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PRELIMINARY RESULTS OF A SEDIMENTOLOGICAL STUDY OF THE CHUOS FORMATION IN THE CENTRAL ZONE OF THE DAMARA OROGEN: EVIDENCE FOR MASS FLOW PROCESSES AND GLACIAL ACTIVITY

by

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ABSTRACT

The Chuos Formation in the study area comprises mainly diamictite with minor interbedded feldspar quartzites and magnetite quartzites. The diamictites and bedded units are thought to have been deposited in a subaqueous environment by mass-flow processes, including mudflows, turbidity currents and grain flows. The widespread occurrence of extra-basinal clasts in the diamictite and the presence of dropstones indicate a glacial influence on Chuos deposition. Syn-sedimentary faulting appears to have caused much of the thickness variations within the Chuos Formation. The depositional setting in the study area appears to have been a half-graben, with a fault-bounded, steepsided south-eastern margin, and a gently sloping north-western side. A model invoking redeposition of glacially-derived material down the slope of a tectonically active basin margin could account for most of the features observed in the Chuos Formation in the area.

1. INTRODUCTION

The study area is in central South West Africa/Namibia, about 80 km north-east of the coastal town of Swakopmund, (Fig. 1). Detailed geological mapping of this area was completed by Smith (1965). This area forms a sector of the Central Zone of the Damara Orogen, where the Damaran rocks have been subjected to high-temperature, medium-pressure regional metamorphism and multiple-phase deformation (Miller, 1983). Peak metamorphic conditions were achieved after cessation of the major phase of deformation. The maximum temperatures and pressures increased from Karibib (590°C and ± 2,5 Kbar) south-westwards towards the coast, where east of Swakopmund they reached 645°C and 3,4 kbar (Puhan, 1983). Locally, extensive partial melting and formation of migmatites have occurred. Strong isoclinal, upright to recumbent D_2 folds (Miller, 1983) produced the pronounced NE trending structural grain. Dome and basin structures are the result of interference of D3 and D_2 structures, with possible modification by diapiric uprising of remobilised basement and lower Damaran metasediments (Sawyer, 1981).

The Chuos Formation was first recognised by Young in 1928 (in Gevers, 1931) and described by Gevers (1931) from the type locality in the Chuos Mountains, about 30 km south-south-west of Usakos. Gevers (*op. cit.*) noted the remarkable similarity between the typical Chuos rocks and the Palaeozoic Dwyka tillite, and interpreted the former as being glacial in origin. Chuos-type rocks were subsequently found in other parts of the Damara Orogen, and the unit became an important marker, interpreted as chonostratigraphic. Studies on the

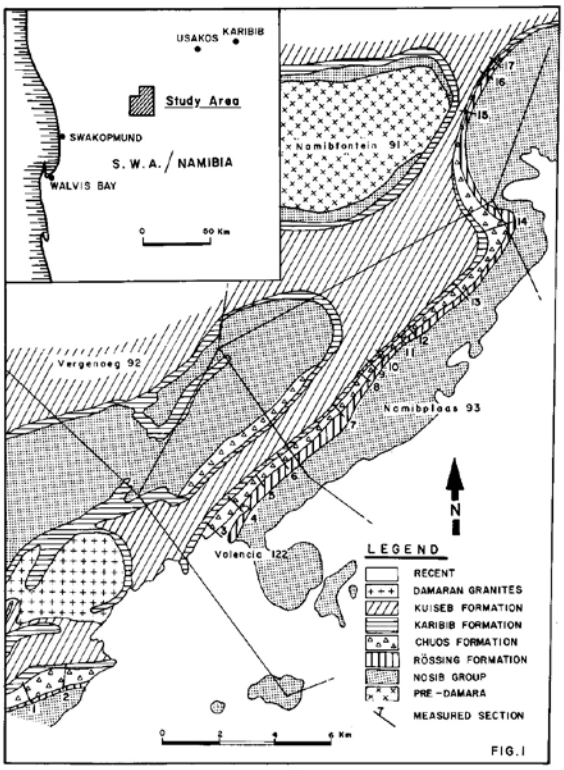

Fig. 1: Simplified geological map of the study area (modified after Smith, 1965) showing location of measured sections.

geosynclinal development of the Damara Orogen have led to a re-interpretation of the Chuos rocks as mainly mass-flow deposits (Martin, 1983; Miller, 1983) which were formed as a result of renewed rifting (Porada, 1983). Carbonate deposition in a marine environment was terminated by rapid basin deepening as rifting proceeded (Porada, 1983). Hoffmann (1983) and Miller *et al.* (1983) in their studies of the southern Damara Belt and the area west of the Brandberg, respectively, provide evidence for glacial activity during Chuos times.

The study involved detailed traverses at 0,1 to 5 km intervals across the Damara Sequence on the south-east limb of a major D_2 synclinal structure (Fig. 1). A pervasive schistosity, which is generally bedding-parallel, resulted from the D_2 deformation (Miller, 1983). Few sedimentary structures are preserved in the more incompetent carbonates and pelitic rocks, whereas some outcrops of competent siliceous rocks show remarkably well preserved sedimentary structures.

The rocks are highly recrystallised, but in some areas, recrystallization has not obscured original grain sizes, especially in siliceous rocks. Fine-grained carbonates and pelites are now coarse crystalline marbles and schists.

Pronounced thickness changes occur in the hinges and limbs of folds (Fig. 1), mainly in the more incompetent marbles and schists. However, the study has shown that some of the thickness variation is due to sedimentological control.

2. STRATIGRAPHY

The stratigraphic sequence in the study area is shown in Table 1.

The basement rocks to the Damara Sequence are mainly quartzo-feldspathic augen gneisses, metavolcanics and metasediments, which outcrop in the Abbabis Inlier (Marlow, 1981).

The Damara Sequence begins with the predominantly arenaceous Nosib Group. This is discordantly followed by the Swakop Group consisting of the carbonate-rich Rössing Formation, the clastic sedimentary rocks of the Chuos Formation, the carbonate rocks of the Karibib Formation and the argillaceous sediments of the Kuiseb Formation.

The dominant rock type of the Chuos Formation is a distinctive diamicite, a clast-bearing pelitic to semi-pelitic rock. During the course of field work, it became apparent that the Chuos Formation could only be recognised and defined as such by its clast-bearing nature. In all traverses, the unit was defined by the first and last occurrences of diamictite.

No interfingering relationships with the adjacent formations appear to exist. The Chuos Formation rests with a sharp contact on the underlying rocks of the Rössing Formation. In places, the Rössing Formation is not developed, and the Chuos Formation rests disconformably on Khan, Etusis or basement rocks.

In Fig. 2, a plot of percentage carbonate per 10 m interval within the Rössing Formation shows that the upper half is carbonate-poor. The figure shows clearly the discordance between the Chuos and Rössing units, implying that the Chuos Formation was deposited on an erosion surface which cut down at least into the Rössing Formation.

The last appearance of clasts defines the upper contact of the Chuos Formation in the area. This may be against a carbonate or non-pebbly biotite schist. The contact is sharp and no interfingering of facies has been observed.

As illustrated in Fig. 3, the Chuos Formation varies in thickness along the length of the synclinal limb. Although the variation may be due partially to thinning during deformation, gross thickness changes could be mainly ascribed to depositional conditions. This is because the rocks are generally competent. Most of the shearing is taken up in zones or units which have a higher pelitic component in the matrix. In general, the Chuos Formation along the south-east limb appears to have been folded without much internal deformation.

