THE UPPER CRETACEOUS SEDIMENTS OF THE GÜRÜN AREA (SIVAS-TURKEY) WITH PARTICULAR REFERENCE TO THE DIAGENESIS AND ENVIRONMENT OF DEPOSITION ~~AND DIAGENESIS~~ OF THE CARBONATE ROCKS

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ABSTRACT

The work is primarily concerned with the study of the petrography, depositional environment and various aspects of diagenesis of the Upper Cretaceous carbonate rocks of the Gürün (SIVAS-TURKEY) area. At the same time the geologic evolution and geology of Anatolia has been reviewed.

Over twenty complete sections of the Gürün carbonate rocks were sampled at various intervals for petrographic study. The environmental suite which can be delineated for the Upper Cretaceous rocks of the province is a typical carbonate platform sequence characterized by shallow-water cyclic carbonate sedimentation extending over large areas. This transgressive marine carbonate sequence is associated with regional Upper Cretaceous transgression. On the outer margins of the shelf knoll reef ramps were developed where rudists had played an important role. Deepwater sediments are characterized by calciturbidites. Ideal carbonate platform model and twenty-four Standar Microfacies Types proposed by WILSON (1975) have been found applicable to the Gürün carbonates and employed in the present account.

Study of almost one thousand thin sections and more than one thousand and five hundred peels reveals that the original carbonate sediments have been modified by a number of diagenetic processes. Biologic diagenesis, formation of micritic envelopes and micritization were contemporaneous with deposition. Two phases of calcite cementation occured in the grainstone facies. The origin of the limestone of the Gürün region involved the deposition of large quantities of carbonate mud over a long interval of geologic time and conversion of this carbonate mud to lithified micrite is the most important aspect of neomorphism. Two stages of dolomitization have been observed. Dolomitization of the Gürün sediments was caused by the mixing of seawater and ground water in the phreatic zone (Dorag dolomitization model). Dedolomitization is common in the NW of the area. Silicification is widespread in the calciturbidites. Where data permit, the author has attempted to draw conclusions regarding time relationships of the various diagenetic processes.

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THE MOUNTAINS MUST INDEED BE EXAMINED WITH THE MICROSCOPE

H.C. Sorby (1856)

CHAPTER I

INTRODUCTION

Previous geologic investigations of the Upper Cretaceous carbonates in the Gürün region have dealt primarily with the local and regional stratigraphy. No detailed information is available on the petrography, depositional environments and diagenesis of the carbonate sediments in the province. The aim of the present study was to establish the depositional environments of the carbonate units and discuss a variety of diagenetic changes observed in these rocks.

Field work and sampling were conducted during the summers of 1972, 1973, and 1974.

In any petrological study which is to involve large areal extends of rocks, it is necessary to devise a plan for taking a representative sample of the rock. For a sedimentary sequence, essentially two possibilities exist. One choice is a study based on a vertical section. The second method , quite naturally, is a lateral facies oriented study. Confronted with an outcrop pattern characterized by a relatively minor amount of stratigraphy extending over a considerable lateral distance as with the Upper Cretaceous succession of Gürün, it is natural to assume that a facies study would be the most profitable. The ultimate decision, however, must rely on the availability of vertical and lateral variations. In the investigated area the lack of lateral variation and the difficulties caused by complex structural complications make a lateral facies study highly impractical. Thus, the project becomes self-determined by its own characters as a petrological study of mainly single vertical stratigraphic sections. Twenty-three sections of the Gürün Upper Cretaceous carbonate rocks were sampled at various intervals. More than 1500 stained peels, and almost 1000 thin sections, also mostly stained, were prepared and studied under a petrographic microscope. Data were recorded on specially prepared sheets (Table I-1).

SAMPLE	Section No.	Locality	Sample No.	Features								
MAJOR COMPONENTS	Grains	Lime Mud	Cement	Authigenic Minerals								
PACKING	Mud Support	Grain Support										
TYPES OF GRAINS	Lithoclasts	Pelletoids	Coated Grains	Fossils								
GRAIN SIZE	> 2 mm.	1-2 mm.	0.5-1 mm.	0.25-0.5 mm	0.62-0.25mm	0.016-0.062						
ROUNDNESS	Angular	Subangular	Subrounded	Rounded	Well-Rounded							
TYPES OF FOSSILS	Red Algae	Other Algae	Foraminifera	Coelenterate	Bryozoans	Brachiopods						
	Gastropods	Pelecypods	Ostracods	Echinoderms	Bioclasts	Others						
TYPES OF CEMENT	Drusy or Fringing Cement	Blocky or Equant Cement	Syntaxial Overgrowths									
NEOMORPHISM	Degrading Inversion	Degrading Recrystalliza.	Porphyroid Inversion	Porphroid Recrystalliza.	Coalescive Inversion	Coalescive Recrystalliza.						
OTHER FEATURES	Stylolites	Pressure Solution	Compaction	Geopetal Structures	Birdseye Fabric	Burrowing						
STANDART MICROFACIES TYPE	1	2	3	4	5	6	7	8	9	10	11	12
	13	14	15	16	17	18	19	20	21	22	23	24
MICROPHOTO.												
NOTES												

Table 1 - 1: Data record sheet

I. GEOGRAPHIC SETTING

The area under investigation is located in east-central Turkey. It is bordered by the 36°45' 00" and 37°37'00"eastern meridians, and 38°30'00" and 39°00'00" northern parallels. (Figure I-1). According to a traditional subdivision of major tectonic units in Anatolia (KETIN 1959 and 1966) this is a part of the Taurid Belt (Figure II-1). It is elongated in NE-SW direction and is 82 km. long and 27 km. wide, and covers an area of approximately 1600 square kilometres (See the Geologic Map). It is the southern part of the Sivas Province. The largest centre in the map area is the town of Gürün (B47*) and is situated in the south east.

Relief is generally lowered when one goes from north to south, and from west to east. The highest point is Gövdeli Dağı with 2675 metres. It is in the northwestern corner of the K37-c1 sheet (just outside of the map area). Other important heights are Arapmuslu Tepe : 2458m. (1 km. NW of Çukuryurt "W11"); KavurmaÇukuru Tepe : 2327m. (V47); Sümbüllü Dağı : 2185m. (Z39); Çekmelikuz Tepe : 2167m. (A25); Karadağ : 2158m. (O60); Devlet Höyük Tepe : 2149m. (3,5km. SW of Yaylacık "K32"); and Bölücek Dağı : 2044m. (J16). The lowest part is in the east near Bıcır (G81) with 1300m.

Due to the continental climate of the region, summers are warm and arid and winters are cold and snowy. The area covered by the K38-d1 and K38-d2 sheets is very dry. There is no running water. The waters of the study area join to three different rivers. The first two draining areas are found in the west and form only a small portion of the total drainage area. The one in the northwest is only a few square kilometres. The waters of this part (NW of Çukuryurt "W11") join to the Zamanti Suyu, main tributary of the River Seyhan, which runs to the Mediterranean Sea. The second drainage area is in the southwest. Şuğul Dere and Bozhöyük Dere run in the west-east direction along the southern part of the K37-c1 and centre part of the K37-c4 sheets, respectively. The former in the north of Güllübucak (L12) and the latter in the south of Akdere (H14) join to the Hurman Dere, one of the tributaries of the River Ceyhan, and again

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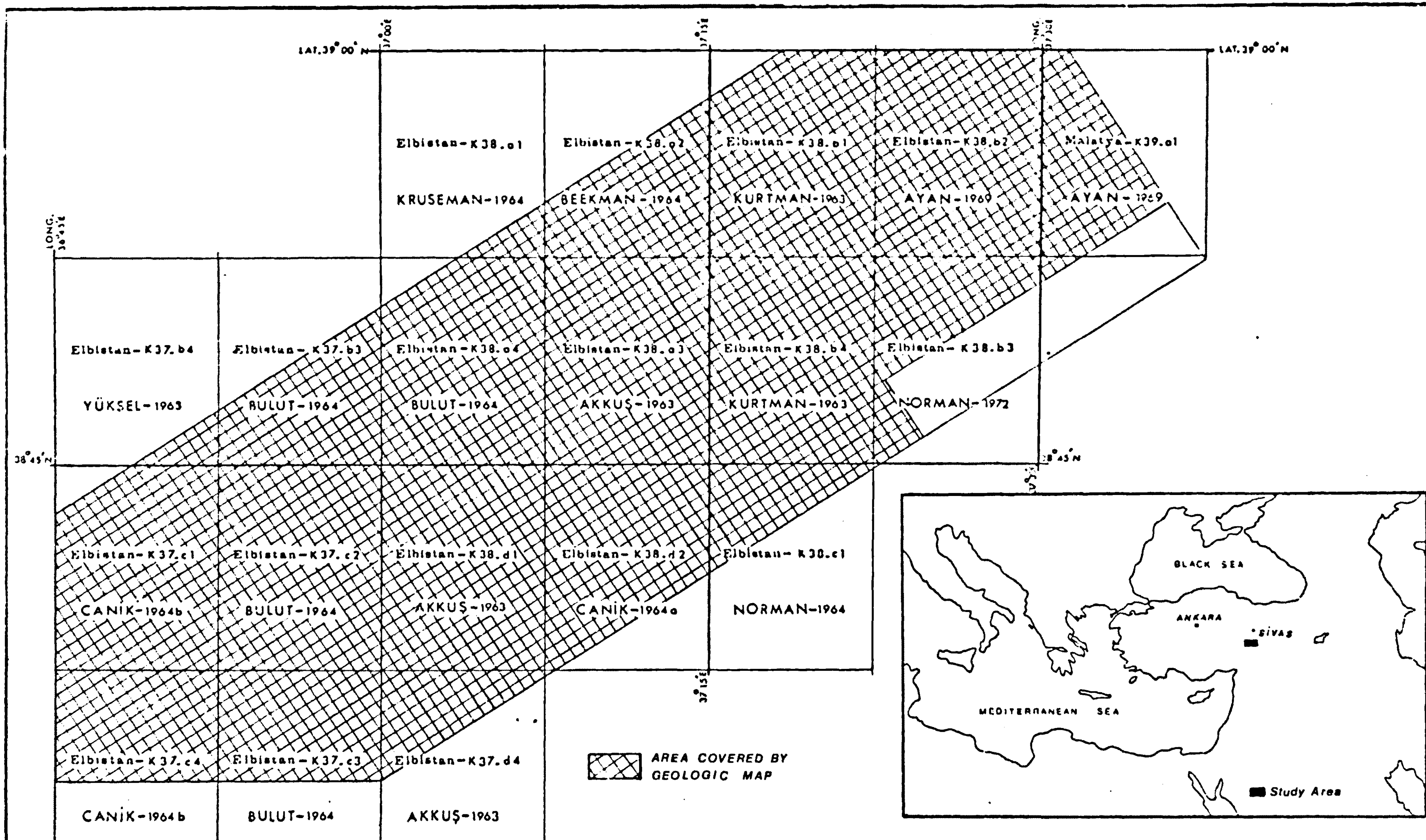
(*) The geologic map is prepared in grid squares which are numbered in order to ease finding the locality names given in the text.



LAT. 38° 30' N

Fig. 1-1. Sheet index showing areas studied by the geologists in connection with the preperation of the 1:25,000 scale geologic map.

reach to the Mediterranean Sea. Most waters of the study area are collected by two streams. One of them, namely Tohma Suyu, flows in W-E direction along the southern borders of the K37-b3, K38-a4 and K38-a3 sheets, then turns to the south-west, cuts the NE corner of the K38-d2 sheet, passes through Gürün (B47) and leaves the area. On the other hand , Tohma Çayı cuts the K38-b2 and K39-a1 sheets across in the NE-SE direction. Both of the streams, Tohma Suyu and Tohma Çayı follow a meandering course. They join at the south east of Darende, some 25 km. east of Gürün (B47) to form the main tributary of the River Fırat (Euphrates) which runs to the Persian Gulf.

The Ankara-Malatya Highway crosses the area. In addition to this highway, the Gürün-Sivas road lies in the south-north direction. The area is very poor with respect to roads mainly due to rough topography caused by the carbonate rocks of different ages. There are a few roads suitable only to field vehicles during the summer period. In the winter, all these roads are completely blocked by heavy snows, and the only connection between towns and villages is by horseback.

II. GEOLOGIC SETTING

In spite of several geologists having worked in the area since the middle of the 19th century, preliminary studies for the geologic mapping of the area started in 1936 by M.T.A. Institute, in 1/100,000 scale. Although some of the results have been published, most of them remain at the Institute Library in the form of reports. For a detailed review of previous works, the reader is referred to the work of BAYKAL (1966).

Between the years of 1962 and 1968 there was an extensive geologic work in the region, mainly due to the oil exploration programs of M.T.A. Institute. Geologic maps in 1/25,000 scale were completed and in connection with this, seven exploration wells were drilled in the Darende Area (25 km. south east of Gürün "B47"). These maps are the base maps of the present work. The 1/25,000 scale sheet numbers, and the geologists who worked in the mapping of these sheets, are shown in Figure I-1. Unfortunately these maps and related reports are confidential. Therefore the geologic map of the area is largely simplified and presented in 1/100,000 scale. In the following paragraphs, the general geology of the region will again be discussed in a very simplified manner.

A. STRATIGRAPHY

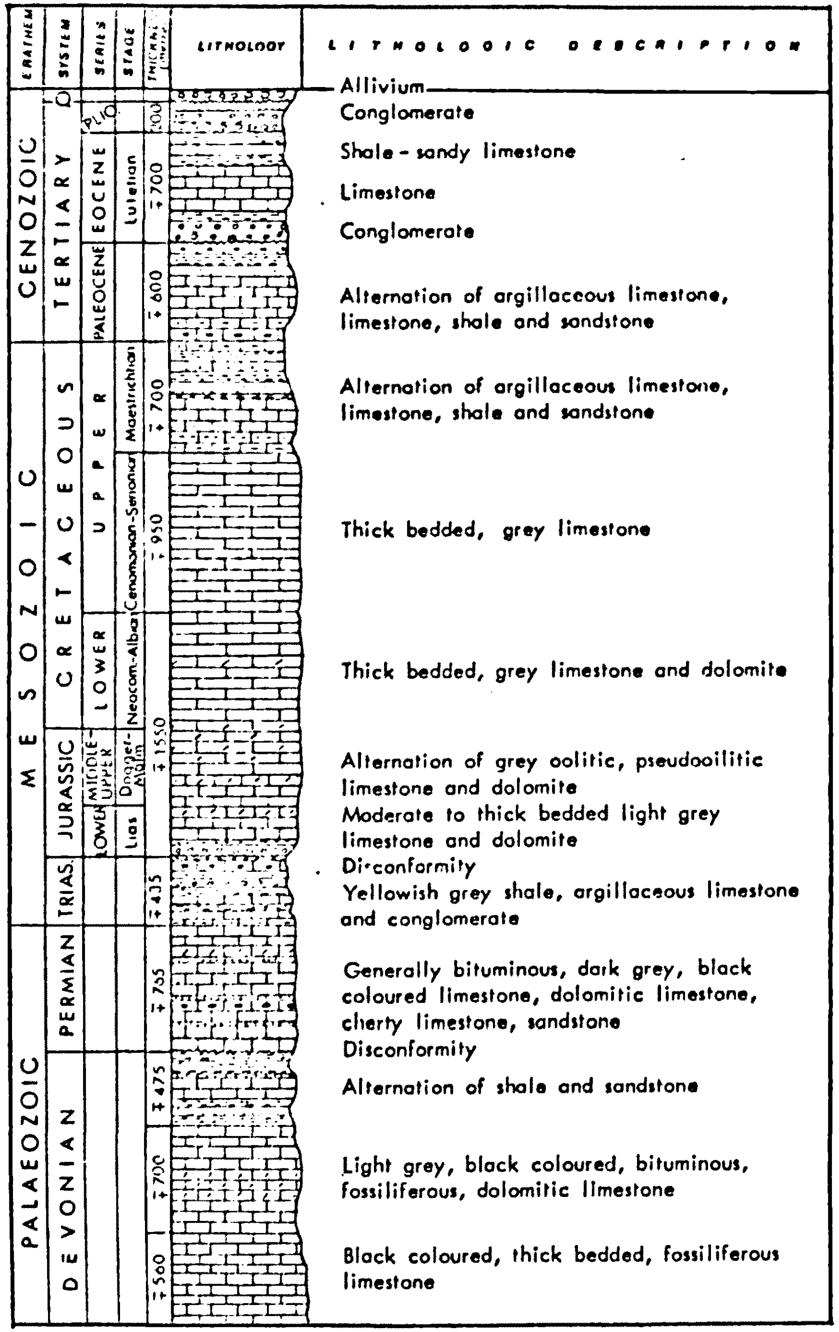
The stratigraphic succession in the map area consists of the Palaeozoic, Mesozoic and Cenozoic formations. Different workers, in different areas, have applied different names for the same formations. Because of the lack of an overall agreement in the nomenclature of formation names, in this study time-stratigraphic (chronostratigraphic) units have been used (AMERICAN COMMISSION ON STRATIGRAPHIC NOMENCLATURE 1970).

There is a striking difference between the east part and the rest of the geologic map. This is true for structural elements, such as fold axis, fault directions, and rock types as well. Approximately the line joining Dürmepınar (S69) and Dayakpınar (L68) forms the border. The differences will be discussed in the appropriate sections.

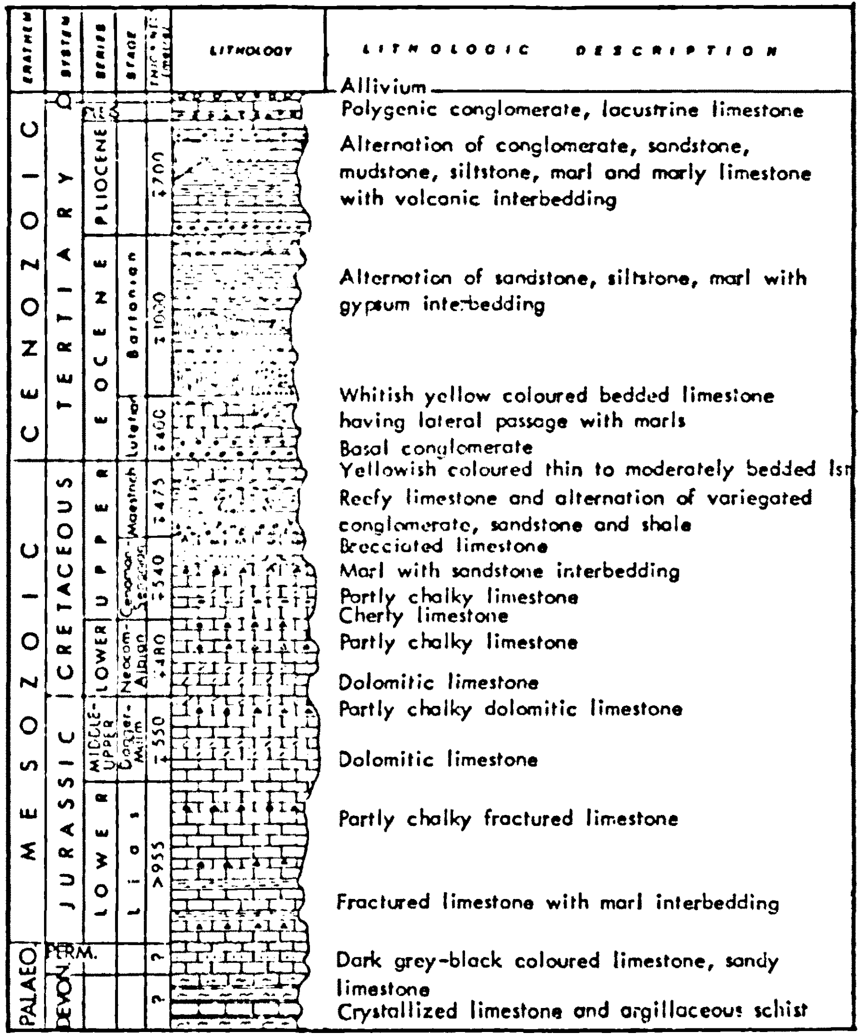
The oldest rocks in the map area are Permian in age, and outcrop in three different localities. From west to east; around the Çukuryurt (W11), north of the Çicekyurt (U40), and between the Kavurma Çukuru Tepe (V47) and the Behram (X53). All three show a general SW-NE trend. Their contacts with the other formations are always along faults. The dominant lithology is limestones with some conglomerate and sandstone intercalations. In the region of Sarız (30 km. further west of the west margin) a section of total 765 m. of Permian has been measured (DEMİRTAŞLI, 1967). Here they overlay, probably with a disconformity, 1735 m. thick Devonian System (Figure I-2A). Everywhere the Permian is bitumenous. Corals are abundant. According to BEEKMAN (1964) some Carboniferous horizons exist. This is confirmed by the present work. Some samples collected from the Kavurma Çukuru Tepe (V47) have shown a Carboniferous age (Determinations by M. UYSAL).

Triassic System does not outcrop in the study area. But in the Sarız region Permian passes to the Triassic without any break in the sedimentation (Figure I-2A). Total thickness of the Triassic deposits is 435 m. (DEMİRTAŞLI, 1967). Shales, argillaceous limestones and conglomerates constitute the main lithology.

The Jurassic-Lower Cretaceous Series cover large areas in the north and south of the map area. It is very widespread in east-central Anatolia (BAYKAL, 1966). The dominant rock type is massive with thick bedded limestones everywhere. It is almost impossible to make a distinction between the Jurassic and Lower Cretaceous limestones on the field. The only criterion is fossil content. Fossil specimens such as Macroporellapigma, Lituola, Valvunella jurassica, Clypeina jurassica, Hensonella, Salpingoporella, Actinoporella, Pseudocyclammina sp. have been encountered by several workers. In one of the bore-holes conducted by



A



B

Fig.I-2: Schematic stratigraphic sections of
A. West of the Gürün
B. East of the Gürün
(Mainly after KURTMAN & AKKUŞ, 1974)

M.T.A. Institute near Darende, more than 2000m. thick Jurassic-Lower Cretaceous succession, mostly carbonates, was drilled (Figure 1-2B). Approximately 1000 metres of it is Lias in age and drilling was stopped in the Lias limestones.

In the west, the Jurassic-Lower Cretaceous sediments passes to the Upper Cretaceous ones. Again a distinction on the field is almost impossible. Some samples containing the Jurassic-Lower Cretaceous fossils have been collected in the 5-6 km. west of Kurucaoba (S12), and 5 km. southwest of Tersakan (N6) which is shown as "Upper Cretaceous Limestones" in the geologic map (CANİK, 1964b). It may be concluded that lower parts of the Upper Cretaceous Series (Senomanian-Turonian) outcrops in this region, whereas in the further east it is represented by the upper parts (Lower Senonian). The thickness of the Upper Cretaceous limestones is approximately 1000m. in the west. On the other hand in the above mentioned bore-hole the thickness is half of this (Figure I-2B) and further east (vicinity of the Malatya Province) no outcrops of this series has been found (KURTMAN and AKKUŞ, 1974). It is represented by massive to thick bedded limestones in the west. In the east the dominant rock type is limestones again, but some marls and brecciated horizons are present. It is very widespread (see the geologic map). The fossil content and distribution is shown in Table I-2A. These limestones and overlaying Maestrichtian deposits are the subject of this study and will be discussed in detail in the following chapters.

Two units can be distinguished in the Maestrichtian Stage. The lower unit cover large areas. It is called "flysch" or "flysch-like" by several geologists. No detailed work has been carried out and it has been described simply as "... alternating beds of sandstone, shale, mark, marly limestone and limestone". This study has clearly shown that they are "calciturbidites" (see p. 90). The fossil content and distribution is shown in Table I-2B and it indicates a deep water environment. The second unit is only seen in the south of Camiiliyurt (P11) and in the vicinity of Konakpınar (R60) and Kuz (N66). The yellowish limestones dominate and usually show a brecciated texture. The fossil content is very poor. Algae are relatively abundant. This unit has been interpreted as "lagoonal" by DEMİRTAŞLI and AYAN (1964).

As mentioned above the east of the map area has a different geology. The ophiolites and related rocks cover large areas. The age and emplacement time of these rocks will be discussed in Chapter II. An Upper Jurassic age for the origin and Senonian age for the time of emplacement is accepted (see p.39-40). Considering the sudden facies change in the relatively short distance, plus the intense tectonics, it can be argued

PRINCIPAL MICROFOSSILS	YUKSEL (1963)	CANIK (1964b)	BULUT (1964)	YOLDAŞ (1970)	AKKUŞ (1963)	CANIK (1964a)	KURTMAN (1963)	AYAN (1969)
Ammobaculites sp.				+				
Barkerina sp.				+				
Beggia cf. Simplex OMARA				+				
" sp.		+		+				
Chrysalidina gradata d'ORBIGNY				+				
" sp.			+	+				
Coscinolina sp.			+		+			
Cuneolina cf. cylindirica HENSON							+	
" cf. hensoni DALBIEZ		+						
" pavonia d'ORBIGNY	+	+	+	+	+	+	+	
" primitiva		+						
" sp.		+	+	+			+	
Dasycladaceae		+				+		
Dicyclina schlumbergeri MUNIER-CHALMAS				+				
" sp.	+	+		+	+		+	
Dukrania sp.		+			+	+		
Haplophragmoides sp.				+				
Lithothamnium sp.		+				+		
Lituola sp.				+		+		
Miliolidae		+	+	+	+		+	
Ophtalmidae		+		+	+	+		
Orbitolina concaca LAMARCK		+						
Ovalveolina ovum d'ORBIGNY		+						
Polygonella sp.		+	+			+	+	
Praeglobotruncana delrioensis PLUMMER		+		+				
Preoalveolina sp.			+					
Pseudocyclammina lituus YOKOYAMA			+					
" cf. vasconica MAYNC				+				
Pseudolituonella sp.				+				
Pseudotextulariella sp.				+				
Quinqueloculina sp.					+	+		
Salpingoporella sp.				+				
Solenoporaceae				+		+		
Spiroloculina sp.					+	+		
Textularia sp.	+	+	+	+	+	+	+	
Textularidae		+	+	+				
Trocholina sp.				+				
Valvulammina piccardi HENSON				+			+	
" sp.		+	+	+				
Valvulina sp.			+	+	+			
Valvulinidae		+	+					
Vermes sections				+				
Verneuilina sp.				+				
Verneuilinidae		+					+	
Vidalina hispanica SCHLUMBERGER				+			+	
" sp.		+				+		

A

PRINCIPAL MICROFOSSILS	YUKSEL (1961)	CANIK (1964b)	BULUT (1964)	YOLDAŞ (1970)	AKKUŞ (1963)	CANIK (1964a)	KURTMAN (1963)	AYAN (1969)
Globigerina bulloides d'ORBIGNY		+						
" sp.		+		+			+	
Globigerinella sp.				+				
Globotruncana arca CUSHMAN		+		+		+	+	
" bulloides VOGLER		+		+				
" calcarate CUSHMAN		+						
" contusa CUSHMAN			+		+		+	
" conica WHITE		+	+	+			+	
" gagnebini TILEV				+				
" fornicata PLUMMER	+							
" globigerinoides BROTZEN		+					+	
" lapparenti coronata BOLLI		+						
" " lapparenti BROTZEN	+	+				+		
" " tricarinata QUEREAU		+	+		+	+		
" leupoldi CANDOLFI				+				
" linnei d'ORBIGNY		+						
" rosetta CARSEY		+		+				
" stuarti De LAPPARENT		+	+	+	+	+	+	
Gumbelina globulosa EHREN		+			+			
" plummerae LOETTERLA			+					
Lagenidae				+				
Lepidorbitoides sosialis LEYM			+					
" sp.		+	+	+	+			
Lituolidae				+				
Marssonella oxycona REUSS							+	
" sp.				+	+		+	
Nodosaria sp.		+						
Omphalocyclus macroporos LAMARCK			+				+	
Operculina sp.				+		+		
Orbitoides apiculate SCHLUM								+
" media d'ARCH	+	+	+	+	+		+	+
" tissoti SCHLUMBERGER	+							+
" sp.		+	+					
Radiolaria sp.				+				
Robulus sp.					+	+		
Rotalia sp.				+				
Siderolites calcitrapoides LAMARCK			+		+		+	+
" sp.		+		+				

B

Table I-2: Content and distribution of microfossils in the Gürün region
A. Cenomanian-Campanian
B. Maestrichtian

that the above border is a microplate boundary. But of course before jumping to such a conclusion, a more regional and detailed study is needed. On the other hand the present author has found the chance to examine the logs of the Darende boreholes. In three of these boreholes at about 1600m. depth a fault has been discovered. Above this fault plane the Lias limestones have been drilled. They are underlain by the Senomanian-Lower Senonian formations. Underneath them the Lower Cretaceous and finally the Jurassic succession have been encountered. This indicates that some catastropic phenomena had occured after the Early Senonian times.

The ophiolites are overlain by coarse clastics in alternations of variegated conglomerate, sandstone and siltstones. They are completely composed of grains of green rocks. Even the matrix is ophiolitic in origin. In the study area no pebbles of the Jurassic-Cretaceous limestones have been found, but AKKUS (1971) reports some occurences of limestone pebbles in the conglomerates in the Darende Area. Conglomerates and sandstones are so poorly cemented that they can be crumbled by hand. A good cross-section of this unit is observed in the Hamalcay Dere Valley (H76). Here, conglomerates and sandstones rest transgressively upon the green rocks on the SE of the section (Figure I-3).

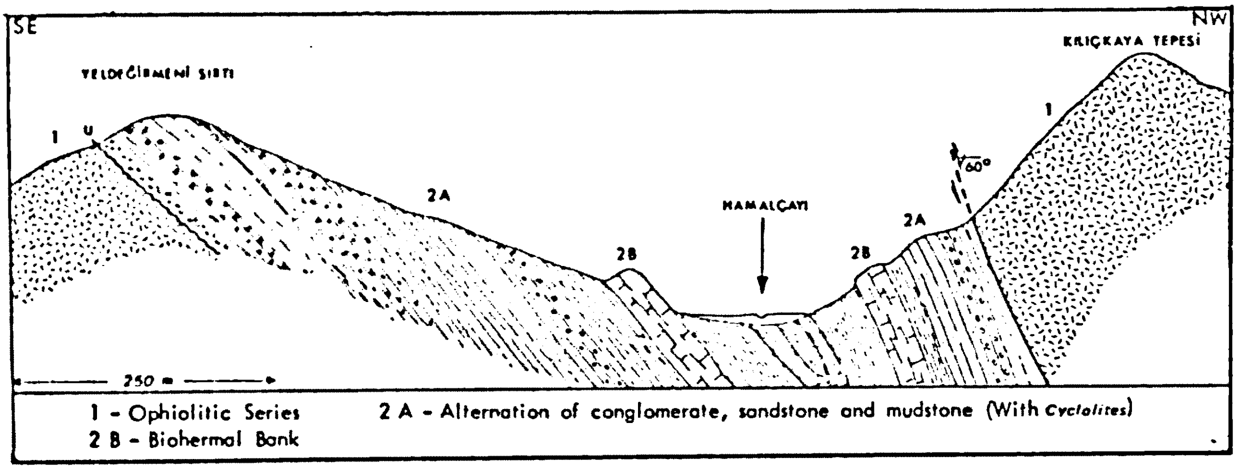


Fig.I-3: Cross-section in the Hamalcay Dere Valley (After AYAN, 1969)

Wholly, it forms an overturned syncline. The most common organism is Cyclobites which characterise the Senonian. Towards the upper parts of the succession some biohermal horizons can be seen. The NW contact is a reverse fault and green rocks overlay this formation. Some similar rocks have been reported along the Tohma Çayı, around Tökler (N75), Sarıca (N77), Sofular (I80) and Bıcır (G81) Villages. The lithology is very similar in the first instance. The biohermal banks have been found in various localities. AYAN (1969) gives a very shallow water origin for all of these occurences. But a close examination reveals that they are turbidites, and so-called "Biohermal banks" are in fact slump deposits (p.109). Also AKKUŞ (1971) has considered the similar deposits as "... detrital coarse clastics which were deposited in a shallow and active water" but his fossil content again shows a deep-water environment (AKKUŞ, 1971; p. 13).

As can be seen on the geologic map the Tertiary sediments extend over a large area. The Paleocene has only been observed in the west. The Upper part of the Maestrichtian calciturbidites is in Paleocene age. There is no lithologic difference. Only the fossil content is different (CANİK, 1946b). They are especially well developed between Arpaçukuru (K8) and Güllübucak (L13), but as they do not show any lithologic difference, they have not been separated on the geologic map. Although some other Paleocene occurences have been reported by several workers from different localities, they are unimportant and there is not any certain palaeontologic evidence. They may represent the lower parts of the Eocene Series.

Almost everywhere the Eocene Series with a basal conglomerate rest upon the Upper Cretaceous and older rocks with an unconformity (Figure I-2). Sediments of this age appear mainly in two types of facies: Limestones and flysch-like formations. A Lower Lutetian and an Upper Lutetian age have been deduced for these formations. In the Darende region thick Upper Eocene (Bartonian) Series have been found (AKKUŞ, 1971). The main lithologhy is alternation of sandstone, siltstone, marl with gypsum inter-beddings. No gypsum has been reported from the map area. The lithology and fossil content indicates the commencement of an uplifting during the Bartonian time. No trace of the Oligocene deposits has been found.

The Pliocene Epoch is represented by fluviatile and lake sediments.

Finally the Quaternary sediments are generally found on plains and valley bottoms. The large area covered by the unlithified Quaternary deposits at Kurucaoba (S12) and the west of Göbekören (T24) are in fact karstic depressions.

B. IGNEOUS ACTIVITY

Both intrusives and extrusives are present in the study area. The instrusives are composed of rock groups like serpantine, gabbro, spilite, etc. These green rocks, called "ophiolitic series", outcrop in the east (K38-b2 and K39-a4 sheets). Near Eskihamal (G71) they clearly cut the

Jurassic-Lower Cretaceous limestones. The Maestrichtian turbidites rest upon them with an unconformity (AYAN, 1969). These two observations support the general "Senonian" emplacement for ophiolitic rocks.

The extrusive rocks show big differences in composition from place to place but it is still possible to group them as trachytes, andesites, and basalts.

The trachytes are found in the NE along the Tohma Çayı. AYAN (1969) gives a Lutesian age based on the field relations; but LEO, MARVIN and MEHNERT (1974) have determined 74.3 and 71.1 million years by the potassium-argon methods which indicates Campanian Age (Geologic Time Table: Compiled by F.W.B. VAN EYSINGA /3rd Edition/1975).

The most widespread extrusives are the andesites. They cover large areas in the south of K37-b3 and in the north of K37-c2 sheets. In the east, Karadağ (060) is an andesitic mountain. In the 5km. southeast of this, another occurrence of andesite can be seen. They usually overlay the Middle and Upper Eocene Series, and are overlain by the Pliocene formations. This field relation is confirmed by the potassium-argon ages obtained by LEO, MARVIN, and MEHNERT (1974). They have determined 18.7 to 16.8 million years for some samples from 2 km. north of Arıkdamı (K81) which correspond to the Early Miocene Epoch.

Most probably the youngest extrusives are basalts around Kaşköy (015). They extend in E-W direction about 12 km. and form a sheet. The maximum thickness is 10 metres (BULUT, 1964). Because they overlay the Pliocene sediments, the volcanic activity is believed to have taken place during the Pleistocene Epoch.

C. STRUCTURAL GEOLOGY

According to Ketin's Classification of Tectonic Units of Anatolia, the area lies within the Taurid Tectonic Unit (Figure II-1). The region was affected by the Alpine Movements in general.

Because of the confidential nature of the data only folds and faults which have been seen in the Upper Cretaceous Series and representative strike and dips of this Series has been shown in the geologic map and will be briefly discussed here.

The Senomanian-Senonian limestones are generally medium to thick bedded and massive. The laminates and thin beds are dominant in the turbidites.

As has been stated earlier there is a striking difference between the west and east parts of the area. The main trend of fold axis and fault planes is generally ENE-WSW whereas in the east it is NW-SE (See the geologic map). The zone dividing the west and east parts lies along

Dürmepınar (S69) and Dayakpınar (L68) and it is thought to be a microplate boundary by the present author. Previously it has been interpreted as a virgation caused by the Keban Massif in the further east (AKKUŞ, 1971; KURTMAN and AKKUŞ, 1974) or an uplift of the Jurassic-Lower Cretaceous limestones (AYAN, 1969).

Generally, the Senomanian-Senonian limestones show broad, open folds, whereas the Maestrichtian turbidites are extremely folded (Plates 1 &2).

Normal and reverse faults are the most abundant fault types. The former is the dominant. Strike slip faults are rare.

Two disconfirmites have been reported from the Sarız region (DEMİRTAŞLI, 1967). They are between Devonian and Permian, and Triassic and Jurassic Systems. In the west of the study are Middle Eocene Series rest upon the Paleocene ones transgressively. In all other places they overlay the Maestrichtian deposits. In the east there is an unconformity between variagated Maestrichtian sedimens and ophiolitic rocks. Another notable unconformity exists between the Middle Eocene and Pliocene Series.

III. METHOD OF STUDY

A. FIELD PROCEDURES

The geographic locations of the measured sections within the map of the area are shown on the location map (Figure I-4). Measurements were made using a 20 metre tape; 100 metres string marked in every 20 metres; and a Brunton compass. For each new station, the following measurements were made and recorded in addition to lithology and characteristic field observations. (1) Slope distance, (2) Strike and dip, (3) Direction of traverse, (4) Angle of slope. In the calculation of thickness, the formula presented by BILLINGS (1972, p. 510) has been employed:

$$T = S\ (\text{Sin a. Cos b. Sin c} \pm \text{Cos A. Sin b})$$

where T = Thickness, S = Slope distance, a = dip of the bed, b = angle of slope, and c = angle between strike of bed and direction of the traverse. (+)was used if the dip of the bed and the slope of the ground were in

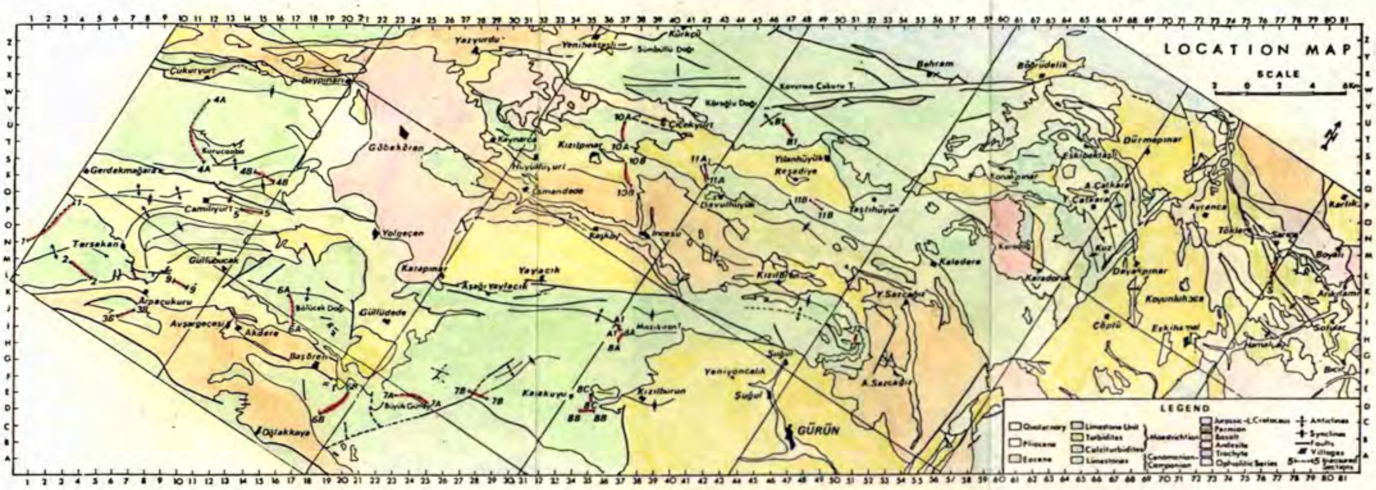


Fig.1-4: Location Map (Simplified from the Geologic Map)

opposite directions, and (-) was used if the dip of the bed and the slope of the ground were in the same direction.

Samples for the petrographic studies were taken contemporaneously with the measurement and description of the stratigraphic succession at intervals determined by the degree of lithologic variation. In addition to the samples taken from each unit as representative of the general lithology an effort was also made to sample local diagenetic phenomena, local lithologic variations, and the contact features between adjacent lithologies (especially in the case of calciturbidites). The detailed measurements of the sections have not been presented in the present account due to the overall volume of thesis, and secondary importance of the subject. However, they can be obtained from the author.

B. LABORATORY PROCEDURES

From each sample collected stained acetate peels were prepared for petrographic studies. Staining has been applied in order to identify carbonate minerals calcite, dolomite, and ferroan calcite (*). Peels are less time consuming, easy to prepare, and the size of the peel can be many as ten times larger than that of the thin section. Consequently, photomicrographs of peels are much sharper. The only disadvantage of the peels is the impossibility to work under the crossed nicols with petrographic microscope, and thus eliminates mineral identification. On the other hand very fine-grained carbonate rocks were found to produce nonsatisfactory peels and in these circumstances thin sections were prepared.

Staining techniques to distinguish iron-free calcite and ferroan calcite in thin sections are described in several papers (FRIEDMAN, 1959; EVAMY, 1963; DICKSON, 1965). This procedure has been expanded so that the stains from the surface of the rock also can be transferred to an acetate peel (KATZ and FRIEDMAN, 1965; DAVIES and TILL, 1968).

The preparation of a stained acetate peel was completed in four stages:
1) Preparation of sample, 2) Etching, 3) Staining, 4) Making the peel.

(1) A slab of rock was cut perpendicular to bedding and one surface grounded with carborandum powder up to 600 grit.

(2) Etching is the most important single process affecting the production of good quality peels. Etching experiments using hydrocholric

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(*) Ferroan calcite is a calcite containing small amounts of FeO (PALACHE, BERMAN, and FRANDEL, 1951).

and acetic acids of different concentrations were conducted. Acetic acid proved to be the most satisfactory etching medium. Ten per cent acetic acid (10 ml. of concentrated acid made up to 100 ml. with distilled water) was found to be of adequate strength as an etching medium. Time is another important factor in etching. While three minutes were adequate for coarse-grained samples, a quarter of an hour was found necessary for very fine-grained calcilutites and dolomitic rocks.

The one face grounded rock sample was placed on a piece of plasticine so that its surface was horizontal. Then the rock surface was flooded with diluted acid. This procedure was repeated several times paying attention to keep a layer of acid on the rock surface during all the etching duration. At the end of the etching time the sample was washed with distilled water without being moved.

(3) The third stage comprises staining. Alizerin Red-S and Potassium ferricyanide were used. Alizerin red-S stains calcite red, whereas potassium ferricyanide reacts with ferrous iron to form a blue stain. When these agents are combined they stain ferrous calcite a colour composed of red and blue. EVAMY (1963) suggested that as the iron content increases the stain would shift from red to mauve to purple. The value of differentiating iron-free and ferroan calcite has been demonstrated in several studies of carbonate rocks (EVAMY and SHEARMAN, 1965; DAVIES and TILL, 1968; EVAMY, 1969). For example, obscure diagentic features such as sequences of pore filling calcite are readily apparent when iron-free and ferroan calcites are differentiated (See Pl.53 A&B).

The procedure for staining the rock sample that has been followed in the present account is a modification of that used by EVAMY (1963), KATZ and FRIEDMAN (1965) and DICKSON (1965 and 1966). Two different solutions were prepared by dissolving 2 grams Alizerin red-S, and 20 grams potassium ferricyanide in one litre of 1.5% Hcl. These two solutions were mixed in the ratio of ARS/PF = 3/2 immediately before use and poured on to the etched rock surface. Two-and-a-half minutes later the solution was intensified by adding alizerin red-S solution. Total staining time was three minutes.

(4) When the time was completed the sample was washed with gently running tap water. Then the sample was allowed to drain but not become dry. This was found to be very important. If the sample was too dry the stain would not transfer to the peel. This finding contradicts with the suggestions of previous workers who found complete drying essential. The nearly dried surface was flooded by acetone and immediately a sheet of acetate was pressed down on the surface. A plain cellulose acetate

film .005 cm thick was found to be the most satisfactory one. Then the sample was allowed to dry completely, a process which generally requires at least half an hour. Complete drying minimizes curling and practically eliminates shrinkage. When it was dry the peel was removed and mounted between two glass plates held together with sellotape.

The procedure for staining thin sections is essentially the same as that for preparation of stained peels. The thin section was etched with 1.5% Hcl and then immersed into the staining solution. The only difference is that a 15 seconds etching and a 1 minute staining time were found to be adequate for thin sections. Above this time there is the danger of loosing the thin section. After the staining time was completed the thin section was rinsed with running water. When the thin section was completely dry it was covered by standard cover glass by conventional method.

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CHAPTER II

GENERAL REVIEW OF THE GEOTECTONIC EVOLUTION AND GEOLOGY OF ANATOLIA

I. INTRODUCTION AND HYPOTHESIS ON THE GOETECTONIC EVOLUTION OF ANATOLIA

Anatolia or so called Asia Minor is situated on the eastern part of the Alpine System, and on the western part of the Himalayan Chain. Expect Asia Minor, the overall geotectonic evolution in the Alpine System, is better understood as compared to the Himalayan Chain. Although there has been intensive geological research in the last two decades, the geotectonic evolution of Anatolia is still not clear. Much of the work is based on short local studies, and scattered data, creating a number of contrasting hypothesis.

The first subdivision of tectonic units of the area was done by NAUMAN (1896) and his subdivision was later developed by ARGAND (1924). They had divided Anatolia into two regions, "Pontique" in the north, and "Tauran" in the south (Table II-1.)

ARGAND 1924	*KOBER 1931*	*ARNI 1939*	*BLUMENTHAL 1946*	*EGERAN 1947*	*KETİN 1959-1966*
Pontique Tauran	Pontids Median Massifs Taurids	Pontids ↗North Branch ↘South Branch Anatolids Taurids Iranids Anatolian-Iranian Border Folds	Pontids Anatolids Intermediate Massifs of Central Anatolia Taurids Iranids Iraqids Syrian-Arabian Block	Pontides Anatolides Zone Intermediare Taurides Egee-Iranides Plis Bordiers Anatoliens-Iraniens Block Arabia	Pontids Anatolids Taurids Border Folds

Table II-1: Tectonic units of Anatolia proposed by different authors

According to them movement of Gondvanaland towards Eurasia was responsible for these mountain chains. ARGAND postulated that the two continents might have had similar heights and isostatic characters in this region, and no overriding occured during the collision of these continents. However, he also stated that this argument did not exclude the possibility of overriding between them.

In 1931, KOBER brought theconcept of "intermediate massifs". He compared the orogenic belt of Anatolia with the various units present in the Alps and accepted the existence of intermediate massifs between the Pontique and the Tauran belts.

Later ARNI (1939) compared the area with West Iran and proposed a new classification based on his research in East and Southeast Anatolia (Table II-1).

ARNI's classification was accepted by BLUMENTHAL (1946), but he added a new unit called "Iraquid" in the very south and separated the Palaeozoic or crystalline bæment from the Mesozoic-Tertiary cover.

EGERAN (1947) also accepted ARNI's classification in principle, but he increased the number of tectonic units to eight. He based his classification on the magmatic activity and metallogenic provinces.

The recognition of the Steinmann Trinity (pillow lava - serpentine - radiolarite) in the form of "Ankara Melange" by BAILEY and McCALLIEN (1950 and 1953) opened a new stage in the hypothesis concerning the geotectonic evolution of Anatolia. They argued that the Central Anatolian region was a "geosynclinal domain" and the thrusting of the Pontid rocks (Kırşehir metamorphic massif) over the Taurids formed the present structure of the region. According to them the crystalline massifs between the Pontids and Taurids were the segments of an enormous thrust sheet (Anatolian Thrust) which had been displaced some 350 kilometres during Late Cretaceous times and preserved in the synclinal areas of the Central Anatolia.

Concept of "long lived geosynclinal" within the region was also accepted by KETİN (1959). Sedimentation was continuous from Palaeozoic to the end of Mesozoic Era. He suggested that the ophiolite series of rocks of Upper Mesozoic System rested comfortably on the metamorphics, so that the age of these could be Mesozoic, rather than Palaeozoic, or even Precambrian. Metamorphism was attributed to the Alpine Movements according to KETİN (1966). He based his division of the tectonic units on their orogenic developments which had migrated from north to south. The units from north to south are:

(1) The Pontids (2) The Anatolids
(3) The Taurids (4) The Border Folds.

The Gürün area which is the subject of this study, lies in the zone of Taurids (Figure II-1). According to his hypothesis the Pontids were a result of Caledonian and Hersinian movements. They had risen above sea level at the beginning of the Mesozoic Era when the other parts of Anatolia were completely underwater. The Anatolids developed at the end of the

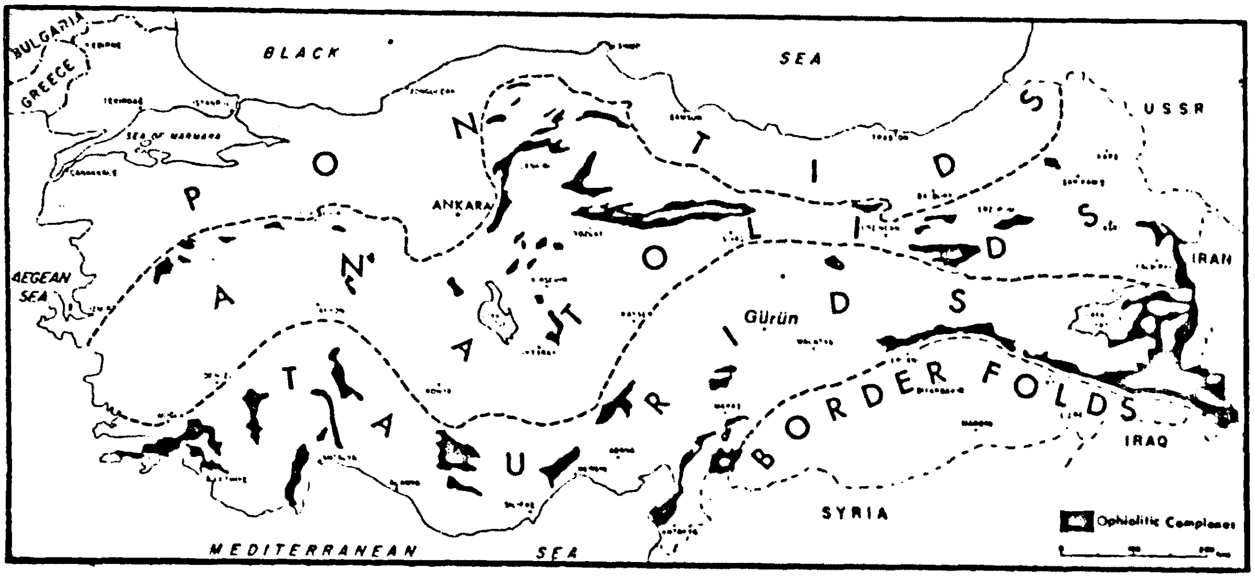
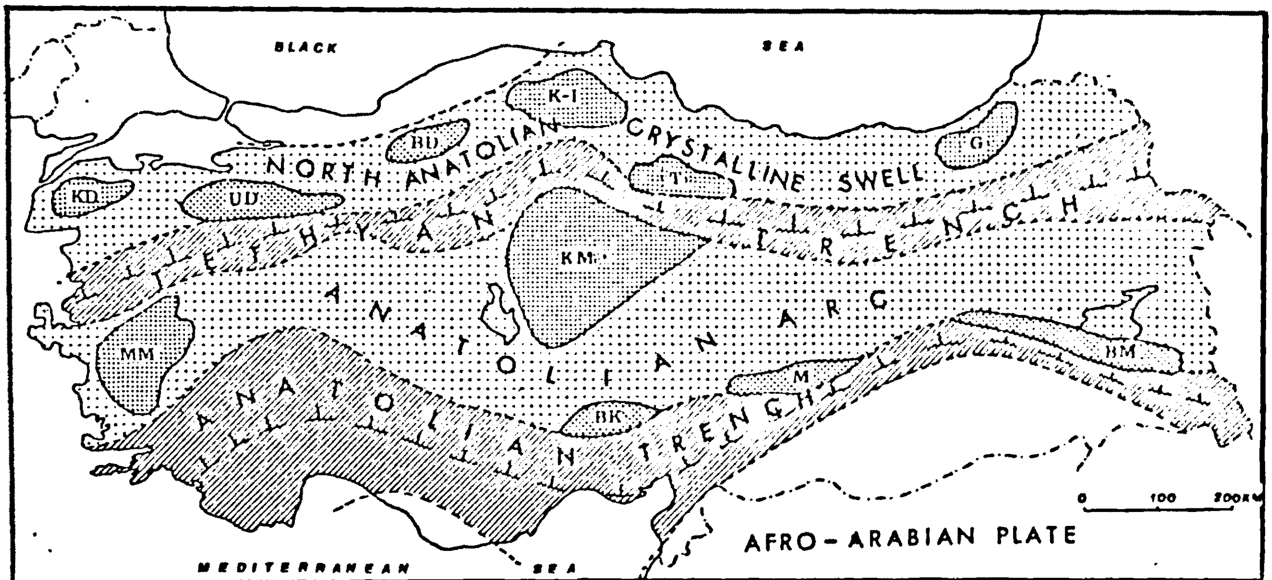


Fig.II-1: Outline of tectonic units of Anatolia and distribution of Ophiolitic Complexes (After KETİN, 1966)



BD - Boludağı Massif; BK - Bolkardağı Massif; BM - Bitlis Massif; G - Gümüşhane Massif; KD - Kazdağı Massif; K-I - Kastamonu - Ilgaz Massifs; KM - Kırşehir Massif; M - Malatya Massif; MM - Menderes Massif; T - Tokat Massif and UD - Uludağ Massif

Fig.II-2: Basement rocks and tentative reconstruction of the Late Palaeozoic palaegeography and plate tectonics of Anatolia (After KETİN, 1966; KAADEN, 1971; OKTAY, 1973, and ŞENALP, 1974)

Early Tertiary Period, and the Taurids at the end of the Oligocene Period. The youngest of all is the Border Folds Unit and its development was completed during the Pliocene times.

MURATOV (1964) approached the problem from an entirely different point of view. In a general synthesis of the tectonic development of the Alpine folded region of the Southeastern Europe and Western Asia, he tried to explain the evolution of this region as a progressive event in time and space rather than to classify geotectonic units as separate entities. Disregarding the Precambrian he recognised two periods in the geologic history of the region:

The 1st Period comprises the whole Palaeozic Era, and is terminated by the Hercynian folding. It represents an ancient period of geosynclinal development, but, because of a limited data no further division has been made in it.

The 2nd Period comprises the Mesozoic and Cenozoic Era and is called "Alpine". The Alpine Period is divided into two major stages:

Stage 1 - Geosynclinal Stage proper: Comprises whole of Mesozoic Era probably up to Early Miocen epoch. MURATOV further subdivided this stage into three phases: Early Phase; maximum development of geosynclinal troughs; and closure of geosynclines.

Stage 2 - Final stage of development and mountain building: includes the Neogene and Quaternary.

The Alpine folded region was developed above the Palaeozoic folded region. The latter outcrops either as median massifs or cores of Alpine folded structures. Median massifs are not affected by Alpine movements, The Menderes and the Kırşehir Massifs can be given as an exaple of this. On the other hand, massifs like the Istıranca, the Boludağ, the Kazdağ, the Tokat, the Ilgaz, the Malatya and the Bitlis Massifs are the cores of Alpine structures (Figure II-2). During the final stage of Hercynian geosynclinal period of development, a relief of mountains and intermontane troughs was formed and the latter were filled by molasse and coal - bearing accumulations of Upper Palaeozoic System (Zonguldak area in the north). From the commencement of the second period (Alpine), denudation of ancient relief began and a number of new basins were formed as a result of this subdued relief. They were filled partly by marine, and partly by lagoonal sediments. Volcanism played an important role in some of these troughs. Upper Triassic deposits were followed by Jurassic carbonate and argillite series, without a break in deposition. In the areas of maximum uplift sediments are generally absent, and these areas correspond to the present day median massifs. The accumulation of carbonate sediments in the early stage of Alpine development, took place in extensive areas of the North and South Anatolia (The Pontus and the Taurus Systems). In some places, diabasic and porphyritic eruptions, tuffs and accompanying siliceous rocks disturbed the uniform carbonate successions; radiolarites, associated with limestones are present at the several parts of the Taurus System. The second phase started with the appearance of the new basins during the Late Jurassic - Early Cretaceous epoch. The beginning and duration of the second phase varies in time, throughout the Alpine System. In the north, the Western Pontus Trough (region of Zonguldak Coal Basin originated in the middle of the Early Cretaceous, whereas in the east, the East Pontus Trough was formed in the beginning of the Late Cretaceous. At the same time two troughs had been generated in the Taurus zone (Figure II-3). The large Ankara Trough, which branches southward from the Pontus Trough, was formed in the Early Eocene time. These geosynclinal troughs of the second phase, were filled by flysch - type sediments (sometimes associated with volcanic rocks) and therefore is called "flyscheogenic" (MURATOV, 1964; p.109). It is the climax of the geosynclinal development of the Alpine region. The third phase started with the uplifting of the geosynclinal troughs. Intensive folding and uplifting, occurred during the Late Eocene - Oligocene epoch and eliminated geosynclinal conditions in Anatolia. From Oligocene time onward only intermontane depressions were left to be filled by molasse. In the final stage of development, extensive mountain uplifts and intermontane and submontane troughs, were imposed on various older structural elements. Volcanic activity also accompanied the uplifting. The tectonic structures of the Alpine geosynclinal area, as proposed by MURATOV (1964) is shown in Figure II-4.

