation are fine grained with an average grain size of 5 to 6μ. Quartz and plagioclase are the most common minerals but it is probable that most of the plagioclase is diagenetic in origin, being nearly pure albite in composition (see chapter four). The quartz grains are occasionally well rounded and it is likely that this represents the original nature of most of the grains but which have now been altered by diagenetic processes such as the development of quartz overgrowths.

Although most of the primary feldspar content has now been destroyed there is still a significant proportion of this mineral left. These grains are distinguished from the authigenic ones by the brownish nature of the alteration and decomposition along cleavage planes. The clay alteration products of these feldspars are difficult to distinguish from detrital ones.

Tourmaline does not appear to be as common in these rocks as it is in the Waaipoort Formation, but rounded zircon, rutile and garnet are abundant. Apatite, sphene and anatase are present in trace amounts but there are other numerous very small, unidentified, heavy mineral grains which are distributed throughout the rock and give it a dusty appearance.

Unidentified plant fragments have also been found in this formation.

2.2.8 Dwyka Tillite Formation :

This unit, which is unmistakeable in the field, consists mainly of a dark blue diamictite which weathers to a dull brown. A minimum calculated thickness of 850 metres for this unit was measured in the south of the study area. This is substantially thicker than the 680m given by Johnson (1966; 1976) for this unit.

Sedimentary structures within the main diamictite are absent (fig.26), but shales and sandstones inter-bedded with the diamictite may show some structures. The shales are very finely laminated and dropstones may be present, but are not common. At the base of the diamictite in the north a crude bedding is present (fig.27). This banding, which is much coarser than that found in varved sediments, is defined by alternating layers of mud and arenaceous to rudaceous material. The sandy layers coarsen upwards, while the shaly ones fine upwards and may have a few large pebbles at the base.

The diamictite is very poorly sorted with a range in grain size from less than 1mm to blocks of granite 1,2m in diameter. Clasts, which may be angular to well rounded, show a wide range in composition with granites, garnet-hornblende gneisses, sandstones, quartzites, cherts, conglomerates, green amygdaloidal lavas, milky quartz and epidote-clinozoisite bearing gneisses being observed. There is also a substantial proportion of matrix in which clasts generally "float". The sandstones are finer grained than the main diamictite and are better sorted, but a wide range of lithologies are still found.

No fossils have been recovered from this formation but Stapleton (1977) has described Permian micro-fossils from the tillite.

Fig.26: View of the main diamictite showing its poor sorting and the lack of sedimentary structures

Fig.27: Hand specimen of crude bedding near the base of the diamictite.

2.3 Palaeoenvironments :

Deposition of the Cape Supergroup sediments occurred in a narrow east-west trending marine basin with the deepest part of this basin being just north of the present southern coastline of South Africa (Rust, 1973). The source area for all these sediments appears to have been to the north of the basin (Rust, 1973; Lock, 1974).

2.3.1 Previous interpretations :

Johnson (1976) recognised that the Upper Witteberg sediments were deposited in a shallow marine environment. He interpreted the Kweekvlei Shale and Floriskraal Formations as representing the offshore muds and lower shoreface sands of a regressive sequence (p.197). The overlying Waaipoort Formation was interpreted by Johnson as having formed on the upper pro-delta slope and the outer delta front platform. Loock (1967) suggested that the Kweekvlei Shale formed in a deep marine basin, but he proposed no specific environment for the Foriskraal and Waaipoort Formations.

The Miller Diamictite has been interpreted as a glacial deposit by Rossouw (1953), who called it the Basal Tillite, and Johnson (1976). Bell (1981) regards this formation as being a debris flow deposit. Johnson (1976) has interpreted the Swartwaterspoort Formation as being a beach deposit which represents either the reworking of glacial outwash or of the Miller Diamictite. The Soutkloof Shale and Dirkskraal Formations have been interpreted by both Johnson (1976) and Loock(1967) as being pro-glacial sediments deposited in a marine environment.

A glacial origin for the Dwyka Formation is now widely accepted (Du Toit, 1921; Stratten, 1968; Theron and Blignault, 1975; Johnson,1976) but differences of opinion do exist as to whether the glacier was terrestrial or marine. Stratten(1968) has interpreted both the northern and southern facies of this formation as being terrestrial, but other workers have suggested that the southern facies is marine (Du Toit,1921; Theron and Blignault,1975; Dunlevey and Hiller,1979). An attempt will be made to resolve this problem.

2.3.2 This Study :

The limited outcrop of the Kweekvlei Shale Formation in the study area means that any interpretation regarding its origin must be largely speculative, and is in this case based on the descriptions of Johnson (1976). According to Johnson this shale is generally massive (p.190), a feature found in pro-delta or off-shore muds with dominantly suspension sedimentation (Wright, 1978). The overlying quartz arenites of the Floriskraal Formation are massively bedded with occasional low angled, off-shore directed cross-stratification, and thin shales. This type of deposit can form in a number of environments, notably the distributary mouth bar of a delta (Wright, 1978), or the lower shoreface of a beach/barrier island system (Davis, 1978; Elliott, 1978). The low angled cross-stratification suggests that the Floriskraal Formation represents the lower shoreface zone of a prograding coast line, whereas the Kweekvlei Shale represents off-shore muds deposited in quieter conditions.

The lack of body fossils in the Lake Mentz and Kommadagga Subgroups is problematical and places constraints on the reliability of the interpretations made. A possible explanation for their absence is that these deposits may have formed in a narrow basin with restricted circulation and as a result salinities were too high for the development of abundant life (Clarkson, 1979).

The heterogeneous Waaipoort Formation sediments have a more complex origin, with the rhythmites indicating that tidal processes have played an important role in the origin of this deposit. Rhythmites are found in all environments where sedimentation is irregular such as the seasonal control of glacial varves and in deep sea muds. However laminations similar to those found in the Waaipoort Formation have been recorded by Reineck (1972) and are attributed by him to tidal action. As the grain size varies between laminations then they must result from regular changes in the hydrodynamic conditions. The laminations are caused by alternating periods of current activity, the sand layers being formed during the active periods of flood and ebb tide, whereas the mud is deposited during the stand-still phases of high and low tides. Recently, however, Hawley (1981) has proposed that rhythmites are possibly related to storm activity more than to tidal processes on their own, although he feels that a tidal flat environment is still the most likely area for the formation of these structures.

Vos (1977) described a mixed tidal flat deposit from the Upper Palaeozoic of Morrocco which shows sedimentary structures similar to those found in the Waaipoort Formation. These deposits are composed of lenticular, wavy and flaser bedded sequences of fine to very fine-grained sandstones, siltstones and silty shales with occasional convolute bedding. A number of different types of ripples (linear, undulatory and interference) are found, while crawling trails are also present. These deposits show little lateral variation over many kilometers, and are up to 10m thick. The rhythmites of the Waaipoort Formation correspond closely to this facies, and are therefore interpreted here as being mixed tidal flat deposits. The medium grained units which show low angled cross-stratification are interpreted as representing barrier beach deposits in front of this tidal flat. Fig.28 shows a mixed tidal flat deposit overlain by barrier/beach deposits with low angled cross bedding. Kraft (1971) described similar deposits from the Oregon coast. The cross-bedded units of the Waaipoort Formation correspond to tidal channels and the impersistent quartz arenite bands represent high energy storm deposits (Vos, 1977). A schematic diagram showing the development of this formation is shown in fig.29.

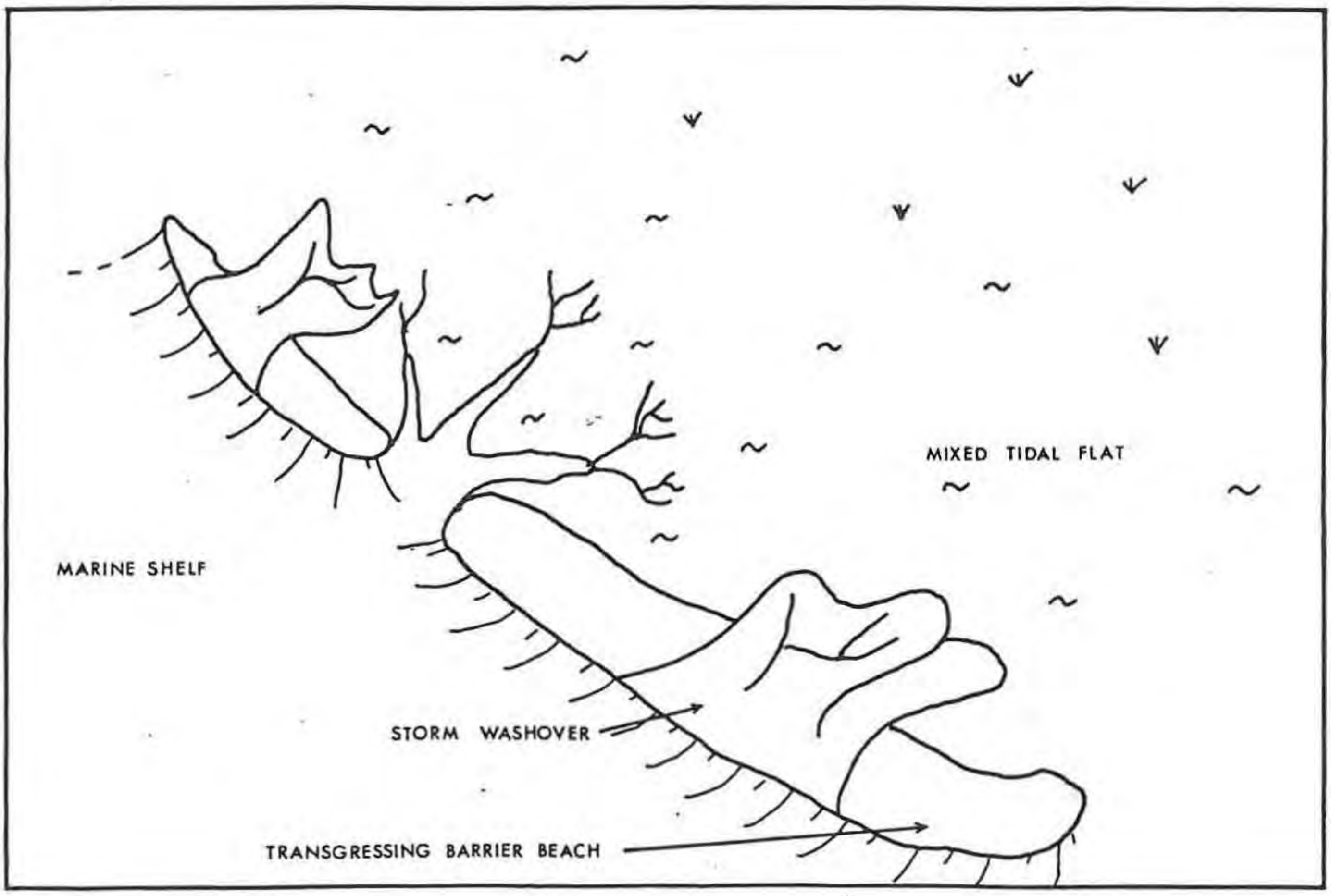


Fig.29: Diagram illustrating the hypothetical development of the Waaipoort Formation (after Vos, 1977,fig.15-c,p.1256)

Fig.28: The sharp contact of the mixed tidal flat facies with the overlying beach deposits

The Soutkloof Shale and Dirkskraal Formations represent a coarsening upward sequence characteristic of a delta. According to Coleman (1976), Wright (1978) and Elliott (1978) pro-delta muds have parallel lamination as the most common sedimentary structure as deposition is from suspension. Coleman also notes that because of the high rates of deposition these muds will not show a great deal of bioturbation. The Soutkloof Shale is characterised by parallel lamination and also lacks intensive bioturbation. Pro-delta muds generally have a high faunal content (Coleman, 1976; Wright and Coleman, 1974) but body fossils are absent from the Soutkloof Shale, possibly for the reasons given above. These shales are therefore interpreted as representing pro-delta muds.

Pro-delta muds are overlain by the delta front environment which is characterised by silts and clays. They have a high degree of bioturbation and small scale cross-laminae, current ripples, scour and fill structures and erosional truncations. The Dirkskraal Formation shows some of these characteristics, but not all. Current ripples and scour features are found but the lack of biological activity is significant. Some bioturbation is present, but is not of the magnitude found in other delta front deposits. However this absence of life may be explained by the reasons given above and the Dirkskraal Sandstone is interpreted as being a delta front deposit.

The Miller Diamictite and Swartwaterspoort Sandstone Formations which underlie the above two formations are more problematical. The diamictite with its lack of sedimentary structures and poor sorting is interpreted as a debris flow deposit (Crowell, 1957; Bell, 1981) off the front of the delta. Delta fronts are characterised by a high depositional slope (Coleman, 1976) with a resulting instability. No evidence for the glacial origin proposed by Rossouw (1953) and Johnson

(1976) was found in this study. The Swartwaterspoort Sandstone is interpreted as representing a reworking of the Miller Diamictite because of its close spatial and mineralogical relationship with the diamictite. The Kommadagga Subgroup is therefore interpreted as representing a prograding delta complex.

The succeeding Dwyka Tillite Formation is unrelated to the underlying rock units in terms of both time and processes of deposition. It was previously believed that in the Eastern Cape the Cape and Karoo Sequences represented uninterupted sedimentation (Haughton, 1969, p.352; Rust, 1973), whereas to the west a disconformable relationship existed between the two sediments (Rust, 1973; Dunlevey and Hiller, 1979). However, as Johnson (1976) notes, the Dwyka Formation does not always overlie the Dirkskraal Formation, but may rather be underlain by the Soutkloof Shale, indicating possible pre - Dwyka erosion in this area as well.

The Dwyka has the characteristics of glacial marine sediments described from the Antarctic shelf by Anderson et al.(1980). According to them three basic types of glacial marine sediment are found in Antarctica and these are:

a) basal till, which results from deposition by grounded ice,

b) compound glacial marine sediments which are deposited from floating ice in a low energy marine environment, and

c) residual glacial marine sediments which are deposited from floating ice in a high energy marine environment.

The main tillite is representative of the first type of deposit, which, as Anderson et al. (1980) point out, is indistinguishable from terrestrially deposited till. The second type is represented by the poorly sorted and crudely stratified units, and the muds represent the distal deposits of this facies (Edwards, 1978). According to Anderson et al. (1980) the third type is rare in modern glacial - marine sediments and has not been observed in the study area, however some of the arenaceous units found near the base of the tillite at Verdun may correspond to this type of deposit. Several stages of transgression and regression, similar to those found in the Western Cape by Theron and Blignault (1975), have been recognised. The exact number of phases is still uncertain, but at least two advances of the ice front have been recognised.

2.4 Provenance :

Rock fragments and heavy minerals indicative of igneous, sedimentary and metamorphic environments are found in the Lake Mentz and Kommadagga Subgroup sediments. Mineralogical evidence for a metamorphic source area is given by the presence of metamorphic rock fragments, rutile, schorl and almandine garnet, whereas an igneous source is indicated by the presence of quartz grains with rutile needles, igneous rock fragments, zoned euhedral grains of zircon and K'feldspar with euhedral zircon inclusions. Numerous chert grains suggest that the source area was a sedimentary one. Grains indicating two possible types of source area may be found in a single specimen suggesting that the river which carried the detritus had a large drainage basin. Johnson (1976) has also proposed a metamorphic - igneous - sedimentary terrain as a source area for this deposit.

Previous workers have suggested that the source area of the Witteberg Group sediments lay to the north of the basin (Loock, 1967; Johnson, 1976; Lock, 1978; Rust, 1973). Although few palaeocurrent indicators were found in this study, those that were found support this conclusion. A suitable source area for this deposit could therefore be the Namaqua-Natal mobile belt, which contains metamorhic, igneous and sedimentary rocks.

CHAPTER THREE

STRUCTURE

3.1 Introduction

The dominant structural feature of the study area is an overturned anticline which plunges at about 9 degrees towards the west. The anticline has the characteristic features of folds described from elsewhere in the Cape Fold Belt (De Swardt et al.,1974) in that it has an overall asymmetry with a gently dipping southern limb, a steeply dipping northern one and a southward dipping axial surface.

The aim of this part of the study was to map all the folds and their associated minor structures and then to deduce the deformational history of the area using this information. No detailed work has previously been done on the area, so all mapping was done using the 1:50 000 topographic map of the area enlarged to a scale of 1:12 500. Lithological and structural information was recorded on the map and in field note books.

As the sequence of sediments is lithologically heterogeneous the contained structures vary considerably from brittle deformation features such as fractures in fold hinges and breccias associated with faults, to ductile flow structures such as necking. The style of deformation is controlled by the coarse grained, well indurated sandstones which are concentrically folded, with disharmonic folding becoming common away from these units.

3.2 Major Structures :

Two structural domains separated by a fault are recognised in the study area. The northern domain is characterised by overfolding, while the southern one has more upright folds. The fault which separates the two domains trends east-west across the area mapped.

3.2.1 Folds :

The fold in the northern portion of the area has the form of a plunging inclined anticline which is overturned towards the north (fig.30, see back folder). Its axial surface dips to the south at an angle of 80 degrees, and strikes 284 degrees (fig.31), while it can be seen from fig.30 that the hinge plunges 9 degrees towards 289 degrees. The amplitude of this first order fold is at least 80 metres, while the wavelength cannot be determined accurately as its full extent is not exposed, but is of the magnitude of 3 000m.

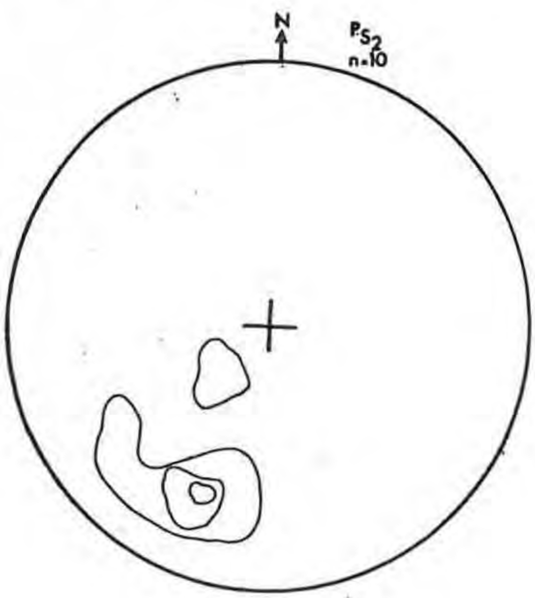


Fig.31 :Stereogram showing poles to axial planar cleavage in the northern portion of the study area.

The fold and its associated minor structures are well exposed in the road and railway cuttings. From these cuttings it is clear that tight to isoclinal folds are common in the Soutkloof Shale, whereas in the Waaipoort Formation the second and third order folds are more open and fracturing is common. Brittle failure is indicated by the presence of breccias and the widespread fracturing found in the hinges of folds. (see fig.32).

Fig.32: Fracturing in fold hinges in the Waaipoort Formation in the northern domain.

In the south the folds are more upright and open, and their axial surfaces also dip southwards. The wavelength of these southern first order folds is 500 meters and the amplitude is 30 metres. In general signs of brittle failure associated with folding are absent in this domain.

3.2.2 Faults :

A thrust fault strikes east-west across the central portion of the mapped area, and is located everywhere at the base of the Floriskraal Formation (see figs.5 and 30, see back folder). The fault plane is itself not exposed, but its presence has been inferred from a number of features. Firstly, the presence of breccia zones near the contact of the Floriskraal and Waaipoort Formations suggests that a fault is present. Secondly the Floriskraal Formation dips steeply to the south, but no beds are recorded as dipping northwards and there is no evidence for isoclinal folding of this Formation. Thirdly, the intensity of deformation increases with proximity to the fault. This change in intensity can be seen in the increase of minor faults and intraformational isoclinal folding in the shales. The presence of widely spaced breccia zones suggests that the fault is not a single plane of faulting, but is rather a zone. Thrust faulting is well known in the Cape Fold Belt (Theron 1969; Hiller and Snowden, 1981; Halbich, 1979; 1981), so this is not an unexpected feature.

The orientation of the fault is difficult to determine accurately because of the lack of exposure but field evidence suggests that it strikes east-west and dips at 20 to 30 degrees to the south. The displacement is also difficult to assess, but it appears to have not been very great as the Floriskraal Formation now overlies the upper portion of the Waaipoort Formation. The estimated magnitude of the throw of the fault is 20 to 30 m.

3.3 Minor Structures :

3.3.1 Tension gashes :

Fine examples of these structures are commonly found in the coarse grained Swartwaterspoort Sandstone (see fig.33), where they range in width from a few millimetres to 2.5 centimetres, and their maximum long dimension is 10 centimetres. The fractures are usually developed in single arrays, but occasional conjugate sets are found, in which case one member of the set is generally less well developed than the other. The fractures are always filled with quartz.

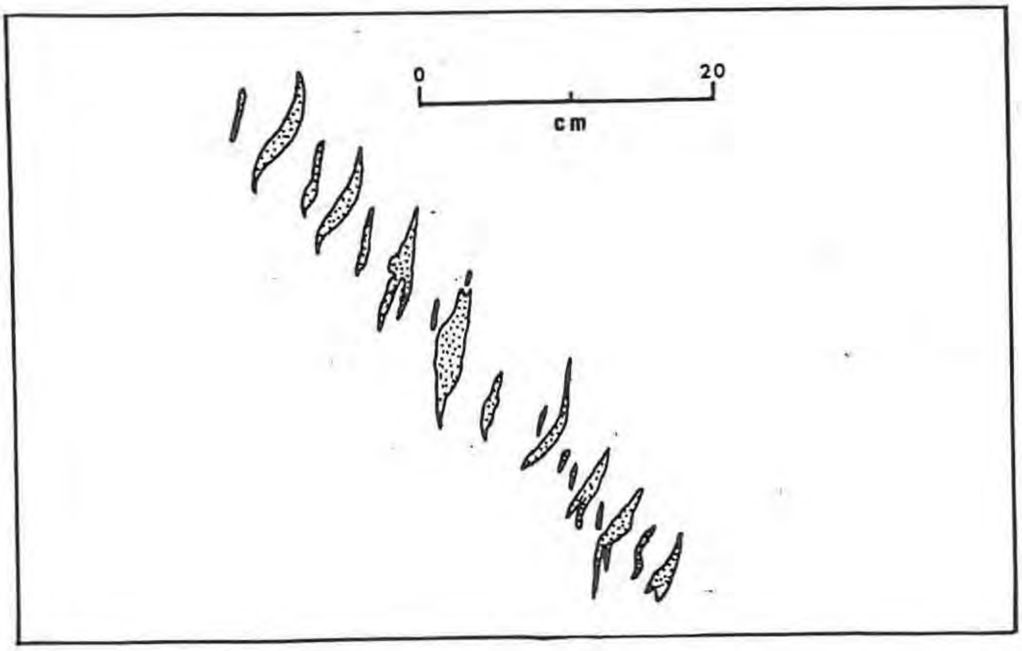


Fig.33: En echelon tension gashes in the Swartwaterspoort Sandstone Formation, drawn from a photograph.

The quartz arenites in which these gashes are found are massive with little topographic relief and consequently three dimensional exposure of these fractures was limited. Nevertheless measurements of the orientation of individual gashes as well as zones of gashes were made.

Several zones of these fractures are exposed on a hill where the Swartwaterspoort overlies the Waaipoort Formation sediments which have moderately well developed cleavage. Models explaining the formation of these gashes suggest that the direction of maximum principal stress (designated σ_1) bisects the acute angle between two conjugate sets of these fractures (see fig.34) (Wilson,1961,p449). The poles to these sets were plotted on a stereographic projection (fig.35) and was calculated to be inclined at 15 degrees and directed from 189 degrees. As axial planar cleavage is commonly believed to generally develop perpendicular to σ_1 , the orientation of cleavages measured in the underlying Waaipoort Formation were also plotted on the stereogram (fig.35). As the cleavage is 89 degrees from σ_1 it therefore seems that the theoretical models can, in this case, explain the origin of the natural occurrences.

Isolated occurrences of these gashes occur in the Dwyka Formation where they are not filled with any material.