TABLE 1: Stratigraphy of the study area (after SACS, 1980)

	GROUP	FORMATION (maximum thickness)	LITHOLOGY
DAMARA SEQUENCE	SWAKOP	KUISEB (3000 m)	Massive biotite schists ± garnet, cordierite
		KARIBIB (700 m)	Mainly calcic and dolomitic marbles, quartz-biotite schists, calc-silicates
		CHUOS (700 m)	Massive diamictite, minor iron formation, pebbly quartzite
		RÖSSING (200 m)	Dolomitic marbles, quartzites, quartz schists, biotite schists, calc-silicates
	NOSIB	KHAN (1100 m)	Greenish pyroxene-amphiboie bearing feldspathic quartzites, biotite-amphibole schists
		ETUSIS (3500 m)	Meta-arkoses, feldpathic quartzites, minor conglomerate
ABBABIS BASEMENT			Quartz-feldspar augen gneisses, metavolcanics and metasediments

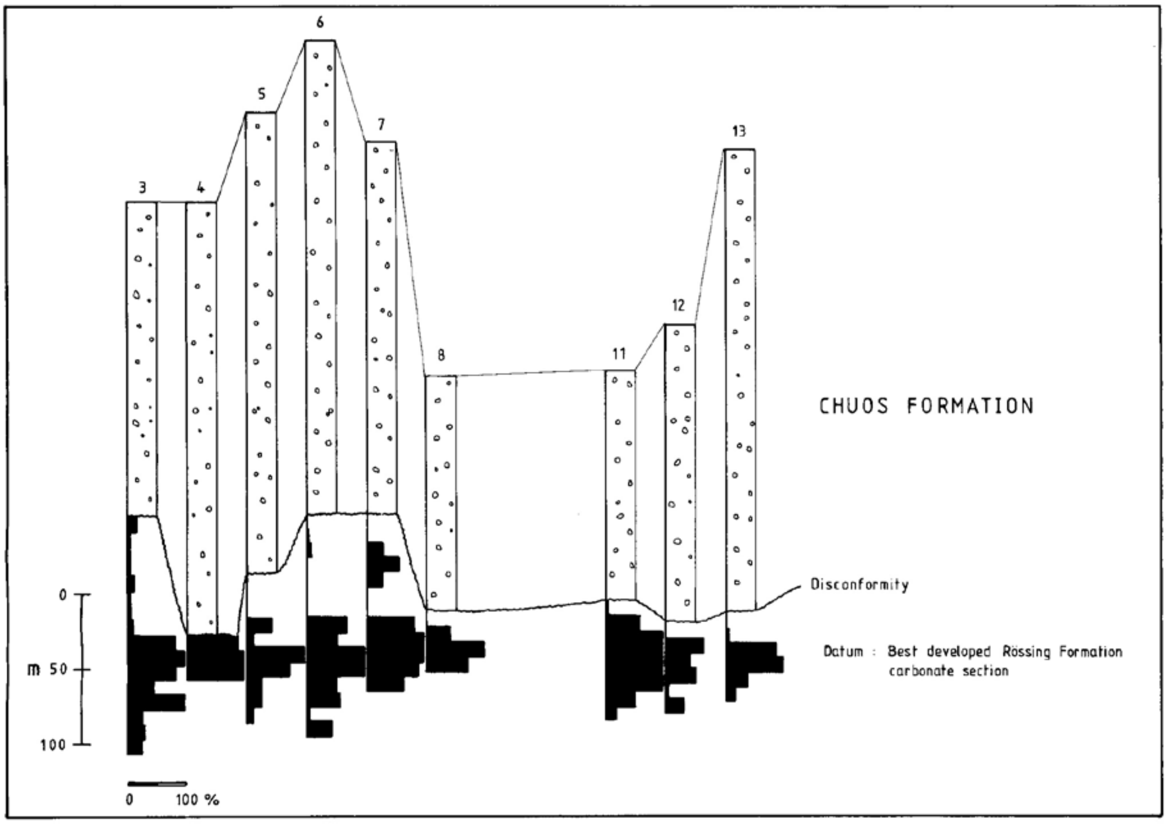

Fig. 2: A plot of the per centage of carbonate in each 10 m interval of the Rössing Formation. A discordance with the overlying Chuos Formation is apparent, with an erosion surface developed on the Rössing Formation. Location of sections shown in Figs. 1 and 3.

North of Namibplaas 93 (Fig. 1), the Chuos Formation thins out rapidly from about 350 m thick to nothing over a strike length of about 8 km. The Khan, Rössing and Karibib Formations thin out as well, with only the latter developed north of the Bergrus anticlinal structure. A fairly uniform thickness is maintained southwestwards over a strike length of 35-40 km.

On the north-west limb of the syncline, the Chuos Formation is generally less than 50 m thick. This limb is strongly deformed as indicated by flattening of clasts, so it is difficult to estimate how much of the thinning is tectonic and how much is primary. From the flattening of clasts, a thinning by about 50 per cent is possible, so it is likely that the original thickness never exceeded 100 m. About 5-7 km further to the north-west, on the southern part of Vergenoeg 92, the Chuos Formation is still found as a thin (less than 25 m thick), highly sheared, pebbly remnant between carbonate-rich sequences; in this area the unit no longer has an airphoto signature, and was not mapped by Smith (1965).

Overall, the Chuos Formation in the study area forms a wedge-shaped, sheet-like unit thinning to the north and north-west.

3. DESCRIPTION OF ROCK TYPES

The composite stratigraphic column shown in Fig. 4 indicates the occurrence of the rock types discussed below within the Chuos Formation.

3.1 Diamictite

This forms the bulk of the Chuos Formation. It consists of a massive, poorly sorted, clast-bearing rock with a pelitic to quartzitic matrix. It is well known in the literature, where it has been variously described as a tillite (Gevers, 1931) or as mixtite (Martin, 1983). The term “diamictite” is used in this paper in preference to “mixtite”, in accordance with Hambrey and Harland (1981, p. 23). In the study area the clasts consist of granite, gneiss, vein quartz, pink, grey, black and colourless quartzite, aplite, and more rarely carbonate, iron-rich quartzite, schist and amphibolite. The clasts are sub-angular to rounded, and are generally sporadically dispersed throughout the rock, forming from less than 5 per cent to over 50 per cent of the rock volume. No relationships between clast-type and abundance, and matrix-type were observed. Shearing in some zones has produced extreme elongations of pebbles into the schistosity. The clasts range in size from 2 mm granules to boulders several tens of metres in long dimension; the average size is generally less than 30 cm. No systematic variation in clast size from bottom to top of the formation was observed on any of the traverses. The largest clasts were found on the south-east limb of the syncline. The matrix consists of quartz, feldspar and biotite. The

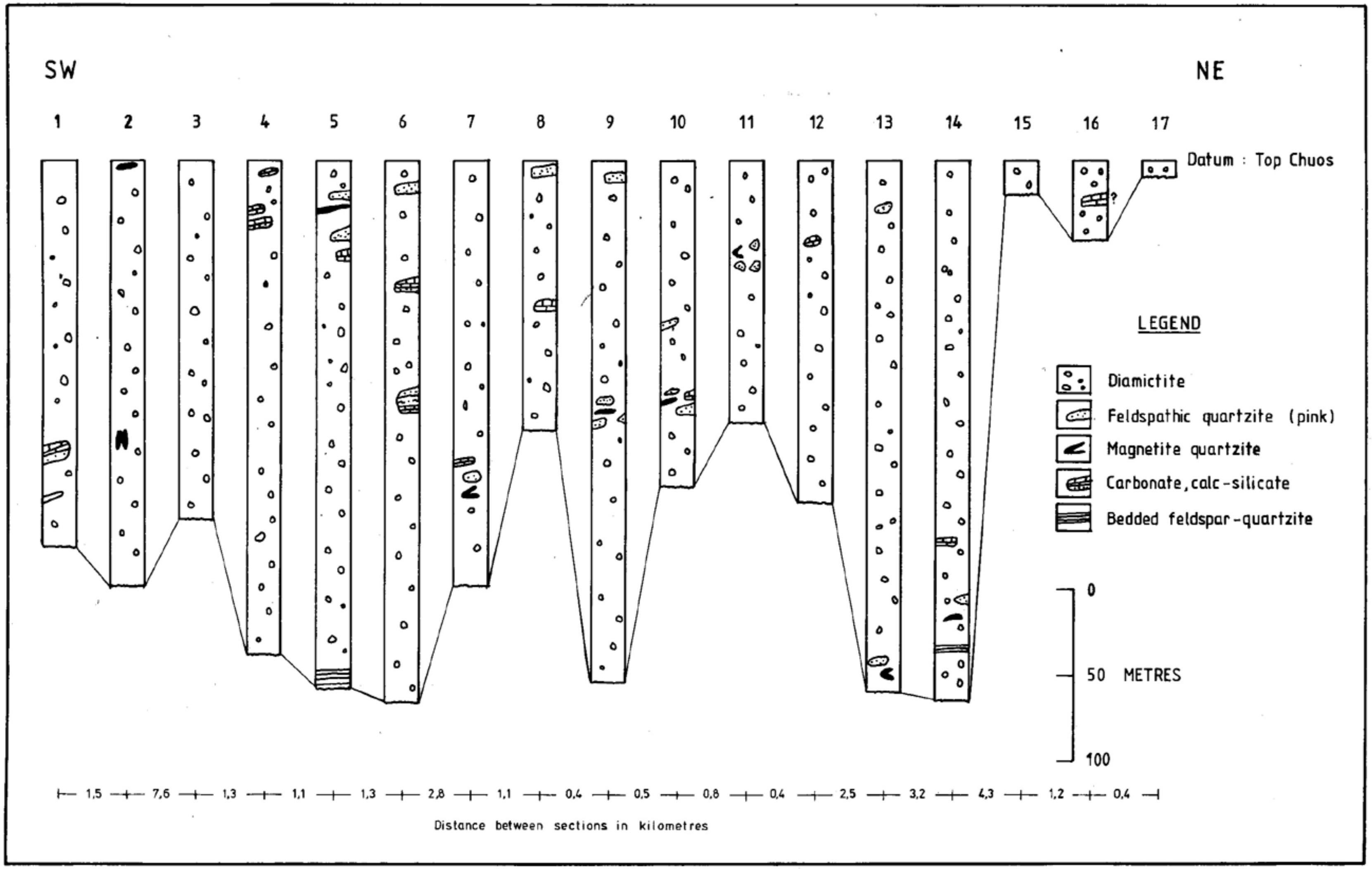

Fig. 3: Cross-sections of the Chuos Formation from 10 km south-west of Valencia 122 to Namibfontein 91 along the main study area, showing vertical and lateral variation.

rock is massive and uniform in appearance (Fig. 5), with a pervasive s_2 schistosity, defined by biotite orientation, well developed in more deformed zones.