In the light of recent theories of plate tectonics, HORSTINK (1971) discussed the geological evolution of Eastern Anatolia during Late Cretaceous and Tertiary times. At the same time he gave a brief review of the Early Mesozoic history of the area. According to present day theories ophiolites are generated as part of an ocean floor during a phase of ocean floor spreading (REINHARDT, 1969) rather than as outflows along fissure zones at the edges of geosynclines (AUBOUIN, 1965). Starting from this point, HORSTINK recognized five megatectonic units in the Mesozoic and Cenozoic development of the Eastern Anatolia, based on the locations of the oceanic troughs in which the ophiolites could have been generated. These units, from north to south, are:

(1). The Pontids: In this unit ophiolites are absent, but he believes that it overlies the ophiolites generated in the Northern Tethys Ocean.

(2) The Northern Tethys Ophiolite Nappe; This unit overlies the Anatolids and its rootzone is between the Pontids and the Anatolids. It consists of rocks which originally formed the floor of the Northern Tethys Oceanic Trough.

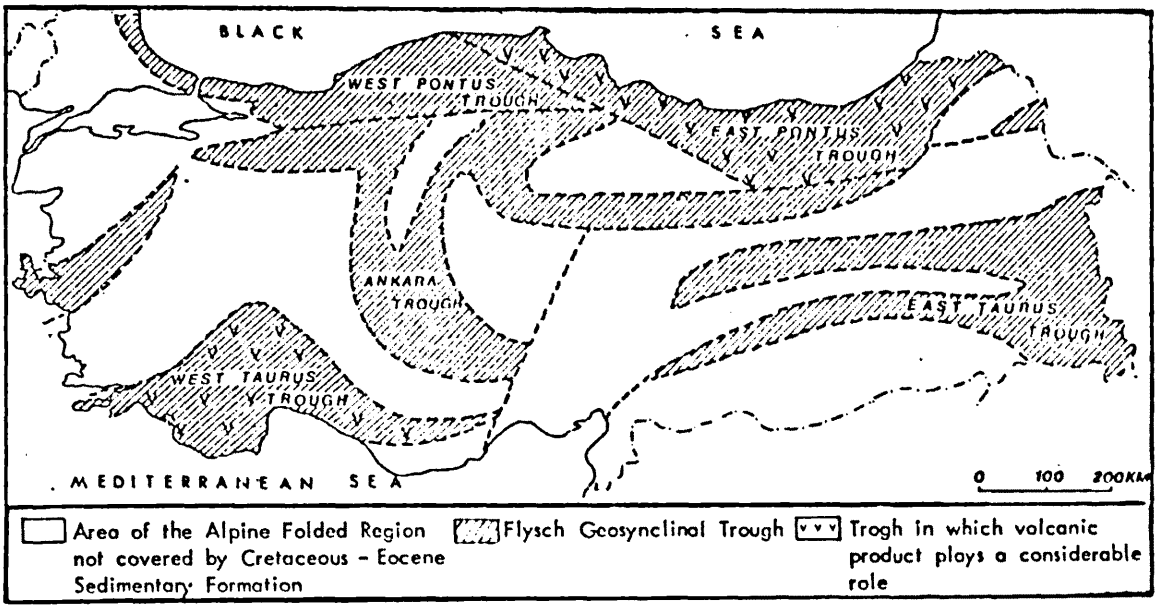


Fig.II-3: The Alpine folded region of Anatolia in the Second (Flysch) Stage of geosynclinal development "Cretaceous-Middle Eocene" (Modified After MURATOV, 1964)

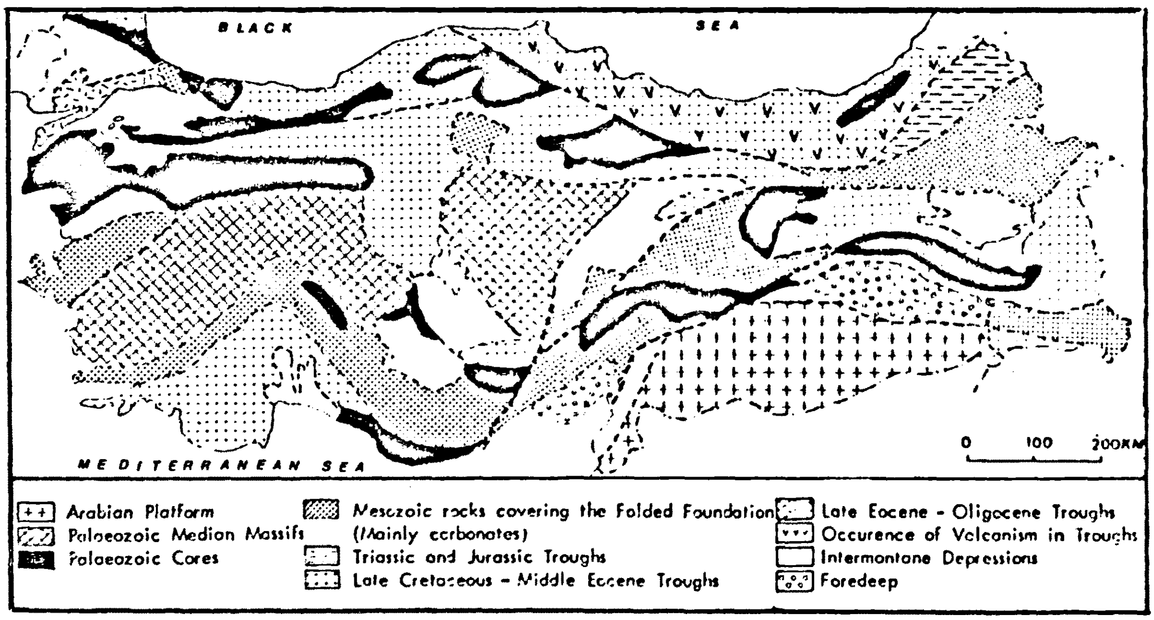


Fig.II-4: Scheme of tectonic structure of the Alpine Geosynclinal area in Anatolia (Simplified After MURATOV, 1964)

(3) The Anatolids: This unit acted as a fairly rigid crystalline massif throughout the Alpine history of the area, and underlies the Northern Tethys Ophiolites.

(4) The Southern Tethys Ophiolite Nappe: It overlies the northern edge of the Arabian Platform. Its rootzone is between the Anatolian and Arabian Platforms and its rocks were originated as the floor of the Southern Oceanic Trough.

(5) The Arabian Platform.

HORSTINK also recognised three major tectonic phases in the Mesozoic and Cenozoic history of the Eastern Anatolia:

A. Pre-Senonian or Pre-Ophilite Phase: During this phase there were three major continental masses. They were Pontia, Anatolia and Arabia, and were separated by the Northern Tethys Oceanic Trough and the Southern Oceanic Trough. They were drifting towards each other (Figure II-5).

B. Senonian or Ophiolite Emplacement Phase: The continuous drifting reached its climax, during the Senonian time, and three continental masses collided. As a result the Northern Tethys ophiolites were pushed southwards as nappes over parts of Anatolia, and the Southern Tethys ophiolites were pushed on Arabia (Figure II-6). More than one oceanic belt idea is also supported by GIANELLI, PASSERINI and SGUAZZONI, 1972; p. 480). HORSTINK argues that the youngest dating below the ophiolites, gives an age of Cenomanian, or probably Turonian, while the overlying sediments are Lower Maestrichtian, and therefore he concludes a Senonian age for the emplacement of ophiolities (HORSTINK, 1971; p. 27). But the age of the ophiolites is still contraversial.

C. Post-Ophiolitic or Late Cretaceous and Tertiary Phase: This phase is characterized by the formation, infilling, and deformation of intermontane troughs. Three sedimentary cycles are distinguishable in this phase:

(1) The Upper Cretaceous-Middle Eocene Sedimentary Cycle: As seen on Figure II-6 there was a stable Anatolia Platform in the centre and it was mainly a land area, but locally some carbonate shelves and shallow restricted embayments, developed on it. An exception to this is the Sivas area where a deeper intermontane trough existed. The Gürün area formed a large carbonate shelf on which a thick succession of shallow marine limestones were deposited. The eastern part of this basin formed an enclosed basin where rudist limestones, coastal and continental clastics

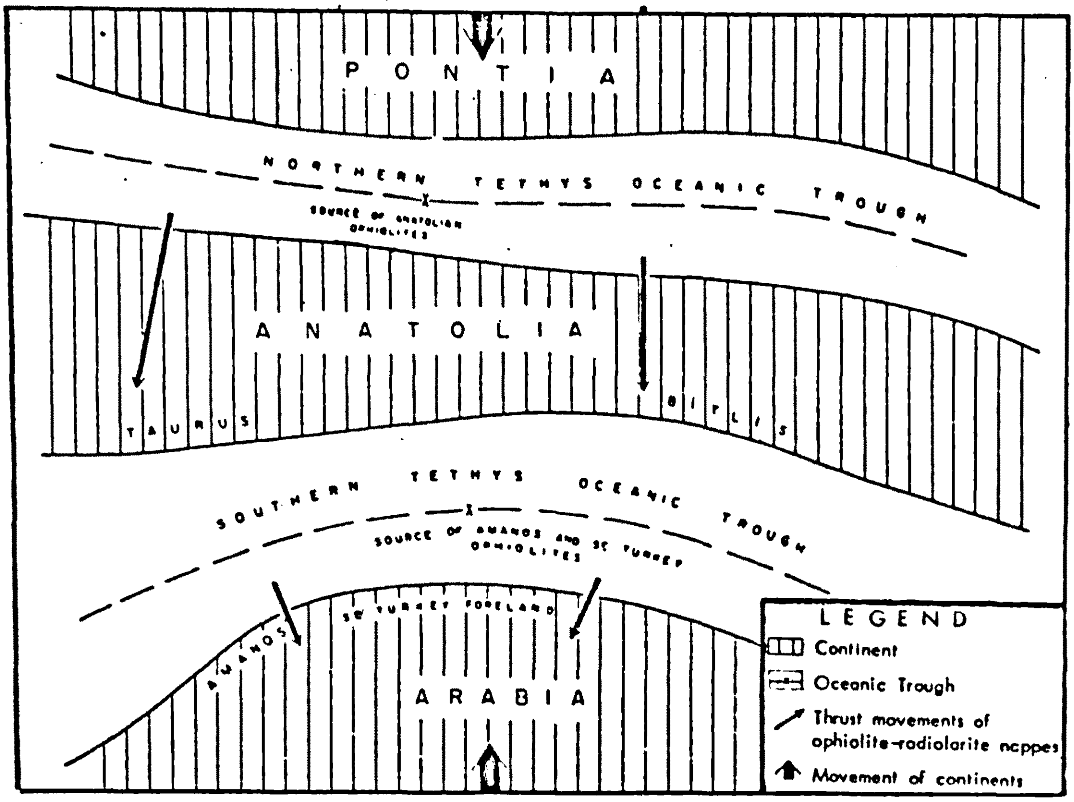


Fig.II-5: Hypothetical Mesozoic palaeogeography of Eastern Anatolia (After HORSTINK, 1971)

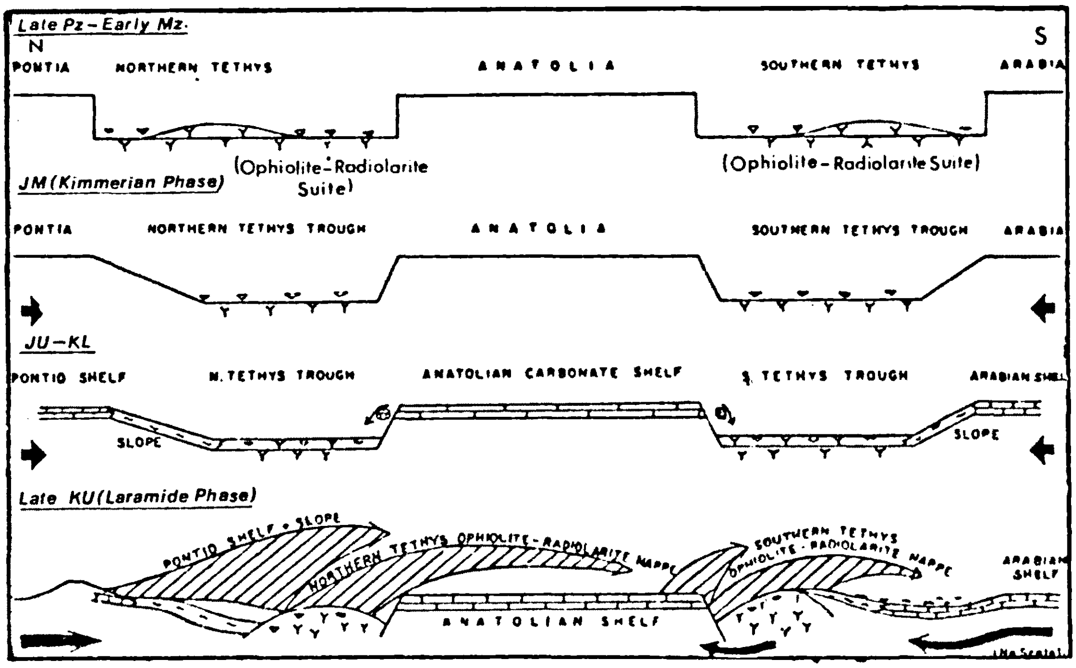


Fig.II-6: Schematic presentation of the "Mesozoic" structural evolution of Eastern Anatolia (After HORSTINK, 1971)

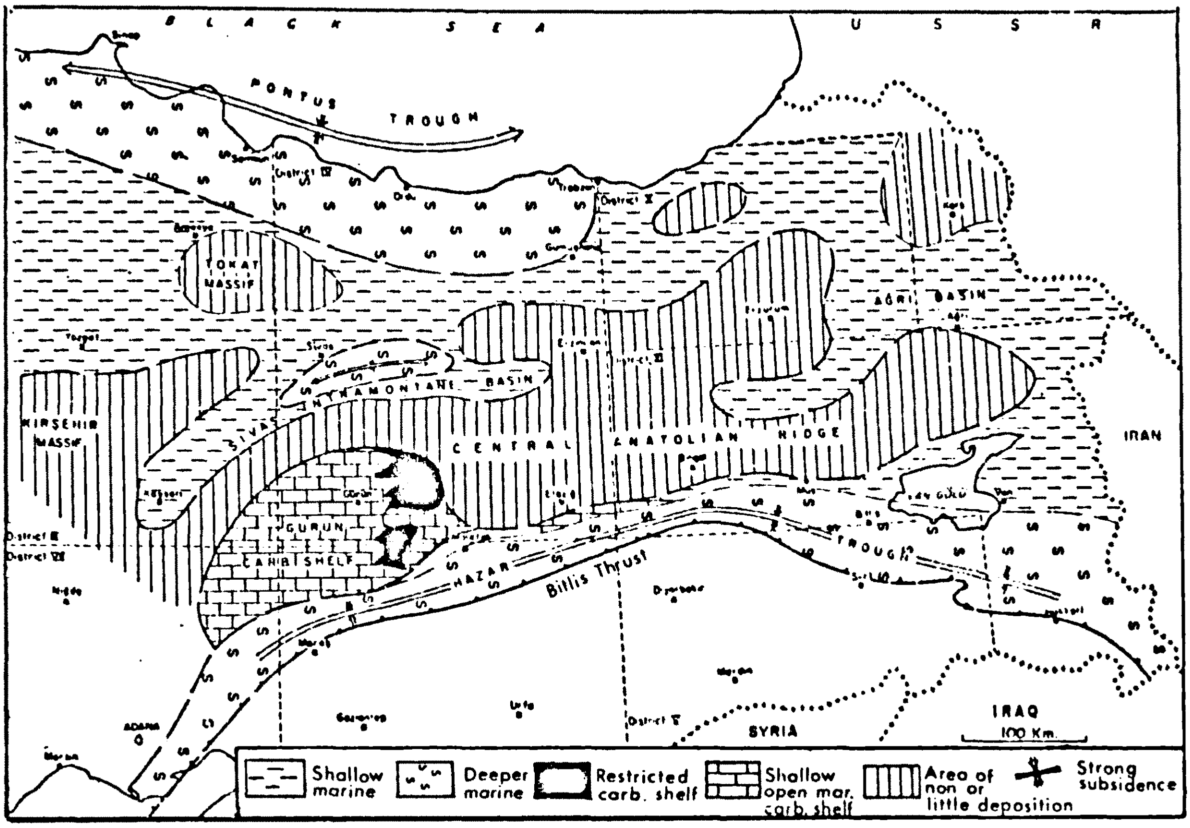


Fig.II-7: Palaeogeography of Upper Cretaceous - Middle Eocene sedimentary cycle in Eastern Anatolia (After HORSTINK, 1971)

developed (Figure II-7). To the north and south, the Anatolian Platform was bounded by deeper marine troughs (namely Pontus and Hazar Troughs).

(2) The Middle Eocene-Oligocene Sedimentary Cycle: Due to continuous compressional forces from the north and south, the Pontus and Hazar Troughs were uplifted. The whole Anatolian Platform was tilted towards the south, and severe downwarping of its southern part developed in an E-W elongated chain of intermontane troughs.

(3) Lower Miocene-Pliocene Sedimentary Cycle: It started with a widespread shallow transgression which covered the whole of the Eastern Anatolia. Regression set in during the Middle Miocene epoch and continental lagoonal sediments were deposited. The latest and strongest tectonic movements in Early Pliocene times resulted in strong deformation of the edges of the Anatolian Platform (the Sivas Thrust in the north, and the Bitlis Thrust in the southeast). Pliocene to recent times have been characterized by tensional movements which led to strong volcanic activity. Today, the greater part of Eastern Anatolia is covered by lavas originated from huge volcanoes, such as Ağrı Dağı (Mount Ararat) and Süphan Dağı.

In a later work, BRINKMANN (1972) concluded that there were narrow troughs running across Anatolia during the Mesozoic Era. His conclusion is based on the close relationship between the large scale topographic features of the earth's surface, and the structure of its crust. He also argued that they would have had similarities to present features, such as the Gulf of California, and the Red Sea. At the same time, he separated the ophiolites into three age groups, ranging from Trissic to Palaeocene and discussed their distribution. But, as pointed out by BRUNN and MONOD (1973), the ages assigned to the ophiolites are highly conjectural.

In a review of the orogenic development of Eastern Anatolia, İLHAN (1974) mostly followed the classification proposed by KETIN (1966) but added a new unit called "The Epirogenic Basins and Young Volcanic Regions".

II GENERAL GEOLOGY AND PLATE TECTONIC HISTORY OF ANATOLIA

In recent years the development of the concept of plate tectonics, which is an outgrowth of the theory of sea-floor spreading, has provided a global explanation for the geotectonic eveolution. The concept of plate tectonics presumes that the outer crust of earth consists of a comparatively strong lithosphere about 100 kilometres thick, overlying a weaker asthenosphere. The lithosphere is divided into a number of rigid plates (6 -20) bounded by trenches, rifts, and megashears. The continents are throught to be superficial passengers on these plates, and continental drift is simply a result of sea-floor spreading during the course of plate tectonic activity. Plates are generated at the rifts by ocean-floor spreading (HEIRTZLER and LE PICHON, 1965, VINE, 1966) and destroyed in the trenches where island-arc systems are formed by a magma, derived from partial melting of the descending plate under high load and temperature (GREEN and RINGWOOD, 1963). Lastly, in the light of recent works, it can be argued that geosynclinal evolution and mountain building, are a direct result of rifting and trench development caused by plate movements.

The Palaeozoic plate tectonic history of Anatolia is poorly known. This is partly due to insufficient data , and partly to the superimposition of the Alpine Mobile Belt on the older streutures. As pointed out earlier (p. 23) the Alpine folded region was developed above the Palaeozoic folded foundation, which is exposed extensively either as core of Alpine folded structures, or as median massifs. The latter was not affected by Alpine Movements. The Palaeozoic foundation is composed from crystalline metamorphics and/or non-metamophic rocks. Because of the lack of detailed stulies, and fossils, their ages are not clearly determined yet.

There is an agreement that a big sea-way, called "Tethys" existed during the Palaeozoic Era. It was situated between Laurasia and Gondwanaland (DIETZ and HOLDEN, 1970, BAIRD, 1971; SMITH, 1971; DEWEY, PITMAN, RYAN, and BONNIN, 1973). In North Anatolia, an E-W trending swell, called North Anatolian Crystalline Swell (BRINKMANN, 1968) was formed. The Tethyan Trench is thought to be the result of the evolution of this swell (OKTAY, 1973). It is believed to be the major cause of Palaeozoic Orogen of North Anatolia (DEWEY and BIRD, 1970; DIETZ and HOLDEN, 1970). The North Anatolian Crystalline Swell was marginal to Pontic Land, and extended along the Black Sea coast from Sinop to the Turkey-USSR border, and was approximately 150 kilometres wide. The Serbo-Macedonian and Transcaucasian Massifs are the west and east extensions, of this zone respectively (Figure II-8).

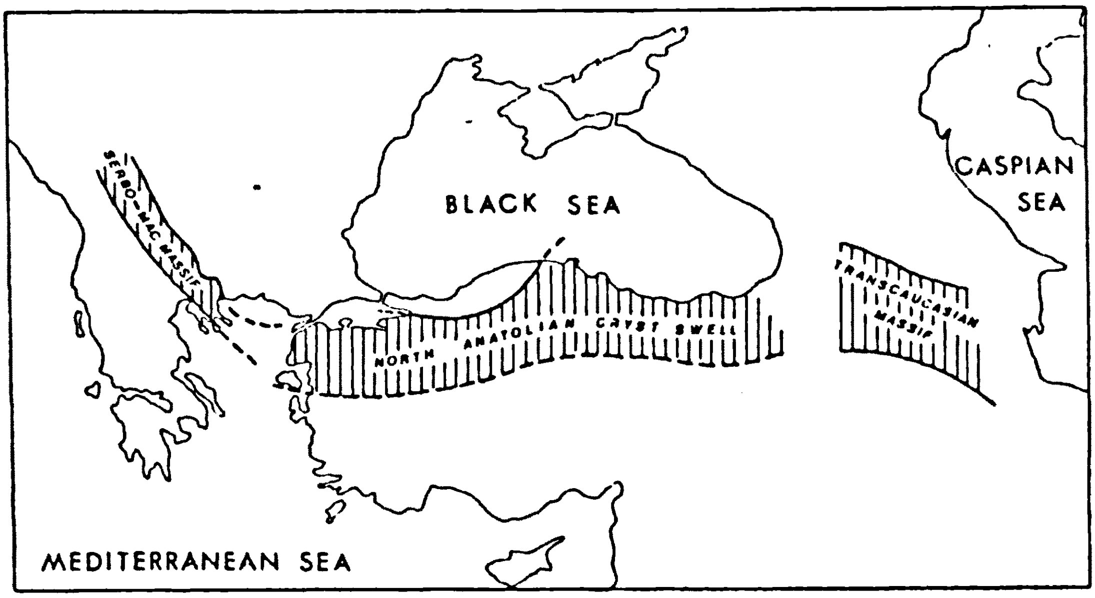


Fig.II-8: The North Anatolian Crystalline Swell and its continuation (After BRINKMANN, 1968)

The North Anatolian Crystalline Swell is composed of a chain of metamorphic massifs, from W to E, the Kazdağ, the Uludağ, the Boludağ, the Kastamonu-Ilgaz, the Tokat, and the Gümüşhane Massifs (Figure II-2). In general, the core of these massifs is composed of catazone - amphiobolite facies metamorphic rocks and is surrounded by a lower grade metamorphic green-schist - glaucophane-schist sequences (KAADEN, 1971). The core and surrounding metamorphic sequences are often overlain by unmetamorphosed Palaeozoic sediments. BRINKMANN (1969) is of the opinion that movements which formed the swell are Variscan (Late Palaeozoic) Movements. On the evidence of stratigraphic relationships with the non-metamorphic Palaeozic sequences, the age of the green-schist - glaucophane-schist is thought to be Precambrian, and the metamorphism is attributed to the Caledonian Movements (KAADEN, 1971). On the other hand, in an earlier study, KETİN (1962) came to a different conclusion. In the light of his observations in the Kastamonu-Ilgaz and Tokat Massifs, he argued that the age of an important part of the widespread metamorphic series, could be Mesozoic rather than Palaeozoic, or Precambrian. According to him, the Alpine movements are responsible from the metamorphism (for further discussion the reader is referred to KETİN, 1962). The age of these massifs was also discussed by BOCCALETTI, BORTOLOTTI, and SAGRI (1966a). The result of their field observations showed that the Alpine Metamorphic Phase of Late Cretaceous had only a secondary significance in the petrogenesis of the Ilgaz Massif. They believed that the rocks of this massif had already got their metamorphic character (in the Hercynian Metamorphic Phase). During the Late Carboniferous, this swell was partly submerged and coal-measures of Zonguldak Area were deposited. In Early Permian the environment of deposition, was rather deeper marine, and extensive flyschlike deposits (greywackes and olistoliths) were formed.

In the Middle Anatolian region another metamorphic belt was created. In this belt massifs like the Menderes, the Kırşehir, the Bolkardağı, the Malatya, and the Bitlis were exposed. The Menderes and Kırşehir Massifs are median massifs of the Alpine folded belt of Anatolia (MURATOV. 1964). The Menderes Massif takes place in the western part of this belt (Figure II-2). The core of it is augen-gneises and forms a dome (SCHUILING, 1962). From the core to the flanks three zones can be distinguished; next to the core is biotite-muscovite-garnet schists; in the second zone diorite-muscovite-epidote schist, and finally thick marbeles (GRACIANSKY, 1966; BRUNN et al, 1971). A lead isotope age determination on pitchblende from the gneises of the Milas region yielded an age of 268 $\pm$ 40 million years (DURAND, 1962) which indicates that the metamorphism of the core is at least Hercynian. Still the age of the surrounding metamorphics is controversial. SCHUILING (1962) suggests a Silurian-Permocarboniferous age; whereas BRINKMANN (1966) supports an age of Post Liassic - Pre Lower Cretaceous.

The second massif in this belt is the Kırşehir Massif. No detailed study is available on the stratigraphy and metamorphism of this massif, but generally it has a similar metamorphic zonation with the Menderes Massif. Gneisses can be observed in the core and lower grade metamorphics surround it (BRINKMANN, 1971). The age of the massif is still unknown.

The rest of metamorphic massifs in this zone is exposed in the mega-anticline of the Bolkardağı-Malatya-Bitlis region. The Bolkardağı Massif is situated at the west of this zone, and has a coarse crystalline marble core surrounded by green-schists and phyllites, and overlain by Permian limestones (OKTAY, 1973).

The Malatya Massif has a core composed of gneises, micaschists, and marbles, which is enveloped by green-schist metamorphics. Unmetamorphosed Palaeozoic deposits overlie the massifs and therefore the age of the metamorphics could be earlier than the Silurian (BAYKAL, 1966).

The Bitlis Massif is similar to the Malatya Massif. The core of paragneisses, amphiobolites, and micaschists is surrounded by chlorite and talc schists and recrystallized limestones (YILMAZ 1971). These sequences are overlain by partly recrystalized limestones of Permian age (BORAY, 1972). The Bitlis Massif was intruded by granites. The age of intrusion was determined by YILMAZ (1971) as 325∓3 million years which corresponds with the Hercynian Movements.

This metamorphic belt is interpreted as an island-arc by OKTAY (1973) and named "Anatolian Arc". The trench, which was responsible for the evolution of this arc is called "Anatolian Trench" (Figure II-2). The consumption of the oceanic crust, and the collision of the Laurasia (Pontic Land), Anadolian Arc, and Gondwanaland by the Tethys and Anatolian Trenches, took place in the Late Palaeozoic time and caused the closure of Tethys and formed the basement of Anatolia (OKTAY, 1973).

In general, it may be accepted that the foundation of the Alpine folded region of Anatolia was formed during the Palaeozoic Era. Both Caledonian and Hercynian Movements played important roles, and the Palaeozoic rocks were deformed into large, open folds. The mega-anticlinoria became the future sites of the trough development, strongly influencing the development of the Alpine folded belt in Anatolia.

The collision of Laurasia and Gondwanaland in the Late Palaeozoic time by moving to a common trench in Tethys, is also suggested by DIETZ and HOLDEN (1970). So the "Pangaea" was created. According to DIETZ and HOLDEN it was broken up again in Triassic, and Tethys was re-opened. Opening during the Triassic is controversial because the Triassic generally represents a tectonic calm in Anatolia. The Early Jurassic is the probable time of opening, because at this time, a regional transgression occurred in Anatolia.

In the northwest of Anatolia, Triassic deposition took place in the shallow basins which were left between the mega anticlinoria. In the İzmit region continental redbeds were deposited. In the west, in the İzmir region, an advancing transgression from Greece and the Aegean Sea covered the area, and neritic limestones were deposited (BRINKMANN, 1971). In the southwest, in the Taurus Belt, Triassic is characterised by dark coloured, neritic, dolomitic, limestones and forms a comprehensive series containing pelecypods Pseudomonotis and Myophora, and the Algal form Diplopora (ERENTÖZ, 1966). Further east, in southern Anatolia, carbonates are still the dominant lithofacies KAMEN-KAYE 1971). Still further east, a standard section of a few hundred meters of carbonate and evaporite deposition, is valid over a large area. This standard section was obtained from a well near Gaziantep (RIGO DE RIGHI and CORTESINI, 1964). Near the Turkey-Iraq border, there are wide variations in the facies. A total of approximately 400 metres of Triassic outcrops here. The lowest part consists of flysch-like sandstone, and alterations of sandy limestones and gray sandstones. Upwards they pass into fine grained, light coloured limestones which continue into the Jurassic (ERENTÖZ, 1966).

With the commencement of Jurassic, most of the structurally low areas were progressively invaded from both the west and east. Probable submergence of the North Anatolian Crystalline Swell, initiated this transgression and silisiclastics, limestone tuffs, and andesitic lavas were deposited in a newly developed trough. This trough was lying along the Black Sea Coast (Figure II-9) and is called "North Anatolian Lias - Dogger Trough" (BRINKMANN, 1971). At the north of this trough the "Pontic Land" was situated. In the Ankara region the Early Jurassic epoch is represented, mainly by calcareous deposits; whereas further south, in the Tauras range, a continuous sequence of partly dolomitized calcareous rocks is found. Some chert nodules are associated with thesse sediments. In the southeast, Jurassic forms a part of a comprehensive Triassic-Jurassic series consisting mainly of carbonate and evaporites (RIGO DE RIGHI and CORTESINI, 1964). Although probably the rifting started in the Early Jurassic, the main phase of sea-floor spreading is in the Late Jurassic time during which basic igneous activity, and abrupt facies changes took place. As a result of the early phase of Alpine Orogenesis, sediments of the North Anatolian Lias-Dogger Trough were folded and emergence occurred (BRINKMANN, 1971). By the Jurassic rifting two Early Alpine troughs were formed. The regions occupied by these troughs had been covered by shallow seas during the Late Palaeozoic and Early Mesozoic times. One of these troughs was located in the North Anatolia (Figure II-10) extending to the Vardar zone in Greece in the west, and to the Lesser Caucasus in the east. This trough has been named the "İzmir-Ankara Furrow" by LISENBEE (1971) and "İzmir-Ankara-Sivas Trough" by OKTAY (1973). Because the first two names characterise the western part of Anatolia only, OKTAY's terminology will be used in future paragraphs. Similarly the development of another trough in southeast Anatolia is marked by radiolarite deposition, and basic igneous activity (RIGO DE RIGHI and CORTESINI, 1964). This trough was extended to the Pindus Zone of Greece in the west and to

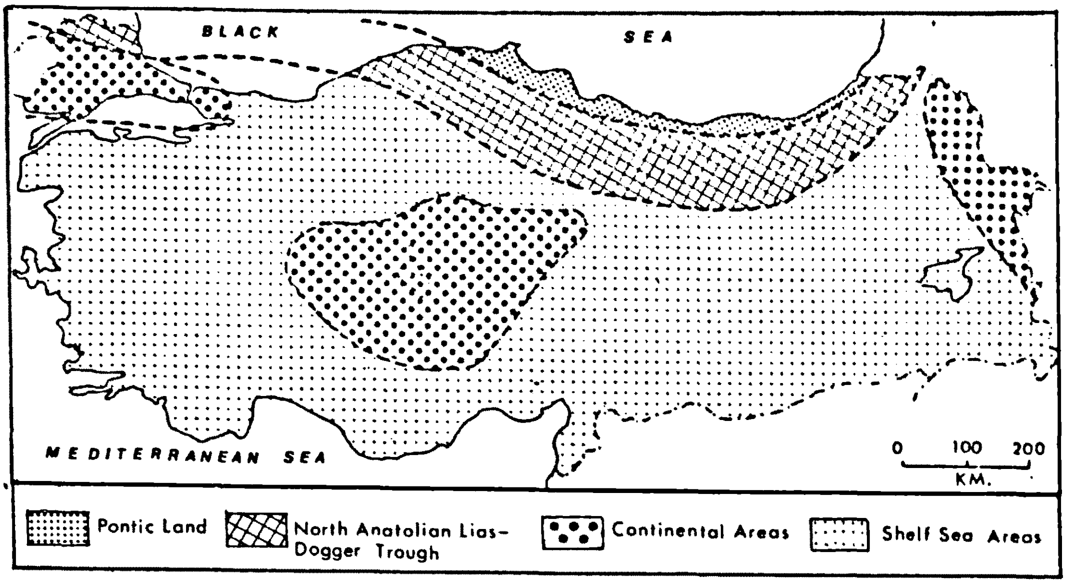


Fig.II-9: Palaeogeography of the Lower and Middle Jurassic in Anatolia (After BRINKMANN, 1974)

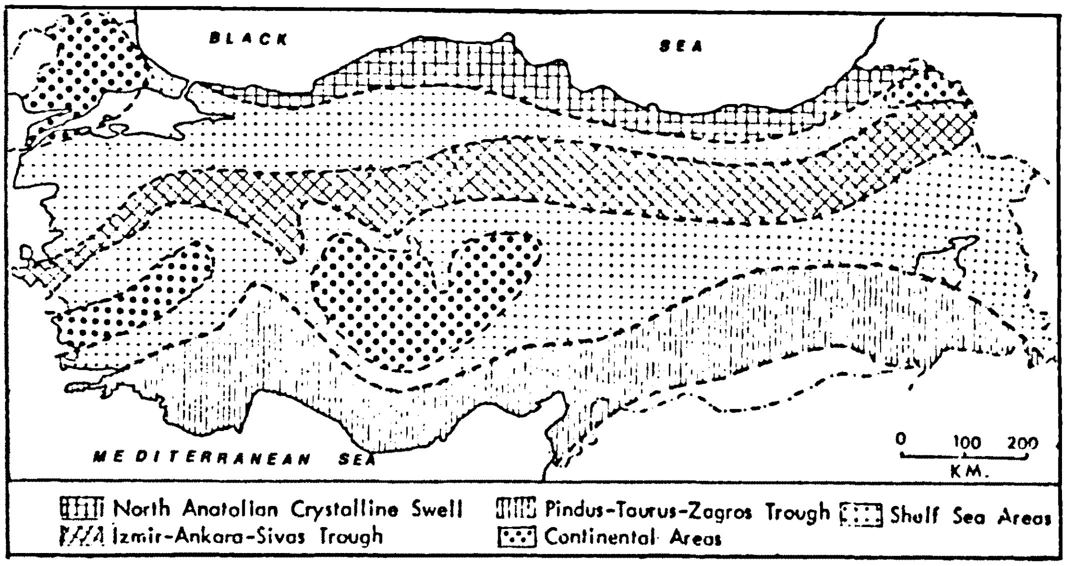


Fig.II-10: Palaeogeography of the Middle Cretaceous in Anatolia (Modified After BRINKMANN, 1974) the Zagros crush zone of Iran in the east. Therefore it has been named the "Pindus-Taurus-Zagros Trough". Further south, near the border of Syria, there is the indication of Late Jurassic-Early Cretaceous period of erosion. Pre - Cretaceous sediments have been eroded from a large area in this zone exposing Lower Palaezoic or even Precambrian formations (TEMPLE and PERRY, 1962).

In Cretaceous times the development of existed troughs reached its maximum. The position of cherty limestones, marls, and radiolarites was frequently accompanied by basic igneous activity. A new phase of development started with the commencement of Late Cretaceous epoch. Enormous masses of chaotic material was deposited in these troughs. The formation of these occurences, which have often been interpreted as gravity slides, involved large scale slumping and they were accompanied by turbidites (RIGO DE RIGHI and CORTESINI, 1964; SESTINI, 1971). However, regional thrusting has been also reported during the evolution of these mèlanges (BRUNN et al, 1971; LISENBEE, 1971). An over 3000 metre thick gravity slide, composed of allochthonous sheets of marls, shales, silisious limestones and radiolarian cherts of probably Mesozoic age, and basic igneous rocks, was found intercalated in an Upper Cretaceous sequence (TEMPLE and PERRY, 1962; RIGO DE RIGHI and CORTESINI, 1964). This drastic change of tectonic environment suggests the evolution of trenches which caused first phase closure and mountain building in the Early Alpine troughs. In the west, south of the Menderes Massif a mobile small basin was left and filled with Palaeocene and Oligocene clastic series (BRINKMANN, 1971). In southeast Anatolia, another mobile basin formed and red beds and limestones were deposited (RIGO DE RIGHI and CORTESINI, 1964). Central Anatolia kept its continental status until the Late Cretaceous time. But, during this time, intensive basic activity possibly initiated by rifting and ocen floor spreading, indicates development of a new trough in the Bolkardağı-Ulukışla-Tuz Gölü region (OKTAY, 1973). In the south, in the Hatay-Cyprus region, another trough developed. This trough was indicated by a sudden change of depositional environment - from shallow water to deep marine - and accompanied intensive basic igneous activity. Folding and emergence of this trough occurred in the veryLate Cretaceous (SCHWAN, 1971). At the same time, in the East Pontids, rapid submergence of the region initiated the development of a new deep trough (TOKEL, 1973). Its trend was parallel to the present day coast line. This trough was ultimately filled with 2000-3000 metres of flysch type rocks intercalated with andesitic volcanics. An uplifted area limiting the Pontic Trough to the north can no longer be recognised on the basis of facies differences in the Late Cretaceous, and most probably, by this time, the Pontic Land had already been submerged (BRINKMANN, 1971; p. 182). Intense tectonic activity in the Late Cretaceous had partly eliminated true geosynclinal conditions of previously developed the İzmir-Ankara-Sivas trough and left a number of small basins. They were generally east-west elongated, long and deep depressions. These were from west to east; the Bolu-Kastamonu Basin; the Haymana Basin; the Çankırı-Çorum Basin; the Sivas Basin, the Erzurum Basin, and the Muş-Hınıs Basin (OKTAY, 1973; ŞENALP, 1974). Dominant rock type is silisiclastic deposits of flysch and molasse facies type. Olistoliths are abundant in the Cankiri-Corum Basin; the Erzurum Basin; the Bolu-Kastamonu Basin; and the Sivas Basin (SESTINI, 1971; SENALP, 1974).

This new tectonic frame of Anatolia persisted until the Middle-Late Eocene times. During this time the second major phase of mountain building occurred. Only the remaining basins of the Southeastern Taurus-Zagros Trough were not influenced by these movements. Intense folding and thrusting can be traced in the İzmir-Ankara-Sivas; the Western Taurus-Pindus and East Pontids trough zones. In Bolkardağı-Ulukışla-Tuz Gölü Trough region some small basins were left and later they were filled by molasse. During Oligocene-Middle Miocene Period an extensive evaporite and clastic deposition took place in the intermontane basins of Central Anatolia. In the Middle-Late Miocene times the collision of Pontic Craton, Mid Anatolian Continental Blocks, and the Afro-Arabian Craton by the further activity of the Late Cretaceous trenches, in the western Taurus-Pindus and Southeastern Taurus-Zagros region, took place. This is the third and final phase of closure and mountain building. At the end of this final phase marine conditions ceased in Anatolia completely.

Since the Pliocene Period only epirogenic movements are active, causing large scale block-faulting. Sedimentation only continued in freshwater lakes. The present day topography of Anatolia started building up. Faulting caused the renewal of volcanic activity. Almost all of Anatolia has been covered by andesitic, dacitic, and basaltic lavas and tuffs. Some volcanoes were active until the modern times.

Today the most active sedimentary process in Anatolia, is denudation although there are extensive areas of continental and plain deposits. To the north and south, deep basins of the Black Sea and the Eastern Mediterranean respectively, are sites of turbidite deposition. Further seawards, fine-grained marly sediments are present in the Black Sea and calcareous oozes are found in the Mediterranean Sea (EVANS, 1971).

Present plate tectonics of the Mediterranean region was reviewed by McKENZIE (1970). According to him three major plates, Eurasian, African and Arabian Plates, and two minor plates are present in the Eastern Mediterranean. One of the minor plates is the Aegean Plate. It covers part of Greece, Crete, and part of the Western Anatolia. The second one is the Turkish Plate which covers most of the Anatolia and Cyprus. (Figure II-11).

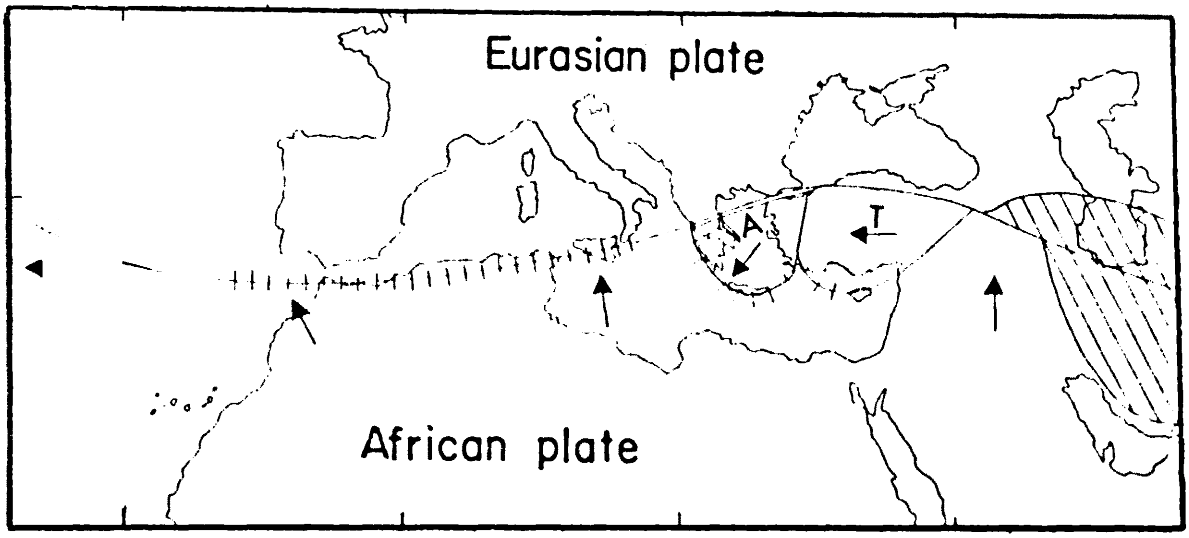


Fig. II-11: Plate tectonics of Mediterranean Area (After McKENZIE, 1970). The arrows show the direction of motion relative to the Eurasian Plate. The nature of the boudaries is indicated as follows: Double lines indicate lithosphere production ; short lines at right angles indicate lithosphere consumption ; cross hatching denotes a region of crustal thickening deduced from fault plane solutions of shallow focus earthquakes. A - Aegean Plate ; T - Turkish Plate

The position of plate boundaries and the magnitude, the direction of relative movements between adjacent plates were based on the location of epicentres and focal mechanisms of earthquakes. The fault plane solutions show that the Aegean Plate is moving towards the southwest relative to the Eurasian Plate, producing extension and strike-slip on the boundary between them. On the other hand, the seismicity of the Turkish Plate and the fault plane solutions shows that it is moving approximately to west relative to both Eurasian and African Plates. The northern boundary of this plate is the North Anatolian Fault which is an active right-hand strike slip fault. The calculated motions of these plates indicates that the closure of the Mediterranean is still continuing at the present time. The African and Arabian Plates are moving towards the north, and being consumed in the Ionian Trench. The future collision of the north east Africa with Greece and Anatolia is inevitable (DEWEY and BIRD, 1970).

It may be said that the geotectonic evolution of Anatolia is originally the result of the movements of Eurasia, Afro-Arabian, and Antolian Plates of the Palaeozoic Era. The Alpine ocean-floor of Tethys is represented by the zones of green rocks which can be observed in Anatolia today (Figure II-1).

III GREEN ROCKS (OPHIOLITES) OF ANATOLIA

Series of ultrabasic rocks and associated sediments are frequent in Anatolia, where their distribution suggests that in this particular aspect, Anatolia is genetically an "Alpine" continuation of Europe. This complex series is represented on the Geological Map of Turkey; 1:500.000, as "Mof" (i.e. Mesozoic Ophiolitic Series). BAILEY and McCALLIEN (1953) called it "Ankara Mélange" while GANSSER (1959) called it "Coloured Mélange". More recently SESTINI (1971) has proposed the term "Ophiolitic Mélange" for the same unit.

Ophiolites are believed to be slices of oceanic crust and mantle tectonically emplaced in orogenic belts and probably generated by axial plate accretion at oceanic ridges, and by diffuse slow spreading in marginal basins, behind and within, island arc complexes. Therefore the distribution of these rocks is quite significant when considering the geotectonic evolution of the region. They have complex internal, igneous, structural, and metamorphic relationships. This is most probably related to processes involved in their generation at ridges and in marginal basins having no significance in terms of processes within the orogenic belts at consuming plate margins either beneath and behind oceanic trenches (subduction zones) or bythrusting onto continental margins (abduction zones) when a continental margin meets a subduction zone (DEWEY and BIRD, 1971). Some other mechanisms are proposed for the emplacement of ophiolites. One of the earlier thoughts is igneous intrusion which has lost its popularity in the light of the plate tectonic theories. It has been shown that some ultramafic masses are tectonically intrusive. This emplacement mechanism is called "Protrusive" (the word "Protrusion" was first used by LYELL (1871) to describe masses of crystalline rock injected into sedimentary beds by tectonic processes). It assumes that ultramafics were slabs extruded from depth in a solid state. A third mechanism has been proposed by LOCKWOOD (1971 and 1972) and it is a non-instrusive, sedimentary process. Here the word "sedimentary" covers all depositional processes including gravity sliding. On the other hand, MAXWELL (1973) believes the "diapiric emplacement" of the ultramafic rocks. He also is of the opinion that ophiolites are generated internally within an oregon (MAXWELL, 1973; p.17). For further discussion on the emplacement of ophiolites see DEWEY and BIRD (1971), MOORES and MACGREGOR (1972), LOCKWOOD (1971), and MAXWELL (1973).

The age of ophiolites is highly controversial. Ages ranging from Palaeozoic to Cenozoic are offered First of all a distinction must be made between the age of the magmatic activity and that of the later tectonic emplacement. Cretaceous or Eocene ages, previously inferred by the geometric position of these ophiolites have been reconsidered because in some cases they concern only the latest emplacement. A general Late Jurassic age is accepted by many workers for the sea-floor spreading, therefore the ages of ophiolites are Late Jurassic. But this does not exclude the presence of older or younger ophiolites.

A. AGE OF SOME OPHIOLITES IN THE ALPINE SYSTEM

In the Northern Appenines a radiometric age corresponding to Lias-Dogger has been measured (BIGAZZI, FERRARA, and INNOCENTI, 1971). For the Alps radiometric ages between 140 and 170 million years are given by BERTRAND (1970) which correspond with Middle-Late Jurassic. In Yugoslavia several occurences of the Upper Jurassic on the ophiolites are reported. In the Othris Region, Eastern Central Greece, K-Ar isotopic ages show the highly differentiated sequence to be at least as old as Lower Jurassic (HYNES et al, 1972). In the Northern Anatolia, in the Pontic Range abundant palaeontological records of Jurassic and rarely Lower Cretaceous on top of the ophiolites are reported (BOCCALETTI et al 1966 a and b; BOCCALETTI and SAGRI, 1968). In the south, in the Western Taurus Range, BRUNN et al (1971) give a Triassic age for the commencement of basic igneous activity. This is based on doubtful stratigraphic and palaeogeographic evidence and contradicts the general Late Jurassic age of the main basic igneous activity in the Pindus Zone (northwest extension of the Western Taurus Range) as proposed by AUBOIN (1965). It also does not agree with the Middle (?) - Late Jurassic age of the basic igneous activity in the rest of Anatolia and with the extensive basic igneous activity in Europe during the same time (SANDER, 1970). In central Anatolia, near Konya, ophiolites are overlain by Jurassic cherty limestones (PASSERINI and SGUAZZONI, 1966). In the Eastern Taurus, RIGO DE RIGHI and CORTESINI (1964) reported Upper Jurassic limestones and cherts associated with the ophiolites.

B. AGE OF EMPLACEMENT

In the Northern Apennines ophiolite olistholiths have possibly been emplaced since the end of the Early Cretaceous (ABBETE et al, 1970). In the Alps, the age of emplacement of ophiolite nappes is believed to be the Late Cretaceous. In Yugoslavia and Greece the main overthrust phases probably occured at about the end of the Early Cretaceous and during the Late Eocene (BERNOUILLI and LAUBSCHER, 1972; HYNES et al, 1972). In Anatolia, in the Pontic Range, ophiolites constitute a huge chaotic complex built by a pile of chaotic nappes of olistrostromes with marly intervals bearing Post Turonian foraminifera. On top of this complex Maestrichtian limestones are deposited unconformably. Therefore, a Late Cretaceous age is deduced (BOCCALETTI et al, 1966a; BORTOLOTTI and SAGRI, 1968; SESTINI, 1971). In the south, in the Western Taurus, BRUNN et al (1970) and GRANCIASKY (1972) cited ophiolite nappes emplaced since the Late Cretaceous. In the Eastern Taurus, ophiolites occur as a large chaotic nappe emplaced in the end of Late Cretaceous (RIGO DE RIGHI and CORTESINI, 1964). For further discussion of the basic igneous activity and emplacement of ophiolites the reader is referred to the work of ABBETE et al (1973).

C. THE GEOLOGIC SETTING AND DISTRIBUTION OF OPHIOLITES

Two types of ophiolites can be distinguished (BRINKMANN, 1968; SESTINI, 1971; and OKTAY, 1973).

1. Mélanges: These are ophiolite masses composed mainly of serpentinite and diabase-basalt, associated with radiolarities, limestones, polymictic breccias, turbidites, and slump deposits. They show a chaotic and complicated internal structure. Mélanges were first reported from the Central Anatolia (BAILEY and McCALLIEN, 1950 and 1953). Later RIGO DE RIGHI and CORTESINI (1964) reported from the southeastern Anatolia on a much larger scale. More recently SESTINI (1971) described ophiolitic mélange from the North Central Anatolia, and he argues that ophiolitic series of this region is generally allochthonous and emplacement took place during the Late Cretaceous epoch. They generally show E-W trending zones. One of these zones is in the north, and the other one is in the south. (Figure II-1).

2. Ophiolite Masses: These are very large (up to 20 x 100 kms in size) masses and composed mainly of ultrabasic and basic igneous rocks. They are occasionally covered by basic lavas, radiolarites, and cherty limestones. Dykes are common. They are characteristic of the eastern and southern parts of Anatolia. They have generally tectonic boundaries, and their bases are never exposed. The origin of these rocks is controversial. Probably they represent the former oceanic floor preserved in thrust sheets (OKTAY, 1973).

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CHAPTER III

PART I

P E T R O G R A P H Y

In contrast to ancient limestones which are now nearly mono-mineralic, being composed of low-Mg calcite; recent limestones are composed of two distinct polymorphs of $CaCo_3$ namely calcite and aragonite. The calcite occurs in both low-Mg and high-Mg forms, which contain 0-5 percent and 11-19 percent magnesium in solid solution, respectively (CHAVE, 1954). Both aragonite and high-Mg calcite, the dominant modern carbonate minerals are metastable under normal conditions, and it is the processes of conversion to stable low-Mg calcite, together with organic influences, that are largely responsible for diversity of fabric types seen in ancient limestones.

The description of carbonate rocks best follows the major subdivision into grains, matrix (carbonate mud), and cement (sparry calcite) used for sandstones by KRYNINE (1948). This primary subdivision then allows the use of eigher at the currently most popular and largely objective carbonate rock classifications, that is, FOLK (1959 and 1962) or DUNHAM (1962) (Figures III-1 and 2). The Dunham classification is generally used throughout this study because of the genetic significance of mud-supported as opposed to grain-supported limestones and in order to avoid the necessary rigidity involved in applying Folk's particle type prefixes. Dunham's textural terms can be combined with names for grain-type classes and mineralogical classes (e.g. Oolite lime grainstone and mixed fossil dolomite wackestone) and they also can stand alone where mineralogy and grain-type are not at issue.

I. PRINCIPLES OF TEXTURAL CLASSIFICATION

Some principles are significant in categorizing texture in carbonate sediments:

A. PRESENCE OR VIRTUAL ABSENCE OF LIME MUD MATRIX

The origin of the limestone of the Gürün region involves the deposition of large quantities of carbonate mud over a long interval of geologic time (Late Jurassic to end of Late Cretaceous). Numerous studies of recent sediments have shown that carbonate mud can originate by a variety of

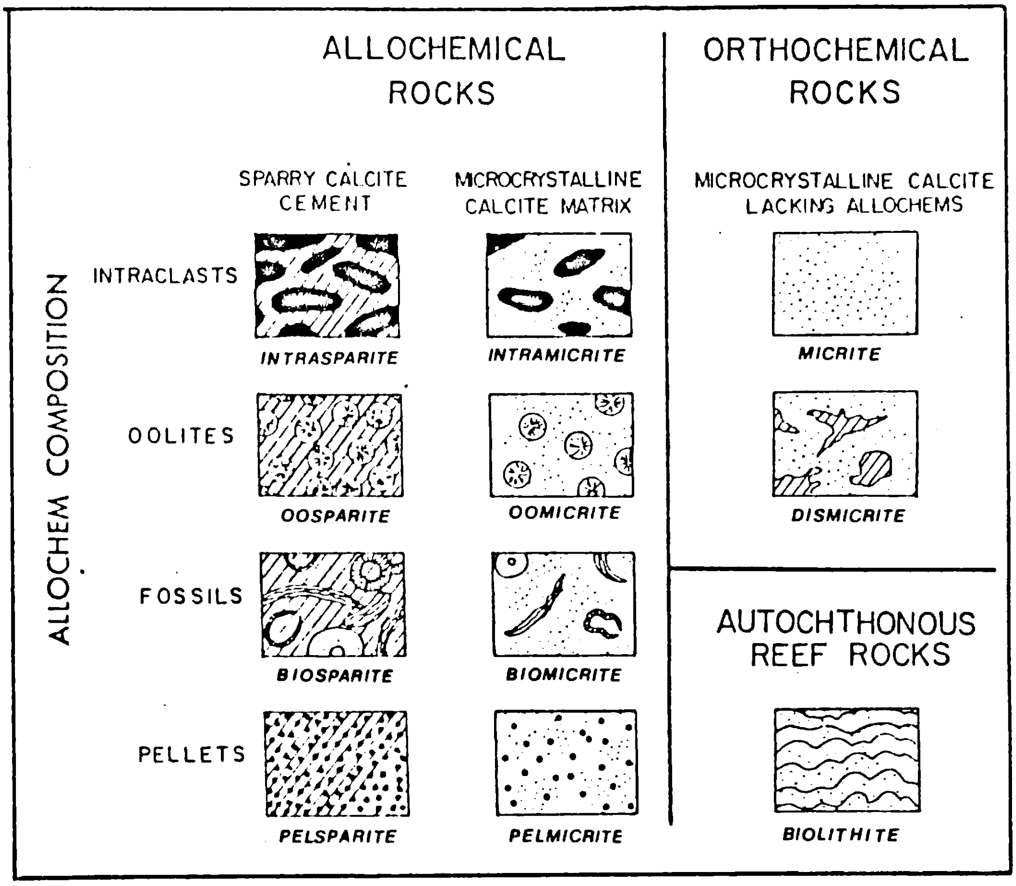

Fig.III-1: Graphic classification table of limestones (After FOLK, 1959)

<table>
<tr><td colspan="5">Depositional texture recognizable</td><td>Depositional texture not recognizable</td></tr>
<tr><td colspan="4">Original components not bound together during deposition</td><td rowspan="4">Original components were bound together during deposition... as shown by intergrown skeletal matter, lamination contrary to gravity, or sediment-floored cavities that are roofed over by organic or questionably organic matter and are too large to be interstices.</td><td rowspan="4">Subdivide according to classifications designed to bear on physical texture or diagenesis.</td></tr>
<tr><td colspan="3">Contains mud
(Particles of clay and fine silt size)</td><td>Lacks mud</td></tr>
<tr><td colspan="2">Mud supported</td><td colspan="2">Grain supported</td></tr>
<tr><td>Less than 10% grains</td><td>More than 10% grains</td><td></td><td></td></tr>
<tr><td>MUDSTONE</td><td>WACKESTONE</td><td>PACKSTONE</td><td>GRAINSTONE</td><td>BOUNDSTONE</td><td>CRYSTALLINE CARBONATE</td></tr>
</table>

Fig.III-2: Classification of carbonate rocks according to depositional texture (After DUNHAM, 1962) processes:

(a) Physicochemical: Many geologists believe that precipitation of most mud-size aragonite needles result from physicochemical processes (e.g. NEWELL and RIGBY, 1957; CLOUD, 1961).

(b) Biochemical: Some workers considered carbonate mud to be mainly biochemical in origin, the precipitation resulting from certain bacteria (LALOU, 1957). On the other hand PURDY (1963) points out that such precipitation is in fact a bacterially induced physiochemical processes.

(c) Mechanical abrasion of carbonate fragments: Abrasion by wave and current action may produce lime mud (HOSKIN, 1963; SWANCHATT, 1965; MATTHEWS, 1966). Such mud will then be trapped by baffle-like organisms or winnowed into adjacent low-energy environments (GINSBURG and LOWENSTAM, 1958).

