With complements
to Prof. W. D. Gill.

Ov. Efeotor.
3/12/74.

To

My late senior brother, Odefieghreaje Efeotor.

SEDIMENTOLOGICAL STUDIES OF THE BIMA SANDSTONES (MURI N.E. AND FUTUK S.E., SHEETS 193 AND 172) NORTHEASTERN NIGERIA.

by

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ABSTRACT

A geological map is presented for the area of study and three formations: the Bima Sandstones, the Yolde Formation, and the Pindiga Formation are recognised, and are based on aerial photographic study, major lithologic changes and tectonic style.

Heavy mineral analyses of the rock samples collected from the Bima Sandstones indicates a suite of heavy minerals characteristic of granitic source rocks, and the presence of certain authigenic species. All heavy mineral species tend to concentrate in the dominant size grade (105μ) of the grain size distribution of the heavy residue. Zircon elongation studies clearly show a similar trend of crystal growth and, therefore, a similarity in the physiochemical conditions under which zircon crystallized, which is an additional indication of a common source rock for the sediments.

Petrographic analysis of the rocks shows a uniform mineralogical assemblage dominated by quartz and feldspars. These minerals do not have any environmental significance. The sandstones are arkoses with secondarily derived clay minerals produced as a result of the degradation of the feldspars during diagenesis.

Sedimentary structures present indicate a preponderance of the planar type of cross-stratification. Paleocurrent analysis indicates that the sandstones were derived chiefly from the northeast, with contribution from the north and northwest. Deformed cross-stratification and other quicksand structures present in the sandstones are described and a conclusion is reached about their probable mode of origin in the light of available experimental data on similar structures described from other parts of the world.

A stratigraphic description of the Bima Sandstones is made for the whole area studied, and it indicates a presistent occurrence of medium- to coarse-grained planar cross-bedded sandstones from the bottom to the top of the exposed part of the formation. These sandstones are equated with the uppermost unit of the three subdivisions of the Bima Sandstones already described from the Standard Section around Dadiya in the Lamurde anticline. Their overall characteristics indicate a braided stream environment of deposition.

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CHAPTER 1

INTRODUCTION

1.1 AIM OF THE PRESENT STUDY

There has been much geological work carried out by many people in Nigeria. Workers in different parts of the country have been motivated by various interests to undertake their studies. The presence of various valuable minerals such as lead, zinc (galena, sphalerite), tin, graphite, marble, etc, promoted initial interest in the crystalline basement rocks and metamorphic rock units in different parts of the country. This in general has led to more work having been done on the crystalline and metamorphic geology, while the geology of the sedimentary sequences remains to be investigated in detail. Initial interest in the sedimentary geology of Nigeria was generated by the search for petroleum, and by the Geological survey of Nigeria. Unfortunately the bulk of information arising from work carried out by the oil companies is not available to the academic world.

At the present time, interest in the sedimentary geology of the country is confined to the southern part where there are well exposed sedimentary units and borehole data. The sedimentary rocks exposed in the northern portion of the country were first studied in detail along the Benue Valley trough by Jones (1962), and around Lokoja by Adeleye (1973). The Geological Survey of Nigeria has compiled geological maps for some of the areas covered by the sedimentary deposits, while in other areas there is no record of either topographic or geological maps. Furthermore, the detailed sedimentological and environmental analysis of these sediments are yet to be studied.

The present area of study was first appraised by Gubbins (1957) for the Mobil Oil Exploration Company of Nigeria. No information was published from his studies. Dessauvagie (1969) later compiled a photogeological map for the area with broad generalizations as to the types of rocks present and their supposed environment of deposition. No contour or recent topographic maps were available for use when field data collection was made for the present study.

The aims of the present study were, therefore, to provide a geological map for the area based on field mapping; to carry out sedimentological work to determine the probable provenence and the environment of deposition of the sediments; and, above all, to stimulate further academic interest in the sedimentary deposits of this portion of the Benue Valley Trough, and elsewhere in the country.

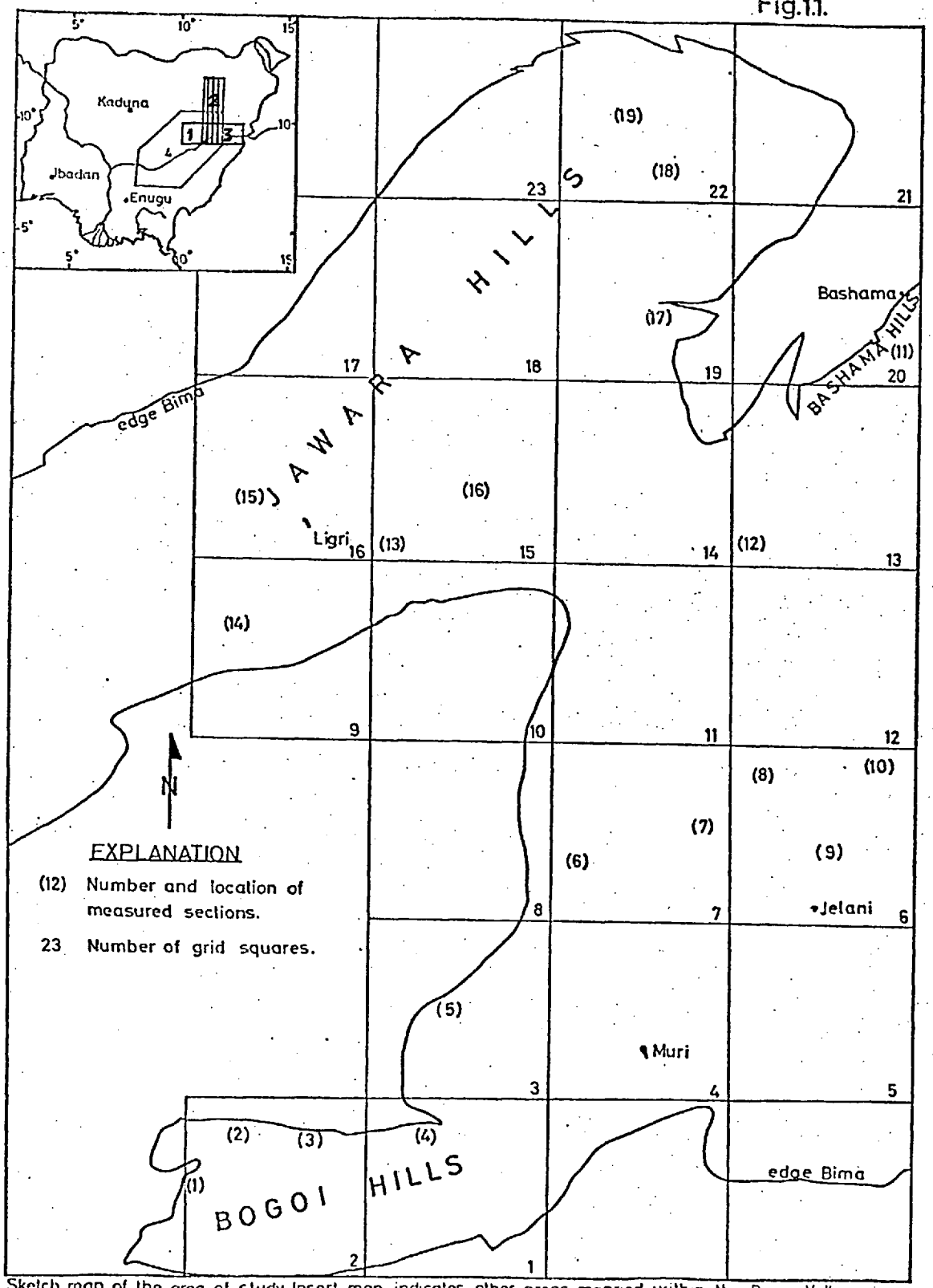
During the course of the present study reference was made to previous work in other areas, and to past and present experimental work being carried out in environmental studies of modern sediments. It is envisaged that by pooling together such a vast array of geological information it will be possible to give a more complete account of the sedimentological processes which may have been in operation at the time the Bima Sandstones were deposited.

1.2 LOCATION AND GEOLOGICAL SETTING

The area of study is situated in the central part of the Benue Valley in northeastern Nigeria (fig. 1.1). It lies between latitudes $9^{\circ}00'$ and $9^{\circ}40'$N, and longitudes $10^{\circ}37'$E and $11^{\circ}00'$E.

The area consists of comparatively flat-lying ground to the south along the present Benue River Valley with its alluvium-covered wide flood

Fig.11.



Sketch map of the area of study. Insert map indicates other areas mapped within the Benue Valley: Area of present study(1:100,000) and Dessauvagie)1969(1:25,000)(1), Carter et al,1963(1:250,000)(2), Jones,1962(3), Cratchley and Jones,1965(1:500,000)(4).

plains. Northward the land rises gently, firstly in undulating ground, and then culminates in the Bogoi Hills to the west, Materum Hills to the east, and the Jawara and Bashama Hills to the north and northeast respectively. There is a marked relationship between topography and the outcrop pattern of the rocks of the different formations present. Thus, the lowest parts of the area toward the south and southeast consist of the Benue Valley Alluvium. The Bima Formation outcrops form the highest ground and consist of a series of linear ridges and hills; whilst the Yolde and Pindiga Formations are restricted to areas of intermediate elevation (fig. 1).

1.3 PREVIOUS WORK

Falconer (1911) studied the general geology of the Muri area and designated the sandstones occurring there as the "Muri Sandstones". Bougue (1946) made some geological notes on the Zurak district. In his work he showed that the ore lode occurred in medium-grained arkosic sandstones with a few interbeds of siltstones and shales. Bougue concluded that the angular feldspar and partially rounded quartz grains "..... are cemented with clay which is mainly kaolinite with small amount of mica". Similar work was carried out by Mackay (1946) on two lead/zinc mines in the Plateau and Ogoja Provinces, and by McConnell (1948) on the geology of Silver/Lead deposits at Arufu and Akwana in Benue Province.

Gubbins (1957) wrote a geological report on the Muri area for the Mobil Oil Exploration Company of Nigeria, and concluded that on account of the occurrence of lead/zinc mineralization, proximity to volcanism, complexity of faulting, and thin sediment cover it was impossible to

envisage commercial accumulation of petroleum in the Muri area.

Carter, Barber, and Tait (1963) studied adjacent areas to and east of the area under investigation. Their work was mainly a general geology of the Crystalline Basement, Cretaceous, Tertiary and Quaternary sediments of these areas. They also attempted to correlate their area of study (fig. 1.1) with other areas in the Benue Valley trough. Cratchley and Jones (1965) made an interpretation of the geology and gravity anomalies of the Benue Valley which included the Muri area. On the basis of data accruing from their study the two authors concluded that the Benue Valley is a rift valley of Cretaceous age.

More recently, Dessauvagie (1969) made an attempt at using photo-geological interpretation to elucidate the stratigraphy and structural complexity of the Muri area. The area he studied (Futuk, sheet 46 of the Geological Survey of Nigeria) covers approximately 12,000km^2 and is situated in the central part of the Benue Valley in northeastern Nigeria (fig. 1.1). Dessauvagie examined a total of 900 photographs over a period of six months and spent four weeks in the field to obtain ground control. Due to the inaccessible nature of the terrane his traverses were restricted to certain areas.

Burke et al. (1970) made gravimetric and magnetic studies of the Benue Valley Trough from which they postulated a hypothesis that a spreading ridge underlay the Niger delta and that this extended a short distance up the Benue Valley trough during parts of the Cretaceous. They further contend that the Santonian folding episode which affected the sediments deposited in the trough was "..... as a result of misfit motion within the African plate due to faster spreading opposite the bulge of Africa than in the south Atlantic".

Work was also carried out by Jones (1962) in other areas in the Benue Valley (fig. 1.1). His analysis of the depositional and structural attributes of the Bima Formation forms the first true sedimentological work in the Benue Valley Trough in northeastern Nigeria. From stereographic plots of cross-bedding directions in the Yola and Lau areas he deduced a two-source origin for the sediments and postulated that evidence from his study areas show that the paleocurrents direction in the Muri area may be in a westerly/southwesterly direction (personal communication, 1970).

1.4 PRESENT WORK

Interest in the area covered by the present work was generated by Dessauvagie's (1969) photogeological analysis of the stratigraphy and structure of the Futuk sheet 46 of the Geological Survey of Nigeria, and the sedimentological work carried out by Jones (1962) in the Yola and Lau areas in northeastern Nigeria. The extent of the present work is determined by accessibility of the area and the distribution of well exposed and measurable units of the Bima Formation.

The work is purely of a reconnaissance nature consisting of field mapping of formation boundaries and tectonic structures to provide a geological map for the area, based on field data; paleocurrent analysis and the preparation of lithologic sections for the sedimentological interpretation of the gross basinal characteristics of this portion of the Benue Valley Trough during Bima times; and heavy mineral and petrographic analyses provide information on the provenance and the development of diagenetic changes on the sediments.

Field mapping was carried out in the dry seasons of 1972 and 1973.

The study of aerial photographs was carried out in the Geology Department of Ibadan University under the guidance of T.F.J. Dessauvagie, and a total of 120 photographs were examined over a period of two months. The present study was carried out on a scale of 1:100,000 using Dessauvagie's photogeological map as a guide. Contoured topographic maps were not available for use as base maps at the time of the field mapping for the present work. All measurements were made in the metric system.

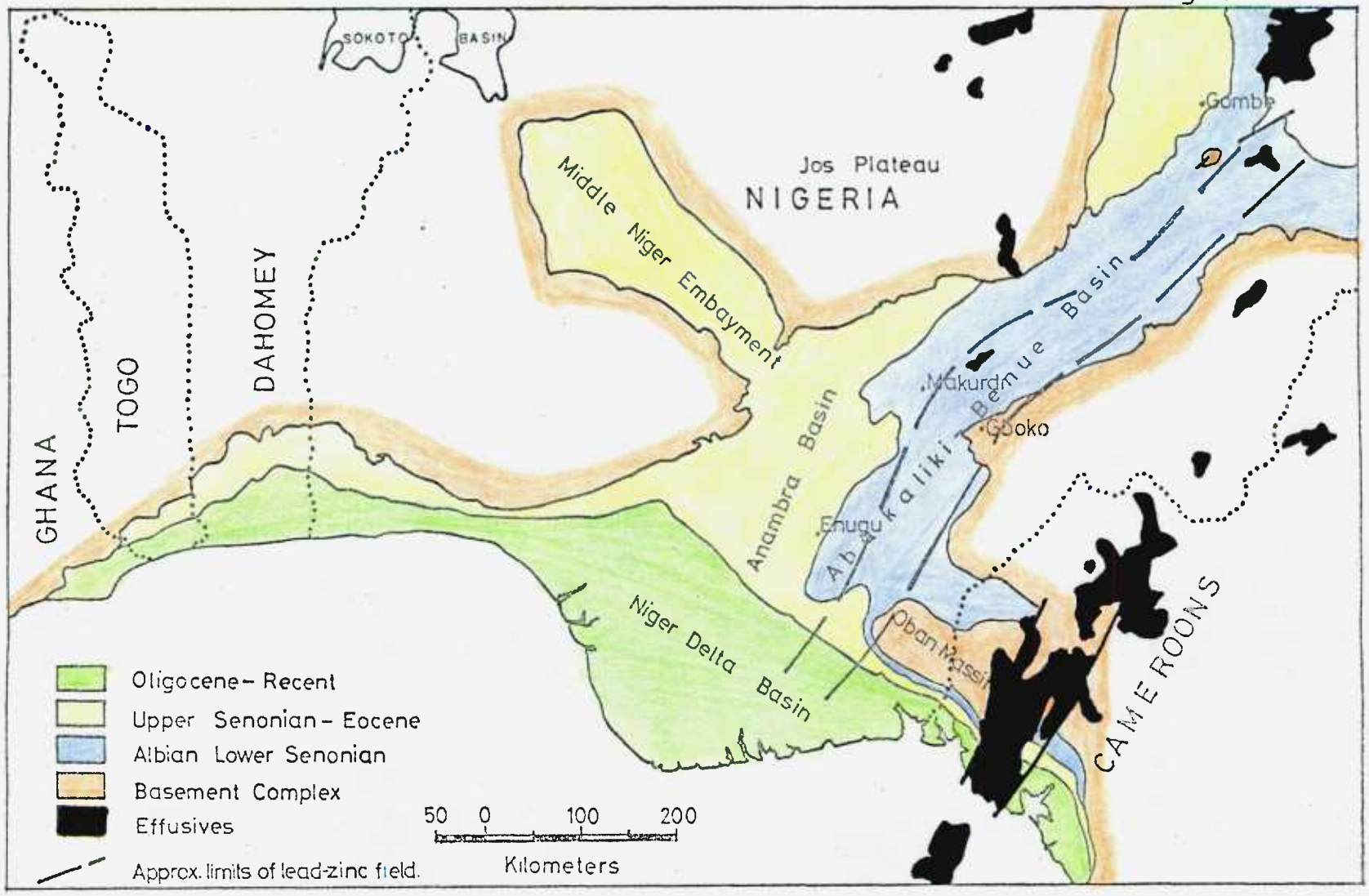
The method of data collection in the field was patterned along lines described by Selley (1968), which involves the production of a continuous field sketch of the sections recorded. Field data were plotted based on the conventional documentation of lithology. In addition, a vertical histogram of grain size was added to one side. The space formed by the latter also carried the sedimentary structures shown symbolically (fig. 3.1, 3.3, and 3.4).

1.5 BASINAL SETTING

The Benue Valley Trough is a linear sedimentary basin which is approximately 130 Kilometers wide, and is one of the three basins which form the Southern Nigerian Sedimentary Basin (fig. 1.2). The Benue Valley Basin (Abakaliki-Benue Trough, Murat, 1970) is filled with Cretaceous rocks of Albian to Maestrichtian age. The basin is limited on either side by granitic and gneissic masses of the crystalline basement complex, believed to be Palaeozoic in age.

The thickness of sediments is variable within the basin because of the irregular topography of the crystalline basement surface. The

Fig.1.2.



Geological sketch map of Southern Nigeria (after Murat, 1970).

sediments represent deposits of both rivers and the sea, with some deltaic deposits in the southern portion of the basin. The continental clastics are very thick in some areas within the basin, and are more abundantly developed in the northern and central portions of the basin.

Marine inundation of the area was initiated by Albian movements along northeast to southwest trending faults. These movements resulted in the formation of the rift-like Abakaliki-Benue Trough. This basin was limited to the northwest by the Benin-Benue Hinge Line (Murat, fig. 3, 1970). The regressive phase which followed the marine transgression resulted in the deposition of continental clastics in the north and central portions of the basin, and in deltaic sequences to the south. The total thickness of Cretaceous sediments that occupy the Abakaliki-Benue Trough represents a series of transgressive and regressive events which culminated in a Santonian movement along the faults established in Albian times. These renewed movements resulted in the uplifting and folding of the sediments deposited within the Abakaliki-Benue Sedimentary Basin.

1.6 REGIONAL STRUCTURAL REVIEW

The Benue Valley forms part of a series of linear sedimentary basins filled with strongly folded Cretaceous sediments. From the width and branching pattern of these basins it has been suggested by many authors that they may be buried Cretaceous rift valleys, and that subsequent erosion and/or later sediment infilling of the troughs have obliterated the boundary faults. King (1950) contends that these boundary faults are to be seen in the north and south scarps of the present extent of the Cretaceous cover in the Benue basin. The absence of obvious normal

faults parallel to the Benue Valley led to doubts about the authenticity of the rift hypothesis. Lees and Farrington (1952) suggest a geosynclinal accumulation of the sediments in these basins. Bouguer anomaly work carried out by Cratchley and Jones (1965) led these two authors to conclude that the Benue trough is a rift valley of Cretaceous age.

The opening and shutting of the Benue Valley Cretaceous Rift System is believed to be an important event in the structural development of the south Atlantic and is of interest as an example of a consdensed orogenic cycle with short-lived episodes of ocean opening and closing (Burke, Dessauvagie and Whiteman, 1971). These three authors have also likened the structural relationships in the eastern parts of the Gulf of Guinea in the early Cretaceous to those prevailing in the southern end of the Red Sea. Accordingly, they interpret the magnetic anomaly pattern which show three positive and three negative anomalies parallelling the trend of the Lower Benue depression in the area of the oceanic crust as being consistent with the hypothesis that an RRR junction underlay the Niger delta in the Cretaceous.

Major folding episodes took place in the Benue depression during Santonian times, the intensities and extent of which have been grossly underestimated. Dips more than 60° have been recorded, and in the Upper Benue over 100 compressional anticlines and synclines, some more than 60 km. long, have been mapped. In the Benue valley the fold belt is more than 800km long and 100 - 150km wide and forms the only major fold belt lying within the African Continent formed since Paleozoic times.

Cratchley and Jones (1965), Burke et al, (1970) and Murat (1970) proposed a sequence of structural events in the Benue Valley during Cretaceous times.

The thin veneer of continental and marine deposits which lies south of the Benue river is less folded than the thicker sandy deposits occurring north of the river. These latter sediments are thrown into numerous anticlinal folds with northeast to southwest trends. In the present area of study these folds reach a maximum length of 60km and a width of 12km with dips of up to 50° and more in some cases. These folds include Dutse (Hausa word for hill or rock) Bogoi in the south, and Dutse Jawara and Bashama in the north and northeast respectively. Faulting is very common in the area of study and is responsible for the complicated nature of the Bogoi and Jawara fold belts. The faults are very obvious in aerial photographs but are obliterated by weathering and the ubiquity of boulders and rock screes in the field.

Cross-bedding structures usually regarded as indicators of fluviatile and/or deltaic environments are a persistent feature of the Bima Sandstones from the base to the top of the sequences. Few trough cross-stratifications were seen and recorded in the field.

Deformed cross-stratification structures are abundantly developed over the entire area covered by the present investigation. Their occurrence appears to be restricted to the upper part of the sequences, and their morphology ranges from minute "rippling" of the forest beds to contortions of different types.

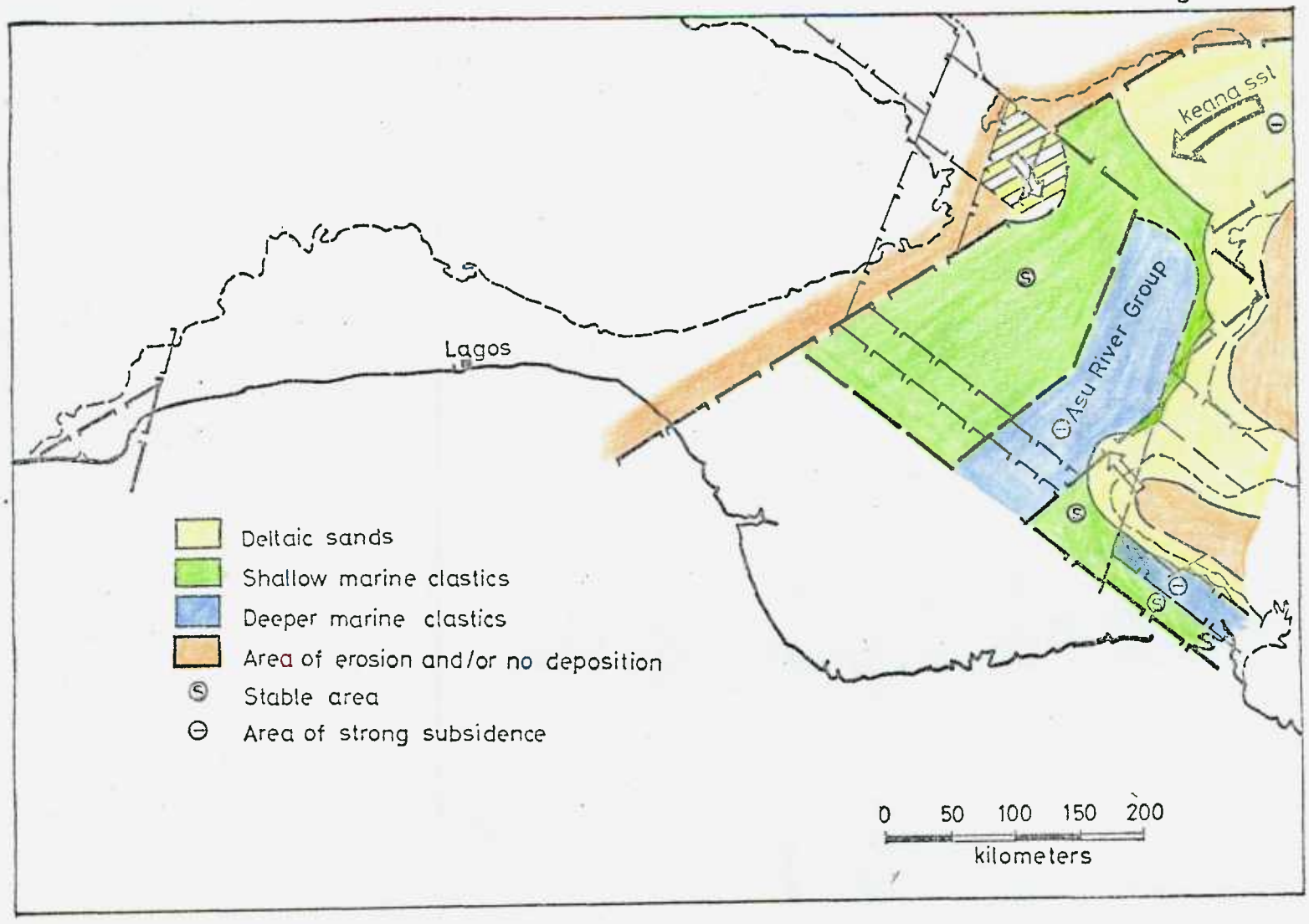
1.7 REGIONAL STRATIGRAPHIC REVIEW

The sedimentation history of the Abakaliki-Benue Sedimentary Basin is divided into three sedimentary cycles by Murat (1970) and is well documented by Cratchley and Jones (1965).

The first cycle (mid-Albian to end-Cenomanian) was transgressive and resulted in the deposition of the Arafu and Gboko Limestones (Reyment, 1965) along the southern edge of the basin. The regressive phase of this cycle is marked by extensive deltaic developments (fig. 1.3) which resulted in the deposition of a thick pile of continental clastics. In the Upper Benue and Lower Gongola area the Bima Sandstones (Carter et al, 1963) were deposited while in the Middle and Lower Benue the "Müri" (Falconer 1911) and Kaena Sandstones were deposited. In the present study the "Müri" Sandstones believed to belong to the same regressive phase (Cratchley and Jones, 1965) and the downdip equivalents of the Bima Sandstones (Murat 1970) are equated with the Upper Bima Sandstone unit of the Standard Section, and will be referred to in this work as the Bima Sandstones. These Sandstones are arkosic, medium- to coarse-grained, feldspathic, and medium-bedded with predominantly planar cross-bedding. The base of the Bima Sandstones rests directly on the crystalline Basement complex wherever their contact is observed in the Upper Benue/Lower Gongola area.

In the southwest portion of the basin sedimentation resulted in the deposition of a marine shale sequence (the Asu River Group). The shales are shallow marine deposits which rest unconformably on the crystalline basement, and consist of dark grey, frequently calcareous shales, subordinate sandstones and limestones. The sequence is strongly deformed.

Fig.1.3.



Palaeogeographical sketch map of the Upper Albian Cenomanian regression (after Murat, 1970).

The second cycle of sedimentation within the basin began in the lower Turonian with the deposition of the Eze-Aku Shales in the Abakaliki portion of the basin. This sequence is rich in pelagic microfauna and consists of flaggy calcareous and bluish non-calcareous shales and siltstones, thin sandy or shelly limestones, and calcareous sandstones. The Lower Turonian transgression which affected areas around Lau in the Lower Gongola areas also extended as far as the present area of study. This event resulted in the deposition of a suite of sediments believed to be equivalent to the transitional Yolde Formation and the overlying marine Pindiga Formation. Around Makurdi the Makurdi Sandstones were deposited, and the deltaic Konshisha River Group were formed along the eastern edge of the Abakaliki trough.

The third cycle (Upper Turonian to Lower Santonian) is marked by the deposition of sedimentary units which were extensively eroded during the formation of the Abakaliki - Benue folded belt (Murat 1970). Limestones belonging to this cycle of sedimentation have been found in Shell-BP's wells around Calabar and in outcrops at Nkalagu near Enugu. Near Makurdi equivalent ammonite-bearing limestones crop out at Wadatta. The Awgu shales which are equivalents of the Eze-Aku shales were deposited in the Abakaliki trough and in the Lower Gongola as the Sekule Formation and the Numanha Shale. The Agbani Sandstone which crops out east of Enugu is believed (Murat, 1971) to be part of a deltaic unit which has been observed in several Shell-BP wells in the Calabar-Enugu-Abakaliki area. (fig. 1.4)

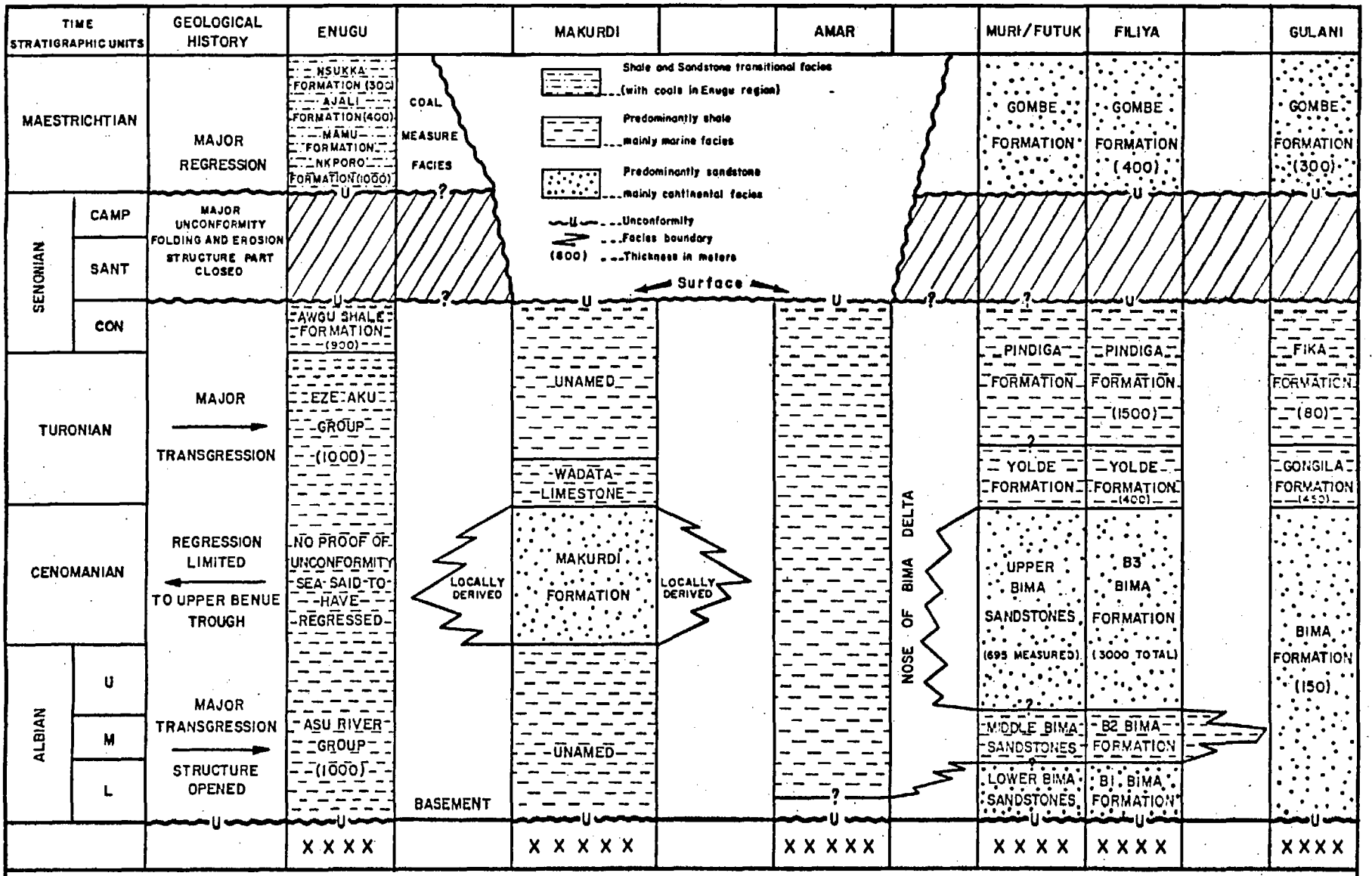


FIG. I.4. General Stratigraphical and Structural history of the Benue Depression.

CHAPTER 2

GENERAL STRATIGRAPHY AND TECTONIC STRUCTURES

2.1 INTRODUCTION

The area of study is made of rocks which have been observed by aerial photographic studies and field mapping to belong to three major formations, which have been assigned on Albian-Upper Turonian age in other areas. These formations are the Bima Sandstones (oldest), the Yolde Formation, and the Pindiga Formation (youngest). Structurally, the area shows evidence for Cretaceous tectonic episodes which affected the rocks in the Benue Valley Trough. The rocks, in the area of study, are intensely folded and fractured, especially in the northern portion where the Jawara Anticline is fractured and displaced into a series of smaller anticlines.

2.2 GENERAL STRATIGRAPHY

The three major formations were identified in the field by lithologic characteristics and textural properties of the rocks. The Bima Sandstones forms the major topographic features in the area. This formation occurs at the base of the sedimentary sequence, and it is predominantly medium- to coarse-grained, planar cross-bedded, feldspathic sandstones. The thickest section measured from this formation in the field is 695 meters, and it is shown in figure 3.4C and D and the location map (fig. 1.1). This section was measured from the Jawara Anticline around Ligri. This thickness is by no means taken to be the total thickness of the Bima Sandstones in this area as its basal contact with the Crystalline Basement Complex and its uppermost limit and contact with the overlying formations were not observed in the field.

The rocks of the other two formations are different both in lithology, textural properties, and in their tectonic style. The Yolde Formation directly overlies the Bima Sandstones. It consists of finer grained material and it is more thinly-bedded compared with the Bima Sandstones. The Yolde Formation is a variable sequence of sandstones and shales which marks the transition from continental to marine sedimentation in this part of the Benue Valley Trough. The Yolde Formation was recognised by Falconer (1911) as the "passage beds", and by Barber et al. (1954) and Reyment (1955) as the "transition beds". The formation consists of very fine-grained grey to pinkish sandstones and greyish sandy mudstones that are thin-bedded and thinly cross-laminated. The youngest of the three formations is the Pindiga Formation which consists of limestones and dark shales. This formation in the area of study is very poorly exposed and it forms lowland where dark soils formed by the weathering of the shales are used for cotton farming. Fig. 1 shows the distribution of the three formations in the area of study. The areas where limestone boulders were observed within dark cotton soils were mapped as Pindiga Formation. Two such places occur south of Muri Village along the road to Didango, and in the synclinal area northeast of Bashama Village.

2.3 TECTONIC STRUCTURES

2.3.1 Introduction

The whole of the Abakaliki-Benue sedimentary basin was subjected to tectonic upheaval towards the end of the Cretaceous, in Upper Santonian times. This series of tectonic movements resulted in uplifting accompanied by folding and fracturing of the sedimentary deposits within

the basin. These events are predominantly expressed in the northeastern part of the area of study, where the Jawara Hills consists of a series of complexly folded and faulted anticlinal structures. In the south the tectonic movements resulted in the formation of the Bogoi anticline which closes to the southwest where it has been severely fractured and displaced.

The major structural features of this part of the basin will now be described and an attempt will be made to undertake a structural analysis of the area to elucidate the pattern of events which gave rise to the structures.

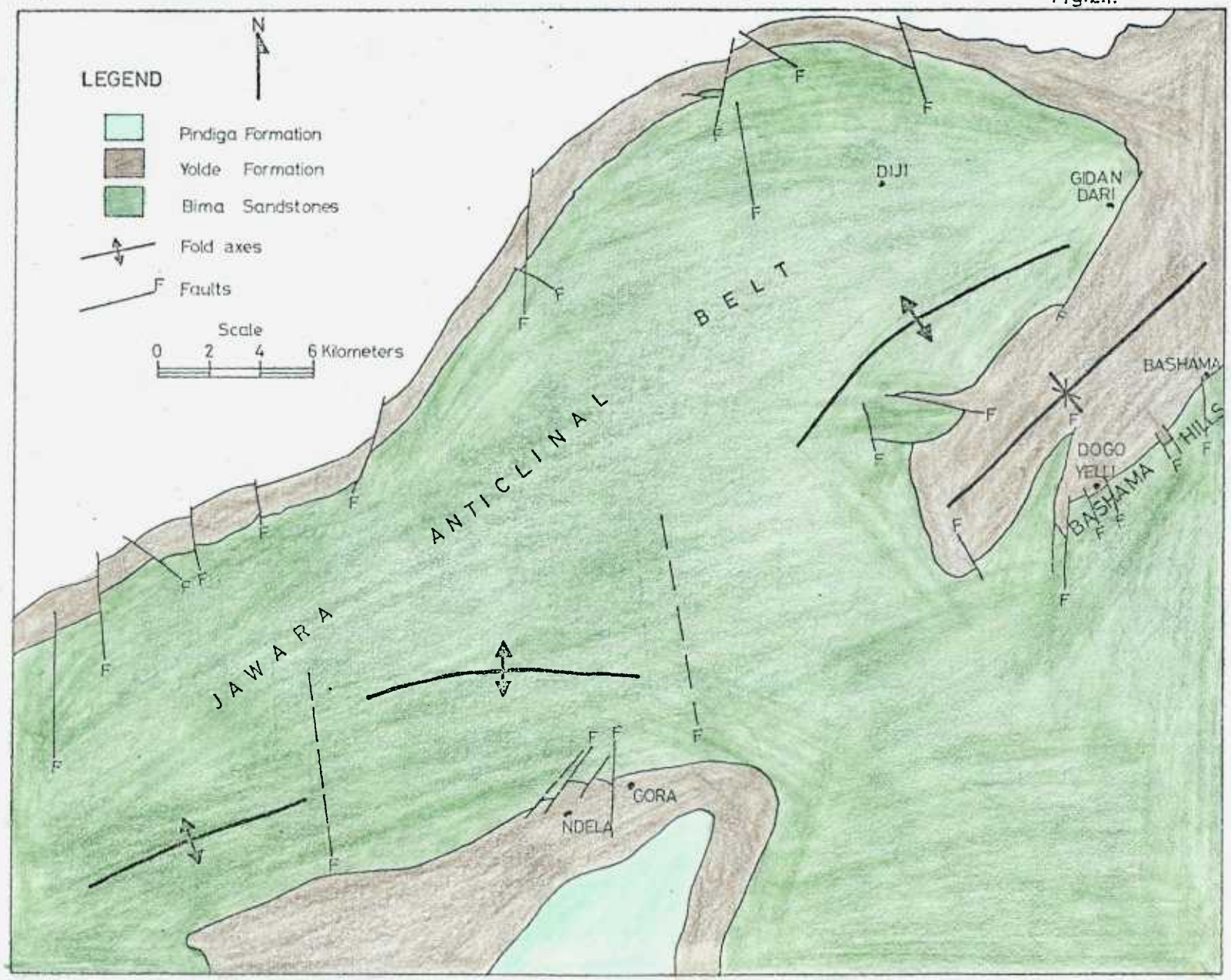
2.3.2 Description of structural features

2.3.2.1 Folds

(a) The Jawara anticlinal belt

The Jawara Hills form a belt of several linearly disposed anticlinal structures which have been severely faulted and rotated with respect to each other. The hills run in a general northeast to southwest direction on which is superimposed a structural trend which appears to have been brought about by the complex faulting, displacement and rotation of an originally single fold axis (fig. 2.1). The folds, a majority of which close westward, are asymmetric with steeper northern limbs. Dip values within the Jawara Anticline range from 45° to 60° in the northern limbs and from 25° to 35° in the southern limbs except in places where the faulting has severely increased the dip values. It is essential to point out the problems posed by the strongly planar cross-bedded nature of the rocks which obscures their true dip and which may

Fig.2.1.



Bima Sandstones structural trends in the northern area

lead to the recording of incorrect dip values.

In the northeastern portion of the Jawara Hills there is a syncline which is a westward extension of the Dadiya Syncline described from the Lau area to the east of the present area of study by Carter et al (1963). The Bashama Syncline is bounded in the north by the Gwana Hills and in the south by the Bashama Hills. The core of the syncline contains limestone boulders and thinly bedded sandstones in the eastern portion.

(b) The Bogoi Anticline

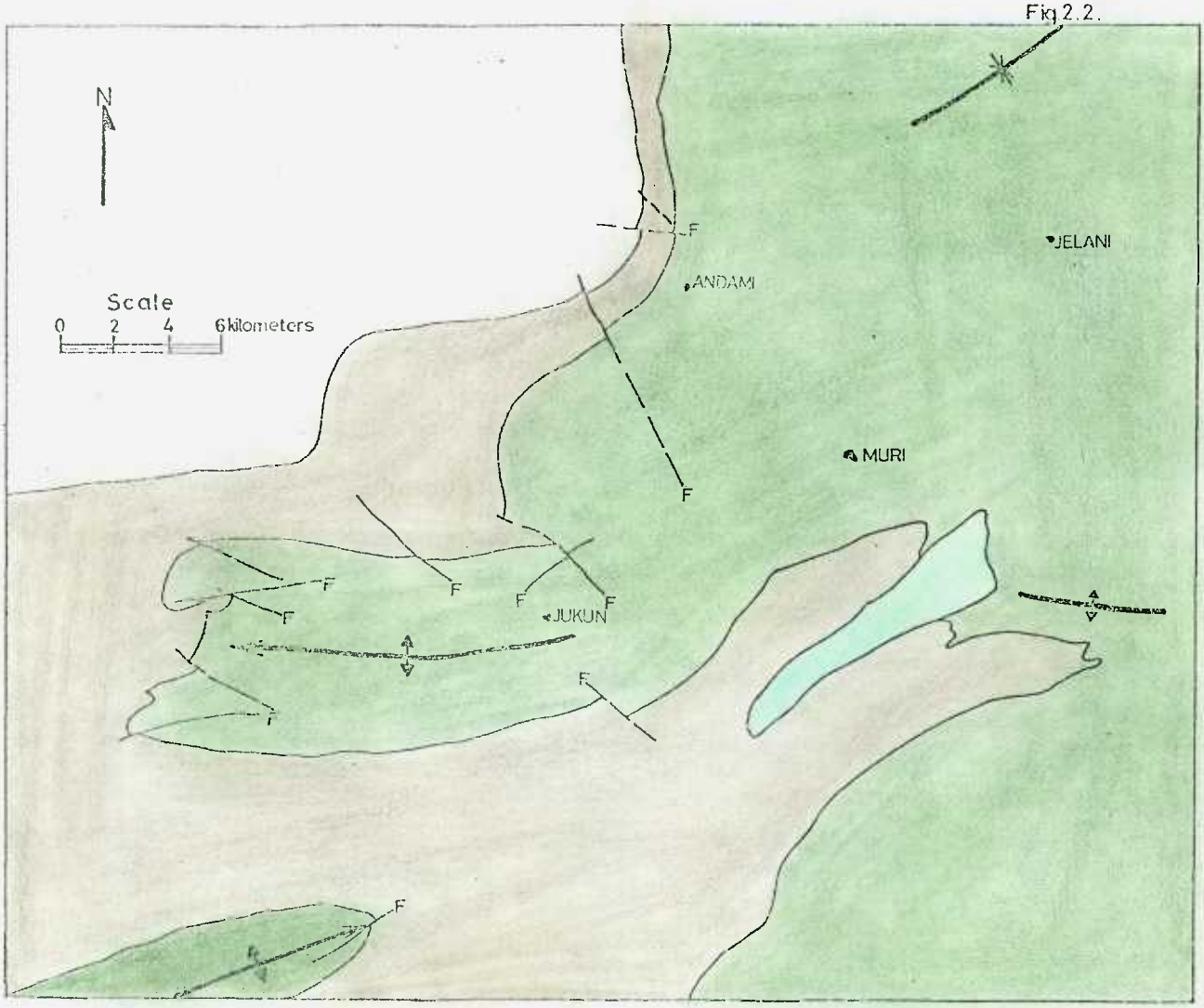
The Bogoi Anticline is a much simpler structure than the Jawara Anticlinal Belt. It lies west of Muri village and southwest of the Jawara Hills (fig. 2.2). The Bogoi Anticline is an asymmetric fold having a more pronounced northern limb and a smaller poorly developed southern limb. The fold closes westward where pitch dips average 50°. As a result of severe faulting along its entire length the northern limb of the anticline appears sinuous. The nose of the fold is severely fractured and rotated to give variable dip and strike values. The lithostratigraphic sequence is poorly exposed in the southern limb which has been deeply weathered.

In the Jelani/Andami area to the east there is a series of hills with rocks which dip northwards in the south, and with a variable dip and orientation in the north as a result of faulting and rotation. The Bashama Hills which form the northern limit of this belt have rocks which dip northwards and form the southern limb of the westward extension of the Dadiya Syncline. The dips in this area average 40°.

2.3.2.2 Fractures

(a) The Jawara Anticline fractures

Fig 2.2.



Bima Sandstones structural trends in the southern area

(for legend see fig.2.1.)

Numerous faults and other fractures criss-cross the Jawara structure with a predominantly north to south trend. Subsidiary trends are in the northeast to southwest, northwest to southeast, and east to west directions. Most of the faults and fractures are minor, and are only obvious or inferred from aerial photographs. The faults are generally difficult to locate in the field as a result of deep weathering of the rocks and deep cover of boulders and rock screes. The position of the faults may occasionally be marked by streams which flow down scarp faces. The monotonous lithology of the Bima Sandstones in the entire area of study accentuates the difficulties of locating the faults in the field, as monotonous lithology makes it impossible to notice a displacement of similar horizons across what may be assumed to be a fault.

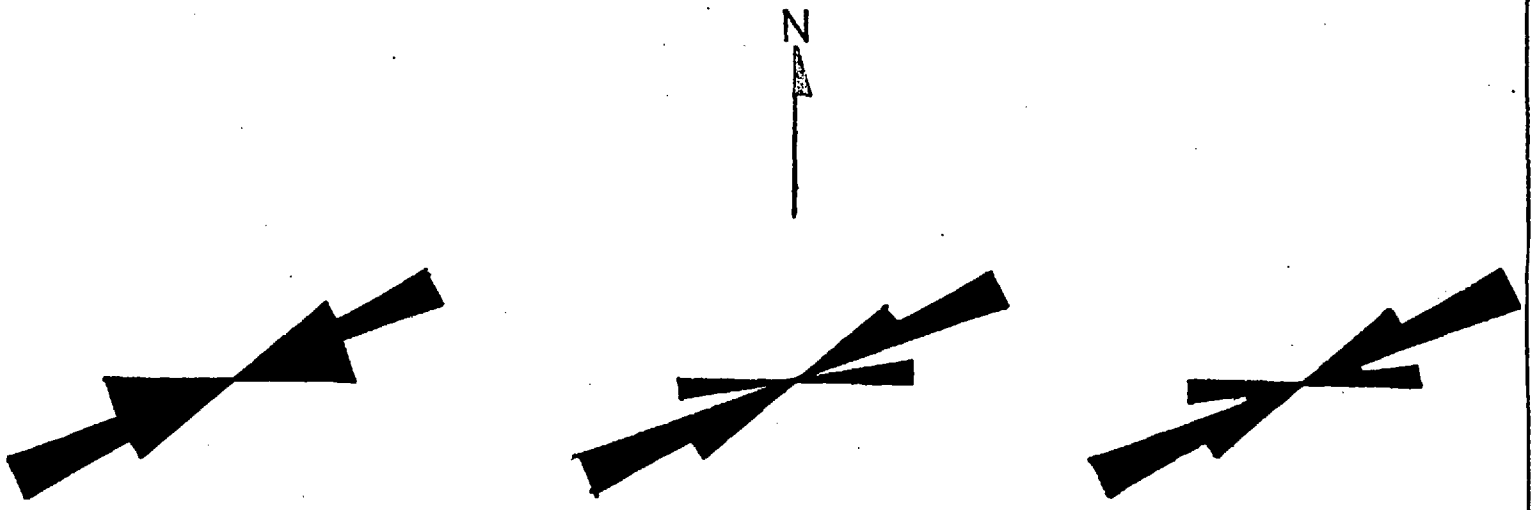
Rapid changes occur in the direction and amount of dips in the Jawara Hills and are due to the severe faulting and rotation of the rocks which affected this part of the basin. In this part of the study area it was possible to observe a few faults in the field, where rocks which dip in different directions are brought into abrupt contact with each other. Sometimes this contact could be traced for some distance along the line of the fault.

(b) The Bogoi Anticline fractures

Faulting and fracturing in the Bogoi Anticlinal structure is less complex than in the Jawara Anticlinal Belt. The Bogoi Anticline faults affect the northern limb and the nose of the fold.

The nose of the anticline is separated into four segments by faults and rotated with respect to each other (fig. 2.2).

Fig. 2.3.



FOLD TRENDS WITHIN THE BIMA SANDSTONES IN THE STUDY AREA.

The Bogoi Anticline fractures are more obvious in aerial photographs than those which affected the Jawara Anticlinal Belt. These fractures are, however, obliterated in the field in the same way as those which affected the Jawara Hills. The faults in the Bogoi Hills area could be inferred from the variable dip and strike directions of the rocks in the different segments of the anticline. Such abrupt changes in dip and strike directions of rocks could only have been brought about by faulting and rotation.

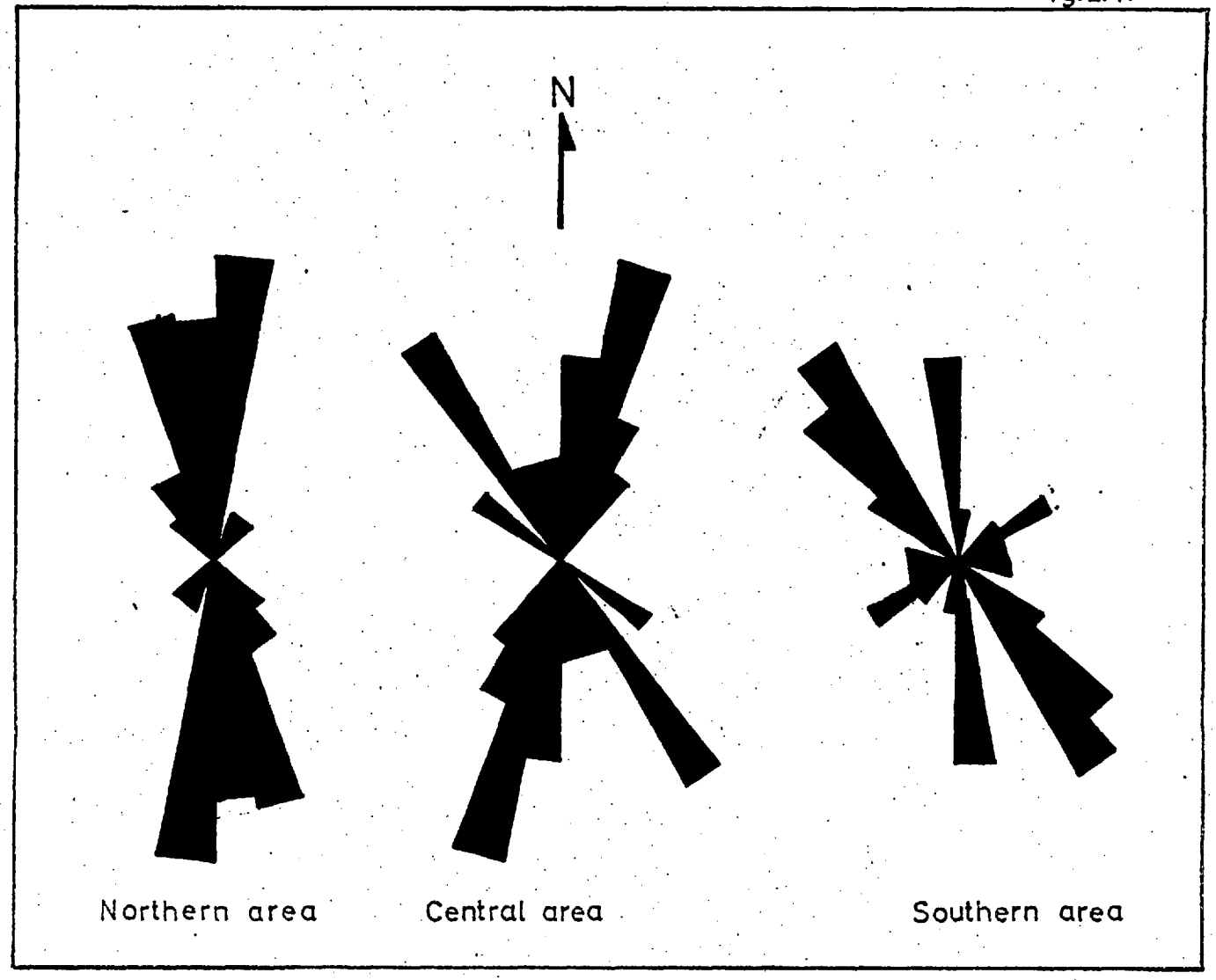
In the Jelani/Andami area, and in areas south of the Bashama Hills the fractures strike in a north to south and northwest to southeast direction. The effect of the faulting in this area also results in variations in the strike and dip of the rocks as in the other two areas already discussed above.

2.4 STRUCTURAL ANALYSIS OF THE TECTONIC TRENDS

An analysis of the structural trends in the area of study will now be attempted based on data collected in the field and from aerial photographic studies. For this purpose the entire area is divided into northern and southern portions. Fault and fold directions are plotted in rose diagrams to illustrate the dominant trend of these structures.

The rose diagrams in figures 2.3 and 2.4 illustrate the fold and fracture pattern in the whole area of study. The trend of the folds is dominantly in a NE-SW direction. The trend of the fractures shows a significant bimodal pattern in the southern portion of the area, while two complementary fracture trends exist in the south and in the central portions. In the northern portion of the area this complementary aspect

Fig.2.4.



Bima Sandstones fracture pattern in the study area.

to the fracture pattern is absent as the fractures trend in a N and NNW to S and SSE directions. A majority of the fractures are normal to the fold axes and the bedding trend. In the south the development of faults parallel to the fold axes occur at the nose of the anticlinal folds. These latter type of fractures which are parallel to the fold axes were referred to as "release joints" by Billing (1954) owing to the fact that they may form due to the release of the principal stress considered to be perpendicular to the fold axes. The development of the major fracture pattern which is normal to the fold and bedding trends was attributed by Pincus (1951) to the existence of the greatest tension axis parallel to the bedding trend.

The folds were formed as a result of compressional forces from the NW and SE. The steep dips on the northern limbs with the corresponding gentler dips on the southern limbs of all the fold structures are probably a testimony of the fact that the compressional forces which operated from the south were greater than those from the north.

Further work is recommended for a more detailed structural analysis of the Bima Sandstones in the area of study. In the main, the structural data presented in the preceding pages of this chapter indicate that only one episode of tectonism affected the Cretaceous sediments in the area of study. Compressive stresses evolved within the basin from the NW and SE resulting in the folding of the rocks, and the development of tensional forces parallel to the fold in post-Bima times during the Cretaceous. An estimate of the fracture trends in these two formations did not show any significant difference from the pattern shown for the Bima Sandstones. The tensional stresses resulted in the fracturing of the rocks along planes normal to the fold axes and bedding trends. The release of the

compressional stresses resulted in the development of the longitudinal fractures which affected the nose of the anticlinal folds in the southern portion of the area.

Chapter 3

STRATIGRAPHIC DESCRIPTION OF THE BIMA SANDSTONES

3.1 GENERAL STATEMENT

Detailed stratigraphic description of the Bima Sandstones has been given by Jones (1962) and Carter et al (1963). The formation is divided into three parts on the bases of lithology and grain size: the Lower, Middle, and Upper Bima Sandstones respectively. The Sandstones are found to underlie the exposed part of the sedimentary succession and directly overlie the crystalline basement rocks in part of the Benue depositional basin. The thickness of the formation is variable within the basin and it ranges from 92 meters to 3,300 meters. The variable thickness arises partly from the uneven nature of the surface of the crystalline basement rocks on which the sediments accumulated, and partly as a result of differing sediment supply to different parts of the basin in the Lower Cretaceous.

3.2 TYPE LOCALITY

The type locality of the formation is at Bima Hill where, although no easily measurable section is available, the general description of the beds was first made along the eastern scarp of the hill. Here, about 4,600 feet ~~meters~~ of the formation is exposed and consists of thick to massive bedded, medium- to coarse- and sometimes very coarse-grained, cream, grey, buff and whitish, feldspathic sandstones. The sediments are often cross-bedded with pebble layers along erosional surfaces. Interbedded with these sandstones are layers of well-sorted, less massive, medium-grained sandstones. Intercalations of thin bands of whitish to dark red siltstones and mudstones occur among the massive sandstones.

3.3 STANDARD SECTION

The Standard Section of the Bima Sandstones is located around Dadiya in the Lamurde anticline where the most complete section of the formation is exposed. In this area the estimated thickness of the formation exceeds 3,300 meters, although the base of the formation and its contact with the crystalline basement is not exposed. The division of the formation into three parts is based on the sequence measured from sections taken along the Dadiya river. The lower part of the formation consists of sandstones and shales; the middle portion is mainly coarse-grained sandstones with interbedded clays and shales, and the upper part is predominantly coarse-grained sandstones. The detailed characteristic features of the three subdivisions are well elaborated in the works of Jones (1962) and Carter et al (1963).

