GEOLOGICAL SURVEY OF NAMIBIA

MINISTRY OF MINES AND ENERGY

THE SEDIMENTOLOGY OF THE ZERRISSENE TURBIDITE SYSTEM, DAMARA OROGEN, NAMIBIA

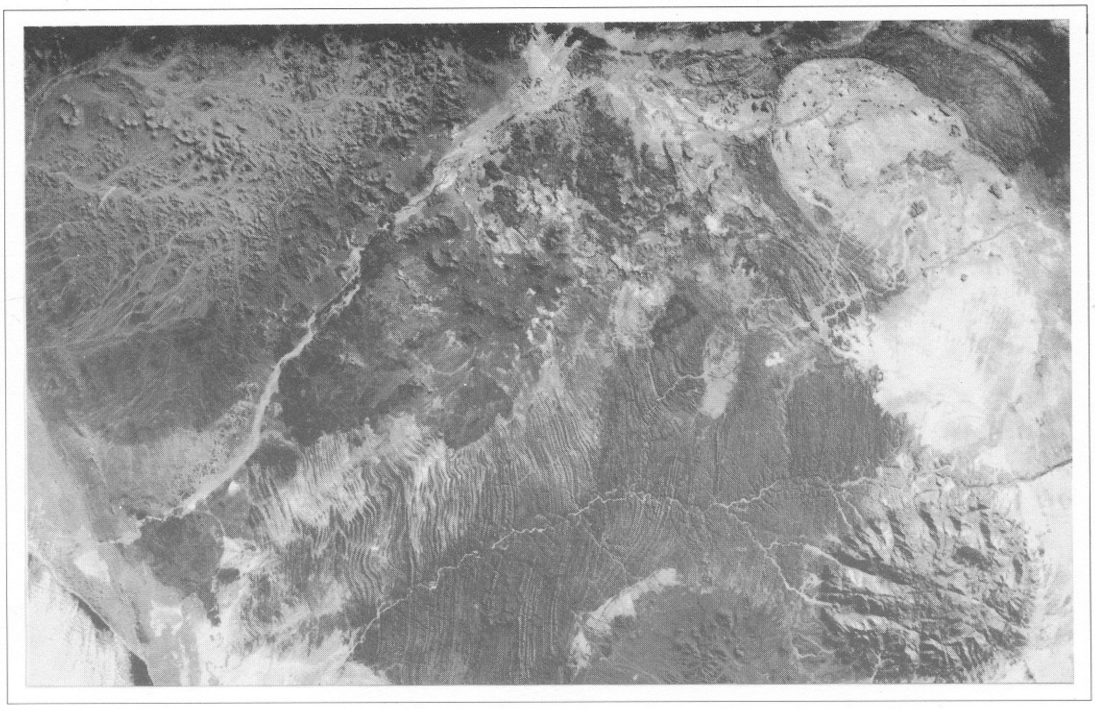
by

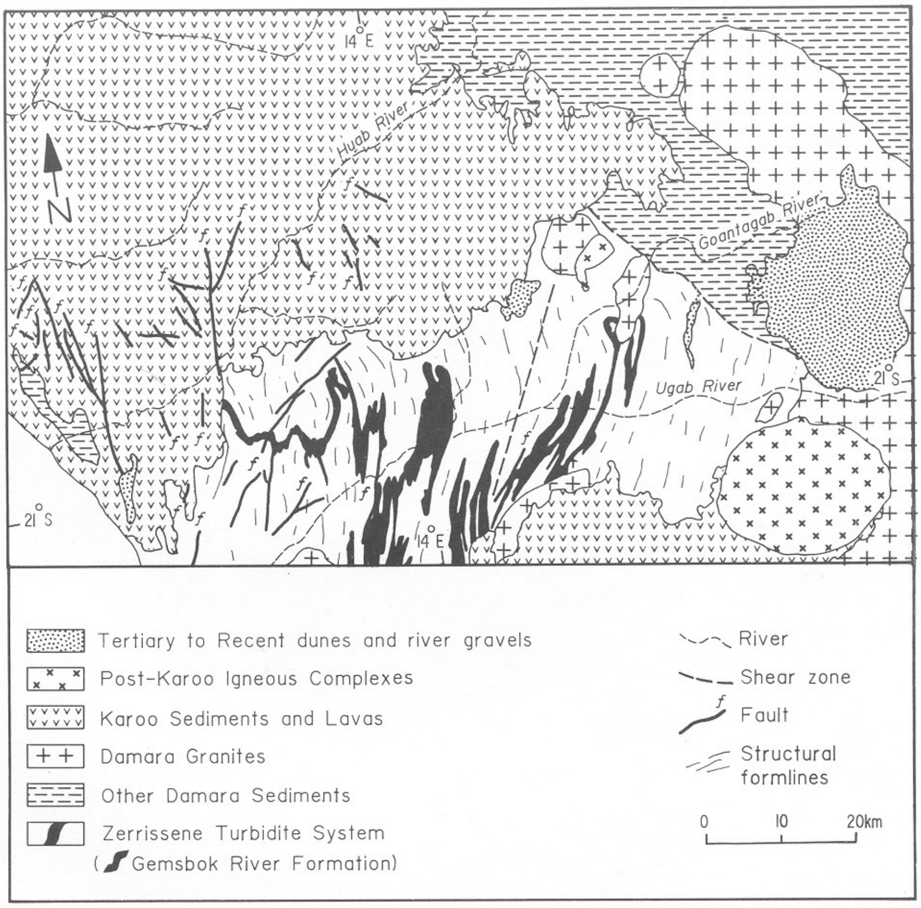
ROGER SWART

MEMOIR 13

1992

Front cover: Oblique aerial photograph, looking south, of the Zerrissene turbidite system showing individual lobe sandstones of the Brak River Formation which form low ridges in the foreground. The Zerrissene Hills are visible in the centre background of the photograph, and a well-developed Ramsay type 2c interference fold, which is developed in the Gemsbok River Formation, is visible in the middle ground. The tight, north-south trending F_1 folds are westward-vergent and have extremely attenuated normal (western) limbs, a characteristic feature of the deformation in the area.





FRONTISPIECE: Landsat image of southwestern Damaraland (top) showing many features which can be recognised on the simplified 1:1 000 000 geological map of Namibia (bottom). The tightly-folded Zerrissene turbidite system, the subject of this memoir, is clearly visible in the lower centre of the photograph. The rugged Brandberg massif, which at 2573 m is the highest mountain in Namibia, is prominent in the lower right corner. Karoo sediments, lavas and basic intrusions dominate the northeastern corner (top left), whereas Damara sediments and granites occupy the top right portion of the image. The tear-shaped dark area just right of the centre of the photograph is the post-Karoo Doros igneous complex.
Photo: Satellite Applications Centre, MIKOMTEK, CSIR, Pretoria.

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Director: R. McG. Miller

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ABSTRACT

The Zerrissene turbidite system of central-western Namibia is a late Proterozoic sequence which consists of dominantly siliciclastic turbidites interbedded with minor turbiditic and hemipelagic marbles. The basin in which these sediments were deposited is located at the junction of the coastal and intracratonic arms of the Pan-African Damara Orogen, and an understanding of the sedimentary evolution of this basin is therefore important to the understanding of the development of the orogen as a whole.

One major and two minor phases of folding have deformed the sediments, but the grade of metamorphism is low and sedimentary structures are often well preserved. Further, the area lies entirely within the Namib Desert and the lack of vegetation cover results in good outcrops providing an unusual opportunity for examining a large Precambrian turbidite system.

The system consists of five formations: three siliciclastic and two mixed carbonate-siliciclastic units. The floor of the system is not exposed, and the oldest sedimentary rocks which outcrop are siliciclastics of the Zebrapüts Formation. This is overlain successively by the Brandberg West Formation (dominantly calcareous), the Brak River Formation (siliciclastic), the Gemsbok River Formation (calcareous) and the Amis River Formation (siliciclastic).

Nine silicilastic turbidite facies have been recognised in the basin. These are facies A_2 (disorganised conglomerates), B_1 (horizontally laminated to massive grewyackes), C_2 (“classical” turbidites), D_1 (sandstone-shale couplets with base cut-out Bouma sequences), D_2 (sandstone-shale couplets with less sand than shale and base cut-out Bouma sequences), E (coarse, discontinuous sandstone-shale couplets), F (slumped units), G (shale) and H (glacial dropstones). Four facies are associated with the carbonate horizons, and these carbonate facies are given the suffix c to distinguish them from similar siliciclastic facies. These are facies Ac (disorganised and graded marble breccias), facies Cc (graded carbonates), facies Gc (hemipelagic marbles) and facies G (pelagic shales).

The basal Zebrapüts Formation is made up of relatively thin packages of thin- to thick-bedded, laterally continuous facies D_1, D_2 and B_1 beds encased in thick envelopes of shale. This type of sequence is typical of a distal lobe fringe, and requires an unconfined basin floor on which it can develop. The overlying Brandberg West Formation consists of a basal portion of interbedded facies Cc and G, followed by a sequence dominated by facies Gc. This sequence is interpreted as representing outer-apron carbonate turbidites, derived from multiple point sources (facies Cc), with background pelagic settling (facies G) overlain by hemipelagic deposits (facies Gc). A reversal back to siliciclastic turbidites followed with deposition of the Brak River Formation. This sequence comprises relatively thick packages of laterally continuous facies B 1, D_1, and D_2 beds sandwiched between facies G shales, a succession characteristic of a lobe to lobe-fringe environment with intermittent abandonment of lobes. An unconfined basin floor adjacent to a passive margin is required for the development of this type of sequence. Glacial dropstones (facies H) are found in the upper portions of this formation, and slumped beds are also present (facies F), but are uncommon. The facies F beds are only found in association with facies H and are therefore considered to be genetically related. Slumping of beds was possibly caused by an oversupply of sediment from ice-rafting which caused instability. The overlying Gemsbok River Formation has a sequence similar to the Brandberg West Formation in that the basal portion consists of interbedded facies Cc and G, which is overlain by a thick sequence of largely facies Gc beds. Minor facies Ac beds occur near the top of the overall sequence. This formation is interpreted as an outer-apron succession with the facies Ac beds representing distal inner-apron deposits, indicating progradation of the system. The youngest unit in the basin, the Amis River Formation, shows strong lateral variation from west to east. In the west the sequence comprises laterally continuous facies B_1, C_2, D_1 and D_2 with rare, discontinuous facies E beds. Facies G is relatively minor in the sequence. In the east the succession is dominated by facies D_1, D_2 and G, and this succession is interpreted as a sequence of distal turbidites which was deposited on a basin plain. The system developed by aggradation rather than progradation, as only minor cycles are developed.

Geochemical and petrological features indicate that the entire siliciclastic system was derived from a granite - recycled orogen terrane. Palaeocurrent data are unreliable because of the deformation, but transport was initially from the southwest, moving later to the west and northwest. The provenance of the carbonates is uncertain as reliable palaeocurrent indicators are rare, but they could have been derived either from South America or from the extensive carbonate deposits developed on the northwestern margins of the basin.

The Zerrissene siliciclastic turbidite system represents the distal portion of a major submarine turbidite system, the more proximal parts of which now lie west of the exposed basin, either under the Atlantic Ocean or in eastern South America. The calcareous deposits developed as an apron adjacent to a multiple point source, the position of which is at present unknown.

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CHAPTER ONE: INTRODUCTION

1.1 THE PROBLEM

The late Proterozoic Damara Orogen of Namibia forms part of the Precambrian network of thermal, tectonothermal and orogenic belts on the west coast of Africa and which is now commonly referred to as the Pan-African system of mobile belts (Kennedy, 1964; Fig. 1.1). The orogen can be divided into two parts; a north-south trending coast-parallel arm which has a minimum width of 150 km, and a 400 km-wide northeasterly-oriented intracontinental arm (Fig 1.2; Martin and Porada, 1977; Miller, 1983). The northern junction between the two arms lies to the southwest of the Kamanjab Inlier (Fig 1.2). An understanding of the tectonic and sedimentological development of this zone is important in the understanding of the orogen as a whole. This study will examine the sedimentological features of this zone. Previous studies of the Damara Orogen have concentrated on its igneous, structural, metamorphic and stratigraphic features, but little is known about the detailed sedimentological development of the orogen, and this study provides new data on the sedimentary history of the Damara Sequence. The area is well exposed and provides an unusual opportunity for examining an ancient turbidite system.

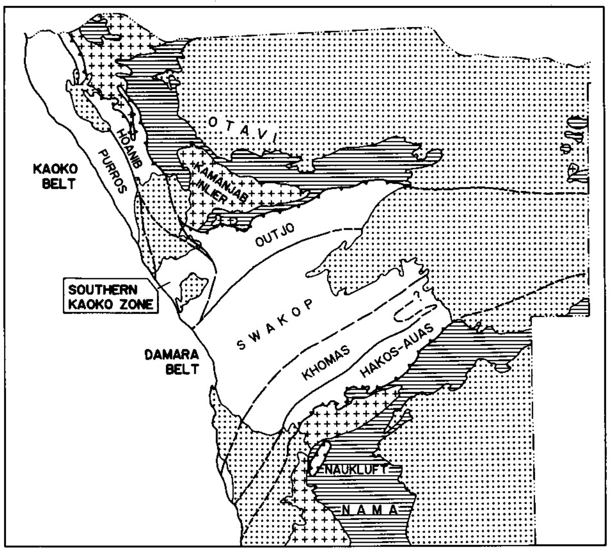

Fig. 1.2: Simplified tectonostratigraphic map of the intracontinental and coastal arms of the Damara Orogen showing the position of the current study area, the southern Kaoko Zone, in relation to other important features (after Miller, 1983 and Hoffmann, 1987). The southern Kaoko Zone lies to the southwest of a major basement inlier.

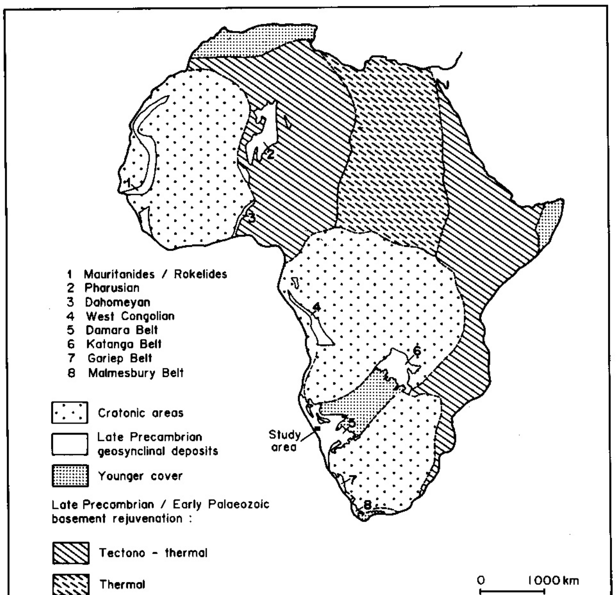

Fig. 1.1: Map of Africa showing the distribution of late Proterozoic Pan-African orogenic belts and related areas of thermal and tectonothermal rejuvenation (after Kennedy, 1964 and Porada, 1983). The Damara Orogen is bordered to the south by the Kalahari craton, the Congo craton to the north and the Atlantic Ocean to the west. A possible eastward extension of this belt is the Katanga Belt.

1.2 REGIONAL GEOLOGICAL SETTING

The northern junction between the two branches of the Damara Orogen was called the southern Kaoko Zone by Miller (1983; Fig. 1.2) and the Ugab Terrane by Hoffmann (1987). As an area east of this zone was called the Ugab structural domain by Weber and Ahrendt (1983) the area currently under study will be referred to as the southern Kaoko Zone (sKZ) to avoid confusion.

The sedimentary sequence in the area comprises a 1.7 km-thick, largely turbiditic, interbedded siliciclastic and carbonate succession (Miller *et al.*, 1983), with minor hemipelagic deposits. The stratigraphy of the area is shown in Table 1.1. As no group name exists for the sequence in this particular area it is here informally called the Zerrissene turbidite system, the name being derived from a prominent range of hills in the west of the zone where the entire succession is exposed (Front cover; Fig. 1.3).

The sediments have been deformed by one major phase (F_1) of deformation which has resulted in westward-vergent, north-south oriented, tight to isoclinal folds (Miller *et al.*, 1983; Porada *et al.*, 1983; Front cover; Figs 1.4 and Plate LA). The normal limbs of these folds are extremely attenuated (Plate 1.A), and the measuring of sedimentological sections on these limbs is therefore not possible. On the normal limbs the associated S_1 foliation is a slaty, bedding-parallel cleavage and this often destroys or modifies the sedimentary structures. On the overturned limbs the cleavage is at a high angle to bedding and where it is not penetrative, sedimentary structures may be well preserved. The overturned limbs are therefore the more suitable for sedimentological work. The second phase of deformation (F_2) refolded S_1 into kink folds, while the third and final deformation event (F_3) locally refolded the F_1 structures to produce Ramsay type 2c interference folds (Front cover; Plate 1.B; Coward, 1981; Ramsay, 1967). The fabrics associated with the last two phases are only locally developed and generally do not affect the sedimentary structures ex-

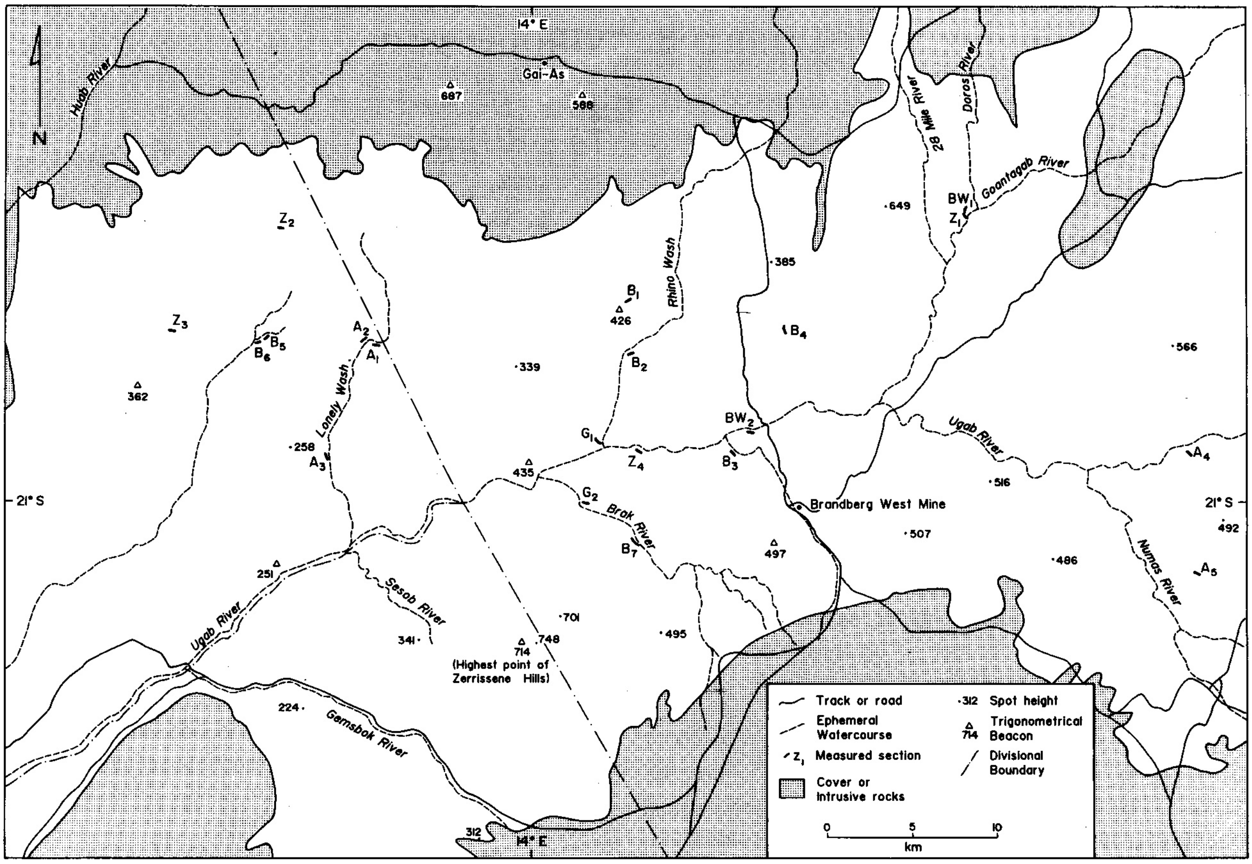

Fig. 1.3: Map of the lower Ugab River area showing major drainage lines, topographic features, routes and localities of measured sections. Section symbols are Z1 - Goantagab River; Z2- Ridge 6; Z3 - Skeleton Coast Park; Z4 - Augur; BW1 - Ulundi; BW2 - Ugab River; B1 - Shelter; B2 - Rhino Wash; B3 - Zebra River; B4 - Gai-as South; B5 - Gemsbok Altar East; B6 - Gemsbok Altar West; B7 - Brak River; G1 - Lookout; G2 - Corner; A1 - Jeppe East; A2 - Jeppe West; A3 - Lonely River; A4 - Bee; A5 - Tributary.

TABLE 1.1: Lithostratigraphy of the Zerrissene Basin and proposed correlations with the Northern Zone of the Damara Orogen (after Miller *et al.,* 1983)

NORTHERN ZONE	ZERRISSENE BASIN	LITHOLOGY	THICKNESS
Kuiseb Fm.	Amis River Fm.	Greywackes and pelites with rare carbonates and quartz-wackes	550 m
Karibib Fm.	Gemsbok River Fm.	Turbiditic and hemipelagic marble and pelite	200 m
Chuos Fm.	Brak River Fm.	Greywacke and pelite with dropstones	500 m
Rössing Fm.	Brandberg West Fm.	Turdibitic and hemipelagic marble and pelite	15-20 m
Okonguarri Fm.	Zebrapüts Fm.	Greywacke and pelite	350 m

cept in the fine-grained lithologies. The grade of metamorphism is low and nowhere exceeds the biotite isograd (Porada *et al.*, 1983). High temperature - low pressure contact metamorphism is developed at the contacts of these rocks with the large intrusive bodies.

A number of large isolated clasts, which are interpreted as glacially-derived dropstones, are found in the Brak River Formation (Table 1.1) and this unit has therefore previously been correlated with the Chuos Formation, an important chronostratigraphic horizon in other parts of the Damara Orogen (Jeppe, 1952; Miller *et al.*, 1983). The underlying and overlying carbonate units have therefore been correlated with marble horizons at similar stratigraphic positions elsewhere (Miller *et al.*, 1983; Table 1.1).

The age of these sediments is at present poorly constrained. Miller *et al.* (1983) suggested on the basis of the correlations described above that the lowermost Zebrapüts Formation is the equivalent of the Okonguarri Formation turbidites to the east of the present study area (Table 1.1), and is therefore older than 750 Ma. A minimum age for the system is given by the intrusion of a granite in the north-eastern portions of the area and which post-dates F_1. This

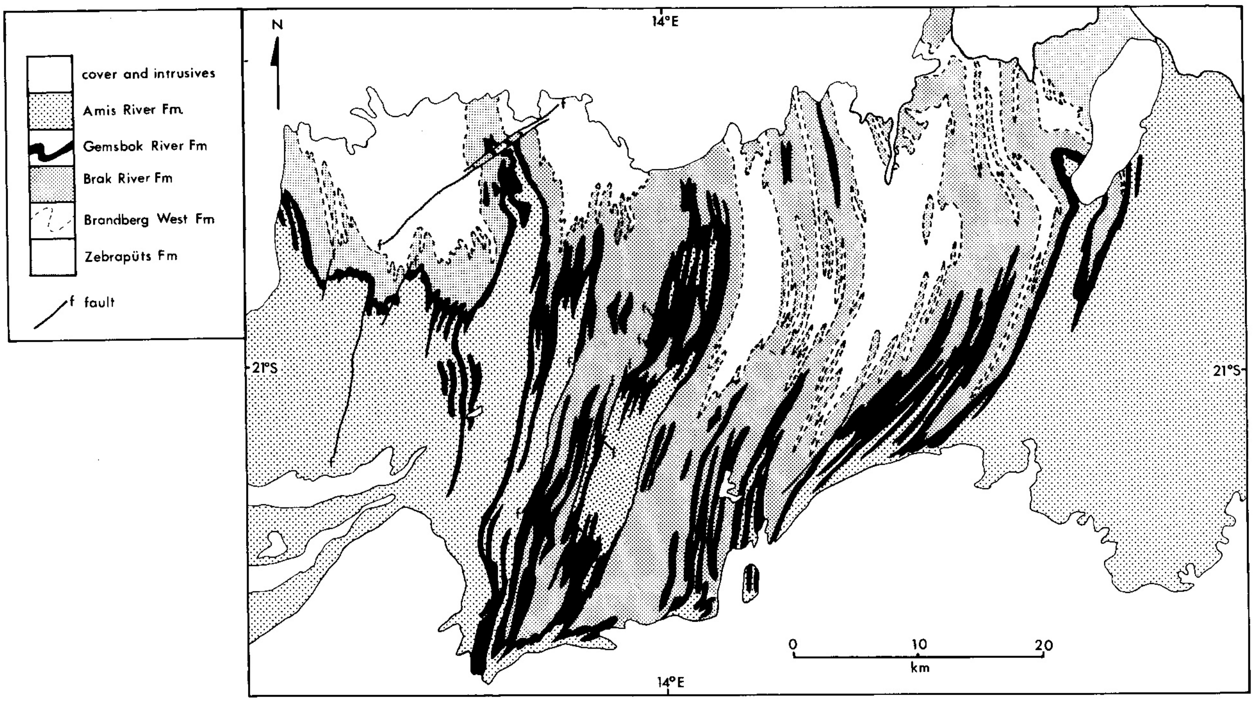

Fig. 1.4: Simplified geological map of the Zerrissene basin (after Geological Map of the Damara Orogen, 1988). The tight folding is clear from the repetition of the various units.

granite has been dated at 572 ± 31 Ma (Kröner, 1982). Evidence from elsewhere in the Damara Orogen suggests that the age of the glacial Chuos Formation is approximately 710 Ma (Hoffmann, pers. comm. 1989). This formation is at present correlated with the Brak River Formation in the current study area (Table 1.1). A mineral cooling age of 490 Ma for the area around the Brandberg West mine has been obtained by Ahrendt et al. (1983).

The exposed outcrop area is approximately 90 km long and 30 km wide. As the tectonic shortening in the area is at least 50 per cent, the undeformed area of the basin must have been at least 5 400 km^2, which makes it one of the largest known preserved ancient turbidite successions (Fig. 1.5). To the east the outcrop is cut off by a major shear zone which is clearly visible on aerial photographs. Previously, the marbles to the east of the shear zone were correlated with those to the west of it (Miller et al., 1983; Geological map of the Damara Orogen, 1988) but reconnaissance mapping has shown that this area is structurally (Freyer and Hälbich, 1983) and sedimentologic ally different to that west of the shear. In the southeast the sediments are intruded by the 125 Ma Brandberg complex. South of the area the exposure is cut off by Damaran granites and Mesozoic Karoo sediments and lavas. Isolated outcrops (Fig. 1.6) of the Zerrissene turbidites are however found southeast of the Messum complex, making the size of the basin even larger than 5 400 km^2. In the western part a major protomylonite belt, the Ogden Rocks Formation (Freyer and Hälbich, 1983), terminates the turbidite outcrops. In the north the Zerrissene turbidite system is covered by Mesozoic Karoo sediments and lavas.

1.3 TERMINOLOGY

The carbonate horizons have all recrystallised and consequently these units can no longer correctly be called limestones, and should rather be referred to as marbles. However, the greywacke and pelitic beds have not recrystallised to the same extent. The pelitic units have previously been referred to as schists (Miller *et al.*, 1983; Porada *et al.*, 1983), but as in many instances the grain size is not coarse enough for these to be correctly classified as schists, these will be referred to as shale or pelite. The matrix of the greywackes has recrystallised, but the greywackes are here referred to without the prefix meta- as many primary features can still be observed.

1.4 PREVIOUS WORK

Prior to the Second World War little was known about the area. Hans Cloos visited the region in 1929, and his name is still painted on the canyon walls of the Ugab River. A sketch of one of the folds in the Ugab River Valley appears in his autobiography (Cloos, 1951).

One of the first reports on the geology of the area is in a 1944 Geological Survey memoir on southern African lead occurrences (Willemse *et al.*, 1944) in which a report by the Solar Development Company is quoted. The report refers to gold "... associated with pyrite and galena in small quartz lenses on the south-eastern side of the isolated mountain south of Brakpüts on the Ugab River. This is the unnamed mountain 20 miles west of the Brandberg and five miles south of the Ugab" (Willemse *et al.*, 1944, p. 165). The mountain is the one now known as the Zerrissene Mountains.

Interest in the area increased during the war years because of the occurrence of tin and tungsten ores in the district. However, Jeppe (1952) was the first person to describe the geology of the area in detail. Much of his work was done by "... pacing and compass work" (Jeppe, 1952, p. 5) as

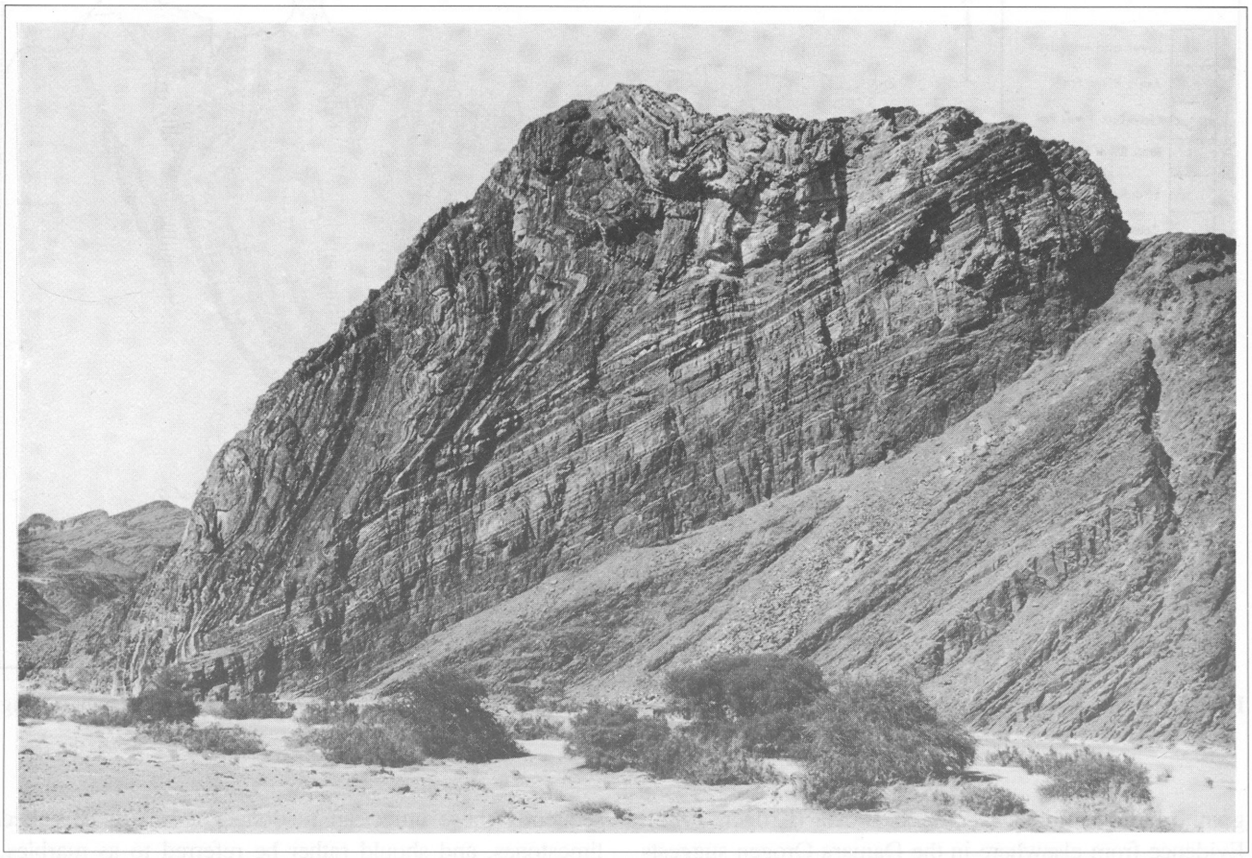

Plate 1.A: The characteristic features of the deformation in the area are the westward-vergent character of the F_1 folds and the strong thinning of the normal limbs of folds such as seen here at Monument Fold in the Ugab River (locality 6.5 km upstream of Salzpütz).

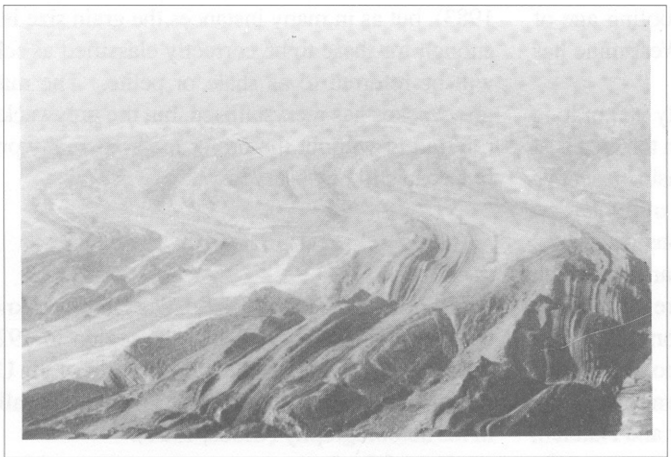

Plate 1.B: Oblique aerial photograph of three Ramsay type 2c interference folds caused by northwest-southeast shortening of the tight north-south F_1 folds. View looking north in the west of the area.