(d) Biological breakdown: This is accomplished in two ways:

(1) The normal disintegration of carbonate skeletal particles because of the decay of their organic binders (LOWENSTAM, 1955; PURDY, 1963; STOCKMAN, GINBURG and SHINN, 1967). BATHURST (1971) pointed out that the breakdown of most skeletal types into their ultimate components would release crystals all under 4 microns.

(2) The breakdown caused by the boring activities or organisms mainly algae (BATHURST, 1966; KLEMENT and TOOMEY, 1967; MARGOLIS and REX, 1971).

(e) Accumulation of the calcareous tests and fragments of pelagic microorganism, especially nannoplankton: modern calcareous deep sea cores are composed in large part of coccoliths. Calcareous nannoplankton evolved during the Jurassic period and have been abundant ever since (see various volumes of Initial Reports of Deep Sea Drilling Projects). Many Jurassic and Cretaceous chalks and mudstones have been shown by electron microscope to consist principally of nanno fossils (FISHER, HONJO and GARRISON, 1967).

HONJO (1969) differentiated two types of micrites based on his electron microscope study on fine grained carbonate matrix. One of them is composed of a subhedral crystalline mosaic and called "orthomicrite". The origianl material of "orthomicrite" was probably inorganically precipitated aragonite mud of shallow water origin which later inverted to calcite (see "Lithification of carbonate mud" p.168). The second type represents concentration of nannofossils and is called "nannoagorite". It is abundant in the chalks and flysch in the Cretaceous and Palaeocene.

Much of the present available evidence from the Gürün carbonates suggest that many of the fine carbonate grains in the matrix were derived from mechanical and biological breakdown of carbonate particles, primarily skeletal fragments. The question is how much carbonate material was produced by breakdown outside of the original depositional environment and transported in; and how much was produced within the depositional environment itself. This question can not be satisfactorily answered based on the data of the present study, but, STOCKMAN, GINSBURG and SHINN (1967) found that the source of most, if not all, the recent lime mud in South Florida sediments was disintegration of fragile algae, especially genus Penicillus, in the wave and current agitated environment. According to these workers the lime mud can not accumulate in the wave and current agitated environments in which they are produced, and, therefore, are swept into the areas of less agitation. A slightly different situation was described by MATTHEWS (1966) from British Honduras. According to him carbonate fines are produced on the shoal and transported into the lagoon. By analogy with studies of areas of recent carbonate deposition, the present author believes that most, if not all, of the lime mud involved in the formation of the Gürün limestones were at least transported some distance to its depositional environment.

BOYER (1972) postulated that cementation in muddy environments might be inhibited by the presence of abundant organic matter. If this is correct, such lack of cementation could make fragments more susceptible to breakdown.

Normally in sediments with some lime mud, the grain/lime mud ratios are so variable, even in the same thin section, it is more important whether some mud exists than to know how much. DUNHAM's (1962) approach to the problem is significant: focus on currents of removal, not delivery.

" ... The distinction between sediment deposited in calm water and sediment deposited in agitated water is fundamental. Evidence bearing on this problem thus deserves to be incorporated in class names. This can be accomplished in several ways. One is to focus attention on average or predominant size, which erroneously assumes that all sizes in a sample are equally significant hydraulically. Another is to focus attention on the size, abundance, and condition of the coarse material brought to the site of deposition. This emphasis on what might be called current of delivery has long been successful in dealing with land-drived sediments, but does not work well in lime sediment because of the local origin of many coarse grains. A third way is to focus attention on the fine material that was able to remain at the site of deposition. This emphasis on what might be called currents of removal seems advisable if we wish to characterize carbonate sediment systematically in terms of hydraulic environment.

Inasmuch as calm water is characterized by mud being able to settle to the bottom and remain there, it seems that the muddy rocks deserve to be contrasted with mud-free rocks, regardless of the amount and size of included coarse material ... (DUNHAM, 1962; p. 111-113)".

B. THE MUD OR GRAIN SUPPORT

The ratio of grain to lime mud, the fundamental of many carbonate rock classifications are very variable. Therefore the packing-framework concept of DUNHAM (1962) is more useful and accurate. He subdivided carbonate rocks according to the basis of abundance of grains. In one of them grains are abundant enough to be predominant, but not so abundant that they support one another. They are called "mud-supported" and "grain-supported", respectively. A grain-supported rock is full of its particular mixture of grains, wheras a mud-supported rock is not. Looser packs would require mud-support. Naturally, the development of the grain framework depends not only on the number, but also on the shape of the grains. A grain-supported rock whose grains are shaped rather like a cornflake would contain a far smaller percentage of grains than would a grain-supported rock whose grains are spherical (see DUNHAM, 1962; p. 112, Pl. II). The difficulty arises because of the need to envision a three-dimensional arrangement of irregular shapes by looking at a two dimensional view. According to DUNHAM, experience gained in examining mud-free carbonates (i.e. grain-supported) is an aid in determining the kind of support in muddy carbonates. Floored interstices, shelter effects, embayed contacts, and overly close packing indicate grain-support.

The grains in grain-supported sediment carry the weight of the overburden and tends to protect it from compaction. On the other hand contact of grains with each other in the grain-supported sediment causes greater grain to grain solution penetration. Any part of the space between grains that is unoccupied by mud, is available for deposition of sparry calcite or, if open, for accumulation of hydrocarbons.

C. GRAIN TYPES

Four major grain categories have been recognised in the Gürün sediments.

I. Lithoclasts: Large particles either derived by desiccation, breakage or burrow disruption of penecontemporaneously deposited carbonate sediment ("Intraclast" of FOLK, 1959), or derived externally from older lithified rocks. Most lithoclasts are recognised by their grain boundaries which may cut constituent particles. Lithoclasts from older outcrops (subaerial or subaqueous) may contain particles which are compositionally different from those occuring in surrounding sediments. In contrast, younger lithoclasts can contain components similar to those in neighbouring sediments.

II. Pelletoids : Pelletoids are round, oblong or cylindirical carbonate grains that have a disoriented or crypotcrystalline granular texture. They are usually between 0.2 and 0.6 mm. in length with diameters between 0.1 and 0.4 mm. The origin of pelletoids has been the subject of some discussion. Because of the diversity of origin, the present author prefers MILLIMAN's (1974) term "pelletoid" since it is descriptive and embaces all grains that are constructed of an aggregate of cryptocrystalline carbonate, irrespective of origin. Many modern examples are clearly invertebrate and vertebrate faeces (see BATHURST, 1971; p. 85) but they can also be produced within stromatolotic algal mats (MONTY, 1967), and even by direct precipitation and cementation of high-Mg calcite lime mud (LAND and GOREAU, 1970). Accretion of mud sized carbonate particles by gentle agitation of bottom sediments has also been proposed as a means of producing pelletoids. HADDING (1958) suggested that clotting of the sediment following the appearance of local centres of crystallization associated with the bacterial decomposition of algae and concurrent deposition of more lime was another origin of pelletoids. WOLF (1965) identified most of the cryptocrystalline calcite grains in the Nubrigyn Reef Complex of New South Wales (Australia), as algal in origin. Although the algal lime mud pelletoid type may be abundant; as noted by the same author (Ibid.; p. 117) if even the slightest recrystallization occured, it is impossible to discriminate algal lime mud pelletoid from other types. Consequently, during the present work, all spherical or ellipsoidal grains, consisting of homogenous, crytocrystalline carbonate, have been called "pelletoid" and no attempt has been made here to determine their origin.

In some pelletoid rocks, the pelletoids tend to merge (BEALES 1965) and produce residual patches of sparry calcite amongst them. This may be responsible mechanism from formation of "grumeleuse structure" which will be discussed in some detail under the heading of "Cementation of Lime Mud" (p.168).

III. Coated grains: The author follows LEIGHTON and PENDEXTER (1962) who define "coated grains" as "grains having concentric or enclosing layers of calcium carbonate around a central nucleus". These coated grains fall into three principal categories, as follows:

A. Oölites: small spherical or subspherical accretionary grains generally less than 2.0mm. in diameter that in thin section display concentric and/or radial structure. A variety in this group is the "superficial oölite" in which the thickness of the accretionary coating is less than the radius of the nucleus.

B. Pisolites: Grains similar to but larger than oölites, and less regular in form; they are generally 2.0 mm. or more in diameter.

C. Algal-Coated grains: These are carbonate grains having a nucleus (generally a skeletal or rock fragment) about which algae have formed encrustations.

Pisolites are absent and the true oölites are rare in the Gürün limestones. The most abundant of the coated grains are the superficial oölites and so-called algal coated grains. These consist of a skeletal or non-skeletal nucleus, that is surrounded by a thin, slightly laminated cryptocrystalline coating. This thin coating is probably algal in origin although it may represent chemical precipitation or mud accretion.

IV. Skeletal grains or fossils: From random thin section study alone, it is not possible to identify in detail many microfossils or fragments of large shells, and the necessary further study of specimens obtained from acid-etched samples or from oriented thin sections has not been carried out. However, the amount of material identified in a general way is so large that clear associations of certain organisms can be seen. These associations combined with petrographic information, offer a good basis for palaeoecological interpretation.

The following descriptions of fossils from Gürün sediments are not basically concerned with taxonomic problems; they are to serve as a basis for appraisal of the environmental significance of the organisms.

ALGAE: Algae can assume a wide range of shapes and structures but with the exception of calcareous algae leave little or no record of their presence in the rock. Thus, the exact role played by algae in the Gürün sediments is not known. But it is the present author's belief that algae were very important, especially in such roles as the trapping and cementation of sediments, and the corrosion and alteration of carbonate particles. Much of the algal material is hard to recognize as it grades into, and has been altered to cryptocrystalline calcite by grain-diminution (WOLF, 1965). In this account discussion is restricted to calcareous algae only.

Laminated coatings and encrustations are commonly observed enclosing grains. Many of these coatings (the oncolites and algal coated grains) are interpreted as having been formed by algae. Though usually structureless, the mat-like layers occasionally show fine filaments or tubules suggesting an algal origin. Besides coating the grains, these algae corrode and alter them as well.

Laminated limestones, common in certain back reef areas, are also interpreted as being formed by algal-mats. These sediments closely resemble the present-day intertidal and supratidal algal mats found in the Persian Gulf, the Bahamas, and the Florida Keys. Non-skeletal green or blue-green algae that were able to trap and bind the sediment to form irregular layers are believed to be responsible for most of the laminated limestones found.

According to their living colour, calcareous algae are subdivided into the divisions of red, green, and blue-green algae (Table: III-1). An excellent up-to-date synthesis of the subject is given by FLÜGEL (1977) and WRAY (1977).

PHYLUM	*FAMILY*
RHODOPHYTA (Red Algae)	Corallinaceae Solenoporaceae Gymnocodiaceae Squamariaceae
CHLOROPHYTA (Green Algae)	Codiaceae Dasycladaceae Characeae
CYANOPHYTA (Blue-Green Algae)	Porostromata Spongiostromata

Table III-1: Divisions of Calcareous Algae
(After HOROWITZ and POTTER "1972"; FLUGEL "1977" and WRAY "1977")

Figure III-3 shows distribution of calcareous algae along an idealized profile of carbonate shelf margin.

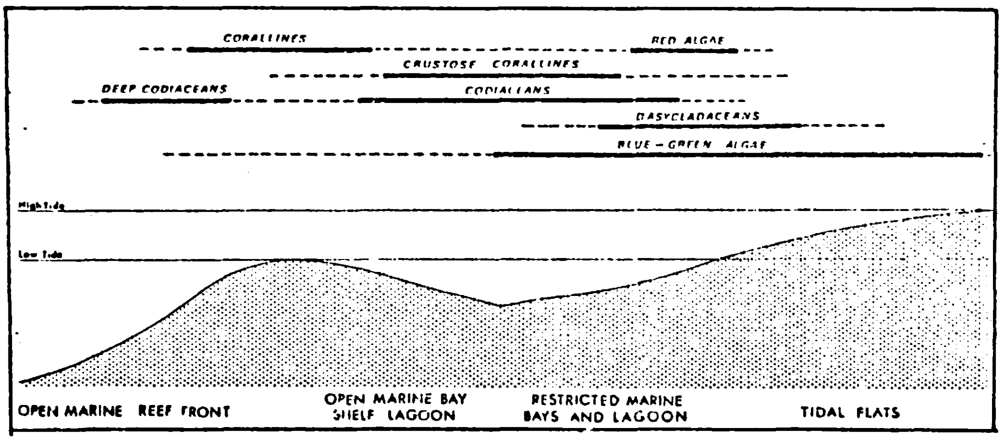

Fig.III-3: Distribution of calcareous algae along an idealized profile of carbonate shelf margin (Simplified from WILSON, 1975)

RED ALGAE: They are the most abundant phyllum found in the Gürün carbonates. Solenoporoid and especially Coralline Algae are common. The latter family can be subdivided into two subfamilies: Articulated Corallines and Crustose Corallines. The articulated coralline algae builds only branched forms which frequently break at their joints so that fossils usually are represented by individual segments of an originally much larger plant. Crustose coralline algae form branches and also crusts and range in thickness from a single layer of cells to many hundred or thousands of cell layers (Pl.45B). The red algae are characterized in thin section by a tiny fabric of cells, 10-30 microns in diameter, which are arranged in lines. Coralline algae contain more pillars than the Solenoporoid Algae (JOHNSON, 1951).

GREEN ALGAE: Codiacean and Dasycladacean Algae have been encountered in some thin sections. Codiacean algae have a central region of branched and interwoven filaments and a peripheral layer of lateral branches. Dasycladaceans are represented as cylindrical segments of branches or stems, which are hollow in the centre and are perforated laterally at right angles to the main stem by numerous pores representing primary or secondary branches.

BLUE-GREEN ALGAE: Porostomata family has simple or branched, unsegmented filaments. Both non-skeletal and skeletal blue-green algae have been active in the deposition of carbonate sediments, although, non-skeletal species have been more important because of their role in the formation of stromatolites and other algal-laminated sediments.

The term "STROMATOLITE" is applied to forms with pronounced vertical relief, whereas forms with flat-lying laminae are generally called "ALGAL-LAMINATED SEDIMENTS" (WRAY, 1977). Some examples are shown on plates 4 to 7. They are biosedimentary in origin; resulting from the interaction of non-skeletal blue-green algae and physical sedimentary processes. The algae have served a sediment stablizing role by mechanically trapping and binding sediment particles on organic films. The term "stromatolite" as used throughout the test refers to "algal stromatolites" definded above. The broad scope of the subject is emphasized by the nearly 800 page volume "Stromatolites" edited by WALTER (1976). Study of recent stromatolites and algal-laminated sediments have shown that the majority were formed in the intertidal or supratidal zone. However GEBELEIN (1969) describes"stromatolites" in the subtidal zone in Bermuda, down to a water depth of 10 metres. A few ancient stromatolites may well have grown many metres below sea-level on the foreslop of carbonate platforms (see examples of PLAYFORD and COCKBAIN, 1969; from the Devonian of the Western Australia and HOFFMAN (1974) from the Lower Proterozoic of

Northwestern Canada). DAVIES (1970) states that flat algal-laminated sediments are the characteristic sediment type in the intertidal zone of sheltered hypersaline embayments in the Shark Bay, Australia.

Several classifications of stromatolites based on geometric and ecologic forms have been published. An excellent summary of stromatolite classification is given by HOFFMAN (1969) and in various papers in WALTER (1976). LOGAN, REZAK, and GINSBURG (1964) proposed a classification based strictly upon morphologic form. Three main growth types were recognized:

1. Laterally linked hemispheroids which contain horizontally continuous layers;

2. Vertically stacked hemispheroids in which the laminae between domes are not connected;

3. Spheroidal structures which are equivalent to oncolites.

Many of the differences between these various morphologic forms can be explained by the environment of accretion. LOGAN, REZAK, and GINSBURG (1964) and DAVIES (1970) found that laterally linked hemispheroids are most common in supratidal and intertidal environments. Vertically stacked hemispheroids are restricted mostly to protected arid areas with considerable tidal range in which sediment preferentially accumulates on the domes.

These findings leads to a generalization supported by HOFFMAN, LOGAN, and GEBELEIN (1971) in recent stromatolites:

" ... Where wave action is strong, discrete columns of up to 1 metre relief occur. Where wave action is moderate, small branching columns and laterally linked domes with less than 15 cm. relief occurs. In quiet water stromatolites are stratiform ... (p.42)".

Studies of recent algae in the Shark Bay, Australia (LOGAN, 1964, DAVIES, 1970) and in the Persian Gulf (KENDALL and SKIPWITH, 1968) have shown that many stromatolites or algal mats are being formed under highly saline conditions. Conditions for preservation of stromatolites formed within environment are favourable. GARRET (1970) stressed the importance of the absence or restriction of burrowing animals in the saline environment. He noted that the burrowing animals are common in the tidal flat and in shallow-marine waters of the Bahama Banks, where they destroy many algal mats or sedimentary lamin ations. Stromatolites then may appear to be restricted to a particular environment because its conditions favoured their preservation, and not because the algae that formed stromatolites flourished only in that particular environment.

ONCOLITES are calcareous accretionary bodies of algal origin. They are spheroidal stromatolites characterized by concentric laminations.

Again, blue-green algae play an important role in the formation of oncolites. They are unattached and can roll about on the bottom. According to LOGAN, REZAK, and GINSBURG (1964) marine oncolites are typical to the deeper parts of the intertidal environments with moderate water agitation and to the area immediately below the low tide.

FORAMINIFERA: These protozoans are almost the most widely distributed fossil group found in the Gürün sediments. Both planktonic and benthonic foraminifera have been encountered in the present study. The planktonic ones (e.g. Globigerina, Globotruncana, Orbulina, Gümbelina) characterize the Maestrichtian calciturbidites. The diameter of these foraminifera rarely exceeds 0.5 mm. A list of most common types based on previous works, has been given in Table I-1B of which most of them have been encountered in the present work too (p. 9). The benthonic foraminifera are widespread and have been seen in many thin sections. Table I-1A shows the common ones noted by many workers. The most abundant group is the Miliolidae and they form dominant grain type in some samples (Pl.10E,11C,D. Also see Section 8A).

TINTINNINES (CALPIONELLA): These are planktonic protozoans with vase-shape shells. Most forms are about 80-150 microns in diameter, and usually characterize the Upper Jurassic and Cretaceous of the Tethys. Some of the circular calcitic bodies seen in the calciturbidite thin sections are thought to be tintinnines.

SPONGE SPICULES; The spicules are either calcareous or siliceous. It is believed that solution of the siliceous ones had played an important role as the source for the formation of chert associated with calciturbidites. Calcitic sponge spicules have been found in some calciturbidite beds.

COELENTERATES: Most coelenterates are dominantly colonial and can build large bioherms. The debris produced by degradation of these colonies usually is confined to the reef environments; limestones composed of coelenterate fragments, are rare outside reef complexes (FÜCHTBAUER, 1974). Two classes, namely Hydrozoa and Anthozoa are important as the rock forming concerned.

Hydrozoans are mostly characterised by stromatoporoids. They are nodular, branching or encrusting colonies. They are constructed of delicate laminae which are arranged parallel to the surface of the colony and are supported by small columns, called pillars, forming rectangular cells with characteristic diameters between 0.05 and 0.5 mm. Systamatic position of "stromatoporoids" is the subject of argument among various workers, but they are generally assigned to the hydrozoan coelentarates. The stromatoporoids occur with normal marine faunas and are also found in the lime wackestones (brachiopod, echinoderm, and gastropods), and with corals and coralline algae. The stromatoporoids are generally poorly preserved in the sediments investigated. They are much more susceptible to calcite recrystallization than are associated red algae or brachiopods. This is probably due to the original skeleton being aragonite rather than calcite.

Anthozoans are represented by CORALS. The ones found in the Gürün sediments belong to Scleractinia (Hexacorals) order of Zoantharia Group. All Hexacorals precipitate aragonite skeletons; therefore recrystallization to calcite, or solution and deposition as calcite have occurred in all specimens. Very often no crystal structure remains and it is difficult identifying corals in thin sections, although fragments of corals are present in a few thin sections, most corals occur as individuals or colonies too large to be studied in thin section (Pl.8)

BRYOZOANS; Bryozoans are rare and have been found mostly in the reef detritus. At one locality, No. 1012, Keslikdere Section, tremendous numbers of bry zoans have been found (Pl.38A). This seems to be an unique locality in the investigated area; no other exposure contains such a concentration of bryozoans. The bryozoans are preserved mostly as small fragments representing fragile fronds or branches. Even small shreds of bryozoans can be easily recognised in thin section by their fibrous structure and brownish colour.

BRACHIOPODS: Shells have a double-layered structure in which the thick inner layer exhibit distinct fibers nearly parallel to the elongation. Brachiopods appear to have been able to adopt themselves to a wide variety of environments. Well-aerated water of normal salinity, however, seems to have been necessary since brachiopods are rare in the back-reef area. The fore-reef and, to a lesser extent, the reef areas with moderately turbulent wave and current action were the most favourable environments for most of the brachiopods. The brachiopod-bearing lime wackestones are lacking in any lamination. This is probably due to environmental conditions favouring burrowers within the mud bottom. The burrowing population could also account for the small size of brachiopod fragments and the rarity of whole shells. Actually, the brachiopods may have been rather sparse and scattered, as were the echinoderms which lived with them.

MOLLUSKS; Mollusks also are represented in the Gürün sediments mostly by fragments, although some small gastropods are preserved nearly intact. Most fragments can not be identified as biologic class from thin section studies, but essentially all mollusks seen, are presumed to be pelecypods and gastropods by analogy with the best preserved specimens seen on the outcrop. Cephalopods and scaphopods are almost absent. The internal structure of mollusk shell is not preserved, a fact attributable to inversion of aragonite in the original shell to calcite (STEHLI, 1956).

GASTROPODS are common, not abundant, in the Gürün sediments. They are frequently recognised as fragmental internal molds, demonstrating that lithoclasts in limestone may commonly be formed in this way after the original aragonitic shell rots from around the semi-hardened lime mud filling. It is impossible to identify generically any of the gastropods. Both high and low-spired (generally very small) forms are present. Low-spired gastropods are common in the lime mudstones whereas high-spired types are assoiciated with coarse lime grainstones of the reef and fore-reef facies. Some of them have been recorded in some of the samples classified as "restricted marine". This is not surprising since gastropods are salinity tolerate. Tidal flat environments of recent carbonate provinces commonly contain lime muds and grainstone which contain no other organisms except gastropods and ostracods. Such associations are also seen in the sediments investigated. Gastropods are easily recognized in thin sections by the succession of whorls around a central axis and by recrystallization of the shell to a blocky mosaic of calcite.

PELECYPODS: In the Mesozoic, Pelecypods of the Pachydont Group (Rudists) become well adapted to life in more or less restricted environments and to both sheltered and rough water (WILSON, 1975). In the Cretaceous rudists evolved into bizarre forms in which one valve rested on the substrate or became attached while the other served as a lid or cap; the valves usually articulated with massive teeth and sockets. When the rudists grew in crowded conditions, the prevalent form was an erect, tall, twisted, slender shell, oriented during growth towards favourable food-bringing currents. Rougher water caused stubby, trunk-like forms to develop. The rudists commonly display a "cellular-prismatic" microstructure characterized by striking rectangular or radial patterns (Pl. 13D,E). The important rock-building caprinid group has a very thick wall containing open canals which occupied from 30 to 75 percent of shell volume (WILSON, 1975). These porous walls, together with a large central cavity, gave the caprinid rudists great original pore space. They grew in abundance and formed moderately large mounds in shallow back reef positions. May specimens appear in gowth position in these mounds (Pls. 8 and 9), rooted in the ever-present mud matrix which accumulated in and around the rudist patches. Many other individuals are overturned and lie in irregular positions in these patches (Pl. 8D). As noted by WILSON (1975) caprinid rudist patches, with some relief, are known to have also formed at shell margins and in downslope positions as well as in back reef areas. Generally, this is the case in the GUrUn area. The shelf margin patches generally have a micritic matrix and are surrounded by lime grainstone flank beds whose particles are exclusively rudist fragments derived from the mound inhabitants. A list of rudists collected by the present author and determined by N. KARACABEY and C. ÖZTEMÜR is given in Table III-2.

Yvaniella cf.maestrichtiensis MILOVANOVIC	Microcaprina sp.(aff.Bulgarica TZANKOV)
Colveraia variabilis KLINGHARDT	Vaccinites ultimus MILOVANOVIC
Sabinia aff.Klinghardti BOHM	Joufia cappodocinsis COX
Radiolites aff.angeiodes PICOT DE LAPEIROUSE	Hippuritella aff.colliciata WOODWARD

Table III-2 : List of rudists collected in the present account (Determinations (Mineral Reseach and Exploration Institute).

OSTRACODS; They have small (0.3 to 0.7 mm. long) convex bivalved tests. Though found in most rock types, they are most abundant in the lime mudstone and lime wackestone of both normal marine and the restricted carbonate facies types. An additional distinctive rock type is the exclusively ostracoda pelletoid grainstone found in the restricted part of the facies belt. This rock type is known in carbonate rocks of various geologic periods. Possibly the associated pelletoids represent ostracod faeces.

ECHINODERMS: Echinoderms are represented by disarticulated fragments but are easily identified because they consist of single large calcite crystals penetrated by a well-preserved internal network. Nearly all the echinoderm fragments consists of crinoid columnals, but spines and plates of crinoid and echinoids are also present. Most echinoid and crinoid fragments could not be differentiated. They are mostly associated with normal marine lime wackestones are rarely found in restricted marine rock types. They are salinity-sensitive organisms. Some echinoderm specimens have been collected from the vicinity of Kızılören (L46) during the present work and identified by Y. SEZGİNMAN:

Hemiaster cf. prunella LAMARCK
Echinocorys cotteaui LAMBERT

On the other hand AKKUŞ (1963) notes the following species from the same locality:

Hemiaster aff. sexaagulatus D'ORB.
Ovulaster aff. Oblusus COTTREAU-BLAYAC
" " suberti GAUTHIER

BIOCLAST: A fragmental particle derived from beakdown of any sort of calcareous shell, test, or skeleton, regardless of whether the breakdown was mechanical or caused by organic agents is called "Bioclast".

This is the definition of WILSON (1975). Some authors have restricted the term to skeletal debris resulting only from organic processes.

D. GRAIN SIZE AND ROUNDING

Grain size has a secondary importance in modern textural classifications. PETTIJOHN (1975) summarizes the development of ideas on grain size classification in clastic sediments and concludes that the WENTWORTH SCALE is at present used by most scientists. FOLK (1959) working on carbonates, used the terms of GRABAU (1904*)but retained the finer division of Wentworth Scale except in the calcirudite range. He put the boundary between "rudite" and "arenite" to 1.0mm. This modification is not based on any particular fundamental. He notes that "... this appears to be more meaningful boundary ... (Ibid., p.16". This boundary causes problems in oolitic limestones which vary in grain size, in the same sample, above and below 1.0 mm. This type of rock would be almost unclassifiable if the rudite and arenite boundary is at 1.0 mm. However, the rudite-arenite boundary of the Wentworth Scale not only makes classifaction easier, but also coincides with the largest of the oolite and the oolite-pisolite boundary; and has some genetic significance. Therefore, in the present study 2.0 mm mark has been accepted as a boundary between rudite and arenite which

64 mm	CALCIRUDITE	VERY COARSE
32		COARSE
16		MEDIUM
8		FINE
4		VERY FINE
2	CALCARENITE	VERY COARSE
1		COARSE
0.5		MEDIUM
0.25		FINE
0.125		VERY FINE
0.062	CALCILUTITE	COARSE
0.031		MEDIUM
0.016		FINE
0.008		VERY FINE
0.004mm		

(*) Early carbonate classicications were primarily based on variations of grain size. GRABAU introduced three major divisions; i.e. calcilutite, calcarenite, and calcirudite. Calcilutites were described as "very fine grained limestones formed of lime mud". Most of these rocks are aphanitic and no original grain structure can be observed even under the microscope. Calcarenites ate described as "limestones composed of small, sand-like calcareous fragments" while calcirudites are reported as "limestone breccia or conglomerates composed of calcareous fragments". This terminology found considerable favour among geologists and is still widely used. KAY (1951) supplemented this classification by introducing the term "calcisiltite" to cover limestones of silt grade, however these sizes are very difficult to resolve even with a petrographic microscope, and the term "calcilutite" is used in this account to include all sizes under the sand grade (< 0.0625 mm) (See Table III-3).

Table III-3: Grain size scale for carbonate rocks (Modified from FOLK½ 1959) originally had been defined by GRABAU (1904) in that way (Table III-3). In determining the grain size name of a carbonate rock only the size of grains is considered; crystal size of microcrystalline carbonate or sparry calcite is ignored (FOLK, 1959; p. 16). The grain size term can be combined with textural terms; such as "coarse-calcarenite lime pack stone (for a grain-supported rock composed of coarse-sand-size grains with some mud)". Grain shape is a useful parameter to describe, provided one recognizes grain type. For example, grains from gastropods and crinoid stems are formed round and not subject to the same consideration as originally angular mollusk shell fragments which have been rounded and coated in agitated water (WILSON, 1975).

E. BIOGENIC PRECIPITATED CARBONATE MASSES

These form an additional textural category treated in all classifications which original components were bound together during deposition. Visible construction or organic framework, stromatolitic la mination contrary to gravity; and the presence of roofed over, sediment floored cavities are interpreted as being signs of binding.

In the last two decades two systems of classification have drawn considerable attention and warrant further comments:

The carbonate classification proposed by FOLK 9195) essentially is based on three end-members; Terrigenous constituents, Allochemical components, and orthochemical constituents. Terrigenous constituent include all materials derived from an adjacent land mass and carried into the depositional basin, such as quartz and chert grains, feldspar and argillaceous material. Allochemical components or "allochems", are carbonate bodies which formed by chemical or biologic processes within the depositional basin. There are, according to FOLK's terminology, four types of allochems; i.e. intraclasts, oölites, fossils and pellets (Figure III-1). Orthochemical constituents, or "orthochems" include all normal precipitates formed within the depositional basin or within the rock itself and show little or no evidence or transport. Microcrystalline calcite ooze matrix, or "micrite", and coarser, clearer sparry calcite are assigned to the orthochems. A third group of orthochems includes minerals formed by recrystallization and dolomitization processes, such as replacement dolomite and recrystallized calcite. Varying proportions of the three end members, as well as the average size of the allochems, determine the name applicable to a particular carbonate. FOLK's classification pays particular attention to the nature of the groundmass. The amount of winnowing, to which the sediment is subjected, depends on the strength of the currents acting in the particular locale. Strong currents result in removing the micrite groundmass and porosity development or subsequent sparry calcite infill will occur. Carbonate deposited in quiet waters contain a large amount of micrite.

The classification by DUNHAM (1962) is based primarily on the depositional fabric of the carbonates, with the subdivisions divided on the degree and nature of the support of constituent particles. Two end-members are indicated, i.e. grain-supported and mud supported carbonates. Grain-supported fabrics are those in which the particles are in contact, therefore offering a supporting framework, with the interstitial spaces occupied by carbonate mud or sparry calcite. Mud-supported fabrics are characterized by floating particles in the carbonate mud. Presence or absence of carbonate mud permits a differentiation between "muddy carbonates" and "grainstone". A further subdivision, based on the abundance of grains and particles, distinguishes between mudstone, wackestone, and packstone. The term "boundstone" is used to describe those rocks which owe their origin to some organic binding mechanism. Figure III-2 presents an outline of DUNHAM's classification.

Definitions of the terms, which forms the fundamental of DUNHAM's classification, with the equivalents in FOLK's classiciation are given below:

<u>MUDSTONE</u>; Muddy carbonate rocks with practically no fossil debris or non-skeletal grains. The term is synonymous with "micrite" and "fossiliferous micrite" of FOLK's terminology. It implies a calm water environment.

<u>WACKESTONE</u>: Carbonate rocks composed of particles (usually unsorted and jumbled) with significant mud matrix to separate and support the grains. For example, FOLK's "biopelmicrite" is a "wackestone" in this classification.

<u>PACKESTONE</u>: Grain-supported muddy carbonate rocks. "Packed biomicrite" of FOLK's classification is an example to this group of rocks.

<u>GRAINSTONE</u>: A carbonate rock free of mud in which the particles from a self-supporting framework and are sorted. This class form the "sparite" group of FOLK's classification.

<u>BOUNDSTONE</u>: Carbonate rock built up by in situ growth in skeletal frame-producing organisms or by lacy intergrowth of skeletons to produce a sediment-trapping mechanism. A "boundstone" can be classified as a "biolithite" in FOLK's spectrum.

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PART 2

MICROFACIES TYPES

I. CONCEPT OF STANDARD MICROFACIES TYPES

The geologic literature is full of different microfacies nomenclature and it is far away form being standard. It is the belief of the present author that a microfacies description and interpretation of a carbonate complex is not of much use unless there is a standard on which it is based. Therefore, with this thought in mind, in the present account 24 Standard Microfacies Types, proposed by WILSON (1975) have been employed. The grouping of microfacies into a limited number of categories is clearly an oversimplication as the range of sedimentological parameters which control deposition in the marine environment is concerned. They also contain some overlapping and inconsistencies and omit many variations. But, as noted by WILSON, " ... the grouping is applicable to enough known geological facies complexes to demonstrate its general accuracy and to show the usefulness of reduction to a limited number of types ... (Ibid., p.64) and his standard microfacies types have been found applicable to the Gürün carbonates. He described twenty-four standard microfacies types which are summarized below. Table III-4 shows these twenty-four types and their position in the "Standard Facies Belts" of a generalized model which will be discussed in detail in the chapter on "Depositional Environments" (p. 71).

SMF-1: Spiculite. Dark, organic rich, and argillaceous lime mudstone. Spicules are generally oriented and replaced by calcite.

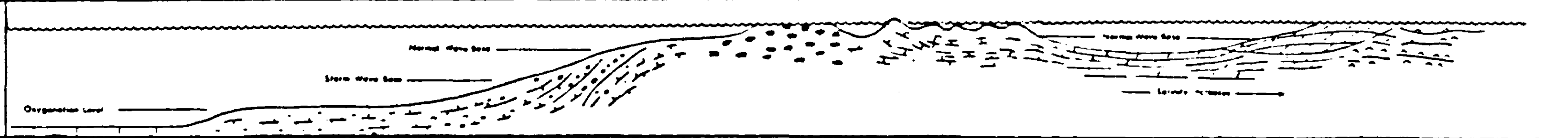
SMF-2: Microbioclastic packstone/grainstone. Most of the grains are in calcilutite range ($< .62\mu$). Fine ripple cross lamination is common.

SMF-3: Pelagic lime mudstone. Lime mud matrix contain scattered pelagic microfissils.

SMF-4; Bioclastic-lithoclastic packstone (microbreccia). This rock type includes both fine talus and coarser debris reulting from turbidites. Quartz and cert grains may be present. The term "allodapic limestone" of MEISCHNER (1965) encompasses this microfacies.

SMF-5: Bioclastic packstone-grainstone. Clasts may be of gravel size ($>$2,0mm). This is a common reef flank facies. Geopetal structures and umbrella effects from unfiltered finer sediment are common.

SMF-6: Coarse gravel grainstone. It is usually encountered within organic buildups formed in zones of high wave energy.

FACIES BELT	BASIN FACIES	OPEN SHELF FACIES	TOE of SLOPE CARBONATE FACIES	FORESLOPE FACIES of CARBONATE PLATFORM	ORGANIC BUILD UP of CARBONATE PLATFORM	PLATFORM EDGE CARBONATE SAND BANKS	OPEN MARINE PLATFORM FACIES	FACIES of RESTRICTED CIRCULATION on MARINE PLATFORM and TIDAL FLATS	SUPRATIDAL and INLAND PONDS FACIES
CROSS SECTION									
LITHOLOGY	Chiefly pelagic; thin calcareous and siliceous sediments; rich in ferric iron	Very fossiliferous lime stone interbedded with marls and chert	Fine-grained limestone; in some places cherty	Variable depending on water energy upslope; sedimentary breccia and lime sand or lime mud-stone	Massive limestone in places consisting solely of organic skeletons	Calcareous lime sand	Variable carbonates	Generally lime muddy sediment with much dolomite	Very fine grain carbonate sediment; algal-mats
COLOUR	Dark brown to black	Gray, green, red	Gray to light gray	Gray to light gray	Light gray	Light gray	Light gray	Light gray to gray	Gray, yellow, red
BEDDING	Very even millimetre lamination	Thin to medium, wavy bedding	Minor lamination, massive beds common	Thin to medium	Massive to thick	Thin to thic bedded	Thin to massive, well bedded	Well bedded, well laminated	Irregular lamination
SEDIMENTARY STRUCTURE	Ripple cross lamination	Burrows, bedding surfaces show diastems	Lenses of graded sediment; lithoclasts and exotic blocks	Slumps in soft sediments, foreset beading, exotic blocks	Roofed cavities; lamination contrary to gravity	Cross bedding; festoons	Abundant burrows	Birdseye structure, algal mats on or near an intertidal zone; stromatolites in very shallow water, graded bedding; cross-bedded sand in channels	Algal-mats
STANDART MICROFACIES TYPE	1) Spiculite; 2) Micro-bioclastic packstone/grainstone; 3) Pelagic lime mudstone	2) Microbioclastic pack-stone/grainstone; 8) Whole fossil wackestone; 9) Bioclastic wackestone; 10) Coated and worn bioclasts in lime mud matrix	2) Microbioclastic pack-stone/grainstone; 3) Pelagic lime mudstone; 4) Bioclastic-lithoclastic packstone	4) Bioclastic-lithoclastic packstone; 5) Bioclastic packstone/grainstone; 6) Coarse gravel grainstone	7) Boundstone; 11) Coated bioclasts in sparry calcite cement; 12) Coquina, bioclastic packstone/grainstone	11) Coated bioclasts in sparry calcite cement; 12) Coquina, bioclastic packstone/grainstone; 13) Onkoidal grainstone; 14) Lag grainstone; 15) Oolite grainstone	8) Whole fossil wacke-stone; 9) Bioclastic wackestone; 10) Coated and worn bioclasts in lime mud matrix; 16) Pelletoidal grainstone; 17) Aggregated lump, pelletoidal grainstone; 18) Foraminifera or dasycladaceae grain-stone	16) Pelletoidal grain-stone; 17) Aggregated lump, pelletoidal grain-stone; 18) Foraminifera or dasycladaceae grain-stone; 19) Pelletoidal lime mudstone-wacke-stone; 20) Stromatolite mudstone; 21) Algal-mat lime mudstone; 22) Onkoidal wacke-stone; 23) Unlaminated homogenous lime mud-stone	23) Unlaminated homo-genous lime mudstone; 24) Lithoclas-bioclast packstone/grainstone

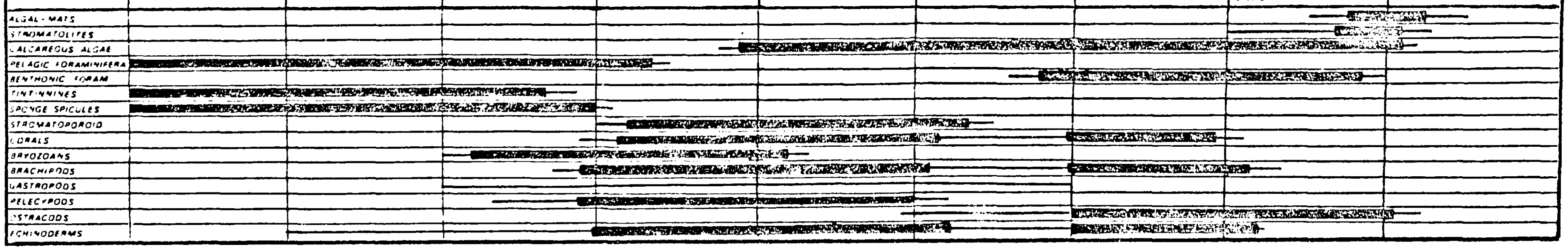


Table III-4: Idealized sequence of Standard Facies Belts and Standard Microfacies associated with each belt (Mainly after WILSON, 1970 and 1975)

SMF-7: Boundstone. In situ organic growth.

SMF-8: Whole fossil wackestone. This type of rock commonly forms below normal wave base, in quiet water environment. Sessile organisms are rooted in lime mud.

SMF-9: Bioclastic wackestone. Grains are fragments of diverse organisms jumbled and homogenized through burrowing and may be micritized. It is formed in shallow neritic water of open circulation at or just below wave base.

SMF-10: Coated and worn bioclasts in lime mud matrix (wackestone/ packstone). Dominant particles are of high energy environments and have moved down local slopes to be deposited in quiet water.

SMF-11: Coated bioclasts in sparry calcite cement (grainstone). This sediment is formed in areas of constant wave action, at or above wave base so that lime mud is removed. Bioclasts may be micritized.

SMF-12: Coquina, bioclastic packstone/grainstone. Some grains may be bigger than 2.0 mm. This sediment characterize an environment of constant wave or current action where mud is removed by winnowing. Some concentrations of special types of organic debris are common. Dasycla dacean grainstones accumulate in very shallow water.

SMF-13: Onkoidal grainstone. Algally coated grains are formed in a moderately high energy, very shallow water environment.

SMF-14: Lag grainstone. Coated and worn particles, in places mixed with oölites and pelletoids which are blackened and iron stained. They are thin deposits and represent slow accumulation of coarse material in zone of winnowing.

SMF-15: Oölite grainstone. It is well-sorted and usually over-packed. Best formed ones are produced on tidal bars.

SMF-16: Pelletoidal grainstone. The pelletoids are derived from organic pelleting of mud and in places are admixed with concentrated ostracod tests and foraminifera. They indicate very slight water movement. Such sediments may grade into pelletoidal wackestone (Type 19) and is common on tidal flats and natural levees. Thick traded laminae and birdseye fabric are common.

SMF-17: Aggregated lump, pelletoidal grainstone. This is a mixed facies of isolated pelletoids, agglutinated pelletoids, some coated particles and lumps which are in part small lithoclasts. This type of facies characterize very warm, shallow water environment with only moderate circulation.

SMF-18: Foraminifera or dasycladacea grainstone. This facies commonly occur as concentrations of tests with pelletoids in tidal bars and channels of lagoons.

SMF-19: Pelletoidal lime mudstone-wackestone. Laminated to bioturbated. Occasionally grades into pelletoidal grainstone of Type 16. An ostracod-pelletoid assamblage is common. Lime mudstone with scatterd foraminifera, gastropods and algae also may occur. The sediment is deposited in very restricted bays and ponds.

SMF-20: Stromatolite mudstone. It is common in the intertidal zone. Fine lime mud is preferentially trapped on the highest areas resulting in a lamination contrary to gravity.

SMF-21: Algal-mat lime mudstone. Characterize tidal ponds.

SMF-22: Oncoidal wackestone. They are typical of shallow water back reef environment, found commonly on edges of ponds or channels.

SMF-23: Unlaminated homogenous lime mudstone. It is developed in saline or tidal ponds.

SMF-24: Lithoclast, bioclast packstone/grainstone. Clasts are generally of unfossilliferous lime mudstone and may be bigger than 2.0mm. Matrix is rare. The sediment is normally termed "intraformational limestone pebble conglomerate" and formed as a lag deposit in tidal channels.

II. CHARACTERISTIC LITHOFACIES

The following rock types are considered to be basic members of the sediments investigated and are plotted on the petrographic logs. Table IV-1 gives a summary of facies, microfacies types and inferred environments (p.113).

A. PURE LIME MUDSTONE (Standard Microfacies 23): This rock type is usually associated with algal-mat lime mudstone. It is a dark gray dense lime mudstone. No visible textural differences exist. The total rock is cryptocrystalline calcite and is essentially without fossils or pelletoids although a few scattered remains of the restricted fauna or vague pelletoids are included in some samples. The interpreted depositional environment for this sediment is very quiet shallow ponds of water with salinities above normal.

B. ALGAL-MAT LIME MUDSTONE (Standard Microfacies 21): Light grey micritic and micritic-skeletal limestone containing zones of dense, laminated, algal-mat fabric. (Pls.4-7) Irregular laminae, which characterize this lithology, represent poorly preserved algal growth which cemented and bounded the fine-grained calcium carbonate mud. It is interpreted as a deposit which probably was formed in the supratiral zone.

C. LITHOCLASTIC PACKSTONE OR GRAINSTONE: This lithology is a variation of Standard Microfacies 24. The clasts are generally of pure lime mudstone (Lithology A) and algal-mat lime mudstone (Lithology

B). (Pl.10 A) They are formed by desiccation wind erosion, and subsequent flooding of the dried surfaces.

D. STROMATOLITE MUDSTONE (Standard Microfacies 20): Dense and closely-spaced growth laminations swelling over protuberances characterize this sediment. (Pl. 6 C) Fine lime mud is preferentially trapped on the highest areas resulting in a lamination contrary to gravity. Such stromatolitic structure is commonest in the intertidal zone (WILSON, 1975).

E. LIME WACKESTONE WITH RESTRICTED FAUNA: This lithology is characterized by light grey lime mudstones containing sparse to common miliolid formaminifera. It is a variation of Standard Microfacies 19. In addition to miliolids other small forams, ostracods, and gastropods are present. (Pl.10E) Pelletoids are also common. Occasionally grades into Type G (Pelletoidal lime grainstone), and indicates restricted circulation on marine platform (Facies Belt 8).

F. MILIOLID GRAINSTONE (Standard Microfacies 18): It is characterized by light-grey skeletal limestone composed primarily abundant tests of miliolid foraminifera. It has a sparry calcite cement and sparselime mud. Miliolid grainstones represent open and/or restricted marine platform. (Pl. 11C).

G. PELLETOIDAL LIME GRAINSTONE (Standard Microfacies 16); The pelletoids of this well sorted grainstone are typically small from 0.1 to 0.3 centimetres and eliptical. Comparatively few ostracod tests and foraminifera occur. Pelletoidal lime grainstone commonly grade into pelletoidal wackestones of Type E and characterize tidal flats. (Pl.11 E). Fenestral fabric is common.

H. LIME WACKESTONE WITH NORMAL MARINE FAUNA: This lithology includes Standard Microfacies 8 (Whole fossil wackestone) and Standard Microfacies 9 (Bioclastic wackestone). The sediment is light grey and contains fragments of diverse organisms jumbled and homogenized through burrowing. (Pl. 12 A). Some lithoclasts and pelletoids also have been observed. Bioclasts and non-skeletal fragments make up from 5 to 40 percent of the rock; but it must be noted that these extreme limits are rare and the average particle content is from 15 to 20 percent. The particles rarely show any sorting; range from 0.1 cm to 1-2 cm in maximum diameter and are mostly fossil fragments of various shapes. The dominant fossil type is mollusks. Rudists from pelecyped group make up most of the bioclasts in many samples. Caprinids are the most abundant rudists. Ratiolitid rudists, colonial corals, stromatoporoids, echinoids, and foraminiferas are common to abundant in some samples. Locally it is stylolitic. This lithology is considered representative of normal salinity open marine environments (Facies Belt 2 and 7).

I. COATED GRAINSTONE (Standard Microfacies 11, 13, and 15): The rock consists of light-grey algal encrusted, bioclastic limestones, oolitic limestones, and composite grain limestones, in which coated grains are the predominant constituent. (Pl.12B). The grains, which are well to moderately well sorted, commonly are cemented by sparry calcite and includes abundant dasycladacean algae, gastropods, miliolids and pelecypeds. Bioclasts may be micritized. (Pl.13 A) It is interpreted as a platform edge carbonate sand bankdeposit that formed on a current or wave agitated bottom.

J. RUDIST LIME MUDSTONE (Stanard Microfacies 7): This lithofacies includes light-grey rudist lime mudstone and rudist lime mudstone-skeletal lime mudstones. (Pl. 13). Caprinids are the most abundant rudists. Colonial corals, gastropods, stromatoporoids, echinoids and foraminifers are common to abundant.

K. BIOCLASTIC PACKSTONE-GRAINSTONE (Standard Microfacies 5): Ligh coloured, fine to medium grained bioclastic limestone with pack-stone or grainstone texture make up common reef flank facies . (Pl.14) The lithology consists principally of rounded, well sorted frag-ments of rudists; gastropods, corals, and foraminifera also are important contributors to the sediment.

L. MICROBRECCIA OR BIOCLASTIC-LITHOCLASTIC PACKSTONE (Standard Microfacies 4): White to light-gray, sand to gravel - and even cobble - sized skeletal limestone with packstone texture characterize this rock type. Rudist fragments make up the most of the rock. Foraminifera, corals, stromatoporoids, and gastropods are also present. Stylolites are common. They are interbedded with pelagic lime mudstones and have been interpreted as debris derived from knoll reefs. (Pl.15). Most debris is derived from breakdown of organisms growing in vast quantity on the tops of reef knolls, and not necessarily from destruction of the organic framework or previously lithified material. Another variety of this lithology is reddish-brown limestone consisting of angular to subangular clasts of consolidated rock of the bioclastic packstone-grainstone, and pelegic lime mudstone lithologies in a micritic skeletal or micritic matrix. Pelagic foraminifera are sparse to abundant. This sediment is throught to have formed by a mixing of lithologies from different environments through a process of rock fall and submarine slumping, in junction with faulting or possible small-scale turbidity flow down the steep slopes.

M. PELAGIC LIME MUDSTONE (Standard Microfacies 3); Brownish red, micritic and argillaceous micritic rocks (pelagic lime mudstone) characterize "Toe of Slope Carbonate Facies". The limestone is replaced locally by chert. (Pl.72E). The fauna comprises common to abundant planktonic foraminifera, tintinnids, and radiolaria. (Pl.15). Fossil assamblage indicates a Maestrichtian-Palaeosen age.

N. MICROBIOCLASTIC PACKSTONE/GRAINSTONE (Standard Microfacies 2): Another characteristic lithology of "Toe of slope carbonate facies" is the mixture of siltsize bioclasts and pelletoids with a very fine grainstone or packstone texture. Fine cross-lamination is common. (Pl. 14).

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CHAPTER IV

ENVIRONMENT OF DEPOSITION

I. GENERAL:

Upper Cretaceous carbonates are widely, and at some localities excellently exposed in the Gürün region, yet detailed environmental interpretations of these rocks have not been given in previous works. Gürün region was a major carbonate province during much of the Mesozoic. Study of the carbonates in the investigated area on the basis of recent depositional models, and the principles of carbonate sedimentology suggests that the Upper Cretaceous in this region, where the record is preserved, is characterized by a platform fringed by rudist knoll-reefs where shallow water carbonate deposition persisted.Calciturbidites represent the deeper facies of the platform. The development of platform is associated with the regional Upper Cretaceous transgression (BRINKMANN, 1976; p.59). Although the sequence is characterized by many oscillations and, undoubtedly, hiatuses, the trend is toward open marine carbonate environments and the development of a carbonate platform. This widespread transgression reached its amplitude during the Maestrichtian time and covered Cenomanian-Campanian shallow water deposits with pelagic limestones interbedded with calciturbidites in many places. Carbonate deposition was replaced entirely by terrigenous deposits in the Early Tertiary as diastrophism accelerated in the region.

Platform facies, with an oversimplification, can be divided into two well-known classical subfacies: "Reef Facies" and the "Back Reef Facies". The reef facies contains many rudists (dominantly caprinids with fewer radiolites) but coarse-grained bioclastic grainstones dominate. Corals, algae, gastropods, and stromatoporoids are minor contributors to the reef facies. Rudist reefs are discontinuous and are thought to be present as knoll-reef ramps (See "Organic Buildup-Facies Belt 5; p.87). This is the pattern of distribution of Recent coral reefs in Florida and Bahamas, as postulated by GRIFFITH, PITCHER, and RICE (1969). A high-energy zone with bioclastic and coated grainstones is present just behind the rudist concentrations at the platform edge.

In the platform interior, the dominant rock-forming particles are milliolid foraminifera, pelletoids, and lime mud. Algal mats and dolomitic intervals, scattered across the platform, probably represent supratidal environment. A comparison of the rock types, sedimentary structures, and faunal assemblages of these rocks with those of recent carbonate sediments indicates that these ancient rocks may be subdivided into lithofacies that are similar to modern tidal flat deposits.

Basin margin facies, the deep water equivalent of the shallow water carbonates in the investigated area, is characterized by pelagic limestones and calciturbidite beds composed of lithoclasts in a matrix of lime mudstone. Clast include fragments of fossils derived from platform and are generally grain supported. The mudstone matrix commonly contains pelagic microfossils. Graded beds are common in some localities. Silica replacement of bioclasts and chart replacement occurs loeally.

