

DIAGENESIS IN THE NEOGENE CLASTIC SEDIMENTS OF THE
BENGAL BASIN, BANGLADESH

A thesis submitted for the degree of Doctor
of Philosophy of the University of London.

by

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ABSTRACT

This thesis describes the petrography and burial diagenesis of the Neogene sandstones and shales in a prograding deltaic sequence in the Bengal Basin of Bangladesh. The Surma group, consisting of marine shales at the base overlain by deltaic to paralic sandstones and shales, is succeeded by the Tipam group - a sandstone dominated unit of the terrestrial fluviatile origin. The sequence appears to be devoid of significant petroleum source rocks.

Petrographically, the sandstones of the Surma group are represented typically by sublitharenites whilst the sandstones of the Tipam group are subarkoses. There are two depth-related diagenetic zones in the sandstones: from surface to 2000 m no cementation has occurred in either the Tipam group or the top of the Surma group; from 2000 m to 4491 m+ significant cementation has affected the potential reservoir properties of many of the sandstones of the Surma group. The paragenetic sequence of diagenetic events in the Neogene sandstones indicates two phases. The first phase represents an alkaline condition of the pore waters, under which chlorite authigenesis, siderite crystallization, early poikilotopic Fe-calcite cementation, illite coating, quartz overgrowth cementation and late Fe-calcite cementation took place sequentially. The second phase marks the change from alkaline to acidic condition of the pore water under which partial dissolution of some rock components and the crystallization of kaolinite took place. Early poikilotopic Fe-calcite, late Fe-calcite and quartz overgrowths are the most important porosity reducing cements. Chlorite rims on detrital quartz, however, effectively retard or inhibit quartz overgrowth

development, thus preserving porosity. Secondary porosity was created by the partial dissolution of the calcite cements, detrital feldspars and some lithic grains in the deeper zone.

The predominant minerals in the Neogene shales are illite, kaolinite, chlorite, illite/smectite, quartz and feldspar, whilst calcite, dolomite and siderite occur in minor to trace amounts. The most important diagenetic change with depth of burial is the gradual loss of the smectite components in the illite/smectite mixed layer clays due to their transformation to illite layers. The diagenetic transformation of smectite to illite has not led to complete illitization of the mixed layer clays at the total depth studied. Decrease of kaolinite at and below 4530 m could be diagenetic, but it could also be due to differential settlement of clay particles due to change in depositional environment. The decrease in calcite below about 3500 m represents diagenetic dissolution of this mineral with burial depth. The diagenesis of shale has taken place without respect to formation boundaries and mainly depends on temperature. Diagenesis of the shales is believed to have influenced the cementation in the adjacent sandstones. The Si^{4+} released from the smectite to illite transformation may have provided a source of silica cement, and the dissolution of calcite in shales could be a source for the late calcites in the adjacent sandstones.

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List of abbreviations used in the thesis

Calc. - Calcite	Kaol.- Kaolinite
Cemt. - Cement	Lacust.- Lacustrine
Cementn.- Cementation	Lith.- Lithic grain
Chlor.- Chlorite	Lst.- Limestone
Crs.- Coarse	Meta.- Metamorphic
Cryst.-Crystallization	Matrx.- Matrix
Dissol.- Dissolution	Overgrth.- Overgrowth
Dol.- Dolomite	Poik.-Poikilotopic
Erup.- Eruption	Precip.- Precipitation
Feld.- Feldspar	Pebl.- Pebble
Ferrg.- Ferruginous	Qtz.- Quartz
Fiss.- Fissure	Ran.- Random
Fm.- Formation	Repl.- Replacement
Gr.- Group	Sst.- Sandstone
Grvl.- Gravel	Sh.- Shale
Ill.- Illite	Sid.- Siderite
Ill/Sec.- Illite/Smectite	Vol.- Volcanic
Interstr.- Interstratification.	

CHAPTER 1

INTRODUCTION

1.1 AIM AND SCOPE OF PRESENT WORK

The aim of the present research work was to study the diagenesis in response to burial that has occurred in the Neogene clastic rocks of the Bengal Basin, Bangladesh.

The Bengal Basin, most of which is occupied by Bangladesh, is located in the zone where three large geotectonic features of the Earth's crust come together, i.e., the Himalayan mega-anticlinorium, the Arakan Yoma geanticline and the Pre-Cambrian Indian Shield (Fig. 1.1), and contains the largest deltaic depositional system of the present world. Yet the geology of the Bengal Basin is not adequately known. Geological concepts here suffer from too much generalization (Khan, 1980). The stratigraphy now in use is not well established and the various rock units are rather loosely defined. Current efforts to establish a better defined stratigraphy are at an early stage (Brunschweiler, 1980; Leitz and Kabir, 1982). No work has been reported on the diagenesis of the sediments, and their petrography is also poorly described. No classification (nomenclature) of the Neogene sandstones has been attempted, neither has the mineralogy of the shales been reported. Geological studies in the Bengal Basin were done almost exclusively in the past in connection with economic mineral explorations, especially petroleum exploration. Since the first drilling in 1910 (760 m), 39 exploratory wells have been drilled in Bangladesh resulting in 12 gas discoveries. The Bengal Basin, one of the largest geosynclines in the world, has not received the attention it deserves from research workers. The present work on the diagenesis of the Neogene clastic sediments may be an important, although small,

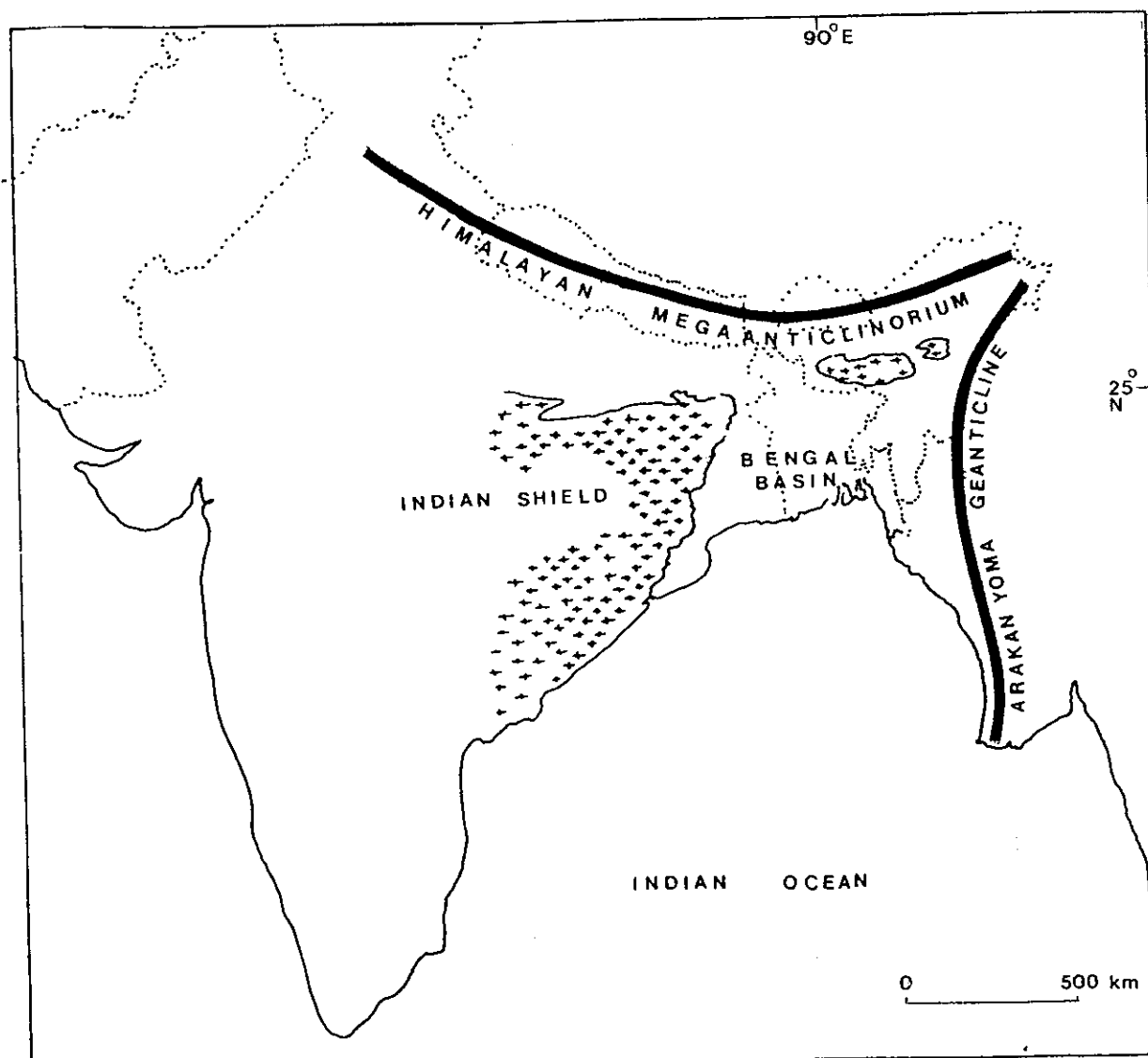


FIG 1.1 BENGAL BASIN AND THE THREE LARGE TECTONIC FEATURES OF THE EARTH

contribution to this otherwise unknown aspect of the geology of the Bengal Basin.

It was originally intended to include a study of the diagenesis of the organic matter in the Neogene shales and to compare the maturation levels of this organic matter with the diagenesis of the clay minerals and the sandstones. The organic carbon percentages of all the shale samples studied, however, were found to be very low and, after macerating several samples, it was found that the organic matter was insufficient to make adequate determination of maturation levels. This led to the decision to concentrate on the inorganic diagenesis that has occurred in the sand and shale of the Neogene sequence.

1.2 DIAGENESIS - A REVIEW

The term "diagenesis" was first introduced more than a hundred years ago by Von Guembel (1868) as a designation for processes which act on sediments after deposition. Since then the term has been applied in varying ways to the post-depositional, pre-metamorphic changes which affect a sediment. It has been used, for example, for (a) all physical and chemical changes that occur within the sediment after deposition; (b) reactions which take place within a sediment between one mineral and another or between one or several other minerals and the interstitial fluids, thus restricting the definition to essentially chemical rather than physical processes after deposition; (c) processes that result in lithification or cementation of the sediments after deposition. An extensive review of the history and development of the study of diagenesis is given by Dunoyer de Segonzac (1968).

In the present study, the term diagenesis is used to include all those processes, chemical and physical, which

affect the sediment after deposition, up to the lowest grade of metamorphism, the greenschist facies (Pettijohn, Potter and Seiver, 1973).

It is recognised that one can only arbitrarily delimit the processes one calls diagenetic from those of metamorphism, since diagenesis and metamorphism represent a continuum in the geologic variables, pressure and temperature. Blatt (1979) indicated that generally the range of physical and chemical conditions involved in diagenesis includes temperatures of 0° C to 200° C, pressure of 1 to 2000 kg/cm. and water salinities from fresh to hypersaline. Winkler (1965) considered that the critical transition temperature between diagenesis and metamorphism was 300° C and was almost independent of pressure. During burial under normal geothermal gradients (30° C/km), this temperature would be reached at depths of about 10,000 m (Muller 1967).

Diagenesis is one of the most important events in the geological history of a sediment. Knowledge of diagenetic processes in sediments is becoming more and more important in applied geology and particularly in petroleum geology. This has led to an increased interest and rapid multiplication of the literature on diagenesis during the last decade. These studies have been recently facilitated by the introduction of advanced techniques such as scanning electron microscopy and Cathodoluminescence.

1.3 SAMPLING

For the purpose of the present study, 40 sandstone and 50 shale samples were selected from five onshore and one offshore petroleum exploration wells (Table 1.1 and Fig. 1.2). All the samples from onshore wells were taken from conventional cores, while most of those from the offshore well are from rock cuttings. These samples range in depth from 500 m to 4550 m below surface and belong to the Tipam

TABLE 1.1: PARTICULARS OF WELLS AND SAMPLES USED IN THE
PRESENT STUDY

Well	Location	Depth range of samples (meter)	Total no. of samples
Muladi-1 (onshore dry hole)	Lat.22° 56'24"N Long.90° 29'16"E	500-4530	33
Muladi-2 (onshore dry hole)	Lat.22° 55'20"N Long.90° 28'29"E	4491-4550	2
Begumganj-1 (onshore gas disc.)	Lat.22° 59'17"N Long.91° 10'03"E	1203-3656	15
Begumganj-2 (onshore dry hole)	Lat.23° 01'00"N Long.91° 08'50"E	3028-3567	6
Titasi-2 (onshore gas well)	Lat.24° 00'20"N Long.91° 07'14"E	2646-2978	22
BODC-2 (offshore dry hole)	Lat.20° 35'43"N Long.91° 47'55"E	3110-4420	12

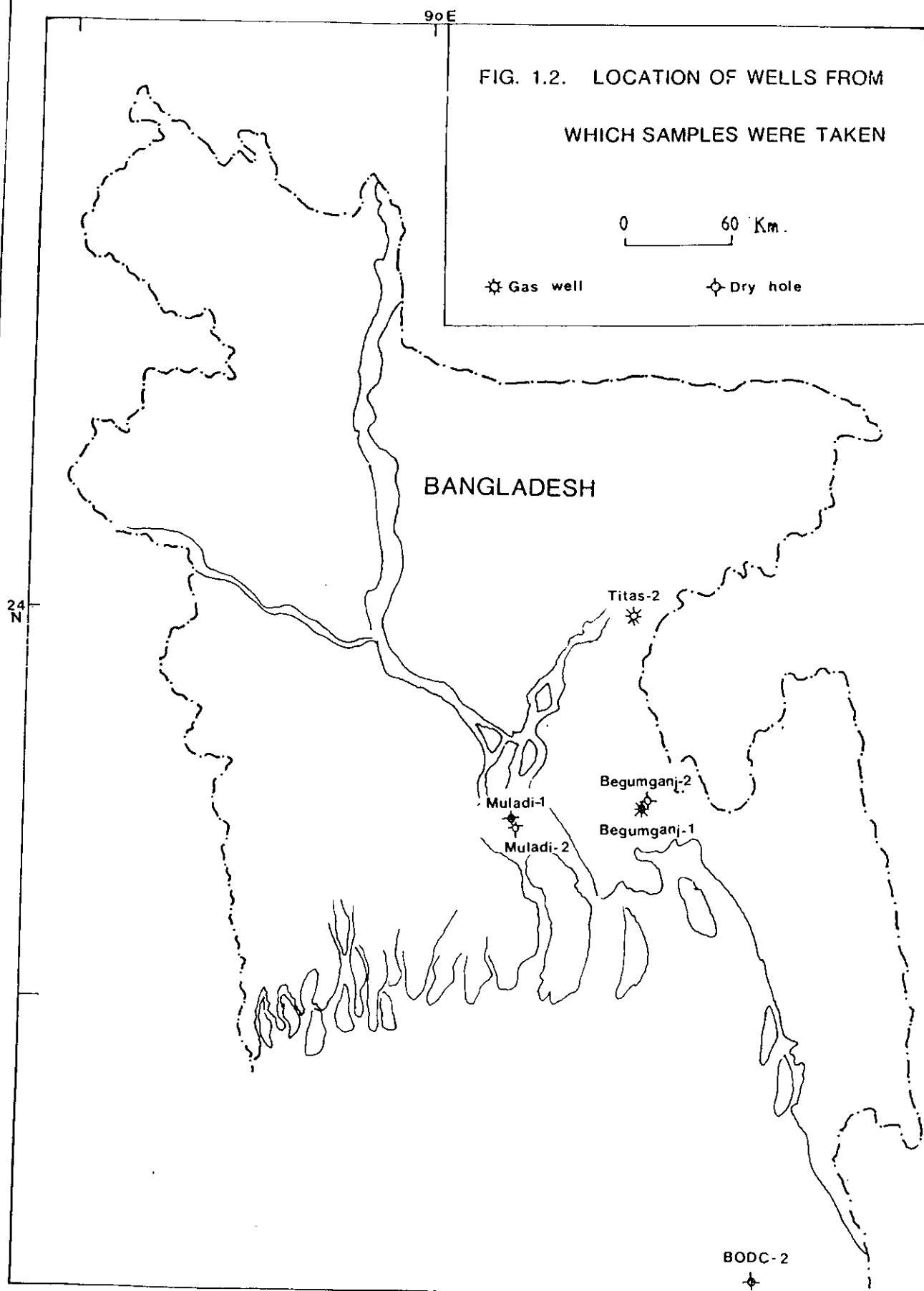


FIG. 1.3 SAMPLE LOCATIONS IN THE WELLS.

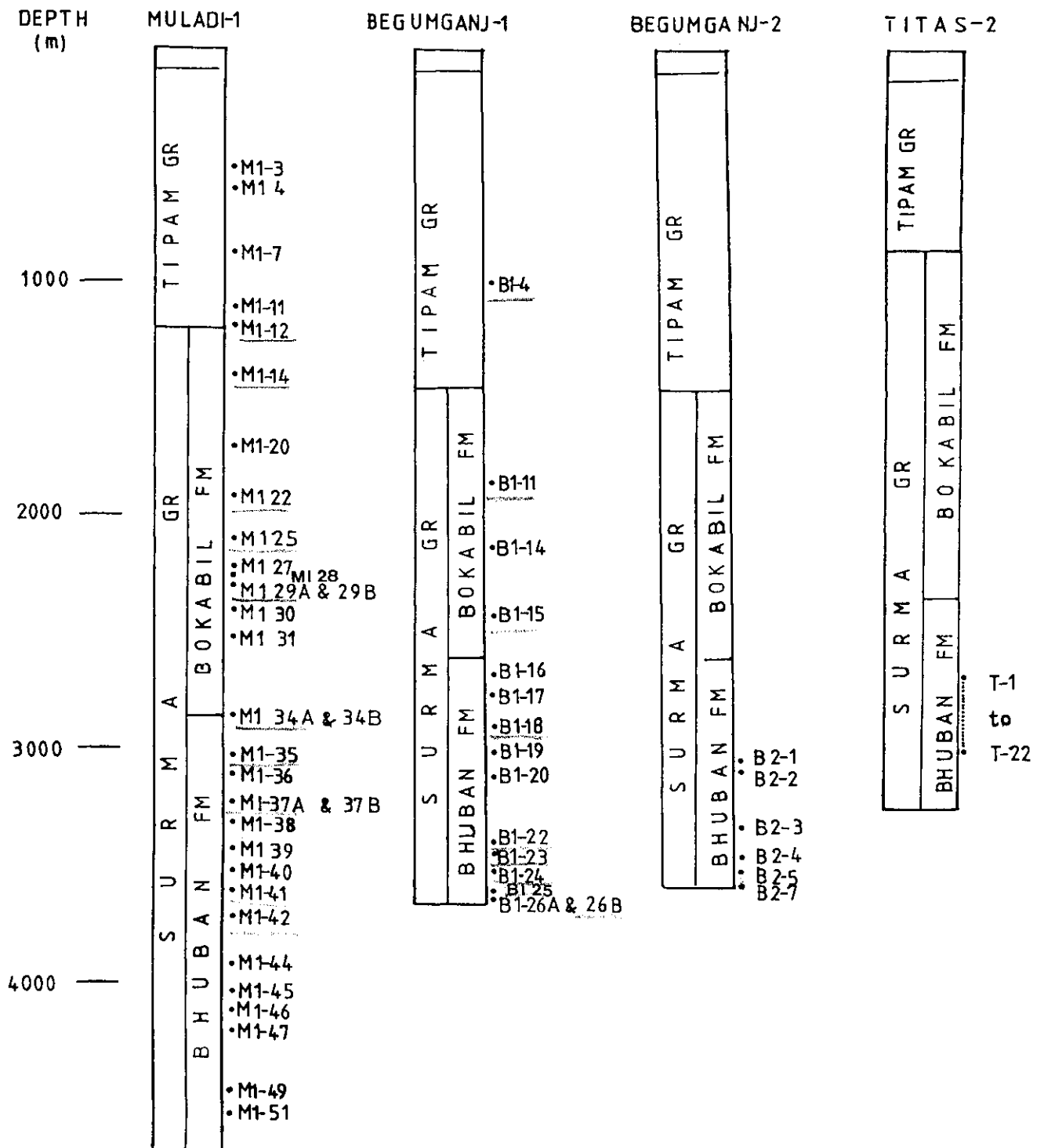
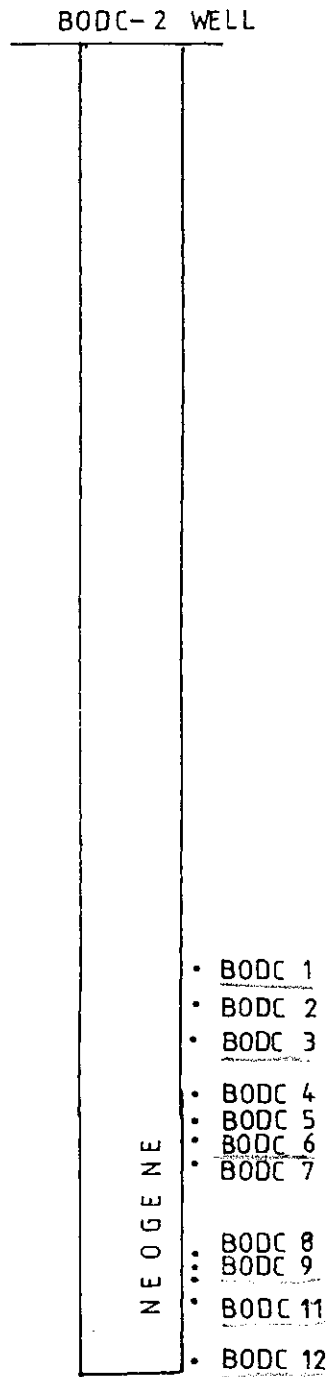


FIG 1.3. (Continued.)



and the Surma Groups of Neogene age (Fig. 1.3). No equivalent stratigraphy has been established in the offshore well and the samples can only be referred to as "Neogene". Study of the Surma Group has been emphasised because of the greater availability of samples from that unit.

The samples were collected as a result of an agreement with Petrobangla, the National Oil Company of Bangladesh, according to which the author went to Dhaka (formerly spelled Dacca) at the beginning of this research project (December 1979). Selection of the wells from which samples could be collected, depended on the material that the company would agree to release. The sampling was made with the aim of covering as much of the depth range as possible in a particular well. However, the distribution of samples was dependent on the depths of the cores/rock cuttings, as well as their availability for the present study. For this reason, for example, only a very small depth interval could be sampled from the Titas-2 well (Fig. 1.3). A detailed description of each of the samples has been included in the appendix. No wireline logs were available for this study and lithologic logs of the wells were made available for consultation only in the Dhaka office of the oil company providing the samples.

The author also made a field trip with a team of geologists from Petrobangla to the northeastern Sylhet district of Bangladesh, in order to collect additional samples of Neogene rocks, exposed along the anticlines of the folded belt. But their use was later abandoned because the intense weathering has made the rocks unsuitable for a study of diagenesis. The author also made a short trip to the headquarters of the Oil and Natural Gas Commission of India at Dehra Dun, India, to explore the possibility of obtaining information on relevant research works in the Indian part of the Bengal Basin.

CHAPTER 2

GENERAL GEOLOGY OF BENGAL BASIN

2.1 INTRODUCTION

The Bengal Basin is situated at the northeastern part of the Indian subcontinent with most of it occupied by Bangladesh, whilst the rest of it lies in the Indian state of West Bengal to the west and Tripura and Assam to the east (Fig. 2.1 and 2.2). The Basin is bordered to the west by the Peninsular Shield of India consisting of an Archaean basement complex with small areas of Gondwana sediments in intracratonic basins plus a few exposures of Upper Mesozoic volcanics (Fig. 2.2). To the north, the Basin is bordered by the Shillong Massif, apparently a northeasterly continuation of the Archaean basement of the Peninsular Shield through the so-called Garo-Rajmahal Gap. The eastern boundary is marked by the Naga-Lushai orogenic belt which consists chiefly of a zone of highly folded Tertiary and possibly Cretaceous deposits, bordered to the northwest by a complex zone of thrust faults. Although the southern boundary has not been delineated, the basin is believed to be open to the Bay of Bengal (Sengupta 1966). The total thickness of sediments in the Bengal Basin exceeds 16,000 m with the Neogene exceeding 5,000 m around the center of the basin (Paul and Lian 1975). The Basin represents a unique combination of a labile environment in a geosynclinal trough under compression with the concurrent development of a megadelta parallel to the north south axis of the trough (Brunschweiler 1980).

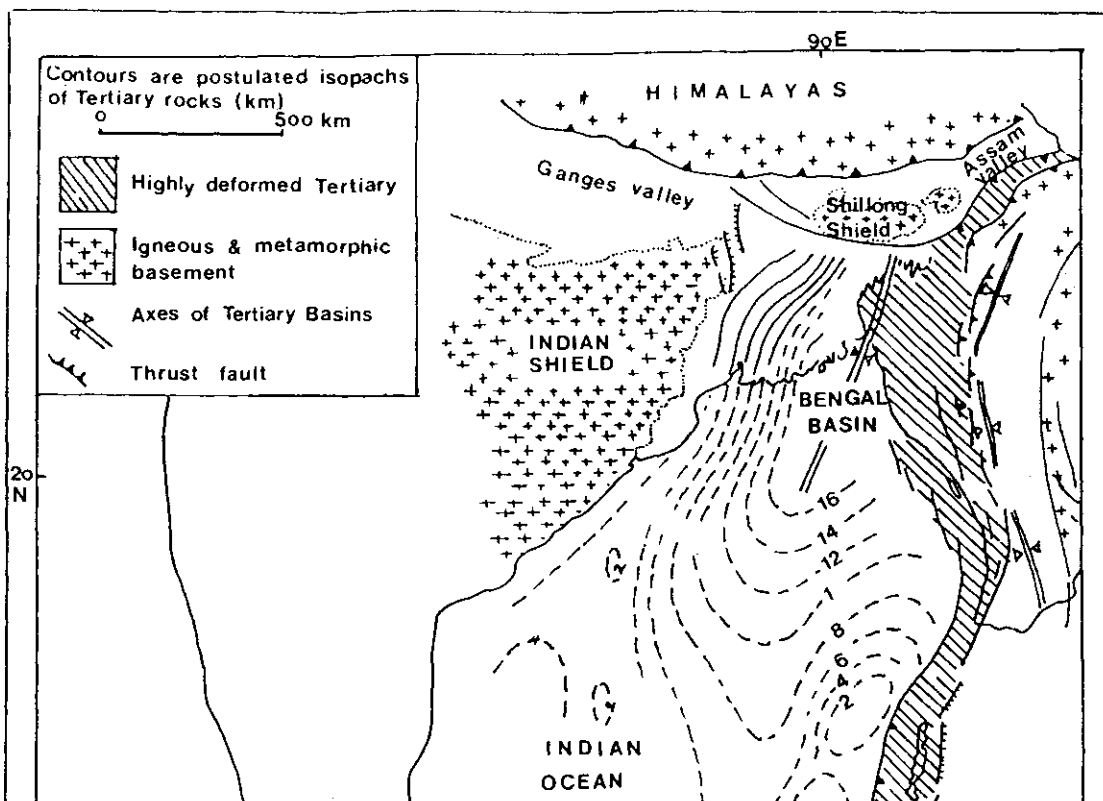


FIG 2.1 LOCATION OF THE BENGAL BASIN

(After Paul and Lian '75)

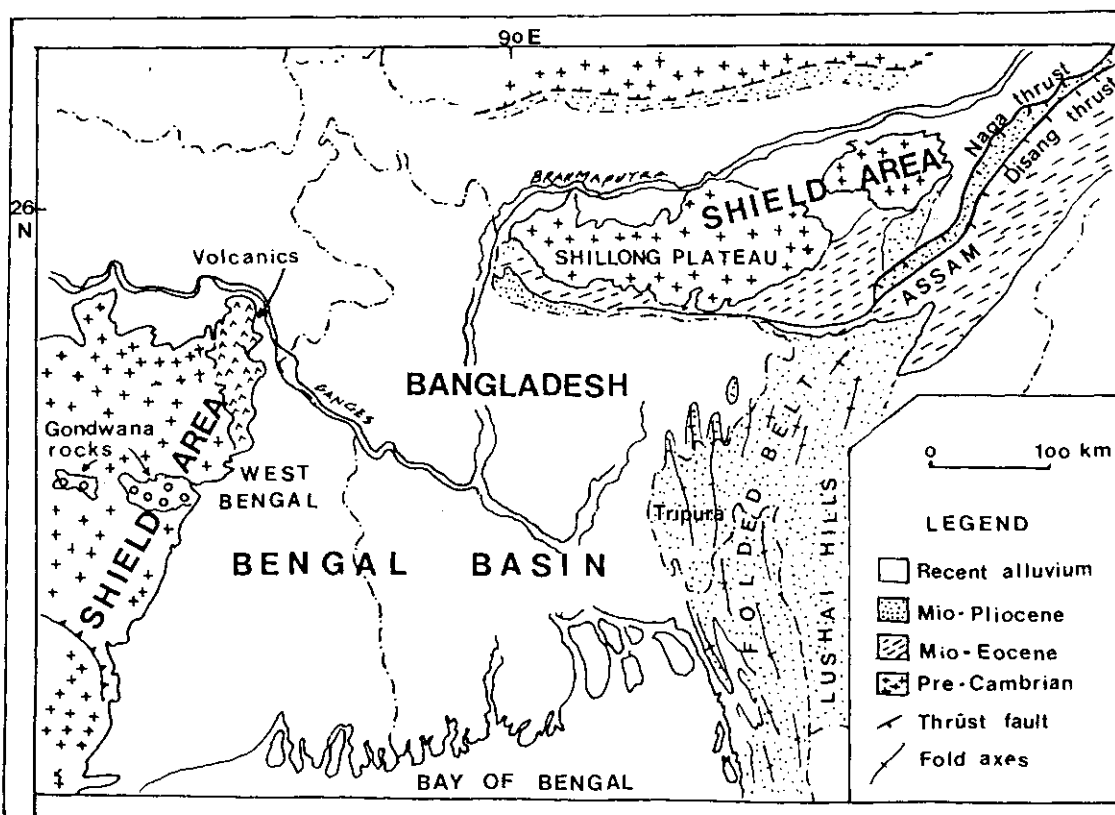


FIG 2.2 BENGAL BASIN AND ADJACENT AREAS

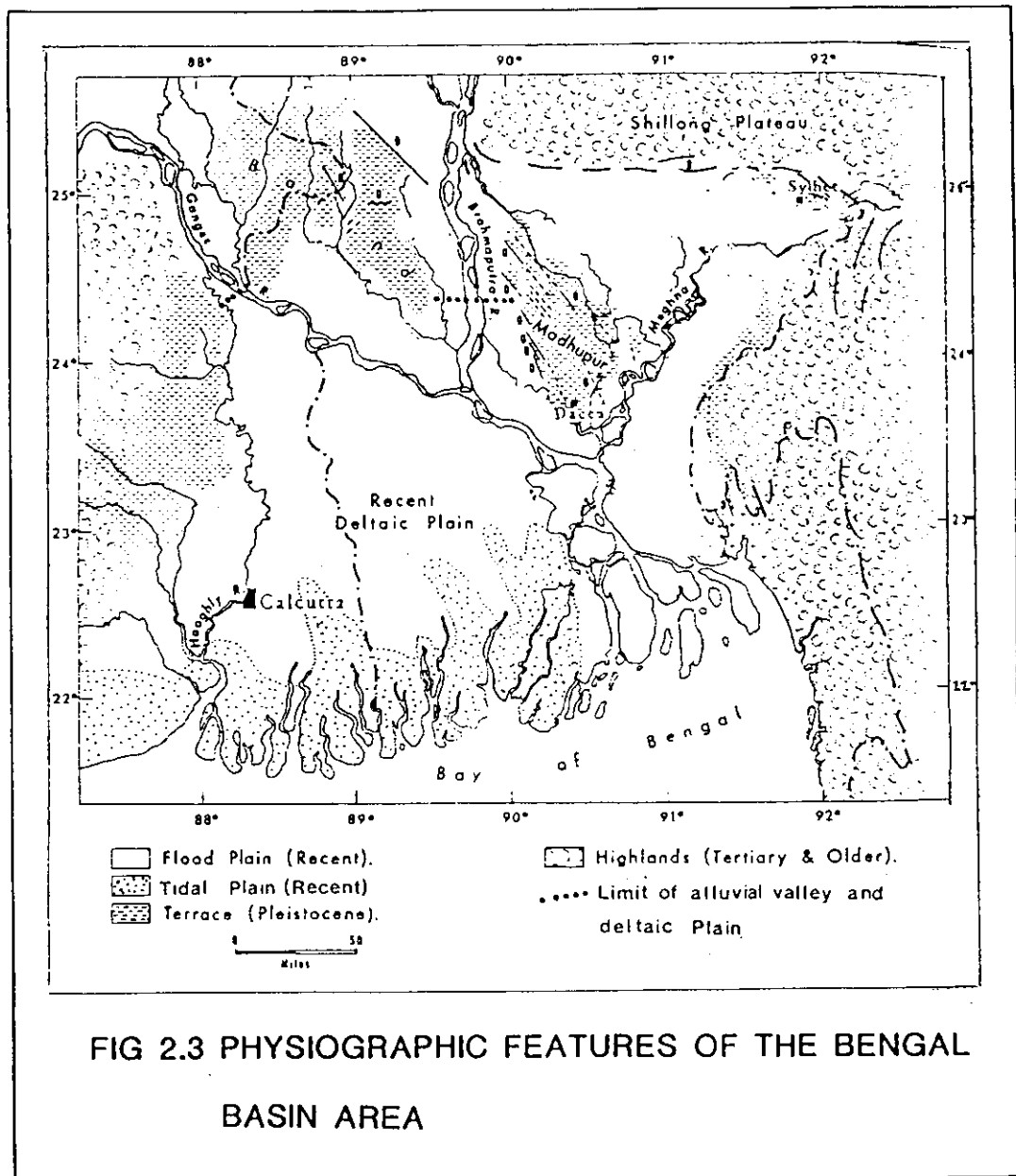
(After Sengupta '66)

2.2 PHYSIOGRAPHIC FEATURES

Except for the Tertiary hill ranges in the eastern part of the basin and a few Pleistocene terraces, almost the whole of the Bengal Basin is occupied by the recent alluvial and deltaic flood plain of the Ganges - Brahmaputra river system (Fig. 2.3). The Ganges and the Brahmaputra rivers, draining the south and the north side of the Himalaya respectively, together have the highest suspended sediment load carried to the sea of all modern river systems (Curry and Moore 1971). This river system together with its numerous tributaries and distributaries has built up the largest delta complex in the world - the Ganges - Brahmaputra delta (Coleman 1969). Part of the sedimentary load arrived to the sea by the Ganges - Brahmaputra river system passed via the "Swath of no ground" submarine canyon into the deeper part of the Bay of Bengal where it forms a vast deep sea fan complex. Both the deltaic platform and the deep sea fan complex represent the largest such depositional features of their kind today (Graham et al, 1975).

2.2.1 Recent fluvial and deltaic plain (Fig. 2.3):

The combined recent alluvial and deltaic flood plain of the Ganges-Brahmaputra river system covers more than 129,500 sq. km and is the largest surface physiographic unit in the Bengal Basin. The braided alluvial valleys of the Ganges and the Brahmaputra, flanked by Recent flood plains on either side, merge southward with the complex deltaic plain with numerous abandoned distributary channels and stream scars. Numerous low, swampy areas occur in the topographic depressions of the deltaic flood plain area. A high tidal range in the Bay of Bengal (10 ft as compared to 2 ft in Mississippi delta) results in strong tidal currents and creates a network of deeply scoured interconnecting channels and a broad tidal plain



in the coastal area (Morgan 1970). The tidal plain is characterised by an extensive dense mangrove forest cover. The Ganges - Brahmaputra deltaic region is a continuously rapidly subsiding area and many instances of rapid subsidence in the Recent deltaic plain deposits have been recorded in the literature (Morgan and McIntire 1959).

2.2.2 Pleistocene terraces (Fig. 2.3):

A few Pleistocene terraces, readily distinguishable from the adjacent Recent flood plain by their higher elevation (6 m to maximum of 30 m above sea level) are present. The sediments composing the terraces are believed to have been deposited as flood plain deposits of earlier Ganges - Brahmaputra river systems and were later tectonically uplifted (Morgan and McIntire 1959).

2.2.3 The Eastern Tertiary Hill range (Fig. 2.3):

To the east of the Bengal deltaic plain and extending south into the Chittagong and Chittagong Hill tract districts rises the hill ranges regarded as the frontal part of the folded belt of the Bengal Basin. The hill ranges are long, narrow, subparallel with a submeridional trend, are separated by low valleys; they consist of Upper Tertiary rocks which have been folded into a series of anticlines and synclines in an en echelon pattern. The elevation of the hill ranges generally increases towards the east with increasing intensity of the folding, the maximum elevation being almost 1,000 m in Bangladesh.

2.2.4 Bengal deep sea fan (Fig. 2.4):

The Bengal deep sea fan extends south about 3,000 km from the continental shelf break and has a width exceeding 1,000 km ; the thickness of sediments over most of the fan exceeds 3000 m with a maximum of over 15,000 m beneath the

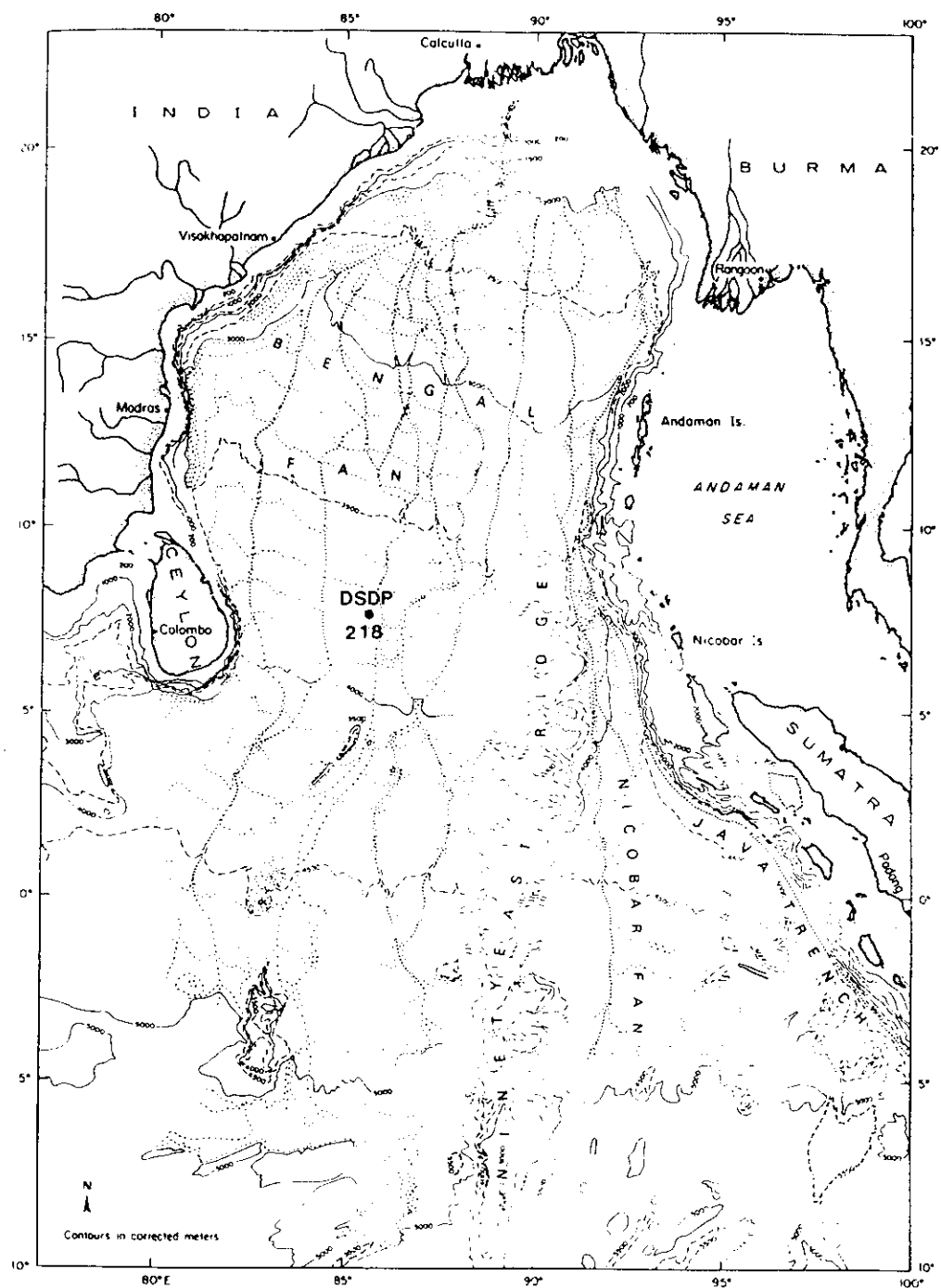


Fig. 2.4 Present topography and channel system of Bengal deep sea fan in the Bay of Bengal. Fan valleys are indicated by dotted lines. (After Curray and Moore 1971).

delta-fan interface (Curray and Moore 1974; Graham et al. 1975). The surface of the fan is controlled by a system of turbidity current submarine channels spreading mostly from the "Swatch of No Ground" submarine canyon at the head of Bay of Bengal. The surface morphology, geological structures, stratigraphy and geological history of the Bengal deep sea fan complex have been studied since 1968 and are well documented in the literature (Curray and Moore 1971; Curray and Moore 1974; Graham et al. 1975; Moore et al. 1974; Curray et al. 1982).

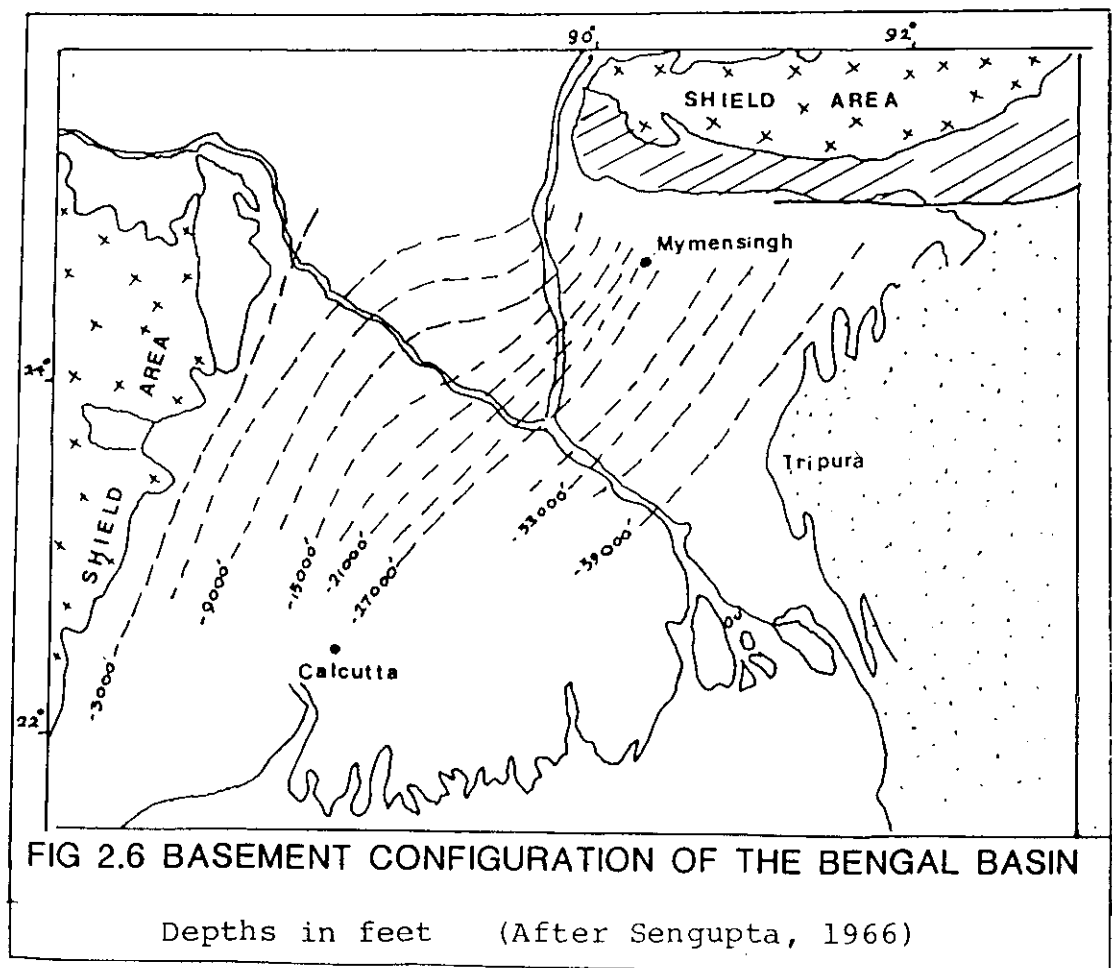
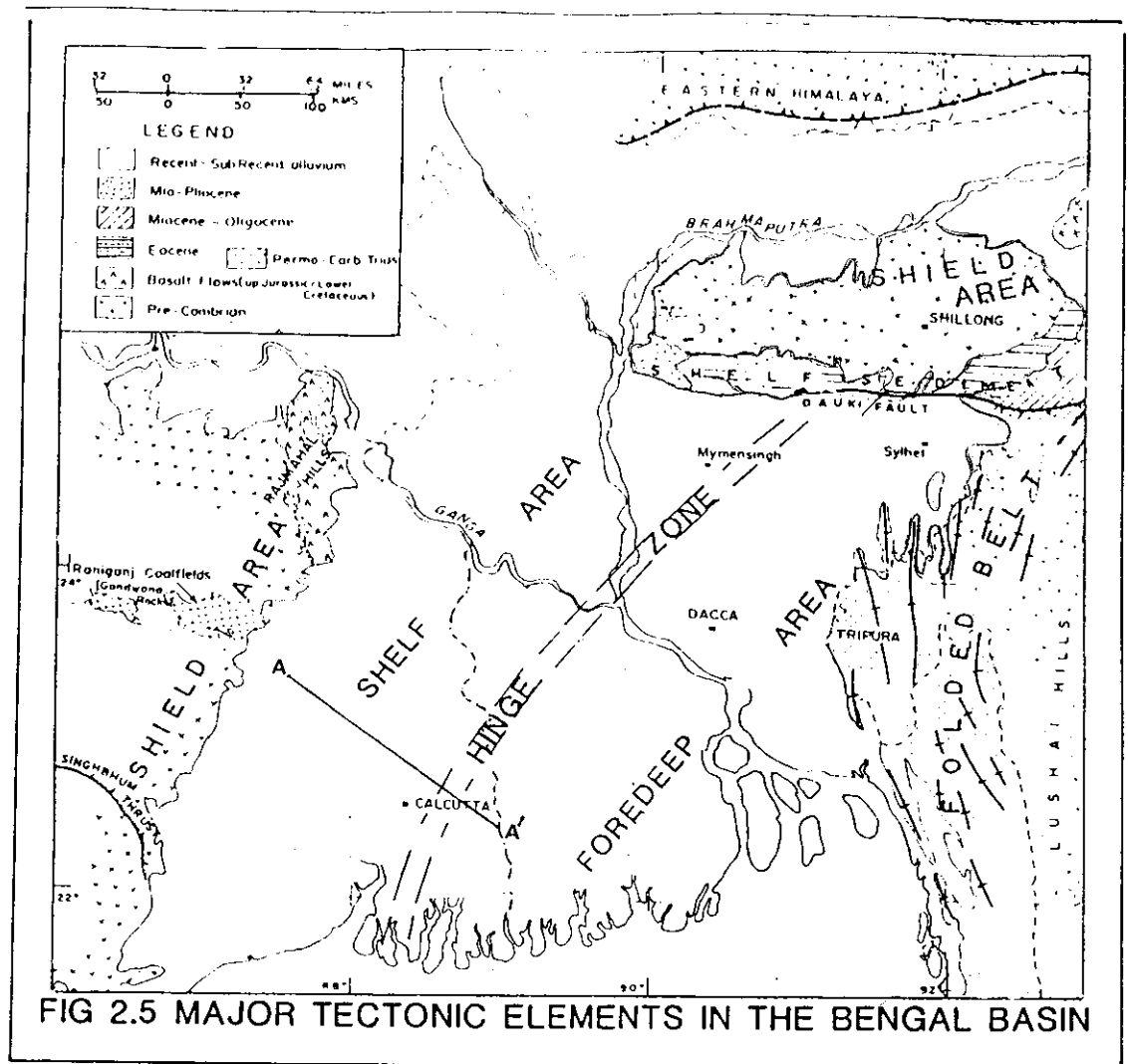
2.3 TECTONIC FRAMEWORK

2.3.1 INTRODUCTION

Tectonically, the Bengal Basin can be broadly divided into two parts (Fig. 2.5). These are (1) The stable shelf in the north and northwest, characterised by comparative tectonic stability and reduced sedimentary sections and (2) The Bengal foredeep in the south and southeast, characterised by greater tectonic mobility and a very thick sedimentary infill. These two tectonic units are separated by a "hinge zone" running in a northeast - southwest direction through the middle of the basin. The structural and tectonic configuration of the Bengal Basin area have been described by Bakhtine (1966), Sengupta (1966), Raju (1968), and Alam (1972).

2.3.2 STABLE SHELF

Bordered to the west by the Pre-Cambrian Indian Shield, the stable shelf occupies most of West Bengal (India) and the northwestern part of Bangladesh and is characterised by a shallow basement surface which dips gently southeast under the increasingly thick cover of sediments. Sengupta (1966) recorded an uniform increase of sediment thickness



from 915 m (3,000 ft) at the northwestern end to about 8,235 m (27,000 ft) at the southeastern edge of the shelf where it meets the hinge zone (Fig. 2.6). Except for a few normal faults the shelf area is structurally relatively undisturbed. A very gentle dip, about 1 degree in the west, increasing to about 3-5 degrees towards the southeast, is indicated by the 'D' reflector, i.e. the top of Sylhet limestone unit of Eocene age which forms a seismic marker horizon throughout the shelf area. The southeastern boundary of the stable shelf is marked at the hinge zone by a sudden flexure of the 'D' reflector (Fig. 2.7). In the Bangladesh part of the Bengal Basin, the area northwest of the hinge zone is regarded as a Pre-Cambrian platform and is divided into three parts (Bakhtine 1966). A shallowest central part, the Rangpur saddle is flanked to the northwest and southeast by the Ruhea slope and Bogra slope respectively (Fig. 2.8). The Bogra slope lying immediately northwest of hinge zone represents the stable shelf area in Bangladesh and is essentially a northeastern continuation of the West Bengal shelf. The Rangpur saddle is, in fact, a shallowly buried basement ridge which is contiguous with the Shillong Shield to the east and the Indian Peninsular Shield to the west; the sedimentary cover of the basement is only about 122 - 150 m thick in places. The Ruhea slope represents the area northwest of the Rangpur Saddle and is characterized by gentle NW plunge of the basement towards the Sub-Himalayan foredeep.