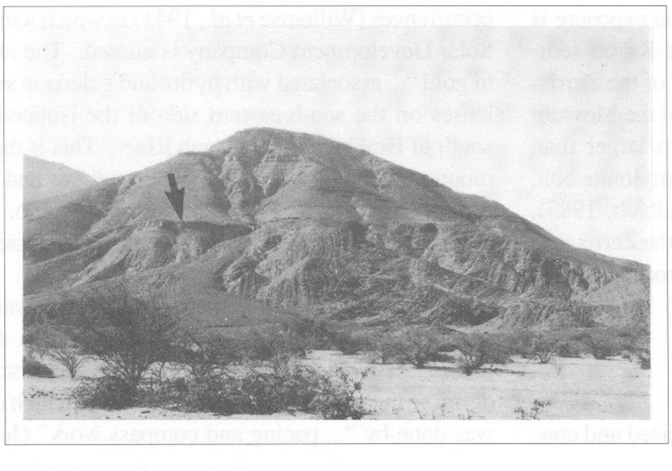

Plate 1.C: The unconformity between the folded Precambrian Zerrissene turbidites and the Mesozoic Karoo Sequence is particularly well developed on the western margins of the Brandberg massif. This photograph shows the unconformity (arrowed) in the Naib Gorge.

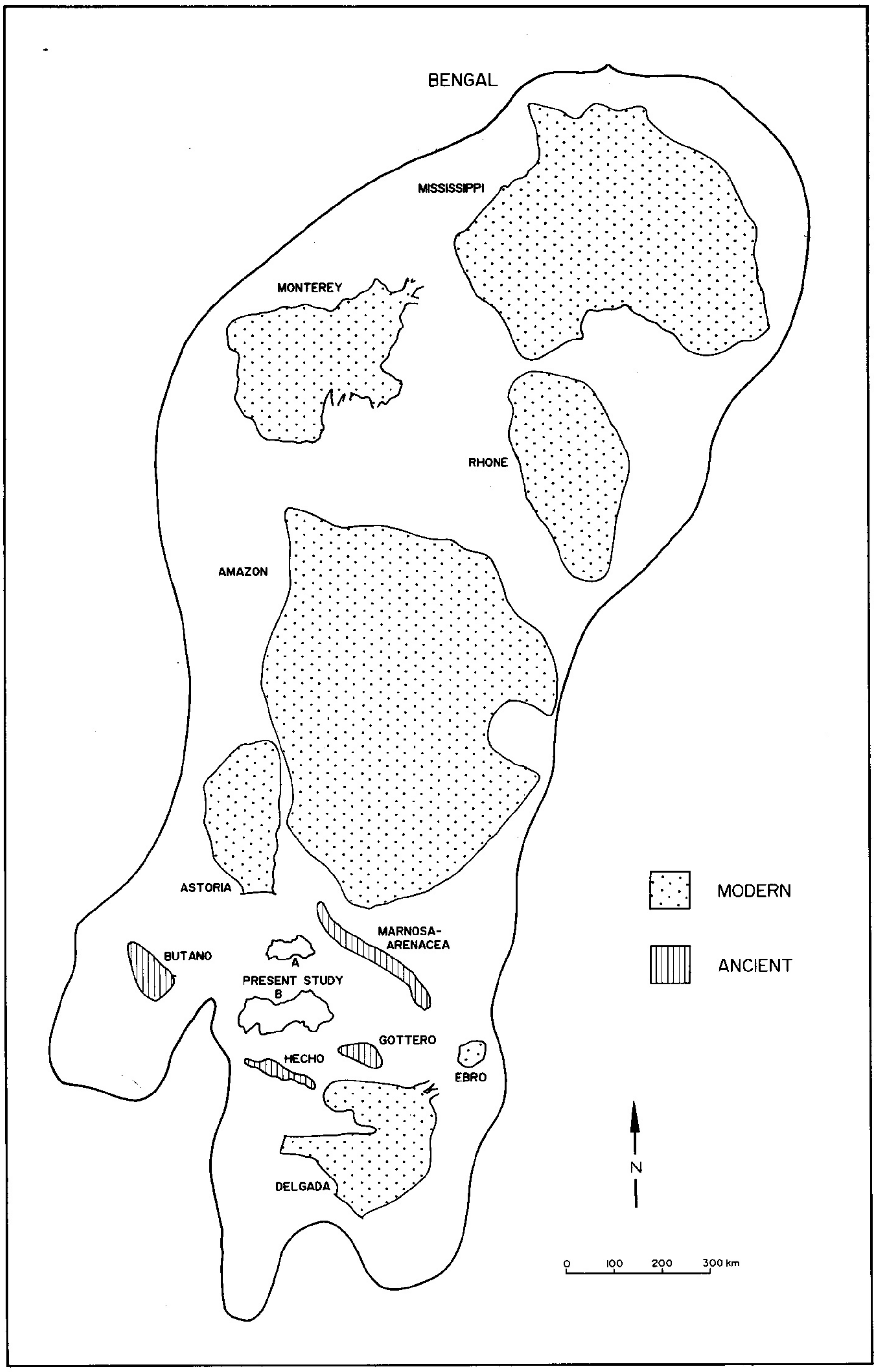

Fig. 1.5: Diagram comparing sizes of modern and ancient fans. The Zerrissene system is depicted showing the current extent of the outcrop area (A) and after unfolding assuming at least 50% shortening (B).

no detailed topographical maps or aerial photographs were available. In many cases his observations and maps of the area south of the Ugab River cannot be improved upon today.

The work of Hodgson (1972) around the Mesozoic Doros Crater was the next detailed study, followed by Miller (1973, unpublished map) who mapped portions of the area in the west. As knowledge of the orogen developed, more interest was taken in the structural and tectonic aspects of this important zone and led to papers by Coward (1981), Porada *et al.* (1983) and Ahrendt *et al.* (1983). Porada *et al.* (1983) suggested that the marbles were not typical of a deep basin and proposed that the area was a basement high during Damaran times. They termed this high the Huab Ridge and they interpreted this as a southerly extension of the Kamanjab Inlier which separated the coastal and intracratonic arms of the orogen (Fig. 1.7). Ahrendt *et al.* (1983) and Porada *et al.* (1983) also suggested that the style of deformation and low metamorphic grade indicated a high structural level. Miller *et al.* (1983) were the first to recognise that the entire sequence was in fact of turbiditic origin. They correlated the sequence with the entire Swakop

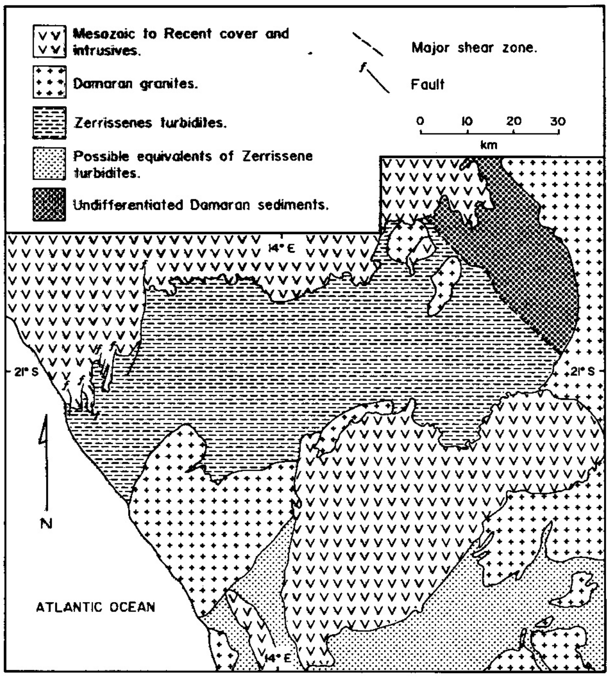

Fig. 1.6: Map showing locality of probable equivalents of the Amis River Formation south of the Mesozoic Karoo cover.

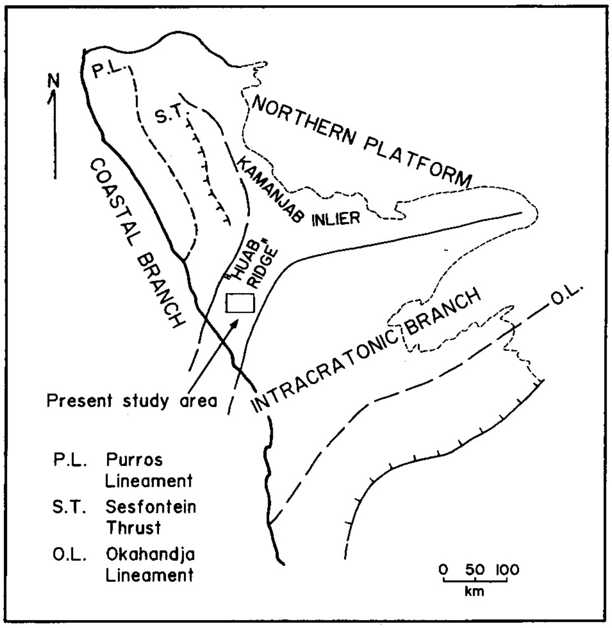

Fig. 1.7: Map showing position of assumed "Huab Ridge" of Porada (1983). This basement high is situated over the current study area.

Group and suggested that the basin was a well-developed ocean by 750 Ma. Currently Freyer (in prep.) is analysing the structural development of the area.

1.5 PHYSIOGRAPHY AND GEOMORPHOLOGICAL DEVELOPMENT

The outcrop area lies entirely within the Namib Desert (Ward *et al.*, 1983) and as there is little aeolian sand there is excellent exposure. However, in many sections sedimentary structures are not visible as extensive weathering and oxide patinas obscure them. A brief review of the physiography and geomorphological history of the area is necessary to appreciate the nature of the exposure.

One of Namibia's major fluvial systems, the ephemeral Ugab River, transects the area from east to west and provides excellent access along its course. Away from the river the tight folding is expressed geomorphologically as a series of low, north-south trending parallel ridges. In the marbles, synclines often form the ridges and anticlines the valleys (Jeppe, 1952). In the drier western parts of the area well-developed quartz-lag deposits (gibber plains) are found. The only major high within the area is the Zerrissene range of hills in the southwest (Front cover; Fig. 1.3). The maximum height of this range is 748 m.

It has long been recognised that the area was exposed at least by the Carboniferous (Jeppe, 1952; Reuning and Martin, 1957). Evidence for this is the existence of an unconformity at the base of the Karoo beds on the exposed margins of the area (Reuning and Martin, 1957; Frets, 1969). This unconformity surface is still apparent in the Damara sediments as a surface of equivalent summit heights or "gipfelflur" (Frets, 1969). The Zerrissene Mountains were however a palaeo-high during the deposition of the Karoo sediments and lavas (Jeppe, 1952) and have remained a high during the exhumation of this surface. From the Carboniferous through to the Permian the area appears to have been a shallow basin in which initially glacial and later shallow marine to lacustrine and fluvial sediments were deposited (Horsthemke and Ledendecker, 1991). A more arid phase in which aeolianites predominated followed on this aqueous phase. Continental break-up heralded a major period of volcanism at approximately 130 Ma which was followed by rapid erosion and up to 2.5 km of sediment was stripped away by the end of the Cretaceous (Brown *et al.*, 1988; Plate 1. C). This erosional period exhumed the pre-Karoo surface (Reuning and Martin, 1957; Frets, 1969). Minor aggradation has occurred since the beginning of the Quaternary and this can be recognised as well-cemented gravel deposits at three different elevations in the Ugab River valley. It is possible that the area has been arid since the early Miocene (Ward *et al.*, 1983; Ward and Corbett, 1990), and this long period of exposure since the Tertiary has allowed the development of oxide patinas which cover many of the rocks and may obscure sedimentary structures. Weathering along cleavage planes also obscures structures, and in many areas extensive scree cover prohibits detailed sedimentological work.

1.6 STUDY METHODS

Detailed sedimentological sections of all the formations were measured at a scale of 1: 100. Initially these were inspected visually for cyclicity. Inspection of the 1 :36 000 and 1 :50 000 aerial photographs was also carried out. Low-level aerial photographs of selected, relatively undeformed areas were also used to facilitate recognition of channel and lobe features.

Palaeocurrent directions were measured in the field and corrected for plunge and folding after the method described by Phillips (1971). Petrographic work comprised routine examination of thin sections. Unfortunately, the metamorphism has obliterated many primary textures, especially in the finer-grained rocks, but grading and primary textures such as clast shape and composition can still be recognised in thin section (see 4.2). Geochemical analyses were done as an aid to understanding the provenance of the sediments. Samples were taken from each of the siliciclastic formations over a wide geographic area and analysed in duplicate on a Phillips 1410 spectrometer at Rhodes University using the method of Norrish and Hutton (1969).

1.7 ACKNOWLEDGEMENTS

This project was initiated and supported by Dr Roy Miller, Director of the Geological Survey of Namibia, and would not have been possible without his backing. Charlie Hoffmann, also of the Namibian Geological Survey, was a constant source of ideas and stimulation.

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Roddy Ward took the set of aerial photographs of the channel described in section 3.3.3.8. His assistance in gaining this important information was invaluable. Hans Nelson and John Slack of the USGS provided references which are not easily obtainable in southern Africa and their willing help is appreciated.

Too many people have encouraged me over the years to thank them all, but John Ward in particular never allowed me to relax and rest on my laurels, and his encouragement in dark and gloomy patches was needed and appreciated. Barbara Curtis and Carole Roberts provided much needed encouragement while Jane Waterman deserves a special thank you for tolerating all my moods.

My mother and brothers were always a source of encouragement and loyalty. Their confidence in me was supportive.

CHAPTER TWO: BRANDBERG WEST AND GEMSBOK RIVER FORMATIONS

2.1 INTRODUCTION

The two carbonate units form a minor proportion (approximately 12 per cent; Table 1.1) of the total succession, but are excellent marker horizons as they are both laterally continuous and easily observable. The lower of the two marbles, the Brandberg West Formation, is only 15-20 m thick but can be traced out across the entire basin along a strike length of 370 km. Although much thicker (150¬200 m), the Gemsbok River Formation has a stratigraphic succession similar to that of the lower marble (see below) and can be traced laterally for over 450 km.

The metamorphism has caused complete recrystallisation of the marbles and consequently many primary fabrics have been destroyed. The extreme attenuation of the normal limbs of the folds and the disharmonic folding of the overturned limbs hamper the measurement of suitable, relatively undeformed sections of the marbles. Variation in bed thickness is therefore not a suitable criterion for distinguishing between various facies in these sediments.

As carbonate rocks respond readily to diagenesis, deformation and low grade metamorphism (Ham and Pray, 1962; Logan and Semeniuk, 1976), it has been suggested that many of the "sedimentary structures" in carbonate rocks, especially those in structurally complex areas, are metamorphic in origin (Logan and Semeniuk, 1976). Although many of the features that were identified as either structural or metamorphic in origin by Logan and Sementuk (1976) have since been shown by Playford (1980) to be primary, it is still essential to review the criteria used for recognising primary sedimentary structures in carbonates because in the southern Kaoko Zone we are dealing with strongly deformed rocks. It is probable that no single criterion is diagnostic, and where possible it is best to use combinations of criteria (Fitches and Maltman, 1978).

The fundamental unit of facies and basin analysis is the bed (Simpson, 1985), and in many carbonate rocks "bedding surfaces" may not be primary in origin. These secondary surfaces have been variously referred to as diagenetic bedding (Ricken, 1986; Bathurst, 1987), pseudobedding (Simpson, 1985; Alvarez *et al.*, 1985) or stylobedding (Logan and Semeniuk, 1976). Precambrian sedimentary sequences lack both body and trace fossils, the deformation of which may be used for recognition of tectonic and compaction structures in Phanerozoic rocks (Hobbs, Means and Williams, 1976; Ricken, 1986). Other methods are therefore needed to detect pseudobedding in Proterozoic sediments. The secondary nature of these surfaces can be identified by a concentration of solution seams in the region of the plane which may be close to, but not neccessarily the same as, the original bedding (Simpson, 1985; Bathurst, 1987). These secondary surfaces may be the planes along which later shear can occur (Logan and Semeniuk, 1976).

The discrimination of sedimentary breccias and slumps from tectonic breccias is also problematic, especially in deformed terrains. Later erosion of the top of a slumped or brecciated unit is conclusive evidence for a synsedimentary origin of the feature (Hobbs, Means and Williams, 1976). Boulders projecting out of beds into overlying strata are further evidence for a syndepositional origin. The presence of allocthonous clasts would also suggest a synsedimentary origin. However, these features are not always present, and observations of the relationship between the feature being examined and other known sedimentary and tectonic structures are required to differentiate them, if it is in fact possible to do so. These criteria are applied here to help in the distinction of sedimentary structures.

2.1.1 Source of the formation names

The Brandberg West Formation derives its name from an abandoned tin-tungsten mine in the area (SACS, 1980; Fig. 1.3). Previously this unit was the lowermost member of the Brak River Formation (SACS, 1980) but was given formation status by Miller *et al.* (1983). The Gemsbok River Formation is named after a tributary on the southern bank of the Ugab River in the western portion of the area (SACS, 1980; Fig. 1.3).

2.2 FACIES TYPES

The facies nomenclature used in this study is based on that proposed for siliciclastic turbidites by Mutti and Ricci Lucchi (1972; 1978) and is described more fully in chapter three. The suffix c is used here to indicate that the facies is calcareous. Three different carbonate facies have been recognised (facies Ac, Cc and Gc) and a single siliciclastic one (facies G).

2.2.1 Facies Ac

This facies consists of matrix-and clast-supported breccias with a dark blue matrix. It is best exposed near the Brak River, some 8 km south of the Ugab River. The layers are laterally discontinuous being only a few metres long, and the maximum thickness of individual beds is 1.5 m. Normal grading is present in some, but not all, layers (Plate 2.A). Lower contacts are sharp and convex downwards, but the upper contacts are often gradational. Clasts vary in length from a few millimetres to 12 cm (Plate 2.B). All clasts are of carbonate and three different types have been recognised. These are a grey-white carbonate, a blue marble and one with a lamination (Plate 2.B). These laminated clasts are not known from the area, thus suggesting an allocthonous source.

The allocthonous clasts and presence of grading suggest a primary sedimentary origin for this facies. Beds of this type are typically interpreted as being the deposits of rock falls, debris flows or of high-density turbidity currents (Mullins and Cook, 1986). Modem reef complexes commonly have major talus aprons on their seaward margins (James and Ginsburg, 1979; Land and Moore, 1979; Mullins and Neumann, 1979; Schlager and Chermark, 1979) and similar features have also been described from ancient

Plate 2.A: Normal grading in a carbonate breccia (locality: Ugab River, 400 m west of Rhino Wash)

Plate 2.B: Carbonate breccia with large clasts up to 12 cm in length. The laminated clast (arrowed) in the lower centre of the photograph has no known lithological equivalent in the area (locality: 10 km south of the Ugab River in the Brak River).

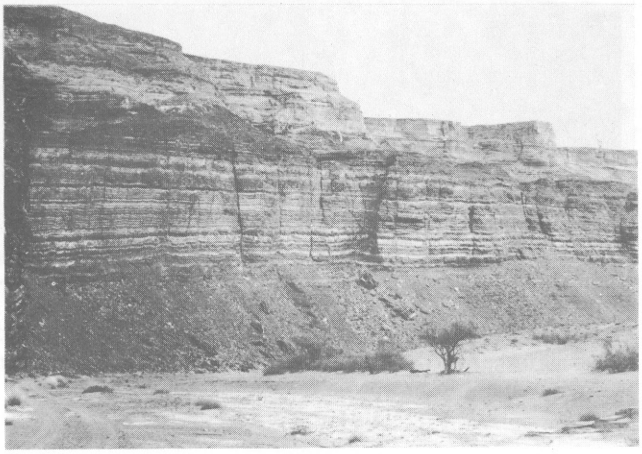


Plate 2.C: Basal portion of the Gemsbok River Formation showing lateral continuity of individual facies Cc (light) and facies G (dark) units (locality: 1 km north of the Ugab River in the Rhino Wash).

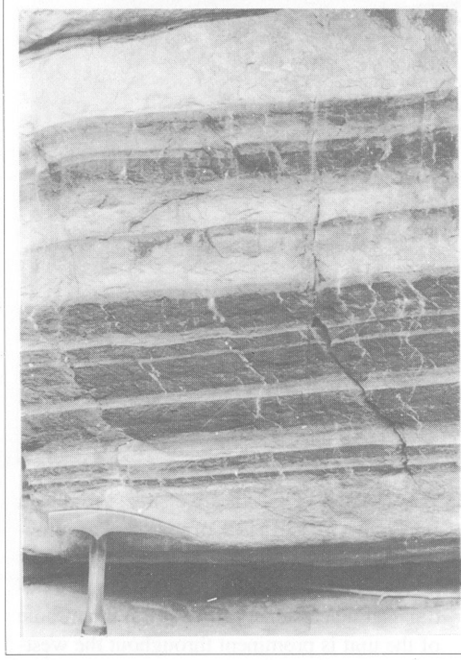


Plate 2.D: Compositional grading in facies Cc from a basal carbonate layer to a more pelitic top (locality: Lookout section).

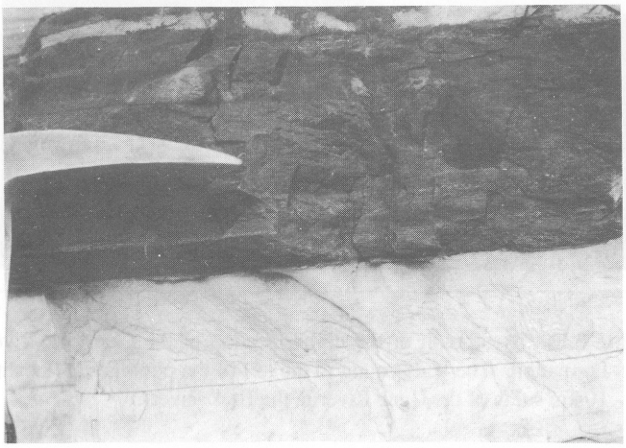


Plate 2.E: In many instances there is no evidence of compositional grading in the white marbles and they rather have a sharp contact with the overlying black schist giving the appearance of a Zebra rock (cf. Plate 2.C).

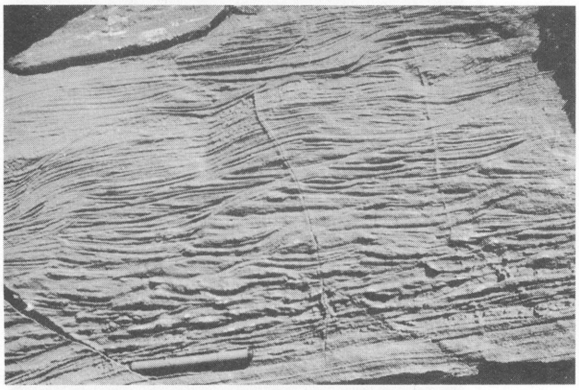


Plate 3.A: Well-developed ripple lamination in silicified facies Cc beds in the Gemsbok River Formation (locality: 2 km south of Ugab River in Brak River).

Plate 3.B: Clasts of white marble in facies Cc beds (locality: 2.4 km east of Goantagab River mouth in Ugab River).

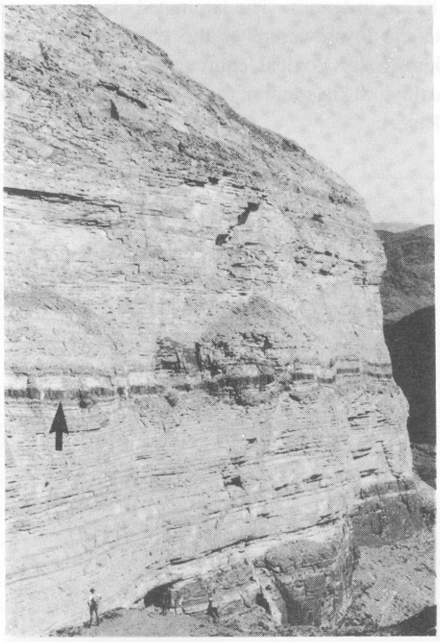


Plate 3.C: Facies Gc beds are characterised by a fine grain size and a lack of sedimentary structures. "Grading" from darker to lighter areas is not a primary grading. The strongly boudinaged black schist (arrowed) in the middle of the unit is prominent throughout the western part of the basin (see text; locality: Lookout Section).

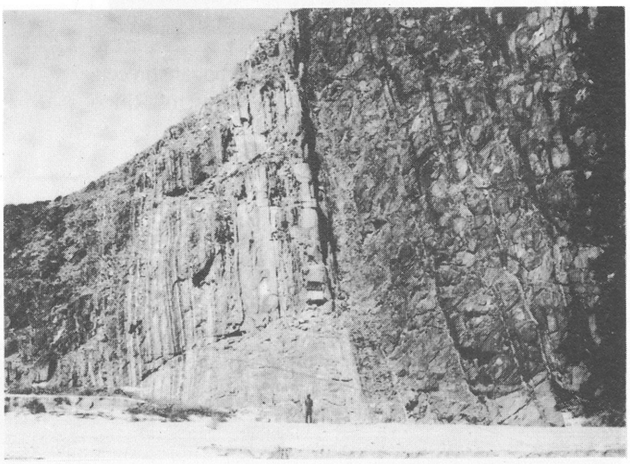


Plate 3.D: Photograph of the Ugab River measured section. Beds young to the right (locality: 2 km downstream of the old Brandberg West well).

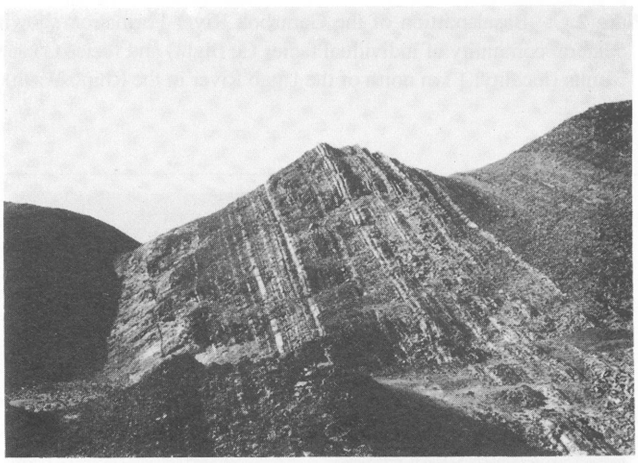


Plate 3.E: Photograph of the corner section of the Gemsbok River Formation. The beds young to the left of the photograph (locality 10 km south of the Ugab River in the Brak River).

sequences (Cook *et al*., 1972; Conaghan *et al*., 1976; Johns, 1978).

2.2.2 Facies Cc

A fine- to medium-grained white marble often with a siliceous lamination comprises this facies. Many of the beds which are included with this facies, in both formations, are possibly of the facies D type, but the metamorphism and deformation make it difficult to distinguish them from facies Cc. As facies C and D are in fact transitional with each other (Mutti and Ricci Lucchi, 1972; 1978), no distinction between the two was made in this study.

The bedding varies from thin to very thick, and individual beds are laterally persistent for many tens of metres (Plate 2.C). Compositional grading from carbonate to pelitic tops occurs in some units (Plate 2.D) and top and bottom contacts are generally sharp (Plate 2.E). Deformed ripples defined by the lamination have been observed (Plate 3.A). Occasionally clasts of white marble are found (Plate 3.B) and these are possibly more common than has previously been realised as it is difficult to distinguish the white clasts set in the similarly coloured matrix.

The lateral continuity of individual beds combined with the presence of clasts, ripples and grading indicates that this facies is a calc-arenite deposited by high- to low-density turbidity currents adjacent to a carbonate margin. This facies is the equivalent of the allodapic limestones of Meischner (1964).

2.2.3 Facies Gc

Generally fine-grained, this facies is a dark to light blue marble with no primary sedimentary structures (Plate 3.C). The fine-grained nature of this facies has been preserved despite the metamorphism. Recrystallisation of the carbonate has possibly been inhibited by the presence of carbonaceous material which is present in this facies in minor quantities. A parallel lamination exists which was previously thought to be primary (Swart, 1987; Miller *et al*., 1983) and the colour difference from a light base to a darker top was taken to be a reflection of sedimentary grading (Miller *et al*., 1983). However, this is not a consistent feature, and it appears that the colour change is a weathering effect, possibly related to an increase either of dissolution seams or of carbonaceous material near the pseudobedding plane, similar to that described by Bathurst (1987). Later slip has occurred along these planes as shown by the offset of quartz veins.

The lack of sedimentary structures and fine-grained character of this facies, even after metamorphism, suggest that it was deposited in a quiet water environment. The “pseudobedding”, which was previously interpreted as primary bedding, is now reinterpreted as being diagenetic in origin. Tucker (1974) has shown that true pelagic carbonates were not formed before the Upper Silurian, and therefore these beds cannot be of this type. They are, however, similar to the hemipelagic deposits described by McIlreath and James (1984). These deposits are typically grey-blue, with a very thin lamination (Cook and Mullins, 1983; McIlreath and James, 1984). Modem equivalents are typically homogeneous and lack obvious bedding (Mullins and Cook, 1986), and the lamination in ancient sediments is probably secondary. These shallow water-derived fine-grained sediments occur on the slope peripheral to a carbonate platform, and were termed “peri-platform oozes” by Schlager and James (1978). They are derived from the shelf where they are either put into suspension by storms or are the result of inorganic precipitation (Shinn *et al.*, 1989) before being carried into deep water where they settle out by gravitational processes. Hemipelagic peri-platform oozes may well have been the most common form of deep water carbonate in the Precambrian (McIlreath and James, 1984).

2.2.4 Facies G

The only non-calcareous beds associated with the Brandberg West Formation and the Gemsbok River Formation, this facies is represented by dark brown, fine-grained calcareous quartz and biotite-bearing pelites. Calcite is present in variable amounts. Bedding is thin to medium, contacts are sharp and no sedimentary structures have been observed. Individual beds are continuous for several hundred metres at least (Plate 2.C).

Miller *et al.* (1983) and Swart (1987) originally interpreted this facies as being turbiditic. However, the fine-grained nature and lack of sedimentary structures in these beds suggest that they were deposited in quiet conditions by pelagic settling. The grading of the facies Cc beds to pelitic tops (Plate 2.D) may be in part due to mixing of the fine-grained portion of a carbonate turbidite with the background pelagic settling.

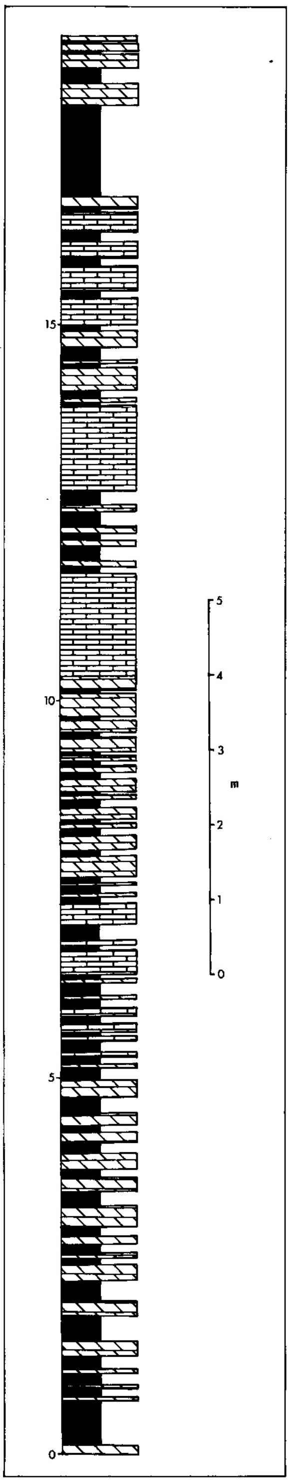
2.3 VERTICAL SECTIONS OF THE MARBLES

2.3.1 Brandberg West Formation

Two sections of the Brandberg West Formation were measured (Ulundi and Ugab River sections; see Fig. 1.4 for localities) and outcrops at various other localities were examined for sedimentary features. The formation varies between 15 and 20 m thick, but because of the deformation it is difficult to determine the true thickness of the formation. Both sections show a similar vertical succession upwards from an association of facies G and facies Cc to facies Gc beds (Plate 3.D; Figs. 2.1,2.5 and 2.6). The proportion of facies Cc beds increases with stratigraphic height (Fig. 2.4). This same succession, although not always well exposed and sometimes highly deformed, can be viewed throughout the entire basin. Lateral variations of the Brandberg West Formation are small, even over great distances, and no beds have been observed pinching out.

2.3.2 Gemsbok River Formation

The vertical succession of facies in the Gemsbok River Formation is the same as the Brandberg West Formation, except in the middle of the upper portion of interbedded facies Gc beds where a temporary return to a type of sedimentation similar to that of the basal portion occurred



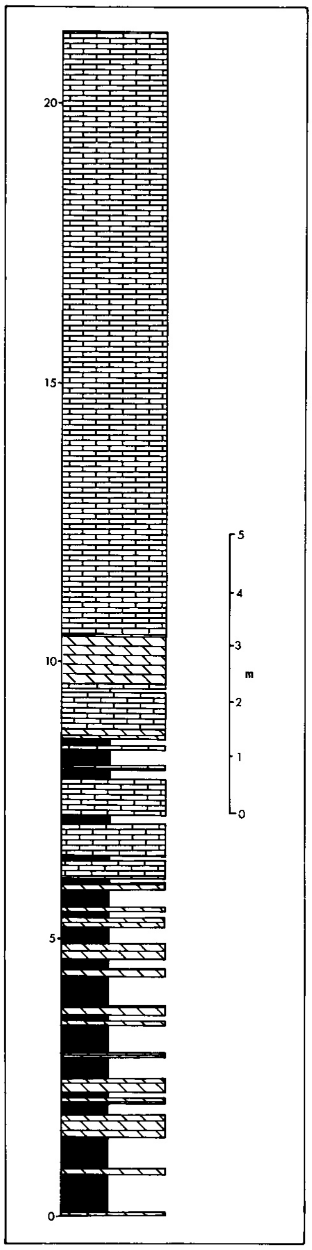


Fig. 2.1 (left): Sedimentary log of the Ugab section (see Fig. 2.3 for key and Plate 3.D for photograph).