3.2 Bedded Feldspar Quartzite and Pebbly Feldspar Quartzite

Pebbly feldspar quartzites (terminology of Winkler, 1974) were found in two traverses, one on the farm Valencia 122 and the other in the northern corner of Namibplaas 93. Detailed sections are shown in Figs. 6 and 7.

The beds on Valencia 122 occur at the base of the Chuos Formation. They have a sharp contact with underlying non-pebbly schist of the Rössing Formation, and are overlain by massive diamictite; they are thus part of the Chuos and not the Rössing Formation. The succession is continuous along strike for 700 m to 800 m, and was observed in only one traverse on Valencia 122, forming a sequence about 15 m thick. Along strike it becomes indistinct and appears to thin out, although it may also be sheared out.

Two main sub-types can be recognised, graded, pebbly feldspar quartzites and non-graded, pebbly feldspar quartzites. The two are interbedded, and are overlain by massive diamictite.

a) The graded, pebbly feldspar quartzites occur in beds from 2 to 15 cm thick. They grade both in grain size and composition. The contacts with overlying and underlying beds are sharp, and individual beds are persistent over several tens of metres (minimum) with a fairly uniform thickness. Macroscopically, the individual beds grade upward from coarse (pebbly) to granule-size grains to very fine grains, with

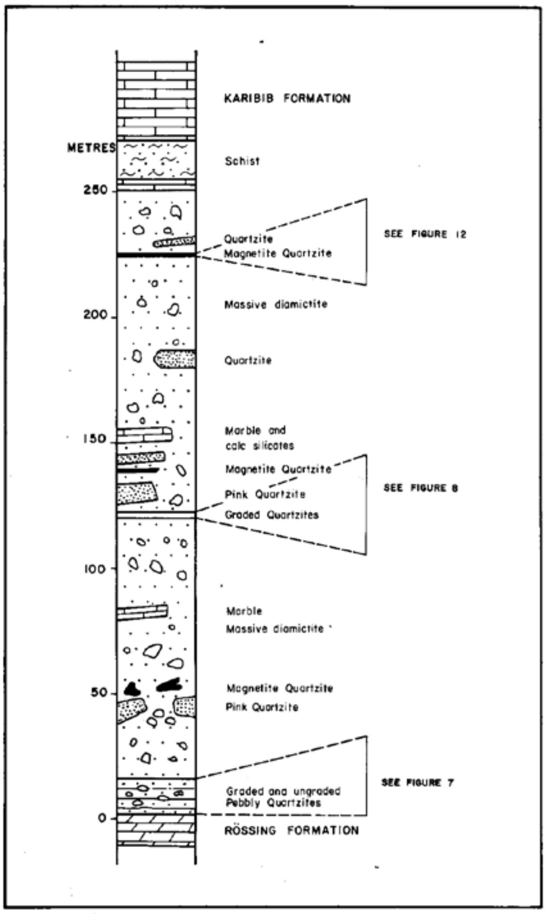

Fig. 4: Composite stratigraphic section of the Chuos Formation showing occurrence of rock types discussed in the text.

Fig. 5: Chuos diamictite in an area about 6 km south-west of Valencia 122. Large boulder is grey quartzite. Smaller clasts include quartz-feldspar gneiss and aplite.

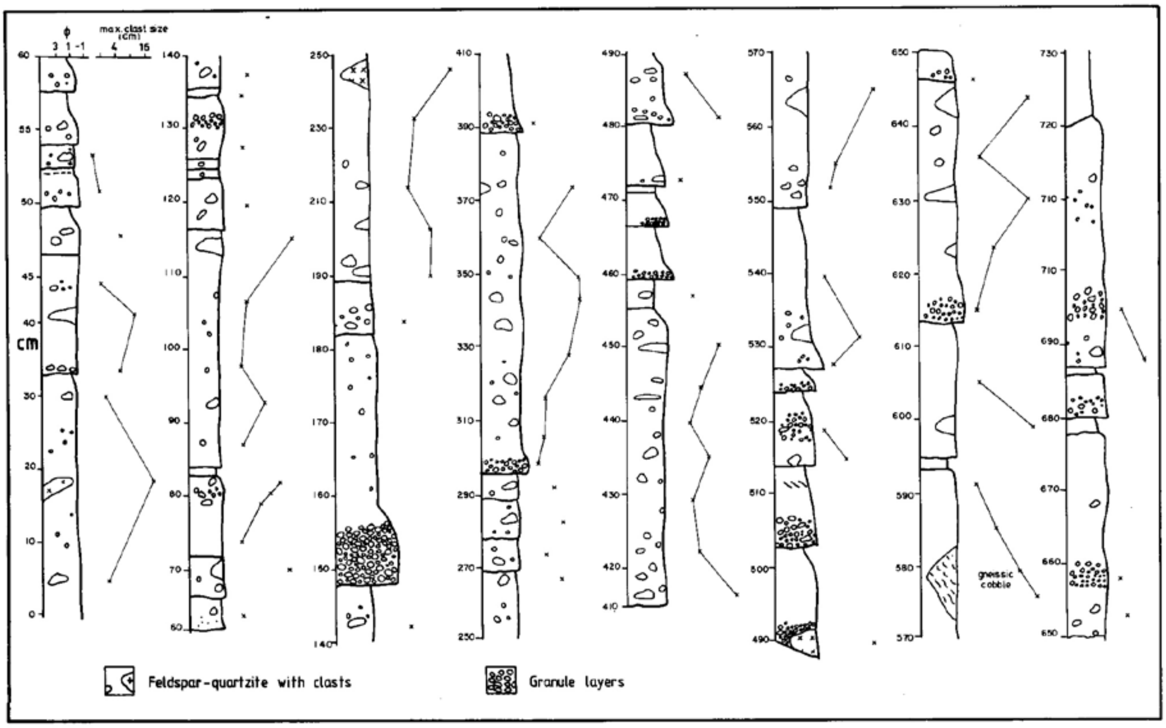

Fig. 6: Detailed measured sections of pebbly feldspar quartzites at the base of the Chuos Formation on Valencia 122. Note the presence of large clasts and graded beds.