2.3.3 HINGE ZONE

The southeastern limit of Bengal shelf is marked by a zone of flexure of the 'D' reflector horizon (top of Eocene Sylhet Limestone unit, Fig. 2.7). This trends approximately N 30° E and coincide with the NE-SW trending Calcutta-Myumensigh gravity high - an important tectonic lineament in the Bengal Basin. The flexure in the sediments is believed to be the shallower expression of a

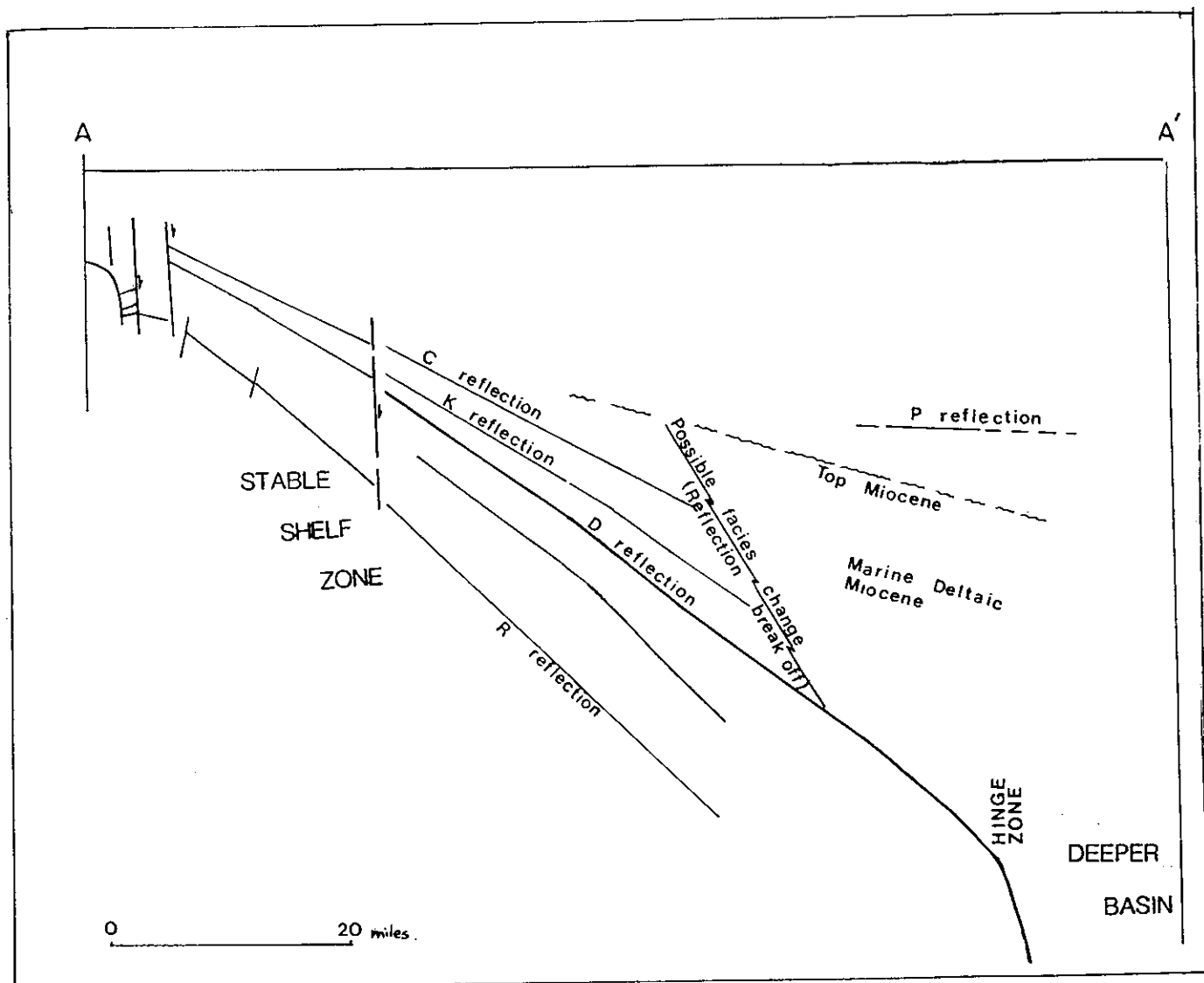


FIG. 2.7 GEOPHYSICAL CROSS SECTION SHOWING THE 'D' REFLECTOR (Sylhet Lst fm) ,BENGAL BASIN. For line of cross section (A-A') see fig.2.5

(Based on Sengupta 1966)

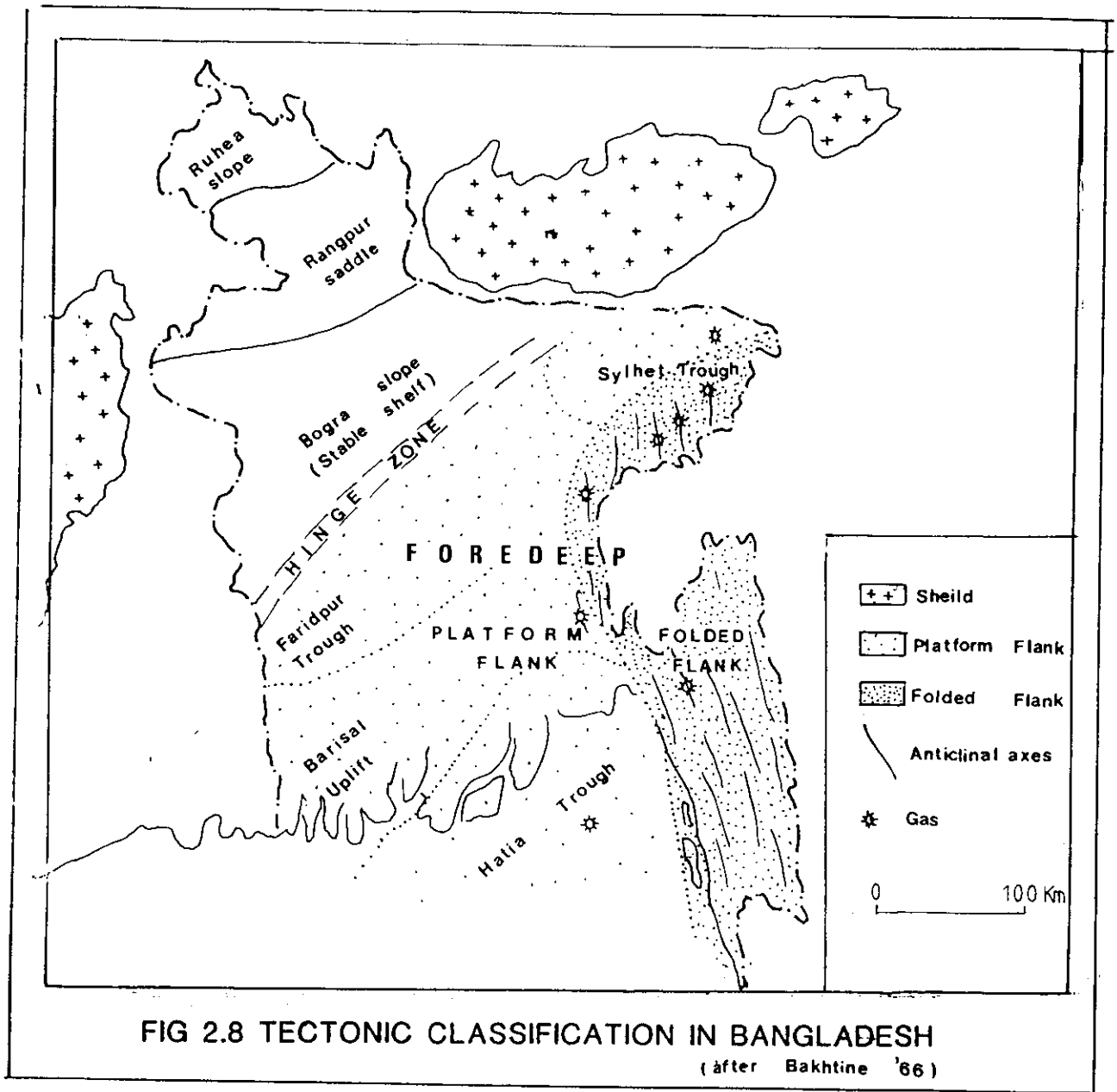
deep zone of basement adjustment, possibly the result of a deep seated basement fault, between the stable shelf to the northwest and the geosynclinal trough to the southeast (Raju 1968). In the shallower subsurface levels of this hinge zone, Sengupta (1966) reported an increased dip in the D reflector, which is 14 times the dip in the shelf area in West Bengal. Similarly a 30 km wide zone of increased basinward dip is reported in the 'D' seismic reflector in Bangladesh (Brunschweiler and Khan 1978). Regionally the hinge zone is seen to continue from Calcutta in a northeasterly direction to Mymensingh and according to Evans (1964) extends further northeast across the Dauki fault to Naga thrust in Assam (Fig. 2.5).

2.3.4 BENGAL FOREDEEP

The area to the east and southeast of the hinge zone is the deeper part of the basin and is referred to as the foredeep or the geosynclinal part of Bengal Basin. The area is characterised by a greater rate of subsidence resulting in greater thicknesses of sediment (in excess of 8,235 m). The maximum sediment thickness in the foredeep area is more than 16,000 m beneath the continental shelf of Bangladesh (Moore et al. 1974). The eastern boundary of the Bengal foredeep with the Naga-Lushai orogenic belt is beyond the eastern border of Bangladesh. A thick sequence of Tertiary sandy-argillaceous sediments constitutes the major part of the sedimentary infill in the foredeep area. Structurally the Bengal foredeep is divided into two parts (Fig. 2.8) as (a) Western platform flank and (b) Eastern folded flank (Bakhtine 1966).

(a) Platform flank of the foredeep

The Western part of the Bengal foredeep constitutes the platform flank and is characterized by simple structures of platform types. It differs sharply from the eastern



folded flank of the foredeep. The difference is reflected in gravity anomaly values: small amplitude isometric anomalies characterize the platform flank, whereas large amplitude and linear-elongated anomalies dominate the folded flank. The structure on the platform flank are broad, gentle and simple, and are mostly concealed beneath the flat surface plains. Geophysical studies in the platform flank indicate a basement topography with several highs and lows; it can accordingly be subdivided into several subunits i.e., the Faridpur trough, the Barisal high, the Sylhet trough (Surma Basin), the Hatiya trough etc. (Bakhtine 1966) (Fig. 2.8). The troughs correspond to the gravity minima and have comparatively greater thickness of sediment. The maximum sediment infill of Bengal Basin is recorded in the Hatiya trough and is mainly a sand-shale sequence of Tertiary age.

(b) Folded flank of the foredeep:

The eastern part of the foredeep comprises the folded belt of the Basin. Here the Upper Tertiary sandy-argillaceous sediments have been folded into a series of long submeridional anticlines and synclines represented in the surface topography by elongated hill ranges and intervening valleys respectively. Most of the hydrocarbon discoveries in Bengal Basin are situated in this belt. The folded structures are characterized by en-echelon orientations with an increasing degree of intensity and complexity toward the east. Accordingly, the folded flank is divided from west to east into three parallel N-S trending zones as follows: (a) The Western Zone - characterized by simple box-like or similar shaped anticlines with steep flanks and gentle crests separated by gentle synclines. (b) The Middle Zone - with more compressed structures with ridgelike and asymmetric anticlines frequently associated with faults and separated by narrow synclines. (c) The Eastern zone - with highly

disturbed narrow anticlines with steep dipping flanks and mostly associated with thrust faults.

2.4 STRATIGRAPHY

2.4.1 INTRODUCTION

The Bengal Basin is characterised by the presence of one of the thickest sedimentary sequences in the world, exceeding 16,000 m in the deepest part of the foredeep or geosynclinal area. The isopach map (Fig. 2.1) shows that the greatest thickness, exceeding 10,000 m, trends northeast from northern end of Bay of Bengal towards the Assam valley. This trend is approximately parallel to the paleo-continental slope, represented by the hinge line of Evans (Curray et al. 1982). To the northwest of this hinge line the relatively thin sedimentary cover of the stable shelf is underlain by continental crust, whereas to the southeast, the huge sedimentary infill of the geosynclinal area is believed to be underlain by oceanic crust (Paul and Lian 1975; Curray et al. 1982) (Fig. 2.11). The hinge zone or paleocontinental slope presumably represents the transition between the continental crust to the northwest and the oceanic crust to the southeast.

Except for the eastern folded belt where the Upper Tertiary sediments are expressed in the long anticlinal hill ranges and the few exposures of Cretaceous - Tertiary rock along the narrow shelf fringing the Southern Shillong plateau, the whole of Bengal Basin is concealed beneath Quaternary alluvium (Fig. 2.9). As a result, the stratigraphic subdivision of the rock sequence are primarily based on the information gained from subsurface drilling. In the stable shelf area, sediments ranging in age from Permian to Recent are found above the

FIG.2.9 GEOLOGICAL SKETCH MAP OF BANGLADESH

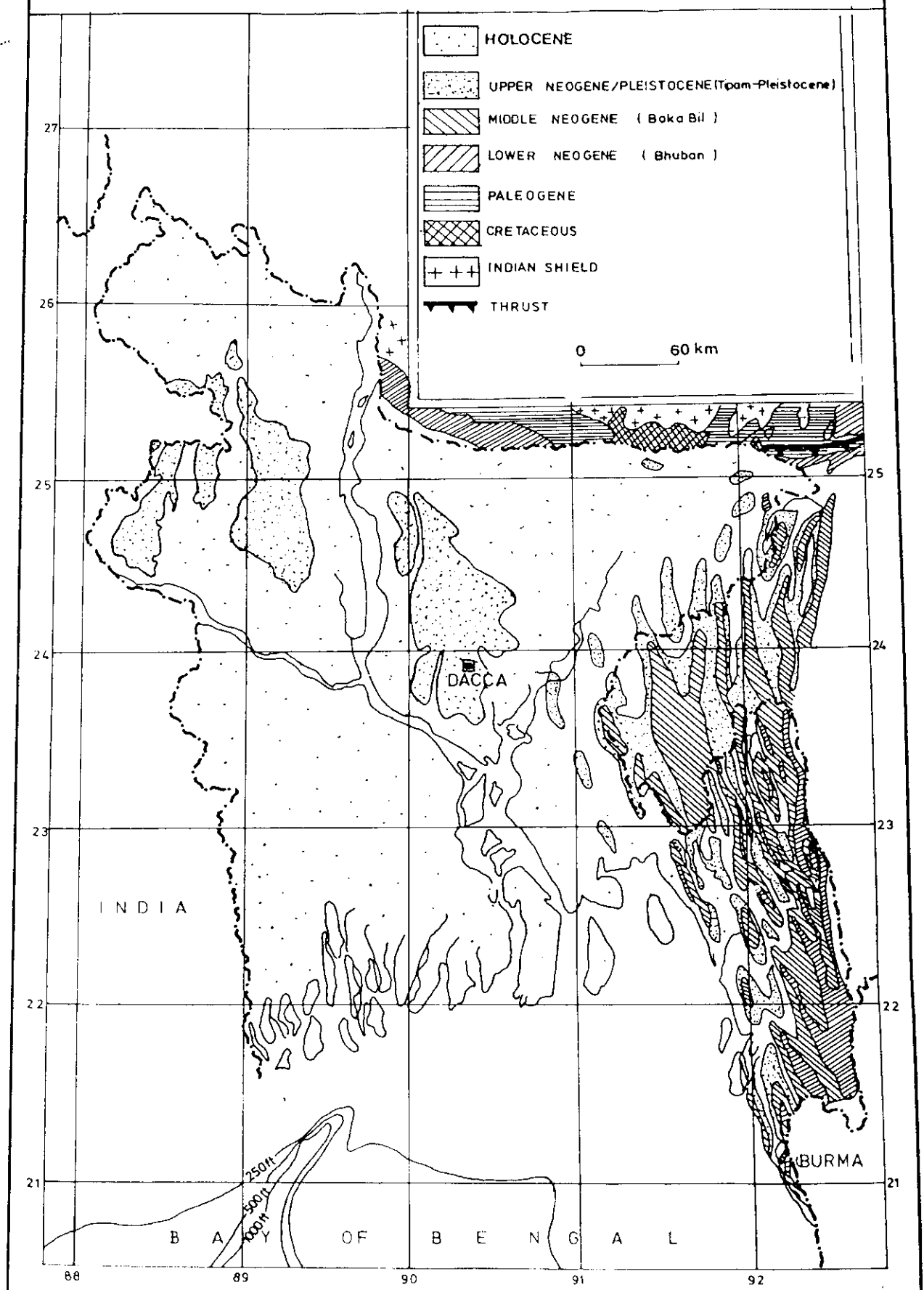


TABLE 2.1 : STRATIGRAPHIC SUCCESSION IN BANGLADESH.

	STABLE SHELF	GEOSYNCLINE	LITHOLOGY	ENV.OF DEPOSIT.
R E C E N T T O O L I G O C E N E	Bengal Alluvium		Sand,silt & grvl	Fluvial.
	Dihing group (200)		Pebbly sst, clay.	Fluvial
	UNCON			
	Tipam gr. (300)	T Dupitila sst fm	Crs. ferrg. sst. with thin pebl.bed	Fluvial
		i p		
		a Girujan clay fm	Mottl.clay with sub.sst.& silt	Lacust.
		m		
	Surma gr. (400)	g Tipam sst fm	Crs. ferrg.sst. sub.shale.	Fluvial
		r		
		(2500)		
E O C E N E	Surma gr. (400)	S		
		u Bokabil fm	Sst,sh & silt.	Deltaic
		r (1300)		
		m		
	Barail gr (160)	a Bhuban fm	Sst,sh,&silt.	Deltaic
		gr. (4000)		
		(5300)		
	UNCON			
	Barail gr (160)	Barail gr	Sh.sst,silt.	Deltaic
		(3350)	-marine	
E O C E N E	UNCON			
	Jaintia gr. (720)	Kopili fm	Sh.minor sst.	Marine
		Sylhet lst fm	Lst.minor sst.	Marine
		Tura sst fm	Sst,minor coal	Marine
	UNCON			
	Sibganj trapwash fm (230)		Crs.Sst,volc,clay.	Lacust.
	UNCON			
	Rajmahal trap fm (610)		Basalt,andesite & serpentinized sh.	Fiss. erupt.
P E R M I A N	UNCON			
	Gondwana gr. (1023)		Sst with sh & coal	Fluvial-
			Sst,with sh & coal	Lacust.
	UNCON			
	Archaeon Basement.		Gneiss & Schist.	
	UNCON			
	Archaeon Basement.			
	UNCON			

UNCON = Unconformity. (THICKNESS) in meter.

Pre-Cambrian basement. However, in the Bengal foredeep, deep drilling has reached no sediments older than latest Oligocene: in the central part of the foredeep area the deepest well drilled so far (Muladi-1 4732 m) bottomed in the Upper Miocene.

Lack of co-ordination between the geological institutions of Bangladesh and West Bengal (India) has led to there being no unified stratigraphical classification for the whole of the Bengal Basin. The following descriptions are based on the nomenclature adopted in Bangladesh (Table 1).

2.4.2 PRE-TERTIARY STRATIGRAPHY

(a) Gondwana Group: (Permian):

Thick continental Gondwana sediments of Permian age were deposited in isolated grabens on the Archaean crystalline basement in the stable shelf area. They are the oldest sediments in Bangladesh, and consist of cross-bedded arkosic sandstone, conglomerates and carbonaceous shales which are locally coal bearing. These sediments were deposited essentially under terrestrial, fluvial to lacustrine environments in a humid climate (Khan 1980). Two formations are distinguished: a lower Kuchma formation mainly of coarse sandstone with thick coal seams and an Upper Paharpur formation which consists of false bedded feldspathic sandstone, conglomerate, carbonaceous shale and coal. Initially the Gondwana basins might have been larger and only the parts down thrown by subsequent faulting have been preserved.

(b) Rajmahal Trap: (Upper Jurassic - Lower Cretaceous):

A thick sequence of basic to intermediate lavas including basalt, dolerite and andesite accumulated unconformably on the Gondwana sediments or on the Archaean crystalline

basement. This is the Rajmahal Trap, named after the Rajmahal Hills of West Bengal (India) where 550 m of these volcanic rocks are exposed. On the Bangladesh shelf, the Rajmahal Trap is represented by 610 m of amygdaloidal basalt, andesite, serpentized shale and agglomerate in the subsurface (Khan 1980). The Rajmahal trap represents subaqueous fissure eruptions through faulted and fractured zones in the basement. (Banerji, 1981).

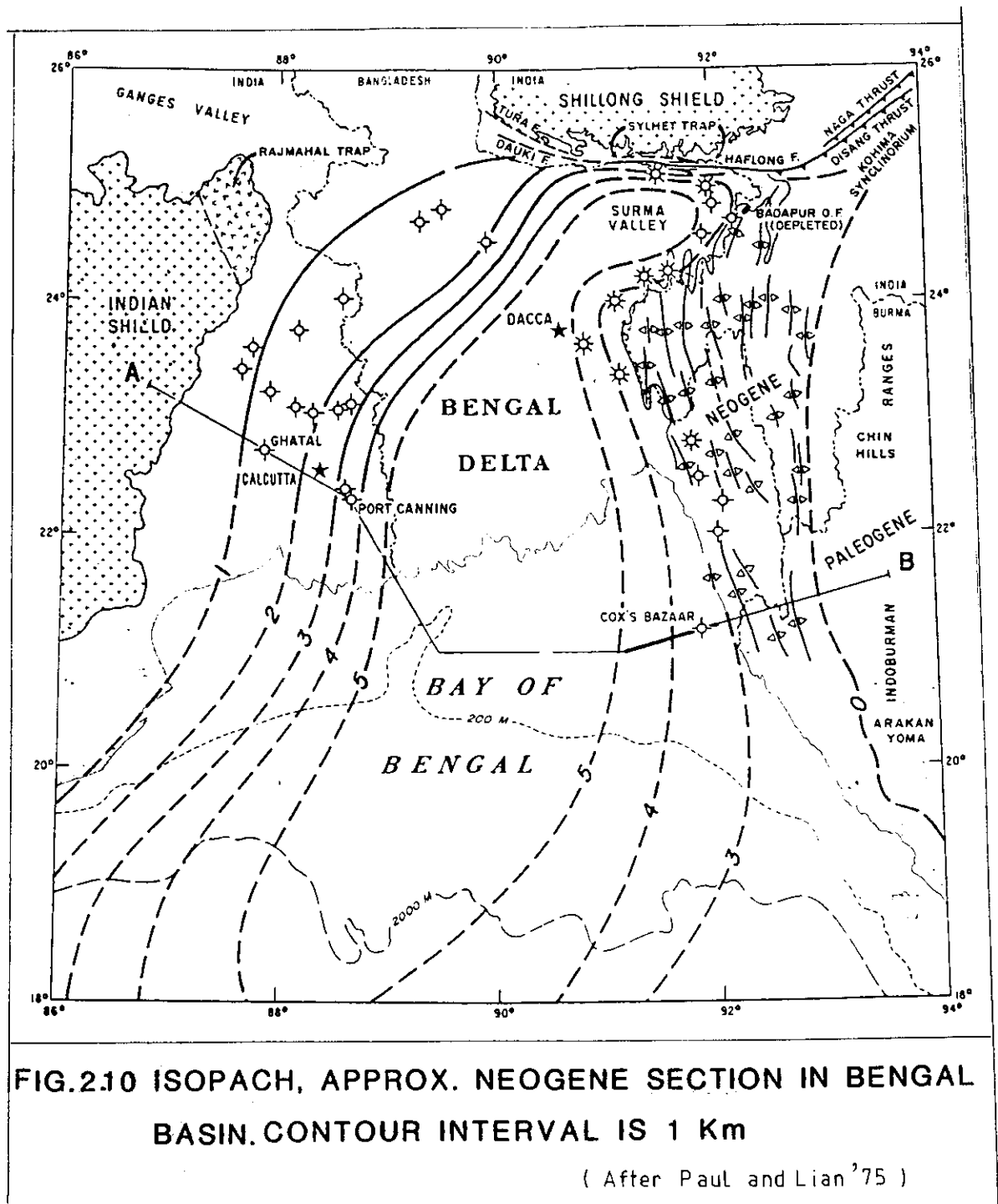
(c) Sibganj Trapwash Formation (Mid-Upper Cretaceous)

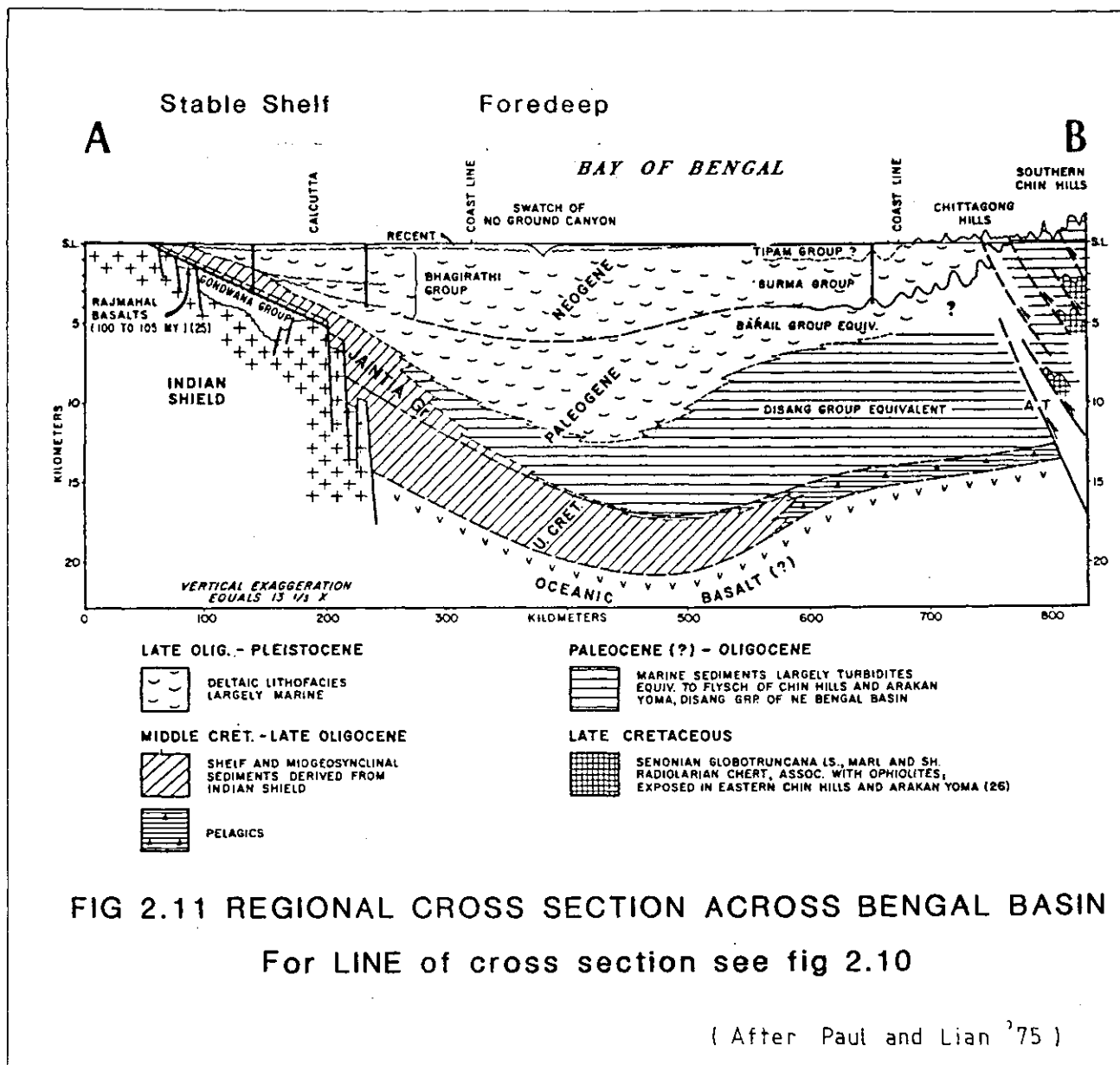
This formation, consisting of coarse, yellow brown sandstone, volcanics and white clay, unconformably overlies the erosional surface of the Rajmahal Trap. It represents trapwash sediments laid down sporadically under continental fresh water and lacustrine environment.

2.4.3 TERTIARY TO RECENT STRATIGRAPHY

The Tertiary deposits of the Bengal Basin probably exceed 16,000 m with the Neogene more than 5,000 m thick along the center of the Bengal foredeep (Paul and Lian 1975), (Fig. 2.10 and 2.11). Except for the Eocene shelf facies (Jaintia group) which represents the last phase of a totally marine regime in the Bengal Basin, the Tertiary deposits are characteristically devoid of fossils (excluding pollen and spores) which could be used for stratigraphic sub-division and recognition. The thick arenaceous and argillaceous sequence, represents a predominantly regressive marine deltaic episode.

Stratigraphically the huge post-Eocene sedimentary infill is a complex intermixing of deposits of various subenvironment in a prograding megadelta, where sedimentary facies characteristically lack horizontal continuity. Frequent facies changes are the rule rather than the exception. Also the sequence suffers from being a





repetition of a limited number of rock types - shale, sandstone and siltstone with few conglomerates, totally lacking any marker horizon. Stratigraphic sub-division of such a sequence is unlikely to be simple, but in Bangladesh, the Tertiary stratigraphy has suffered from over simplification and too much generalization (Khan 1980). Traditionally, the sub-division and nomenclature used here are a straight transcript of the Assam Tertiary that was established by Evans (1932) and later modified by Mathur and Evans (1964). The classification in Assam is based purely on lithology because of the lack of fossils; and has proved to be difficult to extend laterally due to lack of lithological continuity, and the monotonous repetition of limited rock types that is sandstone, shale and siltstone (Mathur and Evans 1964). The application of this stratigraphy to the northeastern part of Bangladesh (Sylhet trough and surroundings) is probably reasonable but its extension into the coastal region of the Bengal Basin is unreliable, although widely used (Khan 1980). Also the diachronous nature of the same lithologies between Assam and the Bangladesh was not recognized in the traditional stratigraphy and led to further confusion (Fig. 2.12A). In the West Bengal (India) part of Bengal Basin, a separate stratigraphic classification of the Tertiary has been established (Biswas 1959, 1963).

Brunschweiller (1980) reinterpreted the traditional stratigraphy in Bangladesh and established the new chrono-lithostratigraphic subdivisions correlated from Assam to the Bengal Basin (Fig. 2.12B). A sequence of parallel diachronous lithostratigraphic units started in the Oligo-Miocene in Assam and extend up into the Quaternary in the Bay of Bengal. This is a regressive sequence resulting from a delta prograding from northeast to southwest. It consists of a sand-shale association (Barail gr. and Surma gr.) of deltaic-coastal marine

Fig. 2.12A Traditional interpretation of Assam-Bengal Basin Stratigraphy (Post Eocene). (After Brunschweiler 1980)

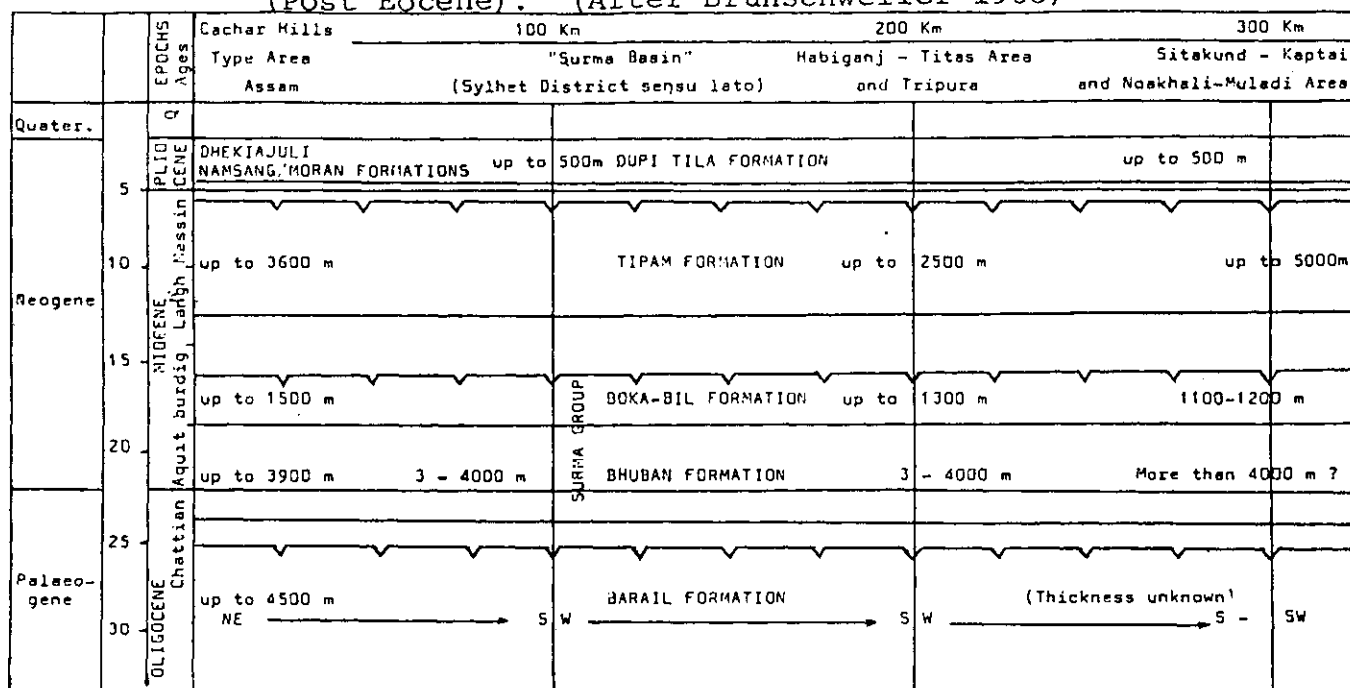
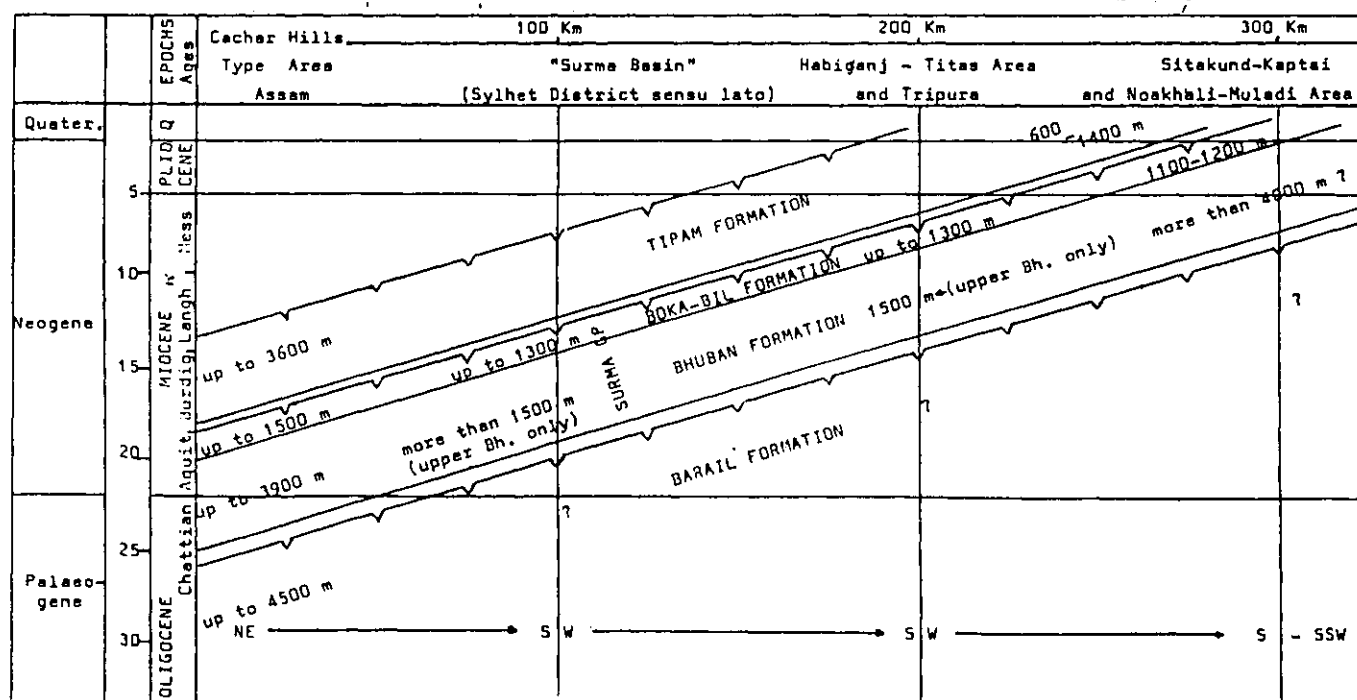
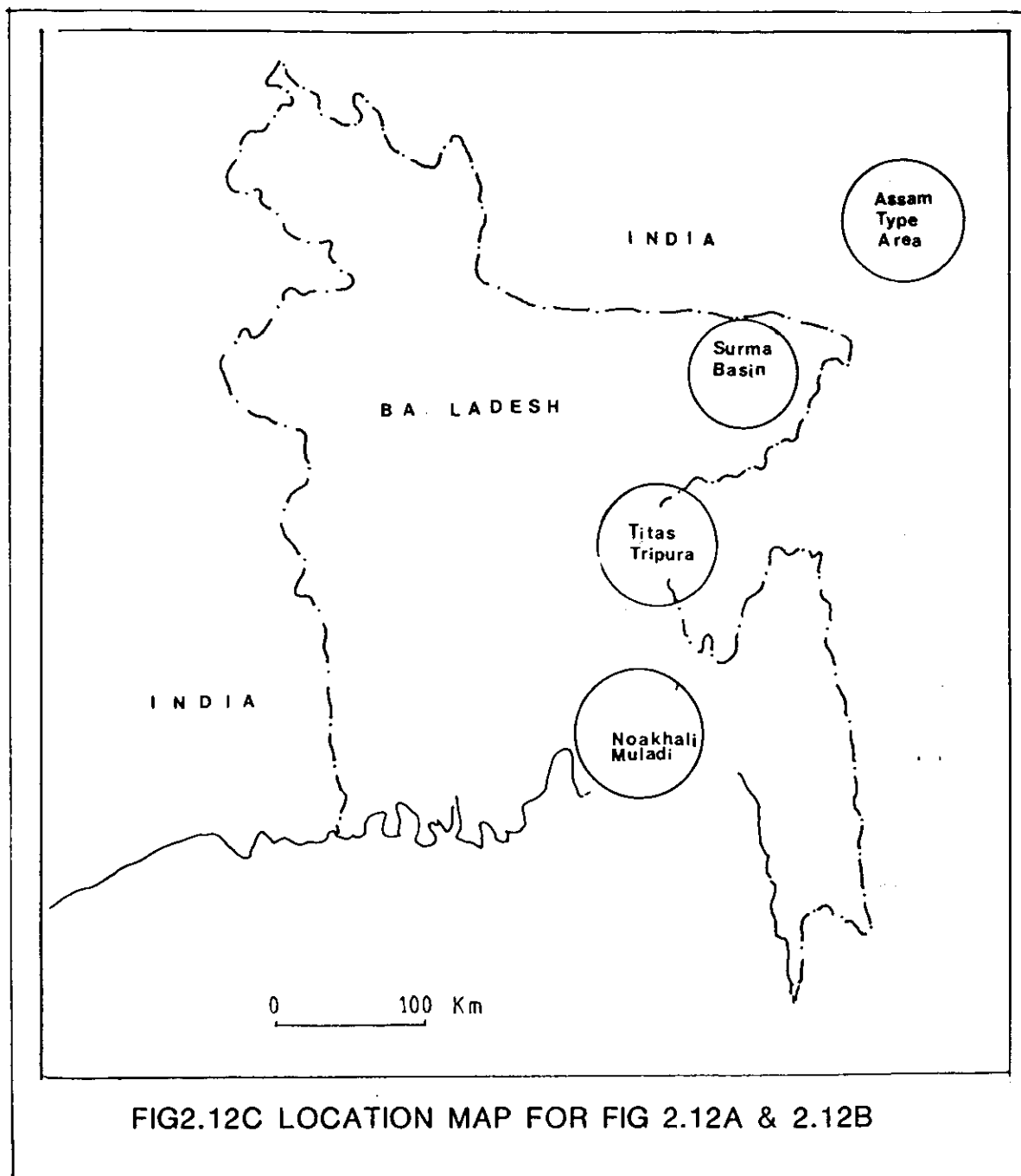


Fig. 2.12B Reinterpretation of traditional stratigraphy of Assam-Bengal Basin (Post Eocene). (After Brunschweiler 1980).





environment at the bottom, passing upward to a predominantly sandy facies (Tipam gr.) of terrestrial environment at the top.

2.4.3.1. PALEOGENE.

(a) Jaintia Group/Disang Group (Eocene)

During the Eocene two distinct facies were deposited on the shelf and in the geosynclinal part of Bengal Basin. On the foreland shelf, represented by Jaintia group, there is a sandstone - limestone - shale sequence, which is encountered in the subsurface throughout the shelf area (Fig.2.11). Although no rocks of equivalent age have been reached in the subsurface of Bengal foredeep, seismic evidence suggests that this sequence changes across the hinge zone into a predominantly shaly facies (Sengupta 1966) referred to as the Disang Group equivalent (Fig. 2.11). The Jaintia group can be divided from bottom upward into three formations: Tura sandstone formation, the Sylhet limestone formation and the Kopilli formation.

The Tura sandstone formation is of early Eocene age. This formation consists predominantly of sandstone with subordinate shale and marl. The sandstones are relatively clean, mature, well sorted and medium to fine grained. The formation was deposited in shallow marine conditions.

The Sylhet limestone formation is of Middle Eocene age. It consists essentially of nummulitic limestone with subordinate shale and thin layers of sandstone. The limestone is highly fossiliferous, grey, massive and hard. The shales are calcareous, moderately hard, and the sandstones are fine to very fine grained, glauconitic, calcareous arenites. The Sylhet limestone formation forms a marker horizon in the subsurface through the Bengal shelf and its top is identified with the 'D' reflector

horizon in the seismic section (Fig. 2.7). The formation was deposited in a shallow, open marine warm water environment.

The Kopilli formation is of late Eocene age. The formation is composed of predominantly of shale with subordinate sandstone and a few marl streaks. The shales are grey to dark grey, fossiliferous, slightly calcareous and moderately hard. The sandstones are calcareous, fine to very fine grained arenites. The formation was deposited in a marine environment and the top marks the end of the wholly marine phase of sedimentation on the Bengal shelf (Banerji 1981).

(b) Barail group: (Oligocene)

The name of the group was taken from the Barail Range in Assam, skirting the southeastern margin of Shillong plateau and formed predominantly of arenaceous rocks to which Evans (1932) applied the term "Barail Series". In the type locality in Assam the group is divided into upper and lower predominantly sandy units separated by a mainly shaly unit. The depositional environment is largely coastal to deltaic.

In the geosynclinal part in Bangladesh the Barail has been reached in the subsurface in only one of the wells in the northeastern district of Sylhet, southwest of the type locality. Only the top of the formation was drilled, but it composed of more shaly facies with a marine fauna (Holtrop and Keizer 1970). In the Sylhet trough (Surma basin) the thickness of Barail group is assumed to range from 2380 m to 3355 m (Holtrop and Keizer 1970). In the stable shelf area of Bangladesh Barail is represented by a reduced thick of about 160 m of siltstone, sandstone and carbonaceous shale. The age of Barail group in the type locality and in northeastern Bangladesh is Oligocene.

The Barail group is the most important oil producing unit in Assam, India; its continuation into the subsurface of Bangladesh has yet to be explored.

2.4.3.2. NEOGENE.

(a) Surma Group (Lower Miocene - Quaternary)

The Surma group essentially consists of a monotonous succession of alternating sandstone, shale, siltstone and sandy shale with occasional thin conglomerate beds. The name of the unit was originally proposed by Evans (1932) as the "Surma Series" for a thick (3,500 m) sequence of shale, sandy shale, sandstone and thin conglomerate unconformably overlying the Barail group (Barail Series of Evans, 1932) in the Surma valley situated in the northeastern part of Bangladesh and adjacent Indian territory of Assam. The Surma group is a diachronous unit which was deposited in the early Miocene in Assam and continued through to the Quaternary in the present Bay of Bengal (Brunschweiler 1980).

The group is well exposed over large areas of the folded belt extending from the Surma valley in the north into the Tripura and Chittagong hills to the south and continuing into the Arakan, Burma (Fig.2.11). It was reached in all the wells drilled in the alluvial covered plain to the west of the folded belt.

The Surma group is divided into two formations - the Bhuban and the Bokabil formations, based on differences in gross lithology. Because of frequent horizontal facies variations, scarcity of fossils and the repetitive nature of the shale and sandstone in the sequence, the correlation of these sub-units from the type area in the Surma valley with those of the Tripura and Chittagong

folded belts and the alluvial covered area of southern and western Bangladesh are speculative.

The Bhuban formation consists of shale, claystone, siltstone and sandstone with occasional thin conglomerates. The sandstones are greyish to yellowish, mostly fine to very fine grained, well sorted to very well sorted, micaceous, well cemented and hard. The shales are dark grey, hard, fissile, laminated to cross laminated, and often silty. In the Surma valley this formation is further sub-divided into two sandstone dominated units (with minor shale and siltstone) separated by a predominantly argillaceous unit in the middle. This distinction disappears in the western part of Surma valley and the whole Bhuban formation becomes a monotonous succession of shale, sandy shale, and siltstone with interbeds of sandstone (Mathur and Evans 1964).

The Bokabil formation is lithologically a passage into the overlying arenaceous Tipam Group (see next section) and consists of shale, siltstone and thick lenticular beds of sandstone. The sandstones are greyish to yellowish, mostly medium to fine grained, micaceous, well sorted, and moderately to poorly consolidated. The shales are bluish grey, frequently silty and moderately hard to rather soft. It is distinguished from the Bhuban formation by absence of well bedded sandstones in it and from the overlying Tipam group by a greater proportion of shale (Mathur and Evans 1964). In the Surma basin, the top of the formation is marked by a marine shale unit named the "Upper marine shale", encountered in all the wells drilled there. Towards the west, the lithology contrast between the predominantly sandy Bokabil and Upper Marine shale becomes vague and undetectable (Holtrop and Keizer 1970). There is much variation in the amount of sandstone in the Bokabil formation (Mathur and Evans 1964).

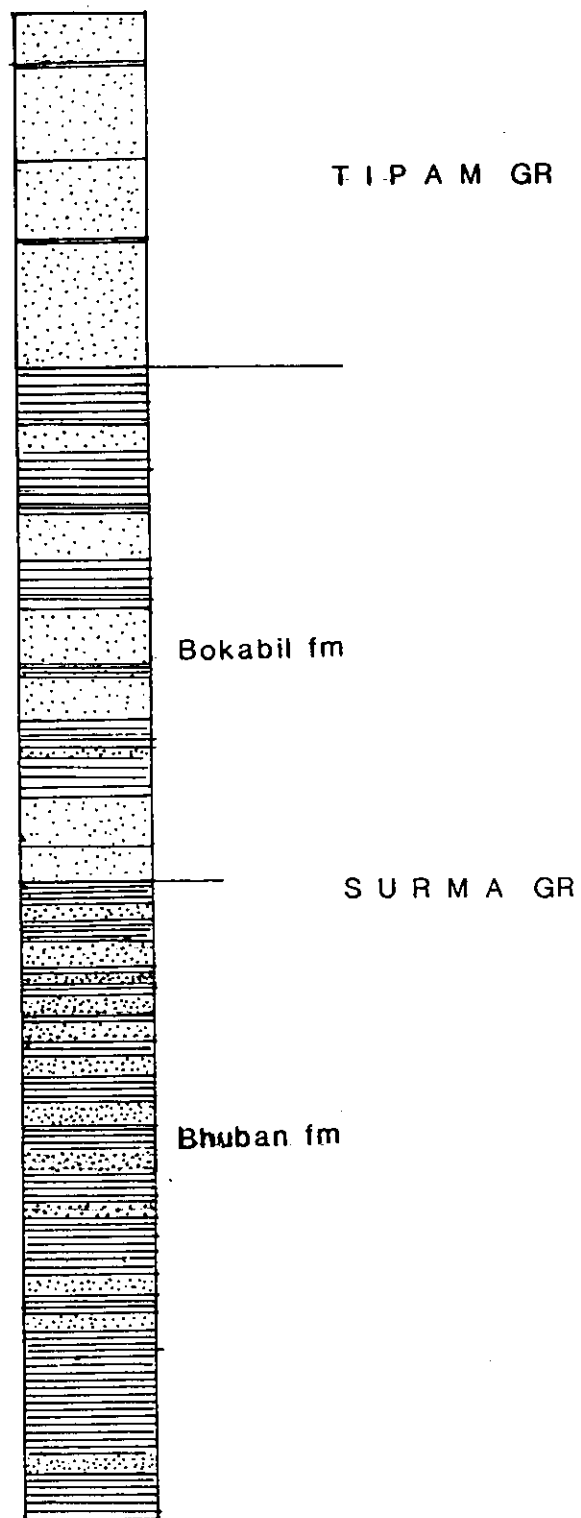


FIG 2.13 GENERALIZED COLUMNAR SECTION OF
THE NEOGENE OF THE BENGAL BASIN

 Shale

 Sandstone

The sediments of the Surma group are indicative of the repetitive deposition of prodelta, delta front and paralic facies with minor wholly marine facies. Holtrop and Keizer (1970) recognized several cycles of the following upward sequence in the lower part of the Surma group: distal fluviomarine, proximal fluviomarine, barrier bar sand, lagoonal clay and multiple channel deposits. The sediments of this group correspond to a major phase of delta progradation punctuated by short periods of marine transgression. The "Upper Marine Shale marking the top of the Bokabil formation in the Surma basin (Sylhet trough), represents the last of such marine transgressions over that particular area. The thickness of Surma group is up to 5,500 m at the Surma valley in Assam. The age here is Lower Miocene. In the subsurface of Bangladesh it is represented by about 2,800 m in the Rashidpur-2 well where it is Lower - Middle Miocene. Further south about 2,800 m (Lower - Middle Miocene), 2,156 m (Upper Miocene - Pliocene) and 3,500 m (Upper Miocene - Quaternary) of the Surma sediments were drilled in Titas-2, Begumganj-1 and Muladi-1 wells respectively although the base was not reached. In the stable shelf area to the northwest, the Surma group is undifferentiated and is represented by a reduced thickness of about 400 m. The Surma group represent the only proved hydrocarbon (gas) bearing sequence in the Bengal Basin.

(b) Tipam Group (Middle Miocene - Quaternary):

The Tipam group is a sequence of predominantly arenaceous sediments of continental origin overlying the Surma group. The name of the unit originated from the Tipam hills of Assam where a thick sequence of sandstones is exposed. It is also exposed over large areas in the fold belts of Bangladesh and is encountered in all the wells drilled in the alluvial covered plain to the west. The Tipam group is

a diachronous unit whose age ranges from Middle Miocene in Assam to Quaternary in Southern Bangladesh.

The Tipam group is divided from the bottom upwards into the following three formations: Tipam Sandstone formation, Girujan clay formation and Dupitila Sandstone formation. The Tipam sandstone formation is made up of thick, massive to cross-bedded, medium to coarse grained sandstones with subordinate (10% -20%) shale interbeds. The sandstone are yellow brown to red brown in colour, ferruginous, non-cemented and friable; they occasionally contain pebbles of shale, granite, quartzite etc. The shales are bluish grey, soft and silty. The Girujan clay formation is essentially an argillaceous facies consisting of mottled clay, sandy shales with subordinate mottled gritty sandstones. This unit is of very local extent and is missing from large areas including the central Surma valley, Chittagong Hill tracts and Tripura Hills. Where the Girujan clay formation is absent, the Tipam sandstones cannot be distinguished from the Dupitila sandstone formation above. The Dupitila sandstone formation consists of coarse, ferruginous sandstone with occasional layers of quartz pebbles and a few clay intercalations. The sandstones are cross bedded, non-cemented, friable and occasionally contain large fragments of petrified wood.

The cumulative thickness of Tipam group is about 2,500 m onshore and about 5,000m offshore. In borehole sections, it is represented by 2,100 m in Kailashtila well (Sylhet trough), 1,400 m in Begumganj-1 well and 1,120 m in Muladi-1 well. In the Bengal shelf area to the northwest the Tipam Group is represented only by the Dupitila sandstone and has a reduced thickness of about 300 m .

The Tipam group was deposited in a terrestrial environment. The conditions were fluvial for the Tipam sandstone formation, possibly lacustrine for Girujan clay

formation and again fluviatile for Dupitila sandstone formation (Holtrop and Keizer 1970). It represents a broad front of fluviatile advance across the entire basin following the retreat of the Surma Group environments towards the South.

(c) Dihing Group (Plio-Pleistocene)

The Dihing Group consists of pebbly sandstones and sandstones unconformably overlying the Tipam group. It has a very restricted areal extent. The sandstone is coarse-grained, ferruginous and unconsolidated. In the entire Surma valley, Tripura and Chittagong region, the Dihing Group has been recognized only in few isolated outcrops.

2.4.3.3. The Bengal Alluvium (Holocene — Pre Holocene)

This unit consists of sand, silt and gravel and occupies extensive tracts of the modern Bengal plain, including the deltaic plain of Bangladesh. The thickness varies from 30 to 300 m. The Bengal alluvium conceals all older rock units in the vast plain area west of the folded belt.

2.5 EVOLUTION AND SEDIMENTATION HISTORY OF BENGAL BASIN

The evolution of Bengal Basin is associated with the collision of two continental plates - the Indian plate and the Eurasian plate. The basin acquired its present configuration during Neogene.

The Indian subcontinent was originally attached to what are now Antarctica, Australia and Africa, forming part of the Supercontinent Gondwanaland. Initial rifting of the Indian block within the Gondwanaland is evidenced by the development of late Mesozoic grabens, within which Upper

Carboniferous to Jurassic sediments have been preserved. The volcanic activity represented by Rajmahal trap of Jurassic to early Cretaceous age is believed to be related to tensional fractures formed during the continental break-up stage: radiometric ages of Rajmahal trap basalts are 100-105 m.y. which is about the time sea-floor spreading data indicate that India broke away from Antarctica (Sclater and Fisher 1974). Following the separation, the Indian continent drifted north and made the longest migration of all the continental masses, approximately 9,000 km (Desikacher 1974). The drifting stage came to an end with the collision of India with the Eurasian continental margin and the uplift of the world's highest mountain range, the Himalayas. The existence of the Bengal Basin as we know it today derives from this collision (Fig.2.14).