Fig. 2.2 (right): Sedimentary log of the Ulundi measured section (see Fig. 2.3 for key).

(Plate 3.C and Fig. 2.5). In addition, in the upper parts of the Gemsbok River Formation there is localised development of facies Ac. Only one complete section (Lookout section) of this formation has been measured (Fig. 2.5) and because of structural effects this does not represent the true thickness of individual beds, but it is at least a correct representation of the sequence of facies and facies associations. A partial section has been measured in the Brak River

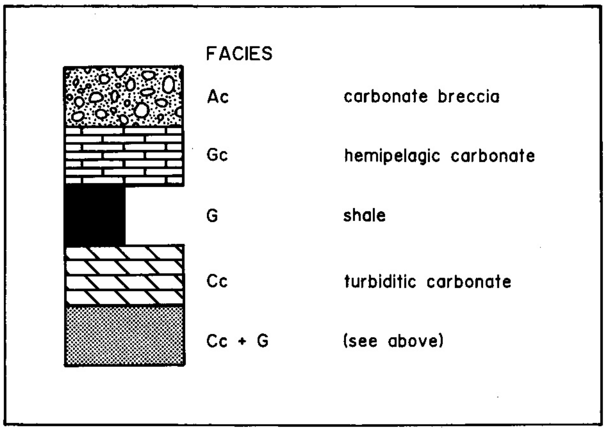


Fig. 2.3: Key to carbonate sections.

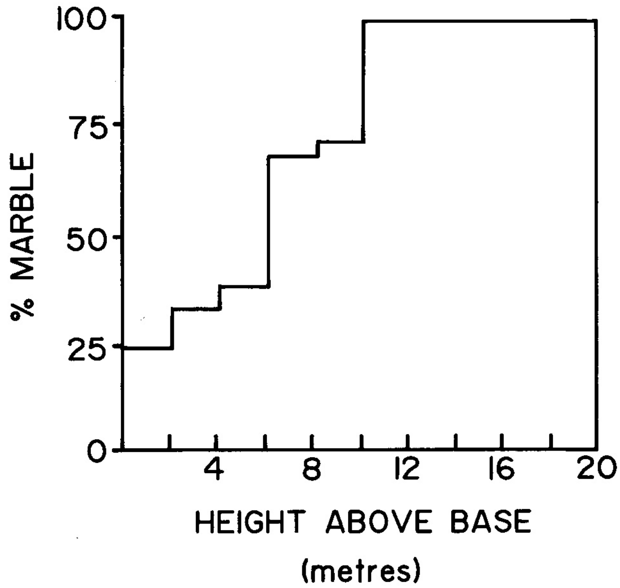


Fig. 2.4: Histogram showing the progressive increase in carbonate content with height in the section depicted in Fig. 2.1.

(Corner section - Fig. 2.4 and Plate 3.£). The succession here is typical of the sequence of facies and facies associations of this formation in that interbedded facies Cc and G are overlain by facies Gc. Near the top of the sequence facies Ac beds are interbedded with facies Gc.

2.4 SYNTHESIS AND DISCUSSION

Both the Brandberg West Formation and Gemsbok River Formation reflect a change from the underlying siliciclastic turbidites to carbonate turbidites (Facies Cc) with interbedded hemipelagic terrigenous muds (Facies G) to hemipelagic settling deposits (Facies Gc) with interbedded carbonate debris flows (Facies Ac). A schematic interpretation of the depositional setting is shown in Fig. 2.7.

Carbonate sedimentation is favoured by higher sea-levels as a broad continental shelf forms on which reefs can develop. There is also a decrease in erosion during higher stands of sea-level (Schlager and Ginsburg, 1981), and consequently the supply of terrigenous sediment to the shelf decreases and the resulting clearer water conditions also favour reef development. Droxler and Schlager (1985)

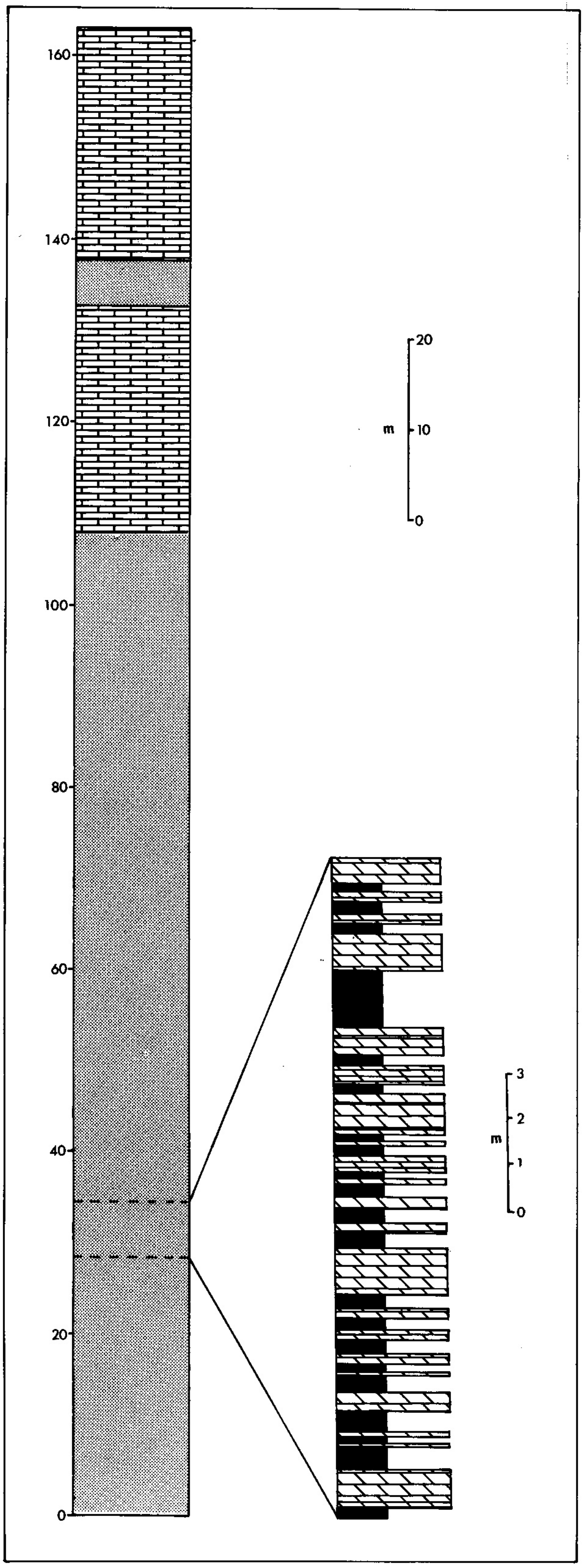


Fig. 2.5: Lookout section of the Gemsbok River Formation.

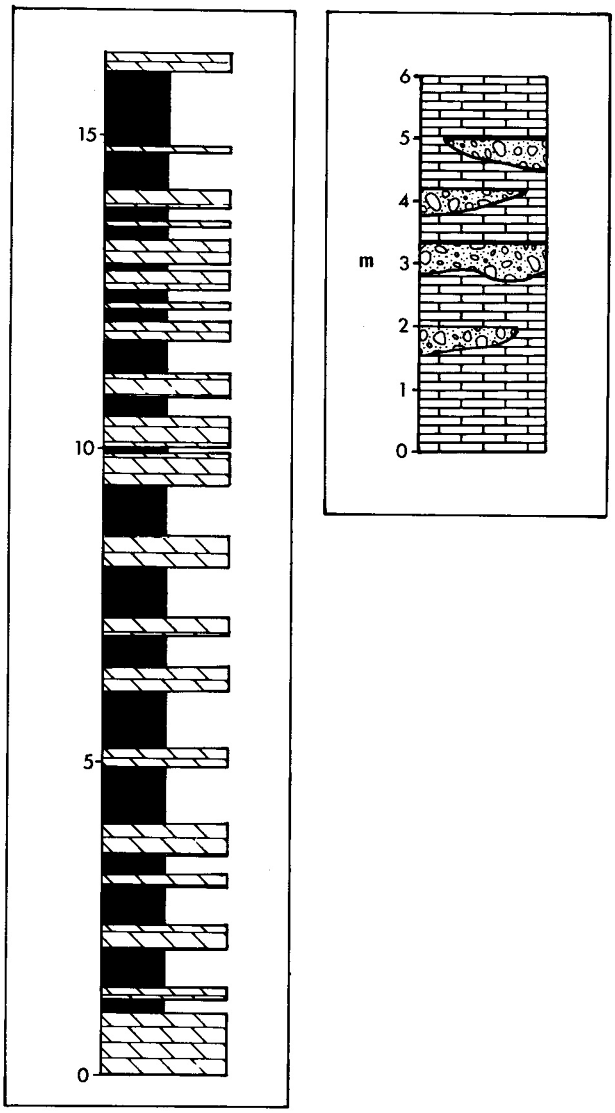


Fig. 2.6: The corner section of the Gemsbok River Formation.

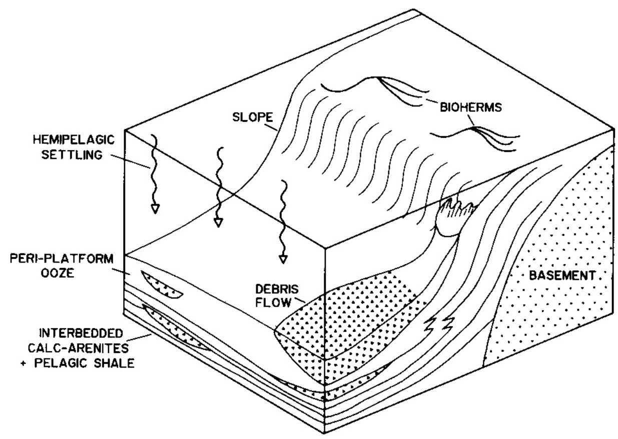


Fig. 2.7: Block diagram illustrating features of the probable palaeogeography during deposition of the Brandberg West and Gemsbok River Formations.

have shown that in the Bahamas the amount of carbonate turbidite activity increased during interglacial high sea-level stands. Deposition of the Brandberg West and Gemsbok River Formations was therefore probably initiated by a rise in sea-level.

The cause of an increase in sea-level at the base of the Brandberg West Formation is unknown, but it was possibly tectonically related. The Brak River Formation which underlies the Gemsbok River Formation has an undoubted glacial influence (see 3.3.3), and it was probably the retreat and melting of the glaciers which caused the increase in sea-level at the beginning of Gemsbok River Formation times.

Carbonate turbidites differ from siliciclastic ones in that sediment is delivered to the slope all along the platform margin by numerous small submarine canyons (McIlreath and James, 1984; Mullins and Cook, 1986). A debris apron results which has a morphology distinct from that of a submarine fan (McIlreath, 1977; Mullins *et al.*, 1984; McIlreath and James, 1984; Mullins and Cook, 1986). These aprons can be further subdivided into inner and outer portions (Mullins and Cook, 1986). The inner apron typi-cally comprises thick resedimented megabreccias (Facies F), thick, coarse-grained turbidites (Facies Ac) and minor peri-platform oozes (Facies Gc). In contrast, the outer apron consists of thinner clast-supported conglomerates (Facies Ac), classic turbidites or allodapic limestones (Facies Cc) and peri-platform ooze (Facies Gc). It is the outer apron association which is found in the Zerrissene Basin. Carbonate turbidites can be deposited adjacent to both depositional or bypass margins (McIlreath and James, 1984), but no distinction between the two can be drawn from the available evidence in the lower Ugab River area. Palaeo-geographic reconstructions of the area, especially identification of the source area, are difficult as palaeocurrent indicators are rare, and where found have been extensively modified by the metamorphism and deformation and there-fore should be treated with caution. This applies in particular to the carbonate layers. The presence of facies Ac beds in the western portions of the area and not in the east would suggest that the easterly parts of the area are more distal and that the source therefore lay to the west. However the area northeast of the Zerrissene basin was covered by extensive carbonate deposits during deposition of both the Brandberg West and Gemsbok River Formations (Hedberg, 1979), and cannot be discounted at this stage as a possible source area.

CHAPTER THREE: THE ZEBRAPUTS, BRAK RIVER AND AMIS RIVER FORMATIONS - DESCRIPTION AND FACIES ANALYSIS

3.1 INTRODUCTION

Each of the three siliciclastic formations (Table 1.1) can be easily identified in the field as they are separated by the distinctive carbonate horizons. In some outcrops the exposures of these formations can be regarded as exceptional according to the graph of Bouma *et al.* (1985; Fig 3.1), and this combined with the good stratigraphic control enables detailed studies of both vertical and lateral facies variations to be carried out. These unusually good outcrops provide an excellent opportunity to examine facies relationships in a submarine turbidite system, a field in which more disagreements than agreements exist at present (Shanmugam and Moiola, 1988).

many cases the emphasis was on attempting to create a single unifying fan model and significant variation between systems was consequently ignored. The factors that control growth of submarine turbidite systems include tectonic setting, sediment supply, basin size, basin morphology and sea-level fluctuations, and a change in one of these will affect many of the others. Consequently no two systems will have exactly the same characteristics, a feature which must be kept in mind when attempting to apply models to a particular system.

The situation is further complicated by the differences in the type and scale of observations between modem and ancient systems (Fig. 3.1). In modem systems our knowledge has, until recently, been confined to the overall morphology

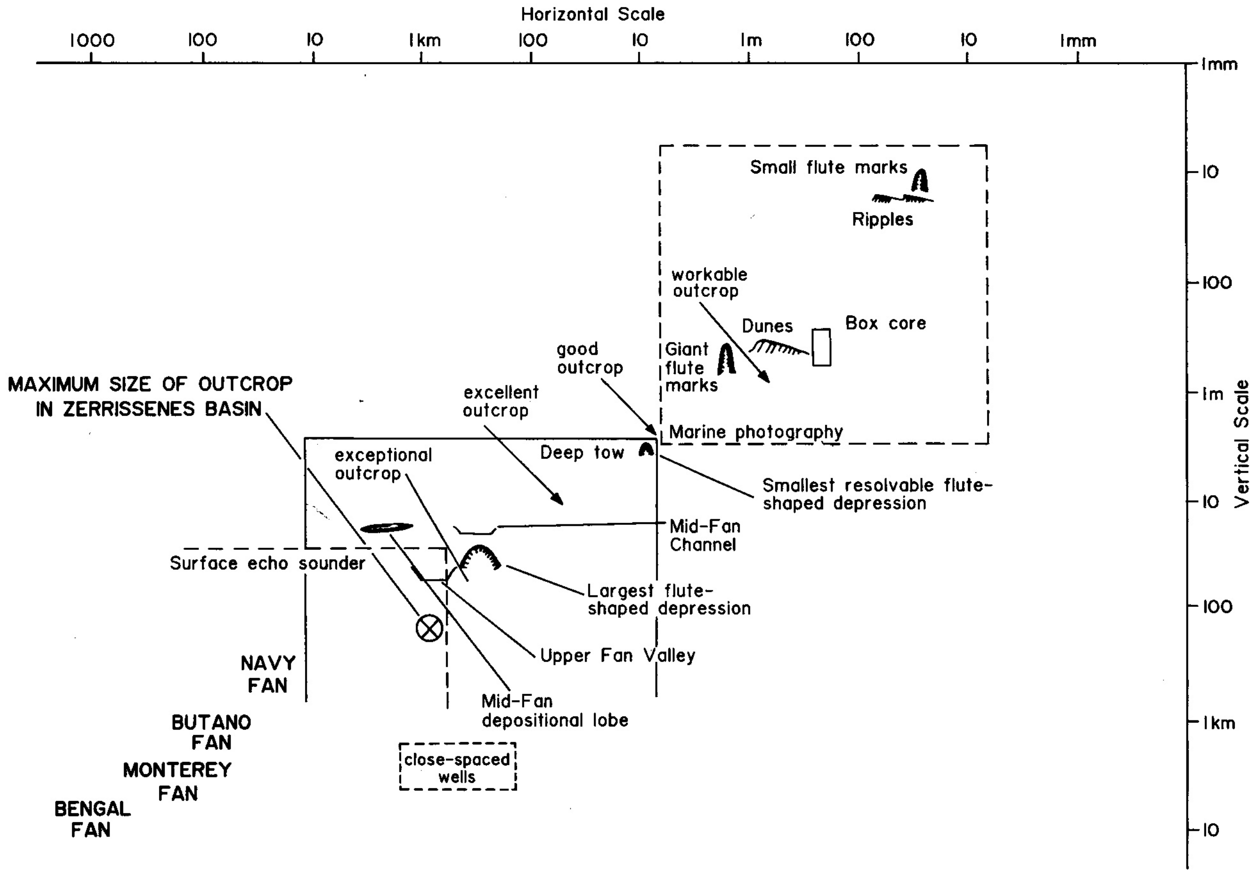


Fig. 3.1: **Graph showing differences in horizontal and vertical scales of observation between modem and ancient fans compared with the maximum size of outcrops in the lower Ugab River area (modified after Bouma *et al.*, 1985). The scale of information obtained from various study methods is also indicated.**

3.1.1 Models of sub)11arine turbidite systems

The submarine fan model was originally proposed by Mutti and Ricci Lucchi in 1972, and an English translation by Nilsen in 1978 brought this important paper to the attention of English-speaking geologists. Subsequent papers have developed and modified this model, but these have often been based on poorly understood systems which are known from only a few outcrops (Bouma *et al.*, 1985). In

of the deposit, and little is known about the internal sequences. In contrast, ancient successions are known almost primarily from vertical sections and morphological features are not easily recognised because of compaction, deformation and lack of three-dimensional exposures. The scale of observation is also important as features which are commonly recognised in outcrop, such as large flute casts and local scours, cannot be detected by the techniques currently used in oceanography, and the huge size of some modem systems prevents adequate sampling (Mutti and Normark, 1987). Modem fans are also commonly much larger than the preserved outcrop area of ancient fans (Fig. 1.5; Barnes and Normark, 1985; Shanmugam *et al.*, 1985; Mutti and Normark, 1987). Temporal differences in the development of turbidite systems also exist (Mutti and Normark, 1987). The differences in scale, both temporal and spatial, and the type of observation are sufficiently great to warrant the suggestion that a moratorium be placed on model building until these differences are resolved (Normark *et al.*, 1985; Mutti and Normark, 1987). Accordingly, no attempt will be made here to compare the Zerrissene system with modem systems, but it will rather be compared with well-described ancient successions.

The terminology of submarine turbidite systems is extensive and non-uniform with the same term being applied to different environments and different terms to the same environments (Walker, 1984; Barnes and Normark, 1985). Some of the problems arise because of use of morphological terms from modem studies to describe ancient deposits where the existence of this type of morphology has not been proven but inferred. It is, however, difficult to discuss these systems without resorting to some of these terms and the terms used in this study are defined here to avoid confusion.

A *lobe* is taken to represent a non-channelised deposition of sand. The use of this term is justified as the three-dimensional exposure in the area is excellent and the sheetlike character of these beds can be recognised.

The *lobe fringe* is the most distal portion of the system and is dominated by thinly-bedded units and possibly compensation cycles (Mutti and Sonnino, 1981). No distinction between the lobe fringe and fan fringe as proposed by Mutti (1977) and Pickering (1981) has been made as they may overlap and the differences between them may be caused by changes in sediment supply rather than their geographic setting on the fan.

The *basin plain* is that portion of the system which is largely free of coarse-grained turbidite deposition and is consequently dominated by pelagic muds, and is uniform over great distances (Mutti, 1977).

3.1.2 Source of the formation names

The Zebrapüts Formation derives its name from a waterhole in the Ugab River and was named as such by Miller *et al.* (1983). It was previously known as the Zebra River Formation (SACS, 1980) but was changed by Miller *et al.* (1983) because of another formation in Namibia with a similar name. The Brak River Formation is named after a tributary of the Ugab River just west of the old Brandberg West mine (SACS, 1980) and previously included the Brandberg West Member which was later given formation status by Miller *et al.* (1983; see 2.1.1). The Amis River Formation was named after a river which drains from the Brandberg (SACS, 1980). The name was taken from the 1 :50 000 topographical map and is unfortunate as the map is incorrect. No outcrops of this formation occur in the actual Amis River, but the name has been retained here to avoid confusion.

3.2 FACIES TYPES

3.2.1 Introduction

The facies nomenclature used in this study is based on the scheme proposed by Mutti and Ricci Lucchi (1972, 1978) and modified by them in 1975. The scheme comprises seven different facies types, designated facies A to G, with various subcategories of these. Their system is useful for facies analysis as relationships between facies and variations of these facies can more easily be described and understood. The system is largely descriptive and is based on the recognition of grain size, bed thickness, bed geometry and internal sedimentary structures (Nelson and Nilsen, 1984). A more recent classification by Pickering *et al.* (1986), defined largely on the grain size and also the internal organisation and composition of beds, unfortunately used the same A to G designations, but for different facies. This can give rise to great confusion. The Pickering *et al.* (1986) scheme is also difficult to apply in the field (Shanmugam and Moiola, 1988) as it has 40 different categories, many of which require recognition of fine grain sizes. The low- grade metamorphism of the rocks studied here may, however, obscure subtle differences in grain sizes, thus making the Pickering *et al.* (1986) scheme difficult to apply. In addition, excellent exposures are required to recognise many of the features. However, certain glacially related facies found in the Zerrissene system are not adequately described by the Mutti and Ricci Lucchi (1972; 1975; 1978) scheme and modifications have been made. These are discussed below.

3.2.2 Facies Az

This facies consists of matrix -supported, coarse-grained greywackes. Beds of this facies have limited lateral extent, have a maximum thickness of 3 m and are commonly amalgamated. Rip-up clasts occur, but no other sedimentary structures are present (Plate 4.A). This facies is therefore equivalent to the disorganised conglomeratic sandstone facies (facies Az) of Mutti and Ricci Lucchi (1975). In the Zerrissene system outcrops of this facies are uncommon and are limited to the Amis River Formation. The best exposures occur in a small gorge just west of the Lonely River (Fig. 1.3)

A poorly exposed outcrop of similar material is found in the eastern part of the area, but these beds do not appear to be very extensive or very thick. An enigmatic coarse-grained quartzwacke is found southeast of the Ugab River near the Brandberg. The wacke is up to 10 m thick and is well graded in places (Miller *et al.*, 1983), but no structures indicating palaeocurrents have been observed. It can be traced out on aerial photographs for several hundred metres and disappears southwards under the Karoo cover, but is again present south of the Messum Complex (Fig. 1.6). No other unit in the entire system is similar in appearance to this quartzwacke.

Mutti and Ricci Lucchi (1972; 1978) ascribed the origin of this type of deposit to highly erosive grain flows. Disorganised, coarse-grained units such as this facies are however now regarded as the result of deposition by high-den-

Plate 4.A: Rip-up clasts of shale resting in a coarse-grained greywacke of the facies A_2 type. Younging direction is to the bottom of the photograph (locality: west of the Lonely River).

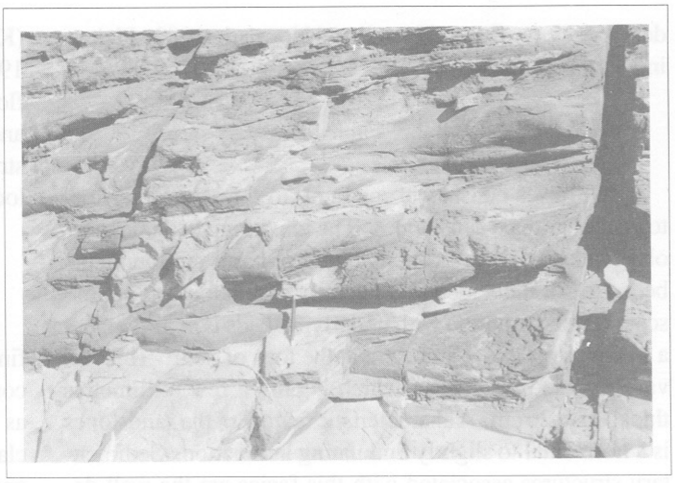


Plate 4.B: Flute casts on sole of a facies B_1 bed. These flute casts give a north-south palaeocurrent (locality: Lonely River).

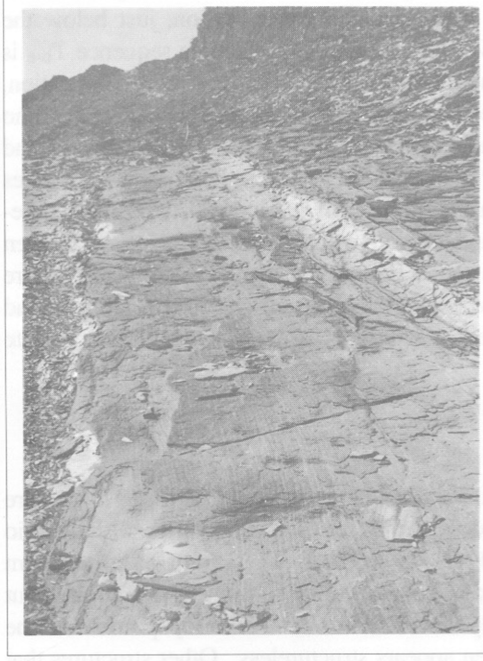


Plate 4.C: 3 m-thick facies B1 bed with well developed horizontal lamination and 20 cm-thick pelitic top (loc. Lonely River section).

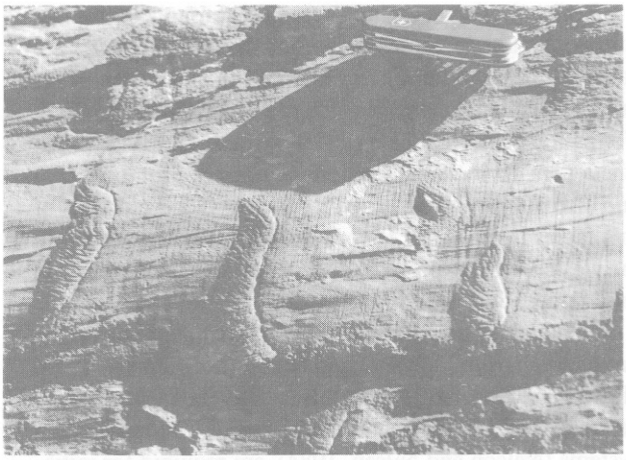


Plate 4.D: Frondescent casts on sole of a facies B_1 bed (locality: Jeppe East section).

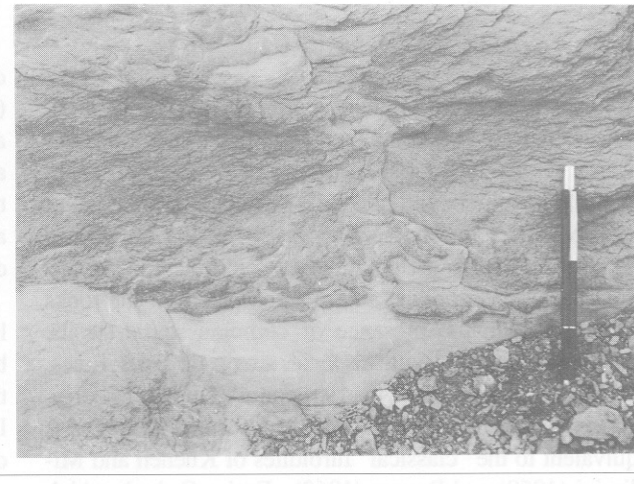


Plate 4.E: Dewatering structures at base of a facies B_1 bed (locality: Lonely River).

sity debris or slurry flows (Howell and Normark, 1982; Pickering *et al.*, 1986).

3.2.3 Facies B_1

This facies consists of medium- to thick-bedded, fine- to coarse-grained sandstones. The maximum bed thickness observed was 11 m. Although erosive bases are found, the beds are parallel-sided and laterally continuous on a gross scale. Amalgamation with underlying beds is common, but a pelitic layer may be present at the tops of beds. This can vary from a thin veneer, which is only a few millimetres thick, up to 20 cm. A characteristic feature of the sandstones is a horizontal to slightly undulating lamination. Sedimentary structures associated with this facies are the well-developed horizontal lamination, dewatering structures, load structures, large-scale cross-stratification, rip-up clasts, frondescent casts and flute casts (Plate 4.B; 4.C; 4.D and 4.E). The characteristics of this facies are similar to facies B_1 of Mutti and Ricci Lucchi (1975).

Beds of this facies are common throughout the sequence. In the Zebrapüts Formation the best exposures are found in the Skeleton Coast Park section in the west of the area. Excellent exposures are found in the Brak River Formation in the Zebra River, Gai-as south, Gemsbok Altar, Shelter and Rhino Wash sections (see Fig. 1.3 for localities) and also at other unmeasured localities. Good outcrops of this facies in the Amis River Formation are only found in the western most exposures of the study area, particularly in the Lonely River area.

Although Facies B_1 was previously thought to be the product of grain flows (Mutti and Ricci Lucchi, 1978; Howell and Normark, 1982), this is unlikely as steep slopes of 18-30°, which are not normally found in submarine fan systems (Nelson and Nilsen, 1984), are required for initiation of these flows (Middleton and Hampton, 1976). Instead they are likely to be the product of high-density liquefied flows which can move down slopes of 3-10° (Middleton and Hampton, 1976). Facies B [beds may grade both vertically and laterally into facies C beds (Dzulynski and Slaczka, 1958; Mutti, 1985).

3.2.4 Facies C_2

Fine- to medium-grained, extremely persistent sandstone which are between 10 and 150 cm thick comprise this facies. Complete Bouma sequences (T_{a-e}) were observed but are uncommon, and generally only beds with T_{abce} or T_{abe} sequences can be seen. No distinction between a turbiditic and a pelagic pelitic layer could be made in this study because of the age and metamorphism of the rocks which preclude the presence of common micro-fossils. Structures associated with this facies are rip-up clasts, flutes, dewatering structures and load structures. This facies corresponds to facies C_2 of Mutti and Ricci Lucchi (1975) and is equivalent to the "classical" turbidites of Kuenen and Migliorini (1950) and Bouma (1962). Facies C_2 beds which are up to 1.5 m thick are found in the Amis River Formation, west of the Lonely Wash.

Facies C_2 is the result of classical turbidity currents (Kuenen and Migliorini, 1950; Middleton and Hampton, 1976), which possibly evolved from slurry or fluidised flows. Erosion at the head of the turbidite will develop flute and groove casts. The vertical succession of sedimentary structures is explained by waning flow over a specific location (Middleton and Hampton, 1976).

3.2.5 Facies D_1

This facies consists of very thinly- to medium-bedded, fine-grained sandstone-shale couplets. Ripples (T_c) are common and the sandstone:shale ratio varies from as much as 12:1 to 1:1. Beds with a higher proportion of shale are classed as facies D_2. Individual beds are graded (Plate 5.A) and laterally continuous. They are commonly found in bundles up to 35 m thick. These beds correspond to the facies D_1 beds of Mutti and Ricci Lucchi (1975), but differ in that they may be as little as 0.5 cm thick, yet are still graded and laterally continuous (Plate 5.B). Associated structures include rip-up clasts, flute casts, load structures, concretions and dewatering structures, some of which are shown in Plate 5.C and 5.D.

Facies D_1 beds are generally regarded as being the deposits of low-density turbulent flows (Howell and Normark, 1982) with deposition from traction being succeeded by suspension settling.

The best outcrops of this facies in the Zebrapüts Formation are in the Goantagab River section, just below the Brandberg West Formation. The Bouma sequence T_{b-e} is developed here (Plate 5.E). In the Brak River Formation, good outcrops of facies D_1 beds are found in the Rhino Wash, Shelter, Gemsbok Altar, Gai-as south, Campsite and Zebra River measured sections. Again, partial sequences are common and rip-ups, dewatering structures and concretions are found. Superb examples of this facies can be seen in the Amis River Formation in the Lonely River where well-exposed examples of rip-up clasts, concretions, load casts and partial Bouma sequences were observed (Plate 5.F).

3.2.6 Facies D_2

Beds of this facies are transitional with facies D_1 and are differentiated from that facies by a lower sand:shale ratio (less than 1:1). Beds vary in thickness from 0.5 cm to 15 cm and are laterally continuous. T_{cde} sequences are found but are relatively uncommon and the sandy portion of the turbidite often appears structureless. Other structures that are found include rip-up clasts, load features, small flute casts and dewatering structures.

This facies also results from deposition from relatively low density flows, with traction deposition being followed by suspension fall-out. Facies D_2 is not as widespread as the above facies but is often found in association with facies D_1 in all three formations, especially in the eastern outcrops of the Amis River Formation.

3.2.7 Facies E

Beds of this facies do not form a significant proportion

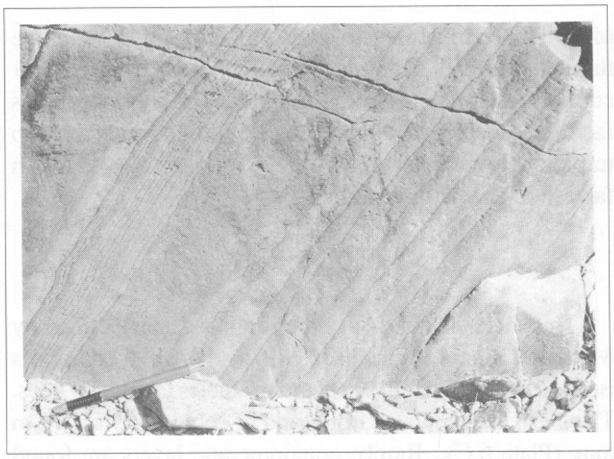


Plate 5.A: D_1 beds in the Brak River Formation which are only 0.5 cm thick but are laterally continuous with very thin pelitic tops (locality: east of the Rhino Wash).