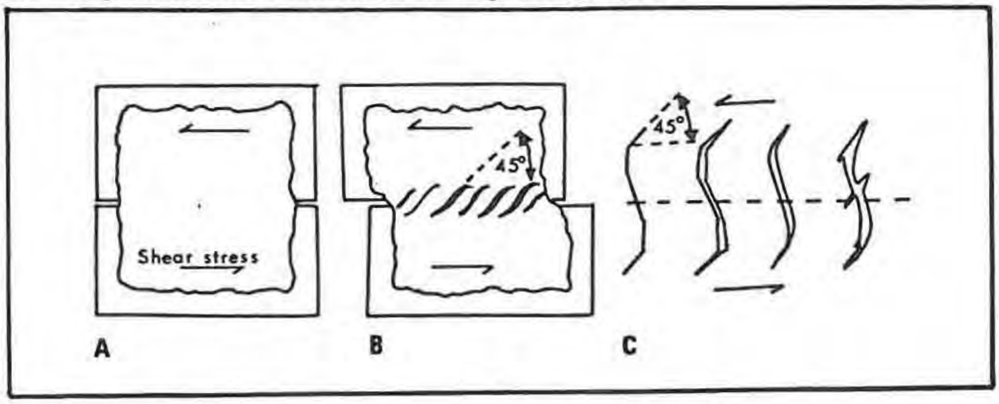


Fig.34: Diagram explaining the formation of en echelon tension gashes (after Wilson,1961,p.449)

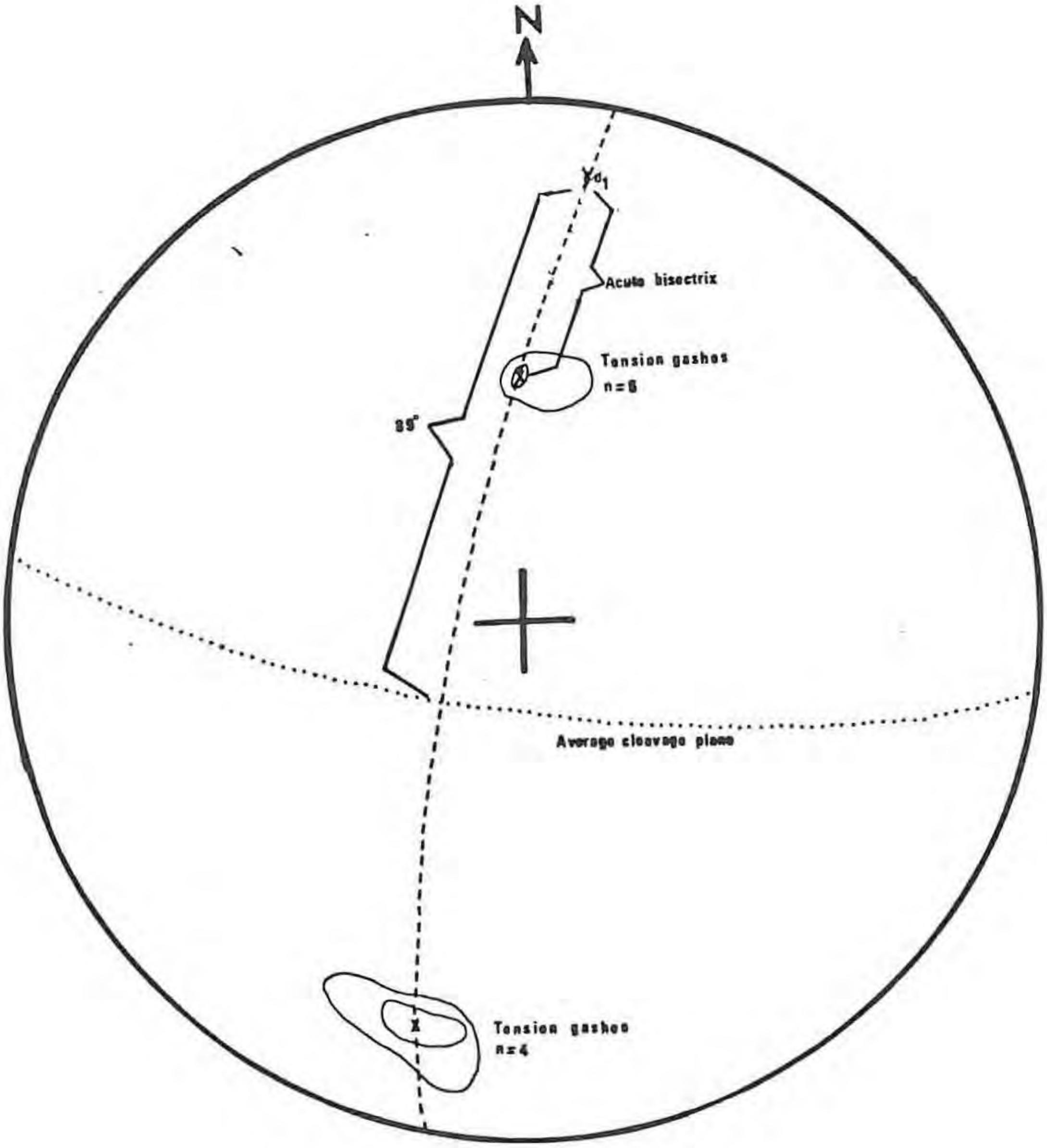


Fig.35: Stereogram showing relationship between cleavage and tension gashes.

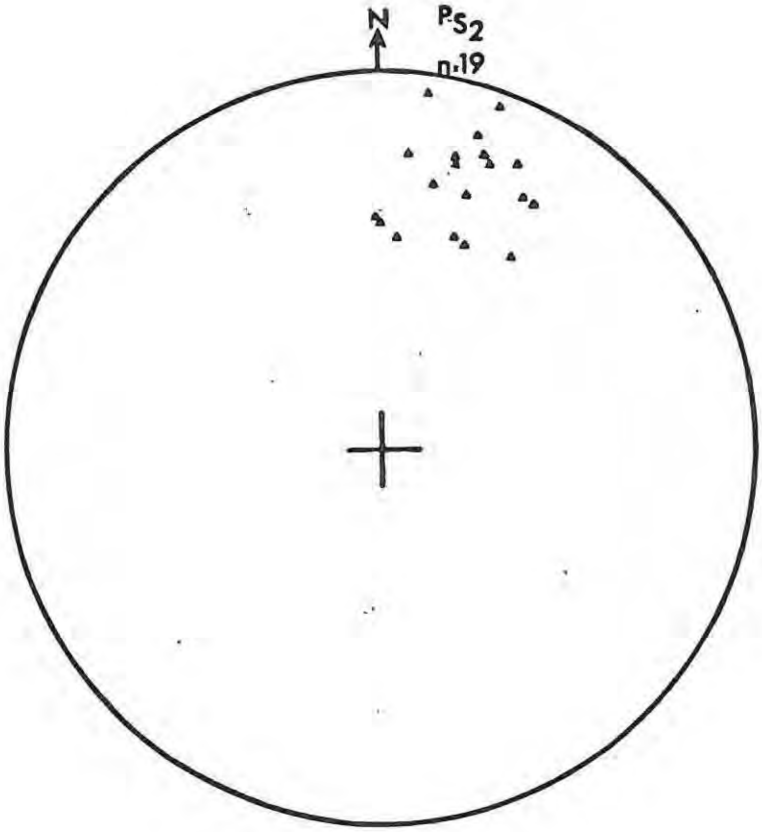
3.3.2 Axial planar cleavage :

Axial planar cleavage is particularly well developed in the finer grained rocks of the study area. The dip of this structure is predominantly 80-85 degrees towards the south (fig.36).

Axial planar cleavage morphology ranges in type from a penetrative cleavage in shales to a spaced fracture cleavage in the arenaceous units, while the coarse grained units contain more widely spaced jointing parallel to the cleavage orientation.

In places the cleavage has caused the original primary bedding foliation, where the micas have adopted a preferred orientation during settling, to crenulate (fig.37). This crenulation should not be confused with that associated with polydeformation, as the earlier foliation is not tectonic in origin but rather sedimentary (Hobbs et al., 1976,p153). Similar cleavage-bedding relationships have been observed in Australia (Williams,1972; Gray,1978) and Maryland, U.S.A. (Geiser,1974).

Thin sections of the Upper Witteberg and Dwyka rocks show that they have a cleavage similar in form to the rough cleavage of Gray(1978) (see fig.38). Gray studied deformed psammitic rocks, metamorphosed to lower greenschist facies, and proposed that pressure solution processes play an important role in the formation of these cleavages. He defined psammite as being a weakly metamorphosed sandstone with abundant clastic material held together by a cement, some matrix material or a combination of the two. He showed that in psammites the nature of the original sedimentary fabric may determine the nature of the ultimate cleavage type. Important features of the sedimentary



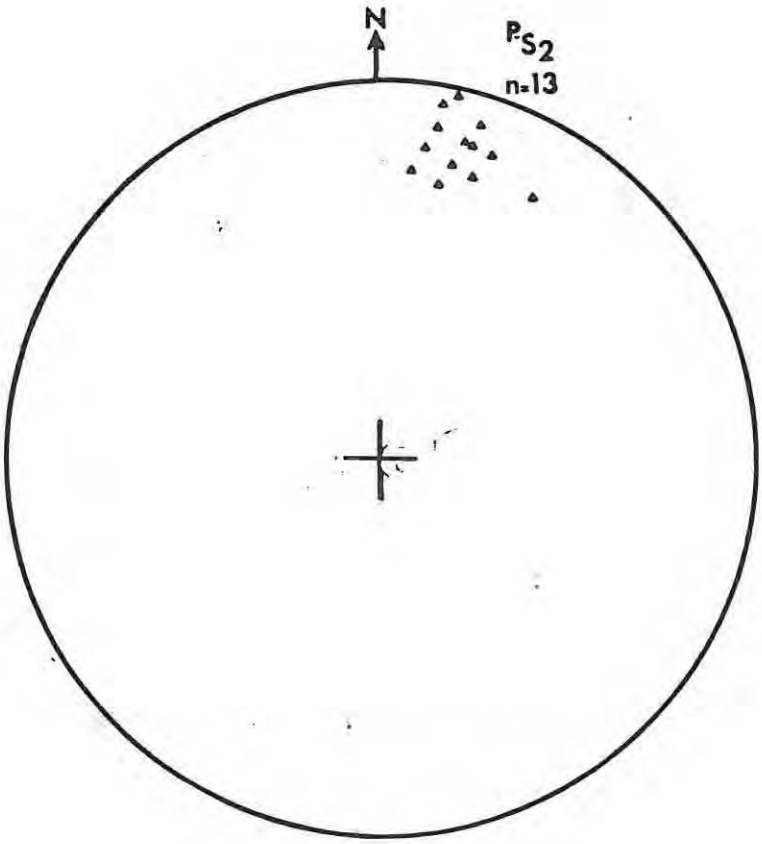
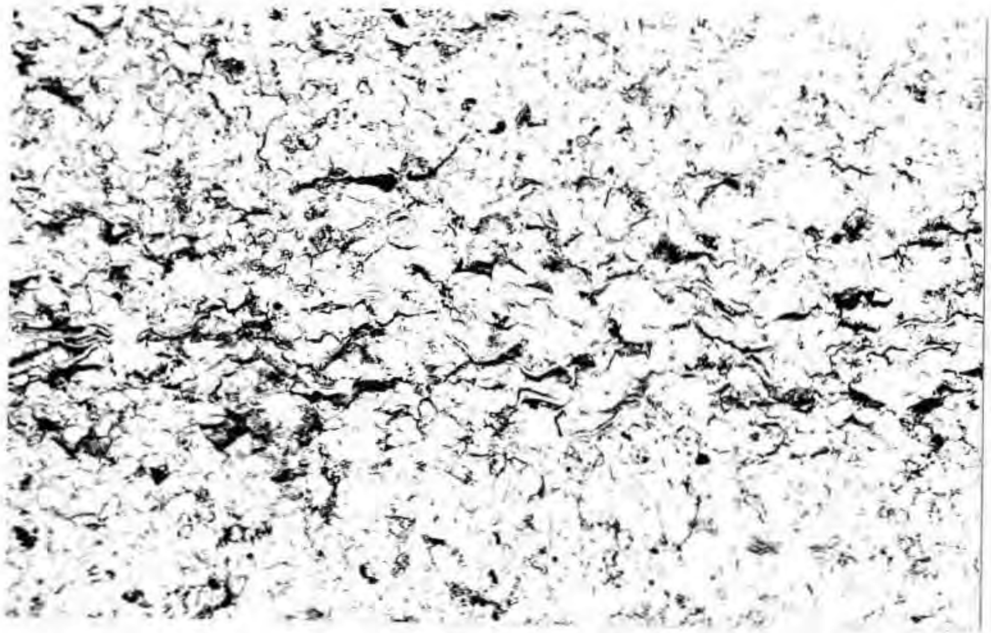


Fig.36: Stereograms indicating cleavage orientation from various parts of the study area



1mm

Fig.37: Crenulation of a primary bedding fabric.

(2.5*; plane polarised light;sample K-24)

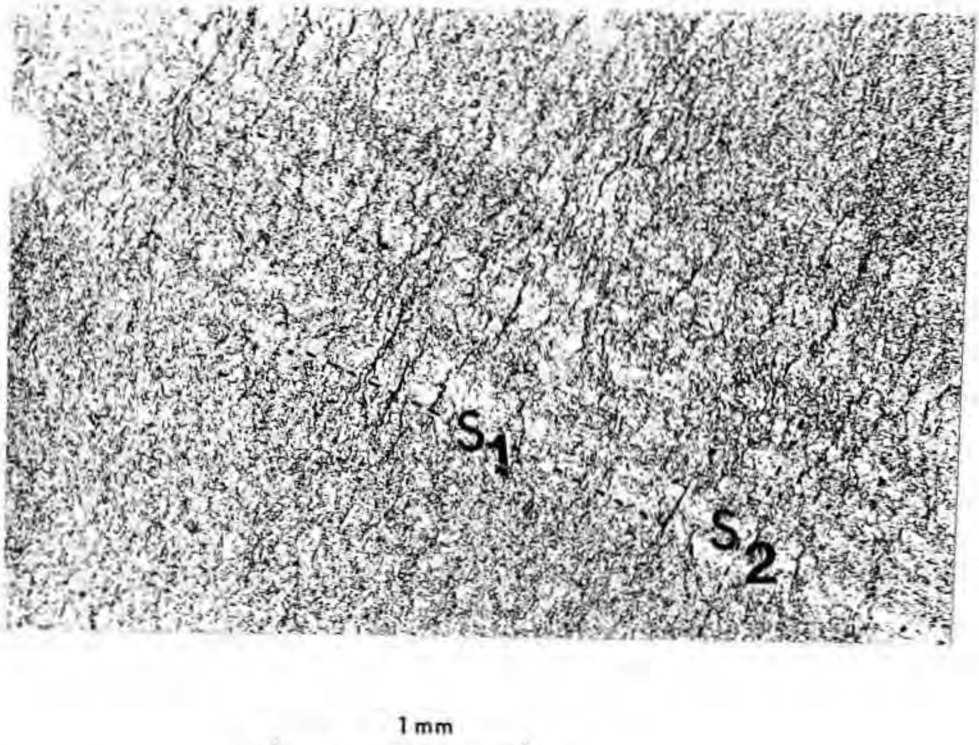


Fig.38(a): Rough cleavage in the Soutkloof Shale Formation

(2.5*; plane polarised light; sample K-57)

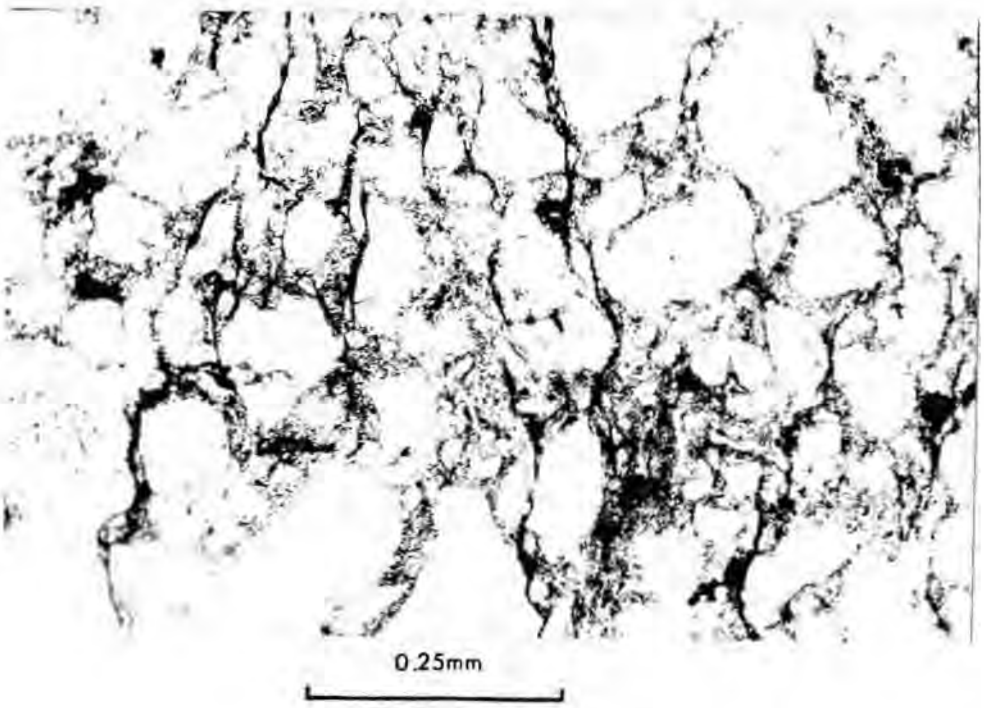


Fig.38(b): Rough cleavage in the Waaipoort Shale Formation.

(10*; plane polarised light; sample K-52) fabric which affect the morphology of the cleavage are :-

1) the nature of the framework,
2) the grain size,
3) the grain shape,
4) grain orientation and
5) packing.

As the nature of the framework is the most important of these factors, he proposed a classification scheme of psammites based on this parameter and it is summarised in fig.39.

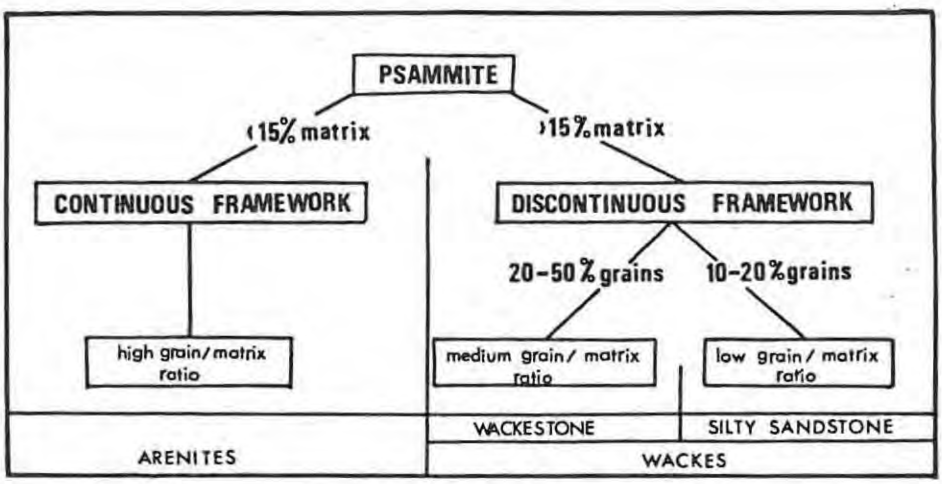


Fig.39: Gray's classification of psammites according to sedimentary framework (after Gray, 1978,fig.1).

Gray found that in psammitic rocks three distinct types of microstructural cleavages occur, and that these were only end menbers of a continuous spectrum of cleavage types (Gray,1978; Means,1975).

These three end members were designated :

1) Type A, with short, discontinous cleavage traces around random or weakly orientated grains.

2) Type B, which have closely spaced, well developed continuous cleavage traces around orientated elongate detrital grains.

3) Type C, with continuous traces in isolated zones of the psammite, with possibly only weakly developed cleavage in the intervening areas. Fig.40 summarises these characteristics.

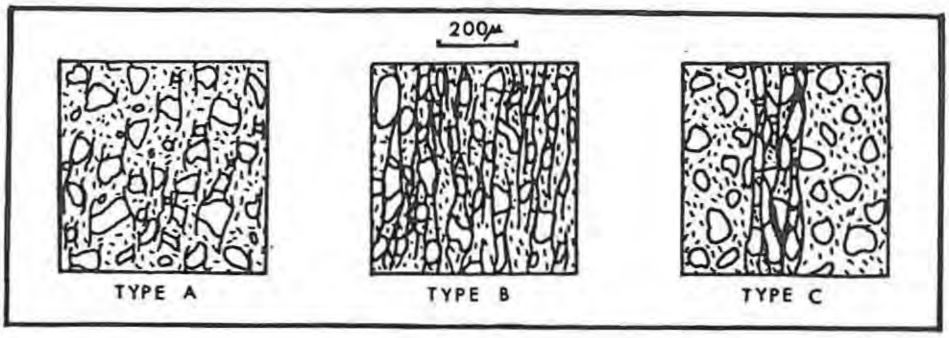


Fig.40: Diagrammatic sketches of rough cleavage fabrics in psammitic rocks (after Gray,1978,fig.6).

The cleavage seams are pressure solution surfaces, perpendicular to which soluble minerals (such as quartz, feldspar and calcite) have been corroded leaving behind dark seams of insoluble residues of clay, micas, carbonaceous matter and heavy minerals (Gray,1978). Geological evidence for these being pressure solution features is that well rounded detrital grains may be truncated on one side by a seam of other residues, while evidence for overgrowth exists elsewhere. These grains often show no other signs of deformation such as undulose extinction or deformation lamellae.

Mathematical models (Durney, 1976; 1978) have been proposed to explain the thermodynamics of pressure solution processes, but will not be discussed here. However an important point is that as solution of grains occurs perpendicular to σ_1, and redeposition parallel to σ_3, the cleavage which results is axial planar.

Durney (1972) feels that pressure solution is more pronounced in pelites than in coarser grained rocks because they have shorter diffusion paths and therefore the rates of transfer are higher. However, cleavage of this nature is not well developed in the very fine grained rocks of the Kommadagga area, and this is probably due to the fact that although the percentage areal change per grain may be greater in fine grained rocks, the total volume of material moved is small in comparison with that in psammitic material. As a result the volume of insoluble residue accumulated in pelitic rocks is not great, and the resulting foliae are very thin. Gray (1978) suggests that rough cleavage in psammites is the morphological equivalent of slaty cleavage in pelitic rocks.

During the development of pressure solution cleavage, modification of grain shape occurs, and therefore grains associated with this type of cleavage either have sutured boundaries or planar truncated sides (Gray,1978). Other possible mechanisms for this grain shape alteration do exist and these are :

1) intracrystalline plastic deformation
2) recrystallisation
3) solution transfer (Durney,1972)

However, as shown in the chapter on diagenesis, metamorphic conditions were not suitable for recrystallisation, and intracrystalline deformation is precluded because many clastic grains in the thin sections, which show this type of cleavage, have no signs of udulose extinction and deformation lamellae.

The tombstone weathering which is so characteristic of the Dwyka Tillite Formation in many of the more arid areas is caused by cleavage of this type. Fig.41-a shows a thin section of Dwyka Formation with well developed rough cleavage of type B. Cleavage is also well developed in the fine grained units of the Dwyka Formation, and thin sections of the stratified tillite show this feature well (fig.41-b). Not all portions of the Dwyka show the tombstone weathering and this is probably due to a combination of climate, aspect of slope and the presence or absence of this type of cleavage. As the Miller Diamictite also shows weakly developed tombstone weathering, it is probable that this feature will only form in massive rocks where the only favourable weakness for erosion and weathering is the cleavage plane.

Evidence for the development of a second cleavage was found in a quarry north of Aalwynspoort. The cleavage is found in a shale from the Waaipoort Formation but as the shales are highly weathered, exposure is poor, and it is difficult to be certain that this cleavage represents a second phase of deformation. No other evidence was found in the study area for polydeformation.