II. IDEAL CARBONATE PLATFORM MODEL AND THE CONCEPT OF FACIES BELTS:

As pointed out by WILSON (1975) recognition of a consistently recurring pattern of limestone facies in the ancient geological record has given birth to the concept of carbonate facies belts. The evolution of the concept has been in progress since about 1950 and has resulted in the development of essentially a single ideal model. NEWELL and Others (1953), EDIE (1958), SHAW (1964), IRWIN (1965) and COOGAN (1969) are the pioneers of the concept (Figs. IV - 1 to IV - 5). On a very gently sloping shelf there is a tendency for a seaward, low energy zone to develop below wave base. A zone of higher wave energy is situated somewhat shoreward where maximum organic productivity occurs. In this zone waves drag the bottom. Another interior low energy zone develops shoreward. These three belts (basin, shelf margin and back reef) may reach a considerable thickness and commonly form a prograding or up-building sequence after a period of marine transgression. Thus a carbonate platform is formed. The hydrologic, climatic and organic controls elaborate these three fundamental environmental belt into nine subenvironments. Its essential outline was published as a discussion of WILSON with W. TYRREL in SEPM Spec. Publ.14 - Depositional Environments in Carbonate Rocks (1969, p.18) and in much more detailed form applied Devonian carbonate complexes of the Canadian Rocky Mountains by DOOGE (1966). WILSON outlined the general principles of the concept in his paper in 1970 and later (1975) discussed them in great detail and applied them to depositional patterns from numerous examples of carbonate facies. In conclusion he notes that "...It is significant that this pattern is so persistent: it offers essentially a single model for prediction of

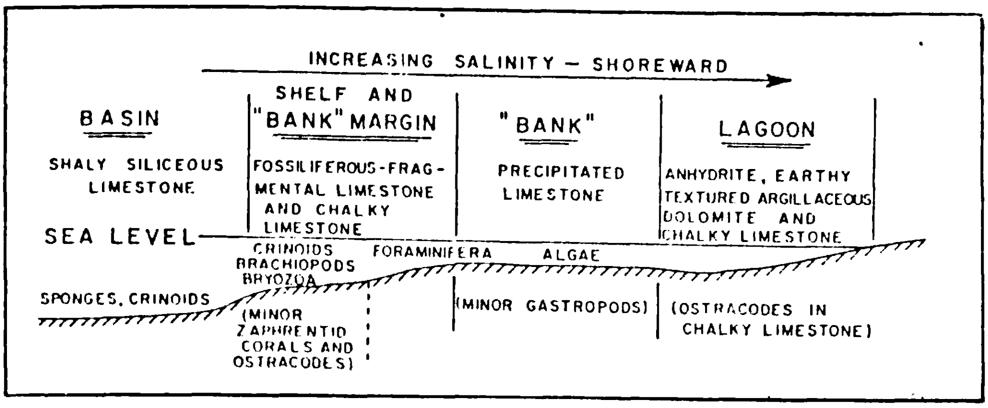

Fig.IV-1: General carbonate model of EDIE (1958)

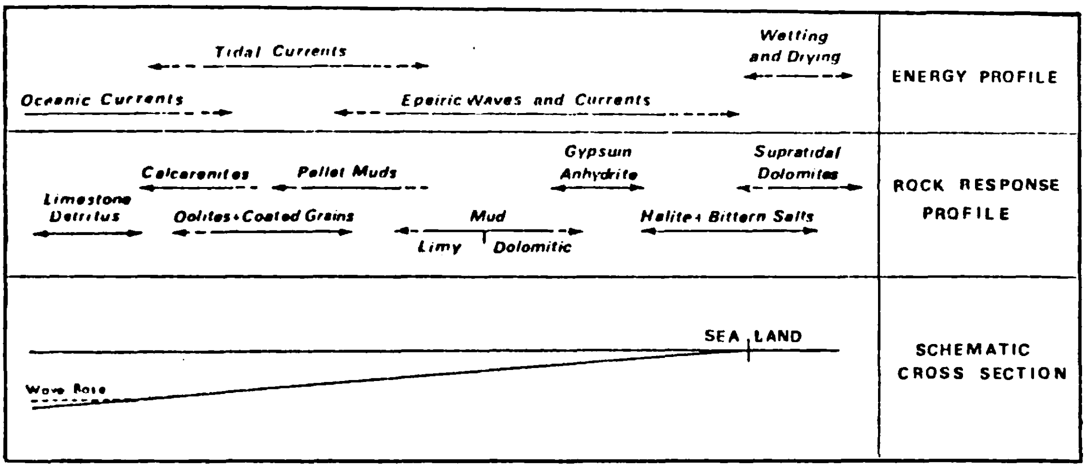

Fig.IV-2: Simplified energy and rock response profile across a stable cratonic shelf having no terrigenous input(Modified After SHAW, 1964)

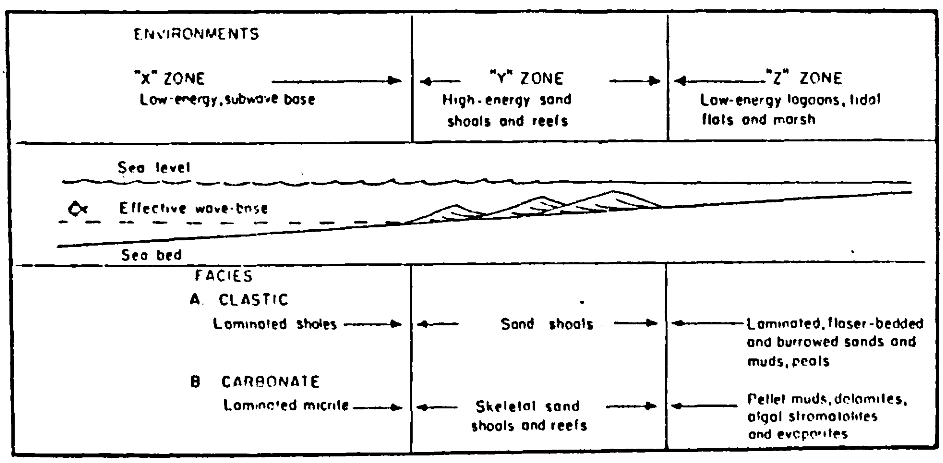

Fig.IV-3: Epeiric sea shelf sedimentary model (Based on ILLING, 1965; After SELLEY, 1976)

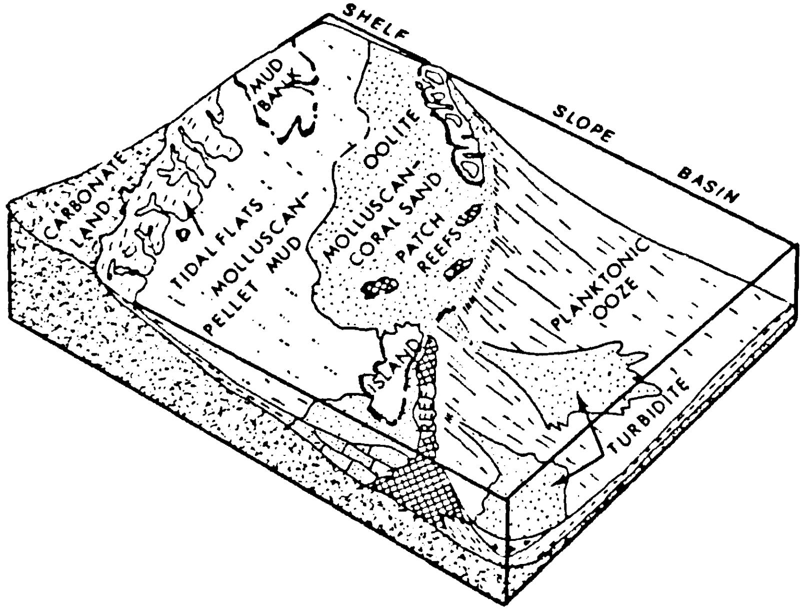

Fig.IV-4

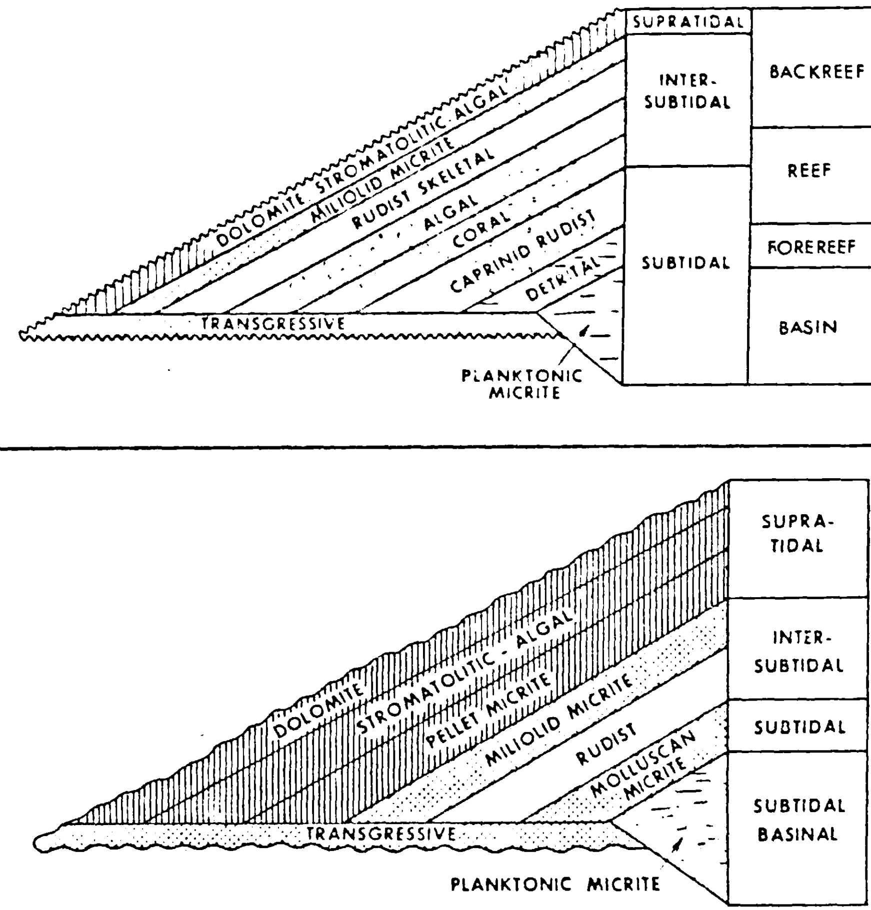

Fig.IV-5

Fig.IV-4 : Generalized framewok of carbonate environments on the shelf, on the slope, and in the basin (After COOGAN)
Fig.IV-5 : Cretaceous reef (top) and non-reef (bottom) cycles (After COOGAN, 1969).

geographic distribution of rock types. It thus becomes a tool for use in practical field mapping, in designation of rock units for correlation purposes, for depositional interpretations, and in the search for petroleum, and for metallic ores whose distribution may be facies - controlled ...(P.24)". Once the uniformity of the model has been established, by comparison of various individual carbonate sequences of widely different geographic setting and age, it is possible to greatly reduce or eliminate the semantic problems and to stress the limited number of parameters of each particular facies. Several studies have shown the basic similarity of carbonate sequences of various age and geographic setting and eliminated the view that each carbonate body is unique. The delineation of the major carbonate facies belts does not mean that all of these must be present in each carbonate sequence. Several facies may be absent. An outline of ideal carbonate platform depositional model, mostly from the work of WILSON (1975), with some modifications, is given below and shown on Table III - 4.

This model was applied to Carboniferous rocks of Arctic Alaska (ARMSTRONG 1974) and has been applied to the Gürün sediments in this account.

FACIES BELT 1 - BASIN FACIES: Water is too deep and dark for benthonic production of carbonates. The sediment is chiefly pelagic and dependant on the amount of influx of fine argillaceous and siliceous material and rain of decaying plankton. Dominant lithology is calcareous and siliceous thin beds, which are rich in ferric iron. The influence of marine currents is negligible, or absent, as witnessed to by the millimetre laminations. Standard Microfacies Types 1 (Spiculite), 2 (Microbioclastic packstone/grainstone), and 3 (Pelagilime mudstone) are dominant.

FACIES BELT 2 - OPEN SHELF FACIES: The depth is sufficient to be below normal wave base but intermittent storms may affect bottom sediments. The water is generally oxygenated and of normal marine salinity, with good current circulation. The sedimentation is quite uniform; main lithology is thin to medium bedded very fossiliferous limestone interbedded with marl and chert. Bedding surfaces show diastems and burrowing is common. Dominant Standard Microfacies Type are Bioclastic (SMF - 9) and Whole Fossil Wackestone (SMF - 8). It is very similar to Facies Belt 7 (Open Marine Platform Facies).

FACIES BELT 3 - TOE OF SLOPE CARBONATE FACIES: This facies is formed at the toe of slope of carbonate producing shelf from material derived from the shelf. It is situated at or below wave base and oxygen level. Pelagic Lime Mudstone (SMF - 3) and Bioclastic, lithoclastic Packstone (SMF - 4) are dominant microfacies types. Some beds are laminated with micrograding of lithoclastic and bioclastic debris from upslope. Chert can be seen.

FACIES BELT 4 - FORESLOPE FACIES OF CARBONATE PLATFORM: Generally the slope is located above the lower limit of oxygenated water and extends from above to below wave base. Composed of debris deposited on an incline, formed seaward as the platform grows. Incline may be as steep as 30^{o}. Sediment is unstable and varies greatly in size and shape depending on the water energy upslope (lime muds and sands, boundstone and sedimentary breccia). Bioclastic - Lithoclastic packstones and grainstones (SMF - 4 and 5) and coarse gravel grainstone (SMF - 6) are the most common microfacies types.

FACIES BELT 5 - ORGANIC BUILDUP OF CARBONATE PLATFORM: The water energy, steepness of the slope, organic productivity, amount of frame construction, binding or trapping, frequency of subaerial exposure, and consequent cementation determine the ecologic character of the organic buildup. WILSON (1974) differentiated three types of linear shelf margin organic buildup profiles. Type I is formed by downslope carbonate mud and organic debris accumulations. Type II includes linear platforms composed of knoll reefs[(x)], commonly forming on gentle slopes at outer edges of shelf margins; organic frame building organisms in isolated clumps or encrusting sheets or organisms growing up to wave base and stabilizing debris accumulations. Type III is frame-constructed reef rims.

The dominant microfacies is masses and patches of boundstone (SMF - 7). Intermound areas consist commonly of grainstone and packstone. Roofed cavities and laminations contrary to gravity characterize this facies belt.

FACIES BELT 6 - PLATFORM EDGE CARBONATE SAND BANKS: They may be in different forms: Shoals, beaches, offshore tidal bars, eolianite dune islands. Depth of such marginal sands range from 5 or 10 m. to above sea level to 5 or 10 m. deep. Although the environment is well oxygenated, because of shifting substrate, is not hospitable to marine life. Rounded and fairly well-sorted grainstones are typical for this facies belt. Some of the grains are coated and oolitic, and some merely rounded bioclasts.

FACIES BELT 7 - OPEN MARINE PLATFORM FACIES: The general term "shelf lagoon" can be applied to this environment. Geographically these are located in straits, open lagoons, and bays behind the outer platform edge. Water depth is generally shallow (a few tens of metres deep at

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(*) Knoll Reef: Isolated more or less circular area of organic frame-built growth in deeper water below wave base.

most). Circulation is moderate and salinity varies from normal to higher. The sediments are variable but contain considerable amounts of lime mud. Burrowing and pelleting of the sediment are common.

FACIES BELT 8 - FACIES OF RESTRICTED CIRCULATION ON MARINE PLATFORM AND TIDAL FLATS: This facies includes mostly fine sediment in very shallow, cut-off ponds and lagoons; coarser sediment in tidal channel and local beaches; and the whole complex of tidal flat environment. The conditions are highly variable and include fresh, salty, and even hypersaline water; areas of subaerial exposure; both reducing and oxydizing conditions; and both marine and fresh-water swamp vegetation. Diagenetic effects are strongly marked in the resulting sediment. Lithology is highly variable too. Lime muddy sediment with much dolomite is prevailing rock type. Grainstone is rare except for thoroughly pelleted sediment. Channels contain lithoclastic grainy sediment; clotted, pelleted mudstones and wackestones. Lamination, bird's eye fabric and stromatolites are common.

FACIES BELT 9 - SUPRATIDAL AND INLAND PONDS FACIES: This facies belt is well developed in an evaporatic climate, and includes the areas of sabkha salines, and salt flats. On normal conditions algal mats is the dominant lithology. Unlaminated homogenous lime mudstones characterize the sediments in the ponds.

In the present account, the author has become able to recognize, describe, and interpret at least five distinct Facies Belts. Distinctive associations of lithologies, textures, structures, and fossils provided the basis for recognizing these five Facies Belts in the sections examined. Excluding basin margin deposits, all the sediments were deposited within a few decimetres of mean sea-level under conditions of shallow water sedimentation, and are comparable with modern, supratidal, intertidal, and subtidal carbonate facies. Description of these Facies Belts, including data on pertinent petrographic detail, fauna, textures and structures are presented further below. Each Facies Belt is interpreted environmentally by the use of a combination of significant petrographic and paleontologic criteria and relations to associated facies.

III. SHALLOW WATER DEPOSITS

A. TIDAL FLAT SEDIMENTS

Many studies of Holocene carbonates (e.g. GINSBURG, 1957; NEWELL and RIGBY, 1957; ILLING, WELLS, and TAYLOR, 1965; SHINN, LLOYD, and GINSBURG, 1969) have provided criteria for recognition of tidal flat environments. Other workers (e.g. LAPORTE, 1967 and 1969); ROEHL, 1967; MATTER, 1967; BRAUN and FRIEDMAN, 1969; MUKHERJI, 1969; TEXTORIS, 1969; YOUNG, FIDDLER and JONES, 1972) have used these criteria in analog environmental interpretation of ancient carbonate rocks. A thorough review of the carbonate shore line and tidal flat environments is that of LUCIA (1972).

In the present account the same criteria have been used to recognize the tidal flat sediments in the Upper Createaceous carbonates of the Gürün region, and to construct a depositional model for these rocks.

LUCIA (1972) notes that the tidal flat environment is the most common shoreline environment in modern carbonate settings. This statement is also valid for the Gürün Carbonates. The study of the measured sections of Cenomanian-Campanian age in the Gürün region has shown the presence of thick intervals of supratidal rock with interbedded intertidal and marine rocks. They consist mostly of carbonate muds. The abundance of stromatolites and algal bedding associated with mud cracks and related features make it apparent that these limestones were deposited in very shallow water and that some were deposited in a tidal flat environment. The petrographic, biologic and primary structural features of the Cenomanian-Campanian limestones are strikingly similar to those found in definite, present day, shallow water carbonate environments. This environment is characterized by a diversity of carbonate rock types and sedimentary structures and represents Facies Belt 8 (Facies of restricted circulation on marine platform) and Facies Belt 9 (Supratidal and inland pond facies). The restriction of sea-water circulation can occur in two ways. First, the low permeability of the tidal-flat sediments provide a restriction to circulation. The sediment in and under the tidal-flat environment is filled with marine water, whose circulation with the ocean is restricted due to the low permeability of the sediment. The second form of restriction is the narrowing of the channel connecting the ocean with a standing body of water (LUCIA, 1972). The restricted marine carbonates were deposited in an environment which was not favourable to the development or survival of biota inferred to require normal marine conditions. The depositional environment of the rocks to be described below is interpreted from the sequence and from a knowledge of tidal-flat sedimentation.

Climate and physical setting determines whether a tidal flat will be evaporitic or non-evaporitic (KINSMAN, 1969). The absence of stratified evaporites or convincing evidence of their former presense in the study area precludes a physical setting during the Upper Cretaceous time conducive to the sedimentation of stratified evaporites out of standing body of water. On the other hand the non-existence of evaporite minerals produced by precipitation from interstitial waters within sediment (i.e. diagenetic) indicates that an arrid climate was not prevailing during that time.

The primary source of the sediment deposited on the tidal flat is the marine environment. The sediment is transported by tidal currents and storms from the sea onto the land (in other words the direction of transport is just opposite of that during alluvial deposition).

Tidal flat environment can be divided into general environments of deposition, based on the action of the tides (LUCIA, 1972).

(1) Supratidal environment: It is defined as that area of tidal flat sedimentation out of reach of the daily tides.

(2) Intertidal environment: Mean low-tide and mean high-tide levels limit this environment.

(3) Subtidal environment: It takes place below mean low-tide level and includes both restricted subtidal and open subtidal."Shallow marine" is synonymous with subtidal.

Between these three environments, there is a continuous gradation from one to another. As a result, particular parts of the investigated area necessarily share characteristics of different facies. Despite the apparent complexity and constant change from one rock type to another, still it is possible to get a rational picture of Cenomanian-Campanian deposition if these three facies are used as interpretive guides.

The algal stromatolites are the most diagnostic organic feature of the tidal-flat sediments (LOGAN, REZAK, and GINSBURG, 1964). In Recent tidal-flats algal stromatolites are found primarily in the high intertidal or the low supratidal environments (ILLING, WELLS, and TAYLOR; 1965 and LOGAN, REZAK, and GINSBURG; 1964). Grazing gastropods and burrowing crustaceans destroy algal mats in modern depositional environments (GARRETT, 1970; FRIEDMAN, AMIEL, BRAUN, MILLER, 1973). Algal mats, grazing gastropods and crustaceans generally prefer environments close to sea-level (CUSEY and FRIEDMAN, 1976). The ecologic control of their distribution in modern environments appears to be salinity. Although algal mats have a remarkable tolerance for water of wide range of salinity, grazers and burrowers which do not have a parallel tolerance for waters of high salinity restrict the mats to the hypersaline conditions. Thus, in the Bahamas cerithid gastropods live in the intertidal zone, whereas mats survive mostly in the supratidal zone (SHINN, 1968). Similar relations have been observed in the present study. While the algal structures are abundant, burrows are infrequent and gastropods are absent; when burrows and gastropods are present, commonly in abundance, the algal structures are scarce or absent.

Supratidal environment lies exposed subaerially for long periods of time and is covered only by spring or storm tides. The resulting sedimentary structures reflect the periodicity of sedimentation as well as the effects of exposure. Mud cracks and fenestral fabric are diagnostic. Alternating wetting and drying lithifies and tends to break the supratidal sediment into lithoclasts. Although shrinkage pores and polygonal structures typical of supratidal rocks appear to be absent in the sediments investigated, intraformational breccias, small erosion pits, and non-polygonal cracks are present and give evidence for periodic subaerial exposure. Because in most cases it cannot be determined whether such periods were diurnal or perhaps associated with spring tides, the general term "tidal" rather than intertidal or supratidal is preferred in this account.

The supratidal facies have been recognized in the study area by the presence of various combinations of the following emergent dessication features:

(1) Fenestral fabric: "Fenestrae" are primary or penecontemporaneous gaps in rock framework, larger than grain-supported interstices. They form by decay of sediment-covered algal mats, shrinkage during drying, or accumulations of pockets of gas. Fenestrae commonly are somewhat flattened and parallel with the bedding planes (TEBBUTT, CONLEY, and BOYD; 1965). Fenestral fabric is not common but some is found in almost all measured sections. Sparry calcite cement is the only fenestral filling. Fenestral fabric in the study area is commonly associated with some lithoclasts. Fenestral fabric in modern carbonate setting is most abundant in supratidal sediments, sometimes present in intertidal sediments, and never present in subtidal ones (SHINN, 1968). Rock units with fenestral fabric in the study area are interpreted to have formed in supratidal or intertidal environments depending on the degree of emergent dessiccation present.

(2) Mud cracks: The cracking of muddy sediment by desiccation most commonly occur during subaerial exposure but may also occur subaqueously when overlying water becomes more saline and draws interstitial water out of the sediment (HECKEL, 1972; p.243). Therefore supporting evidence of subaerial exposure

is desirable for environmental interpretation.

(3) Brecciation: Another fabric of diagnostic of the supratidal zone is the brecciation and formation of lithoclasts. They are commonly associated with the other supratidal deposits, particularly the algal mats; and because many of them are composed primarily of small slabs and chips of such mats they may be related to intense dissication during subaerial exposure. Lithoclasts are abundant and found in beds as well as scattered throughout the irregular laminated supratidal rocks. They range in size from several millimetres to several cantimetres long.

Other environmental criteria used recognizing the supratidal sediments are irregular laminations and scarcity of or lack of fossils.

The supratidal rocks show characteristic irregular laminations. The type of lamination varies from poorly defined to very distinct. They range in thickness from 0.25 to 1 mm. The laminae are composed either of fine grained dolomite or pelletoidal mudstone(*). The lamination is not explainable as due to the settling of sediments in still water, deposition from currents of variable velocity, or periodic chemical precipitation. On the other hand the laminae do not pinch and swell to compensate to relief on the underlying surface, but rather bear an encrusting relationship. According to LAPORTE (1967) the alternating pelletoidal mudstone and fine grained dolomite laminae are recordings of successive periodic flooding of the supratidal environment by unusually high tides, when the waters recede carbonate mud in suspension settles at and mixes with the pelletoids. Algal mats then develop on this newly deposited, wet layer of sediment. In accordance with the work of LOGAN, REZAK, and GINSBURG (1964) the horizontal, slightly wavy, bituminous laminae are interpreted as having been formed by uninterrupted algal mats established in quiet water, just a few cantimetres deep, following intermittent flooding of the supratidal flats. The continued subaerial exposure causes hardening and cracking of the sediment. In some instances, vertical breaks are found crossing laminations and appear to be desiccation cracks (Pl. 4). Evaporation at the algal mat surface brings magnesium rich sea water upward through the sediment by capillary action. As the result of this process dolomite precipitated at the mat surface as a fine grained layer. Some local replacement of carbonate grains by dolomite can be seen below the mat surface. Similar

(*) X-Ray analysis of these rocks indicate their constituent mineralogy to be mostly calcite and dolomite with minor amounts of quartz and illite.

interpretation of the origin of pelletoidal mudstone and fine grained dolomite laminae was proposed by SHINN, GINSBURG, ans LLOYD (1965).

Recently GEBELEIN and HOFFMAN (1973) have suggested that the dolomite layers are secondary and showed that the dolomite layers correspond to the original algal mat layers, and the limestone layers to the sediment layers between the mats. According to the same authors, dolomite is formed by replacement of the surrounding calcium carbonate matrix in contact with the algal layers. Decomposition of the organic residue releases the excess Mg in the micro-environment of the algal layer. The origin of these dolomite laminae is not the scope of this study. However, the absence of gypsum or anhydrite, and the crystal cast pseudomorphs of these minerals in the sediments investigated seems supporting an origin non related with "dense hypersaline brines".

Except for scattered, disarticulated ostracod valves and small forams, fossils are very rare in the supratidal sediments. Some fragments of marine fossils have been observed but they show isolated occurrrences and are thought to represent skeletal materials thrown onto the supratidal flats during abnormally high tides.

In section 8A rust coloured dolomitic-algal stromatolitic mudstones have been observed. By analogy with moder Persian Gulf tidal flats they represent supratidal conditions in the landward side of the "flat zone" of KENDALL and SKIPWITH (1968; p.1046).

Under the petrographic microscope, the texture of the supratidal rocks varies considerably. Floored cavities, small fenestral fabric, brecciation and other features probably resulting from early diagenesis make thin sections of supratidal rocks distinctive. Fossils are scarce, and when present are primarily small forams and ostracods.

The following rock types in order of decreasing relative abundance are characteristic of the supratidal facies:

(1) Algal-mud lime mudstone (Lithology B)

(2) Pure lime mudstone (Lithology A)

(3) Lithoclastic packstone or grainstone (Lithology C)

The intertidal facies of this account consist of carbonate that were deposited between daily mean high and low tide, and therefore were subjected to daily inundation and emergence. The intertidal sediments are not so distinct as the supratidal sediments. Therefore the recognition of the intertidal sediments can be best accomplished through the recognition of supratidal rocks. Once the supratidal environment has been recognized, the intertidal - and marine - environments are inferred to be present. Thus, the sediments immediately underlying the supratidal probably represent the intertidal zone. The following rock types are thought to be characteristic of the intertidal facies in the study area:

(1) Lime wacketstone with restricted fauna (Lithology E)

(2) Pelletoidal lime grainstone with fenestral fabric (Lithology G)

(3) Stromatolite mudstone (Lithology D)

The head-like algal stromatolites attached to the substrate were developed in intertidal areas-by analogy with present day forms - and were exposed to water sufficiently agitated to destroy the mats or inhibit their grwoth between flourishing heads.

Recent intertidal carbonate sediments commonly appear to be pelletoidal packstones with gastropod (cerithidea) shells common. The thickly laminated, pelletoidal unit is identical to intertidal deposits of the Persian Gulf (ILLING, WELLS, and TAYLOR; 1965) and strikingly similar to deposits of the intertidal flats of the Bahama Islands (SHINN, GINSBURG, and LLOYD; 1965) and Floriday Bay (GINSBURG; 1956).

Rocks with current features are found primarily within and immediately below intertidal rocks. The most common type of current feature present is the current lamination. The current laminated rocks have alternating laminae of sorted fine pelletoidal sand and muddy fine pelletoidal sand. These rocks probably represent tidal channel deposits in the Gürün region.

The differentiation between intertidal and restricted subtidal has been found impossible because of the similarity of the sedimentary structures and scarcity of fossils. Some of the pelletoidal mudstones and wackestones with scattered forams and gastropods are thought to represent the restricted subtidal deposits in the investigated area.

On the other hand the transition from intertidal rocks down into normal marine rocks might be represented by a recognizable change in the fossil assemblage. The presence of a normal marine fossil assemblage such as a coral, bryozoan, echinoderm, and stromatoporoid is excellent evidence for normal marine (subtidal) deposits which form Facies Belt 7 (Open Marine Platform Facies). The rock types noted in this facies generally range from lime wackestone with normal marine fauna (Lithology I) to pelletoidal lime grainstone (Lithology G). The stratification which is so well displayed in the other facies of the Gürün region such as lamination in the supratidal facies and alternating layers of pelletoidal lime mudstone/wackestone and pelletoidal lime grainstone in the intertidal facies is usually absent in this facies. As noted above, stratification in the supratidal facies was caused by algal mats and periodic flooding by waters with mud in suspension which eventually settled out. Absence of burrowing organisms in this environment secured the preservation of the lamination. Although some burrowers were present in the intertidal environment, the rate and effects of sediment reworking were greater than burrowing and therefore depositional stratification remained dominant. In the subtidal facies, the rates and effects of burrowing were higher and stratification that was formed by wave and current action was lost. Subtidal deposits of the Gürün region are therefore represented by thick to massive beds.

B. PLATFORM EDGE CARBONATE SAND BANKS:

This facies has been termed as "winnowed platform edge facies" by WILSON (1969, p.18; and 1975) and consists of well sorted, winnowed sands, mostly skeletal debris, occasionally in situ precipitated oolites in which coated grains are predominant constituents. It is inferred to have been deposited along a platform edge separating carbonate forming environments in shallow water and deep water. It is believed that they were formed as subaqueous bars and dune like accumulations such as today occur in the Florida and Bahama carbonate provinces; especially near the south edge of the Tongue of the Ocean. The morphology and composition of sand bodies have been described in detail by BALL (1967). Their internal sedimentary structures, mainly cross-bedding, in the Bahama sands have been described by IMBRIE and BUCHANAN (1965).

The most striking and distinctive feature of this facies is its large-scale cross-bedding (Pl. 12). The thickness of the cross-bedded sets is variable - up to 3 m. individual foresets pass completely through a bed. They are not graded. Both of these characteristics suggest deposition from currents of uniform and continuous competency.

This facies is characterized by the presence of coated bioclasts in sparry calcite cement (Lithology I). Grain support reflects either greater production of skeletal grains relative to production of influx of mud, or sufficient winnowing of muds to produce a residual deposit of larger grains. Slow deposition in an environment subject to longer winnowing and periodic turbulence is indicated by the presence of abraded grains.

There are considerably more skeletal grains in this environment than in the previously described environments. Fossil fragments include crinoids, brachiopods, echinoids and calcareous algae. They make up the bulk of the rock with the other grain types such as lithoclasts and coated grains. Skeletal grains and lithoclasts are the most common nucleu for coated grains. Other grains of minor importance are quartz, glauconite and hematite. Nevertheless, the fossil assemblage of this facies differs from that characterized by the organic buildup (Facies Belt 5). This feature suggests that platform edge carbonate sand banks were not derived principally from the organic buildup; but may have been deposited "upslope" from and in more agitated water than the more muddy platform edge deposits of the organic buildup (TYRELL, 1969).

C. ORGANIC BUILDUP

A conflicting usage of terminology has perhaps caused more problems in carbonate studies than any other single group of genetically associated rocks. To avoid confusion therefore, it is advisable to each author to define his terms and where possible use existing rather than new ones. The terms and definitions used in this study are given below:

Carbonate Buildup: A body of locally formed (laterally restricted) carbonate sediment which possesses topographic relief. It carries no inference about internal composition (WILSON, 1975; p.20).

Bank: A linear deposit consisting of nonfragemental skeletal matter, formed in place by organisms (such as crinoid and braciopods) that lack the ecologic potential to erect a rigid wave-resistant structure (NELSON, BROWN, and BRINEMAN, 1962; p.242).

Reef: A reef is a buildup that displays

(1) Evidence of
  (a) Potential wave-resistance
or (b) Growth in turbulent water which implies wave resistance
and (2) Evidence of control over the surrounding environment (HECKEL, 1974; p.96)

Ecologic Reef: Buildup formed in part by wave resistant framework constructed by organisms (DUNHAM, 1970)

Stratigraphic Reef: Thick lateral restricted masses of pure or largely pure carbonate rock (DUNHAM, 1970).

Knoll Reef: Isolated more or less circular area of organic frame build growth in deeper water below the wave base. In general reef knolls are used to identify individual buildups of shelf margins (WILSON, 1975; p.22).

Study of many varied carbonate complexes in ancient rocks has resulted in recognition of recurring patterns of shelf-margin facies and associated buildups which has been discussed in detail by WILSON (1974 and 1975). Based on the stratigraphic profile, shape and disposition of discrete carbonate bodies along the profile slope, and environmental interpretations through analysis of texture and biologic microfacies, he has differentiated three patterns:

(1) Downslope lime-mud accumulations;

(2) Linear belts of ecologic (i.e. frame build) knoll shaped reefs present on gentle slope at outer edge of the shelf margin;

(3) Organic reef rim of resistant frame work.

Shelf margin of the Gürün Carbonate Platform belongs to type 2 (Fig. IV - 6).

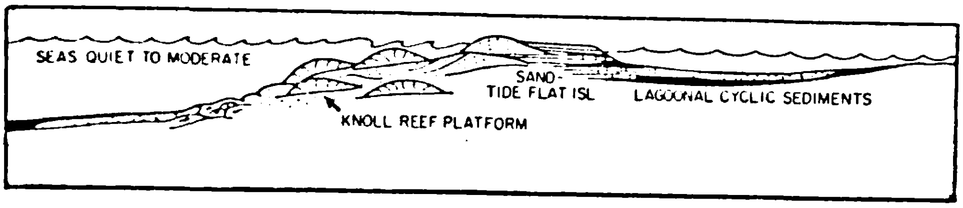

Fig.IV-6: Type 2 Carbonate Shelf Margin (After WILSON, 1974)

Rudists played the most important role in these organic buildups and added immensely to the volume of material associated with them. Because rudists are extinct fossils and are the only apparent reef-building organisms among the bivalves, problems concerning the capability of rudists to form wave-resistant structures are not fully understood at the present. It generally has been considered that rudists were no means true reef-building organisms, but rather "Hard bottom type bank-formers" (LOWENSTAM and EPSTEIN, 1959). However, some authors considered that the rudists must be classified as true reef builders because of their apparent ability to form topographic wave resistant rigid frameworks which, in turn, brought about environmental control. In constrast to recent coral reefs they were almost completely barren of encrusting organisms. Probably they were too large to be easily coated or encrusted. It is suggested that the framework stability of the rudists was achieved mainly by the simultaneous growth of individual organisms in crowded populations. Mutual cementation of interlocking organisms during the secretion of the skeletons also played a part in the construction of framework rigidity.

LOWENSTAN and EPSTEIN (1959) stated, on the basis of palaeo-temperature determinations, that the large and crowded populations of rudists developed in a climatic belt comparable to the present day tropics. They are well known in an equatorial belt from 40°N to 20°S along the Tethyan seaway through Southern Europe, in the Middle East, across

Southern Asia and around the Gulf of Mexico-Caribbean region. The palaeogeographic configuration and the lithofacies assemblages of the rudist knolls in the Gürün region appears to be very much like those of rudist reefs described from margins of other carbonate shelfs (e.g. Gulf of Mexico - FISHER and RODDA, 1969; GRIFFITH, PITCHER, and RICE, 1969; COOGAN, BEBOUT, and MAGGIO, 1972 - Central Apennines - CARBON, PRATURLON, and SIRNA, 1971 - and Middle East - BEIN, 1976). The only remarkable difference is in the age of these buildups. The age of the rudist buildups cited above are Albian-Turonian whereas rudists characterize Maestrichtian Stage in Turkey (p.57 ; and also see BRINKMANN, 1976).

The discontinuous distribution of the rudist knolls appears comparable to recent coral reefs in the Florida and Bahamas Platform. They are accumulations formed by individual mounds of caprinids with some ratiolitids. They consist of a core of whole rudist shells. Some of the rudists are in growth position (Pl.8, 9) but most are not. They are commonly disoriented in a lime mud matrix, because in the region of only moderate wave action, the frame builders offer enough protection to prevent its removal. Along with the rudist dasycladacean algae provided much skeletal detritus to the knolls; and often gastropods, corals, and some large foraminifers (e.g. Orbitolina and Cuneolina) are locally important sediment contributors. Generally carbonate grainstone consists of the inter-knoll area because wave and current action is sufficiently great to remove only the finest debris; hence only lime sand accumulated around the reef cores.

D. RECONSTRUCTION OF THE DEPOSITIONAL ENVIRONMENT:

The carbonate depositional model for the Gürün region represents a dynamic system controlled by rates of subsidence and amounts of carbonate production on the shelf. The Late Creaceous represents an interval of unusually high sea level stand, or unusually low stand of the land. Thus, with time, major part of the land was flooded by relatively deep epicontinental sea. In the Albian, supratidal deposits indicate the early phase of the major flooding which was broadly continued at least until the close of the Maestrichtian. The palaeogeographical picture reconstructed from the study of Cenomanian-Campanian limestones of the Gürün region, by analogy with Recent carbonate environments is an environment very near mean sea-level having great lateral extent with very low relief. A supratidal coastal plain was exposed on the Gürün carbonate platform.

An intertidal environment was parallelling the southwestern edge of the platform and shallow marine environments were located further southwest. The rocks deposited on the platform maintained almost the same thickness over most of the area and everywhere are evenly bedded. Similar rock types form uniform and homogenous belts which extend over kilometres. Modern analogues of carbonate sedimentation on the inner parts of the shallow shelves have been studied in the BahamaIslands and Persian Gulf by PURDY (1963); ILLING, WELLS and TAYLOR (1965); SHINN, GINSBURG, and LLOYD (1965) and LUCIA (1972). The shallow marine shelf is, and has been through geologic time, an area of rapid sedimentation. The study of the measured sections of the Cenomanian-Campanian age in the Gürün region has also shown the presence of thick intervals of supratidal deposits interbedded with intertidal and marine rocks.

Unfortunately, excluding the eastern part of the area, no outcrops or the knoll reefs has been found in the limits of the investigated area. This may be due to two factors. Firstly, these knolls may have been eroded; secondly, they may be overlain by younger sediments. Whatever the case is, the enormous volume of skeletal debris encountered in the calciturbidites strongly indicates the presence of these knolls. Two outcrops of knoll reefs have been found in the eastern part of the area, in the vicinity of Hamalçay (H76). In both cases rudists are in growth position (Pl.8,9). Besides these two occurrences some large blocks of knolls were emplaced into the turbidite succession by slumping (See p.109 and Pl.29C).

Cenomanian-Campanian limestones deposited on the shallow water platform gave way to the Maestrichtian calciturbidites and pelagic limestones which indicates a subsidence after Campanian times into considerable depths. Details of deeper water environments is given in p.95 .

Figure IV - 7 represents a conceptual model of the Gürün Carbonate Platform. It is similar in many respects to carbonate models developed for many platforms in the geologic literature. The representation of Figure IV - 7 in no way characterizes a particular time interval but represents the whole of the late Cretaceous.

IV. DEEPER-WATER SEDIMENTS (CALCITURBIDITES):

A. GENERAL:

Calciturbidites are the carbonate equivalent of the turbidite facies. The term "allodapic" was applied to certain limestones believed

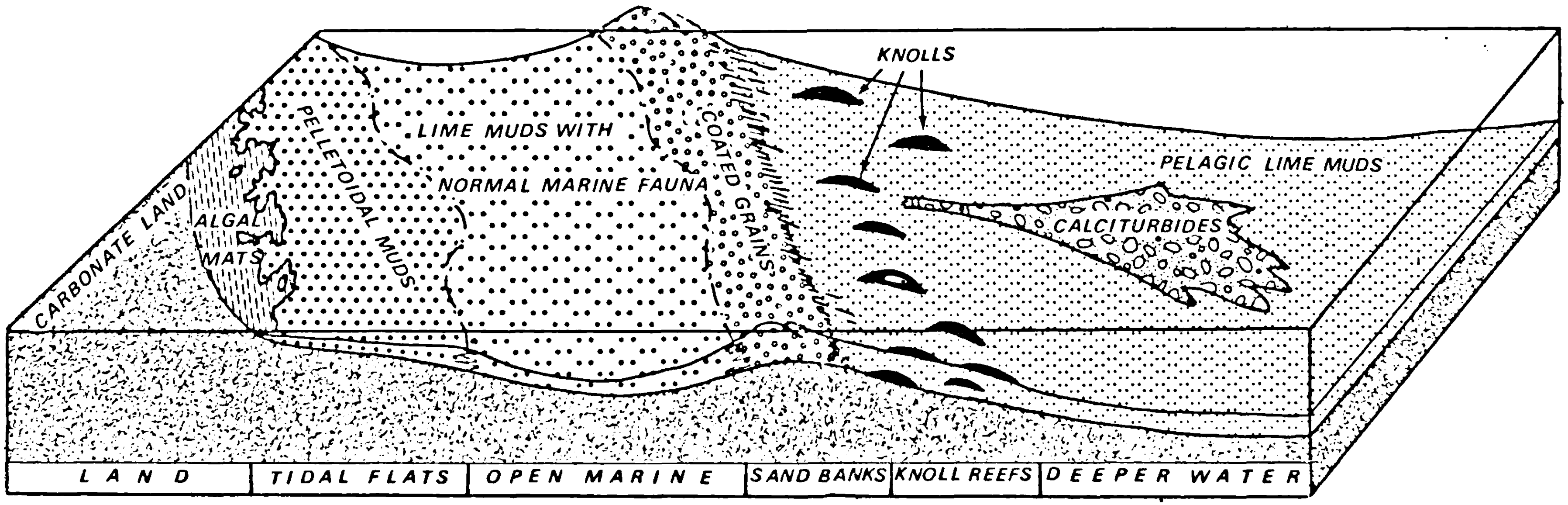

Fig.IV-7 : Diagrammatic reconstruction of environment of deposition of the Gürün carbonates during the Late Cretaceous Epoch, based on the present study.

to have been deposited by turbidity currents in relatively deep water that derived their load from shallow water areas (MEISCHNER, 1964). The beds found in the Gürün region fit into MEISCHNER's definition of "allodapic limestones" if it is somewhat broadened. Differences between his examples of the Kulmplattenkalk and Devonian Flina from Rheinshces Schiefergebirge (Germany) can be seen in that with the limestones encountered in the present study.

(1) The authochtonous pelagic sediments interbedded with the turbidites are calcilutites and not clays;

(2) The pelagic deposits were not soft but largely cemented and thus inhibited erosion by the turbidity currents. Seafloor solution produced indurated pelagic limestone clasts, which were subsequently incorporated into the turbidity current deposits;

(3) Rich bentonic faunas were present in the pelagic beds led the bioturbation of the upper parts of the limestone particles.

Similar beds composed of shallow water carbonate particles occur in Silurian of Indiana, U.S.A. (CAROZZI and FROST, 1966); Devonian of Cornwall, England (TUCKER, 1969); Devonian and Carboniferous of the Rhenish Geosyncline, Germany (EDER, 1970); Pennsylvanian of the Marathon region of Texas, U.S.A. (THOMSON and THOMASSON, 1969); Jurassic of the Northern Alps (GARRISON and FISCHER, 1969). BANDEL (1974) has given examples of pelagic limestones with interbedded calcareous turbidites from the Devonian and Carboniferous of the Carnic Alps (Austria). Extra-ordinary thick allodapic limestones have been described by SCHOOLE (1971) from the Upper Cretaceous Monte Antola Flysch in the Northern Appennines (Italy). Several Upper Cretaceous flysch formations of the East Alps, consisting of an lateration of carbonate rich sand - and silt-sized turbidites interbedded with pelagic sediments have been reported by HESSE (1975) and HESSE and BUTT (1976). Recently ROBERTSON (1976) has reported calciturbidites from the Lower Tertiary of the Troodos Massif, Cyprus. Numerous deep-sea carbonates have been examined as part of the Deep Sea Drilling Project. The results of these studies can be found in SCHOLLE (1974).

Modern examples of deep-sea carbonate turbidites have been reported from the vicinity of the Great Bahama Banks (e.g. RUSNAK and NESTEROFF, 1964; BORNHOLD and PILKEY (1971); HUANG and PIERCE, 1971). Most of these occurrences are in the basins contained within the Baham Islands Complex; i.e. Tongue of the Ocean, Exume Sound, Columbus, and BLAKE Basins. Recent sediments in many respects resembling the calciturbidites from the Gürün region have been described by DAVIES, (1968) from the

Abyssal Gulf of Mexico. In these deep-sea oozes, 2-120cm. thick, light coloured carbonate beds occur. They are graded, show erosional base, often contain micritic pellet at basal layers, are only bioturbated in the uppermost part, and are separated from each other by pelagic oozes. The thickness and composition of these beds is variable and they contain dominantly benthonic shallow water remains mixed with few pelagic components.

In addition to the compositional distinction, the calciturbidites of the Gürün region show some specific characteristics that distinguish them from the oceanic abyssal plain turbidites. Firstly, they are less well sorted and show less grading. This characteristic seems to be related to:

(a) the small size of the basin, which limits the distance over which sorting can occur;

(b) The variety of biogenic debris whose hydraulic behaviours are independent upon shape and effective density in addition to size. Secondly, they are coarser, especially in the lower layer of the sequence. This coarseness seems to be related to the supply from the bank where mechanical destruction of carbonate rock is most common. This process does not produce very much clay-sized particles.

The bedding characteristics and general characteristics of the Gürün calciturbidites can be seen in Plates 16 - 25 and will be described below.

B. DISTRIBUTION: Calciturbidites cover large areas in the SW and central parts of the investigated area (See 'Geologic Map'). Because of the lack of continuous outcrops in the region field studies were confined to three main different sections within the study area. Two of these sections have been measured in the SW of the area. Deveçayırı Section (Section No.5) is located in the north and Aptalpınar Section (Section No.6 in the south of this outcrop (See "Location Map"). The third section characterizes the central part and has been measured along the Keşlik Dere Stream (Section No.10B). Besides these three measured sections some random samples have been collected along the Gürün-Kayseri highway in the vicinity of Kaynarca (T29). The three main sections total approximately 1250 m. in thickness and range from 250 m. to 600 m. each. General uniformity of lithology, lack of any characteristic marker beds, and strong multiphased deformation makes correlation and thickness estimates almost impossible. But from the geologic map and field observations it can be evaluated that a thickness of well over 1000 m. is a reasonable minimum value for these calciturbidite beds (Including the Palaeosen (?) part).

C. ROCK TYPES: Three main lithologic groupings can be recognized in the calciturbidites and almost all sequence was formed by repitition of these three main groups.

(1) A white to very light grey coarse calcarenite: This unit is typical example of Lithofacies L (Microbreccia or bioclastic-lithoclastic packstone) and is characterized by an abundant and diverse shallow water fauna which includes large fragments of corals, algae, stromatoporoids, thick-shelled brachiopods, bryozoan colonies and other organic fragments. Subordinate amounts of feldspar, quartz, and volcanic grains have been noted. Skeletal fragements commonly have micritic envelopes which suggest water depth no greater than 60 m (BATHURST, 1967; WINLAND, 1968) for the original environment of these fragments. Most of the feldspars are plagioclase. Quartz, feldspar, and volacanic fragments show a variety of rounding stages, but are mainly angular to subangular. Other grains are well rounded. Sorting is moderate to poor. No quantitative grain size analysis were made of these rocks largely because of the irregular grain shapes of much of the sediment made grain-size determinations from thin sections essentially meaningless; and all the sediments were too well cemented for disaggregation. Grain size estimates have been based on the measurements of representative grains in the thin sections and/or peels. Some samples show excellent preferred orientation of grains. A large number of grains are of gravel size with finer matrix supporting fabric and can be grouped as Standard Microfacies Type 5 (Bioclastic grainstone-packstone).

(2) A yellowish-white to grey, fine-grained calcarenite-Coarse-grained calcilutite with reduced shallow water fauna admixed with planktonics: Lime mud with pelagic foraminfera, calcitized sponge spicules, molluscs, echinoderms, brachiopod and crinoid fragments form a major part of the rock. This unit is a variety of Lithofacies M and is a typical Alpine basin-slope microfacies (WILSON, 1975; p.264). A characteristic feature of these beds is the presence of rounded to angular lithoclasts up to 2 cm. in diameter. The boundaries of the clasts are generally perfectly sharp, mostly encrusted with ferrogineous minerals and commonly truncate fossils (Pl. 15). Stylolites may or may not form

parts of these surfaces. Chambers of foraminifera, the early whorls of gastropods and borings in skeletal fragments are commonly filled with ferrogineous material. Skeletal fragments may also have a light external coating. This may be due to pH and Eh conditions in the bottom muds, influenced by decomposing organic material and the oxygen consumption of burrowing organisms which cause migration of iron-salts to the surface where oxydizing conditions prevail (BANDEL, 1974). Bioturbation has usually destroyed most of the internal stratification in these beds. Another variety of this group is Microbioclastic Packstone/Grainstone (Lithofacies N). The sediment is composed of off bank debris partly indigenous and partly drifted of a bank. Currents which deposited the grains also winnowed the lime mud. Pelletoids are common.

(3) Vari-coloured argillaceous limestones characterized by a planctonic fauna: This unit is typical Lithofacies M (Pelagic lime mudstone). The colour is variable (grey to light red, pink and reddish brown). FABRICUS (1961) explained the difference on colour between red and grey in pelagic limestones as an effect of different sedimentation rates. Grey colour produced under high rates of sedimentation, where complete decomposition of organic matter could not take place. The upper parts of the pelagic carbonates, especially of red ones, were exposed to sea water until emplacement of a new bed by turbidity currents. During these very long time intervals 'hard grounds' were formed due to lithification, solution and boring. This early diagenetic cementation in the sense of ZANKL (1969) created a bottom sediment with more than one indurated layer, whose upper level consisted of more or less loose pebbles. These were picked up by the following turbidity current and incorporated into the lower part of the newly deposited bed.

D. GENERAL CHARACTERISTICS: Most bedding contacts are sharp and easy to recognise. The coarsest unit is the thickest as well. The average thickness of this unit is in the range of 40-50 cm. Some beds more than 1 m. thick have been observed. Grading in the lower parts of this coarse unit is well defined.

Fine calcarenite-coarse calcilutite beds average 20-30 cm. and never exceed 70 cm. Grading in this finer grained bed is not obvious in the field and often was impossible to discern. However, thin sections of these rocks showed some slight grading to be present.

The pelagic carbonates is the thinnest unit with 10-20 cm. average bed thickness.

Data on the lateral extent of beds are scarce due to poor lateral exposure and complex structure. However, individual beds of constant thickness can be traced laterally for distances over 200 m.

Directional sole markings are extremely scarce in the Gürün calciturbidites. The lack of these features seems to imply either low velocity, low turbulence currents or a type of protective "traction carpet" such as postulated by DZULYNSKI and SANDERS (1962). Organic markings are present on the bases of some beds.

Bioturbation is conspicuous on the upper surface of turbidite beds. Burrows often protrude downwards from the overlying pelagic sediments. into the underlying calciturbidite beds to a level of about 15 cm. As SEILACHER (1962) has pointed out the presence of burrowing down from the top surface, coupled with total lack of burrowing in any deeper parts of the turbidites is strong evidence for rapid "single event" deposition of the unit, presumably by turbidity currents. During deposition no burrowing organism was active, but after cessation of the turbidity current deposition, the new upper surface was populated by lateral migration.

Ripple-drift lamination is widespread in the lower part of Section No.9 (Pl. 17). Some ripples have been observed in the vicinity of Konakpınar (R60) (Pl.16I).

Convolute lamination, with amplitudes of a few cm. and wave length of 15-20 cm. are common. They are found at the level above graded and lower parallel laminated units at which sand-sized material had largely dropped out and a very silty calcareous mudstone was deposited. This zone is thought to be one of which current velocities were high enough and sediments cohesive enough to allow the formation of synsedimentary conditions (SCHOLLE 1971).

The most ubiquitous sedimentary structure is horizontal lamination which is found in both calciturbidites and pelagic beds. Horizontal lamination in the B zone of ideal sequence (See below) is defined by an alternation of light and dark coloured laminae. The upper interval of parallel lamination (Division D) is defined largely by a predominant particle orientation with long axes parallel to bedding and slight differences in grain size. As noted above, this lamination is often much more apparent in thin-section than on the outcrop.

The calciturbidite beds may display partial to complete BOUMA (1962) Cycles (See p.106). HARMS and FAHENSTOCK, 1965; SIMONS, RICHARDSON and NORDIN, 1965) have related Bouma intervals to the standard flow regime sequence, showing a decrease in current velocity from high flow regime at the base, to low flow regime rippling in internal C and deposition of pelitic material from nearly stegnant suspension in interval E (Fig.IV-8).

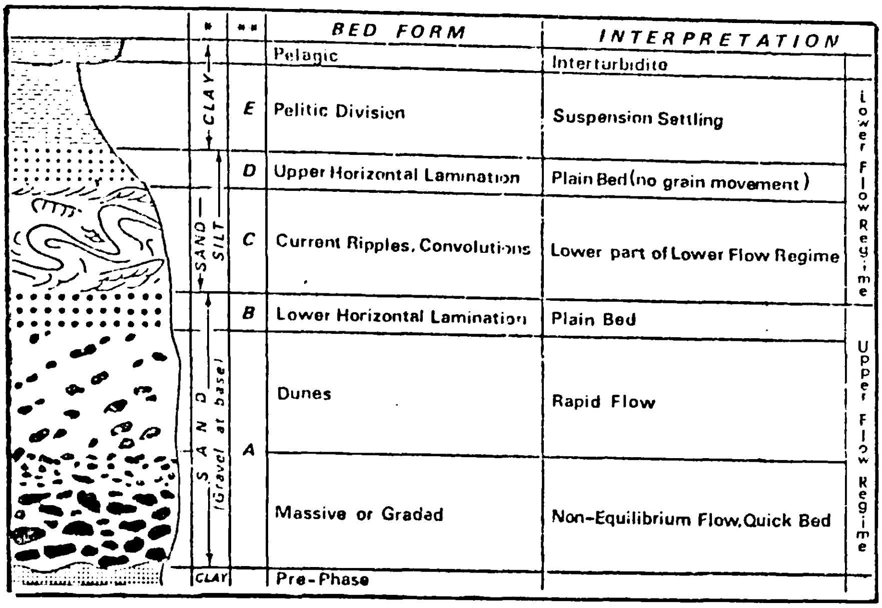

Fig.IV-8 : Ideal sequnce of structures in a calciturbidite bed (After BOUMA, 1962 with modifications from MEISCHNER, 1964; THOMSON and THOMASSON, 1969; and interptetations after HARMS and FAHNESTOCK, 1965; SIMONS, RICHARDSON, and NORDIN, 1965).

The most important modification of the Bouma model for carbonate rocks is the adding of a "pre-phase" to the base of the main graded bed (MEISCHNER, 1964; p.159). It contains very fine-grained, transported skeletal material. This lithology and fauna is quite different from the pelagic material beneath. The other modification of the Bouma Sequence concerns the discovery of a "dune" phase in carbonate turbidites (HUBERT, 1966; THOMSON and THOMASSON, 1969). HUBERT argues convincingly for the inclusion of this bed form in the ideal sequence. THOMSON and THOMASSON (1969; p.72-75) record dunes only in proximal turbidites. This finding has also been observed in the present study. The sequence they record is A-Dune-B-C-D in other words the dunes appear more often below the B laminations than above.

E. TURBIDITY CURRENT ORIGIN: There can be no doubt that these beds owe their origin to deposition from turbidity currents. Turbidity currents are accepted as the specific mechanism of transportation because they are the only type of current which will provide for extensive transport of very large amounts of sediment discontinuously with subsequent rapid deposition. The other evidence for turbidity current origin can be summarized as follows:

1. Each unit is characterized by a vertical size and compositional grading (Upward fining).
2. The units show a large variety of sedimentary structures characteristic of turbidites (e.g. sole marks, ripple drift cross lamination, convolution and slumping features).
3. Flysch-like rhythmic interbedding.
4. Only the uppermost horizons of the carbonates are bioturbated indicating more rapid deposition of the lower member.
5. Absence of sedimentary structures associated with shallow water environments.
6. Associated pelagic calcilutites above and below.
7. Contrast between benthonic shallow marine fauna of coarse grained carbonates and pelagic fauna of the calcilutites.

Turbidity currents probably initiated by a combination of a steep shelf edge and rapid carbonate sedimentation on the shelf and caused the transfer of shallow water carbonates to the adjacent slope/basin environments. Despite the lack of the directional evidence, the carbonate sourse area is undoubtedly the nearby Gürün shelf. The dense flow associated with the lateral transfer of shallow water carbonates from the Gürün shelf resulted in the scouring of the slope-basin sediment surface. With some decrease in the turbidity current velocity the white to very light grey, coarse-grained calcarenites and calcirudites were rapidly deposited. Current velocities were high enough to completely disrupt normal pelagic sedimentation. As current lost their velocities even further, the finer carbonates settled slowly out of suspension and were mixed with pelagic calcilutites. With a cessation of turbidity current, the deposition of pelagic calcilutites represent a return to the characteristic pattern of slope/basin sedimentation. Deposition from successive turbidity currents following one another in rapid succession complicated the vertical succession.

Detailed lithological examination of the measured sections demonstrates that each of these sections are similar in principal, consisting of a vertical repetition of coarse-grained calcarenite, fine-grained calcarenite - coarse grained calcilutite, and pelagic carbonate mudstone. On the other hand section no 10B consists mostly of light grey-white calcarenites while in Section Nos. 5 and 9 the amount of this lithology diminishes and fine-grained calcarenite-coarse grained calcilutite becomes dominant.

The lower part of the Section No.10B shows A and B intervals of the ideal sequence and can be classified as "Proximal Turbidites" (WALKER, 1967) whereas upper parts of this section and almost entire of Section Nos. 5 and 9 have been characterized by alternation of bed forms within the lower flow regime (C,D, and E intervals). They are typical of sediment deposited in the distal environments (WALKER, 1967). Section No.10B shows a transition from proximal to distal environments and Section Nos. 5 and 9 represent distal calciturbidites (Fig. IV - 9). However, from time to time currents were still strong enough to pick up and transport solution clasts and skeletal fragments from the proximal environments and incorporate them into the distal deposits. As pointed out by SCHOLLE (1971) it may become necessary to distinguish between "true" distal turbidites and turbidites derived from very fine-grained source area which show most distal characteristics but have considerable bed thickness. Fine grain size, regular and parallel sided bedding, lack of channels, well developed calcilutite layers between calcarenites, grading and sharp base are indicative of distal origin of these beds.

The occurrence of at least two genetically different kinds of deposites in tur bidite basins, namely turbidites and pelagic layers, which represent entirely different mechanisms of sediment accumulation (e.g. rapid, bottom-following more or less horizontal influx of sediment versus slow vertical settling) has been recognized long since, but criteria for differentiating between the two in ancient turbidite sequences have only recently been compiled (HESSE, 1975). They include Calcium Carbonate content, colour, sequential order of rock types in the vertical succession, distribution of bioturbation and microfossil content.

Theoretically, four principal basin types of turbidite deposition may be distinguished. The distinction is based on a simple criteria, namely the calcium carbonate content of the pelagic mudstone interbedded with the turbidites. Calcium carbonate free pelagic mudstone is interpreted to indicate deposition below the "Calcite Compensation Depth"(*), Carbonate-bearing pelagic mudstone to indicate deposition above the calcite

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(*) In parts of the deep oceans, dissolution of calcium carbonate removes the excess supply of biogenic carbonate particles from surface production. The depth where dissolution first balances supply is known as "Calcite Compensation Depth" below which pelagic sediments are free of calcium carbonate. The actual depth of the compensation level varies considerably in space and time depending on local and temporary factors controlling dissolution and supply rates, (BERGER & WINTERER, 1974). In present day oceans this level lies for the most part, at about 4500-5000 m. On the other hand, HESSE and BUTT (1976) proposed that calcite compensation depth for the Late Cretaceous was around 3500 m. They also postulated that it probably had a similar depth range as elsewhere in the low and mid-latitudes of the Late Cretaceous (Ibid. p.529).