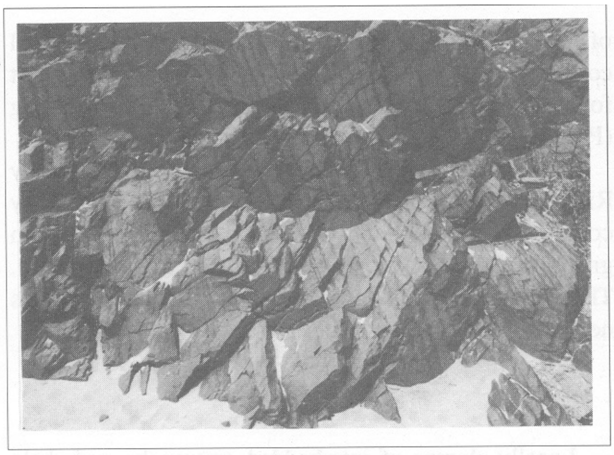


Plate 5.B: Deformed D_1 beds in the Brak River Formation showing good lateral continuity (locality: immediately above the Ugab River measured section of the Brandberg West Formation).

Plate 5.C: Turbidite in process of ripping up a shale clast (locality: Lonely River section).

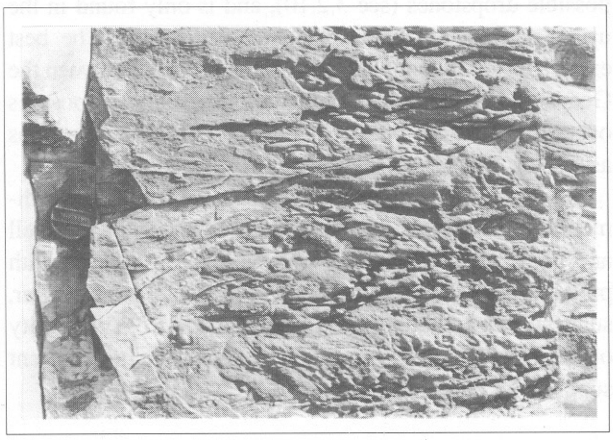


Plate 5.D: Load structures at the base of a thin facies D_1 bed (locality: east of Lonely River Section).

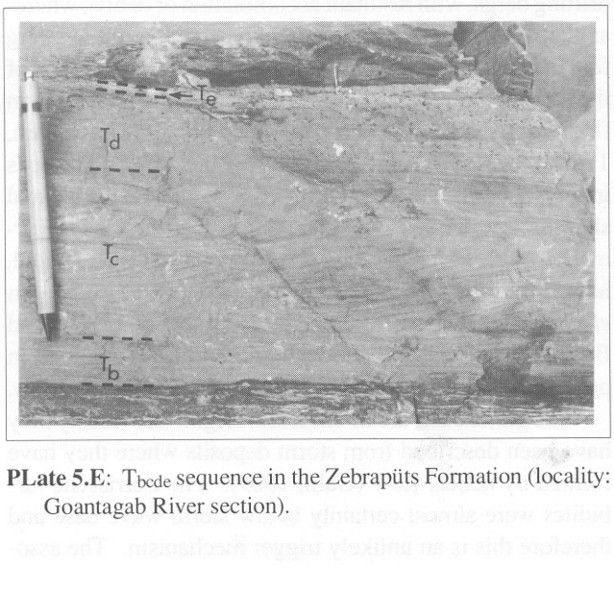


PLate 5.E: T_{bcde} sequence in the Zebrapüts Formation (locality: Goantagab River section).

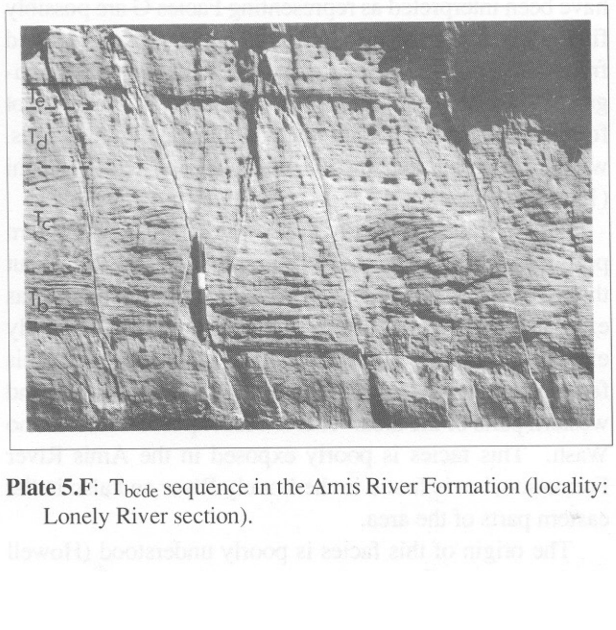


Plate 5.F: T_{bcde} sequence in the Amis River Formation (locality: Lonely River section).

of the sequence. They are best recognised by a lack of lateral continuity of individual units, a coarser grain size than that found in facies D, a high sand:shale ratio and flaser bedding (Plate 6.A). They are commonly 10 to 20 cm thick.

Facies E beds have only been recognised in the Lonely River section in the western most portions of the area. They commonly represent the deposits of high-concentration gravity and traction flows (Howell and Normark, 1982). This facies is normally associated with overbank deposits near mid-fan channels.

3.2.8 Facies F

Locally slumps of interbedded grey wacke and shale occur (Plate 6.B). The beds which are slumped are generally of facies D_1 and D_2, but the sand and mud portions have remained discrete (Plate 6.B). These slumped beds are up to 1 m thick, but are not laterally continuous. Slumped beds of this nature are characteristic of the facies F of Mutti and Ricci-Lucchi (1978).

This facies commonly contains outsize clasts, which are possible dropstones (see 3.2.10), and is only found in the upper portions of the Brak River Formation. The best exposures are found in a gorge east of the road between the Goantagab River and Brandberg West. Other exposures occur in the Rhino Wash, Shelter and Brak River sections as well as in tributaries of the Sesob River.

Facies F is generally believed to have formed by sediment failure near the lower slope or along levees (Howell and Normark, 1982). The association of the facies with probable glacial dropstones (facies H - see 3.2.10) however, suggests that in this case the slumps formed from instability of the beds caused by the increased volume of sediment supplied by fallout from icebergs.

3.2.9 Facies G

In the Zerrissene basin, facies G consists of fine-grained, thin-bedded shales. In the lower Ugab River area the deep weathering along cleavage planes has resulted in poor exposure of this facies. Consequently, many of the beds that have been interpreted as representing Facies G are possibly fine-grained, graded turbidites similar to those described from the Sevier Shale Basin of eastern Tennessee (Shanmugam, 1980). However all fine-grained material that does not form part of other turbidite facies is included in this facies, which is similar to facies G of Mutti and Ricci Lucchi (1978).

Much of the Zebrapüts Formation in the northwestern portions of the area is possibly represented by this facies but the exposure is poor. The Brak River Formation also has extensive development of this facies but again it is poorly exposed. In particular a 30 m-thick development of this facies occurs at the top of this formation in the central and western parts of the area. This is best exposed in the Rhino Wash. This facies is poorly exposed in the Amis River Formation, but is found in the Lonely River and also in the eastern parts of the area.

The origin of this facies is poorly understood (Howell and Normark, 1982) and probably involves the interaction of a number of processes including deposition by dilute turbidity currents, from nepheloid layers or from contour currents (Howell and Normark, 1982). The presence of this facies may represent either a sea-level highstand when no turbidite flows could reach the sea-floor (starved basin facies), or abandonment of a lobe.

3.2.10 Facies H

This facies is not one of the original facies defined by Mutti and Ricci Lucchi (1972; 1975; 1978). It comprises isolated clasts generally set in typical facies D_1 and D_2 beds. The clasts range in size from sand to cobbles up to 1.75 m long (Plate 6.C). Rarely, conglomeratic layers are found (Jeppe, 1952; Miller *et al.* 1983; Plate 6.D). These layers are not laterally extensive, having a maximum strike length of 8 m and a maximum thickness of 2 m. Some layers are pod-like, being convex-up, only 2 m long and up to 1 m thick. The fabric of these conglomerates is generally matrix-supported. Clast types include granite, gneiss, quartzite, marble and gabbro in order of decreasing abundance. An unusual set of U-shaped gutter casts is associated with this facies in the Brak River section (section 3.3.3.2). These gutter casts are found at the base of greywackes and contain abundant clasts up to 8 cm in diameter (Plate 6.E). The gutters have a maximum width of 8 cm and a maximum exposed length of 60 cm.

This facies is only found in the upper portions of the Brak River Formation and is confined to the western and central parts of the area.

The presence of large, isolated clasts resting in thin facies D_1 and D_2 beds is strongly suggestive of glacial deposition, possibly by ice-rafting (Jeppe, 1952; Miller *et al.*, 1983). Further evidence for a glacial origin is the presence of clasts which appear to pierce the underlying beds, but this type of feature may also be caused by compaction (Plate 6.F). Rafting of material in tree roots is precluded by the age of the sediments. The conglomerate bands may be caused either by winnowing of fine material or more probably by the foundering of overloaded icebergs. The latter mechanism is caused by ablation at the surface of drifting bergs, with resultant accumulation of debris, whereas the submarine portion melts more rapidly and the berg founders because of a mass imbalance. The rolling over of the iceberg dumps the ablation debris as a pile on the ocean floor which is later preserved as a lens or pod (Lavrushin, 1968, quoted in Powell, 1983; Ovenshine, 1970). The beds in which debris is accumulating may become unstable and slumping can result. The glacial origin of the Chuos Formation, with which the Brak River Formation is correlated, has recently been questioned (Schermerhorn, 1974; Martin *et al.*, 1985), but recent work by Henry *et al.* (1986) and Badenhorst (1987) has shown that the Chuos is, at least in part, glacial in origin; a conclusion supported by this study.

The gutter casts are of uncertain origin. Previously they have been described from storm deposits where they have formed by helical flow (Galli, 1989). The Zerrissene turbidites were almost certainly below storm wave base and therefore this is an unlikely trigger mechanism. The association of clasts, in some cases as large as the gutters

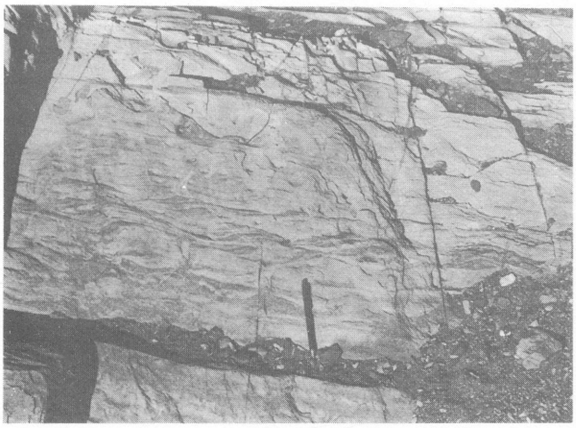


Plate 6.A: Facies E, shown here as flaser structure, is not common in the sequence and is only found in the Amis River Formation (locality: Lonely River section).

Plate 6.B: Associated with facies H (dropstone facies) is facies F which comprises slumped beds of facies D_1 and D_2 in which the mud and sand material have remained discrete from each other (locality: east of road to Brandberg West from the Goantagab River).

Plate 6.C: Granite boulder (outlined), 1.75 m in diameter, resting in thin- to medium-bedded facies D_1 and D_2 beds is good evidence for ice--rafting (locality: tributary of the Sesob River).

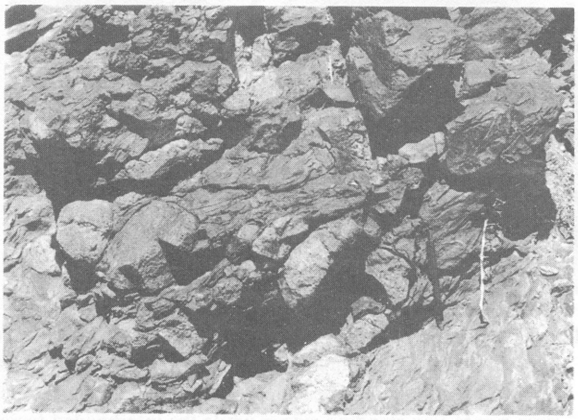


Plate 6.D: Occasional conglomeratic layers are also associated with facies H and represent deposition by foundering of overloaded icebergs (locality: Sesob River).

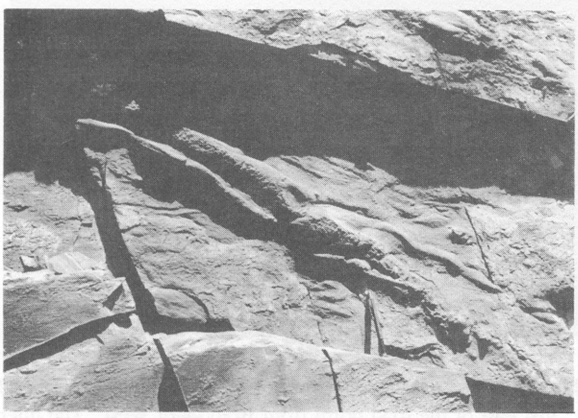


Plate 6.E: Gutter casts containing pebbles such as in this photograph are found near the top of the Brak River Formation (locality: north of the Brak River Section).

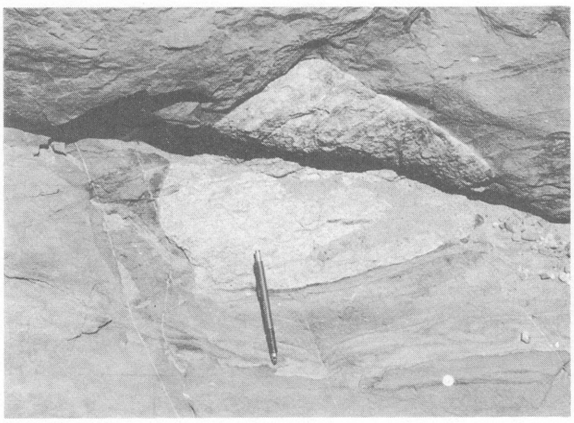


Plate 6.F: Piercing of underlying laminae by pebbles or cobbles such as in this photograph may be caused either by compaction or glacial dropstones (locality: Sesob River).

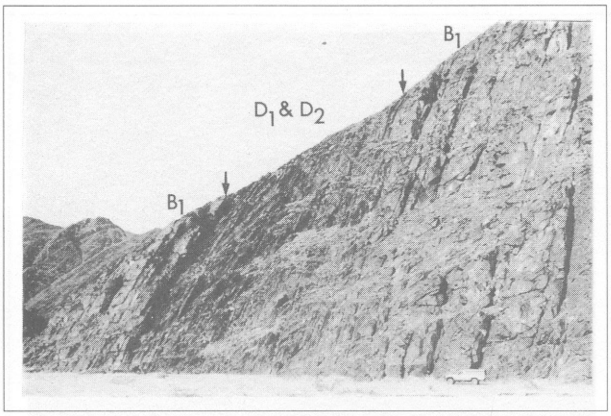


Plate 7.A: Photograph of Augur section depicted in Fig. 3.7. The younging direction is to the right. Note the distinctive package of interbedded D_1 and D_2 beds (front of vehicle), overlain by thick facies B_1 beds (behind vehicle).

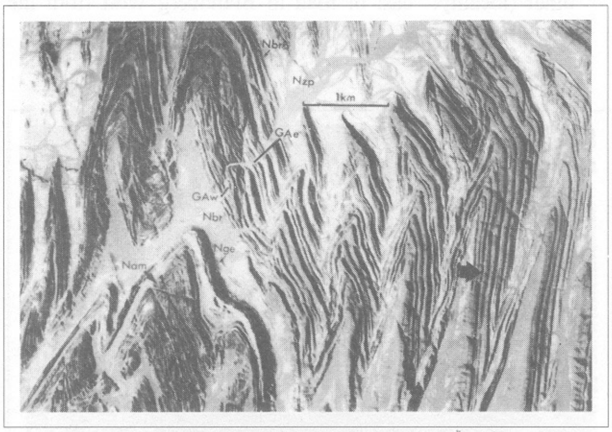


Plate 7.C: Portion of aerial photograph 724/11/434 showing the superimposed river course in which the Gemsbok Altar east and west sections were measured. Individual lobe sandstones give rise to low standing ridges while the interbedded facies G beds form valleys. The channel discussed in section 3.3.3.8 is indicated by the arrow. Reprinted with permission of the Surveyor-General. GAe - Gemsbok Altar east; GAw - Gemsbok Altar west; Nzp - Zebrapüts Fm.; Nbrb - Brandberg West Fm.; Nbr - Brak River Formation; Nge - Gemsbok River Fm.; Nam - Amis River Fm. Scale is approximate.

Plate 7.E: Interbedded lobe sandstones and basin-plain muds just east of the Jeppe east section. This type of sequence is typical of outer fan to basin-plain deposits. Beds young towards bottom of photograph.

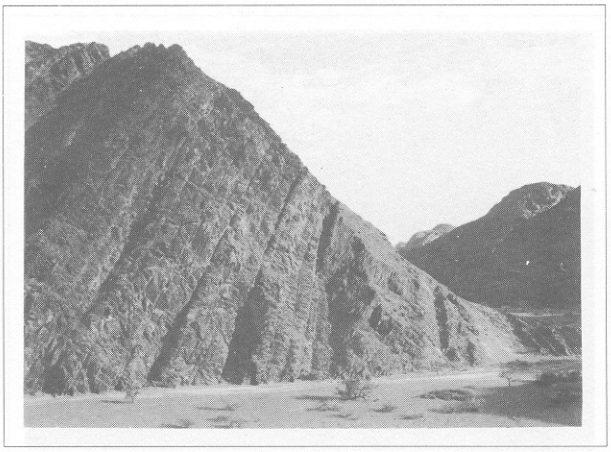


Plate 7.B: Photograph, looking south, of the Rhino Wash section showing the tabular nature of the bedding in this section. The beds are overturned and the younging direction is to the right of the photograph.

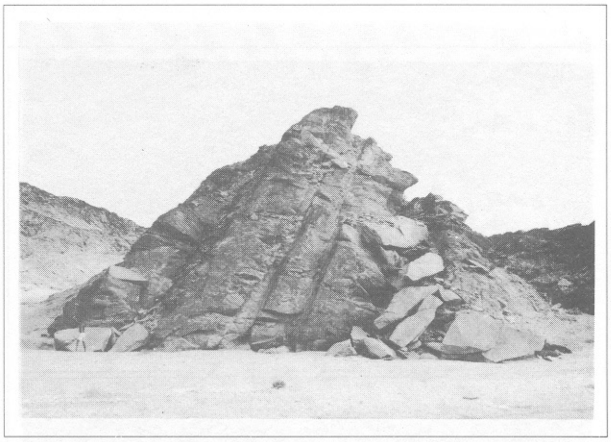


Plate 7.D: Photograph of the uppermost lobe sandstone in the Gemsbok Altar east section. The sandstone package is sandwiched between facies G beds. Note the person for scale in the lower left.

Plate 7.F: View, looking south, from the prominent dolerite hill just east of the road north from Brandberg West. The interbedded nature of the basin-plain sediments, which weather negatively, and the greywackes is again prominent. The lateral continuity of individual beds is also striking. Beds young to the right of the photograph.

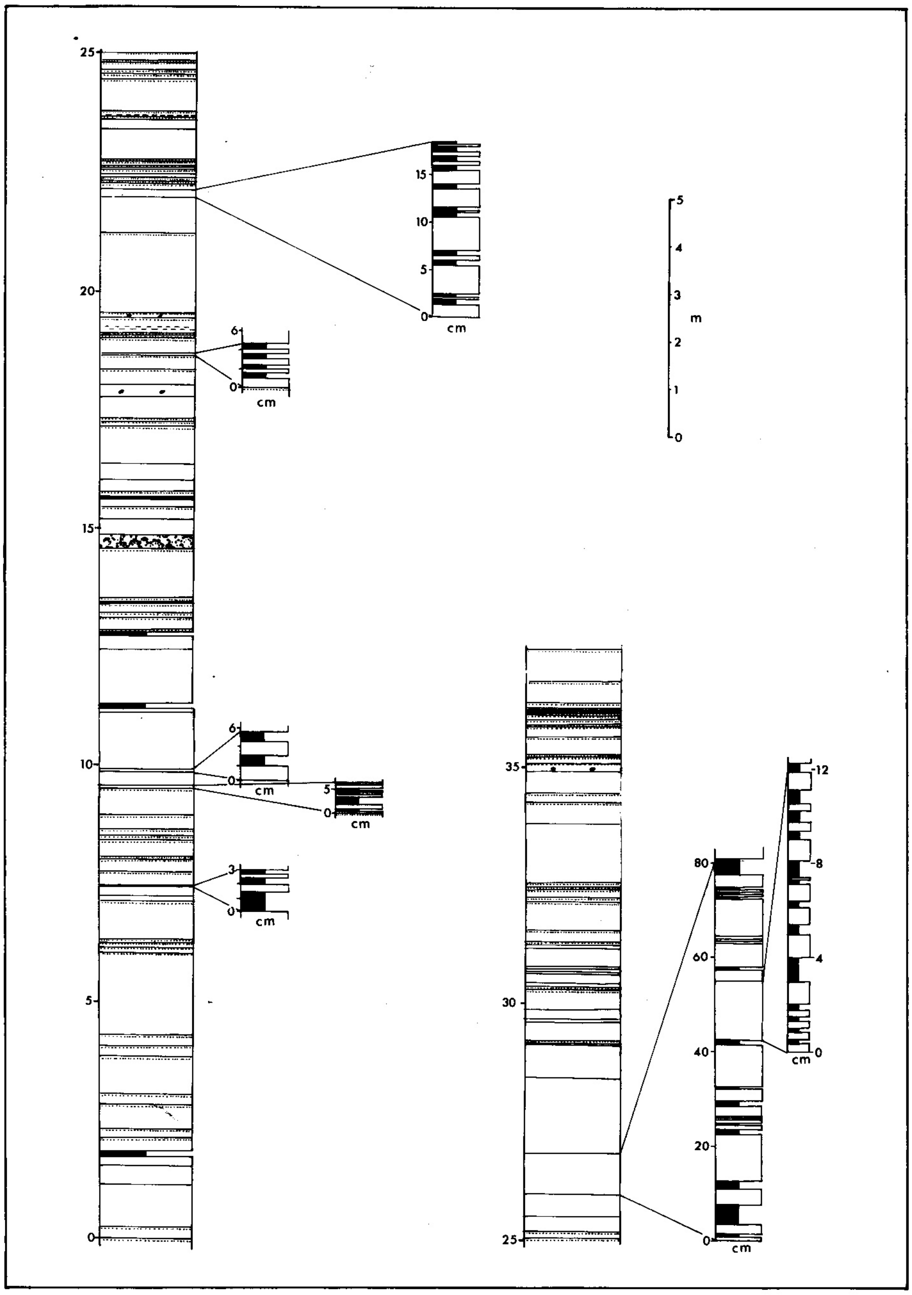


Fig. 3.2: Graphic log of the Ridge 6 section through the Zebrapüts Formation. The section is dominated by facies D_1 with a high sand:shale ratio. Most beds are capped by a very thin pelitic top. For the key to all sections of the siliciclastics see Fig. 3.3.

themselves, suggests that they were influential in the formation of the gutters. The clasts would have been either dropstones on the sediment surface at the time of formation of the gutters, or were entrained in the current. As dropstones would have had a random distribution on the sea-floor, and because the clasts are preferentially concentrated in the gutters, they must have been transported and aligned by the current. The clasts may have been eroded from the sea-floor by the current and their presence in the flow may have changed its characteristics to one of helical flow. This helical flow would have both eroded into the underlying bed and also concentrated the clasts.

3.3 VERTICAL SEQUENCE ANALYSIS

3.3.1 Introduction

Asymmetrical cycles in turbidites are commonly defined on visual features such as field characteristics (Kukla *et al.*, 1988), plots of bed thickness versus bed number (van Vliet, 1978) and graphic logs (Ghibaudo, 1980; van Vliet, 1978). These are subjective techniques and often more than one interpretation can be applied to a particular sequence (Hiscott, 1981; Walker, 1984). Similar equivocal successions are found in the Zerrissene basin such as in the Jeppe east

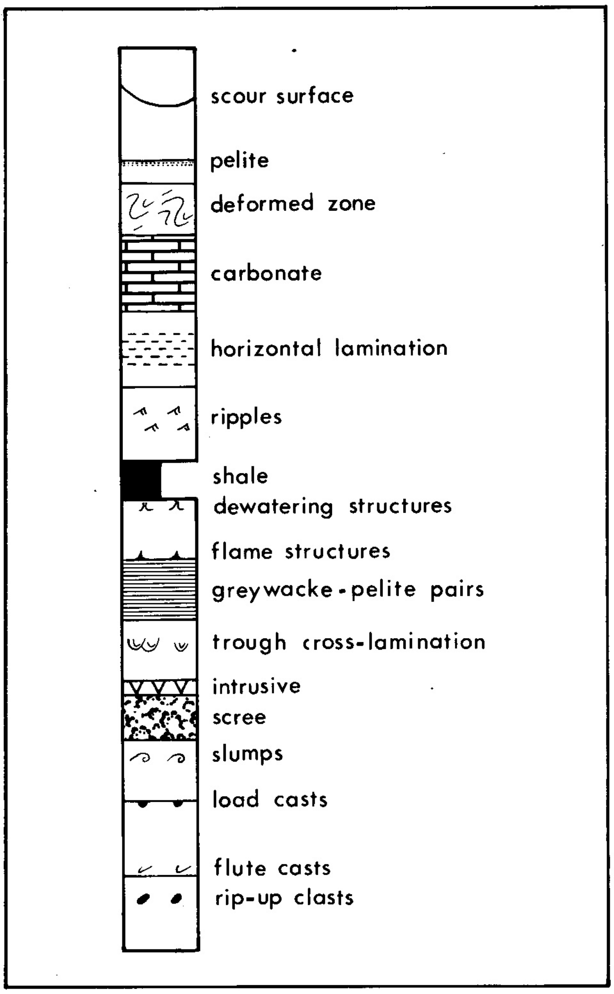


Fig. 3.3: Key to all sections.

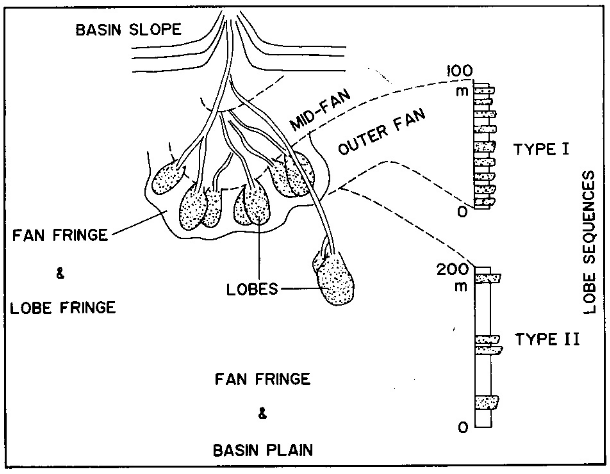


Fig. 3.4: Submarine fan model showing different types of lobe sequences as defined by Pickering (1981). Type I sequences develop by continual switching and stacking of lobes to produce sand-dominated sequences, whereas intermittent, irregular progradation of lobes over the basin plain produces mud-rich sequences, or type II (from Pickering, 1981, his Fig. 9).

section (see 3.3.4.1). Closer investigation of some apparent cycles visible in the field revealed that these were actually non-cyclic successions and consequently the identification of cycles is treated with caution.

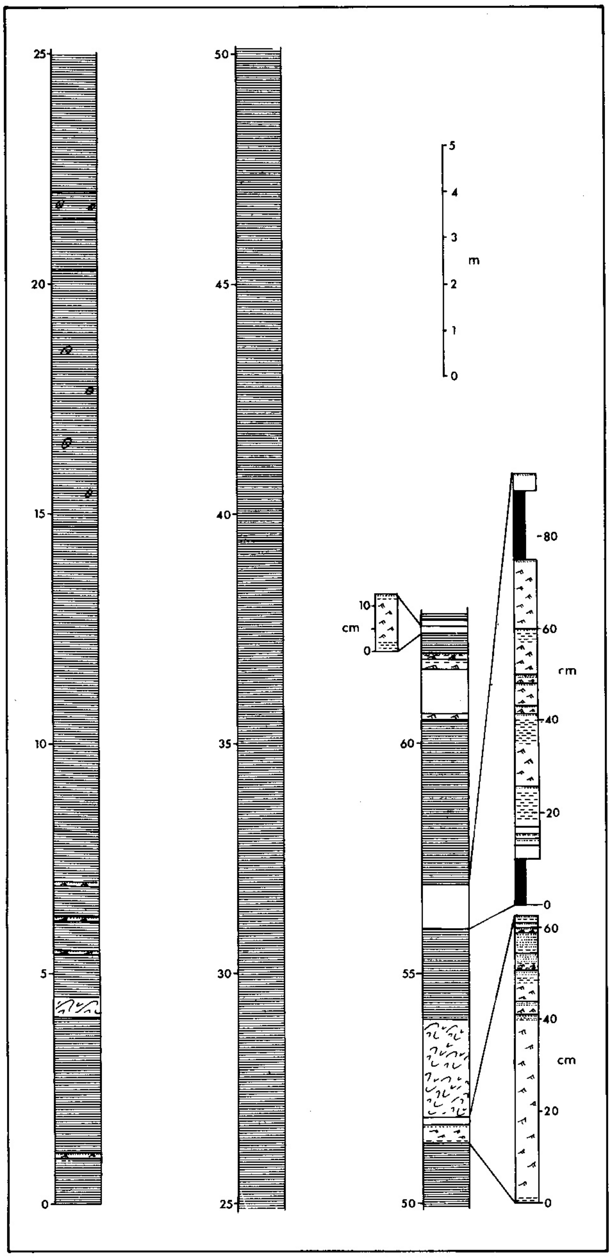


Fig. 3.5: Sedimentary log of the Goantagab River section of the Zebrapüts Formation.

3.3.2 Zebrapüts Formation

Exposures of the Zebrapüts Formation are not well developed because in the northwest, where the majority of the outcrop of this unit occurs, there are no major drainage lines. Further, in the central portions of the area, sedimentary features have been destroyed by a north-northeast trending shear zone. However, four detailed sections from a wide geographic area have been measured and are described below.

The total thickness of the formation cannot be determined as the base is not exposed. A maximum exposed thickness for this formation obtained from field measurements is 350 m, which is somewhat less than the 500 m suggested by Miller *et al.* (1983). The top of this formation is defined as the base of the first marble layer in the overlying Brandberg West Formation, which was used as the datum for correlation of sections.

3.3.2.1 *Ridge 6 section*

This section (see Fig. 3.2 and 1.3 for locality) is dominated by thinly-bedded facies D_1 and D_2 units. The maximum bed thickness recorded was 1.78 m and the average bed thickness was only 0.26 m. Graded greywacke-pelite pairs may be as little as 0.5 cm thick in places, yet they are still laterally continuous. The sandstone:shale ratio is also generally greater than 1: 1 (facies D_1). Rare rip-up clasts and parallel lamination are also present, but no other sedimentary structures were observed.

Minor non-repetitive, thinning- and thickening-upward sequences are present. Individual beds are too thin to be traced out on aerial photographs, but the package is laterally continuous and is encased within blankets of poorly exposed facies G beds.

The thinly-bedded nature of the sandstone packages and their associated fine-grained blankets suggest that this section represents a distal lobe which prograded over a lobe-fringe area (cf. Mutti, 1977; Pickering, 1981; Nelson and Nilsen, 1984). Pickering (1981) distinguished between lobe-fringe (type I) and fan-fringe (type II) lobes. Type I lobe sequences are characterised by a larger proportion of lobe sands and a lower percentage of fan-fringe deposits

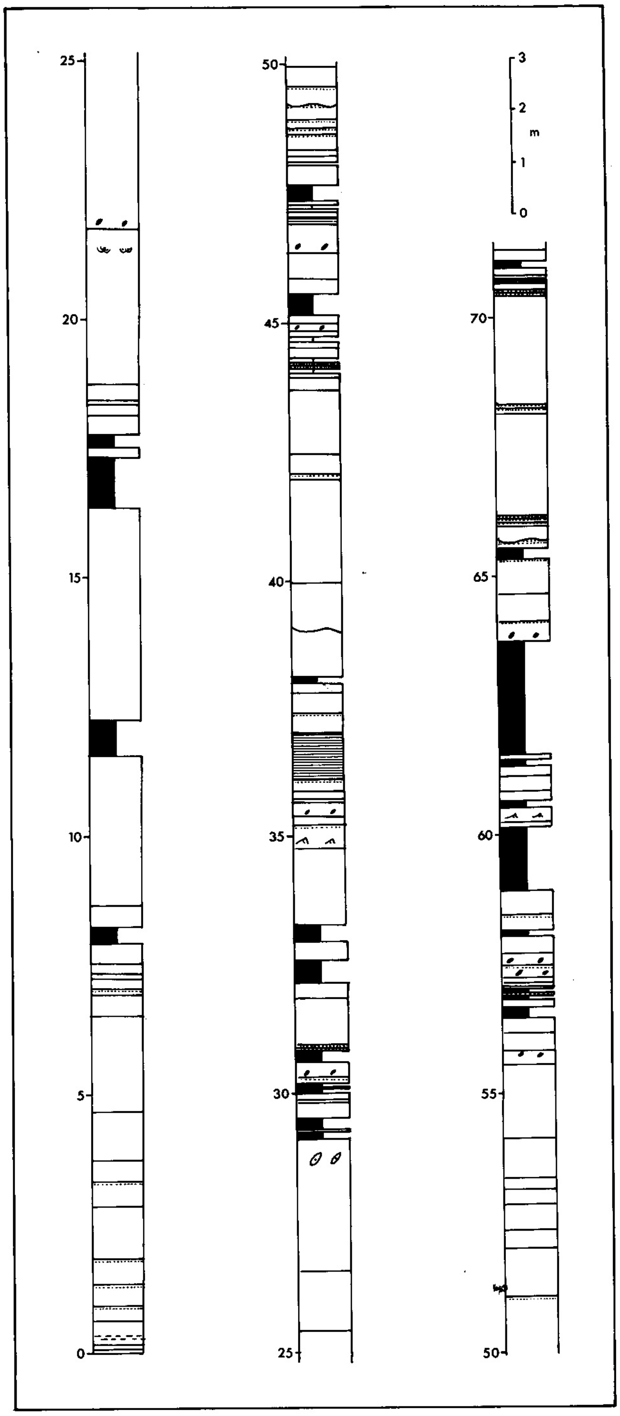


Fig. 3.6: Sedimentary log of the Skeleton Coast Park section of the Zebrapüts Formation, which is dominated by a high sand sand:shale ratio.