The sedimentation history of the Bengal Basin area can be divided into two phases. The first one - a marine transgression following an initial subsidence in the early Cretaceous lasted until the end of the Eocene. This phase preceded the collision of India with Eurasia. It was followed by a regressive phase (with minor transgressions) from the Oligocene onwards.

(a) Phase One: Marine transgression: Pre-collision stage (Late Cretaceous to Late Eocene)

While the Indian plate drifted north from late Cretaceous to late Eocene times, wedges of miogeosynclinal sediments built outward along its eastern shelf. The Eocene Jaintia group of sediments is the most distributed rock unit of this phase. The Tura sandstone formation represents an early Eocene coastal marine to estuarine environment. A further increased transgression in the middle Eocene resulted in calm, open marine conditions under which the Sylhet Limestone was widely deposited. During the late

Eocene Bengal shelf received extensive fine clastics, possibly due to uplift in a distance cratonic area, resulting in the deposition of the Kopilli formation. Open marine condition was replaced by more brackish and estuarine environments. This marked the end of the wholly marine regime of deposition in Bengal shelf.

At this time, the "Eocene hinge" came into existence, extending from present day Calcutta northeast across the whole of the Basin (Desikacher 1974). A large thickness of argillaceous sediment (Disang gr. equiv.) accumulated to the east of the hinge zone, in which direction the shelf was open to the ocean. Through most of this time, the Indian plate was moving north at an average rate of 11 cm/year (Desikacher 1974).

(b) Phase Two: Regressive Phase: Post collision stage:
(Oligocene - Recent)

India collided with the Eurasian plate in the early Oligocene, causing the uplift of the Indoburman range to the east and ancestral Himalaya to the north. The motion of the Indian plate slowed markedly from the Late Eocene to the early Oligocene and then resumed in a north-northeasterly direction. It was during this latter period that the Bengal Basin came into existence (Paul and Lian 1975). Uplift of the Indoburman range caused a marine regression to the south and the new river system produced a delta in the northeast corner of the newly formed basin: the calcareous Jaintia and shaly Disang sediments grade upwards into the sandy and coaly Barail in Assam indicating lagoonal, deltaic and fluvial conditions. Intermittent periods of partial emergence produced the environments in which coals developed, principally in upper Assam (Raju 1968). To the southwest, in the Surma valley, the Barail group is more shaly indicating continued marine sedimentation in this deeper basinal part

in Oligocene times. The present day delta system was initiated in Assam, northeast of Bangladesh, in the Oligocene with a regional paleoslope from northeast to southwest.

In the early and middle Miocene, major phases of uplift in the Himalaya took place due to continued collision between the Indian and Eurasian plates. This resulted in a widespread unconformity between the Oligocene and Miocene sediments. As a result of this orogenic uplift, a large volume of coarse to fine clastic sediment was supplied to the basin. The southwest - northeast trending hinge zone became reactivated with rapid subsidence of the foredeep adjoining the platform. A thick deltaic lap sequence of alternating coarse and fine clastics (Surma group) progressively infilled a vast basinal area, including the Surma Valley, Tripura and the Chittagong hills. The Surma group represents a complex interfingering of deltaic sub-environments with minor marine facies and indicates a major phase of marine regression with minor intermittent transgression. Some of the sediments were transported further south into the deeper part of the ancestral Bay of Bengal, possibly by turbidity currents, to form the Bengal deep sea fan complex. A second phase of major orogeny affecting the Himalaya in the latest Pliocene and Pleistocene is believed to have supplied a large amount of sediment to the fan complex. The Neogene of the Bengal Basin is thus marked by a major delta progradation toward the southwest and south, the apex of the delta lying to the northeast. As the delta prograded the Surma group environments retreated south along a broad front to make room for advancing fluviatile sedimentation, which is represented by the Tipam group. The present Ganges - Brahmaputra river system evolved into Holocene and covered the older sediments with a blanket of fluviatile and delta plain deposits. The progradation of the delta is still continuing and the present Ganges - Brahmaputra delta is

seen as an active continuation of the system that originated at the beginning of the Neogene.

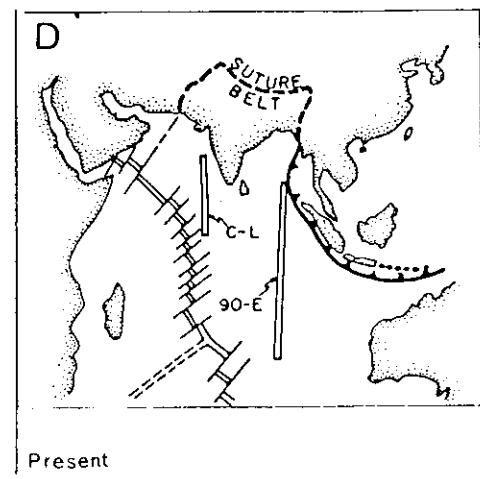
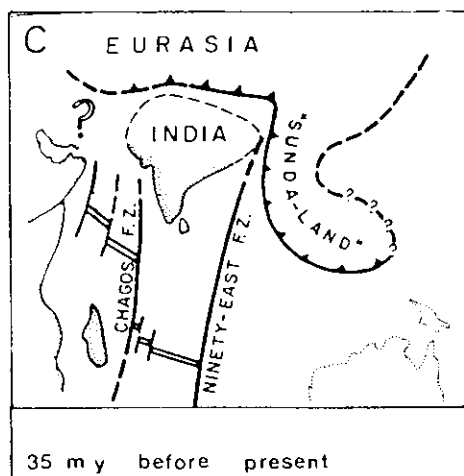
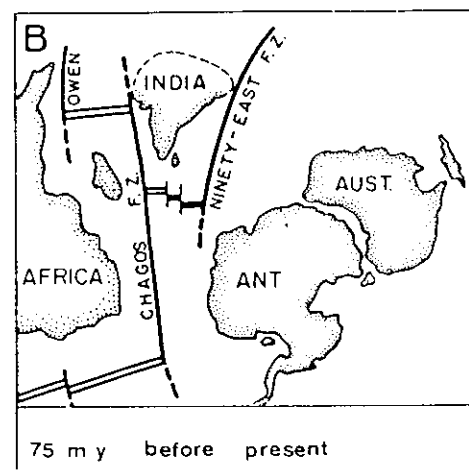
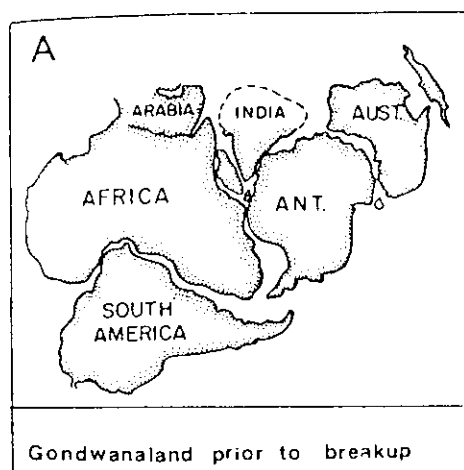


FIG.2.14 BREAKUP AND MIGRATION OF INDIA FROM GONDWANALAND

(After Graham et al '75)

CHAPTER 3

PETROGRAPHY OF SANDSTONE I - DETRITAL PHASES

3.1 INTRODUCTION

The primary mineralogical and textural properties of sandstones are factors, in addition to the pressure, temperature and pore water chemistry, on which the subsequent course of burial diagenesis may strongly depend.

Petrographic analyses were carried out on forty sandstone samples, thirty six from the Surma Group and the remaining four from the Tipam Group of the Neogene sequences. These samples are from five onshore wells, Muladi-1, Muladi-2, Begumganj-1, Begumganj-2 and Titas-2 (Fig. 1.3), and range in depth from 500 m to 4491 m below the surface. Compositional and textural properties of the detrital phases of the sandstones, as observed in this study are discussed in this chapter. The petrography of the diagenetic phases is discussed in the following chapter (Chapter 4).

3.2 ANALYTICAL TECHNIQUES:

(a) Standard thin section technique:

The basic data for mineralogical, compositional and textural properties of the sandstones were collected from standard thin sections. The standard thin sections were prepared from conventional core samples after impregnating the rock with araldite. All the slides were stained with Alizarin Red S and Potassium ferricyanide for carbonate determination according to the table 3.1. (Dickson 1966). The proportions of various rock components in each slide were estimated by point counting, with 600 counts being

made per slide.

For the estimation of primary and secondary porosity (under thin section), a second set of thin sections was prepared after impregnating the rock with araldite stained with red dye to facilitate the identification of primary and secondary porosity. The porosity was estimated by point counting with 2000 counts being made per slide.

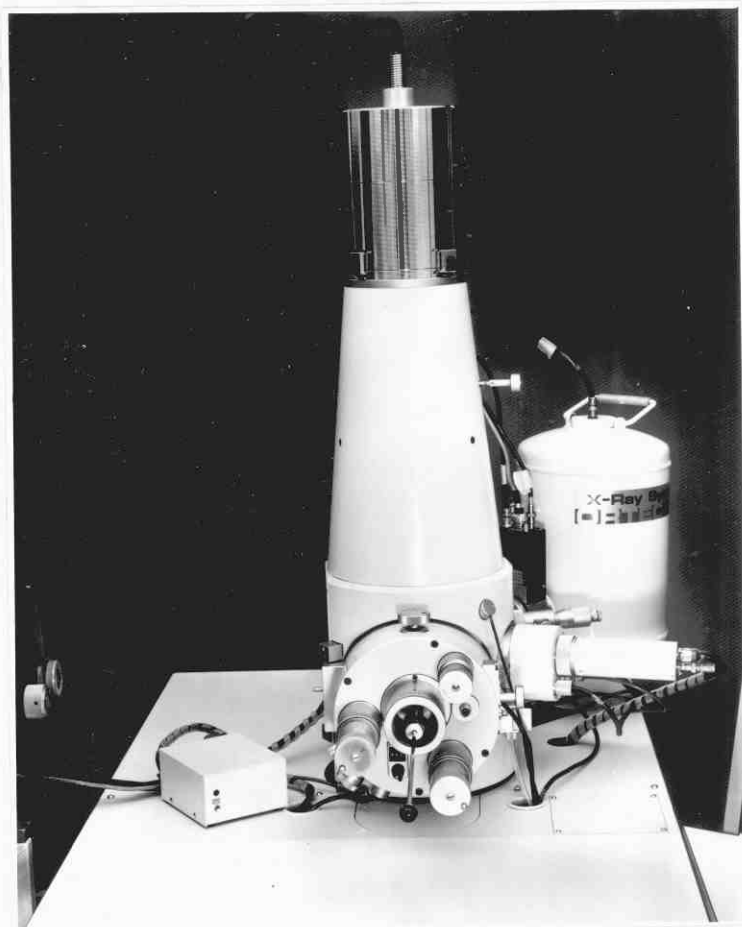
Table 3.1 STAINING FOR CARBONATE MINERALS.				
(After Dickson '66)				
Staining Agents	Calcite	Ferroan Calcite	Dolomite	Ferroan Dolomite
Alizarin Red S	Pink to red	Pink to red	-	-
Potassium ferricyanide	-	Pale to deep turquoise	-	Pale to deep turquoise

(b) Scanning electron microscopy:

The sandstones were examined using a Scanning electron microscope (SEM), JEOL JSM 35, fitted with an energy dispersive x-ray analyser (Fig. 3.1). This technique was most useful in the study of the diagenetic phases especially the authigenic clay minerals. For the purpose of sample preparation, a small cube measuring about 1 cm² was cut from the core, dried and mounted on an aluminium stub with Duraflex glue. A fresh surface of the cube was exposed by splitting the upper half of the cube from the part attached to the stub. The sides of the specimen were



A



B

Fig.3.1.Scanning electron microscope. (A) Controll panel
(B) Specimen chamber.

painted with conductive silver paint. The top exposed surface of the specimen was coated with gold under vacuum using a sputter coater. The sample was then ready for examination under the SEM. Elemental analyses of any phase to aid their identification were carried out, if required, by the use of the energy dispersive x-ray analyzer attached to the microscope.

(c) Cathodoluminescence technique:

This technique was used specifically to distinguish quartz overgrowths from detrital quartz grains. The Cathodoluminescence detector was fitted onto the SEM described above. The sample preparation technique is identical to that for the SEM analysis with the exception that the sample was coated with carbon instead of gold.

(d) X-ray diffraction technique:

X-ray diffraction analyses of powdered whole rock samples were carried out to examine variations in the bulk mineral composition of the sandstones. For the purpose of sample preparation, the sandstone was broken into small pieces (0.5 - 1.0 cm) and crushed to a fine powder using a Tema Mill. The powder was then loaded into the cavity of an aluminium holder and the surface smoothed with a glass side. The specimens were scanned from 2° to 75° in the diffractometer using a $\text{CoK}\alpha$ radiation.

3.3 TEXTURE

The textural properties of sandstones can be expressed in terms of their grain size, sorting, shape (sphericity) and roundness (angularity) parameters. Exact measurements of these parameters such as grain size (grain diameter for example) etc. are complex and time consuming from thin sections, but estimations of these parameters from visual

comparators are usually adequate and can be done fairly rapidly from thin sections using a binocular microscope (Beard and Weyl 1973).

The Wentworths grain size scale (Wentworth 1922) was used as the reference scale for grain size determinations. Grain size estimations from thin sections by visual comparison with reference slides indicated that, of the total number of samples studied, 48 % are fine grained, 13% are very fine grained, and 39% are medium to fine grained (Table 3.2).

Reference photographs for estimating sorting from thin section as provided by Beard and Weyl (1973) were used as visual comparators. According to this study 65% of the total number of samples are well sorted, 30% are very well sorted and 5% are moderately well sorted.

Grain shape (sphericity) and roundness (angularity) reference photographs available from Beard and Weyle (1973, after Fowler 1953) were used as visual comparators. The study showed most of the grains to be of high sphericity and subangular to subrounded.

3.4 POROSITY

Thin section porosity was measured by point counting for each sample and was used for a comparative study of the porosity of the samples. Stanton (1977) has show , from the study of the U.S. Gulf Coast Eocene sandstones, that thin section porosity is consistently lower than the porosity determined by porosimeter, with an average difference of 13.5 percent. In the present study, three porosity values were calculated from each thin section and the mean taken. The results are tabulated in Table 3.2 which shows that the thin section porosity of the studied samples ranges from 0% to 22%.

TABLE 3.2. GRAIN SIZE AND POROSITY OF THE STUDIED NEOGENE SANDSTONES FROM THE BENGAL BASIN.

(Legends: M-F = Medium to fine; F = Fine; VF = Very fine;)

Unit	Sample no. (Gr)	Depth (m)	Grain size	Thin section porosity	Sample no.	Depth (m)	Grain size	Thin section porosity
T	Muladi wells							
I	M1-3	500	M-F	22%				
P	M1-4	602	M-F	21%				
A	M1-7	878	M-F	20%				
M	M1-11	1137	M-F	20%				
	M1-20	1700	VF	16%		Titasi-2 well		
	M1-27	2212	M-F	0%	T-1	2646	F	15%
	M1-28	2216	VF	11%	T-2	2660	F	9%
S	M1-29A	2290	F	0%	S T-3	2668	M-F	9%
U	M1-34B	2865	M-F	12%	U T-4	2673	M-F	8%
R	M1-36	3114	F	9%	R T-5	2678	F	10%
M	M1-37B	3216	F	8%	M T-7	2690	M-F	11%
A	M1-38	3294	F	11%	A T-10	2706	F	9%
	M1-39	3420	VF	0%	T-11	2712	F	
	M1-49A	4458	F	6%	T-12	2718	F	
	M2-3	4491	M-F	4%	T-13	2725	VF	8%
					T-15	2753	M-F	9%
					T-16	2761	M-F	
	Begumganj wells				T-18	2927	F	
	B1-14	2139	M-F	15%	T-19	2935	F	7%
S	B1-16	2680	F	4%	T-21	2967	F	7%
U	B1-17	2758	F	8%				
R	B1-19	2999	F	7%				
M	B2-1	3028	M-F	3%				
A	B2-3	3325	F	7%				
	B2-4	3476	VF	5%				
	B1-26A	3642	M-F	9%				

3.5. DETRITAL COMPOSITION :

From the thin section petrographic study the detrital compositions of the studied sandstones can be subdivided as follows :

(1) Framework grains :

Majors : Quartz
Feldspars
Lithic fragments

Minors : Chlorite
Micas

Minor Accessories : Heavy minerals
Organic matter
Fossils

(2) Matrix

3.5.1 FRAMEWORK GRAINS:

(a) QUARTZ:

Quartz is the most abundant of the detrital rock components. Percentage estimations by point counting indicate that quartz constitutes 51% to 76% of the total rock components in the studied samples (Table 3.3). However, some of the quartz percentage estimates, especially those towards the higher end (70% - 76%), are believed to have included important amounts of authigenic quartz cement, due to not being able to always adequately distinguish authigenic quartz from detrital quartz grains under the ordinary petrographic microscope (see section 4.3.1).

Table 3.3. QUANTITATIVE ESTIMATION OF ROCK COMPONENTS IN
NEOGENE SANDSTONES, BENGAL BASIN.

Gr	Sampl	% Qtz	% Feld	% Lith	% Chlor	% Mica	% Matx	% Calc cent	% Sid cent	% Kao cent
	no.									
T I P A M	M1-3	52	11	10	8	10	5	-	-	-
	M1-4	57	21	9	5	1	4	-	-	-
	M1-7	57	9	9	7	3	11	-	0.1	-
	M1-11	54	15	14	3	4	8	-	0.3	-
S U R M A	M1-20	51	4	8	7	6	13	7	-	-
	M1-27*	50	3	12	2	1	-	30	-	-
	M1-28	55	3	10	3	15	7	3	0.2	-
	M1-29A*	31	0.5	5	2	3	-	58	-	-
	M1-34B	59	10	20	6	3	2	1	-	-
	M1-36	61	8	19	6	5	3	0.8	0.3	-
	M1-37B	61	8	11	5	5	3	5	1	0.1
	M1-38	63	8	20	3	2	3	-	0.6	0.1
	M1-39*	40	-	7	0.5	2	-	50	-	-
	M1-49A	68	5	16	5	1	1	2	0.3	-
	M2-3	55	5	18	2	2	0.4	16	-	-
	B1-14	55	11	21	3	3	4	0.3	-	-
	B1-16*	49	7	8	1	4	2	25	0.3	-
	B1-17	60	6	14	4	6	7	1	0.6	-
S U R M A	B1-19	64	4	11	6	3	8	2	0.3	0.1
	B2-1*	42	2	14	4	1	-	35	-	-
	B2-3	65	5	17	2	1	7	0.6	0.8	-
	B2-4	56	3	16	3	1	3	17	0.4	-
	B1-26A	54	11	18	5	2	5	-	-	2
	T-1	65	10	14	4	2	3	0.3	0.4	1.3
T-2	T-2	60	14	16	3	3	4	-	-	0.4
	T-3	57	10	21	6	1	2	1	-	1
	T-4	63	6	16	1	2	8	1	0.6	1

	T-5	67	6	16	4	3	3	0.3	-	0.3
S	T-7	59	10	20	2	2	3	-	0.6	2
U	T-9	67	10	13	4	2	5	0.1	-	0.4
R	T-10	73	7	15	2	1	3	0.2	-	0.2
M	T-11	60	15	17	2	0.3	3	-	1	-
A	T-12	64	7	18	4	1	5	0.3	-	0.1
	T-13	67	5	17	2	4	3	0.8	-	-
	T-15	60	9	22	2	0.9	5	0.3	1	0.5
	T-18	70	4	11	5	4	4	0.3	0.6	-
	T-19	76	3	12	1	1	1	0.4	0.3	-
	T-21	73	3	11	1	0.5	-	2	1	-

(* Estimated detrital grain percentages
are not representative in these samples
because of high calcite cements)

The detrital quartz grains are predominantly monocrystalline, but polycrystalline quartz constitutes a small proportion in each sample, being more abundant in the coarser grained than the finer grained sandstones. Typical percentages of polycrystalline quartz range from 15% of the total quartz in medium to fine grained sandstone to about 9% in fine to very fine grained sandstone. Normal extinction is shown by the majority of the quartz grains although undulose extinction is common. The undulose extinction is indicative of strained quartz crystals which might have been subjected to deformation either in the course of crystallisation or after formation (Blatt and Christie 1963). Inclusions in the quartz grains include biotite but these are rare. A few quartz grains have abundant vacuoles as indicated by their cloudy appearance. Vacule-rich quartz appears to be derived from hydrothermal vein sources (Scholle 1979). The detrital quartz grains are medium to fine grained sand size and are subangular to subrounded with occasionally well rounded grains.

(b) LITHIC GRAINS

Lithic grains are the second most abundant grain type among the rock components. Lithic fragments make up 8% to 22% of the total rock components in the samples with a mean value of 14% (Table 3.3). The lithic grains include metamorphic, sedimentary, igneous and volcanic rock types but are dominated by metamorphic and sedimentary rocks. The terms used by Ingersoll and Suczek (1979) and Graham, Ingersoll and Dickinson (1976) have been found most convenient in describing the lithic fragments. These descriptions also help in comparing the present results with those of the petrological study of the Neogene sands of the Bengal deep sea fan, south of the presently studied area by Ingersoll and Suczek (1979). The relative abundances of the various types of lithic grains in 10

representative samples have been determined by point counting. They show that the quartz-mica tectonite, quartz-mica-aggregate, argillite shale and chert are the four most abundant (Table 3.4).

The quartz-mica-tectonites are metamorphic lithic grains composed of quartz and mica with a preferred planar fabric. This group of lithics is one of the most abundant type, consisting of 22% to 30% of the total lithic fragments in the representative samples.

The quartz-mica-graphite tectonites are metamorphic lithic grains similar to the quartz-mica-tectonites but containing black-opaque graphitic material in addition to quartz and mica. The constituent components have preferred planar fabric. This group is common in the samples and constitutes 5% to 13% of the total lithic grains.

The mica-tectonites (Polycrystalline mica) are metamorphic lithic grains consisting of mica aggregates which have schistose fabrics. These are a relatively minor component consisting of 1% to 9% of the total lithic fragments.

The Foliated metaquartzites are metamorphic lithics with polycrystalline quartz having a planar fabric. They occur as very minor components representing only 1% to 5% of the total lithic grains.

The Argillite-shales are fine grained, argillaceous sedimentary lithic grains mainly composed of shale, mudstone and argillite fragments with occasional enclosed detrital silt clasts. They are one of the most abundant types of lithic fragments comprising 22% to 29% of the total lithic grains in the representative samples. Their murky appearance and frequent plastic deformation help to distinguish them from other lithics.

TABLE 3.4. RELATIVE ABUNDANCE OF LITHIC GRAIN TYPES IN
REPRESENTATIVE NEOGENE SANDSTONES (SURMA GR)
OF THE BENGAL BASIN

SAMPLE NO.	M1 34B	M1 36	M1 38	B1 14	B2 3	B1 26A	T- 1	T- 12	T- 15
Qtz-mi-tec	22%	24%	23%	29%	23%	26%	24%	30%	23%
Argll-sh	23%	25%	24%	23%	25%	26%	22%	29%	29%
Chert	14%	9%	11%	10%	12%	17%	14%	10%	15%
Qtz-mi-gra- tec.	13%	6%	7%	11%	10%	5%	11%	6%	5%
Qtz-mi-aggr.	8%	10%	9%	9%	11%	12%	12%	9%	8%
Mica-tec.	2%	5%	6%	2%	5%	1%	9%	4%	1%
Foli-metaqtz.	1%	2%	2%	4%	1%	2%	1%	0%	5%
Volcanics	1%	0%	1%	0%	1%	1%	1%	0%	4%
Indeterminate	16%	19%	16%	11%	10%	10%	6%	11%	10%

The quartz-mica-aggregates are lithic grains of possible igneous origin, composed of quartz and mica but without a preferred planar fabric. They constitute 8% to 12% of the total lithic grains.

The quartz-mica-feldspar aggregates are lithic grains of possible igneous origin, composed of quartz, mica and feldspar but without a preferred planar fabric. This is a rare component among the lithics.

Volcancis - Fine grained lithic grains recognised as volcanic by the presence of felsitic, microlitic or lathwork texture. These are very minor components of the lithic grains.

Indeterminate - Lithic fragments of unknown and obscure nature, not falling into any of the above categories.

(c) FELDSPAR

The feldspars are the third most abundant type of mineral grain in the sandstones, comprising 3% to 15% of the total rock components, the mean value being 8%, except for one sample where feldspars constitute 21% of the total rock. Feldspars are in general more abundant in the Tipam Group of sandstones than the Surma Group. Both potassium and plagioclase feldspars are present but the former are more abundant than the latter. The potassium feldspars include orthoclase and microcline. It was found by point counting that the microcline with characteristic cross-hatched twinning constitutes 2% - 7%, orthoclase constitutes 65% - 75%, and plagioclase 20% - 32% of the total feldspars in the representative samples. Both potassium and plagioclase feldspars show evidence of alteration which is occasionally extensive. However both fresh and altered feldspars occur in the same sample. Krynine (1950) and Folk (1959) suggested that association of fresh and

altered feldspars in same samples was indicative of humid weathering conditions and high relief. James et al. (1981) considered this notion misleading because of insignificant difference in the degree of alteration from climate to climate. James et al. (1981) also suggested that, although alteration of feldspars does occur in the weathering profile it could also result from diagenetic reactions after deposition and it is very difficult to distinguish between the two in most cases. Two kinds of alteration are extensively found in the samples. These are (1) Vaculization (Fig. 3.2) and (2) Sericitization (Fig. 3.3). Vaculization refers to alteration whereby feldspars attain a grayish turbid or cloudy appearance. James et al. (1981) assumed that the vacules were voids or pores produced during weathering. Sericitization describes the alteration of feldspar to sericite, very fine grained white micas which appear as moderately birefringent flakes within the feldspars. The presence of vacules and sericite in feldspars throughout the sequence irrespective of depth of burial suggests that they were produced before burial. A third kind of alteration of feldspar - that is to kaolinite - is considered to be diagenetic because of its restricted occurrence below certain depths where pore filling kaolinite is also common. Feldspar grains also show extensive dissolution below depths of about 2000 meters.

(d) CHLORITE:

Detrital chlorite is the fourth most abundant mineral amongst the framework grains, constituting 1% to 8% of the total rock components, with a mean value of 4%. The chlorite occurs as large (0.2 - 0.5 mm) detrital flakes which often have splayed out ends, and is characterised by its green colour and strong pleochroism. The chlorite flakes are often concentrated along partings, laminae or bedding planes.

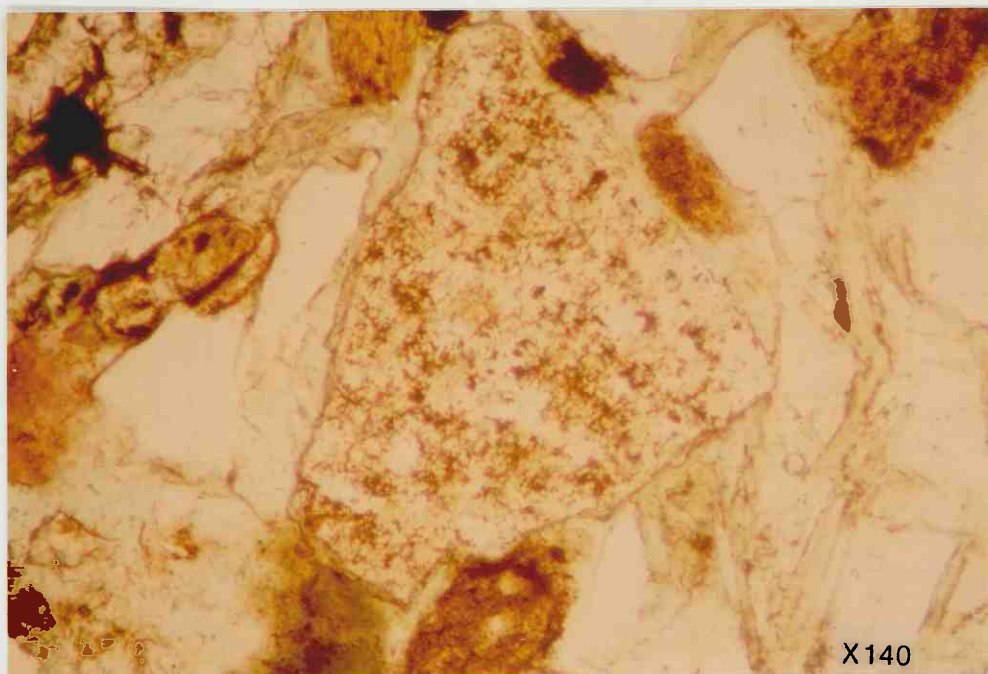


Fig.3.2. Photomicrograph of feldspar with vacuole alteration. Pla. light. Begunganj-1 well. 2139m. Bokabil fm.

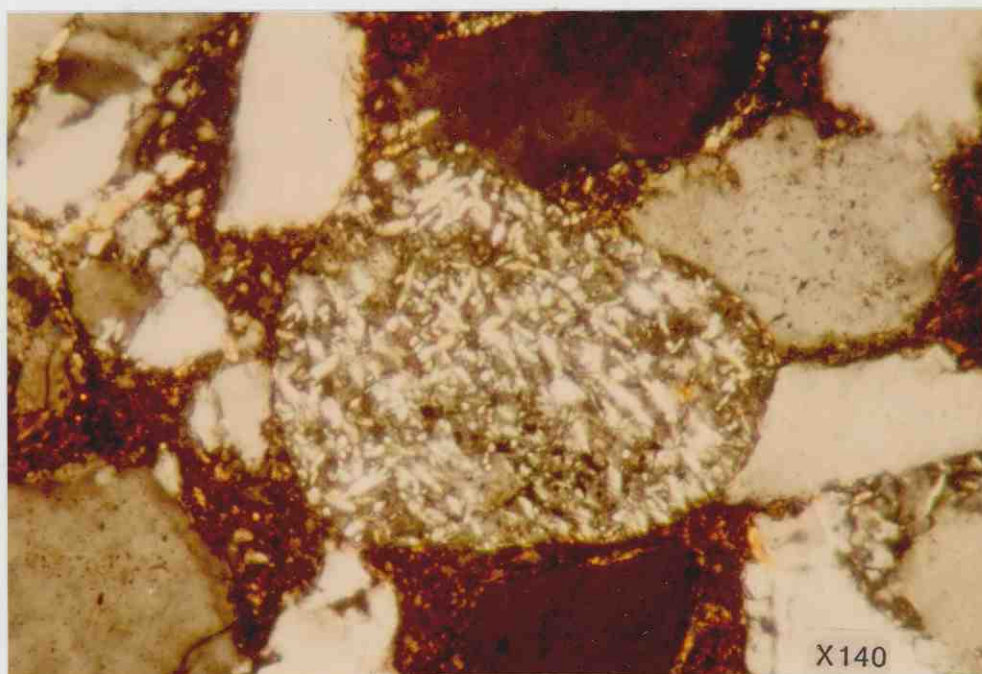


Fig.3.3. Photomicrograph of sericite alteration on feldspar. Crossed nicols. Muladi-1 well. 2216m. Bokabil fm.

(e) MICA:

Both muscovite and biotite occur as detrital flakes, often with splayed out ends. Together, they constitute 0.3% - 6.0% of the total rock components with a mean value of 2% except for two samples where they constitute 15% and 10%. The micas are often concentrated along partings, laminae or bedding planes and frequently show bending due to compaction. The muscovite, colourless and unaltered, is generally more abundant than the biotite. The biotites, brown and strongly pleochroic, commonly show varying degrees of alteration to chlorite, although fresh biotite also occurs.

(f) HEAVY MINERALS:

A wide variety of heavy minerals occur as minor accessory detrital phases, and together constitute 1% or less of the total framework grains. They include garnet, zircon, tourmaline, epidote, rutile, kyanite, sillimanite, hornblende, pyroxene and opaque heavy minerals. Garnet is the most common of the non-opaque heavies and occurs most persistently throughout the sequence. Hornblende and pyroxene are common in the shallower samples of the Tipam group but are rare in the underlying Surma group. Opaque heavy minerals are also common and their presence was noted in every sample. The principal opaque heavies that were recognized are magnetite and hematite.

(g) ORGANIC MATTER:

Detrital organic matter is present as a minor accessory component in many samples. These occur as dark reddish brown to black material with brown translucent edges, commonly as isolated grains and grain clusters or occasionally coalesced into thin laminae (Fig. 3.4). The

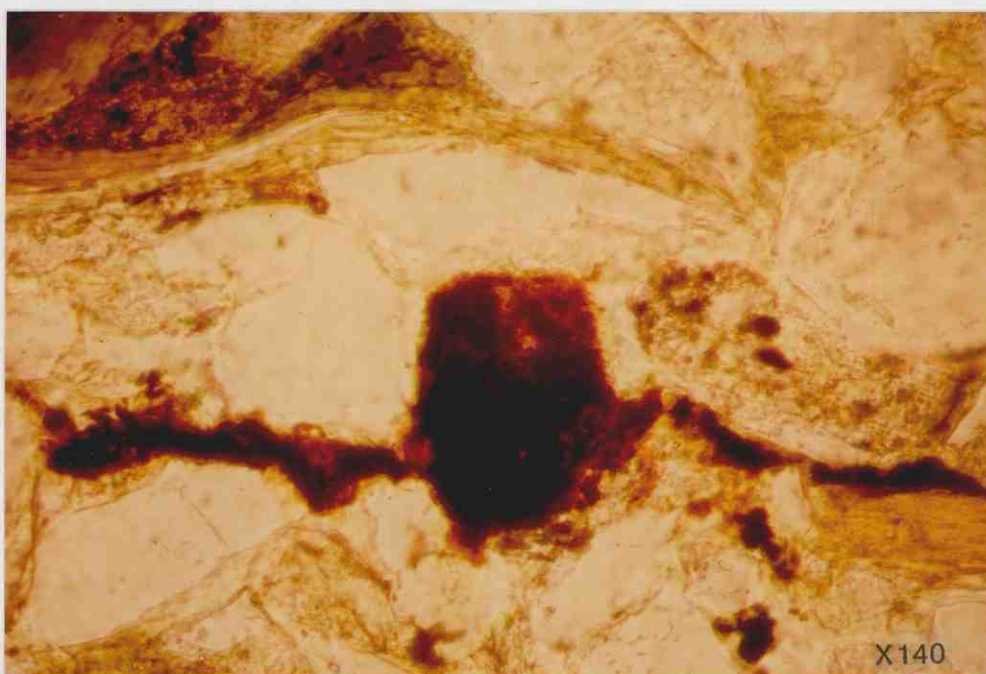


Fig.3.4. Photomicrograph of organic matter in coaly seam in sandstone. Pla. light. Muladi-1 well 2216m. Bokabil fm.



Fig.3.5. Photomicrograph of foraminifera fossil in sandstone. Pla. light. Titas-2 well. 2718m. Bhuban fm.

detrital organic matter is believed to be derived principally from terrestrial woody sources.

(h) FOSSILS:

Fragments or whole foraminifera were found in many of the samples of the Bhu ban formation. The fossil shells are 0.5 to 1.0 mm in diameter, have a circular shape with chambers (Fig.3.5). Authigenic pyrites often fill the chambers, indicating post-depositional reducing conditions.

3.5.2 MATRIX

The matrix refers to the medium silt to clay size grade (Wentworth's scale) allogenic materials in the studied sandstones. The selection of 30 microns as the upper limit for matrix material is adopted from Pettijohn et al. (1972, p. 159). Point counting showed that the matrix content is low for most of the samples and ranges from 1% to 8% of the total rock components with an average value of 3.9%. Only two samples have significantly higher matrix content than this average value (11% and 13%). Squashing and crushing of soft lithic grains, especially shale and mudstone, often produces an interstitial "paste" virtually indistinguishable from matrix materials; they are however, distinguished as ps domatrix following the criteria used by Dickinson (1970). The very fine grained nature of the matrix materials did not allow their compositions to be fully deduced from thin section examination. However, the combination of petrographic, x-ray diffraction and SEM analyses suggested that the major constituents include quartz, kaolinite, illite and chlorite.

3.6 SANDSTONE CLASSIFICATION:

The sandstone classification scheme proposed by Pettijohn, Potter and Siever (1972) has been adopted for the present study (Fig. 3.6). All the samples can be classified as arenites as they all contain less than 15% matrix. The proportions of quartz, feldspar and lithic grains were re-calculated as a percentage and plotted on a standard QFL (Quartz-Feldspar-Lithics) triangular diagram. Two QFL triangular plots are presented for sandstones of the Surma Group and the Tipam Group respectively (Fig. 3.7A and 3.7B). From the plots it can be seen that sandstones from Surma Group can be classified as sublitharenite while those from Tipam Group are classified as subarkose. It is however important to note that no sandstone from Tipam Group was available for this study except for those from Muladi-1 well

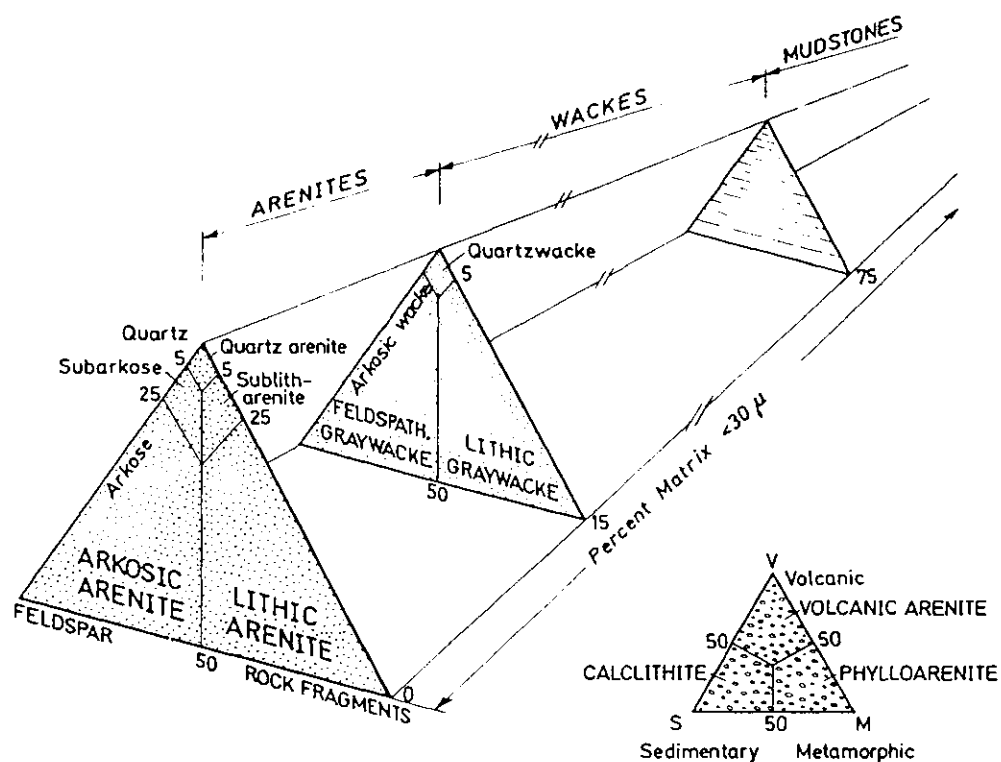


FIG . 3.6 SANDSTONE CLASSIFICATION SCHEME.
(After Pettijohn et.al.1972)

FIG.3.7A QUARTZ FELDSPAR-LITHICS TRIANGULAR PLOT FOR SURMA GROUP SANDSTONES FROM BENGAL BASIN.

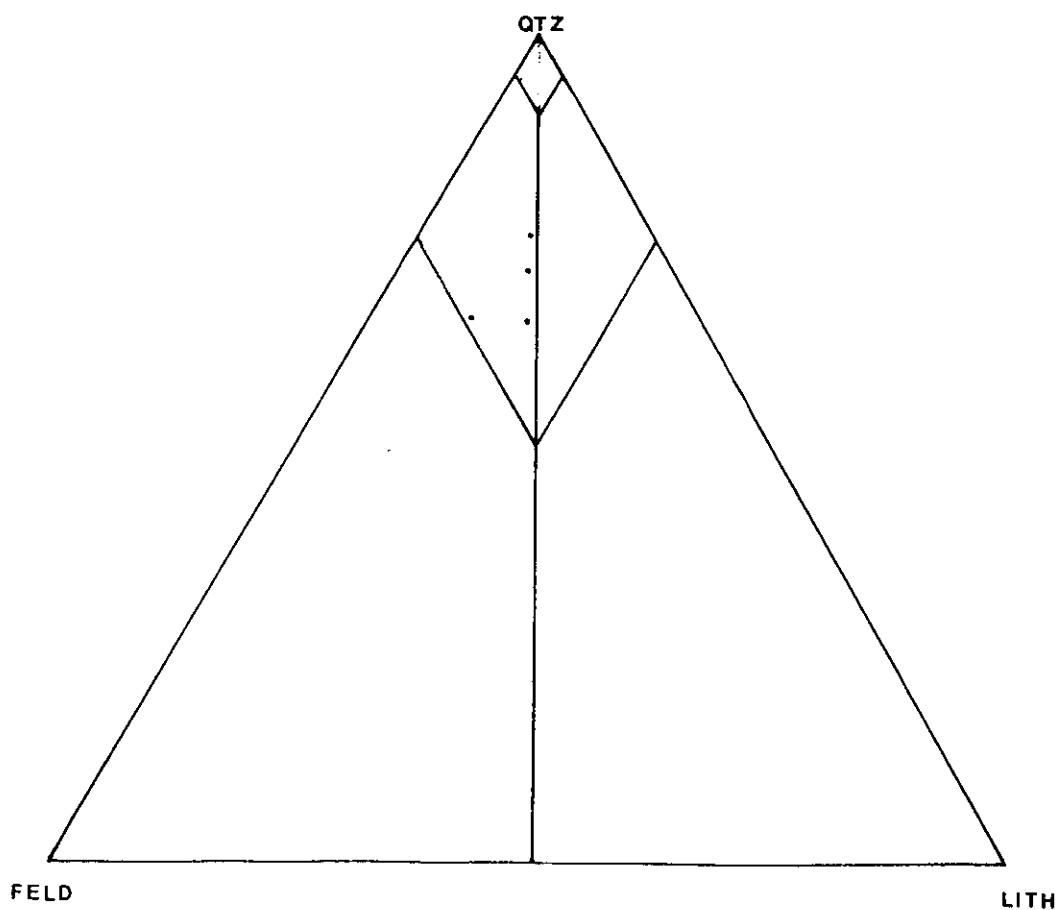
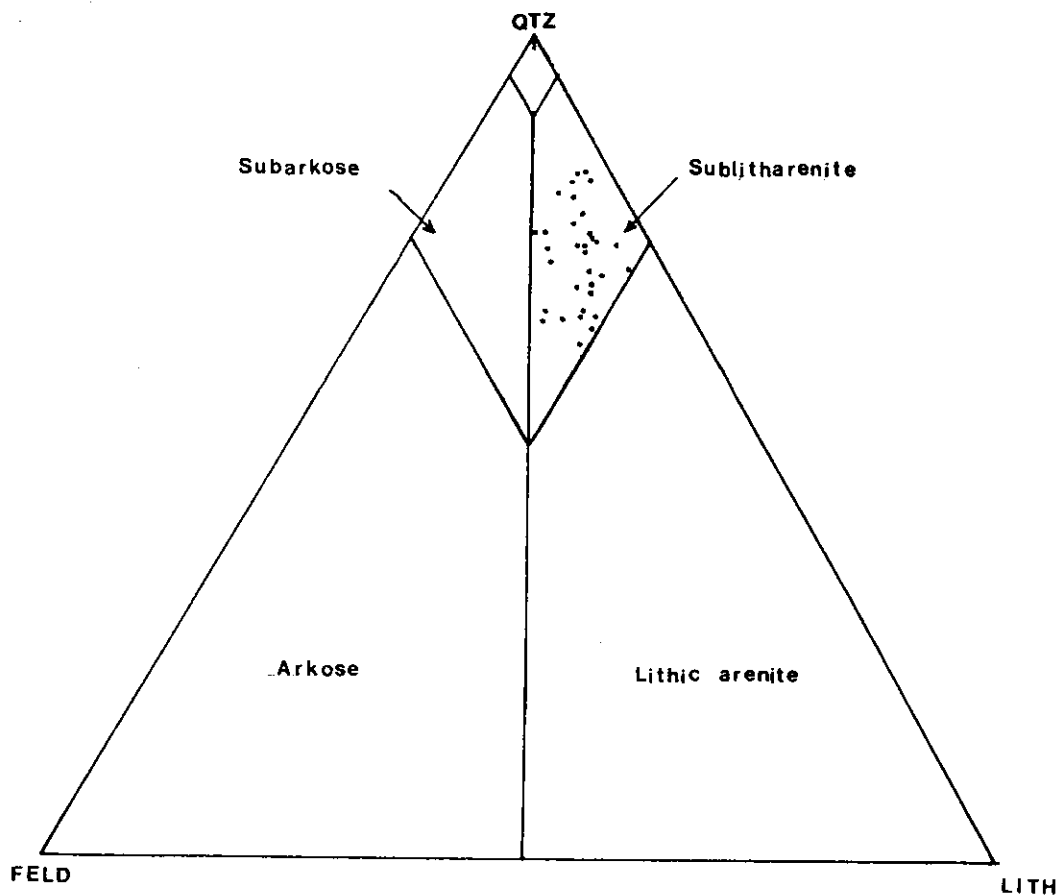


FIG. 3.7B QUARTZ FELDSPAR-LITHICS TRIANGULAR PLOT FOR TIPAM GROUP SANDSTONES FROM BENGAL BASIN.

3.7. DISCUSSION

The Neogene sandstones of the Bengal Basin consist of medium grained to very fine grained, well to very well sorted arenites. The sandstones of the Surma group, deposited in shallow marine to deltaic environments, are mostly fine grained and very fine grained and are classified as sublitharenite. The sandstones from the Tipam Group, deposited in a terrestrial fluviatile environment, are medium to fine grained and are classified as subarkose. The Neogene sandstones are, texturally mature; however their relative mineralogical immaturity is indicated by the relatively lower ratio of quartz plus chert/feldspar plus lithic grains (Blatt et al. 1972).

Little has been published on the petrography of the sands of the Ganges - Brahmaputra delta area. Siddique (1967) studied the mineralogy of the recent sediments from the Bay of Bengal sea floor and reported quartz, feldspar, mica, rock fragments, volcanic glass and heavy minerals in the assemblages. Mallik (1976) in his work on the Recent marine sediments from the shelf area of the Bay of Bengal concluded that the sediments of this area show general compositional similarities with those of the Bengal deep sea fan complex suggesting that the sediments were transported further to the south. Ingersoll and Suczek (1979) reported the petrology of the Neogene sands from the DSDP site 211 and 218 situated at the southern edge of the Bengal deep sea complex (Fig. 2.4). They estimated the percentages of the major rock components of the Neogene sands from these two wells. The data on the recalculated QFL percentages of these sands are compared with those of the presently studied Neogene sandstones in table 3.5, and the standard QFL diagram for these two groups of sands is shown in Fig. 3.8. It is shown in this figure that the Neogene sands of the deep sea fan complex (Ingersoll and Suczek 1979) fall in the subarkose and arkose groups.

These sandstones have, therefore, correlatable relations with the Neogene Tipam Group sandstones (subarkose) of the present study. The Neogene sands from the deep sea fan complex also plot in close proximity to the Neogene Surma sandstones (sublitharenite) of the present study, although they belong to a different group from the latter (Fig. 3.8). There is, however, no mention in the Ingersoll and Suczek's (1979) publication as to the depth or the horizon from which the samples were taken and the study of the Tipam Group sandstone in the present study was limited to one well, because of the lack of samples. It is important to compare the compositional properties of more Tipam Group sandstones with those of the Neogene of the Bengal deep sea fan complex (Ingersoll and Suczek 1979) for a better correlation between the two.

The lithic grain compositions of the Neogene sands from Bengal fan (Ingersoll and Suczek 1979) are compared with those of the representative sandstones from the onshore Neogene sands of the Bengal Basin in table 3.6. The dominant types of lithic grains in both groups are the quart-mica-aggregate, mica-quartz-tectonite,, chert and argillite-shales. Emphasising the role of lithic grain composition in the study of provenance, Ingersoll and Suczek (1979) concluded that the source of the Bengal-Nicobar deep sea fan complex sediments is the uplifted gneissic, sedimentary and metamorphic terranes of the Himalayas. The composition of the presently studied Neogene sand from the onshore Bengal Basin also suggests the metasedimentary, sedimentary and crystalline plutonic and gneissic basement of the uplifted Himalayas as the provenance. That the sediments of the Ganges - Brahmaputra delta area and the Bengal deep sea fan have similar provenances could also be inferred from the fact that the delta is along the dispersal route for much of the fan sediments. One reason for the lesser abundance of the argillite-shale group of lithics in the Neogene sands of

the Bengal deep sea fan complex as compared to those from the onshore Bengal Basin may be the longer distance of transportation to the former site (Table 3.6).