The lower limit of the Bima Sandstones consists of coarse-grained feldspathic sandstones with pebble horizons, and shales with reddish, greyish and greenish clays, thought to be weathering products of basaltic rocks. This unit is also calcareous in parts with well-bedded, medium-grained sandstones and siltstones. The middle unit contains fine-grained materials and consists of very coarse-grained, feldspathic sandstones, thin clays and shales, limestones with numerous indeterminate bivalves. The finer-grained members of this unit are presumed to represent the Albian marine transgression. The upper unit consists of cross-bedded, pebbly, whitish, grey cream and buff coloured medium to coarse-grained feldspathic sandstones.

3.4 STRATIGRAPHIC DESCRIPTION OF MEASURED SECTIONS

The stratigraphic account of the sections measured within the Bima

Sandstones in the area of study will now be discussed, and for the purpose of simplicity the whole area has been divided into six sub-areas: the Bogoi Hills sub-area, Andami-Yamblo sub-area, Jelani sub-area, Gwana/Diji sub-area, Jawara Hills sub-area, and the Bashama Hills sub-area. Figures 3.1, 3.3 and 3.4 show a detailed graphical representation of 18 lithologs compiled from measured sections in the five sub-areas. The exact locations of the sections are shown in fig. 1.1.

3.4.1 The Bogoi Hills sub-area

3.4.1.1 Northern limb of the Bogoi Anticline, locality 1, fig. 3.1A

The recorded thickness of the succession at this locality is about 55 meters. It consists of medium to coarse grained, whitish, pink and buff-coloured, fine to medium-bedded sandstones at the base. These beds lack erosional bases and pebble layers, except along the base of the foreset beds where granule-sized sands grade upwards into finer sand material. The basal beds are succeded by a horizon, about 15 meters thick, which consists of alternating layers of thin-bedded, fine- to medium-grained calcareous sandstones and thin, planar-bedded, fine-grained sandstones. The planar bedding of the fine-grained sandstones is made evident by the presence of alternating bands of pink to reddish, whitish, and buff coloured layers.

Above this horizon is about 10 meters of thin cross-bedded and planar-bedded, predominantly buff coloured, medium to coarse-grained sandstones.

3.4.1.2 Northern limb of the Bogoi Anticline locality 2, fig. 3.1B

The section represents approximately 350 meters of predominantly medium- to coarse and very coarse-grained sandstone with intercalations

of thin bands of fine-grained sandstones.

The basal 53 meters of this section consists of coarse to very coarse-grained grey coloured, pebbly, and thick cross-bedded sandstones. The average thickness of a cross-bedded unit in the basal section is about 70cm. These beds are succeded by about 30 meters of medium to coarse-grained calcareous, cross-bedded sandstones which are reddish brown in colour and contain oyster shells in their upper parts. Above this shelly bed is a 15 meter interval of horizontally bedded very fine-grained sands with intercalated layers of silstone and mudstone. This horizon is deeply weathered and covered with screes and debris. The silty units pass upward into about 4 meters of medium to coarse grained, grey coloured, cross-bedded sandstones containing indeterminate bivalves. The shells occur as casts and moulds which make it impossible to identify them. The rest of the section consists of alternations of thinly laminated fine to medium-grained sandstones and medium to coarse-grained sandstones that are predominantly grey and buff in colour. At the top of the section the rocks are coarse-grained and deeply weathered.

3.4.1.3 Northern limb of the Bogoi Anticline locality 3, fig. 3.1C

The total thickness represented by this section is approximately 250 meters of arenaceous sediments. The basal 90 to 100 meters of the section consists of alternations of medium- to coarse-grained, grey and buff coloured, cross-bedded sandstones, and laminated fine-grained sandstones. The fine-grained material is usually not as well exposed as the coarser-grained sands which are more resistant to weathering. Above this alternating unit is a 15 meter horizon of friable calcareous siltstone and mudstone which is horizontally laminated, becoming sandy

upward and eventually passing into a 12-meter thick laminated fine- to medium-grained sandstone. The friable siltstone and mudstone horizon contains some organic material, probably plant remains which imparts a brownish colour on the siltstones.

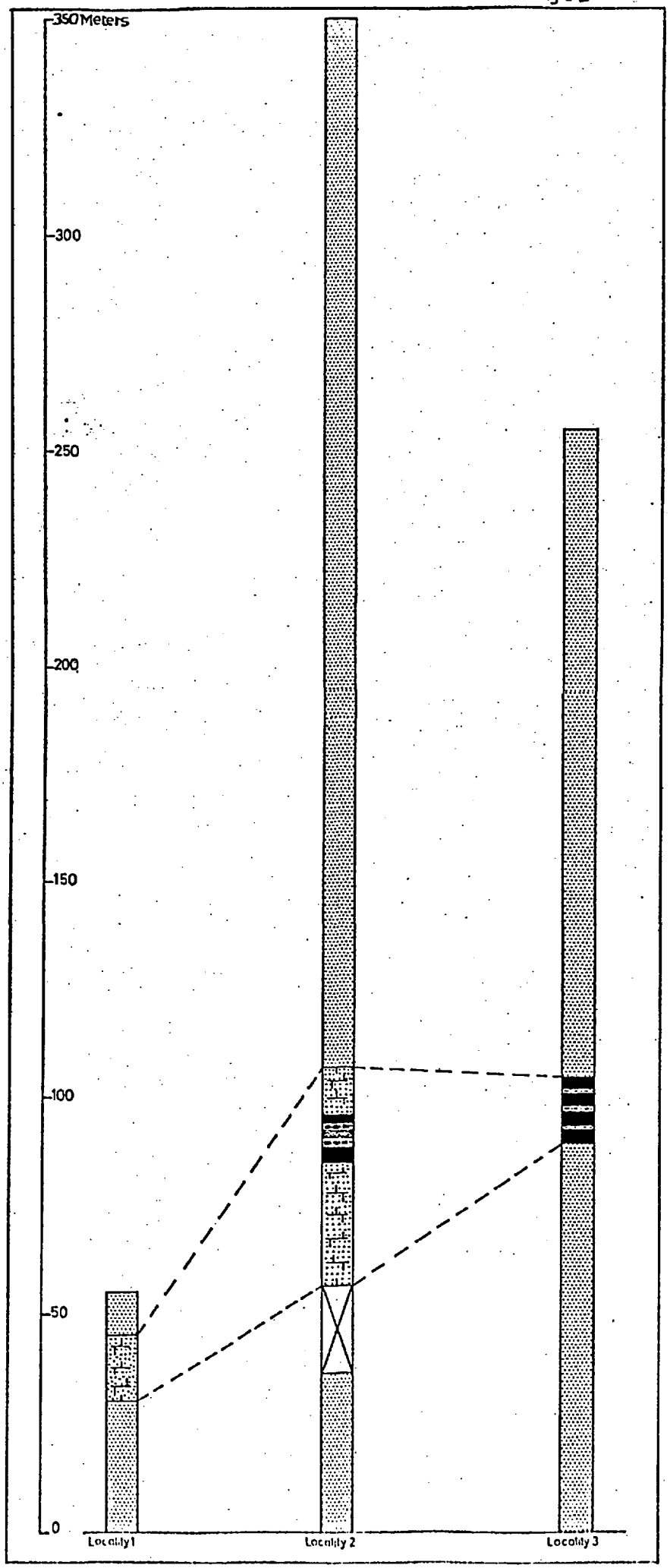
The remaining part of the section consists of fine to medium-grained, and coarse-grained massive beds, and fine to medium and occasionally coarse-grained, cross-bedded sandstones with thin layers of horizontally bedded, medium-grained sandstones. This portion of the sequence contains water ejection structures and minute flexures of the foreset beds. These structures are described in the chapter of sedimentary structures.

3.4.1.4 Discussion

Lithologically, the Bogoi Hills sub-area consists of entirely arenaceous sediments with localised occurrence of shelly calcareous beds. The occurrence of the silty mudstones and siltstones is also highly localised and restricted to only the western portion of the outcrop along the northern limb of the anticline. There is no measurable exposure of the rocks along the southern limb, however, there is a complete absence of the fine grained sediments observed along the northern limb of the anticline.

Figure 3.2 shows the three lithologs taken from the western extremities of the exposure of the Bima Sandstone along the northern limb of the Bogoi anticline. The three logs have been described in some detail in the preceding sections of this chapter. A study of the sections shows that there is a distinct lithologic variation from east to west along the strike of the rocks. Eastward the rocks consist entirely of cross-bedded, horizontally laminated and massive sandstones

Fig 3.2



Correlation between three localities in the Bogoi hills area.

of variable grain size. Westward, along the strike at locality 3, a horizon of calcareous siltstones and mudstones is encountered within the sequence. This horizon which is dark, brownish in colour in places shows evidence of the presence of organic matter, probably plant remains. Samples taken from this horizon effervesced with hydrogen peroxide. Microscopic study of prepared portions of these samples did not show the presence of microfossils. Further westward, at locality 2, a horizon of calcareous shelly (oysters) sandstones, reddish brown and cross-bedded, i s encountered above which a 15 meter thick argillaceous horizon passes upward into a thin layer of calcareous cross-bedded sandstones bearing a suite of inde terminate bivalves. Westward of locality 2, no calcareous or argillaceous horizon is encountered along the sequence until locality one is reached at the nose of the fold, where calcareous sandstones are present.

The poor nature of the shells makes it impossible to determine whether they are marine or fresh water. It is, however, suggested that they are probably fresh water organisms which lived in localised pockets of water where the current flow was relatively moderate but strong enough to prevent the settling of fines, and allow animal life to thrive.

The sandstones of the Bogoi Hill sub-area are highly feldspathic. The weathering of the feldspars produces a greyish powder in between quartz grains. Where the feldspars have not been deeply weathered they impart a pinkish colour to the rocks.

3.4.2 The Andami/Yamblo sub-area

This sub-area lies to the east of the Bogoi Hills, and consists of virtually planar cross-bedded, medium to coarse-grained sandstones throughout the entire thickness of the sequence. The sections are capped by ferruginised sandstones which consist of silica cemented detrital quartz grains to the complete exclusion of feldspars. The iron oxide occurs as a matrix material.

3.4.2.1 Andami west, locality 6, fig. 3.1F

The exposed part of the sequence here is about 280 meters in average thickness. For the most part the bedding units are thin and exhibit only the planar type of cross-stratification. The foreset beds dip constantly to the southwest with an occasional reversal in the dip direction to the northeast. The beds are gritty as a result of granule size pebbles which are dispersed all over the rocks. At the bases of some of the bedding units quartz pebbles concentrate as a thin layer. The concentration of coarser grained material also occurs along the bases of the foreset beds. The forset beds show a definite fining upwards sequence which is absent in the major bedding units.

The ferruginised units capping the sequence in this locality are about 30 meters thick. The contact between the iron-containing units and the sandstones beneath them is gradational. A common feature of the ferruginised units is the complete absence of feldspars from the rocks. The highly ferruginised units are cross-bedded, medium- to coarse-grained, and reddish brown in colour. Ferruginised siltstones were described from the Lau and Yola areas around Tula by Jones (1962). In the area of study the iron oxide is sometimes concentrated along bedding planes or along veins criss-crossing the bedding units.

3.4.2.2 Yamblo (Andami east), locality 7, fig. 3.3A

The exposed part of the sequence in this locality is approximately 415 meters thick. The basal 1.5 meters consist of massive-bedded, medium- to coarse-grained, gritty sandstones of pink to grey colour. The rest of the sequence consists of medium- to coarse-grained sands, occasionally fine- to medium-grained in places. The entire sequence is gritty, and the foreset beds are inclined at a lower angle than elsewhere in the area of study. Most of the cross-bedded units exhibit fine cross-laminations. Toward the top of the section there are beds which show reversals in current direction. Ferruginised layers also occur along bedding planes near the top of the sequence. The ferruginised layers become very thick and reach up to more than 1 meter at the top of the sequence. The iron-containing bands are reddish brown in colour.

Further east from Yamblo at locality 8 where the sequence is 190 meters thick, ferruginised bands also occur at the top of the section. The rocks are also gritty and exhibit planar cross-stratification.

3.4.2.3 Discussion

The occurrence of iron oxide as matrix material in the Bima Sandstones is restricted to the Andami/Yamblo and the Jelani sub-areas. The iron oxide, in all the localities where it was observed, occurs near the top of the sequence as bands along bedding planes and within the beds where it imparts a reddish brown colour to the beds. The mode of occurrence of the iron oxide suggests that it is not primary to the rocks, and that it is probably deposited from percolating surface waters. At locality 8 the iron oxide is powdery with a pinkish white colour.

There is a complete absence of finer-grained material (silts and

clays) and horizontal stratification in this sub-area. The coarse grain size and the exclusion of horizontal stratification in the rocks of this sub-area are attributed to action of strong violent currents of low regime flow during the deposition of the beds.

3.4.3 The Jelani sub-area

This sub-area is represented by two measured sections each about 320 meters in total thickness of the exposed portion of the sequence. The hills formed by the Bima Sandstones in the Jelani area have very steep scarp faces, and it was not possible to measure the section to the top of the sequence in one of the two localities. Ferruginised horizons also occur at the top of the sections in this area.

3.4.3.1 Jelani Village, locality 9, fig. 3.3C

The exposed part of the sequence in this locality is approximately 320 meters. The basal 29 meters of the section consists of massively cross-bedded, medium- to coarse-grained, gritty sandstones that are grey in colour. The beds become more thinly cross-bedded above but still gritty and medium- to coarse-grained. Pink coloured beds also occur and conglomeratic bases occur along bedding planes. At the top of the sequence the beds are ferruginised and coarser-grained. There is a complete absence of fine-grained material in this locality.

3.4.3.2 Jelani northeast, locality 10, fig. 3.3D

Steep scarp faces occur at this locality and the top of the sequence was not reached due to difficulties encountered. The measured portion is 315 meters in thickness. The basal 40 meters consists of cross-stratified units of grey-coloured gritty sands. The rest of the sequence consists of fine- to medium-grained, grey coloured, gritty, cross-stratified units.

Some of the cross strata have asymptotic bases that make them resemble trough cross-strata. Three horizons with carbonate cemented, feldspathic, cross-stratified sands occur near the top of the horizon with greenish colour and fine-grained texture. Near the top of the sequence thin bands of very fine-grained sands occur within medium to coarse-grained, cross-bedded sandstones. The fine-grained sands are also finely cross-laminated.

The occurrence of calcite cement in some of the rocks is thought to be diagenetic. There are no fossils in the rocks and the calcification of the rocks is localised both vertically and laterally along the out-crop. The origin of the carbonate cement probably lies in calcite-rich groundwaters of localised circulation.

3.4.4 Bashama sub-area

The litho-stratigraphic sequence in this sub-area does not differ very much from that already described in other areas. The sequence consists entirely of medium- to coarse grained sandstones with occasional interbeds of finer-grained sands. Water ejection structures occur in the fine-grained sands. Much of the sequence exposed along the Bashama Hill consists of thin-bedded units with cross laminations.

3.4.4.1 Bashama hill, locality 11, fig. 3.3E

The total thickness of the exposed part of the sequence at this locality is approximately 410 meters. At the base the beds are grey, gritty, thin bedded fine to medium grained sandstones; horizontally stratified beds occur within this horizon, and in some could be found water ejection structures. These beds pass upward into medium- to coarse-grained, gritty, buff coloured sandstones. Interbedded with

these coarse-grained rocks are beds of fine- to medium-grained sands. The whole unit becomes greyish coloured at the top, with some of the cross beds exhibiting asymptotic bases. The foreset bedding units also become thicker towards the top of the sequence.

3.4.4.2 Bashama hill, locality 12, fig. 3.3F

The exposed part of the sequence at this locality is about 290 meters thick. The rocks are buff coloured, medium to coarse-grained, cross-bedded sandstones. Interbeds of fine to medium-grained sands are common and these often show horizontal stratification. At the top of the sequence the rocks become very coarse-grained and gritty. Thin layers of quartz pebbles occur along some of the bedding planes.

The occurrence of fine-grained sands within the sequence indicates periods of quieter current conditions from the violent current condition in operation during the deposition of the coarser grained materials. The occurrence of horizontal stratification, on the other hand, indicates periods of high flow regimes during the deposition of the sands in the Bashama sub-area.

3.4.5 Gwana/Diji sub-area

The Gwana/Diji sub-area occupies the northeast portion of the area of study, and consists of highly dissected and displaced hills. The tectonic influence which affected the Bima Sandstones is more pronouncedly expressed in this sub-area than elsewhere in the entire area of study. The part of the sequence represented here consists of fine- to coarse-grained, thin to massive-bedded, planar and trough cross-bedded sandstones. Deformed varieties of cross-bedded sandstones are ubiquitously distributed, especially near the top of the sequence. Water ejection structures are

also commonly developed in this area.

3.4.5.1 Gwana Hill, locality 17, fig. 3.4E

The exposed part of the sequence at this locality is approximately 340 meters, and the rocks are cross-bedded throughout the whole of the sequence. The basal beds are coarse- to very coarse-grained, massive, and grey in colour. Towards the top of the sequence the beds are less massive and are medium- to coarse-grained. Trough cross stratifications occur in the beds exposed at this locality.

3.4.5.2 Diji locality 18, fig. 3.4F

The part of the sequence exposed at this locality is very thin, being roughly 42 meters. The rocks are fine- to medium-grained with no quartz pebbles, finely cross-bedded and grey in colour. Toward the top of the sequence the foreset beds are folded in various manners from simple wavy contortions of the foreset laminae to complicated over-folds. In these deformed beds the foreset laminae maintain their individuality and can be traced throughout the folded areas. The significance and the probable origin of these secondary sedimentary structures have been treated in another chapter (see chapter 7).

3.4.5.3 Diji locality 19, fig. 3.4G

The part of the sequence exposed here is approximately 51 meters thick. It consists of planar cross-stratified beds, some of which have asymptotic bases. The direction of current flow is reversed as opposed to the general flow direction to the southwest in other areas. The rocks are fine- to medium-grained and thin bedded with pebble layers along bedding planes.

The Gwana and Diji areas have several beds showing deformed foreset beds. These deformed strata appear to be restricted to the top of the sequence. Generally, the beds are thin and the rocks are fine- to medium-grained. This attitude of the beds in this area is attributed to their position near the top of the formation where the Bima Sandstones are thin bedded and finer grained. The deformed foreset beds occur more in fine-grained sands than in coarse-grained ones.

3.4.6 Jawara hills sub-area

This sub-area comprises the rocks exposed along the north and south limbs of the Jawara Anticline complex. The Jawara anticline has a nose which closes to the west and it is dissected by faults and fractures and displaced progressively to the southwest. The sandstones which form the hills are gritty in parts, and nearly always medium to coarse-grained in texture.

3.4.6.1 Southern limb of the Jawara Anticline, locality 13, fig. 3.4A

The portion of the sequence exposed here is about 315 meters thick. At the base it consists of an alternation of grey, medium- to coarse-grained, cross-bedded sandstones and fine-grained, horizontally laminated sandstones. Near the top about 7 meters of the beds are trough cross-stratified and contain pinkish, coarse-grained sands. The topmost beds in the sequence are planar cross-stratified, grey in colour and coarse-grained.

3.4.6.2 Southern limb of the Jawara Anticline, locality 14, fig. 3.4B

The exposed part of the sequence in this locality is about 330 meters thick. The section consists of grey, medium to coarse-grained cross-bedded sandstones at the base and at the top. In the middle portion of the

sequence there occurs a 1.6 meter thick horizon of fine to medium-grained, thick cross-bedded sandstones. The rest of the sequence is an alternation of fine- to medium-grained and coarse-grained, cross-laminated sandstones. Pink feldspars present in the rocks are responsible in part for their pink colouration.

3.4.6.3 Northern limb of the Jawara anticline, locality 15, fig. 3.4C

375 meters of the sequence is represented at this locality. The rocks are mostly pink coloured, medium to coarse-grained, cross-bedded sandstones. Like in all the other localities already discussed the beds can only be traced for short distances laterally. Horizons of fine to medium-grained sandstones with quartz pebbles occur along bedding planes and are present in the middle portion of the sequence. Fine-grained, thin-laminated and cross-bedded sandstones also occur within the sequence.

3.4.6.4 Northern limb of the Jawara Anticline, locality 16, fig. 3.4D.1 and 2

The total thickness of the exposed part of the sequence could not be measured at this locality due to the difficulties encountered along numerous steep scarp faces on the exposure surface. The measured portion is approximately 695 meters thick and this represents a section taken along a stream flowing from the top of hill at this locality. The rocks are pink to grey in colour, predominantly medium to coarse-grained in texture, and planar cross-bedded for the most part of the sequence. Fine-grained, cross- and horizontally-laminated sands are interbedded with the medium to coarse-grained sandstones. Trough cross stratifications are also commonly developed, especially towards the top of the sequence.

3.4.7 General discussion of the Bima Sandstone Stratigraphy

The stratigraphic description of the Bima Sandstones at different

localities covering the entire area of study shows that the sequence is very monotonous in lithology and stratification types. The lithology is dominated by sands of medium to coarse grain in texture, and the stratification is predominantly of the planar type of cross bedding. There is a complete absence of argillaceous sediments, and fine-grained sands are sparingly developed except in two localities in the Bogoi hills sub-area where a thin horizon of siltstones and mudstones was recorded. These siltstones are carbonaceous and palynological examination of extracts indicated the presence of a few brown spores. The brown appearance of the spores is an indication of the degree of metamorphism which the rocks have undergone. Detailed palynological studies of these rocks is beyond the scope of this work.

The pink colour of the sandstones is largely due to the presence of pink feldspars, and in part to pigments in the rocks. The grey colour of the rocks is due to the large quartz grains and the alteration products of the degraded feldspars. Iron oxide present in the rocks gives a reddish-brown colour to them, especially at the top of the sequence in the Andami/Yamblo, and Jelani sub-areas, where the sequence is capped by ferruginised sandstones.

Generally, the bedding units of the Bima Sandstones lens out laterally and it is difficult to trace them along the outcrop for more than a few meters. Such lensing attitude of cross-bedded units is attributed to the mode of formation of the beds by the accretion of material on the slip face of a bar and the overlapping of many such bars within a stream channel.

The inaccessible nature of the area of study and the presence of

steep scarp faces on the hills make section logging tedious and slow. The absence of marker horizons and the monotonous lithology both preclude the use of such sedimentary attributes for correlation purpose. There is also a complete absence of microfossils that could be of help in correlation between units.

Basement rocks were not observed to be exposed anywhere in the area of study. The texture, lithology, and sedimentary structures of the sandstones in the whole area reflect deposition in a braided stream channels. The sedimentary attributes of the sandstones compare favourably with those described from the upper units of the standard section by Carter et al (1963), and they lead the present writer to conclude that the sandstones in the present area of study are equivalent to the Upper Bima Sandstones of the standard section.

CHAPTER 4

HEAVY MINERAL ANALYSIS

4.1 INTRODUCTION

The heavy mineral components of terrigenous sediments usually consists of clastic grains with specific gravities in the neighbourhood of 2.9. The heavy mineral content of a sediment or sedimentary rock can be used as a guide for the provenance of the rocks. Hubert (1962) used the combined percentage of zircon, tourmaline and rutile (ZTR index) as an index of mineralogical maturity of sandstones that contain very little feldspars. Heavy minerals also provide information about conditions of diagenesis in the rocks. The processes of intrastratal solution are indicated by the solution or corrosion of the heavy minerals and this may lead to the complete destruction of important minerals such as pyroxenes and hornblende which are good provenance indicators. Intrastratal solution also alters the ZTR index and this affects its reliability as an indicator of the degree of maturity of a rock at the time of its deposition.

Heavy mineral studies have been employed in the solution of many geological problems. Walker (1967) attacked the red-bed problem, in part using heavy mineral studies. From his studies he showed that the Pliocene red beds in California were formed authigenically by intrastratal alteration of hornblende and biotite. Hand (1967) used differences in settling velocities of light and heavy mineral fractions to differentiate between eolian and beach sands. White and Williams (1967) used the differences in the settling velocities of quartz and tourmaline in various parts of a cross-bed set to demonstrate differences in the ratios of

traction to suspension sedimentation in different parts of the structure.

Heavy mineral studies are also employed in the exploration and evaluation of sedimentary ore deposits and the up-stream tracing of hydrothermal plutonic ore minerals.

In the Bima Sandstones studies the heavy mineral analysis is carried out to demonstrate the types of heavy minerals present in the rocks and to use them as a guide to determine the provenance of the unit. An attempt will also be made to use the heavy mineral content of the sandstones as a correlation tool as the rocks are totally devoid of any microfossils and marker horizons.

Jones (1962) in his work on the Bima Sandstones in the Yola and Lau areas mentioned the presence of the following heavy mineral suite in the rocks: pyrite, magnetite, hematite, ilmenite, epidote, zircon, tourmaline and sphene. In the present study the heavy minerals of 39 samples from different localities were separated and identified with the aid of a petrographic microscope and X-ray diffractograms. A maximum of 15 varieties were identified and the results were treated statistically in the samples used for the present study; Limonite, Leucoxene, magnetite, ilmenite, "alterite", zircon, tourmaline, epidote, sphene, rutile, amphibole, apatite, brookite, barytes, and garnet. They are treated under two major headings as opaques and non-opaques. All the mineral grains which have been altered as a result of severe chemical weathering, and were therefore unidentifiable under the microscope are collectively grouped under the category alterite in this study (Van Andel, 1950; Briggs, 1965; and Chang, 1971).

4.2 METHODOLOGY

4.2.1 Sampling

The samples used in the present study were selected from 150 rock samples collected by the writer. The choice of samples for study did not follow any predetermined pattern. However, an average of two samples were selected from each measured section. An attempt was made to limit the number of samples selected for study to a sizeable minimum as each sample was divided into four size fractions as an excess of samples would have resulted in a large number of ultimate samples to be treated and studied.

4.2.2 Laboratory study

4.2.2.1 Crushing procedures

All the samples used in the present study were fairly indurated and they had to be crushed, using different methods, so as to reduce them to small sizes and to ultimately free the individual grains from the rocks. An iron pestle and mortar was first used to break the rock samples to reasonably small fragments. Care was taken not to grind the samples as this could result in the mechanical fragmentation of the individual grains. Thus, only an up and down movement of the pestle on the mortar was used to break the rocks into smaller fragments. Samples with carbonate cement and iron oxide matrix were first digested in dilute hydrochloric acid before crushing. Use of an over concentrated acid solution leads to the etching of the heavy mineral grains, and if etched grains are found in samples treated with acid, they cannot readily be distinguished for natural ones (Hutton, 1950). Another problem posed by acid treatment is the loss of apatite which is destroyed by acids.

4.2.2.2 Size separation

Rubey (1933) demonstrated that heavy mineral grains are hydraulically equivalent to light mineral grains of a larger size, depending on the specific gravity of the heavy mineral. A heavy mineral suite will therefore vary in composition with the grain size of the sample.

Opinion is varied as to the range of size fractions suitable for heavy mineral studies. Young (1966) found little difference in the average reproducibility of grain counts between the 61 to 88 microns and 124 to 175 microns size fractions. He, however, recommended the smaller size fraction on the basis of tests of reproducibility of heavy mineral ratios. Krumbein and Rasmussen (1941) found that sampling and laboratory errors were lower for the coarser fraction of two size fractions they studied. Muller (1967) recommends the study of the 63 to 200 micron size fraction. Hails and Hoyt (1972) studied the 62 to 250 micron fraction; Tieh (1973) the 62 to 125 microns fraction; Stanley (1965) the 100 to 450 microns fraction; and Van Andel and Poole (1960) studied the 63 to 500 microns fraction. From their studies Van Andel and Poole were able to show that heavy mineral distribution patterns are independent of grain size in their samples. Carver (1971) recommended the 125 to 250 microns fraction as this is a class "narrow enough to produce uniform optical effects and reduce or eliminate variation in heavy mineral proportions caused by differences in grain size of the samples".

Chang (1971) sieved his samples at 0.5 phi interval for the heavy minerals studies of the Wash. In the present exercise the 62.5 to 210 microns (2.25 to 4.0 phi) size fraction was chosen for study because preliminary investigation showed that the bulk of the heavy minerals are

concentrated in this size range in the Bima Sandstones in the area studied.

All samples were sieved into four fractions between 62.5 and 210 microns. Samples containing too much silty and clayey materials, that is, particles with diameter less than 62.5 microns, were wet-sieved to remove the fines before the heavy minerals from a single sieve fraction were separated for analysis. The material collected from each sieve fraction was carefully weighed before the heavy minerals separation was carried out.

The percentage of the heavy fraction in each size grade was determined. The grain size distribution of the heavy minerals in each sample was calculated and the results expressed as histograms (fig. 4.1). The percentage of the non-opaque fraction of each size grade was also determined and expressed as histograms for three samples (fig. 4.2).

4.2.2.3 Heavy mineral separation

A wide range of techniques is available for heavy mineral separations. In the present study the heavy liquid method of separation under gravity was employed to concentrate the heavy minerals.

The use of heavy liquids for the separation of minerals depends on differences existing between the specific gravity of the different mineral types. The heavy liquid used for the Bima Sandstones studies is tetrabromoethane ($C_2H_2Br_3$ Sp.gr. 2.96 at 20°C). Tetrabromoethane is readily miscible with acetone and alcohol and can therefore be recovered from either of these liquids by washing with water.

The apparatus used for the separations consists of a steep sided funnel, watch glass, rubber tubes and stopper, glass fitter funnel and

collecting bottles for used liquid. The separation of the heavy minerals was carried but along lines described by Carver (1971) and Allman and Lawrence (1972). A total of 150 size fractions was processed for heavy mineral separation from the 39 rock samples. The heavy minerals from each sieve fraction were weighed before they were mounted on slides with canada balsam for petrographic identification. Heavy mineral identification was carried out with the aid of Milner's (1962, fourth edition) Introduction to Sedimentary Petrology; Kerr's (1959) Optical Mineralogy; Smith's (1968) Minerals and the microscope; and Harker's (1960) Petrology for students.

4.2.2.4 Procedures for mineralogical analyses under the microscope

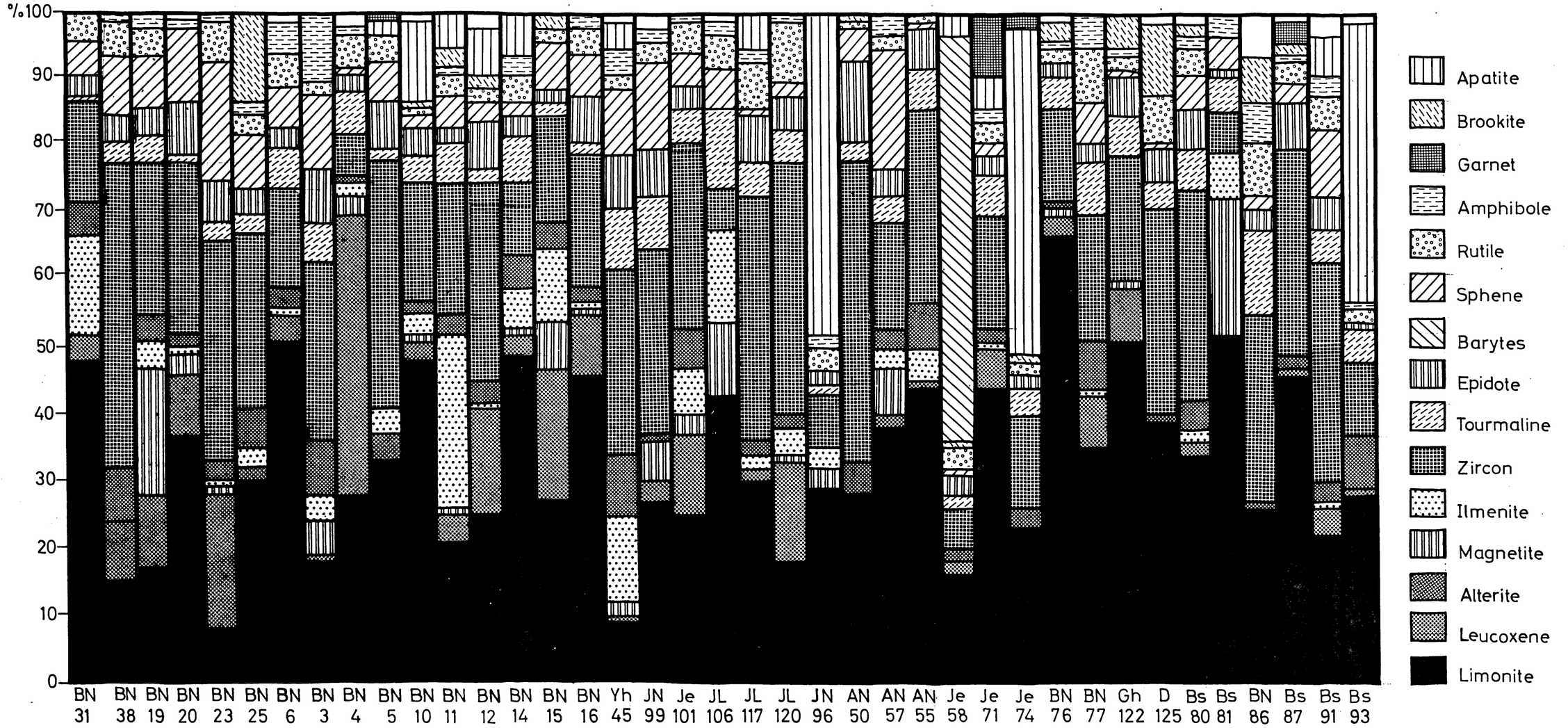
The relative proportions of the different heavy minerals were determined by point counting using a mechanical stage, at a magnification of X 100. The stage was turned slowly clockwise and only those grains that touched the intersection of the cross-hairs were counted. The whole slide was traversed by moving the mechanical stage from left to right along the horizontal and also along the vertical at 1.0 mm intervals. In every slide at least 250-400 grains were counted. From these counts the ratio of the opaque to the non-opaque minerals in each slide was estimated.

The percentages of the various transparent heavy mineral species in each slide were determined by considering the total number of only the non-opaque minerals. Such minerals as zircon, tourmaline, amphiboles, sphene, epidote, rutile, brookite, and apatite, were found to be the predominant non-opaque minerals. Their petrographic identification was confirmed by X-ray diffraction studies. Baryte was found in only one sample and garnet occurred in small quantities in a number of samples (table 4.1).

TABLE 4.1: Mineral percentages of heavy fractions of whole rock sample.

Sample No.		Limonite	Leucoxene	Magnetite	Ilmenite	Alterite	Zircon	Tourmaline	Epidote	Sphene	Rutile	Amphibole	Brookite	Barytes	Apatite	Garnet
BN 31	46	4	-	15	5	15	1	3	5	5	1	1	1	1	1	
BN 38	15	7	-	2	8	46	3	4	9	8	2	-	-	-	-	
BN 19	17	11	19	4	4	23	4	4	8	4	3	-	-	-	-	
BN 20	37	9	3	1	2	26	1	8	11	-	1	-	-	-	-	
BN 23	8	20	1	1	3	33	3	6	18	6	1	-	-	-	-	
BN 25	30	2	-	3	6	27	3	4	8	3	2	13	-	-	-	
BN 6	51	4	-	1	3	15	6	3	6	5	5	-	-	-	-	
BN 3	18	1	5	4	8	27	6	8	11	2	11	-	-	-	-	
BN 4	28	42	3	2	1	6	7	2	1	5	1	-	-	-	-	
BN 5	33	4	-	4	-	37	2	7	6	4	2	-	-	7	2	
BN 10	48	3	1	3	2	15	7	4	2	1	-	1	-	14	-	
BN 11	21	4	1	26	3	20	6	2	5	3	1	3	-	5	-	
BN 12	25	16	-	1	3	29	3	7	3	2	2	-	-	7	-	
BN 14	49	3	1	6	5	11	7	3	2	4	3	-	-	7	-	
BN 15	27	20	7	11	4	16	2	2	7	2	-	2	-	-	-	
BN 16	46	9	1	1	2	19	3	7	6	4	2	-	-	-	-	
YH 45	9	1	2	13	9	28	9	8	10	2	4	-	-	4	-	
JN 99	27	3	6	1	-	29	8	7	13	3	2	-	-	-	-	
JE 101	25	12	3	7	6	28	5	3	5	5	1	-	-	-	-	
JL 106	43	-	11	14	-	6	12	-	6	6	2	-	-	-	-	
JL 117	30	2	-	2	2	38	5	7	1	7	2	-	-	5	-	
JL 120	18	15	11	4	2	38	5	5	2	9	4	-	-	1	-	
JN 96	29	-	3	3	-	8	1	2	-	3	1	1	-	48	-	
AN 50	28	-	-	-	5	45	3	12	5	1	-	-	-	-	-	
AN 51	38	2	7	3	3	16	4	4	18	2	4	-	-	-	-	
AN 55	44	1	-	5	7	29	6	6	1	1	2	-	-	-	-	
JE 58	16	2	-	-	2	6	2	3	1	3	-	1	61	3	-	
JE 71	44	6	-	1	2	17	6	3	2	3	2	-	-	5	11	
JE 74	23	-	-	-	3	14	4	2	-	2	1	-	-	49	1	
BN 76	67	3	1	-	1	14	5	2	-	2	1	3	-	-	-	
BN 77	35	8	-	1	7	19	8	3	6	8	6	-	-	-	-	
GH 122	51	8	1	-	-	19	6	6	1	2	1	5	-	-	-	
D 125	39	-	-	-	1	31	4	5	1	7	-	11	-	-	-	
BS 80	34	2	-	2	4	32	6	6	5	4	2	2	-	1	-	
BS 81	52	-	21	7	-	6	1	1	5	-	2	-	-	-	-	
BN 86	26	-	-	-	1	33	13	3	2	8	6	7	-	-	-	
BS 87	46	1	0	-	2	28	3	7	3	3	1	2	-	-	3	
BS 91	22	4	-	1	3	33	5	5	10	7	3	-	-	6	-	
BS 93	28	-	-	-	8	11	5	1	-	2	1	-	-	42	-	

Fig.4.1



The percentage occurence (by number) of heavy minerals identified in the Bima Sandstones

4.2.2.5 Calculation of data

The percentage of each heavy mineral species in each size grade was calculated for all samples, and the percentages of the mineral species of the whole sample were also calculated. The values obtained for both sets of calculations are summarised in tables 4.1 and 4.2, and figs. 4.1 and 4.2. The percentages of the mineral species in the whole sample were obtained by dividing the total number of each mineral species in all size grades of that sample by the total number of grains counted for the whole sample. This can be expressed by the following formula:

$$P(X) = \frac{\sum_{s=1}^{n} X_s}{\sum_{s=1}^{n} N_s}$$

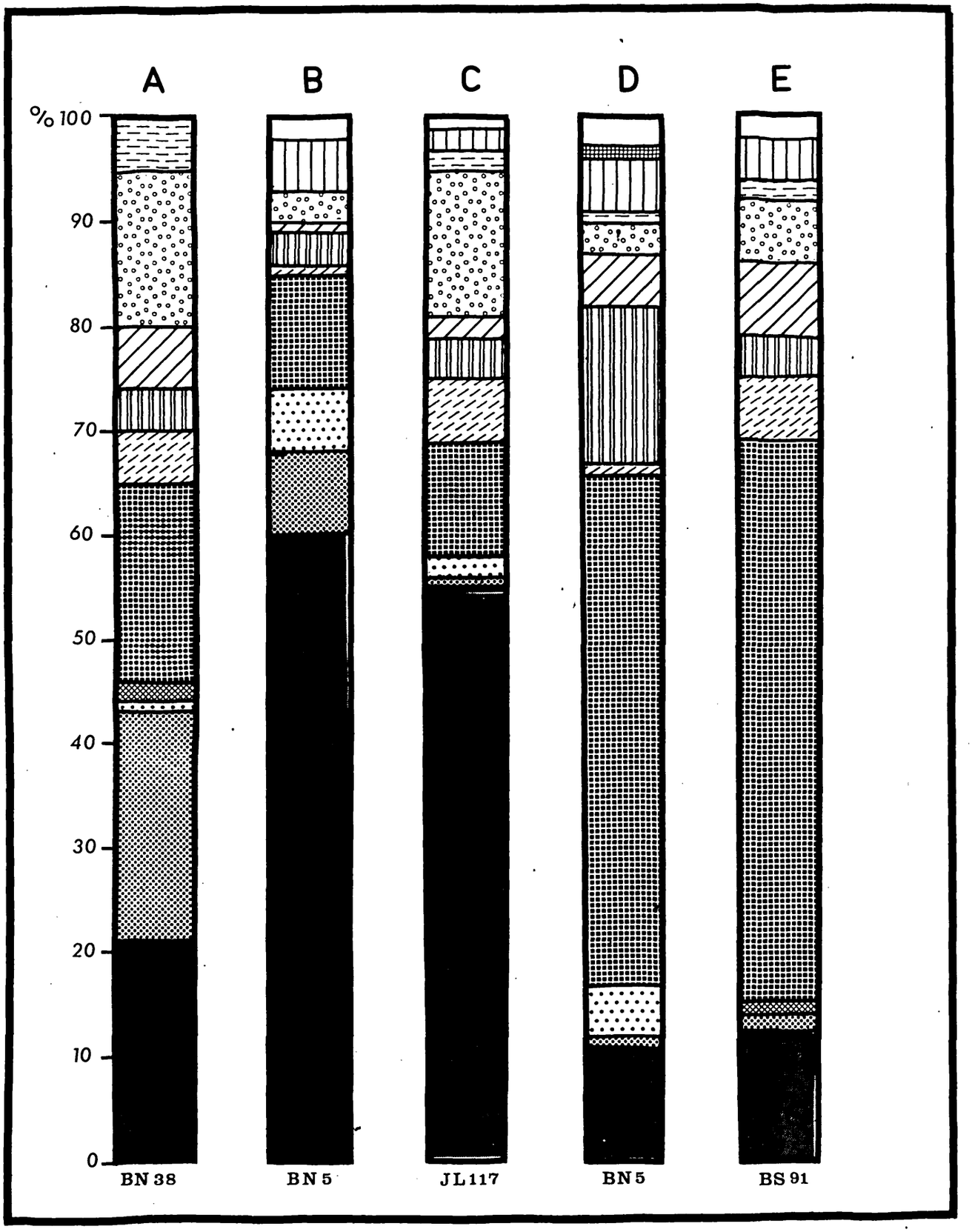
where P(X) = Percentage of a particular mineral in the sample.

X = Number of grains of the specific mineral in a particular size grade.

S = The size grade.

N = Total number of grains of heavy minerals counted in that size grade.

The percentages of the non-opaque heavy minerals in each size grade of each sample and of the whole sample were also calculated using the above formula, after subtracting the total number of opaque minerals counted in each sample from the total number of heavy minerals in that sample.



Columnar diagrams showing the mineralogical composition of one particular grain size in three samples (A,B,C) and the dominant grades (D,E). For legend see fig.4.1

4.2.2.6 X-ray analysis

X-ray analysis was carried out on each of the 39 samples used for the Bima Sandstones heavy mineral study. The aim of the X-ray studies was qualitative, to aid the identification of the minerals under the petrographic microscope. Samples were exposed to the Cobalt K radiation at 1^{o}/minute.

The identification of the heavy minerals present in each sample was carried out on the diffractograms after the conversion of the angles in θ values to d-spacings.

The samples used for the X-ray analysis were ground up to a very fine powder. A small amount of the powder was put on a square piece of glass and made into an even paste with acetone on the glass slide and allowed to dry. Samples that were too small to be handled in the manner described above were treated for X-ray analysis in a different manner using a photographic film.

4.3 THE OPAQUE MINERALS

4.3.1 Introduction

The mineralogical composition of the opaque fraction of the Bima Sandstones samples analysed in this study is generally very simple. The opaque minerals are: limonite, leucoxene, ilmenite, and magnetite. Limonite is the predominant opaque mineral in all the samples, although leucoxene or ilmenite occasionally exceed limonite in percentage occurrence. Limonite, therefore, has a great influence on the grain size distribution of the opaque fraction of the Bima samples. Limonite occurs in higher concentrations in the coarser size grades of the samples analysed.

In almost all the samples analysed in this study the opaque minerals form more than 70% of the heavy minerals in the coarsest size grades. There is only one isolated case in which the total non-opaque minerals exceed the total opaque minerals in the coarsest fraction.

LIMONITE

Limonite occurs as irregular grains, powdery aggregates or as coating on other minerals, mainly quartz grains. It has brown to a brownish-yellow colour in reflected light. In some instances in all the samples the limonite grains are sub-rounded to well rounded according to Powers' Roundness scale (1953). This scale is shown in fig. 5.2. In other instances the grains are dull, polished, with flattened shapes. The limonite grains are sometimes granular being an aggregate of smaller "sandy" grains. Sometimes the grains are botryoidal in shape and form. On the whole the polished smooth grains predominate the different varieties of limonite present in the Bima Sandstones samples in the area of study.

The percentage occurrence of limonite in the samples varies from 8 to 67 in samples Yh 45 and BN 23 respectively, with an average of 32.6 percent in all 39 samples. The amount of limonite in any sample increases with the size grade and it is greatest in the 210 phi size grade in any of the samples with a few exceptions. This behaviour of limonite influences the grain size distribution of the opaque minerals in general which results in a positive skewness for the opaque fraction of the Bima Sandstones in the area of study. In all the samples analysed limonite alone constitutes more than half the total amount of the opaque minerals.

TABLE 4.2: Mineral percentages of the non-opaque fraction

Sample No.	Zircon	Tourmaline	Amphibole	Sphene	Epidote	Rutile	Apatite	Barytes	Brookite	Garnet
BN 31	51	5	-	-	10	17	-	-	-	-
BN 38	64	4	3	12	6	11	-	-	-	-
BN 19	51	8	7	17	9	9	-	-	-	-
BN 20	54	3	3	24	17	-	-	-	-	-
BN 23	48	4	2	27	11	8	-	-	-	-
BN 25	44	7	2	16	7	6	-	-	18	-
BN 6	43	24	9	6	8	10	-	-	-	-
BN 3	42	9	17	17	12	3	-	-	-	-
BN 4	27	31	5	5	8	24	2	-	-	-
BN 5	59	2	2	9	11	6	11	-	-	-
BN 10	33	17	-	4	8	3	32	-	3	-
BN 11	48	8	1	12	3	7	12	-	9	-
BN 12	55	6	3	6	13	4	13	-	-	-
BN 14	27	19	4	9	13	9	18	-	-	-
BN 15	59	6	-	18	6	6	-	-	6	-
BN 16	46	11	2	27	-	15	-	-	-	-
YH 45	42	13	6	16	13	3	6	-	-	-
JN 99	48	11	4	21	11	5	-	-	-	-
JE 101	62	9	1	10	7	10	-	-	-	-
JL 106	19	37	7	19	-	19	-	-	-	-
JL 117	57	7	2	1	13	11	7	-	-	-
JL 120	65	9	3	3	9	8	2	-	-	-
JN 96	8	1	1	-	4	6	77	-	1	-
AN 50	68	5	-	8	18	1	-	-	-	-
AN 51	39	21	15	5	11	9	-	-	-	-
AN 55	71	13	3	2	9	3	-	-	-	-
JE 58	6	1	-	1	3	2	3	81	1	-
JE 71	36	12	4	3	6	5	10	-	-	24
JE 74	20	5	1	-	2	1	68	-	-	2
BN 76	47	15	5	4	13	7	-	-	13	-
BN 77	49	15	8	11	7	10	-	-	-	-
GH 122	43	17	4	3	17	5	-	-	11	-
D 125	51	6	1	2	8	12	-	-	20	-
BS 80	56	10	3	10	10	7	1	-	3	-
BS 81	25	20	15	20	20	-	-	-	-	-
BN 86	45	17	8	3	5	11	2	-	10	-
BS 87	57	6	1	6	15	6	-	-	2	6
BS 91	48	8	4	14	7	11	8	-	-	-
BS 93	18	8	2	-	1	3	68	-	-	-

LEUCOXENE

Leucoxene occurs as white or yellowish white rounded grains with an "unglazed" porcelain appearance. An occasional fragmented grain of leucoxene with a core of ilmenite or limonite could be seen in some of the samples. Sometimes grains could be seen which are predominantly white with patches of brown or yellowish-brown colour or vice versa. These grains were counted among grains of the predominant colour showing on the "variegated" grain. Sometimes the grains could be seen to consist of aggregates of grains of leucoxene and ilmenite, and/or limonite. Leucoxene comes second to limonite in percentage occurrence among the opaque minerals of the Bima Sandstones in the area of study. It ranges from 0 to 42 percent and averages 7.1 percent in 32 of the samples in the whole sample analysis. Leucoxene is absent from 7 of the 39 samples. Leucoxene behaves almost like limonite in being generally concentrated in the coarsest size grades. Its percentages occurrence generally increases towards the coarsest size grades in all samples with a few exceptions.

Polished smooth surfaced grains of leucoxene are abundant in all the samples studied and only a few number of dull grains were encountered.

ILMENITE

Ilmenite occurs as black or crimson-grey grains or as aggregate grains. Most of the grains of ilmenite observed in the samples are flattened and are mostly non glossy in texture being earthy and granular. The grains are nearly always angular in shape and they sometimes show a tendency to alter into leucoxene. In such grains whitish areas of leucoxene are interspersed within the black grain of ilmenite. Ilmenite is third to limonite and leucoxene in abundance of occurrence. In some

TABLE 4.3: Mineral percentages of heavy fraction in the 0.210 mm size fraction

Sample No.		Limonite	Leucoxene	Magnetite	Ilmenite	Alterite	Zircon	Tourmaline	Epidote	Sphene	Rutile	Amphibole	Brookite	Barytes	Apatite	Garnet
BN	31	77	7	-	13	-	3	-	-	-	-	-	-	-	-	-
BN	38	29	17	-	2	-	20	10	5	5	5	7	-	-	-	-
BN	19	41	5	27	-	5	14	9	-	-	-	-	-	-	-	-
BN	20	63	27	-	-	3	7	-	-	-	-	1	-	-	-	-
BN	23	38	51	-	2	-	5	-	-	2	2	-	-	-	-	-
BN	25	94	4	-	-	-	-	2	-	-	-	-	-	-	-	-
BN	26	74	17	-	-	-	6	3	-	-	-	-	-	-	-	-
BN	3	12	-	2	4	4	44	7	9	13	-	4	-	-	-	-
BN	4	41	44	12	2	-	-	-	-	-	-	1	-	-	-	-
BN	5	77	7	-	3	-	7	-	3	3	-	-	-	-	-	-
BN	10	91	9	-	-	-	-	-	-	-	-	-	-	-	-	-
BN	11	56	8	-	36	-	-	-	-	-	-	-	-	-	-	-
BN	12	69	26	-	3	-	2	-	-	-	-	-	-	-	-	-
BN	14	90	-	2	6	-	2	-	-	-	-	-	-	-	-	-
BN	15	75	25	-	-	-	-	-	-	-	-	-	-	-	-	-
BN	16	87	11	-	2	-	-	-	-	-	-	-	-	-	-	-
YH	45	38	3	-	31	10	5	-	-	-	5	8	-	-	-	-
JN	99	65	6	15	2	-	12	-	-	-	-	-	-	-	-	-
JE	101	77	7	12	-	-	2	-	-	-	2	-	-	-	-	-
JL	106	45	-	18	23	-	5	-	-	-	9	-	-	-	-	-
JL	117	70	-	-	10	-	8	1	-	-	8	-	-	-	2	-
JL	120	31	48	-	-	3	9	6	-	-	3	-	-	-	-	-
JN	96	50	-	8	-	-	-	-	-	-	-	-	-	-	42	-
AN	50	85	-	-	-	-	7	7	-	-	-	-	-	-	-	-
AN	51	80	4	7	1	3	1	4	-	-	-	-	-	-	-	-
AN	55	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
JE	58	15	-	-	-	-	5	3	-	-	1	-	2	64	-	-
JE	171	65	-	-	4	-	-	4	-	-	-	-	-	-	26	-
BN	76	73	-	-	-	-	7	10	3	-	7	-	-	-	-	-
BN	77	86	-	-	3	-	7	-	3	-	-	-	-	-	-	-
GH	122	93	-	-	-	-	3	-	-	-	-	3	-	-	-	-
D	125	74	-	-	-	-	15	2	1	-	-	1	7	-	-	-
BS	80	92	-	-	-	-	4	4	-	-	-	-	-	-	-	-
BS	81	65	-	20	12	-	3	-	-	-	-	-	-	-	-	-
BN	86	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
BS	87	84	2	-	-	1	3	2	-	-	-	-	-	-	-	7
BS	91	54	18	-	4	-	16	-	-	-	7	-	-	-	-	-
BS	93	80	-	-	2	-	7	4	2	-	-	-	-	-	4	-
JE	74	76	-	-	-	-	5	2	-	-	-	-	-	-	13	3

samples it is more than either limonite or leucoxene while in other instances it is more than both limonite and leucoxene combined. In some samples ilmenite is totally absent from the whole sample while in others it is present only in the coarsest size grades. Ilmenite therefore ranges in percent occurrence in the whole sample analysis from 0 to 26 percent with an average of 4.9 percent in 30 samples, and totally absent from 9 samples.

MAGNETITE

Magnetite occurs in the Bima Sandstones samples in the area of study as silver-grey coloured grains. The grains show a hackly fracture where it was possible to observe this feature, and some grains show the presence of minute facets. The grains are angular with some showing alteration to limonite. The percentage occurrence of magnetite averages 4.9 and the range is 0 to 21 percent.

There is no particular sequence of occurrence observed for magnetite. Like limonite and leucoxene, which show a marked increase in concentration in the coarsest size grades, magnetite tends to be more concentrated in the coarsest fractions although the tendency is not as marked as it is in the case of limonite and leucoxene.

4.4 THE NON-OPAQUE MINERALS

4.4.1 Introduction

The mineralogical composition of any terrigenous sediment is largely a function of provenance in contrast to the textural peculiarities of that sediment which are almost entirely controlled by history of transportation and the processes in the depositional environment (Van Andel, 1960).

TABLE 4.4: Mineral percentages of heavy fraction in the 0.150 mm size fraction

Sample No.		Limonite	Leucoxene	Magnetite	Ilmenite	Alterite	Zircon	Tourmaline	Epidote	Sphene	Rutile	Amphibole	Brookite	Barytes	Apatite	Garnet
BN	31	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
BN	38	21	22	-	1	2	19	5	4	6	15	5	-	-	-	-
BN	19	18	17	38	-	3	8	2	2	4	7	2	-	-	-	-
BN	20	42	30	-	-	3	14	-	4	7	-	1	-	-	-	-
BN	23	22	51	2	-	-	9	5	-	1	8	2	-	-	-	-
BN	25	73	3	-	-	-	11	5	3	3	2	-	-	-	-	-
BN	6	85	9	-	1	-	3	2	1	-	-	-	-	-	-	-
BN	13	61	-	10	8	4	4	6	-	2	2	2	-	-	-	-
BN	4	35	46	-	5	2	1	9	-	-	-	2	-	-	-	-
BN	5	60	8	-	6	-	11	1	3	1	3	-	-	-	5	1
BN	10	70	1	1	4	2	16	-	-	-	2	-	2	-	1	-
BN	11	35	14	2	40	-	2	-	-	3	1	2	-	-	2	-
BN	60	32	-	1	3	3	-	-	-	-	1	1	-	-	-	-
BN	14	74	4	1	11	-	2	4	-	-	1	2	-	-	2	-
BN	15	53	40	--	-	-	-	2	2	-	2	-	-	-	-	-
BN	16	78	12	2	1	-	3	-	-	2	1	-	-	-	-	-
YN	45	24	2	5	32	4	19	-	-	-	1	7	-	-	6	-
JN	99	50	6	20	2	-	9	-	5	6	-	2	-	-	-	-
JE	101	41	19	5	10	-	14	4	1	-	5	1	-	-	-	-
JL	106	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
JL	117	55	1	-	2	-	11	6	4	2	14	2	-	-	2	-
JL	120	38	38	-	2	5	6	5	1	-	5	1	-	-	-	-
JN	96	33	-	11	-	-	2	-	-	-	-	-	-	-	54	-
AN	50	84	-	-	-	-	9	6	1	-	-	-	-	-	-	-
AN	51	66	2	12	6	-	3	4	2	-	2	3	-	-	-	-
AN	55	82	-	-	3	-	3	9	3	-	-	-	-	-	-	-
JE	58	21	-	-	-	-	3	1	-	-	3	-	-	71	-	-
JE	71	59	11	-	4	1	2	6	-	-	1	2	-	-	5	8
BN	76	83	8	2	-	-	1	5	-	-	-	-	1	-	-	-
BN	77	69	14	-	1	-	9	2	2	2	-	-	-	-	-	-
GH	122	71	-	1	-	-	11	9	3	-	3	2	-	-	-	-
D	125	69	-	-	-	-	18	5	1	-	4	1	3	-	-	-
BS	80	68	-	-	2	1	7	11	2	1	2	3	3	-	-	-
BS	81	58	-	20	8	-	5	2	-	4	-	-	4	-	-	-
BN	86	38	-	-	-	2	17	22	1	-	4	10	5	-	-	-
BS	87	69	5	1	-	-	6	4	2	-	1	1	1	-	-	10
BS	91	55	12	-	1	1	6	6	1	3	8	4	-	-	2	-
BS	93	48	-	-	-	-	7	5	1	-	1	-	-	-	38	-
JE	74	38	-	-	-	-	5	5	1	-	1	-	-	-	49	-

Physical and chemical sedimentary processes such as selective weathering of unstable minerals, selective sorting according to size or density, mechanical wear, and diagenetic processes could change the composition between the source area and the site of deposition.