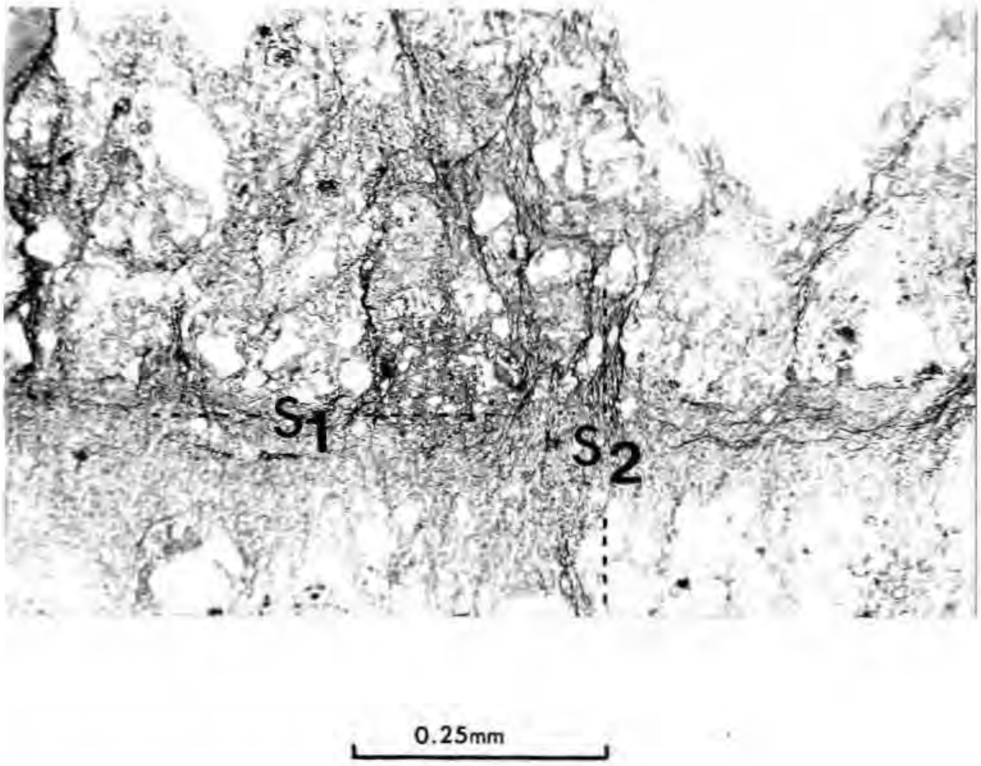


Fig.41(a): Dwyka Tillite Formation with rough cleavage

(10*; plane polarised light; K-66)

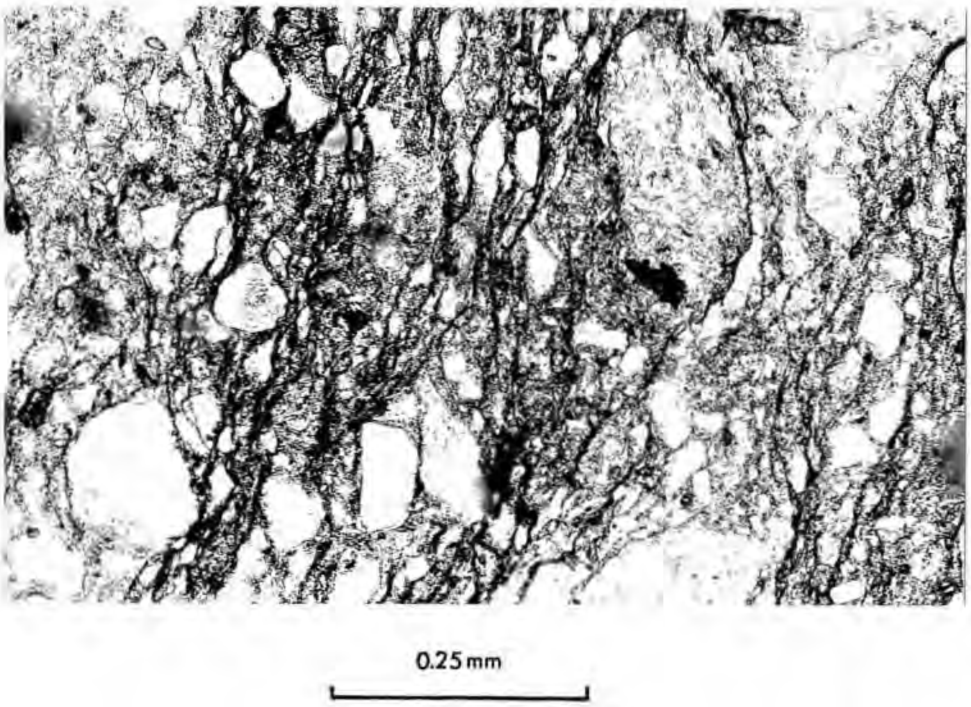


Fig.41(b): Crudely bedded tillite with rough cleavage

(10*; plane polarised light; K-62)

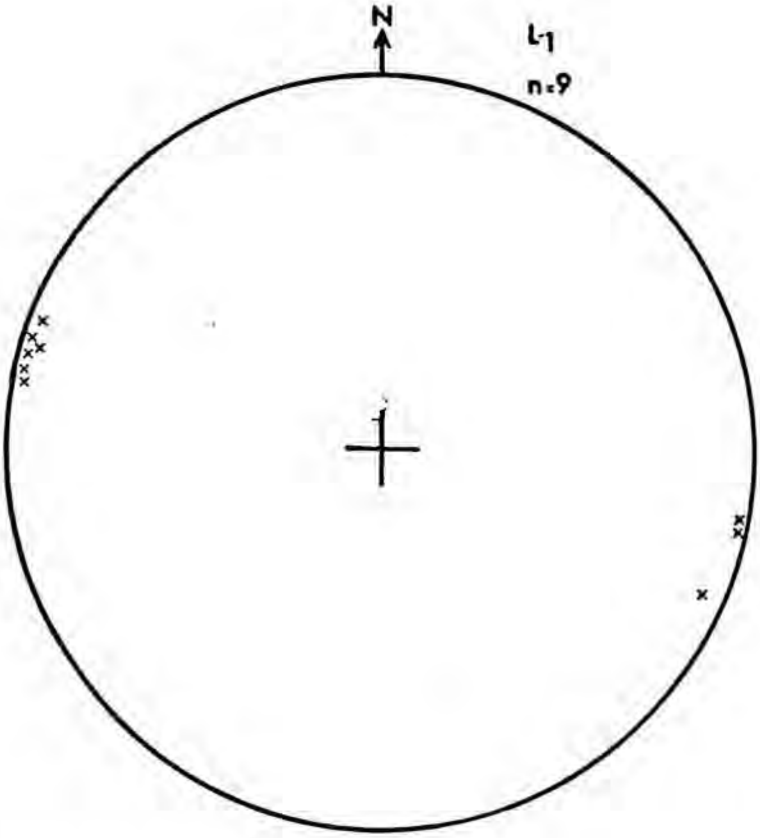


Fig.42: Stereogram showing plunges of lineations

Fig.43: Boudinage of carbonate units in the Soutkloof Shale Formation

3.3.3 Lineations :

In areas where the axial planar cleavage is well developed there is a corresponding lineation, called L-1 here, which is defined by the intersection of S-1 (bedding) and S-2 (cleavage). The plunge of this lineation is generally about 9° towards 286°. Fig.42 shows some stereographic plots of the plunge of L-1. Since S-2 is axial planar this lineation gives the plunge of the major fold structures.

3.3.4 Slickensides :

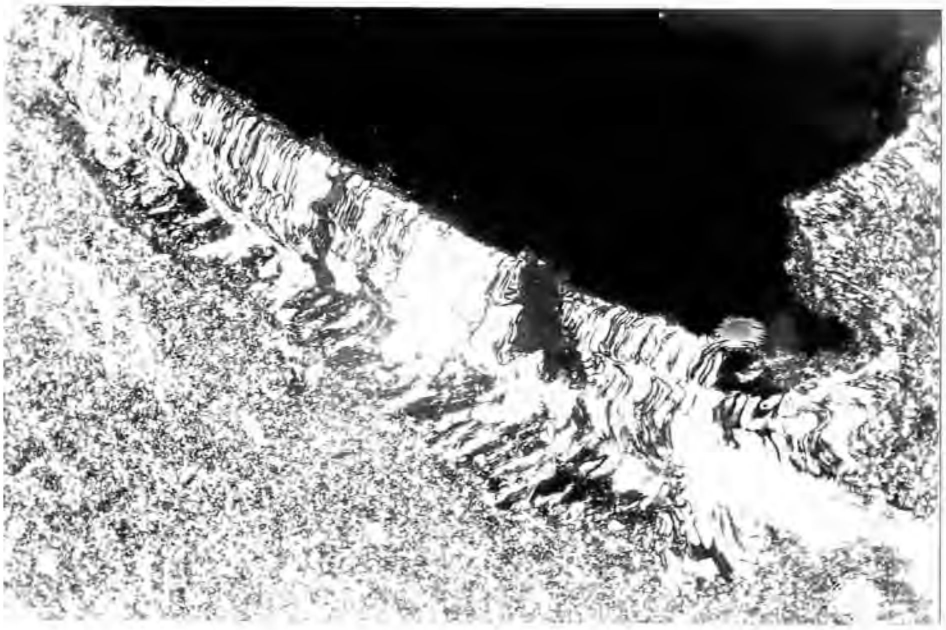
These are common structures on the fracture surfaces and bedding planes of both competent and incompetent units. On some surfaces more than one generation of slickensides were observed, probably caused by changes in the direction of tectonic pressure. On some surfaces steps were encountered which faced in opposing directions, which suggests that at least two phases of movement occurred on that surface. However, in general only one set was developed and this was taken to face the overall direction of transport, despite the fact that experimental work (Paterson,1958; Gay,1970) and natural examples (Tija,1964) suggest that this may not be the case.

3.3.5 Boudins :

These structures are formed only in the carbonate units of the Soutkloof shale, where the carbonate has acted as the competent layer and the surrounding shale has flowed into the zones of necking (see fig.43). These boudins are exposed in a rail cutting north of the Kommadagga siding, but measurements of the plunge of the hinge of these structures was considered to be unreliable because of the effects of soil creep. However the plunge of these linear structures appears to parallel the L-1 lineation.

3.3.6 Microscopic Structures :

Besides the microscopic rough cleavage previously described there are other microscopic deformation structures such as the frequent development of both Bohm lamellae and deformation twinning in detrital grains . Quartz grains never appear to form the pillars behind which pressure shadows can develop, but in one sample of shale interbedded with tillite, a haematite(?) grain has a well developed pressure shadow (see fig.44). The crenulations associated with this presure shadow are not necessarily the result of a second phase of deformation as the axial plane of the crenulation appears to be parallel to the main axial planar cleavage visible in the rest of the slide. Similar pressure shadows have been recorded by Durney (1976,fig.8).



0.25 mm

Fig.44: Pressure shadow formed behind a haematite grain.
(10*; crossed nicols; sample K-57)

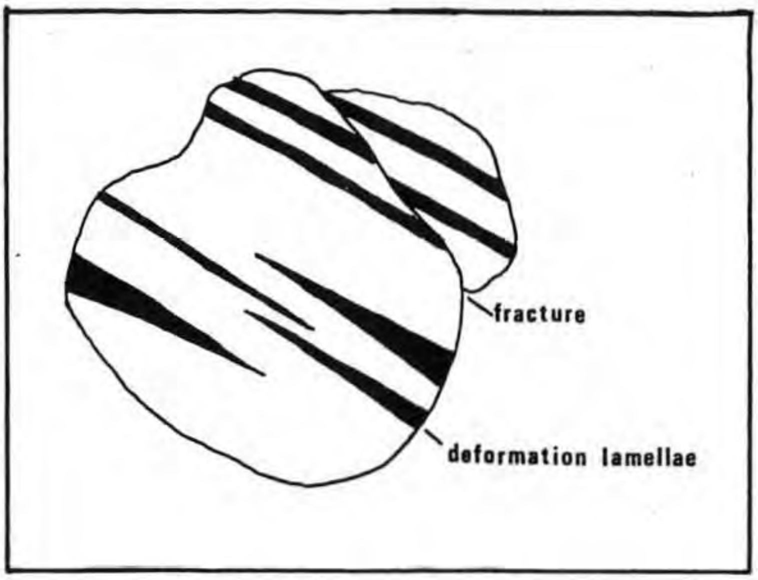


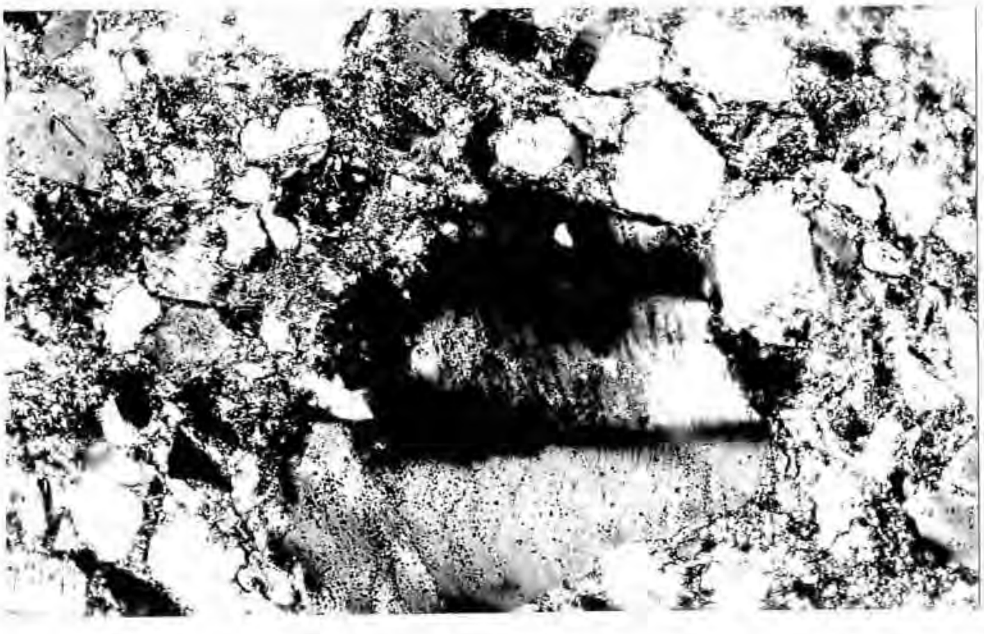
Fig.45: Deformation of plagioclase feldspar

Fig 46: Scanning electron micrograph of deformation of a platy mineral

(1400*; sample K-14)

Plagioclase grains show deformation bands and deformation twinning, while original growth twins may be kinked (fig.45). Mica crystals in places are kinked, especially on a sub-microscopic scale (see fig.46). No detailed work has been done on the orientation of these submicroscopic kinks, and their origin could possibly be non-tectonic.

It was expected that because of the high proportion of matrix in the Dwyka Formation sediments, quartz grains would be unstrained and would not show significant deformation effects. (Siever, 1959; Whinsonant, 1970). However, all cut sections of this unit showed deformation features in the quartz grains (fig.47). A possible explanation is that previously strained grains are not preferentially removed during glacial transport in contrast to fluvial transport where previously deformed quartz grains are mechanically unstable (Blatt and Christie,1963).



0.25 mm

Fig.47: Extreme deformation of quartz in the Dwyka Formation. (10*; crossed nicols; sample K-669)

Suturing is more common in rocks with a low proportion of matrix than in those with a high quantity of matrix. This is in contradiction to the findings of Renton et al. (1969) who concluded from experimental work that the type and amount of matrix has no effect on pressure solution. Whinsonant (1970), however, found an inverse relationship between the amount of suturing and matrix quantity. This apparent contradiction could be caused by a combination of differences in the type of matrix and the amount of diagenesis/metamorphism to which the rock has been subjected. Pressure solution is aided by the presence of a thin film of fluid (Weyl,1959), the composition of which also effects the rate of solution-deposition. Pressure solution occurs most frequently in acid or alkaline solution, but redeposition occurs only in alkaline solutions (Renton et al.,1969). As fluids are released during metamorphism of clays (Fyfe et al.,1978) metamorphism is an important generator of fluids, but no degree of metamorphism will generate water if the minerals did not originally contain water.

CHAPTER FOUR

DIAGENESIS

4.1 Introduction

Diagenesis and very low grade metamorphism are topics which are closely interrelated and have recently received a lot of attention in Europe and North America. This increasing interest in these topics is reflected in the fact that since 1976 two journals have published special issues devoted to diagenesis and the lower grades of metamorphism (J.Geol.Soc.Lond.,135, pt.1, and Can.Min.,12, pt.7), while two books on the subject have recently been reprinted (Larsen and Chilingar, 1979; Berner, 1980). In addition to these publications the proceedings of a number of symposia have been published in book form by the Society of Ecomomic Paleontologists and Mineralogists (Scholle and Schluger, 1979), while the American Association of Petroleum Geologists have reprinted several papers on diagenesis (Ali and Friedman, 1977). Pittman (1979) has recntly reviewed the latest developments in the field but in South Africa these topics have not yet been extensively studied, partly because of the lack of important hydrocarbon reservoirs where knowledge of the effect of diagenesis on porosity and permeability is important.

What is diagenesis? To date no universally accepted definition of the topic has been proposed, and what one geologist classes as diagenesis may be regarded as being metamorphism by another. However, as Blatt, et al.(1980,p332) have pointed out, if the topic is of interest to you one should study it, no matter what it is called. Although definitions are useful in communicating what one is talking about, an understanding of the processes involved is far more important. The literature on diagenesis is also full of jargon, the meaning of which is often clear only to the proposers of these terms.

Three widely differing definitions of diagenesis are quoted below to illustrate some of the complexities involved. Firstly, Coombs (1961) defined diagenesis as being those post-depositional changes which take place at essentially the same temperatures as those prevailing at the time of deposition, while metamorphism involves higher temperatures. Blatt (1979, p.141), however, defined diagenetic reactions as being those that occur in a rock from the time of deposition to the time that there is a significant change in the texture and mineralogy of that rock. This means that the variation in temperature, pressure and chemistry during diagenesis is great. A third definition which involves the mineralogy of the rock is given by Winkler (1979) who feels that the beginning of metamorphism and the end of diagenesis is marked by the formation of mineral assemblages which cannot originate in a sedimentary environment (p.11). This type of definition is very dependant on the bulk rock chemistry which will then control the step from diagenesis to metamorphism (Miyashiro 1973,p.21 ; Ghent and Miller,1974-a; Ghent and Miller, 1974-b), and the role of pressure and temperature is reduced. Diagenesis and metamorphism are therefore part of one continuum and no hard and sharp division can be drawn between them. It is, however, useful to have a name for the processes which one is studying and the term diagenesis will be retained here and will refer to all the post depositional modifications affecting the sediment.

These processes are of interest because of their effect on the porosity and permeability of a sediment. This is important in the exploration for, evaluation and exploitation of hydrocarbon reservoirs. Knowledge of porosity and permeability are also important in groundwater studies, while sandstone uranium deposits are affected by these factors. Further, recognition of post-depositional alteration is important in environmental interpretation as failure to recognise these changes may lead to erroneous conclusions. For example it is now believed by some workers that there is no primary graywacke (eg Galloway,1974; Dickinson, 1971; Whetten and Hawkins, 1970) and that the matrix in these rocks is secondary.

Fyfe (1974) has also shown that low grade metamorphic rocks are important in our understanding of the geochemical and tectonic history of the crust. Sediments rarely contain less than 20% water (Blatt, et al., 1972), the bulk of which exists as pore fluids or in hydrated minerals. Metamorphism of a wet pile of sediments releases the water and by the time the rock becomes a greenschist about 0,5km^3 of water will be lost from every cubic kilometer of rock (Fyfe,1974). The released waters are highly saline, and, as the transport of ore minerals is now commonly believed to be as chloride complexes (Park and MacDiarmid,1975) then the importance of these processes is immense.

What, therefore, are the processes operating after deposition of a sediment? Firstly compaction during burial will be followed by in situ decomposition of clastic grains by circulating ground waters. Secondly new mineral cements may develop either randomly or as oriented overgrowths, whereas existing grains and cement may be removed by solution, and finally new minerals may form.

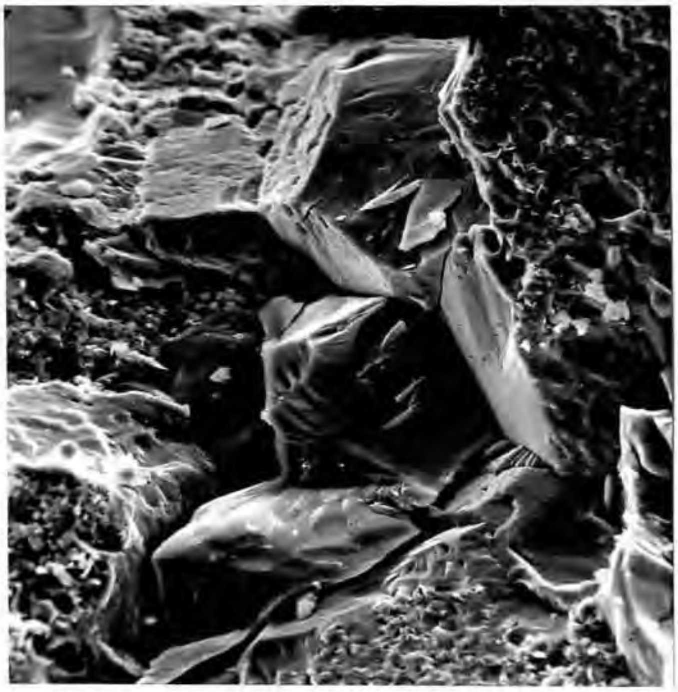
4.2 Methods of study :

Rock thin sections were studied using a conventional petrographic microscope, while rock chips were examined using the scanning electron microscope (SEM) and chemical analyses of individual grains were done using an electron microprobe.

With the petrographic microscope identification of clay minerals is restricted because of their fine grain size. Consequently identification of clay minerals was based largely on their morphology observed under the SEM. The SEM is also useful for distinguishing authigenic clays from detrital ones as the delicate morphology of authigenic clay minerals would not be preserved during weathering, erosion, transport and deposition (see for example fig.62).Very fine grained zeolite crystals, which were not originally seen in thin section, were observed under the SEM. Mineral overgrowths could be recognised by using both the SEM and conventional microscopy.

4.3 Cements :

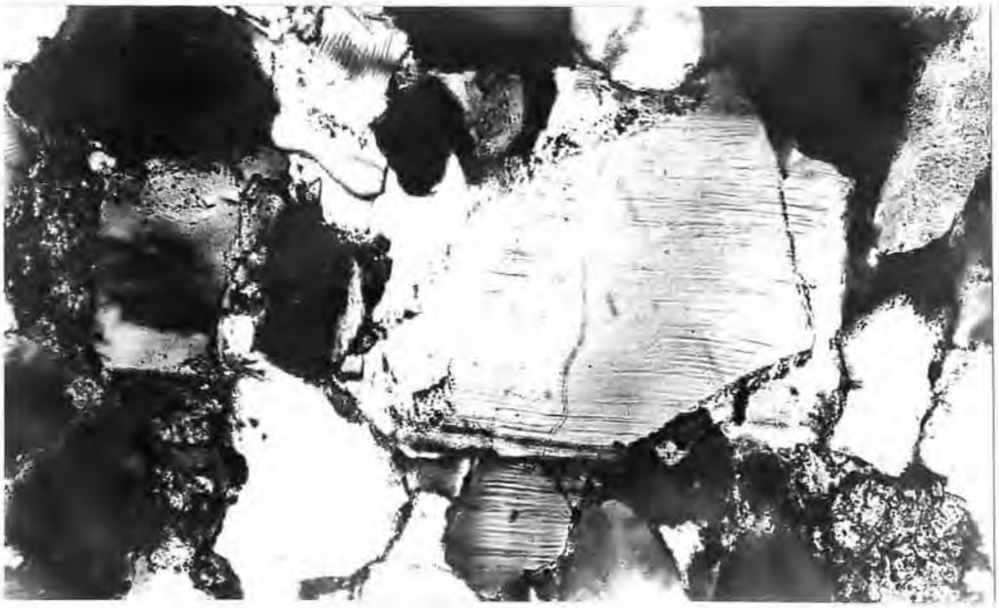
Silica is the most common cement found in the sandstones, occurring most commonly as optically continuous quartz overgrowths on detrital grains (fig.48-a, 48-b). Quartz overgrowths are more common in the quartz arenites than in the feldspathic or finer grained units as the presence of clay minerals and carbonaceous material does not favour the development of overgrowths (Heald,1950; Thomson,1959). These overgrowths can only be seen in thin section when the border of the original grain is defined by either a coating of heavy minerals or a clay layer. The sutured grain boundary contacts seen in many of the purer quartz arenites are probably caused by overgrowths rather than



20µ

Fig.48(a): Quartz overgrowths as seen under the SEM

(600*; sample K-69)



0.25 mm

Fig.48(b):Thin section view of quartz overgrowths

(10*; crossed nicols; sample K-36)

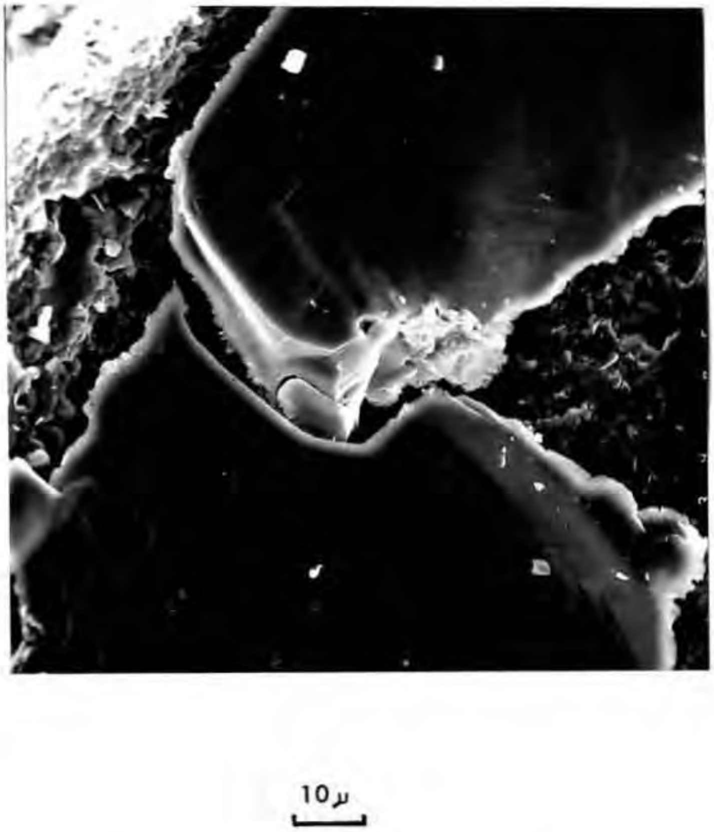


Fig.49(a): Although not common, pressure solution does occur,and is seen here under the SEM.