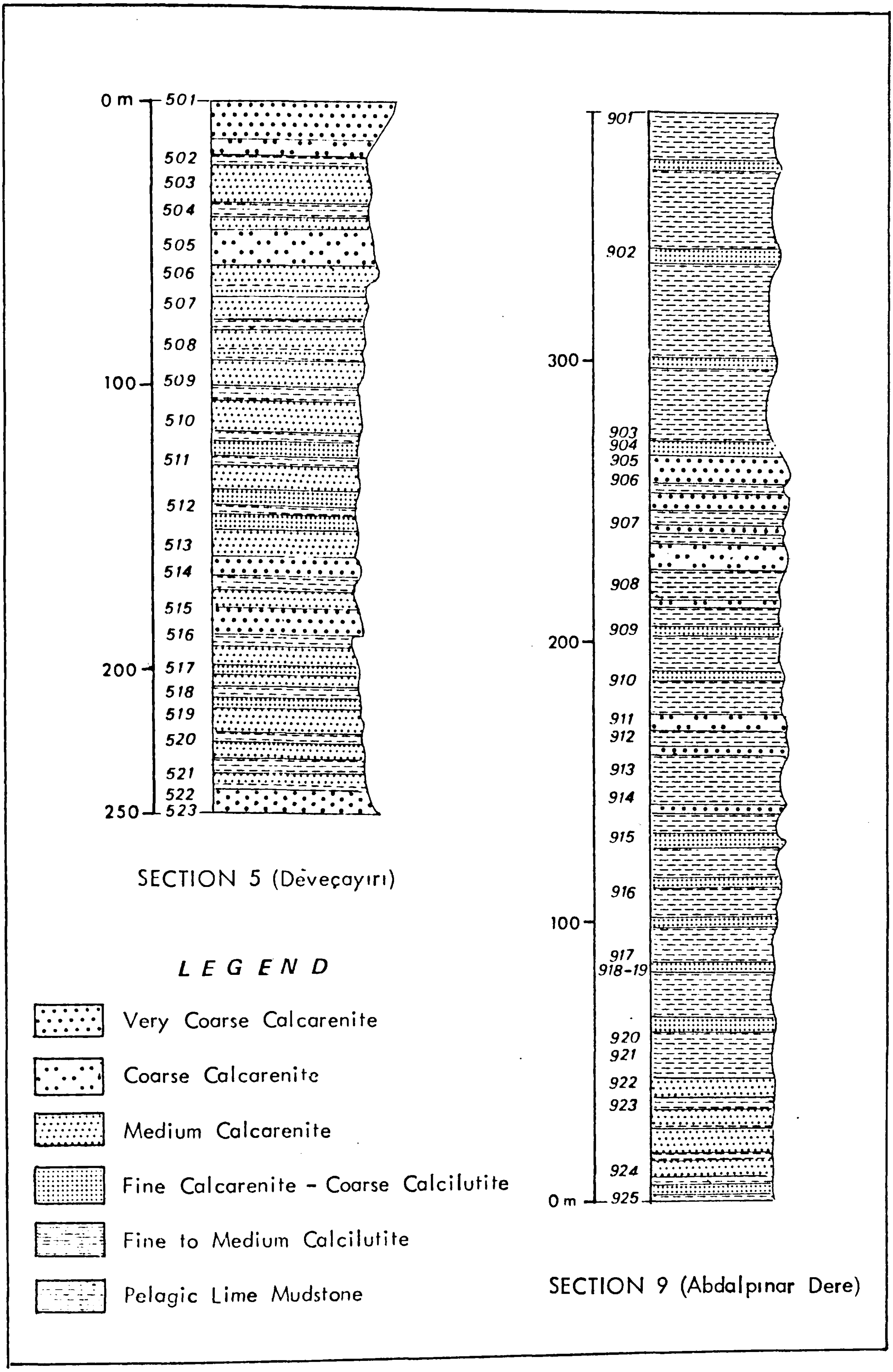


Fig.IV-9 A : Measured sections from calciturbidites

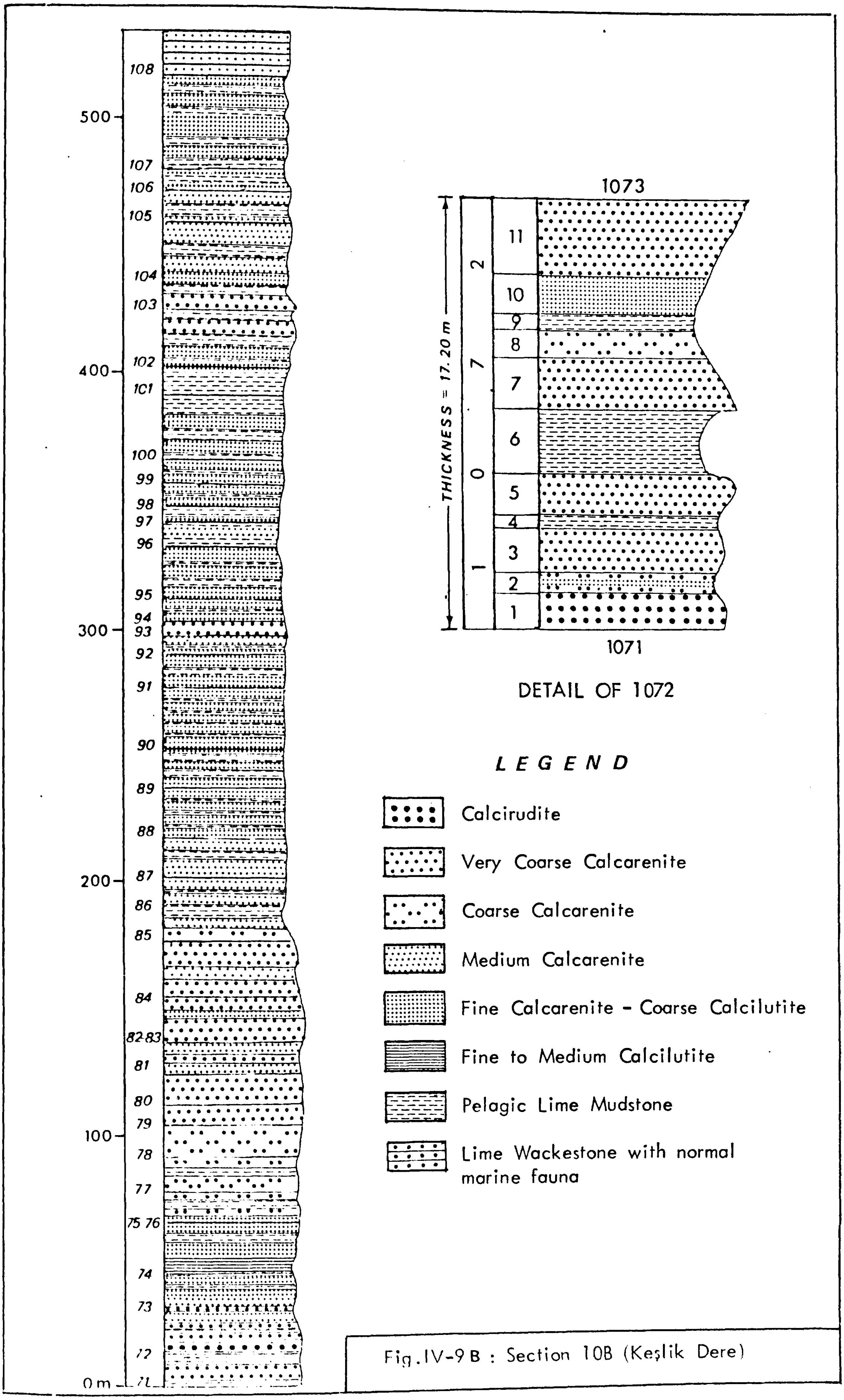


Fig.IV-9 B : Section 10B (Keşlik Dere)

compensation depth. However the deposition of carbonate-free pelagic mudstone may be caused by the influx of terrigenous sediment or climatic changes and that they might occur at shallow depth. The four turbidite basin types are:

(1) Carbonate turbidite deposition below the calcite compensation depth;

(2) Carbonate turbidite deposition above the calcite compensation depth;

(3) Terrigenous (carbonate-poor or carbonate-free) turbidite deposition above the calcite compensation depth;

(4) Terrigenous (carbonate poor or carbonate free) turbidite deposition below the calcite compensation depth. From field and petrographic observations it is certain that deposition of the Gürün calciturbidites took place above the calcite compensation depth. Carbonate-free pelagic layers encountered in the vicinity of Kaynarca (T29) reflects the terrigenous influx and not deposition below calcite compensation depth because they are interbedded with pelagic carbonate mudstones (Pl.24).

F. RECONSTRUCTION OF THE DEPOSITIONAL ENVIRONMENT: Gürün calciturbidites represent slopes and basin facies of the Maestrichtian time in the region. The chaotic nature of tectonics, and the uncertain lateral correlations prevent the reconstruction of a regional picture. Close to shallow-water carbonate platform,periodic turbidity currents, originating on gently basinward sloping carbonate shelf, flowed along the bottom into the basin and deposited graded limestones. Some of these limestones were deposited on the slope under conditions of high energy (proximal turbidites). The others travelled further and resulted ultimately in the well developed graded limestones of the basin facies (Distal turbidites). Calciturbidites are intercalated with pelagic deposition which indicates gradual diminishing of the current velocities. Detailed examination reveals a source area in the Northeast. Two sections measured in the SW of the area are represented by distal turbidites while the one in the centre (Keşlik Dere Section-No.10B)is characterized by coarse-grained calcarenite and calcirudites indicating its proximality. However, in the upper parts of the same section predominant lithology is fine-grained calcarenites and calsisiltites. The topmost part of the section is represented by bioclastic and whole fossil wackestones characteristic of open marine neritic facies. From these observations it can be argued that the Upper Cretaceous transgression continued during the Maestrichtian and even the Paleosen times (evident by some Paleosen microfossils encountered in the continuation of the Abdalpınar Section (Section No.9). (See CANİK, 1964); and the depth of the sea increased with time. These conditions probably prevailed in the region till the closure of Alpine Tethys during the Late Cretaceous and Early Tertiary as suggested by several authors (e.g. ERNST, 1973; HSÜ and SCHLANGER, 1971; OXBURGH, 1972).

G. DIAGENESIS OF THE CALCITURBIDITES: Compaction and silicification seems to be the dominant diagenetic process operating in the calciturbidites of the Gürün region.

The relatively stable original mineralogy and the absence of an early fresh water flush in the Gürün calciturbidites allowed considerable compaction to occur before cementation, yielding a characteristic close-packed texture with relatively small volumes of cement. Evidence for extensive pre-cementational compaction is widespread. The remarkably close packing of most of the sediment indicates extensive compaction. There is considerable grain interpenetration, pressure solution and breakage of fossils (Pl. 38).

Silicification is another common diagenetic feature in the calciturbidites of the Gürün Basin. The subject has been discussed in detail under the heading of "Silicification" (p.203). and will be only briefly summarized here: Silicification has been observed both in large scale, as cherts; and small scale, as void filling cement and replacement of grains and/or matrix. Cherts have been found forming nodules, lenticular bodies and beds. Two types of chert have been differentiated: Granular chert which is usually associated with white to light grey coarse calcarenites (i.e. packstone and/or grainstone facies) and vitreous chert which is common in mudstone and wackestones of distal calciturbidites. Field and petrographic evidence strongly suggests a replacement origin for cherts. The probably sourse of the chert is the dissolution of siliceous organisms. Silicification is a late diagenetic phenomena in the Gürün calciturbidites.

The calcite cement encountered in the calciturbidite samples consists of fine - to medium crystalline anhedral to subhedral equant crystals. No bladed or fibrous textures are present.

The recrystallization of certain zones to microspar and pseudospar

(FOLK, 1965) is the most apparent neomorphic process. Although not so much evident, the conversion of carbonate mud to lithified micrite was no doubt the most important aspect of the neomorphism.

Dolomitization is very scarce and has been observed only in one sample from the Kaynarca region (T29).

All the samples including the pelagic lime mudstone are rich in ferrous iron content. The distribution of ferrous iron in carbonates is not normally observalbe under the petrographic microscope and it has bean revealed by staining with potassium ferricyanide (See p.16). which gives dark blue to purple colour indicating high ferrous iron content. EVAMY (1969) who worked on the precipitational environments of the calcite cements concludes that ferroan carbonates represent precipitation in reducing environments. This is mainly below the water table, where organic matter undergoes decomposition. The conditions which existed during the sedimentation of the calciturbidites no doubt were very favourable for enrichment of sediments in this respect to ferrous iron. Ferromanganese nodules as described by JENKYNS (1970) from Jurassic beds and Ferromanganese crusts as described by TUCKER (1973) from Devonian of South France and Germany and BANDEL (1974) from Devonian-Carboniferous of Austria has not been encountered in the present study.

V. GEOLOGY OF THE EAST PART OF THE INVESTIGATED AREA.

In the east of Dürmepınar (S69) - Dayakpinar (I68) line the predominant lithology is variegated conglomerate, sandstone, siltstone, and mudstone. They are almost completely composed of grains of green rock (Ophiolitic Series). Even the matrix is ophiolitic in origin. Two distinct facies types have been observed. First one is mostly dominated by conglomerates and sandstones and outcrops in the vicinity of Hamalçay (H76). Conglomerates contain a compositionally heterogenous mixture of volcanic, chert-like glassy volcanic, plutonic and metamorphic pebbles, cobbles, or boulders; all contained in a muddy or sandy matrix. Massive cobbles and boulder conglomerates with clasts up to 60 cm. have been observed. No detailed study has been undertaken in this region but under the light of some observations made at outcrops of these facies throughout NW of Hamalçay, it may be postulated that the clasts were derived originally from land and were rounded by river and/or wave activity. Some similar rocks have been observed along the Tohma Çayı around Tökler (N75), Sarıca (N77), Sofular (I80), and Bıcır (G81) villages. The lithology is very similar in the first instance, but a close examination reveals that they are turbidites. Analogous deposits

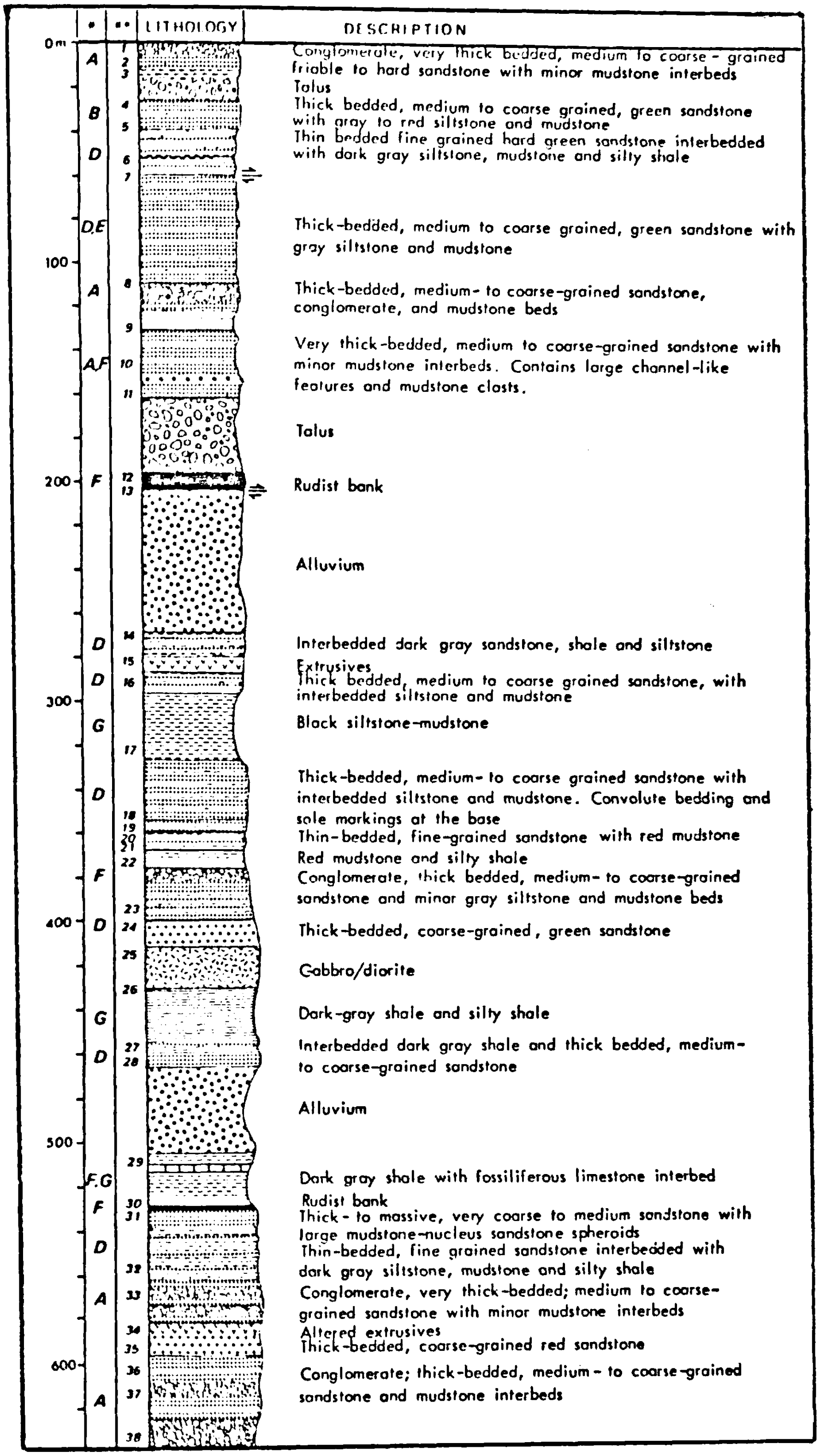


Fig.IV-10: Section: SARICA
(*) RICCI-LUCCHI (1975) facies
(**) Sample Number

have been reported from different localities outside the investigation area (See: BAYKAL, 1966; AKKUŞ, 1971). This turbidites were sampled at a valley in the SE of Sarıça (N77) (See "Location Map", p. 14), Fig IV - 10. Plates 28 to 35 show field appearances and features of the Gürün turbidites.

A. CHARACTERISTICS OF SARICATURBIDITES:

(1) Outstandingly characteristic of these beds is a rapid alternation of fine - and coarse - grained strata (Pl. 28)
(2) Generally the rock types involved are mudstone and sandstone (Pl. 28). Colour differences between the dark grey to red mudstones and green sandstones produce an eye-catching bedding contact in fresh exposures and the unequal weathering properties of the two rock types is spectacular.
(3) The finer-grained sediments are commonly laminated shale, but structureless mudstone is not uncommon and siltstone also may represent this interval (Pl. 31).
(4) The coarse clastic members generally are fine-grained sandstone, but this phase may be represented by sediments as coarse as conglomerates (Pl.28A). They are generally thin bedded, with a sharp contact at the base of each sandstone layer and an even but less conspicuous contact with the overlying fine-grained sediment (Pl. 28).
(5) The lower surfaces of some sandstone beds have sole marks which protrude downward into the underlying fine sediment.
(6) Internally the sandstone beds may show graded bedding at the base, become laminated in a higher interval, show convolution in a yet higher interval and be followed by other divisions showing parallel lamination (BOUMA Divisions).(*) (Pls.32-33) One or more of the division may be missing or may not have been developed.

(*) KUENEN (1953) seemingly developed the concept that there is a fixed characteristic vertical succession of layers in a turbidite and that each layer (ideally) will have a fixed number of sedimentary structures. BOUMA (1962, p.49) worked out a vertical sequence of five divisions that ideally occur in the turbidite sandstones of the French Maritime Alps. (Fig. IV - 11). Named in an upward order they are: (a) a graded division best developed where the sandstones are poorly sorted, (b) a division of parallel lamination which contains some clay particles along with the sand (c) a division characterized by current ripple laminations, some of which are convulute, (d) an upper division of parallel laminations, and (e) a pelitic division with no sedimentary structures. The complete sequence was designated T_{a-e}.

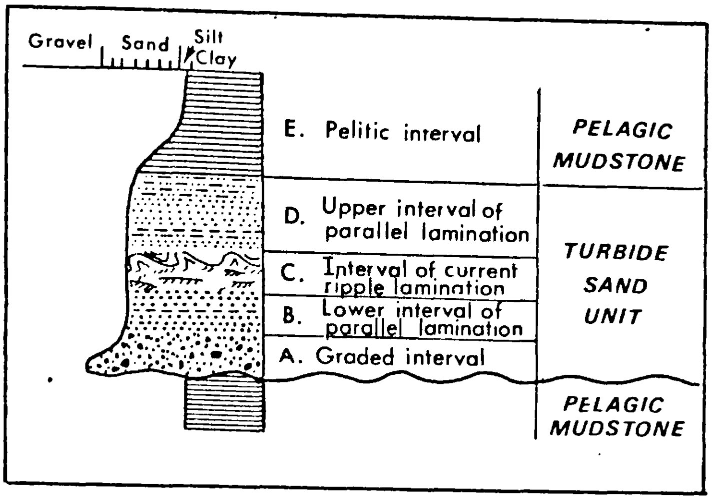


Fig.IV-11 : Turbidite unit showing the complete BOUMA (1962) sequence.

(7) Large scale cross-bedding is virtually absent and so are some other sedimentary structures associated with shallow water environments.

(8) Fossils are scarce in most sandstone beds and those that are present are likely to be fragmental and displaced from shallow water environments.

The most compelling evidence that the sandstones are turbidites are:

(a) They are interbedded with mudstone that is best interpreted as a deep water deposit; (Pl. 75)

and (b) Internal structures suggest that they were deposited by weakening currents that initially had high velocity. In addition, the sandstones have sole marks, graded bedding and convolute bedding that are typical features of turbidites (Pls.32-33)

The extent to which individual sandstone beds have lateral continuity has not been determined in the present account simply because there are not enough exposures and there was not the slightest possibility that enough boreholds would be put down to allow the point to be clarified. In the valley exposures that were available for study, generally only less than 100 metres long individual sandstone beds persists from one end of the exposure to the other, but some sandstones have been observed to pinchout or terminate abruptly. However, from what is known about the lateral continuity of individual turbidite beds from all over the world one would expect some of the Sarıca turbidite beds to have considerable lateral extent.

There is a strong suggestion of cyclicity in strata but until statistical work is done it is impossible to know whether the bedding is cylic or megarhytmic.

The mineralogical composition of the sandstones has been determined partly by author and partly by Dr. Q. JAN, and include feldspar amphibole, pyroxene, chlorite, ores, epidote, quartz, hornblend. The source is partly extrusive and partly intrusives (gabbro/diorite) and most of the sandstones can be classified as "Lithic wackes". (Pl. 33).

B. GENERAL FEATURES OF TURBIDITES: Two end-type may be defined, namely, proximal and distal. These terms are used to indicate their presumed proximity to the source from which they were derived.

DZULYNSKI, KSIAZKIEWICZ, and KUENEN (1959) recognized rocks of intermediate origin between shelf and basin, and referred them as 'fluxoturbidites'. They stated that "It is transitional between the turbidites and the pure slides, because it shows a mixture of features characteristic of the two mechanism of transport... (p.1095). Unfortunately too little is known about the transport mechanicsm involved and understandably, few geologists have been satisfied with a genetic term such as fluxoturbidite. WALKER (1967, p.38) proposed that the term should be abandoned because it has been applied to a bed, to a facies, and to a process. Subsequent writers have referred to rocks of similar facies as "proximal trubidites" and this usage has been adopted in the present account.

The proximal facies is characterized by regularly bedded sandstones ranging from 10 c.m to 1 metre or more in thickness with fairly frequent phenomena of intense erosional channelling. (Pl.30B). The thickness of beds are not always constant; in some cases they are clearly lenticular. The clastics are always very coarse, even pebbly in some cases. The beds are generally graded and frequently show a truncated sequence (Interval A of BOUMA). Laminated and rippled intervals (B and C) are rare. They are in BOUMA terms AE turbidites.

In the distal facies the sands are finer-grained, and beds range from 1 to 10 cm in thickness. They generally begin with the laminated interval (B) or even the rippled phase (C). (Pl. 31).

On the other hand recent work on modern and ancient turbidite basins (e.g. MUTTI and RICCI LUCCHI, 1972; STANLEY and UNRUG, 1972; NELSON and KULM, 1973; NELSON and NILSEN, 1974; MUTTI, 1977) has established that the thin-bedded turbidites are present in a variety of deep-marine depositional environments and are not simply confined to the basin plain or depositional sites located most distant from the source. MUTTI (1977) has provided some general criteria for distinguishing thin-bedded turbidites deposited in different environments.

The following physiographic elements form the framework for dispersion and deposition in many turbidite basins. A near-shore point source of clastics, a submarine canyon or other type of channel funneling them down a slope; a deep-sea fan and a basin plain (Fig. IV - 12). Two zones can be differentiated in the fan: The inner fan is characterized by large channels (Deeper than 2m). The outer fan lacks channels. The sediments of channelized portion of the fan consist of coarse-grained and thick bedded channel-fill sandstone bodies that are enclosed by thin-bedded sandstone, siltstone and mudstone. Inner fan has been further divided by MUTTI (1977) (Fig IV - 13). The outer fan consists of thick-bedded and coarse-grained non-channelized bodies of sandstones that form distinct depositional lobes (MUTTI 1977). The lobes are enclosed by thin-bedded and finer grained lobe fringe deposits (Fig IV - 12). Basin plain consists of fine-grained and generally thin-bedded turbidites, with thin - to very thin interbeds of hemipelagic mudstone. Vertical association of turbidites and sediment flow deposits in channel fills, outer fan depositional lobes and basin plain are shown in Fig. IV - 14. Facies types shown on this figure have been described by RICCI LUCCHI (1975). The thicker and coarser beds are represented by facies A, B, C and F in RICCI-LUCCHI scheme. Similar facies are found in modern deep sea channels (NELSON and KULM, 1973). BOURCART (1964) described interbedded coarse gravel and pelagic mud found at depths as great as 2410 metres on the Mediterannean abyssal plain off the French Riviera. Gravel is distributed from river mouths down submarine canyons onto the abyssal plain. This lithofacies occurs in the both upper and lower parts of the Sarica Section (Pl.28A). It stratigraphically takes place between shallow marine and basin deposits. The rudist bank encountered in the 200th metres of the section is believed to be a slump deposit (Pl.29C). In detail the sandstone beds are usually thick. Signs of amalgamation and composite beds have been seen (Pl.32B). These thick-bedded sandstones have an extremely thick A-B interval (BOUMA 1962, Graded Division). Some directional sole marks have observed on the lower bedding surfaces (Pl.32A). They usually show upward fining of material. This lithofacies is interpreted to be

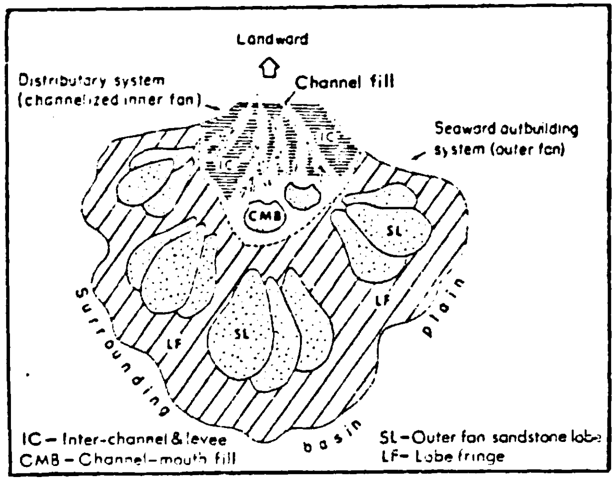


Fig.IV-12: Physiographic model for dispersal of sediment gravity flows. (After MUTTI, 1977)

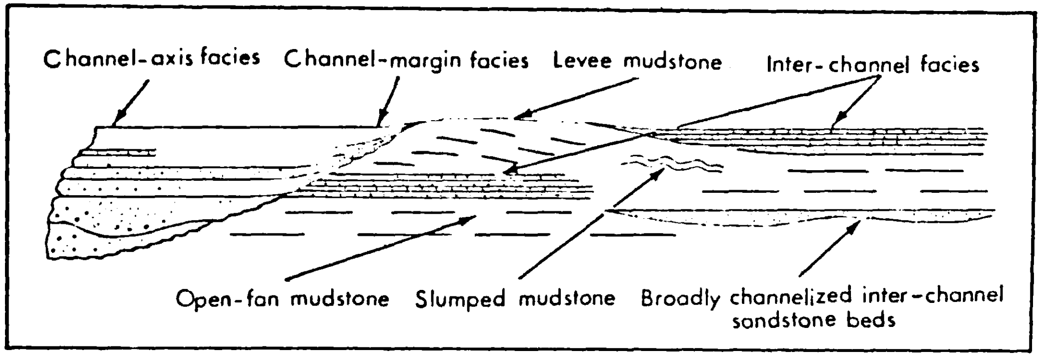
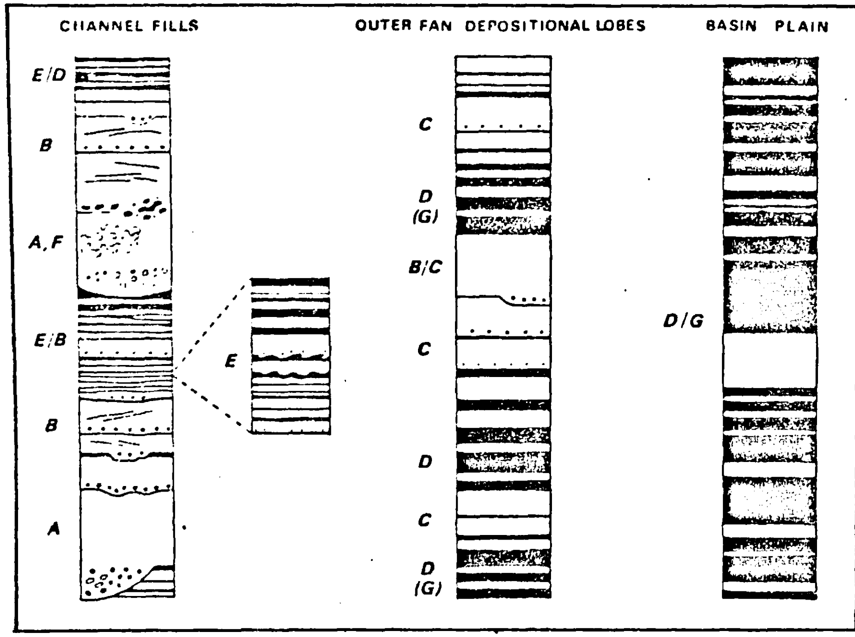


Fig.IV-13: Relationships and terminology of channel and interchannel turbidite facies as observed in the inner fan deposits (After MUTTI, 1977)



TURBIDITE FACIES :

A. Organized conglomerate, pebbly sandstone, very coarse to medium sandstone; finer materials subordinate to absent; variable matrix content; poor sorting; thick to massive bedded; sand/lutite ratio very high to indefinite

B. Coarse to medium-fine sandstone, better sorted than in A; more frequent and thicker lutite partings; thick to massive bedded; sand/lutite ratio high to very high.

C. Sandstone/mudstone interbeddings; coarse to fine sand (immature variety); medium to fine sand (mature variety); sorting poor to good; medium to massive bedded; sand/lutite ratio high to very high.

D. Thin to massive bedded sandstone/mudstone interbeddings; fine to very fine sand, moderately to well sorted; thin to medium bedded lutite (mudstone/shale); sand/lutite ratio low or very low.

E. Sandstone/mudstone interbeddings; coarse to very fine sand, sorting variable; thin to medium bedded; sand/lutite ratio from slightly<1 to very high.

ASSOCIATED FACIES:

F. Chaotic deposits: disorganized conglomerates, pebbly mudstones, debris flow debris flow deposits with extrasize clasts, isolated blocks (olistoliths), slumps, etc.

G. Hemipelagic lutite (more or less sandy marl, siltstone, clay); thin to medium bedded.

Fig.IV-14: Schematic columns showing vertical association of turbidites in channel fill, outer fan depositional lobes and basin plain and associated facies (Modified from RICCI-LUCCHI, 1975) deposited on the slope and the upper part of basin (Channelized inner fan).

Finer and thinner turbidites are referred to as facies D and E. As noted earlier MUTTI (1977) has shown that they may occur in different deep-sea environments such as channel margins, interchannel areas, channel mouth, lobe fringes and also in the basin plain (Fig. IV - 12,13). In detail, the thin bedded and graded sandstone beds display vertical sequences of internal structures (Bouma Sequences) showing upward fining of material (Pl.33).

C. DEPOSITIONAL MODEL FOR THE EAST PART OF THE INVESTIGATED AREA: The general depositional model of the east of Dürmepınar (S69) - Dayakpınar (I68) line has been inferred from the observed facies relationships and refers to the Senonian, probably Maestrichtian time. No attempt to restore the present setting to its palinspastic configuration has been made. The vertical sequence of sedimentary features, facies and distribution patterns suggest that the sediments of this region were described as a progradational deltaic complex to the south that fed, via channels cut on a delta slope, a deeper turbidite basin located to the north. (Fig. IV - 15).

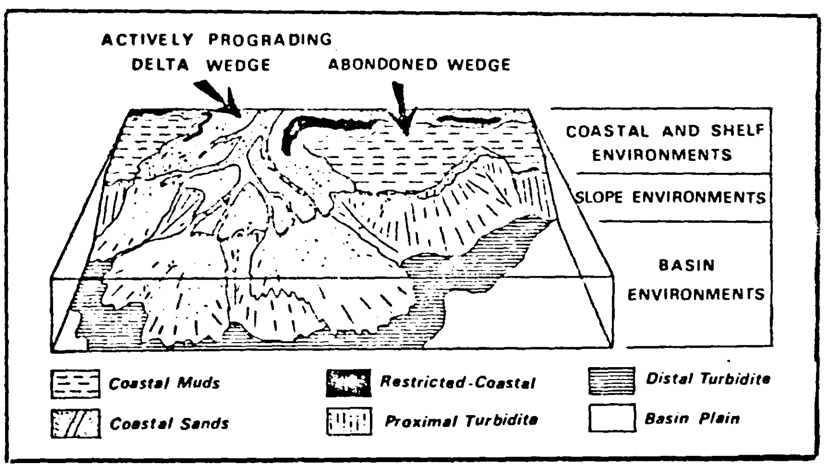


Fig.IV-15 : Depositional model for the eastern part of the investigated area (Modified after FERM, 1970; Fig.3 ; p.250).

River erosion and wave erosion transported late Mesozoic sediments derived from slightly older metasedimentary, volcanic, and ophiolotic source rocks to narrow shelves where further sorting and rounding by surf and sea currents occurred. Here, massive to thick bedded conglomerates with only faint stratification accumulated in a setting probably similar to the east side of Baja California where thick gravels accumulate next to deep water (VAN ANDEL, 1964). Mass flows, turbidity currents and grain flows re-deposited much gravel and sand in relatively quiet trough-bottom environments where pelagic muds normally accumulated. The transition between the turbidite and deltaic intervals is difficult to determine due to complex tectonic history of the region but is thought to be abrupt.

The depositional model presented here associates shallow-marine delta deposition, the submarine fan complex and basin plain deposits in a single explanation of sedimentation. Conglomerates and thick bedded, coarse-grained beds are postulated to be directly related to deltaic deposition and generated by slumping and sliding of sediments toward deeper water as high-density sediment gravity flows on the front and flanks of a delta. Thin-bedded, fine-grained beds are formed by changing of high density sediment gravity flows to low density one as a result of deposition of some of the material.

VI. SUMMARY OF THE DEPOSITIONAL ENVIRONMENT:

Thick sequences of the Upper Cretaceous rocks outcrop in the Gürün region. Detailed study of these rocks indicate the presence of an Early Mesozoic cratonic shelf and a northwards marine transgression. onto this shelf during the Late Cretaceous epoch resulting in the development of a carbonate platform. "Ideal carbonate platform model" proposed by WILSON (1975) has been found applicable to the Gürün carbonates and five different facies belts have been recognized in the investigated area. Tidal flat deposits (Supratidal, intertidal and subtidal) cover large areas and represent Cenomanian-Campanian time interval. Knoll reefs were developed on the outer margins of the platform where rudists played an important role. A high energy zone was present just behind the rudits concentration at the shelf edge. Transgression reached its amplitude after Campanian time and calciturbidite beds associated with pelagic lime mudstones were deposited.

In the eastern part of the investigated area, Senonian, probably Maestrichian time is represented by a progradational deltaic complex to the south that fed, via channels cut on a delta slope, a deeper turbidite basin located to the north.

Based on the petrographic, palaeontologic, textural, and structural criteria 14 microfacies have been recognized. 24 Standard Microfacies types proposed by WILSON (1975) constitute the fundamental of this classification.

Table IV - 1 summarizes facies belts, microfacies types and inferred environments. Figure IV - 7 shows diagrammatic reconstruction of environment of deposition in the Gürün region during the Late Cretaceous epoch.

FACIES BELT	MICROFACIES	INFERRED ENVIRONMENT
TIDAL FLATS	A. Pure lime mudstone	Shallow ponds with high salinity
	B. Algal-mat lime mudstone	Supratidal zone
	C. Lithoclastic packstone/grainstone	Desiccation, wind erosion and subsequent flooding of the dried surfaces
	D. Stromatolite mudstone	Intertidal zone
	E. Lime wackestone with restricted fauna	Restricted subtidal
	F. Miliolid grainstone	Subtidal
	G. Pelletoidal lime grainstone	Tidal flats and natural levees
OPEN MARINE PLATFORM	H. Lime wackestone with normal marine fauna	Quiet water below normal wave base
PLATFORM EDGE CARBONATE SAND BANKS	I. Coated grainstone	Areas of constant wave action
ORGANIC BUILDUP	J. Rudist lime mudstone	Knoll reefs
DEEPER WATER	K. Bioclastic packstone/grainstone	Reef flank
	L. Microbreccia or bioclastic lithoclastic packstone	Basin slope
	M. Pelagic lime mudstone	Toe of slope
	N. Microbioclastic packstone/grainstone	Toe of slope

Table IV-1 : Facies classification in the Gürün region.

The depositional facies of the Gürün region is broadly similar to those of the Golden Lane and Poza Rica Trends in Mexico; but the subsequent geologic histories of the two regions are markedly different. The differences in the post depositional history affect the reservoir characteristics, the type of traps, and the movement of fluids. In eastern Mexico, these features resulted in the development of giant oil reservoirs (See COOGAN, BEBOUT, and MAGGIO, 1972) but no reservoir has been found in the Gürün region so far.

Eastern Mexico was a major carbonate province during much of the Mesozoic. The Albian-Cenomanian is characterised by widespread basinal limestones, and platforms fringed by rudist reefs, where shallow water carbonate deposition persisted. The resultant lithofacies are

(1) a suite of platform carbonates (El Abra Formation),

(2) basin-margin carbonates consisting of platform-derived brecciac and bioclastic rocks, interlayered with pelagic limestones (Tamabra Facies), and

The shelf margin facies of the El Abra consists of coated grain-stones which corresponds to Lithofacies I in the Gürün region. Oolites are more common than at Gürün. Shelf margin rudist mounds are characterised by rudist lime mudstone as in the Gürün (Lithofacies J).

The Golden Lane fields produce from the El Abra Limestone. Movements associated with the Early Tertiary orogeny caused exposure and subaerial leaching producing remarkable porosity in the Golden Lane. The source of the oil is thought by some to be the Tertiary fine-grained clastic rocks, other geologists consider the source of the oil to be Jurassic black shale and limestone, and others believe that the oil might be indigenous to the Albian-Cenomanian limestones. Original intergranular porosity is in the Miliolid grainstone and rudistic limestones. Solution and recrystallisation also developed secondary porosity in limestones that originally had lower porosity.

Coarse lithoclastic-bioclastic facies (Lithofacies K of this study) characterise the Tamabra Limestone. Basinal limestones of the Upper Tamaulipas consist of pelagic lime mudstones and wackestones (Lithofacies M in the Gürün region). Production in the Poza Rica fields is essentially from a stratigraphic trap caused by the facies change from porous bioclastic Tamabra Limestone into dense pelagic micritic Upper Tamaulipas Limestone and overlying non-porous rock of the Upper Cretaceous. Porous interval of the Tamabra Limestone coincides with the presence of coarse and fine bioclastic limestone. Micritic-pelagic limestones generally are not porous. Extensive recrystallisation, dissolution and fracturing have enlarged the intergranular porosity in the bioclastic limestones and developed porosity in the originally non-porous micritic limestones (COOGAN et al., 1972).

Environment of deposition is only one causal force contributing to the origin of a reservoir. Diagenesis, structure and hydrodynamics also have a profound effect. Reservoirs may be produced by causes independent of environments (for similarities and differences between Gürün and Golden Lane and Poza Rica Trend see "Summary and Conclusions", p.225).

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CHAPTER V

DIAGENESIS

PART 1

THE CONCEPT OF DIAGENESIS

Although the term "diagenesis" is more than 100 years old, there is still no generally accepted definition, and many authors have conflicting views as to when it begins and when it ends. Some have used the term in a restricted sense and have limited it to those changes which affect a sediment after deposition and up to, but not beyond lithification. Where as some consider that diagenesis continues post lithification and grade into metamorphism.

"Diagenesis was first introduced by VON GÜMBEL in 1868:

"...With regard to the origin of these various ancient rocks - the gneises may be cited as the chief representatives - it is well known that a variety of opinions exist; they are regarded by some as the earliest crust material and by others as altered schists (metamorphosed formations). However, the most likely interpretation is, by analogy with sediment formation, that they precipitated during the first stage of the earth's formation, from an aqueous envelope under the influence of pressure and heat, and subsequently the amorphous precipitate underwent transformation (diagenesis) into individual crystalline constituents or minerals (from DUNOYER DE SEGONZAC; 1968)."

Diagenesis is thus already defined as a post-sedimentary, non-metamorphic transformation. Later in 1888 he defines it clearly:

"...It can hardly be doubted that chemical processes take place on a large scale on the sea-bed, and it is terms of those processes that we must explain the partial transformation (diagenesis) and the consolidation of the deposited materials by infilling of the voids. The result of these transformations is today seen in the older marine deposits (from DUNOYER DE SEGONZAC; 1968)."

The modern broad meaning of the term is attributed to WALTHER (1894) who wrote:

"...By diagenesis we mean all the physical and chemical changes that a rock undergoes from time of its deposition, other than changes due to the intervention of tectonic compression and volcanic heat (from DUNOYER DE SEGONZAC; 1968)."

It is clearly understood that "rock" was used in the traditional geologic sense, which in this case implies an initially soft, unconsolidated sediment.

The concept of diagenesis is not easily accepted by English speaking authors. TWENHOFEL's (1926) work is the first American reference to diagenesis in the European meaning of the word (DUNOYER DE SEGONZAC; 1968).

"...Between deposition and lithification a great variety of changes may take place. Most of these are included under the term of diagenesis which as here used is intended to embrace those chemical and molecular changes which take place in a sediment prior to solidification while it is still in the environment of deposition or is acted upon by conditions and processes inherited from that environment, but excludes those katamorphic changes which may take place in a sedimentary deposit after it has been uplifted and brought within the range of ground water of meteoric origin (p.82)."

So he establishes a sharp upper limit for diagenesis which is accepted today by the great majority. He limits the "diagenesis" to "prior to solidification" and this is the fundemental American conception of the zone of diagenesis. In spite of the text books by KRUMBEIN and SLOSS (1963), WILLIAMS, TURNER and GILBERT (1954) and PETTIJOHN (1949; 1957; 1975) for the most of American geologists, diagenesis affects only the earliest stages of sediment. These former authors agree with the European definition of term and extend it up to metamorphism. However, TWENHOFEL himself, in the first edition of his "Principles of Sedimentation (1939)" changes his views and gives this definition:

"...Diagenesis includes all modifications that sediment undergo between deposition and lithification under conditions of pressure and temperature that are normal to the surface or outer part of the crust, and in addition, those changes that take place after lithification under the same conditions of temperature and pressure, which are katamorphic in character so that the effect is delithification (p.254)."

This definition eleminates regional and contact metamorphism on one hand and weathering processes on the other.

KRUMBEIN (1942) follows the same definition. He notes more than 30 diagenetic changes, briefly reviews them and consolidates into 6 major processes. They are compaction, cementation, recrystallisation, replacement, different solution and authogenesis. He also discusses the result of these changes in terms of sediment properties - composition, texture, and structure - with emphasis on textural changes.

PETTIJOHN (1957) in the second edition of his book "Sedimentary

Rocks" devotes an entire chapter to lithification and diagenesis. In his discussion of diagenesis he is mainly concerned with chemical changes and defines diagenesis as a process primarily due "to the reactions which take place within a sediment between one mineral and another or between one or several minerals and their interstital or supernatant fluids (p.648)." Such a definition is not entirely adequate, for limestones are frequently monomineralic. He distinguishes between the chemical rearrangement and replacements that take place on the sea-floor from those that occur after the sediment has been removed from direct contact with sea water. He suggests terms "Halmyrolosis" for the former and "Epigenesis" or "Metharmosis" for the latter group, and remarks that diagenesis includes halmyrolosis and metharmosis and by degree passes into metamorphism (p.649). In the 1971 Edition of "Encyclopedia Britannica" he gives two contradictory definitions for diagenesis:

"...Diagenesis, the geologic process by which, after deposition, a sediment may be materially altered or modified. These modifications may occur before burial at the common boundry, or interface, between sea water and the sediment, after burial but before consolidation or after consolidation either at normal temperatures and pressures or at elevated temperatures and pressures" and he goes on "...the term diagenesis is applied to those changes which take place in a newly deposited sediment, prior to its consolidation, either at or below the sediment-water interface. Changes in the sediment produced by higher temperatures and pressures are called metamorphic."

Lastly; in the third edition of "Sedimentary Rocks (1975)" he does not give any definition of diagenesis but in the discussion of crystalline textures on p.80 he notes that "...We have here used the terminology employed in the study of metamorphic rocks, in part because we do not wish to overload an already over burdened terminology, but mainly we hold that diagenesis - recrystallization, replacement, and internal reorganization (neomorphism) - is after all a metamorphic transformation" and later on p.366 he states "...The conversion of a limy sediment to a lithified crystalline body is a species of metamorphism, the resulting textures are not dissimilar to those found in the more conventional metamorphic rocks."

In a review of diagentic environment, DAPPLES (1959) suggested three phases of diagenetic process. The first involves changes in the sediment that occur during the course of its burial (The Initial or Depositional Stage); the second includes the changes that occur in the early stages of moderate burial (The Intermediate or Early Burial Stage); and the last phase is deep burial, sometimes associated with a long time span

(The Late Burial or Pre-Metamorphic Stage).

PACKHAM and CROOK (1960) approached to the problem from an entirely different point. They define diagenesis very broadly:

"...Diagenesis may be subdivided into halmyrolysis, early diagenesis, and epidiagenesis, being, respectively, alteration before and during deposition, immediately after burial, and under deep-seated conditions (p.392)." They put forward a sequence of facies, analogous to metamorphic facies, starting at the surface and passing down into the "low-grade metamorphic facies". Each diagenetic facies is defined as "...including all rocks which have, by a process of diagenesis, developed mineral assemblages that are the result of adjustment to a particular diagenetic environment (p.400)." They distinguished two groups of diagenetic facies. The first one is "Low-rank facies" which corresponds to early stage of transformation and can be subdivided on the basis of Eh, pH, and pressure variations. "Higher-rank facies" is the characteristic of late stages of transformation and forms the second group. Temperature is the most important parameter.

KRUMBEIN and SLOSS (1963) use "post depositional" and "diagenetic" synonymously and give clear and precise definitions:

"...At the instant of deposition, a sediment may consist of loose, detrital particles, crystals, organic fragments, colloidal mud, or mixtures of such substances. As a given lamina of sediment is formed, it becomes the interface between the previously deposited material and the medium of sedimentation.

The 'diagenetic environment' is the environment of post-depositional change. It extends an indefinite distance downward from the depositional interface. The nature of the diagenetic environment and the rapidity of the postdepositional changes depend upon the medium of deposition and the kind of sediment being deposited.

The 'depositional interface' represents an important boundry condition that separates two different physico-chemical regions. As a simple illustration, clay and silt settle through sea water as a group of particles in a liquid medium. When the particles come to rest on the bottom, they form a solid matrix with water-saturated pores. The water has the same composition as the medium above, but marked changes occur once it is sealed from free circulation by confinement in the pores.

As deposition continues, the lamina of sediment posses from the interface to successively lower positions and enters a realm of greater pressure, higher temperature, and of changed chemical and biological conditions. These new conditions promote the consolidation of lithification of the sediment into a sedimentary rock (p.262-3)." Then they discuss the marine and non marine diagenetic environments and give a classification of diagenetic processes which is not very different from KRUMBEIN (1942). To be precise, the only difference is the addition of PETTIJOHN'S (1957) "diagenetic differentiation".

The usage of the term "diagenesis" in Great Britain is very recent. The eighth edition of HARKER's (1954) "Petrology" does not mention it. MILNER (1962) although in the fourth edition of his book describes phenomena of "cementation and authigenesis" ignores the term. The same year READ and WATSON (1962) defines diagenesis as follows:

"...Diagenesis comprises all those changes that take place in a sediment near the earth's surface at low temperature and pressure and without crustal movement being directly involved. It continues the history of the sediment immediately after its deposition and with increasing temperature and pressure it passes into metamorphism."

TAYLOR (1964) presents a very detailed review of the diagenetic phenomena. He follows READ and WATSON's (1962) definition but he feels that "...the whole field of sub-aerial weathering effects..." should be specifically excluded from this definition (p.417). TAYLOR subdivided the diagenesis to "Preburial Stage (including transformations in the freshly deposited sediment with our without subsequent action by the environment - Halmyrolysis)", "Early Burial Stage (transformation before lithification)", and "Late Burial Stage (transformation after lithification)". He gives special attention to the "diagenetic environment" realized at each of these stages and attempts to reconstruct this environment for the principal petrographic facies (clays, sandstones, carbonates, etc.).

In the introduction of "Dolomitization and Limestone Diagenesis" the editors of the symposium (PRAY and MURRAY 1965) give a braod definition:

"...In its broadest sense, diagenesis encompasses those natural changes which occur in sediments or sedimentary rocks between the time of initial deposition and the time - if ever - when the changes created by elevated temperature, or pressure, or by other conditions can be considered to have crossed the threshold into the realm of metamorphism (p.1)."

As has been pointed out earlier there is not yet a universally accepted definition or delimitation of the term "diagenesis". The best example of this can be seen in the "Diagenesis in Sediments" edited by LARSEN and CHILINGAR (1967). In the chapter on diagenesis of carbonate rocks, CHILINGAR, BISSELL and WOLF use the term in a limited sense.

According to them diagenesis is complete when the sediment has been converted to a sedimentary rock. This application is the general practice in the U.S.S.R. The distinction is made between "syngenesis", "diagenesis", "epigenesis", and "metamorphism". However this limitation is not supported by the others of the same book. FAIRBRIDGE, DAPPLES, MÜLLER, and VON ENGELHARDT are in agreement that diagenesis includes all processes between deposition and metamorphism. In the same volume FAIRBRIDGE gives this definition:

"...Diagenesis begins at the moment a sedimentary particle comes to rest, for example, on the sea floor; and it continues to a point in history when either deep burial and orogenic buckling cause the initiation of metamorphism, of when emergence leads to exposure and the initiation of weathering and erosion (p.31-2)." So he puts a clear upper limit and distinguishes metamorphism and weathering from diagenisis. This approach is followed by the present author. FAIRBRIDGE recognizes three phases in the diagenesis. They are:

1 - Syndiagenesis (The Sedimentation Phase) : It comprises two stages:

a) Initial Stage - Marked by oxidizing conditions

b) Early Burial Stage - Marked by reducing conditions

2 - Anadiagenesis (The Compaction - Maturation)

3 - Epididiagenesis (The Emergent - Pre-erosion Phase)

The relationships between these phases are shown in Fig. V - 1.

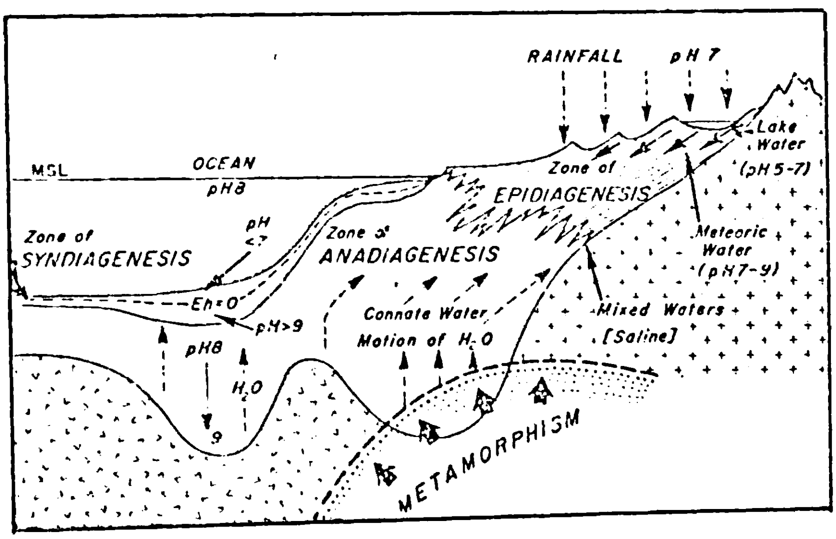

Fig.V-1 : Phases of diagenesis (After FAIRBRIDGE, 1967)

Although BATHURST (1971) devotes considerably part of his book "Carbonate Sediments and Their Diagenesis" to the diagenesis of carbonate rocks, he avoids giving a definition of the term. But it is clear that he applies the term in a restricted sense (i.e. not beyond lithification).

According to BLATT, MIDDLETON and MURRAY (1972) in its broadest sense, diagenesis of carbonate sediments and rocks includes all those processes that act on these materials after their initial deposition but before elevated temperatures and pressures create minerals and structures normally considered within the realm of metamorphism (p.456).

Latest edition of "The Glossary of Geology and Related Sciences, (AGI, 1972)" defines diagenesis as:

"...All the chemical, physical, and biological changes, modifications, or transformations undergone by a sediment after its initial deposition, and during and after its lithification, exclusive of surficial alteration (weathering) and metamorphism".

Also MILLIMAN (1974) and FUCHTBAUER (1974) apply the term in its broadest sense. MILLIMAN defines it as "...alteration and cementation of sediments during the interval between deposition and metamorphism (Marine Carbonates, p.251)". FÜCHTBAUER emphasises that "...diagenesis comprises not only lithification but all alterations which occur within the sediment after deposition (Sediments and Sedimentary Rocks I; p.4)".

Recently, WILSON (1975) uses the term in restricted sense:

"...The alteration of carbonate sedimentary particles and matrix is a continuing process. It begins during deposition and continues long after burial and the first stages of lithification (Carbonate Facies in Geologic history; p.69)".

It is clear from the above examples that there is still no generally accepted definition of the "diagenesis". Since first definition of WALTHER (1894), many authors have defined it differently in one way or another. There have been conflicting views concerning the limits of the concept. Some have applied the term in a restricted sense (i.e. Up to but not beyond lithification). On the other hand, many authors are in the thought of that diagenesis continues after lithification and grades into metamorphism. Table V - 1 summarises the views of some authors concerning the limits of the "diagenesis".

AUTHORS	USAGE OF THE TERM "DIAGENESIS"	
	In limited sense	In broad sense
WALTHER (1894)		+
TWENHOFEL (1926)	+	
TWENHOFEL (1939)		+
KRUMBEIN (1942)		+
PETTIJOHN (1949, 1957)		+
WILLIAM, TURNER, & GILBERT (1954)		+
DAPPLES (1959)		+
PACKHAM & CROOK (1960)	+	
READ & WATSON (1962)		+
KRUMBEIN & SLOSS (1963)		+
TAYLOR (1964)		+
PRAY & MURRAY (1965)		+
CHILINGAR, BISSELL, & WOLF (1967)	+	
FAIRBRIDGE; DAPPLES; MULLER; VON ENGELHARDT (In: LARSEN & CHILINGAR, 1967)		+
BATHURST (1971)	+	
BLATT, MIDDLETON, & MURRAY (1972)		+
GLOSSARY OF GEOLOGY (AGI, 1972)		+
MILLIMAN (1974)		+
FUCHTBAUER (1974)		+
WILSON (1975)	+	
PRESENT STUDY		+

Table V-1 : Concept of "Diagenesis" according to various authors

The principal diagenetic processes to affect sediments of the Gürün region are described in the following sections. These include biologic diagenesis, micritic envelope and micritization, cementation, neomorphism, compaction, pressure solution and formation of stylolites dolomitization, dedolomitization and silicification. Where data permits the author has attempted to draw conclusions regarding time relationships of the various diagenetic processes. A general summary of diagenesis has been given at the end of the chapter.

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PART 2

BIOLOGICAL DIAGENESIS

Studies of recent carbonates have shown that organisms play an important role in the early stages of diagenesis. Although not so easily recognisable, biologic activity was probably an important diagenetic agent of the Upper Cretaceous Carbonates of the Gürün Area as it is in the Recent carbonate sediments. Biological diagenesis includes burrowing and boring. They are holes made by organisms in soft sediments and in hard surfaces, respectively. Recognizable burrows are extremely rare in the field. Under the microscope sediment-filled burrows are fairly common. They differ in colour, texture, and sometimes in composition from thehost rock (Pl. 36). Usually they exhibit a circular or near-circular cross-section, indicating a lack of significant post depositional compaction (Pl.40 F). They are mostly filled with surrounding sediments, but some burrows are filled with sparry calcite. It is thought to be that the host material was, at least partially, consolidated at the time of burrowing.

Certain areas of disrupted primary stratification have been attributed to burrowing organisms (Pl.36A). Borings are common in allochems (Pl. 37). This phenomenon was probably an important process in the disintegration and alteration of skeletal material. In particular, algal corrosion of limestone fragments, especially skeletal remains, is believed to have been important. (See Part 3 - Micritic Envelopes and Micritization; p.132). It is probable that biologic activities were responsible for at least some of broken and abraded fossil fragments, though almost impossible to differentiate them from that broken by physical processes.

The abundance of pellets in some beds may also be taken as indirect evidence of organic activity.

Burrowing, and disruption of stratification, boring and disintegration of allochems by organic activity, and the aggregation of fine sediment into fecal pellets were all contemporaneous with the accummulation of sediment on the sea-floor and the first stage in the diagenesis.

PART 3

MICRITIC ENVELOPE AND MICRITIZATION

Micritic envelopes are layers of fine grained carbonate material constituting an outer layer on fossils or other grains in carbonate sediments. Based on a detailed study of carbonate particles from Bimini Lagoon, the Bahamas, BATHURST (1966) proposed that micritic envelopes form on the sea and their development takes place in three stages:

(1) Algae bore into the skeletal particles;

(2) The algal filaments within the bores die and decay, leaving the bores empty;

and (3) Tiny aragonite particles fill the empty spaces formerly occupied by the algae.

Repeated boring, dying and infilling could result in the formation of an opaque mass of micrite in the outer parts of the skeletal particles, yet within the original outline of the particle. In a recent work KOBLUK and RISK (1977) have used the term "destructive micritic envelope" for this type and shown that the algae also may produce micritic envelopes outside grains by the calcification of exposed dead endolithic filaments. This type has been called "constructive micritic envelope (Fig. V - 2).