The non-opaque minerals of the Bima Sandstones contrast with their opaque counterparts in being more highly concentrated in the finer size grades. The original grain sizes of the heavy minerals of the source rocks help to determine in which size grade the heavies will occur. This fact is also aided by the hydraulic conditions prevalent in the transporting media and the physical properties of the mineral grains themselves in determining the relative abundance of the heavy minerals in each size grade. Briggs (1965) stated that the heavy mineral content of a fluvial sandstone deposited under many different stream regimes is that selected from the entire store of mineral types and sizes by the hydraulic properties of the stream peculiar to the transportation and deposition of that sandstone. He further amplified Colby's (1963) idea that in highly turbulent streams nearly all the heavy minerals in the bed-load are transported and that sediments deposited from such streams contain a representative suite of the heavy minerals in the source material.

In the Bima Sandstones heavy mineral analysis the predominant minerals making up the non-opaque mineral suite are; Zircon, Tourmaline, Amphibole, Sphene, Epidote, Rutile, Apatite, and Brookite. Others occurring in very small quantities in some of the samples are barytes and garnet. Both brookite and barytes appear to be authigenic in origin as a result of their perfect or near perfect euhedral shapes, the lack of rounding of the grains, and if their position in the mineral hardness scale is considered. Individual minerals will now be discussed in some detail.

TABLE 4.5: Mineral percentages of heavy fraction in the 0.150 mm size fraction

Sample No.		Limonite	Leucoxene	Magnetite	Ilmenite	Alterite	Zircon	Tourmaline	Epidote	Sphene	Rutile	Amphibole	Brookite	Barytes	Apatite	Garnet
BN	31	75	3	-	17	-	1	1	1	1	1	-	-	-	-	-
BN	38	13	3	-	1	3	57	2	4	10	6	1	-	-	-	-
BN	19	20	13	22	1	4	24	4	3	5	2	3	-	-	-	-
BN	23	11	38	2	1	-	25	2	4	5	10	2	-	-	-	-
BN	25	45	5	-	4	4	18	5	4	5	4	2	4	-	-	-
BN	6	61	3	-	3	2	8	12	2	-	3	6	-	-	-	-
BN	3	23	-	8	2	14	10	10	4	9	-	19	-	-	-	-
BN	4	35	48	-	-	1	5	8	1	-	1	1	-	-	-	-
BN	5	15	5	-	2	4	48	2	6	6	4	-	-	-	7	2
BN	10	74	7	-	4	-	11	-	-	3	1	-	-	-	-	-
BN	11	37	1	-	48	33	1	1	-	-	1	-	-	5	-	-
BN	12	38	32	-	-	-	8	3	-	3	3	1	-	-	13	-
BN	14	71	1	-	3	5	3	5	1	-	1	1	-	-	9	-
BN	15	36	28	8	13	-	9	-	-	6	-	-	-	-	-	-
BN	16	75	12	-	1	2	3	2	-	3	2	-	-	-	-	-
YN	45	6	-	1	9	18	26	8	11	10	1	5	-	-	5	-
JN	99	28	-	3	-	-	43	2	6	16	-	2	-	-	-	-
JE	101	20	16	1	-	3	42	5	3	3	6	-	-	-	-	-
JL	106	52	-	8	11	-	6	4	2	-	-	-	-	-	-	-
JL	117	37	-	-	1	2	33	8	5	-	7	-	-	-	7	-
JL	120	23	14	-	-	2	38	10	3	1	7	2	-	-	-	-
JN	96	23	-	-	3	3	1	1	-	-	3	-	-	-	66	-
AN	50	29	-	-	-	-	51	5	12	2	1	-	-	-	-	-
AN	51	50	-	9	3	1	14	5	3	3	3	4	-	-	-	-
AN	55	67	1	-	5	-	17	6	4	-	-	-	-	-	-	-
JE	58	22	-	-	-	1	6	1	1	1	2	-	-	65	2	-
JE	71	66	-	-	-	-	4	7	1	-	2	3	-	-	7	10
BN	76	57	-	-	-	1	23	6	4	-	5	2	2	-	-	-
BN	77	53	11	-	-	3	10	9	1	4	3	5	-	-	-	-
CH	122	61	-	-	-	-	18	7	5	2	3	4	2	-	-	-
D	125	25	-	-	-	-	32	8	8	-	16	-	11	-	-	-
BS	80	34	2	-	2	4	32	10	5	3	7	1	-	-	-	-
BS	81	48	-	25	4	-	8	4	4	4	-	3	-	-	-	-
BN	85	31	-	-	-	-	29	14	2	1	11	6	6	-	-	-
BS	87	74	1	-	-	-	11	6	1	1	3	1	-	-	-	2
BS	91	16	2	-	-	5	38	3	6	12	8	2	-	-	7	-
BS	93	15	1	-	-	5	16	6	1	-	2	1	-	-	54	-
JE	74	12	-	-	-	3	2	7	-	-	-	1	-	-	74	-

TABLE. 4.6: Mineral percentages of heavy fraction in the 0.0625 mm S.F.

Sample No.		Limonite	Leucoxene	Magnetite	Ilmenite	Alterite	Zircon	Tourmaline	Epidote	Sphene	Rutile	Amphibole	Brookite	Barytes	Apatite	Garnet
BN	31	31	4	-	17	3	28	1	3	6	7	-	-	-	-	-
BN	38	79	7	2	12	-	-	-	-	-	-	-	-	-	-	-
BN	19	18	7	12	4	5	31	5	5	8	2	2	-	-	-	-
BN	20	26	-	2	-	1	44	1	8	18	-	2	-	-	-	-
BN	23	2	5	-	-	1	50	2	6	30	4	1	-	-	-	-
BN	25	15	2	-	-	6	41	2	4	11	3	1	16	-	-	-
BN	6	43	-	-	-	2	32	8	3	4	5	3	-	-	-	-
BN	3	7	1	2	3	7	40	17	7	14	2	-	-	-	-	-
BN	4	27	47	-	1	1	8	5	2	1	7	-	-	-	-	-
BN	5	11	1	-	5	-	49	1	15	8	3	1	-	-	5	1
BN	10	56	1	-	2	1	16	4	2	1	-	-	-	-	16	-
BN	11	14	3	-	20	2	37	4	2	5	4	-	2	-	6	-
BN	12	12	9	-	-	3	60	3	4	2	2	2	-	-	3	-
BN	14	35	1	-	7	3	25	11	3	2	6	2	-	-	7	-
BN	15	16	16	3	13	6	32	-	-	10	-	-	3	-	-	-
BN	16	28	10	1	-	2	35	3	5	10	3	5	1	-	-	-
YH	45	3	-	2	-	7	61	7	3	9	4	2	-	-	2	-
JN	99	15	-	4	-	-	44	9	10	12	3	3	-	-	-	-
JE	101	15	8	3	4	7	43	6	5	5	6	-	-	-	-	-
JL	106	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
JL	117	9	1	-	-	1	66	2	12	-	4	-	-	-	5	-
JL	120	7	7	1	3	3	60	3	7	2	3	2	-	-	1	-
JN	96	12	-	-	1	-	9	-	2	-	3	1	1	-	73	-
AN	50	6	-	-	-	3	69	2	14	5	1	-	-	-	-	-
AN	51	35	-	-	1	5	28	4	-	20	-	3	-	-	-	-
AN	55	32	-	-	3	3	48	7	4	-	1	1	-	-	-	-
JE	58	9	2	-	-	2	10	1	4	-	1	-	-	67	4	-
JE	71	40	-	-	-	-	29	8	3	2	2	1	-	-	4	11
JE	74	15	-	-	-	-	33	2	4	-	1	-	-	-	45	-
BN	76	66	-	-	-	-	20	3	4	-	-	-	7	-	-	-
BN	77	28	5	-	1	6	32	8	4	6	7	4	-	-	-	-
GH	122	43	1	-	-	1	34	4	10	1	1	1	5	-	-	-
D	125	17	-	-	-	1	49	2	7	1	2	-	21	-	-	-
BS	80	22	-	-	1	5	54	1	7	8	2	-	-	-	-	-
BS	81	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
BN	86	9	-	-	-	1	56	4	7	6	6	1	12	-	-	-
BS	87	23	-	-	-	3	52	2	11	3	4	-	1	-	-	-
BS	91	12	2	-	-	1	54	6	4	7	7	2	-	-	4	-

ZIRCON

The mineral zircon is ubiquitous in all specimens containing material of average grain size less than 0.210 mm. It usually reaches its highest frequency in the finest size grades, and in almost all instances the frequency declines rapidly in the coarser size grades. The mineral may be completely absent in a number of samples in the coarsest size grade (tables 4.3, 4.4, 4.5 and 4.6). Zircon is the most abundant heavy mineral among the non-opaque minerals of the Bima sandstones in the area of study (table 4.2). It makes up to as much as 71 percent of the non-opaque minerals in certain samples. It is lowest in value in samples where the frequency of such minerals as apatite, brookite or barytes are exceedingly high (tables 4.3, 4.4, 4.5 and 4.6).

Colourless zircon is the commonest variety. However, both pink and reddish varieties are occasionally found in some samples but these are only of minor importance when compared to the colourless varieties. The colourless variety is referred to by Hutton (1950) and Anderson (1941) as "normal" zircon or "high" zircon and the coloured varieties as hyacinth (Hutton 1950, p.686; of Anderson, 1941, and Smith 1940).

The forms and habit of the zircon grains are very varied in most samples. There is every gradation from sharply euhedral crystals through slightly abraded crystals to well-worn anhedral forms devoid of any crystals faces. In addition to these variety of forms are variously shaped fragments, flakes and splinters exhibiting conchoidal fracture faces. The latter shapes may have resulted from the crushing of the rocks during sample preparation in the laboratory. However, that some of the flakes and splinters are inherent in the rocks and not products of the crushing mechanism used could be seen in the smoothness of the faces of

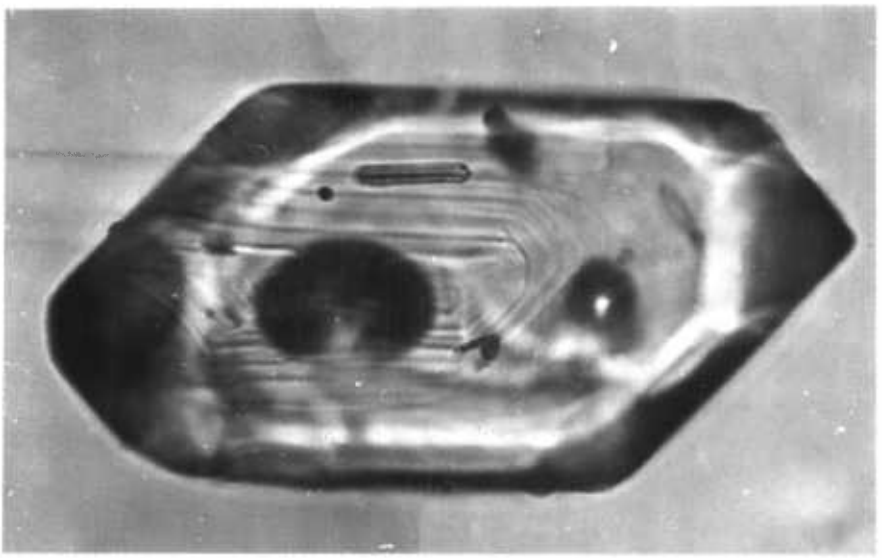


Plate 4.I.I. Euhedral crystal of zircon with well developed pyramidal faces terminating in a small basal plane at one end only. Note the large globular inclusions, the inclusions of small crystals of zircon, and the partial zoning in the center of the crystal. XIOO.

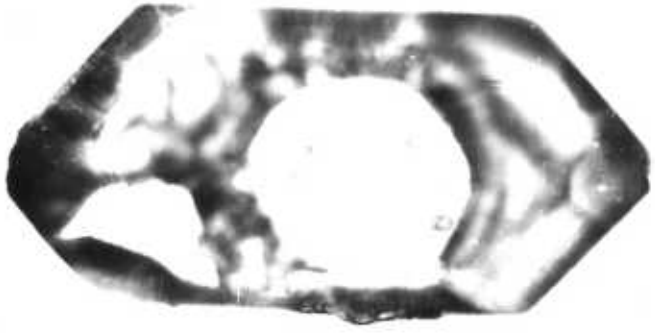


Plate 4.I.2. A euhedral crystal of zircon with well developed pyramidal faces terminated by small basal planes at both ends. XIOO.

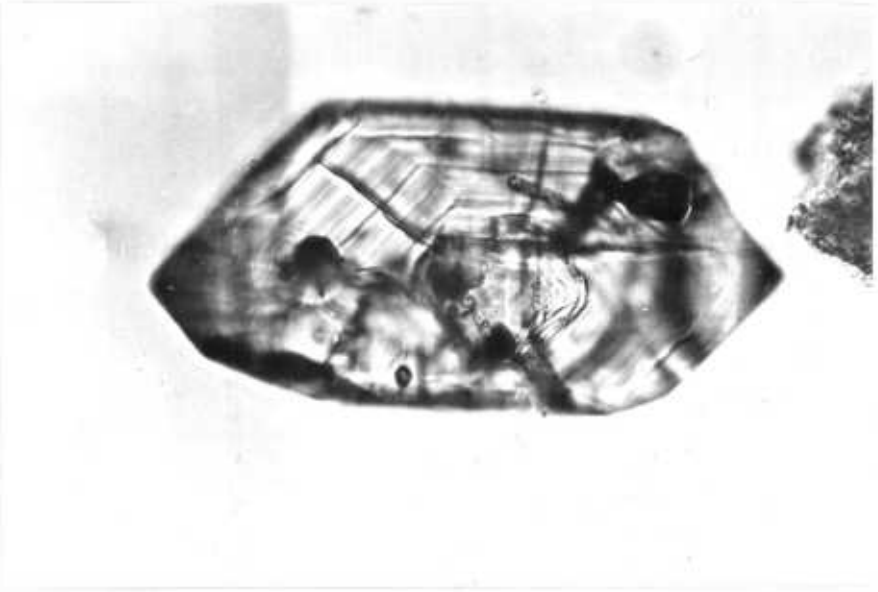


Plate 4.2.I. Stumpy euhedral crystal of zircon with well developed pyramidal faces with no basal planes. Length to breadth ratio, approx. 2:I. XIOO.

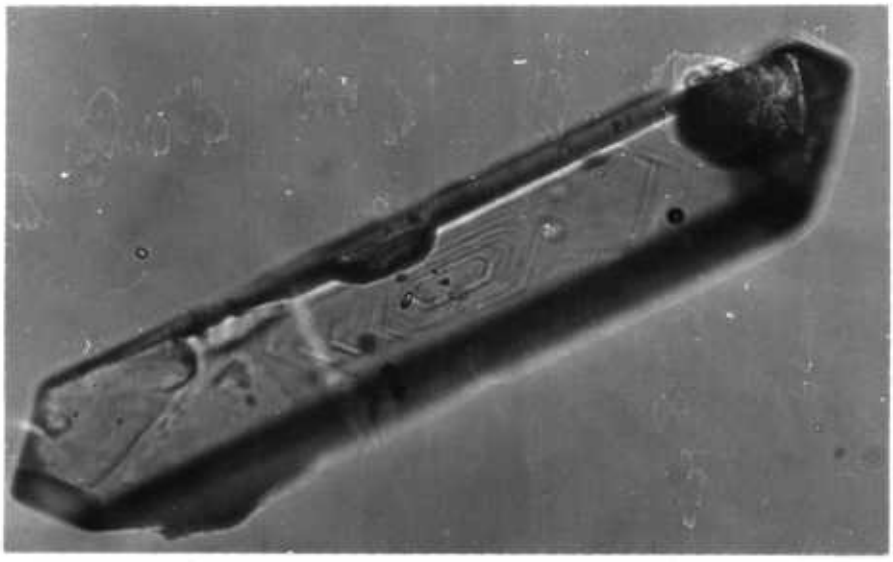


Plate 4.2.2. Slender euhedral crystal of zircon terminating in pyramidal faces (note steep 33I pyramidal faces) and showing partial zoning in the central portion and grooves on the crystal surface. Length to breadth ratio, approx. 3:I. X50.

fractured crystals. Generally speaking, a higher percentage of the unworn euhdral crystals are concentrated in the finer size grades, although no specific relationship was observed between grain size and degree of rounding. Rounded grains as well as perfect euhdral crystals are to be found in all size grades of all samples.

The most commonly occurring habit in the zircon crystals are grains with prisms terminated by pyramidal faces. The [100] and [110] prism faces and [111] pyramid faces are the commonest developed in the euhedral crystal forms. In some cases more than one pyramid face is developed. As was noted by Hutton (1950) in New Zealand sands he studied, distorted crystals with well developed [100] and poorly developed [110] prism faces may be terminated by large first order or second order pyramids at one end, and at the other end of the crystal the pyramid faces may be developed in a reversed order. Sometimes small basal planes are present (plates 4.1.1 and 4.1.2). Euhdral crystals of zircon with well developed steep [331] pyramid faces are common in the finer size grades of all samples, although their occurrence is by no means restricted to these size grades. The ratio length to breadth of the zircons is approximately 2:1, with some very slender crystals giving values of 3:1 (plates 4.2.1 and 4.2.2). In the determination of the length to breadth ratios only complete crystals were measured in all samples.

Inclusions abound in the zircons of the Bima Sandstones studied (plate 4.3.1 and 4.3.2). The inclusions are varied and do not occur in all the crystals. The types most abundant in the samples are:

(a) Euhedral crystals of zircon.

(b) Spherical and variously shaped cavities which are probably gas and/or fluid filled.

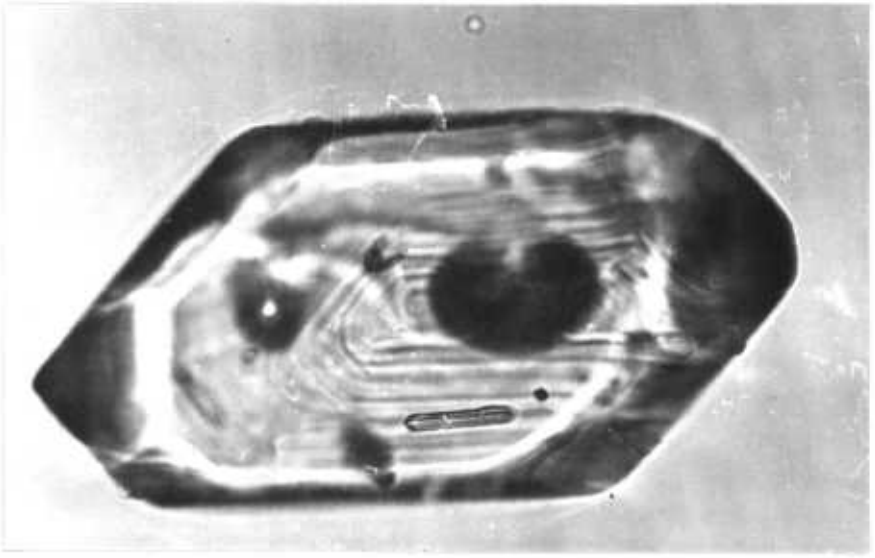


Plate 4.3.I. A stumpy crystal of zircon with different types of inclusions: spherical globules, and a small slender crystal of zircon. XIOO.

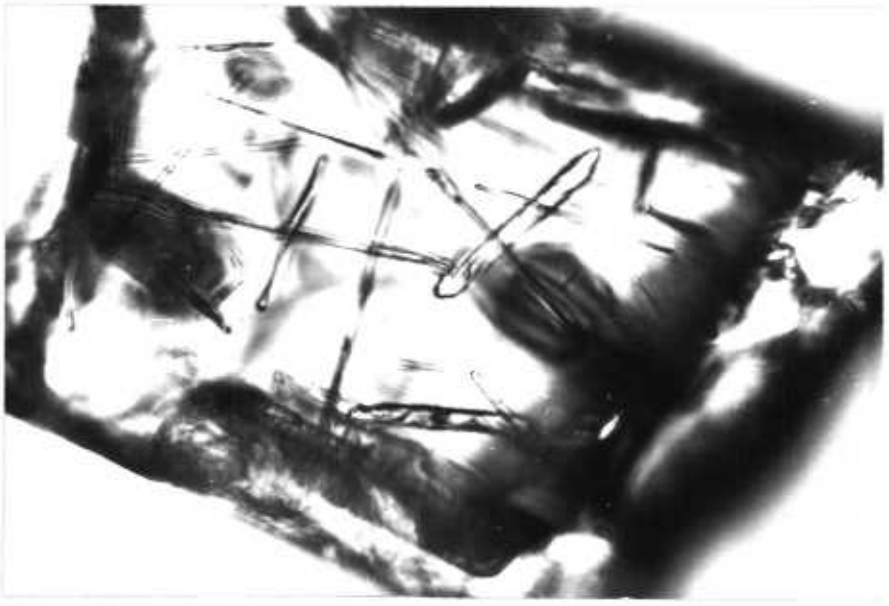


Plate 4.3.2. A fragmented crystal of zircon showing inclusions of slender needles of brown rutile and other minerals. XIOO.

(c) Slender, acicular, colourless crystals.

(d) Brown crystals of rutile.

(e) Other mineral grains.

(f) Clouds or patches of dust-like material.

Most crystals contain only one or sometimes two types of these inclusions. The most singly occurring inclusions in the zircon crystals are either cavities or slender colour crystals. These two are also the commonest combination of inclusions occurring together in one crystal. The small zircon crystals and rutile also occurs singly in some zircon crystals. A few crystals were found to contain many different types of inclusions.

There is no particular preferred orientation of these inclusions observed in the zircon crystals. However, the acicular crystals are sometimes oriented parallel to the vertical crystallographic axis of the parent zircon crystal. The presumably gas or fluid filled cavities may also be oriented in this manner in many crystals. Hutton (1950) listed 10 different types of inclusions occurring in the zircons of the New Zealand sands. His identification of the inclusion types was confirmed by laboratory tests, he carried out on their extracts. Use of warm acids or clerici solution failed to yield good extract results as both zircon and the inclusions behaved similarly in these liquids.

Zoning was observed in some of the zircons (plate 4.4). It consists of very fine thread-like concentric bands in the crystals. As was observed by Hutton (1950) the zoning is only very prominently developed along the periphery of the crystals in some of the zircons, and sometimes only in their central portions. However, there are zircons in which

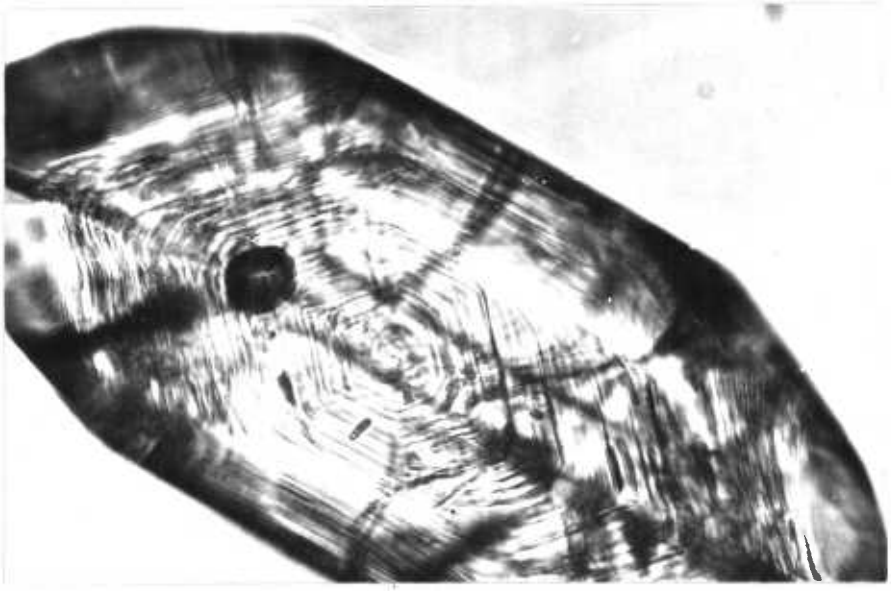


Plate 4.4. A large crystal of zircon showing well developed zoning consisting of very fine thread-like concentric bands all over the crystal from the periphery to the center. X50.

zoning is well developed all over the crystals.

Another prominent feature common in the zircons of the Bima Sandstones is the development of overgrowths. This phenomenon was also observed in the New Zealand sands studied by Hutton (1950); Poldervaart (1955); Poldervaart and Ecklemann (1955). The development of overgrowths in zircon appears to be similar to that in detrital light mineral grains. In grains where it is well developed the primary grain which has been smoothened up by abrasion has the crystal prism and pyramidal faces restored (plate 4.5).

4.5 ZIRCON ELONGATION STUDIES

Many authors have carried out elongation studies on zircon. Poldervaart (1956) summarised the voluminous literature on zircon in igneous rocks. Smithson (1939) and Wyatt (1953) used the method of histograms and frequency curves to represent quantitative elongation data on zircon, Larsen and Poldervaart (1957); Alper and Poldervaart (1957); Hall and Ecklemann (1961); and Spotts (1962) have used the method of reduced major axes with results quite indicative of the fact that zircons from the same source area exhibit the same morphological and elongation characteristics.

In the Bima Sandstones studies, statistical work was carried out on 34 samples by measuring the length and breadth of a large number of zircon crystals in each sample. The method of reduced major axis has been employed to analyse the data obtained. The technique used was developed by Kermak and Haldane (1950) and used by Imbrie (1956) in his statistical studies on invertebrate fossils and has been applied in zircon elongation studies by Larsen and Poldervaart (1957), and Alper and



Plate 4.5. A euhedral form of zircon developed as an overgrowth material on a core of detrital zircon that has been abraided by erosion. Note also the smaller overgrowth material beginning to form on the surface of the crystal. X100.

TABLE 4.7: Statistical data for reduced major axes of Zircon Elongation in Bima Sandstones. Sample localities shown in fig 1.1.

Sample No.	Mean length X (mm)	Mean breath Y (mm)	Sx (mm)	Sy (mm)	Slope (a)	Slope (angles in degrees)
BN 3	0.0231	0.0094	0.0768	0.0313	0.4072	22
BN 4	0.0297	0.0109	0.1073	0.0395	0.3687	20
BN 5	0.0289	0.0130	0.1328	0.0451	0.3390	19
BN 6	0.0261	0.0107	0.0982	0.0401	0.4080	22
BN 10	0.0165	0.0064	0.0596	0.0192	0.3868	21
BN 11	0.0270	0.0109	0.1048	0.0425	0.4052	22
BN 12	0.0232	0.0108	0.1304	0.0766	0.5870	30
BN 14	0.0223	0.0098	0.1182	0.0522	0.4418	24
BN 16	0.0196	0.0098	0.1683	0.0843	0.5010	27
BN 19	0.0235	0.0103	0.0103	0.1197	0.0527	24
BN 20	0.0262	0.0122	0.1403	0.0743	0.5298	28
BN 23	0.0220	0.0105	0.1542	0.0724	0.4696	25
BN 25	0.0249	0.0112	0.1162	0.0526	0.4529	24
BN 31	0.0231	0.0100	0.0830	0.0364	0.4378	24
BN 38	0.0253	0.0112	0.2056	0.0912	0.4435	24
YH 45	0.229	0.0108	0.2272	0.1070	0.4508	24
AN 50	0.0275	0.0127	0.1827	0.0841	0.4604	25
AN 51	0.0231	0.0115	0.1533	0.0963	0.6276	32
AN 55	0.0217	0.0099	0.1150	0.0522	0.4540	24
JE 58	0.0258	0.0122	0.1128	0.0534	0.4733	25
JE 71	0.0227	0.0104	0.1360	0.0626	0.4601	25
JE 74	0.0237	0.0106	0.1590	0.0711	0.4472	24
BN 76	0.0314	0.0133	0.1135	0.0505	0.455	25
BN 77	0.0207	0.0097	0.1118	0.0525	0.4765	26
BS 80	0.0246	0.0115	0.1475	0.0690	0.4678	25
BS 86	0.0266	0.0110	0.0110	0.1988	0.0826	23
BS 87	0.0242	0.0108	0.1302	0.0584	0.4487	24
BS 91	0.0312	0.0119	0.2521	0.0714	0.2832	16
JN 99	0.0264	0.0130	0.1516	0.0747	0.4929	26
JE 101	0.0284	0.0127	0.2050	0.0919	0.4481	24
JL 117	0.0257	0.0120	0.2043	0.0953	0.4665	25
JL 120	0.0217	0.0095	0.1945	0.0856	0.4398	24
GH 122	0.0255	0.0120	0.1509	0.0710	0.4709	25
GH 125	0.0282	0.0125	0.0172	0.0763	0.445	24

Poldervaart (1957), in which they represented the number of crystals measured by N; mean length, $\bar{x}$; mean breadth, $\bar{y}$; standard deviation of length, Sx; standard deviation of breadth, Sy; and slope, a. They defined the statistical measures by the following equations;

Mean of x, $\bar{x} = \frac{(x)}{N}$ ——————— (1)

Mean of y, $\bar{y} = \frac{(y)}{N}$ ——————— (2)

These equations express the central values of the x and y observations.

Standard deviation of x, $Sx = \sqrt{\left\{\frac{(x-\bar{x})^2}{N}\right\}}$ (3)

Standard deviation of y, $Sy = \sqrt{\left\{\frac{(y-\bar{y})^2}{N}\right\}}$ (4)

Equations (3) and (4) are a measure of the absolute variation within the samples.

Slope, $a = \frac{Sy}{Sx}$ (5)

Equation (5) expresses the tangent of the angle at which the fitted line is drawn through the point $(\bar{x}, \bar{y})$ for each sample.

By the use of these statistical parameters; mean length and breadth, standard deviation of length and breadth, and slope of the reduced major axis (table 4.7), samples from a given area can be compared. Differences or similarities between samples may be evident upon visual comparison of the reduced major axis. Length and breadth measurements of euhdral zircon crystals were carried out in each sample with an ocular micrometer and regularly spaced traverses using an integrating mechanical stage. Measurements were made with a magnification of X 100, and up to 100 zircon euhedra were measured from each sample although the number measured varied according to the frequency of zircon euhdra in each sample (table 4.8).

TABLE 4.8: Elongation frequencies of euhedral zircon in Bima Sandstones. Frequencies given as percentages. Sample localities given in fig. 1.1

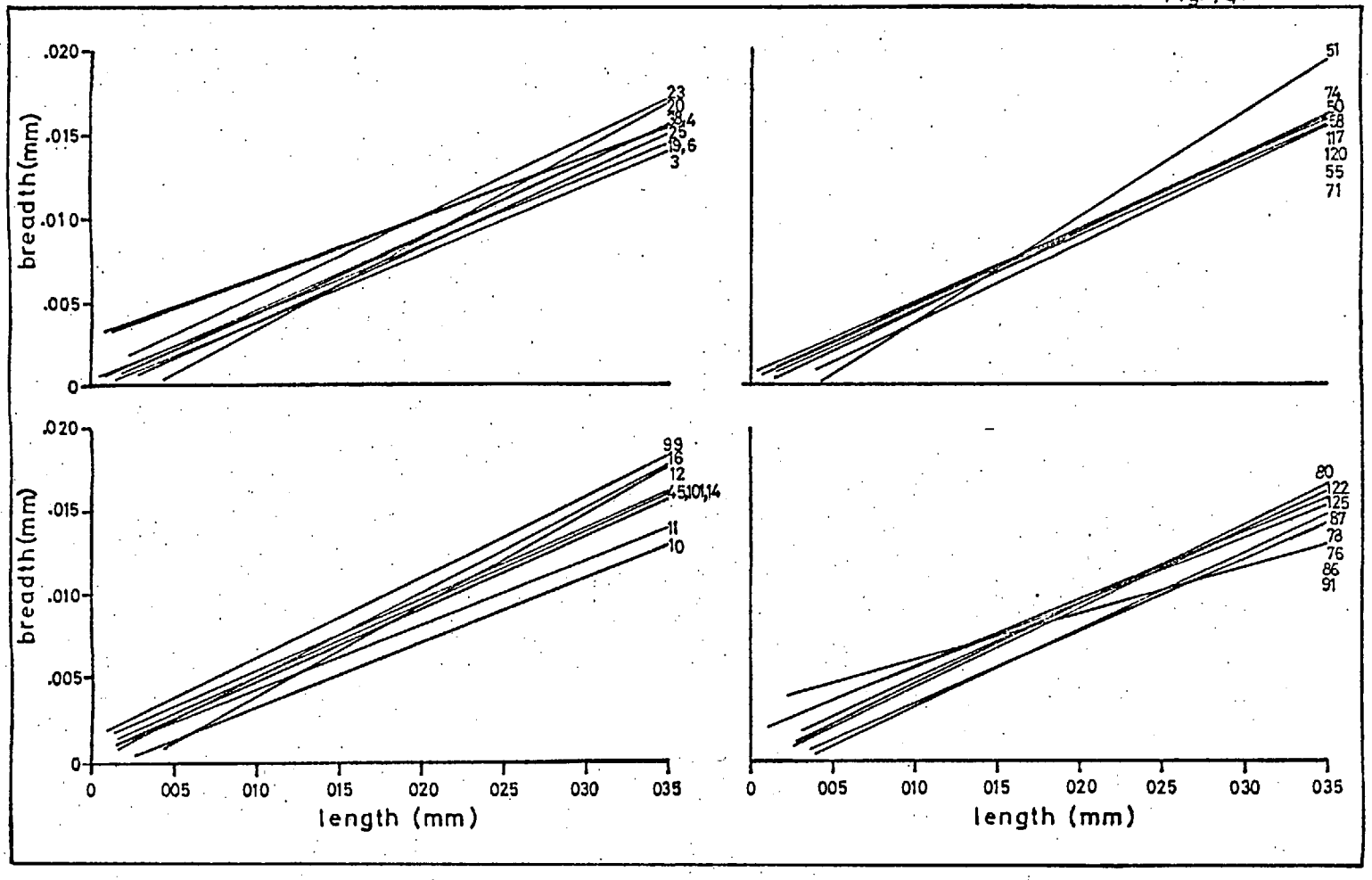
Sample No.	Length		Breath		Ratios		
	1.00-1.49	1.50-1.99	2.00-2.49	2.50-2.99	3.0-3.49	3.50-3.99	4.0 +
BN 3	-	23	15	46	8	8	-
BN 4	-	-	27	33	27	7	7
BN 5	-	30	39	13	17	-	-
BN 6	6	31	13	31	6	-	13
BN 10	-	9	37	9	18	9	18
BN 11	-	35	29	12	12	-	12
BN 12	6	25	42	25	2	-	-
BN 14	3	23	43	17	7	7	-
BN 16	5	43	41	5	3	-	-
BN 19	4	18	43	14	14	7	-
BN 20	-	35	49	11	3	3	-
BN 23	-	29	59	10	-	2	-
BN 25	-	25	58	13	4	-	-
BN 31	-	27	40	20	7	-	7
BN 38	4	19	44	19	12	2	-
YH 45	4	28	50	8	8	1	1
AN 50	4	28	43	15	9	-	-
AN 51	2	43	41	11	2	-	-
JE 58	3	23	43	20	7	3	-
JE 71	9	24	48	9	5	5	-
JE 74	3	26	50	11	11	-	-
BN 76	-	13	60	20	7	-	-
BN 77	-	35	45	13	6	-	-
BS 80	5	32	42	13	5	3	-
BS 86	-	14	40	27	14	3	2
BS 87	-	29	39	19	6	6	-
BS 91	-	11	34	37	16	3	-
JN 99	3	40	37	11	6	3	-
JE 101	11	19	30	28	7	-	6
JL 117	2	32	43	14	6	2	2
JL 120	1	19	52	15	7	3	3
GH 122	-	28	52	17	3	-	-
GH 125	-	23	49	20	5	3	-

The reduced major axis is regarded as a relative growth trend or pattern of crystal growth for a zircon population (Spotts, 1962). The axis passes through the plot point of mean length and breadth for a particular zircon crystal, and the slope represents the relative growth rate parallel to the C-axis (fig. 4.4). Spotts (1962) further states that a population of zircon crystals with fast growth parallel to the C-axis will have a larger standard deviation in crystal length than crystals with faster growth normal to the C-axis. The logical conclusion that can be drawn from the reduced major axis is that two zircon suites with crystals having similar mean lengths and breadths and similar standard deviation ratios have the same crystal growth pattern, and they probably enjoyed similar physiochemical conditions during their crystallization.

In the Bima Sandstones studies, tests of similarities in elongation characteristics were made by elongation frequency diagrams and the plots of reduced major axis were visually compared for similarities. The method of Spotts (1962) was used in the elongation studies. The length/breadth ratio was computed for each crystal measured and converted to frequencies for each sample. The elongation frequencies obtained were then plotted using a frequency polygon type of diagram (fig. 4.5).

Fig. 4.4 shows the relationship between the length and the breadth of zircon crystals in the Bima samples. The slope is defined by the ratio of the standard deviation of the two measurements. The plots indicate a general cluster of the axis based on the slope of the line and its position on the graph with respect to length and breadth. There are a few axes with a common slope and it is quite evident that they fall generally within a fairly narrow zone defined by the slopes. A scatter of the mean length - breadth points along the trend is a function of size or degree of

Fig. 4.4.

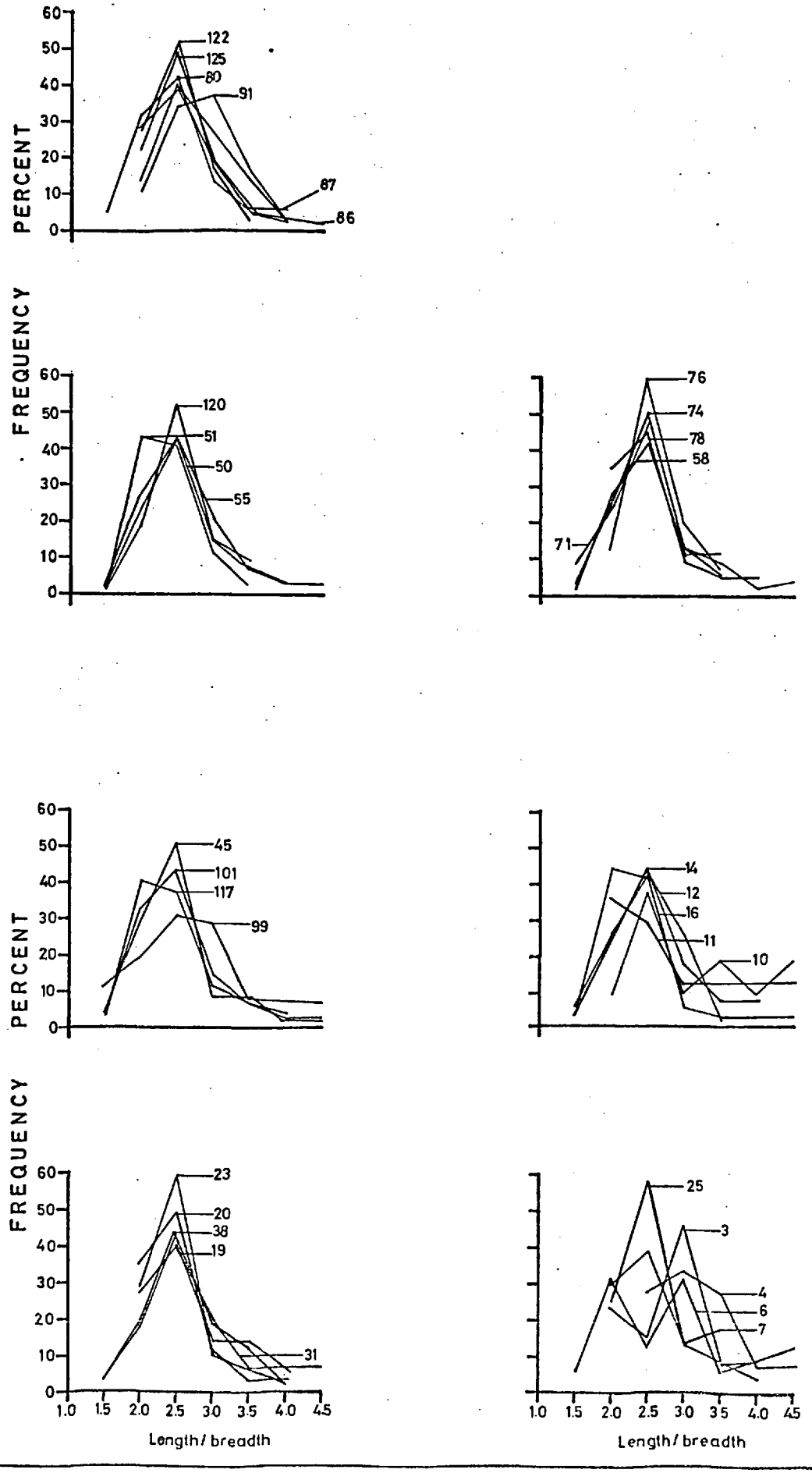


Reduced major axes for zircon elongation measurements.
(plot points represent mean length vs. mean breadth of zircon crystals for each sample; slope of line (reduced major axis) is ratio of standard deviation of breadth to standard deviation of length for zircon population of each sample).

development of a particular growth trend (Spotts 1962). Small deviation shown by some samples in the Bima Sandstones are attributed to the low number of measurements in those samples (Fig. 4.4). Thus there is no significant dispersion in the trends which could be related to differences in physiochemical conditions under which the zircons crystallized. It is therefore satisfactory to conclude here that the reduced major axes of the zircon crystals indicate a uniform physiochemical condition of crystallization and the same growth trends for the zircon crystals, and therefore, a common source area for the Bima Sandstones in the area of study.

Fig. 4.5 represents the elongation frequency curves for the same samples. Ten of the plots show broad peaks while a few others show several maxima. The rest have narrowly defined maxima. Poldervaart (1955) attributed the narrowly defined maxima of zircon elongation frequency curves to good sorting of the sediments, and broad maxima or several maxima to ill-sorting of the sediments. From grain size studies of the Bima Sandstones (Chapter 5, figs. 5.1, 5.4 and 5.5) it was concluded that the Bima Sandstones are ill-sorted. This conclusion is supported by the results of the zircon elongation frequency curves. The average elongation ratio of the zircons in the Bima Sandstones in the area of study is generally more than 2.0. Poldervaart (1955, page 441) came to the general conclusion that the average elongation ratio in most sediments is generally less than 2.0. The Bima samples are more akin to those plotted for granites in the Ben Lomond area of California by Spotts (1962). This relationship may be taken to indicate that the Bima Sandstones were derived from granitic sources and that the sediments were transported only for short distances before they were deposited in the Benue Valley. The short

Fig. 45.



Elongation frequencies for zircon in the Bima Sandstones

(Reduced major axes for crystal elongation for these same samples shown in fig.44.)

distance of transport did not give enough time for sedimentary processes to alter the original characteristics of the zircons of the granitic masses from which the Bima Sandstones were derived.

TOURMALINE

Tourmaline occurs in the samples studied and it forms a sizeable proportion of the non-opaque minerals of the Bima Sandstones in the area of study (table 4.2). Tourmaline ranges in percentage occurrence from 1 to 37 and averages 15 percent of the non-opaque minerals in all samples studied. The percentage occurrence of tourmaline decreases with increase in grain size, with the highest concentration in the finer size grades. In the coarsest size grade tourmaline and rutile rank next to zircon in percentage abundance among non-opaque heavy minerals.

The most common form of tourmaline in the Bima samples in the area studied was as short stumpy prismatic crystals that occasionally showed terminal forms at the crystal ends. Irregular forms showed distinctive conchoidal fractures. The percentage of rounded grains is higher in comparison with that of any other important mineral. The commonest variety is the green to greenish brown coloured types, although blue and black varieties are present in all size grades of all samples. Brown and reddish brown ones are very rare.

The commonest type of inclusions found in the tourmaline crystals are circular vacuoles, zircon, opaque particles, and both granules and needles of rutile. Zoning was never observed to occur in tourmaline crystals of the samples studied.

SPHENE

Sphene is relatively abundant in the Bima Sandstone samples in the

area studied. It is usually limited to the finer grade sizes (table 4.2), and exhibits a range of colour from pale yellow through brown to grey varieties. The sphene crystals are characterised by their lack of extinction. The commonest forms are the spindle-shaped and double wedge-shaped crystals.

Sphene weathers into leucoxene and this could be the reason for the scarcity of sphene in some samples, especially in those with a high leucoxene content. Sphene is also known to form as an alteration product of ilmenite. There was no observable association between sphene and ilmenite or sphene and leucoxene to confirm the possible alteration of one to the other in the Bima rocks.

EPIDOTE

Epidote has the same general distribution as sphene being more abundant in the finer size grades and diminishing considerably in the coarsest size grades (table 4.2). Epidote occurs either as broken fragments or as euhdral crystals showing prismatic faces. Occasionally sub-angular grains that look like chips of broken "bottleglass" are present in some samples. The commonest varieties are the grey ones.

HORNBLENDE

The green type is the commonest variety although brown types, showing very strong pleochroism, are also present. Hornblende is generally present only in small amounts in all the samples. It also shows a tendency to be more concentrated in the finer size grades like all the other non-opaque minerals. It never exceeded 17 percent in occurrence in any of the samples studied. Generally the hornblende occurs in elongated cleavage plates, and as rounded or angular grains. A few bluish grains

were observed to be present in some samples, as occasional grains.

RUTILE

Rutile occurs commonly as reddish-brown prismatic or subhedral grains with rounded extremeties. There are also many broken euhedral and well abraded spherical or near spherical grains in almost all samples studied. Most of the euhedral crystal forms show good pyramidal terminations. Many prismatic grains exhibit characteristic stiations which run obliquely to the prism edges (plate 4.6). This phenomenon is known to be a result of polysynthetic twinning parallel to the [10] face. (Milner, 1962).

APATITE

Apatite is absent from most of the samples which have been studied, but it forms the major constituent of the heavy mineral suite in some. The abundance of this mineral in some samples and its complete absence in others is probably due to granular variation, and therefore has no significance as regards to source. The most common type of apatite in the samples studied from the Bima Sandstones is rounded to well-rounded or elongated prismatic, or egg-shaped grains. Some grains contain minute inclusions aligned parallel to the principal axis. Grains generally show evidence of solution with re-entrants and pitted surfaces, while others have very smooth surfaces. The commonest variety is the white to colourless ones, with straight extinction under crossed nicols. Apatite is liable to attack by acid, and treatment with dilute acid solutions may lead to the loss of apatite from samples. In the Bima samples which were treated with dilute hydrochloric acid there was a considerable quantity of apatite.

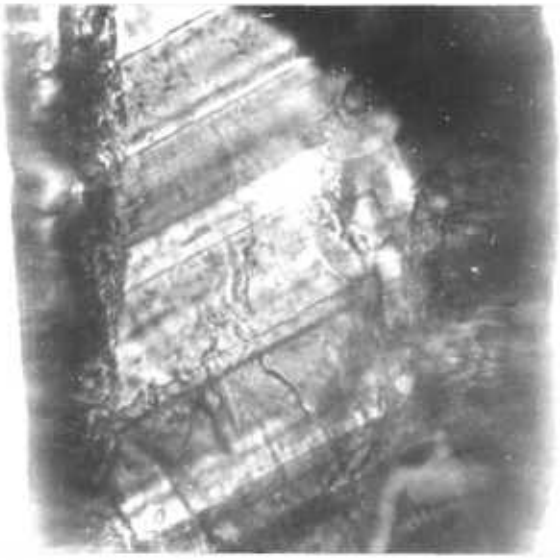


Plate 4.6. Prismatic form of euhedral crystal of rutile exhibiting characteristic striations which run obliquely to the prism edges. X100.

ALTERITES

These are mineral grains which were made unidentifiable under the microscope as a result of severe chemical weathering. Their appearance under the microscope is either as turbid or translucent grains or aggregates of unknown composition. Van Andel (1950) assumed them to be the end products of alteration of various minerals.

BROOKITE

Brookite occurs in a number of the Bima Sandstones samples as yellowish-green, prismatic and tabular forms (plates 4.7.1 and 4.7.2). The grains exhibit peculiar interference colour change on rotation, and have striae parallel to the principal axis. That the grains are detrital is shown by the worn attitude and the coating of iron stain around the edges of the grains. Brookite is a mineral of igneous and crystalline metamorphic rocks but it can also form authigenically by the pneumatolysis and tourmilinization of titaniferous biotite (Milner, 1962).

The other heavy minerals present in the Bima Sandstones in the area of study are barytes and garnet. The former was observed in only one sample where it forms about 80 percent of the heavy mineral constituent. The mineral occurs as colourless euhedral forms exhibiting prismatic and strongly developed basal plane, with characteristic straight extinction. The perfect euhdral nature of the crystals and the position of the mineral in the minerals hardness scale (Hardness 3) indicate an authigenic source for the mineral brookite. Garnet was observed in only four samples as colourless transparent crystals that are sometimes stained yellow. The grains are transparent with jagged edges.

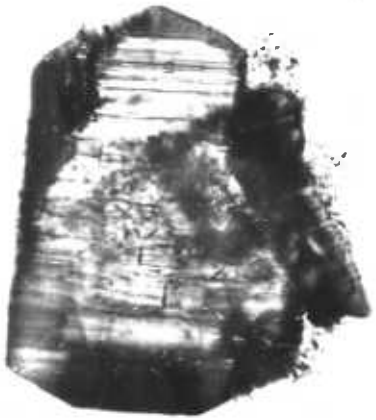


Plate 4.7.I. Euhedral crystal of brookite terminated at both ends by well developed pyramidal faces and showing striae which run at right angles to the edge of the crystal. X100.



Plate 4.7.2. A euhedral crystal of brookite with pyramidal faces at one end and a large basal plane at the other end. The dark areas to the top left of the crystal are iron oxide stains. X100.

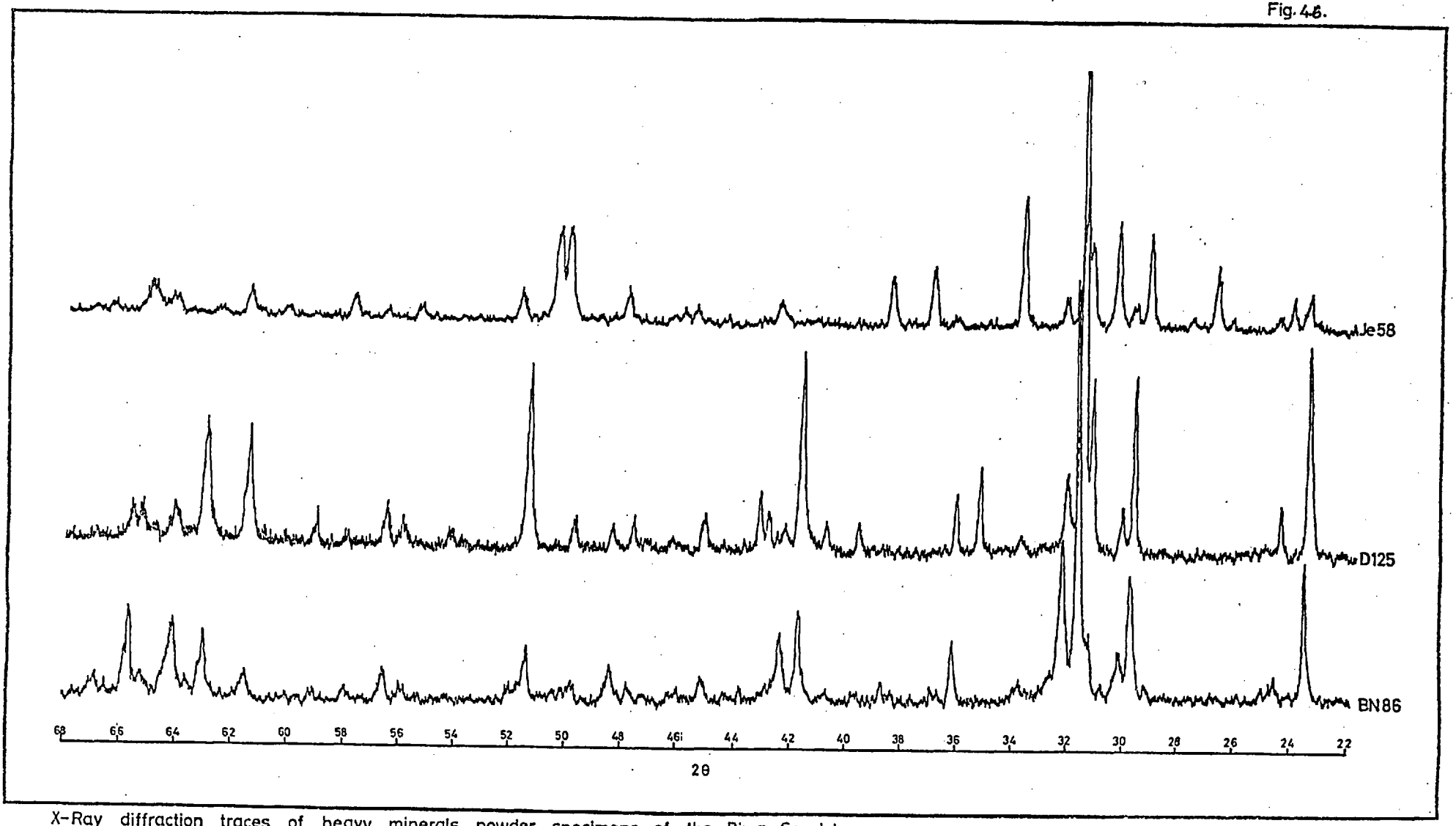
4.6 X-RAY ANALYSIS OF HEAVY MINERALS-DISCUSSION

The X-ray analysis of the heavy minerals of the Bima Sandstones samples was carried out as an aid to microscopic identification of each heavy mineral species. Thus the X-ray diffraction analysis of the heavy mineral assemblage was employed strictly in a qualitative sense to determine the various species present in different samples studied. The detailed quantitative results discussed in this chapter were obtained by normal petrographic analysis. X-ray diffraction traces of the different heavy mineral species was not used for the quantitative analysis. Zircon, sphene, epidote rutile, tourmaline, apatite, brookite, hornblende and barytes were identified from the X-ray diffraction peaks of the analysed samples (fig. 4.6). Zircon peaks are strongest at the 3.30 Å, 443 Å and 2.52 Å; sphene at 3.23 Å, 2.99 Å, and 2.60 Å. Epidote was identified at the 2.99 Å, and 2.60 Å. Epidote was identifed at the 2.90 Å, 2.40 Å; and 1.64 Å, peaks; Brookite at 3.51 Å; 2.90 Å, and 3.46 Å; Hornblende at 1.44 Å, 2.55 Å; and 2.68 Å; Rutile at 3.24 Å, 2.48 Å, and 1.68 Å; Baryte at 3.44 Å, 3.10 Å, 2.12 Å and 4.44 Å; Tourmaline at 4.22 Å 3.51 Å; 2.89 Å, and Apatite at 2.91 Å; 2.92, and 2.81 Å.

4.7 THE GRAIN SIZE DISTRIBUTION OF THE HEAVY FRACTIONS

The grain size distribution of the heavy fractions of the Bima Sandstones in the area of study was investigated on the basis of the relationship between the grain size distribution of the sediments and the heavy fractions, and the grain size of the heavy fractions and their heavy mineral composition. For the study of the grain size distribution of the heavy fractions and the sediments only those samples which have data for the grain size of the light fraction were used. The difficulties encountered in having samples with enough overgrowth material on detrital

Fig. 46.



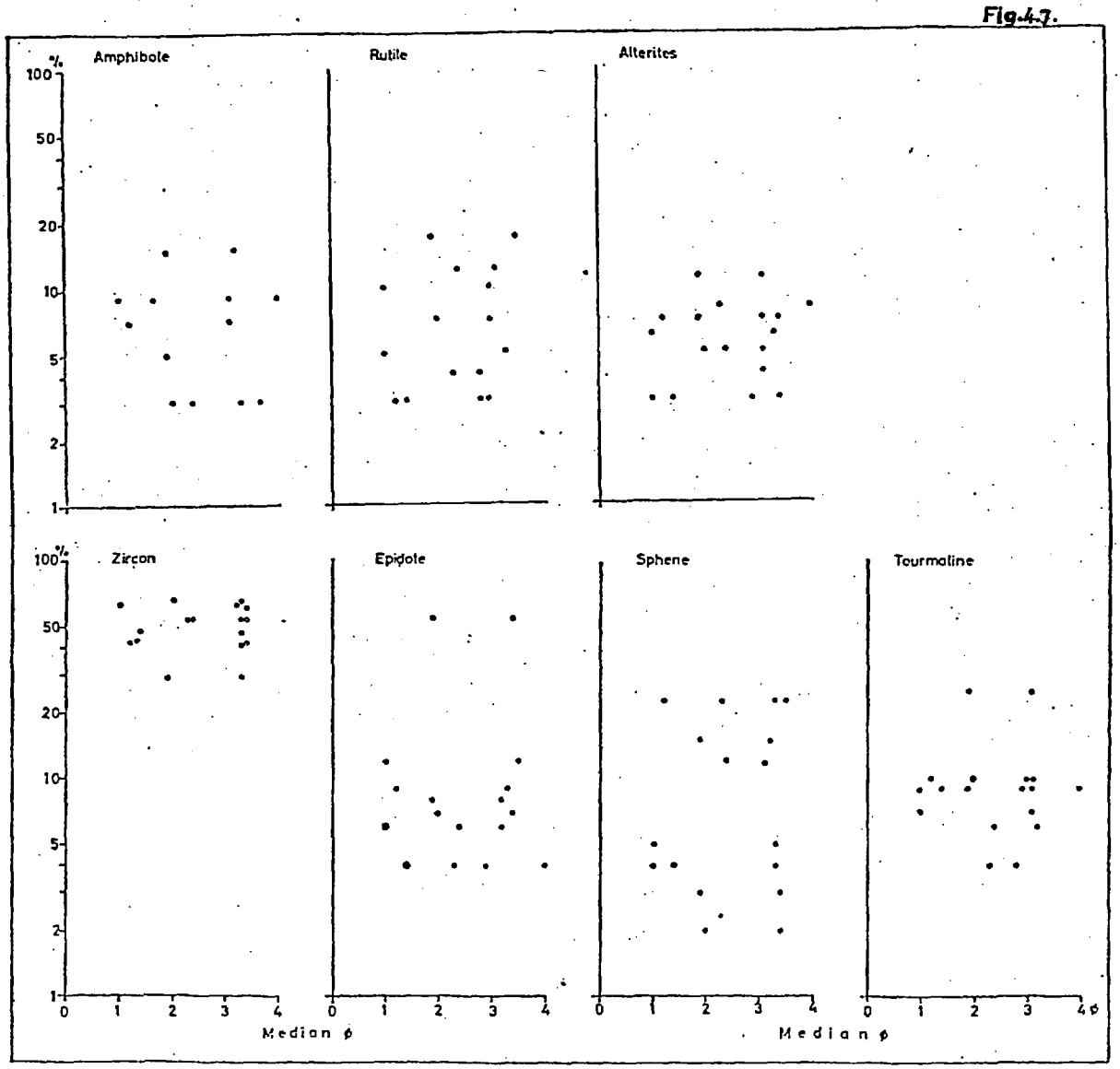
X-Ray diffraction traces of heavy minerals powder specimens of the Bima Sandstones.

grains for grain size measurements of the light fraction limited the number of samples used for this aspect of the study to 9. This number is too small to portray a reasonable relationship between the grain size of the light fractions and the heavy fractions. Thus, the data are too scanty and probably not accurate enough for a detailed investigation as discussed by Chang (1971) and Van Andel (1953).