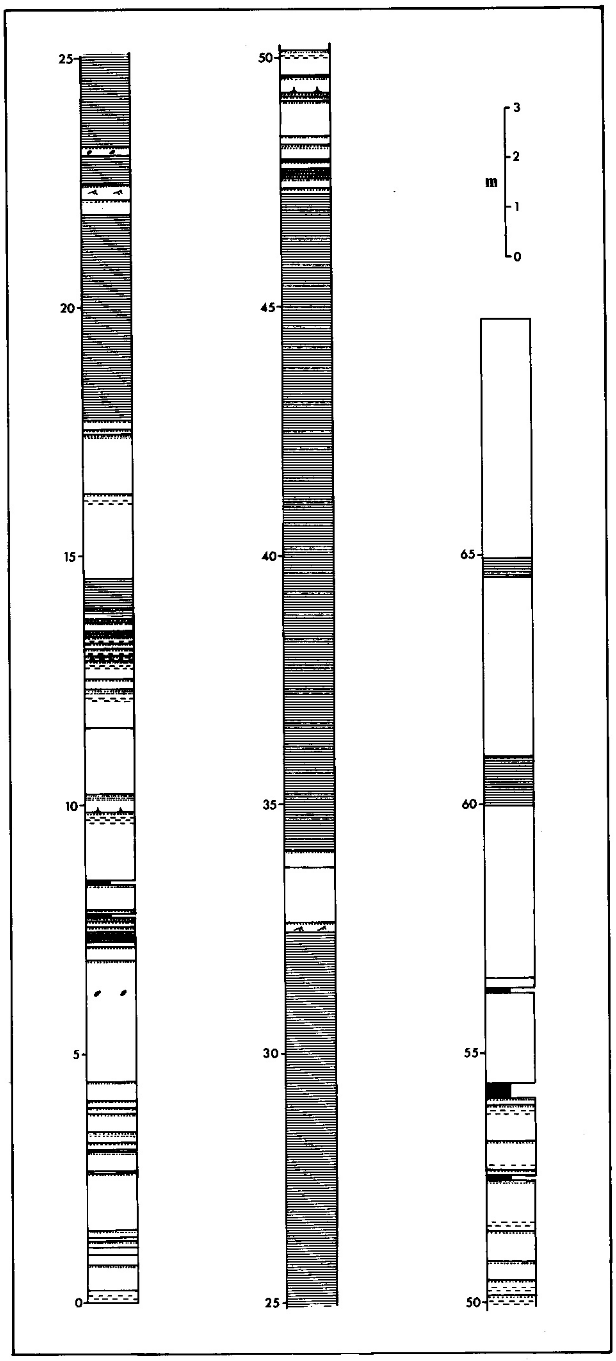


Fig. 3.7: Sedimentary log of the Augur section of the Zebrapüts Formation. See Plate 8.A for a photograph of this section.

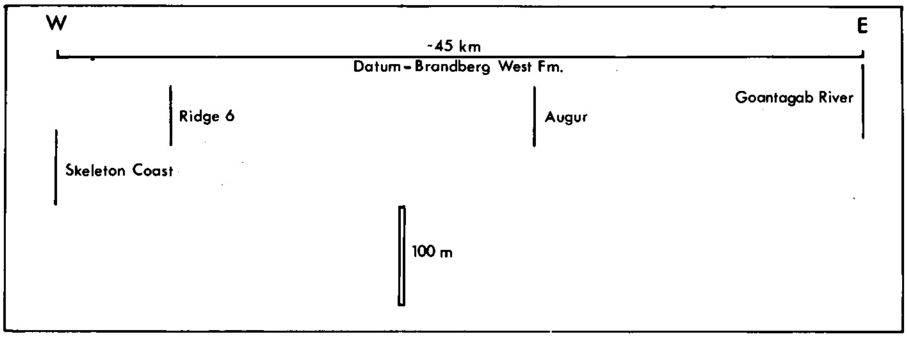


Fig. 3.8: Relative stratigraphic position of the sections measured in the Zebrapüts Formation with the base of the Brandberg West Formation as a datum.

than type II lobes (Pickering, 1981; Fig. 3.4). Pickering ascribes the origin of type I lobes to normal, stable deposition in the basin and type II lobes to changes in the sediment supply caused by either channel switching or changes in the source area. These differences are therefore not caused by relative proximity and can therefore not be used to indicate position on the fan. Accordingly, no distinction between the fan- and lobe-fringe is made here. It is important to note that the type I and II lobe sequences of Pickering (1981) are not the same as the type I, II and III systems of Mutti (1985) which refer to the entire basin-fill and which are discussed in section 3.3.4.6. The relatively small amount of sand in the area, compared to the quantity of pelite, indicates that these deposits are similar to the type II lobes of Pickering (1981).

3.3.2.2 *Goantagab River section*

This section (Fig. 3.5) lies immediately below the Brandberg West Formation in the Goantagab River (Fig. 1.3 for locality). The first 56 m of the section consist almost entirely of facies D_1 and D_2 beds. The sandstone:shale ratio increases upwards from 2:3 to 7:1. Near the top of the section some facies D_1 beds with partial Bouma sequences (T_{b-cde}) are interbedded with the facies D_2 beds. Concretions, which are now oriented parallel to the foliation, possibly developed before the main phase of deformation. The beds underlying this section are all thinly bedded.

The thin bedding found in the entire section, the similar character of the underlying beds and the lack of erosional features all suggest that the beds were deposited in a distal lobe-fringe environment.

3.3.2.3 *Skeleton Coast Park section*

This section, located in the dry western part of the area (see Fig. 1.3 for locality), is useful for showing variations in bed thickness, but not for sedimentary structures as these are generally poorly preserved. The bottom 6.5 m of the section (Fig. 3.6) consist of interbedded facies B_1 and D_1 beds. Many beds are amalgamated and the T_e interval, where developed, is generally very thin. Minor thickening-upward sequences are found in this basal portion. This is followed by an interval of very thick facies B_1 beds (up to 4.13 m) interbedded with minor facies C_2, D_1, D_2 and G beds. Overlying this, the succession is again dominated by facies D_1 beds with minor facies G beds. The proportion of pelitic material in the facies D_1 beds is less than 10 per cent. Rare rip-up clasts and load structures are found. Some of the facies G beds are thinly laminated and appear to be graded. These beds may represent deposition from extremely dilute turbidity currents. The enclosing beds are all thinly bedded and poorly exposed.

The largely thinly-bedded and non-cyclic nature of the section and the lack of erosional features suggest a distal environment of deposition, with intermittent progradation of lobes. The minor thickening-upward cycles are possibly "compensation" cycles (cf. Mutti and Sonnino, 1981) and suggest deposition in a lobe environment. The thicker facies B 1 beds represent incursions of lobe sands over distal de-posits (cf. Busby-Spera, 1985).

3.3.2.4 *Augur section*

The basal 17.5 m of this section consist largely of facies D_1 beds interbedded with minor facies B_1 and D_2 units (Figs 3.7 and Plate 7.A). Horizontal lamination is found as well as sole structures in the form of occasional dewatering and load features. The next 30 m consist largely of facies D_1 beds (Figs 3.7 and Plate 7.A). The greywacke:shale ratio is difficult to observe accurately but is greater than 1: 1. Overlying this is a 10 m succession of facies D_1 beds similar to the lowennost 17.5 m, and this is in turn overlain by a number of very thick facies B_1 greywackes which are up to 4.8 m thick and which are separated from each other by up to 40 cm-thick packages of facies D_2 beds (Fig. 3.7). The beds are all laterally continuous for the length of the outcrop (Plate 7.A) and no major channels are present.

The thinly-bedded nature of the basal portion of the sequence and the lack of evidence for channels suggest deposition in a distal environment such as a lobe fringe.

3.3.2.5 *Synthesis*

All of the described sections, which come from four widely separated localities (Fig. 1.3 and 3.8), indicate that deposition in the Zebrapiits Formation occurred either on distal lobes or lobe fringes. Analysis of aerial photographs, especially of those showing the areas in the vicinity of the Skeleton Coast Park and Ridge 6 sections, indicates that much of the sequence is thinly bedded. Field checking confirmed that most of the areas that appeared poorly exposed on aerial photographs were underlain by thinly-bedded pelites with occasional thicker and coarser-grained greywacke interbeds. The entire Zebrapüts Formation can therefore be regarded as having been deposited in a distal-lobe to basin-plain setting. The dominantly thin-bedded basin-plain succession with occasional development of lobe sandstones is similar to the type II lobe sequences described by Pickering (1981; 1985; Fig. 3.4). Pickering (1981; 1985) interpreted this type of sequence as developing on a passive margin.

3.3.3 Brak River Formation

Sections from a wide range of localities and covering the entire thickness of this formation have been measured. The base of the Brak River Formation is taken as the top of the last blue marble of the underlying Brandberg West Formation. Some marble interbeds which resemble facies Gc are found near the base of the Brak River Formation such as in the Zebra River Section (see 3.3.3.l). The top of the Brak River Formation is taken as being at the base of the first marble band of the overlying Gemsbok River Formation. The maximum thickness estimated in this study is 500 m, substantially more than the 350 m proposed by Jeppe (1952) and Miller et al. (1983).

3.3.3.1 Zebra River section

In this section (see Fig. 1.3 for locality) the first 23 m (Fig. 3.9) are dominated by facies D_1 and D_2 units with mi-

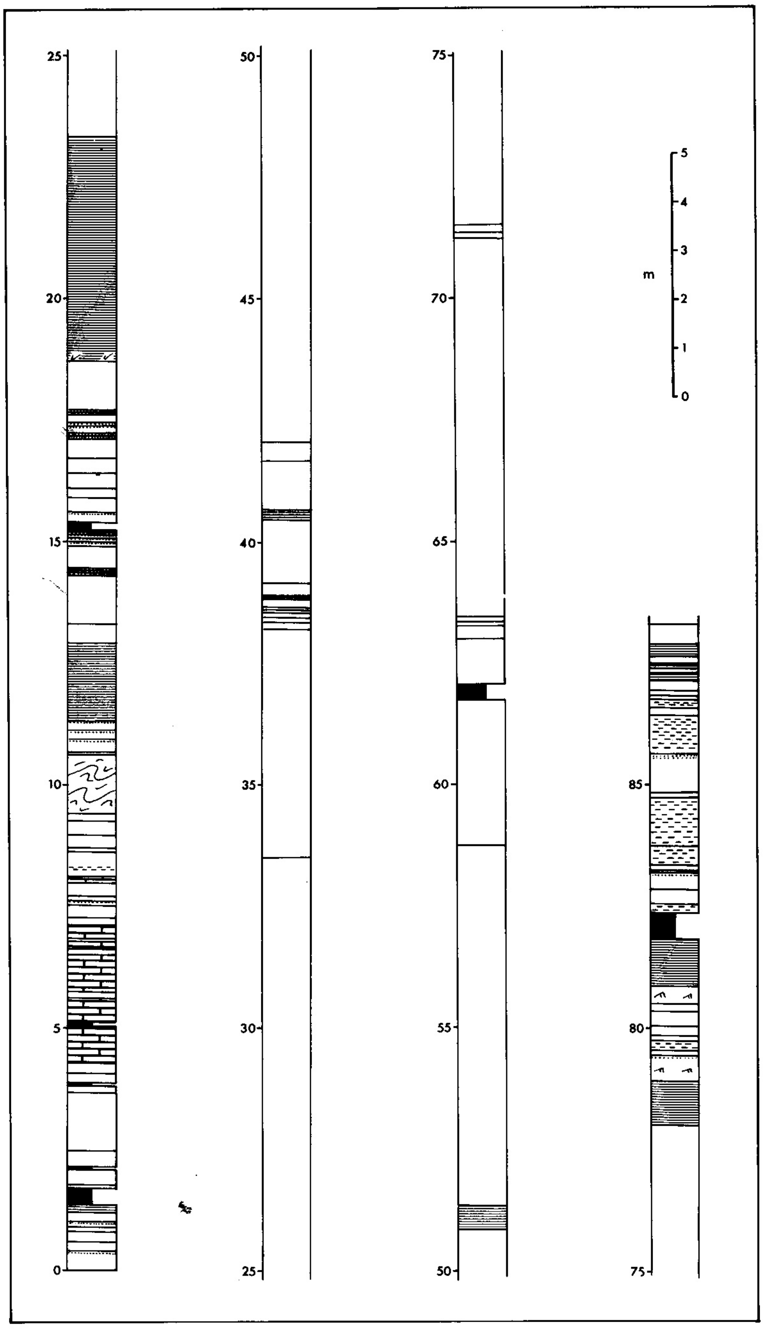


Fig. 3.9: Sedimentary log of the Zebra River section of the Brak River Formation.

nor facies G and Gc beds. The following 55 m are almost entirely made up of thick (up to 10 m) facies B_1 beds. No sedimentary structures have been observed in these beds which are parallel-sided and show no obvious grading. These greywackes appear to be laterally continuous but are difficult to follow out in this area because of the isoclinal folding. Minor facies D_2 beds are interbedded with these thick units. The final 10 m of the section consist of interbedded facies D_1 and D_2.

There are two possible interpretations of the Zebra River section. Firstly, the thick greywacke portions may represent isolated channel-fill deposits, while the thinner bedded associations are levee accumulations (cf. Shanmugam *et al.*, 1985; Nelson and Nilsen, 1984). However, the lack of erosional features such as scours and rip-up clasts, as well as the parallel-sided and continuous nature of the thick greywacke units suggest that they are not channel deposits. The levee deposits should also contain more pelitic material. An alternative explanation is that the thick greywackes represent unconfined flow of extremely large turbidite events over lower-fan areas. The thinner-bedded facies would then represent lobe-fringe and not levee deposits.

3.3.3.2 *Brak River section*

This section (see Fig. 3.10 and Fig. 1.3 for locality) is

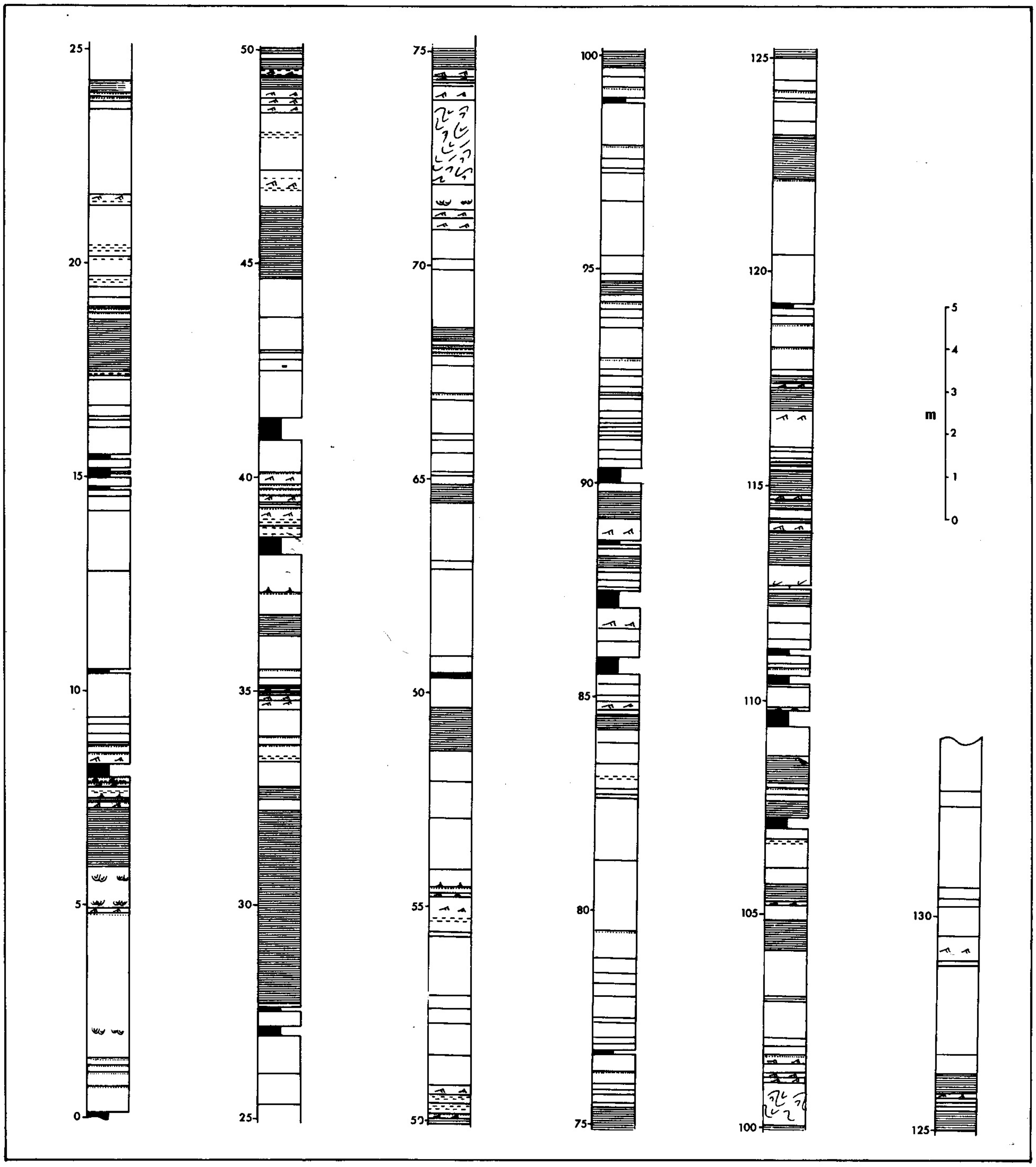


Fig. 3.10: Sedimentary log of the Brak River section.

made up of thinly- to thickly-bedded facies D_1, D_2, and H units. Minor facies G horizons are often graphitic and strongly deformed. Beds are laterally continuous and minor thinning- and thickening-upward cycles are present. However, some of these cycles are open to alternative interpretations and may merely be in the "eye of the beholder" (Walker, 1984). For example there are various interpretations that can be placed on cycles in the interval between 50 and 55 m in Fig. 3.10. Thickening-upward compensation cycles are developed between 117 and 125 m. The facies F beds are only developed near the top of the measured section. Excellent examples of gutter casts are associated with facies H in the interval immediately above this section. Although the section does contain minor cycles, its overall character does not change in the measured portion. However the greywacke sequence is overlain by a pelitic interval with a minimum thickness of 15 m, and this is itself immediately overlain by the Gemsbok River Formation.

The dominance of thin-bedded facies D_1 and D_2 beds in this section indicates that these sediments were laid down in a lower-fan setting. The presence of compensation cycles confirms a distal origin for these beds. Beds of facies H indicate a glacial influence at least in the final stages of the deposition of this formation. The overlying shale, which is overlain by the Gemsbok River Formation, probably indicates a period of glacial retreat with a resulting increase in sea-level, cutting off the supply of terrigenous sediment to the basin. Ultimately, carbonate bioherms with their associated deep-sea deposits started developing, resulting in deposition of the Gemsbok River Formation. Extensive carbonate deposits with well-developed stromatolites are known from this period, in particular from the platform area to the northeast of the basin (Hedberg, 1979).

3.3.3.3 *Rhino Wash section*

This section was measured in the dog-leg of the Rhino Wash (Fig. 1.3) and covers the upper portions of the Brak

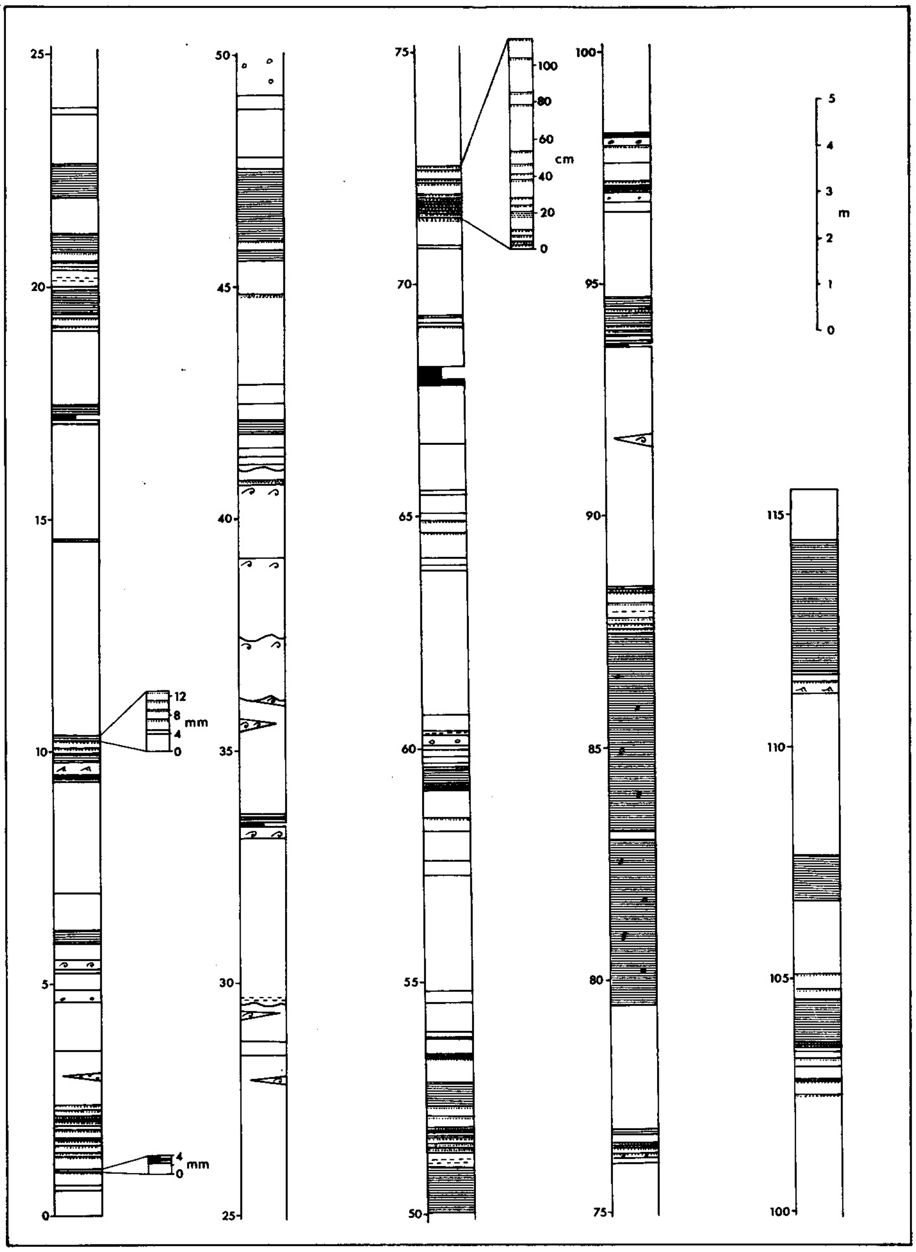


Fig. 3.11: Sedimentary log of the Rhino Wash section. Plate 8.B is a photograph of this section. Note the very thinly bedded facies D_1 units interbedded with facies B_1.

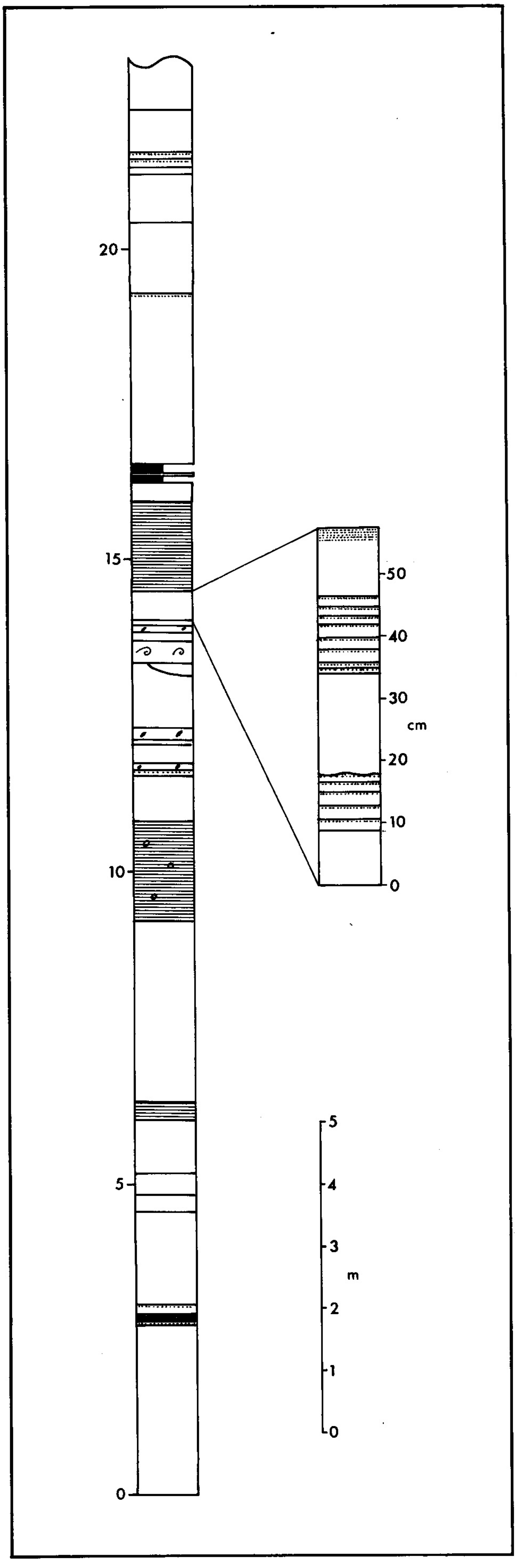


Fig. 3.12: Sedimentary log of the Shelter section of Brak River Formation.

River Formation (Fig. 3.11). A photograph of the section is shown in Plate 7.B. The section comprises a sequence of thin-bedded facies D_1 and D_2 interbedded with thick facies B_1 greywackes and occasional facies F slumps and clast-bearing horizons. Facies G beds are almost totally absent from the measured section. Maximum thickness of the greywackes is 4 m and beds are commonly amalgamated, especially in the interval between 30 and 40 m. The sandstone:shale ratio in the facies D_1 units may be as much as 12:1 for beds which are only 1.3 cm thick (eg. at 71 m; Fig. 3.11, enlarged portion). No obvious cyclicity is apparent in the section. The section is overlain by a thick shale (facies G) horizon.

The lack of cyclicity and abundance of facies D_1 and D_2 is indicative of a distal origin, with deposition on the fan-fringe or distal lobes. A large channel-levee complex is unlikely; although scours are present, the beds are largely parallel-sided and there is no evidence for large-scale channel features (Plate 7.B). The top of the succession is similar to that described from the Brak River section in that the upper portions contain drops tones and slump structures and are overlain by thick shale horizons and the Gemsbok River Formation. A similar interpretation to that invoked for the Brak River section, involving glacial retreat and an associated rise in sea-level, is therefore suggested.

3.3.3.4 *Shelter section*

Located 3.2 km north of the Rhino Wash section and along strike from it, this section (Fig. 3.12) is very similar to that sequence, but individual units cannot be correlated between sections. This is despite the fact that good control on the position of these sections is obtained from the aerial photographs, and the presence of the thick pelitic unit which overlies the succession. The same association of facies B_1 beds, up to 2.9m thick, with facies D_1, D_2, F and H is found.

As this sequence and the contiguous facies are the same as those found in both the Brak River and the Rhino Wash, the same interpretation is applied. The inability to make detailed correlations between sections, such as those made by Ricci Lucchi and Valmori (1980), is problematic because distal turbidite system deposits are thought to be non-channelised and laterally extensive (Mutti et al., 1978; Ghibaudo, 1980; Ricci Lucchi and Valmori, 1980). No evidence for major channel features between the two sections could be recognised on aerial photographs.

3.3.3.5 *Gemsbok Altar east section*

This section was measured in the gorge of a superimposed river course (Plate 7.C). The section (Fig. 3.13) is made up of three distinct packages, each of which is up to 60 m thick, separated by significant (10 m) shale horizons (facies G). The first two packages are made up of 10-25 m-thick bundles of facies B_1, C_2, D_1 and D_2. Individual beds are laterally continuous and parallel-sided. A number of nested thinning-and thickening-upward cycles are present in these bundles, although this is a very subjective interpretation often based on only a few beds (Fig. 3.13). The basal

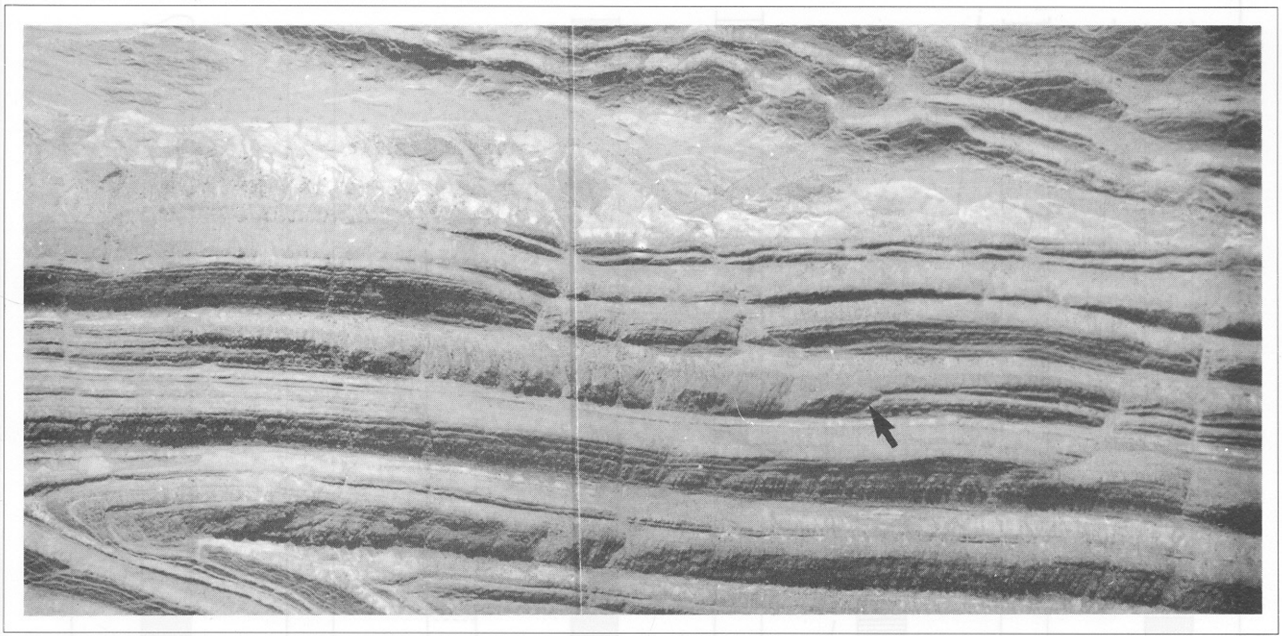


Plate 8.A: Mosaic of a series of vertical aerial photographs of a fold limb in the Brak River Formation showing lateral continuity of beds, except where the broad shallow channel (arrowed) is developed. See Fig.3.19 below for an interpretation of the area surrounding the channel. Photographs by C.J. Ward.

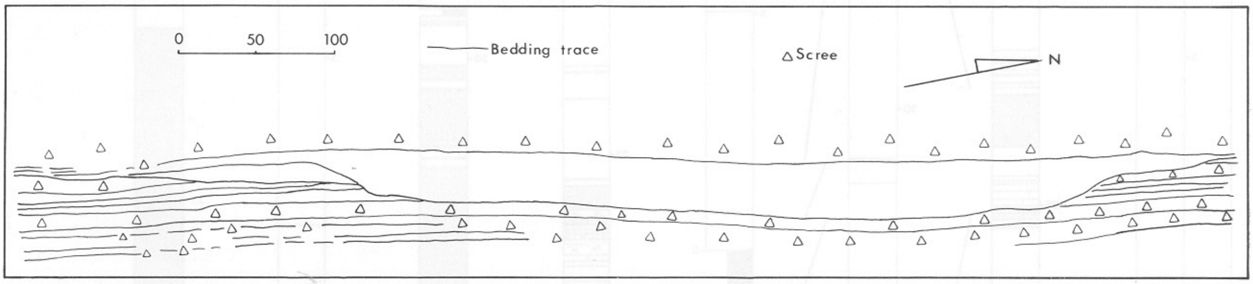


Fig. 3.19: Interpretation of the area around the channel in the aerial photograph shown in Plate 8.A.