TABLE 3.5 : COMPARISON OF THE RECALCULATED QUARTZ-FELDSPAR-LITHIC PERCENTAGES OF THE BENGAL DEEP SEA FAN NEOGENE SANDSTONES WITH THOSE OF THE ONSHORE BENGAL BASIN NEOGENE SANDSTONES

BENGAL DEEP SEA FAN NEOGENE SANDSTONE (INGERSOLL & SUCZEK 1979)

Sample no.	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	Mean
Qtz.	54	56	71	49	63	51	61	49	56	55	54	59	58	64	58	56	56	66	51	52	61	63	57
Feld.	28	26	22	24	23	35	25	38	35	33	33	27	24	18	19	28	31	25	40	26	23	30	28
Lith.	18	18	7	27	13	14	14	13	9	12	12	14	17	17	23	15	13	9	9	21	16	7	14

ONSHORE BENGAL BASIN NEOGENE SANDSTONE (PRESENT STUDY)

Sample no.	M1 3	M1 4	M1 7	M1 11	M1 28	M1 34	M1 36	M1 37	M1 38	M1 49	M2 3	B1 14	B1 16	B1 17	B1 19	B1 26	B2 3	B2 4	T- 1	T- 3	T- 4	T- 7	T- 9	T- 11	T- 12	T- 13	T- 15	Mean	
Qtz.	71	65	75	65	82	65	68	76	69	76	70	63	76	75	80	66	75	74	72	64	74	75	66	74	65	71	75	66	71
Feld.	15	24	12	18	3	12	9	11	9	6	7	13	11	7	5	14	5	4	11	11	7	7	11	11	16	8	6	10	10
Lith.	14	10	12	17	15	23	22	13	22	18	23	24	13	18	14	21	19	22	16	24	19	18	22	15	18	20	19	24	18

• NEOGENE SST FROM
ONSHORE BENGAL BASIN

✱ NEOGENE SST FROM
BENGAL DEEP SEA FAN

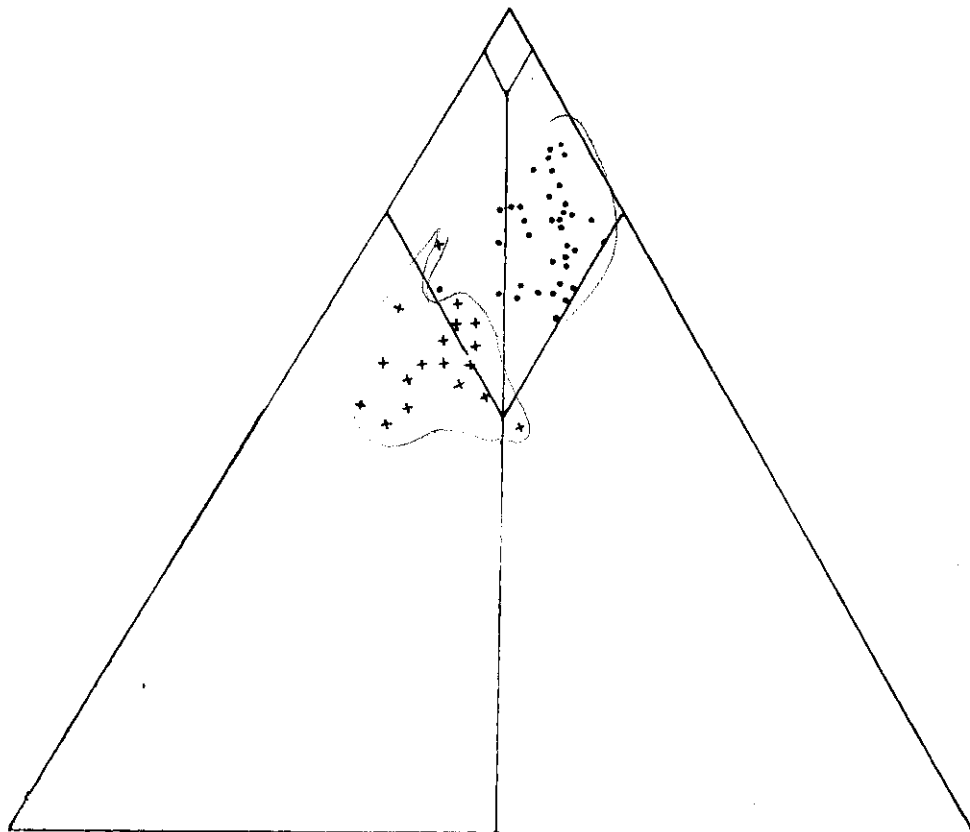


FIG. 3.8 QUARTZ-FELDSPAR-LITHICS TRIANGULAR PLOT FOR THE NEOGENE SANDSTONES FROM THE ONSHORE BENGAL BASIN AND THE BENGAL DEEP SEA FAN COMPLEX

TABLE 3.6 : COMPARISON OF THE LITHIC GRAIN TYPES IN THE ONSHORE BENGAL BASIN NEOGENE SANDSTONES WITH THOSE OF THE BENGAL DEEP SEA FAN NEOGENE SANDSTONES

BENGAL DEEP SEA FAN NEOGENE SANDSTONES (INGERSOLL & SUCZEK 1979)

Sample no.	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22
Volcanics.	4	2	1	4	4	3	4	4	4	4	4	2	2	1	1	3	6	3	3	6	6	7
Mica-tec.	14	10	8	11	4	8	8	14	8	14	11	10	7	9	8	10	14	11	8	8	12	11
Qtz-mica-tec.	31	38	27	35	24	30	28	36	23	30	32	27	28	22	34	37	27	32	35	26	33	35
Qtz-mica-aggr.	25	25	38	27	37	18	29	20	18	25	23	38	32	48	27	26	28	28	27	33	27	24
Qtz-mica-feld-aggr.	5	6	7	4	13	14	9	6	10	7	6	13	9	4	7	6	9	9	12	7	2	6
Chert	5	6	6	3	5	14	11	7	9	8	7	2	3	2	2	4	5	5	4	4	3	4
Argill-shale.	9	9	5	8	5	7	6	7	8	6	10	2	10	5	11	11	8	8	4	9	13	9
Indeterminate.	7	4	7	8	7	6	4	6	20	6	6	7	10	8	10	5	4	4	6	7	4	6

ONSHORE BENGAL BASIN NEOGENE SANDSTONES (PRESENT STUDY)

Sample no.	M1-34B	M1-36	M1-38	B1-14	B1-26A	B2-3	T-1	T-12	T-15
Volcanics.	1	-	1	-	1	1	1	-	4
Mica-tec.	2	5	6	2	1	5	9	4	1
Qtz-mica-tec.*	35	30	30	40	31	33	35	36	28
Qtz-mica-aggr.	8	10	9	9	12	11	12	9	8
Chert.	14	9	11	10	17	12	14	10	15
Argill-shale.	23	25	24	23	26	25	22	29	29
Indeterminate	16	19	16	11	10	10	6	11	10

(* including Qtz-mica-graph-tec.)

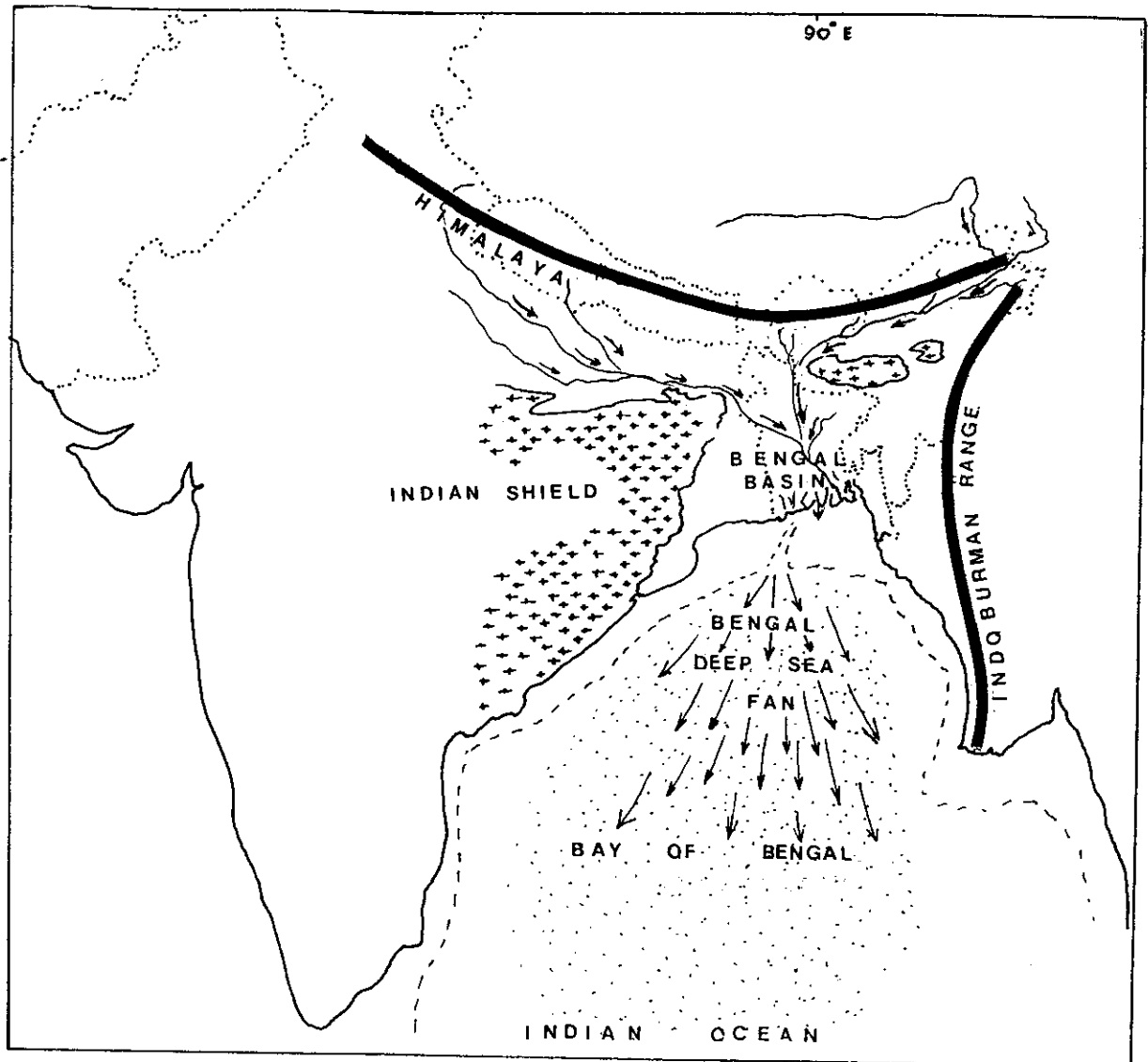


FIG 3.9 DISPERSAL ROUTE OF THE NEOGENE SEDIMENTS OF THE BENGAL BASIN AND BENGAL DEEP SEA FAN

CHAPTER 4

PETROGRAPHY OF SANDSTONE II - DIAGENETIC PHASES

4.1 INTRODUCTION

The diagenetic phases of the Neogene sandstones of the Bengal Basin include authigenic clay and non clay minerals. The clay minerals include chlorite, kaolinite and illite; they occur mainly as pore filling and pore lining agents but also replace grains in few instances. The non-clay minerals include quartz, Fe-calcite, dolomite, siderite, feldspar and pyrite and occur as pore filling and grain replacing agents. Identifications of these authigenic minerals, and study of their textural relationships with each other and to adjacent pores or grains, were made. Quantitative estimations of the major authigenic components were also attempted. The analytical techniques used for the study of diagenetic phases include standard thin section petrography, scanning electron microscopy and cathodoluminescence technique as described in Chapter 3.

4.2 AUTHIGENIC CLAY MINERALS:

4.2.1 CHLORITE:

Authigenic chlorite is a very common diagenetic mineral over the entire depth range studied. It occurs in three different forms: (a) as radial rims around detrital grains, (b) as pore filling materials and (c) as a replacement of detrital grains.

(a) Chlorite Rims

Chlorite rims around detrital grains are the most common of the three forms of authigenic chlorite. They are present, if only around a few detrital grains, in all the samples studied, and are common in most of them with one to two thirds of all the clastic grains having chlorite rims. In thin section, the chlorite rims are light green, non-pleochroic to weakly pleochroic under plane polarized light and show first order weak birefringence colours under cross nicols. The chlorite rims around grains are commonly thin and discontinuous, although thick and continuous rims were also observed in many grains. The latter totally conceal the detrital grain surfaces (Fig. 4.1) and are up to 0.01 mm thick. The chlorite rims line the pores and are absent where the grains touch each other, indicating formation after deposition (Fig. 4.2). However, later compaction has brought some coated grains into tight contact leaving some chlorite between them. The rims on closely adjacent grains have sometimes grown together to attain greater thickness.

The delicate nature of the authigenic chlorite rims is revealed from SEM studies. They are seen to consist of small irregular to hexagonal platelets attached more or less perpendicularly to the surface of the grain. The individual chlorite platelets, which are 2-3 microns in length, are more or less radially arranged with respect to each other in a two dimensional house of cards structure (Fig. 4.1A). This delicate morphology is indicative of their authigenic origin. Elemental analysis by the energy dispersive x-ray analyzer attached to the SEM indicated that the chlorites are iron rich.

(b) Pore filling chlorite:

Chlorite infilling pore spaces is the second most common

Fig.4.1.SEM micrograph of chlorite rims around detrital grains. Muladi-1 well. 3114m. Bhuban fm. Scale bar = 12 microns.

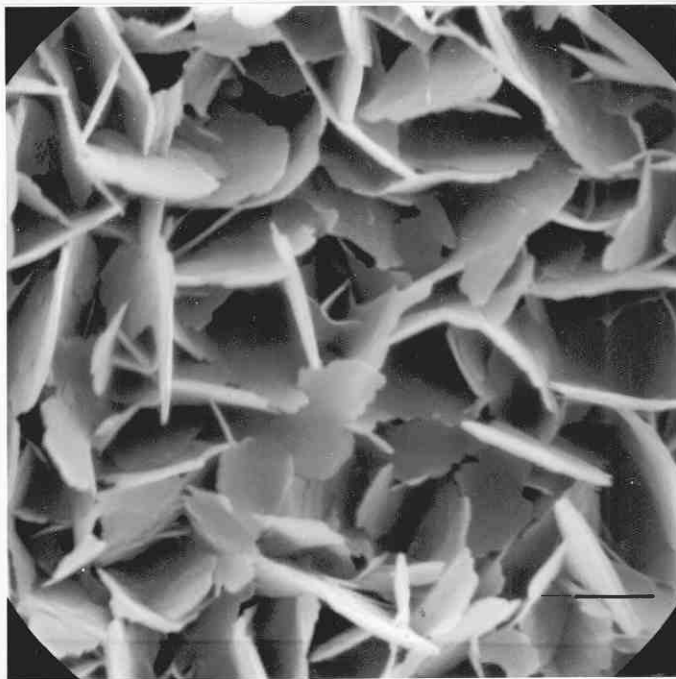


Fig.4.1A.Magnified view of the chlorite rim shown in fig.4.1. Scale bar = 1 micron.

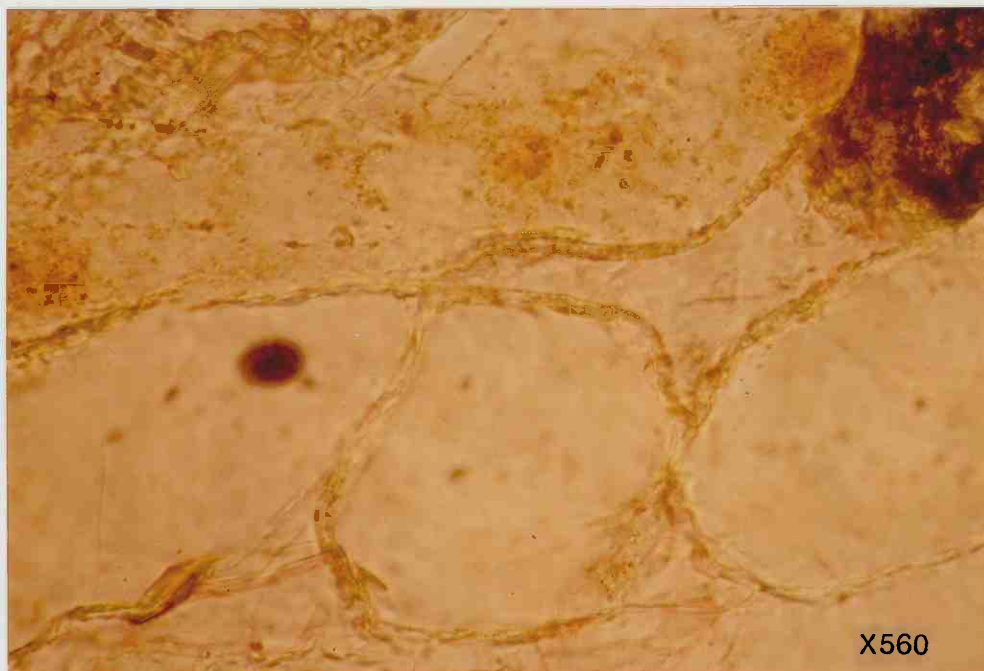


Fig.4.2. Photomicrograph of chlorite rims around detrital quartz grains. Plane light. Muladi-1 well. 3223m. Bhuban fm.

type of authigenic chlorite. It shows more than one type of morphology which are most conveniently studied in the scanning electron microscope. The most common morphology is the cluster of bladed or platy chlorite crystals. Often these clusters take the form of rosettes (Fig. 4.3). In one example, the cluster of chlorite plates was seen to grow into a long chain curving through the pores (Fig. 4.4). A second form of chlorite, though much less common, has a cabbage head-like growth which is also developed within the pore spaces. Though these different forms of pore filling chlorite generally occur individually, intimate associations of these chlorite masses with chlorite rims were frequently observed. The chlorite pore fills are usually most common where the chlorite rims are also abundant. The porefilling chlorite masses are often found resting on top of the chlorite rims and sometimes these were seen to have grown out into the pores from thick chlorite rims (Fig. 4.5). Also observed were chlorite masses infilling individual pore spaces which are thickly lined by chlorite rims on all sides.

(c) Grain replacing chlorite:

Partial replacement of detrital grains by chlorite is the least common of the authigenic chlorite types and was only rarely observed. Chlorite selectively replaces labile grains such as feldspars and lithic fragments. Feldspars have been partially replaced along the cleavage traces (Fig. 4.6). Among the rock fragments, partial replacements were noted in mica-schist, metaquartzite and chert clasts.

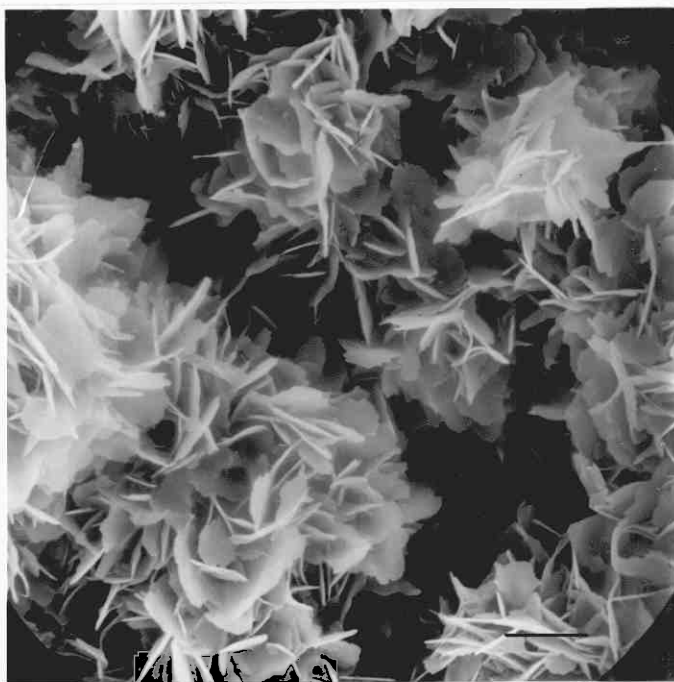


Fig.4.3.SEM micrograph of chlorite rosettes filling pore space. Muladi-1 well. 2871m. Bhuvan fm. Scale bar = 2 microns.

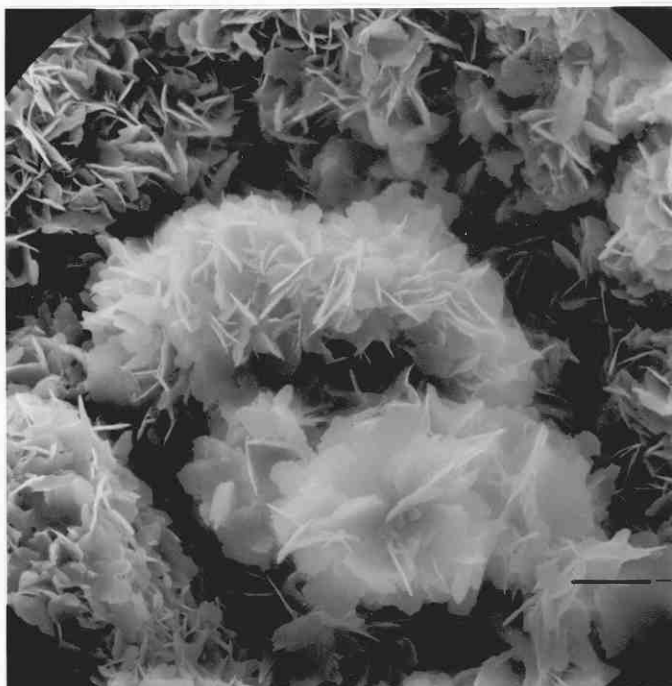


Fig.4.4.SEM micrograph of chlorite rosettes extending into the pore space. Muladi-1 well. 2871m. Bhuvan fm. Scale bar = 3 microns.

Fig.4.5.SEM micrograph of chlorite growing into pore space from thick chlorite rim.Muladi-1 well.2871m.Bhuban fm. Scale bar = 6 microns.

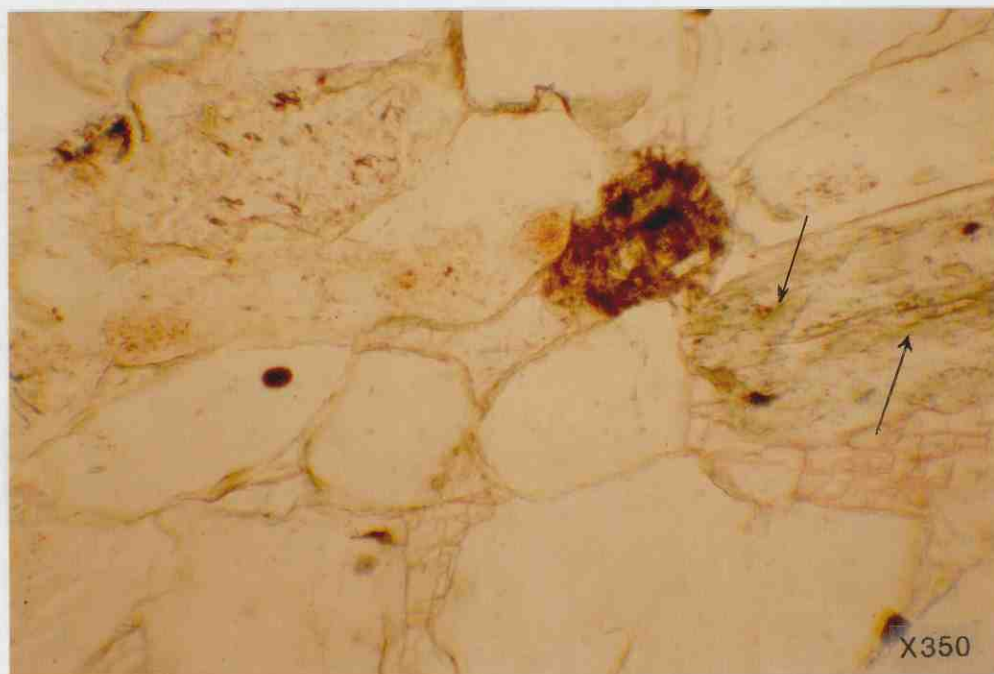


Fig.4.6.Photomicrograph of chlorite (green, arrowed) replacing detrital feldspar. Muladi-1 well. 3223m. Bhuban fm.

4.2.2. KAOLINITE:

Authigenic kaolinite is a common cement in the sandstones of the Bhuban formation in the Titas-2 well. It was also observed in some sandstones of the Bhuban Formation in the Muladi-1 and Begumganj-1 wells, but it is either absent or rare in most of the samples there. Point counting data indicates that authigenic kaolinite constitutes 0.1% to 2.0% of the total rock components in the studied samples. Two modes of occurrence have been observed. These are (a) as pore filling materials, and (b) as in situ alterations of feldspar.

(a) Pore filling kaolinite:

This is not associated with any precursor minerals and fills individual isolated pore spaces in well sorted sandstones. The pore filling kaolinites are believed to have been precipitated from interstitial solutions and are the more common of the two modes of occurrence.

In thin section the kaolinite is identified as colourless, low birefringence interstitial masses of granular microcrystalline aggregates. The aggregates are composed of kaolinite particles averaging several microns in size. The nature of kaolinite was more clearly shown from the SEM studies where it was seen as books formed of pseudo-hexagonal platelets stacked parallel to the c-axis (Fig. 4.7). These books of kaolinite platelets occur in random arrangements infilling the pore spaces partially or completely (Fig. 4.8A and 4.8B). In a few cases, occurrences of kaolinite as a delicate vermicular growth were observed in which a sequence of stacked pseudo-hexagonal platelets of kaolinite extends for some distance into the pores (Fig. 4.9). Individual plates of kaolinite crystal measures up to 1 micron in thickness while the composite lengths of individual books are

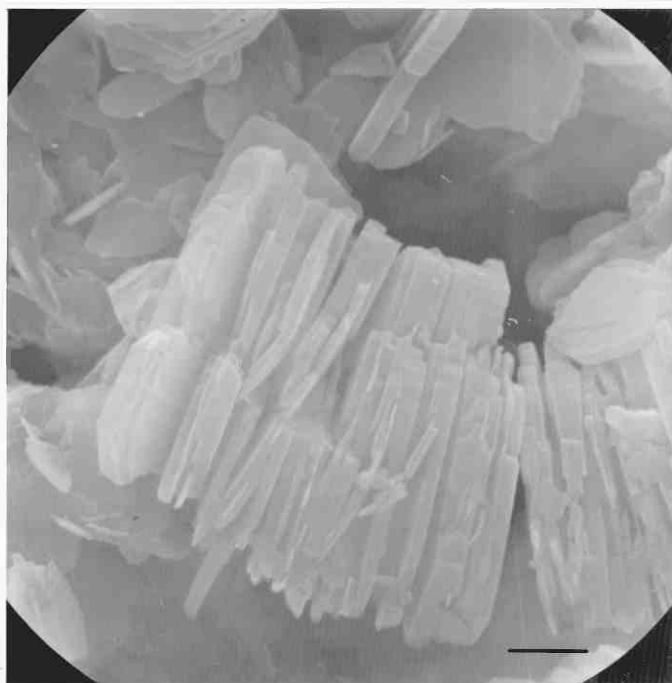


Fig.4.7.SEM micrograph of stacked platelets of Kaolinite in pore space. Titas-2 well. 2646m. Bhuban fm. Scale bar = 1 micron.

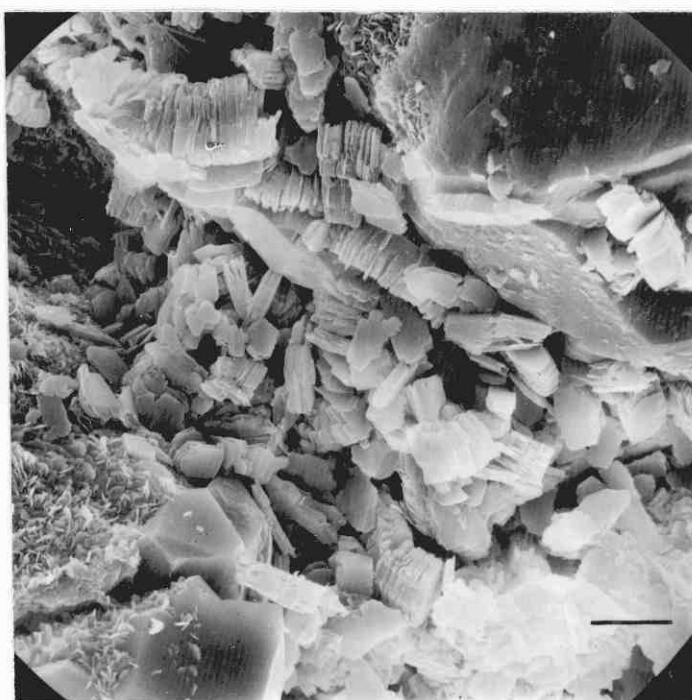


Fig.4.8A.SEM micrograph of books of kaolinite platelets filling pore spaces. Titas-2 well.2646m.Bhuban fm. Scale bar = 8 microns.

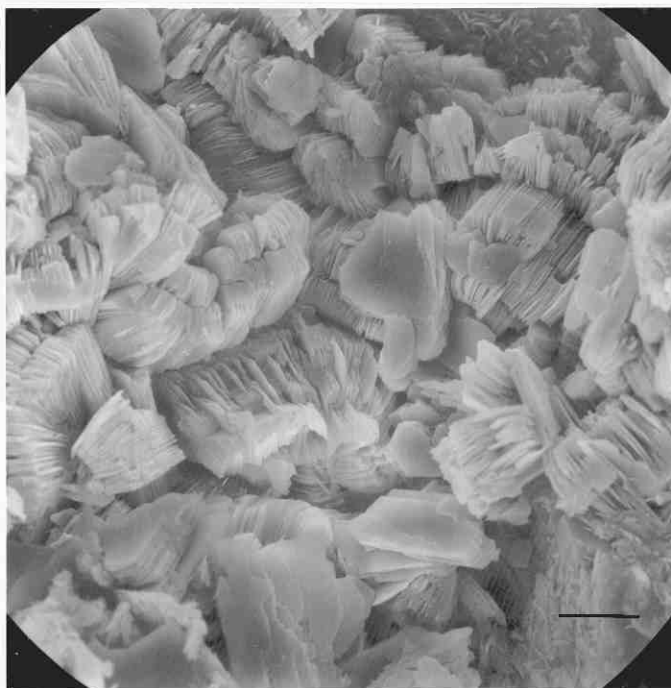


Fig.4.8B.SEM micrograph of kaolinite books chocking pore space. Titas-2 well. 2690m. Bhuban fm. Scale bar = 11 microns.

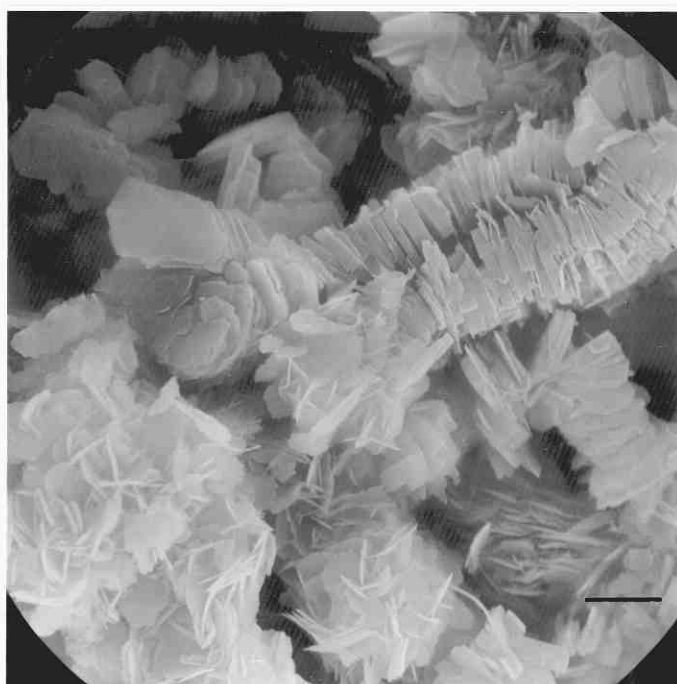


Fig.4.9.SEM micrograph of the vermicular growth of kaolinite into the pore space. Begumganj-1 well. 3642m.Bhuban fm. Scale bar=4 microns.

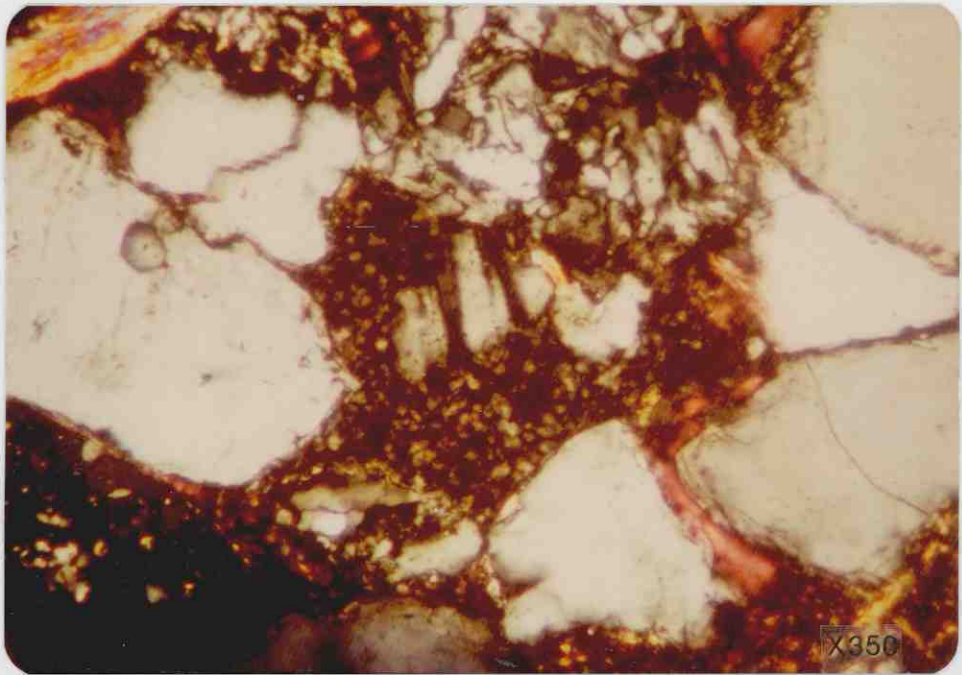


Fig.4.10. Photomicrograph of insitu alteration of feldspar to kaolinite. Crossed nicols. Titas-2 well. 2668m. Bhuban fm.

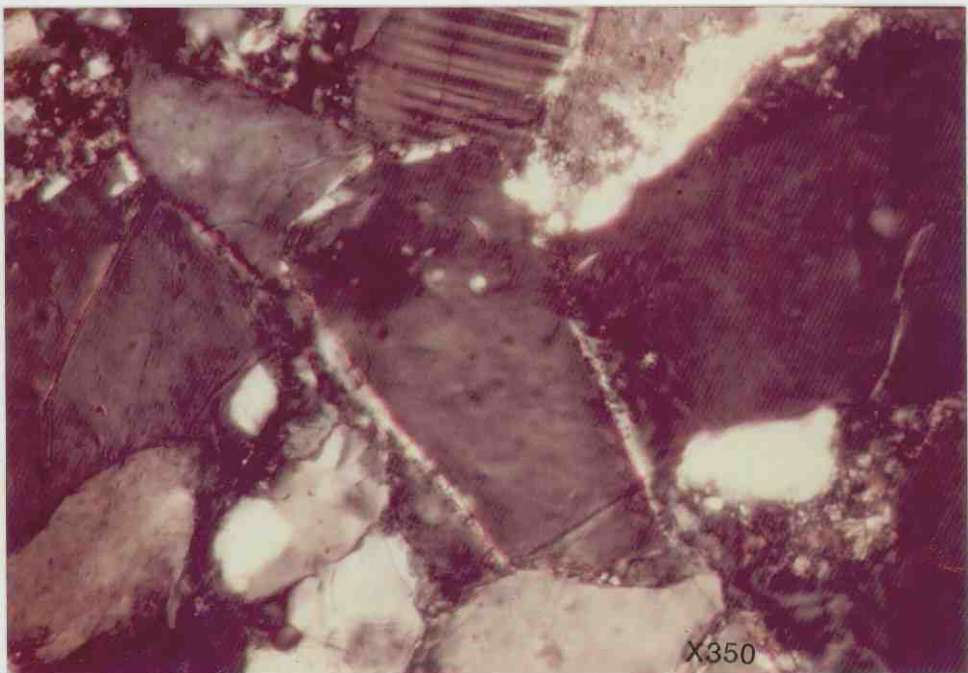


Fig.4.11. Photomicrograph of high birefringent illite coating on detrital grain. Crossed nicols. Begumganj-2 well. 3325m. Bhuban fm.

generally 10 microns to 30 microns. The most important criteria for defining the authigenic or secondary origin of the pore filling kaolinite are: (i) Its euhedral crystalline habit and delicate form. Extensive transportation would definitely have destroyed these delicate habits. (ii) Its occurrence in isolated and scattered pore spaces. A detrital origin would have been expected to distribute the clay evenly throughout the sandstone, and (iii) The monomineralic nature of the kaolinite aggregates.

(b) In situ alteration of feldspars:

This type of occurrence is represented by kaolinite masses enclosing relic feldspar grain (Fig. 4.10) or by kaolinites along the edges of partially decomposed feldspar, both directly implying its origin as the in situ alteration product of feldspars. These were less commonly observed than the pore filling kaolinites, but show similar optical and morphological characteristics under the petrographic and scanning electron microscopes.

4.2.3 ILLITE

Authigenic illite is a common but minor diagenetic clay mineral in the Neogene sandstones of the Bengal Basin. It occurs as thin coatings on detrital grains over the entire depth range sampled. In thin section, it is readily recognized by its high birefringence (Fig. 4.11). The illite coatings are generally discontinuous over the detrital grains. Under the SEM, the authigenic illite appears as tangential sheets or large flaky crystals draped over the grain surfaces or chlorite rims (Fig. 4.12). The thread like or hairy growths of illite laths that have often been described in the literature to form meshworks and bridge the gap between detrital grains and destroy permeability (Guven et al.1980, Wilson and

Pittman, 1977, Hancock, 1978), have not been seen in any of the studied samples.

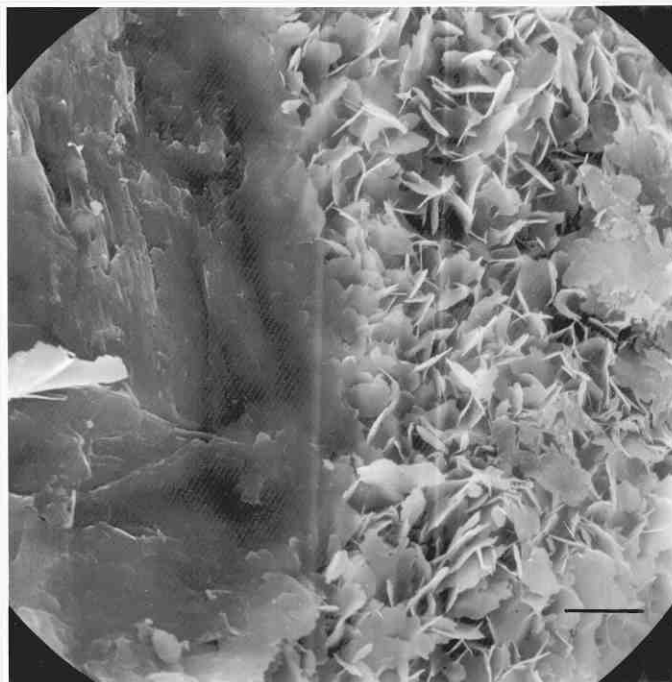


Fig.4.12. Tangential sheet of authigenic illite draping over chlorite rim. Begumganj-2 well. 3325m. Bhuban fm. Scale bar = 6 microns.

4.3 AUTHIGENIC NON-CLAY MINERALS:

4.3.1 QUARTZ

Authigenic quartz, in the form of quartz overgrowths, is a common diagenetic mineral in the more deeply buried sandstones. It is absent or very rare in samples of the Tipam group and the Bokabil formation. In the underlying Bhuvan formation quartz overgrowth is common in most of the sandstones and is abundant to very abundant in many of them where it reduces porosity significantly. A comprehensive study of the quartz overgrowths was made using thin section petrography, scanning electron microscopy and cathodoluminescence techniques.

The recognition of quartz overgrowths in thin section is dependent on the presence of a "dust line" between the detrital grain nuclei and the overgrowths, or by the presence of distinctive well defined crystal faces on the overgrowth. The "dust lines" are commonly formed by a thin film of illite clay which could be recognized by its distinctive high birefringence (Fig.4.13). Apart from clay films or coatings, various impurities between the overgrowths and the nuclei often form the "dust lines". Occasionally distinct euhedral crystal faces, growing into the pore spaces, present unmistakable evidence of quartz overgrowth (Fig. 4.14). These growths of euhedral quartz could only be formed when sufficient space was available for the crystal to develop freely into the pore. When neither of these two criteria is present, it is not possible to distinguish the overgrowth from the detrital grain using standard petrographic thin section analysis. For this reason many of the overgrowths remained unnoticed when examined in thin section. As a result, estimate of the percentage of quartz overgrowth could not be made by point counting whereas most of the other important cementing materials of the sandstones have been quantified

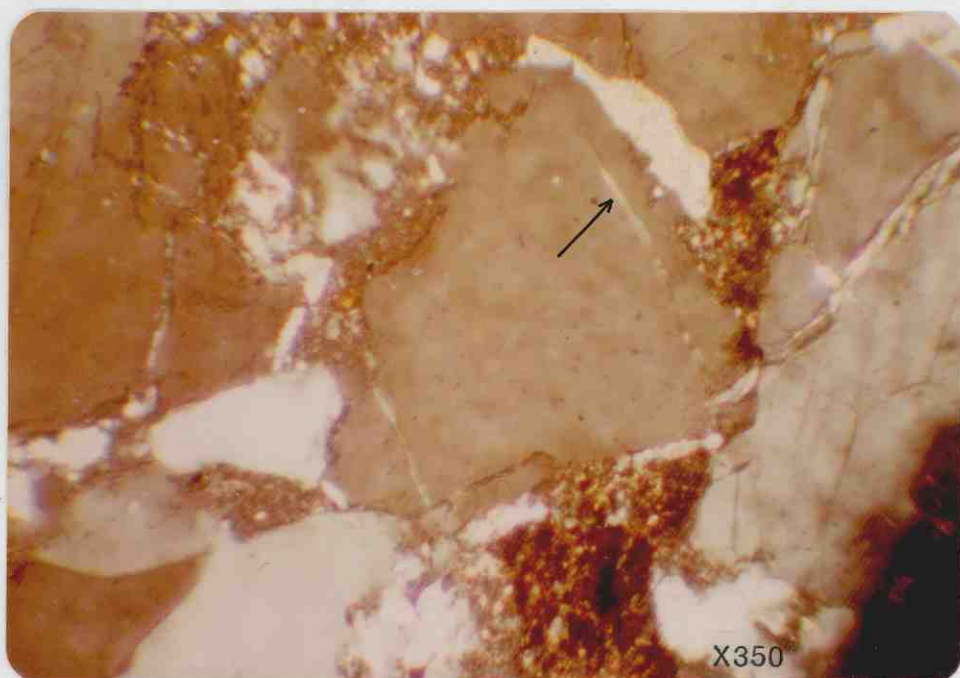


Fig.4.13. Photomicrograph of dust line of illite clay (arrow) between detrital grain and quartz overgrowth. Crossed nicols. Begumganj-1 well. 2758m. Bhuban fm.

Fig.4.14. Photomicrograph of euhedral quartz overgrowth crystal faces enveloped by late calcite cement. Crossed nicols. Muladi-2 well. 4491m. Bhuban fm.

by this method. Modal analyses indicate that the percentages of detrital quartz in samples where quartz overgrowths are absent or not significant (as seen in the SEM) range from about 51% to 66% with a mean value of 60%. However significant variations from this average are found (up to 76%) in samples where quartz overgrowths are very abundant (as seen in SEM). These high values are possibly due to inability to adequately distinguish overgrowths from detrital quartz, and incorporating the overgrowths into the estimates of detrital quartz grains. Assuming the detrital quartz content of these samples to be similar to the highest obtained for the previous ones (66%), the overgrowths may account for as much as 10% of the total quartz in these samples.

Scanning electron microscopy gives the clearest evidence of the nature of the quartz overgrowths. Under the SEM two types of quartz overgrowths can be distinguished: incipient and terminated. Incipient overgrowths describe the small isolated and incomplete growth of quartz crystal faces on detrital quartz grains (Fig. 4.15). These are the result of overlapping and merging of smaller irregular overgrowth sub-units (Pittman 1972). The overlapping nature is indicated where incomplete crystal merging results in holes in the crystal faces of the overgrowths (Fig. 4.16). If the growth of the authigenic quartz had continued, these holes would have been infilled eventually. There often remains an unfilled space between the advancing incipient overgrowths and the quartz grains (Fig. 4.17). These voids may represent the "dust line" with or without being infilled with impurities (Pittman 1972).

A more common and advanced growth of authigenic quartz is represented by the terminated type of overgrowths. Here the overgrowths have developed complete and well developed crystal faces (Fig. 4.18 - 4.22). They can commonly

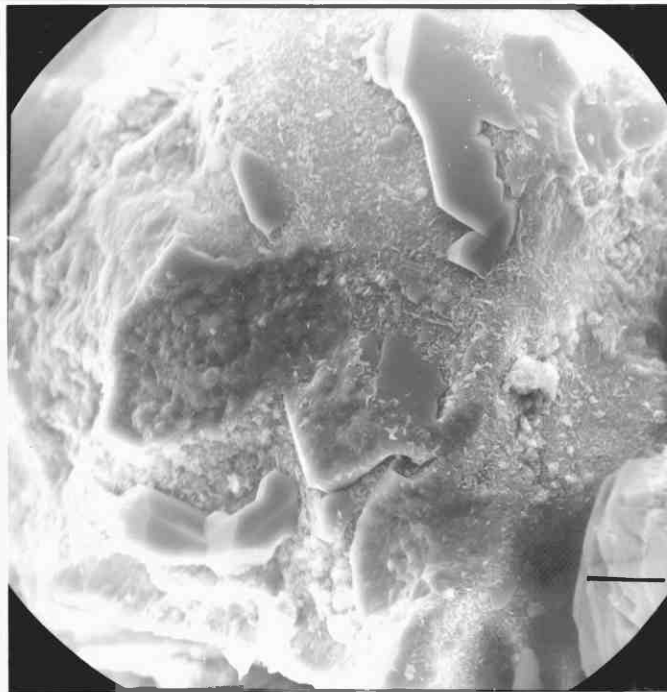


Fig.4.15.SEM micrograph of isolated and incipient quartz overgrowth. Muladi-1 well. 2216m. Bokabil fm. Scale bar= 20 microns.

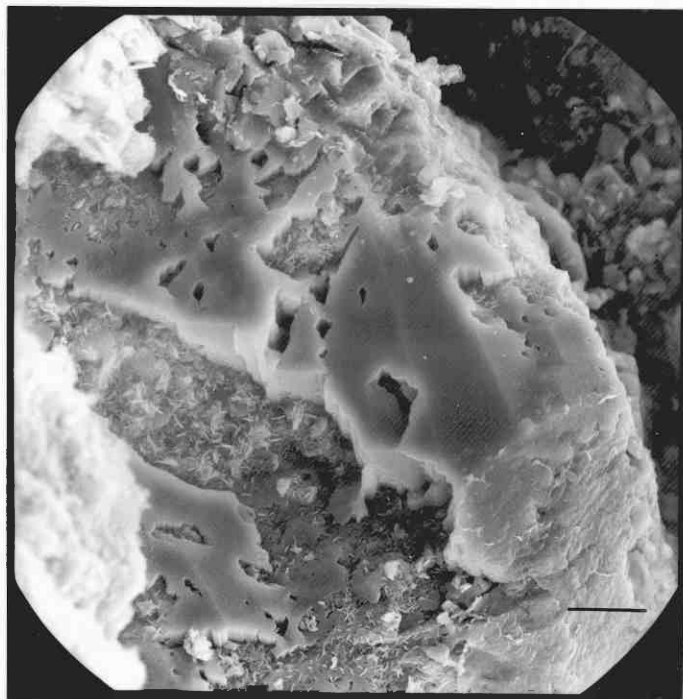


Fig.4.16.SEM micrograph of holes in the quartz overgrowth crystal faces resulting from incomplete mergence of incipient quartz overgrowths. Muladi-1 well. 3294m. Bhuban fm. Scale bar = 15 microns.

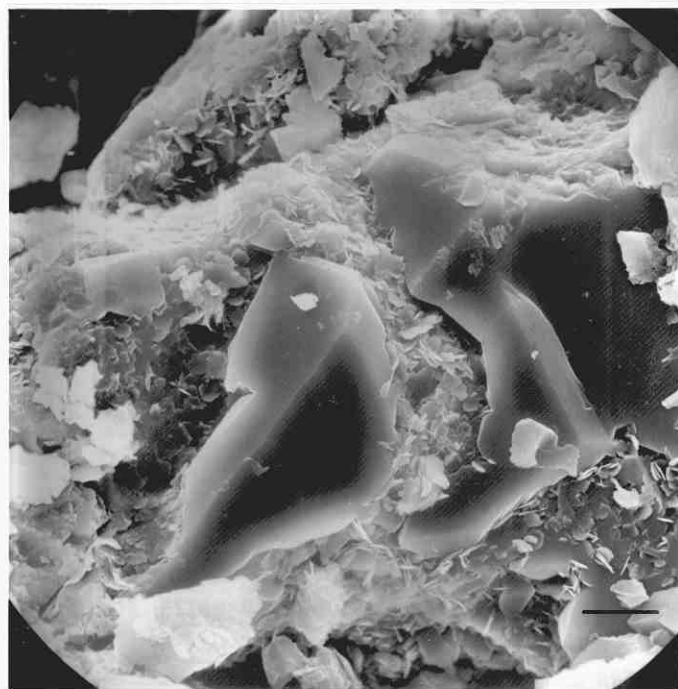


Fig.4.17. SEM micrograph of incipient quartz overgrowth. Note the gap between overgrowth and the detrital grain. Muladi-1 well. 3223m. Bhuban fm. Scale bar=8 microns.

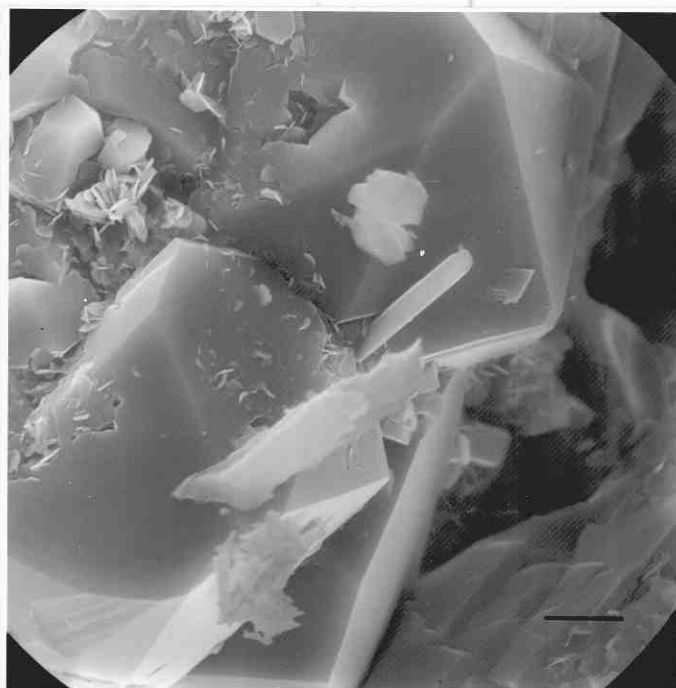


Fig.4.18. SEM micrograph of euhedral crystal faces of terminated quartz overgrowth. Titas-2 well. 2967m. Bhuban fm. Scale bar = 5 microns.

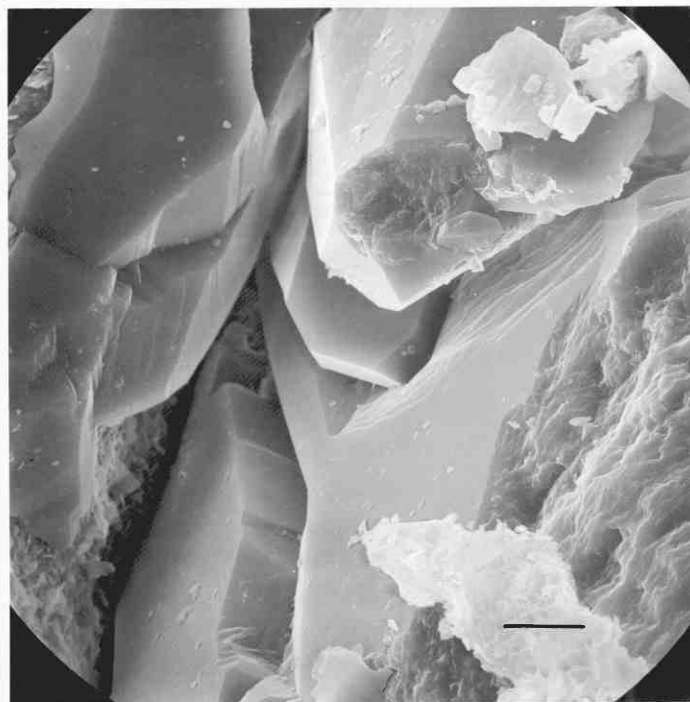


Fig.4.19. SEM micrograph of terminated quartz overgrowths with euhedral crystal faces. Titas-2 well. 2646m. Bhuban fm. Scale bar = 8 microns.

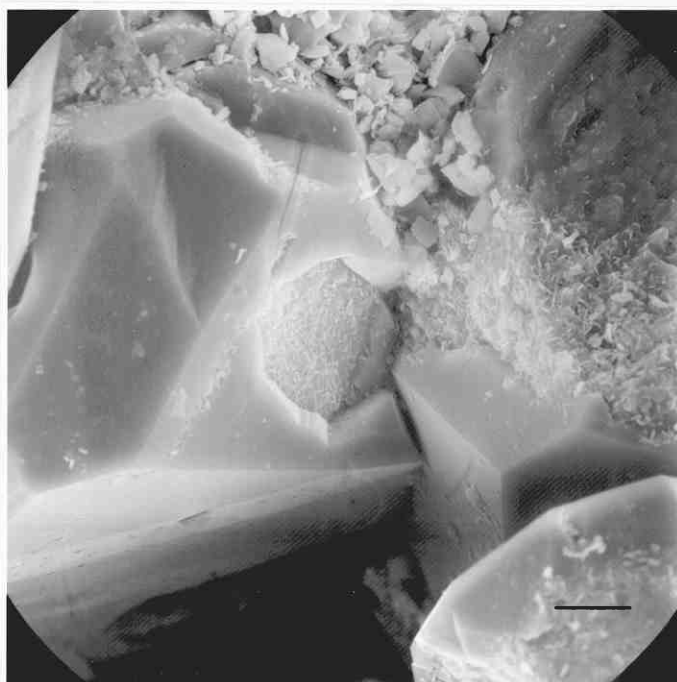


Fig.4.20. SEM micrograph of terminated quartz overgrowths Muladi-1 well. 3294m. Bhuban fm. Scale bar = 14 microns.