In figure 4.7, the medians of the sands are compared with the medians of the heavy fractions. No close relationship of both sets of parameters is evident. However, one significant fact is shown by the graphs which are plotted on logarithmic paper. The heavy minerals always plot on the fine side of their equivalent light fractions. This relationship illustrates the fact that the heavy minerals were deposited along with light minerals of larger grain size. The calculated average difference between the grain size of the light minerals and the heavy fraction using their medians is approximately 1.42 phi. This means,that on the average, the heavy minerals always settled with light minerals 1.42 phi larger than them. The conclusions arrived at in this aspect of the grain size distribution of the heavy minerals are by no means assumed to be strictly authentic for the Bima Sandstones as the data used for the study are too scanty. Nevertheless, they provide a good insight into the general behaviour of the heavy minerals with respect to the light minerals during their transportation and deposition.

The influence of size on the heavy mineral composition of the Bima Sandstones is shown by the relationship between the percentages of the various mineral species and the size distribution of the heavy fraction. This relationship is shown in figure 4.1 and has been discussed under

Fig.4.7.



Relation between medians of sand (black dots) and heavy fraction (open circles) plotted logarithmically

each mineral species. The general trend is that of a decrease in amounts of the heavy minerals with increase in grain size. Only the opaque minerals (black areas in fig. 4.1) show an increase in amounts with increase in grain size. The bulk of the non-opaque minerals are concentrated in the finer size grades.

It is here suggested that a more detailed analysis of the relationship between the grain size distribution of the heavy fractions and the sediments be carried out. This analysis should take into account only those samples with an appreciable amount of overgrowth material on detrital grains to provide information on the grain size of the light fraction, as the Bima Sandstones are highly indurated which prevents the use of sieving analytical methods.

4.8 THE PROVENANCE OF THE HEAVY MINERALS

The Cretaceous Bima Sandstones in the area of study differ little in heavy mineral composition, being principally colourless zircon, tourmaline, green-brown hornblende, green amphiboles, sphene, epidote, rutile, apatite, limonite, leucoxene, magnetite, ilmenite, alterites, and more rare types. Acid treatment of samples containing carbonate cement did not remove acid soluble apatite but this could probably have affected the percentage occurrence of this mineral in the treated samples.

There is a distinctive occurrence of weathering resistant minerals in this formation namely: zircon, tourmaline, and opaques; and minerals of probable diagenetic origin which include brookite, baryte, and leucoxene. There is a preponderance of zircon, tourmaline, sphene, epidote, rutile and hornblende in the samples obtained from both the lateral and vertical extent of the Bima Sandstones in the area of study. This group

of minerals is characteristic of granitic, gneissic, and other metamorphic rocks. They could also be derived from re-worked sediments. The view that the Bima Sandstones were derived from the granitic and gneissic rocks of the Basement complex lying to the northeast and southeast of the present area of study was first suggested by Jones (1962). He came to this conclusion as a result of the paleocurrent analysis of the foreset dip azimuth and the mineral composition of the samples he studied from the Lau and Yola areas. The abundance of a suite of heavy minerals typical of granitic and highly metamorphosed rocks leads the present writer to also conclude that the Bima Sandstones were derived from granitic and metamorphic source rocks, and the fact that the Bima sediments in the present area of study were also derived mainly from the granitic masses to the northeast and southeast is supported by the results of the paleocurrent analysis of the Bima Sandstones in this area (chapter 8). Minor contributions were also made from the Basement rocks lying to the northwest of the present study area.

The lack of noticeable variation in the heavy mineral suite, both laterally and vertically, in the sequence prevents their use as a correlation tool and can be used only in a general way to determine provenance. It is possible to use spatial differences in one particular mineral in colour or form as a means of determining sediment provenance. Callender and Folk (1958) used idiomorphism in zircons of the Lower tertiary sands of Central Texas to determine the source of the sediments. However, such detailed work in heavy mineral analysis is beyond the scope of the present work. A more detailed study of the heavy mineral assemblage of the Bima Sandstones is however suggested to provide a more complete picture for the Cretaceous sediments of the Benue valley.

CHAPTER 5

PETROGRAPHY OF THE BIMA SANDSTONES

5.1 INTRODUCTION

The petrographic study of the Bima Sandstones was undertaken to obtain information on the textural properties of the rocks, the primary detrital minerals (framework minerals), the authigenic minerals, the types of cementing material and the influence of these different properties on the primary porosity and permeability of the Bima Sandstones.

Rock samples collection in the field was done in each measured section of the Bima Sandstones spanning the entire area of study. At each locality samples were taken at an average distance of 10 meters of each other up the sequence. Choice of sampling location was also determined by change in the lithology and texture. A total of 143 samples were collected from the entire area of the present study. All of the samples were made into thin sections for microscope study. Only those which showed a good correlation with the type of petrographic study being undertaken were chosen and used for that particular study. Thus 32 samples were used for grain size analysis, and shape and roundness studies; 29 samples were selected for the modal analysis; and 24 samples were used for the X-ray analysis of clay minerals.

5.2 METHODOLOGY

5.2.1 Modal analysis

The 29 samples selected for this study were chosen from four areas as follows: 7 samples from the Bogoi Hills area, 7 samples from the Jelani area, 7 from the Bashama area, and 8 samples from the Jawara Hills area. Each slide was quantitatively analysed for the different minerals present

using a mechanical stage. Each slide was traversed at a regular interval of 1mm. on the vertical axis at a counting speed of 0.5mm. on the horizontal axis. This is the line method of analysis. The analyses were made at a magnification of X 100. An average of 45 points were counted for each traverse, and 7 traverses were counted for each slide. The minerals identified during the modal analyses were recorded as follows: Quartz was recorded as polycrystalline, undulose, and non-undulose varieties following the terminology of Blatt and Christy (1963); feldspar was differentiated into three varieties: microcline, plagioclase and orthoclase; rock fragments; micas; alteration products; and opaques. The minerals grouped under the umbrella of alteration products are those believed to have been produced by the degradation of primary detrital feldspars.

5.2.2 Mineral identification

Mineral identification was based on optical mineralogy using the different optical properties of the different minerals present. The staining technique described by Evamy (1963) using alizerin Red S was employed to differentiate the types of carbonate minerals present in the samples. The feldspars did not respond to etching with hydrofluoric acid to differentiate orthoclase from quartz. Seven different mineral types were identified optically: feldspars, quartz, carbonates, rock fragments, alteration products, opaques, and accessory minerals.

5.2.3 X-ray mineralogy

Clay mineral analysis was carried out on 24 samples of the Bima Sandstones in the area of study, using X-ray diffraction methods. The aim of the study was to determine the varieties of clay minerals in the Bima Sandstones, and was therefore purely qualitative. The selection of samples for analysis was based on the visual estimation of the clay minerals content of the hand specimens of the rocks.

5.2.3.1 Sample preparation

A total of 24 rock samples known to contain an appreciable amount of clay minerals were selected for crushing and clay mineral extraction. The choice of samples analysed was based entirely on the availability and abundance of clay minerals as seen on the hand specimens of the rock samples, and from the results of the modal analysis of the whole-rock samples. Most of the rock samples selected were friable as a result of weathering. It was therefore possible to crush the samples by an up and down movement of a metal pestil on a metal mortar. Grinding of the samples was avoided as this would have resulted in the fragmentation of the quartz grains and their incorporation in the clay fraction. The clays were separated from the coarse materials by dry sieving twice using a 240mm. mesh sieve.

5.2.3.2. Experimental procedure and analytical technique

An approximate weight of 2 to 3 grams of the clay mineral fraction was taken to which water and sodium hexametaphosphate (Calgon) were added. The mixture was stirred vigorously and allowed to stand for about 10 minutes, to allow materials greater than clay size to settle down. The clay suspension was poured into a filtering device which uses suction to deposit the clay on a ceramic tile. The clay deposit on the ceramic disc was dried in an oven at temperatures up to 60°C until the clay film was firm to the touch. This is necessary, otherwise in the specimen holder of the X-ray apparatus, the clay film would be damaged and the geometry of the diffraction distorted.

X-ray traces were run on the oriented clay mineral samples mounted on ceramic discs in the following manner:

a) Each specimen was air-dried and scanned from 2^{o}-20^{o} 2θ.

b) A thin layer of glycerol was smeared on the surface of the clay specimen and allowed to dry before scanning from 2^{o}-20^{o} 2θ.

c) Specimens were heated to 400^{o}C for one hour in a furnace and then scanned from 2^{o}-20^{o} 2θ.

d) Specimens were finally heated to 500^{o}C for one hour in a furnace and then scanned from 2^{o}-20^{o} 2θ.

An unoriented powder press run was made on one specimen.

The identification of the various clay minerals present in each sample was carried out on the diffractograms after the conversion of the angles in θ values to d-spacings using standard conversion tables. The guide lines laid down by Brown (1961) and Carroll (1970) were followed in the identification of the clay minerals from their peak characteristics.

5.2.4. Textural analysis

Textural analyses of the Bima samples involved the measurements of grain size, roundness and sphericity. These measurements were carried out on thin sections of 32 samples covering the whole area of study. The choice of samples for use in these analyses depended upon the presence of secondary overgrowth material on the primary detrital grains. Thin section analysis was used for the determination of the textural characteristics of the rocks because of the highly indurated nature of the Bima Sandstones, which precluded the use of sieving methods. The chief cementing material in the Bima Sandstones is silica which is difficult to digest chemically.

Textural analyses of sedimentary deposits are an aid to the

determination of the effects of the transporting medium on the detrital grains. The analyses should therefore be carried out on the primary detrital grains. The Bima Sandstones have been cemented by the addition of silica on the detrital grains as overgrowth material which has altered the shape and size of the primary detrital grains. The results obtained on such a material by the use of sieving methods will not be correct. Thin sectional methods are therefore more adequate and reliable in the textural studies of the Bima Sandstones. This was accomplished in the case of the grain size analysis by measuring the intermediate axis of the primary detrital grain, excluding the secondary overgrowth material from the measurements. The intermediate axis was chosen because a great number of grains were encountered having their longest axis many times the size of their shortest axis. The measurement of either only the longest or shortest axis of such grains would not be representative enough of the true size of the grains. At least three measurements were taken along the intermediate axis on each grain to make the results more statistically representative. The results of the measurements given in appendix 1. are the mean values of three measurements for each primary detrital grain of quartz. Measurements were made on only quartz grains because no noticeable secondary overgrowth material was found on other mineral grains.

The magnification used in the measurement of grain size was X 100. An occular micrometer was used for these measurements, after it had been calibrated to the metric system with a stage micrometer.

The actual size represented by the calibrations on the occular micrometer were calculated using a slide micrometer. It was found that 10 divisions on the eyepiece micrometer was equivalent to 1mm. on the

slide micrometer. The results obtained from these measurements were converted into phi units from the graphic conversion tables of Folk (1968).

5.2.4.1 Grain size analysis

A maximum of 100 grains were measured for each sample, although in samples with few obvious secondary quartz overgrowth material a smaller number of grains was counted (see appendix 1). The results of the measurements in phi units were plotted on probability paper for the 32 samples (fig. 5.1) studied and the following statistical measures were determined:

Graphic mean (Folk and Ward, 1957) $Mz = \frac{\emptyset 16 + \emptyset 50 + \emptyset 84}{3}$

Median (obtained from the cumulative curve) $Md\emptyset = \emptyset 50$.

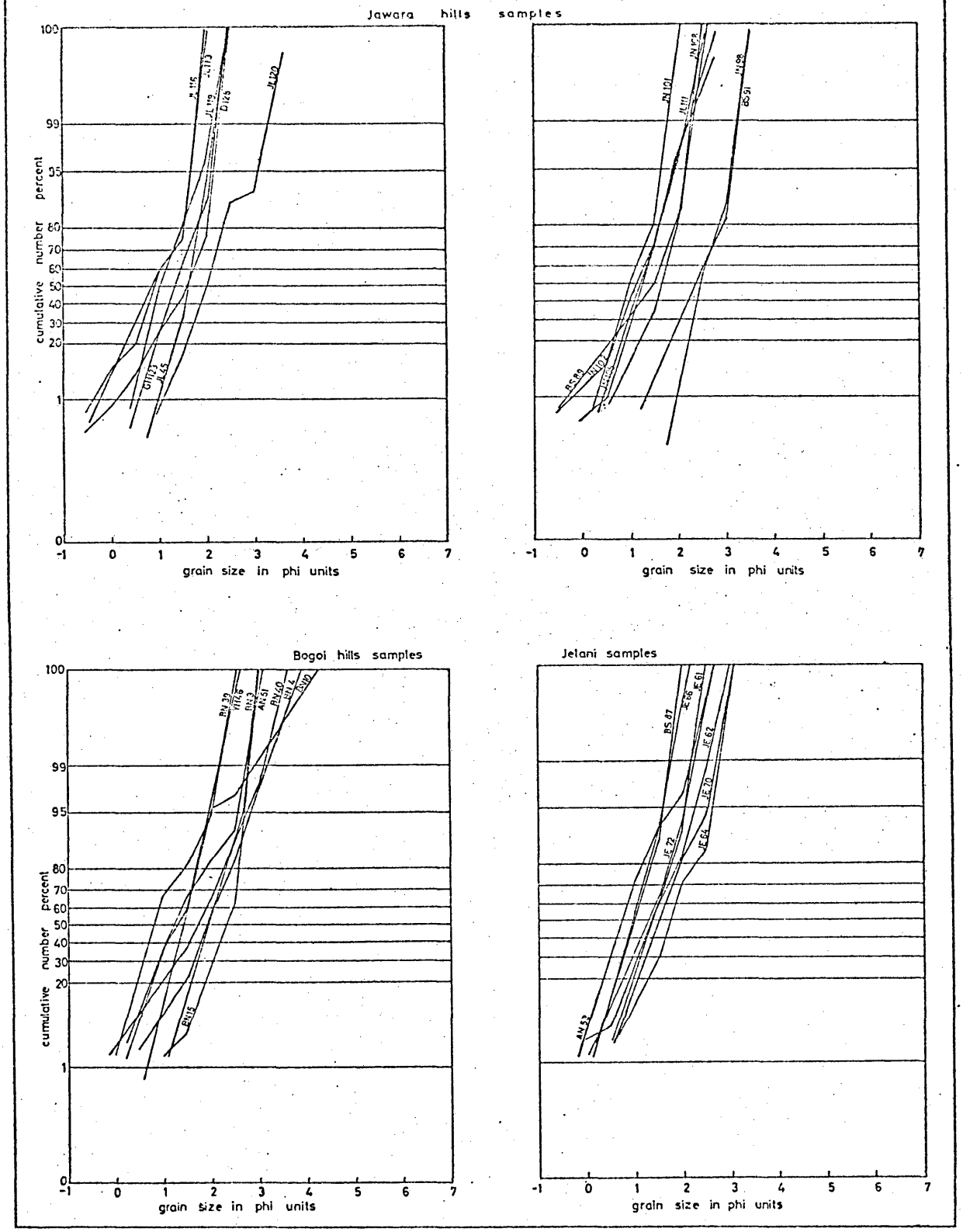
Inclusive Graphic Standard Deviation (Folk, 1968) $O_1 = \frac{\emptyset 84 - \emptyset 16}{4} + \frac{\emptyset 95 - \emptyset 5}{6.6}$

Graphic Kurtosis (Folk, 1968) $K_G = \frac{\emptyset 95 - \emptyset 5}{2.44(\emptyset 75 - \emptyset 25)}$

Inclusive Graphic Skewness (Folk, 1968) $SK_1 = \frac{\emptyset 16 + \emptyset 84 - 2\emptyset 50}{2(\emptyset 84 - \emptyset 16)} + \frac{\emptyset 50 + \emptyset 95 - 2\emptyset 50}{2(\emptyset 95 - \emptyset 5)}$

In the present work both the kurtosis and skewness measures were not used as these two statistical measures are only useful in the tests of sieve analyses with small variations in the mean size and sorting range of the sediments.

Fig.5.1.



Grain size distribution in the Bima Sandstones.

5.2.4.2 Shape (sphericity) and roundness

Wadell (1932) was the first to demonstrate the difference between these two textural parameters. Krumbein (1941) defined the shape of a particle as its form, measurable by the ratio of the surface area of the particle to its volume. He defined the roundness of a particle as the curvature of its corners, expressed as the ratio of the average curvature of the particle as a whole independent of its form.

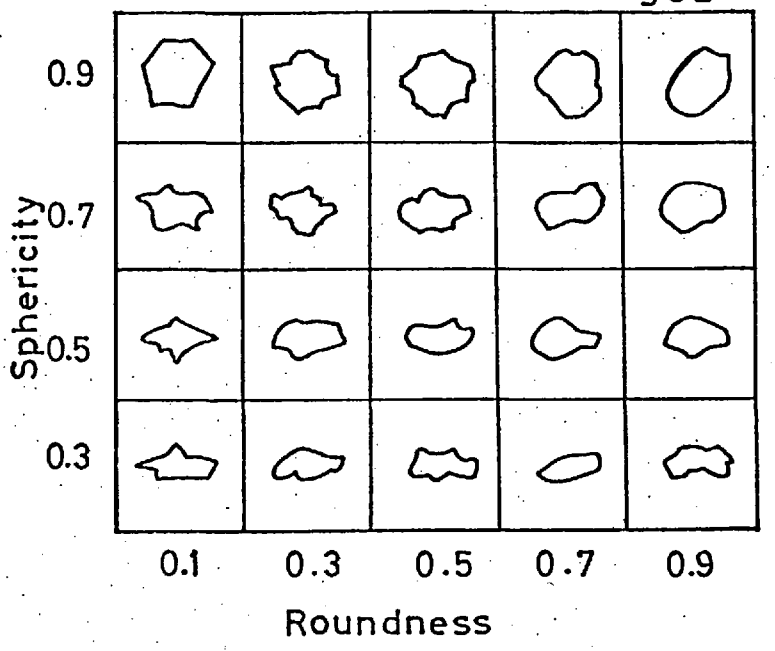
Shape

The shape of the particles of the Bima Sandstones was measured in thin sections of 32 samples (see appendix 2) by the visual estimation method of Krumbein and Sloss (1963). The chart used is shown in figure 5.2. The chart incorporates values for both shape and roundness represented by mathematical values, of from 0.1 to 0.9 for roundness, and from 0.3 to 0.9 for sphericity. In this study, like in the micrometric work, only quartz grains showing secondary overgrowth material were used, and only the primary cores of the grains were measured. A grid pattern of sampling each thin section was strictly adhered to. However, since quartz grains with good secondary overgrowths are few not only the grains that were traversed by the cross-hair of the microscope were measured. Great care was taken not to measure the same grain twice.

Roundness

The procedures for measuring the roundness of grains in thin sectional study were similar to those described for the grain shape measurements. The same comperator chart of Krumbein and Sloss (1963) was used for this study, and the results are given in appendix 2.

Fig.5.2.



Visual estimation chart

5.3.1 Vertical profile of grain size

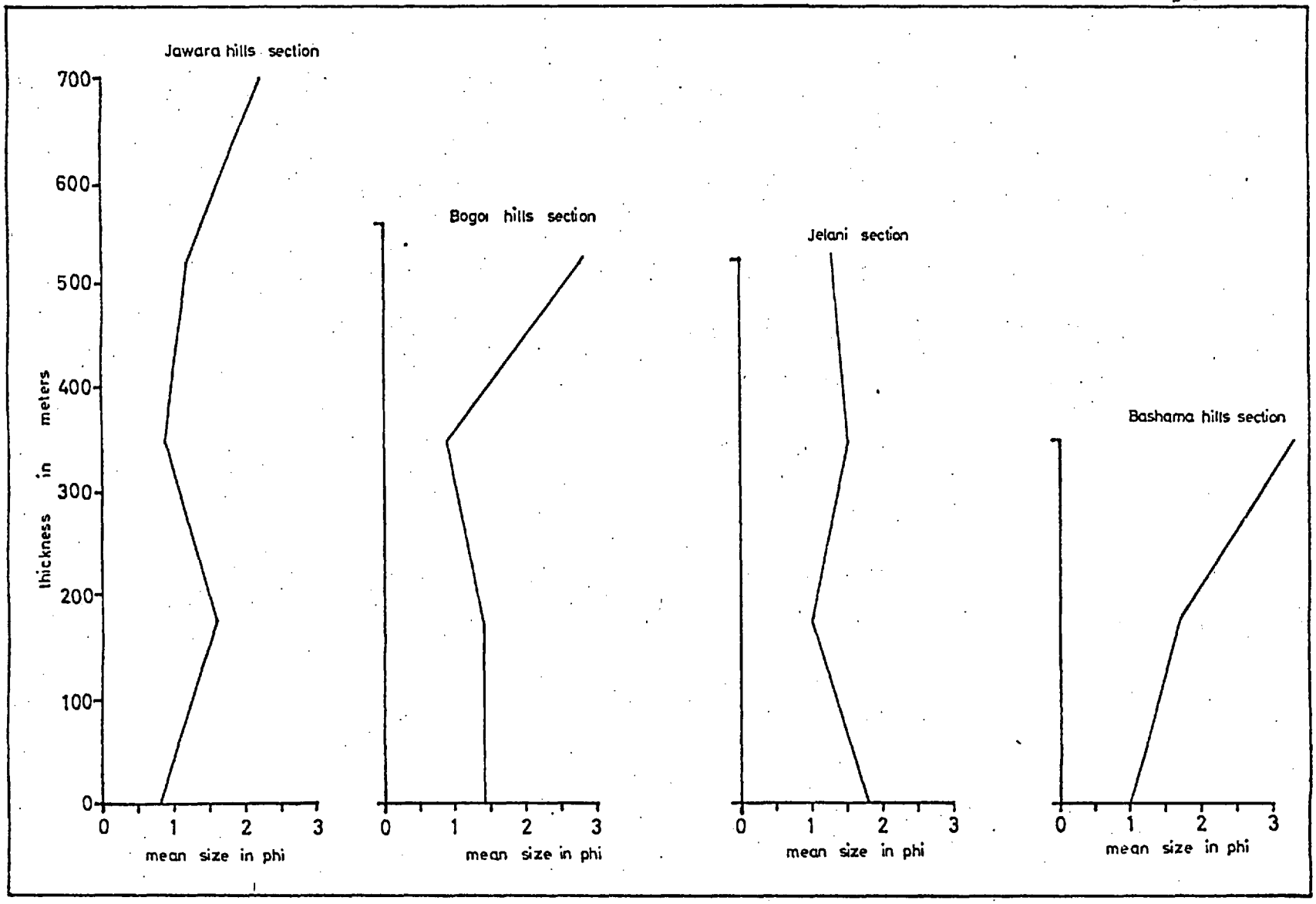
Figure 5.3 illustrates the vertical profile of the grain size of the primary detrital grains. The results used in the plotting of the grain sizes were obtained from the micrometric work carried out in thin section on the quartz grains with secondary overgrowth material. For this study samples were chosen along the sequence in measured sections, and the values of the mean grain size were plotted at the points in the sequence from where the samples were taken. The values used for each sample are the averages of at least 50 grain size measurements, and at least three measurements were made on each grain to ensure that the values obtained are statistically representative of the size of the grains measured.

The picture given by the plot of grain size against the thickness of the sequence at four localities (fig. 5.3) indicates a general upward fining sequence within the Bima Sandstones in the area of study. The upward fining nature of the sediments as shown by this study is in agreement with the conclusion reached in other chapters (see chapters 7 and 8), that the Bima Sandstones are fluvial deposits.

5.4 TEXTURES

For textural studies of the Bima Sandstones 32 samples were examined to determine the grain size, shape (sphericity) and roundness of the detrital components of the rocks. Quartz was used for these studies because it is the only detrital mineral with secondary overgrowth material on the detrital particles. The results of these measurements are shown in appendices 2 and 3.

Fig. 5.3.



Vertical profile of grain size in the Bima Sandstones

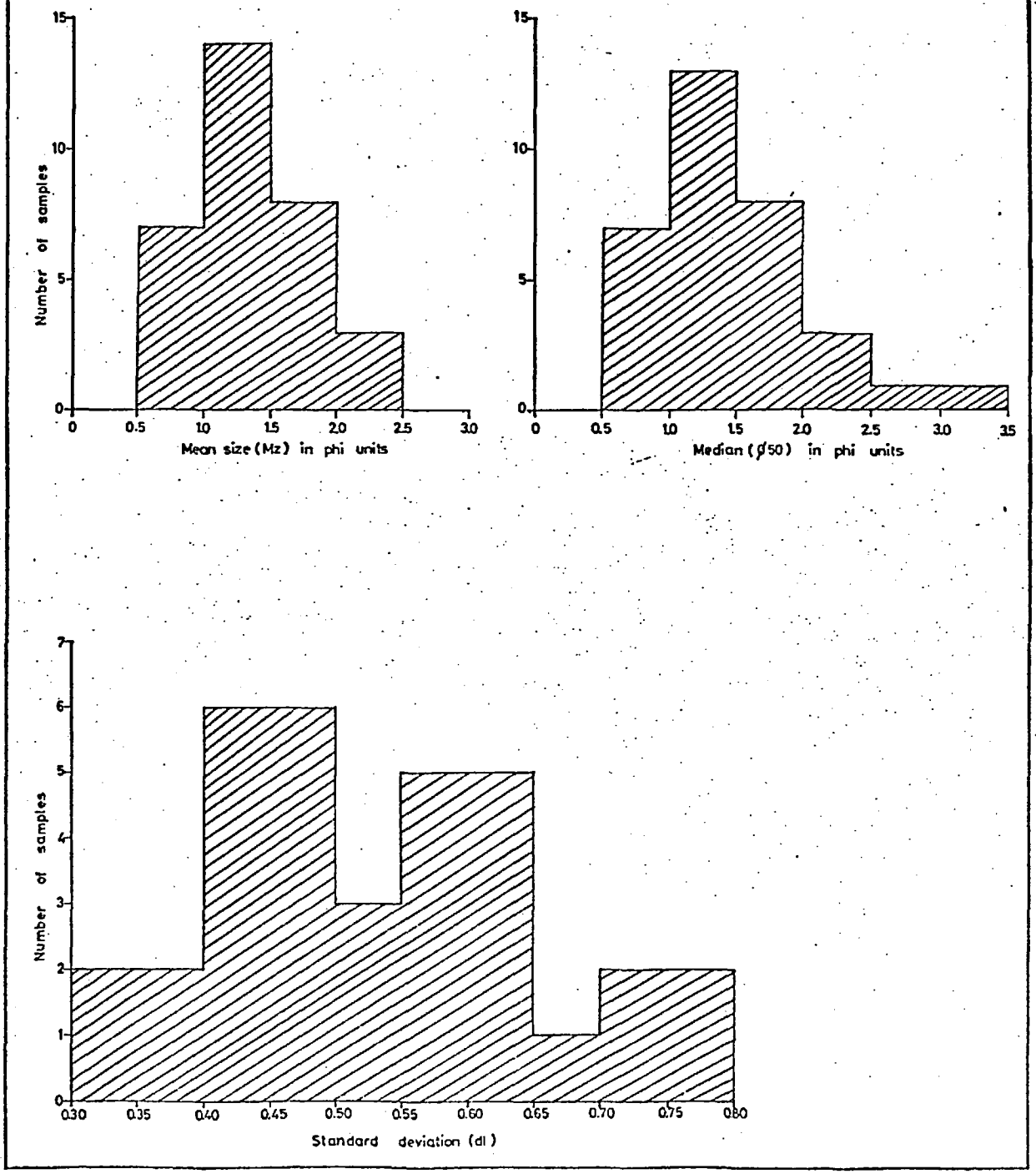
In all the samples studied the mean size (Mz) ranges from 0.75Ø to 2.36Ø, with a standard deviation of 0.31Ø to 0.80Ø. Figure 5.1. shows the cumulative size distribution curves for the 32 samples studied. Probability paper was used for plotting the size data. All samples fall within the range from fine- to coarse- sand. Three samples, Bn 3, Jn98, and Bs91 fall in the fine- to medium- grained sand.

The calculated mean size (Mz) and the median Ø(MdØ) which was obtained from the graphs show that the sandstones range in grain size from fine- to coarse- sands. The values obtained for these parameters are shown in appendix 3, and fig. 5.4. The results indicate that the Bima Sandstones are poorly sorted. The poor sorted nature of the Bima rocks is also illustrated by the polymodality shown by the standard deviation, and the asymmetric distribution shown by the mean and median size of the detrital quartz grains (fig. 5.4). Thus the presence of several peaks in the plot of the standard deviation indicates poor sorting of the Bima Sandstones.

Figure 5.5 shows the relationship between the standard deviation and the mean size (Mz) of the Bima Sandstones. When considered as a whole the sandstones show no definite correlation between mean size and sorting. The plot points range from very poorly sorted to very well sorted sands, and this scatter indicates the general lack of sorting exhibited by the sediments. However, there is a general increase in the standard deviation with decreasing grain size, although some samples deviate from this general trend.

The mean sphericity and mean roundness values for the Bima Sandstones are shown in appendix 2. Figure 5.6 illustrates the relationship

Fig.5.4.



Frequency distribution of mean size, median and standard deviation

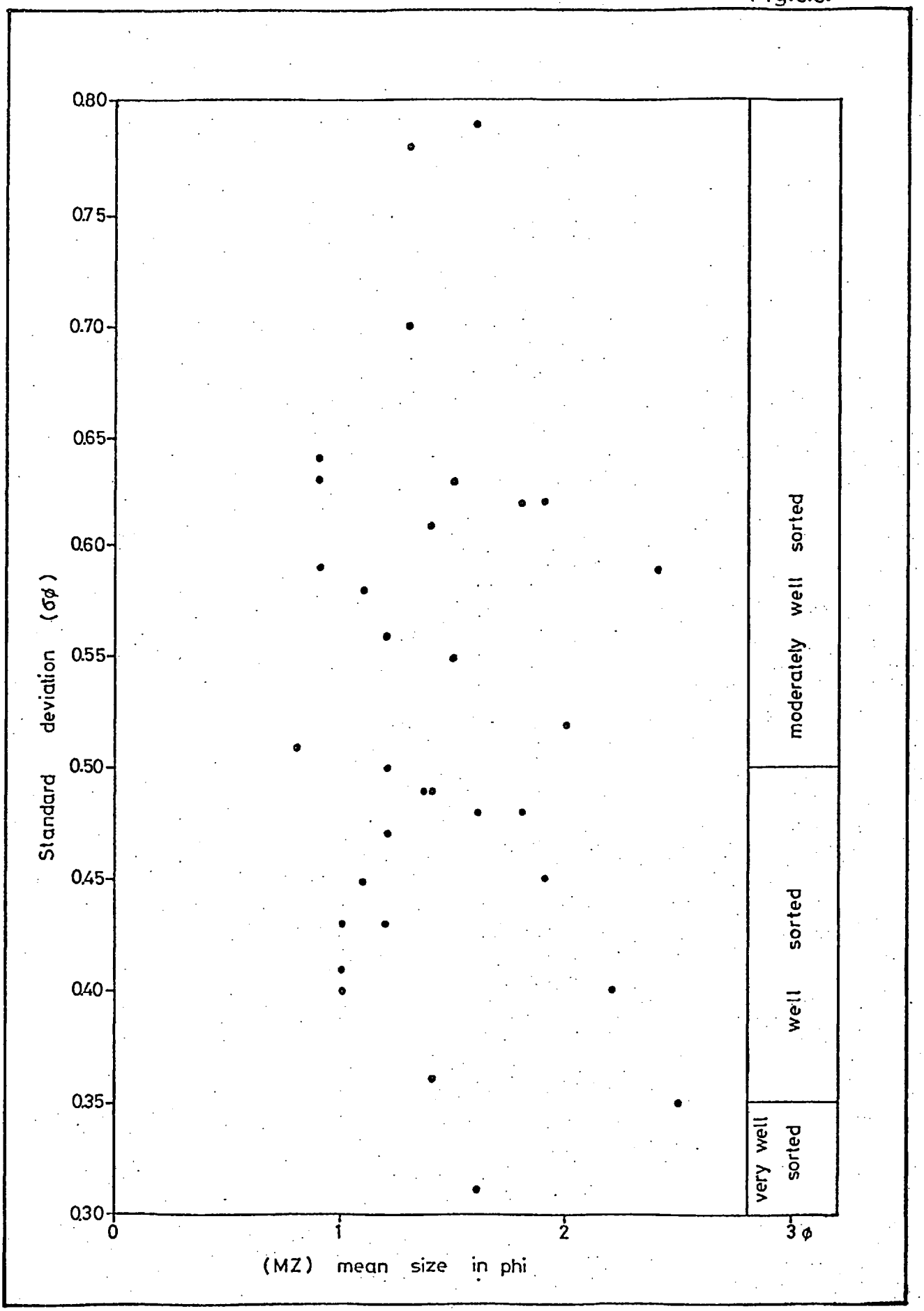
between mean sphericity and mean roundness for the 32 samples studied. Taylor and Russell (1937) noticed that a small change in sphericity is associated with a relatively greater change in roundness, and that this relationship is linear. There is no definite pattern in the plot of the Bima Sandstones roundness and sphericity values. There are samples with high sphericity and low roundness values and others with high roundness and low sphericity values. This lack of pattern shown in these studies indicates that there is no strict correlation between these two parameters in the Bima samples. Pettijohn and Lundahl (1943) suggested that both roundness and sphericity are a function of grain size and that they are also functions of each other. There is also no significant relationship between the roundness and mean grain size of the Bima Sandstones detrital components (fig. 5.6). The Bima Sandstones are well suited to show a correlatable relationship between roundness, sphericity and grain size because of their wide range in grain size. The lack of correlation is therefore suggested to be due to the small number of samples studied and the small number of grains measured in each sample. A study of the textures of the Bima Sandstones involving a larger number of sample and the use of techniques such as cathodo-luminescence may improve on the present results.

5.5 FRAMEWORK MINERALS

5.5.1 Introduction

The qualitative evaluation of the mineralogy of the Bima Sandstones in the area of study was achieved by the use of a petrographic microscope for mineral identification. Staining methods were also employed to differentiate between the types of carbonate minerals and the untwinned

Fig.5.5.



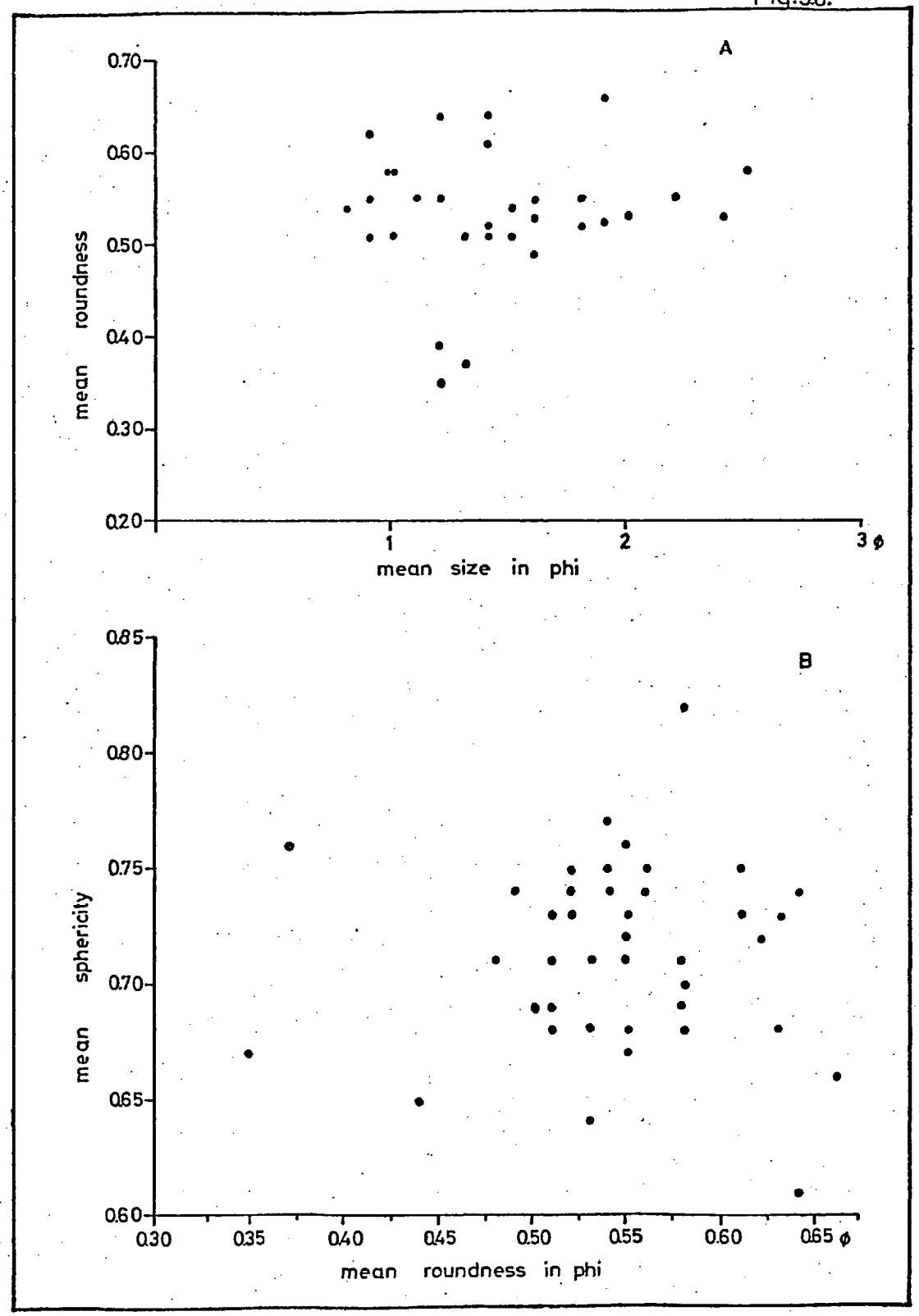
Relationship between standard deviation and Mz mean size in the Bima Sandstones

varieties of feldspars. Quartz and feldspars are the major framework minerals. Other minerals include rock fragments, the accessory minerals, and micas. All the minerals mentioned above form the detrital mineral suites of the Bima Sandstones. Chemically precipitated minerals include carbonates, and reddish brown iron oxide which forms the matrix in some of the samples. Authigenic minerals present in the rocks are mainly kaolinitic clays, sericite, chert, and micas. The authigenic minerals are derived from the degradation of the feldspars. The mode of occurrence of the different minerals will now be discussed in detail.

5.5.2 The detrital minerals

Quartz is the most abundantly occurring detrital mineral in the Bima Sandstones. It forms 41 to 78 percent of the minerals present in the rocks. Three main varieties of quartz are present: undulatory monocrystalline quartz, non-undulatory monocrystalline quartz, and polycrystalline quartz. The terminology applied for the different varieties of quartz is adopted from those formulated by Blatt and Christy (1963). They defined undulatory quartz as a monocrystalline quartz grain which has visible evidence of strain, the zones of undulatory extinction being either gradational into one another or separated by more or less well defined planes. Both types of undulatory quartz are present in the Bima Sandstones, with the latter type being predominant. Bailey et al, (1958) associated this feature with quartz which has been strongly deformed. Non-undulatory quartz is a monocrystalline variety which shows no visible sign of strain with less than one degree of rotation of the microscope stage. A polycrystalline quartz grain is composed of two or more quartz crystal units, each of which may be either strained or unstrained, with either planar, curved or sutured boundaries between adjacent crystal units of the

Fig.56.



Relationship between roundness and mean size in the Bima Sandstones

composite grain. Some of the polycrystalline quartz grains exhibit two distinct morphological sizes of quartz crystals within a grain. This phenomenon is assumed to represent an arrested stage in the recrystallisation of a quartz aggregate with the smaller crystals being the newly formed ones (Blatt, 1967). Another variety of quartz observed in the Bima rocks is chert which is a microcrystalline form of quartz. It will be discussed along with minerals formed as the alteration products of feldspars.

Of the three major types of quartz the undulose monocrystalline variety is the most abundant in all samples of the Bima Sandstones. The non-undulose variety is the least abundant. The abundant development of the undulose variety may be related to the severe deformation of the rocks in the area of study. Conolly (1965) suggested that deformation is a factor responsible for inducing undulatory extinction in non-undulatory quartz grains in areas which have suffered from severe folding and faulting. Inclusions of bubbles, micas, sericite, clay minerals, rutile needles, zircon grains, and dust trains are common in the quartz grains. Nearly all the samples studied from the Bima Sandstones show quartz with secondary overgrowth material. The overgrowths which appear to be restricted to only monocrystalline quartz are made evident by a train of dust or iron stain around the periphery of the primary detrital grain. The secondary overgrowth material is sometimes distinguished from the primary quartz by its clear appearance as opposed to the inclusion-filled primary core. Some of the grains show lines which are visible only under crossed nicols. These may be cracks which have healed as a result of the deposition of secondary quartz material over them.

Feldspars form 11 to 56 percent of the total minerals present. Three major types were observed in the rocks studied: microcline, plagioclase,

and untwinned (orthoclase) feldspars. Microcline is the dominant type in most of the rocks except in some rocks in which plagioclase is the most abundant variety. The low values for orthoclase may be related to the difficulties of identifying this mineral under the microscope. Staining methods did not yield satisfactory results. It could also be argued that the low percentages in the values obtained for microcline and orthoclase in certain samples is due to the effect of selective alteration of these varieties of feldspar. No overgrowth material was observed on the feldspars. The plagioclase grains range in size from medium- to coarse-grains, with a few fine grains especially in samples containing carbonate cement. The feldspars usually alter into clay minerals, sericite, and chert. The amount of fresh feldspars in the carbonate cemented rocks is higher than that in the silica cemented rocks.

Rock fragments of metamorphic origin are present in the Bima Sandstones. They form as much as 10 percent of the detrital mineral grains in some of the rock samples, while they may be totally absent in others. The metamorphic rock fragments consist mainly of quartz and feldspar aggregates. The presence of rock fragments in a sandstone has been used to suggest the relative distance of transport of the sediments from their source to the depositional basin.

Mica grains of detrital origin are present in the rocks and they never exceed one percent of the minerals present in the rocks. Muscovite is the only detrital form of mica found in the Bima Sandstones in the area of study. The mica is colourless under plane polarised light, with a platey form and cleavage traces running parallel to the long axis of the grain. In hand specimens of the rocks the muscovite occurs as whitish flakes or specks. Detrital micas occur more abundantly in the finer-grained

rocks.

Accessory minerals present in the Bima Sandstones are zircon, rutile and apatite. These are the only accessory minerals identifiable under the microscope in thin sections. The accessory minerals are characterised by their high refractive indices and interference colours under crossed nicols of the microscope. Some of the accessory minerals occur as inclusions of different types in the quartz grains. The accessory minerals are discussed in more detail in the chapter of heavy minerals (see chapter 4).

5.6 THE AUTHIGENIC MINERALS

The authigenic minerals present in the Bima Sandstones in the area of study are those formed from the alteration of the feldspars. The authigenic minerals are kaolinite, illite, sericite, micas and chert.

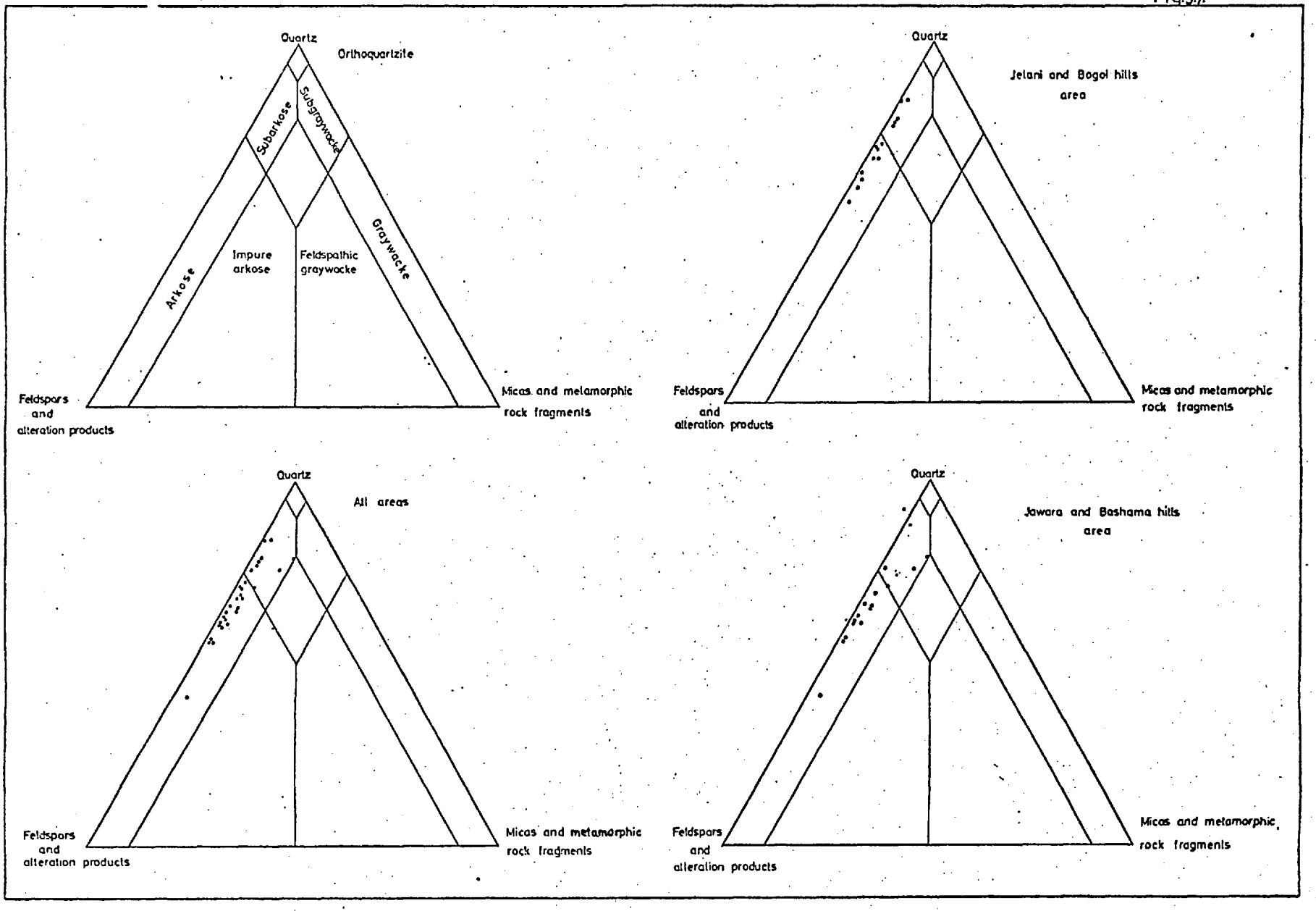
The clay minerals occur as a grey powder in the spaces between detrital grains and in the cores of altered feldspars. The clay minerals are dominated by the presence of kaolinite which occurs as ill-formed vermicular aggregates in some cases. Their petrographic identification was made on material scratched from the hand specimens of the rocks and mounted on slides in water. The petrographic identification was confirmed by the X-ray analysis of the clay fraction of the rocks. Other clay minerals present include illite, mixed layer clays and montmorillonite.

Sericite occurs as small lath-like aggregates between detrital grains and within the clay material in most cases. Sericite is fine-grained muscovite, and in many samples "booklets" of sericite aggregate to form muscovite flakes.

Chertification is another common process in the degradation of the feldspars in the Bima Sandstones. The chert forms as small aggregates of microcrystals of quartz within the clay and sericite materials. The processes of kaolinization, sericitification, and chertification of detrital feldspar grains are observable in all the samples of the Bima Sandstones. These processes will be discussed in more detail under the chapter of diagenesis. In some instances the chert, sericite, and clay materials act as cement to the detrital grains. The presence of chert, sericite flakes and clay materials along twin lamellae and cracks in feldspars, and the relicts of the twin lamellae in altered material containing chert, clays and sericite, are obvious testimonies of the formation of these minerals as the alteration products of the feldspars in the Bima Sandstones.

5.7 SANDSTONES CLASSIFICATION

The sandstone classification used in the present work is adopted from that proposed by Folk (1954). Folk's method involves the use of a three end-member ternary diagram, with quartz and chert; metamorphic quartz, micas and rock fragments; and feldspars and volcanic rock fragments as the end-members. In the present work the classification of the sandstones is based on the modal analysis of 29 samples taken from four areas (fig. 5.7), and the end-members used are quartz and chert; micas and metamorphic rock fragments; and feldspars plus their alteration products. The inclusion of the alteration products in the feldspars end of the ternary diagram is justified by the fact that many samples show feldspar grains in different stages of alteration and in association with the alteration products. A review of the literature on sandstone classification has been extensively treated by McBride (1963) and Pettijohn et al. (1972).



Framework minerals composition and Bima Sandstones classification

Samples in which carbonates, accessory minerals, and opaque minerals occur were recalculated for the three end-members used in the present classification to 100 percent. Fig. 5.7 illustrates the composition of the Bima Sandstones, using a modified form of Folk's ternary diagram. Most of the samples from the Jawara Hills and Bashama Hills areas fall in the arkose class, with only three samples in the sub-arkose class. One of these samples falls very close to the feldspathic sub-graywacke class because of its high content of rock fragments. The Jelani and Bogoi Hills samples have five samples plotting within the sub-arkose class and the rest of the samples fall within the arkose class.

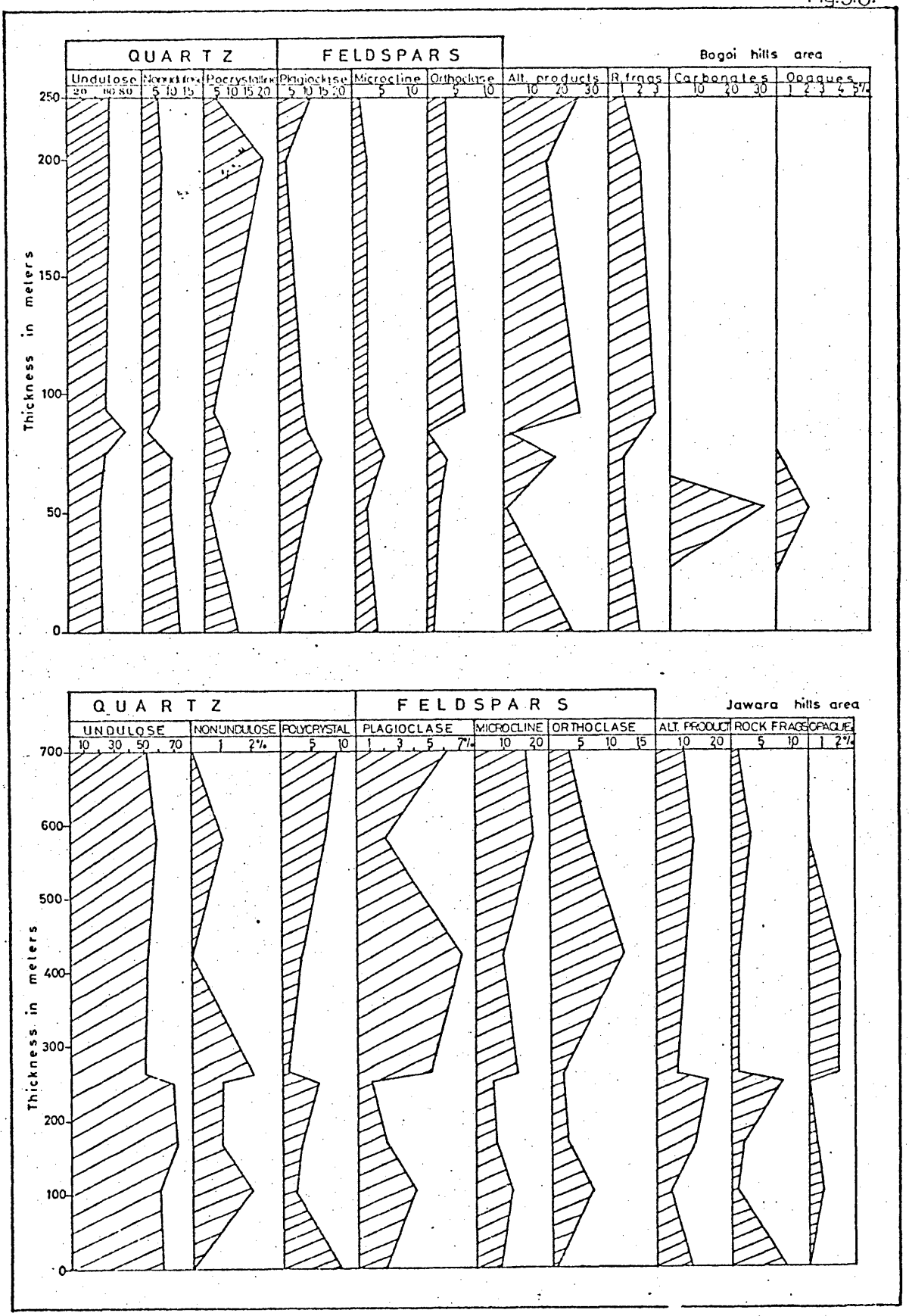
5.8 VERTICAL VARIATION IN MINERALOGY AND MINERALOGIC RELATIONSHIPS

The modal analysis of 29 samples of the Bima Sandstones showed the presence of three major detrital minerals: quartz, feldspar, and rock fragments, and others such as the alteration products of the feldspars, accessory minerals, micas, carbonates, and opaque minerals. At least 7 samples were randomly selected from the total thickness of four measured sections and the different minerals present in them were point counted to provide a statistical information on their percentage frequency occurrence. The results of this operation are shown in figs. 5.8 and 5.9, and appendices 4.1, 4.2, 4.3 and 4.4. From the plot of the data obtained from the modal analysis and the thickness of the sections from which the samples were taken the following observations are made:

a) The absence of orthoclase is accompanied by a decrease in the amounts of alteration products.

b) The presence of carbonates influences the amounts of alteration products.

Fig.5.8.



Vertical variation in mineralogy of the Bima Sandstone

c) A sharp increase in the amounts of feldspars present is accompanied by a significant increase in the amounts of alteration products.