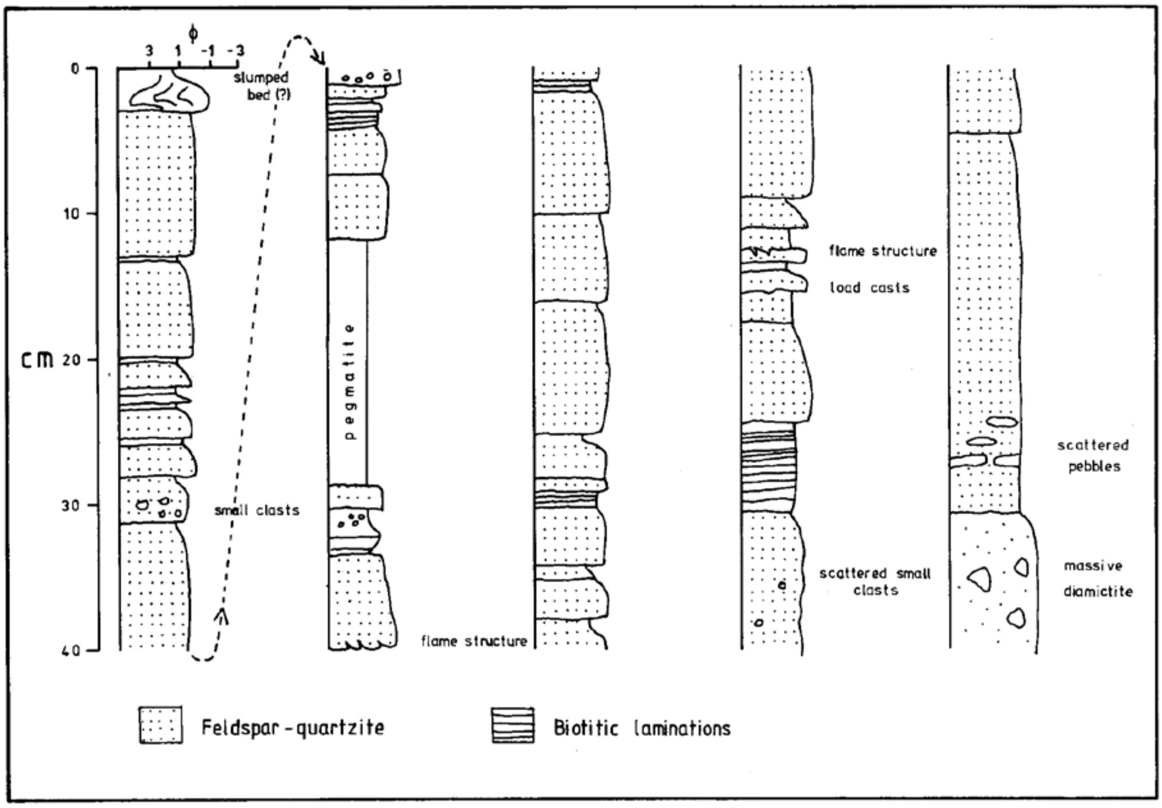

Fig. 7: Detailed, measured section of graded and ungraded feldspar quartzite beds on Namibplaas 93.

a concomitant increase in biotite. Rounded, oval-shaped cobbles, mainly of quartz and quartz-feldspar rock, are common at or near the bases of the beds, but may also occur near the middle or at the top. Thin beds may contain clasts which appear to be oversized relative to the thickness of the bed, e.g. a 5 cm diameter pebble in a 7 cm thick bed. Over the length of the measured section, the graded beds appear to thicken upwards.

Under the microscope, the graded feldspar quartzite comprises 70-75% quartz, 15-20% feldspar, mainly plagioclase, and 5-10% biotite, concentrated near the top of the beds, with accessory tourmaline, apatite and opaque minerals.

The quartz grains form a regular, polygonal mosaic with the feldspars, with well developed triple junctions, indicating recrystallisation. Biotite occurs among quartz and feldspar grains, and its orientation defines a pronounced schistosity. In one thin section, the grain size of the quartz decreases regularly from about 0,6 mm to 0,15 mm over the length of the section, with an accompanying increase in biotite. The generally altered plagioclase grains also decrease in size, and in abundance, over the entire length of the section. This indicates that recrystallisation has not completely destroyed the original sedimentary grain size grading.

b) The non-graded, pebbly feldspar quartzites generally occur in beds from 25 to 140 cm thick. The upper and lower contacts of individual beds are sharp, and thicknesses stay uniform over several tens of metres. Clasts, consisting mainly of granite, gneiss, quartz and quartz-feldspar rock, occur sporadically within the beds. Overall, there are no systematic changes in clast size with height; one 10 cm thick bed shows an inverse grading of small clasts (Fig. 6). Several beds have a concentration of granule size clasts near the base or near the centre of the bed. The clast size bears no relationship to the bed thickness. The tops of several beds show an apparent rapid fining of grain size and increase in biotite content over less than 1 cm; this fine-grained fraction serves to separate individual ungraded beds where they occur in a stacked sequence.

The non-graded beds are similar in composition to the graded beds, consisting mainly of quartz and plagioclase with minor biotite and accessory opaque minerals, apatite and tourmaline.

Grain size averages about 0,6 mm; the fine-grained fraction is less than 0,15 mm.

Fig. 8 illustrates some of the features described above. Of interest is the large granitic clast which appears to have been deposited between two graded units, and representing a component of neither. Note the wrapping of bedding and schistosity around the clast.

The graded feldspar quartzites on Namibplaas 93 occur between massive diamictites. The sequence is about 2 m thick and could be traced intermittently along strike for 40-50 m before it appears to thin out, or be sheared out. Although the outcrop studied is heavily intruded by pegmatites, a well preserved section was available for measurement.

The graded beds may be sub-divided into two types:

(a) Finely bedded (1 to 5 cm) rocks which grade rapidly both compositionally and in grain size into dark, very fine-grained, biotitic rock (Fig. 9), and

(b) more massive bedded (7 to 10 cm) rocks which show a slight gradation from bottom to top of the beds; these form stacked sequences.

The thin, graded beds have sharp upper and lower contacts. One 4 cm thick bed contains small quartz-feldspar clasts; another 8 cm bed displays flame structures at its base.

The thicker beds appear to grade rapidly in the bottom few centimetres, and then maintain a fairly constant grain size upwards. Very fine-grained, biotitic lay-

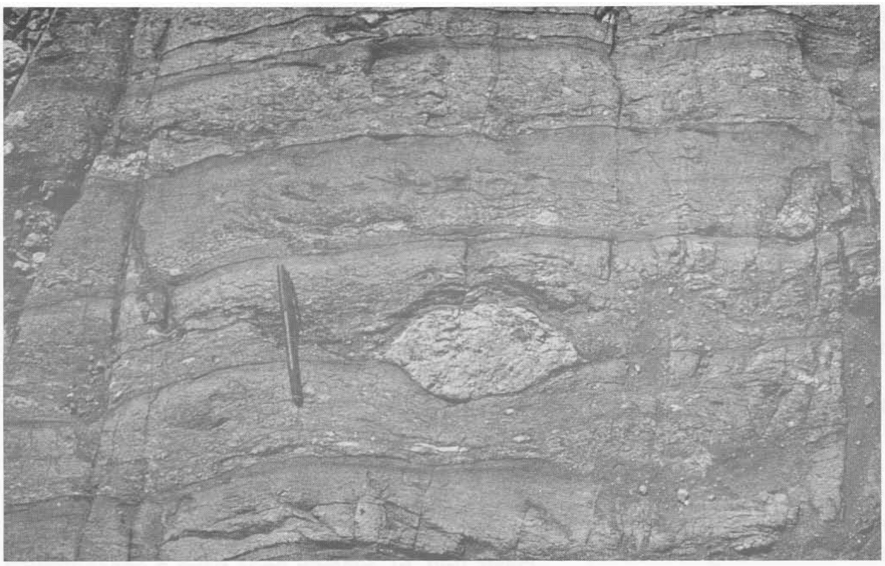

Fig. 8: Graded and ungraded pebbly feldspar quartzites in the Chuos Formation on Valencia 122. Note the gneiss cobble between the two 9 cm thick beds. Pen is 14 cm long.

Fig. 9: Bedded feldspar quartzites on Namibplaas 93, showing thin biotite laminations.

ers occur between beds. A 30 cm thick, ungraded bed occurs at the base of the measured section. It contains scattered, small (1 cm) pebbles near the base. The beds consist mainly of quartz, feldspar and biotite. Notable differences with the beds on Valencia 122 are the scarcity of clasts, and the more abundant biotite content.

The sequence is immediately overlain by a 1,5 m thick chaotic unit, which appears to be a slumped bed. Massive diamictite rests with a sharp contact on this bed.

3.3 Iron-rich Quartzitic Rocks

Iron-rich, usually magnetic, quartzitic rocks occur at several stratigraphic levels within the Chuos Formation (Fig. 10). The rocks form conspicuous, blocky outcrops because of their generally dark colour and their resistance to weathering. They form sharp contacts with the enclosing diamictites, and are generally of limited lateral extent. Several of the "beds" are folded, with steeply plunging fold axes. They are laterally discontinuous, with one exception, and cannot be traced much over 100 m along strike. They pinch out abruptly into diamictite. The thickness of the units varies from 10 cm to about 2 m, forming apparently conformable beds (see later discussion) within the diamictite.

The characteristic rock type is a dark coloured, conspicuously to faintly laminated, magnetic, quartzitic rock, which may be termed a banded magnetite quartzite or a banded iron formation.