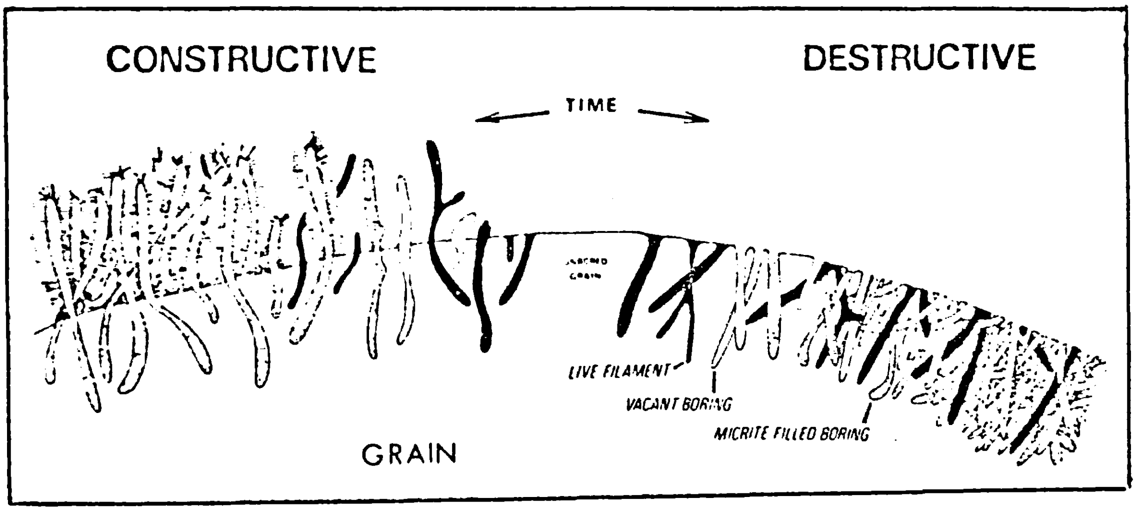

Fig.V-2 : Formation of constructive and destructive micritic envelopes (After KOBLUK and RISK, 1977)

Under the petrographic microscope these envelopes commonly are very difficult to distinguish from destructively generated envelopes (KOBLUK and RISK, 1977; p.1074). Micritic envelopes are a result of incomplete micritization. Through micritization, skeletal fragments and oolites may be transformed into structureless lithoclasts, the origin of which is no longer visible (Pl.37). FOLK (1965b; p.152) considers this process as a degrading neomorphism so does PETTIJOHN (1975; p.368). On the other hand according to MILLIMAN (1974) the process is an intrgranular cementation. More than one mechanism has been reported to take in the micritization process; for example, recrystallization (PURDY, 1968), dissolution - reprecipitation within an organic mucilagenous sheath (KENDALL and SKIPWITH; 1969), and repeated boring by endolithic algae, followed by precipitation of micrite in the vacated borings (BATHURST; 1966 and 1971; WINLAND, 1968; FRIEDMAN, GEBELEIN and SANDERS, 1971). As a result of the latter process carbonate grains are gradually and centripetally replaced by micrite. An excellent review of this subject has been given by ALEXANDERSSON (1972).

Micritization is often concentrated in slowly agitated lagoonal environments. However, micritization may also occur in deeper water where algae are absent; FISCHER and GARRISON (1967) found micritized subrecent Globigerinae included in crusts, in sites of low sedimentation. Subaerial weathering also may cause micritization (FRIEDMAN, 1964; p.806).

Micritization is generally absent in mud-supported Gürün Limestones but relatively common in the grain-supported platform edge carbonate sand banks.

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PART 4

COMPACTION

Compaction is a reduction in bulk volume of sediment, produced by the increasing weight of oberburden as the sediment is buried. In the late fifties it was thought that carbonate muds might compact mainly by water loss, similar to the near surface compaction of argillaceous muds and by recrystallization. Furthermore, it was believed that carbonate sands probably compacted very little, if at all, owing in part to the widespread phenomenon of early cementation in carbonate environments (See review of WELLER, 1959). The fact that carbonate sediments can, and do compact has been established in recent years. Cores of modern carbonate muds taken in the Bahamas and in Florida Bay usually show a porosity decrease from 80 to about 65 percent in the first few cm.s of burial. In a summary paper on carbonate diagenesis CHILINGAR, BISSELL, and WOLF (1967) note that compaction affects mainly unlithified sediments, that fine-grained sediments undergo the highest degree of compaction in the first 30 cms because of high initial water content (Average 80-90%) and that autochthonous limestones such as reefs, do not undergo much compaction. Intergranular spaces of allochthonous carbonate deposits are said to be eliminated by closer packing, crushing, deformation, expulsion of interstitial fluids and possibly by grain corrosion. They conclude that the rate of cementation, degree of compaction, and pressure solution, are closely related in the examples cited by them from the Nubrigyn-Tolga reef complex, and related basinal rocks of New South Wales (Australia).

In the first stage of compaction the interstitial water is forced out until the particles are brought into contact with each other. As compaction proceeds, there may be some rearrangements of grains with a resultant development of closer packing.

Field observations on carbonate rocks often indicate that very little compaction has occurred (e.g. MURRAY and PRAY, 1965; BEALES 1965; BATHURST, 1971 and FRIEDMAN, 1975). Other workers such as NEWELL and Others (1953) and POWERS (1962) however, have suggested that the fine-grained limestones, at least, have undergone considerable compaction. PRAY (1960) suggested that little evidence of compaction could generally be found in carbonate muds, and amongst other criteria cited, the common absence of the crushing of delicate fossils as one feature possible indicative of this. But experiments of BROWN (1969) on compacting pelecypod valves and whole shells in a carbonate mud matrix at pressure up to 1050 kg/cm^2 showed that where the shells of valves are mud supported and fluid is present then the pressure remains uniform around them, during compaction and only rarely did any breaking of shells take place. Thus, individual well-preserved shells or valves surrounded by carbonate mud are not evidence of lack of compaction. Another criterion against the compaction is pellets without any squashed or flattened appearance. FRIEDMAN (1975) states that "...if compaction was operative, the pellets would be squashed". However, well-formed micritic pellets are sometimes found preserved within shells where they were protected from compaction, where as the surrounding matrix contains perhaps only faint relics of merged pellets (MURRAY, 1965; p.317). Many uniform textured micrites may have originated by squashing of pellets. Well preserved pellets are probably due to early cementation and not the lack of compaction.

Some features found in the sediments investigated may be taken as evidence of lack of important compaction in these sediments. They can be summarized as follows:

(1) Borings and other voids now commonly filled with sparry calcite show no evidence of early diagenetic distortion, or compaction.

(2) The relatively undisturbed appearance of the algal-laminated sediments from original growth position.

(3) The similarity between the sediment under shells or in other uncompacted cavities and that of the surrounding unprotected areas.

(4) Scarcity of drag or penetration effects where rigid clasts or fossils occur. On the other hand, there are certain features found in the limestones that tend to indicate considerably compaction, such as:

(1) The presence of numerous stylolites, especially in the fine-grained, more argillaceous rocks. (Pl.40).

(2) Micritic limestone which on close examination can be seen consist largely of pellets and intraclasts that have been compressed together to the extent that their boundaries have merged and are no longer clearly distinguishable. (Pl.38E).

(3) Presence of crushed fossils.

(4) The degree of orientation of the bioclasts within a mud supported matrix with respect to bedding planes (Pl.38A). (Fossil orientation may be produced by the current sorting action as well).

A direct relationship seems to exist between the increase in compaction and the increase in the insoluble residue of the limestones. The upper limit of this content, beyond which compaction would take place or below which no significant compaction would have taken place, is just round 2% (ZANKL, 1969). It has been observed in the present study that more argillaceous limestones contained abundant stylolites. The lack of evidence of compaction in a considerable part of the Gürün sediments confirms that they become lithified prior to sediment burial. Limestones showing evidence of compaction observed to be rich in insoluble residue which prevented them being lithified at earlier stages. MARSCHER (1968) suggested that insoluble constituents tend to form envelopes around carbonate grains and inhibit their recrystallization or enlarement. He concluded that the carbonate grain size decreases with increasing clay mineral fraction. This is also observed in the present study (Pl.56B,D).

Contrary to most of the field observations, laboratory experiments on the compaction of carbonate sediments have shown that volume loss due to compaction in carbonates is important. TERZAGHI's (1940) experimental compaction of modern carbonate mud seemed to support her view regarding to the over steeped dip of the reef flanking beds, which she attributed to greater compaction of these beds as compared with the reef core.

FRUTH, ORME and DONALD (1966) described the effects of high pressure compaction on various carbonate sediments, including skeletal muds. They succeeded producing artifical limestones containing features present in natural limestones. Their experiments also revealed that most of the sediment types show a marked initial compaction below about 31 kg/cm^2 pressure, but above 360 kg/cm^2 the compaction curves are nearly parallel.

ROBERTSON (1967) compressed aragonitic Bahamian sediments, and indicated that size and shape of the carbonate grains, rate of loading and initial water content all affected the amount of compaction that took place; similar properties infact to those that are important in controlling the compaction of terrigenous muds.

In an article on the compaction of carbonate sediments EBHARDT (1968) discussed factors in addition to overburden pressure, which affect compaction including grain size and pore fluid composition. He compacted Recent carbonate sediments of different grain size and with different carbonate content under pressure of up to 670 kg/cm^2. His experiments revealed that coarse fraction shows intense compaction, where as the finer sediment needs a longer time for compaction and elevated temperature cause more compaction.

In a study of compaction within and around a Devonian bioherm MOSSOP (1972) found a close agreement between estimates of the amount of compaction made from an analysis of the irregular trend surfaces of horizons that were originally horizontal and the direct measurement of the amplitudes of the stylolites.

Petrographic study of the Gürün samples show that some plastic deformation occurred in the sediments investigated. There is no doubt that during plastic deformation the sediment was still soft and it took place in the early burial stage, with little overburden pressure. Some features thought to be the result of plastic deformation are as follows:

a. Merging of the boundaries of two grains. This feature is normally seen to occur between two pellets or adjacent ooliths. In the case of the latter the oolitic laminae in the area of contact become diffuse or disappear. The grains appear to have been "stuck" together (Pl.36).

b. Contact penetration of two grains. This is seen when two grains are in contact more than tangentially. A concova-convex contact is formed. Evidence of pressure-solution is absent: therefore it is interpreted as a plastic deformation (Pl.36).

c. Deformation of grains. In some samples, grains which have lost their external appearance are being observed. It is usually caused by the squashing of grains against solid objects (Usually shell fragments). Again no indication of pressure-solution has led the author to conclude that they are caused by plastic deformation.

Data obtained from the Deep Sea Drilling Project support the field evidence rather than experimental findings. That is to say, lime mudstones suffer only minor compaction compared with terrigenous mudstones. MATTER (1974; p.441) notes that "...compaction plays a minor role in the diagenesis of pelagic carbonates" in the Arabian Sea samples. In the light of petrographic evidence and findings of the Deep Sea Drilling Project.Cruises the present author can see no reason to assume that compaction had played an important role in the Upper Cretaceous carbonates of the Gürün area.

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PART 5

PRESSURE SOLUTION

AND

FORMATION OF STYLOLITES

Another phenomenon, seen frequently, due to compaction is pressure solution and formation of stylolites. In this account discussion is restricted to pressure solution and stylolites seen in carbonate rocks only.

Pressure solution is a preferential solution which takes place in the highest stressed parts of a grain and deposition of matter on surfaces with lower potential energies. The pressure is supplied by overburden. The most significant aspect of the pressure solution process is the bulk volume reduction. It is generally regarded to be effective in the reduction of initial porosity due to local precipitation of the pressure solution derived cement. BATHURST (1971) is of the opinion that pressure solution is the most probable source of second generation carbonate cement (Ibid; p.459). Simultaneously, pressure solution is considered to be an effective agent of pore reduction through grain corrosion and the resultant closer packing. The factors that control the absolute and relative pressure-solubility of carbonate grains are not completely understood at present. TRURNIT (1968) has shown that among other things, shape of grain, size of grain, surface tension, temperature and pore fluid composition are important parameters. TRU RNIT has also given a detailed classification of pressure-solution contacts based on their geometry and genesis. The most abundant type encountered in the present study is the penetration of two adjacent grains (Pl.38C). This phenomenon is considered to be due to different relative pressure-solubility of the grains (TRURNIT, 1968). Sutured contacts resulted from equivalent relative pressure-solubility and described by the same author are relatively rare in the samples studied. (Pl.39B).

BATHURST (1971) believes that grain-to-grain pressure solution can take place after the precipitation of drusy calcite cement, because this cement does not prevent grain movement. After formation of a second generation of cement, grain-to-grain pressure solution is impossible because this later cement prevents grain movement. (Ibid; p.465) and the rock yields to solution process by the formation of stylolites. Thus an early diagenetic origin is postulated for the pressure solution process.

NEUGEBAUER's (1973 and 1974) work on the Chalk of Northern Europe has advanced the theoretical understanding of pressure-solution. Theoretically, burial to 1500 m. or more should lead to substantial loss of porosity by pressure-solution process, yet in the Chalk this is not the case. He constructed a model, relating overburden, particle geometry, mineralogy, the ratio Mg^{++}/Ca^{++} in the pore solution and evidence of dissolution and precipitation. He also argued that in favourable circumstances, an overburden of at least 2000-4000 m. is required for the complete lithification of the Chalk by pressure-solution mechanism. Deep Sea Drilling Project data support his conclusion. NEUGEBAUER's finding that any large scale release of calcium and carbonate ions to give cement can only accompany the advanced stage of pressure-solution, is also supported by the theoretical model of MANUS and COOGAN (1974) who conclude that "...if pressure is a major source of pore-filling cement, a very large degree of bulk volume reduction must accompany it somewhere in the system (Ibid. p.470)".

In a recent work, DEELMAN's (1975) experiments on the dry compaction of mixture of mineral fragements with different hardness have given rise to grain contacts commonly interpreted as being the result of the pressure-solution.

PARK and SCHOT (1968b; p.66) have defined stylolites as "...irregular planes of discontinuity, along which two rock units appear to be interlocked or mutually interpenetrating". Mostly the planes are characterized by the accummulation of relatively insoluble residues, mainly clay and ferruginous material, and called "stylolitic seam" (Pl. 39). They can be traceable for long distances, up to several metres. Stylolites in the Gürün sediments are generally parallel to bedding, although there are stylolites which are transverse or vertical to bedding planes (Pl. 40). They show a variety of forms from gently undulose at one extreme, to very steep-walled at the other (Pl.40A). In some samples it has been observed that some stylolites truncate the others, implying two generations of stylolitization (Pl.40B,C) A large literature does exist on the subject which is discussssed in detail by PARK and SCHOT (1968 a;b). Two theories, regarding the origin and formation of stylolites have been supported by many workers. One of them is the "contraction-pressure theory" which proposes the formation of stylolites through the plastic flow of a thin clay layer over a lime mud when both are in a plastic state. This theory was advocated by MARCH (1868), ROTHPLER (1900) and SCHAUB (1939)(*).

(*) For references see PARK and SCHOT (1968 a).

This theory fails to explain the presence of insoluble residual material in the stylolite seams and the pressure-solution effects between grains along them. These features indicate that stylolites are not a simple plastic flow due to overburden; but the result of solution (PARK and SCHOT, 1968a; p.187).

The other theory was advocated by WAGNER (1913) and STOCKDALE (1922)(*) and is known as "solution-pressure theory". According to this theory stylolites originate through differential solution and pressure along a fracture or mechanical plane, after the hardening of the carbonate host rock. But some difficulties do exist for this theory too. They are formulated by PARK and SCHOT (1968 b) as "... the advocation of the presence of a joined surface, fracture, or mechanical plane along which stylolitization could have initiated and differential pressures on such surfaces or planes; the lack of microtectonic adjustment planes due to volume reduction during stylolite formation; the occurrence of generally flat grain boundary sufaces betwen late drusy mosaic calcite in indurated carbonates; and the lack of scarcity of observations of stylolites in folded carbonate rocks in the field and in the literature".

At present, it is generally accepted that stylolites are formed as a result of solution caused by pressure, and many probably result owing to the effect of compaction. The main problem concerning the phenomenon is not on the processes forming the stylolites; but on the time at which they were formed. Some workers (e.g. PARK and SCHOTT, 1968 a;b) believe that stylolitization commences early in the diagenetic history of a sediment and has a pre-lithification origin. Starting from the point that mechanical treatment of stylolite formation in virtually completely cemented carbonate rocks fail to account for the volume reduction which accompanies stylolitization (up to 35-42%). They came to the conclusion that a majority of stylolites and stylolitization is one of the important factors in supplying cement. They argue that the process is thought to end, when cementation reaches its final stages and porosity has virtually been eliminated. Some other authors (e.g. BROWN, 1959; DUNNINGTON, 1954, and 1967; BATHURST 1971; PETTIJOHN 1975) have presented evidence that stylolites develop after lithification of the sediment. In the present study it has not been possible to date precisely the time at which stylolites began and ceased to form. However, the fact that sylolites transect structures that were solid at the time of deposition (i.e. oolites, fossil fragments etc) and cut post-lithification features such as veins suggest that stylolitization is a relatively late

(*) For references see PARK and SCHOT (1968 a) diagenetic event although it may be, as suggested by PARK and SCHOT (Ibid.), a long drawn-out process, the length of time depending on the manner in which pore space is eliminated, by cementation, in adjacent sediments.

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PART 6

CEMENTATION

I. GENERAL

The cement in carbonate rocks is a non-skeletal, void filling precipitate on an intra-granular or intra sedimentary free surface (BRICKER, 1971). Rock components included in this definition are:

(1) The common intergranular cement as opposed to similar crystals believed to be neomorphic;

(2) Precipitated crystalline fillings in the body and wall chambers of fossils and within porous non-skeletal grains;

(3) Precipitated crystalline fillings of vugs, molds, caves, fractures, and other openings of post-depositional origin.

There are two types of crystal fabric of pore filling. In one of them the framework does not react with the solution from which the cement was precipitated. In this case the precipitated materials form a drusy implantation of crystals on the pore walls. These tend to grow outward into the pore; some crystals become dominant and suppress the growth of the others (Fig. V - 3A).

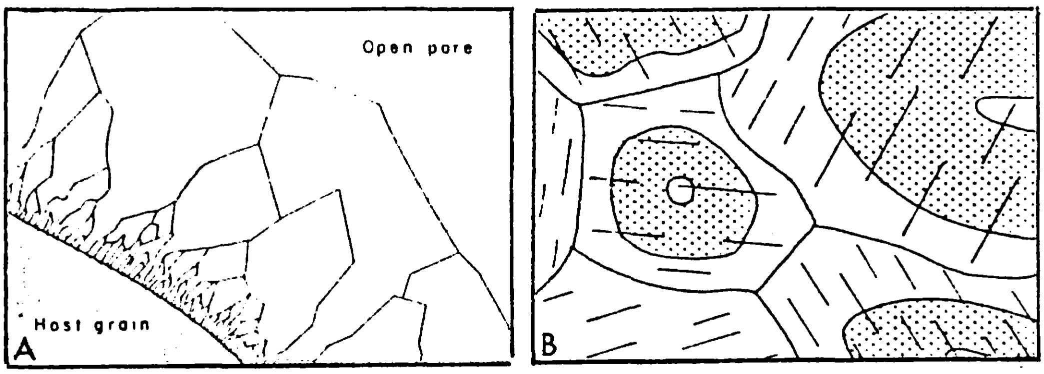

Fig.V-3 : A - Crystal fabric of pore filling (After PETTIJOHN, POTTER, and SIEVER, 1972, Sand and Sandstones)
B - Overgrowth cementation. The cement is optical continuity with the host grain.

Ultimately the crystal growing from various surfaces meet and the pore is filled. When the framework grains are reactive, they grow or are added to by precipitation from the pore solutions. This is the case with overgrowth of echinoderm fragments. The cement is in optical continuity with the host grain (Fig. V - 3B).

The predominant constituent of the Gürün sediments is micrite; whereas sparry calcite is not very common. The only carbonate cementing mineral in the sediments investigated is low-Mg calcite and for simplicity in this account it is called "calcite". Calcite has been seen to exist in three main crystal morphologies: Micrite, fibers and coarser, euhedral to anhedral, equant crystals (FOLK, 1974).

The term "micrite" was coined by FOLK (1959) as an abbreviation for the term "microcrystalline calcite" which referred to calcite crystals smaller than 4 microns. Later, LEIGHTON and PENDEXTER (1962) re-defined micrite as being finer than 30 microns, but in the present account the original definition of the term is accepted. It nearly always looks like extremely fine, semi-opaque "mud" in the petrographic microscope, no matter whether it exists as a bulk-rock component, or as a thin crust on grains. The distinction between micrite resulting from inorganic precipitation of a high-Mg micrite cement and micrite resulting from algal disintegration and diminution is difficult. This is confirmed by ALEXANDERSSON'S (1969) observations in the lithified carbonate sediments of the litoral and sub-litoral zone of the Mediterranean Sea. He found that high-Mg micritic cement is identical to the micrite in a lithified algal framework (Ibid.,p.47).

Fibrous calcite includes crystals with one large and two small dimensions. FOLK (1965) defined fibrous calcite as having crystals with a length to width ratio greater than 6:1. Although the same author used terms "bladed" for crystals with length/width ratio between 1 1/2:1 and 6:1 and "equant" for crystals with less than 1 1/2:1 length/width ratio LINDHOLM's (1974) work on pore filling calcite in septarian veins revealed that a natural boundary between "acute" and "bladed" lies near the length to width ratio of 3:1. The present author follows LINDHOLM's boundary between "equant" and "bladed". Therefore in this account "equant" is used for crystals with length/width ratio less than 3:1, "bladed" for crystals between 3:1 and 6:1 ratio and "fibrous" for crystals with a length to width ratio greater than 6:1.

Sparry calcite forms coarser euhedral to anhedral, equant crystals. It is generally accepted that sparry calcite is at times difficult to distinguish from neomorphic calcite. Criteria for differentiating two forms will be discussed further below.

The basic mechanisms of carbonate cementation are inherently chemical process and cement precipitation is genrally controlled by physiochemical and organic processes.

Factors such as increasing temperature, decreasing pressure (most likely in local pressure gradients adjacent to grain contacts), evaporation (increasing salinity), mixing of low and high salinity solutions, and loss of CO_2 may cause carbonate-bearing solutions to be supersaturated. However, supersaturation alone does not cause precipitation. Supersaturated solutions by as much as 500 percent are commonly reported from the marine environment; and even higher levels of supersaturation have been observed in meteoric water environment (BRICKER, 1971).

Decomposition of organic matter in carbonate sediments serves to increase alkalinity, dissolved ammonium and bisulphade ions in the interstitial waters. The primary mechanisms for this process are metabolism of the sulphate reducing and deaminating bacteria. This can result in the formation of excess bicarbonate which may cause precipitation of $CaCo_3$. On the other hand, certain bacterial cells and sheath of algae appear to concentrate some ions and may induce the precipitation of $CaCo_3$. However, the general quantitative significance of $CaCo_3$ precipitation by bacteria as compared to $CaCo_3$ originally supplied to sediments as biogenic particles is probably slight. The amount of dissolved calcium in interstitial water is usually much less than the amount of calcium present as biogenic carbonate. The most favourable site for the calcium carbonate precipitation by bacterial activity is near the sediment-water interface, where diffusion of appreciable amounts of calcium into the sediment from the overlying sea water is possible. Slow deposition favours this process.

II. ENVIRONMENT OF CEMENTATION:

Studies on Recent and Pleistocene carbonates have shown that carbonate sediments can become cemented in many different environments. Figs. V - 4 and 5 show these environments and carbonate cement morphologies in these environments. In the following paragraphs calcite cement types in each of these different environments will be discussed under the light of recent investigations.

A. METEORIC CEMENTATION: The cement of many marine limestones is inferred to have been deposited by percolating fresh water under subaerial conditions (SCHLANGER, 1963). This statement is later supported by geochemical evidence (FRIEDMAN, 1964; BERNER, 1966; and GAVISH and FRIEDMAN, 1969). The term "meteoric water" is preferred to the "freshwater" since water of meteoric origin can become "salty". Meteroric water diagenesis may take place in the vadose-subaerial or in the fresh-

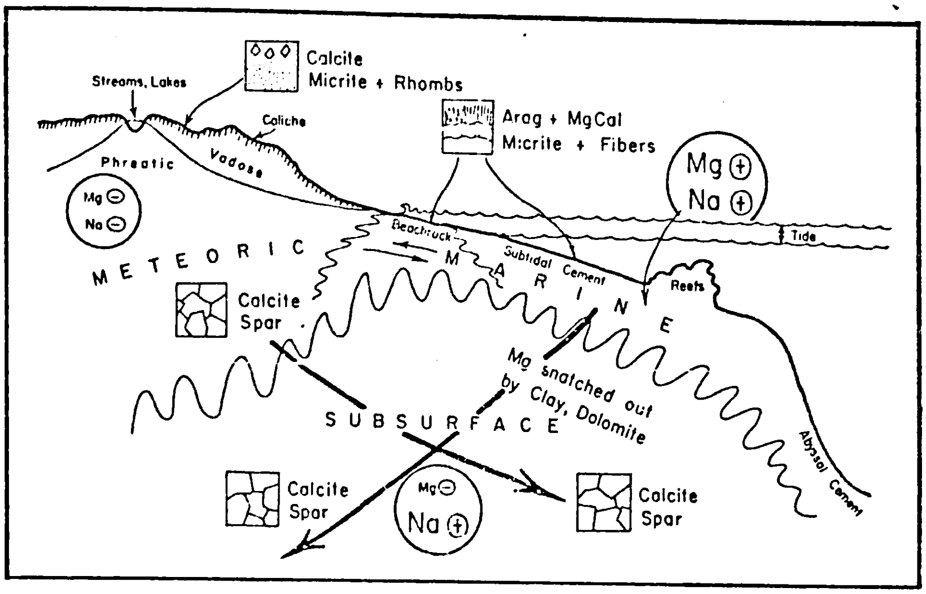

Fig.V-4: Schematic diagram showing the relationship between the mineralogy and morphology of carbonate cements and their diagenetic environments (After FOLK, 1973)

water-phreatic environments. The vadose diagenetic environment is situated above the water table. Pore spaces in this environment are occupied by both meteoric water and atmospheric gasses. The phreatic environment lies below the watertable where pore spaces are occupied entirely by water.

Subaerial exposure and diagenetic alteration in the vadose zone have been recognised widely as the major early diagenetic alteration environment of the carbonate sediments (SCHLANGER, 1963; FRIEDMAN, 1964; PURDY, 1968; BATHURST, 1971), but recent data suggests that the fresh water phreatic environment may be an equally important site of carbonate diagenesis (MATHEWS, 1974; LAND, 1970 and 1973; PINGITOR, 1976). STEINEN and MATHEWS (1973) suggested that carbonate diagenesis is more rapid in the freshwater phreatic environments than the vadose environments.

Criteria for meteoric vadose diagenesis in the Gürün sediments are as follows:-

(1) Deposition of vadose internal sediment "Vadose silt (DUNHAM, 1969)". Pls.41-42.Diagnostic features of vadose silt are:

(a) More or less equigranular calcite mosaic with average diameter of 10-25 micron. Extra ordinarily well sorted;

(b) They always occupy the lowest part of a former cavity and produces inclined floors or complete filling indicative of current transport;

(c) They occur both in primary and secondary voids;

(d) They postdate early calcite cement and predate blocky cement.

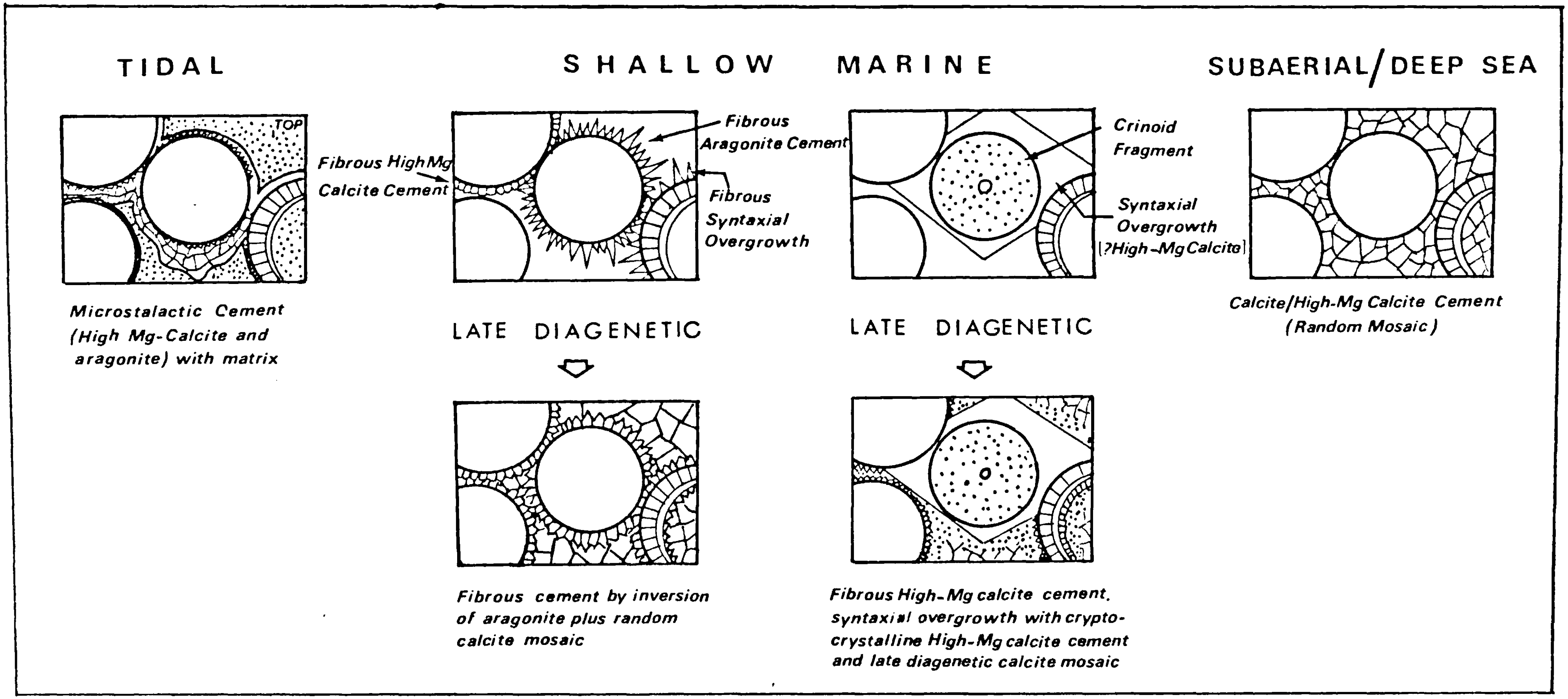

Fig.V-5 : Cement types(Modified from FÜCHTBAUER, 1969)

(2) Gravitational cement (MÜLLER, 1971). It forms after the bulk of the mobile water has drained out of the pores but still leaves a thick water film at the lower surface of the grains (Pl.41D.) This type of cement has also been found in the experiments of BADIOZAMANI, MACKENZIE, and THORSTENSON (1977) as characteristic of vadose zone;

(3) Micrite rim cement (Including DUNHAM's vadose pisolites);

(4) Crystal size is finer in vadose than in phreatic cements (LAND, 1970).

Although needle-fiber cement composed of randomly oriented, straight, elongate fibers has been noted by JAMES (1972) and WARD (1975) as characteristic feature of vadose zone, it has not been observed in the present study. It may be because of the erosion of the uppermost part of the limestones. Needle-fiber fabric would have formed close to the original ground surface. On the other hand the lack of "meniscus cement" in the inferred vadose zone may reflect lack of grainstone facies in the sediments of this area. Meniscus cement is an equant cement and fills the corners of pore spaces; e.g. the points of grain contacts, thus rounding the pores and characterize the vadose zone (DUNHAM, 1971). It is best developed in sediments with primary porosity (e.g. grainstones).

Criteria for meteoric phreatic diagenesis in the Gürün sediments are:

(1) Calcite cement crystals are larger in phreatic than in vadose zone. The experiments of BADIOZAMANI, MACKENZIE and THORSTENSON (1977) have shown that increased salinity or temperature are major factors in increasing crystal size of cement formed in the phreatic zone.

(2) Ferroan calcite cement precipitation can be seen in reducing environments of phreatic zone (BATHURST, 1971).

Vadose cement seems to be the most abundant type encountered in the present study (Pls.41-42).

B. MARINE CEMENTATION: Marine cementation occurs within two general environments; Intertidal and subtidal.

In the intertidal zone the predominant rock type is a layered calcarenite cemented with calcium carbonate and is called "beach rock". It may be lithified in the intertidal plus sea-spray zones, whether on high or low energy beaches, or even on broad tidal flats and tidal channels (BRICKER, 1971). It is cemented by simple pore filling and may occur in two main morphologies (micritic coatings and fibrous bladed or equant crusts) involving two minerals (aragonite and high Mg calcite).

(See Table V - 2).

AUTHOR	LOCALITY	CRYSTAL HABIT	MINERALOGY
GINSBURG (1953)	Florida	Fibrous	Aragonite
ILLING (1954)	Bahamas	Fibrous	Aragonite
STODDART & CANN (1965)	British Honduras	Fibrous	Aragonite
FRIEDMAN (1968)	Gulf of Aqaba (Red Sea) and Persian Gulf	Micritic	High-Mg calcite
GAVISH & FRIEDMAN (1969)	Mediterranean Coast of Israel	Micritic Fibrous	High-Mg calcite Aragonite
ALEXANDERSSON (1969)	Mediterranean	Micritic	High-Mg calcite
TAYLOR & ILLING (1969)	Qatar Peninsula (Persian Gulf)	Micritic Fibrous	High-Mg calcite Aragonite
TIETZ & MÜLLER (*)	Fuerventura (Canary Islands)	Equant and micritic	High-Mg calcite
MOORE (*)	Cayman Islands (BWI)	Micritic and fibrous	Aragonite
FRIEDMAN & GAVISH (*)	Gulf of Aqaba (Red Sea) and Mediterranean	Micritic Fibrous	High-Mg calcite Aragonite
SCHMALZ (*)	Eniwetok Atoll (Marshall Islands)	Fibrous Micritic	Aragonite High-Mg calcite
MULTER (*)	Dry Tortugas (Florida)	Fibrous	Aragonite
TAYLOR & ILLING (*)	Qatar Peninsula (Persian Gulf)	Fibrous and micritic	High-Mg calcite and aragonite
MOORE & BILLINGS (*)	Cayman Islands (BWI)	Fibrous and micritic	Aragonite
SIBLEY & MURRAY (1972)	Bonaire (Netherlands Antilles)	Fibrous	Aragonite
MOORE (1973)	Grand Cayman Island (BWI)	Equant and micritic Fibrous	High-Mg calcite Aragonite

(*) In: O.P.BRICKER (Editor), 1971, Carbonate Cements, The John Hopkins Univ.Studies in Geology No.19

Table V-2: Examples of various shallow water cementation

The aragonite cement is usually fibrous although cryptocrystalline aragonite also has been noted. In most instances high-Mg calcite is present as micritic coating. In their work in lagoonal, intertidal and suratidal sediments surrounding the Qatar Peninsula, Persian Gulf, TAYLOR and ILLING (1969) found stalactitic cement indicative of alternating wetting and drying in the intertidal environment. The same type of cement has also been observed in the Middle Jurassic limestones of the Paris Basin (PURSER, 1969). It is very similar to gravitational cement described by MÜLLER,(1971) indicating the vadose zone of the subaerial diagenetic environment.

Beach rocks form very rapidly, as evidenced by the incorporation of World War II skulls and military debris in Pacific beach rocks. Most workers seem to favour the precipitation of cement by sea water. The exact mode of lithification may involve perculation and lithification at depth or sea water evaporation and heating at the surface MILLIMAN, (1974). GINSBURG (1953) theorized that porous sands and high temperatures were necessary for the evaporation of sea water and subsequent cementation of beach rocks. Similar theories have been proposed by MOORE and BILLINGS (1971) and TAYLOR and ILLING (1971).

As pointed out by BATHURST (1971) recognition of fossil carbonate beach rocks is a difficult task. If the aragonite has been dissolved and recrystallized as calcite,(*) than its one-time existence may be impossible to prove. In the case of inversion(*) some relic fabric may remain but has not been documented in the geologic literature yet.

On the other hand micritic beach rock cement is very common in the Recent sediments but almost entirely absent in ancient rocks BRICKER, (1971). It is the opinion of the present author that beachrock cementation played an important role in the lithification of large amounts of intertidal sediments lying across a vast area in the Gürün region during Upper Jurassic-Createaceous times. It is most probable that at least some of fibrous calcite crusts observed as the first pore-filling, followed by random equant calcite mosaic, represent inverted beach rock cement preserving the original texture. (Pl. 43)

Subtidal environments consist of shallow water and deep water limestones. The former is restricted to depth of less than 100 metres while the latter can occur at depth exceeding 3000 metres.

Two distinct types of shallow water cementation have been found. The first one involves the cementation of loose grains and the second type is reef limestone in which primary cavities and biogenic burrows have been filled with internal sediment and subsequently lithified (MILLIMAN, 1974). Many workers have noted the occurrence of shallow water cementation from very different localities (See Table V - 3.) GINSBURG, SHINN and SCHROEDER (1967) reported a high-Mg calcite cement in algal reefs of Bermuda. Following this discovery MACINTYRE, MONTJOY and D'ANGLEJAN (1968) described calcite cement in fecal pellets and skeletal sand on a submerged barrier reef off Barbados. TAFT, ARRINGTON, HALMOUITE, MACDONALD and WOOLHEATER (1968) documented a wide spread aragonite cement in non-skeletal sands on Eatern Great Bahama Banks. SHINN (1969) discovered extensive areas in the Western Persian Gulf where lime sands were cemented by aragonite and high-Mg calcite. ALEXANDERSSON (1969) reported a high-Mg calcite cement within the coralline algal crusts from several localities in the Mediterranean; LAND and GOREAU (1970) found high-Mg calcite cement in a Jamaican coral reef; SCHROEDER (1973)

(*) See Part On Neomorphism (p. 166).

AUTHOR	LOCALITY	CRYSTAL HABIT	MINERALOGY
GINSBURG, SHINN & SCHROEDER (1967)	Bermuda	Fibrous	High-Mg calcite
MACINTYRE, MOUNTJOY & D'ANGLEJAN (1968)	Barbados	Fibrous	High-Mg calcite
TAFT, ARRINGTON, HAIMOUTA, MACDONALD & WOOLHEATER (1968)	Bahamas	Fibrous	Aragonite
ALEXANDERSSON (1969)	Mediterranean	Micritic	High-Mg calcite
SHINN (1969)	Persian Gulf	Fibrous Micritic	Aragonite High-Mg calcite
LAND & GOREAU (1970)	Jamaica	Fibrous	High-Mg calcite
FRIEDMAN, SANDERS, GAVISH & ALLEN (*)	New Jersey	Fibrous and micritic	Aragonite
GINSBURG, SCHROEDER, & SHINN (*)	Bermuda	Fibrous	High-Mg calcite
LAND (*)	Jamaica	Micritic and fibrous	High-Mg calcite
SHINN (*)	Persian Gulf	Fibrous Micritic	Aragonite High-Mg calcite
HOSKIN (*)	Alacran Reef (Mexico)	Fibrous	Aragonite (?)
PINGITORE (*)	Barbados	Fibrous	Aragonite (?)
GLOVER & PRAY (*)	Bahamas	Fibrous	High-Mg calcite and aragonite
ROBERTS & MOORE (*)	Grand Cayman Island	Micritic	Aragonite
MACINTYRE, MOUNTJOY & D'ANGLEJAN (*)	Barbados	Micritic and fibrous	High-Mg calcite
MACINTYRE & MILLIMAN (*)	SE USA	Micritic	High-Mg calcite
MARLOWE (*)	Caribbean Sea	Fibrous	High-Mg calcite
FRIEDMAN, SCHNEIDERMANN & GEVIRTZ (*)	Red Sea	Fibrous	Aragonite
JAMES, GINSBURG, MARSZALEK, and CHOQUETTE (1976)	Belize (British Honduras)	Micritic and bladed	High-Mg calcite
MACINTYRE (1977)	Caribbean Sea	Fibrous	High-Mg calcite
(*) In: O.P.BRICKER (Editor), 1971, Carbonate Cements, The John Hopkins Univ.Studies in Geology No.19			

Table V-3 : Examples of various shallow water cementation

AUTHOR	LOCALITY	CRYSTAL HABIT	MINERALOGY
FRIEDMAN (1964)	Atlantis Sea Mount	Micritic	High-Mg calcite
MILLIMAN (1966)	Atlantic Sea Mounts	Fibrous	High-Mg calcite
GEVIRTZ & FRIEDMAN (1966)	Red Sea	Fibrous	Aragonite
FISCHER & GARRISON (1967)	Barbados	Micritic	High-Mg calcite
" " "	Eastern Mediterranean	Micritic and sparry	High-Mg calcite
BARTLETT & GREGGS (1969)	Mid-Atlantic Ridge	Sparite and micritic	Calcite
MILLIMAN, ROSS & KU (1969)	Red Sea *	Micritic	High-Mg calcite
MÜLLER & FABRICUS (1970)	Ionian Sea *	Micritic	High-Mg calcite
MILLIMAN & MÜLLER (1973)	Eastern Mediterranean *	Micritic	High-Mg calcite

Table V-4 : Examples of various deep-sea cementation
(*) Semi-enclosed basins.

has shown that cements recognized in Recent Bermuda cup reefs can be traced into the past and the additions to the diagenetic sequence can be distinguished. The varieties of habits of the aragonite and high-Mg calcite cements the time sequences of their precipitation and the influences of the various substrates on cement mineralogy and fabric have been described by GINSBURG, MARSZALEK and SCHNEIDERMANN (1971); FRIEDMAN, AMIEL, and SCHNEIDERMANN (1974); and GOREAU and LAND (1974). On the other hand the first results of a continuing study of the Belize (British Honduras Reef Complex) (See JAMES, GINSBURG, MARSZALEK and CHOQUETTE 1976) have revealed that early shallow water cement is restricted to marginal facies of the reef complex and cementation is not continuous below a depth of 1 metre. The predominant cement is high-Mg calcite as micritic and bladed crusts. Aragonite is found only in coral pores. They have come to the conclusion that the presence of well-sorted lime silts as internal sediment is the most important factor governing the presence or absence of early shallow marine cement. MACINTYRE (1977) has also found mainly high-Mg calcite coating various reef components in the Carribean fringing reefs. He suggested that cementation in modern reefs is a near-surface phenomenon which is intensified by exposure of the reef-water interface to agitated normal marine waters.

There is virtually no obvious difference in morphology or mineralogy between beachrock and shallow water cementation. An exception is the "stalactitic cement" described by TAYLOR and ILLING (1969). Fibrous cement seems to be the most typical cement type as it is in the inter-tidal environments.

Another characteristic of shallow water environments is the presence of a pelletoidal texture. The pelletoidal texture thought to be the result of differential micro-growth rather than organic pellet forming processes (ALEXANDERSSON, 1972).

The concept of submarine cementation of carbonate sediments in the deep sea has gained increasing attention in recent years. FRIEDMAN (1964) noted lithified micrite at 300 metres depth from mid Atlantic ridge (Ibid., p.806). MILLIMAN (1966) described in detail fibrous high-Mg calcite crusts and lithification of carbonate sediments on two North Atlantic Guyots. The limestones were found to be in oxygen isotopic equilibrium with deep marine water, thus, indicating that submarine cementation does occur. Very slow rate of deposition grants cementation. GEVIRTZ and FRIEDMAN (1966) took cores from 1370 and 1700 metres in the Red Sea and found fibrous aragonite crusts. Following these discoveries many more locations have been found (Table V - 4). There are certain differences in cementation between shallow water and deep water environments. In most instances the cement of the submarine limestones is high-Mg calcite, and is usually microcrystalline, the component crystals being smaller than 10 microns. Aragonite cement is very rare in the deep sea. Random calcite mosaic which does not form during early diagenesis in shallow water environments has been found in the deep sea. Deep sea limestones also tend to contain planktonic components.

In the modern deep seas submarine cementation does appear to be quantitatively unimportant. Exceptions are the semi-enclosed Red Sea and Easter Mediterranean Sea basins where conditions are sufficiently restricted to facilitate inorganic precipitation and lithification of calcium carbonate (MILLIMAN, 1974).

C. SUBSURFACE CEMENT: FOLK (1973) is of the opinion that when to sample and study will become possible that subsurface cement "...will eventually be recognized as the most abundant type in ancient limestones (Ibid., p.131)". He suggests that subsurface cement is mainly a coarsely crystalline sparry calcite cement mosaic.

As conclusion, it may be said that sea water produces micritic or fibrous aragonite or high-Mg calcite cement while fresh water produces micritic cement and sparry calcite mosaic. Subsurface waters also produce sparry calcite (Fig. V - 4). These relations are best explained by the influence of Mg and Na ions as discussed by FOLK (1973 and 1974). Sea water contains abundant Mg and Na ions. This Mg ions will selectively poison side faces of the growing calcite crystals so that only micrite or short fibers may develop. Fresh water contains little Mg. Thus low-Mg calcite forms. Depending on the rate of precipitation micrite or sparry calcite can form (Rapid precipitation results in micritic calcite and continuous supply of fresh water produces sparry calcite as seen in the meteoric phreatic zone). Magnesium is also low in deeply buried connate waters. This is because of the capture of magnesium ions by clay minerals or the growth of replacement dolomites. As a result of this, sparry calcite will crystallize in the pores of deeply buried carbonate rocks. BADIOZAMANI, MACKENZIE, and THORSTENSON'S (1977) experiments confirm conclusion of FOLK concerning cement genesis. They have found that Mg ion was the main morphological modifier in their experiments and presence of Mg ion in the solution favoured precipitation of small, elongated, needle-type crystals of aragonite as opposed to euhdral, bladed to equant, and relatively large crystals of calcite that formed in solutions containing little or no magnesium ion.

Origin of cements and the conditions under which they originated are still highly debatable in the ancient rocks. Especially identification of shallow marine cements in post-depositional non-marine environments is problematics. Biologic and sedimentologic characteristics prove that great parts of the Gürün sediments to be deposited in an environment very near to mean sea level. From numerous extensive studies on eary diagenesis and lithification of Recent sediment it follows that subaerial exposure and shallow water conditions during deposition of the Gürün sediments favoured early diagenetic cementation with calcite. MEYERS (1974) has applied cathodoluminesance technique to an examination of the calcite cements in the Mississippian Lake Valley Formation of New Mexico and revealed 4 zones based on the content of Mn ions in calcite crystals. It became possible to corralate these zones through 100 m of vertical section and over a horizontal distance of 16 km. He has also discussed the possible sources of carbonate, both autochtonous and allochthonous.

III. CALCITE CEMENT TYPES:

The terminology for classification of the observed cements is adapted from FOLK (1965) who differentiated three types according to the foundation which the cement is built. They are (a) Overgrowth; (b) Crust; and (c) Random. The properties of these cement types are equal to those of BATHURST's (1958) "rim cement", "granular cement", and drusy mosaic"; ORME and BROWN's (1963) "syntaxial cement rim", "drusy fibrous calcite", "granular cement" and "drusy mosaic".

One of the most important problems of calcite cement studies is the distinction between sparry calcite formed by cementation and neomorphism(*). Therefore, before dealing with the cement types criteria for distinguishing two types of calcite will be discussed below:

A. THE CHARACTERISTICS OF CEMENT AND NEOMORPHIC CALCITE: It is now known that many of the criteria that have been listed for the recognition of primary calcite developed in a cavity apply as well to neomorphic calcite (FOLK, 1965; CHANDA, 1967). The neomorphic origin of the sparry calcite as opposed to the chemical precipitation of the calcite into pores is generally quite evident throughout most areas of the thin sections and peels studied. However, in some places, a neomorphic origin for the sparry calcite is not apparent because neomorphic sparry calcite closely resembles sparry calcite that is precipitated directly into a cavity.

(*) See Part On Neomorphism (p.166)

BATHURST (1958; 1971), FOLK (1959; 1965), HARBAUGH (1961), STAUFFER (1962), ORME and BROWN (1963) are among the authors who have described the criteria for recognition of pore fillings sparry calcite and neomorphic sparry calcite. The following types of evidence for the origin of sparry calcite have been applied to the present study, always keeping in mind that most criteria "...are equivocal, merely giving indications and not certainties (FOLK, 1965; p.45)".

(1) Neomorphic calcite is more typical random in orientation because it can start anywhere, whereas cement must of necessity start by encrusting a surface;

(2) In the carbonates investigated, a major character of the cement fabrics encountered is the marked size differences between the first formed crystals and the next generation. It is only seen in cement fabrics; (Pl. 43).

(3) The presence of absence of relic structures within sparry calcite. When these features occur, there is no doubt regarding the neomorphic origin of the fabric;

(4) Neomorphic sparry calcite transects grains and sometimes some large crystals at the margins of the sparry masses embay the adjacent microcrystalline matrix;

(5) Preferred orientation of crystal axes normal to a pore wall along with longest diameters normal to the wall; increase in crystal size and decrease in the number of crystals away from the wall; overgrowth around grains in optical continuity with host; wide variation in size of crystals are criteria of grain morphology which usually indicate cementation origin but they can also occur in neomorphic situations (FOLK, 1965; CHANDA, 1967);

(6) Sparry calcite found in the well rounded, well sorted and generally grain supported rocks is usually of the cementation origin. The sorting and rounding both indicate current movements, which probably would have removed any fine interstitial material originally present, living open pore spaces, which are readily filled with sparry calcite. On the other hand, loosely packed, poorly sorted and angular grains indicate a lack of current activity and probably an original microcrystalline calcite matrix which later converted to microspar or pseudospar;

(7) Sparry calcite mosaic in the upper part of a cavity whose lower part is filled by a more or less flat-topped internal sediment is the result of cementation process;

(8) If micrite is abundant but scattered then neomorphism fabrics are suspected. If micrite occurs only in selected areas, then cement fabrics are more likely.

Common cement and neomorphic calcite types have been shown on Figure V-6.

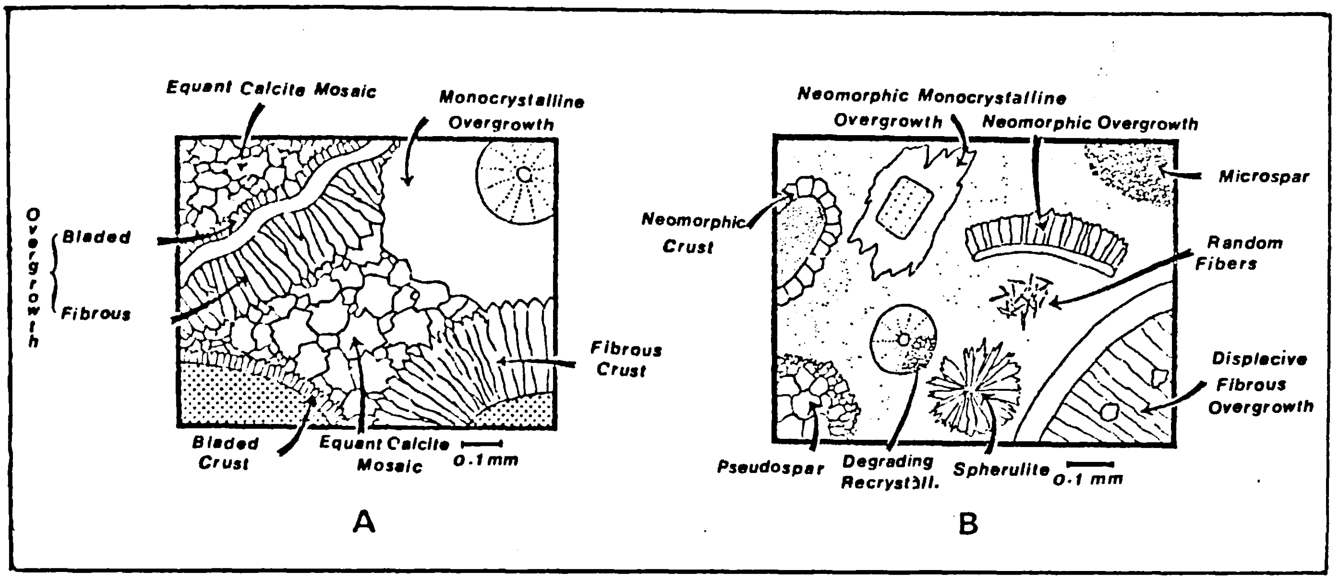

Fig.V-6 : A - Pore filling calcite types (Simplified after FOLK, 1965)
B - Neomorphic calcite types (Simplified after FOLK, 1965)

B. OVERGROWTH CALCITE: This type of calcite refers to the calcite cement that is deposited on grain (best developed around crinoid fragments) as an optically oriented (syntaxial) overgrowth). This type of cement, although common in the geologic literature, has not been observed in the sediments investigated. This is probably due to partly the scarcity of grain-supported rocks and partly the lack of echinoderm fragments in these grain-supported rocks. Echinoderm fragments are relatively abundant in some wackestones and the overgrowth observed around them is believed to be of neomorphic origin. (See p. 175 and Pl.54A,B).

C. CRUSTS: They consist of bladed or fibrous crystals that line cavities and void- or form "fringes" around large particles. The elongated crystals tend to be oriented with their long axis perpendicular to the surface of deposition (Pls.43,44) This is usually the first cement to be deposited in a rock and is believed to be an early diagenetic event. The contact between the early and later cements is generally sharp.

There is no transition between the crust and overlying coarser sparry calcite. The abrupt change from bladed or fibrous calcite cement to equant calcite mosaic with the extremely sharp contact defined by the crystal faces of the former suggest a break in calcite precipitation (WHITCOMBE, 1970, p.337; TALBOT, 1971, p.262). There is also a marked difference in size between crystals forming crust and equant sparry calcitemosaic. It was thought by BATHURST (1964) to be due to either lower nucleation density in the late stages, or due to a reduces supply of dissolved carbonate effected by a lowering of the permeability during early cementation. OLDERSHAW and SCOFFIN (1967) agree with BATHURST, stating that because of the reduced permeability, precipitation could cease until interstices were reopened by slight fracturing during compaction. They further state that the crystal size difference between two generations of cement is due to the attainment of perfect shape by the first formed crystals. EVAMY (1969) considered a drop in the concentration of the solutions to be important where as OLDERSHAW and SCOFFIN (1967) suggested that if the source of the first stage become exhausted, then calcium carbonate would not be available for precipitation until the sediment was subjected to a new environment. The present author agrees with EVAMY because a drop in concentration of the solutions is important. First a drop in the concentration of the solutions would produce larger crystals because the crystals would take longer to grow. Secondly, a drop in the concentration of solutions would mean less calcium carbonate available for deposition.

In the intertidal zone the beachrock cement is either aragonite or high-Mg calcite (p.150) whereas meteoric environment produces a low-Mg calcite crust (BATHURST, 1971, p.368). Although the fabrics recorded from recent limestones are comparable with those of the Gürün Region it is not possible to assess the original mineralogy of the first phase cement in the sediments investigated. A fossil example comparable to that found in the Gürün sediments is the "short, closely packed, calcite druse" described by PURSER (1969) from the Middle Jurassic of Paris Basin and it is suggested as being originally calcitic. In Recent limestones these early crusts are part of Stage II of LAND's six stages of cementation (LAND, MACKENZIE and GOULD 1967) usually formed by meteoric water above the water table (e.g. vadose zone). It is thought that the Gürün Basin, like the present day Bahama Platform, was extensively shallow with a depth of only a few metres over a large part of the platform and slight sea-level oscillations or extrmely minor fluctuations of the basin may lead to exposure of the entire basin with resultant fresh water diagenesis of the carbonate sediments. Although this early cement has textural analogies with recent examples from a vadose zone, the possibility of its being formed in an intertidal or subtidal (shallow water) environment can not be ruled out. It is possible that this cement originated as a precipitate of calcium carbonate from super-saturated sea water as described by TAYLOR and ILLING (1969) and SHINN (1969) from the present day Persian Gulf.

Another type of cement which can be attributed to this class is the "micritic cement". This type of cement has not been admitted so often and has mostly been considered as mechanically deposited material. However work on lithification and diagenesis of Holocene and Recent sediments has shown that a dark, dense cryptocrystalline material was found cementing carbonate sediments as well as filling or lining cavities. In the sediments investigated they have also been observed to be present as thin laminae within bladed or fibrous crusts (Pl.45). This is 'irregular micritic rim' of HAVARD and OLDERSHAW (1976). AL-HASHIMI (1977) has observed similar type of cement in the upper part of the intertidal zone. HARRISON and STEINEN (1978) have also reported "thin, somewhat arcuate 'sheaths' of elongate (?) micritic crystals (Ibid., p.394)" from Mississippian of Kentucky, U.S.A., and Recent Barbados. They have also noted that these sheaths "consists of tiny bladed to needlelike crystals oriented with their long axes mutually parallel and in the direction of sheath elongation (Ibid., p.394)". Their interpretation is that they represent coatings precipitated on the surface of rootlets and root hairs that once occupied cavities. The only impropriety between this type of cement described by the above authors and the one observed in the present study is that all the above examples have been observed in the intertidal or meteoric environments; whereas the one observed by the present author is associated with the calciturbidites. The only possible explanation is that it represents a very late diagenetic cement (after the emergence while the pore spaces were still available) in the meteoric environment probably vadose zone.

Another type of "micritic cement" is the "microcrystalline calcite cement" of PURSER (1969) which is much thicker. In the sediments investigated it generally overlies crystals forming the crusts (Pl.46). According to PURSER it represents synsedimentary cementation in marine environment (Ibid., p.228). On the other hand it is well possible that it may represent the result of degrading neomorphism of existant crust.

D. RANDOM CALCITE MOSAIC: Mosaic of calcite with equant grains fill the depositional voids and secondary or late diagenetic cavities not already been occupied by the earlier formed crusts. The mosaic shows an increase in crystal size outward from the pore wall. Sometimes it exhibits a transition from iron poor to iron-rich calcite (Pl. 53). This second phase cement presumably formed after the conversion of all metastable carbonate minerals to low-Mg calcite; part of LAND's Stage IV of cementation (LAND, MACKENZIE, and GOULD, 1962). Some equant sparry calcite mosaic fills smooth bottomed, unsheltered cavities known as "stromatactis",(*) "bird's eye structure" or "fenestrae". Sometimes this calcite cement is underlain by internal silts similar to those described by DUNHAM (1969) from the Permian Limestones of New Mexico and ascribed to the vadose zone, suggesting that at least some of this phase cement developed in vadose zone. On the other hand it is generally accepted that iron rich calcite is produced under reducing conditions (EVAMY, 1969; BATHURST, 1971) and that such conditions are generally encountered only below the water table (e.g. phreatic zone).