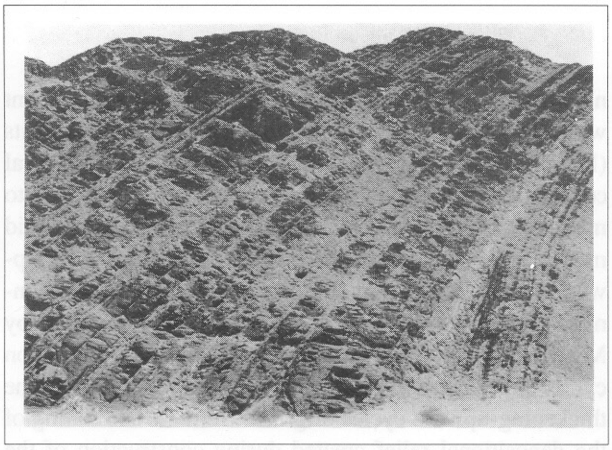


Plate 8.B: Photograph of the upper portions of the Jeppe east section of the Amis River Formation. The Amis River Formation differs from the Zebrapütz and Brak River Formations in that there is a lower proportion of pelitic material in the sequence (cf. Plates 7.D, 7.E and 7.F). Beds young to the right of the photograph.

Plate 8.C: Basal portion of the Lonely River measured section showing packages of facies D_1 beds overlain by facies B_1 and more D_1 (bottom of B_1 package arrowed).

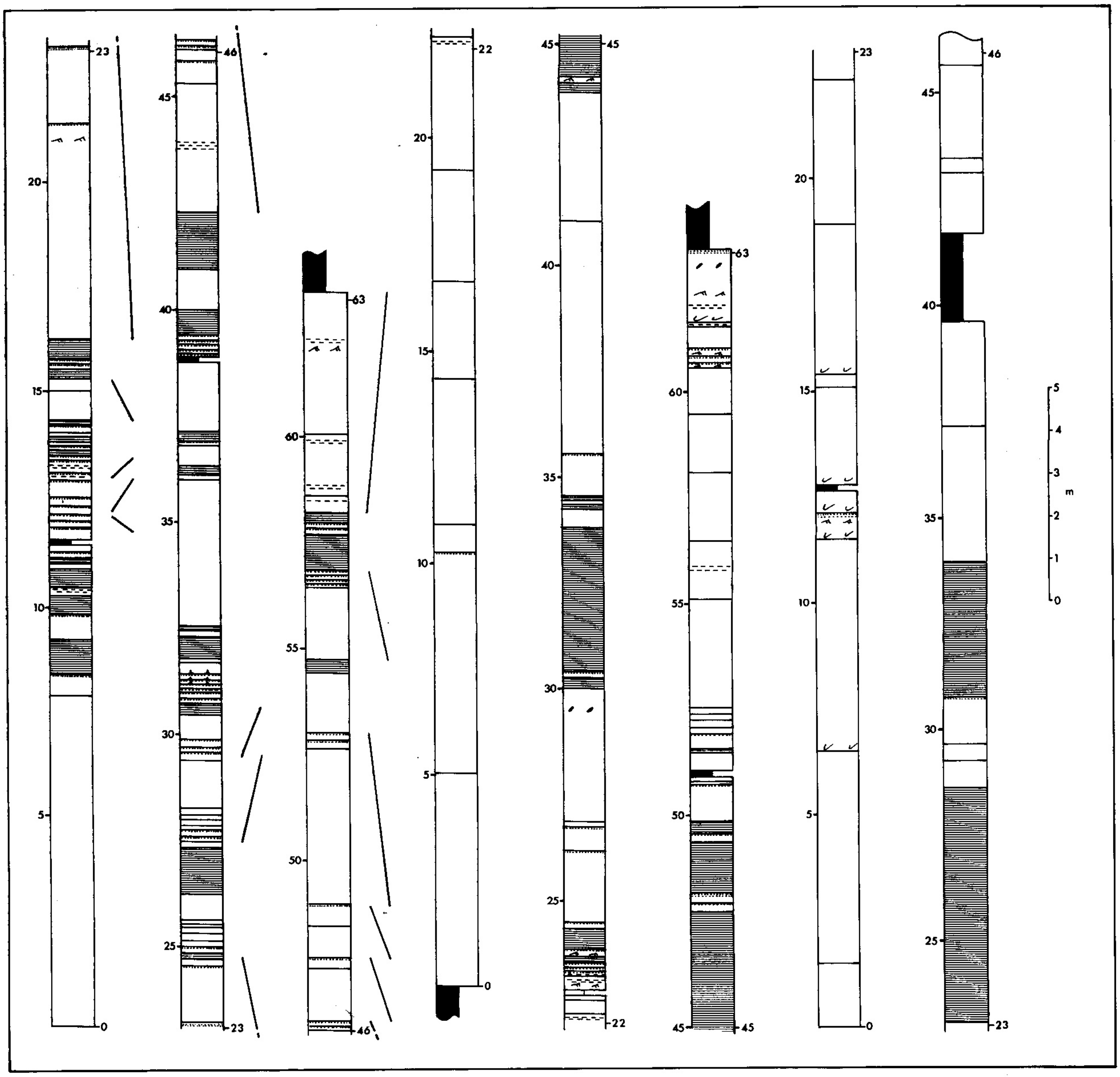


Fig. 3.13: Sedimentary log of the Gemsbok Altar east section. Only the lobe sandstone portions are shown as the outcrop of the facies G beds is poor. \ Thinning-up; / Thickening-up.

23 m of the final package are dominated by thicker units, but are overlain by thinner facies D_1 and D_2 units towards the top. A photograph of this uppermost package is shown in Plate 7.D.

The outer-fan deposits of the modem Mississippi fan, which is the only modem fan which has been drilled, are characterised by a high overall proportion of sand (41 and 64 per cent for the two fan-lobes drilled into), and maximum bed thickness is 10 m (O'Connell *et al.*, 1985). Although it is dangerous to compare modem and ancient systems, the long distance of transport of thick sand layers in the modem example is compatible with the thick beds found in the lobe deposits measured in the Gemsbok Altar east section.

Each of the 60 m-thick packages can be identified on aerial photographs (Plate 7.C). From the photographs it is clear that individual lobe sandstones are laterally continuous for many kilometres. Similar lateral continuity can be observed elsewhere (Plates 7.E and 7.F).

The classical interpretation of numerous stacked thinning-upward cycles is one involving channel abandonment with the thinner-bedded units representing levee deposits (Nelson and Nilsen, 1984). However, in this case the lateral continuity of individual facies B_1 beds and their fine- to medium-grained character suggest that this is not a valid model. These cycles are the reverse of the thickening-upward "compensation" cycles described by Mutti and Sonnino (1981). A variation of the mechanism proposed by Mutti and Sonnino (1981) for the origin of the compensation cycles is therefore suggested here. Their model is that the thickening-upward cycles originate from the smoothing of the deposition of relief created during construction of the lobe (Fig. 3.14). If the currents were consistently deflected to one side by the depositional bulge, then vertical sections through certain portions of the lobe would consist of thinning-upward sequences (Fig. 3.14). Subsidence of the basin would allow repetition of these sequences till eventually lobe abandonment occurred. This is marked by the presence of the thick facies G portions.

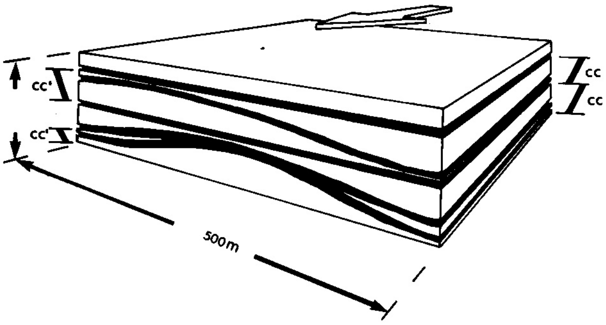


Fig. 3.14: Development of thickening-upward compensation cycles according to Mutti and Sonnino (1981), with modifications indicating how the same set of beds can develop thinning-upward cycles (**cc'**).

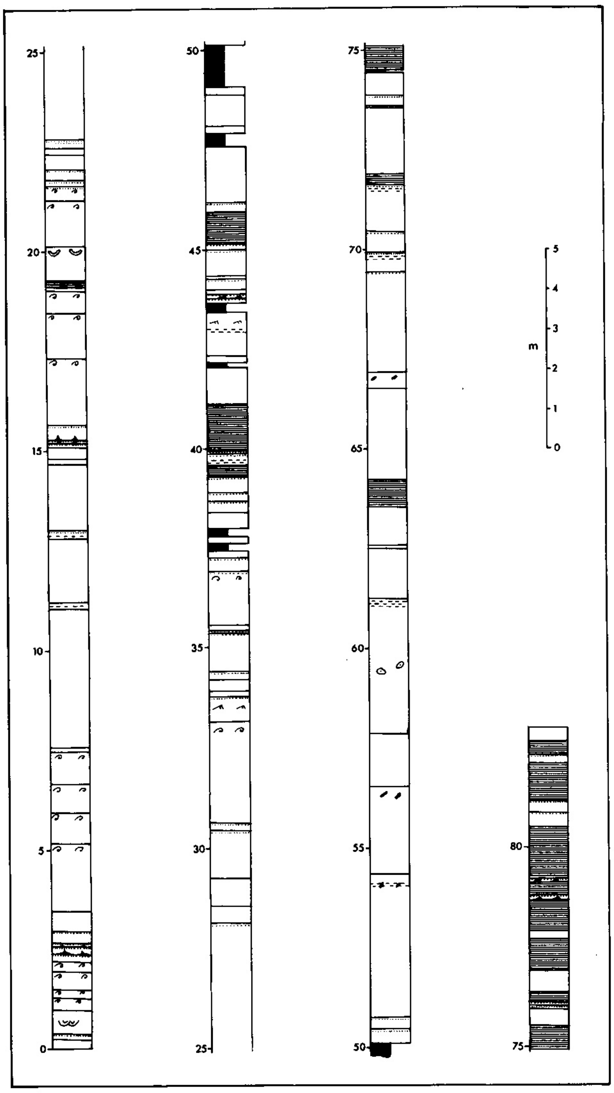
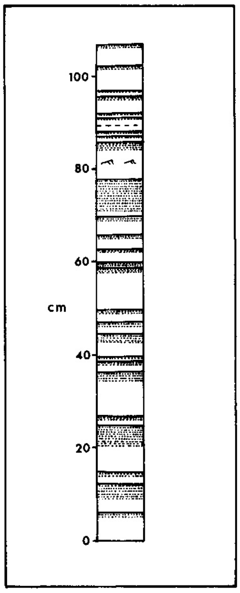


Fig. 3.15: Sedimentary log of the Gemsbok Altar west section of the Brak River Formation. The sequence here is characterised by a lack of pelitic material.



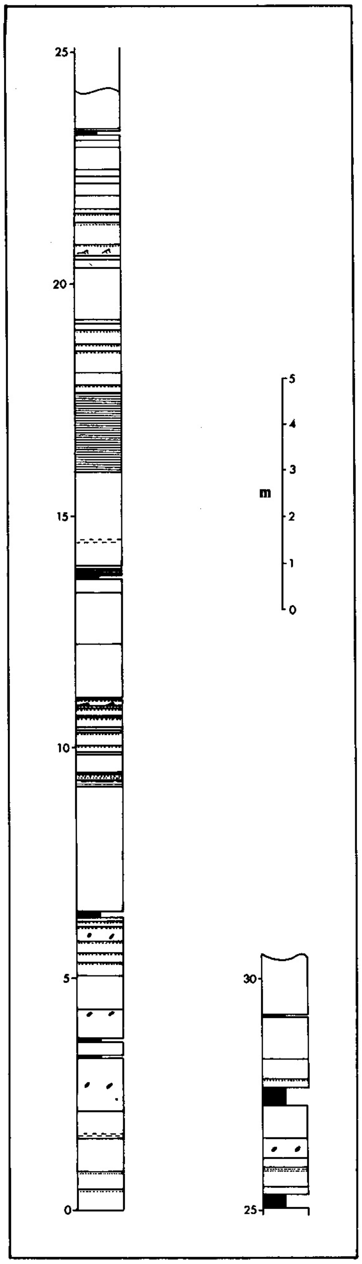


Fig. 3.16 (left): Detail of the portion between 40-41 m of Fig. 3.15 showing the high sand:shale ratio in typical thin-bedded facies D1 beds.

Fig. 3.17 (right): Sedimentary log of the Gai-as south section of the Brak River Formation.

3.3.3.6 *Gemsbok Altar west section*

This section (Fig. 3.15), which lies stratigraphically above the one at Gemsbok Altar east, is separated from it by a zone of deformation. The first 39 m of the section are dominated by facies D_1 and D_2 units in thinning-upward cycles and rare thickening-upward ones. This is overlain by 11 m of non-cyclic facies D_1 and D_2. Detail of a portion of the section between 40-41 m is shown in Fig. 3.16. This detailed 1 m of section illustrates the high sand:shale ratio of the majority of beds, despite their very thin nature. The following 25 m are again dominated by facies D_1 and D_2 with thinning-upward cycles. The final 8 m are dominated by facies D_1 beds.

Overlying the section is a poorly-exposed sequence of interbedded facies D_1, D_2 and H. Clasts in facies H are up to 30 cm in long dimension. Pods of clast-rich material are also found and a gritty unit up to 12 cm thick is present. This sequence is in turn overlain by a thick (> 10 m) shale (facies G) horizon.

The section is interpreted, in a similar manner to the one

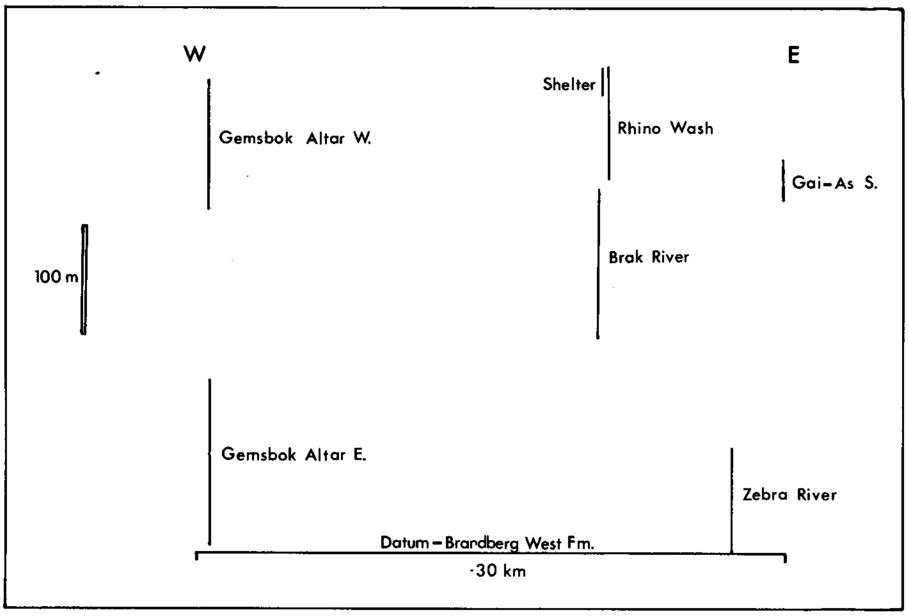


Fig. 3.18: Relative stratigraphic positions of the various measured sections of the Brak River Formation, using the uppermost blue marble of the underlying Brandberg West Formation as a datum, and projected onto an east-west line. Unfolding of the north-south oriented folds will significantly increase the distance between sections

below it, as a sandstone lobe. The presence of facies H indicates a glacial influence in the later part of the deposition of the sequence. The overlying sequence of facies G beds represents basin-plain deposition after retreat of the glaciers.

3.3.3.7 *Gai-as south section*

Exposures in a gorge south of the road between the old Brandberg West mine and the Gai-as waterhole were measured for this section which covers a portion of the middle of the Brak River Formation (Fig. 3.17). Facies D_1 and B_1 dominate the section. The beds are parallel-sided and laterally continuous. No major erosional features are present, but occasional rip-up clasts are found.

The sedimentological features observed in this section are similar to those described from the Gemsbok Altar east and west sections (see 3.3.3.5 and 3.3.3.6), and using the same arguments for its origin a distal-lobe to lobe-fringe environment is therefore suggested.

3.3.3.8 *Synthesis*

The present disposition of these sections relative to each other is shown on Fig. 1.3 and their relative stratigraphic positions in Fig. 3.18. Unfolding of the sequence would result in a major increase in distance between the sections in an east-west orientation, but little change in the north-south direction, as the major fold axes trend north-south (Fig. 1.4). The sections which show well-developed lobe deposits are located in the western parts of the area (Gemsbok Altar, Rhino Wash and Shelter sections), whereas typical fan-fringe deposits with incursions of large flows over them are found in the sections to the south and east (Brak River, Zebra River and Gai-as south). The abundance of glacial dropstones decreases eastwards indicating that the source of the glacial material lay to the west.

Deposits of lobe sands such as these are believed to be sheet-like, but outcrop limitations normally prevent examination of the three-dimensional character of the deposit. In the Brak River Formation, in the vertical sections which have been described here, the beds always appear tabular (eg. Plates 7.A and 7.B) However, channels, where they are known to exist in lobe areas, are subtle and do not form deep conduits. An example of this is the cut-down described by Mutti *et al.* (1978) in which a bed was incised by 5 mover a lateral distance of several hundred metres. Also, channels on the lower-fan area of the modem Mississippi fan have a depth to width ratio of 5:200 (Barnes and Normark, 1985 - wall-chart). In the far west of the study area the outcrop is suitable for detailed examination of aerial photographs. Existing aerial photographs are at too small a scale to allow examination of individual lobes and therefore a set of detailed aerial photographs was taken of a relatively undeformed fold limb at a flying height of 4000 m above ground level, using a conventional 35 mm SLR camera. On this set of photographs individual lobes can be followed out for the length of the fold limb, but a shallow channel can be identified in the middle portions (Plates 7.B and 8.A). The channel has an approximate depth of 25 m and a width of 1050 m. The exposure is almost certainly of an oblique cut across the channel and the true width may be significantly less than this. However, the broad, shallow nature of this channel is similar to those described from the surface of modem (Barnes and Normark, 1985) and ancient outer-fan areas (Mutti *et al.*, 1978), but would be too small to be detected in the subsurface by multi-channel seismics. Internally the channel consists of a non-graded, fine- to medium-grained greywacke. No pebbles are present, but concretions are common. No bedding-plane surfaces were observed and unfortunately, scree cover prevents examination of the lateral margins of the chapnel.

The shale which blankets this formation was deposited

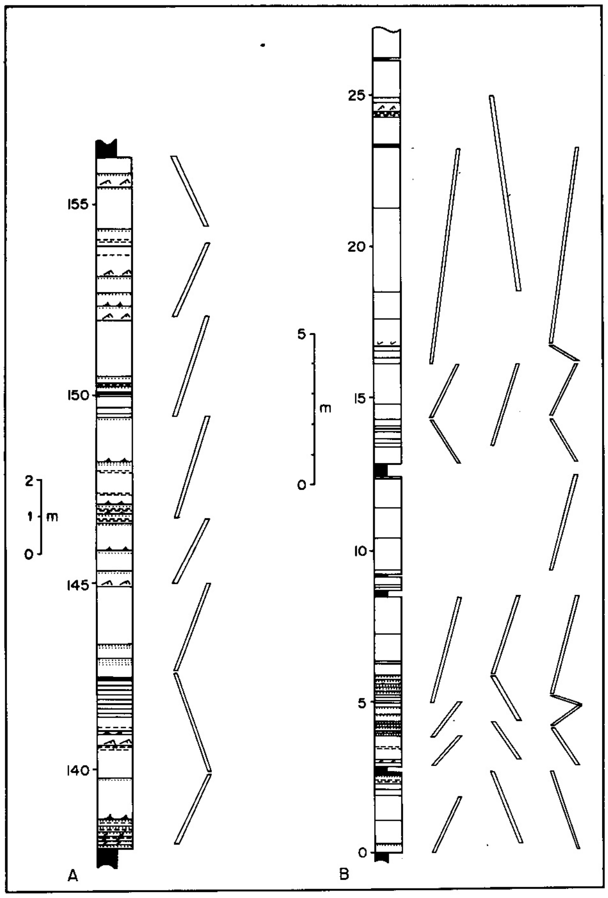
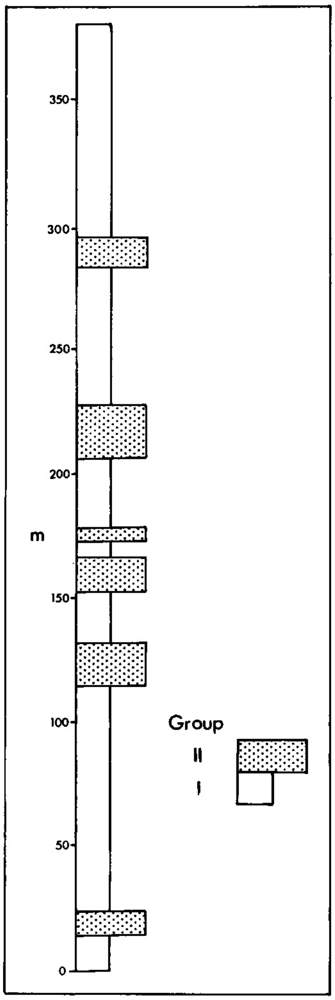


Fig. 3.21: A: Portion of Fig. 3.20 showing asymmetric cycles which comprise only a few beds. These are similar to the compensation cycles of Mutti and Sonnino (1981). **B**: Basal portion of the Jeppe east section showing different possible interpretations of cycles. Cycles shown as in Fig. 3.13.



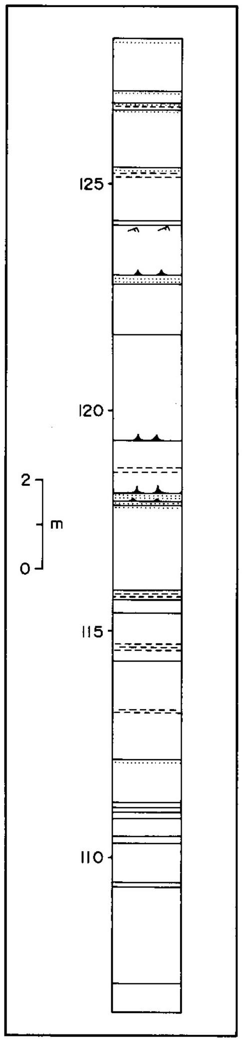


Fig. 3.22 (left): Two common facies associations are found in the Jeppe east section. Group I comprises largely facies D_1, D_2 and G, while group II is made up of B_1 and D_1.

Fig. 3.23 (right): The group II association, a typical example of which is shown here, comprises largely beds of facies B_1 and D_1.

during the period of raised sea-level which followed the retreat of the glaciers. Complete abandonment of the turbidite system occurred as terrigenous sediments were trapped on the shelf. This was followed by carbonate de-velopment on the shelf with associated calcareous turbidites (the Gemsbok River Formation).

3.3.4 Amis River Formation

Five sections have been measured in the Amis River Formation over a wide geographic area (Fig. 1.3). Three of these sections have been measured in the well-exposed western part of the area and between them cover nearly the entire interval. The remaining two sections which were measured in the east (Fig. 1.3) are not as comprehensive, as the deformation in this region is more intense. No sections could be measured in the central part of the area, as exposures of the Amis River Formation in this area are near the noses of folds (see Fig. 1.4) and are highly deformed. Younger cover obscures other possible sections.

The base of the first siliciclastic unit overlying the hemipelagic blue marble of the Gemsbok River Formation is taken as the base of the formation. In certain areas (eg. Brak River), some carbonate turbidites are found interbedded with the basal portions of the Amis River Formation. The top of the formation is not exposed in the study area and the maximum exposed thickness of the unit in the western part of the area is 450 m.

3.3.4.1 *Jeppe east section*

This profile (Fig. 3.20) highlights some of the problems encountered in the recognition and interpretation of asymmetric cycles in turbidite sequences. "Cycles" often comprise very few beds (Fig. 3.21A) and alternative interpretations can be made for many of the sequences (Fig.3.21B). These are similar to those shown by Walker (1984; his Fig. 22). However, in certain portions of the succession minor thickening-upward sequences can be recognised (Fig. 3.21A). These are similar to the "compensation cycles" of Mutti and Sonnino (1981). No cycles of the order of ten metres or more are developed at all and many are based on only three beds. It is the repetition of the minor cycles, rather than their vertical extent, which suggests that there must be some inherent depositional control. An overall view of

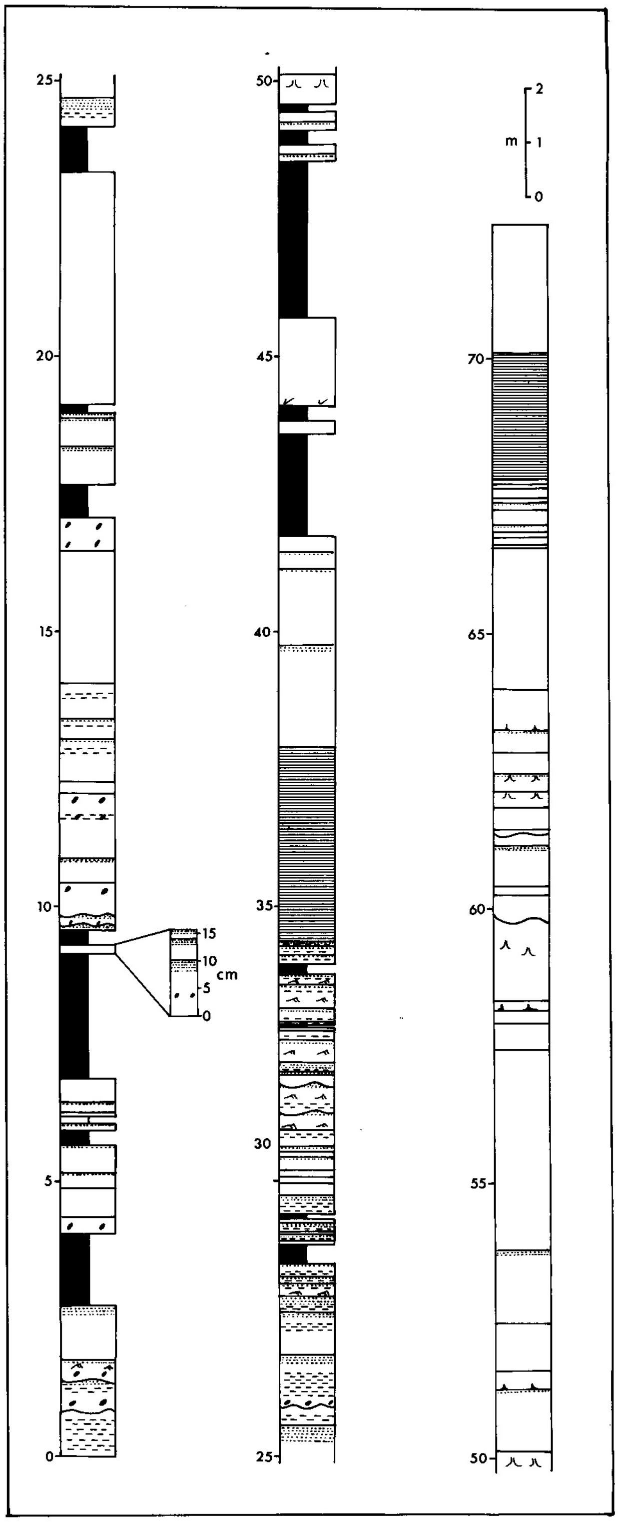


Fig. 3.24: **Sedimentary log of the Jeppe west section. This section is dominated by facies D_1 and D_2, with rare facies B_1 beds.**

the upper portion of this section is shown in Plate 8.B.

Two distinct facies groups make up this section (Fig. 3.22). The first group (I) is an association of facies D_1, D_2 and G beds, which commonly thicken upward (Fig. 3.21A). The second group (II) comprises facies Bland D_1 beds with occasional D_2 and G beds (Fig. 3.23). Both thickening- and thinning -upward cycles are found in this group.

The dominance of thickly- to thinly-bedded facies D_1 and D_2 turbidites of group I, the high sand content and lack of slump features strongly suggest a lobe to lobe-fringe environment for this group (Pickering, 1982; 1985; Mutti, 1985). If the short, coarsening-upward cycles are real and are the equivalent of compensation cycles, then according to Mutti and Sonnino's (1981) model, facies group I represents outer-fan lobe deposits. The interbedded thicker facies B_1 beds in group II possibly represent incursions of lobe channels over their distal deposits (Busby-Spera, 1985).

3.3.4.2 *Jeppe west section*

This section is only separated from the previous section by a short interval of deformed beds. The section (Fig. 3.24) is similar to that which underlies it, but only facies group I, with isolated interbeds of facies B_1, is developed.

As this section is similar to the previously described one, a lobe-fringe environment is suggested, with very rare incursions of facies B lover the distal deposits.

3.3.4.3 *Lonely River section*

This section is dominated by laterally continuous facies D_1 beds with two thick packages of facies B_1 beds (Fig. 3.25). Facies G, D_2 and E are minor in the sequence and the amount of sand in individual beds is generally greater than 70 per cent. Minor thinning-upward cycles are apparent in the profile. Facies groups (Fig. 3.26) include thinning-upward bundles of facies D_1 and E beds (facies group I) and thick non-cyclic packages of facies B_1 beds which are commonly erosive (facIes group II). Facies D_1 and E are also associated with this group, but make up a minor proportion of the sequence.

The thinning-upward group I packages represent non-channelised deposition in a proximal outer-fan lobe environment (cf. Nelson and Nilsen, 1984). The high percentage of sand and thinning-upward cycles preclude this from being a fan fringe. Breaks in turbidite deposition are indicated by the facies G units. The group II association suggests intermittent channel switching or non-channelised deposition of turbidity currents of large volume over the fan - or lobe-fringe.

3.3.4.4 *Tributary section*

This is a short section in which sedimentary structures are not very well exposed, but in which bedding thicknesses can readily be determined. The succession (Fig. 3.27) is made up entirely of thinly-bedded facies D_2 beds with a sandstone:shale ratio of less than one. Rare facies D_1 units occur, but no cycles could be identified. The beds which are adjacent to these are similarly thin-bedded.

The dominance of facies D_2, the low sandstone:shale ratio and the overall thinly-bedded nature of the sequence is characteristic of a distal outer-fan lobe or fan-fringe (Mutti, 1977; Nelson and Nilsen, 1984).

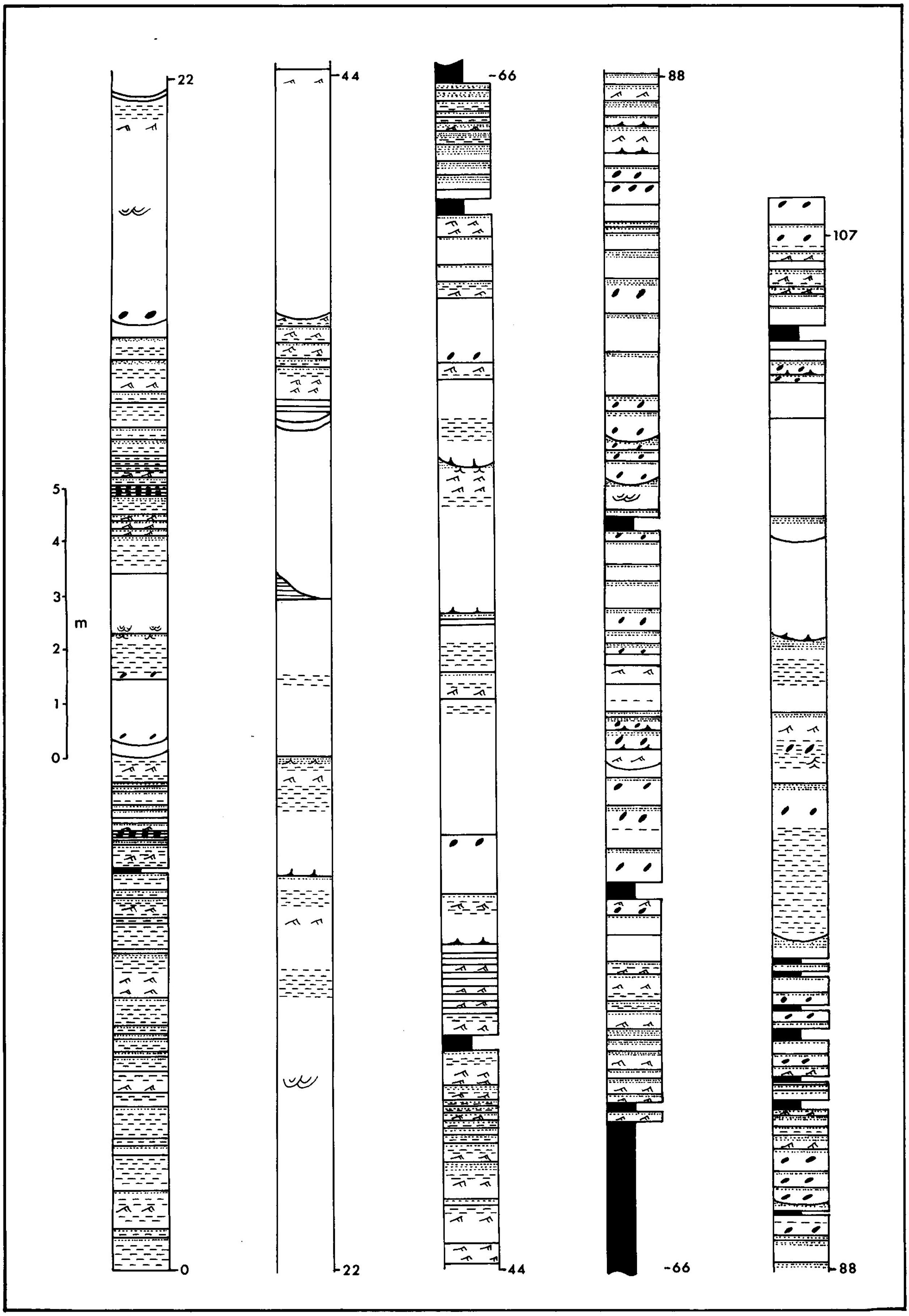
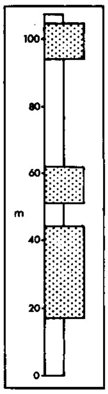
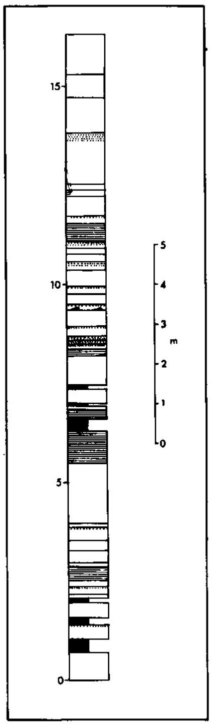


Fig. 3.25: The best exposures in the area are found in the Lonely River section, shown here. The section is dominated by facies D_1 beds with packages of facies B_1. Facies D_2, E and G are minor in the section. Fig. 3.26 shows the various facies groups. A photograph of the basal part of this section is shown in Plate 8.C.