Microscopically, a typical example of the faintly laminated magnetite quartzite is seen to consist mainly of quartz and magnetite. Quartz-rich layers alternate with magnetite layers. The grains are less than 0,01 mm in size, and occur as thin (0,1 mm) laminations. The magnetite is present as sub-idioblastic, interlocking grains, with minor interstitial quartz. Xenoblastic quartz grains are slightly larger. Accessory grains of a greenish amphibole are less than 0,1 mm in size. The more thickly laminated rock consists of 1 to 10 mm thick alternations of iron oxide and quartz. Several samples are not magnetic, indicating the presence of hematite rather than magnetite. This may be due to surface alteration.

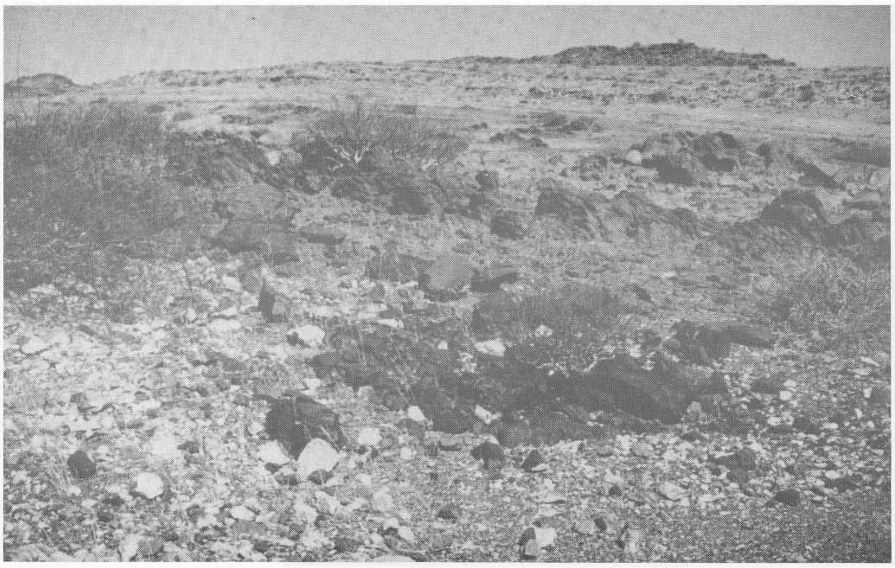

Fig. 10: Folded magnetite quartzite lens, probably a clast, in Chuos Formation diamictite. Namibplaas 93.

On Valencia 122, an iron-bearing quartzitic unit occurs near the top of the Chuos Formation, several metres below the contact with non-pebbly schists of the Karibib Formation (Fig. 4). The bed overlies diamictite, and passes upwards into pebbly, quartzitic rocks. A section through the beds is given in Fig. 11 and is illustrated in Fig. 12. Banded magnetite quartzites are found, as well as rusty-coloured feldspar quartzite and quartzitic schist. A hand specimen shows small blebs of sulphide in a highly recrystallised quartz matrix. Minor amounts of iron oxide are found. Beds of graded and non-graded, pebbly quartzite are interbedded with the iron-bearing rocks. Microscopically, the rock consists of quartz (70-75%) as recrystallised grains, from 0,1 to 0,05 mm in size, showing triple junctions, biotite as irregular laths generally less than 0,05 mm in length along grain boundaries (20%) and iron oxides forming about 5% of the rock. Sillimanite is present as accessory laths intergrown with biotite. No grading is evident in the F thin sections studied. A lonestone occurs at the top of the graded unit. It has penetrated the upper part of the bed, and disrupts the overlying, thinly laminated layers in the manner of a "splash" (Fig. 13). The bed is continuous along strike for over 800 m, and is easily followed because of its colourful, gossanous, weathering, in contrast to the grey to dark grey schists which overlie it.

3.4 Carbonates

Carbonates occur in two forms in the Chuos Formation: as thin, lenticular beds and as clasts.

At several stratigraphic levels, 0,5 to 2,0 m thick carbonate-rich beds are developed. These beds are in sharp contact with diamictite, and are continuous for 100 to 200 m along strike. The thickness of the beds varies,

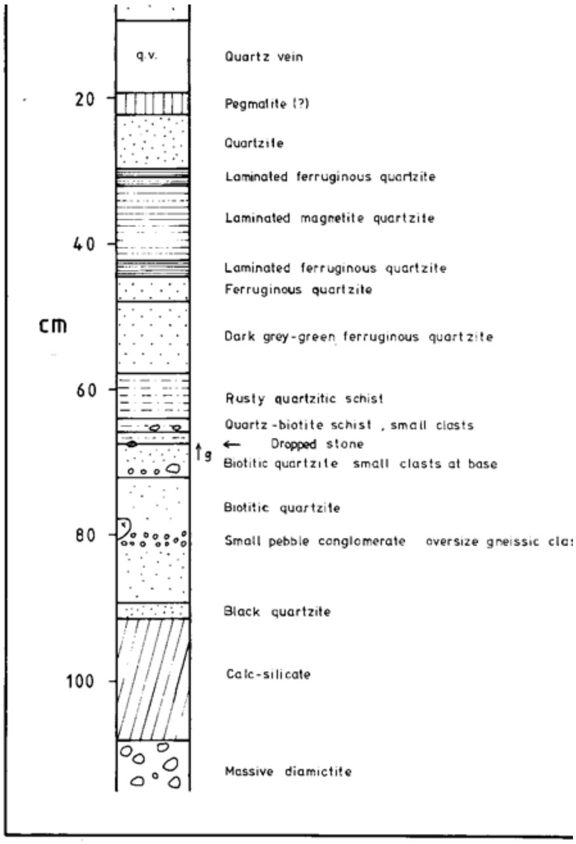

Fig. 11: Detailed sections of the magnetite quar beds near the top of the Chuos Formatio Valencia 122.

Fig. 12: Magnetite quartzite bed on Valencia 122. Hammer head at base of black magnetite quartzite bed, handle on graded feldspar quartzite bed.

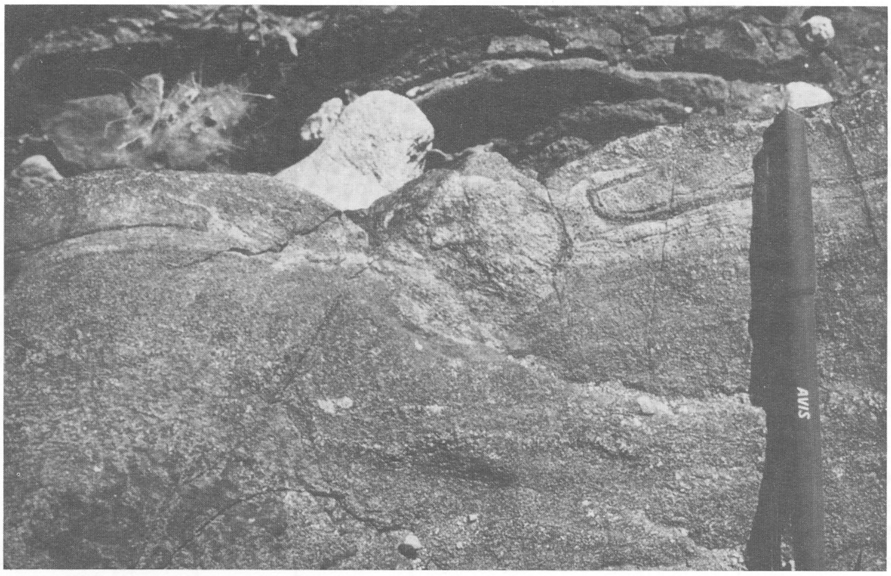

Fig. 13: Lonestone (considered a drop-stone), in feldspar quartzite of the Chuos Formation on Valencia 122. Note the disruption of the thin, light and dark layers, producing a splash effect.

and they pinch out abruptly against diamictite. No gradations are developed between carbonate-rich beds and diamictite. Two samples of carbonate examined indicate that they are mainly calcic marbles. The twinned calcite grains form a recrystallised mosaic, and vary from 0,1 to 0,5 mm in size. One sample shows an elongation of grains, probably parallel to the dominant schistosity. Thinly-bedded calc-silicate-pelite-marble transition units are uncommonly developed above or below the carbonate; they are clast-free and are in abrupt contact with diamictite.

Carbonate clasts are extremely rare. One clast 30 em in diameter was observed; another 10 em clast has almost completely altered to a greenish-white calc-silicate assemblage, with only a 1 cm core of marble remaining. The larger clast is also rimmed by calc-silicate.