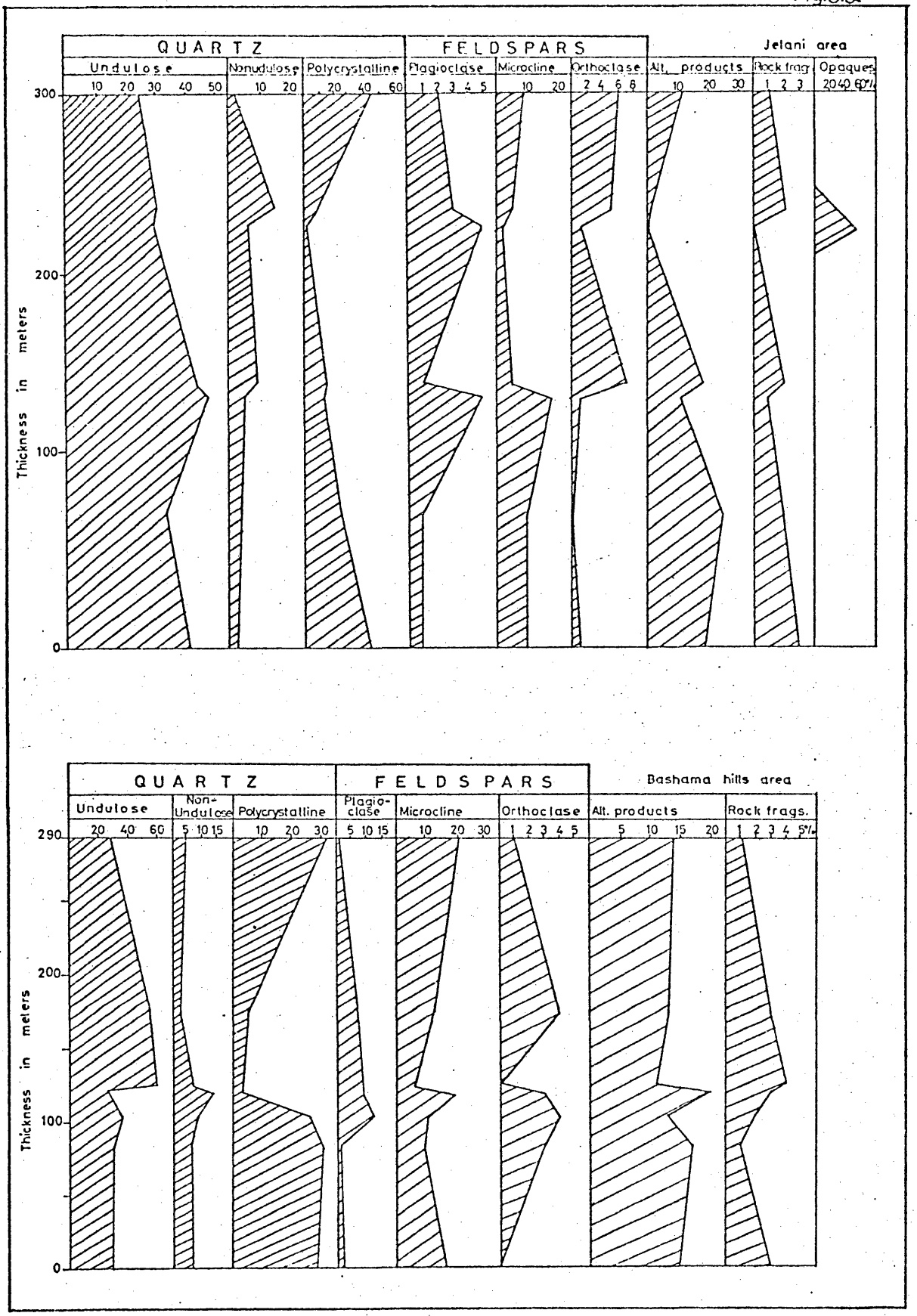
All the three observations made above point to the fact that alteration products were derived from the degradation of feldspars, especially the orthoclase and microcline varieties.

d) A small but significant relationship exists between the different varieties of quartz. An increase in polycrystalline quartz is accompanied by a decrease in the two varieties of monocrystalline quartz. The possibility that the polycrystalline variety breaks down to contribute to the monocrystalline varieties as shown by this relationship cannot be overlooked.

e) The preponderance of the undulose variety of monocrystalline quartz and the low values obtained for the non-undulose variety in all the samples studied reflect the influence of tectonism on the nature of the detrital quartz grains. Conolly (1965) suggested that close proximity of faulting and folding induces undulose extinction in non-undulose quartz grains. This phenomenon is believed to have strongly influenced the quartz grains of the Bima Sandstones in the area of study, as the rocks are severely faulted and folded. This same phenomenon is also evoked to explain the high values obtained for the undulose variety of quartz and the corresponding low values obtained for non-undulose quartz.

f) Althoughdetrital micas are present in the Bima rocks, their scarcity precludes them from the final results of the modal analysis. This scarcity is not unexpected as the rocks are believed to be channel deposits with strong traction currents which would preclude the deposition of fines and micas.

Fig.5.9.



Vertical variation in mineralogy of the Bima Sandstones

g) There is no significant variation in the mineralogy of the sandstones from the four areas selected for this study. This uniformity in mineralogy both laterally and vertically in the sequence indicates a common source area for the Bima Sandstones in the area of study.

5.9 ORTHOCHEMICAL CEMENT AND MATRIX

The principal cementing agents in the Bima Sandstones in the area of study are silica and calcite, while iron oxide (probably hematite) and clay minerals form the matrix. The silica cement occurs as secondary overgrowth material on detrital quartz grains. The secondary overgrowth material manifests itself in the cases where the detrital grains have a coating of iron oxide around them. The development of overgrowths around detrital grains gives rise to a situation in which the grains become tightly welded together, with straight compromise boundaries. The end result of such a cementation process is the total loss of primary porosity. The silica cement may have been precipitated from the circulating ground water, from the pressure solution of detrital quartz grains at their points of contact, or from chert formed as an alteration product of the feldspars.

Carbonate cemented sands are rare in the area of study. Carbonate cement was observed in rocks from the Bogoi Hills and the Jelani areas, as very thin horizons. In certain instances the carbonate cement results in a considerable loss of grain to grain contacts between detrital grains.

Iron oxides and clay minerals with sericite form the matrix in most of the sandstones. Iron oxide matrix is more abundant in rocks from the top of the sequence in the Jelani area where they impart a brownish red colour to the rocks.

5.10 CLAY MINERALOGY

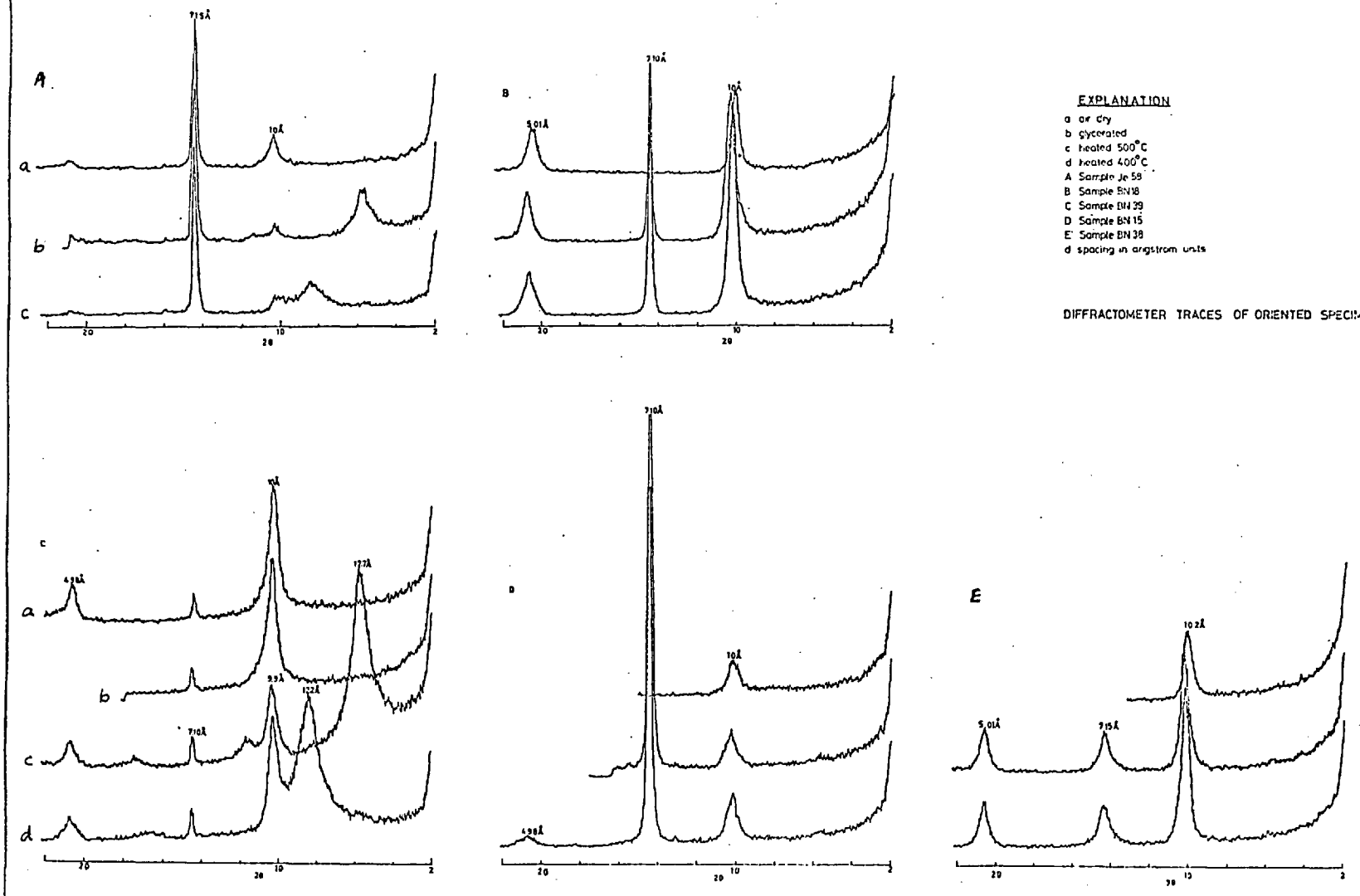
In all the samples that were used in the clay mineral analysis the predominant clay minerals are kaolinite and illite, whilst montmorillonite was observed in only two samples. Figures 5.10 and 5.11 illustrate the X-ray diffractogram traces of some of the samples analysed. The different varieties of clay minerals will now be discussed.

Kaolinite was identified by its characteristic diffraction peaks at 7.1Å, 3.6Å and 2.3Å. The identification of kaolinite could become complicated by the presence of chlorite. However, chlorite was not detected in any of the samples. The presence of chlorite could be confirmed in a sample by heating to 550°C which destroys the reflections due to kaolinite, and changes the structure of chlorite. The presence of chlorite could also be detected by boiling the sample in 10% hydrochloric acid which completely destroys it. The identification of kaolinite was also aided by petrographic microscope study of material scratched from the hand specimens of the samples.

Illite was identified by its characteristic basal reflection at 10Å. Illite was also noted in samples during thin sectional studies in association with kaolinite and sericite. The identification of illite under the microscope was based on its higher interference colours compared with those of kaolinite which are grey.

Montmorillonite diffraction peaks occurred at 12Å and 12.2Å. The identification of montmorillonite was confirmed by heat treatment of the two samples from which its occurrence was observed. In both instances the presence of montmorillonite was indicated by a basal (001) peak at 12Å in the air dried untreated specimen. Glyceration of the samples produced

Fig 510



EXPLANATION

a air dry
b glycerated
c heated 500°C
d heated 400°C
A Sample Je 58
B Sample BN 18
C Sample BN 39
D Sample BN 15
E Sample BN 38
d spacing in angstrom units

DIFFRACTOMETER TRACES OF ORIENTED SPECIMENS

an expanded spacing of 17.7Å and 18Å respectively for both samples (fig. 5.10). These peaks disappeared on heating to 440°C and 550°C for a period of two hours. The collapse of the montmorillonite structure on heating resulted in an increase in the illite peak in both cases, indcating the presence and contribution to the illite peak of "collapsed" montmorillonite.

Discussion

The presence of clay minerals in the Bima Sandstones is attributed to the alteration of the feldspars. The association between the feldspars, especially microcline and untwinned varieties, and the clay minerals and chert indicates that the feldspars have altered into the clay minerals and chert according to the following equation:

Feldspar ⟶ clay minerals + chert (1)

Kaolinite can also alter into illite in the presence of

K^+ ions: Kaolinite + K^+ ⟶ illite (2).

Montmorillonite can also form as an alteration product of other minerals such as glass:

Glass ⟶ montmorillonite + chert (3)

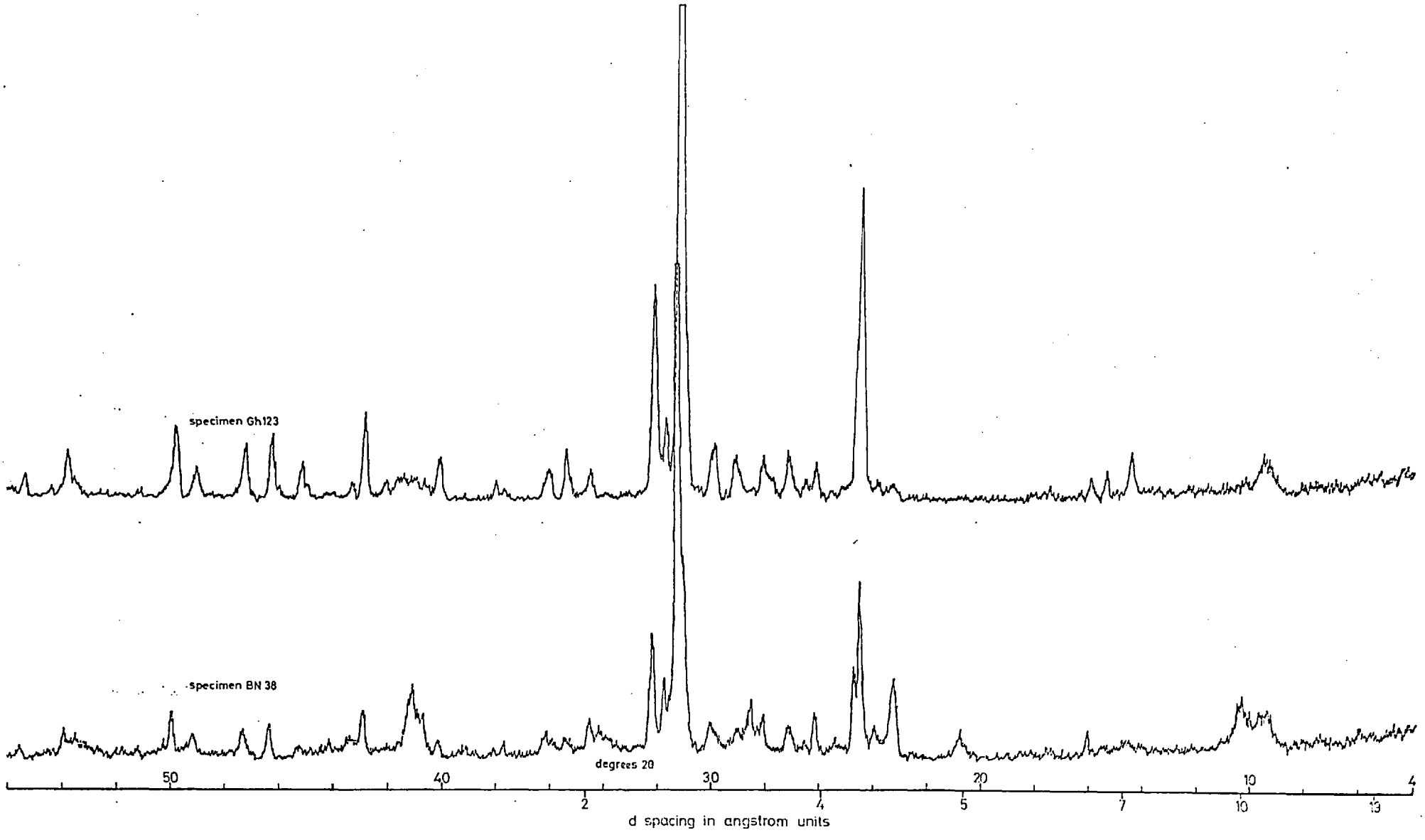
Feldspars can also alter directly into sericite:

Feldspars ⟶ sericite (4)

Equations (1), (2) and (3), and equation (4) belong to the redoxomorphic and the phyllomorphic diagenetic stages respectively, of Dapples (1962).

Another consideration which strongly favours the clay minerals as secondary products of feldspar alteration is the proposed environment of deposition and the strength of the currents which deposited the Bima

Fig. 5.11.



DIFFRACTION TRACES OF RANDOMLY ORIENTED POWDER SPECIMEN IN CAVITY MOUNT

Sandstones. These rocks are channel deposits of strong traction currents which would preclude the deposition of materials finer than fine-grained sand.

CHAPTER 6

DIAGENESIS OF THE BIMA SANDSTONES

6.1 INTRODUCTION

The term diagenesis was first used by Von Gumbel in 1888 and it has been variously defined by many workers since then. Dunoyer de Segonzac (1968) gives a complete review, history and the terminology of the subject of diagenesis; and Pettijohn, et al. (1972) have a complete dossier on the concepts for which the term diagenesis has been used. Correns (1950) states that no simple and non-arbitrary dividing line exists between diagenesis and metamorphism in terms of pressure and temperature. Sujkowski (1958) suggests that diagenesis is the introduction to metamorphism and that diagenesis cannot be separated from weathering. The definition of the term diagenesis as stated by Sujkowski (1958) and Pettijohn, et al. (1972) that diagenesis includes all those processes, physical and chemical, which play on a sediment from the time it is deposited to the onset of metamorphism, is the one adopted in this study. As no published information is available on the diagenetic changes of the Bima Sandstones, it is proposed to describe and discuss in some detail the processes which have changed the primary composition and textures of the sandstones.

6.2 DIAGENETIC MODIFICATIONS TO THE SANDSTONES

6.2.1 General statement

The diagenetic changes of the Bima Sandstones will now be discussed under the following headings:

a) Physical changes

b) Physiochemical changes

c) Chemical changes

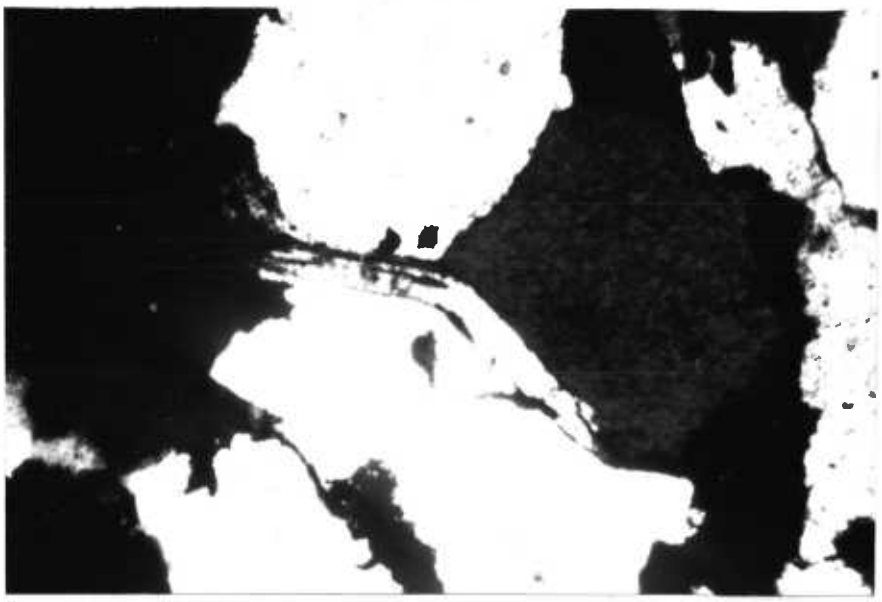


Plate 6.I.I. A mica grain that has been crushed and tightly wedged between detrital grains of quartz and feldspars. Note the straight contact between the mica and the quartz grains. X50, crossed nicols.

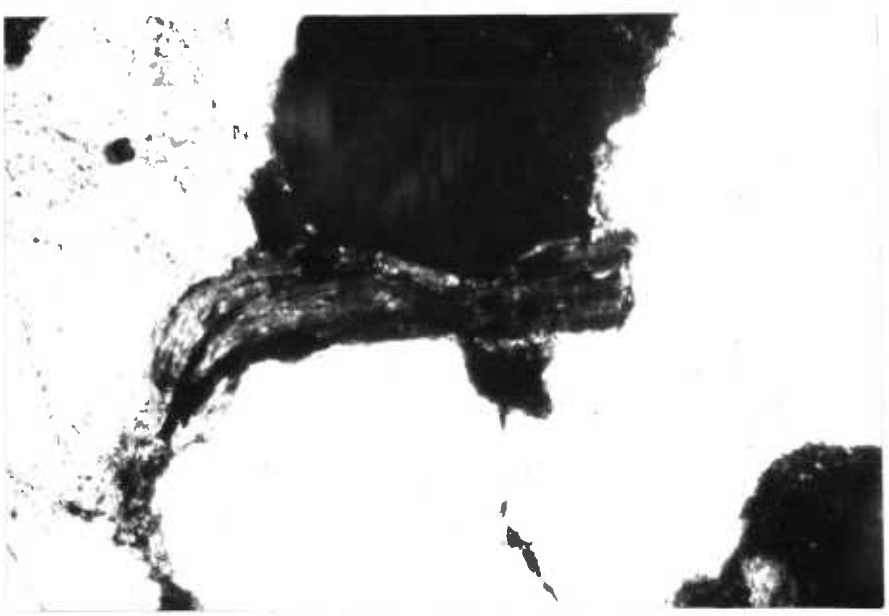


Plate 6.I.2. A detrital grain of mica that is tightly wedged between other detrital grains. Note the jagged contact indicating the alteration of the mica along its point of contact with the quartz and feldspar grains. The alteration product is clay minerals and sericite. Crossed nicols. X50.

6.3 PHYSICAL CHANGES

The physical changes observed in the Bima Sandstones in the area of study are grouped into two major events:

(i) Grain bending

(ii) Grain fracturing.

6.3.1 Grain bending

This phenomenon is most commonly exhibited by detrital grains of muscovite. Long flakes of musoovite in contact with other detrital grains such as quartz become bent around the quartz grains (Plates 6.1.1 and 6.1.2). In many instances the muscovite flakes are bent into a smooth S-shaped curves while in other instances bending is very severe, and distinct kinks are formed at the points of bending. This phenomenon results in grain fracture in other cases. The forces which probably gave rise to the bending of the muscovite flakes are thought to be associated with pressure as a result of the compaction of the sediments. The muscovite flakes respond to compaction pressure by bending. The flakes remain straight at points where they are in contact with detrital grains, and at points where the flakes are not in contact with any detrital grain they yield to the forces of compaction by bending. Occasionally the ends of muscovite flakes where they abutt on detrital grains show a splaying out. This may be as a result of pressure normal to the long axis of the flake.

6.3.2 Grain fracturing

Grain fracturing is commonly developed in most of the detrital grains. Among the mica flakes severe cases of grain bending culminating in fracturing is common. Sometimes, the muscovite flakes are split along their lengths and are invaded by either clay minerals or carbonates

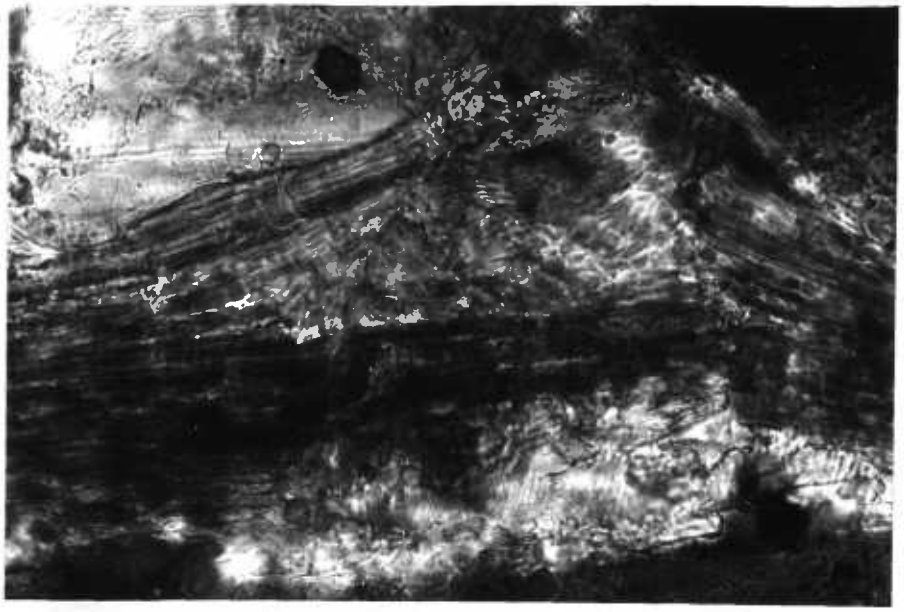


Plate 6.2.I. A fragment of detrital mica that has been fractured, split and broken. At the point of fracture the mica has been invaded by clay minerals. Crossed nicols. X50.

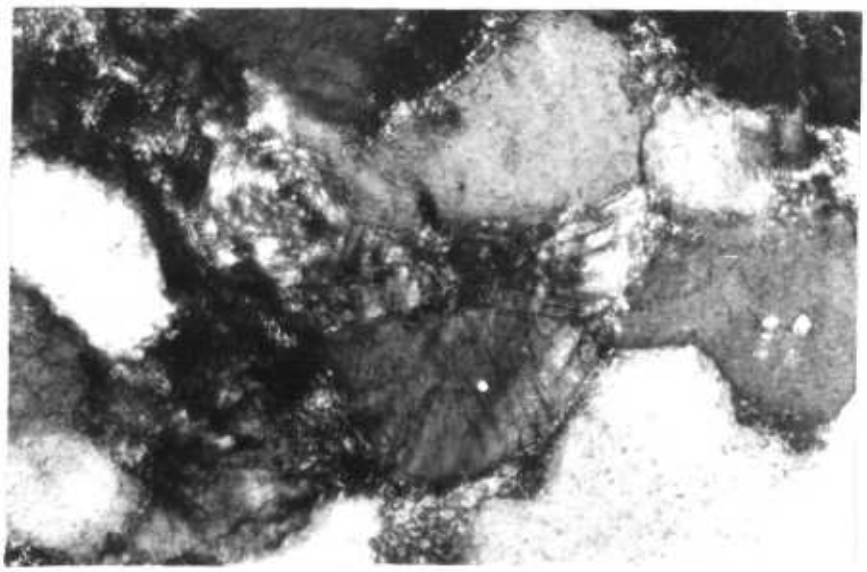


Plate 6.2.2. A detrital grain of mica that has been split along its lenght at both ends and intruded by fragments of quartz and partially altering into clay minerals. Cross nicols. X50.

(Plates 6.2.1, 6.2.2, and 6.3.1).

Fracturing is also exhibited by both plagioclase and microcline feldspars. Fracturing in these grains usually manifests itself by the displacement the twin lamellae or an invasion of the fracture zone by carbonate or silica cement. Rotation of the fragments of the fractured feldspar occurs and it is indicated by lack of uniform extinction behaviour by the fragments.

Fractured quartz grains usually show a distinct displacement and small variations in extinction as a result of rotation of the fractured halves (Plate 6.3.2).

In many instances intense fracturing results in the fragmentation of the mineral grains into several small pieces which exhibit different orientation and extinction behaviour.

6.4 PHYSIO-CHEMICAL CHANGES

These changes involve the interplay of both physical and chemical processes on mineral grains. In all cases observed in the Bima Sandstones the physical processes occur before the chemical processes, and they both impart a new texture and association between detrital grains. The following physio-chemical changes were observed in the samples examined in this study:

(i) Introduction of cement along fractures.

(ii) Introduction of cement along cleavages.

(iii) Contacts between minerals grains.



Plate 6.3.I. A large fragment of detrital mica that has been fractured and split, showing alteration into clay minerals and sericite. Note the mass of clay minerals and sericite in continuity with the mica at its upper end. Crossed nicols. X50.



Plate 6.3.2. Two halves of a fractured grain of quartz (dark areas in the center and right of the photomicrograph) showing slight difference in extinction behaviour under crossed nicols as a result of slight rotation of the two halves with respect to each other. Both halves are separated by a mass of clay minerals and sericite. Crossed nicols. X50.

6.4.1 Introduction of cement along fractures

The healing of cracks and fractured grains is a common occurrence in the Bima Sandstones. This phenomenon is more obvious when the cementing material is of a different chemical composition from the detrital grains. The common fracture cementing material are carbonates, silica, sericite and clay minerals. In the carbonate cemented rocks many grains with fractures intruded by calcite abound (Plates 6.4.1 and 6.4.2). It is probable that the force of crystallization of the calcite cement in these fractures is responsible for the pushing apart of the fractured halves of mineral grains in those cases where a wide gap exists between fragments of fractured grains. Sometimes the fractures are first invaded by clay minerals and sericite some of which are later replaced by calcite in some of the carbonate cemented rocks (Plate 6.4.2). Introduction of cement along fractures is more common in the feldspars than in quartz and micas. In quartz grains which have been fractured and recemented by quartz the cement which heals the fracture is usually found to have a different extinction behaviour from the fractured grains.

6.4.2 Grain contacts

Different types of grain contacts are exhibited by the mineral grains present in the Bima Sandstones. Concave-convex, sutured, straight and wavy contacts are the most common types. All four types of contacts between grains are present in every slide studied petrographically regardless of whether secondary overgrowth materials on the detrital grains is made obvious or not by the presence of a dust ring around the detrital grains.

(i) Concave-Convex contact between grains (Plate 6.5.1)

Taylor (1950) suggested that this type of contact between grains may

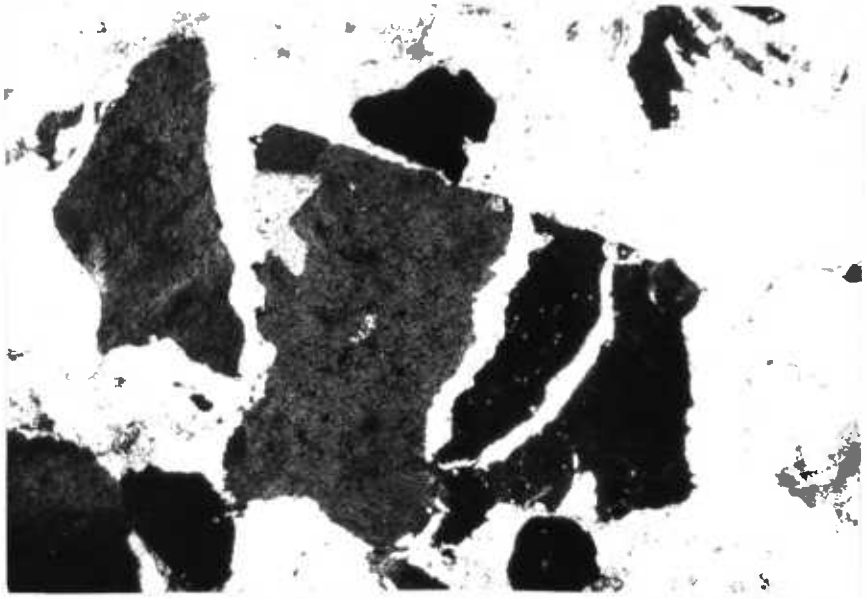


Plate 6.4.I. A detrital grain of quartz fractured and invaded by calcite. Both halves of the original quartz grain still show the same extinction behaviour under crossed nicols. X50.

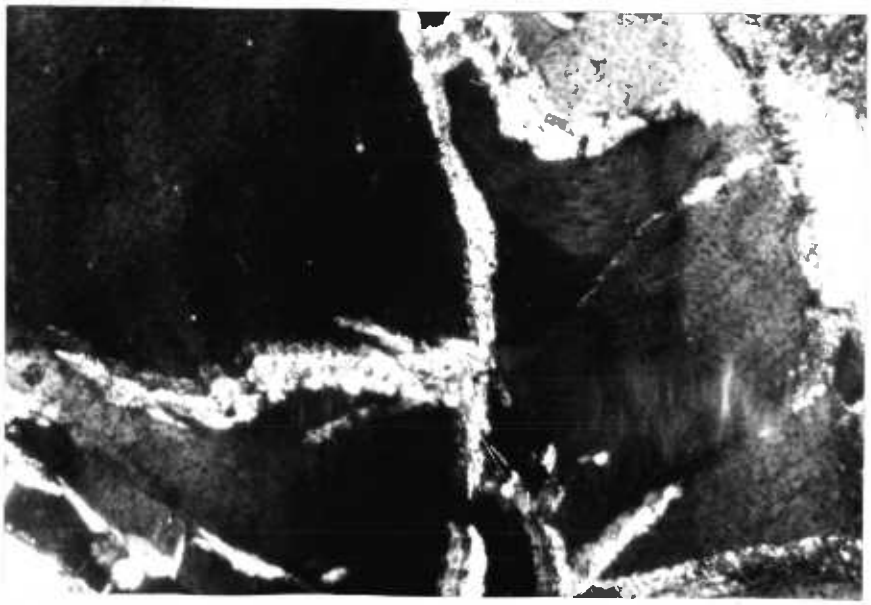


Plate 6.4.2. A large detrital grain of quartz that has been fractured and subsequently invaded by calcite along the fractures. The development of re-entrants on the detrital grain indicates that the calcite is slowly replacing the detrital grain. Crossed nicols. X50.

be produced by solid flow of the yielding grain or by solution at the points of contact between grains. Pressure is believed to be the major factor in initiating concave-convex contact between grains. The pressure developed as a result of compaction of sands gives rise to solution of grains at their points of contact with other grains. The dissolved material may be moved to adjacent areas with lower pressure conditions and deposited as secondary cement on detrital grains.

The factors that may determine which grain shall yield were given by Taylor (1950) as crystallographic orientation, chemical structure, physical structure, original shape, type of pressure to which the grains are subjected, and the amount of heat that has been supplied to the sand.

(ii) Sutured contacts (Plate 6.5.2)

This type of contact between grains, as observed in the samples of the Bima Sandstones, is developed along the compromise boundary formed by the secondary overgrowth material between two detrital grains. The sutured appearance of the compromise boundary results from the differential growth rates of the secondary overgrowth material on two adjacent detrital grains. Sutured contacts may also arise as a result of local solution at the points of contact between grains. A continuation of this process, under increased pressure and increased differential solution may result in the formation of stylolites (Taylor, 1950). No stylolitic development was observed in any of the samples of the Bima Sandstones.

(iii) Straight contacts

Straight contacts occur between all types of grains and may be related to the matured stage of the development of secondary overgrowths on detrital grains (Plate 6.6.). Perfect straight contacts between mineral grains (Quartz and Feldspars) is usually used as an indication



Plate 6.5.I. Grains of detrital quartz in concavo-convex contacts. Note the termination of the secondary overgrowth material on the other detrital grains. Crossed nicols. X50.

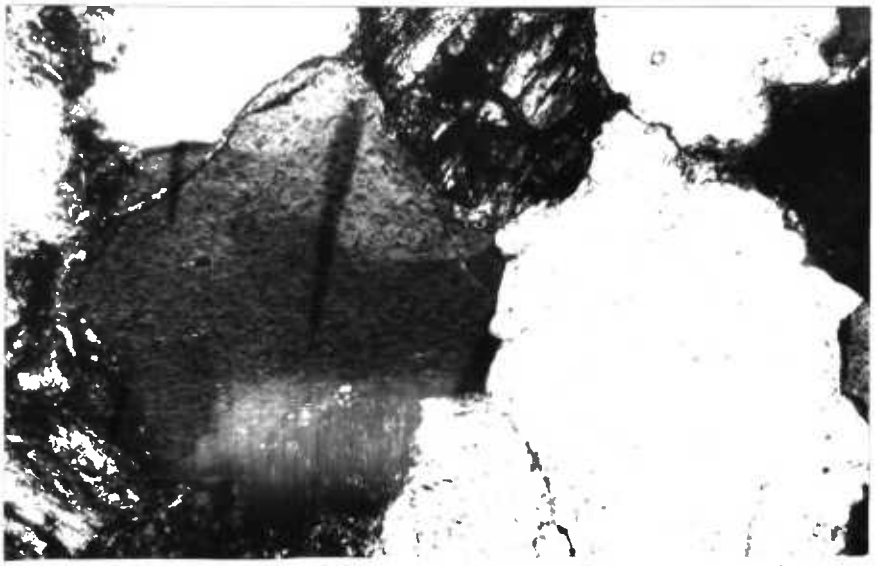


Plate 6.5.2. Partial sutured contacts between quartz grains. The grain on the left of the photomicrograph shown in extinction has secondary overgrowth material around it. Crossed nicols. X50.



Plate 6.6. Fairly straight contacts between detrital grains of quartz. The grain under extinction has secondary overgrowth material around it. Crossed nicols. X50.

of the presence of secondary overgrowth materials in rocks that do not show a dust ring around the detrital grains (Shearman, oral communication).

Straight contacts also exist between quartz and feldspar grains, and micas, (Plate 6.7.). Such straight contacts may result from the original packing of grains with straight sides or edges (Taylor, 1950). Straight contacts are formed when yielding grains are pressed against adjacent rigid grains having relatively straight edges.

Introduction of cement along fractures is more common in the feldspars than in quartz and micas. In quartz grains which have been fractured and recemented by quartz the cement which heals the fracture is usually found to have a different extinction behaviour from the fractured grains.

6.4.3 Introduction of cement along cleavages

This phenomenon is restricted to the twinned feldspars. Most of the feldspar grains show signs of alteration. This is manifested by the presence of clay minerals and sericite along twin planes and some fractures within the detrital grain. In silica-cemented rocks the cleavage lines are invaded by clay minerals and sericite which eventually results in the feldspar grains being broken into several fragments. With the introduction of more material along the cleavage lines and coupled with further alteration the broken fragments of feldspar become pushed further apart (Plate 6.8.1). In the carbonate-cemented rocks calcite is the major material invading the cleavage lines of twinned feldspars. The force of crystallization of the calcite eventually pushed the fragments of feldspar apart (Plate 6.8.2).

The intrusion of cement along cleavage lines in micas is not very common. However, in cases where mica grains have been split length-wise

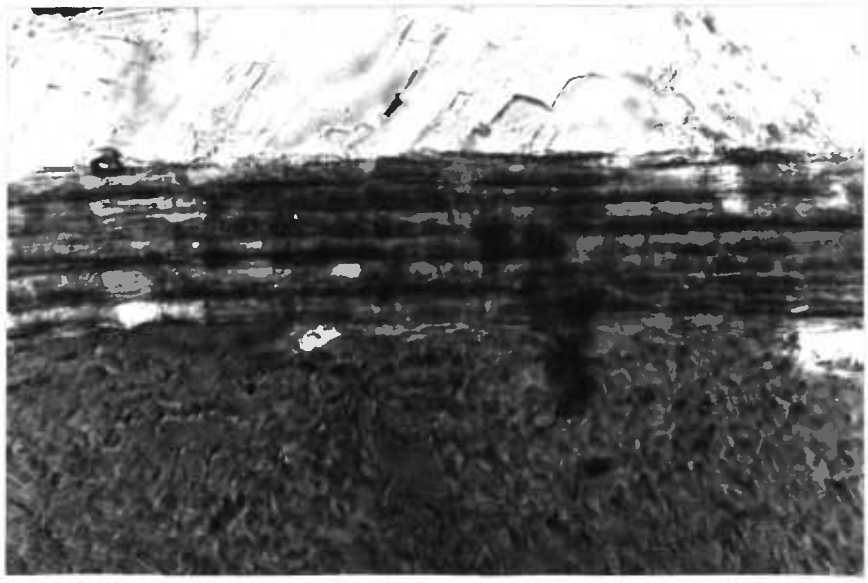


Plate 6.7. Straight contact between a long slender grain of muscovite and two grains of detrital quartz. Crossed nicols. X50.

parallel to cleavage lines calcite was noted to have intruded the fracture zone.

6.5 CHEMICAL CHANGES

The chemical changes which affect any sedimentary rock have a significant influence on the overall texture and mineralogy of the rocks, and are therefore more important than the physical changes. Chemical influences result in the cementation and lithification of the detrital grains, by the precipitation of binding agents. The chemical changes which the Bima Sandstones in the area of study have undergone will now be considered in detail under the heading of the development authigenic minerals, carbonate cementation, and the alteration of detrital mineral grains.

6.5.1 Development of Authigenic Minerals

The development of authigenic minerals is common in the Bima Sandstones and is restricted to detrital quartz among the minerals of the light fraction, and to zircon in the heavy fraction. These secondary materials are usually noticeable as overgrowths on the primary detrital grains. The development of authigenic zircon was discussed in Chapter 4. Mineral authigenesis as it affects the texture of the rocks in pore space reduction will now be discussed and it is restricted to two minerals:

(a) Secondary quartz overgrowths.

(b) Authigenic micas.

6.5.1.1 Secondary quartz overgrowths

The development of secondary quartz overgrowths is fairly common in the Bima Sandstones in the area of study. Its presence is indicated by a thin line of dust which marks out cores of primary quartz material in

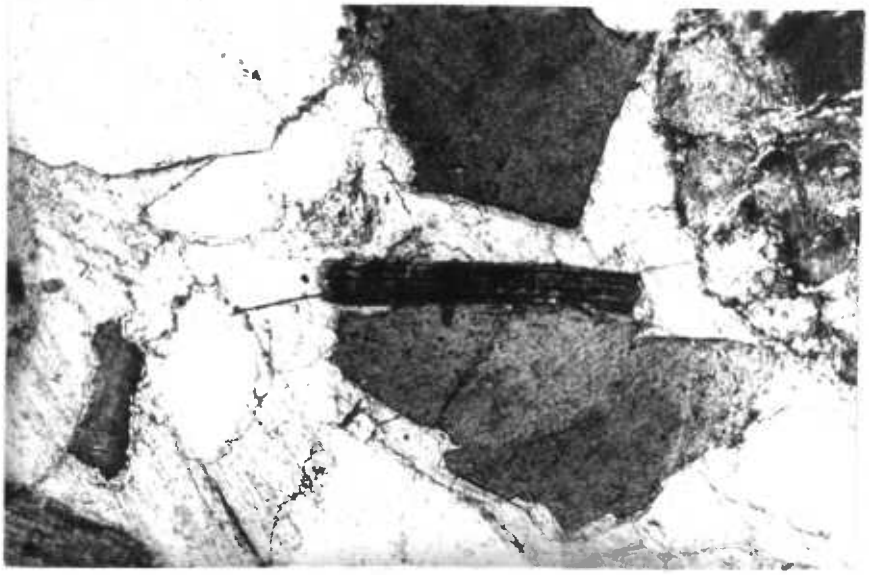


Plate 6.8.I. Calcite intruding into areas between detrital grains. The mica grain has been completely separated from the quartz grain on one side while separation has only been initiated along its contact with the other quartz grain. Crossed nicols. X50.



Plate 6.8.2. Fragments of microcline feldspar (dark areas) invaded by calcite (white areas) and displaced to give different extinction behaviour to the feldspar fragments. Crossed nicols. X50.

a mosaic of quartz crystals, (Plates 6.9.1 and 6.9.2). In areas where a dust line has failed to develop around the detrital grains it is difficult to differentiate between the primary core and the secondary quartz material. Sippel (1968) described the cathodoluminescence method of detecting overgrowth material on such grains. The overgrowth materials are usually added to the detrital grains in optical continuity with them, and it is an attempt by the grains to attain and restore the euhdral crystal form which was lost during the transportation of the sediments to the depositional basin. Waugh (1970) described perfect examples of quartz overgrowths in the Penrith Sandstone of northwest England.

Where secondary quartz overgrowth material of two contiguous grains meet a compromise boundary is formed. This boundary, in the Bima Sandstones, is usually straight. Rocks in which the detrital grains are cemented by secondary quartz resemble a meta-quartzite because the grains are tightly welded together. The development of quartz overgrowths results in a considerable loss of porosity.

Secondary quartz overgrowths are not a common feature of the polycrystalline variety of detrital quartz grains. The development of authigenic quartz as overgrowths on detrital grains appears to be restricted to the moncrystalline varieties, as seen in the petrographic microscope studies. It is probable that the use of cathodo-luminescence techniques may reveal the presence of overgrowth material on the polycrystalline quartz grains.

6.5.1.2 Authigenic micas

One of the products of the alteration of feldspars in the Bima

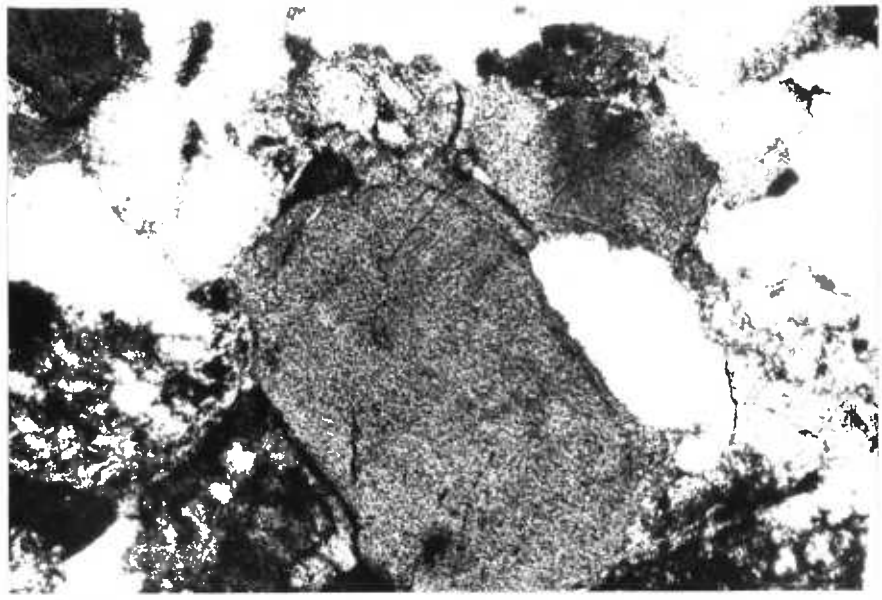


Plate 6.9.I. Detrital grains of quartz showing well developed secondary overgrowth material around a core of primary quartz. Crossed nicols. X50.

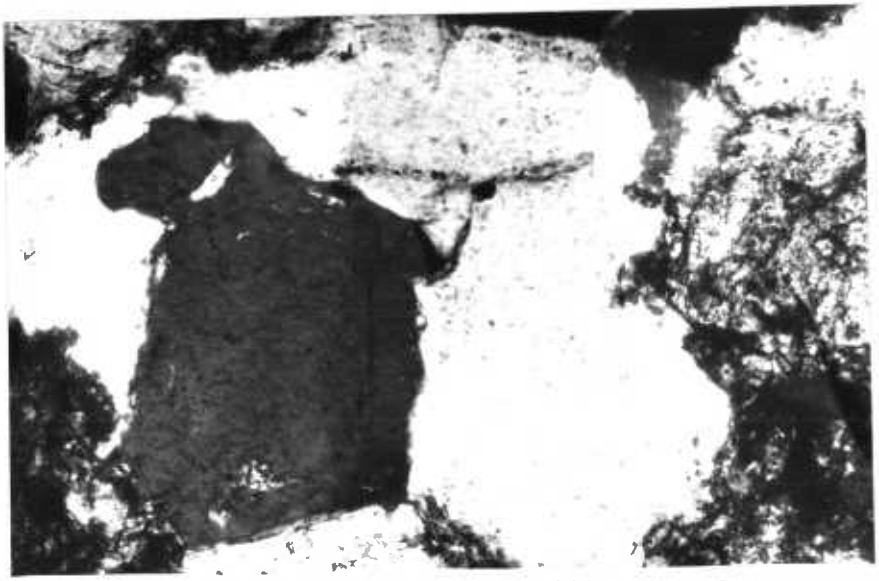


Plate 6.9.2. A detrital quartz grain with secondary overgrowth material. The overgrowth is made evident by the presence of a dust line of iron oxide around the periphery of the primary grain. Crossed nicols. X50.

Sandstones is sericite which is microcrystalline muscovite. In the rock samples used for the petrographic studies of the Bima Sandstones sericite flakes are abundantly developed, and they can be seen in different stages of transformation to musoovite (Plate 6.10.). Authigenic micas are not very common in the Bima Sandstones, but that some of the muscovite flakes in the samples are authigenic can be inferred from the association between sericite and muscovite. In a few instances where sericite "booklets" could be seen there is a high degree of transformation into muscovite. Some of the muscovite could also have formed directly from clay minerals (Dapples, 1962).

6.5.2 Development of carbonate cement

Carbonate-cemented rocks are rare in the area of study. They are restricted to a few horizons in the Bogoi Hills and the Jelani areas. At least four types of carbonate development are recognised in the Bima Sandstones:

(a) Patchy development of carbonate cement.

(b) Formation of carbonate cement with increase in volume.

(c) Poikilotopic development of carbonate.

(d) Carbonate replacement of detrital grains.

6.5.2.1 Patchy development of carbonate cement

In some of the carbonate-cemented rocks of the Bima Sandstones the carbonate cement is only developed in patches. The principal cementing agent in these rocks is silica with a matrix of clay minerals. The development of carbonate cement is rather late in these rocks as their occurrence is within the clay minerals which are formed from the alteration of the feldspars. This fact is further supported by the microcrystalline nature of the carbonate cement in places where it occurs in

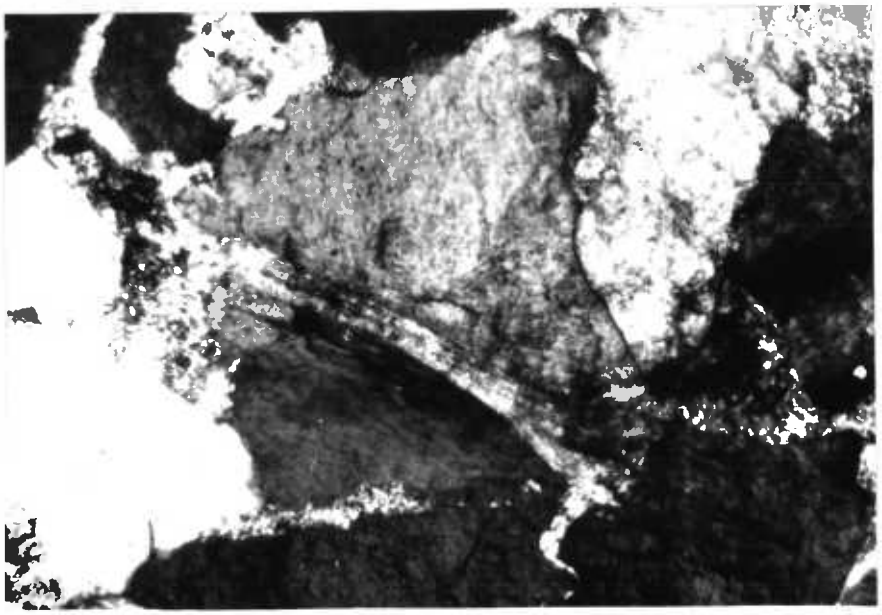


Plate 6.IO. Authigenesis in micas illustrated by a grain of mica which is tightly wedged between two quartz grains along its entire length except at the ends where the mica grain is being added onto by sericite with which it is in contact. Note the splaying out of the mica at the ends. Crossed nicols. X50.

association with the clay minerals (Plate 6.11.). This patchy development of the carbonate cement is therefore not in primary pore spaces.

6.5.2.2 Formation of carbonate cement with increase in volume

This type of carbonate development results in the destruction of the grain-supported framework of the rocks. It is probable that this type of carbonate cementation is a more advanced stage of the patchy type of carbonate cementation. First the carbonate cement is introduced into the rocks in areas of "defeldspathaziation" until the carbonate material completely replaces the clay minerals. This is followed by the fracturing and pushing apart of the detrital grains by the force of crystallization of the carbonate cement. The final stage is represented by a rock (Plate 6.12.) devoid of clay minerals and in which the detrital grains have been fractured and isolated from each other by the carbonate cement. This argument is supported by the presence of authigenic micas and the complete absence of clay minerals in such rocks. The feldspars and clay minerals show re-entrants which have been invaded by the carbonate cement. These re-entrants in the feldspars and the mica flakes probably indicate areas previously occupied by clay minerals which have been replaced by the carbonate cement.

6.5.2.3 Poikilotopic development of carbonate cement

Poikilotopic development of carbonate cement is a common feature of the predominantly carbonate-cemented rocks of the Bima Sandstones. In these rocks the carbonate cement is represented by calcite with its well-developed characteristic twin lamellae (Plate 6.12.2). The poikilotopic texture is developed as a result of a single large crystal of calcite enclosing a number of detrital grains. The continuity of the twin lamellae of the calcite across detrital grains indicates a single crystal of calcite.

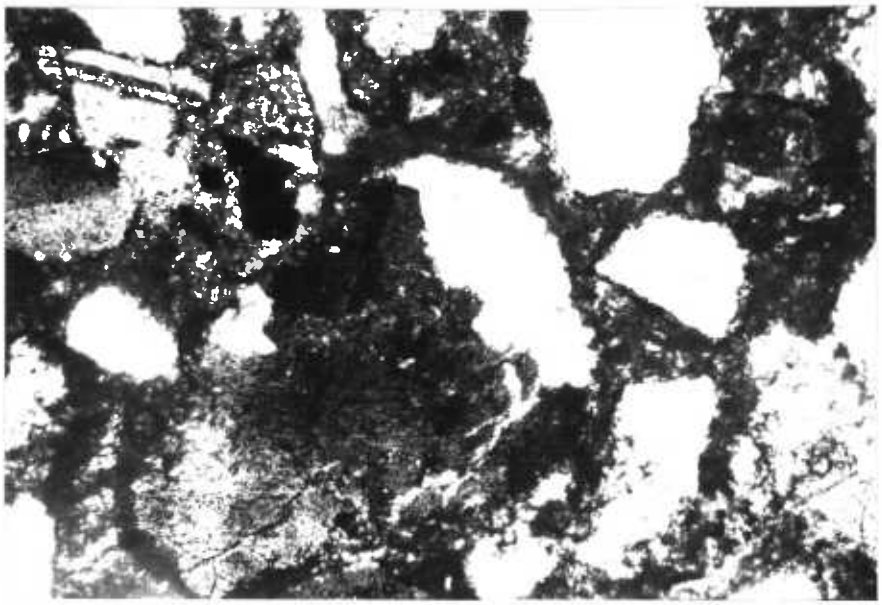


Plate 6.II. Patchy development of calcite (some of the white areas). The cacite first appaers within the alteration products of the feldspars which it replaces. Crossed nicols. X50.

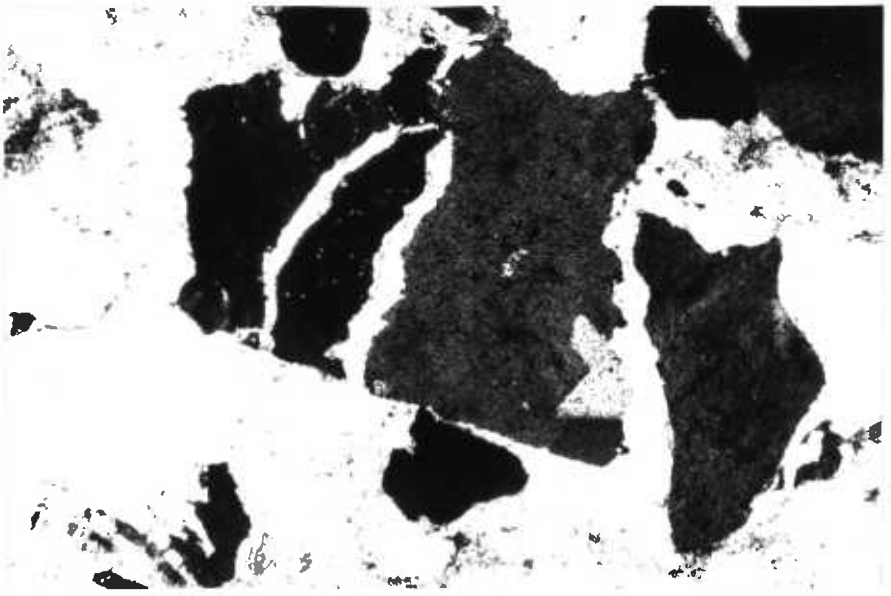


Plate 6.I2.I. Fragments of detrital grains of quartz pushed apart and isolated by calcite cement (white areas). Crossed nicols. X50.

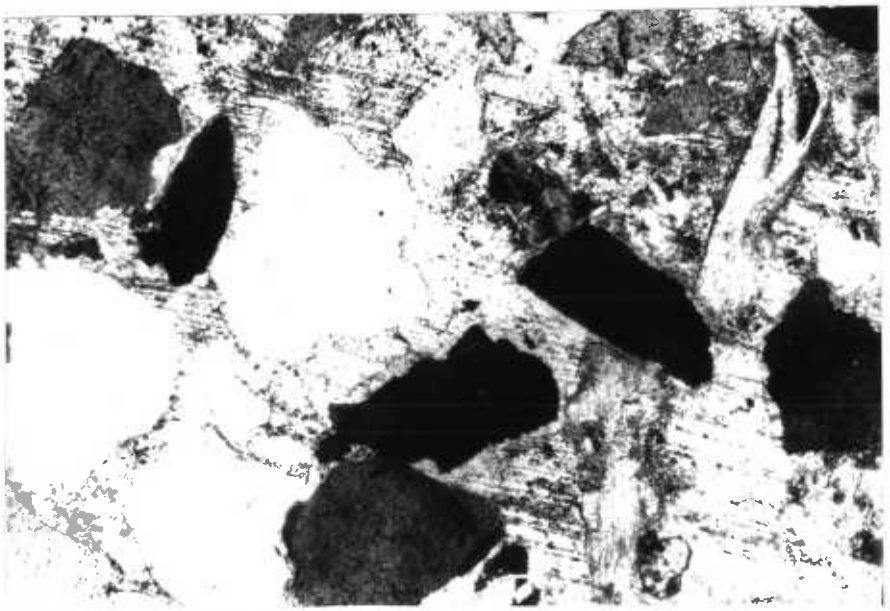


Plate 6.I2.2. Poikilotopic texture. Note several detrital grains floating within a single large crystal of calcite with its characteristic twin lamellae continuous across the detrital grains. Crossed nicols. X50.

Samples in which there is a development of the poikilotopic texture the detrital grains show very little grain to grain contact. Occasionally clusters of detrital grains could be found within the carbonate cement. No secondary overgrowth material was observed on the detrital grains in these rocks. There is a considerable degree of replacement of the detrital grains by carbonates. Some of the feldspars show signs of alteration. This indicates that the carbonate cementation occurred as a result of the replacement of clay minerals from between quartz grains followed by displacive cementation of the quartz grains by calcite. The ultimate product of these processes is the loss of grain to grain contact between detrital grains. Plate 6.15.2 illustrates an altered grain of feldspar with a core of chert plus calcite. The areas occupied by calcite represent the initial amount of clay minerals formed from the alteration of the feldspar, while the bulk of the alteration product is represented by chert.