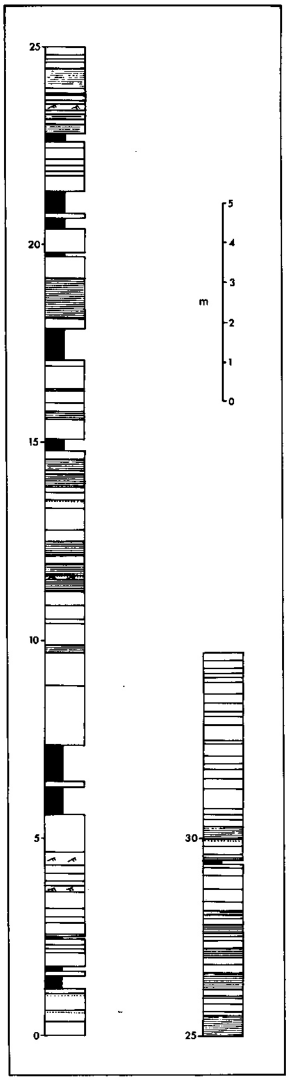


Fig. 3.26 (left): The same facies groups found in the Jeppe east and west sections are found in the Lonely River section. These groupings are shown here. See Fig. 3.22 for key.

Fig. 3.27 (middle): The Tributary section of the Amis River Formation (shown here) is dominated by thin- to medium-bedded facies D_1 and D_2 beds. Facies B_1 is virtually absent in the area.

Fig. 3.28 (right): The Bee section is similar to the Tributary section in that it is dominated by facies D_1 and D_2 beds.

3.3.4.5 *Bee section*

This section (Fig. 3.28) is dominated by thinly-bedded facies D_1 and D_2 units with no cyclicity and an overall sandstone:shale ratio less than one. Immediately adjacent to the section is a 3 cm-thick interval of coarser grained, discontinuous beds up to 10 cm wide which are interbedded with finer-grained units (Fig. 3.28). The other contiguous beds are all thinly bedded. A clean, very coarse-grained quartzwacke is also found in this area. This unit is not laterally continuous and consists of up to five well-graded layers, each with a maximum thickness of 50 cm. No other sedimentary structures have been observed in this quartzwacke and consequently no palaeocurrents could be determined for this unit.

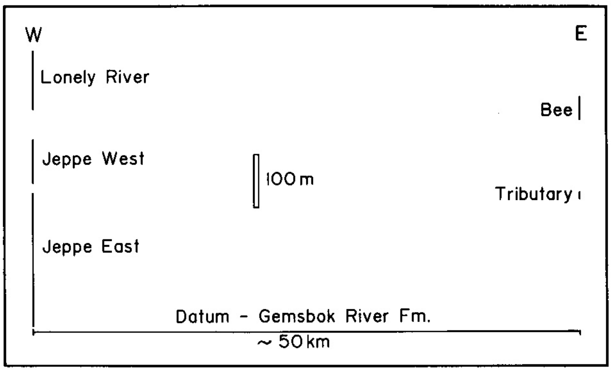


Fig. 3.29: From the above diagram it is clear that the sections measured in the west represent an almost continuous stratigraphic sequence. These show only minor vertical variations. The sections in the east represent a small proportion of the sequence, but the succession is thinly bedded in the area (see text).

The facies association and low sandstone:shale ratio is again characteristic of either a distal outer-fan lobe or fan-fringe environment. The thinly-bedded, coarser-grained material interbedded with these distal sediments is interpreted as being the result of winnowing by contour currents.

The presence of multiply-graded layers in the quartzwacke and their very coarse grain size preclude this from being a contourite. The compositional and grain size differences between this unit and the enclosing rocks suggest that it is either a megaturbidite or was derived from a different source area which was more proximal than that supplying the bulk of the sediment. The lack of lateral continuity and the thin nature of these beds compared to described megaturbidites, which are generally laterally continuous and very thick (eg. the Contessa bed of Ricci Lucchi and Valmori, 1980; Bouma, 1987), suggest that the quartzwacke is not a megaturbidite. The probable source of these beds must therefore be different to the main source and was possibly lateral to the margin of the basin. The lack of a fine-grained matrix suggests that the beds have been reworked, possibly by wave action on a beach, before slumping into the basin.

3.3.4.6 *Synthesis*

Although no significant large-scale vertical cycles are present in the Amis River Formation, minor compensation cycles are present and significant lateral facies changes occur from west to east. In the west the sediments are thinly- to very thickly-bedded, have a high proportion of sand and show minor erosional features. The relative stratigraphic and geographic positions of the various sections are shown in Fig. 3.29. The deposits in the east are generally finer grained, thin- to medium-bedded and do not show erosional structures (lobe-fringe). Maximum bed-thickness in the east is 3 m but beds this thick are rare. Clearly, the sediments in the west are relatively more proximal than those in the east. If the thinly-bedded schists to the south of the Karoo cover are lateral equivalents of the Amis River Formation as suggested by Miller (1988; Fig. 1) then these units are the most distal basin-plain equivalents of the Amis

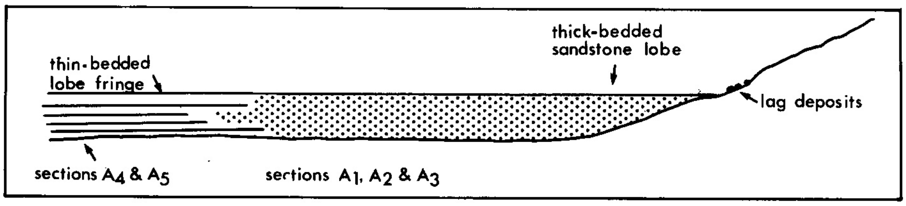


Fig. 3.30: Proposed model for the deposition of the Amis River Formation. The system developed by aggradation, not progradation, and is similar to the type I system of Mutti (1985). This type of system requires a major input of sand.

River Formation. Swart (1988) proposed that the lack of major cycles in the Zerrissene system indicated a debris apron made up of many overlapping small fans. However, the large size of the system in comparison to known debris aprons, which are generally less than 10 km in diameter (Nelson, 1983), and the presence of the small-scale cycles would suggest that this model is not valid for the Amis River Formation.

Macdonald (1986) has described major lateral facies changes without significant vertical variation in the Cumberland Bay Formation of South Georgia and these were attributed by him to tectonically controlled aggradation of the system rather than progradation. Surlyk (1978) has also described a fault-controlled sequence from the Mesozoic of Greenland which does not show progradational features. The Amis River Formation is different from these two systems in that no inner- or mid-fan deposits are present in the exposed portions of the basin. The development of over 600 m of largely thickening-upward lobe-fringe to lobe deposits suggests that some progradation did occur during development of the system. However:, the classic profile of a prograding fan with major thickening- and thinning- upward cycles representing lobe and channel deposits respectively is not developed at all. Sequences of the type shown by the Amis River Formation develop by accretion and are similar to the type I system of Mutti (1985; Fig. 3.30). In type I systems the sand is largely deposited in non-channelised bodies (lobes) which are detached from the associated channel-fill sequences. A voluminous sediment supply is required for the development of this kind of system which is characteristically associated with large elongate troughs in flysch basins. A narrow basin with marginal input of coarse- grained material is suggested by the presence of the quartzwacke which was clearly derived from a different source area to the majority of the turbidites.

The original tectonic setting of the Amis River Formation is unfortunately not known at present. The base of the Zerrissene system is not exposed and it is therefore impossible to classify the type of basin according to the scheme proposed by Mutti and Normark (1987) as their method requires know ledge of the nature of the crust underlying the system.

CHAPTER FOUR: PROVENANCE OF THE SILICICLASTICS

4.1 INTRODUCTION

Provenance studies of ancient sedimentary sequences require an integrated knowledge of palaeo currents, composition of the framework grains, heavy-mineral composition and geochemistry of the sediments. Other useful techniques for determining direction of the source area include facies mapping and determination of grain-size trends (Miall, 1984). Facies and grain-size variations have already been discussed in chapter three. This chapter will deal with the petrography, geochemistry and palaeocurrents of the Zerrissene terrigenous turbidites.

4.2 PETROGRAPHY

4.2.1 Introduction

The composition of the framework grains in greywackes has been used to infer the tectonic setting of depositional basins (Crook, 1974; Schwab, 1975; 1981; Potter, 1984; Valloni and Maynard, 1981; Dickinson, 1970; 1982; Dickinson and Suczek, 1979; Dickinson and Valloni, 1980; Dickinson *et al.*, 1982; 1983; Valloni, 1985; Zuffa,1987). Various studies have measured the proportions of quartz (both monocrystalline and polycrystalline), feldspar (plagioclase and K-feldspar ratios) and rock fragments (igneous, metamorphic or sedimentary). Plots of these ratios are then used to infer the tectonic setting. Unfortunately, in many greywackes much of what appears to be a primary matrix may in fact be degraded labile rock fragments (Cummins, 1962; Hawkins and Whetten, 1969: Galloway, 1974), hence meaningful point-counting on such sediments is difficult.

A further complication in the Zerrissene greywackes is the deformation, metamorphism and calcitisation. Firstly, the deformation has caused many clasts to become highly elongate and this feature, which is superimposed on the effects of compaction, makes recognition of original rock fragments nearly impossible in many thin sections. This is particularly so for volcanic and shale fragments which are highly susceptible to modification. In addition the low-grade metamorphism has caused recrystallisation of the matrix, and possibly also of original fine-grained volcanic and sedimentary rock fragments, making distinction of these clasts from the matrix very difficult (cf. Bhatia, 1983). Later calcite cement has also, in some cases, replaced both the matrix and framework grains. Detailed point counting was therefore not a viable technique, but samples were carefully examined as to the nature of the quartz clasts, type of rock fragments present where recognisable, feldspar composition and the heavy mineral population.

4.2.2 Zebrapüts Formation

Rocks of this formation are generally very fine-grained in keeping with their basin-plain origin, and are not suitable for petrographic studies. Secondary recrystallisation is strong in all specimens, but in the coarser-grained samples original grains of both monocrystalline and polycrystalline quartz are common. No definite lithic fragments have been observed, but both altered and unaltered plagioclase are found. A single grain of perthite was also identified, indicating a probable granitic source. The heavy-mineral suite comprises tourmaline, zircon and apatite. The durability of these minerals in igneous, metamorphic and sedimentary environments means that they are not useful indicators of the nature of the source terrain. Secondary minerals include biotite, chlorite, calcite, sericite and opaque oxides.

4.2.3 Brak River Formation

Mono- and polycrystalline quartz grains are the most common framework elements. The polycrystalline grains consist of chert and metamorphic clasts which show well-developed triple junctions. Plagioclase is common and clasts with myrmekitic intergrowths and ones with oscillatory zoning are found. K-feldspar is uncommon. Rock fragments are particularly common in this formation because of the presence of drops tones. They are useful for identifying the source area of the glaciers but cannot be used for determining tectonic setting as the glaciers could have transported material across tectonic boundaries (Zuffa, 1987). The dropstones include granitoids and rare mafic igneous fragments. The mafic clasts appear to have been derived from hypabyssal rocks. The heavy-mineral suite consists of zircon, tourmaline, apatite and monazite. Secondary minerals include epidote, chlorite, biotite, sphene, opaques, calcite and sericite. Some calcite is possibly of detrital origin as the grains are well rounded and are composed of material which is coarser than that found in the matrix, but the recrystallisation makes positive recognition of a primary origin very difficult.

4.2.4 Amis River Formation

Mono- and poly-crystalline quartz are the most common framework grains. Chert and metamorphic quartz rock fragments are common and a fragment of a quartz-arenite with well-developed quartz overgrowths on rounded grains also occurs (Plate 9.A). A single garnet-biotite-quartz schist fragment was also observed (Plate 9.B), but no individual garnet grains have been found. Another rock fragment contains feldspar and three subhedral grains of tourmaline (Plate 9.C). Myrmekitic intergrowths are present in some grains. Mafic fragments are uncommon. Carbonate rock fragments are also present, but difficult to distinguish from secondary carbonate. The heavy-mineral suite consists of apatite, zircon and tourmaline. Some tourmaline grains show overgrowth features (Plate 9.D). Secondary minerals present are biotite, chlorite, calcite, epidote, sericite and sphene.

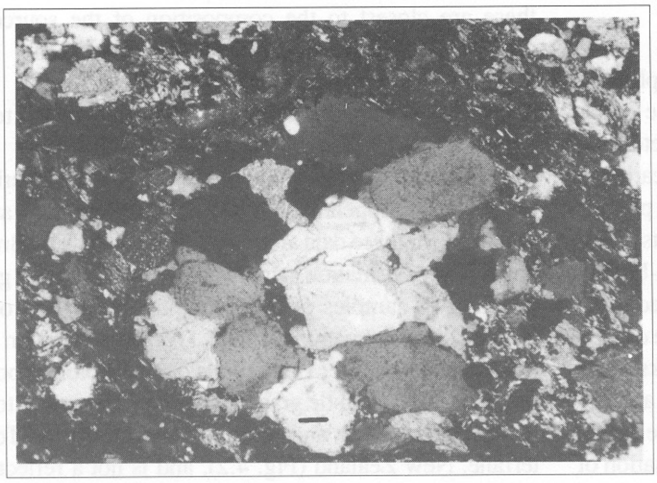


Plate 9.A: Photomicrograph of a quartz-arenite clast which shows well-developed, optically continuous overgrowths on individual quartz grains (sample UR 6/87). Scale bar 0.1 mm.

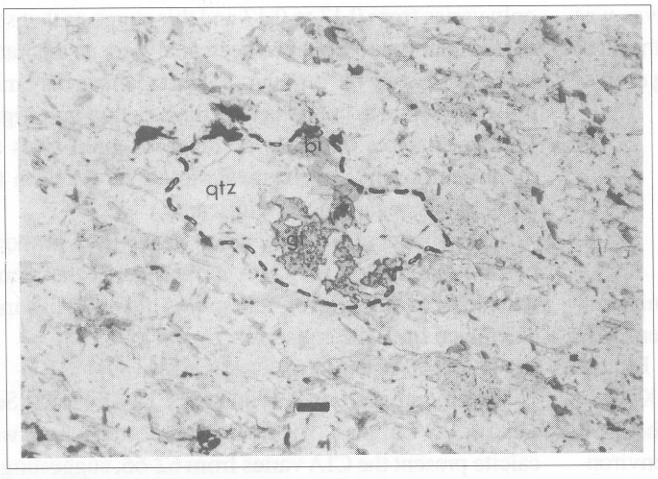


Plate 9.B: Although no individual detrital garnet grains have been observed, a single clast of quartz-biotite-garnet schist (outlined) is present and this, combined with other metamorphic, igneous and sedimentary clasts, indicates a basement - recycled orogen source terrane. bi - biotite; gt - garnet; qtz - quartz. Scale bar 0.2 mm.

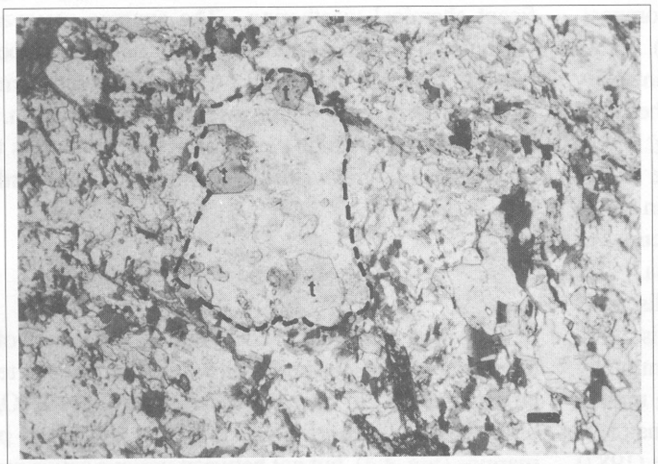


Plate 9.C: Tourmaline is the most common detrital mineral, but is not useful for provenance studies as it is polycyclic. However rare inclusions of tourmaline grains in a granitic fragment (outlined), such as these shown here, indicate that they were derived from a granitic source (sample UR 20/87). t - tourmaline. Scale bar 0.2 mm.

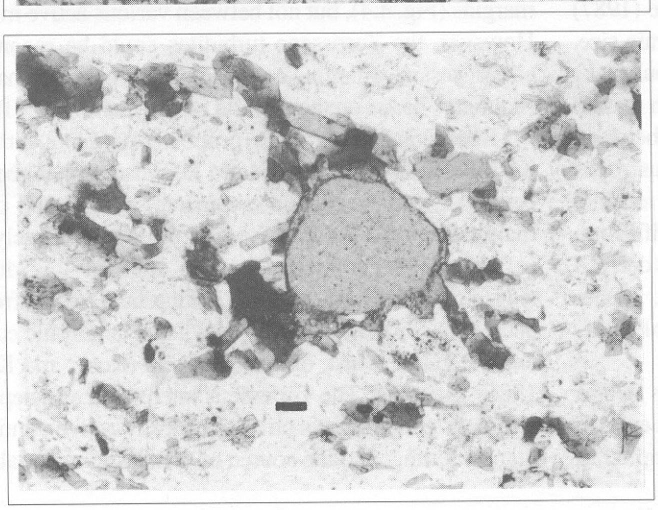


Plate 9.D: Overgrowth of tourmaline on a well-rounded detrital tourmaline grain. Scale bar 0.2 mm.

4.2.5 Discussion

Although meaningful point-counting of the samples was not possible (see 4.2.1), the identification of single grains diagnostic of particular environments is useful in determining the nature of the source terrain. Features indicative of plutonic rocks include myrmekitic intergrowths, perthite clasts, tourmaline-bearing grains and granitoid clasts. Indicators of a metamorphic source include the garnet-biotite-quartz schist clast and the ones with well-developed triple junctions. Carbonate clasts, chert fragments and the quartz-arenite clast are all derived from sedimentary rocks. The presence of these typical plutonic, metamorphic and sedimentary rock fragments indicates that the source was a granite-gneiss terrane, recycled orogen or a combination of these. No petrological evidence for volcanic arc-related sedimentation such as glass shards or andesitic lavas was observed. Also, no significant differences between the three siliciclastic formations were recognised, suggesting derivation from the same source area.

4.3 GEOCHEMISTRY

4.3.1 Introduction

The use of framework mineralogy for discriminating tectonic environments has been complemented recently by geochemical techniques (Bhatia and Taylor, 1981; Bhatia, 1983; 1985; Maynard *et al*., 1982; Roser and Korsch, 1986; Floyd and Leveridge, 1987). These geochemical criteria are based on the premise that the tectonic environment has a fundamental control on sedimentary composition (Maynard *et al*., 1982; Pettijohn *et al*., 1987). Other factors affecting geochemical composition include intensity of weathering and sorting (Sawyer, 1986), but these are themselves partly controlled by the tectonic setting.

Geochemistry is potentially of particular use in sedimentary rocks, such as the Zerrissene turbidites, in which the nature of the framework grains has been obscured or modified by deformation and metamorphism (Bhatia, 1983; Floyd and Leveridge, 1987). The use of tectonic discriminant diagrams in determining setting of sediments is, however, still in its infancy and no technique is commonly accepted as correct. For instance, the Al_2O_3/SiO_2 ratio is regarded by Bhatia (1983) as a reliable discriminator of tectonic environments, while Floyd and Leveridge (1987) are of the opinion that this ratio is dependent on grain size. Attempts at identifying the tectonic setting of igneous rocks from trace element compositions have met with varied success and it is possible that there is no simple link between composition and tectonic environment for igneous rocks (Hawkesworth *et al*. 1986). If the recognition of the tectonic environment of igneous rocks is difficult, then identification of the setting of sediments, which are in many cases derived from these igneous rocks, is even more difficult. Other complications would include glaciations and complex tectonic settings (Zuffa, 1987).

Samples were taken from all three formations over a wide geographic area (see Appendix 1 for localities). In general only coarse-grained greywackes were sampled as these are closest to the composition of the source rock (Taylor and McLennan, 1985).

4.3.2 Geochemistry of the Zerrissene turbidites

The major element analyses for the Zerrissene turbidites are given in Appendix 2A and 2B, and Harker variation diagrams for these data are shown in Fig. 4.1. The Brak and Amis River Formation samples plot in two distinct groups, but the two samples of the Zebrapüts Formation overlap both of these formations (Fig. 4.1). The separation of the Brak and Amis River Formations on these plots is possibly partly caused by subtle grain-size variations similar to those described by Roser and Korsch (1986) from the Torlesse terrane, New Zealand (Fig. 4.2), and is not a reflection of changing source areas. The Al_2O_3/SiO_2 ratio, however, only varies from 0.13 to 0.17 and this suggests that samples of a similar grain size have been analysed, as this ratio will be controlled by the amount of quartz versus clay minerals present. This was confirmed by thin-section investigations. Further analyses over a wide range of grain sizes are required before the geochemical variations between formations can be regarded as significant.

The chemical index of alteration (CIA) proposed by Nesbitt and Young (1982) is a measure of the amount of chemical weathering of the source rock. This index is calculated from the equation:

$(Al_2O_3/(Al_2O_3 + CaO + K_2O + Na_2O)\} \times 100$

Fresh granites typically have a CIA of 45-55 (Sawyer, 1986). In samples of the Zerrissene turbidites with no calcite present the CIA varies from 62-66, suggesting significant chemical weathering. The average greywacke of Pettijohn (1963) has a similar CIA of 64. It must be borne in mind that this index will be influenced by sorting and should therefore be interpreted with caution. The K_2O/Na_2O ratio which varies from 0.61 to 1.00 is typical for Palaeozoic greywackes as defined by Maynard *et al*. (1982) (Fig. 4.3).

4.3.3 Tectonic discrimination

Maynard *et al*. (1982) compared the chemistry of modern and ancient deep-sea sediments from known tectonic environments. On the basis of SiO_2/Al_2O_3 *vs* K_2O/Na_2O plots they could discriminate between active and passive margins (Fig. 4.4), but not between various active margins. However, the Zerrissene turbidites could be regarded as either passive margin, continent margin-arc or back-arc sediments based on the method of Maynard *et al*. (1982).

Roser and Korsch (1986) proposed that a plot of the ratio K_2O/Na_2O vs SiO_2 can be used as a discriminant between oceanic island-arc, active continental margin and passive margin settings. The Zerrissene turbidites plot largely within the active continental margin field of this diagram (Fig. 4.5). These data plot in a similar field to data from the accretionary Torlesse terrane and its metamorphic equivalent, the Haast schist, of New Zealand (Roser and Korsch, 1986; Fig. 4.5). The Torlesse is interpreted as a deposit of sediment gravity flows on a tectonically active continental margin setting, and the source of the clastic material was a continental volcanic-plutonic arc

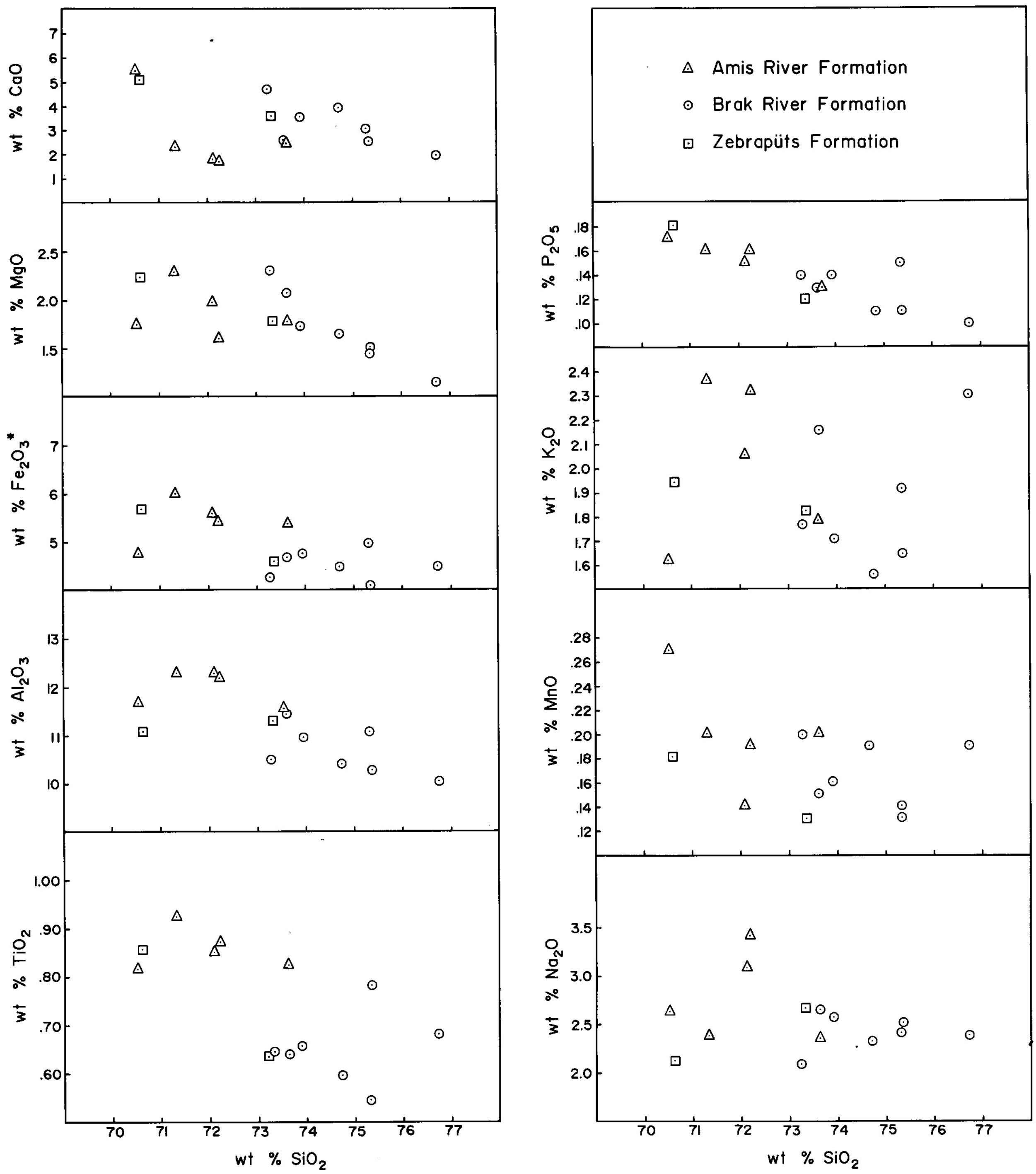


Fig. 4.1: Harker variation diagram for the analysed samples of Zerrissene greywackes.

(MacKinnon, 1983).

Bhatia (1983) suggested that plots of (Fe_2O_3* + MgO) versus TiO_2, K_2O/Na_2O, Al_2O_3/SiO_2 and $Al_2O_3/(CaO + Na_2O)$ could be used to discriminate between oceanic island-arc, continental island-arc, active continental margins and passive margins. The data from the Zerrissene turbidites are, however, ambiguous when plotted on his discrimination diagrams (Fig. 4.6). The data plot in two adjacent fields (oceanic and continental island-arc; Fig. 4.6A), no field at all (Fig. 4.6B), partly within two overlapping fields (continental island-arc and active continental margin; Fig. 4.6C), or almost entirely within a single field, but still takes in two fields (continental island-arc; Fig.4.6D). The only field in which the data do not plot at all is that of the passive margin environment.

4.3.4 Discussion

Variations between the composition of the Brak and Amis River Formations could be due either to changes in source rock composition or to differences in grain size between samples.

The various discrimination diagrams proposed by Maynard *et al.* (1982), Bhatia (1983) and Roser and Korsch (1986) are inconclusive in determining the specific tectonic environment of deposition of the Zerrissene turbidites. The

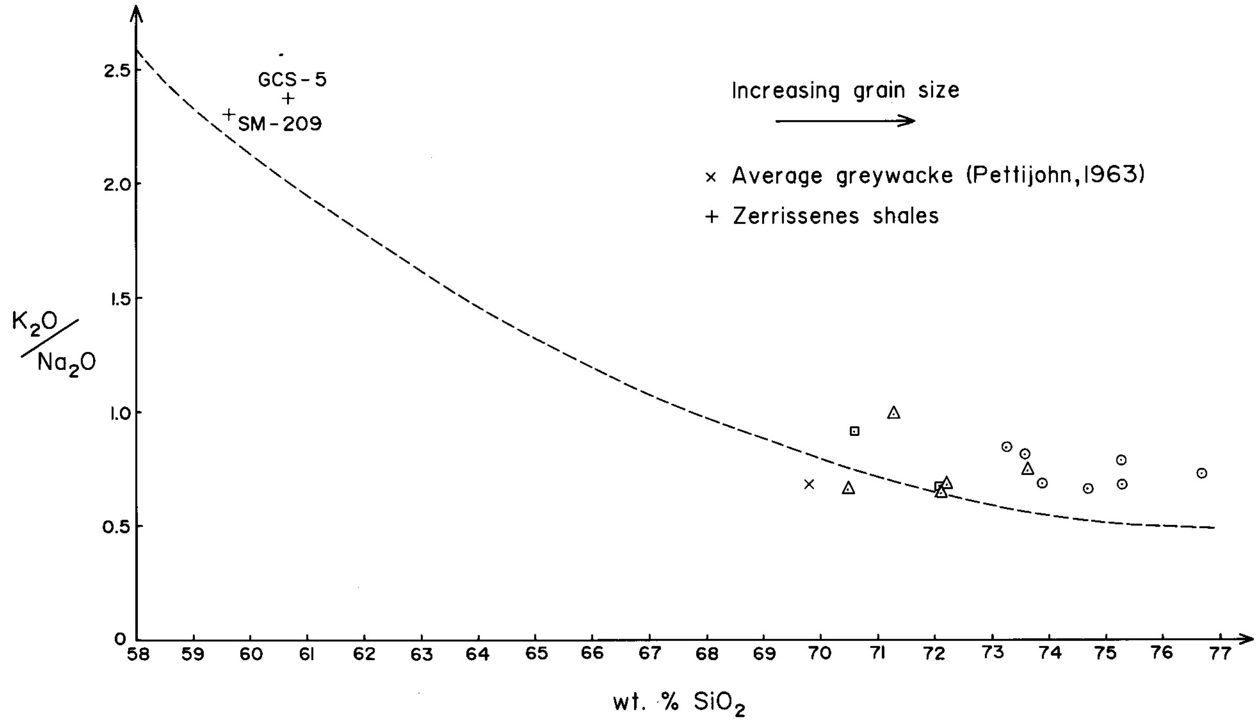


Fig. 4.2: Grain size variations strongly affect the geochemical composition of sediments, with decreasing grain size corresponding to a decrease in SiO_2, and an increase in the K_2O/Na_2O ratio (adapted from Roser and Korsch, 1986). Symbols as for Fig. 4.1.

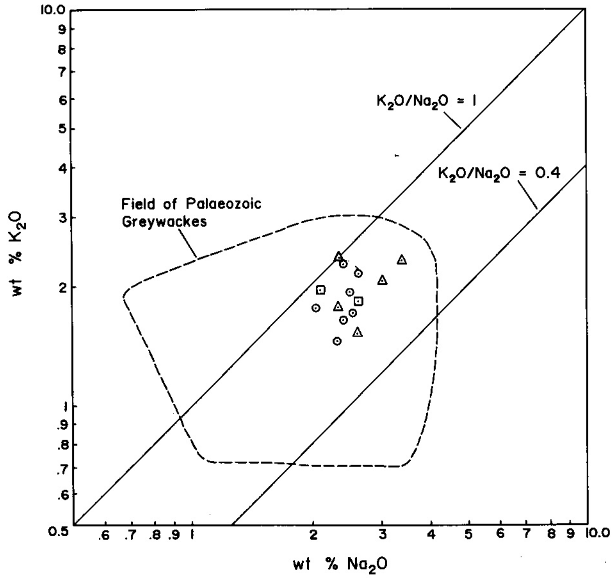


Fig. 4.3: The composition of the Zerrissene greywackes is typical for Palaeozoic greywackes (modified from Crook, 1974 and Floyd and Leveridge, 1987). Symbols as for Fig. 4.1

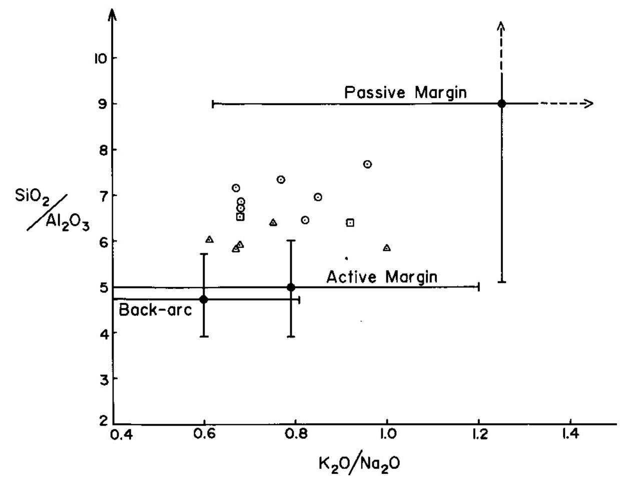


Fig. 4.4: Chemistry of modern deep-sea sands compared with that of the Zerrissene greywackes. Extensive overlap of modern sands from different tectonic settings occurs and it is therefore difficult to categorise the ancient greywackes. The error bars on the modern analyses are one standard deviation each side of the mean (from Maynard *et al.*, 1982). Symbols as for Fig. 4.1.

plots of Bhatia (1983) and Roser and Korsch (1986) exclude the passive (trailing) margin environment, but cannot differentiate between different types of active margin. The plot of Maynard *et al.* (1982) is also inconclusive in discriminating between various active margins, and cannot clearly differentiate these environments from that of a passive margin.