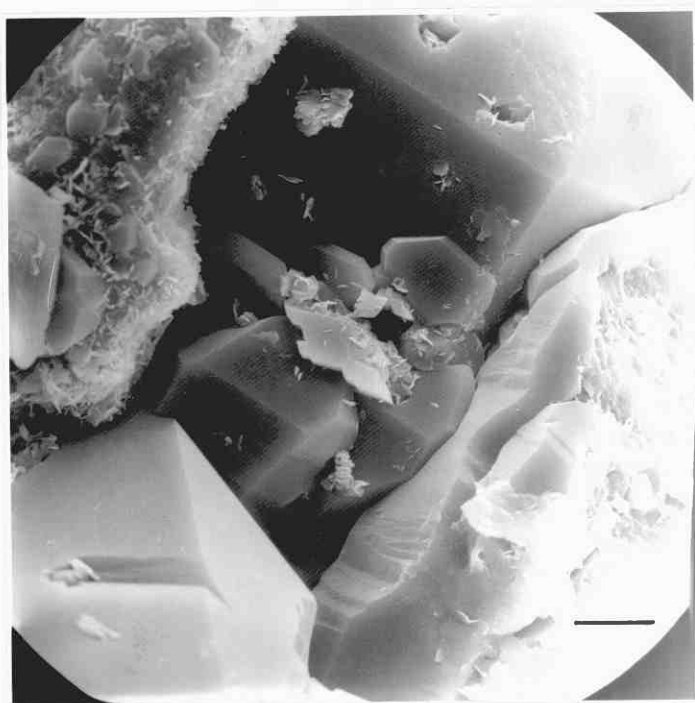


Fig.4.21.SEM miccograph of interlocking quartz over-
growths. Begumganj-1 well. 2758m. Bhuban fm
Scale bar = 9 microns.

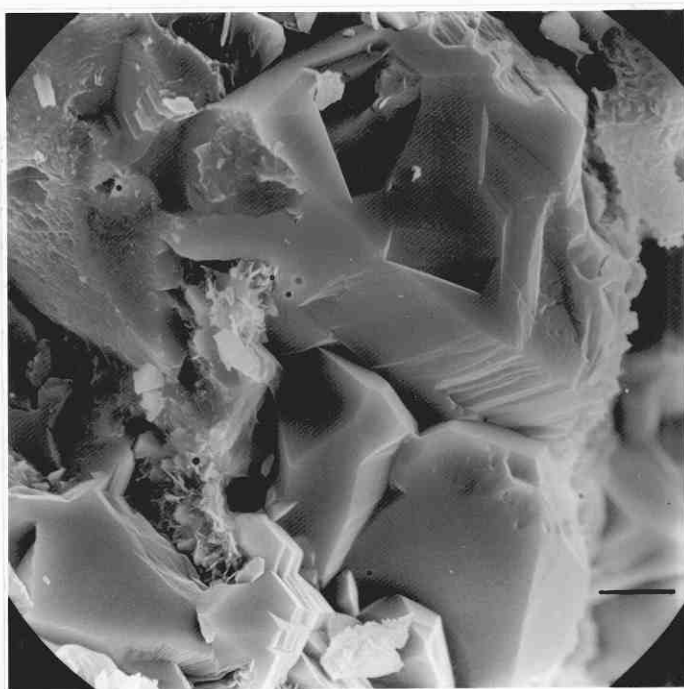


Fig.4.22.SEM micrograph of interlocking quartz over-
growths. Begumganj-1 well. 2680m. Bhuban fm.
Scale bar = 8 microns.

envelope the detrital grain surfaces completely. Depending upon the availability of space during growth, terminated overgrowth crystals can take the form of perfect euhedra. Such euhedra grow into the open pore space and often form hexagonal prisms with rhombohedral pinacoids. However, more commonly, due to the lack of available space required for such free growth, the crystal faces interact with one another to form compromise boundaries. This results in an interlocking texture of the quartz and in such cases porosity is reduced significantly. Such compromise boundaries between quartz overgrowths are illustrated in Fig. 4.21 and 4.22. These types of boundaries often resemble sutured contacts and in the absence of "dust lines" could be mistaken as grain to grain pressure solution in thin section. This problem, however, could be resolved by cathodoluminescence studies. Cathodoluminescence technique was useful in distinguishing between quartz overgrowth and detrital quartz, because of the different luminescent characteristics of the two. This enables one to distinguish between interlocking quartz overgrowth and pressure solution, both of which can be represented by sutured contacts. In Figure 4.23 and 4.24, interpenetrating grain contact which result from pressure solution, are marked 'P' and those which result from interlocking quartz overgrowth on compromise boundaries between overgrowths and detrital grains are marked 'O'. Such distinction is possible because of the darker tone of the quartz overgrowth in the print due to its differing luminescence character compared to the detrital quartz grain.

Pittman (1972) critically reviewed the development of the quartz overgrowths and indicated that development proceeds from initial isolated incipient types to the advanced terminated types by the gradual overlapping and merging of former smaller units into single large crystals, provided suitable physico-chemical conditions prevailed. In the

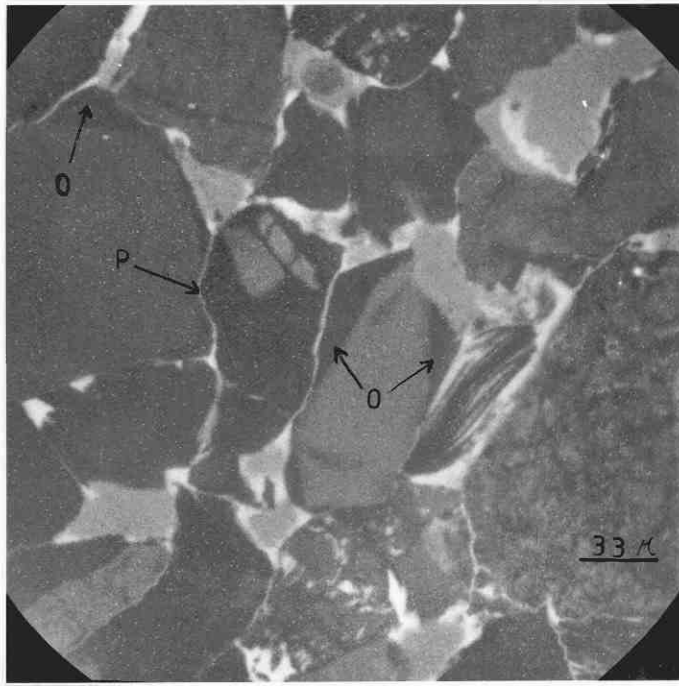


Fig.4.23.Luminescence view of sutured and concavo-convex grain contacts.Quartz overgrowths are distinctly darker than the detrital quartz. O=overgrowth, P=Pressure solution.Muladi-1 well.Bhuban fm.

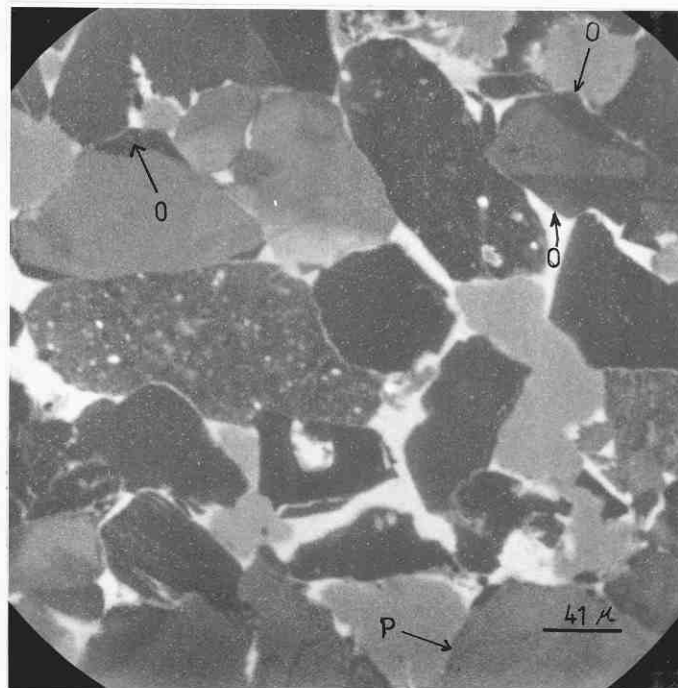


Fig.4.24.Luminescence view of sutured and concavo-covvex grain contacts.O=Overgrowth. P=Pressure solution. Muladi-1 well. Bhuban fm.

present study also such gradual development is indicated by the presence of different stages of crystal growths, ranging from incipient crystal faces covering isolated parts of detrital grains, through incomplete merging of these smaller units, to the terminated overgrowths completely enveloping the detrital grains. An inverse relationship has been observed between the growth of quartz overgrowth and that of the chlorite rims (Table 4.1). Thick chlorite rims tend to inhibit quartz overgrowths. Advanced, terminated and interlocking quartz overgrowths are only developed where chlorite rims are absent or very sparse (sample B1-16, B1-17 for example). Thin and discontinuous chlorite rims generally allow incomplete and incipient quartz overgrowth to develop. In discontinuously rimmed grains, quartz overgrowth could nucleate in the gaps, or the exposed part of the grain, and could then engulf those parts of the grains covered by the thin chlorite rims. Thick and continuous chlorite rims totally inhibit quartz overgrowth development (sample M1-34B and M1-36). Similar relationships between quartz overgrowth and chlorite rim development have also been reported by others (Pittman and Lumsden 1968; Heald and Larese 1974; Thomson 1979).

TABLE 4.1.: RELATIONSHIPS BETWEEN THE DEVELOPMENT OF QUARTZ
OVERGROWTH AND CHLORITE RIMS IN NEOGENE
SANDSTONES OF THE BENGAL BASIN.

Sample no	Depth (m)	Unit	Quartz overgrowth		Chlorite rims
M1-3	500	T I P A M	Nil		Thin & discontinuous
M1-4	602		Nil		Thin & discontinuous
M1-11	1137		Nil		Thin & discontinuous
M1-20	1700	B O K A B I L	Nil		Sparse
B1-14	2139		Very rare		Generally thick and continuous
M1-28	2216		Rare		Thin & discontinuous
B1-16	2680	B H U B A N	Abundant	C E M E N T A T I O N O F S I L I C A Z O N E	Very sparse
B1-17	2758		Abundant		Very sparse
M1-34B	2871		Nil		Generally thick and continuous
B1-19	2999		Abundant		Very sparse
M1-36	3114		Nil		Generally thick and continuous
M1-37B	3223		Very common		Thin & discontinuous
M1-38	3294		Very common		Thin & discontinuous
B1-26A	3642		Common		Thin & discontinuous
M1-49	4458		Abundant		Sparse
M2-3	4491		Very common		Thin & discontinuous

4.3.2 FERROAN CALCITE:

Ferroan calcite is a very common diagenetic mineral in the sandstones of the Bokabil and Bhuvan Formations. The ferroan nature of the calcite is indicated by staining with potassium ferricyanide with which it stains blue. The calcite stains pink with Alizarin red S. Under the petrographic microscope the calcite cement is colourless, and is characterized by rhombohedral cleavage, variable relief with rotation of the stage, and high order pink and green birefringence colours.

The ferroan calcite occurs in the following textural varieties in the studied sandstones:

- (I) Isolated pore filling and replacive cements
- (II) Poikilotopic cements

(I) Isolated pore filling and replacive Fe-calcite cement:

Fe-calcite cements infilling isolated pore spaces and also replacing detrital grains are the most common types of carbonate cements in the sandstone studied (Fig. 4.25). The first occurrences of these cements were observed in samples from depths below about 1700m and their presence continue down to the deepest samples studied from 4491 m. Point counting data indicate that the cements in general constitute about 0.3% to 7.0% of the rock volumes infilling isolated pore spaces and also marginally replacing few detrital grains. In some samples, however, grain replacement is very widely developed, thus increasing the amount of Fe-calcite in the samples to 17%-35% of the total rock components. In such cases, almost total replacement of grains by Fe-calcite is occasionally found, leaving only a relic skeletal grain (Fig. 4.26). Fe-calcite replaces quartz grains most extensively but replacement of feldspars and rock

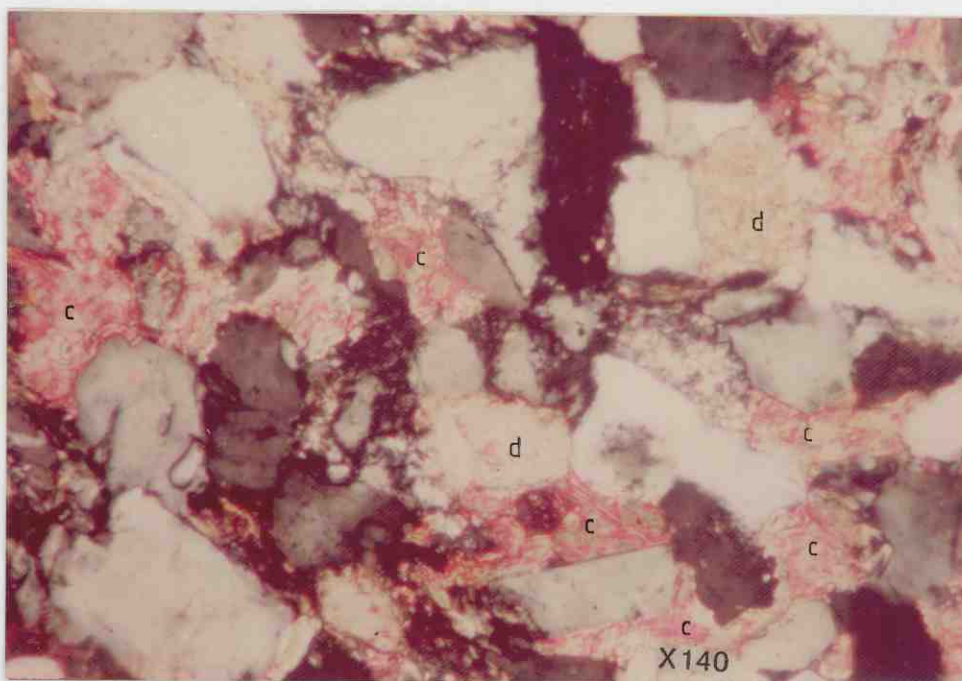


Fig.4.25. Photomicrograph of late calcite (c) and dolomite (d) filling pore and replacing grains. Sample stained with Alizarin red S. Begumganj-2 well. 3476m. Bhuvan fm.

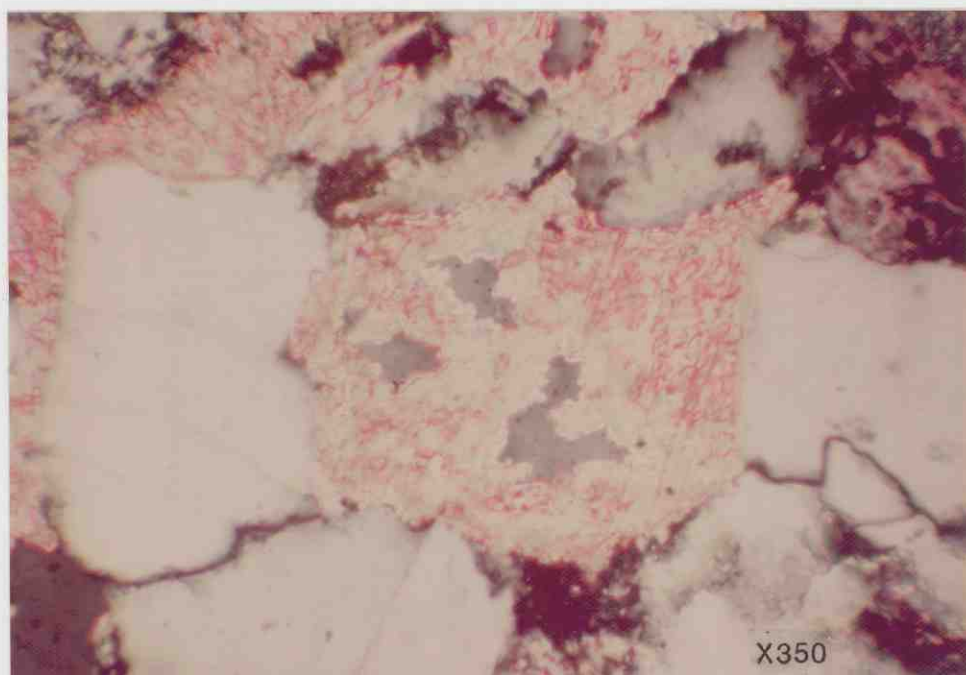


Fig.4.26. Photomicrograph of advance stage of grain replacement by calcite leaving relic quartz grain within replacing calcite. Muladi-2 well. 4491m. Bhuvan fm.

fragments is also common. Replacive calcite occasionally cross cuts grain contacts, including sutured contacts in a few samples.

(II) Poikilotopic ferroan calcite:

Poikilotopic ferroan calcite is present only in a very few samples - only 5% of the total number of sandstones studied. However, where it occurs, it occupies the total pore space reducing the porosity and permeability to zero. Point counting data indicate that poikilotopic calcite cement constitute 35%-50% of the total rock components. These cements are composed of interlocking mosaics of individual euhedral calcite crystals ranging in size from 0.5 cm to more than 1.0 cm. The individual nature of the ferroan calcite crystals could be noted by the different positions of their extinction with respect to adjacent crystals. Within the mosaic of calcite cements, the detrital grains are embedded with few or no grain contacts. The detrital components essentially occur as floating grains in the calcite cement and are separated beyond the normal grain supported framework which would have been established during deposition (Fig. 4.27). It has been suggested that crystallization of such poikilotopic cements involves expansion of intergranular space and consequent rearrangements in the packing of the detrital grains, increasing the separation between them (Dapples 1971). There is evidence of minor corrosion along the outer margins of some detrital grains.

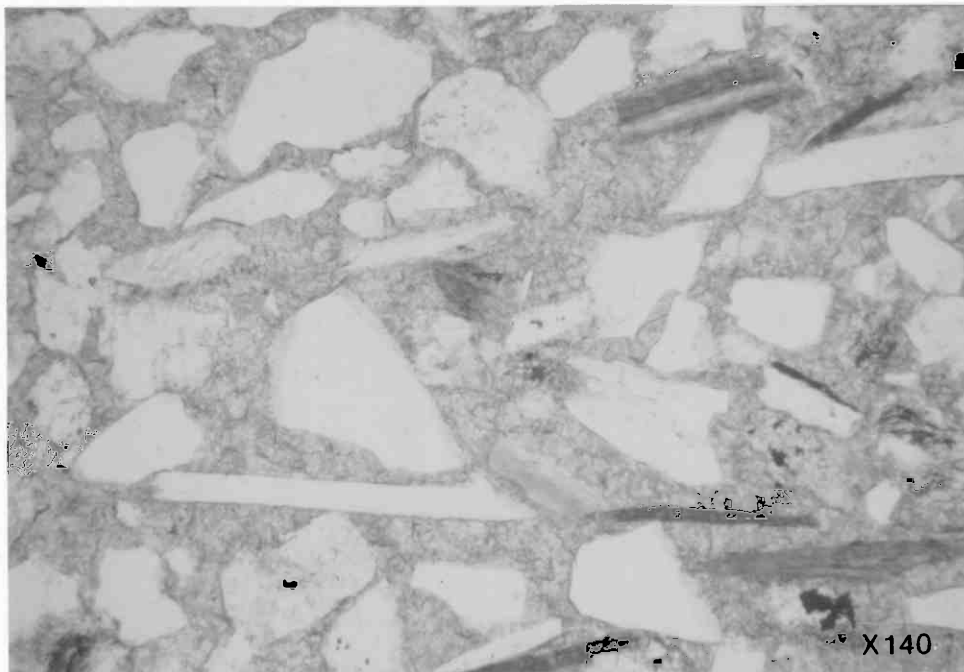


Fig.4.27. Photomicrograph of early poikilotopic calcite cement with "floating" grains. Sample stained with Alizarin red S. Plain light. Muladi-1 well. 2212m. Bokabil fm.

4.3.3 DOLOMITE

Dolomite occurs as a very minor authigenic component in some samples and is mostly associated with isolated pore filling Fe-calcite cement. It was distinguished from the latter in the studied samples by staining with Alizarin Red S., which stain calcite pink/red but does not affect dolomite. Dolomite commonly occurs along the outer edges of the calcite pore filling cement, indicating a replacive origin. However, individual pores filled only with dolomite have also been found (Fig. 4.25); this could be the result of the total replacement of calcite by dolomite, or it could be due to direct precipitation of dolomite in the pore spaces. The latter seems unlikely as the majority of the dolomite appears to be replacive after Fe-calcite.

4.3.4 SIDERITE

Authigenic siderite is a commonly occurring diagenetic mineral in the Neogene sandstones of the Bengal Basin. It is present in almost all of the samples examined, although only in very minor amounts in most of them. Point counting data indicates that it constitutes only trace to about 1.0% of the total rock components in the studied samples, with a mean value of 0.4%. The siderite occurs as minute euhedral to subhedral rhombic crystals, either singly or more commonly as aggregates, partially filling pores (Fig. 4.28). Less commonly, siderite replaces grains varying from only marginally to almost total replacement in rare cases. The mineral is colourless to pale brown and is characterized by high relief and extreme birefringence. The siderite is often masked by yellow brown to dark brown staining due to partial alteration to iron oxide.

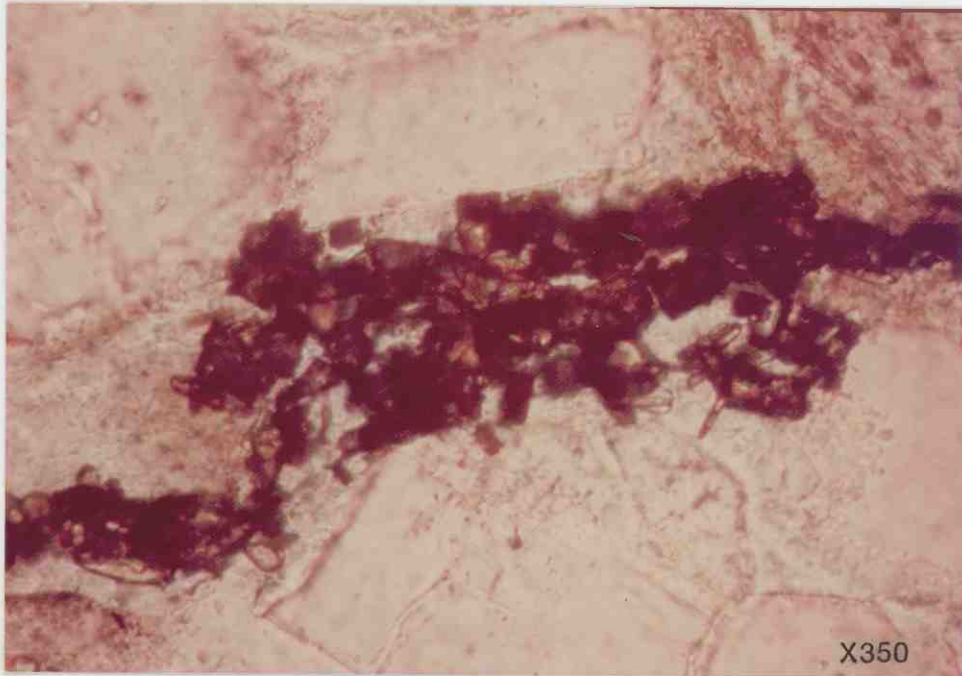


Fig.4.28. Photomicrograph of euhedral and subhedral authigenic siderite crystals filling pore space. Titas-2 well. 2753m. Bhuban fm. well. 2212m. Bokabil fm.

4.3.5. FELDSPAR

Authigenic feldspar in the form of overgrowth occurs rarely in the Neogene sandstone of the Bengal Basin. An initial stage of formation of feldspar overgrowth is represented by small oriented rhombohedral crystals on detrital feldspar (Fig.4.29). The term "Adularia habit" has been used to describe the morphology of such incipient feldspar overgrowth (Ali and Turner 1982). Merging and growing of these rhombic crystals may lead to the formation of larger overgrowth crystals (Ali and Turner 1982). Feldspar overgrowth partially enveloping detrital grain represents a more advanced stage of development and was seen on very few occasions in the studied sandstones.

4.3.6. PYRITE

Trace amount of pyrite are present in few samples as scattered cubes and masses. Pyrite was also observed in rare cases within the chambers of fossil foraminifera. The mode of occurrence of pyrite indicates an authigenic origin.

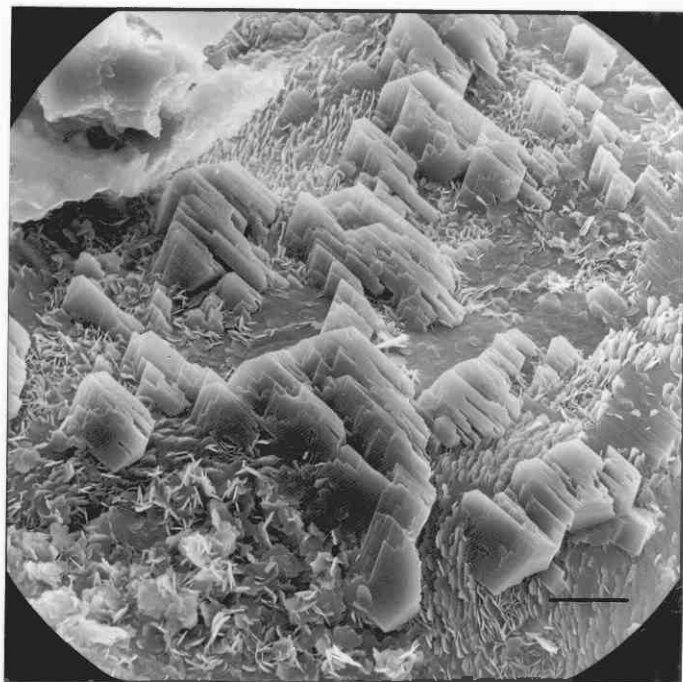


Fig.4.29.SEM micrograph of incipient feldspar overgrowth.
Begumganj-1 well. Bhuban fm. 3325 m. Scale
bar= 5 microns.

CHAPTER 5

DIAGENETIC PROCESSES IN SANDSTONE

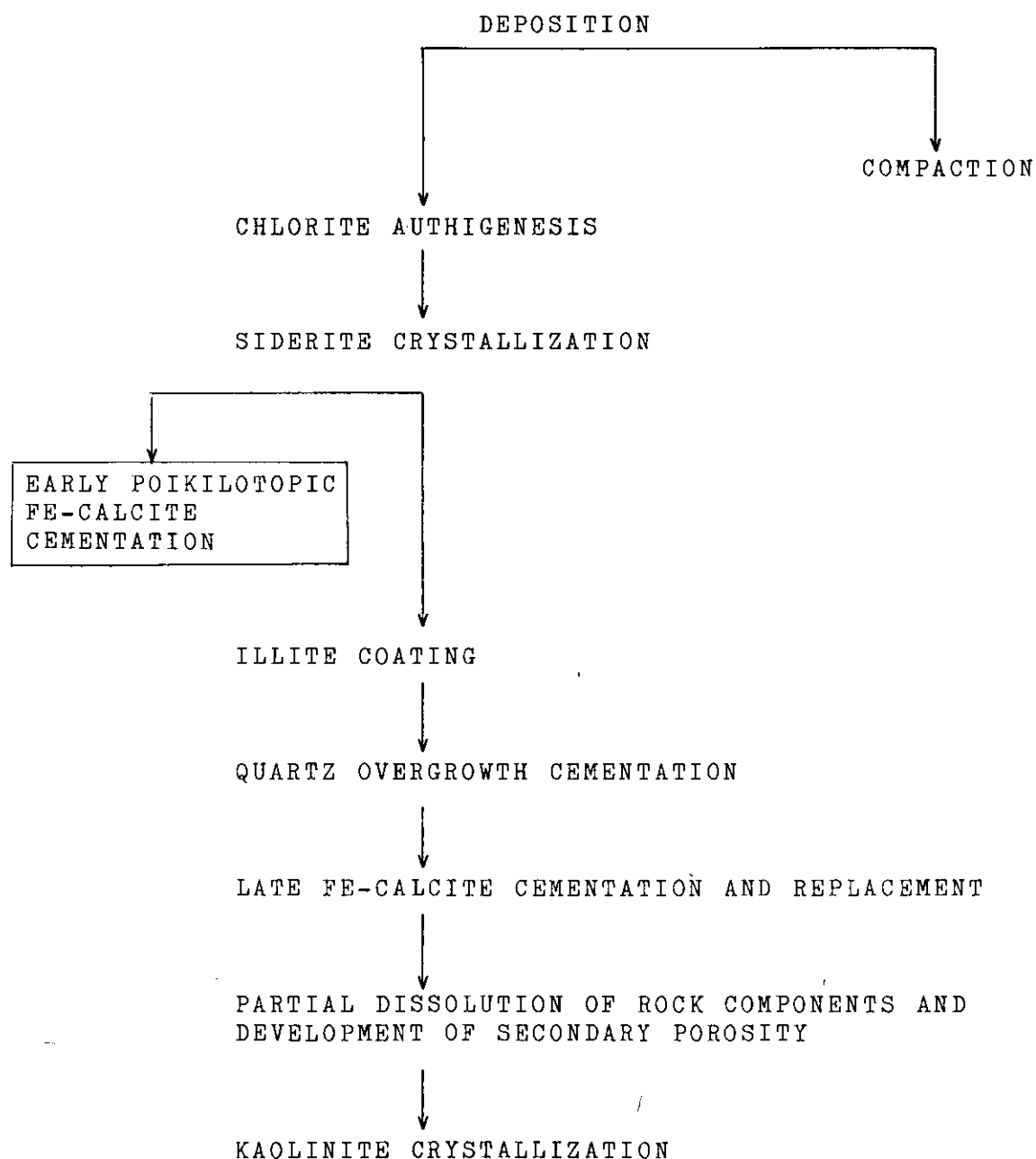
5.1 INTRODUCTION

Physical, physico-chemical and chemical processes have all been involved in the diagenesis of the Neogene sandstones in the Bengal Basin. The Neogene sequences, represent a huge thickness of deltaic clastic sediments deposited in a rapidly subsiding basin. The sediments of the studied area have never been significantly uplifted since deposition and are presently near or at their maximum depth of burial. The sequence therefore provides an ideal opportunity to study the processes of burial diagenesis in a sand-shale section that is active today. The following discussion of the various diagenetic processes affecting the sandstones is mainly based on the study of diagenetic phases described in detail in the previous chapter.

5.2 PARAGENETIC SEQUENCE:

The paragenetic sequence of the major diagenetic events can be deduced from the mutual textural relationships between diagenetic minerals and the depth at which diagenetic features first appear in the drilled section (Fuchtbauer 1981), and is shown in Fig.5.1. In addition to the above, feldspar overgrowth formation and pyrite crystallisation are the two diagenetic events which have taken place in the samples but whose position in the paragenetic sequence could not be established because of the insignificant amount and limited distributions of these two minerals.

FIG.5.1 : PARAGENETIC SEQUENCES (WITH OVERLAPPING OF SOME EVENTS POSSIBLE) OF DIAGENESIS IN THE NEOGENE SANDSTONES OF THE BENGAL BASIN.



5.3 COMPACTION OF SEDIMENTS

Compaction of sediments by overburden pressure started as soon as the sediments were buried below the sediment/water interface. Unless early cementation arrested the diagenetic process, the sediments continued to undergo compaction as they were being buried deeper and deeper. Compaction was the only process active in the consolidation of sandstone of the Tipam group where no cementation has taken place. It remains an important process in the lithification of underlying Surma group sandstones along with other chemical cementation processes which were active in this unit. Evidence of compaction in Neogene sandstones includes physical and physico-chemical processes and these are discussed in detail below.

5.3.1 Physical Processes:

(a) Packing adjustment of grains:

Packing re-arrangement of detrital grains started with initial burial and was the first response of physical compaction to the overburden stress. The processes involved grains slipping past each other and grain rotation. These were only important in the early stages of burial, as seen in the Tipam group and top of the Bokabil formation of the Surma group. These processes resulted in an increased closed packing of the grains and are evidenced by the change in the nature of grain contacts; samples at shallower depth with predominantly point contacts between grains gradually change to ones with predominantly long contacts between grains with increasing depth of burial. Point contact and long contacts thereafter continue to decrease in number (below 2216 m of burial depth) as processes other than packing readjustment became important resulting in an increase in concavo-convex and suture type contacts between grains.

(b) Ductile grain deformation:

Plastic deformation of ductile or labile grains has been observed in samples below about 2140 m (in Surma group sandstone) and the degree of intensity of such deformation increases with increasing depth of burial. The ductile grains are the soft lithic fragments and include shale, claystone, mica schist, phyllite, and slates in the studied samples. The plastic deformation occurs as the labile grains are squeezed by adjacent rigid grains in response to the overburden pressure. Such squeezing often leads to the flowage of soft lithics into the adjacent pore spaces (Fig. 5.2 and 5.3). If this process has proceeded to an advanced stage, the soft grains lose their original shape and may take the form of an interstitial material resembling matrix. This is referred to as the pseudo-matrix and the following criteria have been used for its recognition and distinction from true matrix (Dickinson 1970), (Fig. 5.4).

- (a) Flame-like wisps of crushed lithic grains extend into the narrowing orifices between undeformed rigid grains.
- (b) The internal fabrics of the lithic fragments are deformed by plastic flow and commonly conform to the margins of surrounding rigid grains.
- (c) Large "matrix" filled "gaps" between the framework grains suggest a pseudo-matrix.

The amount of porosity lost by the deformation of soft lithic grains depends on their abundance and amount of overburden. McBride (1978) indicated that it is common for sandstones to lose 5% to 15% of their porosity in this way. Hayes et al. (1976) reported that grain squeezing alone has reduced the porosity to nearly zero in the



Fig.5.2. Photomicrograph of pseudomatrix resulting from plastic deformation and flowage of soft lithic grain. Plain light. Muladi-2 well. 4491m. Bhuban fm.

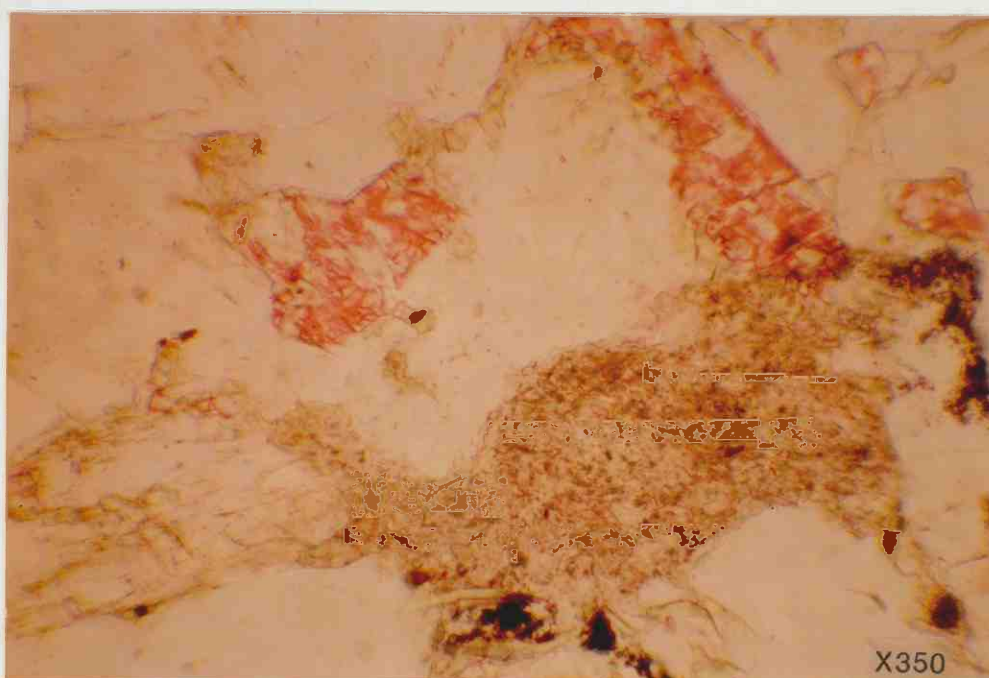


Fig.5.3. Photomicrograph showing plastic deformation and flowage of soft lithic grain into adjacent pore space. Plain light. Muladi-1 well. 3294m. Bhuban fm.

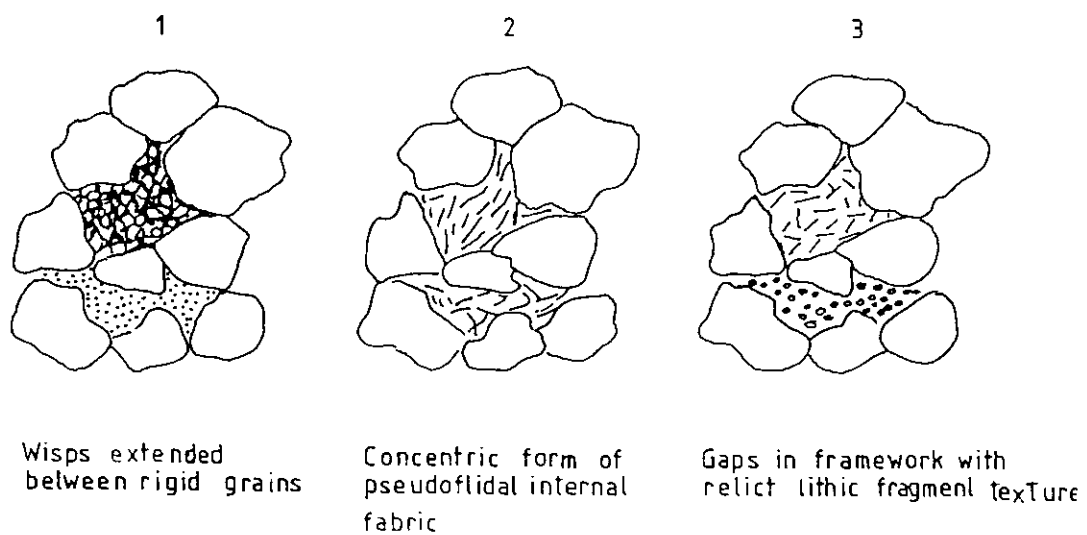


FIG. 5.4 CRITERIA FOR RECOGNITION OF PSEUDOMATRIX
(After Dickinson 1970)

lithic sandstones of Beluga formation, Alaska. In the present study, modal analysis indicated that 90% of the samples are sub-litharenite, the rest being sub-arkose and the average (mean) amount of lithic grains in these samples is 14% of the total rock components. Soft lithic grains including shale, claystone, siltstone and mica schist constitute about 52% to 66% of the total lithic grains in these samples with argillaceous sedimentary lithics (shale, siltstone, claystone) being one of the most abundant types (Table 5.1). Plastic deformation of these grains with flowage into adjacent pore spaces is common below about 3000 m (in the Bhuban formation) and significant porosity reduction has taken place below this depth via this process.

(c) Grain bending and fracturing:

A third and common feature of mechanical compaction of the Neogene sandstones is the bending of flexible minerals such as mica and chlorite and fracturing of brittle grains like the feldspars. Depending on the degree of overburden stress, mica bending ranges from gentle curving of long flakes into more extreme cases where the mica is bent into a zig-zag shape with some fracturing. Feldspars are the most susceptible to fracturing which takes place principally along their cleavage planes.

TABLE 5.1 : RELATIVE ABUNDANCE OF SOFT LITHIC GRAINS AMONG
THE TOTAL LITHICS IN THE NEOGENE SANDSTONES
(SURMA GR.) FROM THE BENGAL BASIN.

Sample no.	Shale clayst. siltst. (A)	Low grade meta. (B)	Total soft lith. (A)+(B)	Chert	Igneous, high grade meta, vol. & indetermi- nate.
M134B	25%	35%	60%	14%	26%
M1-36	25%	35%	60%	9%	31%
M1-38	24%	36%	60%	11%	29%
B1-14	23%	42%	65%	10%	25%
B1-26	26%	32%	58%	17%	25%
B2-3	25%	38%	63%	12%	25%
T-1	22%	44%	66%	14%	20%
T-5	27%	25%	52%	10%	38%
T-10	35%	19%	54%	12%	34%

5.3.2 Physico-chemical process - Pressure solution

Pressure solution in the sandstones studied principally involves the solution of quartz grains at the points of contact as a result of stress. Pressure solutions between detrital grain result in overly tight packing of the grains and increased compaction of the rock.

Pressure solution is usually recognised petrographically by the presence of concavo-convex or sutured boundaries between grains. However, in more recent studies using cathodoluminescence techniques it has been found that such contacts could equally be produced by the mutual interlocking of quartz overgrowths or by a compromise boundary between quartz overgrowths and detrital quartz grains where no pressure solution is involved (Sippel 1968; Sibley and Blatt, 1976). To always interpret the presence of sutured and concavo-convex contacts as indicative of pressure solution may thus be misleading. This has been evident in the present study from observations on the nature of grain contacts and presence of quartz overgrowth in the sandstones in the Muladi-1 and Begumganj-1 wells using petrographic, and scanning electron microscopes. The results of these observations are tabulated in table 5.2 It is indicated in the table, for example, that although samples from 2,680 m and 2,758 m have predominantly sutured and concavo-convex contacts, pressure solutions in these samples are minor and that most of these contacts result from interlocking quartz overgrowths.

TABLE 5.2 : RELATIONSHIP BETWEEN NATURE OF GRAIN CONTACT PRESSURE SOLUTION AND QUARTZ OVERGROWTHS IN THE NEOGENE SANDSTONES, BENGAL BASIN.

Sample no.	U	Depth (m)	Nature of grain contacts	Quartz overgrth.	Pressure solution	Remark
M1-3	T I P A M G R	500	Point contact predominates; Long contact subordinate; No sutured or concavo-convex contact.	Nil	None	—
M1-4		602		Nil		
M1-7		878				
M1-11		1137				
M1-20	B O K A B I L FM	1700		Nil		
B1-14		2139	Long contact predominate; Point contact subordinate; Few sutured & conc-conv. contacts.	Very rare	Little	zone of minor pressure solution Most of the interpenetrating grain contacts result from interlocking qtz overgrowth
M1-28		2216		Rare		
B1-16		2680	Sutured and concavo convex contact predominates. No point contact.	Abundant	Little	
B1-17		2758		Abundant		
B1-19	B H U B A N F M	2999	Sutured and concavo-convex contact predominates. No point contact.	Abundant	Common to very common	Zone of important pressure solution; however many of the interpenetrating grain contacts result from interlocking qtz overgrowths.
M1-37B		3223		Very common		
M1-38		3294		Very common		
B1-26A		3642		Common		
M1-49A		4458		Abundant		
M2-3		4491		Very common		

Under the scanning electron microscope, solution pits between the grains can be observed which indicate pressure solution. In Fig. 5.5 and 5.6 such presolved surfaces are shown where one of the grain has been removed during sample preparation exposing the surface of contact between them. A more magnified view of these surfaces indicates that they can be very irregular with ridges and depressions within them (Fig. 5.6A). The use of cathodoluminescence technique in distinguishing pressure solution from interlocking quartz overgrowths has been discussed in section 4.3.1.

From the above observations, it is concluded that:

(1) There is evidence that pressure solution has taken place in Neogene sandstones, although it is less abundant than is apparent from simply observing the interpretation of grain boundaries in petrographic microscope. Cathodoluminescence and SEM studies have shown that many such contacts are the results of interlocking of quartz overgrowths rather than pressure solution.

(2) Pressure solution increases with depth and has progressively contributed to the compactional processes in these sandstones. Pressure solution appears to be absent above 1700 m of burial (in the Tipam group and the upper part of the Bokabil formation.). Sandstones of the Bokabil formation (lower part) show some evidence of pressure solution. Pressure solution is common to very common in the Bhuban formation as seen in samples below about 2,758 m.

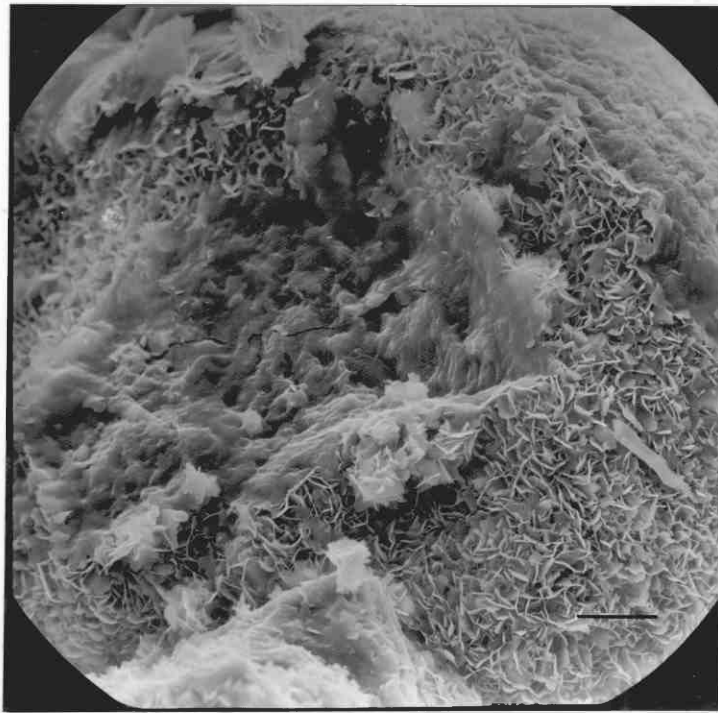


Fig.5.5.SEM micrograph of solution pit on quartz grain surface indicating pressure solution.Begumganj-2 well. 3325m. Bhuban fm. Scale bar=14 microns.

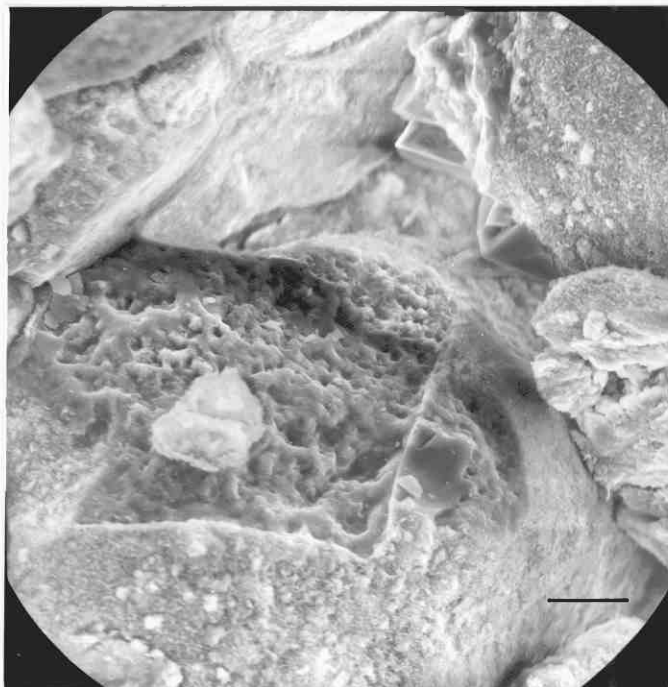


Fig.5.6.SEM micrograph of solution pit on quartz grain surface.Note ridges and depressions on the presolved surface.Titas-2 well.Bhubanfm.2712m. Scale bar=20 u.

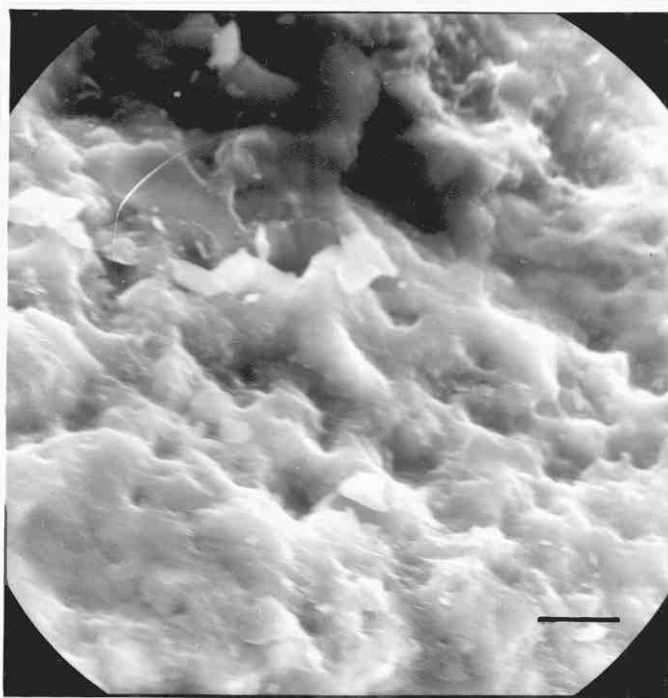


Fig.5.7. Magnified view of the presolved surface shown on fig. 5.6. Note the irregularities of the presolved surface. Scale bar = 5 microns.

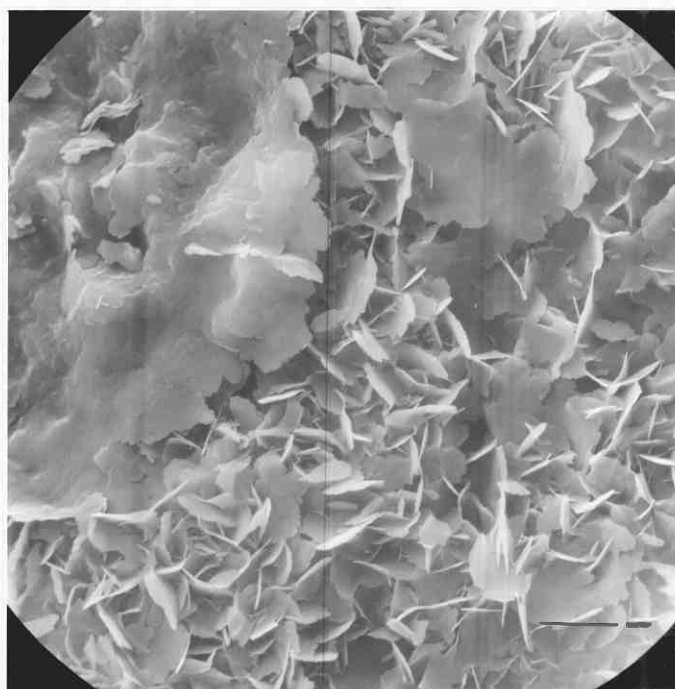


Fig.5.8. SEM micrograph of tangential flaky crystals of illite draping over radial chlorite rim. Begumganj -2 well. 3325m. Bhuvan fm. Scale bar = 4 microns.

5.4 CHLORITE AUTHIGENESIS

The development of authigenic chlorite is the earliest chemical diagenetic event recognised in the sandstones studied. The formation of chlorite rims predated all other diagenetic mineral formation. This is evidenced from their textural relationship with other diagenetic minerals. For example, illite overgrows the chlorite rims (Fig. 5.8), quartz overgrowths occasionally engulfs the rims and the chlorite rims are present in the very early poikilotopic carbonate cemented samples where further diagenesis was arrested following the carbonate cementation. Furthermore their presence in the shallowest sample studied (500 m) where most of the other diagenetic minerals are absent testifies to their very early origin in the diagenetic history. Though the other two forms of authigenic chlorite (i.e., pore filling chlorite and chlorite replacements) differ in habit from the chlorite rims, their intimate textural relationship and close association with chlorite rims indicate that their formation was more or less contemporaneous with the rims. They probably represent a single episode very early in the diagenetic history with an environment in which chlorite formation and stability was favoured over other silicates.

Thomson (1979) noted a close correlation between the presence of volcanic detritus and the formation of chlorite rims in the Tuscaloosa Sandstone (Silurian, U.S.A.). He suggested that the authigenic chlorite rims were the result of alteration of volcanic detritus and concomitant precipitation of chlorite crystals on detrital grain surfaces, although the manner in which the alteration and precipitation took place was not known. Volcanic rock fragments represent trace to very minor fractions of the detrital components in the studied samples and are unlikely to be the principal source of the very common authigenic chlorites. It has not been possible

to observe any likely precursor minerals of the authigenic chlorites in the sandstones studied. The delicate morphology and texture of the chlorite rims and the chlorite pore fills indicate that they are neoformed. Mineral components other than volcanic detritus may also constitute the progenitors of authigenic chlorite (Hayes 1970). An alkaline conditions of the pore water is suggested which favours the formation of authigenic chlorite (Shaw 1980). The corrosion of the lithic and other mineral components including the volcanic clasts may contribute to the alkaline geochemical pore water environment rich in Fe and Mg required for the formation of the chlorite. However, there is no evidence to suggest that these lithic or mineral phases were directly altered to the chlorite.