6.5.2.4 Carbonate replacement of detrital grains

The replacement of detrital grains by carbonate material is a fairly common feature in the carbonate-cemented sandstones, (fig. 6.13.1). The feldspars show a greater degree of susceptibility to replacement by calcite than quartz or muscovite. Grains of feldspars which have been completely replaced but which still maintain the crystal boundary abound in many of the samples. Sometimes, small islands of feldspathic material form relicts within the calcite which has replaced the detrital grain. The high susceptibility of feldspars (especially the twinned varieties) to calcite replacement leads to their being improvished in the rocks.

The replacement of quartz by calcite is less common and this is

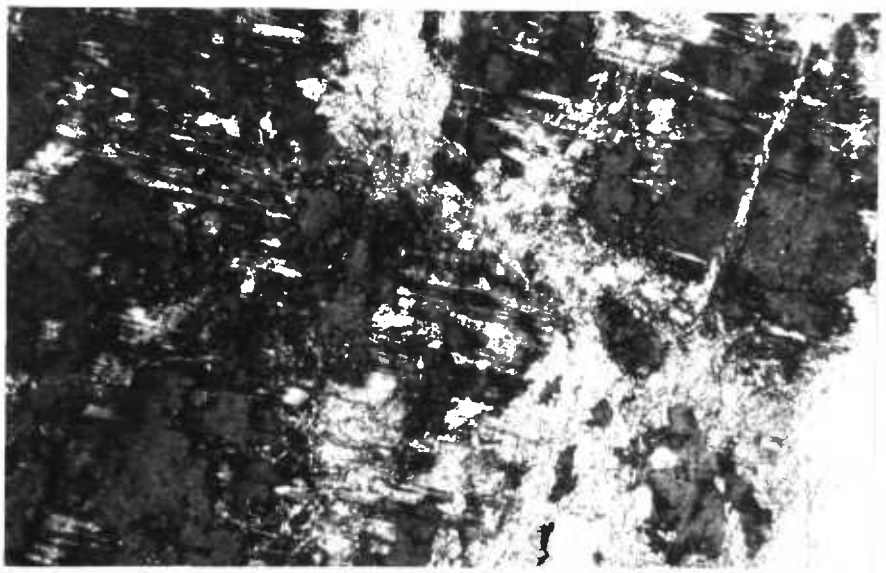


Plate 6.I3.I. A large detrital grain of microcline being replaced by calcite (white areas). The twin lamellae of the microcline are still visible and have been attacked by the invading calcite. Crossed nicols. X50.

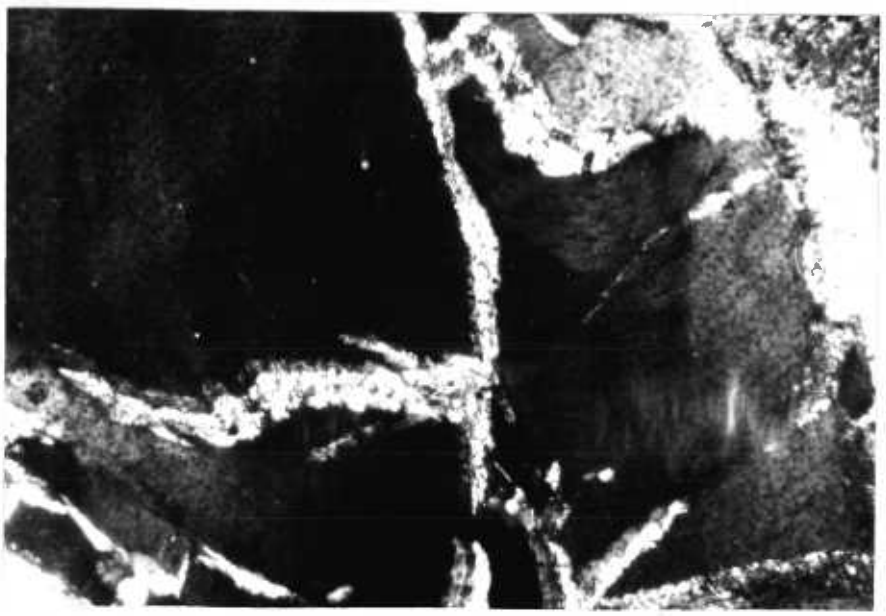


Plate 6.I3.2. Detrital grain of quartz invaded by calcite along cracks and fractures. Crossed nicols. X50.

probably due to the higher stability of quartz compared with feldspars. The replacement of quartz by calcite is initiated along cracks in the detrital quartz grain. The introduction of more calcite into the crack leads to the eventual breakup of the quartz to smaller fragments which show re-entrants invaded by calcite (Plates 6.13.2 and 6.14.).

6.5.3 Alteration of detrital mineral grains

The role played by chemical alteration in the diagenesis of sediments is the appearance of completely new minerals by the reaction of circulating formation water with detrital mineral grains. In the Bima Sandstones the final products of such an event are kaolinite, illite, sericite and chert from feldspars.

(a) Development of kaolinite and illite

Clay minerals were observed in all the sandstones studied from the Bima samples. X-ray analysis of the clay minerals indicated the presence of kaolinite and illite. The sharp peaks of the clay minerals in the X-ray diffractogram traces shows that they were fairly well crystallized.

The clay minerals usually occur in association with sericite and chert within grains of altered feldspars. The clay minerals appear as irregularly shaped masses of low birefringence in association with brownish flakes of sericite.

(b) Development of sericite

In all the slides studied sericite is always associated with clay minerals and/or chert. It is actually difficult to separate the clay minerals and sericite except by their markedly different optical properties. The sericite flakes usually form aggregates or "booklets" along twin planes of feldspars and within authigenic chert crystals. In some

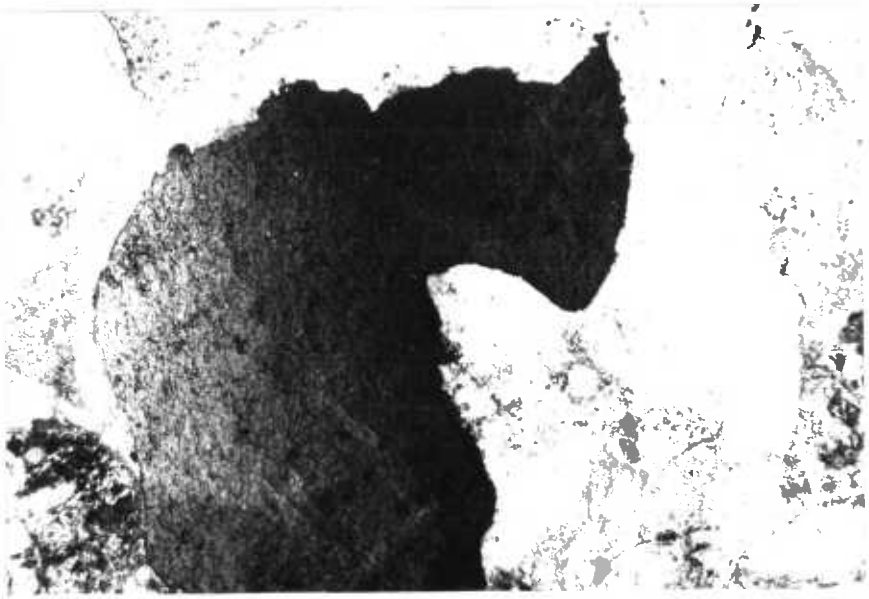


Plate 6.I4. A grain of detrital quartz showing re-entrants where calcite cement is encroaching on it. Crossed nicols. X50.

samples large "booklets" of sericite aggregate to form muscovite (Plate 6.15.1).

(c) Development of chert

Within the masses of clay minerals and sericite flakes small aggregates of microcrystalline quartz could be seen. Sometimes these microcrystals of quartz are dispersed all over the altering feldspar grain. Where the microcrystals are the dominant alteration product of the feldspars they form visible crystals of chert, (Plate 6.15.2). Occasionally, the chert crystals are elongated into interspaces between detrital quartz grains, and the presence of chert crystals altering into coarsely monocrystalline quartz is not uncommon in the Bima samples.

6.6 Mineral paragenesis

The mineral paragenesis in the Bima Sandstones in the area of study is fairly a straight forward one. In the silica-cemented sandstones the earliest developed mineral is silica as secondary overgrowths on detrital quartz grains. This event was followed by the alteration of the feldspars into clay minerals, sericite and chert. The final stage of this alteration event is the development of muscovite from sericite and clay minerals.

In the calcereous sandstones the introduction of calcite postdates the development of alteration products in the feldspars. This fact is indicated by the presence of calcite in areas within altered grains of feldspar and the presence of relicts of clay materials within crystals of calcite. Although the clay minerals have been replaced by the calcite cement their microcrystalline form is still preserved in the calcite.



Plate 6.I5.I. Authigenic mica forming from an aggregate of sericite and adding on to the ends of a grain of detrital mica wedged tightly between quartz grains. Crossed nicols. X50.

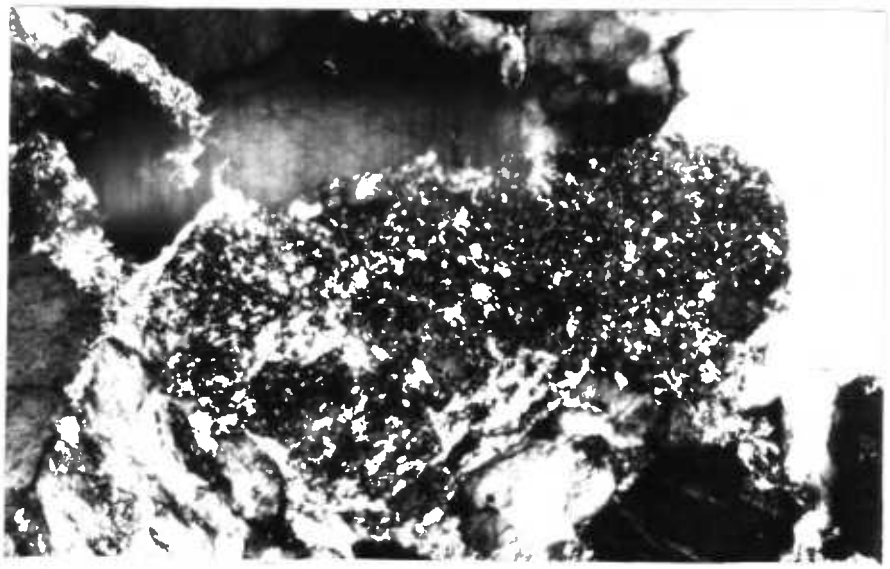


Plate 6.I5.2. Microcrystalline quartz (chert) forming as an alteration product of feldspar. Remnants of the feldspar are still visible around the chert. Crossed nicols. X50.

Dapples (1967) suggested that such clay residues within the calcite can be recovered on solution of the carbonate.

6.7 STAGES OF DIAGENESIS

Three stages of diagenesis can be recognised in the Bima Sandstones in the area of study. The alteration of the feldspars to clay minerals and chert is characteristic of the redoxomorphic stage of Dapples (1962), while the development of quartz overgrowths on detrital quartz grains and the replacement of silica by carbonates belong to the locomorphic stage. The phyllomorphic stage is characterised by the formation of muscovite from clay minerals. These three stages of diagenesis are referred to by Fairbridge (1967) and Siever (1959) as syndiagenesis, anadiagenesis, and epidiagenesis; and early, middle, and late stages respectively.

6.8 ORIGIN AND FORMATION OF MINERALS

6.8.1 The formation of secondary quartz

Secondary quartz as overgrowths on detrital grains forms the primary cementing material in most of the sandstones. From the mineral paragenetic sequence established for these rocks it is believed that the development of silica cement is early in the diagenetic history of the rocks. This view confirms Dapples' (1959) thesis that silica diagenesis is characterised by the formation of quartz overgrowths during early diagenesis. During early burial the interstitial fluids are in contact with individual grains long enough for the two to react, and the pH and temperature conditions of the environment are conducive for the precipitation of silica from the interstitial fluids. Many authors have put forward several possible sources for the silica precipitated as

overgrowths on detrital quartz. Among these authors, Waldshmidt (1941), Heald (1956), Thomson (1959) and Fuchtbauer (1967), held the view that silica was derived from the pressure solution of detrital grains. Goldstein (1948) believed that fine-grained quartz may be the source of the silica, while other workers such as Siever (1959) considered groundwater to be a major contributor of the silica found in sandstones. Von Engelhardt (1967) held the view that silica could be transported in solution from deeper areas and precipitated as quartz near the surface where lower temperature conditions obtain.

The poor development of concave-convex contacts between detrital grains discredits pressure solution as an important source of the silica in the Bima Sandstones in the area of study. The abundance of quartz overgrowths on detrital grains point to the other sources, which supplied silica in copious amounts, for the derivation of the silica. Possible sources for the silica are: (a) the decomposition of feldspars, and alteration of clay minerals; (b) the precipitation of silica from circulating interstitial waters, and silica-rich solutions derived from the compaction of shales in adjacent areas.

In the Bima Sandstones silica cementation is extensive resulting in the interference of the overgrowths with each other such that euhedral faces are rare, and straight line contacts (Compromise boundaries) are developed between grains.

Two major stages are proposed for the secondary development of quartz. The first stage is that of the formation of quartz as overgrowths on detrital grains. The silica involved in this development is

believed to have been precipitated from silica-rich interstitial fluids. The second stage of the formation of secondary quartz is the decomposition of the feldspars. This process resulted largely in the formation of chert which is seen in close association with clay minerals and sericite within decomposing grains of feldspar. This source may also have contributed to the silica precipitated on detrital quartz grains as overgrowths.

6.8.2 The formation of clay minerals

Clay minerals abound in the Bima Sandstones in the area of study. These clay minerals are known from petrographic and X-ray studies to be mainly kaolinite and illite. The petrographic evidence for authigenic origin of the clay mineral is unequivocal as a grain that has all of the geometry of a detrital twinned fragment of feldspar is seen to be composed of entirely small aggregates of kaolinite, illite and sericite. Such a composition would be highly unlikely to survive as an original detrital grain. Dapples (1962) Von Engelhardt (1967), Pettijohn et al. (1972) and Hawkins (1972) mentioned similar associations between clay minerals and feldspars in sandstones. McLaughlin (1955) suggested that a weathering series exists whereby K-feldspars alter into sericite, illite and finally kaolinite.

The chemical conditions which tend to favour the kaolinization of feldspars have been stated by Pettijohn et al. (1972) to be the presence of a groundwater low in dissolved solids, including being low in K^+ and H_4Sio_4 and having a relatively low pH that is probably controlled by dissolved H_2CO3. Such conditions probably exist a few hundred meters of the surface and they are thought to have existed in the Bima Sandstones during the kaolinization of the feldspars.

The montmorillonite present in some of the samples may be related to the late ?Tertiary Volcanicity associated with the Bima Sandstones. Montmorillonite is known to form from volcanic glass (Dapples, 1962; Pettijohn, et al, 1972).

6.8.3 The formation of authigenic micas

The predominant variety of mica present, both as detrital grains and as secondary derivatives, is muscovite. The evidence for the authigenic origin of some of the muscovite flakes is manifested by the association between sericite flakes and well-formed flakes of muscovite that grade into clay minerals and/or well developed sericite clusters. The crystallization of muscovite from clay minerals is favoured by an oxidising environment and generally neutral solutions (Dapples, 1967). The complete absence of biotite in the Bima Sandstones in the area of study indicates that the prevalent conditions were oxidising and that associated solutions were generally neutral.

6.8.4 The formation of carbonate cement

Petrographic staining techniques (Evamy, 1969) indicated that the carbonate cementing material in the Bima Sandstones in the area of study is calcite. It is found both as uniformly and patchily distributed cement in the samples in which carbonate cement is present. Poikilotopic textures are present in some of the samples in which detrital grains of quartz and feldspar "float" in a single large crystal of calcite.

A dual source is suggested for the calcite cement of the Bima Sandstones in the area of study. In the oyster-beds and pelecypod

containing horizons in the Bogoi Hills area the calcite cement is thought to have been derived from the pressure solution of the oyster and pelecypod shells. Petrographic evidence for this derivation of the calcite from this source is found in the rock samples. The shells are found to have a concave-convex relationship with the detrital grains (espeically quartz) with which they are in contact. The shells being less resistant than quartz have yielded and dissolved to provide the calcite which is redeposited as secondary cement within the same beds.

It is envisaged that the decomposition of the feldspars preceded the solution of the shells, and that the calcite cement first started to precipitate within the clay minerals, formed by the alteration of the feldspars. by a replacement process.

In the other sandstones in which calcite forms the secondary cementing material the source of the calcite is believed to be in localised Ca-rich groundwaters. This may be the reason for the localised effects of the carbonate cementation.

6.8.5 The replacement of detrital grains by calcite

The replacement of the detrital grains of quartz and feldspars by calcite is of common occurrence in the carbonate cemented sandstones. The replacement reaction is characterised by the presence of etched detrital grains and relicts of quartz and feldspars in the calcite cement. Detrital grains in different stages of replacement are abundant in samples. Such replacement reactions have been noted by Greensmith (1957) and Dapples (1962-1967). The replacement of feldspar by calcite is not restricted to the potash varieities as plagioclase feldspars are

also involved in this phenomenon in the Bima Sandstones samples observed in the area of study. The replacement of this type is interpreted as resulting from a process by which Ca^{+} and $CO3^{-}$ rich solutions destroy the feldspar lattice by causing the silica tetrahedral units to go into solution under the high pH which characterises calcite-precipitating solution (Dapples 1967).

6.8.5 The replacement of clay minerals by calcite

Calcite replacement of the authigenic clay minerals is an obvious phenomenon in the carbonate cemented sandstones. Sometimes this replacement is so complete as to give the impression that the rocks involved never contained clay minerals. In other instances the clay remains as a residue within the calcite. Dapples (1967) suggested that the mechanism for such a replacement may lie in the fact that illite and kaolinite (in the case of some greywackes) are flocculated by Ca-ions and occupy less space which is filled by the precipitated carbonate.

6.9 CONCLUSION

The diagenetic sequence proposed for the Bima Sandstones in the area of study follows the process of the development of the quartz overgrowths on detrital quartz grains. The decomposition of the feldspars to form clay minerals and chert is followed by the precipitation of calcite cement in the carbonate cemented sands and the formation of muscovite from the clay minerals.

The silica necessary for the development of the secondary overgrowths is believed to have been brought chiefly by silica-rich groundwaters and from the silica released by the decomposition of the feldspars. The source of the calcite is the shells in the shell-bearing sands and

pressure solution is believed to be the mechanism responsible for the release of the calcite from the shells. Another source of calcite is localised calcium and carbonate-rich groundwaters. The nature of the proposed sources of the calcite is believed to be responsible for the localised occurrence of the calcareous sands in the area of study.

CHAPTER 7

SEDIMENTARY STRUCTURES

7.1 INTRODUCTION

The sedimentary structures observed and recorded in the field are separated into two major groups:

(a) Primary sedimentary structures, and

(b) Secondary sedimentary structures.

The primary sedimentary structures are those which are inherent in the rocks containing them, and are a result of the depositional processes. These structures are therefore described as syndepositional structures. The secondary sedimentary structures were formed after the rocks had been laid down and are therefore postdepositional.

7.2 PRIMARY SEDIMENTARY STRUCTURES

Four distinct types of primary sedimentary structures characterise the Bima Sandstones in the area of study. These primary stratification types have been described from modern and ancient fluvial sediments, by many authors: Harms, Mackenzie and McCubbins (1963) in deposits of the Red River; Frazier and Osanik (1961) in deposits of the Mississippi River; Jones (1962) in the Bima Sandstones in the Yola and Lau areas; and Selley (1963) in the Torridonian rocks of northwest Britain. The nomenclature used in the present study is descriptive and is based on the attitude of the strata and the shapes of the sets of strata. The term cross-stratification is used where the laminae are inclined at an angle to the horizontal and to the boundaries of the sets containing them, and such modifying adjectives as trough and tabular or planar are used to denote

the shape of the sets. Horizontal stratification is used to describe sets of laminae in sand-sized material within which the laminae are horizontal or nearly so. Parallel stratification is used to describe sets of laminae in clay- and silt-sized material in which the laminae are parallel to the bounding surfaces of the sets.

The following primary sedimentary structures observed and described in the Bima Sandstones in the area of study will now be discussed:

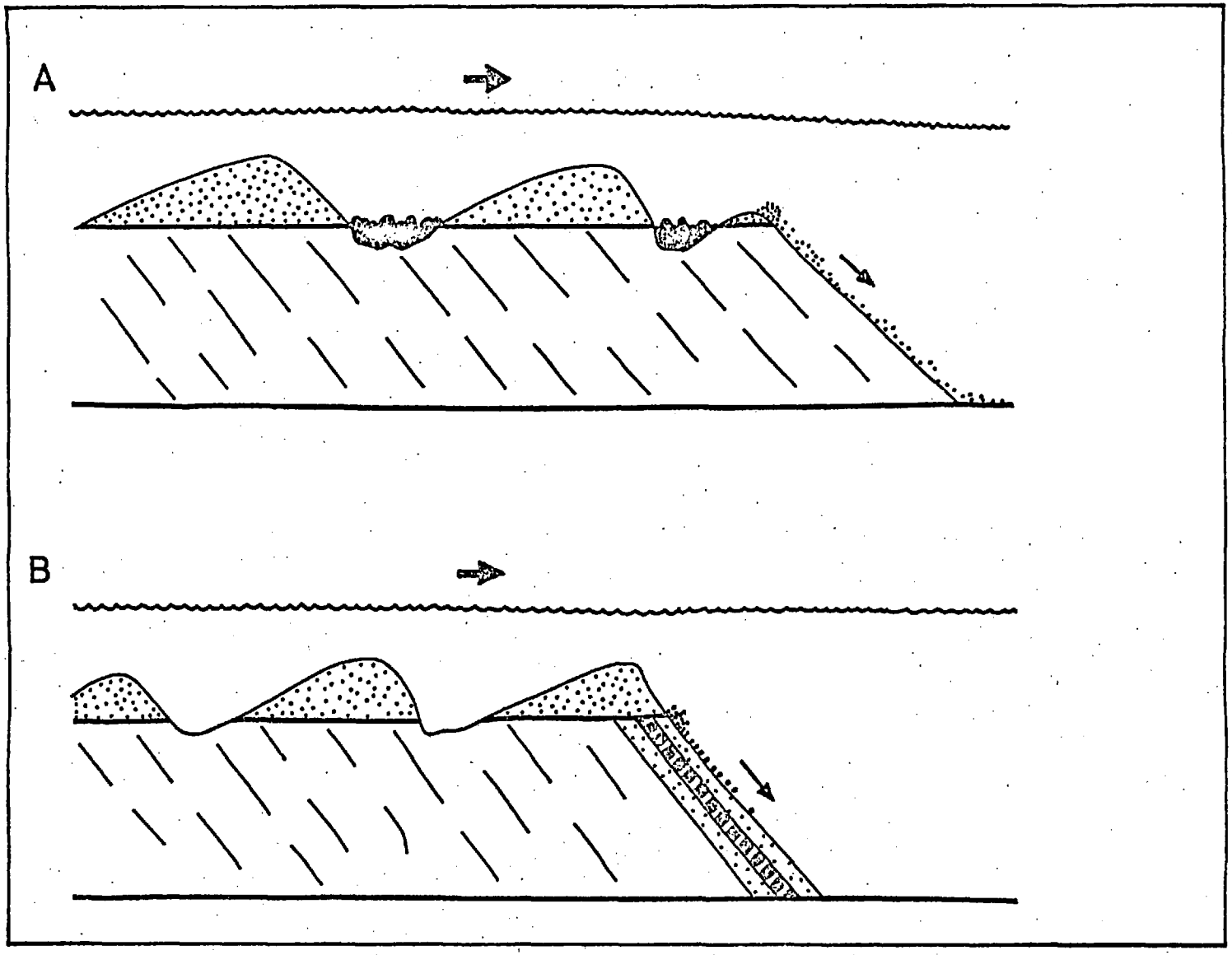
1. Planar (tabular) cross-stratification.
2. Trough (festoon) cross-stratification.
3. Horizontal stratification.
4. Parallel stratification.

7.2.1 Planar (tabular) cross-stratification

Planar or tabular cross-stratified sets are those that have broad lateral dimensions in relation to the set thickness and they are not bounded by trough-like erosional surfaces. Their bounding surfaces are planar and parallel although this is not strictly so in all cases as some bounding surfaces are convergent or undulatory in some places.

Planar cross-stratification is volumetrically the most abundantly developed primary sedimentary structure in the Bima Sandstones in the area studied. It occurs in all the sandstones ranging from medium- to coarse-grained varieties. The foreset planes of the cross-stratifications are plane surfaces which dip relative to the base of the bed containing them. The lower surfaces are planar or gently curved, but the shape and disposition of the laminae within the tabular set do not appear to be controlled by the relief on the lower surface of the containing bed (Harms and Fahnestock, 1965). The direction of dip of the foreset beds is taken

Fig.7.1.



Origin of coarse and fine foreset laminae by continuous avalanching(after Smith,1966)

to be the direction of flow of the currents which deposited the sediments. The angle of foreset dip varies and it is possible to correlate this variation with grain size. Foreset dip angles are higher for the coarser grained sandstones than for the medium grained ones. Dip values up to 15 degrees were recorded for the medium grained sandstones whereas coarser grained ones showed values as high as 35 degrees and even more in some exceptional cases where tectonism has increased the dip of the foreset beds.

Grading was observed to be a constant feature of the cross-stratified sandstones. The foreset beds are composed of alternating layers of coarse and fine sand (fig. 7.1, and plate 7.1.1). This feature has also been recorded by several workers in their studies of sedimentary deposits (Selley, 1963 photo 35; Bassumallick, 1966; and Smith, 1972 fig. 4). In his studies Smith noted that the alternating coarse-fine laminations in the low discharge bars of the Platte River were formed as a result of avalanching of sediments from the bar slip face, and he offers a good explanation for the grading. Scour pockets formed by lee-side eddies usually precede lunate and lingoid dunes and ripples. These scour pockets contain the coarsest material in transport while the finer-grained sediments are contained within the bed forms themselves. As the bed form and its scour pocket reach the edge of the bar, the coarse material spills over the lee-side face. This is followed by a continuous deposition of the finer sediments in the bed form over the coarse material. The repetition of this process by the movement of ripples and dunes over the slip face of the bar, results in a coarse- and fine-grained alternations within the cross-stratified unit. The thickness of the foreset laminae depends on their length and the size of the bed form plus its scour pocket (fig. 7.1).



Plate 7.I.I. Upward fining of material within a foreset bed of a cross-bedded unit. Note that there is coarse-grained material at the base of each foreset bed and fining up toward the base of the next foreset bed. The hammer head for scale.



Plate 7.I.2. Trough (festoon) cross-stratification in transverse sectio showing the parallel to sub-parallel laminae within the troughs. The hammer for scale. Gwana Hills section.

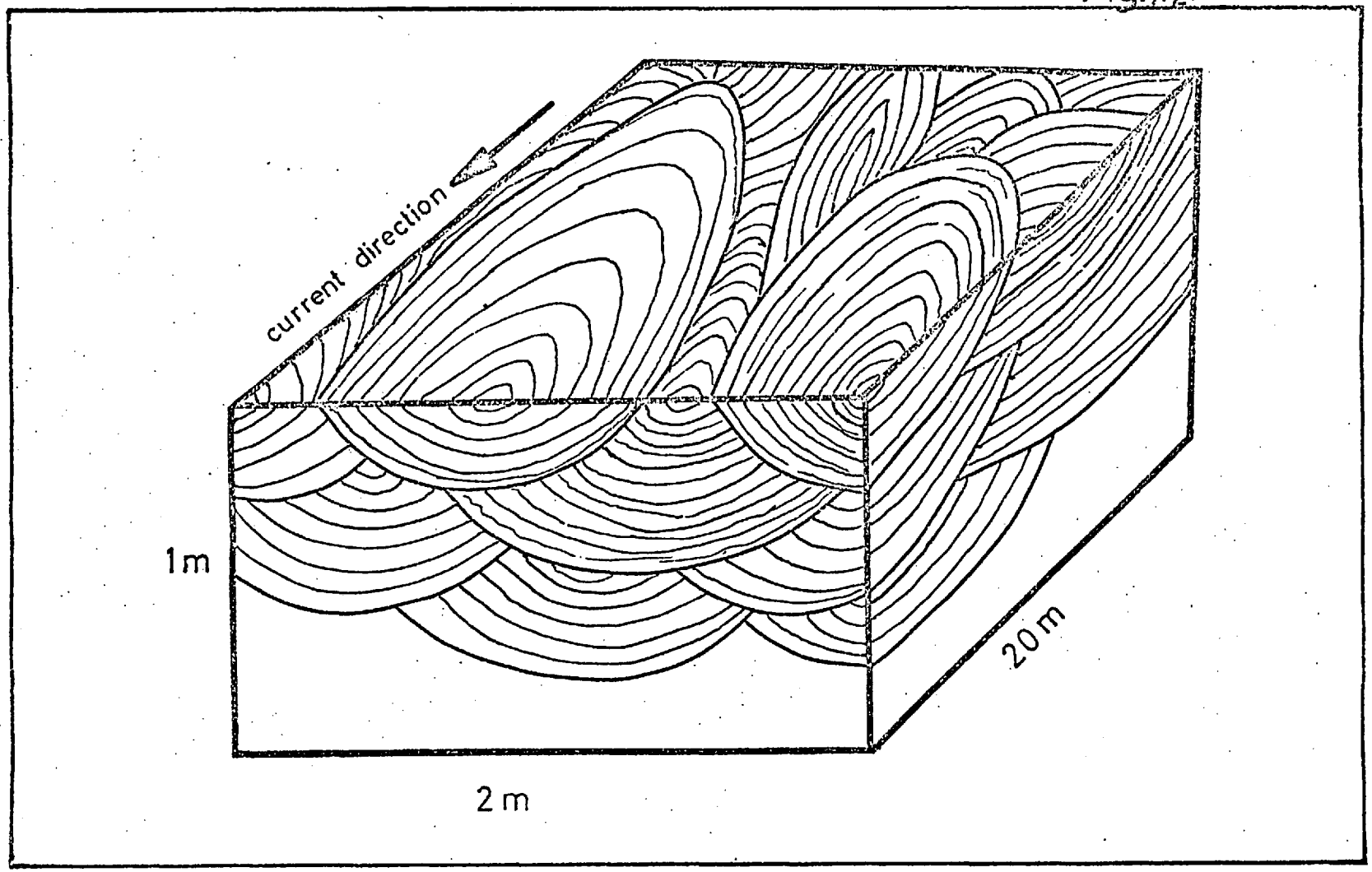
At least two types of cross-stratification were observed in the field based on the thicknesses of the foreset laminae and the cross-bedding unit itself. The planar cross-stratified units are therefore classified into two categories as thin- or thick-bedded. The thin-bedded units range in thickness from 4cm to 15cm, and thick-bedded ones range up to 1.5 meters. The thickness of the foreset laminae varied from 3mm to 5mm in the samples from which measurements were taken.

7.2.2 Trough (festoon) cross-stratification

Trough or festoon cross-stratifications were observed in the field and rank next to planar cross-stratification in abundance. Wherever they are observed in the stratigraphic sequence they occurred in cosets and never as isolated sets. They occur in medium- to coarse-grained and very coarse-grained sandstones. They were not seen in plan view in the field. However, they have been extensively studied and described by many authors. In plan view each trough cross-shaped set is an elongate erosional scour which is infilled with curved laminae. These troughs are aligned with their longaxis parallel to the direction of the currents which deposited the beds containing them. In plan view the scoop-shaped infilling laminae plunge in a down-current direction. In longitudinal section the laminae curve asymptotically toward the base (fig. 7.2). In transverse section the laminae run parallel or sub-parallel to the curved base of the trough (plate 7.1.2). The maximum depth of these troughs seen in transverse section seldom exceeds 14cm in the largest ones recorded in the field.

The infilling laminae of the trough sets have dips that average 28 degrees and range from 15 to 35 degrees. In some cases the laminae are symmetrically disposed within the trough cross-strata while they are

Fig.7.2.



Diagramatic representation of a three dimensional view of trough cross-stratification.

asymmetric in other troughs. Harms and Fahnestock (1965) suggest the explanation that the avalanche faces that filled the troughs were curved, and that the current direction or the volume of sand supplied to different parts of the trough are responsible for the asymmetry of laminae about the trough axis.

7.2.3 Horizontal stratification

Horizontal stratification as used here is restricted to tabular sets of laminae occurring in sand-sized material in which the laminae are horizontal or nearly so. This delimination is based on the definition of Harms and Fahnestock (1965). Horizontal stratification is not a very prominent feature of the Bima Sandstones in the area studied, but they form a distinctive physical unit. The sets of horizontally stratified units are thin, ranging from a few millimeters to 10cm and average 5cm. They extend laterally for tens of meters. The upper and lower boundaries of each set are erosional, horizontal and planar or sometimes irregular, where the boundaries are seen in the field.

The horizontally stratified sets are laminated and these are visible where weathering has not defaced the rock surfaces (plate 7.2). The laminations usually form bands of alternating pink and grey coloured sands. Such colour banding is a common feature of the stratified units of the Bima Sandstones. The laminae are persistent and can be traced laterally for several meters. Horizontal stratification due to textural differences is also common. The horizontally stratified units usually occur within a sequence of stratification types.



Plate 7.2. Thin horizontally bedded fine- to medium-grained laminated sandstone. Jelani area. Hammer for scale.

7.2.3 Parallel stratification

This terminology is used here for sets of laminations which occur in clay- and silt-sized material and following the definition outlined by Harms and Fahnestock (1965). Parallel stratification is the least common of all the primary sedimentary structures. This fact reflects the general scarcity of silts and clays in the stratigraphic sequence of the Bima Sandstones in the area of study. Silts and clayey silts were encountered only twice in the Bogoi Hills area (logs 2 and 3). In these two instances the beds were deeply weathered and covered with screes and loose rock fragments. Lamination in the silt and clayey silt beds was revealed only by shallow pits dug on the face of the rock outcrops. The silts are dark brown in colour and black in places where they contain petrified organic material. They were probably deposited from suspension by slowly moving or stagnant waters. In log 2 the silty beds overlie a horizon of sandy oyster beds and they are about 15 meters in estimated thickness. In log 3 the beds also average 15 meters in total thickness and are interbedded with laminated fine-grained sandstones.

7.3 MODE OF FORMATION OF THE PRIMARY SEDIMENTARY STRUCTURES IN THE BIMA SANDSTONES

The origin and geological significance of cross-stratification has been studied by many authors. The proposed mode of formation of the primary sedimentary structures described above for the Bima Sandstones will now be discussed.

The deposition of planar cross-stratification by avalanching of material over a bar slip face has been described from recent alluvial deposits of the Platte River by Smith (1972). This seems a reasonable

mode of origin for the planar cross-stratified Bima Sandstones. The mechanism of this process and the geometry of the resultant cosets have been described by Bassumallick (1966), Harms and Fahnestock (1965), and Smith (1972). The dips of the foreset beds directions in the Bima Sandstones is reasonably unimodal (chapter 4). It does not seem likely therefore that the foresets could have been formed by deposition on the banks of laterally migrating channels as was proposed by Wright (1959). Planar cross-stratification has been described from both ancient and modern sediments of various environments: the fluvial deposits of the Platte River (Smith, 1972), the Rio Grande (Harms and Fahnestock, 1965), the Bima Sandstones (Jones, 1962), and the marine Lower Greensand (Schwarzacher, 1953). It is concluded here from the morphology of the planar cross-stratified Bima Sandstones that they were deposited by fast flowing traction currents of known direction of flow, and that they were formed by the avalanching of material from dunes and ripples migrating over a bar slip face.

Knight (1929) was the first to describe trough cross-stratification. He explained them as the products of the scouring of channels which are later filled by sand laminae parallel or sub-parallel to the channel walls. McKee (1957) experimentally produced isolated examples of this type of troughs formed in the manner described by Knight. This mechanism is not adequate to explain the occurrence of cosets of trough cross-strata of uniform size like those observed in the Bima Sandstones in the area of study. Frazier and Osanik (1961) postulated that this form of cross-bedding is a product of migrating pockets of deeper scour associated with large sand waves on their downstream side. Allen (1963) suggested migrating

lingoid or lunate dunes as being responsible for the deposition of trough cross-stratification. Harms, Mackenzie, McCubbins (1963) related the trough forms to spoon-shaped scours occurring on point bars. Harms and Fahnestock (1965) pointed out the fact that the scour could not have been formed by eddies generated downstream on the lee face of dunes, as the turbulence related with the lee face of a dune is usually associated with low velocities and currents directed up the face. This will cause deposition rather than erosion. The highest velocities likely to cause scour will occur in areas between two adjacent dunes where an avalanche face is very low or totally absent. Such scouring and later infilling takes place at different times, with the infilling occurring at a time later than the scouring when local flow conditions are much changed.

Small-scale trough cross-stratifications on the other hand are related to asymmetric lingoid or lunate mega-ripples in their mode of origin (Allen, 1962). They are formed and preserved where an avalanche face follows the scour of a lingoid ripple occurring below the general level of erosion (Harms and Fahnestock, 1965). Both lingoid and lunate ripples have been described from recent sediments of the Niger River (Nedeco, 1959); in the tidal channels of the Wadden Sea and the Arcachon basin (van Straaten, 1953); in the Bahama Banks (Illing, 1954); in the Mississippi where they are associated with trough-bedded sands (Frazier and Osanik, 1961); and in the Platte River recent sediments (Harms and Fahnestock, 1965).

From the above considerations it can be concluded that the cross-bedded sands of the Bima Sandstones were formed by traction currents, and their occurrence in coarse-grained sands suggest deposition by very violent currents.

Horizontal stratification occurring in sand-sized material is usually formed by plane bed or low standing wave transport. The internal laminations of the horizontally stratified beds of the Bima Sandstones in the area of study indicate the absence of any prominent topographic feature on the bed during deposition. Horizontal stratification in the Bima Sandstones occurs as small units between cross-stratified beds. This indicates the fact that the currents which deposited the Bima Sandstones rarely maintained the upper flow regime conditions for a long period of time and that conditions were quickly changed to lower flow regime as shown by the abundance of great thicknesses of cross-stratified units compared with the thin horizons of horizontal stratification and massive bedding in the area of study.

Parallel stratification which occurs in clay- and silt- sized material is a product of slow settling from suspension as opposed to stratification formed by traction currents. Parallel stratification is formed by, currents of a flow regime lower than that necessary to transport sand bed material, and turbulence is weak enough to permit the deposition of suspended load (Harms and Fahnestock, 1965). Parallel stratification in the Bima Sandstones in the area of study is very scarce in occurrence and it was observed only in the Bogoi Hills area. This restricted occurrence could be attributed to the absence of conditions conducive to slow settling from suspension during the deposition of the Bima Sandstones in the area of study.

7.3 SECONDARY SEDIMENTARY STRUCTURES

7.3.1 INTRODUCTION

A major feature of the Bima Sandstones is the presence of cross-bedded units. The cross-bedding is either of the planar or trough (festoon) type,

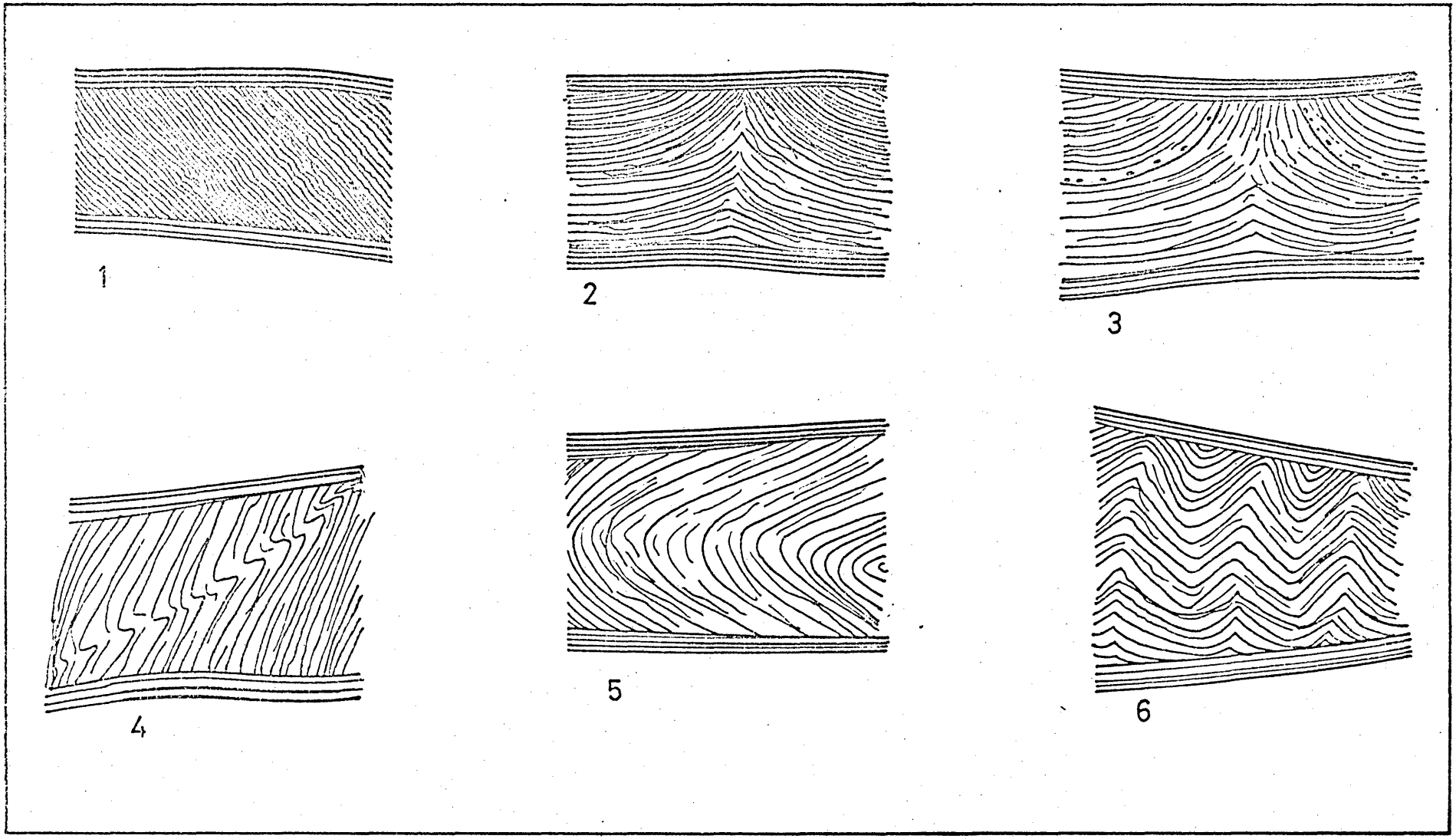
with the former predominant and ubiquitously developed throughout most of the vertical and lateral extent of the formation in the present area of study.

Peneconteporaneous deformation of bedding has been recorded from the Bima Sandstones in the past. Jones (1962) described deformed bedding in the Bima rocks of Lau, Yola and adjacent areas. Similar deformed structures in sandstones have been described by many authors in other parts of the world (McKee, 1938; Rice, 1939; Fuller, 1955; Robson, 1956; Potter and Glass, 1958; Sutton and Watson, 1960; 1964; Stewart, 1963; Selley and Shearman, 1962; Selley, Shearman and Watson, 1963; Selley, 1964, 1969; Williams, 1966; and Rust, 1968), and various deformation processes have been advanced for their origin. Attempt has been made by many authors to simulate these structures in laboratory experiments (McKee and others, 1962; Selley, 1969; Graf Peterson, 1967). Opinion is divided as to the definite mode of origin of these structures and various theses have been put forward to explain their origin. The cosensus of opinion of these authors discredits the suggestion that these structures are slumps, and favours the idea that they are "fossilised" quicksands, formed by the vertical expulsion of water from unconsolidated sands. The mechanism thought to be responsible for triggering off the prosses has been cited by many authors to be earthquakes, hydrostatic pressures and current action. However, all three mechanisms are possible causes for generating quicksands as proved by laboratory work carried out by Selley (1969).

7.3.2 Morphological description

At least two distinct types of deformational structures were recorded by the present writer in the present area of study. These are:

Fig. 7.3.



Types of deformation structures in the cross-bedded Bima Sandstones (after Jones, 1962).

(1) cuspate structures, and

(2) overturned foreset beds.

7.3.3 Cuspate structures

Similar structures have been described from the Torridonian rocks of northwest Scotland by Selley (1963). The Bima Sandstones structures were observed to occur only in horizontally bedded strata, usually in medium- to coarse-grained sandstones. These structures (plates 7.3.1 and 7.3.2) appear to be more commonly developed in the Bogoi Hills area, and in areas around Muri and Jelani, while their occurrence is very rare in the Jawara Hills area. Their occurrence was therefore recorded from almost all localities from the northern to the southern extent of the Bima Formation outcrop in the present area of study.

The presence of the cuspate structures is marked at their bases by an upward doming of the horizontal laminae of the sandstones in the areas where they occur. This upward doming generally gives way to a situation in which the laminae are ruptured and upturned toward the base of the next bed above. In some cases the individual laminae of the deformed bed could still be traced through the entire extent of the disturbed zone while in other cases there is a chaotic mixing up of the material toward the top of the structure to form a slurry (Plates 7.3.2 and 7.4, fig. 7.3.3), with all primary bedding characteristics having been lost.

From the field observations of these structures a possible sequence of development is proposed. The first stage in the sequence is marked by a gentle doming of the laminae within the bed. This is followed by the next stage in which the doming becomes more pronounced, and a final stage in which the apex of the dome becomes ruptured. It was observed in most



Plate 7.3.I. Cuspate structure in cross-bedded, medium-grained sandstone Diji area. Note the transition from undisturbed foreset laminae at the base of the bed to completely deformed laminae at the top of the bed. Hammer head for scale.



Plate 7.3.2. Cuspate structure in medium-grained sandstone in which all the laminae are disturbed with the core of the structure showing highly deformed layers. Hammer for scale.



Plate 7.4. Cuspate structure in cross-bedded sandstone in which there is a chaotic mixing up of the material in the center and top of the structure to form a slurry. Bashama area. Hammer head for scale.

cases that the original laminae kept their individuality within the structure, and that only in certain extreme cases was the core of the cusps observed to be massive at the upper part of the structure (plate 7.4).

7.3.4 Overturned foreset beds

Deformed cross-beds are the most abundantly developed of the two types of secondary sedimentary structures recorded in the Bima Sandstones in the present area of study. They were recorded from various levels in the stratigraphic sequence although they appear to be restricted to the upper part of the sequence in the Diji, Gidan Dari and Gwana areas.

At least five types of deformed cross-bedding were recorded in the field. Each type of the deformational structures is assumed to be a stage in the deformation sequence of the cross-beds. Thus, there is a distinct trend from minor deformations of the foreset laminae in the form of "rippling" to complex overfolds and contortions of various kinds (Fig. 7.3). The minor "ripplings" show some degree of angularity at the peaks of their convex upward features (Plate 8.1). . In another instance only the upper parts of the foreset laminae are folded. The deformation trend culminates in forms in which the foreset laminae are folded such that they become doubled up and lie on their sides (plates 7.5.1 and 7.5.2). In other more complex and advanced stages in the deformation of the foreset laminae the foreset beds become contorted into a series of "anticlines" and "synclines".

In all the 65 cases observed overturning is toward the dip direction of undeformed cross-beds within the same set, and with the deformation axes mostly perpendicular to this direction. This regularity



Plate 7.5.I. Severely deformed foreset bedding in which the individual laminae are completely doubled up on themselves. Jawara Hills area. Hammer head for scale.



Plate 7.5.2. An isolated deformed cross bed in which the individual laminae are doubled up on themselves and lying between undeformed cross-bedded units. Diji area. Hammer for scale.



Plate 7.6. Two disturbed cross-bedded units separated by undisturbed cross-bedded units. Note that the disturbed units have the foreset laminae folded in the direction of dip of the foreset laminae of the undisturbed bed. Diji area. Hammer for scale.

in the orientation of the deformational axes of the foreset beds shows that there is a relationship between deformation and current direction.

The maximum length of an overturned bed ranges up to 150 cm. In certain cases the distance throughout which deformed beds could be seen is about 3 meters, whereas in other cases the complete range of deformation occurs within a few centimeters. The largest distance observed in the field was about 3.5 meters. In the areas where the deformation of the foreset beds was severe and complex it was impossible to see them in situ because of the weathering of the rocks which destroys the deformed structures. Several such cases were observed in the Gidan Dari areas of the Jawara Hills.

A unique feature of these types of deformational structures is the degree to which the foreset laminae maintain their individuality no matter the complexity of the folding. Another remarkable observation about these deformed structures in the Bima Sandstones in the are of study is the complete absence of fractures, and the undisturbed nature of the cross-stratified units above and below them where they occur in isolated cases (plate 7.6).

A particular interest was taken in observing the nature of the contacts between the deformed beds and the undisturbed units above and below them. Two types of contacts were recognized. Those contacts in which the deformed cross-beds are truncated by the base of the overlying undisturbed cross-stratified units (plate 7.7.1) and those contacts in which there is no evidence of truncation (plates 7.7.2 and 7.8.1).

It was of a rare occurrence to see two contiguous layers of deformed cross-beds. In most cases recorded in the field the deformed beds rest



Plate 7.7.I. The top of a deformed cross-bedded unit truncated by the base of an undeformed bed with an erosional base. Bashama area. Pencil for scale.



Plate 7.7.2. Completely deformed foreset units of a cross-bed which show no sign of truncation by the base of an overlying bed. Jawara Hills area. Hammer for scale.

on and are overlain by undisturbed strata. There was only one isolated example of two deformed cross-beds lying one on top of the other found in the present area of study (plate 7.8.2). In this isolated case the top of the lower unit is truncated by the upper unit.

7.3.5 The mechanism of deformation

The mechanisms thought to be responsible for the deformation of unconsolidated sediments have been extensively discussed in the geologic literature. Many authors have put forward various theories, from the light of experimental work carried out by them to explain these deformed structures. McKee (1938) considered the deformation structures as resulting from slumping after the withdrawal of the depositing water. Rice (1939) held the view that the contortions in the Triassic sandstones of northwest England were formed as a result of sub-aqueous gliding and slumping. Fuller (1955) thought that they were non-orogenic in origin and he attributed them to the creeping of the cross-bedded sands down the depositional slope. Robson (1956) suggested friction due to over-riding sand-laden water currents as being responsible for the deformed structures in the Carboniferous Fell Sandstones of northwest England. Gill (unpublished work) supports the idea that the release of pore water following closer packing of the sand grains after seismic shock was responsible for the formation of the water ejection structures occurring at the top of the Tullig Sandstones at Killard in County Clare of Ireland. Jones (1962) ascribed the overturned cross-bedding in the Bima Sandstones to gravity slumping induced by earth tremors. Selley (1969) concluded after a series of experiments that earthquakes, hydrostatic pressures, and currents action could generate quicksand structures.



Plate 7.8.I. Two units of a completely deformed cross beds separated by layers of undeformed cross beds and showing no signs of truncation. Diji area. Hammer for scale.



Plate 7.8.2. Two contiguous units of deformed cross beds with the upper bed truncating the top of the lower bed. The head of the hammer marks the point of contact between the two deformed beds. Bashama area.

The field associations, general morphology, and their restriction to the upper part of the Bima Sandstones all lead to the conclusion that the deformed sedimentary structures described in this chapter for the Bima Sandstones were formed by a combination of the action of earth tremors and that of sediment laden currents.

CHAPTER 8

PALEOCURRENT ANALYSIS

8.1 INTRODUCTION

Cross-bedding and other primary current structures have been used in reconstructing regional paleoslopes (Potter and Olson, 1954), fixing the trend of ancient shorelines (Tanner, 1955), and in provenance studies of the probable source areas of the sands forming cross-bedded strata (Potter and Siever, 1965). Paleocurrents reconstruction thus provides certain details which other parameters such as lithofacies and isopach data cannot resolve. Cretaceous Bima Sandstones primary current structures have been mapped in the Yola and Lau areas by Jones (1962). It is therefore the purpose of this study to extend the work to the present area of study, so that the depositional history of the Benue Valley Sedimentary Basin in Bima times in northeastern Nigeria may be understood more completely.

The Bima Sandstones crop out fairly abundantly to the north of the area mapped (fig. 1), and to the southwest along hills which sometimes have very steep scarp faces. In the central, western and southeastern areas the Bima beds are completely covered beneath younger sediments of the Yolde and Pindiga Formations.

The Bima Sandstones outcrops form long linear ridges and hills on the flanks of gently plunging anticlines and synclines. The Bima Sandstones consist of coarse- and very coarse-grained cross-bedded, feldspathic sandstones with occasional basal conglomerates. Scattered through the section are pebbles of vein quartz and occasional clay lenses.



Plate 8.I. Cross-bedded units of the Bima Sandstone showing the planar type of cross-bedding. Bashama southwest area.

8.2 METHOD OF INVESTIGATION

8.2.1 Cross-bedding

Much of the cross-bedding measured in the field are probably of the planar type (plate 8.1). The foreset beds of the cross-bedding units are planes which are parallel to one another and meet the overlying and underlying true bedding at sharp angles. Occasionally, some are gently curved and tangential to the base with concave side upward. In some places several cross-bedded units occur in vertical succession and are called "cosets" (McKee and Weir, 1953) or "grouped sets" (Allen, 1963). In most places, the cross-bedding occurs as a single sedimentation unit between "normal" bedded strata. Although some cross-bedded units are wedge-shaped, most layers are uniformly thick. In most cross-bedded units foreset beds are graded, with coarser grains below and becoming finer upward toward the next foreset bed above.

In the cross-bedded units measurement and record were made of the following:

(1) Dip and strike of the "normal" bedding;

(2) Dip and strike of the foreset beds;

(3) Thickness or "scale" of the cross-bedded units; and

(4) Length of the foreset beds.

Good exposures enhanced the measurement of the true position of the cross-lamination planes in space. However, in places where suitable cross-bedding were not exposed strike and dip readings were made in accordance with the method outlined by Pelletier (1958). In this method a notebook is aligned with the traces of the cross-bedding on two non-parallel rock surfaces. These cross-bedding traces are lines and as two intersecting lines form a plane, the notebook is assumed to be in

the plane of the cross-beds. The thickness of cross-bedded units and foreset lengths were measured at the place where the strike and dip of the contained laminations were recorded.

The cross-bedding data collected in the field were re-oriented. Some workers advocate a graphic record of field data and their re-orientation on the spot. In this work re-orientation of data was carried out on return from the field chiefly because of the adverse climatic conditions of the area. In areas where there is a low structural dip correction for tectonic tilt is unnecessary and 5° is taken as the critical value. In areas where the beds have suffered secondary tilting in excess of this value due to tectonism the cross-bedding azimuths and dips were corrected by re-orienting the true bedding to horizontal. This is done on the assumption that the true bedding of the sandstones was essentially horizontal at the time of their deposition. It is also further assumed that by returning the true bedding to horizontal, the associated cross-bedding would be returned to its original attitude of deposition.

The corrections for tectonic tilt were carried out using the polar stereographic projections of Phillips (1954). The strike and dip values of both true bedding and cross-bedding were plotted on a transparent overlay revolving above the fixed net. By rotating the pole of the cross-bed through the angle of dip of the bed itself about the strike of bed the plane of the stratum is restored to horizontal. Compensation was not made for the plunge of folds, which would have involved an additional rotation preceding that about the strike of bedding, because the estimated average fold plunge in the Muri/Futuk area is less than 10° (cf Ramsay, 1961).

Unless the strike of the cross-bedding coincides with that of the true bedding the correction for tectonic tilt would result in a complete alteration of the strike values of the cross-bedding. The values obtained for the cross-bedding after re-orientation represent their strike and dip prior to disturbance and, therefore, give the current direction and the inclination of the contained material in that cross-bed. All azimuths, each of which is assumed to indicate a down-current direction, were corrected for tilt, and summarised in circular histograms (figs. 8.4 and 8.5). Visual inspection of these histograms gives an estimate of the variability in cross-bed orientation at each outcrop, thereby eliminating statistical analysis to demonstrate the existence or validity of the pattern thus revealed. Maps were prepared of the azimuthal variability so that regional trends in orientation could be interpreted.

To enhance interpretation on a regional scale, the vector mean current direction at each locality was calculated trigonometrically according to Curray (1956), and the resulting current "arrows" were plotted on maps. In order to emphasize regional trends over local and erratic directions, a moving average of the mean directions was calculated. The graphic grid method (Pelletier, 1958) was employed and the resulting map was used as the best estimate of the regional current pattern.

8.2.2 Maximum pebble size

The regional variation in grain size was investigated in the Bima Formation in the area of study. The maximum diameter of the 10 largest quartz pebbles were measured at each locality, and a map of the mean diameters was prepared. The general concept of using average maximum

particle size for provenance studies is based on Hack's (1957) theory that particle size change downstream is determined partly by the initial size of the material, and partly by the relative resistance of the material to abrasion and breakage.

In the Bima Formation studies the maximum diameter of 10 vein quartz pebbles was measured at each locality and the measurements were averaged. Areal-moving averages were calculated from previously recorded averages based on the moving average grid system. The resulting grid-point values were contoured.