Geochemical methods of discriminating tectonic environments are therefore unsuccessful in uniquely identifying the setting of the Zerrissene turbidites.

4.4 PALAEOCURRENT ANALYSIS

4.4.1 Introduction

Palaeocurrent analysis of turbidite systems rarely indicates the source area adequately (Bouma *et al.*, 1985). In the Zerrissene system the tight folding further complicates the evaluation of palaeocurrent data and corrections need to be made for these tectonic effects. Reliability of the data is further reduced by the rarity of flute and groove casts. Rip-

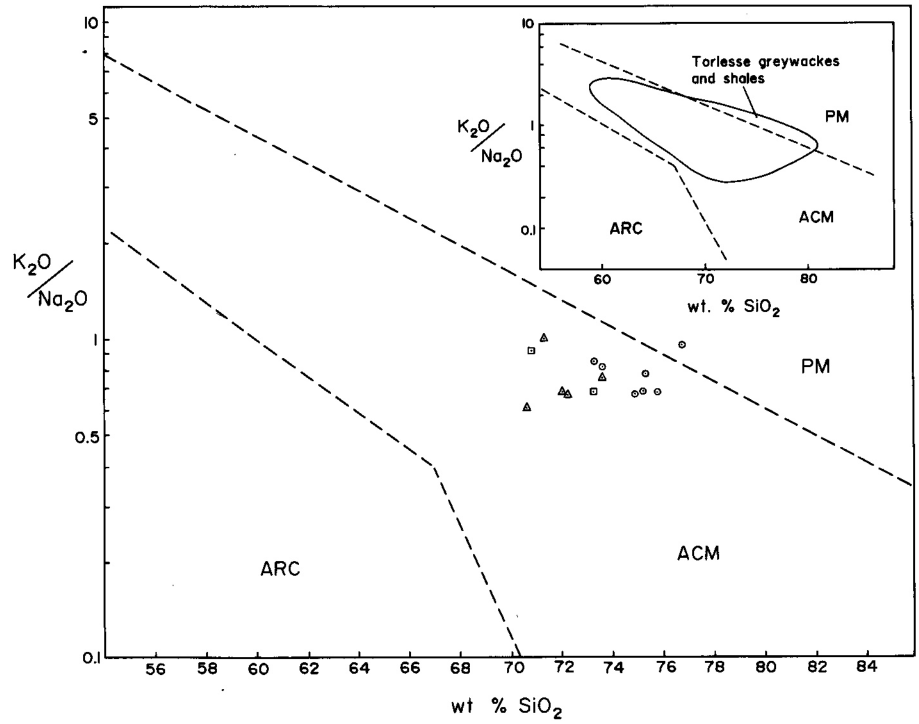


Fig. 4.5: Tectonic discrimination diagram for the Zerrissene greywackes. Note the similarity between the compositions of the Zerrissene greywackes and those from the Torlesse terrane of New Zealand (inset) (modified from Roser and Korsch, 1986). Symbols as for Fig. 4.1. PM - passive margin; ACM - active continental margin; ARC - volcanic island-arc.

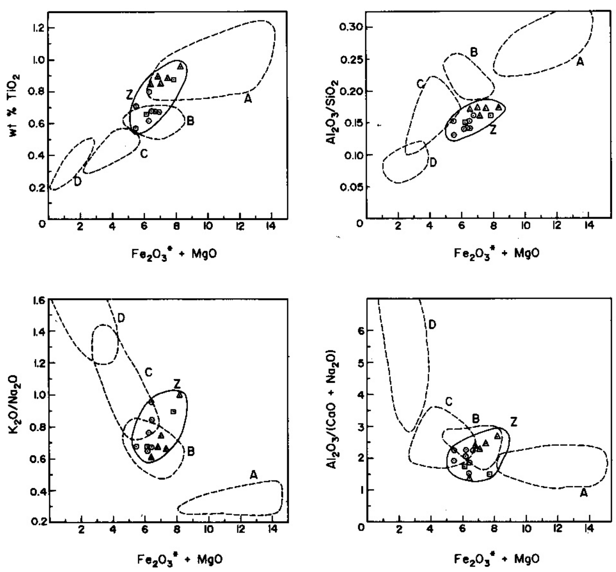


Fig. 4.6: Various plots of major element ratios are used for tectonic setting discrimination of greywackes. Shown here are the Zerrissene data plotted on the diagrams proposed for this purpose by Bhatia (1983). Symbols as for Fig. 4.1. The key to the tectonic setting is: A - oceanic island-arc; B - contnental island-arc; C - active continental margin; D - passive margin; Z - Zerrissene greywacke. See text for discussion.

ples, which are not reliable palaeocurrent indicators (Miall, 1974), are the most common form of directional structure found, and were the structure most often measured. No palaeocurrents were measured in areas of interference folding as reconstruction of these is unreliable. The results of the palaeocurrent analysis should therefore be treated with caution.

4.4.2 Palaeocurrent data

The palaeocurrent data for the three siliciclastic formations are shown in Figs 4.7 to 4.9. Limited data are available for the Zebrapüts Formation and the only measurements obtained were from the Goantagab River section. A southwesterly source area is however suggested by the data from this locality (Fig. 4.7).

The Brak River Formation appears to have been derived predominantly from a northwesterly source (Fig. 4.8A and 4.8B). These data are derived from both the entire area and most of the stratigraphic interval of this formation, except the basin plain. The nearly 90° spread of the data is common in turbidite systems as the currents tend to fan out on the basin floor.

The Amis River Formation is similarly derived from the northwest (Fig. 4.9). The data are again obtained from the entire outcrop area, including the distal deposits in the east. No palaeocurrent data are however available for the quartzwacke unit as no directional structures have been observed.

4.4.3 Discussion

Although only limited data are available, both the Brak: and Amis River Formations appear to have been derived from the northwest, whereas the Zebrapüts Formation had a southwesterly source area. Previously Miller *et al.* (1983) and the Geological map of Cape Cross (1988, rock-relation

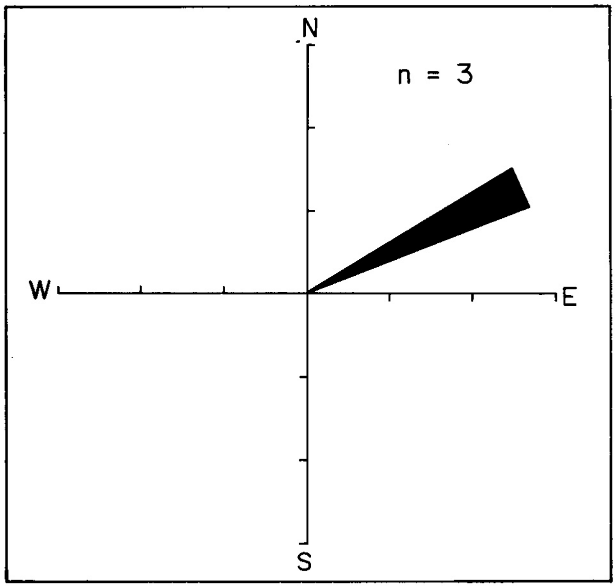
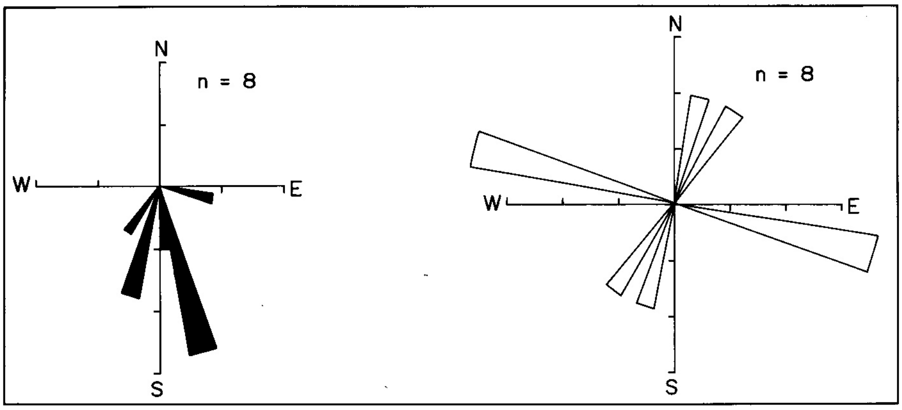


Fig. 4.7: Palaeocurrent rose diagram (in 10° sectors) for the Zebrapüts turbidites. This formation has few directional structures.

diagram) suggested that the succession was derived from the east. This was based on the evidence of a few cross-bedding measurements, the presence of the quartzwacke in the east and on the overall distribution of facies. The quartzwacke is however a local feature and appears to have a different source area to the main succession (see section 3.3.4.5). The facies descriptions given by Miller *et al.* (1983) are superficial and a new interpretation has been presented here (see section 3.3.4.6). Further, ripples are generally unreliable indicators of the source area (Miall, 1974) and palaeocurrent studies rarely accurately indicate the source area (Bouma *et al.*, 1985). A westerly source area for the Brak River Formation is suggested by the increase in dropstones westwards. A northeasterly source area is unlikely as during deposition of the siliciclastics the region to the northeast was covered by thick carbonate sequences (Hedberg, 1979). The facies analysis of the Amis River Formation indicates that this formation is more proximal in the west. A similar source area for these two formations is confirmed by their similar major element compositions (see section 4.3.2).

4.5 SYNTHESIS

The geochemical and petrographic characteristics of the Zerrissene turbidites suggest that the succession was derived from a continental basement - recycled orogen type of terrane which was possibly exposed on an active margin. This contradicts the sedimentological evidence which indicates that the Zebrapüts and Brak River Formations developed on an unconfined basin floor adjacent to a passive margin. The margin, whether active or passive, lay somewhere to the north and west of the current outcrop area as the palaeocurrent evidence supports the conclusion ob-tained from detailed facies analysis that the sediments in the west are more proximal than those in the east. Possible source areas are discussed in chapter 5.



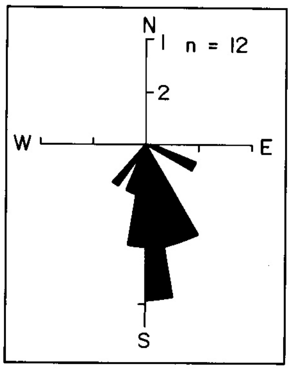


Fig. 4.8 (left): Palaeocurrent rose diagrams (in 10° sectors) for the Brak River Formation. **A:** Directional structures; **B:** Trend only. A strong component of transport from the north and west is apparent.

Fig. 4.9 (right): Palaeocurrent rose diagram (in 10° sectors) for the Amis River Formation. The direction of sediment transport is dominantly from north to south.

CHAPTER FIVE: SEDIMENTOLOGICAL DEVELOPMENT

5.1 INTRODUCTION

Two widely contrasting views on the nature of the Zerrissene basin were proposed by Miller *et al.* (1983) and Porada *et al.* (1983). Miller *et al.* (1983) suggested, on sedimentological and stratigraphical evidence, that the basin was a deep-water one by at least 750 Ma, whereas Porada *et al.* (1983) proposed that the sedimentological characteristics of the sediments were indicative of a shallow-water origin, and that the structural features of the area could only be explained by the existence of a basement high which they termed the Huab Ridge. However, in a more recent paper Porada (1989) did not discuss the existence of the Huab Ridge and included this region in a portion of the early Damaran rift system (his Fig. 4). The new sedimentological evidence presented in chapters two and three supports the proposal by Miller *et al.* (1983) that the area was a deep-water basin for all of the exposed sedimentary record.

The correlation of the Zerrissene sequence with the entire Swakop Group as suggested by Miller *et al.* (1983) is critical with regard to palaeogeographic models. The correlation is based largely on the presence of large isolated clasts of glacial origin in the Brak River Formation, a unit which is both underlain and overlain by carbonate units (Table 1.1). This unit was therefore correlated with the Chuos Formation, a diamictite of probable glacial origin (Gevers, 1931; Henry *et al.*, 1986; Badenhorst, 1987; section 3.2.9) which outcrops over a wide area of the Damara Orogen and which is similarly sandwiched between two carbonate units (Table 1.1). This correlation is accepted here but should, however, be regarded with caution as at least two major glaciations have been recorded from late Proterozoic successions in Africa (Harland, 1983), and long distance correlations are difficult as there is little age control (Crowell, 1983).

Palaeogeographic reconstructions of the Zerrissene basin are limited by the lack of knowledge about terranes adjacent to the exposed parts of the basin at the time of deposition as the area is surrounded by structural breaks, intrusives or younger cover (see section 1.2). The lack of macro- or microfossils for dating purposes makes time inferences difficult; nor do we have detailed knowledge of sea-level variations in this area during the late Precambrian. Global sea-level variations recorded elsewhere may be negated by local tectonic activity and may therefore not be applicable to the Zerrissene basin.

5.2 SEDIMENTARY EVOLUTION OF THE ZERRISSENE SYSTEM

The oldest sediments exposed in the basin are distal outer-fan lobe, lobe-fringe and basin-plain deposits of the Zebrapüts Formation (see section 3.3.2.5). These sediments were derived from the west to southwest and were deposited on an unconfined surface in what was already a deep-water basin (Miller *et al*., 1983). Large, unconfined surfaces of this type are most likely to develop adjacent to a passive margin. The lack of interbedded volcanic material also indicates that the sediments were deposited near a passive margin. The lithological characteristics of the floor of this basin are, however, at present unknown. Intermittent hiatuses in turbidite deposition, indicated by facies G beds, are possibly caused either by lobe abandonment or a decrease in sediment supply related to changing sea-levels. Lobe abandonment is the most likely cause of regular breaks in sand deposition in a submarine turbidite system. This type of mud-dominated system with interbedded lobes is similar to the type II lobes described by Pickering (1981; 1985; see Fig. 3.4) from the Precambrian Kongsfjord Formation of northern Norway.

This period of siliciclastic sedimentation was followed by carbonate turbidites with interbedded pelagic muds (basal portion of the Brandberg West Formation). The change from siliciclastic to carbonate activity must reflect a relative rise in sea-level, as development of carbonates is favoured by higher sea-levels (Schlager and Ginsburg, 1981; Mullins, 1983; Droxler and Schlager, 1985). No glacial sediments are known to be associated with the Zebrapüts Formation and the rise in sea-level is therefore probably tectonically, rather than eustatically, controlled. The source of the carbonates cannot be defined adequately as there are no reliable palaeocurrent indicators. Progradation of the system resulted in the development of a blue-grey, structureless, peri-platform ooze at the top of the Brandberg West Formation.

A return to predominantly terrigenous sedimentation followed (Brak River Formation). The presence of some hemipelagic blue marble near the base of the Brak River Formation indicates that the carbonate source was still active for a period after the initiation of siliciclastic sedimentation. The floor of the basin appears to have been unconfined as the sandy lobe deposits are interbedded with pelagic muds (facies G), indicating lobe switching. A relatively higher proportion of sand is, however, present in the Brak River Formation than in the Zebrapüts Formation. The basal portion of this system is similar to the type I lobes of Pickering (1981; 1985).

The first glacial drops tones appear in the upper parts of the Brak River sequence, indicating a climatic change with an associated decrease in sea-level. The turbidites, however, do not need to have the same source as the dropstones as icebergs can move large distances from the glacier or ice-shelf from which they are derived. For example, the furthest north that an iceberg, derived from the Antarctic ice-shelf, has been reported is 26°30'S; 25°40' (May, 1988), and it is therefore feasible that drops tones could also have been transported this far. This is substantially further than sediments are now, or ever have been, transported from the Antarctic continental shelf by gravity flows. Later melting of the glaciers resulted in a rise of sea-level, preventing terrigenous material from reaching the basin. A period of non-turbidite deposition resulted and is represented by the shale unit at the top of the Brak River Formation. This unit could represent a substantial time period.

The climatic warming and the associated rise in sea-level again favoured development of carbonate bioherms and their related slope and basin deposits (Gemsbok River Formation). Palaeomagnetic evidence from other portions of the Damara Orogen indicates that the area was in low latitudes during this period (McWilliams and Kröner, 1981), thus further favouring the development of carbonates. A sequence similar to that which occurred during deposition of the Brandberg West Formation resulted. Distally, carbonate turbidites with interbedded pelagic shales were deposited, whereas hemipelagic peri-platform oozes settled out adjacent to the shelf. Progradation of this system resulted in the vertical profile observed today.

A fall in sea-level is indicated by a return to siliciclastic sedimentation (Amis River Formation). The increase in sand content in the Amis River Formation compared to the Zebrapüts and Brak River Formations suggests that the basin was possibly either more restricted than previously, or that there was an increased sediment supply to the basin. The basin was also possibly fault-controlled on its landward margin, as the system is dominated by a non-cyclic succession of greywackes with less interbedded shale in the west, and thinner beds in the east. This is similar to facies variations described from the Cumberland Bay Formation of South Georgia in which sedimentation was fault-controlled (Macdonald, 1986; see 3.3.4.6). The quartzwacke which is found in the east of the outcrop area was possibly derived from a more proximal source on the lateral margins of the basin. A cartoon illustrating the sedimentological development of the siliciclastics is shown in.Fig. 5.2.

5.3 A MODEL

Detailed published models of the Damara Orogen are largely "fixist" in nature in that the various basins are regarded as having formed *in situ* (see reviews by Miller, 1983 and Martin, 1983). These models also concentrate largely on the intra-continental arm of the orogen. For example Miller *et al.* (1983) recognised that the Zerrissene system was largely turbiditic, but, in his plate-tectonic model for the evolution of the Damara Orogen, Miller (1983) does not suggest any explanation for the development of the Zerrissene basin or southern Kaoko Zone. The intracratonic model of Porada *et al.* (1983) assumed the presence of the so-called Huab Ridge which was proposed by Porada *et al.* (1983), and they therefore do not discuss the origin of the basin. Porada (1979; 1989) proposed that the Ribeira Orogen of eastern Brazil was the western portion of the coastal branch of the Damara, and that the sediments exposed north of the current field area represent the foreland zone of an eastward-dipping subduction zone. The suture area now lies west of the present coastline under the southern Atlantic Ocean. This model, although adequate for the northern area, does not explain the structural characteristics of the Zerrissene basin as the folds are westward-vergent in this area, compared to eastward-vergent for the area to the north.

The evidence from palaeocurrents and facies variations presented in chapters three and four suggests that the majority of the Zerrissene siliciclastic turbidites were derived from a source to the west of the basin, in contrast to the easterly source proposed by Miller *et al.* (1983). The large size of the system and the high overall proportion of sand require a substantial supply of sediment, which was most likely provided by a major fluvial system, possibly similar to the modem-day Mississippi River.

The points that therefore need to be accounted for in developing a palaeogeographic model for the area include the initial southwestern source area during deposition of the Zebrapüts Formation, which later moved to the northwest, the large size of the basin and the passive nature of its margin, which later became more restricted and possibly fault-bounded. The shift in source area could have been caused either by glacio-eustatic adjustments of the crust during and after the Chuos glaciation, or by tectonic processes related to the developing orogen, or by a combination of these two processes. Porada (1989) has proposed a reconstruction of the Pan-African and Brasiliano rift systems which would accommodate this model in which the sediments are derived from the west (Fig. 5.1).

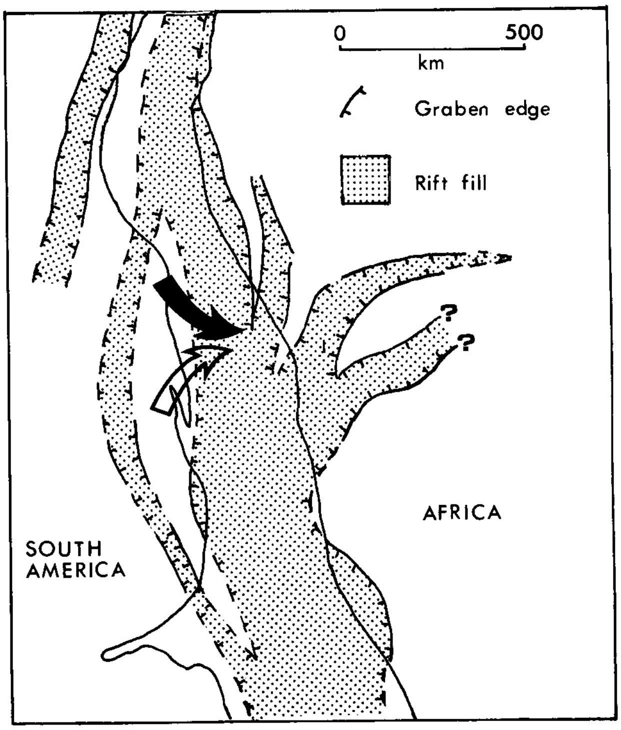


Fig. 5.1: Map showing the Pan-African and Brasiliano rift systems which developed on the present southwest coasts of Africa and southeastern America (after Porada, 1989). The direction of transport of the Zerrissene sediments is shown by the arrows; the open arrow representing the Zebrapüts Formation and the solid one the Brak River and Amis River Formations.

This model is also fixist, but more recent tectonic models for the evolution of the coastal branch of the Damara Orogen however suggest that major left-lateral movement occurred on the west coast of Africa during the late Precambrian (Hoffmann, 1987; Hälbich *et al.*, 1988). Hoffmann (1987) proposed that this basin (his Ugab terrane) was ac-

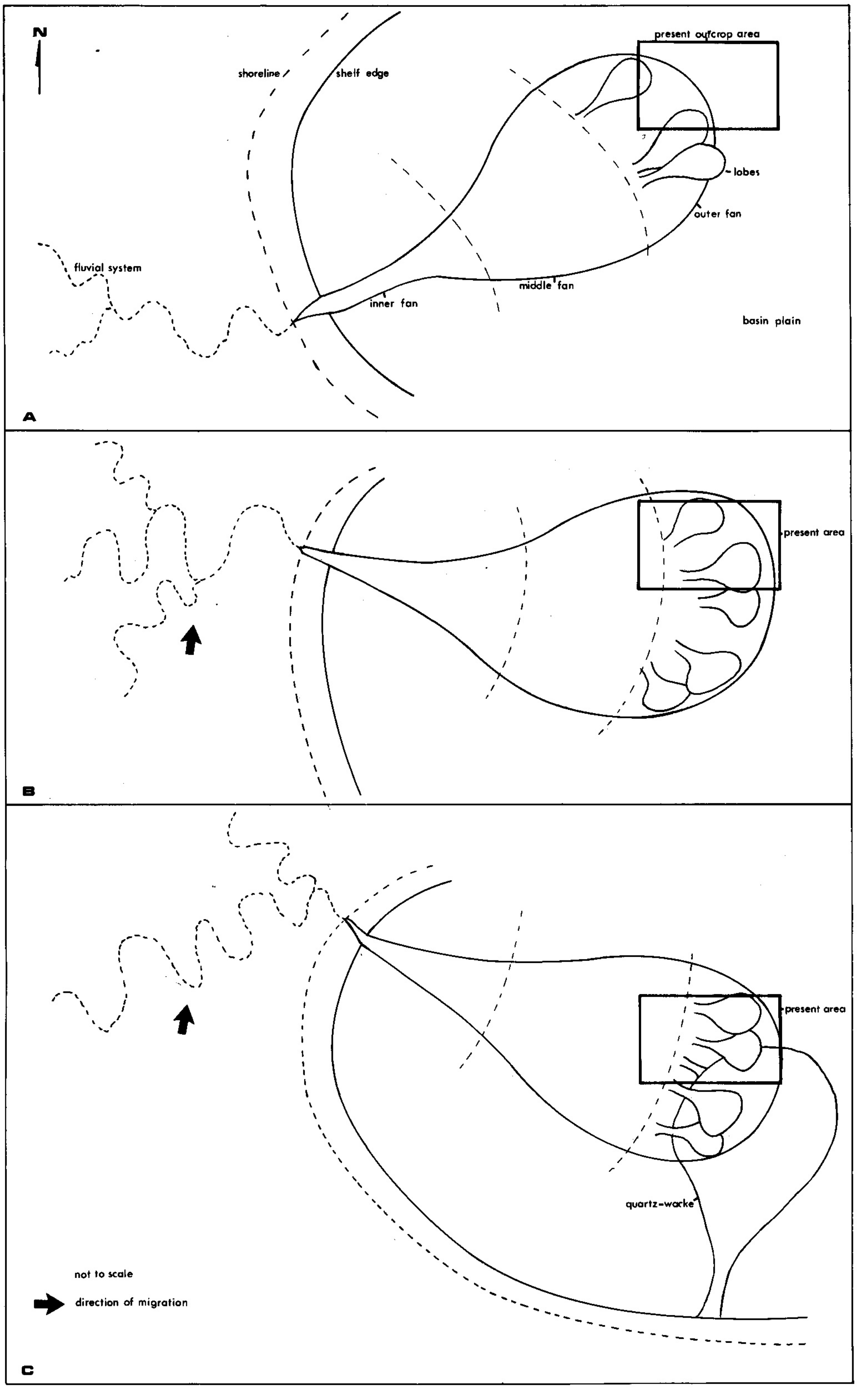


Fig. 5.2: Cartoon illustrating the envisaged evolution of the Zerrissene siliciclastic deposits. The lack of reliable palaeocurrent indicators in the carbonates precludes development of a similar model for their development. Notably only the distal parts of the siliciclastic system are preserved in the lower Ugab River area. Firstly, an open ocean appears to have been already fully developed by the time that the turbidites of the Zebrapüts Formation were deposited (a). The next episode (b) of siliciclastic sedimentation, the Brak River Formation, also occurred on an unconfined ocean floor, and was accompanied in its later stages by a glacial period. The final stage (c) of exposed sedimentation possibly developed in a more restricted basin with a voluminous sediment supply. Indications of lateral sources are the presence of the coarse-grained quartz-wacke beds interbedded with the distal sediments in the east.

creted to the Congo craton during continental collision. No palaeontological evidence is available to test this model, but the minor amount of carbonate present in the succession compared to the estimated 5 km on the adjacent shelf (Hedberg, 1979) supports the proposal that these dominantly siliciclastic sediments could not have been derived from this source.

Potential source areas are therefore now either under the Atlantic Ocean or are located on the east coast of South America. Further detailed studies on the tectonic history of the Zerrissene basin are required before a better understanding of both the sedimentological and structural development of this interesting area is acquired.

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APPENDIX 1

Geochemical sample localities

SAMPLE	LATITUDE	LONGITUDE	FORMATION
GCS-4	21°02′13″S	14°04′00″E	Brak River
GCS-5	21°01′08″S	14°01′39″E	Brak River
GCS-6	20°59′50″S	14°01′31″E	Amis River
GCS-7	20°59′47″S	14°01′24″E	Amis River
GCS-8	20°51′19″S	14°14′53″E	Zebrapüts
GCS-11	21°00′07″S	14°01′39″E	Brak River
GCS-12	21°02′08″S	14°04′00″E	Brak River
GCS-13	21°02′11″S	14°05′15″E	Amis River
GCS-14	21°03′32″S	13°55′45″E	Brak River
GCS-15	21°05′29″S	13°48′05″E	Amis River
GCS-16	20°50′28″S	14°15′50″E	Zebrapüts
1/87	21°01′18″S	14°27′13″E	Amis River
5/87	21°02′08″S	14°04′22″E	Brak River
6/87	21°02′09″S	14°04′21″E	Brak River
12/87	20°55′31″S	13°49′51″E	Brak River
*SM-209	20°56′02″S	14°07′14″E	Brak River

*S.C. Milner (pers. comm., 1988).

APPENDIX 2A

Major element analyses of Zerrissene turbidites

	GCS-4	GCS-5	GCS-6	GCS-7	GCS-8	GCS-11	GCS-12	GCS-13	GCS-14	GCS-15	GCS-16
SiO_2	75.99	72.49	71.74	68.22	68.12	72.53	74.42	70.55	71.97	71.65	72.03
TiO_2	0.67	0.57	0.85	0.78	0.82	0.64	0.77	0.91	0.63	0.86	0.62
Al_2O_3	9.90	10.08	12.19	11.27	10.62	10.70	10.15	12.09	11.18	12.07	11.06
$Fe_2O_3^*$	4.42	4.35	5.51	4.62	5.45	4.62	4.86	5.92	4.56	5.35	4.46
MnO	0.19	0.18	0.14	0.26	0.17	0.16	0.13	0.20	0.15	0.19	0.13
MgO	1.12	1.59	1.97	1.69	2.16	1.70	1.43	2.25	2.01	1.59	1.76
CaO	1.96	3.82	1.80	5.53	4.98	3.40	2.43	2.09	2.51	1.64	3.56
Na_2O	2.38	2.25	3.07	2.55	2.05	2.52	2.48	2.33	2.59	3.40	2.62
K_2O	2.29	1.51	2.05	1.57	1.88	1.68	1.90	2.34	2.11	2.30	1.80
P_2O_5	0.10	0.11	0.15	0.16	0.17	0.14	0.15	0.16	0.13	0.16	0.12
H_2O^-	0.39	0.35	0.33	0.44	0.36	0.22	0.34	0.23	0.85	0.19	0.31
LOI	1.85	2.14	0.80	3.66	3.91	2.08	1.71	0.98	1.82	0.95	2.33
TOTAL	101.24	99.44	100.58	100.75	100.69	100.39	100.75	100.04	100.50	100.34	100.80

	1/87	5/87	6/87	12/87	SM-209
SiO_2	71.22	70.00	59.20	73.39	55.56
TiO_2	0.51	0.61	0.89	0.82	0.78
Al_2O_3	10.44	10.01	16.63	11.38	16.00
$Fe_2O_3^*$	3.84	4.05	7.67	5.34	7.10
MnO	0.13	0.19	0.20	0.20	0.11
MgO	1.42	2.20	4.44	1.77	4.38
CaO	3.02	4.61	2.91	2.43	5.19
Na_2O	2.30	2.00	1.60	2.36	1.36
K_2O	1.56	1.69	3.81	1.78	3.13
P_2O_5	0.10	0.14	0.18	0.13	0.22
H_2O^-	0.27	0.33	0.22	0.31	0.16
LOI	4.23	3.94	2.33	1.00	6.47
TOTAL	99.02	99.77	100.09	100.89	100.46

1. $Fe_2O_3^*$ - all Fe as Fe_2O_3.
2. SM-209 - S.C. Milner (pers. comm., 1988).
3. LOI - loss on ignition.

APPENDIX 2B

Major element analyses of Zerrissene turbidites
(normalised to 100 per cent)

	GCS-4	GCS-5	GCS-6	GCS-7	GCS-8	GCS-11	GCS-12	GCS-13	GCS-14	GCS-15	GCS-16
SiO_2	76.74	74.77	72.12	70.58	70.65	73.94	75.38	71.37	73.63	72.22	73.38
TiO_2	0.68	0.59	0.85	0.81	0.85	0.65	0.78	0.92	0.64	0.87	0.63
Al_2O_3	10.00	10.39	12.25	11.66	11.01	10.91	10.28	12.23	11.44	12.16	11.27
$Fe_2O_3^*$	4.46	4.48	5.54	4.78	5.65	4.71	4.92	5.99	4.67	5.39	4.54
MnO	0.19	0.19	0.14	0.27	0.18	0.16	0.13	0.20	0.15	0.19	0.13
MgO	1.13	1.64	1.98	1.75	2.24	1.73	1.45	2.28	2.06	1.60	1.79
CaO	1.98	3.94	1.81	5.72	5.16	3.47	2.46	2.11	2.57	1.65	3.63
Na_2O	2.40	2.32	3.09	2.63	2.13	2.57	2.51	2.36	2.65	3.42	2.67
K_2O	2.31	1.56	2.06	1.62	1.95	1.71	1.92	2.37	2.16	2.32	1.83
P_2O_5	0.10	0.11	0.15	0.17	0.18	0.14	0.15	0.16	0.13	0.16	0.12

	1/87	5/87	6/87	12/87	SM-209
SiO_2	73.68	60.70	73.30	75.33	59.67
TiO_2	0.82	0.91	0.64	0.54	0.84
Al_2O_3	11.43	17.05	10.48	11.04	17.18
$Fe_2O_3^*$	5.36	7.86	4.24	4.06	6.86
MnO	0.20	0.21	0.20	0.14	0.12
MgO	1.78	4.55	2.30	1.50	4.70
CaO	2.44	2.98	4.83	3.19	5.57
Na_2O	2.37	1.64	2.09	2.43	1.45
K_2O	1.79	3.91	1.77	1.65	3.13
P_2O_5	0.13	0.18	0.14	0.11	0.24

1. $Fe_2O_3^*$ - all Fe as Fe_2O_3, except SM-209 where as FeO.
2. SM-209 - S.C. Milner (pers. comm., 1988).