Another variety of this type cement is the "clear calcite pore infills". Single clear calcite crystals may develop in the pore spaces between closely packed grains (Pl.48A). It is possible that the development of clear infills depends upon the amount of space needed to be filled and the strength of the solution. A similar type of cement has been observed in some cavities. Only a few large crystals, normally not more than three, fills the cavity (Pl. 46).

IV. CALCITE SHELL FABRICS:

Organic structures in life may be composed of high-Mg calcite, low-Mg calcite, or aragonite. Both aragonite and high-Mg calcite are metastable under normal conditions and convert to stable low-Mg calcite.

(*) HECKEL (1972) defines "stromatactis" as "...carbonate spar filled cavity in calcilutite which has a typical flat to smoothly curved base and an irregular top that is not sheltered by larger rigid grains (Ibid., p.8)". Their origin is still unkown. Both inorganic and organic processes have been used as an explanation. BATHURST (1959) pointed out that similar features termed "reef tufa" or "stromatactis" in the Carboniferous of England were resulted from decay of soft organisms. This idea was supported by PHILCOX (1963) and developed further by LEES (1964). On the other hand SCHWARZACHER (1961) argued that stromatactis were opened inorganically by sediment creep, slumping or compaction. SHINN (1968) concluded that the more nearly spherical openings were caused by trapped gas bubbles and that the less regular planar forms were caused by shrinkage. According to HECKEL (1972) stromatactis cavity can be formed by nonuniform compaction during initial unmixing and segragation of water from an originally partly water-supported accumulations of pure lime mud.

SORBY (1879) described the in situ replacement of aragonite by calcite. FRIEDMAN (1964) described the in situ replacement of high-Mg calcite and the dissolution of aragonite in modern shells. BATHURST (1958; 1964 and 1971) showed that both solution of aragonite and in situ replacement of aragonite by calcite in the molluscan shell occurs. BATHURST (1964) further considered that the process of in situ replacement is either by direct inversion, or across a thin water film and this process seems, at least in some limestone, to have begun before the sediment was cemented to a rigid framework, while shell fracture was still possible. On the other hand FOLK (1965) suggests that an original calcite fabric can undergo recrystallization, resulting in the birth of a completely new fabric WILSON (1967) recognised these difficulties and suggested that genetic classifications of fabrics erected by BATHURST (1958); ORME and BROWN (1963) and FOLK (1965) were too difficult to apply to shell fabrics. He therefore erected a special threefold division of fossil shell fabrics based on the geometric shape and arrangement of their constituent grains. Type 1 is characterized by inclusions showing to a varying degree the original structure of the shell with irregular sutured grain boundaries. Type II shell fabrics are similar to Type I but the inclusions are scattered and does not show original structure. In Type III fabrics grain boundaries are straight and grain size increases away from the shell wall. They all show the genetic features of the void filling calcite. No genesis is implied in WILSON's classification and doubt is therefore cast on its use.

Under the light of present literature it can be argued that an original aragonite shell can travel along two different diagenetic paths, that is solution-redoposition as calcite and in situ inversion. The latter is thought to be a process of neomorphism and will be discussed under that heading (p.174). DODD (1966) distinguished four types end product in aragonite solution and redeposition as calcite process depending on the relative timing of the solution of aragonite skeletons and skeletal fragments, lithification of carbonate mud matrix, and deposition of the sparry clcite cement. If solution of the aragonite skeletons and skeletal fragments occurs before lithification of the carbonate mud matrix, the resulting void will collapse. This process could normally leave no evidence observable in thin section. An exception can be seen if the outside of the skeletons were coated with an non-aragonitic encrusting organism or a thick micritic envelope which did not dissolve with the enclosed fossil (Pl. 49). A second type occurs before the lithification of carbonate mud again, but after the sparry calcite infill of the original voids with the solution of aragonite. Resultant void will collapse as in above type but will leave patches of sparry calcite within matrix. This type is possible but very difficult to prove in thin section. Solution-redeposition may occur after lithification of the carbonate mud matrix but before any sparry calcite has formed in the originally existing voids within fossils; resulting in fossil replacement without preservation of the outlines of internal structure. This is the most common inversion process described in th geologic literature and also the most encountered one in the present study (Pl.).

If sparry calcite had started to form in original voids within fossil before the solution process begins replacement retaining outlines of internal cavities will result. This type of inversion is characterized by sparry calcite growing outwards in opposite directions from a common surface (Pl.).

TALBOT (1972) has studied the preservation of Scleractinia Corals (Hexacorals) and came to the conclusion that preservation was influenced by differences in the structure and chemical composition of the skeleton and by diagenetic effects.

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PART 7

NEOMORPHISM

FOLK (1965) introduced the term "neomorphism"which refers to all transformations between one mineral and itself or a polymorph; and embraces the processes of inversion (aragonite to calcite), recrystallization (calcite to calcite) and strain-recrystallization (strained calcite to unstrained calcite). This term does not include the process of cementation (simple pore filling) or replacement (change in gross composition). In ancient carbonate studies, such as the present one, this term is particularly useful, since it is almost impossible to tell which of the lithological constituents were originally aragonite and which were calcite.

Neomorphism usually results in increase of crystal size (aggrading neomorphism) such as when a carbonate mud goes to neomorphic calcite(*). Occassionally, the crystal size decreases (degrading neomorphism) as when oolites or fossils degrade to microcrystalline calcite, but this process is much more rare. Aggrading neomorphism is the major neomorphic process in the Gürün sediments. Almost the whole column of FOLK's aggrading neomorphic calcite types participates in the compsotion of the sediments investigated. Aggrading neomorphism may be either porphyroid or coalescive (FOLK 1965). In porphyroid neomorphism a few large crystals grow at the expense of a static matrix, thus a porphyroid fabric with large and small crystals is in existence during intermediate stages of the process. In coalescive neomorphism most of the crystals are either grwoing or being consumed, i.e. the whole mass is changing like the growth of bubbles in a soap foam;

(*) BATHURST (1958) called this process "grain growth" which is a term used in metallurgy. Grain growth is defined by metallurgists and acts in monomineralic fabrics of low porosity. The intergranular boundaries migrate, causing some grains to grow at the expense of their neighbours. The reaction takes place in solid state, ions being transferred from one lattice to another without solution. Larger grains tend to replace smaller grains and a fine mosaic is gradually replaced by a coarser one. As grain growth proceeds, many of the enlarged grains are themselves replaced by their more successful neighbours. FOLK (1965) presented a convincing argument for abandoning the term, which is purely metallurgical, on the ground that this process does not take place during diagenesis. Instead he recommended the use of the term "aggrading recrystallization" or "aggrading neomorphism" to describe this type of grain enlargement. BATHURST (1964; 1971) abandoned the use of the term "grain growth" and its use has declined; although some authors are still using it (e.g. CHILINGAR, BISSELL & WOLF; 1967; HOROWITZ & POTTER, 1971; PETTIJOHN, 1975).

a uniform crystal size is maintained throughout the mass though the average crystal size gradually grows larger (Fig. V - 7).

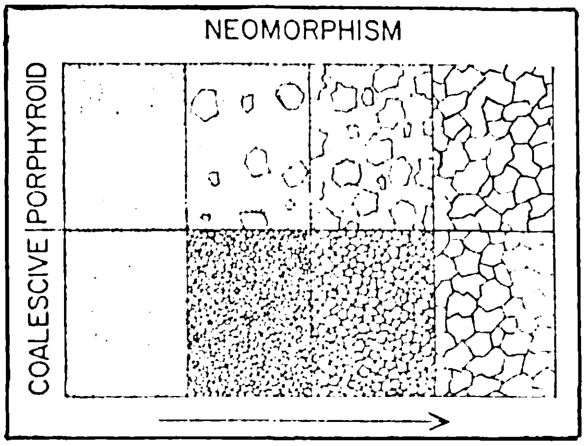

Fig.V-7 : Types of neomorphism (After FOLK, 1965)

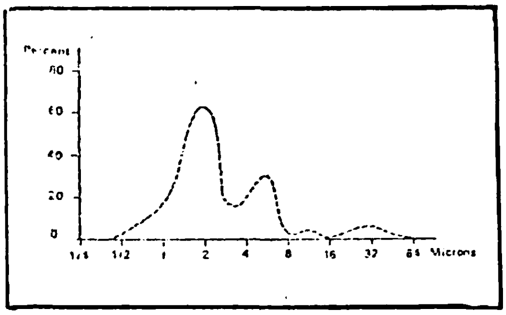

Fig.V-8 : Grain-size of the carbonate mud matrix (Simplified from FOLK, 1965)

Porphyroid neomorphism appears to be a much more common process, but coalascive neomorphism may be the mechanism by which microspar(x) is formed. (It is usually very equigrained)(FOLK, 1965). The importance of calcite-to-calcite recrystallization is an unsolved point. In FOLK's opinion this occurs important in some localities and at some stratigraphic horizons, but its overall volumetric importance is minor.

As noted above, in general, neomorphism leads to the coarsening of the rock fabric. But in a few special cases diagenesis leads to "grain diminution" or the conversion of larger crystal elements to a

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(x) An arbitrary division can be made between two distinct types of neomorphic sparry calcite on the basis of crystal size. The first category includes the "microspar" which ranges between 5-30 micron in size. The name may be seen rather unsatisfactory. "Spar" has for long described freely grown calcite precipitated from solution. FOLK (1959, 1962) himself adopts this meaning in his limestone classification. People not familiar with FOLK's work assume it refers to small crystals of sparry calcite. To knowledge of the present author, so far a more satisfactory term has not been offered for "microspar" and it seems meaningless to increase the term jungle. Therefore in the account to follow the term "microspar" will be used for the neomorphic calcite in the range of 5-30 micron size. Normally the size of "pseudospar" (second category) ranges between 30-100 micron.

mosaic of much smaller grains. Some evidence has been found in the Gürün carbonates for the alteration of algal masses to micrite. WOLF (1965) called this type of diagenetic formation of micrite, "grain diminution". Though the exact mechanisms involved in algal-grain dimunition are not certain, solution and degrading recrystallization are the two most likely processes responsible for the change. Plate shows the change in algal colonies from clear recognizable algal material to micrite in which few or no algal features are visible. In the same way, although less evident, some of the micrite observed in the sediments investigated may also be due to grain-diminution of algal colonies. However, as pointed out by WOLF (1965) it is usually difficult to tell whether a former cellular or filamentous algal colony changed diagenetically to micrite, or the micrite is a direct algal precipitate. The abundance of micrite postulated to have originally been algal in origin suggests that algal-grain diminution may have been a very important process.

Three aspects of neomorphism are important:

(1) Conversion of carbonate mud to lithified micrite;

(2) Formation of microspar calcite;

(3) Development of coarser neomorphic calcite (pseudospar).

I. CONVERSION OF CARBONATE MUD TO LITHIFIED MICRITE: Modern carbonate muds and presumably ancient ones are originally laid down as soup of tiny calcite (as described by DAVIES (1970) from Shark Bay Western Australia) or aragonite grains (CLOUD, 1962; The Bahamas), much of the latter in the form of 1-4 micron needles. The normal course of neomorphism is for the aragonite to invert to calcite, for magnesium calcite to expell the magnesium ion; and for the entire sediment to be converted to a mass of subsequant 1-3 micron calcite polyhedra. This process involves digestion of the vast majority of aragonite particles of similar length but much more slender than the polyhedra. FOLK (1965) is of the opinion that porphyroid neomorphism is responsible from the conversion of carbonate mud to lithified microcrystalline calcite, where grains starting from widely-spaced nuclei grow to replace the original mass of tiny needles of plates as well as filling up the pores in between. According to the same author, this process is one that probably goes on in all limestones that contained any original carbonate mud; he notes that "... even the finest of carbonate mudstones, or micrite are the result of important neomorphism (Ibid.p.28)". The arisen question is whether did the former aragonite needles inverted to calcite while retaining the needle form, and later fettened, or did the former aragonite needles fattened and then inverted to calcite. HATHAWAY and ROBERTSON's (1961) experiments shed light upon this problem. They subjected wet aragonite mud from the Bahamas to various temperatures (up to 400 C^{o}) and pressures (up to 3500 kg/cm^{2} which is equivalent to an overburden of about 10 km. of average crust), in a cylinder from which surplus pore water could escape as the mud compacted. They found that the general needle form was retained after the mud had inverted to calcite, although the ends became somewhat blunted. Therefore it is likely that inversion takes place while the needle mud is rather porous.

The grain size of lithified carbonate mud or micrite was questioned by FOLK (1965). His measurements of the grain size of the carbonate mud matrix in 100 samples of limestones, each one from a different names stratigraphic formation revealed that the distribution was bimodal. The most common grain size of the micrite was 1.5 to 2 micron. He also found a secondary peak at 4-6 micron and a lack of grains from 3 to 4 micron in size (Fig. V - 8). This gap has been termed as "micrite curtain" by FOLK (Ibid.). The magnesium ions expelled from the high-Mg calcite during course of neomorphism are retained interstitially in the rock, probably attached to the surfaces of calcite polyhedra; and form a sort of "cage" around each calcite crystal. Their distortion of the lattice prevents normal growth beyond the equilibrium size of 2-3 micron (FOLK, 1974). Thus, small amounts of interstitial magnesium ions cause the world wide uniformity of crystal size in micrite. Removing this cage is the key to forming microspar.

If the mud were originally calcite, the process probably was similar except for the inverstion stage.

Work of WACHS and HEIN (1974) on Franciscan Limestones (California, U.S.A.) showed that the micrite of these rocks was composed of two primary parts; that includes coccoliths or coccolith fragments and angular to subrounded calcite grains (Nannoagorite of HONJO 1969; See p. 46). The similar size of coccolith plates and average diameter of calcite grains; 2.5 and 2.6 micron respectively, suggests that these plates are the source of calcite grains. Therefore they conclude that calcite grains in the micrite are probably recrystallized coccolith fragments. They also showed coccoliths in various stages of recrystallization. Original pore space between coccoliths is reduced or totally filled by growth of coccolith plates during recrystallization and a uniform calcite mosaic is formed. An advanced stage of re-crystallization is characterized by lack of recognizable coccoliths and the micrite is composed of calcite crystals with well developed cleavage (WACHS and HEIN, 1974; p.1223).

A texture thought to be related with lithification of the carbonate mud is the "grumeleuse texture". It was first described by CAYEUX (1935, p.271) and is characterized by clots or clusters of clots of microcrystalline carbonate surrounded by a matrix of sparry calcite. Usually clots touch one another in many places. It may resemble the clotted texture caused by dedolomitization and described by SHEARMAN et al (1961) and EVAMY (1967) (See p.199). In some cases the development of texture did not expand significantly and were arrested for some reason, resulting in relatively coarsely crystalline but isolated centres surrounded by the unaltered matrix. The boundaries of micrite clots are not sharp. The question is whether microcrystalline islands are modified pelletoids in a sparry calcite cement or are they the result of incomplete recrystallization of a fine-grained limestone. CAYEUX (1935) suggested that grumeleuse texture was the result of the growth of calcite crystals throughout the mass of an original homogenous microcrystalline calcite and gradual differentiation forming a coarsely crystalline matrix with discrete residual clots of microcrystalline calcite. On the other hand, BEALES (1956 and 1958), based on his work on palaeozoic pelletoidal limestones of Alberta (Canada), defended the contrary view. According to him, many pelletoids appear to have merged on recrystallization into a homogenous microcrystalline rock which is not easy to differentiate from calcilutites. The grumeleuse texture problem has been discussed in some extent by BATHURST (1969 and 1971). Like CAYEUX, he is of the opinion that grumeleuse texture initiates by the growth of sparry calcite in a homogenous carbonate mud (or mudstone). On the other hand diagenetic processes reduce the number of pelletoids and it is followed by merging of soft pelletoids. The relatively coarsely crystalline but isolated centres surrounded by the unaltered matrix have been observed in the present work. This texture is probably the result of incomplete course of grumeleuse texture. If this assumption is true then it can be argued that the evolution of grumeleuse texture follows the path described by BATHURST (1969 and 1971). In other words the clots are microcrystalline calcite relics in a neomorphic sparry calcite and not mechanically deposited pelletoids in cement.

II. FORMATION OF MICROSPAR: Mechanism responsible from the formation of microspar has caused much contraversy in the last fifteen years. In his outstanding review of recrystallization in ancient limestones, FOLK (1965) supported a neomorphic origin. He wrote "...microspar is not simply a coarser mechanically deposited silt, but is definitely a product of neomorphism (Ibid.,p.40). According to him, neomorphism i.e. replacement in the solid sate, probably with the aid of interstitial solutions is the only possible process to account for destruction of vast number of grains, and their incorporation into larger crystals. His evidence for neomorphic origin can be summarized as follows:

(a) The often "patchy" appearance of this type of calcite, which usually bears no relation to the bedding. Sometimes it is graditional with micrite (Pl.56D), at other times the contact is quite sharp.

(b) Microspar commonly forms an auresole around the various grains with preferred orientation of crystals. Surely they cannot be detrital. On the other hand they simulate void filling fabric. According to FOLK"... the only fundamental thing is that any process starting from a surface and progressing outward should yield such enlarging and systematically oriented grains; no matter whether it forms crystal growing in a pore or encrusting a sand grain, or recrystallization in a carbonate mud starting from quartz grain, glauconite pellet, shell fragment or intraclast as a nucleus. (Ibid., p.44).

(c) Some faecal pellets embedded in microspar have been replaced by identical microspar so that the only remaining evidence of their existence is an elliptical, brown organic stain.

(d) The usual occurence of clay-grade non-carbonates, generally clay minerals, as impurities in microspar indicates a hydraulic equivalent to clay, FOLK argued that microspar was originally deposited as a clay-sized carbonate and not silt-sized carbonate. The validity of this evidence has been questioned by BATHURST (1971) who states that "...one of the special features of carbonate sediments is that they are (or were) never mixtures of grains in hydraulic equilibrium,... being locally formed, and not dependant on water transport...(Ibid., p.514). On the other hand according to BATHURST (Ibid) microspar which is deposited on the floors of fenestral cavities in bird's eye type micrite (including DUNHAM's (1969) "syndiagenetic vadose silts") is not a product of neomorphism; but a mechanically deposited silt. The mosaic of microspar occupies the floor of a cavity, is covered by a pore filling calcite cement (Pl.42) and the fillid cavity is enclosed by micrite. But FOLK (1974) defended his view points regarding the neomorphic origin of all microspar, based on the fact that in the fresh water environment little magnesium ion is present. The favoured crystal size of calcite in a magnesium-free environment appears to be about 5-10 micron. Above this it becomes more anhedral. FOLK noted that euhedral to subhedral calcite rhombs of about 5-10 micron diameter are typical in some nodular caliches; they crystallize on the surface of some creeks and as algal crusts in streams; occur as vadose cements in some eolian carbonates and he also argued that DUNHAM's (1969) vadose silt is apparently made up by 5-10 micron size and perhaps because it is recrystallizing in a solid, unyielding mass only forms imperfect, loafish rhombs (FOLK, 1974; p.51). CHOQUETTE (1968) showed that microspar had isotopic compositions related to fresh water, while micrite in unaffected limestones had normal marine ratios.

As noted by FOLK (1974) if for any reason the surrounding becomes very low in magnesium during the conversion of carbonate mud to lithified micrite, the calcite crystals are freed of their imprisoning cage of magnesium ions. They can burst through the micrite curtain; and by porphoroid neomorphism transform to microspar or even pseudospar. Magnesium ions may be removed from the micrite by several processes, all of which rely on the presence of interstitial fluids as a transport medium. Probably the most important mechanism for removing magnesium ions from the micrite involves clay minerals. Several clays, particularly chlorite and montmorillonite are known to absorb magnesium ions from sea water (DUNOYER DE SEGONZAC, 1970) and apparently can also absorb magnesium ions in the subsurface environment during diagenesis. Presence of these clay minerals in micrite greatly facilitates the formation of microspar by removing the "cage" of magnesium ions from the calcite crystals in the micrite. Some samples, formed by carbonate mud mixed with clay clasts during a period of turbulence, most probably a storm, show results of this process. By acting as a magnesium ion absorber, the clay liberated the 2 micron calcite grains in the micrite and allowed them to recrystallize to coarser size, microspar and even pseudospar. Grain size of the recrystallized micrite gradually decreases away from the clay clast because the ability of the clay to attract magnesium ions decreases with distance. Smaller clasts have been observed to have smaller neomorphosed aureoles surrounding them (Pl. 57).

Usually microspar in the lime mudstones of Gürün sediments tends to be concentrated in burrows and in cluster of fossil fragments. In this case there is no indication of absorbtion of magnesium ions. Porosity in the microenvironment of the burrows and clusters of fossils may have been the controlling factor. Percolating fresh water was probably more active in the burrows and clusters of fossil fragments and facilitated flushing of the magnesium ion "cages". On the other hand, deposition of the original sediment in a brackish environment with initially low magnesium is another cause of surroundings in low magnesium and produces microspar (FOLK, 1965).

CHAFETZ (1972) has described the importance of outcrop weathering in neomorphism of micrite, and FOLK (1974) considers this to be one of the most important processes forming microspar. Although many Gürün outcrops have suffered extensive weathering, only minor neomorphism can be attributed to weathering. Fresh samples contain microspar nearly identical to that in deeply weathered samples.

The possibility of microspar being the crystals which are released by the breakdown of skeletal carbonate to its ultimate components has been ruled out because the component crystals are finer than microspar. An exception to this is the echinoderms with some of the larger molluskan prisms which may be mistaken for microspar product of neomorphism.

Recognition of microspar from normal micrite is simply by its crystal size (4 micron is a convenient boundary). Electron microscope observations by SHOJI and FOLK (1964) showed that there was no difference in crystal morphology between microspar and normal micrite, except the grain size. Both types form subequant rounded to polyhedral blocks with slightly curved to plane faces (See (FOLK, 1965; Fig.10,p.33). Criteria for distinguishing microspar from coarser pore-filling sparry calcite have been given in the preceeding section (See p.155).

III. PSEUDOSPAR: Neomorphic pseudosparry calcite mosaic is encountered only in grain rich rocks with minor amounts of interstitial carbonate matrix. The pseudospars are equant to anhedral in shape and in some cases mimicks closely the appearance of pore-filling sparry calcite, (See p.160). Pseudospar growth leads to partial or complete obliteration of original structures and textures in the rock. They show highly irregular distribution of crystal size in the grains and surrounding matrix (Pl.56) being small in the former and larger crystals in the latter. Diagenetic studies of Recent carbonate sediments indicate that non-skeletal aragonite undergoes very slow inversion to low magnesium calcite as compared to its skeletal counterpart (LAND, 1967). The difference in crystal size of pseudospars in grains and surrounding matrix is thought to be due to variation in composition of the $CaCo_3$

polymorphs in the original sediment. Details of pseudospar genesis are given by FOLK (1965) who distinguished two different types: Pseudosparry neomorphic calcite formed by a) Invers ion of aragonite skeletons b) Neomorphism of carbonate mud.

A. PSEUDOSPAR FORMED BY INVERSION OF ARAGONITE SKELETONS:

Bathurst (1964) suggested two processes for the inversion of aragonite to calcite. I.- In situ transformation (In the presence of liquid films) and II. - Solution-deposition. He also suggested petrographic criteria for the recognition of each type.

The process by which in situ transformation is achieved has been discussed by FYFE and BISCHOFF (1965) in great detail and it does not include a large void stage. It is suggested that the catalytic influence of the water is great and reduces the activation energy which must be exceeded for the inverion to proceed. At first the reaction accelerates. This is consistent with a twofold process involving (1) the steady production of nuclei; and (2) continuous growth of existing nuclei. FYFE and BISCHOFF noted that the first calcite to nucleate was always in the prism zone of the aragonite. In shells that have undergone this type of inversion, the calcite generally is medium to very coarsely crystalline, commonly elongate, and crystal boundaries are usually irregular. Relics of original shell structure may still be visible in the form of inclusions within the calcite mosaic (Pl.53D).

Inversion by solution-deposition does involve a void stage but is complex. Details of it can be found in BATHURST (1971). This type inversion has been considered in the realm of cementation and discussed in that part (See p.161). In fact, the grain size, shape, fabric, and orientation of calcite formed by invers ion of aragonite skeleton fragments by solution-deposition process very closely resembles that of void filling calcite.

B. PSEUDOSPAR FORMED BY NEOMORPHISM OF CARBONATE MUD.

Carbonate muds can and do continue to recrystallize (or invert) to pseudospar. FOLK (1965) distinguished three indistinct evolutionary paths. First variety of pseudospar has crystals that are equant and uniform and resembles microspar closely (Pl. 56). 30 micron boundary is accepted by many workers although it does not have any genetic significance. The second kind pseudospar is characterized by very irregular boundaries between individual crystals which sometimes may be suture-like and saw toothed (Pl. 59). Usually twin-lamellae and cleavage surfaces are bent. These properties are typical of a "strained calcite" described by DEMIRMEN and HARBAUGH (1965, p.143). In one sample individual pseudospar crystals have been seen showing the outlines of former crystals which indicates the conversion of small crystals to larger ones (Pl.59C). A third type pseudospar closely resembles pore-filling spar (ie. sparry calcite). They may form neomorphic overgrowths and crusts on fossils and other grains (Pl. 54). Sometimes they form a completely random mosaic of pseudospar. Crystal boundaries are straight and they may show features of pore-filling calcite. Criteria for cementation and neomorphic origin are given in p.155. Neomorphic origin can be proven if the crystal can be shown to transgress grains and matrix more or less indiscriminately.

IV NEOMORPHIC CALCITE FABRIC TYPES:

As described by FOLK (1965) neomorphic calcite, like sparry calcite formed by cementation, may be found in three different forms in the sediments: Overgrowth, crust and random. The crystals may be equant, bladed or fibrous.

A. NEOMORPHIC OVERGROWTH:

Some neomorphic calcite occurs as obviously optically oriented overgrowths on skeletal grains where the organic fragment and the neomorphic calcite extinguish together. It may be formed on monocrystalline nucleus such as echinoderm fragments; or on regularly fibrous fossils such as trilobites, ostracods, and brachiopods (as described by FOLK 1965) or on calcitized molluscan shell wall. Although the first type is common (ie. neomorphic overgrowth on monocrystalline nuclei), the second type has rarely been observed in the present study.

As pointed out above, neomorphic overgrowth on monocrystalline nuclei is the abundant type and is usually associated with the wackestones. This process has been called "Syntaxial Rim (BATHURST, 1958 and 1971)" and "Syntaxial Replacement Rim (ORME and BROWN, 1963)". BATHURST (1958) described syntaxial rims around echinoderm debris in micritic Dinantian Limestones of North Wales and Yorkshire, and considered the rims had formed by grain growth (Aggrading neomorphis of this account). ORME and BROWN (1963) also concluded that syntaxial replacement rim was product of recrystallization resulting in grain enlargement. On the other hand EVAMY and SHEARMAN (1965) observed echinoderm fragments with overgrowth in fine-grained matrix similar to ones described by BATHURST; ORME and BROWN; and encountered in the present study. But they came to the conclusion that they were of cementation origin and the fine-grained material was entered the sediment subsequently. Although neomorphic echinoderm overgrowth resembles superficially an echinoderm overgrowth formed by cementation; it has quite different fabric relationships.

(1) The neomorphic overgrowth, or the host where it does not have an overgrowth, is in contact with a matrix of microcrystalline calcite (Pl.54A) or with overgrowths of other skeletal fragments (Pl.54B).

(2) The neomorphic overgrowth has irregular outer boundary and may be produced into sharp spines.

(3) Patches of the matrix occur as inclusions within the overgrowth (Pl.54).

On the other hand its association with wackestones rules out the possibility of them being cementation products because the echinoderm fragments float in a lime mud, and the mud can not has arrived after mechanical deposition and partial cementation of the echinoderm fragments. Therefore the overgrowths observed on the echinoderm fragments embedded in the microcrystalline matrix are not of cementation origin but of neomorphic origin.

Neomorphic overgrowth outside echinoderm fragments is very very rare in the Gürün sediments. The unique example which could be interpreted as neomorphic overgrowth on polycrystalline nucleus has been found in a thin section of silicified limestone. Here, neomorphic overgrowth (?) crystals in the shell wall extends syntaxially into the adjacent matrix which is now completely silicified (Pl.54D). Neomorphic overgrowth in calcitized molluscan shell has been described from Pacific Atolls, Eniwetok and Guam (SCHLANGER; 1963 and 1964) and FUNAFUTI (See BATHURST 1971 p.353).

Under the light of observations of the present work the author concludes that the formation of neomorphic overgrowth is an insitu process which can be attributed to aggrading neomorphism resulting in an increase of crystal size.

B. NEOMORPHIC CRUST:

In some samples neomorphic equant and/or bladed calcite has been found encrusting some surfaces forming a fringing crust and replacing micrite. It is physically oriented but is not obviously in optical continuity with the nucleus. This type of neomorphic calcite is well illustrated by FOLK (See his figs. 13 A and B. p.43; 1965). Similar calcite crystals have been observed in the present study (Pl.55A).

FOLK (1959) argued that this type of neomorphic calcite (usually pseudospar) is devoted to early phase of recrystallization process and completion of process would yield rock with grains floating loosely in neomorphic calcite. Sample 723 seems to support FOLK'S hyphothesis. In this sample all the grains are fringed by neomorphic pseudospar, which also can be seen growing towards the centre of skeletal fragments. Some of the skeletal fragments are completely neomorphosed by pseudospar (Pl.55A). The outline of former shells have been preserved as micritic envelopes. Micritic matrix in places has been replaced by random neomorphic calcite (Pl.55). This sample is thought to characterize a more advanced stage of neomorphism.

Some fibrous calcite has been observed replacing carbonate mud to produce crusts around some grains, even microspar. This type of fibrous calcite has been discussed in detail by ORME and BROWN (1963) and clearly shown that it is a product of neomorphism and not a direct precipitate. Among the other evidence they cited the cross-cutting relationships of fibrous calcite to organic and inorganic structures; and grading from micrite through silt size equant grains to fibrous calcite (Ibid. p.55). Very similar type of fibrous calcite has been found in the present study (Pl.55). Although any cross-cutting relations or grading has not been evident, their overall appearance and association with lime mudstones strongly suggests a neomorphic origin. They are generally elongate and broad featured. This is probably due to the fact that the growth of crystals outward into the micrite is faster than the lateral or inward growth at the expense of nuclei. Sometimes this type of fibrous calcite has been found to exhibit no visible nuclei (Pl.55). The crystals start to grow from a centre of crystallization and expand out to form a spherulite (See: MUIR and WALTON, 1957; FOLK, 1965, p.26; PETTIJOHN, 1975, p.368). A special type of neomorphic fibrous calcite, with a core of equigranular microspar, surrounded by elongate, sparry crystals with their long axis radially arranged, has been termed as "stellate" by BATHURST (1971; See his fig. 333, p.482) and has also been observed in some samples of present study (Pl.56A).

C. NEOMORPHIC RANDOM CALCITE:

Much of neomorphic calcite is not evidently oriented by any surface, it is a simple random mosaic, and it can start anywhere. Crystal size of randomly oriented neomorphic calcite may change abruptly. Sometimes micrite, microspar and even pseudospar can be seen in the same thin section (Pl.56D). MARSCHNER (1968) studied the distribution of calcite grain size in some Triass Limestones from N.W. Germany and concluded that crystal size is inversly proportional to clay mineral content suggesting that the relationship is "...... probably due to the protective envelopes of the clay minerals during recrystallization. Organic substances and envelopes of metallic hydroxides have a similar effect (Ibid. p.56 - 57)". A limestone relatively free of insoluble residue would be able to recrystallize completely and produce a coarser mosaic, where a limestone relatively rich in insoluble residue would have less chance of crystal-enlarging due to the material around the calcite grains acting as an obstacle to enlargement. In the sediments investigated, it has been observed that where neomorphic calcite is formed by large crystals (ie. pseudospar) the appearance of crystals is clear; where as the small neomorphic calcite crystals are seen cloudy under the petrographic microscope (Pl.).

In one of the samples, No. 441 (Section 4A: Kurucaoba), some euhedral calcite crystals have been observed arranged in a line (Pl.). They are quite different from the pseudospar encountered in other samples. The present author is of the opinion that these calcite crystals are of very late diagenetic origin and formed by neomorphism. Probably a joint in the rock gave way to the percolating fresh water and facilitated flushing of the magnesium iron "cage".

V TIME OF NEOMORPHISM:

The relative times at which various neomorphic changes took place in the Gürün sediments can not be established with any degree of certainty on the basis of present data. BATHURST (1971; p.481) suggests that the growth of neomorphic sparry calcite begins in partly consolidated sediments. The inversion of aragonite to calcite is known in Recent carbonate sediments. It is possible that the development of neomorphic sparry calcite, the inversion of aragonite to calcite in colites and aragonitic shells and neomorphism of carbonate mud of the Gürün sediments were relatively early diagenetic events. Subsequent recrystallization of colites and the recrystallization of shells may have occurred somewhat later.

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PART 8

DOLOMITIZATION

I. TERMINOLOGY:

In this study the term "dolomite" refers only the mineral species dolomite, Ca Mg $(Co_3)_2$. Although the same term has been widely applied by many workers (e.g. BLATT, MIDDLETON and MURRAY 1972) to the carbonate rocks that contain more than 50% of the mineral dolomite, the use of the same word for the mineral and rock may cause confusion. Therefore a rock containing more than 50% of mineral dolomite is in this account called "dolostone". On the other hand "dolomitization" is a process by which limestone is wholly or partially converted to dolomitic limestone or dolostone.

II. FIELD CRITERIA:

Although not abundant, dolomite is an important mineral in certain parts of the investigated area. Some field criteria are useful for dolomite identification. They may be summarized as:

1. Yellowish coloured weathering surfaces which are usually associated with iron rich dolomites;
2. The colour of bed: Reddish, yellowsih colour usually indicates the presence of iron-rich dolomites;
3. Texture: Most dolomite crystals are coarser compared with the finely crystalline micritic limestones. They usually show perfect rhombohedra which may be easily seen in the field by a hand lense.
4. Lack of fossils: This criterion is only valid for dolostones where intense dolomitization erases texture of former fossils. One should be careful applying this criterion, because unfossiliferous limestones are not rare. In this case, the staining techniques and treatment with dilute HCl are useful.

III. TYPES OF DOLOMITES:

Dolomites found in the investigated sediments consist of scattered, euhedral rhombs in some of the micritic limestones and in the internal sediments of certain voids or shell cavities. This dolomite is believed to represent either primary precipitation or very early diagenetic event. Another type of dolomite occurs as scattered crystals or a mosaic that replaces matrix, allochems, and fossil fragments. This dolomite is definitely a replacement mineral. Proof of replacement origin of dolomite consists of:

1. Skeletal remains now composed of dolomite, which were originally secreted as aragonite or calcite;

2. Dolomite crystals which preserve relics of distinctive calcium carbonate sediments (e.g. oolites, pellets, etc.).

3. Dolomite patches in limestones (FRIEDMAN and SANDERS, 1967).

The earliest formed dolomite is usually referred to as "syngenetic" which is formed penecontemporaneously in its environment of deposition (i.e. Prior to lithification) as a dolomite mud. The other form of dolomite is secondary. It is usually called "diagenetic" and may occur during early stages of diagenesis or later in the history of the rock. (Fig. V - 9).

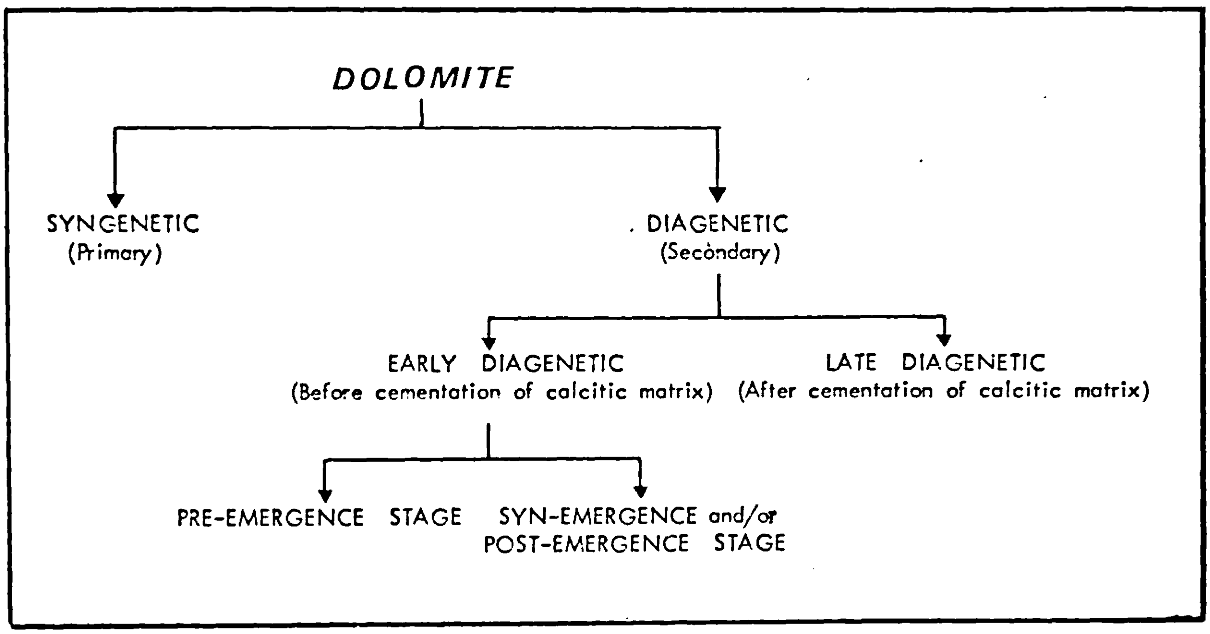

Fig.V-9: Genetic classification of dolomite formation (Mainly after BADIOZAMANI, 1973).

It is formed by replacement of the pre-existing $CaCo_3$ sediments. It is most likely that the majority of dolomites have a secondary replacement origin and only a small fraction of dolomitic rocks are considered to be primary precipitation. The distinction between syngenetic and diagenetic dolomites is often difficult to make. If dolomite had formed penecontemporaneously by replacement of grains or the tests of organisms, it would be considered "diagenetic"; but, if dolomite had replaced an aragonitic matrix between the grains or tests, criteria for recognition would be non-existent, and the distinction between two classes of dolomite breaks down (FRIEDMAN and SANDERS, 1967).

A. SYNGENETIC DOLOMITES: Petrographic criteria suggestive of syngenetic origin have been discussed by various workers (See BLATT 1976). Some of the criteria are listed below:

1. Very fine and uniform grain size (Often less than 10 μ.) However, it can be increased by a process of neomorphism.
2. Homogenous nature of the individual crystals and rock itself.
3. Association with evidence of intertidal environments (Algal mats)
4. Restriction to one horizon.

Dolomite does not precipitate direct from sea water although the water is many times supersaturated with respect to the dolomite. Also dolomitization of carbonate sediments in direct contact with sea water is rare (SONNENFELD, 1964; p.101). On a large scale it is only known in ancient rocks. On a small scale syngenetic dolomites form today mostly in isolated lakes and lagoons. Only recently was actively forming primary dolomite described from the Courong Area in South Australia by the members of the Department of Geology, Adelaide University, especially A.R. ALDERMAN, H.C.W. SKINNER, C.C. VON DER BORCH and B.J. SKINNER (For references see BATHURST, 1971 p.543). It is followed by discoveries of primary dolomite in supratidal carbonate crusts in Florida Bay (SHINN, 1964) and the Bahamas (SHINN, GINSBURG and LLOYD, 1965); in the sabkha flats of the Persian Gulf (ILLING, WELLS and TAYLOR, 1965) and on the marginal flats of Lake Balkhash (STRAKHOV, 1958).

B. DIAGENETIC DOLOMITES: Secondary or diagenetic dolomites may be divided into two categories: 1. Early diagenetic; 2. Late diagenetic. Early diagenetic dolomites form before cementation of calcitic matrix and late diagenetic dolomites form after cementation of calcitic matrix (BADIOZAMANI, 1973).

a. EARLY DIAGENETIC DOLOMITE: Two stages may be differentiated: Pre-emergence Stage and Syn-emergence and/or Post emergence Stage. The Pre-emergence Stage is defined as the product of early diagenetic metasomatism of calcium carbonate by magnesium ions taking place in freshly deposited original calcareous sediments and in which modified sea water is the medium of its formation. Dolomite crystals may start to form as replacement of both aragonite and high-Mg calcite during the process of stabilization of these two minerals. This is presumably favoured by the increase in the Mg/Ca ratio of the interstitial waters (LAND, 1967) as a result of incongruent dissolution(*) of the high-Mg calcite. At this

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(*) Dissolution of magnesium calcite can be either complete or incongruent. In the case of incongruent dissolution, some textural information is retained. It yields a replacement product of calcite plus a solution enriched in magnesium.

stage according to LAND, dolomite would be calcium rich. This is further supported by the discovery of calcian dolomite in the clay fraction of the bottom sediments in Lake Balaton (Hungary) by MULLER (1970) where precipitation of high-Mg calcite is taking place. Recently, MULLER and FORSTNER (1975) have reported the formation of recent dolomites from non-marine environments. In the lacustrine and cave environments dolomite formation seems to take place only at elevated Mg/Ca ratios if high-Mg calcite is available in the sediment. They argue that no trace of dolomite could be detected in the lake sediments, where aragonite is the only primary carbonate although the Mg/Ca ratio of these lakes could be up to 180. The same authors have explained the mechanism as follows:

Stage 1 - Due to their crystallographic similarity, the surface of high-Mg calcite acts as a matrix for dolomite nucleation after a Mg/Ca ratio favourable for the dolomite formation has been achieved.

Stage 2 - Further dolomite growth involves dissolution of high-Mg calcite and subsequent precipitation as dolomite (p.105).

Petrographic criteria suggestive of a Pre-emergence Stage of dolomitization are:

1. The characteristically poor preservation of fossils and the drusy calcite mosaic filling of former organic cavities or chambers, both indicating their original calcareous nature.

2. Relative coarseness of grains and moderately uniform texture of dolomite rhombs. The nucleation rate is high due to the easy access of Mg ions through the soft sediment.

Syn-emergence and/or Post-emergence Stage dolomites are formed in semi-consolidated sediments. They are characterized by the heterogenous texture and non-uniform grain size, varying between 100 μ. and several mm; existence of individual crystals, often zones and have more than one phase of dolomite formation. It is believed that although they were initially in communication with the overlying supernatant seawater, subsequently they were largely occupied by the pore solution which acted as a medium of formation. Petrographic characteristics suggest the existing of non-uniform conditions for dolomitization such as lack of uniform permeability which developed in later stages of diagenesis due to burial and over load.

b. LATE DIAGENETIC DOLOMITE: It is defined as replacement of limestone with the dolomite being localized by post-depositional structural elements. They are closely related to faults and fractures in carbonate rocks and sometimes are genetically associated with metallic ore deposits such as lead and zinc minerals (FRIEDMAN and SANDERS, 1967). Downward percolating meteoric solutions or rising hydrotermal solutions may be responsible from the formation of late diagenetic dolomites. They are easily identified in the field since they are restricted to local dolomite veins.

IV. SELECTIVE DOLOMITIZATION:

Two stages of dolomitization are apparent in the Gürün sediments. The earlier stage of dolomite occurs as silt sized, subhedral to anhedral grains that selectively replace fine-grained calcitic matrix. Little or no dolomite has replaced biogenic fragements in this stage (Pl.60). Evidence of selective dolomitization is found in partially dolomitized limestones and in intercaleted dolomitized and undolomitized rocks. The strong affinity between dolomite and mud-size carbonate particles in ancient rocks, has also been noted by MURRAY (1960), LUCIA (1962), POWERS (1962) and MURRAY and LUCIA (1967). SCHMIDT (1965) observed the following susceptibility sequence for calcite replacement to dolomite, in Jurassic Limestones in Germany (In order of decreasing susceptibility): 1. Matrix, 2: Aragonitic bioclasts, interclasts, pellets and oolites; 3. Magnesium calcite bioclasts. Original calcitic bioclasts such as foraminfera and echinoderm were least susceptible to dolomitization. He used this sequence to argue for contemporanity of the dolomitization process; that is, dolomitization took place before the aragonite had been inverted to calcite. In applying SCHMIDT's susceptibility sequence to ancient carbonates, one must assume, on the basis of analogy to modern shallow water carbonates (CLOUD, 1962), that the fine grained matrix was aragonite and high-Mg calcite, both metastable carbonates. If dolomitization was contemporaneous with deposition, or took place while the sediment was still soft (aragonite rich, intraclastic lime mud) one could expect aragonite to be the first mineral to be replaced by dolomite, as now occurs in the Persian Gulf. This may not always be a strict order because if part of the mud consisted of high-Mg calcite, then it would probably be dolomitized after the coarser aragonitic intraclasts and bioclasts. Beside the pre-dolomitization mineralogy, some other factors may control distribution of dolomitization. One of the factors is the particle size. Generally the finer particle is more susceptible to dolomitization than the coarser ones. On the other hand permeability is an important factor in causing selective dolomitization. Highly permeable parts of a sediment will allow the movement of interstitial fluids with consequent dolomitization of these parts. Generally the permeability of carbonate sediments is reduced, during subsequent cementation and compaction. Then the movement of interstitial fluids is limited and dolomitization is confined to bedding planes which consist of a permeable zone. SHINN (1968) found that relatively non-dolomitic limestone nodules float in lithified dolomitic sediments on a supratidal mud flat in the Lower Florida Keys. Storm tides periodically deposit layers of lime mud above normal high tide level, which dry and crack to form typical mud crack polygons. These polygons erode into flattened nodules that subsequently become buried in relatively more porous and permeable sediments. Magnesium-enriched brines concentrated through evaporation, are more readily transmitted through the more permeable sediment, producing features bearing a strong similarity to sedimentary structures commonly found in the geologic record.

Another factor controlling dolomitization is probably organic matter. It is believed that organic remains may play a significant role as serving as centres for dolomitization. The role of organic matter is manifold. Their decay may create locally favourable pH-Eh conditions. They may be food for the sulphate reducing bacteria, consequently producing local alkalinity. Another feature of organisms was discovered in the last decade. They precipitate dolomite. The dense axial zone of echinoid teeth contains calcian dolomite (SCHROEDER,DWORNIK, and PAPIKE, 1969). Organic processes may also effect dolomite formation in other ways. Supratidal dolomites at Andros Island are usually associated with stromatolitic algal mats (SHINN, LLOYD and GINSBURG, 1969). The blue-green algae within the mats are capable of concentrating magnesium, bringing the Mg/Ca ratio to 3 or 4 times that of sea water, thus offering an additional source of magnesium for dolomitization (GEBELEIN and HOFFMAN, 1971). This would also explain why some ancient stromatolites are only dolomitized in their darker layers (MILLIMAN, 1974).

The second stage of dolomite is composed of ferroan dolomite spar that selectively replaces grains. This suggests that the dolomitization of the second stage took place during a later time in the history of the rock. If dolomitization took place in an early stage (i.e. Pre-emergence Stage) it would be expected that the micritic matrix (mainly aragonitic) would be the first component to be dolomitized. The observed resistance of matrix to dolomitization suggests that the matrix was stable (calcitic) and cohorent when replacement began.

V. MECHANISM OF DOLOMITIZATION

The conditions necessary for dolomitization are:

1. Sufficiently porous and permeable calcareous sediment to act as host for the magnesium replacement.

2. A fluid of the correct chemical composition to react, capable of dissolving $CaCo_3$ and releasing magnesium.

3. A long enduring supply of magnesium.

4. A hydrodynamic head, to force great volume of water through the sediment (WILSON, 1974).

Several processes probably produce the dolomites observed in the geologic record. A summary of recent thoughts on dolomitization mechanism is given in BADIOZAMANI (1973). Most of the processes proposed so far are clearly associated with emergent platforms. As noted by MILLIMAN (1974) most modern dolomites occur on supratidal flats, such as those in the Persian Gulf, Florida and the Bahamas where the dolomite is restricted to sediments above the high tide level. Mechanisms proposed by various workers are roughly similar. Periodic floodings supply water to the supratidal flats. The waters sink into the sediments, but subsequent capillary action causes an upward movement of pore water. This capillary activity is caused by high evaporation rates at the surface. The most popular mechanism was first proposed by ADAMS and RHODOS (1960) and was called "seepage refluxion". Based on observations they had made on the Permian Reef Compex (Texas and New Mexico, U.S.A.) they postulated that evaporation in the overlying watermass, produces dense brine, which flows outward along the bottom. If the return flow along the bottom (reflux) of this brine is prevented by natural barriers, such as reefs or sills, then its only route of escape is downward interstitial flow through porous underlying carbonate sediments. The brine is inferred to have been an active agent of dolomitization. On the other hand DEFFEYES, LUCIA and WEYL (1965) have suggested that the seepage of dense lagoonal brines into the sea through porous limestones have dolomitized the pre-existing limestones on Bonaire (Netherlands Antilles). But LUCIA's (1968) later observations in the same island are at variance with the seepage refluxion hypothesis. A bore hole made through the floor of the lagoon showed that no dolomite was formed in the subsurface, and the pore water was normal sea water at a depth where a hypersaline brine would have been expected. He suggests that this is probably due to the presence of the impermeable clay layers and he notes that "... although reflux of hypersaline water out of the bottom of the lagoon is well documented, the flow path is still unknown (p.852)".

HSU and SIEGENTHALER (1969) postulated a reverse process of dolomitization which they called "evaporative pumping". Their assumptions are based on a series of preliminary experiments in the laboratory.

The authors concluded that in an arrid coastal area the interstitial water from the uppermost layer of the sediment evaporates and provides the energy for the landward movement of subsurface waters.

The role of mixtures of fresh water and sea water in affecting dolomitization of calcite is the basis of a new model for dolomitization put forward by BADIOZAMANI (1973). It is called "Dorag dolomitization model". According to this model fresh phreatic water with only a minor amount of magnesium, when combined with marine waters forms a fluid which may be undersaturated with respect to calcite, where as dolomite saturation increases continuously. He has shown that in brackish water "...in the range of 5-30% sea water, the solution is undersaturated with respect to calcite and many times supersaturated with respect to dolomite (p.969)". Therefore calcite can be replaced by dolomite. This model can explain the genesis of dolomites that lack evidence of supratidal origi n and are not associated with evaporites. BADIOZAMANI (Ibid) concludes that "...the fact that the mixing of fresh and sea water is readily attainable under many circumstances suggests that this type of dolomitization should be widespread and substantial volumes of ancient dolostones should be attributed to this category (p.979)". The present author is in total agreement with him and believes that dolomitization of Gürün sediments was caused by the mixing of sea water and ground water in the phreatic zone. Extensive sea water evaporation and high Mg/Ca ratios in solution are not likely for the sediments investigated.

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PART 9

DEDOLOMITIZATION

I. TERMINOLOGY:

Dedolomitization is a process of replacement of dolomite by calcite. The replacement may be complete or only partial. Dedolomitized rocks, or dedolomites, are known to be widely distributed and have attracted considerable attention in the last two decades (Notable examples are in SHEARMAN, KHOURI and TAHA, 1961; EVAMY, 1965; KATZ, 1968; FOLKMAN, 1969; AL-HASHIMI and HEMINGWAY, 1973).

SHEARMAN and Others (1961) trace the origin of the term "dedolomitization" from VON MORLOT, who identified it as the replacement of dolomite by calcite (1847). Later TEALL (1903) used the term for the metamorphic transformation of dolomite under sufficiently low pressure and high temperature (contact metamorphism) and it has since become established in metamorphic petrology. On the other hand, as noted above, the dedolomitization has also been observed by various workers in the field of sedimentary petrology, to occur in dolostones and dolomitic limestones at a relatively low temperature and pressure (non-metamorphic conditions). The use of the term "dedolomitization" has been accepted as a valid and expressive term in literature for the mechanism of dolomite replacement or destruction by calcite. But SWETT (1965) and SMIT and SWETT (1969) have questioned the validity of this term and offered to abandon it for the following reasons:

1. To have a replacement process on the basis of the replaced mineral instead of the replacing mineral, requires multiple names for a single replacement process. For example, a single phase of dolomitization may replace calcite, silica, glauconite, etc. To name this single replacement process by the host or replaced minerals would require names such as decalcitization, desilicification, deglauconitization and so on for the single process of dolomitization.

2. It is inconsistent with accepted nomenclature for other processes (for example, dolomitization, silicification, glauconitization, feldspathization, etc.).

3. It is ambiguous because a replacement process named by the host mineral does not directly imply which of a variety of mineral is forming at the expense of another. Several processes could conceivably be treated as dedolomitization.

4. It is geochemically misleading (Dedolomitization is effected by

the removal of magnesium not dolomite).

They also proposed the use of the term "calcitization" as a more logical, meaningful, and acceptable name for the replacement process of dolomite by calcite. The term "calcitization" proposed by SMIT and SWETT to replace "dedolomitization" was defined by TAYLOR (1964) as "implying an introduction of calcium carbonate as a replacement of another mineral which could be any mineral susceptible to replacement under certain conditions". It is clear from this definition that "calcitization" is a very general term which includes a variety of replacement processes. Calcitization of siderite, chert, glauconite, phosphate etc., are well known examples. The term "dedolomitization" represents only one type of calcitization. "Glossary of Geology and Related Sciences (AGI, 1972)" defines "calcitization" as follows:

"...The alteration of existing rocks to limestone, due to the chemical replacement of mineral particles by calcite, such as dolomite in dolomitic rocks or of feldspar and quartz particles in sandstones".

Therefore, "dedolomitization of dolomitic limestone" is more meaningful than saying "calcitization of dolomitic limestone". Because calcite is already present in the rock and the term here will be somewhat misleading.

On the other hand the term "dolomitization" is defined as "... The process whereby limestone is wholly or partially converted to dolomite rock or dolomitic limestone by the replacement of the original calcium carbonate (calcite) by magnesium carbonate (mineral dolomite)...(AGI,1972)". It is apparent from the above definition that the replacement by dolomite of any mineral other than calcite cannot be in any circumstances called "dolomitization". Dolomite replacing silica, glauconite etc., could not be accepted as dolomitization and therefore SMIT and SWETT's (1969) first argument is not valid.

The definition of "dedolomitization" is given as follows:

"...A process whereby, presumably during contact metamorphism at low pressure, part or all of the magnesium in a dolomite or dolomitic limestone is used for the formation of magnesium oxides, hydroxides, and silicates, resulting in the enrichment in calcite (AGI, 1972)". This is the original definition of TEALL (1903). It has also been added to this definition that "the term was originally used by VON MORLOT (1847) for the replacement of dolomite by calcite during diagenesis and chemical weathering". Irrespective of the conditions given in the definition, the term implies the destruction of only the mineral "dolomite". The alteration or destruction of dolomite always involves calcite as a main product. Removal of dolomite by silica, phosphate, glauconite, feldspar, or by solution, cannot be called "dedolomitization" contrary to the argument given by SMIT and SWETT (1969).

Lastly, the use of the term "dedolomitization" in published works (excluding metamorphic dedolomitization) has been fairly persistent. The present author is of the opinion that retaining the use of this term will avoid any complication or misunderstanding regarding this replacement process.

II. ENVIRONMENT AND CAUSES OF DEDOLOMITIZATION:

Several authors have quoted VON MORLOT's formula (1847) which describes the reaction as follows:

$$Ca\ Mg\ (CO_3) + Ca\ SO_4.2H_2O \quad 2CaCO_3 + MgSO_4 + 2H_2O$$

The dedolomitization reaction preceeds in the presence of sulphate ions. These ions could be derived from solution of gypsum or anhydrate (VON MORLOT 1847; SHEARMAN and Others, 1961; GOLDBERG, 1967) or from oxidation of pyrite (EVAMY, 1967; FOLKMAN, 1969). On the other hand some different types of dedolomitization which are not based on the dolomite-calcium sulphate reactions have been discussed by KATZ (1968, 1971) FOLKMAN (1969) and AL-HASHIMI and HEMINGWAY (1973). KATZ considered that calcian dolomites were partially susceptible to dedolomitization. In the case of Middle Jurassic Mahmal Dolomites of Israel, he suggested that dedolomitization took place either after the formation of a calcian dolomite crystal or at definite intervals during the growth of a dolomite crystal. In the latter case, the replacing calcite would constitute rhombehedral cores or zones within the dolomite. KATZ gave reasons for the preferential dedolomitization of calcian dolimites or of zones or cores in dolomite crystals which are relatively rich in calcium. Firstly, there is an increase in the free energy of calcian dolomites compared with "normal" dolomites. Secondly, there is ease of replacing calcium ions in the microcrystalline aggregate through the intercrystallite boundaries.

FOLKMAN (1969) described a type of dedolomitization in the sub-Recent to Recent weathering crusts of limestones (caliche) which forms in the meteoric weathering zone under certain climatic conditions. According to him, solution of dolomite by cold rain water and deposition of calcite from capillary water causes the formation of dedolomites.

The other form of dedolomitization was put forward by AL-HASHIMI and HEMINGWAY (1973). It was encountered in the Recent and sub-Recent, rusty looking weathering crusts of the ferroan (iron rich) dolomites of the carboniferous rocks of Northumberland. Ferroan dolomites are unstable under subaerial conditions and tend to break down by the action of circulating sea or fresh waters. The result of breakdown is formation of iron hydroxides and magnesian calcite in a porous aggregate.