8.3 STATISTICAL APPROACH TO INVESTIGATION

The sampling design used in this study was adopted after a field season during which the accessibility of the area, the abundance and variability in size and types of exposures, and time available were carefully considered. The Bima Sandstones, in the area of study, has a reasonably continuous exposure pattern and exposures are usually large enough to furnish satisfactory data on cross-bedding azimuth and inclination. However, certain areas were highly inaccessible and could not be sampled completely, or at all in some cases.

The number of azimuth readings therefore varies from area to area depending upon the accessibility of the area, and the state and abundance of the exposures.

Care was taken to avoid unequal and therefore biased sampling between the well exposed and numerous scarp faces on hilly areas and the fewer poorly exposed outcrops on lowland areas. The sampling locations were chosen arbitrarily and as evenly spaced as possible on

the map before field data collection was commenced. The whole area was divided into 5' (minutes) lines of latitude and longitude to give a total of 25 squares in a grid pattern. Depending on the availability of well exposed cross-beds and accessibility of the area 1-4 exposures were examined in each square, and 1-4 cross-bedded units were randomly selected on each exposure. Two foreset planes were measured per unit as there was no prior information available on the local variability of cross-bedding azimuth in the rocks in this area. Potter and Pettijohn (1963) suggested the measuring of one foreset planes from each cross-bedded unit. Although no strict control was imposed on the distance between exposures within the squares and between squares because of the difficult terrain, an attempt was made to make the sampling locations as near equidistant as possible.

8.4 CROSS-BEDDING

8.4.1 Introduction

Sorby (1853) was the first to carry out any systematic study of cross-bedding, although his measurements were neither compiled as rose diagrams nor recorded on maps. Most of his information on several formations were documented in notebooks. The first cross-bedding measurements to be plotted on a map were those of Rubey and Bass (1925).

Several authors have since such modest beginnings, measured and mapped cross-bedding in different places. Among the many who have measured and presented their cross-bedding data in a frequency distribution or current rose, are Knight (1929, figs. 30 and 31), McKee et al. (1953) figs. 3, 4 and 5). Agaston (1954), Edwards (1955, fig. 26), Dott (1955,

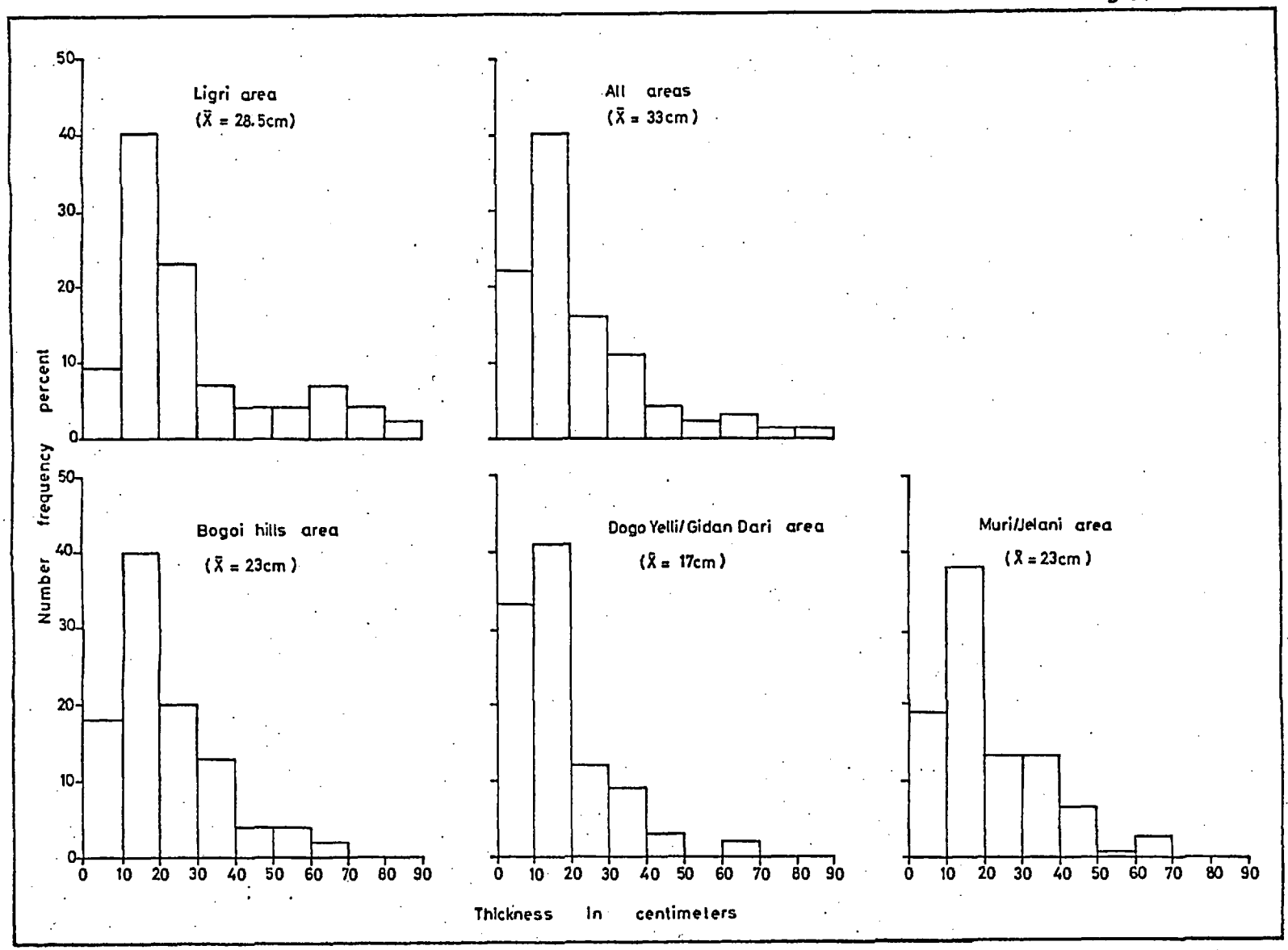
fig. 10), Tanner (1955, figs. 3 and 7), Whitaker (1955, fig. 2), Mellen (1956, fig. 30), Crook (1957), Pettijohn (1957), Kaneho (1958, figs. 1 and 2), Gradzinski et al. (1959, fig. 5), Tanner (1959), Bassett and Walton (1960, fig. 9a), Ball and Dinley (1961), Hamblin (1961b, fig. 9), Potter and Pryor (1961), Allen (1962, fig. 18), Stewart (1962), Yeakel (1962), Hoffman (1966), Agterberg and Trettin (1967), Casshyap (1970) Barret (1970), Casshyap and Jain (1970), Casshyap and Qidwai (1971).

These studies show that cross-bedding is ubiquitously distributed in every major environment: fluvial, littoral, marine and eolian. Cross-bedding occurs in sediments that range from Precambrian to Recent and has been measured from clastic deposits varying from silt to coarse conglomerates, and in sandstones ranging from graywackes of turbidite origin to the cleanest quartzose sand.

8.4.2. Thickness of cross-bedded units (Scale)

284 measurements were made on the thickness or scale of the cross-bedded sedimentation units from 73 exposures covering the whole area of study. The data were grouped in 1cm-class Intervals and the whole area divided into 4 subregions. Thickness frequency histograms were plotted for each of these sub-regions and one for the whole area (fig. 8.1). The thickness of the cross-bedding for each locality was determined by averaging the thicknesses of all the units measured at that locality. The averages for the localities in each sub-region were then converted to number frequency percent. The thickness of the sedimentation units ranges from 4 cm to 86 cm with an occasional high value of up to 130 cm in some places. These exceptionally high values were not included in computing the average thickness of cross-bedded units. The average

Fig. 8.1.



Histograms showing frequency distribution of cross-bedding thickness in the Bima Sandstones.

thickness in the four sub-regions are as follows:

SUB-REGION	AVERAGE THICKNESS (CM)
Bogoi Hills	23
Muri/Jelani	23
Dog Yelli	17.9
Ligri	28.5

Schwarzacher (1953) concluded that the thickness of cross-bedded units are log normally distributed. The thickness of Bima Sandstones cross-bedded units shows a normal distribution (fig. 8.1), with the highest concentration in the 10-20 cm ranges.

The significance of the scale of cross-bedding has been discussed by many authors (Schwarzacher, 1953). The scale of cross-bedding may be correlated with coarseness of grain size. Thus silts exhibit "small-scale" cross-bedding whereas gravels show "large-scale" torrential bedding. In the Bima Sandstones the scale of cross-bedding may be classified as "medium-scale", and can be correlated with the grain size at any locality. Most of the cross-bedding thicknesses recorded fall into the "medium-scale" type while a few are "large-and small-scale" in thickness.

The cross-beds are thickest in the Ligri area where they average 29 cm, and thinnest in the Dogo Yelli/Gidan Dari area where they average 18 cm. In the Bogoi Hills and Muri/Jelani areas they average 23 cm respectively. Thus the main direction of thinning is northeastward. Schwatzacher (1953) found that cross-bedding thickens toward the source. However, in the Bima Sandstones this concept does not apply as the cross-bedding units thin toward the source which lies to the northeast and

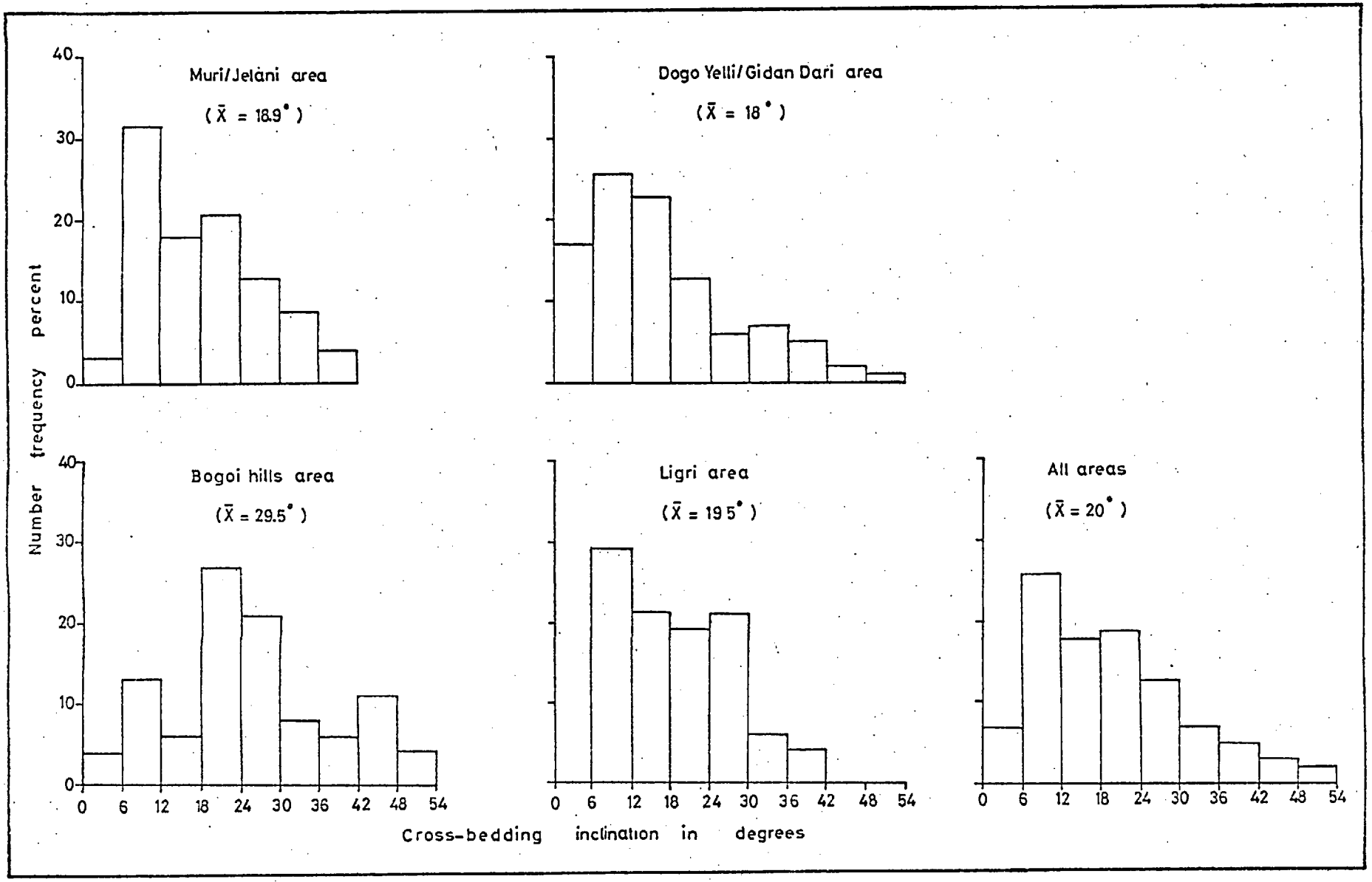
southeast. The reason for this deviation from Schwarzacher's concept is probably due to the lack of lithologic variation and the absence of marker horizons for stratigraphic control of sample locations in the Bima Sandstones in this area. The Bima Sandstone becomes younger and finer grained northward and northeastward in the area of study. Both factors could have strongly influenced the results obtained for the cross-bedding thicknesses.

8.4.3 Inclination

Most of the cross-bedding traces measured in the field are plane surfaces tangential to the base and truncated by the top of the bed containing them. The inclination of the cross-bedding is therefore taken as the angle between the plane of the foreset beds and the plane of the true bedding. Much variation occurs in the tilt-corrected inclination of the Bima Sandstones foreset beds. In the Dogo Yelli/Gidan Dari area the foreset inclination ranges from 2° to 54° and averages 18°. In the Ligri area the cross-bedding are inclined between 7° and 41° and average 19.5°. In the Muri/Jelani and Bogoi Hills areas the foreset beds are inclined 4° to 54°, and 5° to 71° with an average inclination of 18.9° and 29.5°, respectively. The average inclination for the whole area is 20° (fig. 8.2).

The inclination of the foreset beds in the Bima Sandstones varies over a wide range as may be seen from the comparative data chart (table 8.1). The predominant modal class is that of 6-12 (classes have 5° intervals) The mean value is always higher than the modal class in all the sub-regions.

Fig. 8.2.



Histograms showing cross-bedding inclination in the Bima Sandstones.

1969

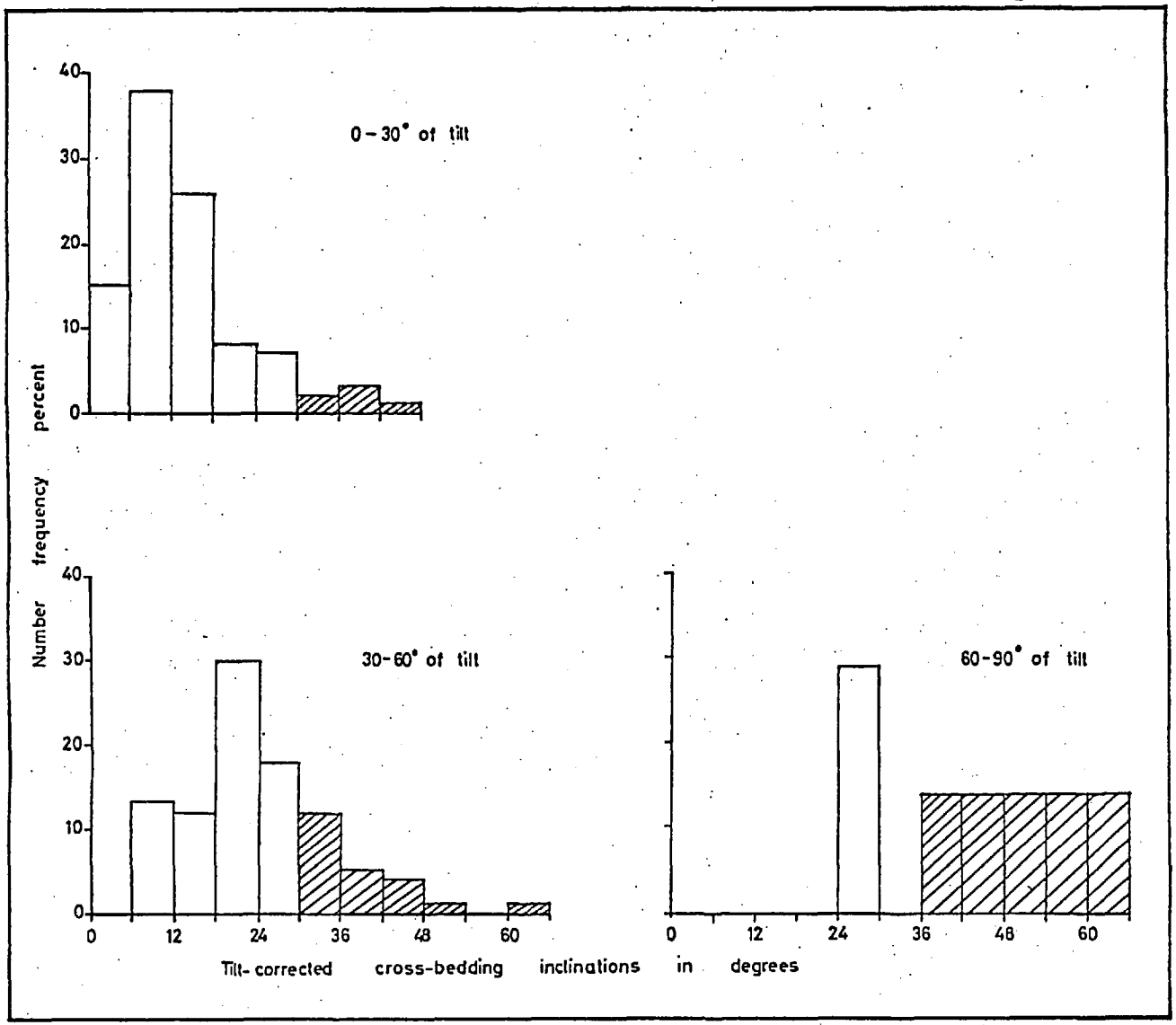
Table 8.1 Comparative sub-regional tilt-corrected dip inclination data

Sub-region	Dip inclination			
	Maximum	Minimum	Mean	Modal class
Dogo Yelli	54	2	18.0	6-12
Ligri	41	7	19.5	6-12
Muri/Jelani	54	4	18.9	6-12
Bogoi Hills	71	5	29.5	18-24

The inclination of foreset beds is taken and believed to be the angle of repose of the sand grains at the time of their deposition. Bagnold (1941) puts the upper limit of the angle of repose of dry sand at 34°. The angle of repose is found to be 27° to 28° for water-laid sandstones (McKee, 1940). The angle is higher for eolian than for sub-aqueous deposits. This was found to be true for cross-bedding measurements of water-laid and eolian sandstones in the Grand Canyon region (McKee, 1940). These assertions are supported by experiments conducted by KcKee (1955) which showed that the same sand had a lower angle of repose in water than it does in air.

The inclination of foreset beds is a function of the amount of folding or tilting of the normal bedding, and the relationship between the orientation of the cross-bedding planes and the direction of tilt of the normal bedding (Pelletier, 1958). Anomalously high values of foreset inclination occur in the Bima Sandstones. These high values of foreset inclination can be correlated with the high degree of folding and faulting common in the Bima beds in these parts of the Benue Valley Trough. Brett (1955), Pelletier (1958) and many other workers have observed anomalously

Fig. 8.3.



Frequency distribution of tilt-corrected cross-bedding inclination of the Bima Sandstones grouped according to different amounts of tilt.

Table 8.2 Variation of mean cross-bedding inclination in other formations

GEOLOGICAL FORMATION	NO. OF MEASUREMENTS	MEAN DIP OF BEDDING	MEAN INCLINATION OF CROSS-BEDDING	AUTHOR
Pennsylvanian (Illinois and vicinity)	531	0°	16.5°	Potter and Olson (1954)
Lorrain Quartzite (Bruce mines, Ontario)	119	19°	20.2°	Pettijohn
Baraboo Quartzite (Wisconsin)	283	46°	23.0°	Brett (1955)
Lorraine Quartzite (Lacloche, Ontario)	32	77°	25.8°	Pettijohn (1957)
Sturgeon Quartzite (Michigan)	80	Ca 90°	28.1°	Trow (1948, Ph.D. Thesis, University of Chicago).
Pocono (Maryland & Pennsylvania)	3512	15°	17.15°	Peiletier (1958)
Pocono "	791	45°	18.63°	"
Pocono "	462	75°	19.39°	"
Pocono "	206	90°	25.83°	"
Bima Sandstones	126	15°	14.7°	This work
Bima Sandstones	138	45°	23.8°	"
Bima Sandstones	8	75°	49.0°	"

high foreset inclination especially in tectonically deformed rocks as in the Bima Sandstones. Pettijohn (1957b) suggested the interpretation that such foreset beds with high deep values have been rotated by deformation.

A mean cross-bedding inclination was computed for the whole area in three classes of folding: 0-30° 30-60° and 60-90°. The three means obtained show a sharp rise in average inclination as the amount of deformation is increased. The Bima Sandstones data are similar to data from other areas and Formations (table 8.2). It is not possible to make exact comparison, although there is a similar increase in inclination with increasing deformation in the Bima Sandstones.

Histograms to illustrate the Bima Sandstones data in Table 8.3 are shown in fig. 8.3. Shaded areas represent cross-bedding inclinations greater than 30°. The percentage of inclinations in excess of 30° is calculated for various degrees of folding (Table 8.3). These calculations also show the increase in inclination with increase in deformation.

Table 8.3 Percent of cross-bed with tilt-corrected inclinations greater than 30° (Bima Sandstones)

Amount of folding	Percent of cross-bedding inclination exceeding 30°
0 - 30°	6.35
30 - 60°	22.46
60 - 90°	75.00

The histograms for the sub-regional averages and those grouped according to different amount of tilt compare very closely with data from

the Pocono Formation (Pelletier, 1958), the Mississagi, Espanola, Serpent and Lorrain Formations (Casshyap, 1968). The histograms however show many high readings exceeding the accepted angle of rest for fluvial deposits. Such an anomaly was also noted by Brett (1955), Wright (1956) Pelletier (1958), Jones (1962), and Casshyap (1968). The presence of pebbles exceeding 1cm. in diameter, however, preclude an aeolian origin of the Bima beds. The probable cause of the high foreset inclination values may be due to deformation as a result of differential movement during folding, at the time of deposition or after deposition. Some of the high inclination values may be due to sampling error.

8.4.4 Azimuth

Cross-bedding azimuthal studies were carried out by measuring the variations in the dip directions of the cross-beds, and their paleogeographic significance in Sedimentary Basins as indicators of Paleocurrent directions. The tilt-corrected azimuthal data for each locality were grouped into four sub-regions and rose diagrams or circular histograms plotted for them. The number frequency percent was shown for 30° degree class intervals. The azimuthal data of all the exposures were finally grouped, number frequency percent computed, and a composite circular histogram was plotted for them. These regional sub-divisions were made to illustrate any local influence on the cross-bedding azimuths.

In most histograms the cross-bedding azimuths, including the principal class, are generally in the western quadrant. In the Ligri area the spread is much larger. Some histograms show reversals, with azimuth direction to the east. Although none of them shows a maximum in the reversed directions, the Ligri area indicates a polymodal source for the

sediments, with contribution from northern, northwestern and northeastern quadrant does not manifest itself as strongly as those from the other quadrants. In the Dogo Yelli/Gidan Dari area the cross-bedding azimuths show a stronger contribution from the northeast and a weaker southeastern component, and could therefore be described as unimodal. In the Muri/ Jelani area the cross-bedding azimuths are bimodal with currents flowing from the northeast and southeast. In the Bogoi Hills area the current direction is essentially to the southwest and is therefore unimodal (fig. 8.4). The histograms constructed for the northern and southern portions of the whole area of study show the current pattern on a more regional scale. In the north the cross-bedding azimuths are bimodal with currents from the northeast and southeast, whilst in the south the currents are from the east thus indicating a unimodal pattern (fig. 8.5). The composite rose for the whole area indicates that the source areas for the Bima sediments laid to the east, with a distinct southwesterly current flow, and a subsidiary trend to the northwest. The variability in current direction shown by the dip azimuths by reversal in direction as demonstrated by some histograms is to be expected in fluvial deposits. This is a result of the fact that rivers meander change course or join a major stream.

Many authors (Jizba, 1953: Chayes, 1954, Curray 1956, Pincus 1956, Pelletier 1958) have pointed out the problems of applying linear techniques to circular data. Thus the circular histograms reveal local deviations from dominant current directions and therefore tend to mask the regional trend. The vector summation method which is independent of choice of origin and presupposes that each observation is a vector with both magnitude and direction, was employed. The method used was first

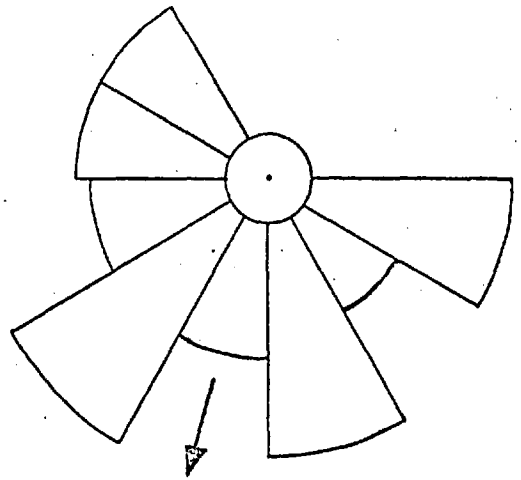
Fig. 8.4.

Ligri area

13 localities

52 readings

Mean = 195°

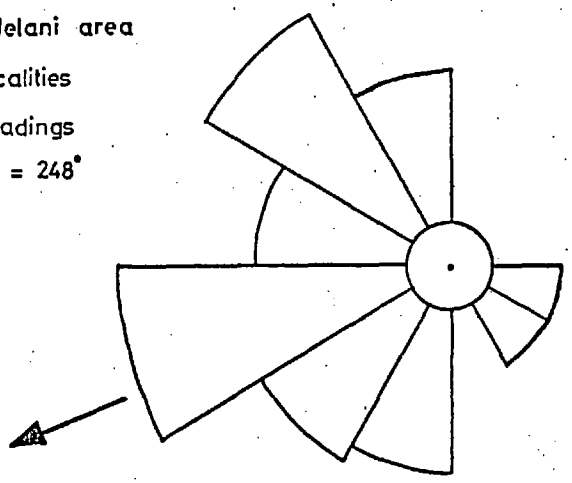


Muri/Jelani area

19 localities

76 readings

Mean = 248°

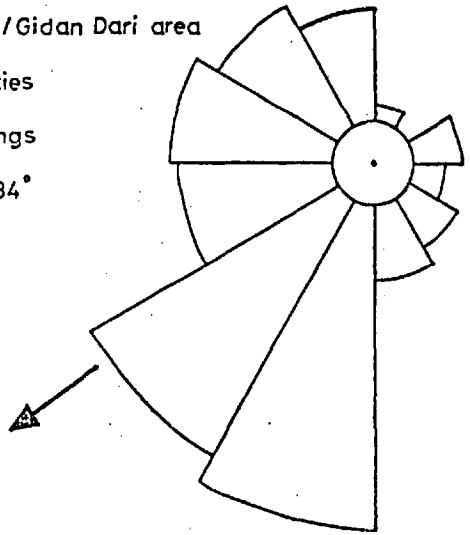


Dogo Yelli/Gidan Dari area

25 localities

100 readings

Mean = 234°

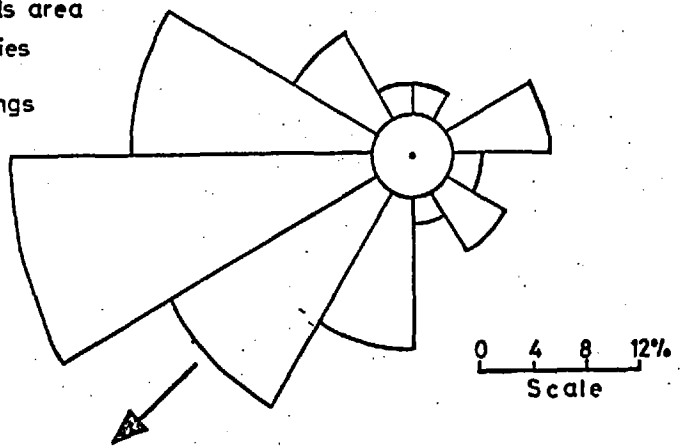


Bogoi hills area

12 localities

56 readings

Mean= 227°



Cross-bedding azimuth distribution in the Bima Sandstones in the area of study.

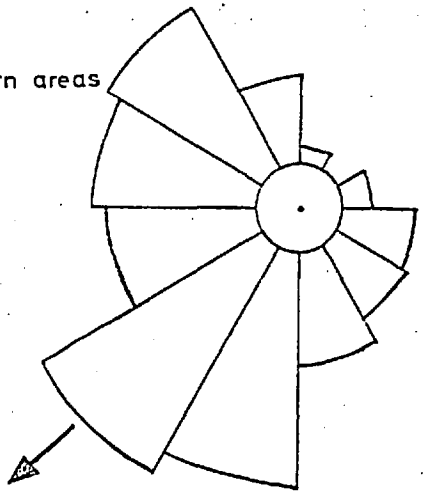
Fig. 8.5.

Jawara hills and northern areas

41 localities

164 readings

Mean = 227°

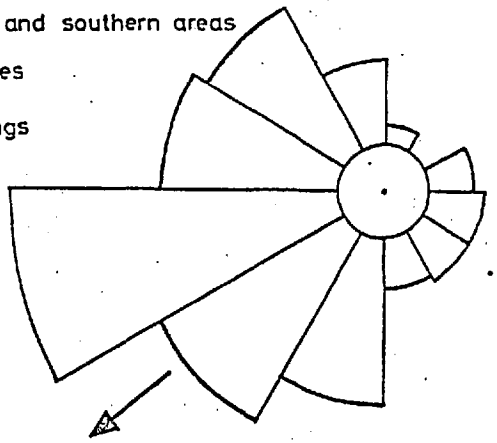


Bogoi hills and southern areas

28 localities

112 readings

Mean = 231°

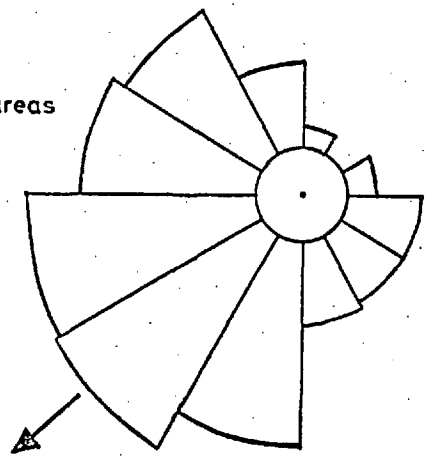


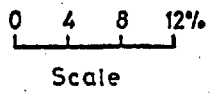
Composite rose for the all areas

69 localities

276 readings

Mean = 229°





Cross-bedding azimuth distribution in the Bima Sandstones in the area of study.

introduced by Reiche (1938) and elaborated upon by Curray (1956), and Pincus (1956). It involves the summations of the vector components derived from the sines and cosines of the azimuths (see appendix for details). The vector mean which is obtained from the above method is believed to be the best measure of central tendency for circularly distributed data (Curray, 1956; Pincus, 1956; Potter and Pettijohn 1963; Pettijohn et al., 1956; and Casshyap, 1968). Sometimes cross-bedding azimuths show a wide scatter at certain localities but are confined to a narrow zone elsewhere. Vector magnitude (Curray, 1956), vector strength (Pincus, 1956), or consistency ratio (Reiche, 1938) is used to measure this degree of scatter. Pelletier (1958) defined the consistency ratio as the length of the resultant divided by the number of the individual unweighted vectors. The vector magnitude is directly related to the standard deviation by the equation:

$$S = \sqrt{\frac{(\sum X - X)^2}{n - 1}}$$

where S is the standard deviation, X is the observations, and n is the sample size. Results of the calculations are given in Table 8.4. There is an apparent relationship between vector spread and vector magnitude. The highest standard deviation is 99.7 and the lowest is 24.6. The number of readings at each locality was never greater than 16 and the least was 8. Schwarzacher (1953) and Pelletier (1958) made their calculations using localities with more than 40 readings. In the Bima Sandstones studies the plan of the work precluded the possibility of taking more than 16 readings at each locality. However, the results obtained compare favourably with those obtained by Pelletier (1958) for the Pocono Formation.

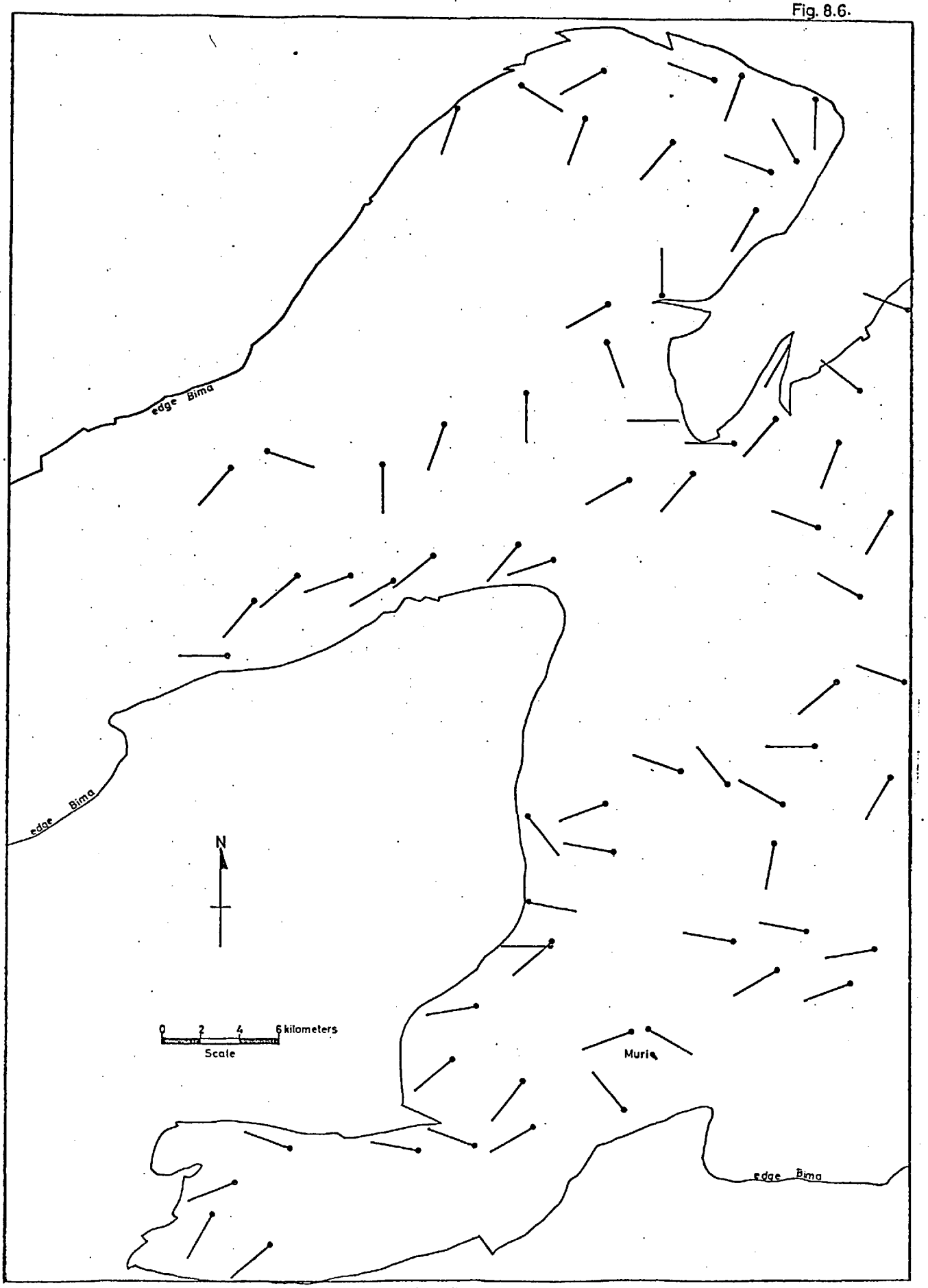
Table 8.4 Computation of mean, variance, vector magnitude in the Bima Formation

Squares	Mean	No. of readings	Variance	Standard deviation (degrees)	Vector magnitude (percent)
17	217	16	3465.6	58.9	60.0
10	302	12	605.3	24.6	32.4
18	278	16	4688.0	68.4	31.5
25	290	16	8870.0	94.1	25.6
1	217	16	3202.0	56.6	57.3
2	251	16	2525.0	50.2	36.0
19	251	12	9910.0	99.5	33.5
24	214	16	2353.1	48.5	70.4
16	225	16	3160.0	56.2	60.0
9	205	16	5120.0	71.6	45.4
5	251	16	4251.7	65.2	40.0
6	225	16	6973.3	83.5	23.9
7	213	16	2777.6	52.7	75.3
8	255	16	2573.3	50.7	54.0
12	216	16	7131.7	84.5	15.0
13	240	16	2526.6	50.3	46.0
14	138	8	6698.6	81.2	67.5
15	187	12	1235.2	35.1	83.0
23	161	8	2618.2	51.2	73.0

Fig. 8.6 is a plot of the mean current vectors for each locality and it shows the local current variability in the Bima Sandstones in the area of study. The length of the arrows is proportional to the vector magnitude of the individual readings. Because of the general apparent nature of the current pattern no statistical analyses were made to interpret a trend. There is less variability in current direction in the south and northeast. These results agree with the work done by Jones (1962) in areas to the east. The behaviour of the currents in the Muri/Futuk area indicates the convergence of two major currents flowing from the northeast and southeast to the southwest.

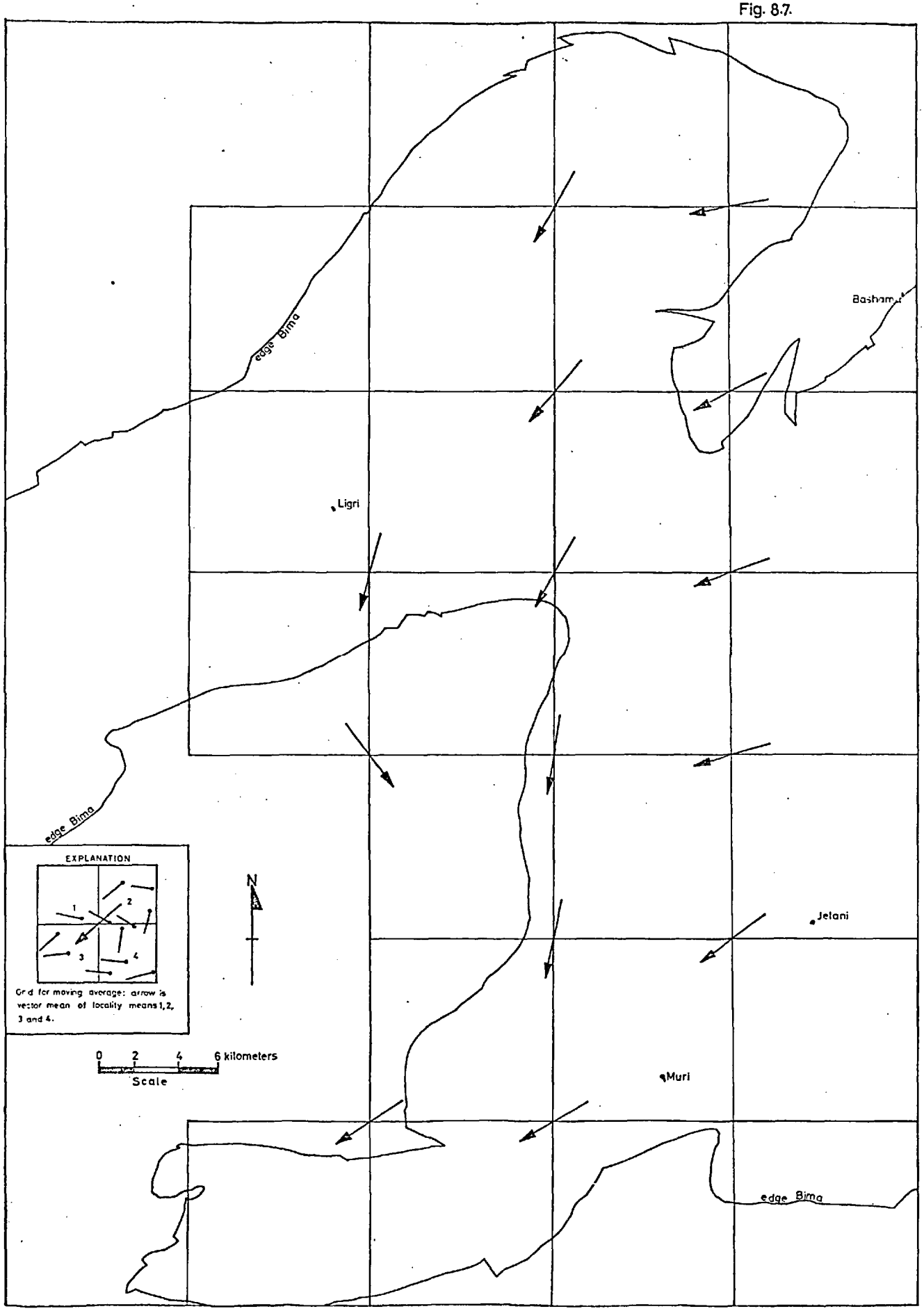
The vector mean at each locality gives the local current pattern, whilst the grand and super vector means show more clearly the regional current patterns. The individual pattern could be combined by means of moving diagrams to give a general paleocurrent direction for the whole area. The resultant vectors are called moving averages which have been used to show areal variation of grain size as well as orientation of directional structures (Potter, 1955; Schlee, 1957; Pelletier, 1958; Jones, 1962; Fahrig, 1961; Potter and Pettijohn, 1963). Moving averages are used to smooth out local variations and therefore provide a simple way to show a major sedimentary trend. In constructing maps of moving averages the whole area is arbitrarily divided into a grid pattern of squares. The vector means of four adjacent squares are then averaged and the result plotted at the intersecting line between them. This process is extended over the entire outcrop area, utilizing each vector mean four times. Fig.8.7 illustrates the resultant current pattern of moving averages in the Bima Sandstones. The general paleocurrent direction in the whole area could be seen to be predominantly in the

Fig. 8.6.



Locality averages of Bima Sandstones cross-bedding dip direction

Fig. 8.7.



Moving average of Bima Sandstones cross-bedding vector means

southwest.

Table 8.5 Mean cross-bedding variability in the Bima Sandstones in the area of study.

Subregion	A	B	C	D	E	F
Number of readings	56	76	100	52	164	120
Formation mean	189	248	234	195	227	211
Standard deviation	59.9	62.2	68.2	63.4	63.8	63.6
Variance	3589	3872	4654	4027	4083	4043
Vector strength	57	39	51	53	50	49

A = Bogoi Hills area

B = Muri/Jelani area

C = Dogo Yelli/Gidan Dari area

D = Ligri area

E = Northern area

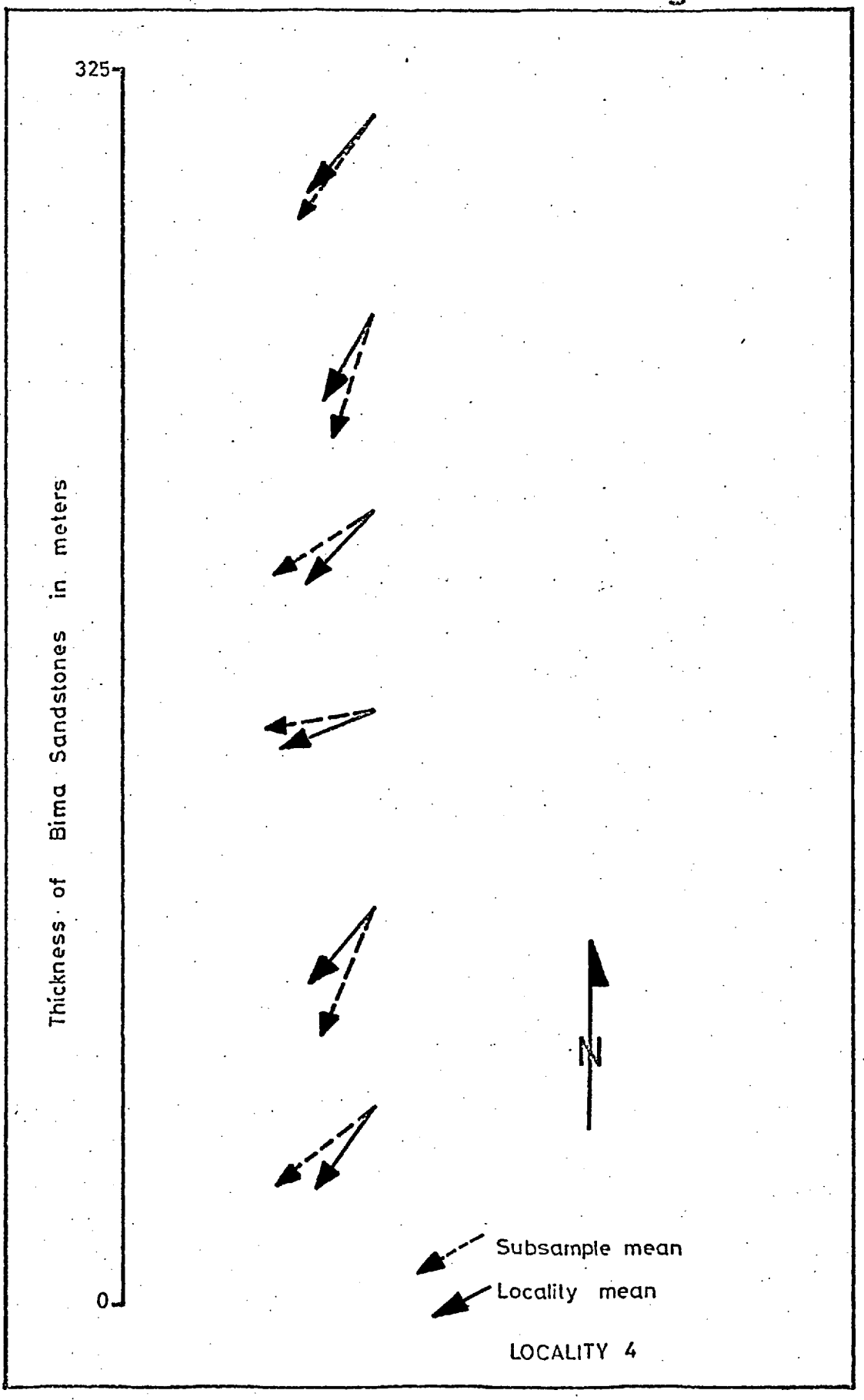
F = Southern area

8.4.5 Vertical Variation in cross-bedding orientation

The tilt-corrected cross-bedding dip azimuths were plotted at successive 54 meters stratigraphic intervals in the Bogoi Hills section (locality 4, fig.8.8). The general current direction is southwesterly in the 324 meters vertical interval. The lack of precise stratigraphic control of the Bima study, therefore, does not appear to alter significantly the paleocurrent pattern deduced from cross-bedding and maximum vein quartz pebble diameter measurements.

A vector mean was obtained from each subsample and compared with vector mean of the whole section. The deviation of the submeans from

Fig. 8.8.



Vertical variation of Bima Sandstones cross-bedding orientation

the locality mean is generally less than 25°.

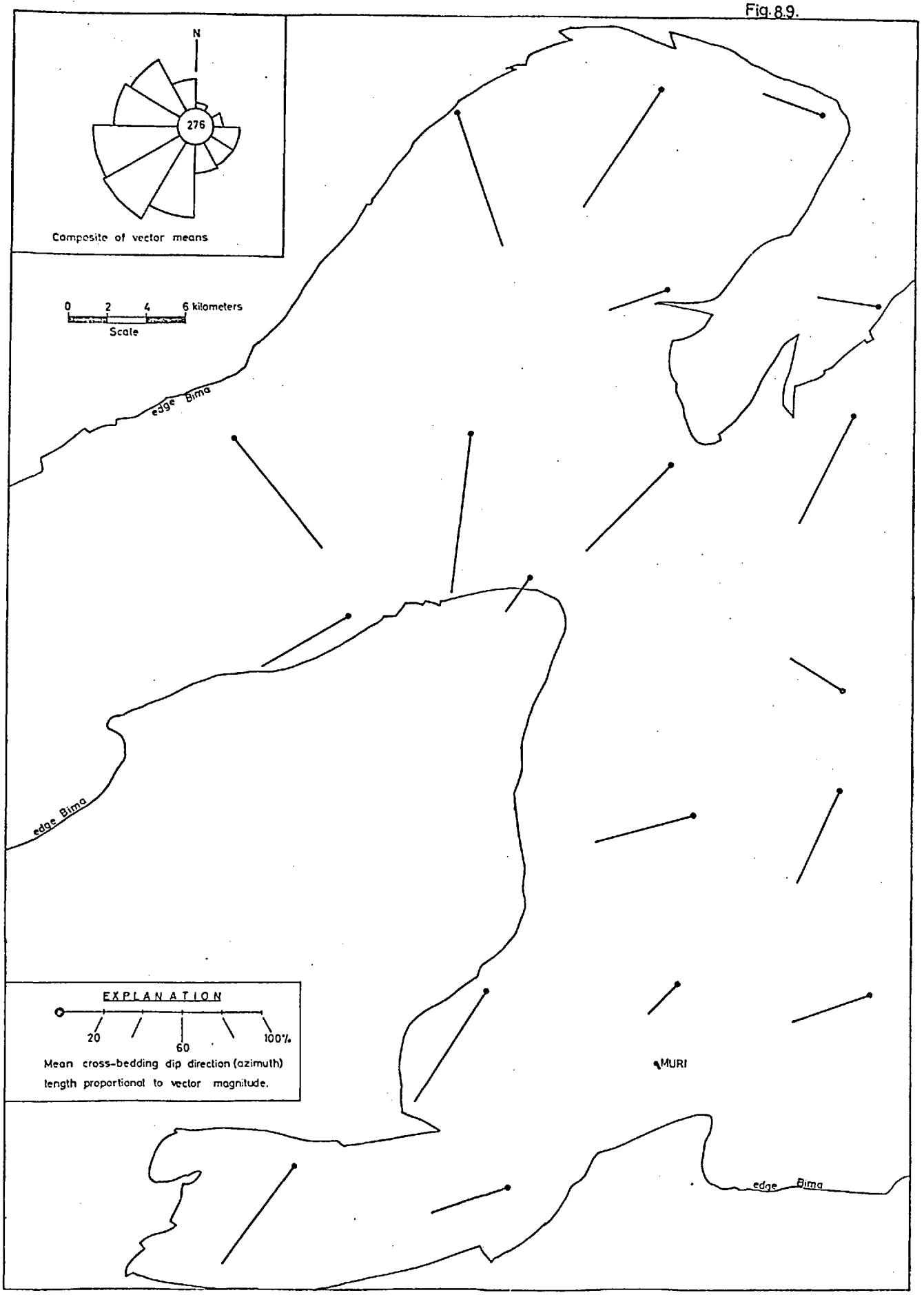
The conclusion could therefore be made that at a given geographic location there are generally consistent current directions through Bima time. Cross-bedding measurements of parts of a section are thus adequately representative, and fairly reliable indicators of current direction for the entire section.

8.4.6 Cross-bedding variability

Statistical analysis of variance of cross-bedding allows for the estimation of the reliability of mean sediment transport direction in a particular region. Many authors have studied the dispersion or variability of cross-bedding (Ramblin, 1958: Pelletier, 1958: Pryor, 1960: and Casshyap, 1968). Potter and Pettijohn (1963) discussed the relationship between cross-bedding variance and environment, and concluded that an overlap exists between the variances of all three major environments: fluvio-deltaic, marine, and aeolian, and that the thickness of the stratigraphic unit does not have any significant effect on regional variance.

The paleocurrent map (fig.8.9) shows by means of arrow heads the mean cross-bedding directions at each exposure. Arrow length indicates the vector strength or degree of dispersion. The mean azimuth direction in most areas shows low variability, especially in the south, whilst in the north the current behaviour appears to have been more erratic. On the average the resultant direction indicates transport from north, northeast, northwest, and southeast. Such a variation in azimuthal direction indicates the local behaviour of the depositing currents. The mean current direction is more consistent on a sub-regional

Fig. 8.9.



Bima Sandstones cross-bedding vector means

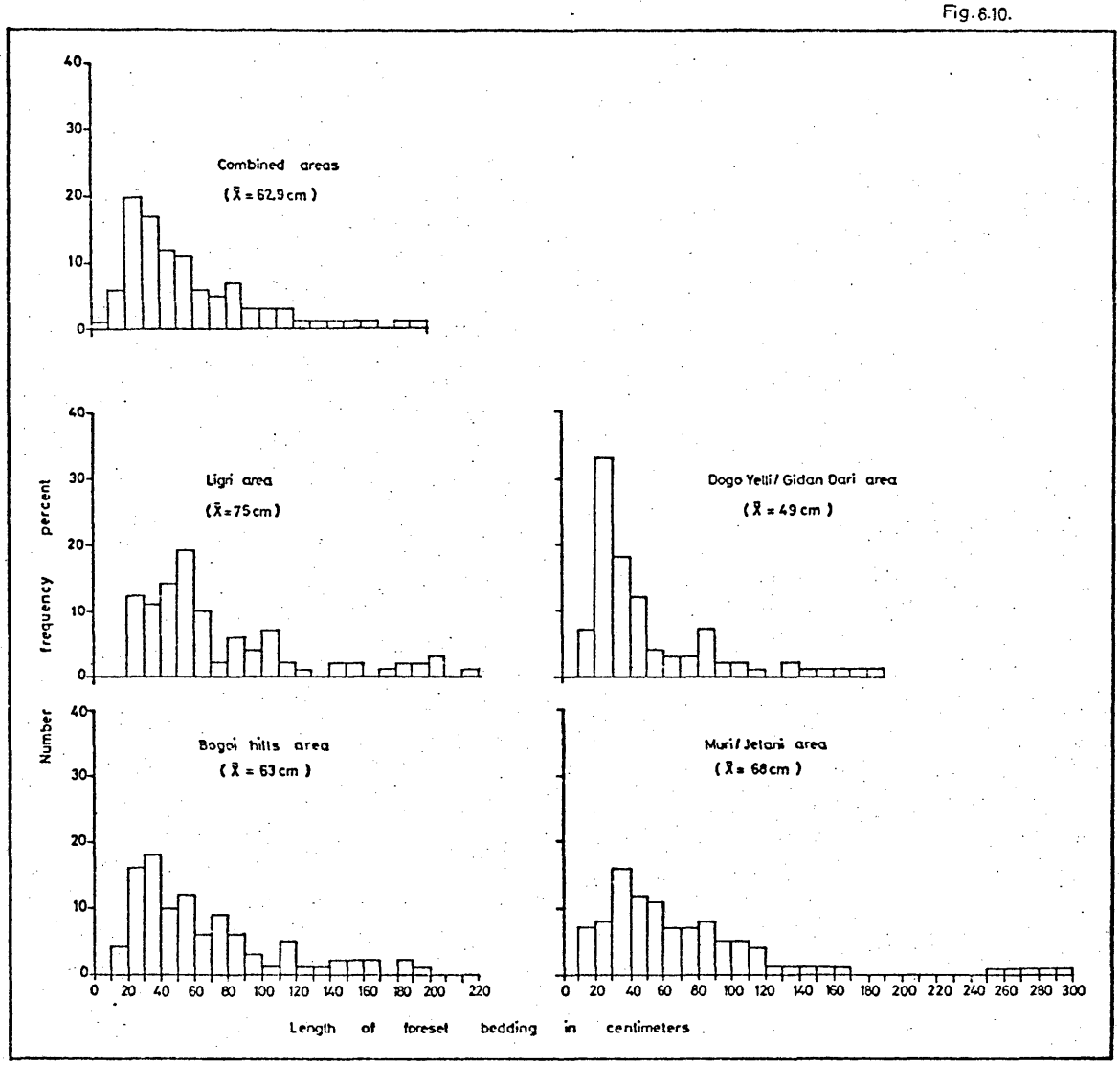
level (fig.8.9). In all the four sub-regions the azimuthal direction is to the south-west. The direction is also to the southwest in both the northern and southern sub-regions. The vector strength is generally moderate (39 % - 57 %), and there is a small range in variance (3589 to 4654). This small range suggests that the current variability was relatively low and more uniform over the whole area of study.

8.4.7 Length of foreset beds

The length of a foreset bed as referred to here may be taken as the distance along the foreset between the base and the top of the containing cross-bedding unit. Casshyap (1968) measured the length of foreset beds and showed their frequency distribution by means of histograms. A similar work was carried out in the present area of study. The lengths of the foreset beds were measured at the places where the azimuth and the scale of cross-bedding were recorded. A total of 568 measurements were taken for the whole area. The length of foreset beds shows a wide range: 7cm to 300cm, with most lengths lying between 10 and 90cm. The calculated average foreset length of the Bogoi Hills, Muri/Jelani, Dogo Yelli, and Ligri areas is 68cm, 49cm and 75cm respectively (fig. 8.10). The average length for the entire area's foreset beds is 62.9cm.