(3000*; sample K-69)

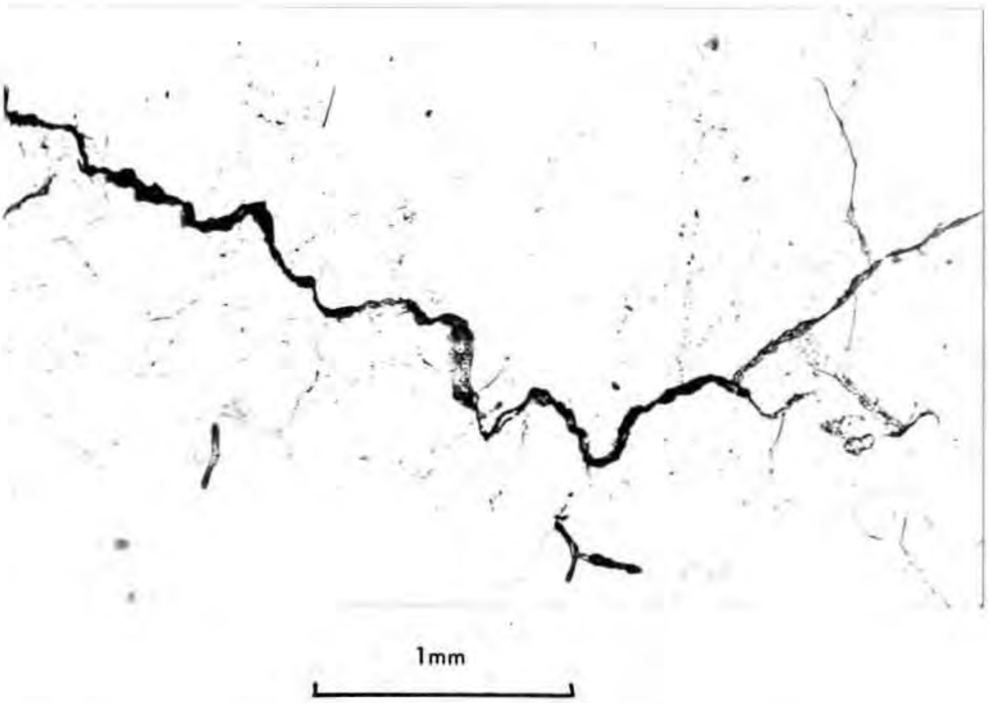


Fig.49(b): Photomicrograph of microstylolites in a quartz arenite.

(2.5*; plane polarised light; sample K-72) by pressure solution. Sibley and Blatt (1976) showed that the interdigitation of adjacent quartz overgrowths caused this type of texture, whereas prior to this it was believed that pressure solution was the cause. Pressure solution does however occur, and may be seen under the SEM (fig.49-a), or as microstylolites (fig.49-b) in thin sections of quartz arenites.

Silica cement also occurs in the form of opal (fig.50). Stanley and Benson (1979) have described opal cement from Tertiary aged continental sediments, and the identification of opal in this study is based on the similarity in morphology of the crystals in fig.50 with that of crystals which they identify as being opal (Stanley and Benson, 1979, fig.13, p.411).

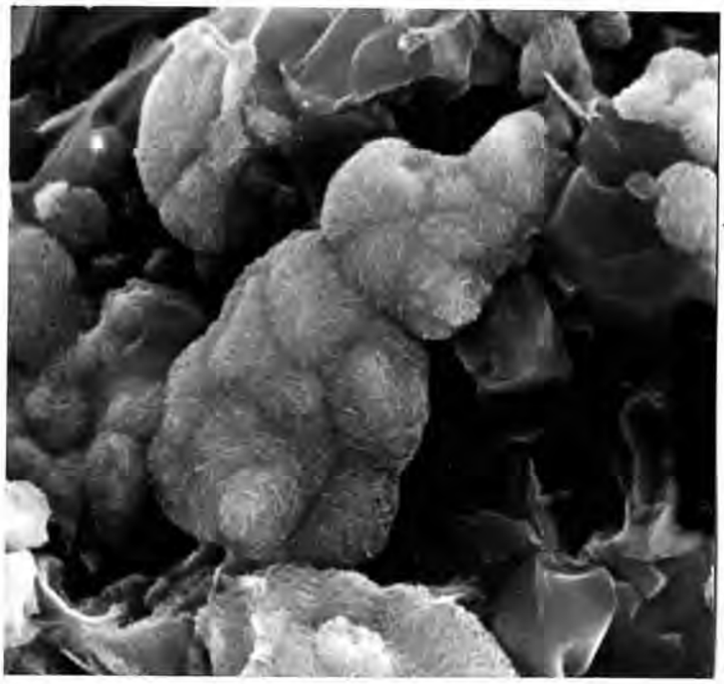
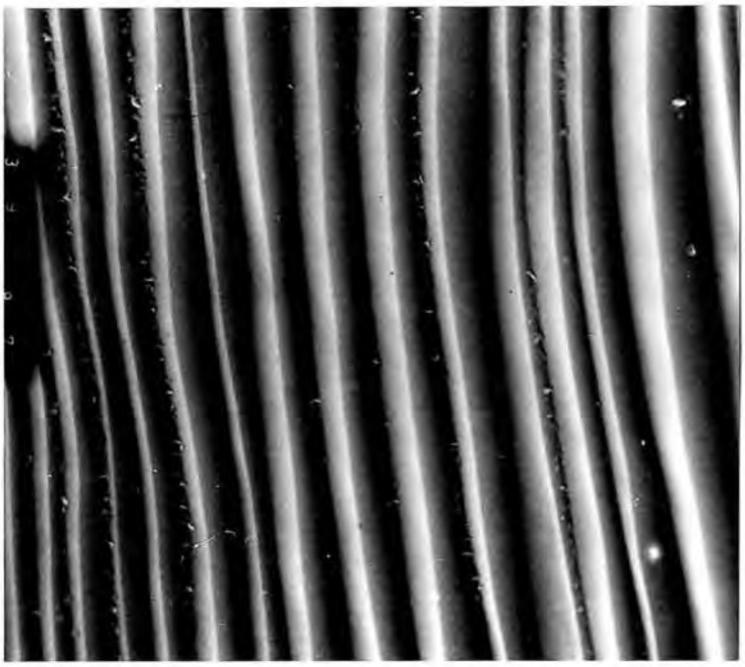


Fig.50: Scanning electron micrograph of an opal cement. This opal is similar in form to opal described by Stanley and Benson, 1979. (8 500*; sample K-24)

The presence of opal in rocks which are Devonian in age is problematical as opal is an unstable mineral, and is not known from rocks older than 60Ma. It is likely that formation of this cement occurs very late in the history of the rock. Fyfe and others (1974 ; quoted in Stanley and Benson, 1979) have shown that opal can form by the alteration of montmorillonite. Although this type of replacement has not been seen using the SEM, it has been seen under the petrographic microscope. It is probable that the opal in these rocks is only an intermediate product in a long sequence of post-depositional changes, and given time will itself be replaced. Post-depositional modification of a sediment can therefore occur over very long periods of time, even after lithification.

Ridge like forms observed on quartz grain surfaces (fig.51) have previously been interpreted as indicating glacial action (Rao, 1981, fig.4; Krinsley and Margolis, 1976, plate 1). However as the sediments in which the marks shown in fig.51 have apparently never been subjected to glacial action then the formation of these structures must be unrelated to these processes. Further evidence for a non-glacial origin of these structures is the fact that these marks have been observed on undoubted quartz overgrowths (see fig.48-a and fig.52 which is an enlargement of part of fig.48-a). Similar features have also been observed on large quartz crystals (fig.53) extracted from a geode. Although the exact mechanism of formation of these structures is still uncertain, it is probably related to crystal growth, either being the pattern which results from the interpenetration of growing crystals or as a result of growth defects. Similar growth related steps have been observed in NaI crystals(Buckley, 1951). The interpretation of glacial activity based on the presence of this type of feature should therefore be approached with caution. Grains in



20μ

Fig.51: Ridge like forms observed on quartz grains

(3000*; sample K-116)

20μ

Fig.52: Overgrowths showing the ridge features. This is an enlargement of part of fig.48-a.

(6000*; K-69)

0.2 mm

Fig.53: Quartz crystal showing the same features shown in figs.51 and 52. (50*; quartz crystal)

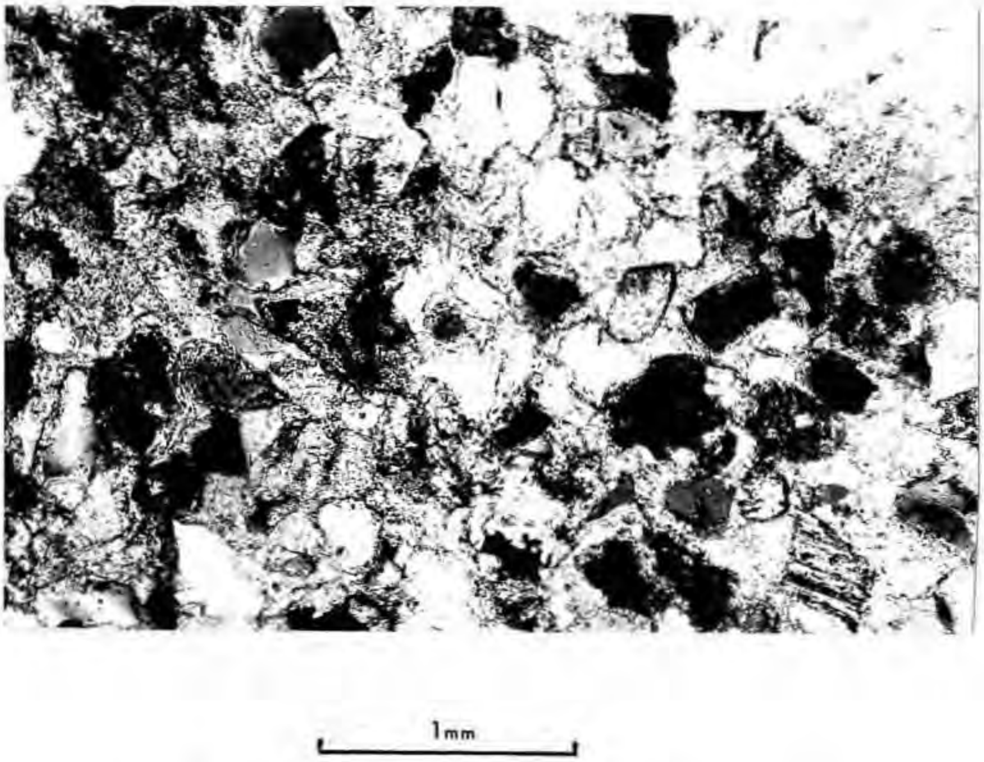


Fig.54: Blocky calcite cement in thin section

(2.5*; crossed nicols; K-17)

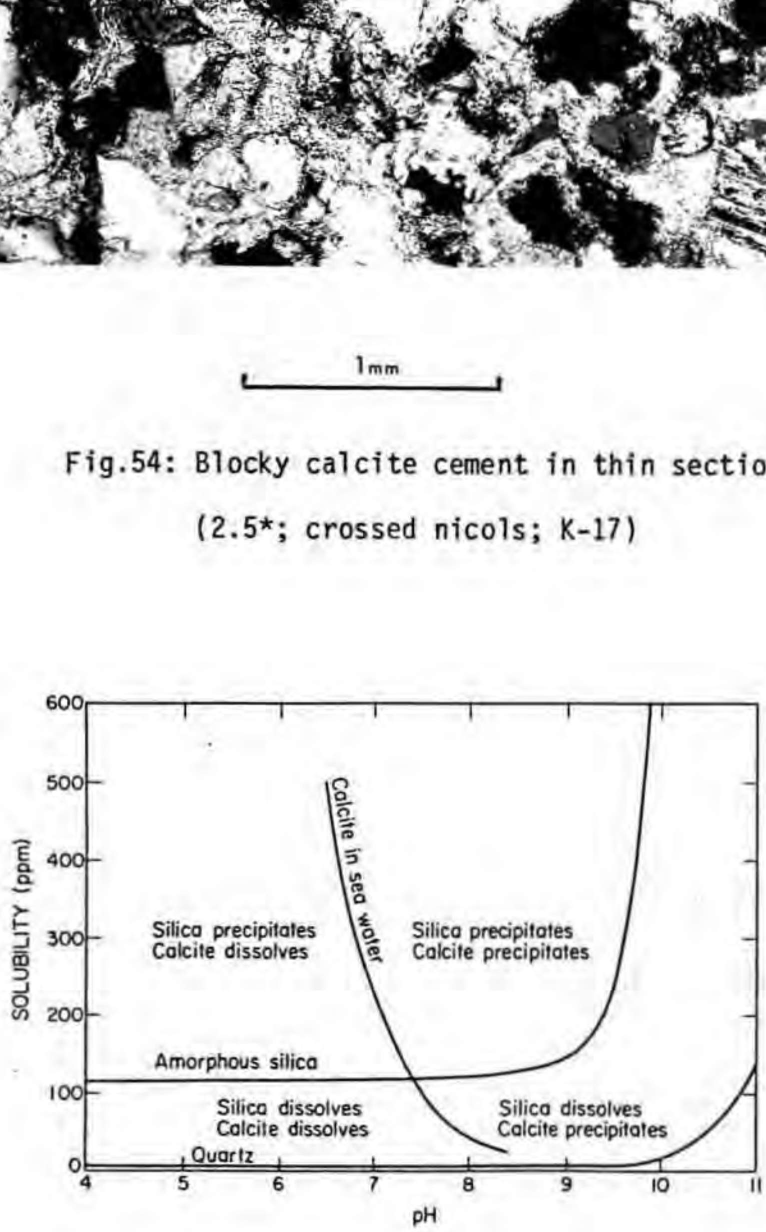


Fig.55: Relationship between pH and the solubilities of

quartz,calcite and amorphous silica

(after Blatt et al.,1980,fig.9.7,p.348)

confirmed glacial deposits with this type of feature might actually be showing a relict structure. In general, environmental interpretations should not be based on grain surface features alone, and should be considered along with other lines of evidence.

Calcite is the mext most common form of mineral cement, occurring as a blocky type. It may bind a rock completely, leaving only grains of quartz 'floating' in the cement (fig.54), or it may only partially replace the groundmass material. Dissolution of quartz by calcite has been observed and this means that the pH of the circulating ground waters must have been greater than eight (fig.55).

4.4 Authigenic minerals :

4.4.1 Feldspars :

Albite is a common authigenic mineral in the lithic arenites and graywackes of the sequence. Microprobe analyses of a number of these grains are given in table 2. Some of these authigenic grains can be quite large with a maximum diameter of 60μ. A SEM micrograph of an authigenic feldspar is shown in fig.56. These feldspars are characterised by a very pure end member composition (Ab_{97-100})and show simple Albite twinning. Kastner and Siever (1969) state that overgrowths are the most common form of authigenic feldspar, while discrete grains are relatively uncommon. In this study the reverse was found to be true, although plagioclase overgrowths are occasionally found (fig.57).

Why do authigenic feldspars occur as low temperature phases in

	1	2	3	4	5
SiO_2	68.81	69.06	68.52	67.97	68.94
Al_2O_3	20.14	19.45	19.40	19.08	19.27
Fe_2O_3	0 17	0.00	0.09	1.00	0.02
CaO	0.34	0.10	0.17	0.13	0.10
Na_2O	10.38	11.52	11.59	11.61	11.93
K_2O	0.08	0.06	0.05	0.22	0.04
	99.92	**100.19**	**99.82**	**100.01**	**100.30**

Number of ions on the basis of 32 oxygens

Si	11.97	12.02	11.99	11.92	12.01
Al	4.13	3.99	4.00	3.95	3.96
Fe	0.03	0.00	0.01	0.13	0.01
Ca	0.06	0.02	0.03	0.02	0.02
Na	3.50	3.89	3.93	3.95	4.03
K	0.02	0.01	0.01	0.04	0.01

Molecular Proportions

An	1.68	0.51	0.76	0.50	0 49
Ab	97.76	99.23	98.99	98.50	99.26
Or	0.56	0.26	0.25	1.00	0.25

TABLE 2: Electron microprobe analyses of authigenic albite in the Kommadagga Subgroup.

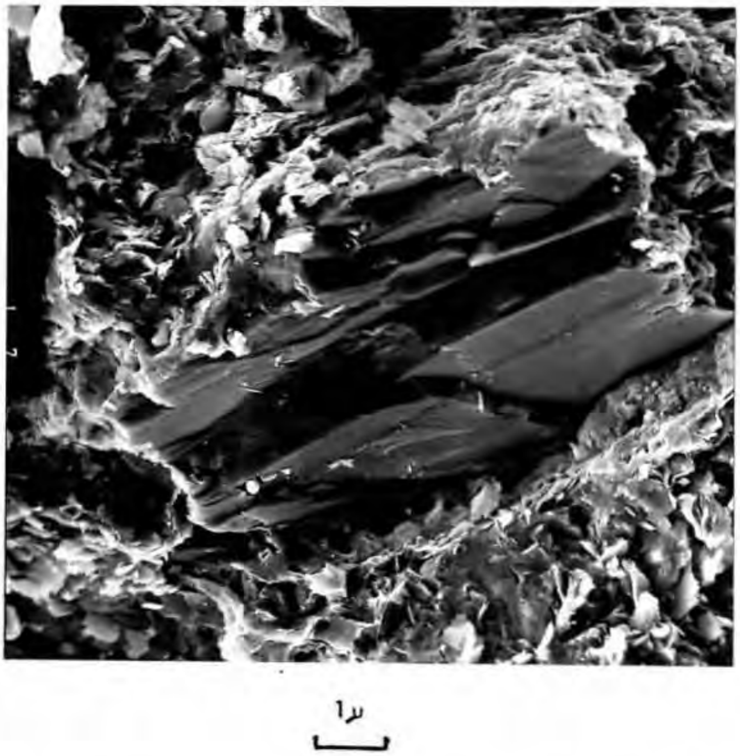


Fig.56: An authigenic feldspar crystal with

well developed crystal faces

(1000*; sample K-39)

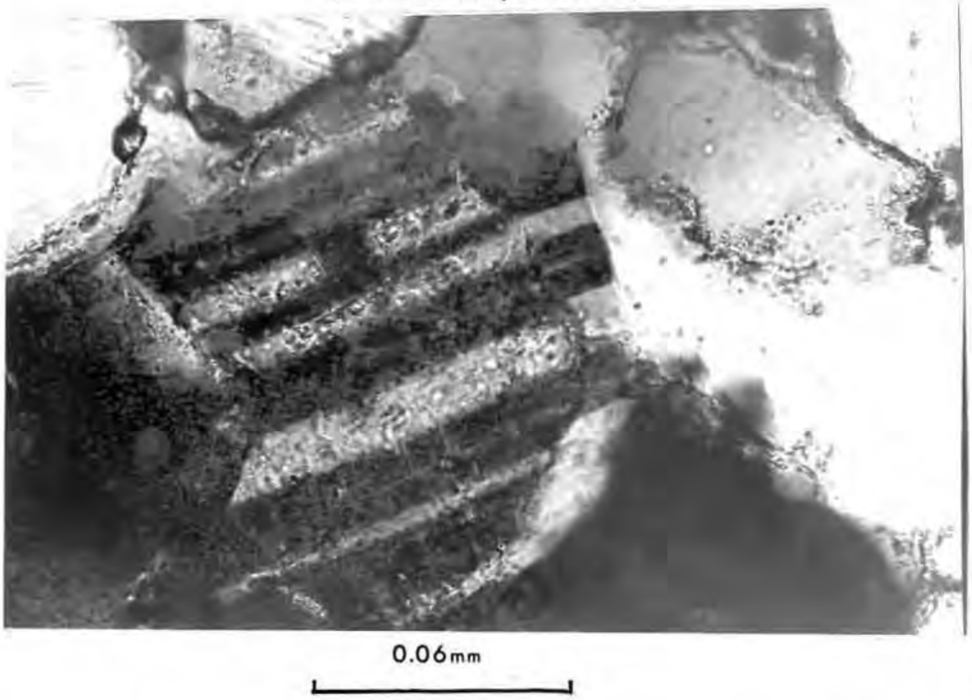


Fig.57: Photomicrograph of a plagioclase overgrowth in optical

continuity with the host grain

(40*; crossed nicols; sample K-20)

sedimentary rocks? A schematic phase diagram for feldspars at temperatures less than 150^{o}C is shown in fig.58,

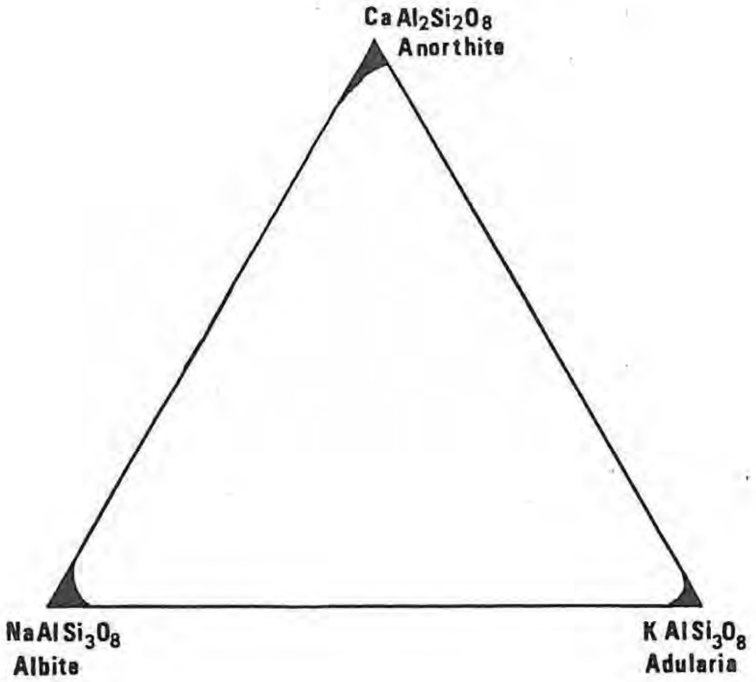


Fig.58: Schematic subsolidus phase diagram showing the stability fields of the feldspar group of minerals at temperatures less than 150^{o}C.
(after Pettijohn et al., 1972, fig.10.3, p.391)

from which it is clear that only the pure end members of the group are stable at these temperatures. The thermodynamic factors controlling low temperature feldspar formation are very poorly understood, but Kastner and Siever (1979) have proposed three isochemical models for the formation of authigemic feldspars in three different geochemical environments. These environments are:

(a) marine carbonate/shale, nonvolcanic

(b) marine, oceanic volcanic and

(c) continental, non-marine volcanic.