The effect of chlorite authigenesis on reservoir properties is principally related to the development of the chlorite rims. The pore filling chlorites and chlorite replacements are volumetrically insignificant and have little noticeable effect on the overall porosity or permeability. The chlorite rims also occupy only small volume of the total pore space and thus do not reduce porosity in any significant way, but they can be detrimental to permeability by reducing the pore throat diameter (Galloway 1979). However, the most important effect of chlorite rims development is the preservation of porosity by inhibiting the growth of quartz overgrowths. The inhibition of quartz overgrowth is most effective where the grain rims are thick and continuous. Nucleation of quartz overgrowths on such grains could not take place at all. However, if fracture exposes any part of the quartz grain, small, isolated prismatic quartz overgrowth develops on the fractured surface. Such cases indicate that the geochemistry of the pore fluids was favourable for quartz overgrowth development and that porosity

presumably could have been reduced appreciably by silica cementation in these samples if chlorite rims had not been present. It has also been suggested that thick and well developed chlorite rims could retard the normal process of compaction (Thomson 1979). Samples M -34B and M -36 (Surma group) from Muladi well provide excellent illustrations of the effect of chlorite rims in inhibiting the formation of quartz overgrowth. These two samples with thick and continuous chlorite rims have developed no quartz overgrowth (Fig. 4.1), although closely associated samples with sparse or no chlorite rim are well cemented with quartz overgrowth. The semi-friable nature of these two samples with respect to their depth (2,871 m and 3,114 m respectively) is the result of lack of silica cementation and possibly, at least in part, due to undercompaction in response to the thick chlorite rims.

5.5 SIDERITE CRYSTALLIZATION:

A very early generation of authigenic siderite is postulated from the textural relationship of the mineral with other diagenetic phases and also from its occurrence in the shallowest samples. Siderite crystallization pre-dates the early poikilotopic calcite as indicated by the presence of siderite euhedral and subhedral crystal aggregates within the cement. Chlorite authigenesis is the only chemical diagenetic phase which is believed to have preceded the siderite crystallization. The stability field of siderite in the Eh-pH diagram (Fig. 5.9) shows that the mineral forms in a reducing environment. Berner (1981) indicated that siderite forms in anoxic non-sulphidic environments i.e. it needs to have a limited supply of sulphate; otherwise the reduction of sulphate to sulphide would produce pyrite. Pettijohn et al. (1972) indicated that the most probable reducing agent of iron is organic matter and linked the formation of siderite to non-marine environments in which organic matters are abundant, such as stagnant basins or estuarine environments with high organic productivity. Siderite in the studied samples is an insignificant/very minor phase occurring throughout the total depth range studied and no apparent correlation has been noted between environment of deposition and siderite occurrence. Siderite crystallization has no important effect on the reservoir properties of sandstones because of their insignificant amount.

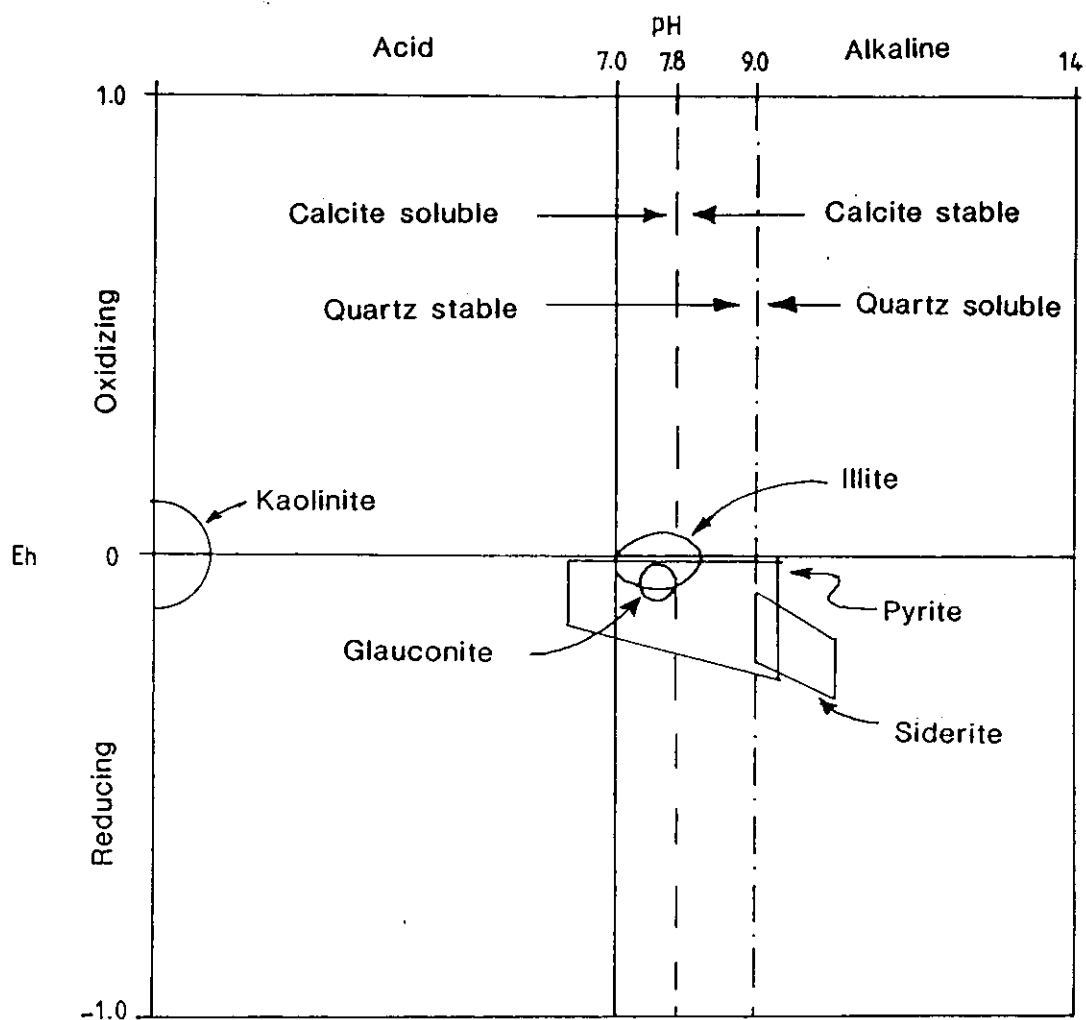


FIG. 5.9 Eh-pH DIAGRAM SHOWING APPROXIMATE STABILITY FIELD OF IMPORTANT SEDIMENTARY MINERALS .
(After Selley 1982).

5.6 EARLY POIKILOTOPIC FE-CALCITE CEMENTATION

Two phases of Ferroan calcite cement generation - an early phase and a late phase - have been identified in the present study. The early phase is represented by the poikilotopic Fe-calcite cement which affects a few samples and has totally destroyed their porosity and permeability. A very early origin of the poikilotopic ferroan calcite is indicated by the loose packing, the "floating" character of the detrital grains and the lack of compactional features such as grain bending, grain contacts etc. in the rock. Also, the absence of most of the diagenetic minerals in these rocks is believed to be related to the total filling of all the available pore spaces by ferroan calcite cement at a very early stage of diagenetic history, arresting any further diagenesis following poikilotopic Fe-calcite cementation. Dapples (1971), however, suggested that crystallization of the poikilotopic cement involves some expansion of the intergranular spaces, thus increasing the separation between detrital grains. The early poikilotopic Fe-cementation is most effective among all the diagenetic events in destroying porosity and permeability. The porosity and permeability of the effected sandstones are reduced to zero by this process as seen in samples no. M1-27, M1-29 and M1-39 (Table 3.2).

The most common origin of the early generation of calcite cement is the dissolution of shell materials and re-precipitation of calcium carbonate cement in pore spaces (Almon and Davis 1979; Blatt 1979). Although rare carbonate shells or particles are associated with poikilotopically cemented sandstones, it is believed that the localised nature of these early calcite cements is probably related to a local concentration of shell debris in the sandstone beds. Dissolution and recrystallization of the carbonate has meant that all trace of the original

material has been lost and one can only offer conjecture as to what it originally might have been (Dapples 1971). The stability field of calcite in an Eh-pH diagram indicates that crystallisation of calcite takes place under alkaline condition (Fig. 5.9). The ferroan nature of the calcite also indicate a reducing environment.

5.7 ILLITE CRYSTALLIZATION:

A relatively early origin for the authigenic illite in the Neogene sandstones is indicated from the presence of illite in samples from very shallow depths. However, it post-dated chlorite authigenesis as is obvious from Fig. 5.8, where tangential illite flaky crystals are seen to drape over chlorite rims. Frequent occurrence of illite as "dust lines" between detrital quartz and authigenic quartz overgrowths testifies to its origin being earlier than the quartz overgrowth (Fig. 4.13).

It has been shown by previous studies that illite could form from the in situ alteration of pre-existing minerals such as feldspar, kaolinite, and mica, the recrystallization of clay minerals or by neoformation (Hancock and Taylor 1978; Huggett 1982). In the present study there is no evidence to suggest the origin of illite by the alteration or replacement of kaolinites, micas or feldspars. A neoformational origin has been suggested which would require K-rich alkaline pore water conditions to be established early in the diagenetic history.

5.8 QUARTZ OVERGROWTH CEMENTATION

Quartz overgrowth formation postdated the authigenic illite formation which the quartz overgrowths envelope but it preceeded late Fe-calcite cementation. Its earlier origin than the late Fe-calcite cement is indicated by the euhedral quartz overgrowth crystal faces being enveloped by the pore filling late Fe-calcite cement (Fig. 5.10). The replacement of quartz overgrowth by late calcite also suggests its earlier origin than the replacive calcite (Fig. 5.11).

The source of silica for quartz overgrowth cementation in sandstone has variously been suggested as follows (Pittman 1979): (1) Pressure solution; (2) Diagenesis of clay minerals; (3) Decomposition of feldspars; (4) Replacement of silicates by carbonates (5) Precipitation from downward percolating ground water; (6) Dissolution of siliceous organisms such as sponges, diatoms, radiolarians etc. (7) Hydration of volcanic glass.

During the past two decades pressure solution of the detrital quartz grains has become the most generally accepted mechanism for supplying dissolved silica into pore waters and stimulating the growth of quartz overgrowths. However, use of cathodoluminescence petrography has shown that pressure solution as a source of silica is less than previously thought (Sippel 1968). Sibley and Blatt (1976), using cathodoluminescence petrography, indicated that pressure solution could not have supplied more than one third of the silica present as quartz overgrowths in the Tuscaloosa sandstone (Silurian). It has previously been mentioned that pressure solution in the present sequence is common below about 2758 m, although it is less abundant than is apparent from simply

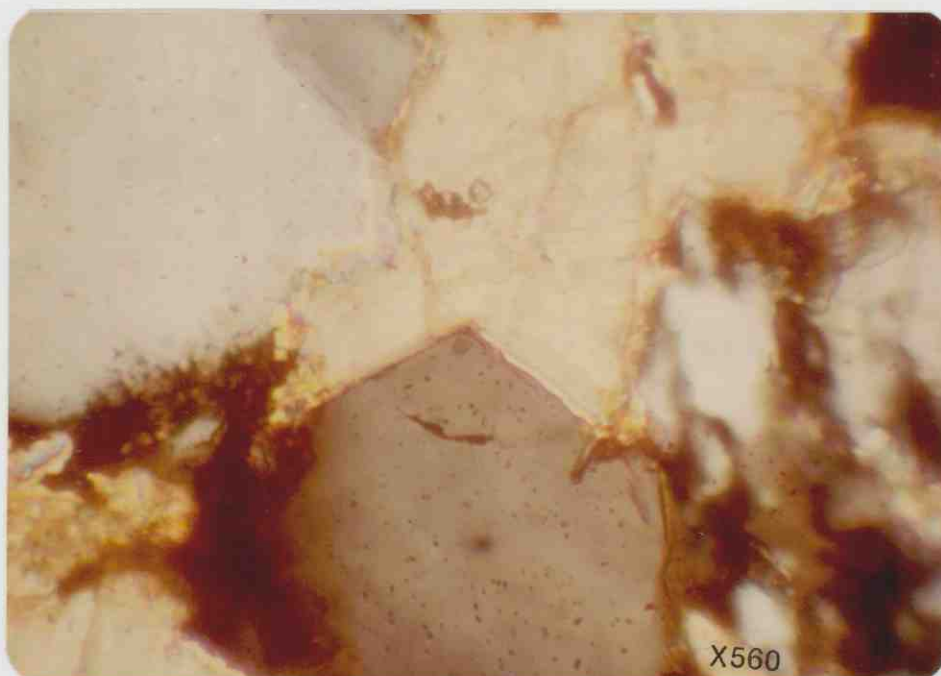


Fig.5.10. Photomicrograph of euhedral quartz overgrowth crystal faces enveloped by late calcite cement. Crossed nicols. Muladi-2 well. 4491m. Bhuban fm.

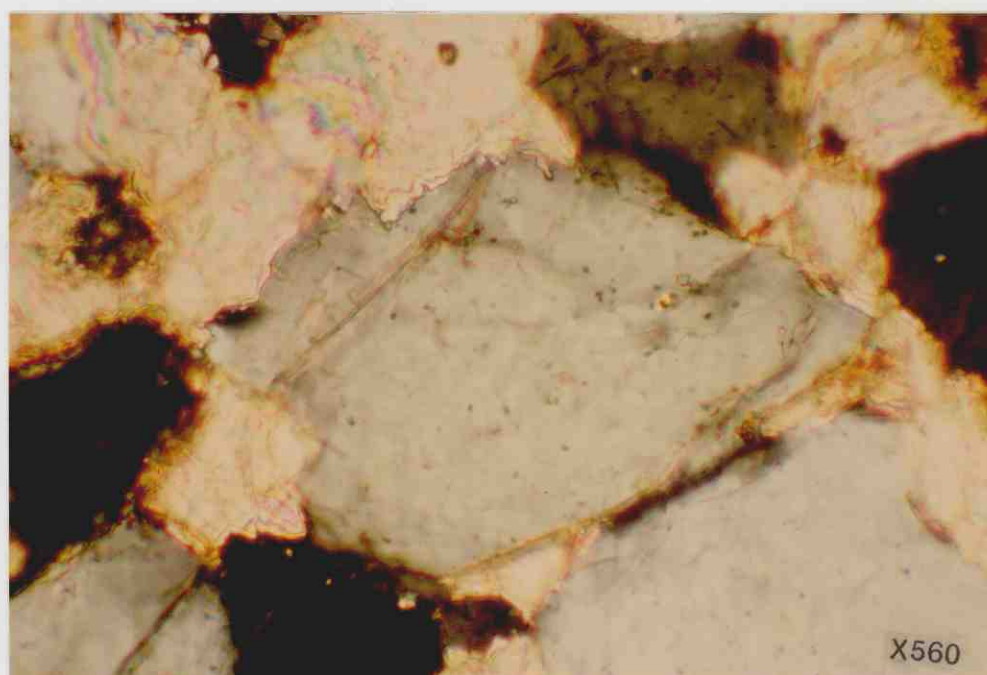


Fig.5.11. Photomicrograph of quartz overgrowth partially replaced by late calcite cement. Crossed nicols. Muladi-2 well. 4491m. Bhuban fm.

observing the interpenetration of grain boundaries under the petrographic microscope. Pressure solution is, therefore, likely to be an important source of silica cement below this depth. However, other processes must have been active as sources of silica cement to generate the abundant quartz overgrowths in the some samples above this depth where pressure solution is minor.

The transformation of smectite to illite during the diagenesis of shales releases Si^{4+} ions as a by-product which could migrate to adjacent sandstone beds during shale dewatering and could cause quartz cementation (Towe 1962, Boles and Frank 1979, Lahann 1980, Boles 1981). The amount of Si^{4+} ion released into the pore water system would depend on the availability of shale containing smectite or illite/smectite and the degree of diagenesis. Illite/smectite clay is one of the major clay components in the Neogene shales of the Bengal Basin and progressive diagenetic transformation of the smectite to illite layers takes place increasing depth of burial (Chapter 6). Quartz overgrowth is a deeper subsurface phases and is significant only in the Bhuban formation. The increasing abundance of quartz overgrowths below about 2220 m correlated with an increased shale/sand ratio within the rock sequences and an increasing degree of diagenetic transformation of smectite to illite in the shales. Also, increased quartz cementation has been observed toward the shaly margins of sandstone beds in general and is presumably related to the greater availability of migrating Si ions from adjacent shales. It is thus suggested that illite/smectite clay diagenesis is one of the important sources of silica cement in these sandstones.

Replacement of the silicates such as quartz and feldspars by calcite and dissolution of feldspars are both considered to be potential sources of silica cement, as both events would release silica into the pore water (Pittman 1979). In the present samples, although marginal replacement of silicate grains by calcite cement is observed in many samples, it is extensive only in a few sandstones. Dissolution of feldspar is also important only in limited areas and is not always correlatable with the major occurrence of quartz overgrowths. Textural evidence suggests that these events post-date quartz overgrowth formation. However, it is possible that some of the overgrowths occur after Fe-calcite replacement of quartz and feldspar dissolution. Thus some of the silica cement may have been derived locally from these sources but generally they are not thought to be important contributors to silica cementation in the Neogene sandstones of the Bengal Basin.

It has been suggested that descending groundwater of normal SiO_2 concentration could provide the necessary flux through sandstone to explain large scale subsurface silica cementation (Bjorlykke 1979). In the studied Neogene section of the Bengal Basin, the sediments appear to have undergone continuous burial since deposition. It is difficult to envisage a regional descending groundwater (meteoric) circulation system as Bjorlykke (1979) has suggested, because the net movement of water in the studied region is believed to have been upward and out of the basin with increasing compaction since deposition. Quartz cementation here is a relatively deep subsurface phenomenon (significant only in the Bhuban formation) and there is no evidence of early quartz cementation in the shallower sediments (Tipam group) in what might be envisaged as the zone of meteoric water circulation. It is thus believed that quartz cementation from descending ground water is not applicable in the studied sandstones.

The possibilities of quartz cementation from dissolution of siliceous shells or from hydration of volcanic glass are also remote, since there is no evidence to suggest the occurrence of these materials in the samples studied.

It is thus concluded that pressure solution and smectite to illite clay diagenesis are the two important sources of silica cements in the studied sandstones, while dissolution of feldspar and replacement of silicates by carbonate may have been a source of silica to a minor extent.

The effect of quartz overgrowth formation on the reservoir properties is most important in Bhuban Formation where significant amounts of porosity and permeability are lost in many sandstones by quartz cementation. This is specially true for samples where quartz overgrowths form interlocking textures. The result of the loss of permeability by quartz cementation is reflected by the absence of authigenic kaolinite and feldspar leaching in highly quartz cemented sandstones, which is presumably due to the pore water not being able to penetrate these sandstones.

5.9 LATE FE-CALCITE (WITH MINOR DOLOMITE) CEMENTATION:

A late stage of ferroan calcite cementation (with minor dolomite) followed the quartz overgrowth formation and partially infilled the available pore spaces and replaced the detrital grains. The late origin of these carbonates is indicated by the nature of the pore filling and their relationships with other diagenetic features. Unlike the early poikilotopic calcites in which grains essentially float in the cement, the late Fe-calcite infills isolated pore spaces remaining after compaction has established a denser packing of the framework grains. The late Fe-calcite cement often cross-cuts grain contacts including sutured contacts indicating its formation after the observed compaction and pressure solution. The occurrence of euhedral quartz overgrowths crystal faces against pore filling Fe-calcite and the replacement of quartz overgrowth by Fe-calcite indicates that the Fe-calcite post dates quartz overgrowth formation (Fig. 5.10 and 5.11). Important amount of porosity and permeability has been lost by late Fe-calcite cementation. This was seen to be most prominent where pore-filling Fe-calcite accompanies major grain replacive Fe-calcite.

Precipitation of calcite requires an alkaline condition of the pore water, high activity of Ca^{++} and $\text{CO}_3^{=}$, low total CO_2 and high temperature (Stanton 1977). The solubility of quartz is increased greatly at pH greater than 9.0 (Fig. 5.9). It is suggested that while alkaline condition of the pore water prevailed during precipitation of late Fe-calcite cement, strong alkalinity (pH 9.0) was established where quartz replacement by calcite took place. The ferroan nature of the calcite also indicates a reducing environment.

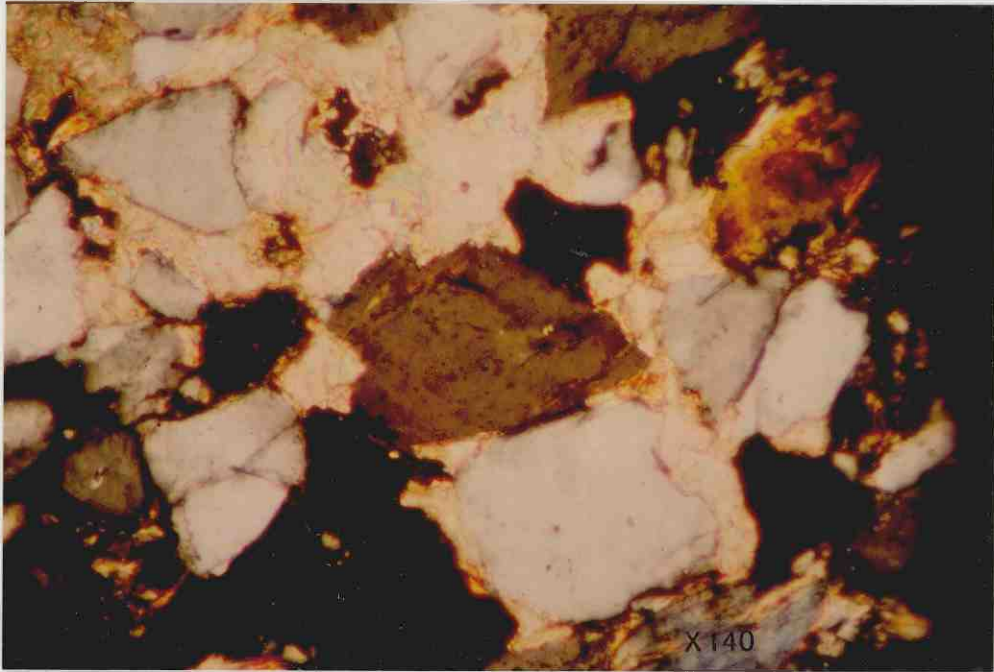


Fig.5.11A. Photomicrograph of late calcite cements.
Begumganj-2 well. Bhuban fm.

Dissolution of calcite shell fragments or particles may be a source of late calcite cement but the available evidence suggests that most of the studied sections are devoid of shell material or carbonate particles and such a source is, therefore unlikely in the present case.

Sources of carbonate cements from outside the sandstone layers have been suggested by many workers. Curtis (1978) indicated that carbonate cement generated in mudstone sequences could migrate in pore water solutions to adjacent sandstones during compaction and dewatering of the mudstones before re-precipitating as calcite cements in the sandstone. Netto (1979) suggested that adjacent shale beds were the source of carbonate cement in Cretaceous sandstone studied in Brazil. Perry and Hower (1970) noted a decrease in Ca^{++} and Mg^{++} in shale due to solution of carbonate with increasing depth. In the present study a decrease in the abundance of calcite in the shales with increasing depth of burial and its ultimate disappearance is indicated in Neogene section (Chapter 6, Section 6.3.1.). This probably reflects progressive dissolution of carbonate cement in the shales with depth. The dissolved calcite could have migrated to adjacent sandstones and re-precipitated as calcite cement.

The diagenetic transformation of smectite to illite in the Illite/smectite clay in shale could also release Ca^{++} , which could subsequently migrate during shale compaction to the adjacent sandstones and serve as a source of calcite cement (Boles and Frank 1979). Such diagenetic transformations of the Illite/smectite clay have taken place in the Neogene shales of the Bengal Basin (see Chapter 6) and Ca^{++} thus released could have migrated to adjacent sandstones to form calcite cement.

The Dolomite which occurs as a very minor diagenetic phase in the sandstone is mostly associated with the Fe-calcite. The textural relationships of the minor dolomite with Fe-calcite indicates a replacive origin of the dolomite after Fe-calcite.

5.10 PARTIAL DISSOLUTION OF ROCK COMPONENTS AND DEVELOPMENT OF SECONDARY POROSITY

Partial dissolution of detrital feldspars, lithic grains and authigenic calcite cements is a common diagenetic feature in the deeper subsurface Neogene sandstones of the Bengal Basin. Dissolution features are found most extensively in the Titas well throughout the cored depth of 2,646 m to 2,967 m (Bhuban formation). Dissolution is also noted in the Bokabil and the Bhuban formations in the Muladi-1 and Begumganj-1 wells but is much less extensive in these two wells.

This stage of diagenesis is thus characterized by development of secondary porosity. The fact that the impregnating resin used during sample preparation has easily filled the dissolution voids in the grains indicates that these constitute part of the effective porosity (Fig. 5.12 and 5.13). Point counting analysis showed that secondary porosity constitutes as much as 37% of the total porosity in some samples of the Bhuban formation.

(a) Dissolution of feldspars:

Feldspars show the most extensive degree of dissolution amongst the rock components. The dissolution of feldspars accounts for almost 75% -80% of the secondary porosity developed in some samples. Among the groups of feldspars, orthoclase shows the most extensive leaching, but dissolution of plagioclase feldspars was also observed.

Microcline feldspar, however, does not seem to have suffered any dissolution and remains "fresh. In many cases where leaching has been very pronounced, 60% - 70% of the individual feldspar grains may have been removed. Solution proceeds irregularly throughout the grains and the solution voids interconnect with primary porosity, leaving only remnant grains. These kinds of leached grains have been termed "honeycombed grain" (Schmidt and McDonald 1979) or "skeletal grain" (Roswell and De Swart 1975) and are illustrated in Fig. 5.12 and 5.13. However, in rare cases almost total leaching of the feldspar grain has taken place resulting in the development of oversized pore space amongst an otherwise closely packed texture (Fig. 5.14). Giles and Marshall (1983) suggested that significant increase in the total porosity as a result of feldspar leaching is not likely in deep subsurface because an almost identical volume of primary porosity is filled by products of feldspar leaching.

Regional variation in the degree of feldspar leaching have been noted. The most extensive feldspar leaching is seen in the Titas-2 well throughout the cored depth range of 2,646 m to 2,967 m (Bhuban formation). Feldspar leaching appears to be much less well established in the Bhuban and in the Bokabil sandstones of Begumganj-1 and Muladi-1 wells.

(b) Dissolution of lithic grains:

The partial dissolution of lithic grains is less commonly observed than the dissolution of feldspars. Amongst the lithic grains which do show dissolution are chert, quartzite, various kinds of argillites etc. The dissolution of lithic grains may have occurred via the selective dissolution of a particular component in them, such as dissolution of carbonate cement filling the primary pores of a chert grain.

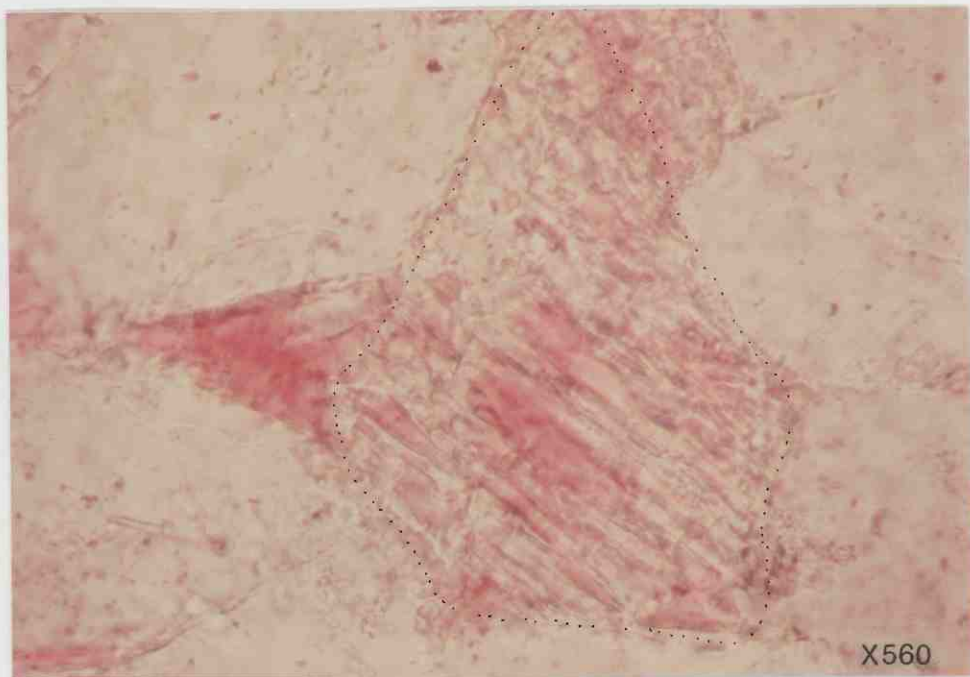


Fig.5.12.Photomicrograph of skeletal grain with secondary porosity resulting from partial dissolution of feldspar. Plate - light. Sample impregnated with red stained media to emphasise porosity(red).Titas-2 well.2646m.Bhuban fm.

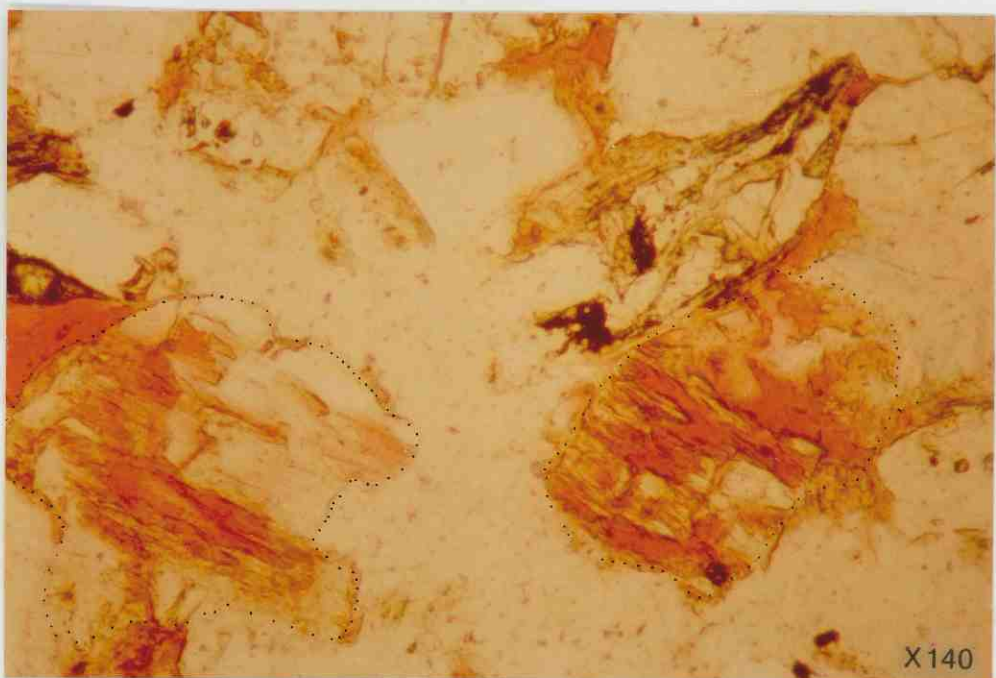


Fig.5.13.Photomicrograph of skeletal grains with secondary porosity resulting from partial dissolution of feldspars. Plate - light. Sample impregnated with red stained media to emphasize porosity(red).Titas-2 well. 2690m.Bhuban fm.

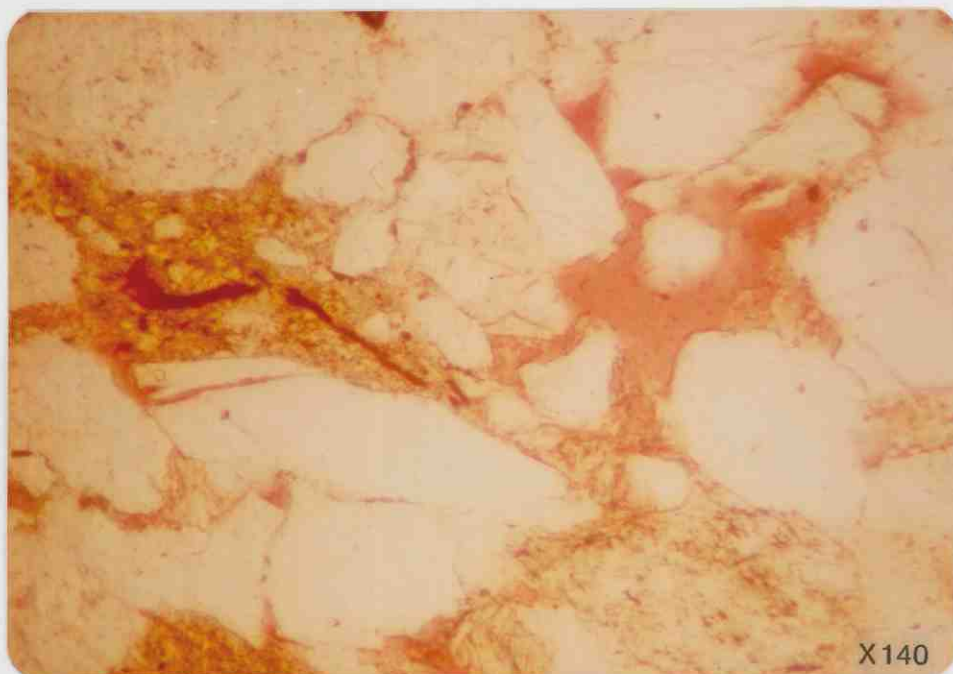


Fig.5.14. Photomicrograph of oversized pore (red) resulting from dissolution of feldspar grain. Plane light. Sample impregnated with red stained media. Titas-2 well. 2690m. Bhuban fm.

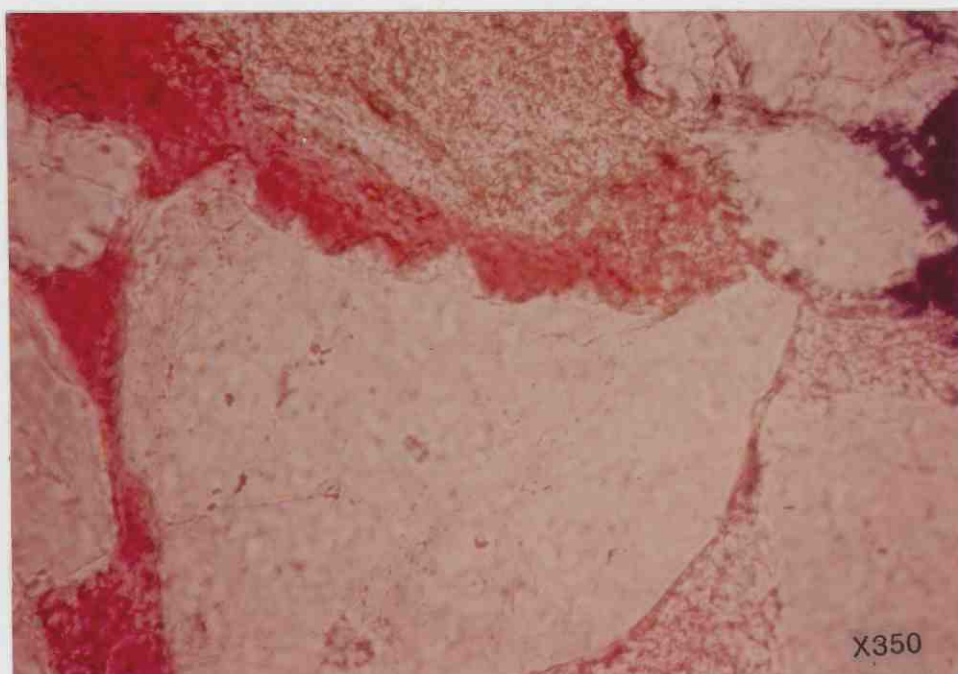


Fig.5.15. Photomicrograph of corroded grain adjacent to pore space (red). Plane light. Sample impregnated with red stained media. Titas-2 well. 2704m. Bhuban fm.

(c) Dissolution of Fe-calcite cements:

Partial dissolution of Fe-calcite cements has been observed in most of the sandstones which were cemented by late Fe-calcite cements. The absence of any evidence of dissolution in the early poikilotopic Fe-calcite cement is probably because the pore waters responsible for dissolution could not percolate into the rock as a result of total destruction of porosity by the early poikilotopic Fe-calcite cement. The evidence by which the partial dissolution of Fe-calcite cements is established mainly includes intracement pores (Schmidt and McDonald 1979). Occasionally corroded grain boundaries associated with intergranular pores serve as evidence of cement dissolution (Fig. 5.15). This results from dissolution of soluble calcite cement that replaced and corroded the margin of grain and filled the adjacent pore prior to dissolution.

The dissolution of calcite cements and some aluminosilicates requires acidic pore water conditions. Introduction of fresh meteoric undersaturated water into the formation is believed to be one way of establishing an acidic environment. This might result either after the uplift and subaerial exposure of the formation or by a fresh meteoric water flow connection from the surface source to the subsurface formation. There is no evidence to suggest that the Bokabil and Bhuban formations, in the sections where dissolution has been observed, have been uplifted significantly or subaerially exposed after burial. The Neogene section here is dominated by a continuous burial history. No meteoric water flow connection exists, nor is believed to have previously existed in these sandstone bodies since their deep burial. Pore water movement is essentially upward and out of the basin as a result of continuing basinal burial and compaction of the sediments. It is therefore very unlikely

that surface meteoric/fresh water is responsible for the acidic condition in the formation.

A second process by which acidic pore fluid conditions may be established at greater depth is from CO_2 released by the maturation of organic matter in shale (Schmidt and McDonald 1979). It is believed that the process of thermal decarboxylation of kerogen in shales leads to the generation of CO_2 . The dissolved CO_2 lowers the pH of pore fluids migrating into the sandstone, thus providing an acidic solvent for the carbonates, and the aluminosilicates (feldspars). Humic kerogen, high kerogen percentages in the shales (source rocks) and high shale/sand ratios favour this process (Bjorlykke 1980). The nature of kerogen in the present Neogene shale is humic (Khan and Hussain 1980), and the shale/sand ratio is high in the more deeply buried sequences. However, the total organic carbon of the studied shales only range for 0.26% to 0.45% (Table 5.3) and the Neogene shale does not form a source rock for the natural gas found in the Neogene reservoirs (Khan and Hussain 1980). To what extent, if any, the maturation of kerogen in Neogene shale contributes towards the establishment of acidic environments in the sandstones is not known. However, there is a correlation between the increased dissolution of carbonates and feldspars in the Titas-2 well samples (Surma Basin areas), as compared to the Muladi-1 and Begumganj-1 well samples, and increased maturity of the sediments within the Surma Basin (Khan 1980). The source of the Neogene gas reserves lies in the underlying Paleogene shales (Lietz and Kabir 1978) and it is possible that CO_2 evolved during maturation of these source rocks could have migrated upward into the Neogene sandstones and established acidic pore water conditions.

TABLE 5.3 : TOTAL ORGANIC CARBON PERCENTAGES IN THE
NEOGENE SHALES (SURMA GROUP) OF THE
BENGAL BASIN.

Sample no.	Percentage organic carbon
B1-15	0.40
B1-18	0.33
B1-20	0.42
B1-22	0.38
B1-23	0.36
B1-24	0.35
B1-25	0.38
B1-26B	0.40
B2-2	0.36
B2-5	0.31
B2-7	0.35
M1-31	0.36
MI-34A	0.27
M1-35	0.26
M1-40	0.30
M1-41	0.37
MI-44	0.36
MI-45	0.44
M1-47	0.36
M1-49	0.40
T-6	0.41
T-8	0.39
T-14	0.45
T-17	0.39

5.11 KAOLINITE CRYSTALLIZATION

Crystallization of authigenic kaolinite is the final event in the sequence of diagenetic mineral formation in the Neogene sandstones. This is indicated by the following textural relationship: (i) Authigenic kaolinite crystals rest on top of quartz overgrowth euhedra, indicating their later origin. (ii) kaolinite occasionally occupies oversized pore spaces which were formed by dissolution of rock components, indicating an origin later than the dissolution stage.

It has been suggested that the precipitation of kaolinite requires a low pH (acidic) of pore water and the availability of a source for Al^{3+} and Si^{4+} (Bucke and Mankin 1971; Shaw 1980). It has also been pointed out that the process of acidic dissolution and decomposition of feldspar could produce authigenic kaolinite by reaction similar to $2KAlSi_3O_8 + 2H^+ + H_2O = Al_2Si_2O_5(OH)_4 + 4SiO_2 + 2K^+$ (Hurst and Irwin 1982, Bjorlykke 1980). Formation of kaolinite by the alteration of other minerals, such as muscovite, and chlorite, has also been described (Sims 1970; Huggett, 1982).

In the present study the genetic relationship of authigenic kaolinite and detrital feldspar is indicated by (i) Evidence of in situ alteration of feldspar to kaolinite, and (ii) Close correlation between the presence of pore filling authigenic kaolinite and the occurrence of partially dissolved feldspar grains. Kaolinite is most common in the sandstones of the Titas well where feldspar dissolution is at its maximum. On the other hand in the samples from the Begumganj-1 and Muladi-1 wells, where feldspar dissolution is much less common, authigenic kaolinite has a relatively low abundance and restricted distribution. The above observation indicates that feldspar is the likely precursor mineral of authigenic

kaolinite, either by dissolution providing the necessary nutrients (Al^{3+} and Si^{4+}) to the pore water, or by in situ alteration.

The flushing of formations by acidic fresh or meteoric water is widely believed to be responsible for such a process of kaolinitization in sandstones (Blanche and Whittaker 1978, Almon and Davis 1979, Bjorlykke et al. 1979). Hurst and Irwin (1982) used the term "flux controlled kaolinite authigenesis" for this process, but indicated that, in shale-enclosed sandstone bodies not interconnected to the earth's surface, such fresh water fluxes through the sandstones are unlikely. They invoked a process of "diffusion controlled kaolinite authigenesis" to account for such environments of kaolinite authigenesis.

Kaolinite is found only in the deeply buried Bhuvan formation. There is speculation as to how acid pore water condition was established in this formation. It was explained in the previous section (Section 5.10) that the flushing of the deep formations by surface meteoric fresh water is not considered to be a likely process in the studied Neogene section of the Bengal Basin because they have never experienced significant uplift and subaerial exposure and there has never been a meteoric water flow connection between them and the surface after deep burial. However, acidic pore waters could have been generated by the CO_2 evolved from the maturation of kerogen (section 5.10); this, however, is conjectural.

Kaolinite precipitation is favoured under acid condition where cations (Na^+ , K^+ , Ca^{++} , Mg^{++}) are not present in sufficient quantity to produce cation aluminosilicate (Curtis 1983); very low cation concentrations at relatively moderate silica values favour crystallization of kaolinite (Sims 1970). Kaolinite crystallization was

the last event in the paragenetic sequence of diagenesis of the Neogene sandstones in the Bengal Basin. The diagenetic events which preceeded it included chlorite authigenesis, siderite crystallization, illite crystallization, carbonate cementation etc. These processes would progressively remove the cations from the pore waters and leave a residual formation water depleted in cations but having an increased $\text{Al}_2\text{O}_3/\text{SiO}_2$ ratio. Conditions would thus have been established which would have encouraged the formation of authigenic kaolinite from the pore water.

Kaolinite crystallization result in loss of porosity, although to a lesser extent as compared to that by quartz cementation or Fe-calcite cementation. Kaolinite constitutes 0.2% to 2.0% of the total rock components, the mean value being 0.8%. Recalculation of these values as a percentage of pre-cement porosity (Dutton 1977) in representative samples indicate that authigenic kaolinite is responsible for reducing total porosity by about 1% - 14% (Table 5.4). This may represent an equivalent amount of porosity that may have been produced secondarily by feldspar leaching (Giles and Marshall 1983).

TABLE 5.4 : POROSITY REDUCTION BY KAOLINITE CEMENT IN THE
NEOGENE SANDSTONES (BHUBAN FORMATION) FROM
THE BENGAL BASIN

Sample no.	%Porosity (thin section)	%Pre-cement porosity (thin sect.)	%Kaolinite cement	Porosity reduction by Kaolinite cement
T-1	15.5	17.3	1.33	8%
T-2	9.4	9.8	0.4	4%
T-3	8.6	10	1.4	14%
T-7	10.8	12.9	1.5	11%
T-10	9.2	12.4	0.2	1%

CHAPTER 6

MINERALOGY AND DIAGENESIS OF SHALE

6.1 INTRODUCTION

An understanding of diagenesis in shales can be gained from the critical examination of their bulk mineralogy using x-ray diffraction methods of analysis. Amongst the principal modifications that affect argillaceous sediments during burial diagenesis are the changes in their clay mineral assemblages. One of the first detailed investigations of shale diagenesis based on the examination of clay mineral assemblage was carried out by Burst in 1959, who reported the gradual change of the swelling smectite clay mineral to a non-swelling illite clay with increasing depth of burial. Since then many investigations have reported that progressive lithification of shales with depth of burial coincides with the progressive loss of swelling clay minerals in many geological sequences. The decrease and disappearance of kaolinite, appearance and increase in chlorite and loss of K-feldspars, are amongst other diagenetic changes recorded in shale with increase of temperature and pressure by burial depth (Dunoyer de Segonzac 1970, Perry and Hower 1970)

To date no studies on the mineralogy or the diagenesis of the Neogene shales in the Bengal Basin have been reported. The Neogene sequences in the Bengal Basin are characterised by subequal proportions of sandstones and shales except for the sandstone dominated Tipam Group at the top. In certain sections, however, particularly in the lower levels, shales are much more abundant. The shale:sand ratios can be as high as 8:1 and individual

shale bodies can have thicknesses of 200 m to 300 m. The uninterrupted burial history of these shales make the Neogene sections ideal for shale diagenesis studies.

For the purpose of the present study a total of fifty shale samples were analysed from the Muladi-1, Muladi-2, Begumganj-1, Begumganj-2, BODC-2 and Titas-2 wells covering a depth range from 1197 m to 4550 m below surface. These samples belong to the Tipam Group and the Surma Group (Bhuban and Bokabil Formations) of Neogene age. Detailed mineralogical analyses of these samples were carried out to see the diagenetic changes these shales have suffered in response to burial.

6.2 METHODS OF STUDY:

The present study of shales is based on analyses using standard x-ray diffraction methods (Brindley and Brown, 1980). The x-ray diffraction analyses were carried out using a Phillips x-ray diffractometer using CoK_{α} radiation at 36 Kv and 29 mA. Each samples was analysed for its bulk mineralogy and its detailed clay mineralogy.

6.2.1 Whole rock analysis.

For determining the bulk mineralogy of the shales, unoriented mounts of the powdered whole rock sample were prepared. For this purpose, 5 to 10 g of sample was washed, dried and broken into small pieces (1-5 mm) using a pestle and mortar, then pulverized using a tungsten carbide mortar and a Tema Mill. The fine powder thus obtained was then packed tightly into the cavity of an aluminium holder and the surface smoothed over by pressing the powder with a glass slide. The whole rock powder samples were then analysed in the x-ray diffractometer over an angular range of 5 degrees 2θ to 75 degrees 2θ with a chart speed of 1 degree per minute and count rate

of 400 counts per second.

6.2.2 Clay mineral analysis:

The clay mineral analyses were carried out on oriented mounts of the clay ($2\ \mu$) fractions. The oriented mounts were prepared from dispersed suspended clay fraction in distilled water. To separate out the clay fraction the samples (15 g - 20 g) were gently broken into very small fragments (1-2 mm) using a pestle and mortar and then put into an ultrasonic bath in a beaker for half an hour or more depending on the hardness of the rock. The suspended sediment thus obtained was then used to obtain $2\ \mu$ fraction by separation under gravity (Turner and Jackson 1947; Brindley and Brown 1980). The oriented mounts of the clay fractions were prepared by filtering the clay fractions suspended in distilled water through ceramic discs using the suction-onto-ceramic-tile method outlined by Shaw (1972). A layer of clay of sufficient thickness thus deposited on the ceramic disc, was allowed to air dry before analysis in the x-ray diffractometer. In addition to the use of oriented mounts in the air-dried state, the orientated clay specimens were also analysed after special treatments to allow complete identification of clay minerals. The clay discs were thus scanned routinely as follows.

- (a) In air dried state.
- (b) After solvation with ethylene glycol and/or glycerol.
- (c) After heating to 300 degrees centigrade for 1-hours.
- (d) After heating to 550 degrees centigrade for 1- hours.

Care was taken to ensure that the same half of the clay disc was analysed after each pretreatment. The samples were scanned from 2 degrees 2θ to 25 degrees 2θ using scan speeds of 1 degree per minutes or $\frac{1}{2}$ degree per minute and count rates of 1000 cps and 2000 cps. In addition to the

above, the region between 16 degrees 2 θ to 21 degrees 2 θ was scanned with a count rate of 400 cps in order to enhance and identify the (002)₁₀/(003)₁₇ reflection of the illite/smectite clay.

6.3 MINERALOGY OF THE SHALES

Kaolinite, illite, chlorite, illite/smectite mixed layer clay, quartz, and feldspars are the predominant minerals (though not necessarily in this order of abundance) in the Neogene shales of the Bengal Basin. Calcite, dolomite and siderite occur in minor to trace amounts. All the shales studied are non-pyritic. The bulk mineralogy of representative samples from the Muladi-1 well covering the total depth range studied is shown in table 6.1. Diffraction profiles of the whole rock of some of the representative shale samples are shown in figure 6.1 and 6.2.

6.3.1 NON-CLAY MINERALOGY:

(a) Quartz:

Quartz is one of the most abundant minerals in all the samples and is consistently present throughout the depth range studied.

(b) Feldspars: Potash and plagioclase feldspars are present in the samples throughout the depth range studied. The principal reflections used for identification of plagioclase feldspars is at 3.18 Å and for potash feldspar at 3.25 Å - 3.28 Å. Plagioclase is also indicated by moderate reflections at 4.02 - 4.03 Å, 3.75 - 3.767 Å and 3.66 Å. There has been no indication of any systematic change in the relative peak intensities of plagioclase or potash feldspar with depth of burial which would indicate a depth dependent change in their concentration.