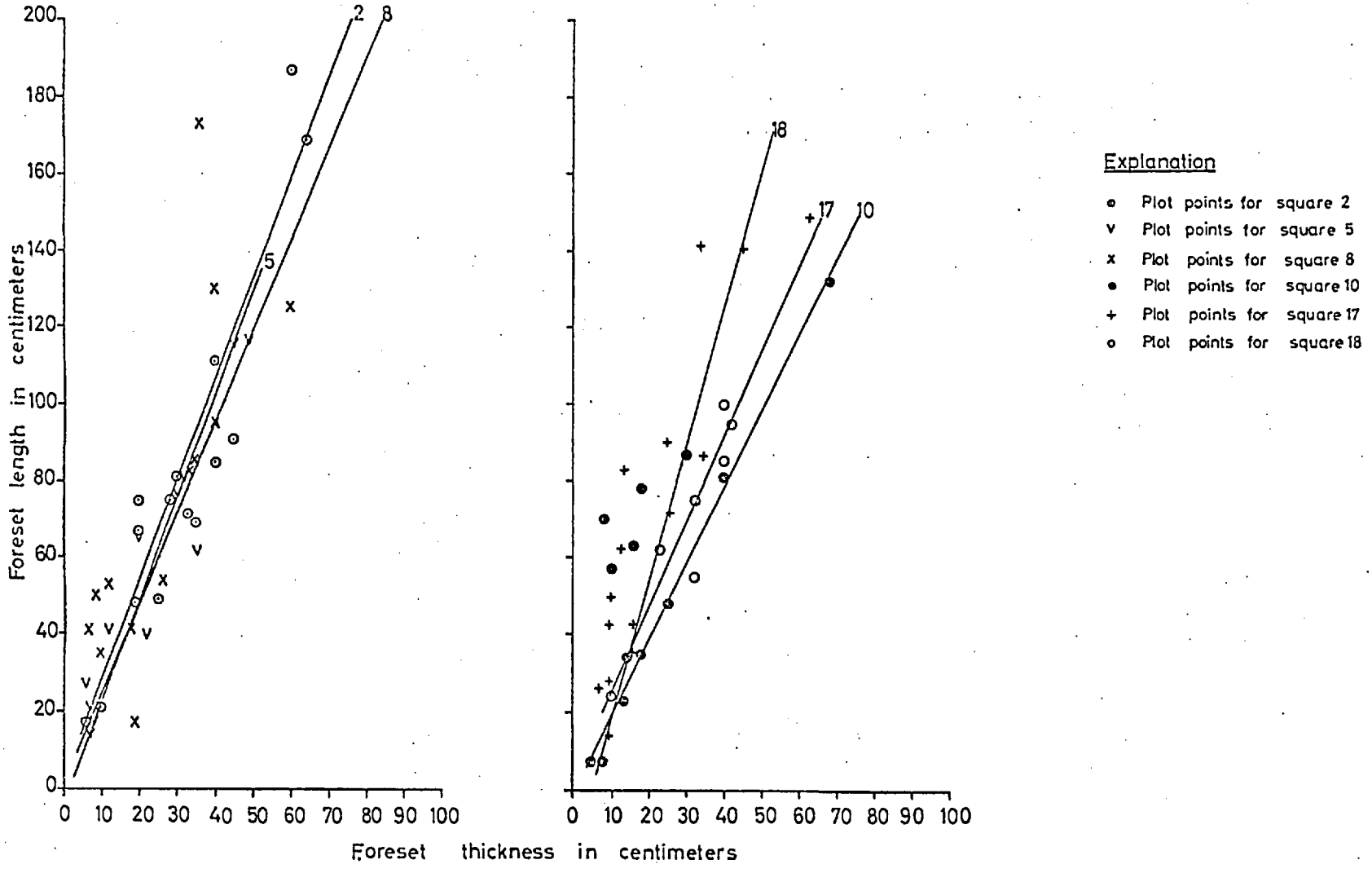
The relationship between the length and the thickness of foreset beds is linear (fig. 8.11). There is a demonstrable relative increase in foreset length with increase in thickness of foresets. Casshyap (1968) mentioned the fact that the length of a foreset bed could be related to the geometry of the depositing current and the thickness of the cross-bedding unit. There is no demonstrable relationship between the length and the inclination of foreset beds. However, for a fixed

Fig. 6.10.



Histograms showing frequency distribution of length of foreset beds in the Bima Sandstones

Fig. 8.11.



Graphs showing the relationship between length and thickness of foreset beds in the Bima Sandstones

thickness of a cross-bedded unit the length of the foreset bed should be a function of its inclination.

8.4.8 Genetic significance of cross-bedding

Sorby (1859) was among the first to observe that ripples are products of currents and that cross-bedding is formed as a result of the deposition of material on the lee side of ripples. Other conditions which have been advanced for the formation of cross-bedding include the down-current formation and migration of large ripples or megaripples (Kindle 1917), sand waves (Pettijohn 1957, and Pettijohn et al. 1972), movement of dunes (Gilbert 1899), and the formation of deltas and alluvial fans. Pettijohn et al. (1972) attributed the formation of small-scale cross-laminations to ripple migration, medium-scale structures characteristic of most fluvial deposits to the migration of sub-aqueous dunes, and large-scale cross-bedding to the migration of large dunes of either subaqueous or eolian origin. Cross-stratification has been described from sediments of various environments, but no authentic criteria, either from theoretical considerations or empirical observations, have been advanced to relate cross-bedding to the agent or environment of deposition.

From the considerations of the cross-bedding types present in the Bima Sandstones (Chapter 7) it was concluded that the planar cross-stratifications were formed by the migration of ripples and dunes over a bar slip face, and the trough cross-strata are products of the infilling of scours associated with dunes.

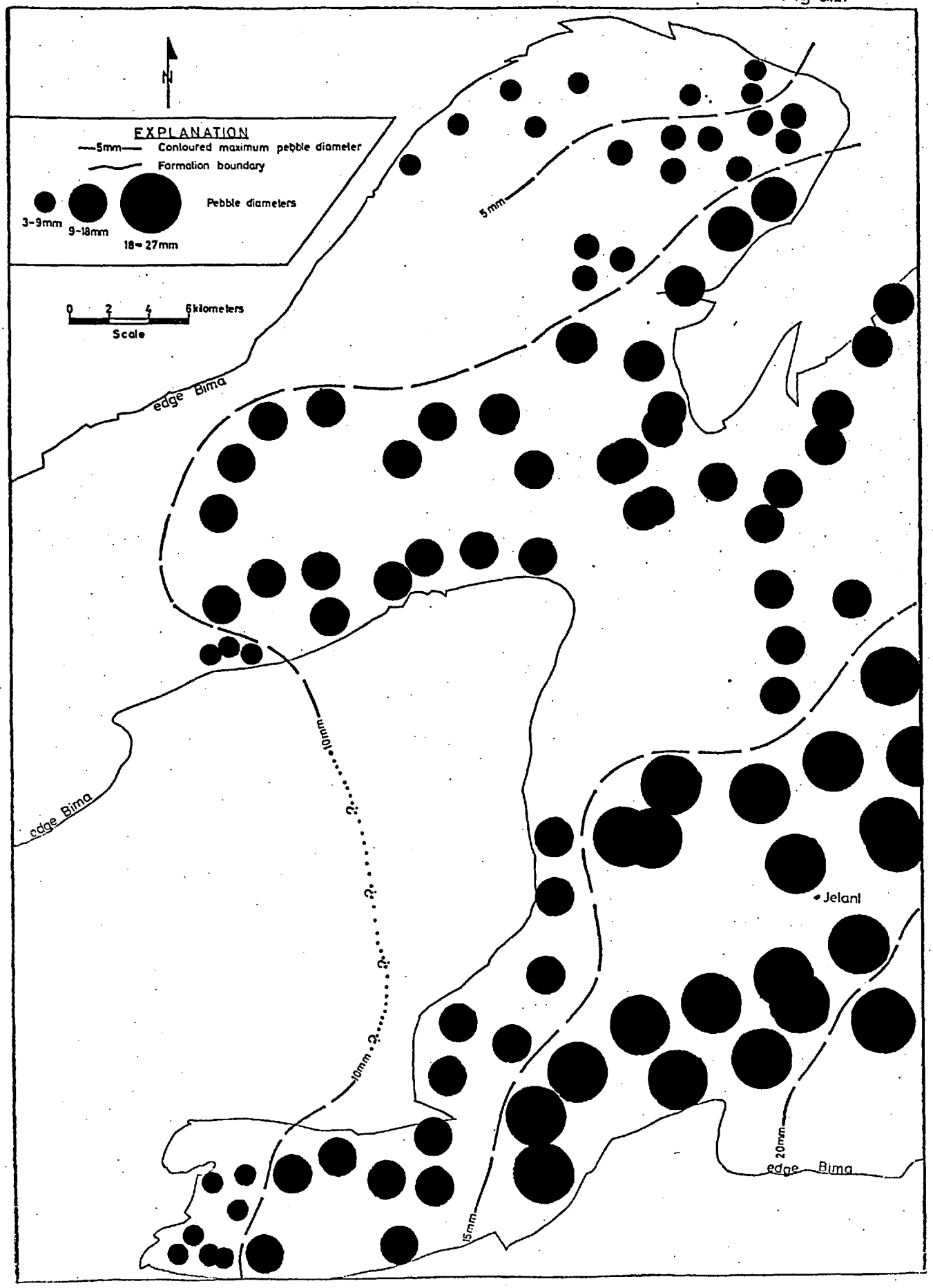
8.5 MAXIMUM PEBBLE SIZE

Decrease in particle size with distance of transport has been used as a paleocurrent indicator in predominantly fluviatile deposits. Studies on particle size have been carried out by Sternberg (in Barrell, 1925), Plumley (1948), and more recently by Pelletier (1958) and Yeakel (1962). Such a correlation of particle size with distance of transport may be largely due to selective transport on mineral grains carried by the depositing medium. The use of particle size as paleocurrent indicator was first suggested by Pettijohn (1957a) who stated that the mean or maximum size of clastic particles could be useful in the determination of the direction and distance of transport of such sediments. This view is supported by studies carried out on fluvial gravels by Potter (1955) and Schlee (1957), on Pocono conglomerates by Pelletier (1958), and on the Tuscarora Formation by Yeakel (1962).

The average of the ten largest quartz pebbles in the Bima Sandstones in the study area was recorded for each locality. The largest pebbles occur in the Jelani area and have an average length of 18mm. In the southwestern and the northeastern portions of the area mapped the pebbles averaged 3 to 6mm.

The diameter of all the pebbles measured has been plotted (fig.8.12) in three major classes: 0-9mm, 9-18mm, and 18-27mm. A progressive decrease in pebble size from the Jelani area westward and southwestward is apparent. There is also a decrease in pebble size northeastwards. Such a decrease is not expected on account of the results obtained from the cross-bedding azimuth. These studies revealed a current flowing in a general west to southwest direction, and therefore a source area lying somewhere to the east of the area of study. A similar deviation

Fig. 8.12.

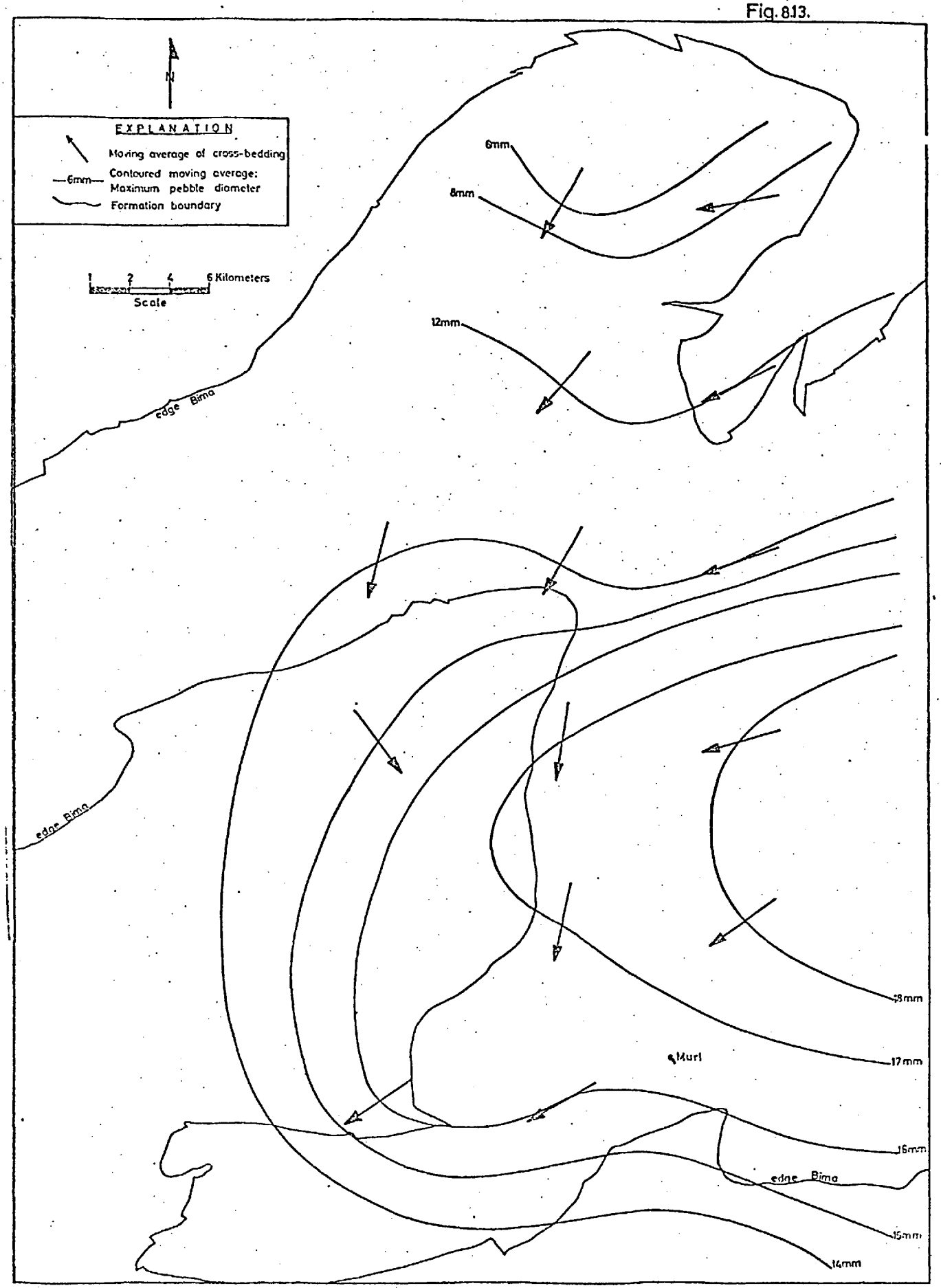


Bima Sandstones depositional framework.

from the expected result was also observed in the Bima Sandstones cross-bedding studies in this same portion of the study area. These anomalous results are probably due to the lack of stratigraphic control on sampling locations, arising from the absence of marker horizons in the Bima Sandstones in these parts of the Benue Valley Trough. At its top the Bima Sandstones get finer-grained. Both factors could have strongly influenced the results obtained on the cross-bedding thickness and pebble sizes in the northeastern part of the area of study.

The Bima Sandstones data indicate a source area to the east with currents that travelled westward and southwestward down an ancient regional slope. The progressive decrease from 20mm in the east to 5mm in the southwest (fig. 8.12) indicates that the ancient regional slope was highest east of the present area of study and was declining toward the west and southwest. Fig. 8.13 shows contoured moving averages based on a 5-minute grid and plotted at the grid intersections. Although contoured moving averages show more clearly the regional pebble size trend of an area, the Bima Sandstones data are too scanty and do not cover enough of the Bima outcrop area to provide reliable interpretation of the size trend. The contours drawn from the moving averages in the study area have a general east to southwest trend, and pebble size decreases to the southwest. The parallel relationship between contour trend of maximum pebble size, pebble size trend, and foreset dip direction is more apparent in fig. 8.12. The contours are drawn at irregular intervals, using the locality pebble size averages. The resulting transport direction as shown by the contours is from northeast to southwest. A detailed sampling of the whole area for the maximum

Fig. 8.13.



Bima Sandstones depositional framework.

pebble diameter over the entire extent of the Bima Sandstones outcrop is recommended to provide a greater number of control points for the establishment of more reliable contoured moving averages of the pebble sizes.

8.6 PALEOGEOGRAPHY AND PALEOENVIRONMENTAL RECONSTRUCTION

8.6.1 General statement

The paleogeography and paleo-environmental discussion of the Bima Sandstone in the area of study is based on field evidence, such as the sequence of sedimentary structures, bedding characteristics, textural changes, paleocurrents analysis, and heavy minerals study. The discussion is based on material in this thesis and also on various other sources.

8.6.2 Origin and environment of deposition of the Bima Sandstone

The evidence for the probable origin of the Bima rocks is based on results obtained from petrographic and heavy minerals study in conjunction with the results of the paleocurrents analysis. The petrographic studies indicate a high feldspar content of the rocks and the presence of metamorphic rock fragments. Both types of mineral grains are characteristic of arkoses and their presence indicates short distance of transport and rapid deposition in a rapidly subsiding basin. The predominant heavy mineral suite is characteristic of granitic source rocks.

The nature of sediment accretion within fluvial channels was suggested by Visher (1965) to be dependent on the size and the quantity of sediment in transport. Large volumes of coarse detritus are

typically deposited in broad channels, with low, flat point bars, while fine detritus is typically transported in deep narrow channels with steep point bar slopes (Schumm, 1963). The predominance of coarse-grained material and the thickness of the depositional units within the Bima Sandstone lead to the conclusion that these rocks were deposited in broad channels with low flat point bars, and that the sediment accretion was intermittent. The material deposited during a single event or flood was of a maximum of 1.5 meters.

The lack of variation in the lithology and sedimentary structures in the rocks suggests that the physical processes operating during the deposition of the Bima Sandstones in the area of study must have been uniform over a long period of time. The coarse- to medium-grained size and the predominantly planar cross-bedding nature of the sandstones suggest conditions of deposition under prevailing strong current activity. The presence of numerous fresh grains of feldspar indicates that the sediments did not travel far from their source areas to the depositional site.

The coarseness of grain size, the presence of medium- to large-scale planar cross-stratification and deformed cross-stratification suggest that transportation and deposition were brought about by subaqueous traction currents of high velocity. The occasional presence of fine-grained sands and silts suggests that periods of violent current conditions passed into brief periods of quieter circumstances.

The features mentioned above are characteristic of fluvial deposits. Jones (1962) also noted these features in his work on the Lau and Yola areas. He noted that much of the middle and upper parts of the Bima

Sandstone are fluvial deposits of coarse- and medium-grained sands, and that river and stream deposits make up by far the major portion of the Bima Sandstone.

Kartashov (1961) stated that deposits of a braided river are made up of essentially coarse-grained channel facies, and that fine-grained material is carried further on and not deposited on flood plains and oxbow lakes which are typically absent. Murat (1970) identified deformed cross-beds as structures of braided stream deposits. Jones (1962) concluded that the Bima Sandstones are deposits of braided streams of the inland part of a delta or of an alluvial plain. In the present area of study the rocks show characteristics of braided stream deposits. Field and laboratory evidence shows that the sandstones are predominantly coarse-grained, with large-scale cross-stratifications, some of which are deformed. From these observations it can be concluded that the Bima Sandstones are deposits of a fluviatile, braided stream environment. The deposition of the sediments probably occurred in many separate channels, resulting from the migration by a process of scour and fill currents, as justified by the presence of a layer of quartz pebbles along erosional planes throughout the vertical and lateral extent of the formation.

A small variance value for the cross-bedding dip azimuths has also been used as an indication of a braided stream origin of some fluvial deposits. Selley (1963) and Barrett (1970) used the unimodal pattern of the current bedding dip azimuths as indicating a braided stream origin for the Red Facies of northwest Britain, and the Fairchild Formation in Antarctica respectively. The Bima Sandstones

show variability of cross-bedding dip values which are to a large extent unimodal on a regional scale. The low scatter in the dip azimuths of the Bima Sandstones lends further support to their braided stream origin.

8.6.3 Paleogeographic synthesis

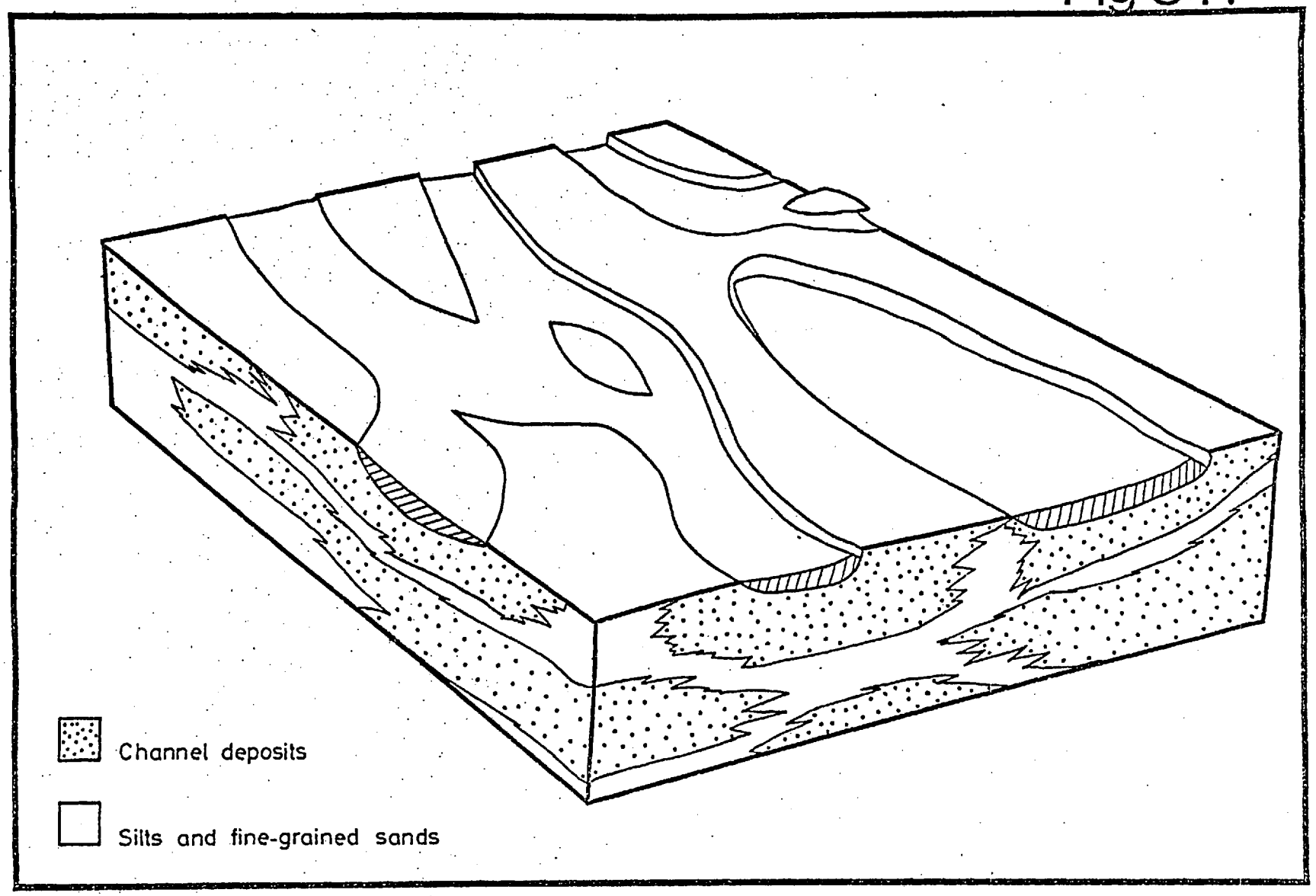
The general trend of events during the Cretaceous in Nigeria is that of alternating transgression and regression of the sea. There was only one major and well-marked regressive phase and it was responsible for the deposition of a vast thickness of continental clastics, the Bima Sandstones. The regressive phases are marked by the development of deltaic sequences (Murat, 1970) to the south of the present area of study. The earliest known marine Cretaceous deposits in the Benue trough occur in aread around the Lamurde anticline (Jones, 1962).

Field evidence and the work of many authors, e.g. Jones (1962), Cratchley and Jones (1965), Carter et al. (1963), and Murat (1970), show that the sequence of depositional events during Bima times was that of marine and deltaic accumulation of sediments in the southern portion of Nigeria, while the present area of study was being copiously supplied with terrigenous clastics from inland areas to the north, northwest and northeast. The picture was probably that of a number of braided streams supplying sediments to the Cretaceous coastline (Fig.8.14) lying to the south of the present area of study. It is, therefore, envisaged that during Bima times the sea did not extend to the present area of study. Jones (1962, page 130) reported marine deposits in the lower parts of the Bima in areas to the south of the Yola and Lau areas, while in the present area of study the entire exposed

portion of the Bima Sandstones (Upper Bima Sandstone, Jones, 1962; Carter et al., 1963) represents fluvial deposits of a braided stream depositional environment. This observation supports Jones' conclusion that a longer period of continental conditions upheld in the north than in the southern portion of the basin. Post-Bima times experienced another major marine transgression which is represented in the area of study by the transitional Yolde Formation and the marine Pindiga Formation. Thus, to sum up the depositional events in the area of study, there was an early pre-Bima marine transgression (not exposed), followed by a major regressive phase, resulting in the deposition of the continental Bima Sandstones, and a post-Bima marine transgression represented by the Yolde and Pindiga Formations.

The volcanic episodes which are represented by basaltic cones and plugs in the area of study probably belong to a post-Bima stage of the Cretaceous, or they may be Tertiary in age. Cratchley and Jones (1965) suggested a Tertiary age for them. The final event which affected the sediments in the area of study is that of mineralization associated with tectonism. In the Ligri area mineralization is associated with the volcanic cones and plugs. Elsewhere in the area of study the mineralization is associated with igneous intrusions which occur along a belt, about 350 miles long, extending from southwest of Abakaliki to the Jawara Hills in the northeast (fig. 1.2). The most commonly occurring minerals along this zone in the area of study are galena, sphaerelite, and native salt (trona).

Fig.8.14.



Diagramatic representation of the lithofacies and the depositional environment of the Bima Sandstones

CHAPTER 9

CONCLUSIONS

From the present study the following concluding statements are made concerning the Bima Sandstones in the area investigated:

(1) The Bima Sandstones is the oldest of the three formations recognized in the area studied. The other two are the Yolde and the Pindiga Formations.

(2) Structurally the Bima Sandstones and the other Cretaceous sediments in the area of study were affected by one episode of tectonism. The fold structures were produced as a result of compressional stresses operating from the northwest and the southeast. The compressional stresses from the south were probably greater than those from the north, as indicated by the presence of steeper dip values in the northern limbs with correspondingly gentler dips in the southern limbs of all the fold structures. A majority of the fractures are oriented normal to the fold axes and the bedding trends. These fractures are formed by tensional stresses parallel to the fold axes. The longitudinal fractures occurring at the nose of the folds in the southern portion of the area are attributed to the release of the principal stresses that folded the rocks.

(3) The Bima Sandstones stratigraphy in the area of study indicates a monotonous lithologic sequence and primary stratification types. The sandstones are predominantly grey and pink in colour. The texture, lithologic sequence and the primary

sedimentary structures compare closely with those described for the upper unit of the Bima Sandstones in the Standard Section. This observation and the stratigraphic position of the Bima Sandstones with respect to the other two formations in the area lead the writer to conclude that the Bima Sandstones in the area of study are equivalents of the Upper Bima Sandstones recognized by Carter et al. (1963) and Jones (1962) in the Standard Section.

(4) The heavy mineral assemblage of the sandstones is dominated by a suite of minerals typical of granitic and highly metamorphosed rocks. The results of the non-opaque minerals study show little variation of the major components in the area studied. Since most of the dominant heavy minerals are common in all the sections, both vertically and laterally, their presence proves to be of little use in correlation. They are, therefore, used only in a general way to determine provenance.

Zircon elongation studies show no significant dispersion in the trends of the reduced major axes which could be related to differences in physiochemical conditions under which the zircons crystallized. It is, therefore, inferred from the foregoing considerations that the physiochemical conditions of crystallization and growth trends were the same for the zircon crystals, and that they, and consequently the sediments containing them, have a common source.

The results of the elongation frequency curves for the Bima Sandstones specimens indicate that the rocks are ill-sorted,

which agrees with the conclusion reached from a consideration of the grain size of the light minerals in the petrographic studies. The similarities of the Bima Sandstones zircon average elongation ratio to that obtained for granites in the Ben Lomond area of California (Spotts, 1962) indicate that the Bima Sandstones were derived from granitic source rocks, and that the sediments were transported only for short distances. The short distance of transport did not give enough time for sedimentary processes to alter the original characteristics of the zircons of the granitic masses from which the Bima Sandstones were derived.

(5) A classification of the Bima Sandstones based on the results of the modal analysis, and using Folk's three end-member ternary diagram, shows that the rocks are arkoses. The clay minerals present are kaolinite and illite, which are authigenic products of feldspar degradation during the weathering of the rocks, probably at the outcrop level. Subsurface data are needed to determine the extent of the weathering processes and the type of clay minerals present in the rocks in the unweathered rock material.

Textural studies indicate that the sandstones are predominantly medium- to coarse-grained in size, and the overall grain size range is from fine to coarse. These results indicate the poorly sorted nature of the Bima Sandstones in the area of study.

(6) The diagenetic sequence proposed for the Bima Sandstones is as follows: the development of quartz overgrowths on detrital

quartz grains is followed by the decomposition of the feldspars to form clay minerals, chert and micas, and the precipitation of calcite cement in the carbonate cemented sands. The source of the secondary silica is believed to be silica-rich groundwaters and silica released during the decomposition of the feldspars. The carbonate cement was probably derived from the dissolution of the shells in the shelly sands and from the percolating groundwaters that are locally rich in calcium and carbonate ions.

(7) The morphology of the dominant type of cross-stratification and the coarse grain size of the sediments present in the area of study lead the writer to conclude that the sandstones were deposited by fast flowing traction currents, that they were probably formed by the avalanching of material from dunes and ripples migrating over a bar slip face. The restricted occurrence of fine-grained and argillaceous materials with the scarcity of stratification types typical of weak turbulence indicate the absence of conditions conducive to slow settling from suspension during the deposition of the Bima Sandstones.

The deformed variety of cross-stratifications present in the sandstones in the area of study are believed to have been formed by a combination of the action of earthquakes and sediment-laden currents.

(8) The cross-bedding azimuth for the sandstones is generally in the western quadrant, and the source area for the sediments lies generally to the east. A vertical variation in cross-bedding orientation indicates a generally consistent current direction

throughout the Bima times to the southwest.

The maximum pebble size and the cross-bedding thickness corroborate the conclusion that the source of the Bima sediments lies generally to the east of the area of study, and that the current direction was generally southwesterly.

A consideration of all the results obtained for the Bima Sandstones study leads the writer to conclude that the sandstones were deposited in a braided stream environment as channel deposits within a basin that subsided rapidly while receiving a copious supply of sediments from the granitic masses in the hinterland to the northeast.

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APPENDIX 1: Statistical parameters of grain size measurements in the Bima Sandstones

Sample No.		ϕ5	ϕ16	ϕ25	ϕ50	ϕ75	ϕ84	ϕ95	Sample locality
BN	3	1.20	1.49	1.61	1.90	2.22	2.39	2.69	Bogoi Hill
BN	4	0.61	1.28	1.54	1.90	2.33	2.42	2.83	"
BN	10	0.89	1.02	1.16	1.41	1.66	1.78	2.00	"
BN	15	1.36	1.79	1.92	2.30	2.55	2.60	2.70	"
BN	39	0.09	0.38	0.50	0.80	1.26	1.58	2.01	"
BN	40	0.10	0.80	1.13	1.72	2.18	2.36	2.77	"
AN	46	0.34	0.68	0.82	1.30	1.63	1.74	1.95	Andami
AN	47	-0.20	0.59	0.74	1.11	1.55	1.67	1.91	"
AN	51	0.25	0.61	0.80	1.21	1.73	2.01	2.57	"
AN	53	-0.03	0.28	0.40	0.72	0104	1.25	1.80	Jelani
JE	61	0.56	0.89	1.30	1.38	1.70	1.86	2.18	"
JE	62	0.17	0.80	0.98	1.38	1.81	2.00	2.28	"
JE	64	0.65	1.15	1.40	1.74	2.19	2.44	2.62	"
JE	66	0.27	0.54	0.68	0.95	1.23	1.38	1.60	"
JE	70	0.62	0.98	1.15	ϕ.50	1.86	2.05	2.52	"
JE	72	0.21	0.65	0.85	1.29	1.65	1.79	2.04	"
BN	79	0.82	1.31	1.51	1.80	2.08	2.21	2.50	Bashama
BS	87	0.25	0.55	0.70	1.00	1.29	1.41	1.50	"
BS	89	0.61	1.10	1.31	1.68	1.95	2.03	2.19	"
BS	91	1.31	1.78	1.92	2.34	2.82	3.01	2.14	"
JN	98	1.98	2.17	2.27	2.47	2.78	2.91	3.10	Ndela
JN	101	0.29	0.58	0.60	0.90	1.35	1.50	1.64	"
JN	105	0.50	0.80	0.93	1.24	1.55	1.68	1.90	"
JL	108	-0.31	0.40	0.75	1.55	1.87	2.01	2.21	Ligri
JL	111	0.41	0.71	0.87	1.20	1.55	1.68	1.93	"
JL	113	-0.23	0.20	0.40	0.88	1.50	1.57	1.68	"
JL	115	0.99	1.28	1.40	1.61	1.79	1.88	2.04	"
JL	116	-0.40	0.27	0.58	0.90	1.48	1.55	1.68	"
JL	119	0.42	0.68	0.78	1.10	1.40	1.58	1.91	"
JL	120	1.10	1.51	1.68	2.01	2.30	2.41	3.10	"
GH	123	0.58	0.86	1.00	1.32	1.70	1.80	2.10	Gwana
D	128	0.11	0.74	1.00	1.60	1.90	2.04	2.15	Diji

APPENDIX 2: Roundness and sphericity values of samples from various localities in the Bima Sandstones.

Sample No.		Sphericity Mean value	Roundness Mean value	Sample Locality
BN	3	0.66	0.66	Bagoi Hills
BN	4	0.74	0.52	"
BN	5	0.71	0.58	"
BN	10	0.75	0.61	"
BN	14	0.75	0.56	"
BN	15	0.73	0.55	"
BN	16	0.68	0.63	"
BN	39	0.72	0.62	"
BN	40	0.74	0.49	"
BN	42	0.71	0.48	"
YH	46	0.56	0.39	Andami
YH	48	0.69	0.58	"
AN	51	0.77	0.37	"
AN	53	0.74	0.54	"
JE	61	0.71	0.51	Jelani
JE	62	0.73	0.52	"
JE	64	0.75	0.52	"
JE	66	0.74	0.56	"
JE	70	0.69	0.51	"
JE	72	0.71	0.55	"
BN	79	0.67	0.55	Bashama Hills
BS	87	0.70	0.58	"
BS	89	0.71	0.53	"
BS	91	0.64	0.53	"
BS	94	0.65	0.44	"
JN	98	0.68	0.58	Ndela
JN	101	0.82	0.58	"
JN	102	0.73	0.63	"
JN	105	0.61	0.64	"
JL	106	0.68	0.58	Ligri
JL	108	0.68	0.51	"
JH	109	0.68	0.51	"
JL	111	0.67	0.35	"
JL	112	0.69	0.50	"
JL	113	0.68	0.51	"
JL	114	0.77	0.54	"
JL	115	0.68	0.55	"
JL	116	0.72	0.55	"
JL	119	0.76	0.55	"
JL	120	0.68	0.53	"
GH	123	0.74	0.64	Gwana hills
D	127	0.73	0.61	Diji
D	128	0.75	0.54	"

APPENDIX 3: Grain size measurements of representative samples from the Bima Sandstones.

Sample No.	Median	Mz Graphic Mean	Standard Deviation (dl)	Skewness (SKL)	Kurtosis (kg)
BN 3	1.90	1.93	0.45	0.07	1.00
BN 4	1.90	1.87	0.62	-0.12	1.15
BN 10	1.42	1.40	0.36	0.02	0.91
BN 15	2.31	2.23	0.41	-0.33	0.87
BN 39	0.80	0.92	0.59	0.28	1.04
BN 40	1.74	1.62	0.79	-0.20	1.04
AN 46	1.30	1.24	0.51	-0.18	0.81
AN 47	1.11	1.12	0.59	-0.10	1.07
AN 51	1.21	1.28	0.70	0.16	1.02
AN 53	0.72	0.75	0.52	0.14	1.17
JE 61	1.37	1.38	0.49	-0.01	1.66
JE 62	1.37	1.39	0.62	-0.06	1.04
JE 64	1.75	1.78	0.62	-0.01	1.02
JE 66	0.96	0.96	0.41	-0.00	0.99
JE 70	1.51	1.51	0.56	0.05	1.10
JE 72	1.28	1.24	0.56	-0.15	0.94
BN 79	1.80	1.77	0.48	-0.13	1.21
BS 87	1.00	0.99	0.40	-0.12	0.87
BS 89	1.69	1.60	0.47	-0.30	1.01
BS 91	3.34	2.36	0.60	-0.04	0.83
JN 98	2.47	2.52	0.35	0.16	0.90
JN 101	0.98	0.99	0.43	0.20	0.74
JN 105	1.24	1.24	0.43	-0.03	0.93
JL 108	1.50	1.32	0.78	-0.45	0.92
JL 111	1.20	1.20	0.47	-0.03	0.92
JL 113	0.87	0.88	0.63	-0.78	0.71
JL 115	1.62	1.59	0.31	-0.14	1.10
JL 116	0.90	0.91	0.64	-0.12	0.95
JL 119	1.20	1.12	0.45	0.08	0.98
JL 120	2.15	1.98	0.53	-0.01	1.32
GH 123	1.34	1.36	0.49	0.07	0.89
D 128	1.59	1.46	0.63	-0.39	0.93

APPENDIX 4.1: Modal analysis of bulk mineral assemblages for the Bogoi hills area (First sample values represent the highest samples of the sequence).

Sample No.	QUARTZ			FELDSPARS			Rock Frags.	Alteration Products	Carbonates	Opaques	Total No. of counts
	Poly-crystalline	Undulose	Non-undulose	Micro-cline	Plagio-clase	Ortho-clase					
BN 16	3.76	51.23	4.70	0.49	9.87	3.29	1.41	23.97	-	-	210
BN 15	19.80	51.60	6.05	1.65	1.65	3.30	1.65	14.30	-	-	185
BN 12	2.94	47.53	4.90	1.96	8.33	5.88	2.45	24.99	-	-	202
BN 9	1.68	42.00	8.40	2.24	8.68	1.68	0.84	0.84	29.40	1.40	347
BN 4	5.68	75.14	2.04	2.72	9.08	0.34	2.04	1.36	-	0.34	291
BN 3	8.19	44.73	8.61	4.83	14.28	2.94	1.05	16.59	-	0.41	484
BN 1	11.16	45.72	11.88	3.96	-	1.80	2.52	23.76	-	-	280

APPENDIX 4.2: Modal analysis of bulk mineral assemblages for the Ligri area
(First sample values represent the highest samples of the sequence).

Sample No.	QUARTZ			FELDSPARS			Rock frags.	Alteration Products	Opaques	Total No. of counts
	Poly-crystalline	Undulose	Non-undulose	Micro-cline	Plagio-clase	Ortho-clase				
JL 120	9.02	50.43	-	16.40	5.74	3.28	1.23	11.48	-	238
JL 119	6.12	56.52	0.36	18.36	2.16	5.40	3.24	4.68	-	274
JL 118	3.08	50.96	0.28	9.24	6.72	11.76	0.56	12.88	1.68	350
JL 117	0.56	49.28	1.96	18.48	4.76	1.96	0.84	17.08	2 24	357
JL 116	6.44	67.48	1.12	6.44	0.84	1.96	7.84	7.00	0.28	353
JL 115	3.10	70.06	0.93	6.82	2.17	3.10	1.55	10.23	0.31	317
JL 114	1.40	58.24	1.68	12.32	3.92	6.72	1.12	12.04	0.84	344
JL 113	8.99	61.07	-	7.75	1.55	0.93	9.30	9.30	-	319

APPENDIX 4.3: Modal analysis of bulk mineral assemblages for the Jelani area (First sample values represent the highrst samples of the sequence).

Sample No.	QUARTZ			FELDSPARS			Rock frags.	Alteration Products	Carbonates	Opaques	Total No. of counts
	Polycrystalline	Undulose	Non-undulose	Microcline	Plagioclase	Orthoclase					
JE 75	44.75	24.52	1.84	8.58	1.84	6.13	1.23	11.03	-	-	163
JE 73	10.71	30.70	15.35	4.64	2.86	4.64	1.79	1.43	27.85	-	280
JE 71	2.12	30.09	6.02	1.77	4.60	1.06	-	-	54.16	-	282
JE 68	14.67	44.00	8.88	4.63	1.16	6.56	2.32	17.76	-	-	258
JE 67	12.32	48.16	2.80	17.36	5.04	1.12	0.56	11.20	-	-	176
JE 66	23.94	34.03	3.74	10.10	0.75	0.37	1.87	25.06	-	-	267
JE 64	33.48	36.58	1.55	9.61	1.24	0.62	2.48	14.26	-	-	322

APPENDIX 4.4: Modal analysis of bulk mineral assemblages for the bashama hills area (First sample values represent the highest samples of the sequence).

Sample No.	QUARTZ			FELDSPARS			Rock frags.	Alteration Products	Opaques	Micas	Total No. of counts
	Poly-crystalline	Undulose	Non-undulose	Micro-cline	Plagio-clase	Ortho-clase					
BS 94	30.54	27.90	4.15	21.11	-	0.75	1.13	14.33	-	-	265
BS 93	5.19	53.63	1.73	12.46	7.27	3.46	2.77	13.15	-	0.35	289
BS 91	3.22	24.99	12.90	20.15	12.09	3.22	3.22	20.15	-	-	124
BS 89	25.76	34.96	8.28	11.04	0.92	3.68	1.84	12.88	-	-	217
BS 88	30.19	30.19	5.90	10.41	2.43	3.12	0.69	17.00	-	-	288
BS 87	28.42	29.92	6.73	16.83	-	0.37	2.99	14.59	-	-	267

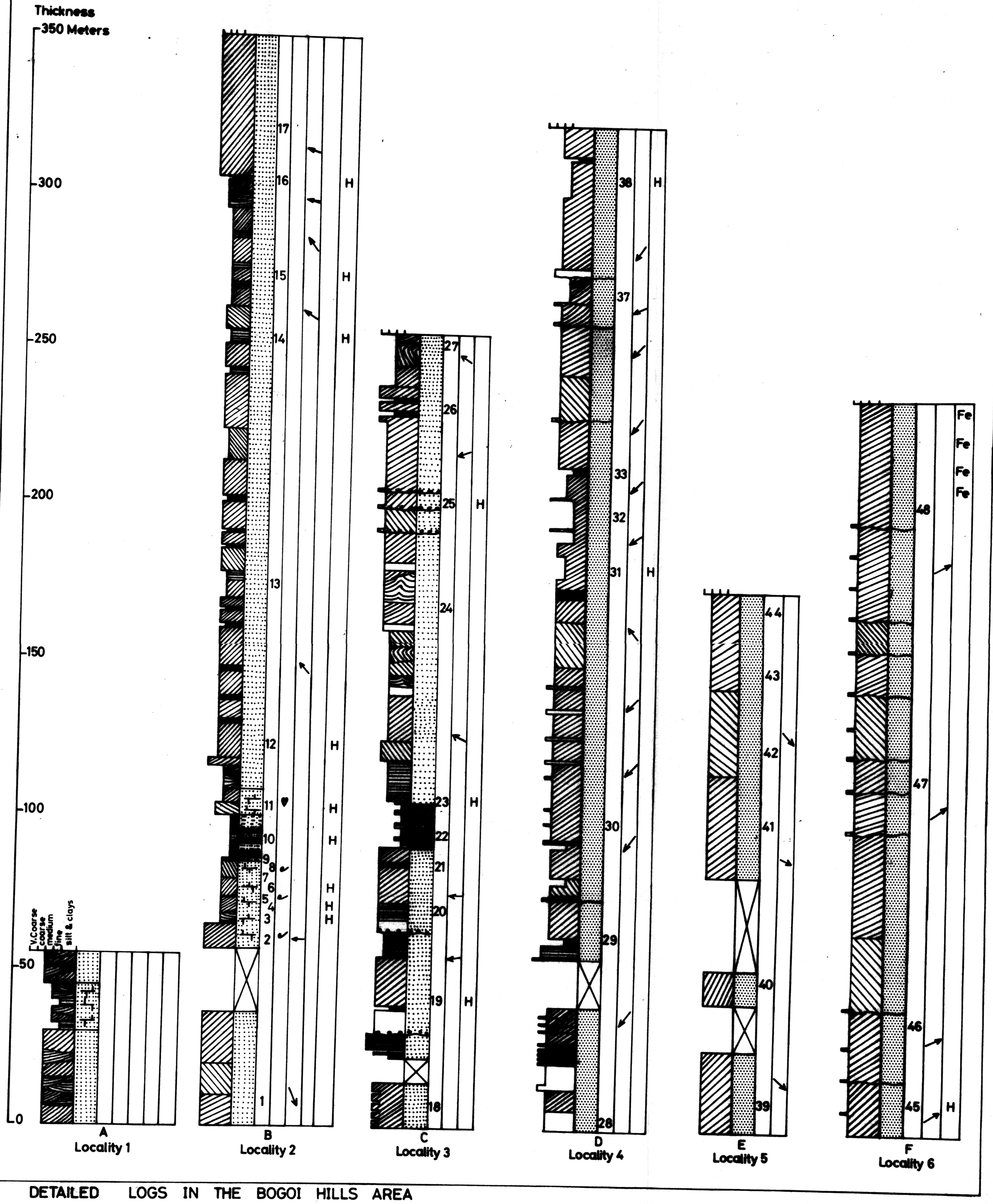
Appendix 5. Mathematical parameters used in paleocurrent analysis.

Point Number	Mid Point Azimuth	Cos - Sin angle	Sign of Cos	Sign of Sine	Cos	Sine
I	I5	I5	+	+	0.966	0.259
2	45	45	+	+	0.707	0.707
3	75	75	+	+	0.259	0.966
4	I05	75	-	+	-0.259	0.966
5	I35	45	-	+	-0.707	0.707
6	I65	I5	-	+	-0.966	0.259
7	I95	I5	-	-	-0.966	-0.259
8	225	45	-	-	-0.707	-0.707
9	225	75	-	-	-0.259	-0.966
I0	285	75	+	-	0.259	-0.966
II	3I5	45	+	-	0.707	-0.707
I2	345	I5	+	-	0.966	-0.259

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Fig. 3.1 (Sheet 1)

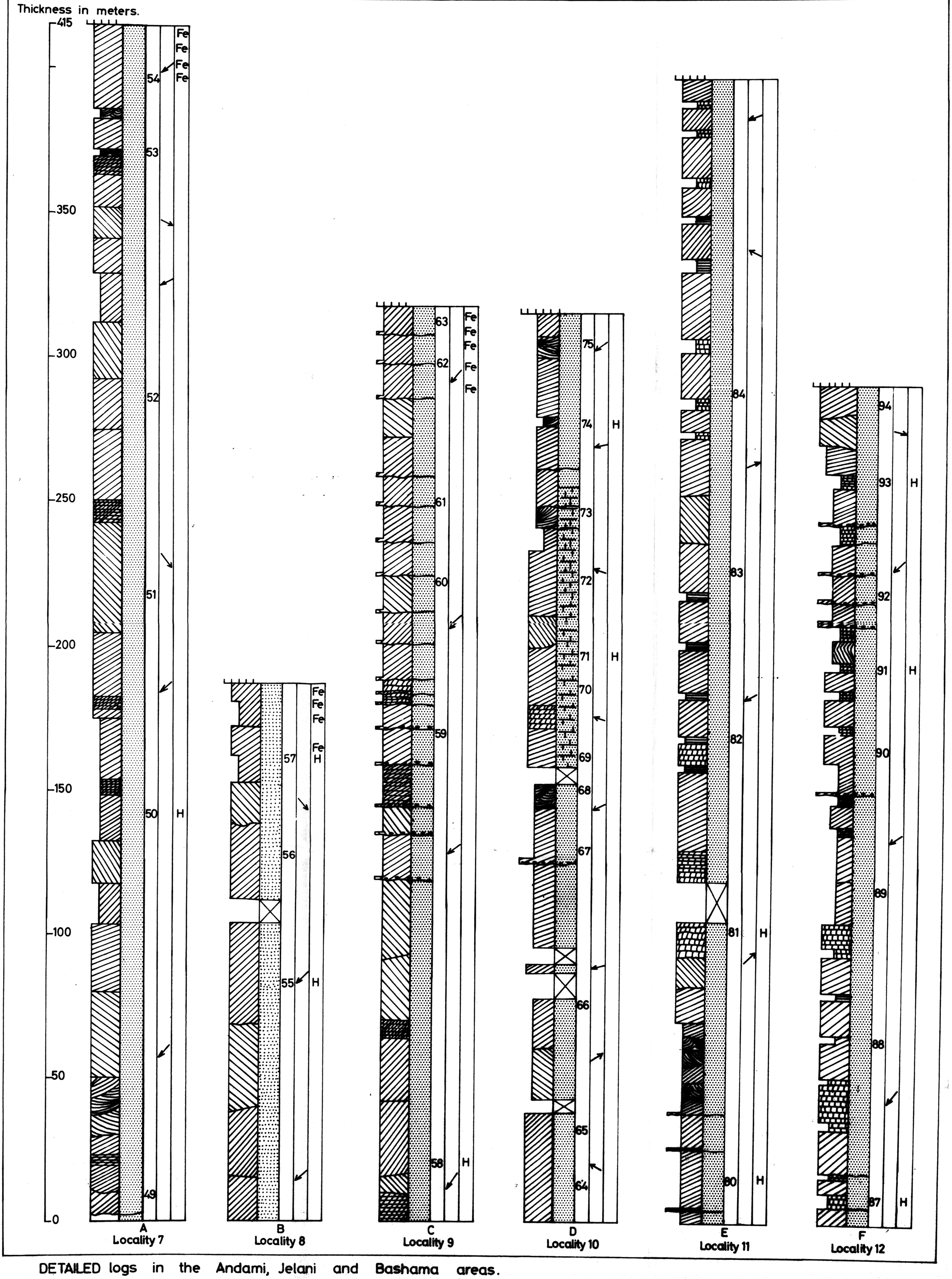
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DETAILED LOGS IN THE BOGOI HILLS AREA

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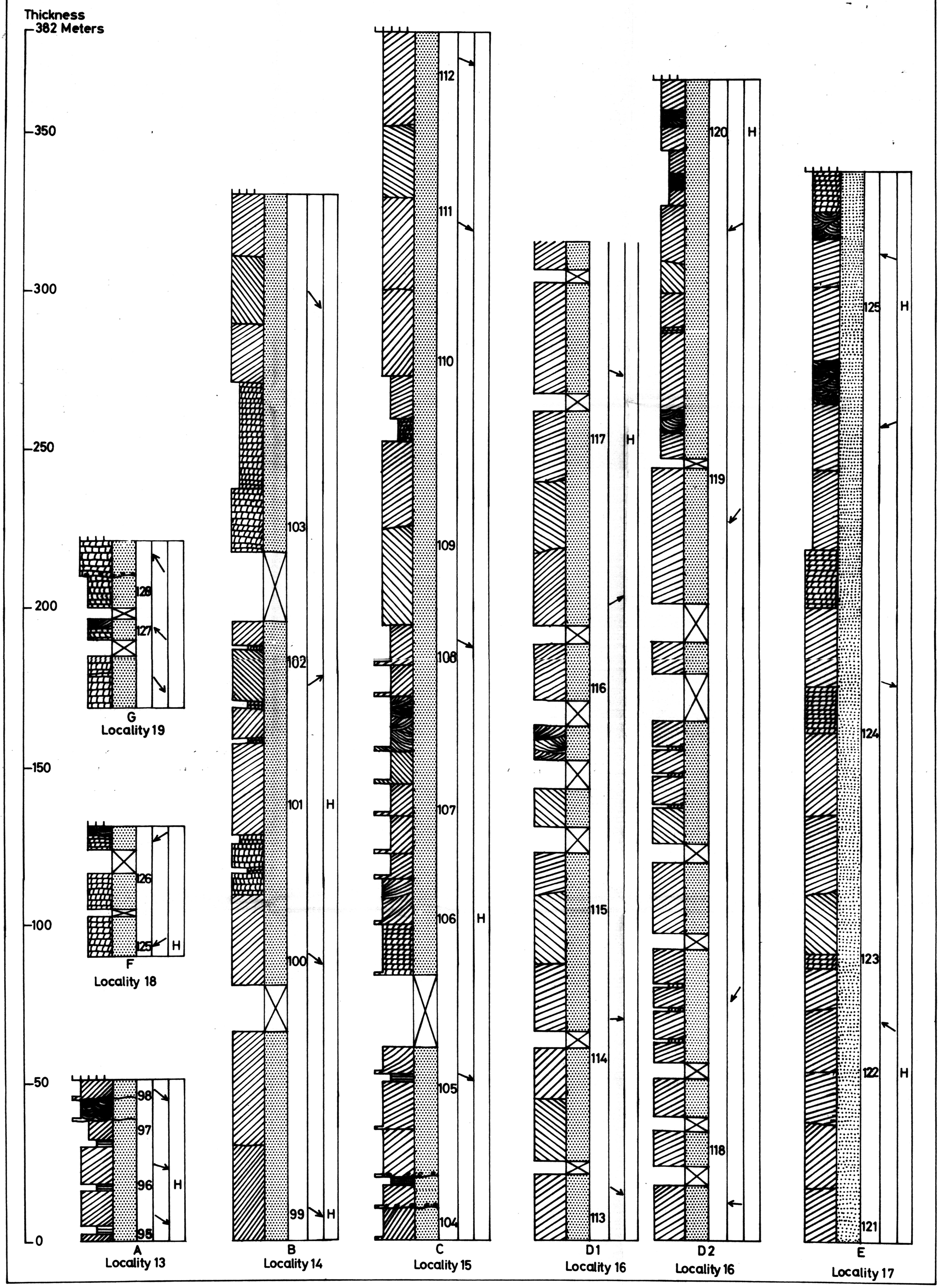
Fig.3.3 (Sheet 2)



DETAILED logs in the Andami, Jelani and Bashama areas.

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Fig. 3.4 (Sheet 3)



DETAILED logs in the Diji, Gwana and Jawara hills areas.

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GEOLOGICAL MAP OF MURI N.E. AND FUTUK S.E. SHEETS 193 & 172 NIGERIA

EXPLANATION

- River Alluvium
- Pindiga Formation
- Yolde Formation
- Bima Sandstones
- Dip and strike of normal bedding
- F Faults and fractures
- Anticlinal axis
- Synclinal axis
- Volcanic plugs
- Ss Saline series and springs
- Lst Limestone

SCALE 1:100,000

0 2 4 6 8 10 kilometers

E10°30′ 10°35′ 10°40′ 10°45′ 10°50′ 10°55′ 11°00′E
N9°45′ 9°40′ 9°35′ 9°30′ 9°25′ 9°20′ 9°15′ 9°10′ 9°05′ N9°00′

SEDIMENTS
YOLDE
POST
JAWARA HILLS
GWANA HILLS
BASHAMA HILLS
BOGOI HILLS
BENUE RIVER

Diji
Gidan Dari
Bashama
Dogo Yelli
Ligri
Gora
Ndela
Jukun
Andami
Jelani
Muri
Lst
Ss
F

Fig.1

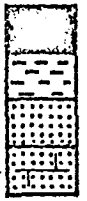
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DETAILED LOGS: SYMBOLS AND EXPLANATION.

BODY FOSSILS.

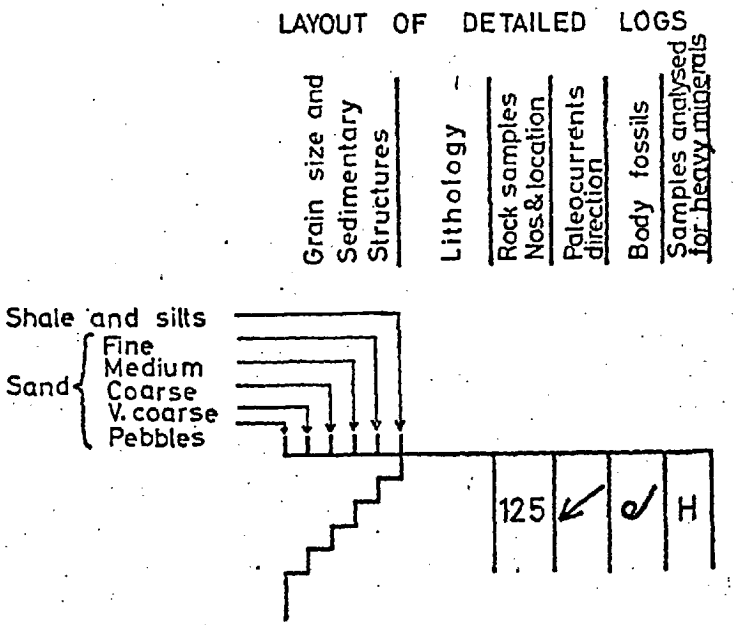
- Bivalves
- Oysters

LITHOLOGIES IN DETAILED AND COLUMNAR LOGS.



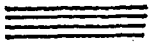
- Shale
- Silts and siltstone
- Sandstone
- Calcareous sandstone

LAYOUT OF DETAILED LOGS



SEDIMENTARY STRUCTURES.

Laminations:



strongly regular.

Foresets:



Festoon foresets.



Thin cross-beds



Thick and medium crossbeds

BED BASE TYPES.

Abrupt and horizontal



Abrupt and inclined



Erosional