Only the first of these models is relevant here, and is summarised in fig.59.

The starting assemblage in this model is a marine sediment with a mixture of detrital clay, silicates from crystalline sources, biogenic carbonate and silica, and organic matter with sea water trapped in this mixture. This assemblage will react with decomposition of organic matter leading to an increase in P_{CO} and a decrease in pH, a change that may or may not be permanent. The biogenic silica will then start to dissolve and this leads to an increase in the concentration of H_4SiO_4. Hurd (1973) found that biogenic opal was the most soluble silica mineral present in sediments. Feldspars are soluble in bottom sea water (fig.60) so all the detrital feldspars dissolve. The production of Na from the dissolution of these feldspars is neglible when compared to the high concentration (0.47M) of this element in sea water. The H_4SiO_4 may be sorbed onto the clay minerals and if the amount of clay is large relative to the quantity of biogenic silica that is dissolving, then the concentration of H_4SiO_4 will not increase sufficiently to permit feldspar formation. If however the amount of clay is not large then the concentration of H_4SiO_4 will increase, and the stability field of albite will be reached, and it will precipitate out of solution. Some clay is neccessary to provide the aluminosilicates required for feldspar formation. A typical reaction leading to the formation of authigenic feldspars is :

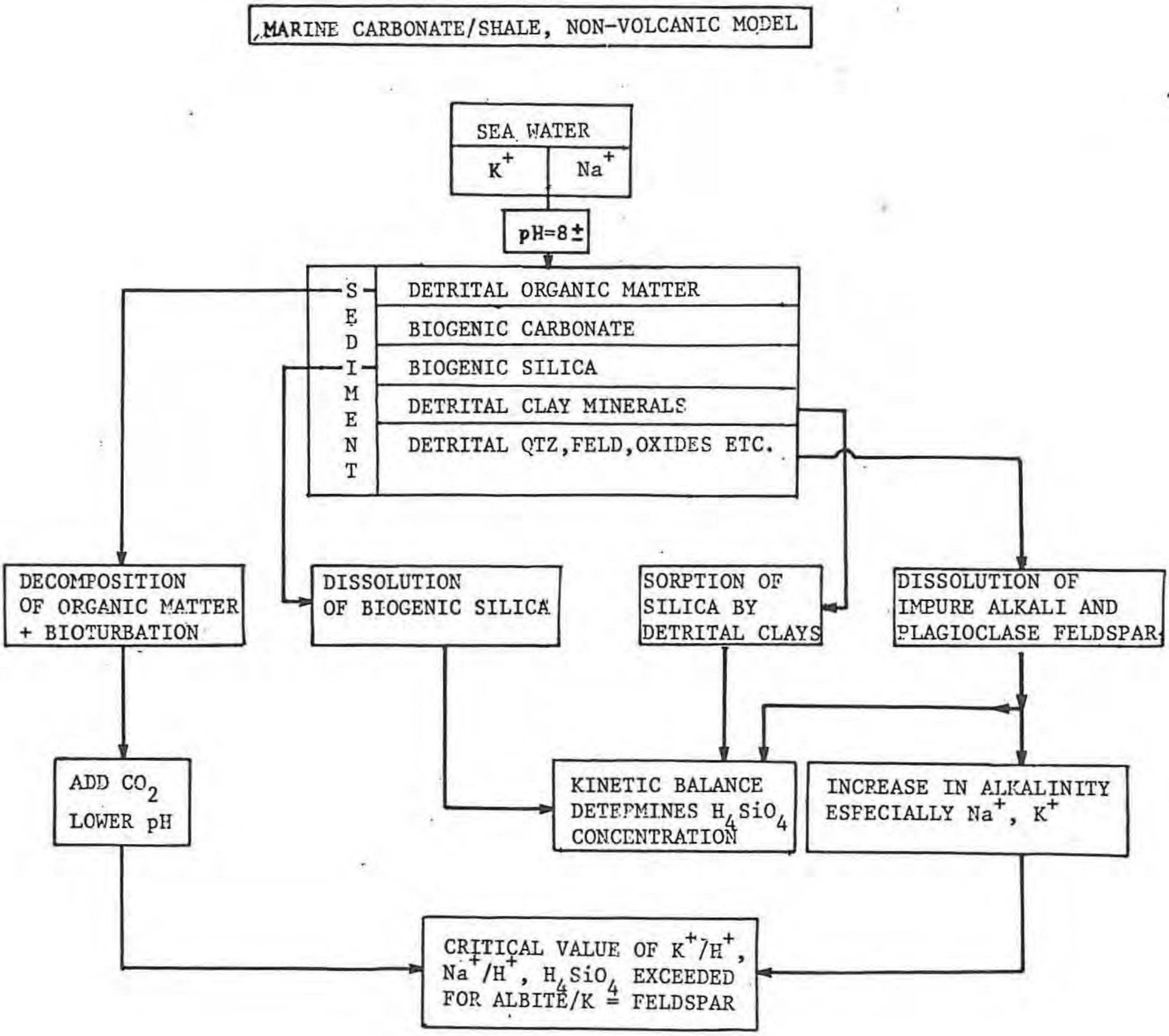


Fig.59: Diagram explaining the main features of the Kastner and Siever model for the low temperature formation of feldspars in a marine carbonate/shale environment (after Kastner and Siever,1979,fig.7,p465)

$$(3NaAlSi_3O_8.2CaAl_2Si_2O_8) + 4H^+ + 2CO_3 + 2H_2O \longrightarrow 3NaAlSi_3O_8 + 2Al_2Si_2O_5(OH)_4 + 2CaCO_3$$

This reaction explains the association of authigenic feldspar and calcite cement seen in some rocks. Heald (1950) described a similar association from the Triassic Newark Series (West Virginia).

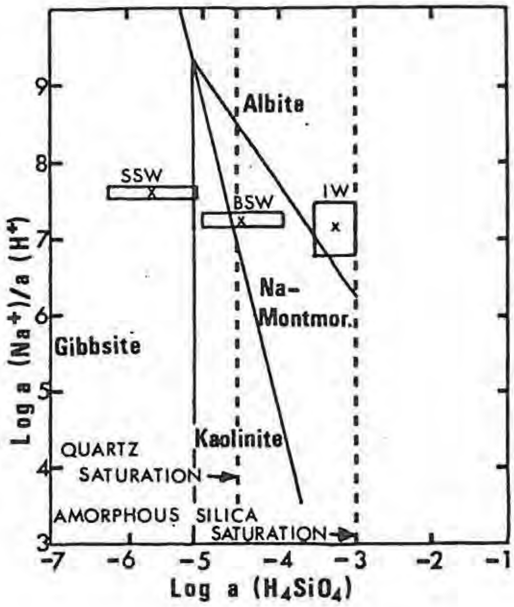


Fig.60: Activity diagram for the system Na_2O - Al_2O_3- SiO_2- H_2O at 0^oC, unit activity of water and latm.

(from Kastner and Siever,1979, fig.7, p.465)

In summary the requirements for the formation of authigenic feldspars are a sediment with an abundance of biogenic silica and some clay. The other models of Kastner and Siever (1979) have volcaniclastic material as the source of silica, and are not applicable to the Upper Witteberg rocks. The concentration of H_4SiO_4 in the fluid phase must also increase sufficiently for albite to become stable.

4.4.2 Prehnite :

Prehnite also occurs as an authigenic mineral (fig.61), forming discrete grains, characterised by moderate to high relief, second order interference colours, tabular form and small grain size (average 30µ). The optical orientation measured on the larger grains is biaxial positive with a moderate 2V. Table 3 gives an electron microprobe analysis of this mineral, the composition of which corresponds closely to the clear variety of prehnite described by Boles and Coombs (1977).

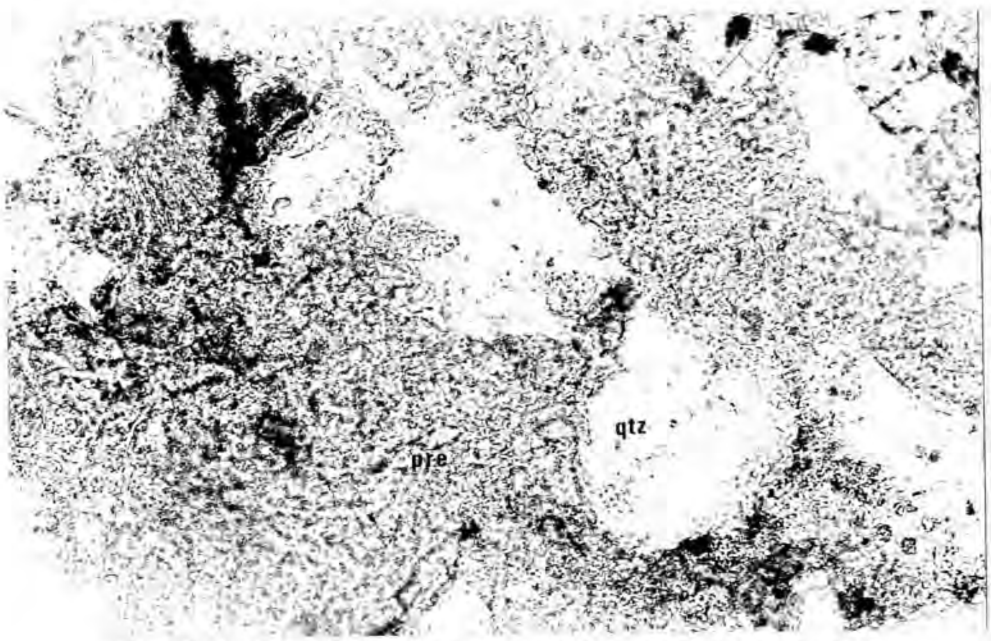


Fig.61: Prehnite from the Dirkskraal Formation.
(40*: plane polarised light; K-87)

The prehnite has only been found in the mottled units of the Dirkskraal Formation, occurring as clusters of grains within each mottle. Each cluster of grains, although apparently consisting of a number of discrete grains, will often extinguish as a whole indicating some sort of crystallographic relationship. Prehnite has been found

	1(a)	1(b)	2	3	4
SiO_2	43.53	47.92	43.3	43.22	43.68
Al_2O_3	23 65	21.81	23.2	24.42	20.05
Fe_2O_3*	0.3	0 17	2.0	0.81	5.52
CaO	26.93	24.08	26.0	27.55	25.11
	94.54	**93.98**	**94.5**	**96.00**	**94.36**

Number of ions on the basis of 22 (O,OH)

	1(a)	1(b)	2	3	4
Si	6.05	6.57	5.28	5.94	6.16
Al	3.88	3.53	3.40	3.95	3.33
Fe	0.04	0.02	5.67	0.08	0.88
Ca	4.01	7.08	2.02	4.05	3.80

(1) Present study.

(2) Surdam, 1969, table 1, analysis 2.

(3) Liou, 1977,table 1, analysis 2(b).

(4) Boles and Coombs, 1977, table 2, analysis 5.

* Determined as FeO, converted to Fe_2O_3 by multiplying by a factor of 1.1113

Table 3: Partial electron microprobe analyses of prehnite.

associated with similar mottles in the Ecca Group at Ecca Pass near Grahamstown (Mitchell,1977).

Prehnite has previously been described from a number of low grade metamorphic environments (Franks 1974; Papezik,1974; Boles and Coombs,1977) where it is commonly associated with pumpellyite. However, no pumpelleyite was observed in any thin section in this study. The chemical composition of the sediments in which most prehnite is found is similar to that of a tholeiitic andesite (Franks,1974). The low K_2O content (1.22%) and high Na_2O/K_2O ratio (4.90) are significantly different to those of an average graywacke ($(K_2O) = 2.00\%$ and $Na_2O/K_2O = 1.45$) given by Pettijohn (1963).

The origin of the prehnite in the graywackes is therefore likely to be different to its origin in the volcanics. It is often found in the same thin section as a calcite cement but never in close association with it, and authigenic albite also occurs with the prehnite. It is therefore possible that a modification of the model given above for the formation of albite will also explain the origin of the prehnite as they are chemically similar. If the P_{CO} is too low then calcite cannot form and if the concentration of H_4SiO_4 does not rise sufficiently for albite precipitation, then instead of forming calcite we could form prehnite by the following reaction :

$$(NaAlSi_3O_8.4CaAl_2Si_2O_8) + 2SiO_2 + 6H_2O \longrightarrow$$
$$NaAlSi_3O_8 + Al_4Si_4O_{10}(OH)_8 + 2Ca_2Al(AlSi_3O_8)(OH)_2$$

The thermodynamic properties of prehnite are incompletely known so it is difficult to be sure if this reaction does occur, but the petrographic evidence suggests that it might.

4.4.3 Stilpnomelane :

Stilpnomelane is a common constituent of many of the quartz arenites and graywackes. It is common in many low grade metamophic rocks throughout the world (Zen,1960; Mather and Atherton,1965; Mather,1970) including metamorphosed iron formations (Klein,1974 ; Miyashiro,1973). It was originally thought that stilpnomelane was only common in high pressure rocks (Miyashiro,1973,p263) but now appears to be common in rocks formed in all pressure regimes (Winkler,1979,p211).

It occurs as two varieties, both varieties of which are found in the study area, one being strongly pleochroic green to colourless (ferrostilpnomelane, rich in Fe^{2+}) while the other is pleochroic brown (ferristilpnomelane, containing Fe^{3+}). Mather and Atherton (1965) reported similar varieties from the Dalradian rocks of Scotland. Zen (1960) proposed that the ferrostilpnomelane is primary and that the ferric variety is a weathering product, a conclusion supported by Brown (1971) and Graham (1976).

The stilpnomelane occurs as randomly orientated sheaves (maximum long diameter 80μ) which are generally interleaved with a white mica, possibly muscovite (fig.62). Some of the stilpnomelane may have occasionally recrystallised into the cleavage (S-2) planes. No evidence for the nucleation of these grains on to K-feldspar cores was observed, although this has been described elsewhere (Mather,1970).

The genesis of this mineral is problematical. An important feature which should be kept in mind when considering its genesis is its close spatial relationship with white mica. The chemical composition of stilpnomelane is also important as it is most common in rocks which are rich in iron. It has been observed in granophyres (Wager and Deer,1939), epidiorites (Wiseman,1974), graywackes (Mather and Atherton,1965 ; Mather,1970), slates (Zen,1960) and in Banded Iron Formation (Klein,1974). It also appears to be stable over a wide range of pressure and temperature (Mather,1970).

0.06mm

Fig.62: Photomicrograph showing the interleaving of stilpnomelane and muscovite. (40*; plane polarised light; sample K-45)

Frey (1973) inferred from microscopic evidence that stilpnomelane forms from glauconite by the reaction :

$$\text{glauconite} + \text{quartz} + \text{chlorite} \longrightarrow \text{stilpnomelane} + \text{K-feldspar} + H_2O + O_2$$

Although it is quite possible that the upper Witteberg sequence once contained glauconite there is no evidence for the presence of K'feldspar now. Also this type of reaction does not explain the close relationship of stilpnomelane and white mica. However if glauconite was present, and it is possible that it was, then the following reaction may explain the genesis of the stilpnomelane :

glauconite + quartz $\longrightarrow$ stilpnomelane + muscovite.

4.4.4 Chlorite :

Chlorite is an uncommon mineral being found only in accessory amounts, and it is possibly a detrital constituent although it is difficult to be sure of this. Alternatively it may have formed diagenetically, and a possible reaction (after Frey in Winkler,1979, p9) is :

disordered mixed layer illite/montmorillonite $\longrightarrow$
Al-rich chlorite + illite + quartz + H_2O.

4.4.5 White Mica (muscovite) :

Muscovite/sericite is a very common constituent of these rocks. It has not crystallised in any preferred orientation, and some of it may be detrital. Diagenetic muscovite can form by a number of reactions, but an interesting relationship is found in the samples containing stilpnomelane. Grains of muscovite, not in contact with the stilpnomelane, may contain numerous tiny inclusions of a red mineral, possibly haematite, which might suggest that some of the stilpnomelane has altered, expelling iron, with the resulting formation of haematite and muscovite.

4.4.6 Clay minerals :

Clay minerals are a common constituent of all of the rock types found in the study area. Identification of all the clay minerals present was not always possible as XRD was beyond the scope of this study, but identification of some minerals was possible on the basis of morphology. It must be remembered however that identification on the basis of morphology alone is dangerous. It is likely that many of these clays are not primary, but are secondary minerals formed as direct precipitates from pore waters or by reaction of pre-existing minerals with pore fluids (Galloway,1974; Whetten and Hawkins, 1970). Wilson and Pittman (1977) have summarised the criteria for distinguishing secondary clays from primary ones. The criteria which are all summarised in table 4, fit into one of five categories, these being:

(a) composition
(b) morphology
(c) structure
(d) texture and
(e) distribution.

Various clay minerals that have been identified under the SEM and in thin section will be described in relation to the criteria of Wilson and Pittman (1977).

Illite is a common clay mineral being found as flakes from which

A: COMPOSITION

(1) Absence of impurities.
(2) Clay assemblage monomineralic.
(3) Significantly different from associated shales.
(4) Concentric colour zoning.

B: MORPHOLOGY

(5) Crystal outlines.
(6) Delicate projections.
(7) Undeformed.
(8) Pseudomorphous replacement.

C: STRUCTURE

(9) Low temperature polytypes.
(10) High degree of crystallinity.

D: TEXTURE

(11) Gap in grain size distribution (no silts).
(12) Large particle size.

E: DISTRIBUTION

(13) Porelinings missing at grain contacts.
(14) Scattered pore fillings.
(15) Fracture fillings.
(16) Absent in early diagenetic concretions.
(17) Cover early formed diagenetic components.
(18) Laminae with abrupt lateral terminations.
(19) Radial alignment of individual flakes.
(20) Medial sutures.
(21) Bridges between detrital grains.

Table 4: Criteria for recognition of authigenic clay (after Wilson and Pittman, 1977, table 3)

delicate needles project. These projections are unlikely to survive transport and must therefore be secondary. In thin section this mineral can be recognised by its interference colours which are higher than those for the common clay minerals (Wilson and Pittman, 1977). It often forms coatings around quartz grains, and because of the higher interference colours this is highlighted under crossed nicols. No work has been done on the illite to determine its crystallinity.

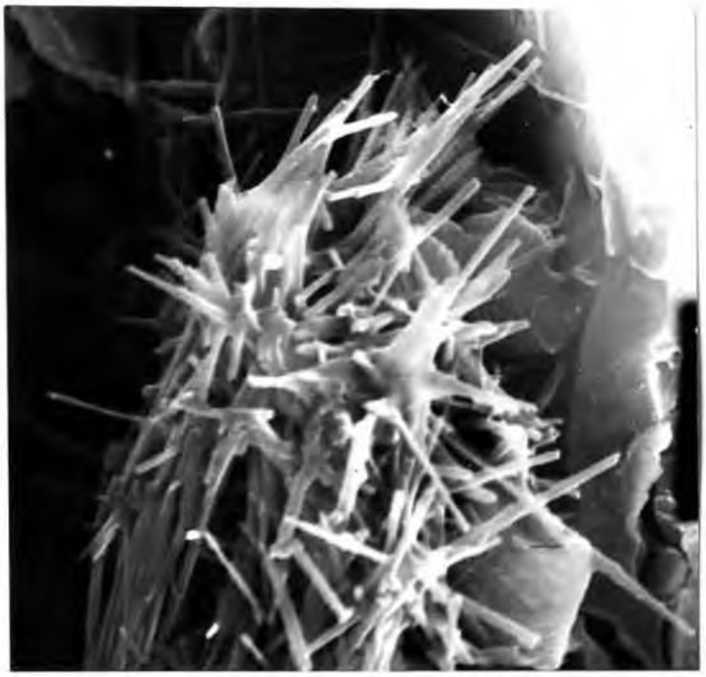
Halloysite, a member of the Kandite group of clay minerals is a common constituent of Waaiport Formation rocks. It is characterised by its hollow, acicular form which is shown in fig.63. The crystal forms this tube because water is taken up in the crystal lattice causing the structure to swell with subsequent curling of the grain (Deer, Howie and Zussman,1962, Vol.3, fig.48, p119).

The most common clay mineral in sedimentary rocks, kaolinite, is not found in its characteristic book-like form but rather as discrete aggregates of grains which are rarely more than 1μ thick or 10μ in diameter (fig.64).

Smectite is also found in a form which resembles the morphology of chlorite (fig.65), but it can be distinguished from chlorite by the fact that no individual crystals can be recognised (Wilson and Pittman, 1977). In thin section it can generally be recognised by a yellow colour in plane polarised light.

4.4.7: Zeolite minerals :

Zeolite minerals are a common fine grained constituent and can only be seen under the SEM. Their maximum grain size is 5μ and identification



1μ

Fig.63: Scanning electron micrograph of halloysite

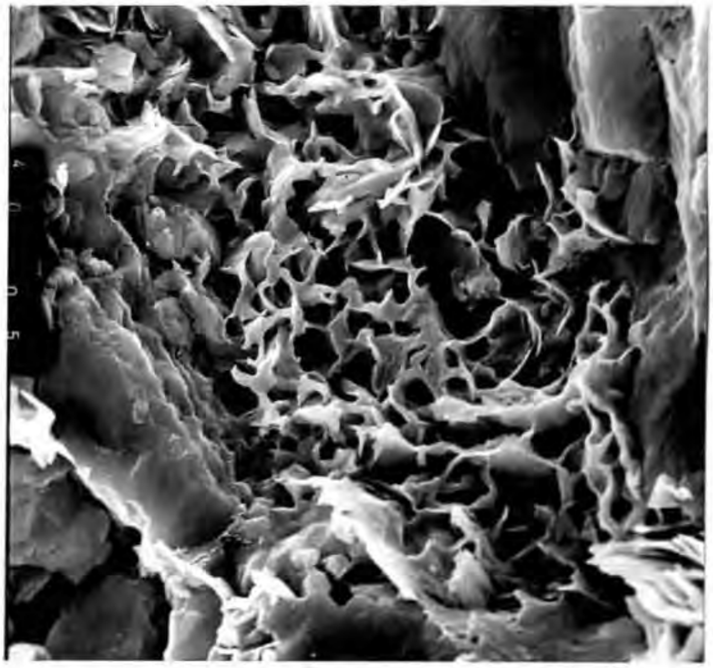
showing its tubular structure

(12000*; sample K-37)

1μ

Fig.64: Discrete grains of kaolinite

(8000*; sample K-42)



1µ

Fig 65: Scanning electron micrograph of smectite

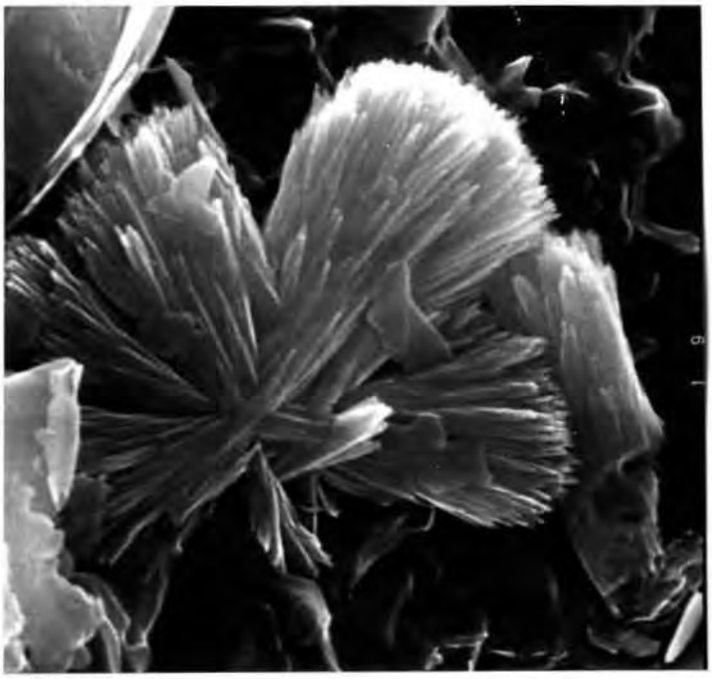
(8000*; sample K-24)

was based mainly on the morphology of these grains. They have been observed in the tuff bands of the Collingham Formation in the lower Ecca Group (fig.66) and in the Waaipoort Formation (fig.67). It is probable that they are more common than previously realised as their small size prevents identification while using the normal petrographic microscope. Gottardi and Obradovec (1978) in a study of sedimentary zeolites in Europe report that the only two zeolites found in sediments with no volcanic component were analcite and clinoptolite. As the zeolites shown in fig.67 come from rocks with no apparent volcanic contribution, it is likely then that they are either analcite or clinoptolite. The latter has a prismatic habit (Scholle,1979,p140) and thus the stellate mineral found during the present study is more likely to be analcite. This identification may be confirmed by further work.