3.5 Pink Feldspathic Quartzites

Pink quartzite "beds" occur at different levels within the Chuos Formation. They are from 1 to 30 m thick and lenticular in shape, form blocky outcrops and pinch out abruptly against diamictite within 120 m or less along strike. The quartzite is generally very fine- to medium-grained, and consists of quartz, pink K-feldspar and minor mica. Primary sedimentary structures are poorly preserved; cross-bedding and bedding planes are rare and are outlined by heavy mineral layers. The "beds" are parallel to the schistosity in the diamictite, and form sharp contacts. Similar rocks occur as metre-size or smaller clasts within the diamictite; these appear to be more abundant immediately above or below the "bedded" units, but are also found throughout the formation. The rock strongly resembles the feldspathic quartzites of the Etusis Formation.

4. VERTICAL AND LATERAL VARIATIONS WITHIN THE CHUOS FORMATION

Fig. 3 illustrates some of the variation within the Chuos Formation along the main study strip. The positions of the sections are shown in Fig. 1, and the top of the Chuos Formation has been taken as the datum on which the sections have been hung.

4.1 Vertical Variations

In any traverse across the Chuos Formation, the diamictite appears to be remarkably uniform, with no systematic or regular change in clast density, size and sorting, and type, nor in matrix type. The clast density varies erratically and gradually from scattered pebbles forming less than 1% of the rock volume, to 40-50%; no discernible breaks between zones with high and low clast population were observed. The size and sorting of clasts bears no relationship to height in the stratigraphy. Poorly sorted clasts from pebble to boulder size occur throughout. The clast population is polymictic, even in zones where there is a buildup of clasts of a particular type.

4.2 Lateral Variations

The Chuos Formation diamictites appear not to vary in nature throughout the length of the main study area. The formation thins out north-east of section 14 (Fig. 3) against Nosib Group quartzites. The clast population density is still irregular where the diamictite thins out, although there is an apparent lack of large clasts (over 30 cm in long dimension). The clasts remain poorly sorted and polymictic.

Magnetite quartzites were encountered in nine of the sections. Six of the occurrences are restricted to the lowermost half of the Chuos Formation, regardless of the thickness developed, two near the middle and one near the top of the sequence. The lower magnetite quartzites are not continuous from section to section; sections 9 and 10, about 500 m apart, have magnetite quartzites at the same level measured from the top of the Chuos Formation. Although the magnetite quartzites in each section are discontinuous and could not be traced laterally for much over 100 m, they appear to occur at the same stratigraphic level. A thick sequence of diamictite appears to have been developed below the magnetite quartzite horizon along section 9, in contrast to that developed in section 10. The magnetite quartzite bed near the top of the Chuos Formation on Valencia 122 (section 5) is a minimum of 800 m long. It was not found in adjacent sections; however, a black, ferruginous quartzitic bed which occurs near the top of the Chuos Formation in section 2, 10 km to the south-west of section 5, could be a correlate.

Pink quartzite lenses or "beds" occur near the top of the Chuos Formation in five sections, especially between sections 5 and 9. They are also developed at other levels in the sequence. Although difficult to prove, the discontinuous lenses near the top of the formation could be developed at the same stratigraphic level, and hence indicate coherent, widespread lenses within the diamictite.

Carbonate-rich layers were found in five of the traverses. They are traceable for less than 200 m

along strike before they pinch out abruptly against diamictite. The carbonate layers occur at different stratigraphic levels (Fig. 3), and cannot be correlated. The occurrence of carbonate is uncommon, and it is not possible to determine their distribution.

The bedded, sometimes pebbly, feldspar quartzites were found in two sections, 5 and 14, about 13 km apart. They are localised and do not appear to correlate. The beds on Valencia are continuous for about 800 m before they become distinct. The thinner beds on Namibplaas 93 were traced for about 40-50 m.

4.3 Lateral Variations Outside The Main Study Area

Nine traverses undertaken in the more highly deformed area to the north-west of the main study area show that the Chuos Formation is still developed, although generally difficult to recognise. Thicknesses were difficult to measure because of numerous pegmatitic intrusions.

The Chuos Formation diamictite in the area is generally highly sheared, and is normally developed between two carbonate-rich zones. The formation appears to thin gradually from south-east to north-west (tectonic effects taken into account), from 250-300 m along the main study area, to less than 100 m about 3 km to the north-west and gradually to less than 10 m about 8 km further north-westwards. The clasts in the diamictite are apparently smaller than in the south-east, with clasts greater than 30 cm in size rare. The diamictite is polymictic, even where it has thinned out. Magnetite quartzites, pink quartzites and carbonates were not found in the diamictite.

5. DISCUSSION

5.1 Previous Theories Regarding the Origin of the Chuos Formation

5.1.1 Glacial

A number of authors have advocated a glacial origin for the Chuos diamictite deposits because of:

(i) the massive, unbedded nature of the diamictite and its resemblance to the Palaeozoic Dwyka tillite (Gevers, 1931);
(ii) the sub-angular to sub-rounded clasts, some of which appear to be facetted (Smith, 1965);
(iii) the variety of intrabasinal and extrabasinal clasts;
(iv) the extremely poor size sorting of the clasts;
(v) the local occurrence of thinly bedded metasediments, akin to varves, near the top of the Chuos Formation (Gevers, 1931);
(vi) the widespread distribution of the diamictite in the same, or similar, stratigraphic position throughout the Damara Orogen (Miller, 1983);
(vii) the occurrence of a large granitic clast, interpreted as a dropstone, in distal turbidites of the Brak River Formation, the equivalent of the Chuos Formation in the area west of the Brandberg (Miller *et al.*, 1983).

5.1.2 Mass flow deposits

A non-glacial origin of the Chuos Formation has been proposed by Martin *et al.* (1985) and Porada (1983). The diamictites or "mixtites" are interpreted by the authors as mass flow and slurry flow deposits based in particular on the following observations (Porada, 1983):

(i) local derivation of clasts;
(ii) facies transitions between the mixtite matrix and underlying and overlying rocks;
(iii) locally, relation of mixtite deposits to synsedimentary faults;
(iv) in the "pebbly schist" type of mixtite in the Southern Zone, the bedding is well preserved.
(v) in "pebbly schist" the repeated occurrence of locally graded interbeds of quartzites, which are interpreted as turbidites.

In addition, Miller and Hoffmann (1981) list the following features as being incompatible with a glacial origin for the Chuos Formation:

(i) the local occurrence of "mixtites" in all stratigraphic units from the Etusis to the Karibib Formations;
(ii) the occurrence of warm water deposits (carbonates) above, below and within the Chuos Forma-

tion;

(iii) at least three short periods during which "mixtite" deposition ceased and chemical deposition took place instead (iron formation, dolomite).

5.2 Features Supporting a Mass-Flow Origin

5.2.1 Summary of types of mass-flow deposits

Lowe (1979) discussed the different types of sedimentary gravity flows (mass-flows) and the problems associated with their nomenclature, classification and applicability to natural systems. He proposes a classification modified after Middleton and Hampton (1976) based on the rheology of the flow and the dominant coarse particle support mechanism. Flows are classified first by their rheology, with fluidal flows and debris flows distinguished by their fluid and plastic behaviour, respectively. These two basic flow types are further subdivided into five types on the basis of the dominant coarse-particle support mechanism: turbidity currents (turbulence), fluidised flows (fluidisation), liquified flows (escaping pore fluid), grain flows (dispersive pressure of grain collisions) and mudflows (matrix cohesiveness).

Middleton and Hampton (1976) and Walker (1979) discussed the sedimentological characteristics of the different types of sediment gravity flows and their depositional settings. A slope-basin configuration is necessary for sediment transport and deposition, the angle of the slope being the main control on the type of flow generated.

(a) Bedded feldspar quartzites

The bedded, pebbly and non-pebbly feldspar quartzites in the Chuos Formation on the farms Valencia 122 and Namibplaas 93 are interpreted to be sediment gravity flow deposits of two types:

(i) the graded beds have the characteristics of the A-division in a Bouma turbidite sequence (Middleton and Hampton, 1976; Walker, 1979); the fine-grained, more biotite-rich top could represent the E-division. Hence the beds could have been deposited by turbidity currents. Walker (1979) describes AEAE sequences as proximal deposits. The presence of lonestone clasts, however, is unusual in a turbidite sequence;

(ii) the non-graded, pebbly feldspar quartzite resemble debris flow deposits. The lack of a substantial pelitic component indicates that they may be grain flow deposits. In such deposits, the coarse particles and clasts are supported by dispersive pressure caused by grain-to-grain interaction during flow. The thin, inversely graded bed (near 80 cm) in Fig. 6 was probably deposited by grain flow. The slopes required for grain flows are greater than that for turbidity currents, (Lowe, 1979).