TABLE 6.1 : BULK MINERALOGY OF REPRESENTATIVE SHALES FROM
THE MULADI-1 WELL, BENGAL BASIN

(Legend: ma=major, mi=minor, tr=trace, -=absent)

Sample no. (Depth-m)	Non-clay minerals					Clay minerals				
	Qtz	Feld	Calc	Dol	Sid	Ill	Kaol	Chlor	17Å Refl.	
						(Ran. Interstr. Ill/Smec)				
M1-14 (1402)	B O K A B I L	ma	ma	mi	tr	tr	ma	ma	ma	ma
M1-25 (2107)		ma	ma	mi	mi	tr	ma	ma	ma	ma
M1-30 (2402)		ma	ma	mi	tr	tr	ma	ma	ma	ma
M1-34A (2866)		ma	ma	mi	mi	tr	ma	ma	ma	ma
M1-35 (3020)		ma	ma	mi	tr	tr	ma	ma	ma	ma
M1-37A (3217)		ma	ma	mi	mi	tr	ma	ma	ma	mi
M1-40 (3502)		ma	ma	tr	-	tr	ma	ma	ma	mi
M1-41 (3602)	B H	ma	ma	-	tr	tr	ma	ma	ma	mi
M1-42 (3692)	U B	ma	ma	tr	tr	-	ma	ma	ma	-
M1-44 (3930)	A N	ma	ma	-	tr?	tr?	ma	ma	ma	-
M1-45 (4007)		ma	ma	-	tr	tr	ma	ma	ma	-
M1-46 (4099)		ma	ma	tr	tr	tr	ma	ma	ma	-
M1-47 (4205)		ma	ma	tr	tr?	tr?	ma	ma	ma	-
M1-49B (4458)		ma	ma	-	-	tr	ma	ma	ma	-
M1-51 (4530)		ma	ma	-	-	-	ma	ma	ma	-

DEGREES 2θ

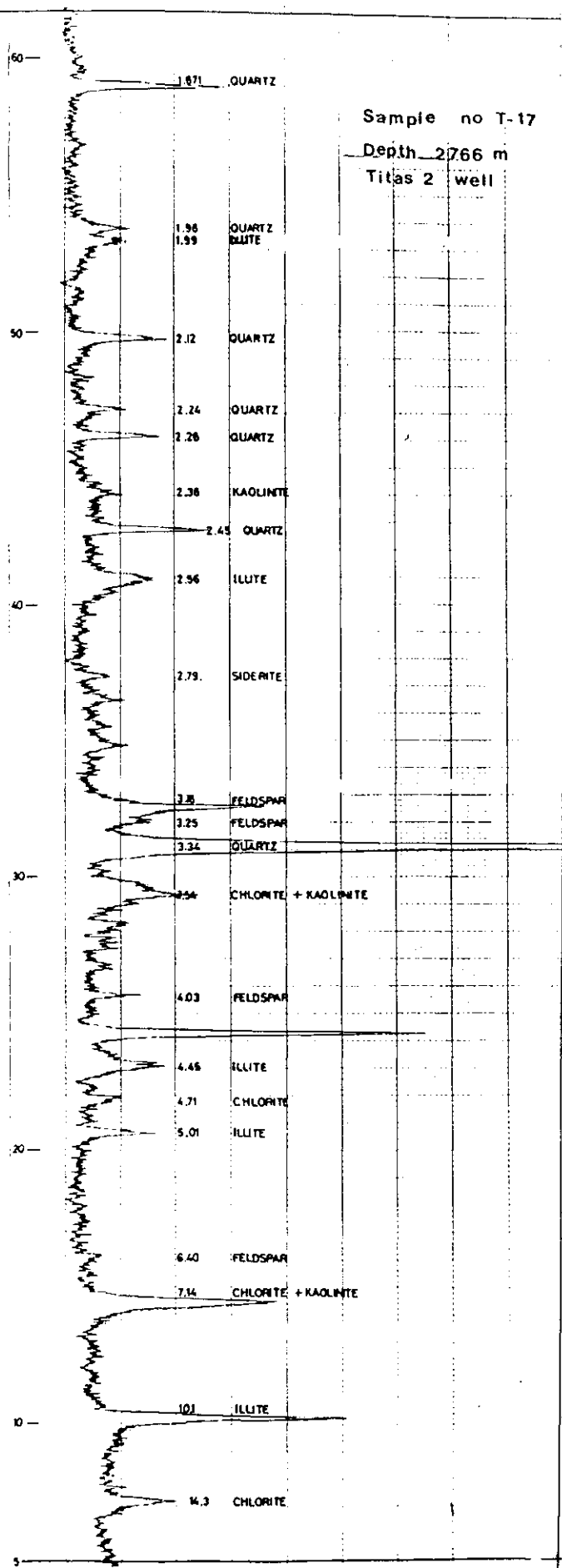
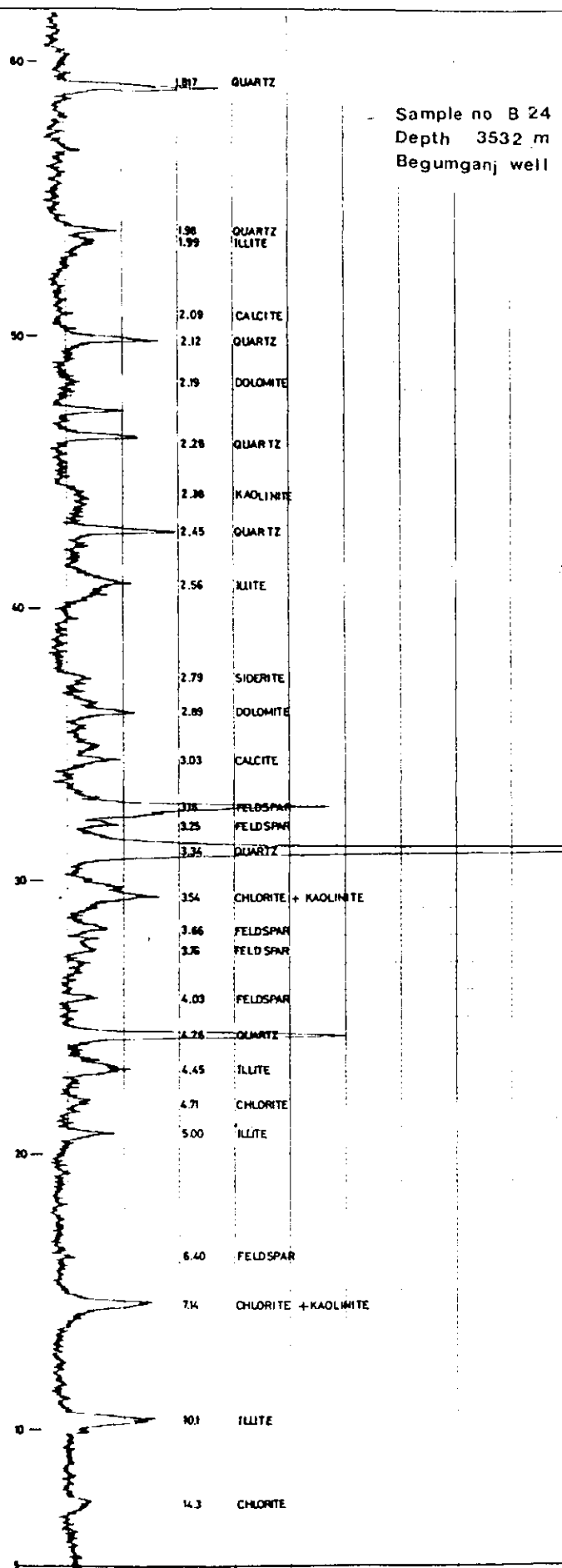


FIG.6.1 X RAY DIFFRACTION PATTERNS SHOWING BULK MINERALOGY OF REPRESENTATIVE NEOGENE SHALES FROM THE BENGAL BASIN. UNORIENTED MOUNT

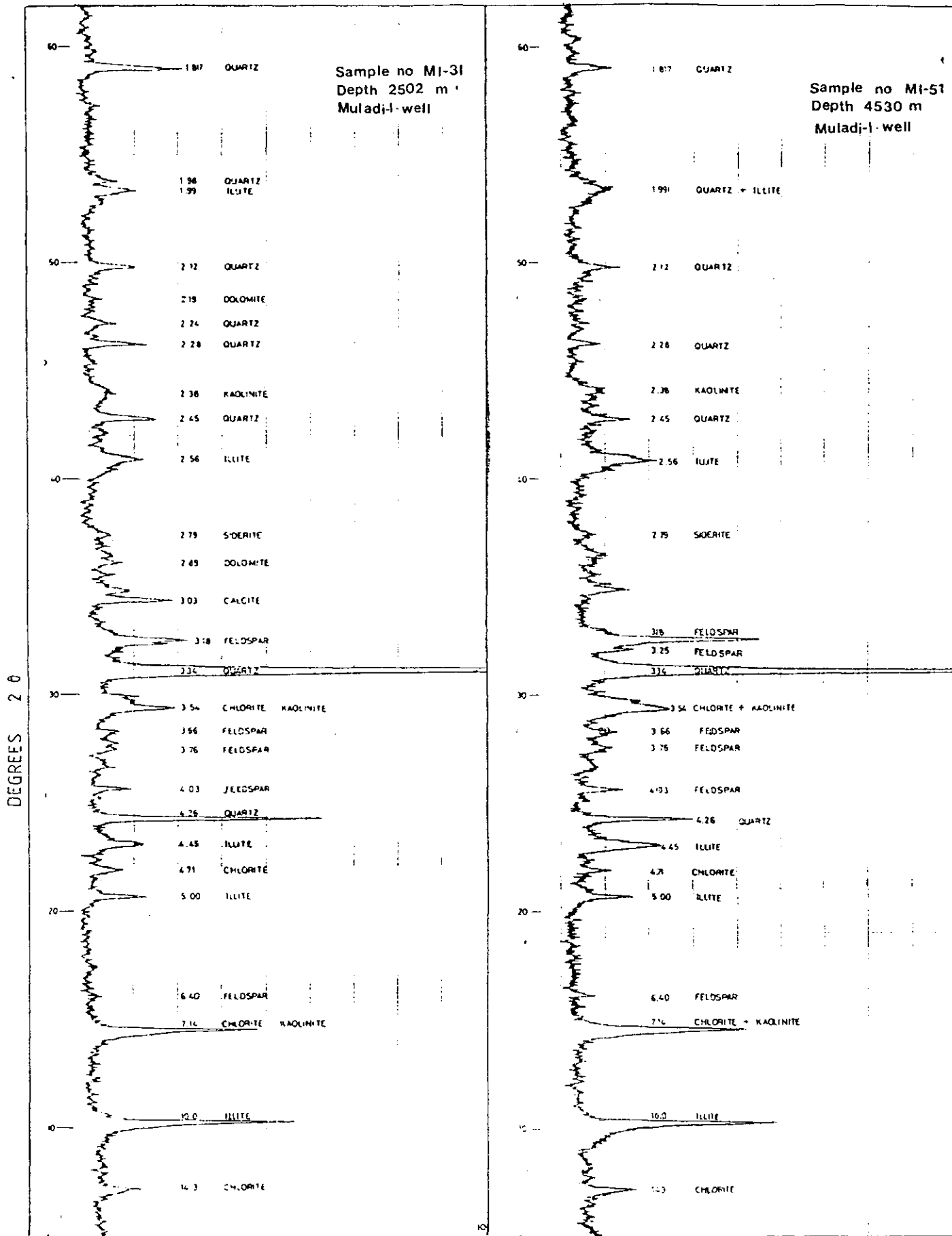


FIG.6.2.X-RAY DIFFRACTION PATTERNS SHOWING BULK MINERALOGY OF REPRESENTATIVE NEOGENE SHALES FROM THE BENGAL BASIN. UNORIENTED MOUNT.

(c) Calcite:

Calcite is a minor constituent of the shale down to a depth of about 3,500 m but below this depth calcite is present only in trace amount or absent indicating a decrease in abundance with depth (Table 6.1).

(d) Dolomite:

Dolomite is present in very minor to trace amounts in most of the shales.

(e) Siderite:

Siderite is present in trace amount in most of the samples throughout the sequence.

6.3.2 CLAY MINERALOGY

(a) Illite:

Illite is one of the most abundant phases in all the samples and is represented in the diffraction traces of oriented mounts by a series of prominent reflections at $9.97 - 10.1 \text{ \AA}$, $4.98 - 5 \text{ \AA}$, $3.3 \text{ \AA}$ and $1.98 - 1.99 \text{ \AA}$. There has been no change in the peak intensities with increasing depth of samples and uniform abundance of this mineral is indicated throughout the depth range.

(b) Chlorite:

Chlorite is a major phase in all the samples and is readily recognized by its basal reflections at $14.1 - 14.3 \text{ \AA}$ and $4.72 - 4.73 \text{ \AA}$. The basal reflections at $7.0 \text{ \AA}$ and $3.56 \text{ \AA}$ could not be used directly for identification of

chlorite because of the interference with kaolinite reflections. No change in chlorite peak intensities has been noticed with depth of burial which could be attributed to a change in concentration of this mineral.

(c) Kaolinite:

Kaolinite is a major constituent in the studied samples. The presence of kaolinite and chlorite together in all the samples made it difficult to recognise kaolinite on the basis of its major reflections at $\sim 7\text{\AA}$ and $3.58\text{\AA}$ (001 and 002 reflections respectively) in most of the samples because of the overlapping with chlorite reflections occurring in this region. In these cases the most reliable indication of kaolinite is the 003 reflection at $2.38\text{\AA}$, although this is a weak reflection. Leaching of chlorite with warm hydrochloric acid without affecting the kaolinite, however, allows the 7 $\text{\AA}$ reflections to be used for kaolinite study (Fig. 6.3) and to monitor the relative abundance of kaolinite with depth of burial. Such study indicates that the $7\text{\AA}$ reflection for kaolinite shows a relative decrease in intensity for the two deepest samples in the present study, that is at 4,530 m (Muladi-1 well) and 4,550 m (Muladi-2 well) (Fig. 6.4). This is interpreted as a decrease in abundance of kaolinite at or below 4,530 m of burial.

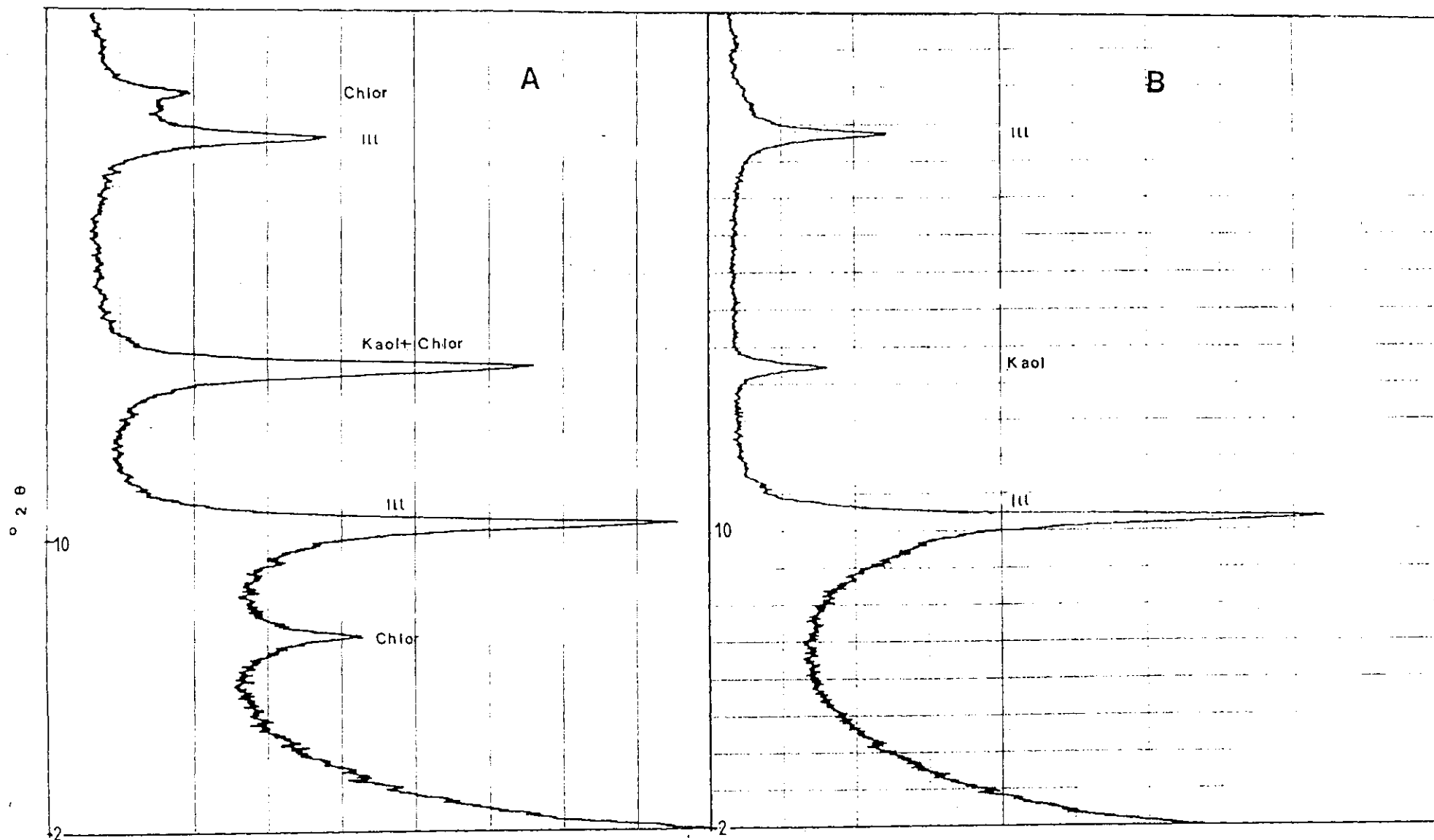


FIG.6.3 DIFFRACTION PROFILES OF CLAY FRACTION OF NEOGENE SHALE BEFORE (A) AND AFTER(B) WARM HCl ACID TREATMENT. CHLORITE IS DISSOLVED AWAY BY THE TREATMENT. SAMPLE FROM BHUBAN FM. DEPTH 3656m

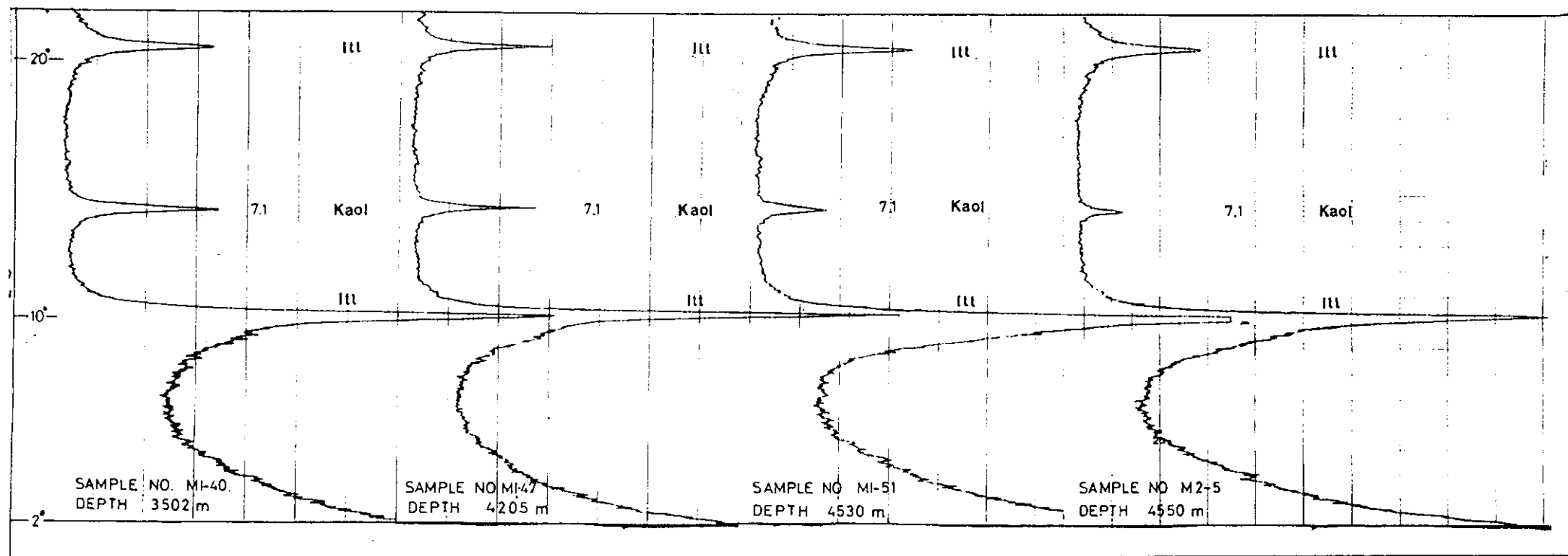


FIG. 6.4 DECREASE OF KAOLINITE WITH DEPTH IN NEOGENE SHALE.WARM HCl ACID TREATED SAMPLES FROM MULADI 1 WELL BENGAL BASIN.

(d) Illite/smectite mixed layer clay:

The illite/smectite mixed layer clay is represented in the present study by reflections at $16.8 - 17 \text{ \AA}$ (hereafter referred to as $17 \text{ \AA}$) for ethylene glycol treated samples and by reflections at $17.8 - 18.0 \text{ \AA}$ (hereafter referred to as $18 \text{ \AA}$) for glycerol treated ones (oriented mounts in both cases). These reflections disappear with simultaneous increase in the $10 \text{ \AA}$ reflections when the sample is heated to 300 degrees centigrade.

In the early publications, the $17 \text{ \AA}$ peak for an ethylene glycol treated sample was identified as discrete smectite and the mixed layer illite/smectites were thought to have basal reflections between $10 \text{ \AA}$ and $17 \text{ \AA}$ depending on their composition (Burst 1959, Burst 1969). However, Reynolds and Hower (1970) showed that a reflection at $17 \text{ \AA}$ after solvation with ethylene glycol is not sufficient to indicate the presence of discrete smectite. The $17 \text{ \AA}$ reflection is also given by randomly interstratified illite/smectite clays (with 40% or more smectite layers) for ethylene glycol solvated sample. The first basal reflection at $17 \text{ \AA}$ for a random mixed layer illite/smectite clay thus differ little from that of a discrete smectite (100% smectite), but the higher order reflections do differ (Reynolds and Hower 1970). The reflection designated as $(002)_{10} / (003)_{17}$, resulting from the 2nd basal order of the illite layers and the 3rd basal order of the smectite type layer is particularly important in this respect. For a discrete smectite this peak occurs at $5.62 \text{ \AA}$ but as the non-expanding illite type layer is incorporated or its proportion increases, the peak position is shifted toward $5 \text{ \AA}$, the 002 of discrete illite (Hower 1981) (Table 6.3). For most of the presently studied samples this $(002)_{10} / (003)_{17}$ reflection is too weak to be identified and measured, but where present (mostly shallower samples) it is found to be intermediate between

5.62 and 5.00 Å (Fig., 6.5). For the above reasons the 17 Å peaks for ethylene glycol treated samples (18 Å for glycerol treated ones) are interpreted in the present study as representing randomly interstratified illite/smectite mixed layer clay mineral.

The illite/smectite mixed layer clay phase is readily recognized as a major constituent in the shallow to moderately buried shales. The 17 Å reflections (18 Å for glycerol treated samples) with significant intensities are noted in the shallower samples from all the studied well sections, but the intensities gradually decrease with increasing depth of burial until they can no longer be identified as distinct reflections. This is followed at greater^{depth} by the complete absence of reflection in the 17 Å region (Fig. 6.6, 6.7 and 6.8). In the Muladi-1 well section the 17 Å reflection is present down to 3,602 m but is absent from all samples below this depth (Fig. 6.6). In the Begumganj-1 well the 17 Å reflection is shown for samples down to a depth of 3444 m but is absent for all samples studied below this depth (Fig. 6.7). In the BODC-2 well section the samples were pre-treated with glycerol and the 18 Å reflection is shown down to a depth of 4,082 m below which it is absent from all samples studied (Fig. 6.8).

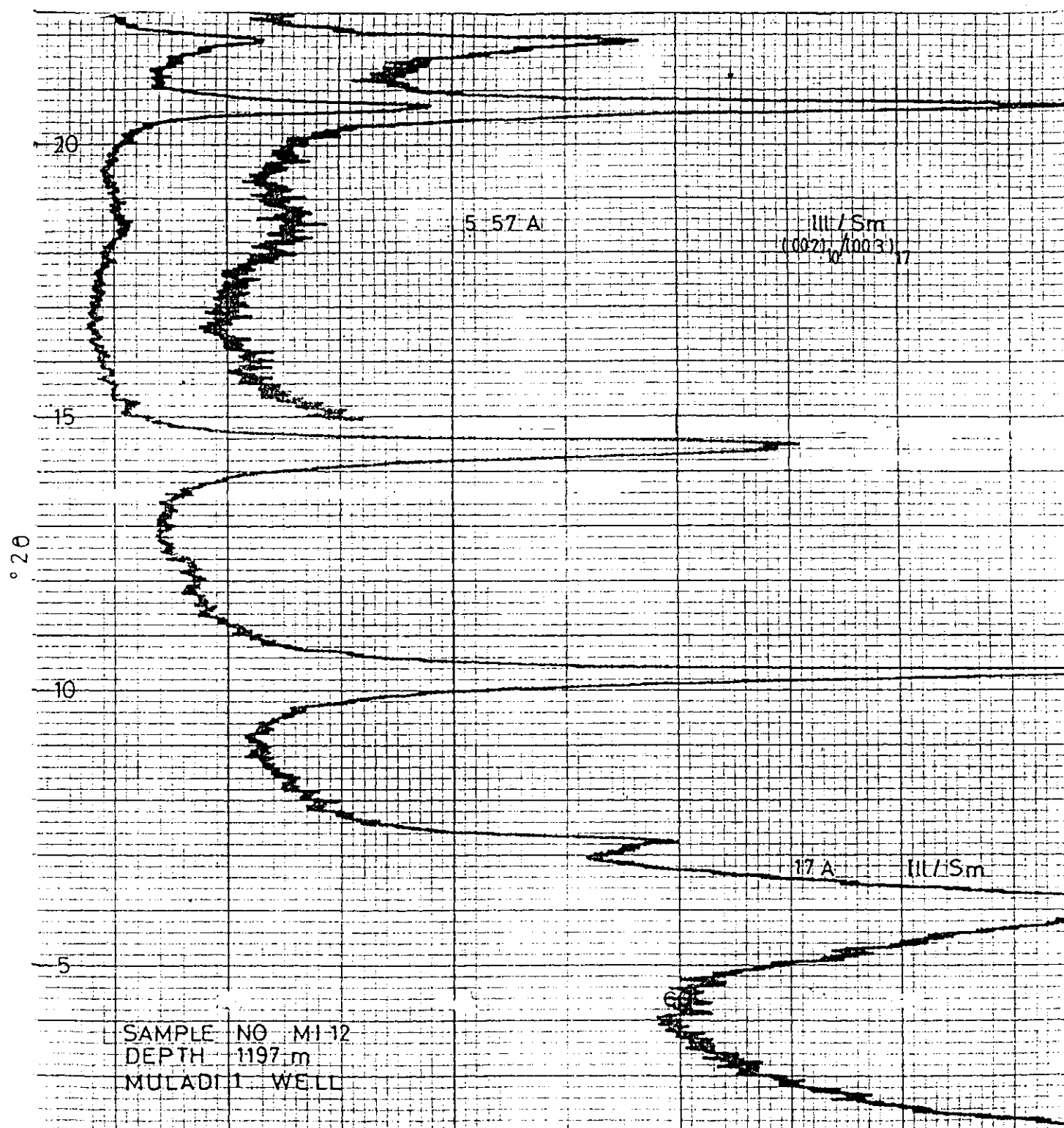


FIG.6.5. IDENTIFICATION OF ILLITE/SMECTITE MIXED LAYER CLAY. X RAY DIFFRACTION PROFILE OF CLAY FRACTION OF NEOGENE SHALE, BENGAL BASIN. SAMPLE SOVATED WITH ETHYLENE GLYCOL.

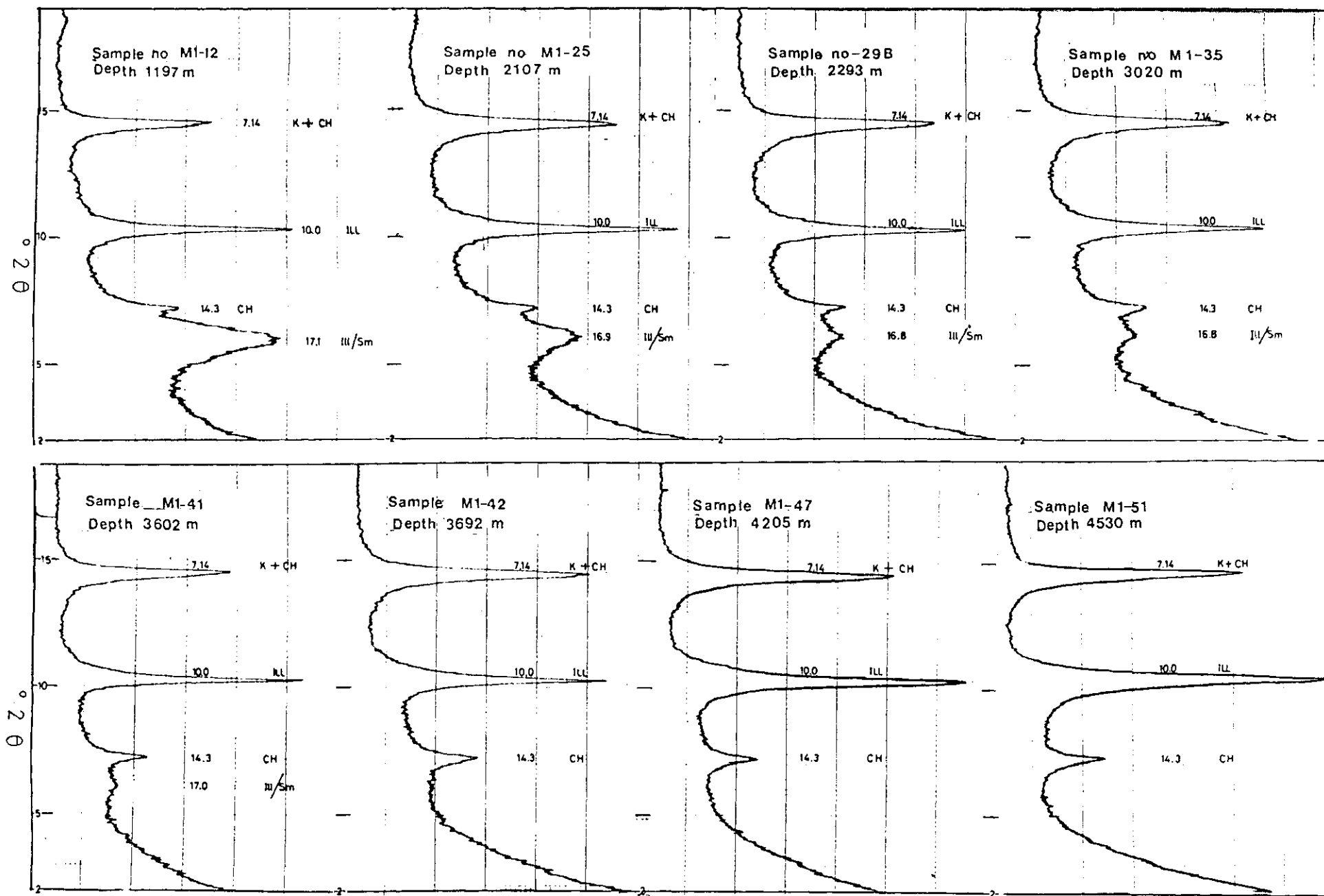


FIG.6.6 X RAY DIFFRACTION PROFILES OF CLAY FRACTIONS OF THE NEOGENE SHALES FROM MULADI-1 WELL BENGAL BASIN SHOWING GRADUAL LOSS OF 17Å PEAK INTENSITY WITH INCREASING DEPTH OF BURIAL. SAMPLES SOLVATED WITH ETHYLENE GLYCOL Legend : Ill/Sm-Illite/Smeectite Ch- Chlorite Ill- Illite K-Kaolinite

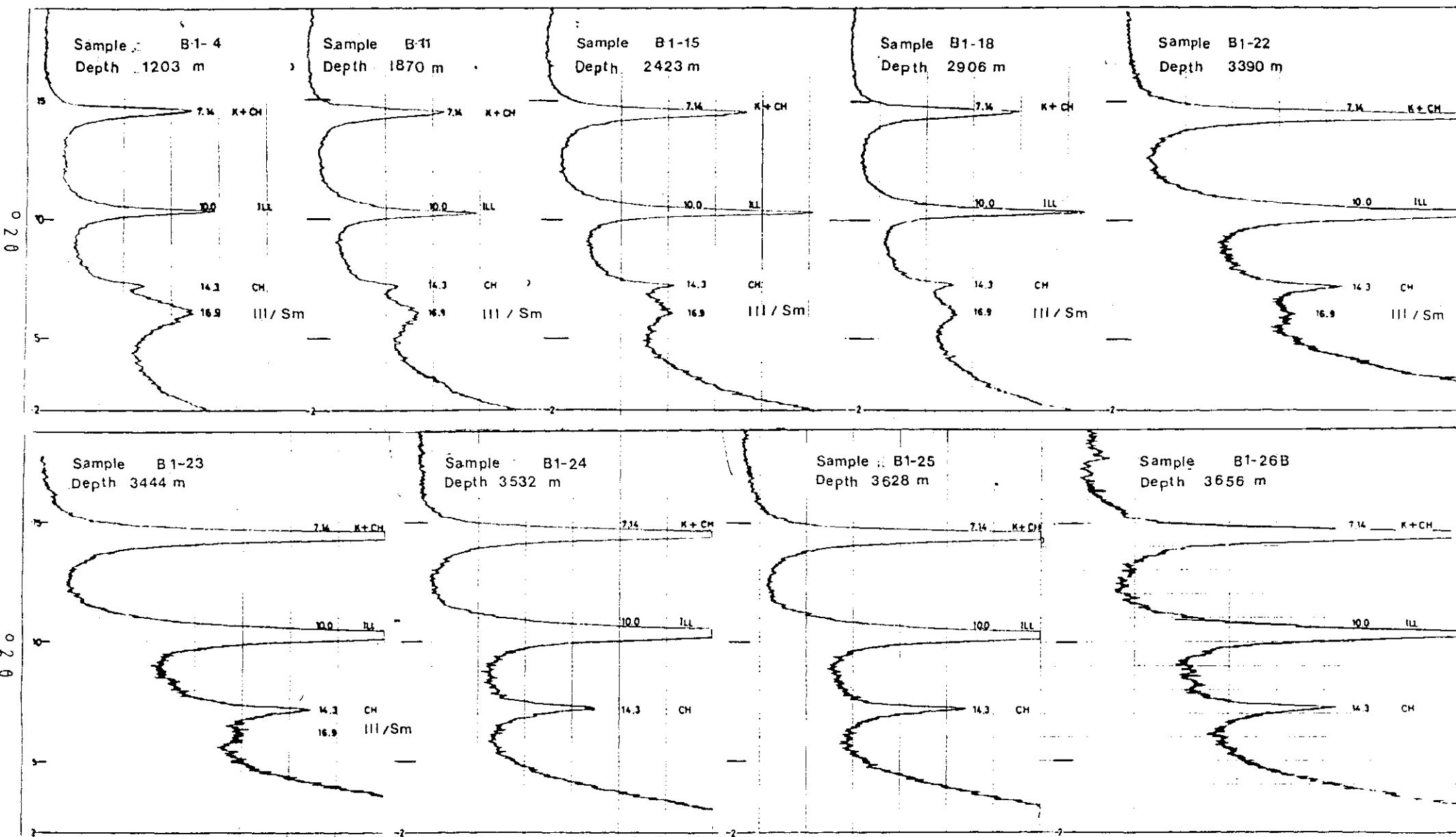


FIG. 6.7 X RAY DIFFRACTION PROFILES OF CLAY FRACTION OF THE NEOGENE SHALES FROM BEGUMGANJ 1 WELL BENGAL BASIN SHOWING GRADUAL LOSS OF 17Å PEAK INTENSITY WITH INCREASING DEPTH OF BURIAL. SAMPLES SOLVATED WITH ETHYLENE GLYCOL. Legend: Ill/Sm- Illite/Smectite, Ch- Chlorite, Ill- Illite, K-Kaolinite.

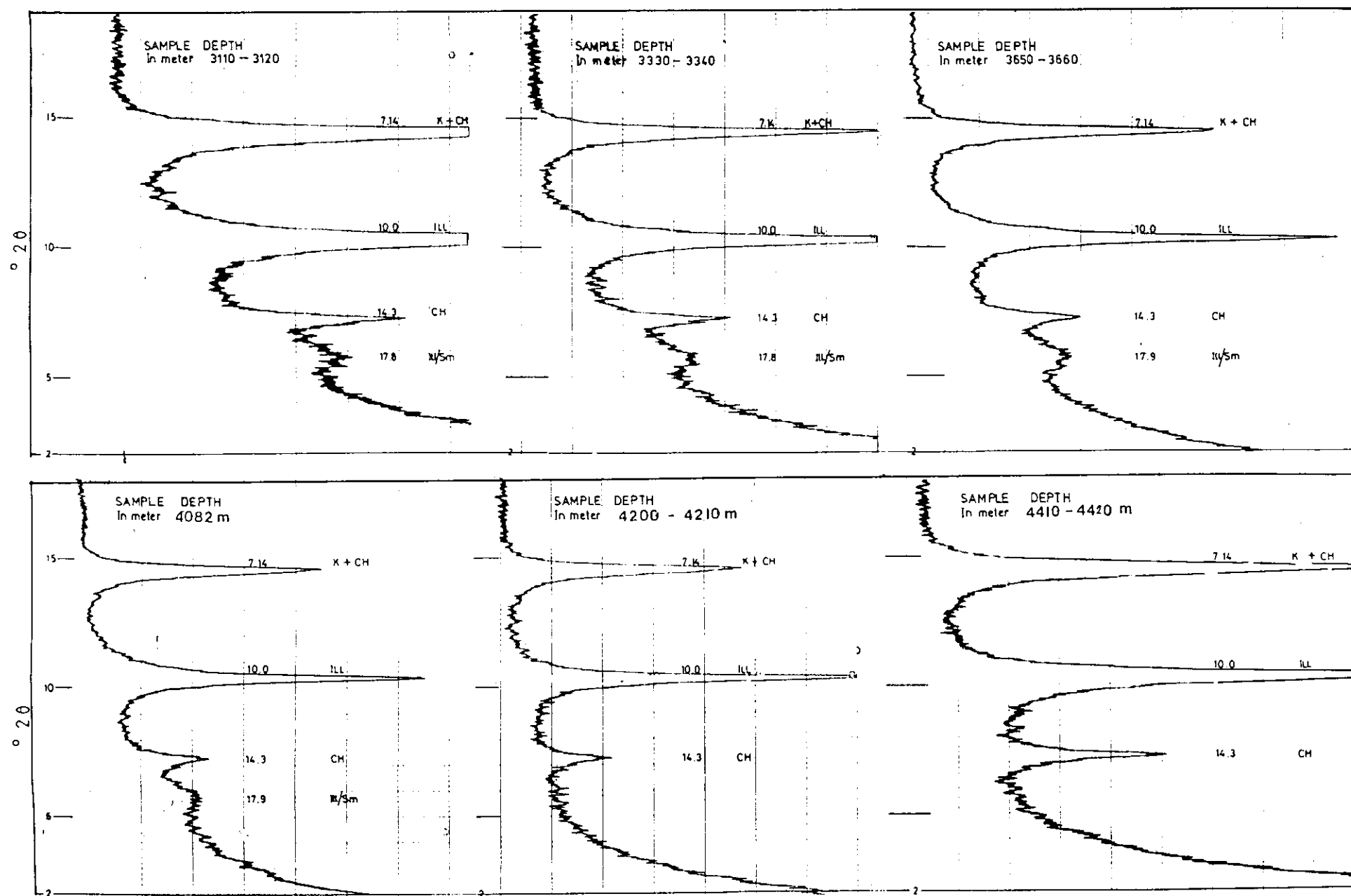


FIG.6.8 X RAY DIFFRACTION PROFILES OF CLAY FRACTIONS OF THE NEOGENE SHALES FROM BODC 2 WELL BENGAL BASIN SHOWING GRADUAL LOSS OF 18 Å PEAK INTENSITY WITH INCREASING DEPTH OF BURIAL. SAMPLES SOLVATED WITH GLYCEROL . Legend: Ill/Sm-Illite/Smectite, Ch-Chlorite, Ill-Illite, K-Kaolinite.

6.3.3 NATURE OF ILLITE/SMECTITE MIXED LAYER CLAY

The illite/smectite is volumetrically the most important mixed layer clay mineral in sedimentary rocks (Weaver 1956). This has also been the most widely studied group of clay minerals from the standpoint of diagenesis and there is significant evidence of diagenetic modification of these clays with increasing depth of burial (Burst 1969; Perry and Hower 1970; Dunoyer de Segonzac 1970; Hower et al. 1976; Hower 1981B). The diagenetic transformation of illite/smectite involves gradual loss of smectite layer due to conversion to illite with increasing temperature. This also results in a change in the nature of interstratification of the illite and smectite layers from a random to an ordered fashion (Perry and Hower 1970). In randomly interstratified illite/smectite no discernable pattern exists in the sequence of layer types while an ordered interstratification implies a periodicity in the sequence of layers. Reynolds and Hower (1970) indicated that the illite/smectite mineral with smectite of above 35% - 40% are almost always randomly interstratified and those of lower expandibilities (smectite %) have ordered interstratification. The compositional dependence of the manner of interstratification of illite/smectite is shown in Table 6.2 (Hower 1981A).

Various methods have been described for determination of the proportions of illite and smectite in the randomly interstratified illite/smectite mixed layer clay. Most of these methods, are however, affected by various other factors inherent in naturally occurring clays such as the mixture of illite/smectite with other clay minerals, variations of crystallite size of the mineral etc. It is therefore difficult to use most of these methods as an absolutely reliable measure of the composition of most naturally occurring illite/smectite clays and some degree

Ill/Sm composition	Ill/Sm Type
0~60% Illite	Random Ill/Sm
~65—70~85% Illite	*allevardite ordered Ill/Sm
~85—100% Illite	^{Kalkberg} ~~Kalkberg~~ ordered Ill/Sm
TABLE 6.2 COMPOSITIONAL DEPENDENCE OF MANNER OF INTERSTRATIFICATION OF ILLITE/SMECTITE . (From Hower 1981A)	

of error is likely to be incorporated in the result unless sufficient control over these factors can be established. In the presently studied samples illite/smectite occur with kaolinite, illite and chlorite clay minerals each of which is a major component of the shales and separation of these clay minerals is not possible. It was, therefore, decided to use three methods to determine the proportion of illite and smectite in the illite/smectite mixed layer clay, instead of relying on one, to determine any meaningful trends in the results.

These methods are:

- (a) $\sim 5\text{\AA}$ peak migration (Reynolds and Hower, 1970).
- (b) "Saddle"/ $17\text{\AA}$ intensity ratio (Rettke 1981).
- (c) $17:10\text{\AA}$ ratio (Schultz 1978).

Samples from the Muladi-1 well have been selected for this study because they cover the greatest depth range and also because of the more regular depth intervals of the available shales from this well.

(a) Peak migration method:

One of the most widely accepted methods to determine the composition of illite/smectite clay is to determine the $(002)_{10} / (003)_{17}$ peak position for glycolated clay sample (Reynolds and Hower 1970). An increase in the proportion of illite layers in the illite/smectite clays causes this peak to migrate towards $5\text{\AA}$. Reynolds (1980) tabulated the reflection positions for corresponding illite percentages in illite/smectite clay (Table 6.3). It is shown in the table that the $(002)_{10} / (003)_{17}$ peak migrates from $5.62\text{\AA}$ to $5.01\text{\AA}$ corresponding to an increase of illite layer from 0% to 100% in the illite/smectite mixed layer clay mineral.

TABLE 6.3 : RELATIONSHIPS BETWEEN COMPOSITION OF
ILLITE/SMECTITE MIXED LAYER CLAY AND
THE $(002)_{10} / (003)_{17}$ PEAK POSITION.
(After Reynolds 1980).

% Illite	% Smectite	d ($\text{\AA}$).
0	100	5.62
20	80	5.57
40	60	5.50
60	40	5.37
80	20	5.16
100	0	5.01

One significant limitation of applying this method to the present study was the very weak intensity of the $(002)_{10} / (003)_{17}$ reflections in most of the specimens so that the peak position could not be measured. Schultz (1978) also indicated in his study of the illite/smectite mixed layer clay in the Cretaceous Pierre Shale from U.S.A., that owing to the weak intensity of this reflections and some interference from other minerals, only about one tenth of the total number of samples could be analysed by this method.

In the present study of the Muladi-1 well samples only three samples show this reflection distinctly enough to be measured and used for the purpose of peak migration observation. It was observed that for sample at 1197 m this reflection appears at $5.57 \text{ \AA}$ and at 2,866 m this reflection appears at $5.50 \text{ \AA}$. Using Reynolds's's (1980) table (table 6.3), this is interpreted as a decrease in expandability in illite/smectite clay from 80% smectite at 1197 m to 65% smectite at 2866 m in the Muladi-1 well.

(b) Saddle/ $17 \text{ \AA}$ intensity ratio:

This method, based on Reynolds and Hower (1970), involves the measuring of the intensity ratio of $17 \text{ \AA}$ "peak" and the "Saddle" on the low angle side of the peak : the percentage of smectite is read from a working curve. The intensity of the "Valley" (Saddle) on the low angle side of the $17 \text{ \AA}$ peak relative to the intensity of the $17 \text{ \AA}$ peak is a function of the proportion of illite layers in the mixed layer illite/smectite clay. However, the "saddle to $17 \text{ \AA}$ peak intensity ratio is not only a function of illite: smectite proportion in illite/smectite clay, but it also depends on the crystallite size distribution of the clay minerals. (Srodon 1981, Hower 1981). The saddle:peak ratio is also different for a pure illite/smectite and for a mixture containing

illite/smectite (Srodon 1981). For the above reasons, this method allows only an approximate measure of the composition of the illite/smectite clay (Srodon 1981; Hower 1981).

Rettke (1981) made a simple modification of the original Reynolds and Hower (1970) method to include mechanical mixture of discrete illite in various proportions with illite/smectite clay. Rettke (1981) produced a graph of the saddle/17 $\overset{\circ}{\text{\AA}}$ intensity ratio as a function of illite/smectite expandibility for pure illite/smectite clay and for mixture of discrete illite with illite/smectite in 3:1, 1:1 and 1:3, illite:illite/smectite proportions (Fig. 6.9). The graph shows a gradual increase of saddle/17 $\overset{\circ}{\text{\AA}}$ ratio with a decrease in expandibility of the illite/smectite clay.

In the present study "saddle"/17 $\overset{\circ}{\text{\AA}}$ peak intensity ratio were measured for samples of the Muladi-1 well and the results are tabulated in table 6.4. From these data it can be seen that the "saddle"/17 ratio increases from 0.58 at 1197 m to 0.92 at 3,602 m. Assuming an illite:illite/smectite proportion of 3:1 for the present samples and using the curve prepared by Rettke (1981), such an increase in Saddle/17 $\overset{\circ}{\text{\AA}}$ ratio would correspond to decrease from about 75% to about 58% smectite layers in illite/smectite clay within the depth interval (Table 6.4).

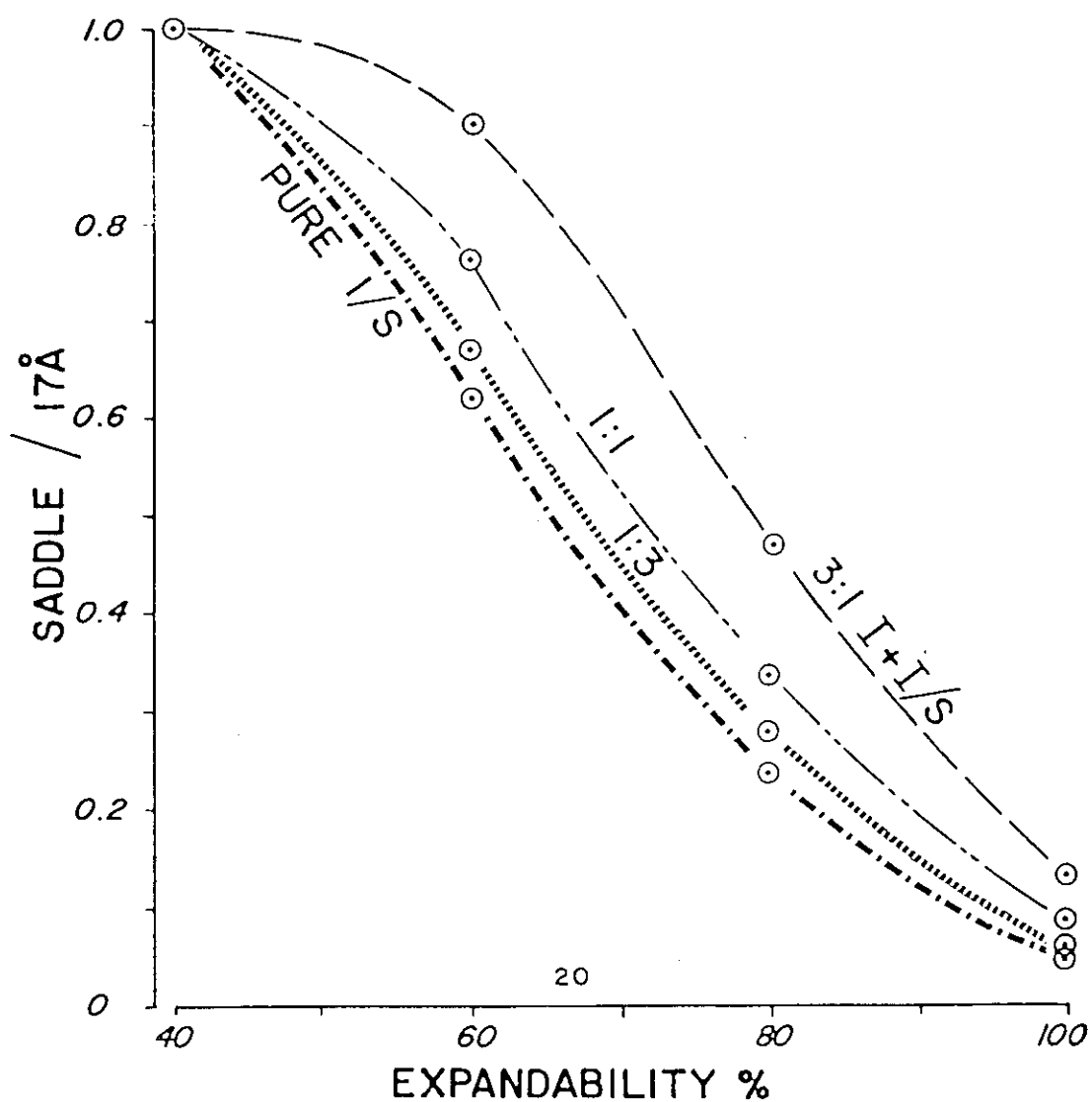


FIG.6.9 GRAPH OF 'SADDLE' / 17Å⁰ INTENSITY RATIO AS A FUNCTION OF ILLITE/SMECTITE EXPANDIBILITY (After Rettke 1981).

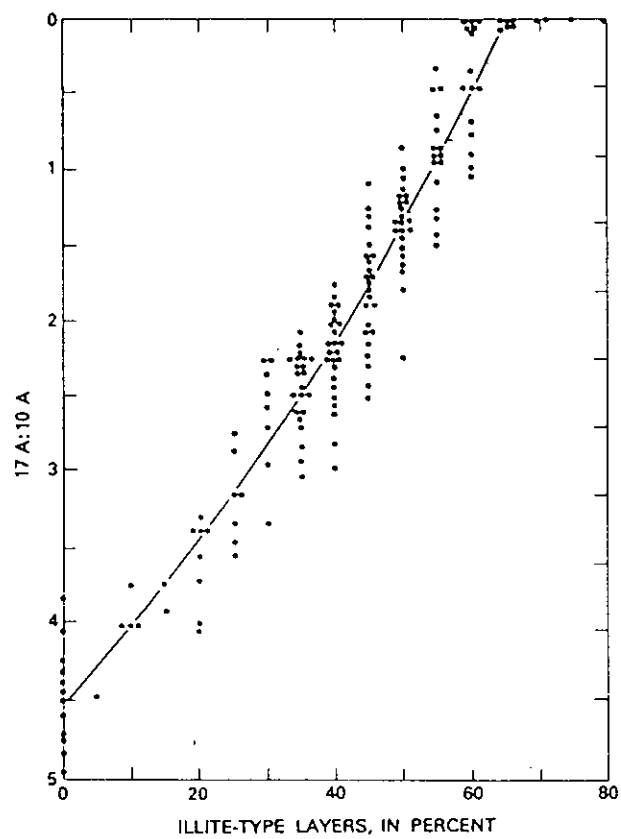
TABLE 6.4: ESTIMATION OF SMECTITE PERCENTAGES IN THE ILL/SMEC CLAY BY THE SADDLE/17A⁰ RATIO METHOD. SAMPLES ARE FROM MULADI-1 WELL, BENGAL BASIN.

Sample no.	Depth (m)	Saddle/17A ⁰ ratio	%Smectite
M1-12	1197	0.58	75
M1-14	1402	0.64	72
M1-22	1918	0.80	66
M1-25	2107	0.77	68
M1-29B	2293	0.82	65
M1-34A	2866	0.84	64
M1-35	3020	0.87	62.5
M1-37	3217	0.88	62
M1-41	3602	0.92	58
M1-42	3692	No 17A ⁰ reflection	

(c) $17:10 \text{ \AA}$ ratio method:

This method was proposed by Schultz (1978) and involved measuring the intensities of the basal $17 \text{ \AA}$ and the $10 \text{ \AA}$ reflections from glycolated and heated (300 degrees centigrade) samples respectively. Schultz (1978) produced a calibration curve showing the percentages of illite type layer in illite/smectite clay compared to the $17 \text{ \AA} : 10 \text{ \AA}$ ratio and showed that the ratio decreases from about 4.5 for completely expanding smectite to 0 where illite type layer make up 60% - 65% of the layers (Fig. 6.10). He concluded that the $17 \text{ \AA} : 10 \text{ \AA}$ ratio can be used in conjunction with the calibration curve to estimate the approximate proportion of the smectite and illite in the mixed layer phases. The $17 \text{ \AA} : 10 \text{ \AA}$ ratio is equivalent to $I(17\text{\AA})_{\text{glyc}}/I(10 \text{ \AA})_{300^\circ} - I(10 \text{ \AA})_{\text{gly}}$.

This method was applied to shale samples from the Muladi-1 well and the results are tabulated in table 6.5. The results show that the $17 \text{ \AA} / 10 \text{ \AA}$ ratio decrease from 1.32 at 1,197 m to 0.21 at 3,602 m. Using the curve produced by Schultz (1978) it is noted that such a decrease in the $17 \text{ \AA} : 10 \text{ \AA}$ ratio corresponds to a decrease of expandibility in illite/smectite clay from 50% smectite layer at 1197 m to 37% at 3,602 m .



17/100
 FIG 6.10 GRAPH OF 10:17 Å RATIO AS A FUNCTION OF
 ILLITE/SMECTITE EXPANDABILITY.
 (After Schultz 1978)

TABLE 6.5: ESTIMATION OF SMECTITE PERCENTAGES IN THE ILL/SMEC CLAY BY THE 17:10Å HEIGHT RATIO METHOD. SAMPLES ARE FROM MULADI-1 WELL, BENGAL BASIN.