4.5 Pressure and temperature of burial :

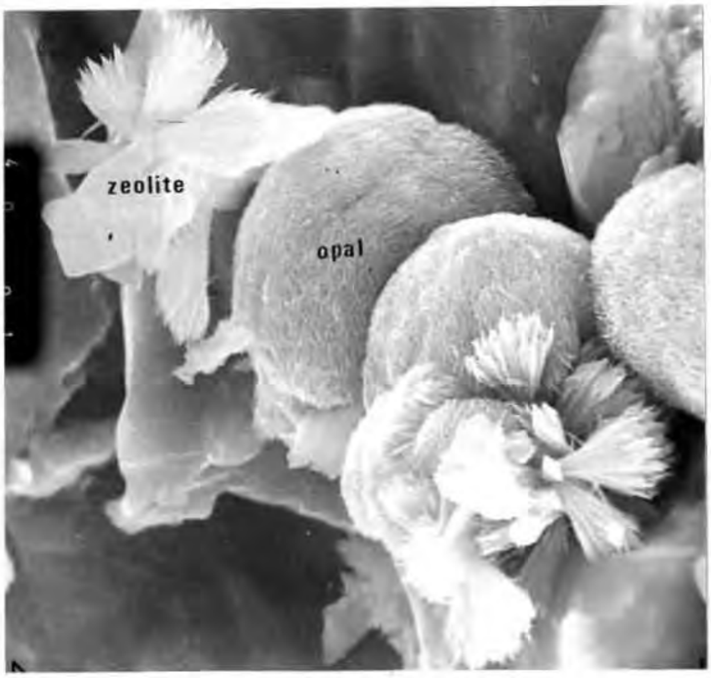
Estimation of the pressure and temperature of burial is difficult because of the fine grain size of the rocks and also because of the incompleteness of the reactions which have occurred. Further, no suitable indicator minerals are found, and deductions which are made rely heavily on the presence of minerals such as prehnite and stilpnomelane.

The stability relationships of prehnite have been summarised by Liou (1971) who found that the high temperature stability limit of prehnite is 400°C at a P_{H_2O} of 2-5kbar. Boles and Coombs (1977) have shown that prehnite can form at temperatures as low as 90 -120 C.Prehnite therefore appears to be stable over a wide range of temperature and cannot be used to give a reliable indication of



1μ

Fig 66: A zeolite (probably analcite) from tuff bands in the Collingham Formation, Ecca Group (1500*; sample E-2)



1μ

Fig 67: A zeolite (probably analcite) from the Waaipoort Formation (1500*; sample K-24) temperature. Winkler (1979) points out that prehnite is not a good indicator of pressure.

The coexistence of stilpnomelane and muscovite indicates that we are on the low temperature side of the reaction :

stilpnomelane + muscovite $\longrightarrow$ bi + chl + quartz + H_2O

(Winkler, 1979, p.212, reaction 3).

This means that according to fig.68 the maximum temperature during burial must have been 400 C. This reaction is however a very poor indicator of pressure (see fig.68). Unfortunately the low temperature stability limits of stilpnomelane are still incompletely known.

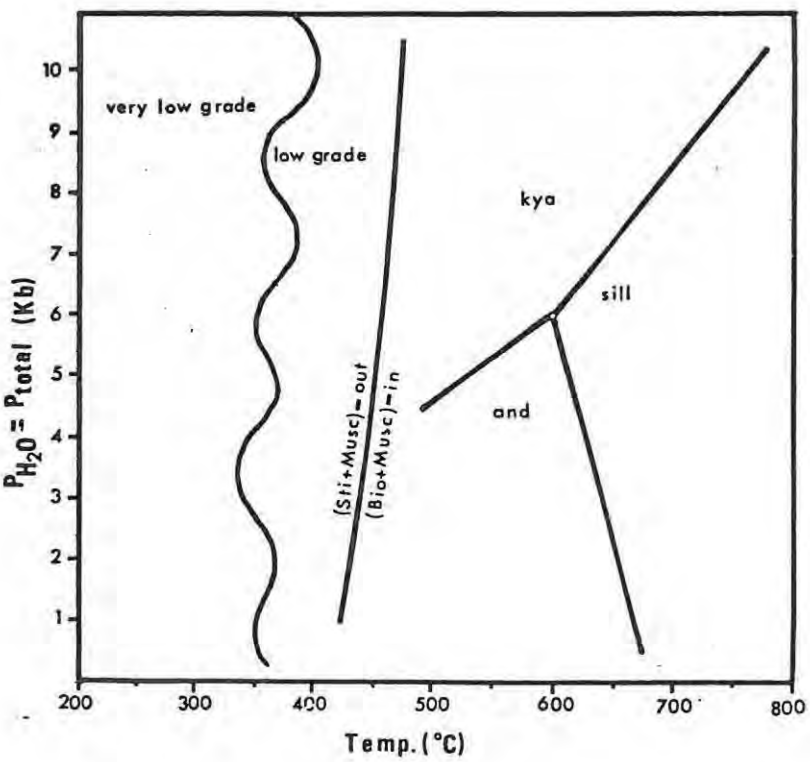


Fig.68: Diagram illustrating the temperature and pressure at which stilpnomelane and muscovite react to form biotite and chlorite.

The zeolite minerals are not good indicators of temperature and pressure being common in sedimentary rocks. Laumontite, the only true metamorphic zeolite, does not occur at all, although this could be a function of composition, and not temperature.

Halbich (1981) reports that the intensity of metamorphism increases southwards in the Cape Fold Belt with increasing depth of burial and intensity of deformation. However, it has been shown that tectonic overpressure plays an insignificant role in metamorphism (Ernst,1971; Brace et al.,1970 ; Heard and Raleigh,1972). The depth of burial is difficult to estimate from sedimentological evidence as the Karoo Basin wedges out southwards, making calculation of the thickness of the cover rocks of the Cape Supergroup very difficult.

The highest grade of metamorphism found by Halbich and co-workers is lower greenschist facies with a pressure of 2,5kb and a temperature of 350°C (Halbich,1981,p.6). These conditions were achieved during the developmemt of a second cleavage, 260Ma. However, metamorphic grades never appear to have reached this stage in the study area. The lack of recrystallisation and the dominance of sedimentary minerals and textures would suggest that the temperature never rose significantly over 100-200°C, the temperature which, according to Winkler (1979), marks the onset of metamorphism. This is despite the presence of prehnite which Winkler regards as being diagnostic of very low grade metamorphism, and it is only the coexistence of prehnite with minerals such as pumpelleyite that is important.

From the preceeding discussion it is clear that the composition of these rocks prevents us from obtaining a clear idea of the pressures and temperatures to which they have been subjected. Work on the crystallinity of illite would help to unveil the mystery. The evidence found so far suggests that the maximum temperature was 150°C, and if we assume a 'normal' geothermal gradient of 25°C/km, then the maximum depth of burial was about 4-5km. These conclusions are still highly speculative.

CHAPTER FIVE

Synthesis

The final stages of the infilling of the Cape Basin appear to have been complex, involving a number of phases of transgression and regression (Rust, 1973). These are reflected in the patterns of sedimentation as shown in a schematic composite log of the Lake Mentz and Kommadagga Subgroups (fig.69).

Firstly deposition of the Kweekvlei Shale and Floriskraal Formations occurred during a regressive phase (1 in fig.69) and resulted in a coarsening upward sequence of off-shore muds overlain by barrier/beach quartz arenites. This phase was followed by a transgressive period (2 in fig.69) during which the mixed tidal flat/barrier deposits were formed, and which probably represent the reworking of a pre-existing sedimentary complex (compare Vos,1977). This phase was followed by a regressive one (3 in fig.69) during which time the coarsening upward sequence of the Kommadagga Subgroup was deposited. After this regression the Cape Basin was nearly completely closed (Lock, 1978; Rust, 1973) and the sedimentation ceased.

The sedimentation of the final stages of the Cape basin appears to have been confined to the eastern member of a pair of troughs which were separated by a positive topographic feature which was orientated north-south in the area of Willowmore (fig.70). Depositional isopachs of the Witteberg Group indicate that two separate troughs existed (fig.70) and secondly, west of this arch no members of the Kommadagga Subgroup are found. It also appears to have influenced sedimentation of the Dwyka Tillite (Stratten, 1968).

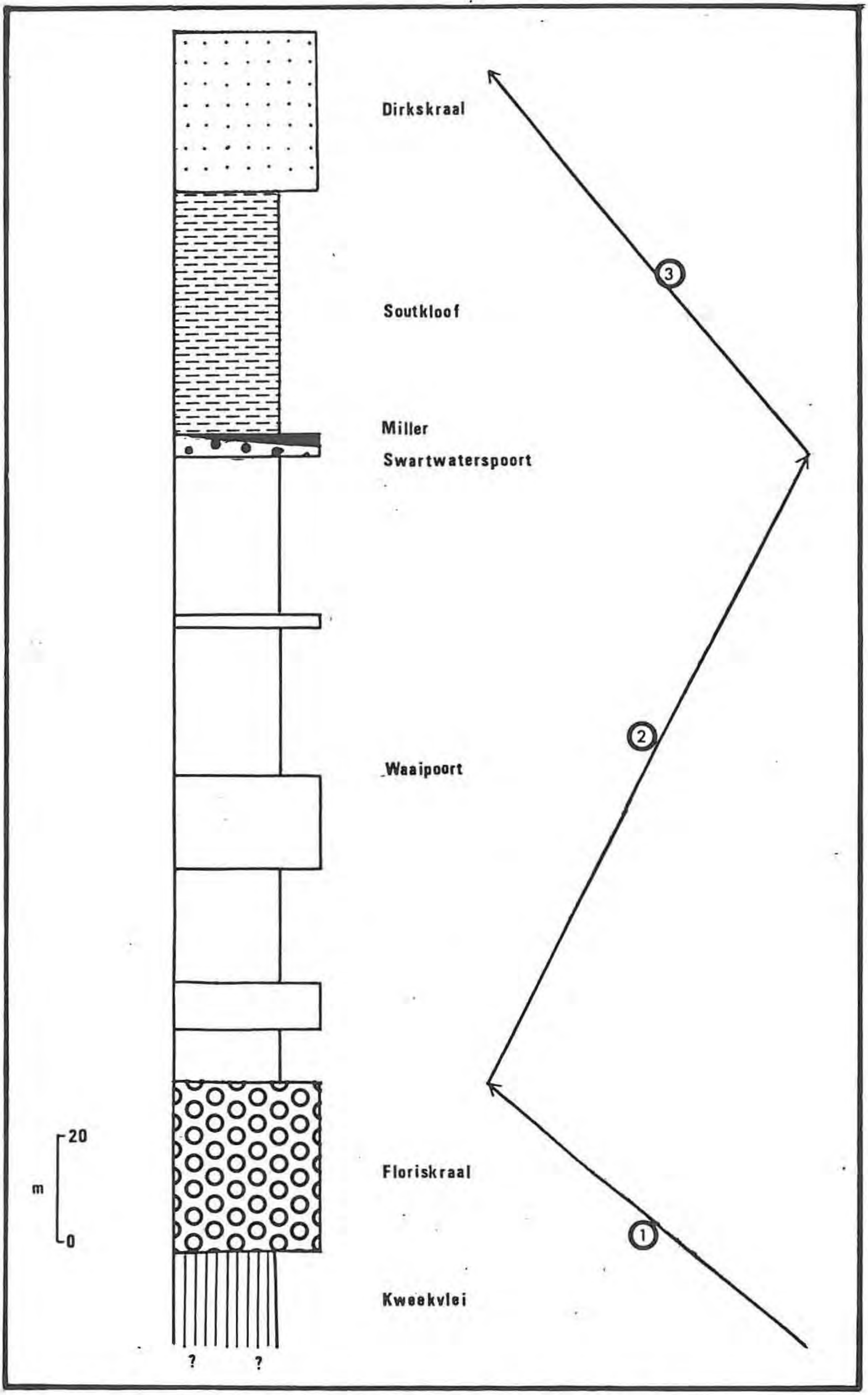


Fig.69: Schematic composite log of the Kommadagga and Lake Mentz Subgroups.

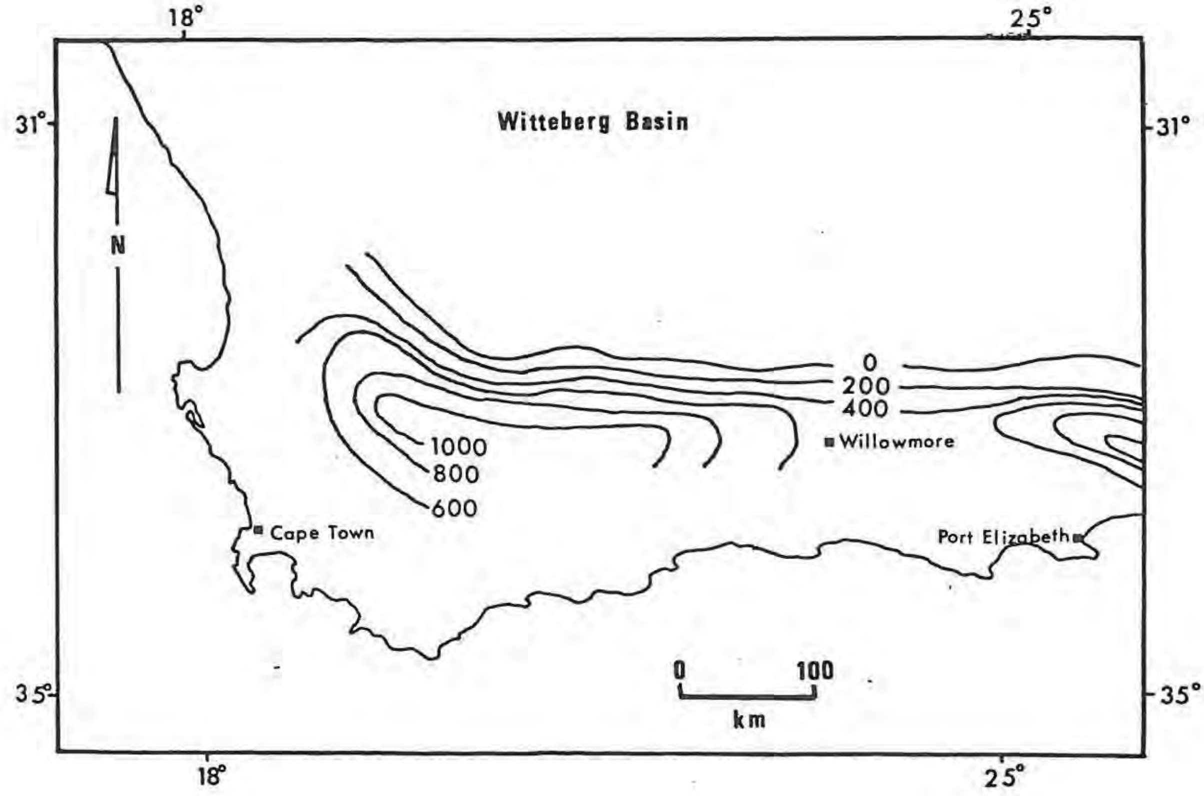


Fig.69: Depositional isopachs of the Witteberg Group demonstrating the existence of two troughs (after Rust, 1973).

Sedimentation of the Dwyka Tillite Formation occurred in a number of pulses which could have been either tectonically or climatically controlled. Each advance and retreat of the ice sheet is reflected in the sedimentary record and can be recognised by the presence of various pro-glacial features interbedded with the main diamictite. In the study area the tillite shows the similar features to marine tills currently forming on the Antarctic shelf and has therefore been interpreted as representing a glacial-marine sequence.

Closure of the basin caused a horizontal space problem which could only be solved by vertical adjustments. Many models have been proposed to explain the origin of the Cape folding ranging from gravity models

(Newton 1973, 1974) and plate tectonic models (Newton,1974, 1980; De Beer et al., 1974; Lock, 1980). Although this study was not sufficiently regional to permit final conclusions to be drawn on this topic, some features are pertinent to this problem. Firstly brittle deformation characterises the study area and secondly the mineralogical evidence suggests that the pressures and temperatures of burial were not very high. Also the deformation in the study area occurred until at least after the deposition of the Lower Ecca strata.

Diagenetic changes have occurred at all stages in the history of these sediments. Quartz overgrowths provide a good example of this feature as some overgrowths show strong deformational features while others are completely undeformed. The formation of opal at a late stage in the history of the Waaipoort Formation sediments also suggests that diagenetic changes occur continuously in a sediment, beginning immediately after deposition of the sediment and continuing throughout the history of that sediment.

References :-

Ali, S.A. and Friedman, G.M., 1977, Diagenesis of sandstone, AAPG reprint series no.20.

Anderson, J.B, Kurtz, D.D., Domack, E.W. and Brookshaw, K.M., 1980, Glacial and glacial marine sediments of the Antarctic continental shelf, Jour. Geology, 88, 399 - 414.

Bell, C.M., 1981, Soft sediment deformation of a sandstone related to the Dwyka glaciation in South Africa, Sedimentology, 28, 321 - 329.

Berner,R.A., 1980, Early diagenesis: A theoretical approach, Princeton University Press, Princeton, pp.241.

Blatt, H., 1979, Diagenetic processes, in Aspects of Diagenesis (Scholle, P.A. and Schluger,P.R., eds.), SEPM Spec. Publ. 26, 141 -157.

Blatt, H. and Christie, J.M., 1963, Undulatory extinction in quartz of igneous and metamorphic rocks and its significance in provenance studies of sedimentary rocks, Jour. sediment. petrol., 33, 559 - 579.

Blatt, H., Middleton, G. and Murray, R., 1972, Origin of Sedimentary Rocks, Prentice-Hall, New Jersey, pp.634.

Blatt, H., Middleton, G. and Murray, R., 1980, Origin of Sedimentary Rocks, Prentice-Hall, New Jersey, pp.782.

Boles, J.R. and Coombs, D.S., 1977, Zeolite facies alteration of sandstone in the Southland syncline, New Zealand, American Jour. Sci., 277, 982-2012.

Brace, W.F., Ernst, W.G. and Kallberg, R.W., 1970, An experimental study of tectonic overpressure in Franciscan rocks, Geol. Soc. Am. Bull., 81, 1325-1338.

Brasier, M.D., 1980, Microfossils, George Allen and Unwin, London, pp.193.

Brenchley, P.J., 1969, Origin of matrix in Ordovician graywackes, Berwyn Hills, North Wales, Jour. sediment. petrol., 39, 1297-1301.

Brown, E.H., 1971, Phase relations of biotite and stilpnomelane in the greenschist facies, Contrib. Mineral. Petrol., 31, 275-299.

Buckley, H.E., 1951, Crystal growth, Wiley, New York, pp.571.

Canadian Mineralogist, 1974, Low grade metamorphism, Canadian Mineralogist, 12, 437-543.

Clarkson, E.N.K., 1979, Invertebrate palaeontology and evolution, George Allen and Unwin, pp.323.

Coleman, J.M., 1976, Deltas: Processes of deposition and models for exploration, Continuing Education Publishing Company, Champaign, pp.102.

Coombs, D. S., 1961, Some recent work on the lower grades of metamorphism, Australian Jour. Sci., 24, 203-215.

Crowell, J.C., 1957, Origin of pebbly mudstones, Geol. Soc. Am. Bull., 68, 993-1110

Davis, R.A., 1978, Beach and Nearshore Zone, in Coastal Sedimentary Environments (Davis, R.A.(ed.)), Springer Verlag, New York, 237-286.

De Beer, J.H., van Zijl, J.S.V. and Bahnemann, 1974, Plate tectonic origin for the Cape Fold Belt, Nature, 252, 675-676.

Deer, W.A., Howie, R.A. and Zussman, J., 1963, Rock Forming Minerals, Vol.3, Longmans, London.

De Swardt, A.M.J., Fletcher,O. and Toschek, P.,1974, Note on orogenic style in the Cape Fold Belt, Trans. geol. Soc. South Africa., 77, 53-57.

De Swardt, A.M.J. and Rowsell, D.M., 1974, Note on the relationship between diagenesis and deformation in the Cape Fold Belt, Trans. geol. Soc. South Africa., 77, 239-245.

De Villiers, J., 1944, A review of the Cape Orogeny, Ann. Univ. Stell., 22-A,183-208.

Dickinson, W.R., 1970, Interpreting detrital modes of graywacke and arkose, Jour. sediment. petrol., 40, 695-707.

Dunlevey, J.N. and Hiller, N., 1979, The Witteberg-Dwyka boundary in the south-western Cape, Trans. geol. Soc. S.Afr., 82, 251-256.

Durney, D.W., 1972, Solution transfer, an important geological deformation mechanism, Nature, 235, 315-317.

Durney, D.W., 1976, Pressure solution and crystallisation deformation in A Discussion on Natural Strain and Geological Structure (Ramsay, J.G. and Wood,.,eds), Phil. Trans. Royal Soc. of London, A283, 229-240.

Durney, D.W.,1978, Early theories and Hypotheses on pressure-solution - redeposition, Geology, 6, 369-372.

Du Toit, A.L., 1921, The Carboniferous unconformity in southern Africa, Trans. geol. Soc. S. Afr., 24, 188-227.

Du toit, A.L., 1926, The geology of South Africa, Oliver and Boyd, London

Edwards, M.B. 1978, Glacial environments, in Sedimentory environments and facies (Reading H.G., ed.), pp 416-418, Elsevier, New York.

Elliott, T., 1978, Deltas, in Sedimentary Environments and Facies (Reading, H.G., ed.), Elsevier, New York, 97-142.

Ernst, W.G., 1971, Do mineral parageneses reflect unusually high pressure conditions of Franciscan metamorphism?, Amer. Jour. Sci., 270, 81-108.

Folk, R.L., 1965, The petrology of the Sedimentary Rocks, Hemphills, Austin, pp.159

Force, E.R., 1980, The provenance of rutile, Jour. sediment. Petrol., 50, 485-488.

Force, E.R., 1976, Metamorphic source rocks of titanium placer deposits - a geochemical cycle, U.S. geol. Surv. Prof. Paper 959-B, pp.16.

Franks, S.G., 1974, Prehnite - pumpellyite facies metamorphism of the New Bay Formation, Exploits zone, Newfoundland, Canadian Mineralogist, 12, 456-462.

Frey, M and Hunziker, J.C., 1973, Progressive niedriggradige metamorphose glauonitfuhrender horizonete in den Helvitischen Alpen der Ostchweiz, Cont. Min. Petrol., 39, 185-210.

Fyfe, W.S., 1974, Low grade metamorphism: some thoughts on the present situation, Canadian Mineralogist, 12, 439-444.

Fyfe, W.S., Price, N.J. and Thompson, A.B., 1978, Fluids in the earths crust, Elsevier, Amsterdam, pp.383.

Galloway, W.E., 1974, Deposition and diagenetic alteration of sandstone in north east Pacific arc-related basins: Implications for graywacke genesis, Geol. Soc. Am. Bull., 85, 379-380.

Gardiner, B.G., 1969, New palaeoniscoid fish from the Witteberg Series of South Africa, J. Linn. Soc. London., 48, 423-452.

Gay, N.C., 1970, The formation of step structures on slickensided shear surfaces, Jour. Geol., 78, 523-532.

Geiser, P.A., 1974, Cleavage in some sedimentary rocks of the Central Valley and Ridge Province, Geol. Soc. Am. Bull., 85, 1399-1412.