The interbedding of proximal turbidites and debris flows suggests sediments gravity flow activity in an area close to the source of the sediment.

(b) Large blocks of quartzite, carbonate and iron formation

The large, discontinuous pink quartzites and some of the iron formations and carbonates that occur within the diamictite are interpreted to be blocks which have been externally derived and not formed *in situ*. The transport of such large blocks could be explained by glacial rafting or by sediment gravity flow. Hiscott and Middleton (1979) discuss the mechanics of transport of large slabs of rock in sandy turbidite beds. They propose a mechanism whereby blocks are supported partly by turbulent flow and by debris (matrix) strength, otherwise unrealistically steep depositional slopes are required for the movement of the blocks.

In the Chuos Formation, the presence of large slabs within the diamictites might be explained by processes similar to those discussed by Hiscott and Middleton (1979). The support mechanism, in this case, however, would be that of matrix strength only, because of the high mud content of the diamictites. The flows would have been, firstly, viscous, which would account for support of the slabs and, secondly, laminar, which would tend not to disrupt these long, narrow slabs.

The origin of the large slabs implies that the diamictites can be, in part, explained as mudflow deposits, and not as tillites. Such an interpretation would not be inconsistent with the characteristics of the diamictite, as discussed by Schermerhorn (1974).

(c) Diamictites as mudflow deposits

Diamictites are unsorted rocks which contain very few features which can be used to determine their origin (Schermerhorn, 1974). The rocks have to be viewed in the context of their depositional environments before interpretation can be made.

On Valencia 122, turbidites and debris flow deposits pass conformably upwards into massive diamictites, indicating that the latter are wholly subaqueous deposits and not tillites by definition (Hambrey and Harland, 1981). As discussed above, the diamictites are most likely mudflow deposits in a submarine environment. The overall remarkable lithologic uniformity of the diamictites implies stacking of mudflows one on top of the other, the sediment coming from heterogeneous sources.

(d) Iron formation as clasts and as an *in situ* bed

The iron formations are interpreted to be of two types with regard to their origin: (i) transported clasts and (ii) *in situ* sediments.

(i) The blocky, folded, discontinuous iron formations appear to be large slabs deposited within diamictites. The transport of large slabs is discussed in the preceding section. The origin of the clasts is uncertain: the source could be iron formations within the basement (Marlow, 1981); iron formations associated with amphibolites in the Southern Margin Zone of the Damara Orogen (Hoffmann, 1983); or reworking of iron formation deposited within the basin (see below).

(ii) The laterally continuous iron formation near the top of the Chuos Formation on Valencia 122 is interbedded with well-stratified, graded to non-graded quartzites. It appears to have formed between periods of diamictite deposition, during which other mass-flow processes were still operating. The type of iron formation, whether sedimentary or exhalative, has not yet been determined. However, the gossanous weathering of the diamictites and other beds near the iron formation indicates an abundance of sulphide, minor amounts of pyrite occur in one sample examined.

Sawyer (1981) found ferruginous quartzites interbedded with schist, pebbly schist, mica quartzite and metarudite in the Chuos Formation in the area south-east of Walvis Bay, but did not discuss their origin. From Sawyer's (*op. cit.*) description, the ferruginous quartzites appear to be beds which were deposited *in situ* in between periods of diamictite deposition. Hoffmann (pers comm., 1984) found a lonestone in an iron formation in the Chuos Formation near the confluence of the Khan River and the Khan Mine Gorge, about 10 km south-east of the study area, which he interprets as a dropstone.

5.3 Features Supporting a Glacial Origin

5.3.1 Lonestones

As illustrated in Fig. 6, the bedded sequence at the base of the Chuos Formation on Valencia 122 carries clasts which appear to be oversized relative to the thickness of the beds in which they occur. The large clast in Fig. 8 rests on the underlying bed, and the overlying bed appears to drape over it. This implies that the clast was dropped separately, and not transported along with the underlying and overlying beds. The outsize clasts in other beds can be explained as lonestones dropped onto a previously deposited turbidite or grain flow bed, which were subsequently reworked by the next sediment gravity flow event. The lonestones were incorporated into the sediment gravity flow and transported along, probably for a short distance before the flow energy dissipated and deposition occurred. The striking lonestone shown in Fig. 13 is interpreted to be a dropstone. It has disrupted a thinly laminated bed, producing a 'splash' effect, before settling into the top of the underlying bed. Ice-rafting is strongly suggested, especially in the Precambrian when rafting by logs or other floating plant material can be excluded.

Studies on sedimentation off the Antarctic coast by Anderson *et al.* (1979) indicate that glacial activity is difficult to recognise in a glaciomarine environment, with only the presence of dropstones being diagnostic.

5.3.2 Abundance of Basement-derived Clasts

Basement-derived clasts, mainly granites and gneisses, occur throughout the Chuos Formation diamictites. These extrabasinal clasts occur together with inferred intraformational clasts (carbonate, quartzite), and are not confined to any particular area. A source for the mudflows which gave rise to the diamictites wholly within the basin is difficult to reconcile with the widespread occurrence of basement clasts. Syn-sedimentary faulting up slope could generate mudflows which, carrying intrabasinal material, move downslope and deposit sediment here or on the basin floor. This could account for the pink quartzite 'beds' and the carbonates.

5.4 The Problem: Glacially Derived or Slope Derived?

The Chuos Formation consists of a sequence of sediments inferred to be deposited mainly by sediment gravity (mass) flow processes in a submarine environment. The sources of the sediments are partly intrabasinal and partly extrabasinal; neither alone can account for the heterogeneous clast population of the whole formation, nor the localised aggregates of quartzitic and carbonate 'beds'.

In a relatively immature rift setting such as that proposed by Martin *et al.* (1985), many of the features of the Chuos diamictites could be accounted for, with the exception of the presence of widespread basement clasts and dropstones. The suggestion by Martin *et al.* (*op. cit.*) that the diamictites are redeposited mass flow deposits from delta fronts is difficult to follow as with the exception of fan deltas, deltaic sediments are generally very fine-grained with no development of rudites (Miall, 1979). In addition, deltaic environments have, as yet, not been recognised in the Damara geosyncline.

Resedimentation of glacially deposited sediments on a basinal shelf or slope by mass-flow processes can also account for the features observed in the Chuos Formation. Contemporaneous glacial activity can explain the dropstones as ice-rafted debris.

Glacial activity would create a heterogenous, unsorted sediment termed till or moraine (Hambrey and Harland, 1981). Deposition of such sediment in a submarine environment could provide an abundant supply for later reworking. The bulk of the Chuos diamictites is interpreted to have formed from the reworking downslope of such glacially-derived material. This is similar to the process proposed by Hoffmann (1983).

In addition to the studies by Anderson *et al.* (1979) on sedimentation off the Antarctic coast, recent work on Pleistocene (McCabe *et al.*, 1984) and Palaeozoic Dwyka glaciomarine deposits (Visser, 1984; Von Brunn and Gravenor, 1983) as well as on the upper Precambrian diamictites of the Port Askaig Formation in Scotland (Eyles and Eyles, 1983), have greatly improved our understanding of glaciomarine sedimentation. The studies have shown that sediment gravity (mass) flows are the most important depositional mechanisms in a subaqueous glacial environment. Direct deposition of ice-rafted material forms a minor percentage of the preserved sediment pile, whereas downslope reworking of material deposited at the head of the glacier as mass flows accounts for most of the deposit. Slumping and generation of mass flows could be initiated by instability of the sediment pile due to rapid deposition, or by earthquakes due to syn-sedimentary tectonism.

A more direct mechanism for the deposition of diamictons below floating ice shelves can also be considered. A floating ice shelf deposits debris in the subaqueous outwash below it, seaward from the ice-grounding line, onto the sea floor. Rapid deposition of debris could cause slumping and redeposition by sediment gravity flow processes (Edwards, 1978). Slumps can also be initiated by a high density underflow caused by the rapid mixing of fresh water and sea water. Such a mechanism could occur on a passive basin margin, and a deep water setting is not necessary for the sediment gravity flow deposits. Tidewater glaciers off Spitsbergen are depositing sediment as glaciomarine diamictons (Edwards, *op. cit.*), which provides a modern analog for ancient glaciomarine sedementation.