Sample no.	Depth (m)	$17\overset{\circ}{\text{Å}}:10\overset{\circ}{\text{Å}}$ ratio	%Smectite
M1-12	1197	1.32	50
M1-14	1402	0.94	45
M1-22	1918	0.55	41
M1-25	2107	0.66	42
M1-29B	2293	0.41	39
M1-34A	2866	0.39	
M1-35	3020	0.27	38
M1-37A	3217	0.28	
M1-41	3602	0.21	37
M1-42	3692	No $17\overset{\circ}{\text{Å}}$ reflection	

Although the three methods used above produced different values of smectite percentage for respective samples, all show progressive decreases of smectite layer proportion in the illite/smectite mixed layer clay with increasing depth. It has been pointed out that these methods of measuring the smectite layer percentages in illite/smectite clay have certain limitations. These arises from the fact that in addition to the indigenous component layer composition in illite/smectite clay, the peaks for such clay could be affected by various other factors and that each method is likely to have some degree of error in its result. The saddle: $7 \overset{\circ}{\text{\AA}}$ peak intensity ratio depends strongly on the domain size distribution of clay minerals and is different for pure illite/smectite and for a mixture containing it (Rettke 1981, Srodon 1980). Similarly the $10:17 \overset{\circ}{\text{\AA}}$ ratio method give ideal measurement where the clay assemblages include only illite and illite/smectite clay (Schultz 1978). The main drawback of $\sim 5 \overset{\circ}{\text{\AA}}$ peak migration method in the very weak reflection of $(002)_{10} / (003)_{17}$ for which perfect measurement of peak position was not be always possible. It is therefore concluded that the significant geologic trend is the gradual reduction of smectite percentages in illite/smectite clay with burial depth; this is more important than the values obtained for individual samples in the present study.

6.4 DIAGENESIS OF SHALE:

6.4.1 Illite/smectite mixed layer clay:

The most important change with burial depth in the clay mineral assemblage of the Neogene shales of the Bengal Basin is shown in the diffraction profile of glycolated samples by a gradual decrease in $17 \text{ \AA}$ peak intensity ($18 \text{ \AA}$ for glycerated ones) of the illite/smectite clays and its apparent disappearance between 3,532 m (Begumganj-1 well section) to 4100 m (BODC-2 well section) of burial (Fig. 6.6, 6.7 and 6.8). There may be two possible causes for such a change in $17 \text{ \AA}$ (or the $18 \text{ \AA}$) reflection in these shales.

(a) A gradual decrease in the abundance of illite/smectite clay in the deeper shales during deposition in which case "absence" of $17 \text{ \AA}$ peak would correspond to an "absence" of illite/smectite clay in the original detrital input.

(b) A gradual diagenetic change of the illite/smectite mixed layer clay in response to increasing burial.

The possibility of a decrease in abundance or absence of illite/smectite clay in the original detrital input during deposition of deeper shales seems unlikely for the following reasons. (1) There has been no indication of a change of provenance during the Neogene period which could have produced a change in detrital clay mineral composition. Since the inception of the proto Ganges-Brahmaputra delta system in early Neogene times, there has been a uniform input of sediments from the uplifted Himalayan terrain in the north ; sedimentation has continued throughout the Neogene, building out and extending the delta southward. Therefore a major change in the amount or composition of clay minerals due to a change in provenance is unlikely. (2) A variation in composition

of the original detrital clay minerals deposited in various sub-environments of the prograding delta, could have been possible due to a physical sorting of clay minerals by size. That is to say, a differential sedimentation process could have caused a lateral change in clay minerals composition during deposition (Gibbs 1977). However, in such a case the smaller sized clay particles should be concentrated in the deeper part of the basin farther away from the shore. The illite/smectite mixed layer clay would be the smallest sized component of the clay fraction (Srodon 1981). In the Bengal Basin stratigraphy, for any given stratigraphic section, the more deeply buried shale would be generally deposited further offshore than the shallower shales. As a result deeply buried shale should have been richer in smaller sized clay minerals including illite/smectite mixed layer clay than the shallower ones. The absence of $17 \text{ \AA}$ reflections from the deeply buried shales of the Bengal Basin can not therefore be readily explained as a result of absence of illite/smectite clay minerals in the original detrital input due to the sedimentation processes according to the well established differential settling model (Gibbs 1977).

There is substantial evidence in the literature of the diagenetic modification of expanding clay minerals (Burst 1969; Perry and Hower 1970; Dunoyer de Segonzac 1970; Perry and Hower 1972; Hower et al. 1976; Hower 1981B, Boles and Frank 1979). These suggest that there will be a gradual decrease in the proportion of smectite in the illite/smectite clay due to transformation of smectite to illite layers with increasing depth of burial. The most distinctive character of the diffraction profile of the randomly interstratified illite/smectite (smectite layer constituting more than 40%) is that the $17 \text{ \AA}$ peak does not shift its position as illite layers are incorporated in the illite/smectite clay but the effect of an increase in

illite layers is that the $17\text{ \AA}$ peak intensity decrease progressively until it almost completely disappears (Hower 1981A, Schultz 1978).

From the evidence presented above it is concluded that the gradual loss of $17\text{ \AA}$ peak intensity in the diffraction profiles of the clay fractions of Neogene shale with increasing burial depth represents a gradual decrease in the percentages of smectite layers in illite/smectite clay due to the diagenetic transformation of smectite to illite. Study of the nature of illite/smectite clay carried out on the Muladi-1 well samples also shows a decrease in smectite percentages in illite/smectite clay with increasing depth. The three methods used to determine the compositions of the illite/smectite clays give different smectite percentages but a general decrease of smectite layers of about 13%-17% has been indicated over a depth interval from 1197 m to 3,602 m in the Muladi-1 well (Table 6.4 and 6.5). It is believed that the different values of smectite percentages in the illite/smectite clay obtained by these different methods is related to the inherent problem of applying these methods to a natural mixture of clay minerals and the lack of control on various parameters such as crystallite size distribution of each clay mineral components which also affect the calculations.

The gradual decrease in the intensity of the $17 \text{ \AA}$ ($18 \text{ \AA}$ for glycerated samples) reflections with depth is followed by a complete disappearance of this peak at greater depths. Such apparent absence of $17 \text{ \AA}$ reflections is noted at or below 3,692 m in the Muladi-1 well (Fig. 6.6), at or below 3532 m in the Begumgang-1 well (Fig. 6.7) and at or below 4,100 m in BODC-2 well (Fig. 6.8). There may be two possible reasons for the disappearance of $17 \text{ \AA}$ in a diagenetic sequence.

(1) Total loss of the smectite component due to complete illitization of illite/smectite clay.

Or

(2) Decrease of smectite percentages in illite/smectite clay to below a level at which the illite/smectite may not show a reflection at $17 \text{ \AA}$ but instead its position may be shifted toward higher angle (i.e. lower d-spacing). This may happen, in contrast to the randomly interstratified illite/smectite, when the illite layers are added in an ordered fashion (Hower 1981A). Perry and Hower (1970) showed that the interstratification in illite/smectite clay change from random to ordered fashion as the smectite layer percentage decrease to about 35% - 40% in response to the depth of burial (Fig. 6.11).

From the measurement of the smectite proportions in the illite/smectite clay mineral in the shales from Muladi-1 well, it seems unlikely that the gradual decrease of smectite percentages from 75% to 58% (Section 6.3.3 b) or from 50% to 37% (Section 6.3.3 c) within the depth interval of 1197 m to 3602 m would be followed by complete illitization of illite/smectite clay (total loss of smectite component) in the next sample at 3692 m where no

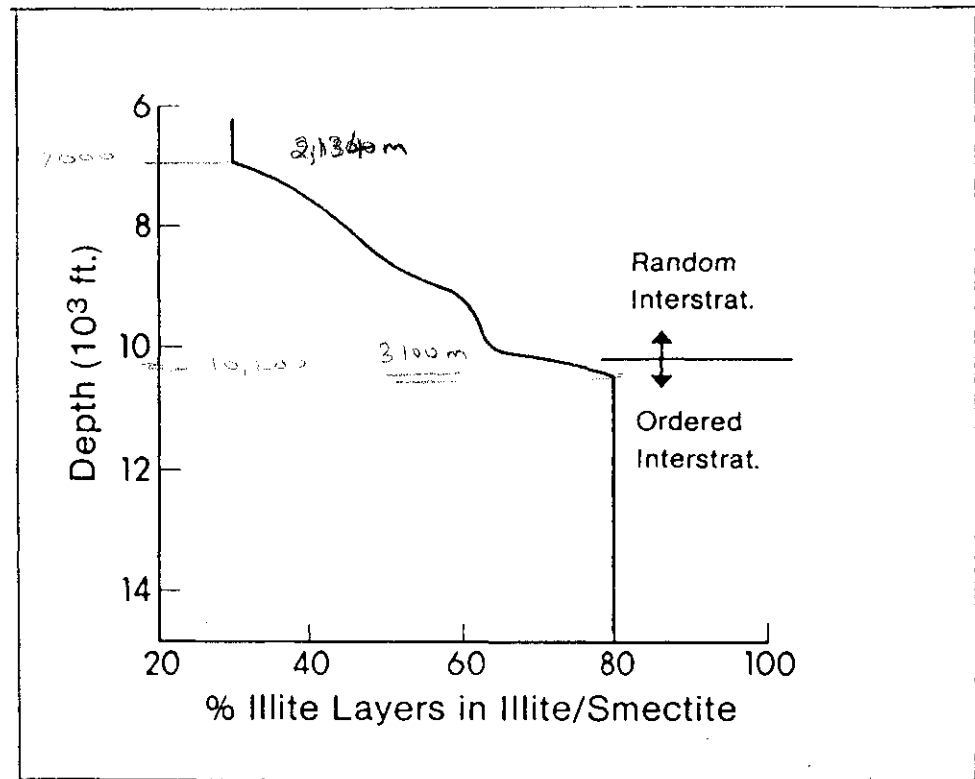


FIG.6.11 THE RELATIONSHIP BETWEEN % ILLITE LAYERS IN Ill/Sm AND DEPTH OF BURIAL IN GULF COAST BASIN, (% SMECTITE IN Ill/Sm = 100 - % Ill). HORIZONTAL LINE DEMARKS THE TRANSITION FROM RANDOM TO ORDERED INTERSTRATIFICATION.

(From Hower 1981)

17 Å reflection was present (Table 6.4 and 6.5). It is more likely that the proportion of smectite in illite/smectite clay in the depth levels where 17 Å reflection is not shown, has decreased below a level which resulted in a shift in 17 Å peak toward the higher angle as the random mixed layer became regular or ordered. The shifted peak, however, could not be seen or resolved presumably because of the presence of a reflection for chlorite at 14 Å under which it may be concealed.

6.4.2 Other Minerals

Two other minerals that show depth dependent variation in concentration are kaolinite and calcite. The decrease of kaolinite concentration at and below about 4,530 m of burial is indicated by relative decrease in 7 Å peak intensity of HCl acid leached samples at these depth levels (section 3.6.2 c; Fig. 6.4). The decrease in kaolinite concentration at and below 4,530 m may be diagenetic. However, it may also be due to lower abundance of Kaolinite in original detrital input of these deeply buried shales, deposited further offshore, as a result of the differential transport of clay minerals (Gibbs 1977).

Calcite concentration also show sharp decrease below about 3,500 m of burial, as evidenced by its absence or presence only in trace amount below this depth. Such a loss of calcite is beleived to be the result of diagenetic dissolution of this mineral with depth (Hower et al. 1976).

No change in the relative abundance of K-feldspar has been noted with depth of burial in the present study, similar to the one noted by several workers on the U.S. Gulf Coast Tertiary sequence (Hower et al '76 Boles & Frank '79).

6.4.3 DISCUSSION

The diagenesis of shales and especially of their clay mineral assemblages, has been a subject much studied in recent years. These studies have shown that the gradual loss of smectite due to conversion to illite in smectite or in illite/smectite mixed layer clay minerals is amongst the most important diagenetic changes that take place in shales with such expandable clay minerals (Dunoyer de Segonzac 1970; Perry and Hower 1970; Hower et al. 1976, Boles and Frank 1979; Hower 1981B). Perry and Hower (1970) showed a monotonic decrease in expandability in illite/smectite clay from about 80% to 20% smectite layer with increasing depth in U.S. Gulf Coast Tertiary sequence; they also indicated that the nature of interstratification in illite/smectite changes from random to ordered at about 35% smectite layer (Fig. 6.11). Such transformation are thought ultimately to lead to complete illitization of the smectite (via illite/smectite) or the illite/smectite mixed layer clay minerals. Among the other diagenetic changes that are commonly recorded in shales are decrease and disappearance of kaolinite (Dunoyer de Segonzac 1970; Boles and Frank 1979; Boles 1981; Shaw 1980), increase in the proportion of chlorite (Hower et al., 1976; Boles 1981), decrease and disappearance of calcite (Perry and Hower 1970; Boles 1981; Hower et al. 1976) and decrease of K-feldspar (Hower et al. 1976; Boles and Frank 1979; Boles 1981).

The conversion of smectite to illite would involve significant change in chemical composition among which fixation of K^+ in interlayer space of smectite is very important. Hower et al. (1976) suggested the following general reaction for such conversion: $\text{Smectite} + \text{Al}^{3+} + K^+ = \text{illite} + \text{Si}^{4+}$. The actual reaction is believed to be more complicated involving loss of H_2O , Ca^{++} , Na , Mg^{++} , and Fe^{++} (Hower et al. 1976; Boles and Frank 1979; Boles 1981). The

source of K⁺ for this change has variously been suggested as being via the breakdown of potassium feldspars (Hower et al. 1976; Boles and Frank 1979), breakdown of detrital mica (Perry and Hower 1970) or breakdown of discrete illite (Freed 1981). Temperature is the most important factor controlling clay diagenesis and decrease in expandibility (smectite proportion) in clay minerals seem to be more rapid in wells having higher geothermal gradients. Pressure and geologic age (reaction time) are less important (Perry and Hower 1972). Pore water chemistry (alkalinity and acidity) however, could also play vital role in controlling clay diagenesis (Shaw 1980). The conversion of smectite to illite appears to begin at about 60° C - 80° C (Boles 1981) and complete illitization of the illite/smectite clay takes place at temperatures between 130° C and 180° C (Shaw 1980). Hower et al. (1976), in their study of the U.S. Gulf coast Tertiary sequence reported that complete illitization of illite/smectite clay has not taken place at total depth of 5,770 m and a burial temperature of 175° C reached in the well. The temperatures at which kaolinite becomes unstable are variable (Dunoyer de Segonzac 1970) (Fig. 6.12). The stability of kaolinite is also a function of geochemical environment; with the increase of pH, for example, from acid to alkaline condition of the pore water the stability of kaolinite decrease (Shaw 1980).

In the Neogene shales of the Bengal Basin, illite/smectite mixed layer clays form a major group of clay mineral and there are distinct indications that the diagenesis has taken place involving gradual loss of the smectite with increasing depth of burial. This change can not be correlated with the formation boundaries or with the age of the sediments and seems to be totally depth dependent which in turn determines the temperature. The gradual decrease of the smectite proportion in the randomly interstratified illite/smectite clay is reflected by a

gradual decrease in the 17° Å (or 18° Å) peak intensity until it can no longer be identified. The total loss of this reflection takes place at greater depth (4100 m) in BODC-2 well than in the Begumganj-1 (at 3,532 m) and Muladi-1 (at 3,692 m) wells. This is believed to be due to lower geothermal gradient in the former well than the latter ones. The disappearance of 17° Å peak is not believed to be due to complete illitization of illite/smectite clay but is probably due to the reduction of smectite proportion below a level at which 17° Å is shifted toward higher angle (lower d-spacing) as the interstratification in illite/smectite clay change from random to an ordered fashion. The shifted peak probably could not, be detected or resolved because of the interference from chlorite peak at 14° Å under which it might be concealed. In the Muladi-1 well which has been studied in the greatest detail, the onset of smectite to illite conversion is noted at about 60° C. The greatest depth studied in Muladi-1 well (4,530 m) and the Begumganj-1 well (3,656 m) both have burial temperature of about 120° C - a temperature below that considered to be required for complete illitization of illite/smectite clay. This is in agreement with the above explanation.

The decrease in kaolinite is also noted around temperature of about 120° C in Muladi-1 well, that is in the deepest sample studied at 4,530 m and may be diagenetic. There is no indication of a change in either K-feldspar, plagioclase feldspar, illite, or chlorite with depth which could be attributed to a diagenetic breakdown of these minerals and the source of K⁺ for the smectite to illite transformation is not apparent in the present study. The decrease and disappearance of calcite with depth results from the diagenetic dissolution of this mineral.

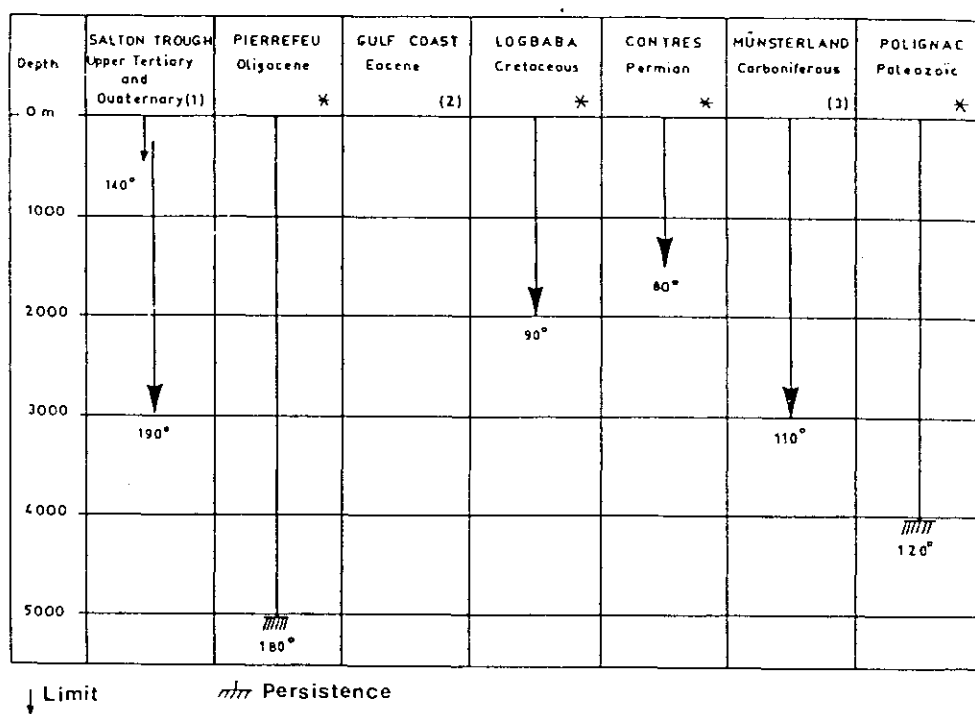


FIG. 6.12 VARIABLE THERMAL STABILITY OF KAOLINITE
(After Dunoyer de Segonzac 1970)

CHAPTER 7

CONCLUSIONS

Progressive diagenesis of the Neogene deltaic clastic sediments of the Bengal Basin has taken place with increasing depth of burial. The diagenetic sequences of the sandstones and the shales, as shown by the present study correspond to a simple burial history.

7.1 SANDSTONE DIAGENESIS

The following three aspects of sandstone diagenesis have been high-lighted by the present study.

(a) STAGES OF DIAGENESIS:

The diagenesis of the sandstones in the Neogene sequence may be divided into two stages according to the depth of burial as: (I) Shallow subsurface stage (0 to 2000 m), and (II) Deep subsurface stage (2000 m to 4491 m +).

(I) The Shallow subsurface stage of diagenesis (0-2000 m) is generally represented by the friable and poorly consolidated sandstones of the Tipam Group and the top of the Bokabil Formation. Except for the very few lenses of sandstones with early poikilotopic calcite cementation, the formation of authigenic phases in this stage plays insignificant or no role in the lithification of the sandstone. Chlorite is the only common authigenic mineral. Siderite is also present but only in insignificant amounts. Quartz overgrowths are extremely rare and no growth of late calcite (or dolomite) or kaolinite has been noted. Physical compaction is the only diagenetic process

which generally reduces the porosity, although to minor extent. Thin section porosity of the sandstone in this stage is reduced from 22% at 500 m to 20% at 1137 m and to 16% at 1700 m (Fig. 7.1).

(II) The Deep Subsurface stage of diagenesis (>2000 m) is generally represented by the moderately to highly indurated sandstones of the Bokabil Formation (lower part) and the Bhuban Formation. This stage is characterised by the active role of chemical diagenetic processes in the lithification of the sandstones. Though physical processes of compaction continue during this stage, chemical processes of diagenesis (cementation) become the major controlling factors in determining the reservoir properties of the sandstones. Quartz overgrowths and late calcite cements are common in most samples and are developed in significant amounts in many samples significantly reducing porosity and permeability. Authigenic kaolinite also occurs as a pore filling cement. The physico-chemical process of pressure solution also becomes important in this stage. Pressure solution of quartz is common to very common below about 2,758 m (Bhuban formation) and this contributes to increased lithification of sandstones both by overly tight compaction and by acting as a source of silica for quartz overgrowths. One other important feature of this stage of diagenesis is the development of secondary porosity below about 2,000 m by the partial dissolution of detrital feldspars and some lithics, as well as authigenic calcite cements. However, generally, the thin section porosity of the sandstones is reduced from 11% at 2216 m to about 4% at 4,491 m in this stage (Fig. 7.1).

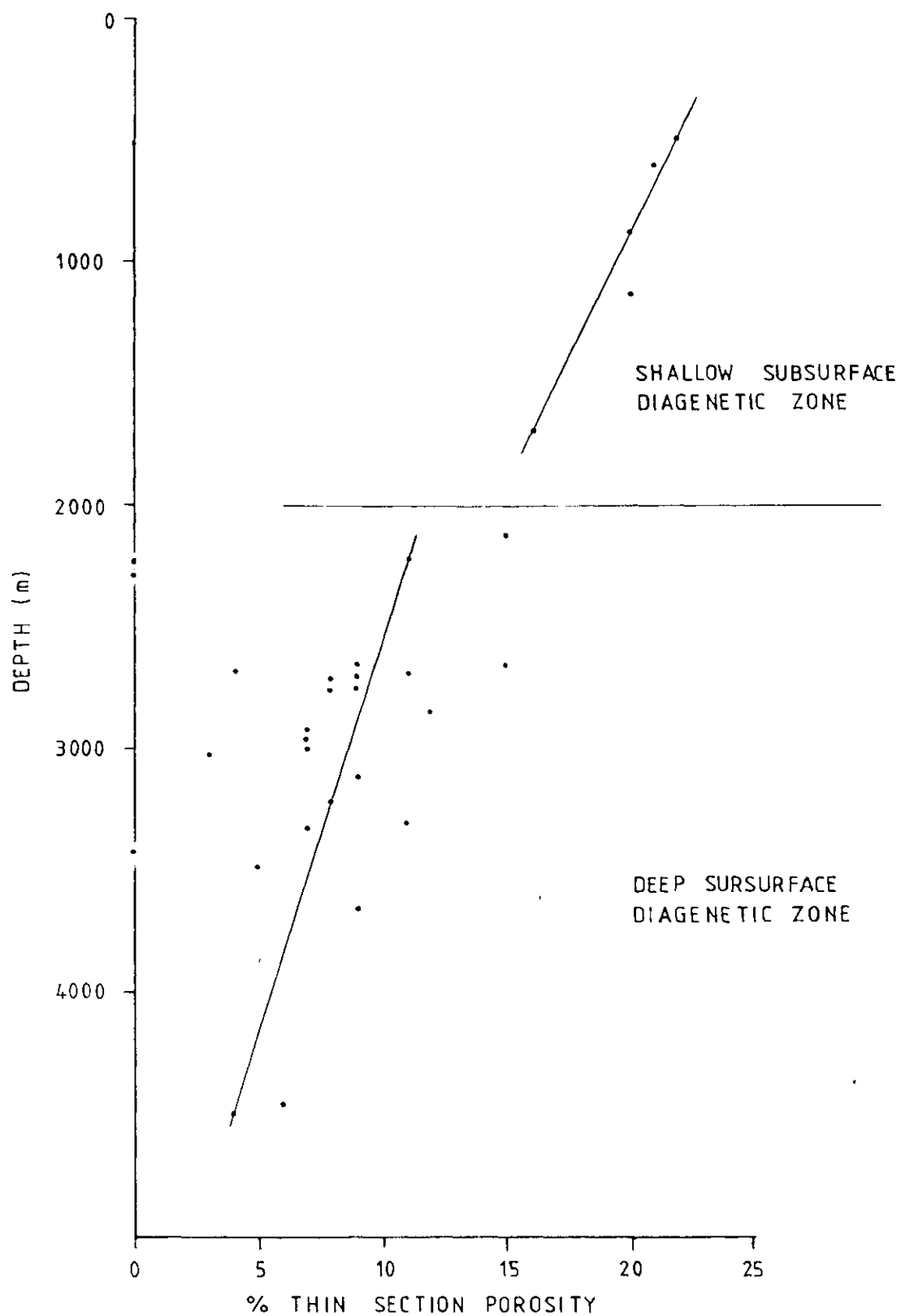


FIG. 7.1 DECREASE OF POROSITY WITH DEPTH OF BURIAL IN THE NEOGENE SANDSTONES OF THE BENGAL BASIN.

(b) PARAGENETIC SEQUENCE AND PORE WATER CHEMISTRY:

The paragenetic sequence of the sandstone diagenetic events suggests that the formation pore water chemistry has passed through two phases: alkaline and acidic (Fig. 7.2). The primary alkaline condition of the pore fluids was established with the initial burial, under which chlorite authigenesis, siderite crystallization, early Fe-calcite cementation, quartz overgrowth cementation, and late Fe-calcite cementation and replacement took place. Later in the diagenetic history, the alkaline pore water was replaced by an acidic pore water environment. This led to the partial dissolution of carbonate cement and detrital feldspars (with minor lithic grains). Acidic condition of the pore water persisted and authigenic kaolinite was then developed.

The establishment of acidic pore water environments by flushing the subsurface units (Bokabil and Bhuban formations) by meteoric water is considered unlikely because: (1) No meteoric water flow connection exists or is believed to have existed with these units after deep burial. (2) These units have never been significantly uplifted and subaerally exposed following deep burial. It is more likely that the establishment of acidic water conditions is related to the vertical migration of fluids from below. Such acidic formation waters may have been generated by dissolution of CO_2 evolved through the maturation of organic matter in the underlying source rock.

FIG.7.2. DIAGENETIC PROCESSES AND THE PORE WATER
CONDITIONS IN THE NEOGENE SANDSTONES OF
THE BENGAL BASIN.

Diagenetic Processes	Pore water condition
Chlorite authigenesis	ALKALINE
Siderite crystallization	
Early Fe-calcite cementation	
Illite coating	
Quartz overgrowth cementation	
Late Fe-calcite cementation	
Partial dissolution of rock components	ACIDIC
Kaolinite crystallization	

(c) INFLUENCE ON RESERVOIR PROPERTIES:

The diagenetic processes in the Neogene sandstones can be classified into three types with respect to their influence on reservoir properties. These are:

(1) Porosity reducing processes :

- (i) Early poikilotopic calcite cementation
- (ii) Quartz overgrowth cementation
- (iii) Late carbonate cementation
- (iv) Kaolinite cement crystallization
- (v) Compaction.

(2) Porosity preserving process:

Chlorite rim formation

(3) Porosity enhancing process:

Partial dissolution of some rock components and development of secondary porosity.

Early poikilotopic calcite cement is the most effective in destroying porosity and permeability. Though it affects only a few samples it reduces porosity and permeability to zero. Late Fe-calcite (with minor dolomite) is the second most important porosity reducing cement. Quartz overgrowths, common in most of the sandstones in the deep subsurface diagenetic stage, act as a significant porosity and permeability reducing agent where they are sufficiently abundant to develop interlocking texture. Kaolinite also reduces porosity but to a lesser extent. It probably reduces a significant, if not identical, volume of porosity that had been secondarily produced by feldspar dissolution. The effect of compaction in porosity reduction (principally by grain deformation and pressure

solution) increases with burial depth. Thick chlorite rims act as a porosity preserving agent by effectively inhibiting quartz overgrowth formation. Partial dissolution of detrital feldspars, lithic grains and authigenic calcite cements leads to the development of secondary porosity in the deep subsurface diagenetic stage, thus increasing the total porosity of the sandstones.

7.2 SHALE DIAGENESIS:

The diagenesis of the shale in the Neogene sequence is principally reflected in the change in the mixed layer illite/smectite clay mineral composition with increasing depth of burial. Progressive decrease in the proportion of smectite (expandable component) in the mixed layer illite/smectite clay with increasing depth has resulted from transformation of smectite to illite layers. This is recorded in the diffraction profiles of the clay fractions in all the Neogene well sections by a gradual loss of $17\text{\AA}$ glycol peak intensity until it is no longer detected. The diagenesis has taken place without respect to formation boundaries and mainly depends on temperature, which in turn is determined by the geothermal gradient. The greater depths at which the expandable clay component disappear below their detection level in the BODC-2 wells as compared to the Muladi-1 and Begumganj-1 wells is believed to be related to lower geothermal gradient in the former well relative to the latter wells. The diagenetic conversion of smectite to illite is not believed to have led to a complete illitization of the illite/smectite mixed layer clay mineral at the total depth studied, but the apparent disappearance of the reflection from the expandable component possibly represents a shift of this reflection toward higher angle which could not be detected because of interference with the chlorite peak. There is also a decrease in the abundance of kaolinite at and below

about 4530 m of burial which is probably diagenetic. The shale sequence also shows a decrease in the abundance of calcite via diagenetic dissolution below about 3,500. No depth dependent change has been noted in the illite, chlorite, feldspars or any other mineral content in the Neogene shale sequence.

The shale diagenesis in the Neogene sequence of the Bengal Basin may have important effects on the diagenesis of the interbedded sandstones, especially in the generation of authigenic cements. The Si^{+4} released into solution as a result of the smectite to illite transformation in the mixed layer illite/smectite clay may have provided a source of silica cement for the adjacent sandstones. The Ca^{++} that may also be released from this diagenetic reaction could be a source for the late calcite precipitated in the sandstones. The dissolution and transfer of calcite from shale to sandstone is also a likely source of calcite cement in the sandstones. Mass transfer of these cementing materials from shale to sandstone is believed to have taken place with the expelled water during shale compaction. The importance of the role of shale diagenesis in sandstone cementation is supported by the fact that the zone of sandstone cementation lies below about 2,000 m in the lower part of Bokabil formation and the Bhuban formation which are also characterised by an increased shale:sand ratio and where an increased degree of shale diagenesis has taken place (Fig. 7.3).

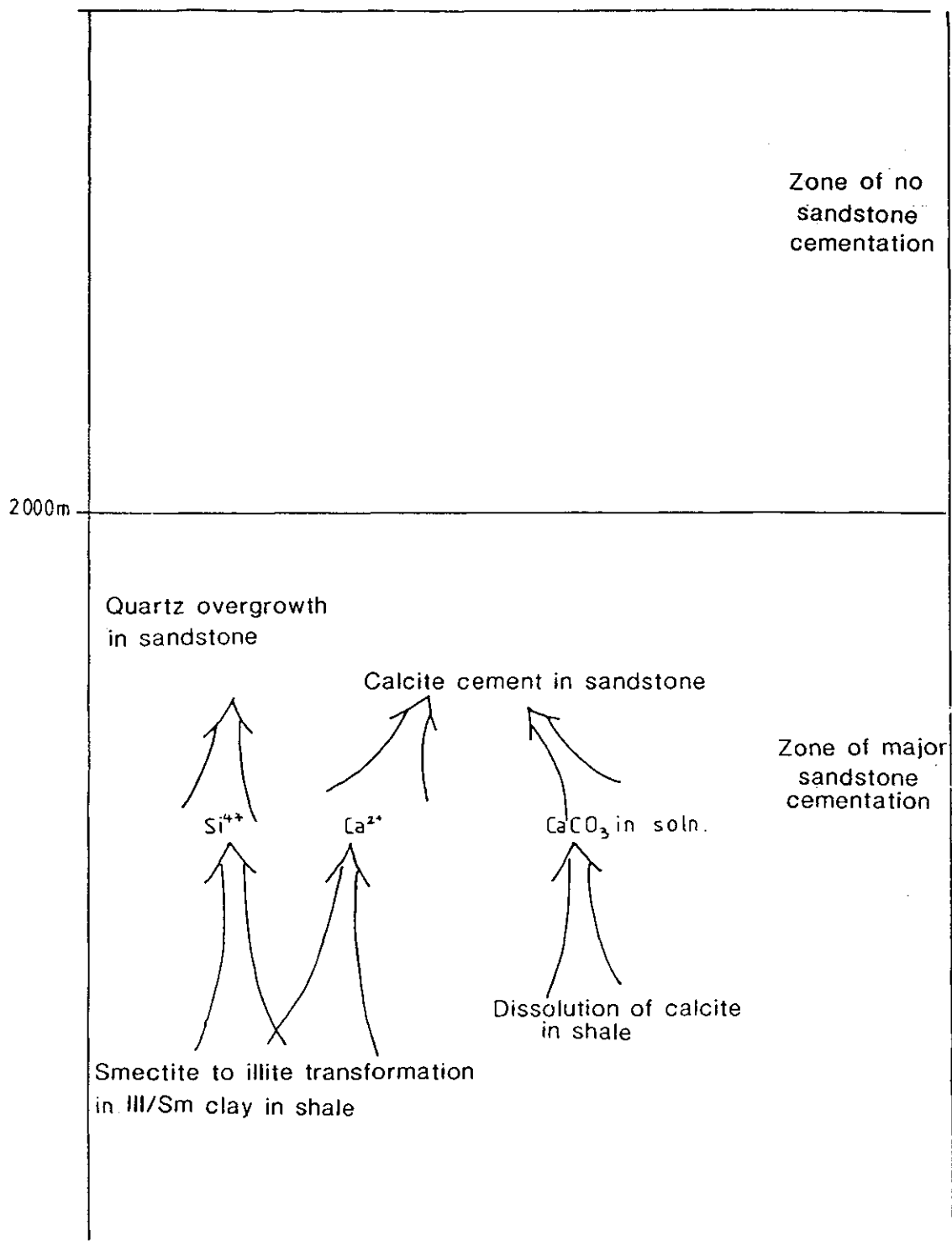


FIG. 7.3 SCHEMATIC DIAGRAM OF THE INFLUENCE OF SHALE DIAGENESIS ON THE CEMENTATION OF ADJACENT SANDSTONES IN THE NEOGENE SEQUENCE OF THE BENGAL BASIN

7.3 COMPARISON WITH OTHER DIAGENETIC MODELS

Several accounts of the diagenesis of deltaic clastic rock sequences are available in the literature. The U.S. Gulf Coast Tertiary sequence is probably the most studied area in this respect.

In Table 7.1, a comparison is shown of the sandstone diagenetic model (paragenetic sequence) of the presently studied Neogene sandstones of the Bengal Basin with those of the Tertiary deltaic sandstones of the Gulf Coast Basin, U.S.A. (Loucks et al. 1977; Stanton 1977), the Tertiary deltaic sandstone of Niger Delta (Lambert 1981), the Pennsylvanian deltaic sandstone of North Central Texas, U.S.A. (Land and Dutton, 1978), and the Cretaceous shelf sandstones which accumulated south of a delta in the Denver Basin, Colorado, U.S.A. (Porter and Weimer 1982). Except for the Tertiary Niger delta, several common aspects of the sandstone diagenetic models in these deltaic settings can be noted. Chlorite rim formation on the detrital grains is recognized as a very early diagenetic feature in most of these sandstones and quartz overgrowths always post-date the chlorite rims. Late calcite cementation and replacement invariably occur after the quartz overgrowth formation, followed by a phase of partial dissolution of calcite cement, detrital feldspars and some lithic grains. This in turn is followed by kaolinite precipitation in all the sequences. A three fold subdivision of the diagenetic history of these sandstones can generally be recognized:

- (1) a phase of cementation and porosity reduction
- (2) a phase of rock component dissolution and secondary porosity development
- (3) a phase of recementation.

TABLE 7.1 : COMPARISON OF THE PARAGENETIC SEQUENCE OF THE DIAGENETIC EVENTS IN THE NEOGENE SANDSTONES OF THE BENGAL BASIN WITH THOSE OF SOME OTHER DELTAIC SANDSTONES.

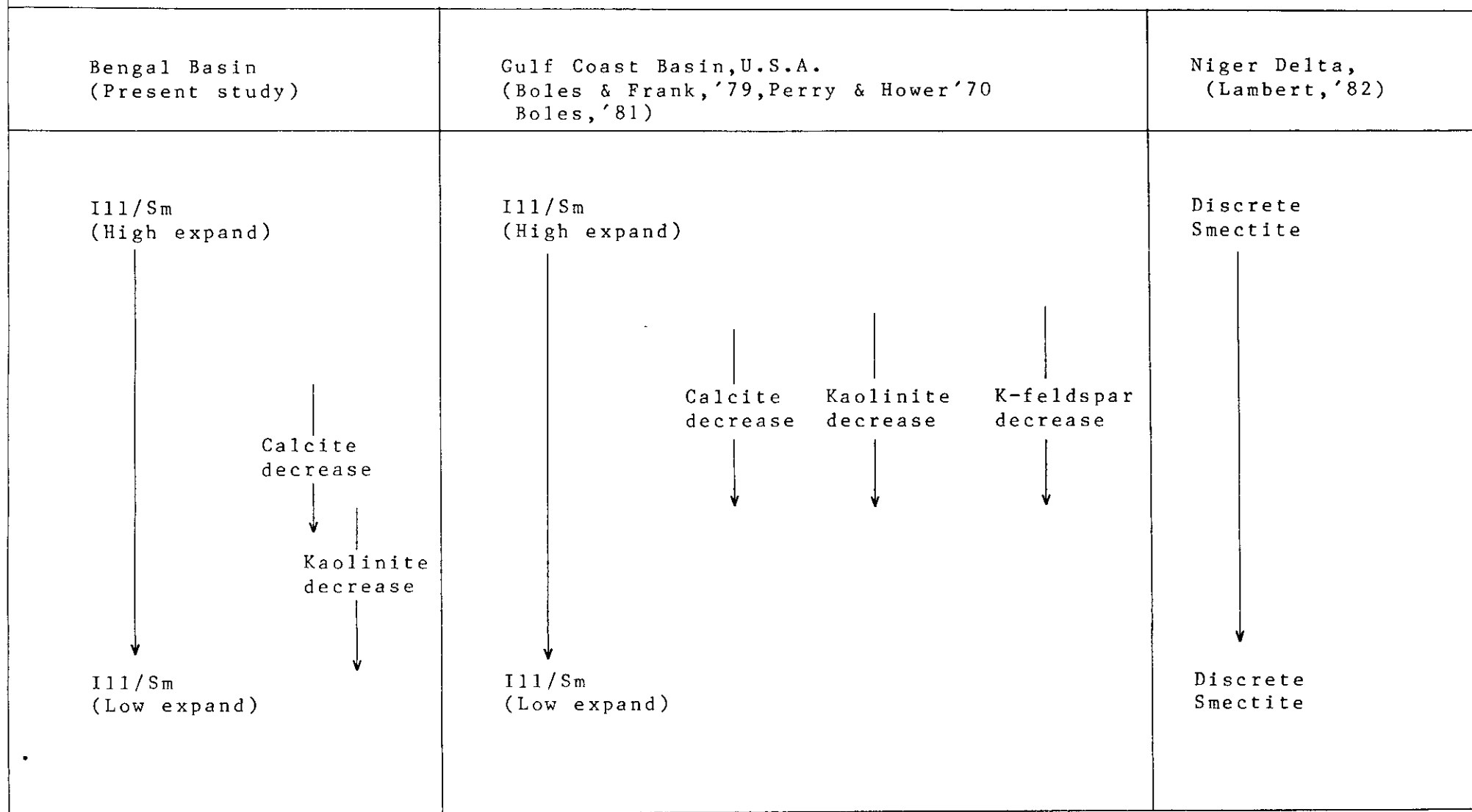
Bengal Basin Neogene sst. (Present study)	Tertiary deltaic Sst: Wilcox fm: Gulf Coast Basin U.S.(Stanton'77)	Tertiary deltaic Sst: Frio fm: Gulf Coast Basin, U.S.(Loucks et al. '77)	Cretaceous shelf Sst:Denver Basin Colorado,U.S. (Potter & Weimer '82)	Pennnsylvanian deltaic Sst: Texas,U.S. (Land & Dutton '78)	Tertiary deltaic Sst: Niger delta. (Lamberth '81)
Compaction	Compaction	Feld.leaching	Compaction	Compaction	Kaol. precip.
Chlorite rims	Qtz overgrowth	Poik.calcite cementn.	Chlorite rims	Chlorite rims	Kaol unstable
Siderite cryst.	Calcite cementn. & replacement	Compaction.	Qtz overgrowth	Qtz overgrowth	Smectite form. Calc.prepic.
Poik.calcite. cementn.	Dissol.of calcite	Clay rims	Calcite cementn. & replacement	Calcite cementn.	Dissol.of calcite
Illite coat	Precip.of Kaol.	Feld. overgrowth	Soln.of calcite	Dissol.of calcite	Qtz overgrowth.
Qtz overgrowth	Precip.of Fe-calc.	Qtz overgrowth	Kaol.precip.	Kaol.precip.	Siderite cryst.
Late calcite cementn. & repl	Precip of minor dolomite	Calcite cementn.		Barite precip.	Pyrite precip.
Dissol. of calc.cement, feld,etc.		Dissol.of calcite cements ,feld & lithics.			
Kaol. cryst.		Kaol. cryst.			
		Fe calc cementn.			

Figure 7.4 summarises the important aspects of shale diagenesis in the Tertiary deltaic sequence of the Bengal Basin (present study), the U.S. Gulf Coast Basin (Boles and Frank 1979, Boles 1981, Perry and Hower 1970) and the Niger Delta (Lamberth 1982). The clay mineral compositions of Tertiary shales from the Bengal Basin and the U.S. Gulf Coast Basin are similar, with illite, kaolinite, chlorite and mixed layer illite/smectite as the clay components. The clay minerals of the Tertiary shales of the Niger Delta, however, include kaolinite, illite and smectite. Progressive decrease of the smectite proportion of the illite/smectite clay with increasing depth of burial is the most important diagenetic change in Tertiary shales of Bengal Basin and U.S. Gulf Coast Basin. In the Niger delta, on the other hand, such progressive loss of expandability of the smectite component with depth of burial is not always observed. Decrease of kaolinite at greater depth has also been noted in both Bengal Basin and in the U.S. Gulf Coast. Calcite also shows a decrease in abundance with depth in both the Bengal Basin and the U.S. Gulf Coast Basin. There is also a decrease in abundance in K-feldspar with increasing depth in the Gulf Coast Tertiary sequence but this has not been observed in either the Bengal Basin or the Niger Delta.

The diagenetic history of the Tertiary sandstones and shales of the Bengal Basin shows general resemblances with those of the Tertiary Gulf Coast Basin. However, the Tertiary sandstone and shales of the Niger Delta appears to differ significantly in their diagenetic history. The sandstones of the Niger Delta appear to have suffered relatively little diagenetic change resulting in poor cementation and undercompaction (Lamberth 1981). The little quartz overgrowth formation that has occurred is preceded by calcite and kaolinite formation. Kaolinite appears to be the first authigenic mineral to form in these sandstones. In contrast, in the Bengal Basin and

Gulf Coast Tertiary Basin, quartz overgrowth always predates calcite cementation and kaolinite precipitation, with the latter being one of the last authigenic minerals to form. The effects of burial diagenesis on the clay minerals of the Tertiary shales in the Niger Delta have also been reported to be minimal (Lamberth, 1982). The discrete smectite clay mineral in Akata shale is preserved at depths of about 12,000 feet (3660 m) and burial temperature of 120 degrees centigrade, at which one would expect it to be transformed to a mixed layer illite/smectite phase. However, whilst burial temperature is the principal control in clay mineral diagenesis, other factors such as overburden pressure, pore water chemistry may also influence such diagenetic reactions (Shaw 1980). It has been suggested that under compaction of the shale might allow the smectite to be preserved in some Tertiary shales of Niger delta at temperatures in excess of what might be expected from the accepted model of clay diagenesis (Lamberth 1982). The lack of diagenetic alteration of smectite clay which could have released silica into pore solution are considered to be one of the principal causes for the poor development of silica cement in the sandstone of the Tertiary sequence in Niger Delta (Lamberth 1982).

FIG.7.4. SCHEMATIC DIAGRAM COMPARING SHALE DIAGENESIS IN THE BENGAL BASIN WITH THOSE OF THE GULF COAST BASIN AND THE NIGER DELTA.



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APPENDIX

Sample No.	Depth (m.)	Core No.	Description
MULADI- 1 WELL			
M1-3	500	3	Sandstone; yellowish grey, loose medium to fine grained.
M1-4	602	4	Sandstone; yellowish grey, loose, medium to fine grained.
M1-7	878	7	Sandstone; light grey, loose, medium to fine grained, argillaceous.
M1-11	1137	11	Sandstone; light grey, loose, medium to fine grained, argillaceous.
M1-12	1197	12	Shale; light grey, soft, silty, massive, calcareous.
M1-14	1402	14	Shale; light grey, rather soft, massive, calcareous.
M1-20	1700	20	Sandstone; light grey, friable very fine grained, argillaceous.
M1-22	1918	22	Shale; bluish grey, compact, rather soft, massive, calcareous.
M1-25	2107	25	Shale; bluish grey, moderately

indurated, massive.

M1-27	<u>2212</u>	27	Sandstone; greyish white, very hard, medium to fine grained, very calcareous.
M1-28	2216	28	Sandstone; light grey, moderately indurated, very fine grained.
M129A	<u>2290</u>	29	Sandstone; greyish white, very hard, fine grained, very calcareous.
M1-29B	<u>2293</u>	29	Shale; bluish grey, moderately indurated, contains few very thin silty streaks.
M1-30	2402	30	Shale; bluish grey, moderately indurated, massive, calcareous.
M1-31	2502	31	Shale; bluish grey, moderately indurated, massive, slightly calcareous.
M1-34A	2866	34	Shale; bluish grey, fairly indurated, contains few very thin discontinuous very fine sand/silt laminations.
M1-34B	2871	34	Sandstone; grey, moderately indurated, friable, medium to fine grained.
M1-35	3020	35	Shale; light grey, indurated, some discontinuous, thin, very fine grained sand/silt lamination; mica concentration along bedding plane.

M1-36	3114	36	Sandstone; grey, moderately indurated-friable, fine grained.
M1-37A	3217	37	Shale; dark grey, indurated, few discontinuous, thin, silty laminations.
M1-37B	3223	37	Sandstone; greyish white, hard, fine grained, calcareous, contains few clay platelets 2mm-5mm long and 0.5mm thick.
M1-38	3294	38	Sandstone; light grey, fairly indurated, fine grained; contains few clay platelets 2mm-5mm long and 0.5mm thick; concentration of mica parallel to bedding planes.
M1-39	3420	39	Sandstone; greyish white, very hard, very fine grained, very calcareous.
M1-40	3502	40	Shale; dark grey, hard; few very fine sand laminations.
M1-41	3602	41	Shale; dark grey, hard, fissile; very few thin silt laminations.
M1-42	3692	42	Shale; dark grey; hard, fissile, few thin laminations of very fine sand and silts.
M1-44	3930	44	Shale; dark grey, hard, fissile, few thin laminations of very fine sand and silt.

M1-45	4007	45	Shale; dark grey, hard, fissile; few thin laminations of very fine sand and silt; concentration of mica along bedding planes.
M1-46	4099	46	Shale; dark grey, hard, fissile; discontinuous laminations of very fine sand and silt.
M1-47	4205	47	Shale; dark grey, hard, fissile; thin discontinuous laminations of very fine sand and silt.
M1-49A M1-49B	4458	49	Very thinly interbeds of fine grained sandstone and shale; the sandstone is greyish white, hard and slightly calcareous; the shale is dark grey, hard and fissile.
M1-51	4530	51	Shale; dark grey, hard, fissile, thinly laminated; laminations are of very fine sand and silts and are discontinuous.

MULADI-2 WELL

M2-3	4491	3	Sandstone; greyish white, hard, medium to fine grained, calcareous.
M2-5	4550	5	Shale; dark grey, hard, fissile.

BEGUMGANJ-1 WELL

B1-4	1203	4	Shale; light grey, rather soft, massive, calcareous.
B1-11	1870	11	Shale; light grey, moderately indurated, massive, calcareous.
B1-14	2139	14	Sandstone; grey, poorly indurated-friable, medium to fine grained.
B1-15	2423	15	Shale; grey, fairly indurated, silty, few very fine sand laminations.
B1-16	2680	16	Sandstone; greyish white, hard, fine grained, calcareous.
B1-17	2758	17	Sandstone; light grey, fairly indurated, fine grained.
B1-18	2906	18	Shale; dark grey, hard, fissile; few very fine sand and silty laminations.
B1-19	2999	19	Sandstone; light grey, fairly indurated, fine grained; dark mineral concentration along some planes.
B1-20	3099	20	Shale; dark grey, hard, fissile, thin laminations of silt.
B1-22	3390	22	Shale; dark grey, hard, massive, silty.

B1-23	3444	23	Shale; dark grey, hard, massive, silty.
B1-24	3532	24	Shale; dark grey, hard, massive; few silty calcareous silty bands
B1-25	3628	25	Shale; dark grey, hard, fissile; thin discontinuous laminations of very fine sand and silts, concentration of mica along the bedding planes.
B1-26A	3642	26	Sandstone; whitish grey, hard, medium to fine grained.
B1-26B	3656	26	Shale; dark grey, hard, fissile, very few thin laminations of very fine sand and silts.

BEGUMGANJ-2 WELL

B2-1	3028	1	Sandstone; greyish white, very hard, medium to fine grained calcareous.
B2-2	3042	2	Shale; dark grey, hard, fissile, few very fine sand and silt laminations.
B2-3	3325	3	Sandstone; light grey, fairly hard, fine grained.
B2-4	3476	4	Sandstone; greyish white, very hard, very fine grained, calcareous.

B2-5	3523	5	Shale; dark grey, hard, fissile; few very fine sand and silty thin laminations.
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B2-7	3567	7	Shale; dark grey, hard, fissile; few thin laminations of silt and very fine sand.
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TITAS-2 WELL

T-1	2646	1	Sandstone; light grey, fairly indurated, fine grained.
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T-2	2660	1	Sandstone; light grey, fairly indurated, fine grained.
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T-3	2668	2	Sandstone; light grey, fairly indurated, medium to fine grained.
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T-4	2673	2	Sandstone; light grey, fairly indurated, medium to fine grained; few clay clasts 0.5mm to 1mm long.
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T-5	2678	2	Sandstone; light grey, fairly indurated, fine grained.
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T-6	2686	3	Shale; dark grey, hard, fissile.
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T-7	2690	3	Sandstone; light grey, fairly indurated, medium to fine grained.
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T-8	2698	3	Shale; dark grey, hard, fissile.
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T-9	2704	6	Sandstone; light grey, fairly
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indurated, fine grained.

T-10	2706	6	Sandstone; light grey; highly indurated, fine grained.
T-11	2712	6	Sandstone; light grey; fairly indurated, fine grained.
T-12	2718	7	Sandstone; greyish white, hard, fine grained.
T-13	2725	7	Sandstone; greyish white, hard fine grained.
T-14	2728	7	Shale; dark grey, hard, fissile few very fine lamination of silts
T-15	2753	10	Sandstone; light grey, moderately indurated, medium to fine grained.
T-16	2761	10	Sandstone; light grey, moderately indurated, medium to fine grained.
T-17	2766	10	Shale; dark grey, hard, fissile very few thin laminations of silts.
T-18	2927	12	Sandstone; greyish white, hard, fine grained.
T-19	2935	12	Sandstone; greyish white, hard, fine grained.
T-20	2954	13	Sandstone; light grey, hard, fine grained.
T-21	2967	14	Sandstone; greyish white, hard,

fine grained.

BODC-2 WELL

BODC-1	3110-3120	-	Shale; rock cuttings, dark grey hard.
BODC-2	3220-3230	-	ditto
BODC-3	3330-3340	-	ditto
BODC-4	3500-3510	-	ditto
BODC-5	3610-3620	-	ditto
BODC-6	3650-3660	-	ditto
BODC-7	3730-3740	-	ditto
BODC-8	4076	1	Shale; dark grey, hard, fissile.
BODC-9	4082	1	ditto
BODC-10	4100-4110	-	Shale; rock cuttings, dark grey, hard.
BODC-11	4200-4210	-	ditto
BODC-12	4410-4420	-	ditto.