Ghent, F.D. and Miller, B.E., 1974-a, Zeolite and clay-carbonate assemblages in the Blairmore Group (Cretaceous), northwestern Alberta, Canadian Mineralogist, 12, 542.

Ghent, E.D. and Miller, B.E., 1974-b., Zeolite and clay-carbonate assemblages in the Blairmore Group (Cretaceous) southern Alberta Foothills, Contr. Mineral. Petrol., 44, 313-329.

Ginsberg, R.N., Tidal deposits, Springer-Verlag, New York, pp.428.

Gottardi, G. and Obradovic, J., 1978, Sedimentary zeolites in Europe, Fortsch. Min., 56, 316-366.

Graham, C.M., 1976, A note on some Dalradian stilpnomelane and their oxidation, Mineral. Magazine, 40, 467-472.

Gray, D.R., 1978, Cleavages in deformed psammitic rocks from southeastern Australia: their nature and origin, Geol. Soc. Am. Bull., 89, 577-590.

Griffiths, J.C., 1967, Scientific methods in the analysis of sediments, McGraw-Hill, New York, pp.508.

Hallbauer, D.K., 1975, The plant origin of the Witwatersrand carbon, Minerals Sci. Engng., 7, 111-131.

Halbich, I.W., 1977, Fold profiles and tectonic shortening in the Cape Fold Belt, Trans. geol. Soc. S. Afr., 80, 253-265.

Halbich, I.W., 1979, Disharmonic folding, detachment and thrusting in the Cape Fold Belt, Geokongress '79, Abstracts, Geol. Soc. S. Afr., 167-169.

Halbich, I.W., 1981, Cape Fold Belt, Excursion guide, Geocongress '81, Geol. Soc. S.Afr., 48pp.

Haughton, S.H., 1969, Geological History of Southern Africa, Geol. Soc. S. Afr., Cape Town, pp.535.

Hawley, N., 1981, Flume experiments on the origin of flaser bedding, Sedimentology, 28, 699-712

Heald, M.T., 1950 Authigenesis in West virginia sandstone, Jour. Geol., 58, 624-633.

Heard, H.C. and Raleigh, C.B., 1972, Steady state flow in marble at 500 to 800 C, Geol. Soc. Am. Bull.,77, 741-760.

Heinrich, E.W., 1965, Microscope identification of minerals, McGraw-Hill, New York, pp.414.

Hiller, N and Snowden, P.A., 1981,Structural and stratigraphic relationships in the Cape Fold Belt south of Steytlerville, Geocongress '81, South African Geodynamics Project, Abstracts, 111-112

Hobbs B.E, Means, W.D. and Williams, P.F., 1976, An Outline of Structural Geology, Wiley, London, pp.571.

Hurd, D.C., 1973, Interaction of biogenic opal, sediment and sea water in the central equatorial Pacific, Geochem. Cosmochim. Acta, 37, 2257-2282.

Johnson, M.R., 1966, The stratigraphy of the Cape and Karoo Systems in the Eastern Cape Province, Unpubl. M.Sc. thesis, Rhodes University.

Johnson, M.R., 1976, Stratigraphy and sedimentology of the Cape and Karoo sequences in the Eastern Cape Province, Unpubl. Ph.D. thesis, Rhodes University.

Journal Geological Society of London, 1978, Diagenesis, 135, Part 1, 1-156.

Kastner, M. and Siever, R., 1979, Low temperature feldspars in sedimentary rocks, American Jour. Sci., 279, 435-479.

Klein, C., 1974, Greenalite, stilpnomelane, minnesotaite, crocidolite and carbonates in a very low grade metamorphic Precambrian iron formation, Canadian Mineralogist, 12, 475-498.

Kraft, J.C., 1971, Sedimentary facies patterns and geologic history of a Holocene marine transgression, Geol. Soc. Am. Bull., 82, 2131-2158.

Larsen, G. and Chilingar, G.V., 1979, Diagenesis in sediments and sedimentary rocks, Developments in Sedimentology, 25A, Elsevier, Amsterdam, pp.579.

Liou, J.G., 1971, Synthesis and stability relationships of prehnite, (Ca Al Si O (OH)), Am. Mineral., 56, 507-531.

Lock, B.E., 1978, The Cape Fold Belt of South Africa; tectonic control of sedimentation, Proc. Geol. Assoc., 89, 263-281.

Lock, B.E., 1980, Flat plate subduction and the Cape Fold Belt of South Africa, Geology, 8, 35-39.

Loock, J.C., 1967, The stratigraphy of the Witteberg-Dwyka contact beds, Unpubl. M.Sc. thesis, University of Stellenbosch.

Marais, J.A.H., 1973, Fossil fish from the Upper Witteberg beds near Lake Mentz, Jansenville district, Cape Province, Annals geol. Surv. South Africa, 2, 191-202.

Margolis, S.V. and Krinsley, D.H., 1974, Processes of formation and environmental occurrence of microfeatures on detrital quartz grains, Amer. Jour. Sci., 274, 449-464.

Mather, A.D., 1970, The biotite isograd and the lower greenschist facies in the Dalradian rocks of Scotland, Jour. Petrol., 11, 253-275.

Mather, A.D. and Atherton, M.P., 1965, Stilpnomelane from the Dalradian, Nature, 207, 971-972.

Means, W.D., 1975, Natural and experimental microstructures in deformed micaceous sandstone, Geol. Soc. Amer. Bull., 86, 1221-1229.

Meyer, W., 1965, The Geology of a portion of south-western Albany, Unpubl. M.Sc. thesis, Rhodes University.

Mitchell, A.A., 1977, Burial metamorphism in the Ecca Pass Formation, Unpubl. B.Sc. (Hons) project, Rhodes University.

Miyashiro, A., 1973, Metamorphism and metamorphic belts, George Allen and Unwin, London, 492pp.

Mountain, E.D., 1945, Problems of Eastern Province geology, Proceedings of the Geol. Soc. of S.Afr., XLVIII, pXXI -XXXVII.

Newton, A.R., 1973, A gravity folding model for the Cape Fold Belt, Trans. geol. Soc. S.Afr., 76, 145-152.

Newton, A.R., 1974, Nature of South Africa's Cape Fold Belt, Nature, 248, 499-500.

Newton, A.R., 1980, Intra-plate nature of the Cape Fold Belt, S. Afr. Jour. Sci., 76, 272-273.

Onesta P.A. and Verhoef,P., 1976, Annual rainfall distribution for 80 rainfall districts in South Africa, S. Afr. Journ. Sci., 72, 120-122.

Papezik, V.S., 1974, Prehnite-pumpelleyite facies metamorphism of Late Precambrian rocks of the Avalon Peninsula, Newfoundland, Canadian Mineralogist, 12, 463-468.

Park, C.F. and MacDiarmid, R.A., 1975, Ore deposits, W.H.Freeman, San Francisco, pp.529.

Paterson, M.S., 1958, Experimental deformation and faulting in Wombeyan marble, Geol. Soc. Am. Bull., 69, 465-476.

Pettijohn, F.J., 1963, Chemical composition of sandstone excluding carbonate and volcanic sands, in Data of Geochemistry, 6th edition, U.S. geol. Surv. Prof. paper 440-S, pp.19.

Pettijohn, F.J., 1975, Sedimentary Rocks, Harper and Row, New York, pp.628.

Pettijohn, F.J., Potter, P.E. and Siever, R., 1972, Sand and Sandstone, Springer Verlag, New York, pp.618.

Pittman, E.D., 1979, Recent advances in sandstone diagenesis, Ann. Rev. Earth Planet. Sci., 7,39-62

Plumstead, E.P., 1967, A general review of the Devonian fossil plants found in the Cape System of South Africa, Palaeont. Afr., 10, 1-83

Rao, C.P., 1980, Criteria for recognition of cold water carbonate sedimentation: Berriedale Limestone (Lower Permian) Tasmania, Australia, Jour. Sedim. Petrol., 51, 491-506.

Reineck, H.E., 1972, Tidal flats, in Recognition of ancient sedimentary environments, (Rigby, J.K. and Hamblin, W.K. eds.) SEPM Special publication 16, 146-159.

Reineck, H.E. and Singh, I.B., 1973, Depositional sedimentary environments, Springer-Verlag, Amsterdam, pp.439.

Renton, J.J., Heald, M.T. and Cecil, C.B., 1969, Experimental investigation of pressure solution of quartz, Jour. sedim. petrol., 39, 1017-1117.

Rossouw, P.J., 1953, The Southern Karroo, in Haughton,S.H., Blignault, J.J.G., Rossouw, P.J., Spies, J.J. and Zagt. S., Results of an investigation into the possible presence of oil in Karroo rocks in parts of the Union of South Africa, Mem. geol. surv. S. Afr, 45, 14-36.

Rust, I.C., 1973, The Evolution of the Palaeozoic Cape Basin, southern margin of Africa in The Ocean Basins and Margins, Vol.1, (Nairn,A.E.M. and Stehli, F.G.), The Ocean Basins and Margins, Plenum Publ. Corp., New York, 247-276.

Scholle, P.A., 1979, A colour illustrated guide to constituents, textures, cements and porosities of sandstones and associated rocks, AAPG memoir, 28, Tulsa, Oklahoma, pp.201.

Scholle, P.A. and Schluger, P.R., 1979, Aspects of Diagenesis, SEPM special publication, 26, pp.443.

Sibley, D.F. and Blatt, H., 1976, Intergranular pressure solution and cementation of the Tuscarora orthoquartzite, Jour. sedim. petrol., 46, 881-896.

Siever, R., 1959, Petrology and geochemistry of silica cementation in some Pennsylvanian sandstones in Silica in sediments (Ireland,H.A. ed.) SEPM spec. publ. 7, 55-79.

Sohnge, A.P.G. and Halbich, I.W., in prep., Geodynamics of the Cape Fold Belt, Geol. soc. S. Africa.

South African Committee for Stratigraphy (SACS), 1980, Stratigraphy of South Africa, pt.1 (Comp. L.E. Kent), Lithostratigraphy of the Republic of South Africa, S.W.A./Nambia, and the Republics of Bophuthatswana, Transkei and Venda, Handbook geol. Surv. S. Afr., 8.

Stanley, K.O. and Benson, L.V., 1979, Early diagenesis of High Plains Tertiary vitric and arkosic sandstone, Wyoming and Nebraska, in Aspects of Diagenesis (Scholle, P.A. and Schluger, P.R., eds.,) SEPM Spec. publ. 26, 401-423.

Stapleton, R.P., 1977, Carboniferous unconformity in southern Africa, Nature, 268, 222-223.

Stratten, T., 1968, The Dwyka glaciation and its relationship to the pre-Karroo surface, Unpubl. Doctoral Dissertation, University of the Witwatersrand.

Surdam, R.C., 1969, Electron microprobe study of prehnite and pumpelleyite from the Karmutten Group, Vancouver, British Columbia, Geol. Soc. Amer. Bull., 84, 1911-1922

Theron, J.N., 1962-a, An analysis of the Cape folding in the district of Willowmore, Cape Province, Annals of the University of Stellenbosch, 37, 347-421.

Theron, J.N., 1962(b), The occurrence of fish remains in the Witteberg-Dwyka transition zone, Annals geol. Surv. S.Afr., 1, 263-266

Theron, J.N, 1969, The Baviaanskloof Range -a South African nappe, Trans. geol. Soc. S. Afr., 72, 29-30.

Theron J.N. and Blignault, H.J., 1975, A model for the sedimentation of the Dwyka glacials in the south western Cape, in Proc. 3rd Gondwana Symposium, 347-356.

Thomson, A., 1959, Pressure solution and porosity in Silica in sediments (Ireland, H.A. ed.) SEPM spec. publ. 7, 92-110

Tjia, H.D., 1964, Slickensides and fault movements, Geol. Soc. Am. Bull., 75, 683-686.

Vos, R.G., 1977, Sedimentology of an Upper Palaeozoic river, wave and tide influenced system in southern Morrocco, Journ. sedim. petrol., 47, 1242-1260.

Wager, L.R. and Deer, W.A., 1939, The petrology of the Skaergaard intrusion, Kangerderlugssuaq, East Greenland, Meddr. Gronland, 105, no.4.

Weyl, P.K., 1959, Pressure solution and the force of crystallisation, - a phenomenological theory, J. Geophys. Res., 64, 2001-2025.

Whetten, J.T. and Hawkins, J.W., 1970, Diagenetic origin of graywacke matrix minerals, Sedimentology, 15, 347-361.

Whinsonant, R.C., 1970, Influence of texture upon the response of detrital quartz to deformation of sandstone, Journ. sedim. petrol., 40, 1018-1025.

Williams, P.F., 1972, Development of metamorphic layering and cleavage in low grade metamorphic rocks at Bermagui, Australia, American Jour. Sci., 272, 1-47.

Wilson, M.D. and Pittman, E.D., 1977, Authigenic clays in sandstone: Recognition and influence on reservoir properties and environmental analysis, Journ. sedim. petrol., 47, 3-31.

Wilson, G., 1961, The tectonic significance of small scale structures, and their importance to the geologist in the field, Ann. Soc. Geologique de Belgique, 84, 423-548.

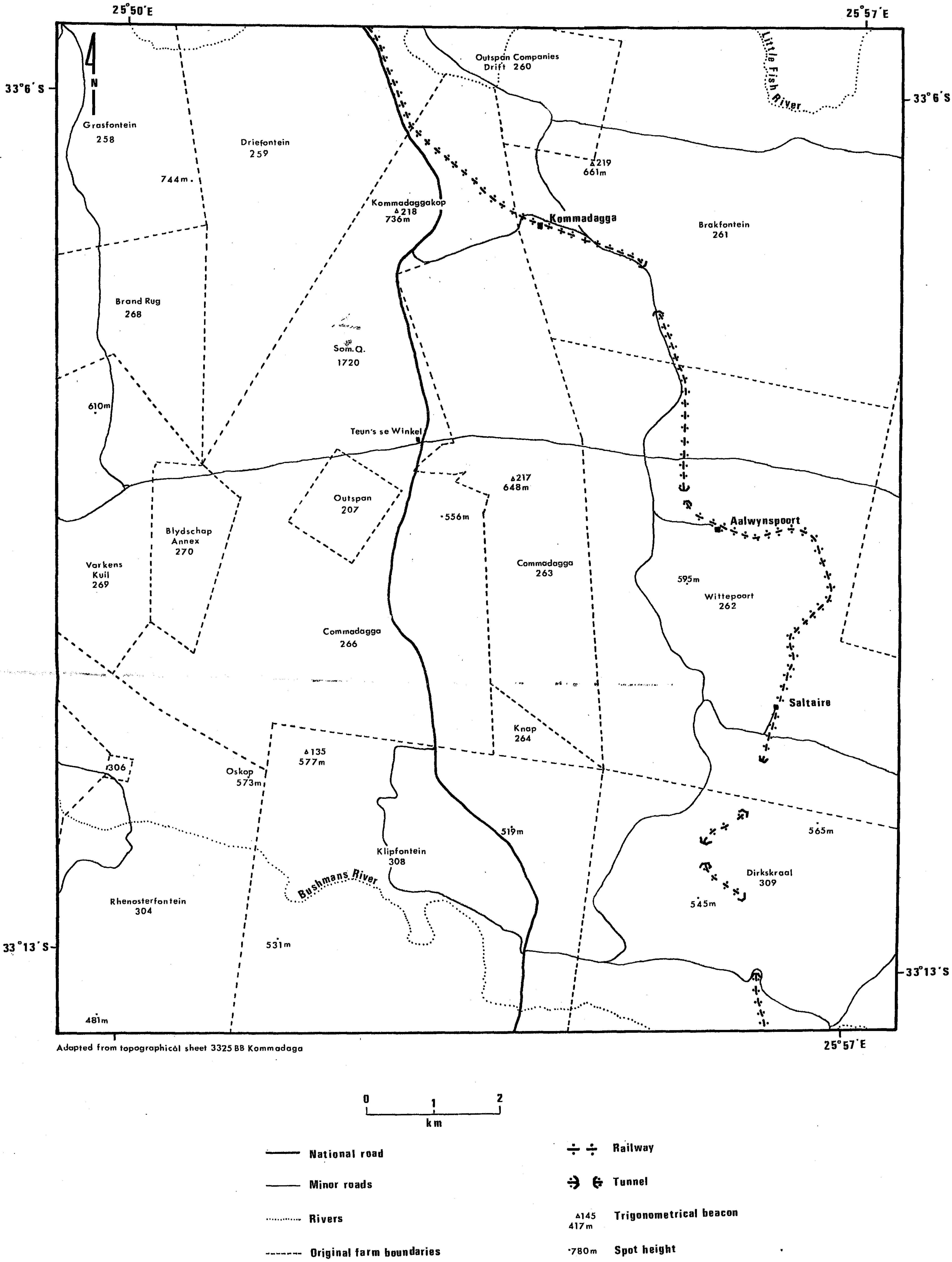
Winkler, H.G.F., 1979, Petrogenesis of metamorphic rocks, Springer Verlag, New York, pp.348.

Wiseman, J.D.H., 1934, The central and south western Highland epidiorites: a study in prograde metamorphism, Quart. Journ. geol. Soc. London, 90, 354-417.

Wright, L.D., 1978, River Deltas in Coastal Sedimentary Environments (Davis, R.A., ed.), Springer Verlag, New York, 5-68.

Wright, L.D. and Coleman, J.M., 1974, Mississippi River mouth processes: Effluent dynamics and morphologic development, Jour. Geol., 82, 751-778.

Zen, E-An, 1974, Burial Metamorphism, Canadian Mineralogist, 12, 445 - 455.



Adapted from topographical sheet 3325 BB Kommadaga

Topo-cadastral features of study area

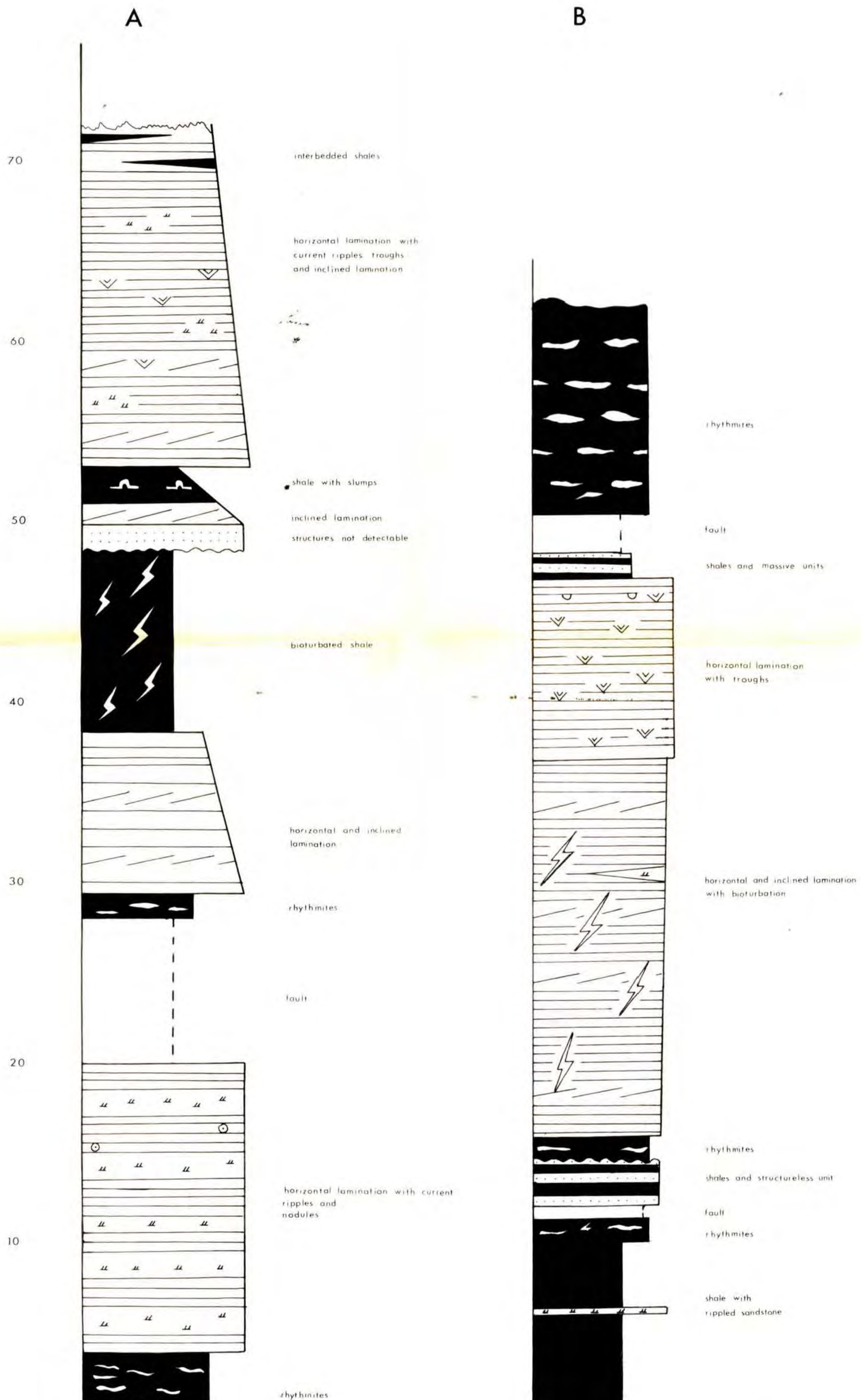
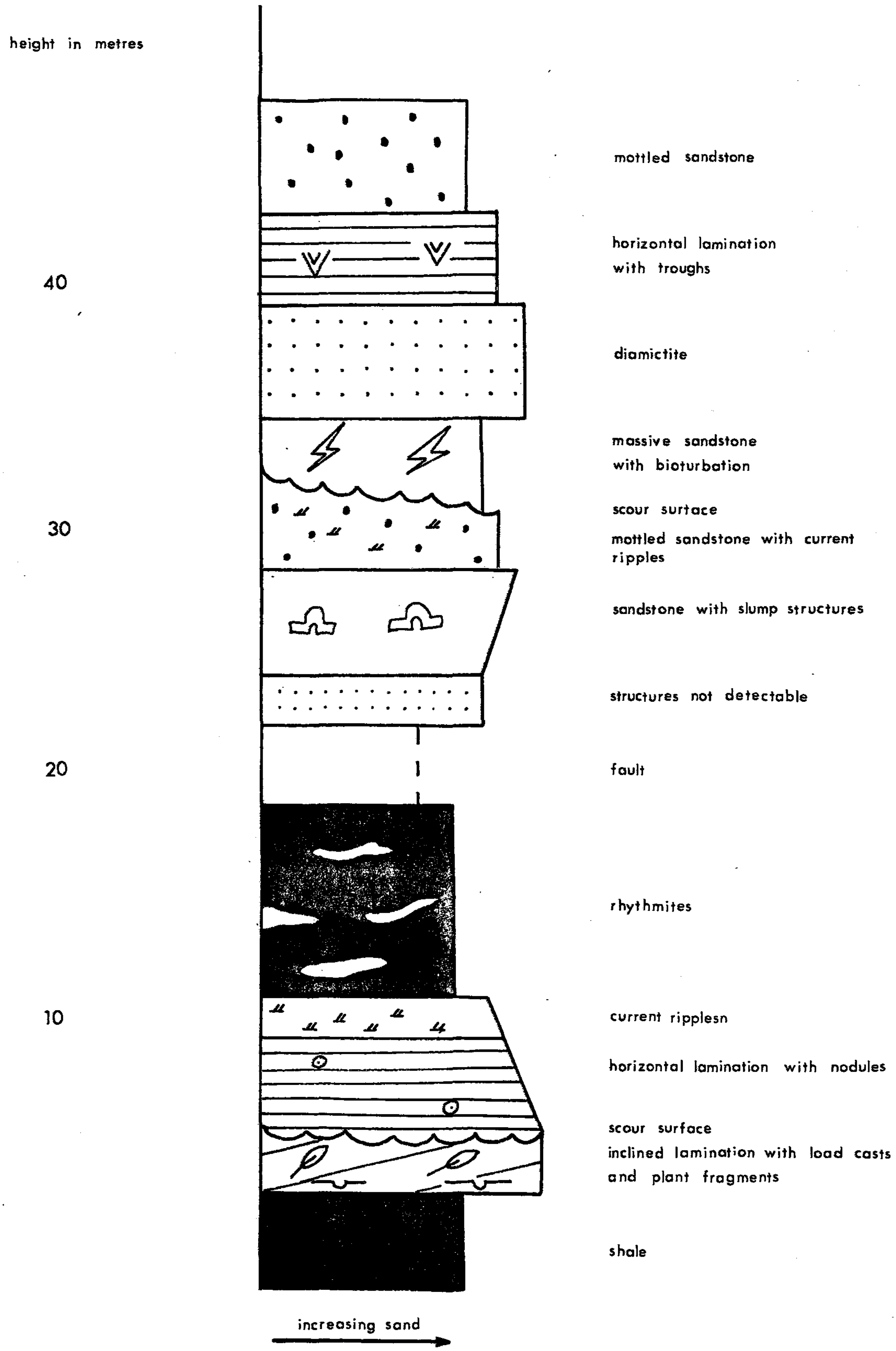