A glaciomarine setting would best explain most of the features of the Chuos Formation in the Central Zone. A similar setting is proposed for the Chuos Formation in the Southern Margin Zone of the Damara Orogen by Hoffmann (1983), who envisaged deposition of the Chuos Formation in a deep water, passive continental margin environment. Glaciogenic debris laid down initially as shallow marine till on the continental shelf and redeposited by mass-flow movements into the adjacent slope region would account for the Chuos diamictites (Hoffmann, *op. cit.*). Within the study area, the occurrence of abundant intrabasinal clasts of the same type at certain stratigraphic levels in the Chuos sequence suggests synsedimentary tectonism within the basin. This tectonic activity, probably faulting, would give rise to locally derived mass-flow deposits. These could incorporate previously deposited, unconsolidated sediment and thus account for the polymictic character of the diamictite as a whole.

6. TECTONIC RECONSTRUCTION AND PALAEOTOPOGRAPHY

The distribution and thickness variations of the Chuos Formation in the study area allow a preliminary reconstruction of the palaeotopography of the depositional basin in the study area during the deposition of the Chuos Formation, and the influence of tectonism on sedimentation.

6.1 Domal structures - Synsedimentary or Tectonic?

The study area is underlain by four major structural domains (Fig. 1):

(i) A large, NE trending synclinal structure, which is flanked to the north-west by
(ii) an E-W elongate, 11,5 x 6,5 km oval dome on Namibfontein 91, and
(iii) an elongated, NE trending domal structure, 26-28 km in length by 5-7 km in width, situated west of Namibplaas 93.
(iv) The south-east limb of the syncline forms part of the large Bergrus anticlinal structure, which plunges to the south-west, and is cored by basement rocks.

The domal structure on Namibfontein 91 is cored by remobilised basement and lower Damara metasedimentary rocks. The margins of the dome are highly sheared, and the structure is thought to be an interference fold, possibly modified by diapirism (Sawyer, 1981).

The elongate domal structure west and south-west of Namibplaas 93 is complex. The margins are highly sheared, basement and lower Damara rocks in the core appear to be remobilised and the structure is thought to be a 02-03 interference fold. Highly sheared Chuos diamictites were recognised on both sides of the structure.

The structures discussed above appear to be post-sedimentation in origin. However, the overall thinning of the Chuos Formation in a northerly to north-westerly direction indicates a basin margin in that direction.

The entire Swakop Group thins out on the eastern side of Namibfontein 91, along the north-west side of the Bergrus anticlinal structure (Fig. 1). The Chuos Formation overlaps a thinned Rössing Formation, and both formations pinch out against Nosib Group meta-quartzites. The Nosib Group, which forms extensive outcrops on either flank of the Bergrus anticlinal structure, appears to thin out against basement rock about 12-14 km to the north-east of Namibfontein 91 (off the map in Fig. 1). The Karibib Formation overlaps the underlying Chuos Formation and continues uninterrupted to the north-east, maintaining a fairly uniform thickness.

The thinning of the pre-Karibib formations towards the north-east and the overlap relationships strongly suggest that the area now occupied by the Abbabis Inlier was a basement high appears to have had a profound effect on the deposition of the Etusis, Khan, Rössing and Chuos Formations. Periodic re-activation and uplift of the basement would account for the pre-Karibib disconformities in the Damaran stratigraphic succession. The erosion of the Rössing Formation prior to subaqueous Chuos deposition may have been caused by vertical tectonic movements which could be produced during block faulting. The paraconformable contact between the Rössing and Chuos Formations indicates that the lower sequence was not deformed prior to Chuos deposition.

6.2 Evidence for Synsedimentary Faulting

The confinement of the large intrabasinal clasts of quartzitic and carbonate composition in the Chuos Formation to the south-east limb of the synclinal structure indicates a south-eastern source ared of the clasts. This would suggest that the south-eastern part of the area was a high during Chuos deposition; this area appears to have been fault-bounded, the faults being periodically re-activated to generate intrabasinal slumps and cause mass-flows.

The sawtooth pattern outlined by thickness changes in the Chuos Formation in Fig. 3 between sections 6 and 11 could be due to post-sedimentary folding, infill of depressions in the Rössing Formation floor or to synsedimentary faulting during Chuos deposition. The former is unlikely since the Rössing and Karibib Formation do not show sympathetic thickness changes. The thickness of the Chuos Formation appears not to be directly related to the amount of Rössing Formation rocks removed during pre-Chuos erosion, as shown in Fig. 2. Hence synsedimentary faulting appears to have controlled Chuos deposition to a large extent, possibly forming local depressions where thicker piles of sediment accumulated.

6.3 Palaeotopography

The above discussion allows a preliminary reconstruction of the Chuos depository in the area. The thick development of the Chuos Formation adjacent to a high and its thinning to the north and north-west suggests a half-graben configuration for the Chuos depository. The south-eastern margin of the basin would have been fault-bounded and periodically active, giving rise to occasional intrabasinal slumps. The north-western margin would have been passive, and gently sloping to the south-east, as shown in Fig. 14.

7. PRELIMINARY CONCLUSIONS

The Chuos Formation in the study area comprises a variety of rock types, diamictite forming the bulk of the sequence. Magnetite quartzite and bedded, occasionally pebbly, graded or ungraded feldspar-quartzite beds occur locally. The diamictites contain large clasts of pink quartzite, carbonate and magnetite quartzite, which appear to form “beds” , or clasts at different stratigraphic levels.

The thickest development of the Chuos Formation appears to be on the south-eastern side of the study area, adjacent to a palaeotopographic high. The formation thins to the north and north-west, forming a wedge-shaped, blanket-like unit.

The diamictite and bedded feldspar quartzites are interpreted to have been deposited by mass-flow processes in a subaqueous environment. Mudflows were generated on the slopes of the basin and deposited the sediment which later formed the diamictite. Turbidity currents and grain flows deposited the graded feldspar quartzite beds, and the non-graded, pebbly, feldspar quartzite beds respectively. The depositional basin may have been a half-graben, with a relatively steep-sided, fault-bounded south-eastern side, and a more gentle, south-east sloping, north-west side.

Synsedimentary faulting appears to be responsible for the increased accumulation of diamictite in the south-east. The faulting could have triggered slumping on the

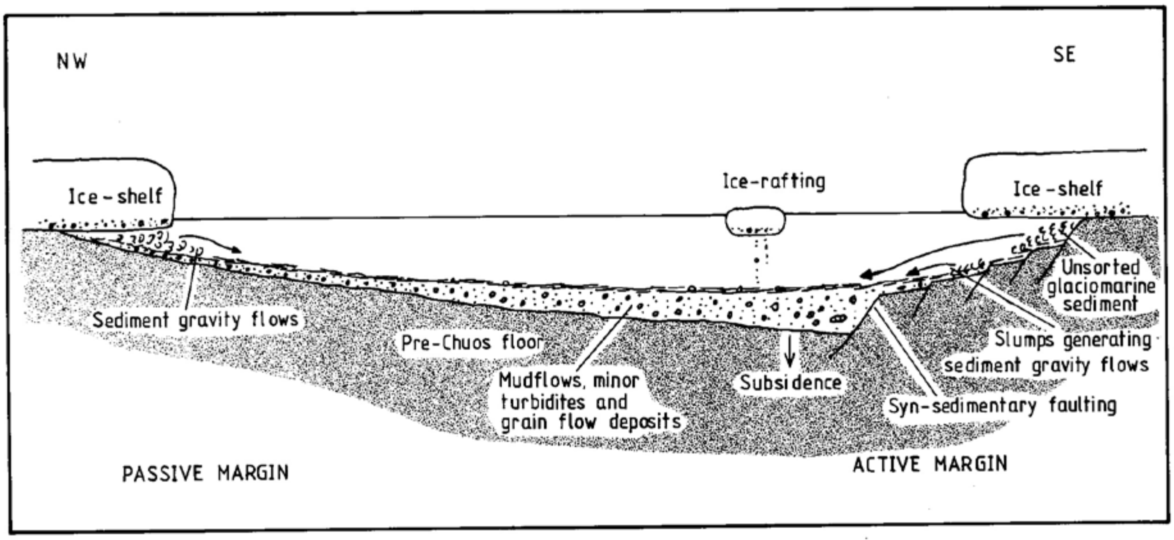

Fig. 14: Schematic reconstruction of the palaeotopography and depositional environment during Chuos times. Vertical scale exaggerated. Section approximately from Vergenoeg 91 to the north-western margin of the Abbabis Inlier.

slope and given rise to the mass flows.

The widespread occurrence of extrabasinal clasts in the diamictite and the presence of dropstones implies a glacial influence on Chuos deposition.

A model invoking re-deposition of glacially-derived material down the slope of a tectonically active basin margin could account for most of the features observed in the Chuos Formation.

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