A
B
70
60
50
40
30
20
10
interbedded shales
horizontal lamination with current ripples troughs and inclined lamination
shale with slumps
inclined lamination
structures not detectable
bioturbated shale
horizontal and inclined lamination
rhythmites
fault
horizontal lamination with current ripples and nodules
rhythmites
rhythmites
fault
shales and massive units
horizontal lamination with troughs
horizontal and inclined lamination with bioturbation
rhythmites
shales and structureless unit
fault
rhythmites
shale with rippled sandstone

Fig. 7: Explanation of the symbols

- after Boersma in Ginsburg, 1975



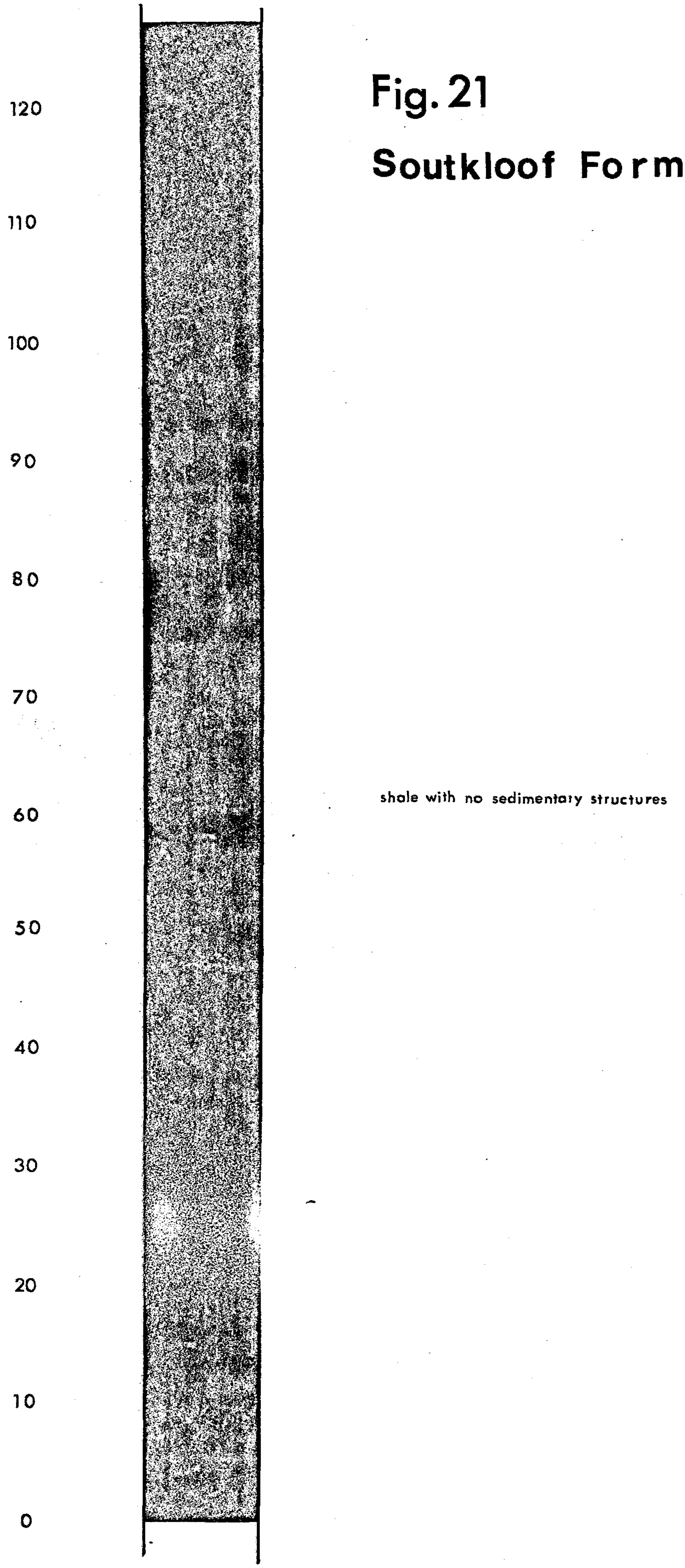
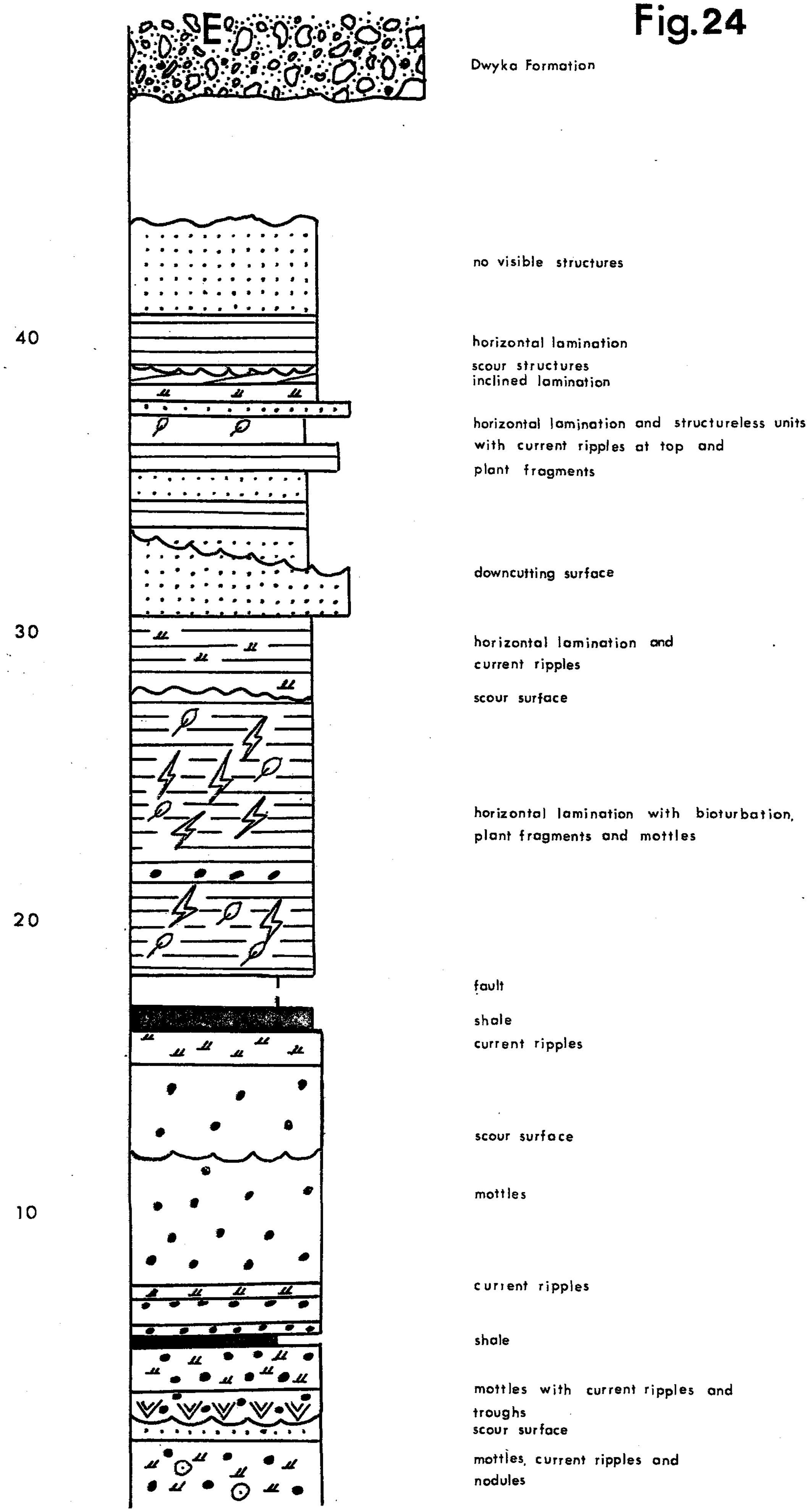


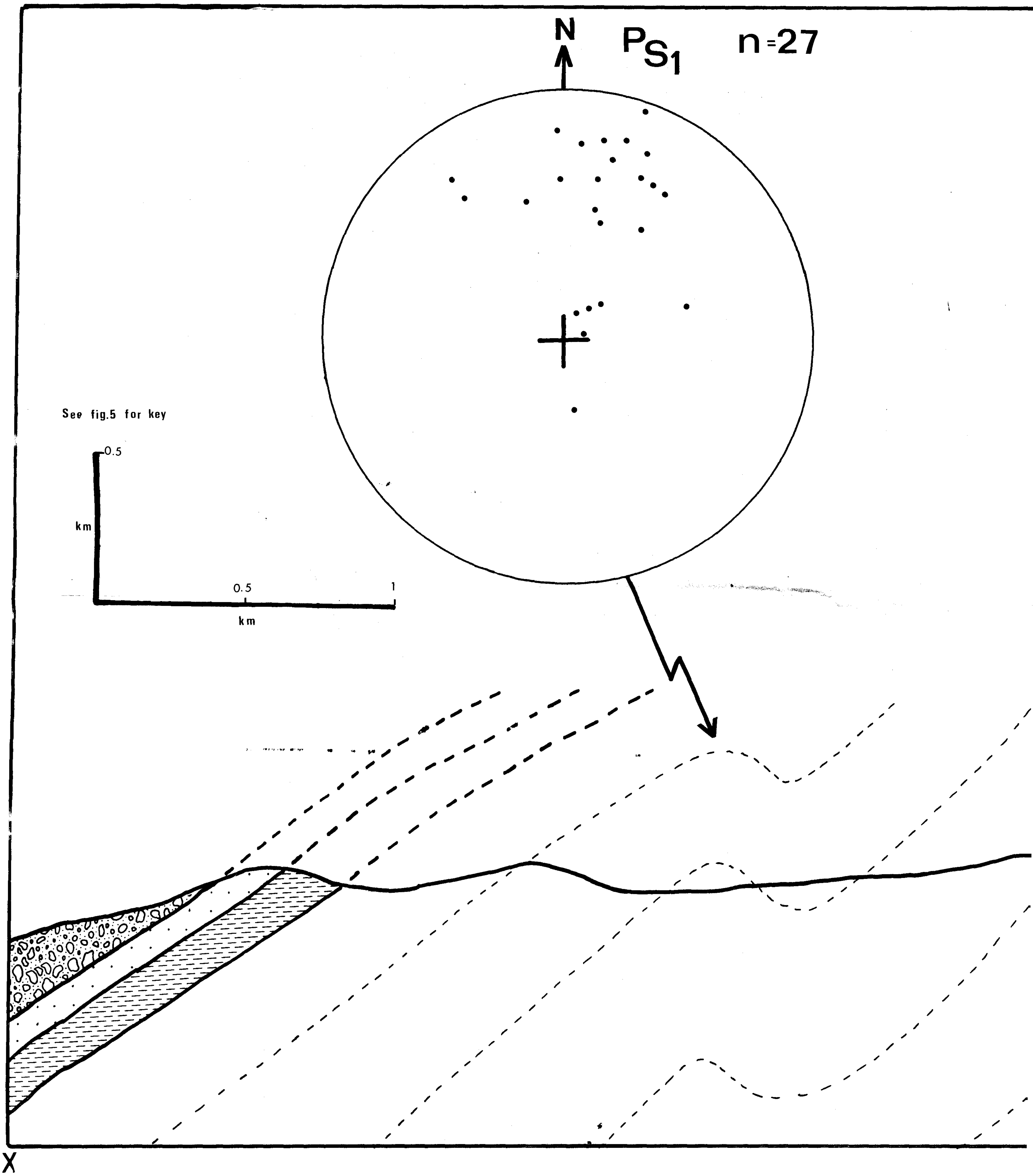
Fig. 21

Soutkloof Formation

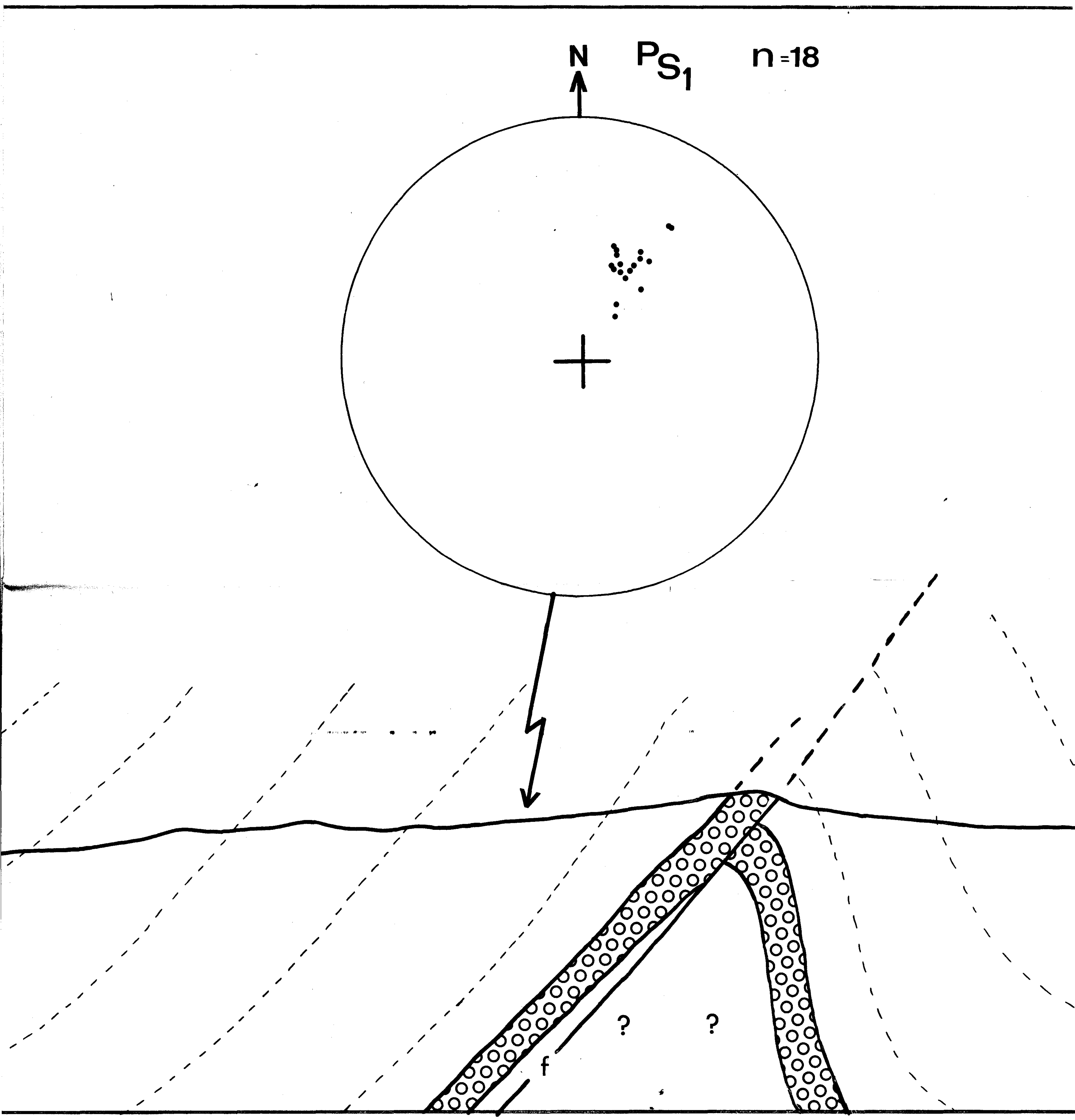
Fig. 24



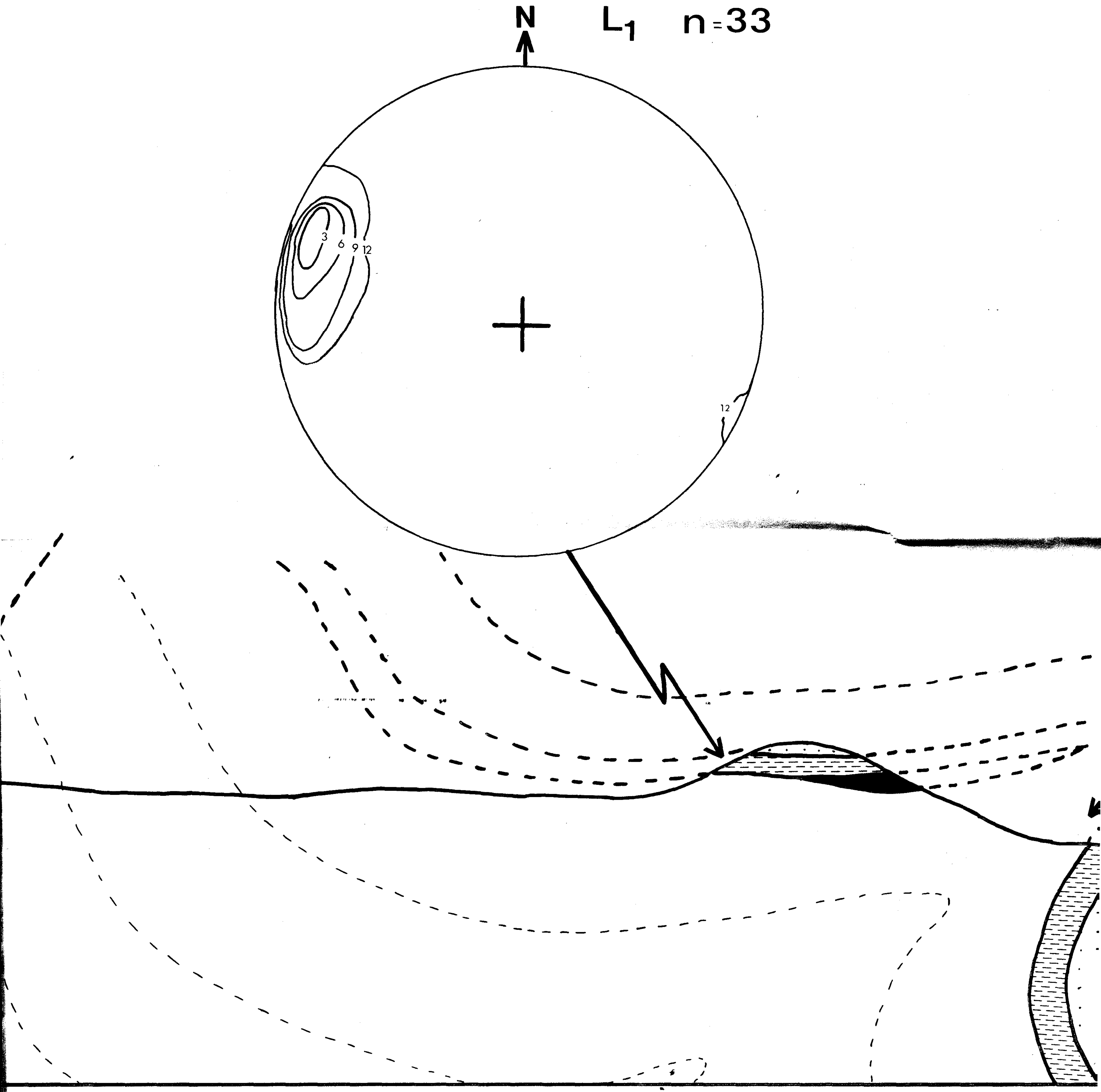
Dwyka Formation
no visible structures
40
horizontal lamination
scour structures
inclined lamination
horizontal lamination and structureless units with current ripples at top and plant fragments
downcutting surface
30
horizontal lamination and current ripples
scour surface
horizontal lamination with bioturbation, plant fragments and mottles
20
fault
shale
current ripples
scour surface
mottles
10
current ripples
shale
mottles with current ripples and troughs
scour surface
mottles, current ripples and nodules



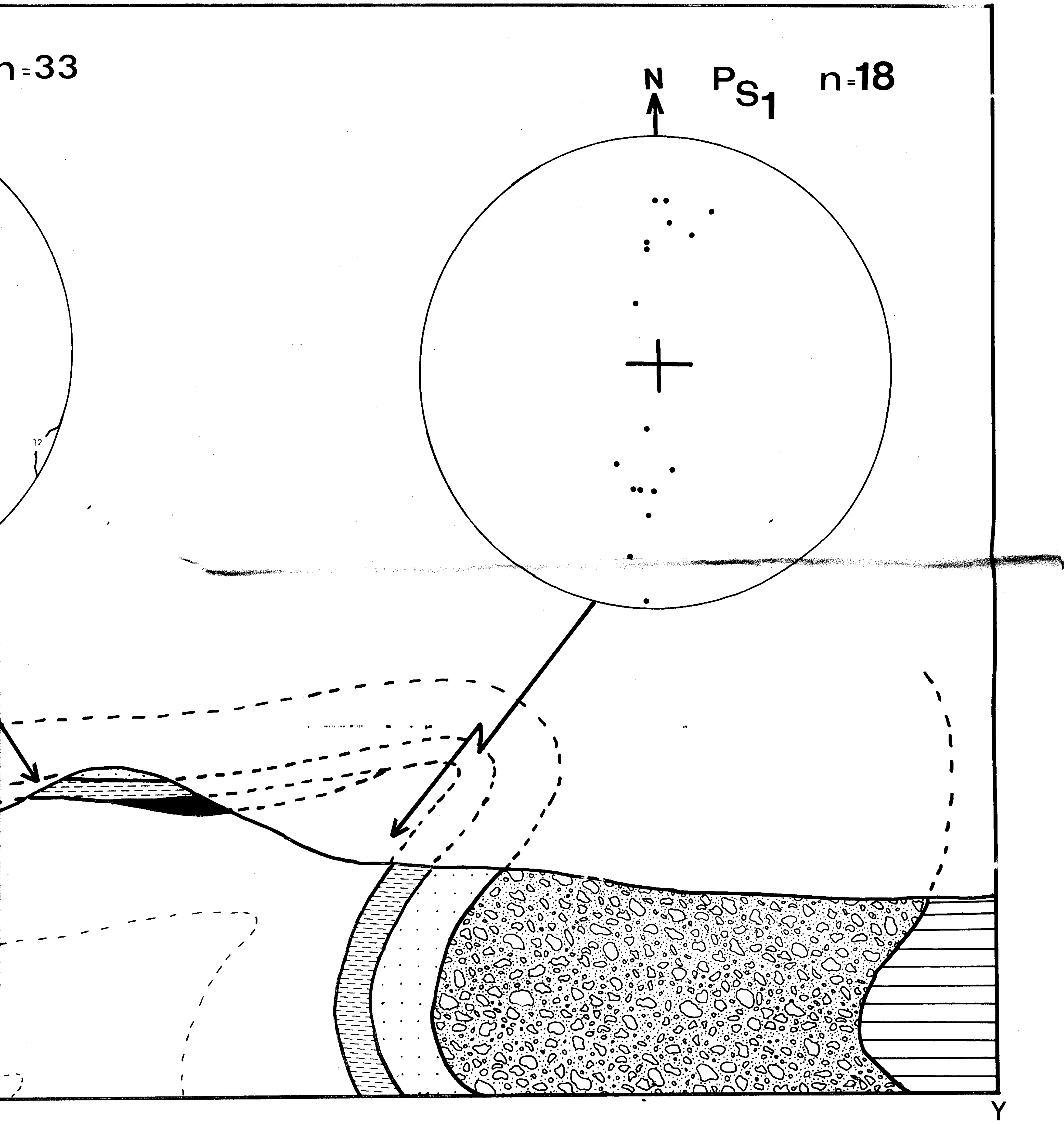
N
P_{S_1} n=27
See fig.5 for key
0.5
km
0.5
1
km
X



N
P_{S_1}
n =18
?
?
f



N
L_1 n=33
3
6
9
12
12



n=33
N
P_{S_1}
n=18
Y