Excluding the one proposed by KATZ (1968,1971) all other forms of dedolomitization are late diagenetic surface and near surface phenomenon. Experimental work of DE GROOT (1967) supports surface environments. He found in laboratory dedolomitization experiments that aragonite replaced dolomite under surface conditions. He concluded that for "effective dolomitization" it is necessary to have:

(a) a high ratio of water-flow to remove Mg^{++} and keep the Ca^{++}/Mg^{++} ratio high.

(b) a low CO_2 partial pressure, similar to that of the atmosphere and (c) a temperature below 50°C (p.1220). In other words, dedolomitization is most readily accomplished under surface conditions.

It was also noted by SHEARMAN and Others (1961, p.11) that TATARSKIY (1949), discussing the Russian occurrences of dedolomitized limestones stated that it is always a near-surface phenomenon and that borehole samples do not show evidence of dedolomitization at depth. FRITZ (1967) also found that the isotropic ratios in dedolomites from the German Jurassic are such as to indicate that the alteration was promoted by meteoric waters and that it is probably taking place at the present day.

Even if dedolomitization is accepted as being restricted to surface outcrops and is related to periods of exposure it is not normally known whether the dedolomites are of Recent or an ancient origin. MAKHALAEV (1957) found dedolomites in boreholes at depths of several hundred metres, but only in areas where they were overlain unconformably by younger sediments. SCHMIDT (1965) also found dedolomites in wells near an area where Lower Cretaceous shales transgressively overlie Jurassic beds and discovered that dedolomitization in the Jurassic rocks increases as the unconformity is approached. BRAUN and FRIEDMAN (1970) pointed out the importance of zones of dedolomitization as a possible indication of unconformity surfaces.

III. OCCURRENCE OF DEDOLOMITIZATION

During the study of carbonate rocks of the Gürün Area evidence of the existence of dedolomitization has been found in some of the dolomitic limestones. The intensity of dedolomitization appears to vary considerably. It is very common in the Kurucaoba Section and relatively scarce or absent in other places. At the same time the intensity of dedolomitization varies locally as well. While it is abundant in one bed it may be totally absent in the other one. It has been noted that the same bed which shows strong evidence of dedolomitization even within a short distance, has absolutely no trace of dedolomitization. Therefore it can be argued that dedolomitization varies both vertically and horizontally. Usually the degree of dolomitization, and related to this, the degree of dedolomitization, is not very intense in the sediments investigated. For this reason in most cases recognition of the existence of both phenomena were only accomplished by microscopic study of stained peels and thin sections. Wherever dedolomitization is intense, some field criteria are helpful detecting this phenomenon.

1. The presence of heavy stains of iron oxides in the weathering crusts, along bedding planes, joints, fractures. The colour of these stains is either limonitic yellowish-brown or haematitic red. It is thought that these colours are produced by oxidation of iron-rich dolomites.

2. Gradual transition from dense dolomitic limestone into highly porous rock. The transition may be both laterally and vertically. Pores are the result of leaching of dedolomites which create rhombic voids (EVAMY, 1967).

According to SHEARMAN and Others (1961) the following criteria are valid to recognise the process of dedolomitization in peels and thin sections under the microscope:

1. Relic dolomite crystals partly replaced by calcite;
2. Presence of pseudomorphs of calcite after dolomite;
3. Presence of rhombohedral aggregated within the new generation of calcite.

IV. TYPES OF DEDOLOMITIZATION:

Thin section observations have shown that all gradations occur between rhombs completely composed of dolomite, through composite rhombs to rhombs completely composed of calcite. Composite rhombs are two types: They have either calcite centres or calcite outside margins. The latter case is very rare. Calcite rhombs are composed of aggregates of calcite crystals which are coarser grained than the micritic matrix (Pl.67D). Basically there are two distinct types of dedolomitization based on the calcite replacement mosaic and they have been discussed in detail by SHEARMAN and Others (1961). In the one type each dolomite rhomb is replaced by a mosaic of finer-grained calcite, and in the other the dolomite rhombs are replaced by a more coarsely crystalline mosaic of calcite crystals (Fig. V - 10 and V - 11). The first type effects a variety of rocks ranging from dolomitized lime stone in which the dolomite rhombs are scattered throughout the limestone groundmass, to those which were entirely a mosaic of dolomite crystals. The rhombs have a sharply

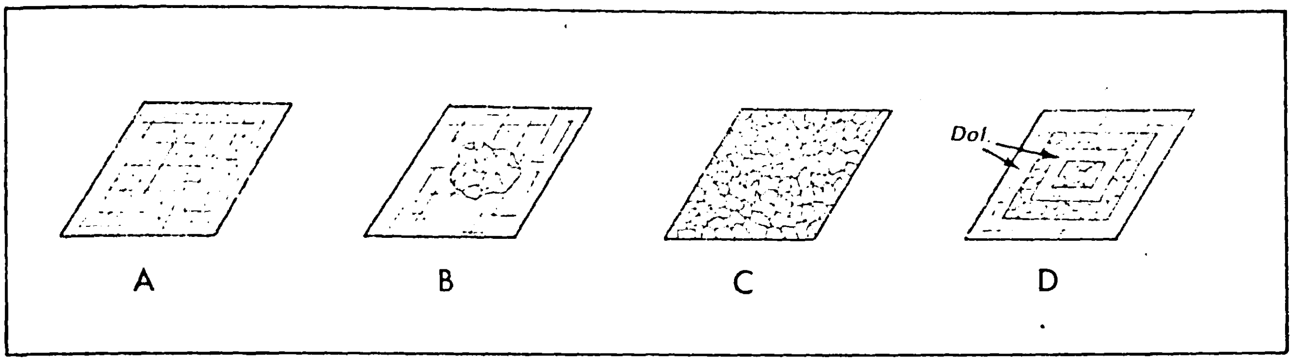

Fig.V-10: The stages in the "centrifugal replacement" of dolomite by by calcite.
A - Dolomite rhomb; B - Dolomite rhomb with calcite core
C - Composite calcite rhomb (Dedolomite rhomb);
D - Zoned dedolomite rhomb. Zones shown by arrows are composed of dolomite.

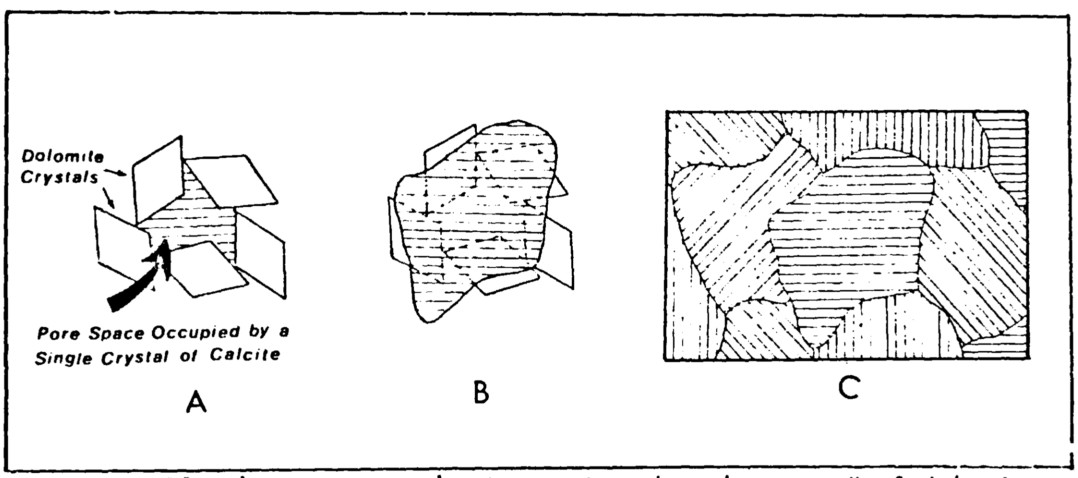

Fig.V-11: The stages in the " centripetal replacement" of dolomite by calcite.
A - Dolomite crystals with rhombo-mouldic porosity filled with a single crystal of calcite;
B - Enlargement of calcite crystal at the expense of dolomite crystals;
C - Limestone after dolomite (The calcite crystal at B is in the centre and shown by horizontal lines)

defined crystal boundary. Although primary rhombic calcite has been considered to be rare (SHEARMAN and Others, 1961; p.3; PETTIJOHN, 1975, p.323) PERKINS (1968) reported a wide occurrence of rhombohedral calcite as void linings in limestones ranging in age from the Lower Palaezoic to Recent, and suggested that rhombohedral calcite is a doubtful criterion for detecting dedolomitization. During the present work, this type of calcite infilling has not been observed. PERKINS' calcite rhombohedra (Ibid. p.1372-3; Figs. 1-6) were present as cavity linings, whereas in the material examined, most calcite rhombohedra do not line the cavity wall, and are enclosed by the sparry calcite anhedra or micritic matrix (Pl. 57A).

Intermediate stages have been observed between dolomite rhombs and rhombs completely composed of calcite. Dolomitic rhombs having cores of fine grained calcite could have been originated in two ways: Either by incomplete dolomitization of the matrix or by dedolomitization of previously formed dolomite rhombs (SHEARMAN and Others, 1961). Study of calcite cores has shown that crystals in them are much larger than the ones seen in the matrix and therefore it can be argued that finer grained calcite cores are not relics of the original limestone; but a new generation of calcite which has replaced the dolomite.

When the rocks have only been partially dedolomitized, it is the central parts of the rhombs which are replaced. The reason for the selective replacement of calcite in the central, rather than the outer parts of the dolomite crystals at earlier stages of dedolomitization is probably due to relic inherent inclusions of calcite which is restricted to the central part of dolomite crystals and causes the cloudy appearance of it (Pls 60 - 61) Crystal rim is free of inclusions (MURRAY, 1964, p.399-403). Therefore at earlier stages of dedolomitization, nuclei for calcite growth are available in the core but not in the rim. As a result, growth of calcite would commence at the core and spread towards the rim until completely replaces entire dolomite crystal at later stages of dedolomitization. This type of dedolomitization is called "Centrifugal Replacement" (Pls 66-67). Another variety of this type replacement is "Zonal Dedolomitization". In this type, some rhombohedral zones within the dolomite crystals are selectively replaced. As noted earlier (p.193) KATZ (1968 and 1971) suggested that the calcian dolomite zones of "zoned dolomite crystals" are more susceptible to replacement by calcite. Resultant product is a dolomite rhomb with planes of calcite parallel to the rhomb faces (Fig. V - 10D). This variety of dedolomitization has not been encountered in the present study.

The second type only takes place and affects thos rocks which were essentially dolostones and consisted of tightly packed aggregates of dolomite crystals (not in dolomitized limestones). These rocks have a rhombo-mouldic porosity and some of these pore spaces are occupied by a single crystal of calcite. Dedolomitization proceeds by enlargement of these calcite crystals at the expense of dolomite (Centripetal Replacement). In the partially dedolomitized rocks, corroded dolomite rhombs are poikilitically(*) enclosed in relatively large crystals of calcite (SHEARMAN and Others, 1961). When the process has gone to completion the rock is coarsely crystalline granular limestone from which all the evidence of the former dolomite has been lost (Fig. V - 11; Pl. 68). In thin sections these rocks have all the general features of metamorphosed limestones (SHEARMAN and Others, 1961; p.6).

Dolomite crystals with calcitic outside margins cause problems. They are "calcite rim" of ZENGER, who concludes that they are formed by "centripetal replacement" of dolomite crystals (1973; p.123). This conclusion contradicts with the nature of "centripetal replacement" which is always associated with dolostones (SHEARMAN and Others, 1961;p.9) and not with dolomitic limestones which calcite rims are found. Therefore the present author is of the opinion that they are either passive precipitation of calcite on dolomite nuclei (i.e. "Calcite Envelopes" of GOLDMAN's (1952)) or a special type of "zoned dedolomitization" discussed earlier and not a centripetal replacement. Because of the very limited occurrence it is impossible to conclude which one is the responsible mechanism.

(*) Friedman (1965, p.652) states"... the term poikilitic and poikiloblastic should be avoided in the study of sedimentary rocks, since poikilitic refers to igneous rocks". He suggests that "poikilotopic" be used instead, and defines it as "A term which relates to the fabric of inequigranular sedimentary rocks which have undergone recrystallization, or which have formed by precipitation, in which the constituent crystals are of more than one size, and in which larger crystals enclose smaller crystals of another mineral (p.651)". But present author is not in agreement with this definition because limestones are essentially monomineralic aggregates. It is therefore more important to have a term which describes the relationships between crystals and grains. For this reason, the term poikilotopic is no further used. The relationship whereby a mineral enloses smaller crystals of the same or different minerals, or different grain types of essentially the same mineralogy has been termed "poikilitic".

V. REGENERATION OF LIMESTONE TEXTURE:

As a result of dedolomitization, the predolomitization limestone texture may be regenerated wholly or in part. SHEARMAN and Others (1961) showed an example of regeneration of the calcite overgrowth on echinoderm fragment. EVAMY (1967) illustrated several examples of regenerated fabric. He argues that when the rocks were dolomitized minute relics of the original calcite remained unreplaced within the dolomite and these relics served as nuclei about which the latter replacement calcite was initiated. It has also been shown by EVAMY that a "Clotted" or "Grumeleuse" texture may result from cloudy centered, clear rimmed dolomites. MURRAY (1964) explained the formation of this type of dolomite by the theory of a local source of carbonate ions and suggested that the included material in dolomite crystals appears dark where it was derived from mud-sized crystals and relatively light where derived from coarser spar. According to EVAMY (1967; p.1209) dark centres of dolomite rhombohedra were likely the product of dolomitization of mud. He concluded that dolomitization and subsequent dedolomitization may, therefore, alter a depositional mudstone to a clotted pseudo-micropelletoid grainstone or packstone (Fig. V - 12; P166E). Grumeleuse textures found in the dedolomitized rocks are characterised by:

(a) Islands or clots of coarse calcite crystals present in a matrix of micrite;

(b) Islands or clots of micrite present in a matrix of coarse calcite.

These textures are very similar to those produced by recrystallization of mudstones. However, grumeleuse texture produced by the dedolomitization process can be distinguished from those produced by recrystallization process by:

1. The presence of serrated boundaries between the internal sediment and the host rock;

2. The presence of ghosts of former dolomite rhombohedra with ill-defined boundaries.

VI. RHOMBOHEDRAL PORES:

Another feature which appears to be related to dedolomitization is the production of rhombohedral voids. Some rhombohedral pores have been observed in the sediments investigated. Usually they are restricted to the highly weathered surfaces, joints, and bedding planes. Some cemented pores have also been observed. The calcite cement differ from calcites of the composite rhombohedra and they exhibit a drusy texture

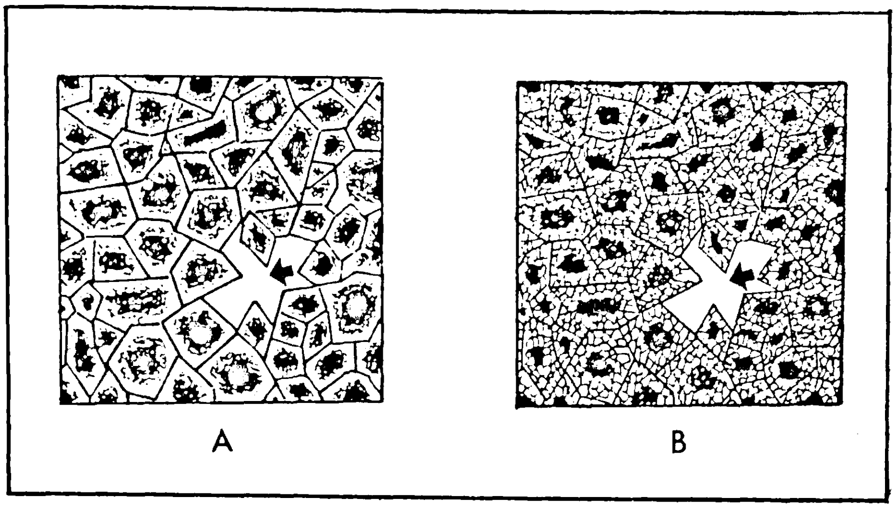

Fig.V-12: Development of pseudo-pelletoidal texture (Modified after EVAMY, 1967)

A. Dolomite crystals with dark centres, rich in calcite inclusions, surrounded by clear rims almost free of inclusions (Arrow indicates pore space);

B. During dedolomitization growth of secondary calcite on the centrally disposed included nuclei gives rise to clots of finely crystalline calcite (shown in black). The clear rims of the former rhombohedra are altered to a relatively coarsely crystalline mosaic of calcite. Compare with Plate 66D.

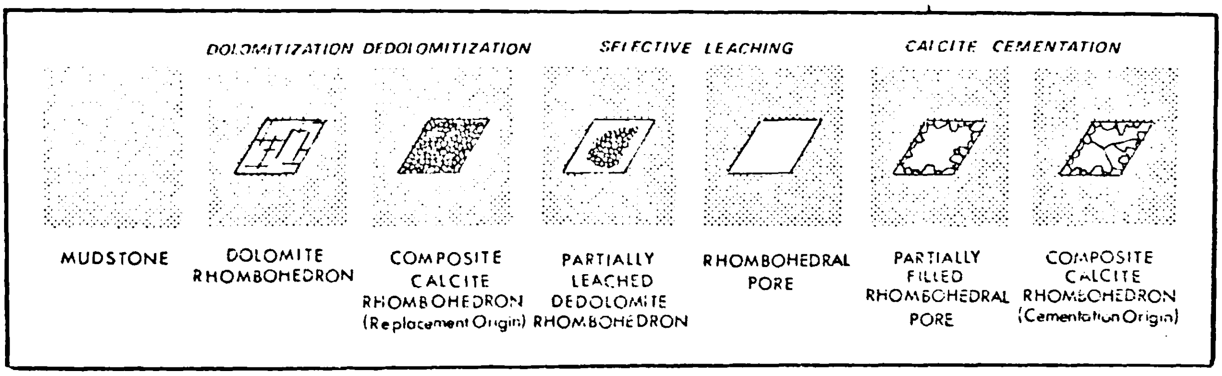

Fig.V-13: Diagenetic history of certain dedolomitized rocks (Modified after EVAMY, 1967) which probably indicates that those rhombohedra had suffered dissolution; resulting pore later being filled by a drusy calcite (Fig. V - 13; Pl.). At first sight it would appear that the voids resulted from solution of dolomite, but EVAMY (1967) who described several examples argued that the voids were produced by selective leaching of calcite which replaced dolomite. He based his conclusion on the experiments of DE GROOT (1967) who noted that dolomite dissolves without decomposing to form calcite only at high PCO_2. Under such conditions, however, it is expected that calcite will dissolve in preference to dolomite. Hence the presence of rhombohedral pores in a preferentially undissolved calcite matrix is unlikely to be the result of the simple dissolution of dolomite at depth. On the other hand the partial solution of some calcitic rhombohedra which are now filled by calcite cement indicates that the production of the early type of rhombohedral pores was not the result of preferential solution of dolomite as an intermediate stage in the near-surface dedolomitization but the result of solution of calcitic rhombohedra. In the same time the presence of rhombohedral pores in the weathering crust of dolostones, and their complete absence in the fresh parts of the same dolostones support EVAMY's conclusion. EVAMY also raised two objections based on textural grounds against the preferential solution of dolomite. The first objection is that if dolomite solution had been complete, any relic inclusions from the original limestone would have been eliminated, and no trace of the predolomitization texture could have been regenerated by dedolomitization. The second objection is that composite calcite rhombohedra would always be secondary to rhombohedral pores, and, therefore, should all exhibit cementation mosaics (Ibid, p.1211). The presence of mosaic cementation in only part of the composite rhombohedra can be considered as infillings of rhombohedral pores after dedolomite, not dolomite (Pl.). The resulting cementation mosaics are entirely different from the textures of direct dedolomitization. It is important to distinguish between fabrics of composite calcite rhombs which result from dedolomitization on one hand and those produced by infilling of rhombic voids on the other.

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PART 10

SILICIFICATION

In this account, silicification refers to silica replacement of carbonate rocks as well as silica cementation and filling of pores. Usually it is difficult to differentiate between silica replacement and cementation. Silica replacement of fossils is proven only if some traces of organic structures are preserved. If not, it is also possible that the organisms were dissolved and the voids filled with silica. During the course of this investigation silicification has been observed in large scale, as cherts; also in small scale, as void filling cement and as replacement of grains or matrix. Therefore silicification will be discussed under two different headings: Cherts and small scale silica fabrics.

A. CHERTS

Cherts of the Gürün Area are almost always associated with the calciturbidites. They are very rare in the shelf carbonates. They form nodules, lenticular bodies and beds(*). So far a general scheme of chert classification has not been adopted and in this account ROBERTSON's (1977) approach has been followed and cherts have been divided into two categories: Granular Cherts and Vitreous Cherts. Granular chert is a grainy siliceous rock with a matt lustre and vitreous chert is a tough compact cryptocrystalline siliceous rock with a vitreous lustre. The former is often impure and breaks with an irregular hackly fracture. These terms are entirely non-genetic field terms.

Granular cherts are a more common type in Keşlik Dere Sectiòn and in Kaynarca (T29) vicinity. They usually form beds and elongated lenticular bodies (Pls.70-71). Brownish grey, grey and dark grey are the dominant colours. Host rock is thick bedded compared to vitreous cherts and main rock type is packstone and grainstone.

(*) Although field appearance suggests that some cherts form "beds" they are not the true form of "bedded cherts" in the meaning of recent literature on cherts. DUNBAR and ROGERS (1957) and HEATH and MOBERLY (1971) define "nodular cherts" as those which occur in carbonate host rocks, where as "bedded cherts" are associated with siliceous sequences and never with the pelagic carbonates.

Vitreous cherts have mostly been encountered in Abdalpınar Dere and Devecayırı Sections. Host rock is frequently thin bedded mudstone and wackestone. They are usually black in colour and form individual lenticular bodies up to a few metres long and 0.20 m. thick (Pl. 70). Most of the lenses are in the range of a few cm. even mm. (Pl.73B). Vitreous cherts are usually restricted to the base of calciturbidite beds and show an elongation of varying degrees parallel to bedding. Another occurence of vitreous chert is as nodules up to 20 cm. in diameter. They are rarely observed and therefore a generalization about their distribution is impossible. Both nodules and lenticular bodies are separated from the surrounding rock matrix by an intermediate light grey zone (Pl.71C). This zone is composed of granular chert and has an average thickness of 1 cm. Its contact with the vitreous chert is highly irregular, tending to be transitional in most cases. On the other hand, the contact of this zone with the surrounding rock matrix is usually smooth and sharp. Therefore the sharpness of the nodule-rock contact observed in the field is caused by the latter and not by the early contact. Some of these nodules and lenses exhibit banding which is alternation of light and dark coloured layers, approximately conformable with the outer surface of nodules of parallel with the margins of more elongated bodies (Pl.72D). They are usually called "Liesegang Rings" in literature (For a review of "banding in cherts" see ORME; 1974). It is believed that the bands are closely correlated with the growth of the nodules and mark the previous growth surfaces. They are caused by expelled insoluable impurities during the stages in chert growth.

In thin sections under polarized light the chert consists of a yellowish-brown matrix, full of inclusions and impurities, and enclosing colourless area, clear and free of any impurities. The matrix consists of microcrystalline and cryptocrystalline quartz which make up the bulk of the cherts in mud supported rock. Also they are a major component of chert masses in grain supported rocks. The fact that micro and cryptocrystalline quartz pseudomorphes mud, crinoids, bryozoans, brachiopods, and corals proves that it is a replacement and not a cement. The fossil fragements are replaced by silica coarser than that of the matrix (Usually in the range of 10-30 μ.); while the cavities within these fossils are lined by fibrous chalcedony which grades towards to the centre of the cavity into a mosaic of megaquartz (Pl.76C,D and E). They correspond to the colourless areas in thin sections.

Most susceptible to replacement are small thick-shelled brachiopods, bryozoans and mud. These all have about the same susceptibility and within chert masses are usually totally replaced. Least susceptible to replacement is syntaxial calcite cement rim on crinoids, followed by crinoids themselves. Within a chert mass typically all other skeletal grains are chertified yet crinoids and their syntaxial cements commonly remain largely unreplaced or only partly replaced (P1 87). Thick brachiopod shells may remain only partially replaced along with crinoids. Silica forms only a thin fringing rim around calcareous fossils resistant to chertification (Pls.79 E , 80 C).

I. MINERALOGY

Four silica varieties (phases) have been distinguished during the Deep Sea Drilling Cruises. They are opal, lussatite (disordered cristobalite), chalcedony, and quartz. The mineralogical and genetic characteristics of which are summarized in Table V-5

SILICA VARIETY	MINERALOGICAL STRUCTURE	OCCURENCE IN SEDIMENTS	GENESIS
OPAL	Highly disordered, nearly amorphous natural hydrous silica	Tests of siliceous organisms	Original organic skeleton
LUSSATITE (Disordered Cristobalite)	Disordered stacking of low-crystobalite and low-tridymite layers	Porcelanitic chert; peripheral parts of quartzose chert nodules; partially silicified carbonate siliceous or shaly rocks	Early diagenetic origin at low temperature from skeletal opal, volcanic glass etc.; authigenic precipitation from pore solutions
CHALCEDONY	Optically fibrous quartz	Void filling and fracture filling in both quartzose and porcelanitic cherts	Late diagenetic
QUARTZ	a. Cryptocrystalline quartz (< 1 µ. No orientation) b. Quartzine (< 1 µ. Fibrous, length-slow) c. Microcrystalline quartz (> 1 µ.)	Quartzose cherts, fracture filling in quartzose chert	Late diagenetic origin in sediments older than 50 million years

Table V-5: Characteristics of silica varieties of deep-sea cherts (Mainly after LANCELOT , 1973; VON RAD & ROSCH, 1974)

The results of these cruises have shown that deep-sea cherts can be divided into two categories, namely porcelanitic cherts and quartzose cherts. The first group is predominantly composed of lussatitite (disordered cristobalite); is found in clayey layers and occur as thin beds. Where as the second group is composed of quartz and/or chalcedony and is found as nodules in carbonates.

The cherts encountered in this study have been examined only by means of petrographic microscope and only a distinction between microcrystalline quarts and chalcedony has been verified. But recently, ROBERTSON (1977) by the aid of X-ray diffraction technics and scanning electron microscope, has shown that disordered cristobalite (lussatite) is the main component of the both vitreous and granular cherts of Cyprus. The same work has also revealed that disordered cristobalite shows some morphologic differences. Granular cherts are usually composed of small, partly coalescent microspherules and micrognanules, with numerous vermifor, or fibreous bodies. Vitreous cherts are seen to consist of coalescent microgranules and microspherules.

II. ORIGIN OF SILICA:

The origin of the silica found in chert has been often debated, and no clear answer to the problem has been provided. Generally two different origins are considered.

1. Organic Origin: The almost exclusive connection of nodules within biogenic carbonate rocks is circumstantial evidence that it is the organisms that are the primary source of the material. The occurence of siliceous spong spicules in the chert nodules and the lack of it in the adjacent limestones supports this hypothesis (SIEVER; 1962). If siliceous fossils are to be available for later production of chert, important requirements would seem to be a relatively silica-rich bottom water, and the presence of reasonable high burial rate. These conditions would tend to prevent the dissolution of siliceous skeletons at the sediment interface. If this burial can be achieved without considerable dilution of opaline skeletons the chances for later chert formation would appear much improved (VON RAD and RÖSCH; 1974). The association of Gürün cherts with the calciturbidites is in favour of this hypothesis. The high sedimentation of turbidite layers has probably caused the quick burial of siliceous skeletons.

2. Inorganic Origin: Geologists have long recognized that volcanic glass is a possible source of silica, because its devitrification is a possible source of silica into solution. Another possible inorganic source of silica for chertification is the transformation of zeolite to montmorillonite, and montmorillonite to illite or kaolinite during diagenesis. This is supported by the decrease in abundance of montmorillonite in older sedimentary rocks (SIEVER; 1962).

From both field and microscope studies it is clear that the only volumetrically significant source of silica available for chert formation in the Maestrichtian Calciturbidites in the Gürün Area is the dissolution of siliceous microorganisms. Cavities formed by dissolution of radiolaria which is later filled by sparry calcite are seen in thin sections (Pl.79C). Most probably diatoms have played an important role in the contribution of silica, but they are never preserved in this sediments. Lastly, a biogenous silica source for deep-sea nodular cherts is now generally accepted (WISE and WEAWER; 1974).

III. ORIGIN OF CHERTS:

Field and petrographic evidence suggests a replacement origin for cherts and not a direct precipitation from the sea water.

Evidence for the replacement:

1. Chert is present as irregular masses randomly scattered within host rock. Even those observed along bedding planes do not persist laterally and may disappear in an adjacent locality. The observed concentration of chert nodules along bedding planes does not necessarily mean a primary origin. The bedding plane may have had a strong influence on the growth of chert nodules, due to the ease of solution movement along these bedding planes.
2. The presence of transitional zone from chert to host rock is a strong indication of its replacement origin.
3. The distribution of the unassimilated intraclasts and fossils within the chert and the transitional zone is identical to that of the surrounding matrix (Pls.74A;75 A,D).
4. The presence of replaced fossils (Pl.76 C-E).
5. The preservation of textures and structures.
6. The presence of fossil ghosts inside the chert (Pl.79A).

IV. MECHANISM OF CHERT FORMATION:

The mechanism of chert formation has been most extensively investigated by the participants of the Deep Sea Drilling Project Cruises. Based on these works, two basic hypothesis of chert formation have been put forward. One of the hypothesis is called "Maturation Hypothesis". It is first outlined by ERNST and CALVERT (1969) and later advanced by HEATH and MOBERLY (1971). According to this hypothesis, chert nodule formation begins with dissolution of biogenous opal and reprecipitation as finely crystalline disordered cristobalite (lussatite). The lussatite is either deposited as interstitial matrix, or replaces pre-existing calcite or montmorillonite and through time converts to chalcedony and/or quartz. The end product is a classic vitreous chert. Four stages have been recognized in the formation of chert nodule (HEATH and MOBERLY; 1971):

1. Early diagenetic precipitation of chalcedony and/or quartz in the empty chambers of foraminifera with preserved calcite test.

2. Rapid replacement of the micritic ground mass of the carbonate rock by extremely fine crystalline lussatite (disordered cristobalite).

3. Replacement of foraminiferal tests by chalcedony and/or quartz.

4. Final infilling of all the voids and pore spaces in the rock with silica by inversion of the disordered cristobalite matrix to quartz.

The second hypothesis is LANCELOT's (1973) "Quartz Precipitation Hypothesis". This hypothesis considers primary quartz precipitation to be the driving mechanism behind nodule formation with disordered cristobalite a by-product of the process. It is suggested that the mineralogical nature of the chert is directly influenced by the composition of the sediment in which it develops and the roles of foreign cations and permeability are important.

According to LANCELOT (1973; p.397 - 398) the sequence is as follows:

1. Quartz precipitates directly in the carbonate matrix where clay minerals are rare and dispersed.

2. The nodule develops outwards by accretion, and in the process all the clay minerals and dissolved cations that cannot be accommodated in the quartz structure are excluded and move along a 'quartzification front'. This can be achieved only because of the high permeability of the sediment, allowing relatively free circulation of the interstitial waters.

3. At the periphery of the quartz nodule the concentration of clay minerals and foreign cations increases, while that of dissolved silica decreases because of limited supply. These conditions favour precipitation of disordered cristobalite that makes the rim commonly observed around the nodule. This concentration process can be visualized in thin section by observing the distribution of dark impurities near the contact between quartzose chert and cristobalitic rim.

Recently ROBERTSON (1977) has suggested "Impurity-Controlled Maturation Mechanism" for the nodular cherts of Cyprus. According to the concept of this hypothesis, insoluble impurities may theoretically be expelled during the two diffent stages in the chert growth:

1. During replacement of calcite or other host sediment by deposition of disordered cristobalite (lussatite). The disordered cristobalite lattice is unable to accommodate much of the impurities such as clays and other small insoluble particles.

2. Expulsion of impurities which are mostly cation bound structurally within the lattice of the disordered cristobalite prior to recrystallization to chalcedony.

Impurity zones which are associated with the replacement of calcite by disordered cristobalite is marked by a diffuse optically translucent material, mostly of clays; where as, the replacement of cristobalite to quartz shows a concentration of optically opaque microgranules, mainly iron oxides.

V. POSSIBLE PRIMARY CAUSE OF THE CHERT FORMATION:

In the last few years, an extensive geologic work has been carried out, especially by the scientists of the Deep Sea Drilling Project Cruises, but the cause of the chert formation is still an open question. Fundamental to this problem is the fact that no marine chert has been observed in the process of formation in Recent sediments.

This has been considered in two different ways:

1. Today's environmental conditions are different from the ones existing during the Early Tertiary.

2. Formation of chert is such a slow process that it has not had enough time to form since the Oligosen.

The former explanation is more likely (For discussion of this problem see LANCELOT 1973).

Because chert resulted from the silicification of the host rock by silica-rich solution circulating through these rocks, a special mechanism must be responsible for attaining such concentrations. Organic matter appears to play an important role in causing local concentrations of silica (EMERY and RITTENBERG 1952). They have suggested that lowering of the solubility of silica by absorption of organic matter might occur at the site of organic decay. The influence of organic matter has also been discussed by SIEVER (1962) who has suggested that organic matter was responsible in producing local but important pH changes during their decay cycle. He has also proposed the "salt-sieving mechanism" as a possible cause of obtaining a high silica concentration with the clay minerals acting as a semipermeable membrane. According to him, if dissolved silica does not pass through the membrane as the ionic strength increases below the membrane, the silica becomes less soluble, eventually building up to supersaturation and precipitating, while carbonate becomes more soluble and dissolves. Above the membrane the solutions are undersaturated and only solution can take place. Another possibility is the role of skeletons producing 'nuclei' for silicification. This has been mostly accepted among the European geologists (FUCHTBAUER 1974).

All present day seas and oceans are undersaturated with respect to silica and there is no reason to assume that seas of older geologic ages were saturated or supersaturated with it. However, supersaturation might be achieved in very special environments and under extreme conditions, which may ultimately result in the inorganic precipitation of silica. The inorganic precipitation of silica is reported to be taking place in ephemeral lakes associated with the Coorong Lagoon of South Australia, following the drying of these lakes (PETERSON and VON DER BORCH; 1965). There, active photosynthesis by indigenous algae cause pH values greater than 10 which detrital quartz grains and possibly clay minerals are corroded, saturating the lake waters with respect to amorphous silica. The algal activity fluctuates seasonally and, in addition, the lake basins are dry during part of the year. This resulting decrease in pH causes the silica in solution to precipitate, forming an "amorphous" gel containing cristobalite crystals.

As explained by LANCELOT (1973) interstitial water is generally undersaturated with respect to amorphous silica and supersaturated with respect to the crystalline forms (p.387). This implies that amorphous silica can be selectively dissolved and reprecipitated at any of the crystalline forms.

VI. DOLOMITE IN CHERT:

The presence of dolomite in chert has been recorded by many authors (for example PITTMAN, 1959; DIETRICH, HOBBS and LOWRY, 1963). PITTMAN suggested that this is evidence for post dolomite origin of the chert. He supposes that the dolomite was formed from the original limestone, the remains of which were subsequently replaced by silica.

This idea was also supported by DIETRICH and Others (1963) who believed that the dolomite crystal occurring within the chert were formed before silica entered to replace, the undolomitized portion of the host rock. Their assumptions were based on the "clearness" of the dolomite crystals in contrast to the "dirtiness" of the chert.

Occurrence of dolomite in cherts is very scarce in the Gürün samples and therefore it is difficult to come to a conclusion. In some samples the dolomite in chert is euhedral while in others the rhombs are strongly corroded and traces of silicification of dolomite rhombs can be detected (Pl.87D). On the other hand the dolomites in the crinoid rich cherts are concentrated along the edges of the chert. The rhombs and patches of rhombs tend to be elongated parallel to the direction of cleavage of both crinoid plate and its syntaxial overgrowth. The distribution of the dolomite can be related to that of a "reaction rim" in which the dolomite is a product of chert formation. The source of magnesium for the dolomite is almost certainly the high-Mg calcite of the original crinoid skeleton.

B. SMALL-SCALE SILICA FABRICS

I. GENERAL: Quartz in sedimentary rocks occur in two main forms: Megaquartz and microquartz (FOLK and WEAVER, 1952). Megaquartz usually has coarse, equant crystals greater than 20 μ. and includes drusy quartz, quartz overgrowth and authigenic quartz crystals (PITTMAN 1959). Microquartz has two end members: Microcrystalline quartz and chalcedonic quartz. Minute equant crystals, usually in 1-4 μ. range, form microcrystalline quartz. It shows pinpoint extinction (FOLK and PITTMAN, 1971). Chalcedony is a delicately fibrous form of microquartz. Chalcedonic quartz seen with the polarizan microscope, seems to be made up of radiating fibres, only a few microns in diameter and up to several hundred microns in length; but FOLK and WEAVER (1952) using scanning electron microscope did not detect the fibrous nature of chalcedony under extremely larger magnifications. They showed this mineral variety to possess a "spongy surface" due to theinclusions of water filled cavities within the mineral, thus causing its brown colour when examined under polarizan microscope. In contrast, microcrystalline was composed of numerous polyhedral blocks with slightly curved faces. Three different forms of fibrous quartz have been noted:

1. Normal length-fast chalcedony (normally known as chalcedony) with c-axis (slow ray) perpendicular to the fibres. It forms fibrous crusts and regular spherulites of delicate fibres neatly arranged in a precise radial pattern with all fibres of the same length. It is almost always a cavity-filling (FOLK and PITTMAN, 1971; p.1050).

2. Quartzine: c-axis is parallel with the fibres (length-slow). Fibres are thicker, cruder, and the radial pattern is not so neat. Quartzine occurs both as a cavity-filling and as a replacement.

3. Lutecite: It is length slow chalcedony with c-axis at approximately 30^{o} to the fibres. Lutecite usually occurs as a

replacement of calcitic shells, intraclasts, calcite cement and evaporite crystals and nodules. Its most common occurrence is a symmetrically-extinguished chevron-like grid of coarse fibres crossing at about 60^{o} and grading into quartz (FOLK and PITTMAN 1971). The same authors came to the conclusion that length-slow chalcedony occurs almost exclusively in association with sulphates and evaporites. On the other hand SIEDLECKA (1972) presented supporting evidence to the observations of FOLK and PITTMAN (1971). In addition she cited examples of length-slow chalcedony replacing fossils in a limestone in which no trace of sulphates was found. Recently JACKA (1974) present examples of length-slow chalcedony selectively replacing fossils of the Middle Permian calciturbidites which were deposited in deep water of the Delaware Basin (Texas, U.S.A.). No indication of the former presence of evaporites were observed in the deep water facies which includes the calciturbidites. In a discussion, HATFIELD (1975) supported JACKA's conclusion. He noted occurrence of the length-slow chalcedony in the Middle Devonian of Ohio (U.S.A.) which is in no way associated with evaporites. These observations are in agreement with the findings of this account. Numerous examples of length-slow chalcedony selectively replacing fossils in the Maestrichtian calciturbidites of the Gürün Basin have been encountered during the examinations of the thin sections. As in the Delaware Basin and Ohio examples no indication of the former presence of evaporites has been found. In reply to HATFIELD's (1975) discussion (FOLK (1975) agreed the occurrence of length-slow chalcedony as replacement of fossils in non-evaporitic rocks and suggested that probably magnesium ion in some way was responsible from the crystallization of length-slow chalcedony. His assumption was based on the presence of 1-4 mol percent magnesium carbonate in the fossils which were replaced by length-slow chalcedony. He also noted the high magnesium content in saline brines.

In recent years two important papers dealing with the silicification and associated silica fabrics have been published. One of them is WILSON's (1966) work on the Upper Jurassic limestones of Southern England. His work based on the previous studies of SOSMAN (1927) and WALKER (1962). Based on a simple morphological and optical criteria, he distinguished five major fabric types. They are: Mosaic quartz, chalcedonic overlays, spherulitic chalcedony, lutecite and microcrystalline granular quartz. All the fabrics, except chalcedonic overlays may be of replacement origin. On the other hand the first three of these, may be cementation fabrics. He also demonstrated two major stages of silicification: pre-calcite cement and post-calcite cement fabrics.

The second important work on silicification is of ORME's (1974). His work on the Visean Limestones of Derbyshire was mainly dealt with silica fabrics, chert, small-scale silica replacements, and quartz rock. He distinguished void filling, replacement, and replacement-recrystalliza·tion fabrics. The last group is mostly related with quartz rocks (The product of low temperature metasomatism). He also suggested that silica might have been developed at more than one time.

Based on the works of WILSON (1966) and ORME (1974) and the classification already discussed (See p.211), small-scale silica fabrics encountered in the Upper Cretaceous Carbonates of the Gürün Area can be divided into two genetic groups: a) Cavity-filling fabrics and b) Replacement fabrics. The first group is very rare in the sediments investigated and comprises megaquartz and chalcedonic crusts. Replacement fabrics include megaquartz, spherulites of length-slow chalcedony, microcrystalline quartz and authigenic quartz crystals (Pls. 77, 78).

II. DESCRIPTION OF SILICA FABRIC TYPES:

A. MEGAQUARTZ: It is a mosaic of equigranular anhedral grains usually bigger than 20 μ. It corresponds to the "mosaic quartz" of WALKER (1962); WILSON (1966) and "drusy quartz mosaic" of ORME (1974). It shows the characteristic increase in grain size away from the margin of the allochems. Although this is a criterion for recognition of calcitic cement (BATHURST 1958; 1971) here it does not necessarily imply the growth of crystals into free spaces. A fabric resulting cavity-filling megaquartz can form by replacement as shown by WILSON (1966; p.1043). Sometimes calcite ghost fabrics are preserved as inclusions which is a clear evidence of replacement origin. But in some other cases inclusion of calcite is absent and the true nature of such a fabric is revealed by its position to adjacent obvious cavity-filling fabrics and allochems. If they pass directly into a peripheral zone of chalcedonic crust a primary origin is more likely. The most of the megaquartz encountered in this study is thought to be of replacement origin. This is "pseudodrusy quartz mosaic" of ORME (1974). Some of the rocks are composed of large allochems, usually brachiopod and/or crinoid fragments which show good sorting. They are partly or completely silicified. Matrix is megaquartz and rarely shows remnants of a former micritic ground mass. These rocks resemble bioclastic limestones. Larger allochems have only been peripherally replaced while smaller ones have been entirely silicified. Calcite is only present as inclusions. This suggests that replacement started from the margins of the allochems and has proceeded inwards (Pl.79 E, 80 C)

B. CHALCEDONIC CRUST: It usually forms rims around silicified allochems (Pl. 77) and bands lining primary or secondary voids. It also fills the veins (Pl 77F). It is in the same meaning with WALKER's (1962) "fibrous chalcedonic quartz overlays"; WILSON's (1966) "chalcedonic overlays" and ORME's (1964) microdrusy chalcedony.

C. SPHERULITES OF LENGTH-SLOW CHALCEDONY: They are commonly found replacing calcitic shell fabrics. Evidence for the replacement is always clear, because they show cross cutting relations with the fabrics in which they occur. If they are well developed they show a subcircular or pseudohexagonal outline (Pl.83 B,C).

D. MICROCRYSTALLINE QUARTZ: This is the commonest fabric encountered in this study. It is composed of minute equant crystals with pin-point extinction. Microcrystalline quartz can sometimes be so finely crystalline that it is difficult to resolve under ordinary microscope. In some cases they may have an isotropic mass appearance.

E. QUARTZ EUHEDRA: Two samples of limestones have been found containing 100-500 micron size authigenic quartz crystals which tend towards a euhedral form. ORME (1974) notes the presence of detrital quartz cores (Ibid. p.87). This is not the case in the samples studied. None of the euhedra found with the detrital quartz core. The cores encountered in these samples are composed of a cloudy material in dark appearance. Microprobe results show that they are calcite. On the other hand, in one core, it has been found that the shell structure which was replaced by quartz euhedra was forming the core of quartz crystals (Pl.86E). GERMAN (1968) notes that the growth of euhedral quartz is favoured by saline conditions. Saline conditions in the Upper Cretaceous in the Gürün region may have been produced during diagenesis by percolating brines responsible for diagenetic cementation.

III. ORDER OF REPLACEMENT:

Fossil remains are always silicified in preference to the enclosing carbonate mud matrix. Also within the replaced skeletons the silica isusually found as a replacement of the actual shell parts, rather than occurring within the voids of pores. Why the silica preferentially replaced the shells and skeletal parts rather than the matrix and voids is not known. It is generally noticeable that replacement by silica begins earlier in skeletal grains which have a relatively massive fabric, than in those which have a very fine-grained wall fabrics such as foraminifera and calcareous algae. Bryozoans, brachiopods and corals are generally more sensitive to replacement by silica than gastopods, cephalopods and echinoderms. This order of susceptibility is accepted by many workers (See DAPPLES 1967). On the other hand DAPPLES seems to believe that the order of preferential silica replacement is a function more of the distance from centres of silica precipitation, and the quantity of silica being precipitated, rather than the crystal habit or composition of the carbonate constituting the fossil shell. This conclusion seems unlikely, because the observed order of replacement is the same in all sediments investigated in this study. The susceptibility of any fossil to silica replacement during early diagenesis may be determined by its mineralogy. But silicification of late diagenetic stage such as encountered in the Gürün sediments would not be dependant on the mineralogy of the fossils, because there will only be low magnesian calcite. It is though that the original habit of the fossil shell controls the order ofsilicification. The fibrous calcite of bryozoans, brachiopods and corals is usually more susceptible to silica replacement than the monocrystalline calcite of the crinoid ossicles and echinoderm plates.

IV. TIME OF SILICIFICATION

WILSON (1966) concluded that two periods of silicification occurred in the Upper Jurassic sediments of South England: The first prior to calcite cementation, the second after calcite cementation. In the Gürün sediments no evidence of a pre-cement age for silicification has been found. Contrary, all interrelations between silicification fabrics and host grain suggest a late diagenetic stage for the formation of silica fabrics. Some of the observations indicating a late diagenetic stage can be summarized as follows:

Silica replaces both fossils, matrix and/or carbonate cement as well. Isolated crystals of carbonate are usually found floating within the matrix of the replacing silica indicating that silicification took place after rock cementation.

In some cases silica replaces carbonate cement of veins which is clearly late diagenetic origin.

The megaquartz usually fills either the upper or central part of cavities and is underlain or enclosed by a mosaic of carbonate cement, thus representing the later stages of cementation and cavity filling (Pl.83D).

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PART 11

SUMMARY OF DIAGENESIS

Although more than thirty different diagenetic process have been described in the literature there is, at present, no universally accepted definition of the term "diagenesis". For the purpose of this account "diagenesis" includes all the chemical, physical, and biological changes that sediments undergo from time of deposition until the present exclusive of weathering and metamorphism. The principal diagenetic processes to affect carbonates of the Gürün region are described in the above sections and includes biologic diagenesis, micritization, cementation, neomorphism, compaction, pressure solution, stylolitization, dolomitization, dedolomitization and silicification.

Modification of the carbonates investigated by burrowing organisms was contemporaneous with deposition and the first stage of diagenesis. Boring organisms, mainly algae, formed micritic envelopes and caused micritization of the grain-supported platform edge carbonate sand banks in the early stages of diagenesis.

Studies on Recent and Pleistocene carbonates have shown that carbonate sediments can become cemented in meteoric, marine and subsurface environments. Generally cementation in sea water produces aragonite or high-Mg calcite which is fibrous or micritic; whereas cementation in meteoric or subsurface waters is characterized by mostly equant calcite crystals. Meteoric environment includes vadose (above water table) and phreatic (below water table) zones. Many characteristics of the vadose zone diagenesis have been observed indicating the important role of vadose diagenesis in the sediments investigated. Marine environments include intertidal and subtidal. The second category consists shallow marine and deep sea environments. There is virtually no obvious difference in morphology and mineralogy between intertidal and shallow water cementation. Both types are characterized by high-Mg calcite or aragonitic fibrous crusts or micritic coatings. It is believed that fibrous calcite crusts observed as the first pore filling followed by random equant calcite mosaic, represent inverted intertidal or shallow marine cement preserving the original texture. In most instances the cement of the deep sea limestones is microcrystalline high-Mg calcite. Recent studies have shown that deep sea cementation is quantitatively unimportant and there has been no evidence showing that it was important during the Cretaceous Time. Contrary, evidence has been found showing that cementation of calcilturbidites took place in the vadose zone, following the emergence in the later stage of diagenesis. Although not abundant, calcite cement types encountered in the present study include crusts and random calcite mosaic. Fibrous crystals form a fringe on carbonate grains and is the first phase cement. High-Mg calcite debris probably lost its Mg ions at this time, almost invariably on a piecemeal basis allowing the original fabric to be retained in low-Mg calcite.

Distinction between sparry calcite formed by cementation and neomorphism is not an easy one. Criteria for distinguishing two types have been discussed. Neomorphism includes inversion (aragonite to calcite) and recrystallization (calcite to calcite) and is either aggrading or degrading. The former results an increase in the grain size while in the latter grain size decreases. The origin of the limestones of the Gürün region involved the deposition of large quantities of carbonate mud over a long interval of geologic time and conversion of this carbonate mud to lithified micrite is the most important aspect of neomorphism in the sediments investigated. Another important product of neomorphism is the formation of microspar and pseudospar. Microspar is a result of coalascive (?) neomorphism initiated by the removal of imprisoning cage of Mg ions surrounding calcite crystals and has a crystal size between 5-30 microns. Pseudospar consists larger neomorphic calcite crystals and are formed either like microspar or by inversion of aragonitic skeletons. The latter process involves either in situ transformation or solution and redeposition as calcite. In in situ transformation relics of organic shell structure may still be visible in the form of inclusions within the calcite mosaic. Second type inversion closely resembles cementation and can be divided depending on the relative timing of the solution of aragonitic skeletons, lithification of carbonate mud matrix and deposition of sparry calcite cement.

Some evidence suggests that clots of grumeleuse texture are microcrystalline calcite relics in a neomorphic sparry calcite and not mechanically deposited pelletoids in a cement. Lithification of carbonate mud to lithified micrite was relatively early diagenetic event. It is difficult to establish the relative time of other neomorphic processes. The growth of neomorphic sparry calcite and most types of inversion of aragonite to calcite in skeletal grains probably took place some time later, at least after the lithification of carbonate mud.

Evidence for compaction can be found, though it is believed to have generally not been too severe excluding calciturbidites, Stylolites are a common feature and are believed to have had more than one mode of formation.

Two stages of dolomitization are apparent in the Gürün sediments. The first stage is characterized by fine-grained dolomite that formed either as early phase of supratidal/intertidal sedimentation from original carbonate mud or from the alteration of lagoonal carbonate muds. The second stage of dolomite occurs as coarse rhoms, many larger than 100 microns, that developed in the grain-stone facies and partly replaced the clasts. These rhombs tend to be zoned as a result of variation in iron content. The observed resistance of matrix to dolomitization suggests that the matrix was stable (calcitic) and cohorent when replacement began. The present author is of the opinion that dolomitization of Gürün sedimetns caused by the mixing of sea water and ground water in the phreatic zone (Dorag dolomitization model). Extreme sea water evaporation and high Mg/Ca ratios in solution are not likely for the sediments investigated.

Dedolomitization (replacement of dolomite by calcite) has widely been observed in the north-west parts of the investigated area. There are two types of dedolomitization. In one type each dolomite rhomb is replaced by a mosaic of finer grained calcite, and in the other the dolomite rhombs are replaced by a more coarsely crystalline mosaic of calcite crystals. Dedolomitization is a late diagenetic surface and near surface phenomenon.

Silicification is a common diagenetic feature in the calcilturbidites of the Gurun Basin. It has been observed both in large scale, as cherts; and small scale, as void filling cement and as replacement of grains and/or matrix. Cherts have been found forming nodules, lenticular bodies and beds. Two types of chert have been differentiated; Granular Chert and Vitreous Chert. The former is usually found with packestones and grain stones while the latter is common in mudstones and wackestones. The source of the chert in the Gürün region is probably the dissolution of siliceous microorganisms. Field and petrographic evidence strongly suggests a replacement origin for cherts. Small scale silica fabrics can be divided into two genetic groups. (a) Cavity filling fabrics and (b) replacement fabrics. Replacement fabrics are common in the samples studied and includes megaquartz, spherulites of length-slow chalcedony, microcrystalline quartz and authigenic quartz crystals. Both chert formation and small scale silica fabrics are late diagenetic phenomena in the sediments investigated.

The relative time relationships of the main diagenetic processes and products observed in the carbonates of Gürün region are summarized in Table V - 3. No attempt has been made to establish boundaries between the stage of diagenesis - syndiagenesis, anadiagenesis and epidiagenesis - defined by FAIRBRIDGE (1967. Also see p. 126), since the stage at which the individual processes and products occur cannot be determined on the basis of available data. The considerable overlap of the processes shown in Table V-3 may reflect bed-to-bed or regional differences in the time at which these changes took place and, as CHILINGAR, BISSELL and WOLF (1967, p.180) state, "Two or more processes may be active simulataneously. They may overlap or the termination of one may mark the commencement of another process; and still other alterations may occur independently in both space and time".

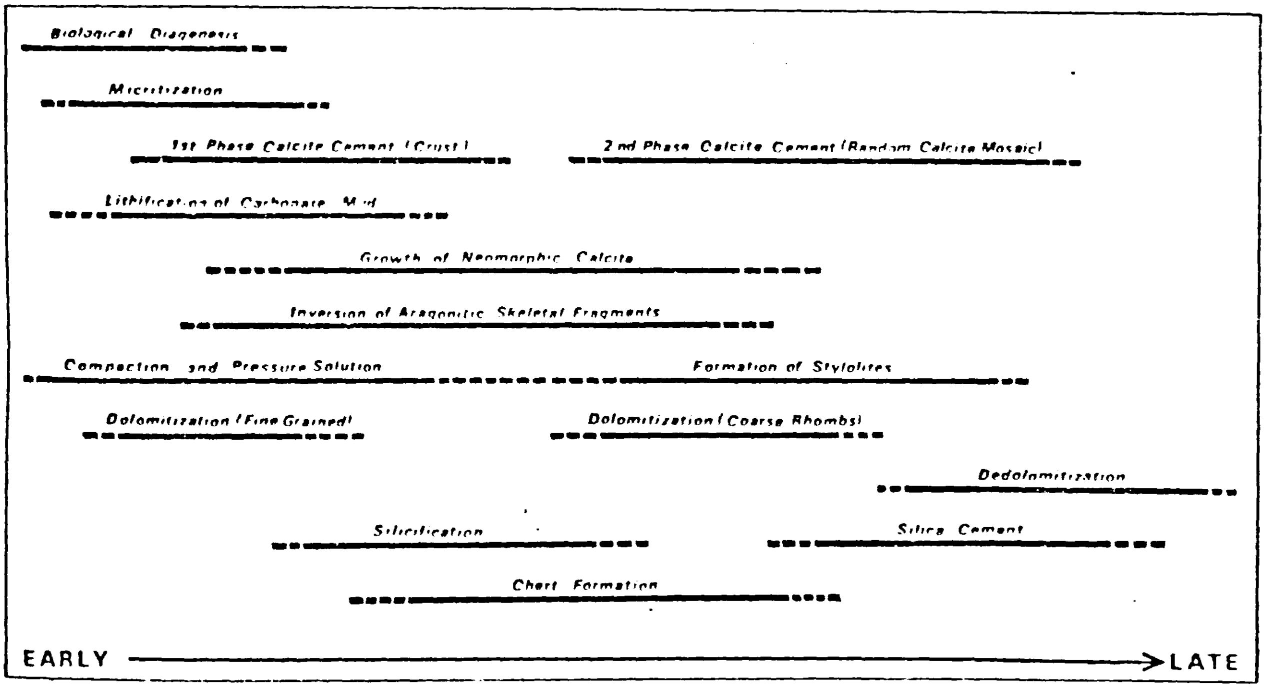


Table V-3: Relative time relationships of diagenetic changes in the Gürün carbonate rocks.

CHAPTER VI

SUMMARY AND CONCLUSIONS

Upper Cretaceous carbonates are widely exposed in the Gürün region. An area of approximately 30 x 80 kilometres was investigated in the present study in order to determine petrography, depositional environments, and diagenetic modifications of these rocks for the first time.

Upper Cretaceous in the region is characterized by a platform where shallow water carbonate deposition persisted, fringed by rudist knoll-reefs.

The development of the platform is associated with the regional Upper Cretaceous transgression.

Transgression reached its maximum amplitute during Maestrichtian time and calciturbidites deposited in the region.

Carbonate deposition was replaced by terrigeneous deposits in the Early Tertiary as diastrophism accelerated in the region.

Most of the sediments were deposited within a few decimetres of mean sea-level.

DUNHAM's (1962) carbonate rock classification has been employed, and four major grain types have been recognized: Lithoclasts, pelletoids, coated grains, and skeletal grains.

Ideal carbonate platform model and 24 standard microfacies types proposed by WILSON (1975) have been found applicable to the Gürün carbonates and with some modifications, employed in the present study. Based on the petrographic, palaeontologic, textural, and structural criteria 14 Micro-facies types have beeb recognized, Five distinct facies belts have been recognized which include, tidal flats, platform edge carbonate sand banks, knoll reefs, foreslope and toe of slope carbonates.

Tidal flats include supratidal, intertidal, and subtidal (both restricted and open marine) environments. Supratidal facies has been recognised by the presence of fenestral fabric, mud cracks, brecciation, and irregular laminations formed by algal-mats. Pure lime mudstone and algal lime mudstone, and lithoclasts composed small slabs of these lithologies characterize these deposits. Intertidal and restricted subtidal deposits are characterized by pelletoidal wackestones/packstones with abundant miliolid foraminifera. Stromatolites characterize intertidal deposits. Normal marine (subtidal) deposits are recognized by fossil content and scarcity of stratification due to extensive burrowing.

Coated grains, mostly bioclasts, in a sparry calcite cement is the typical lithology of platform edge carbonate sand banks. The most distinctive feature of this facies is its large scale cross bedding.

Belts of knoll shaped reefs were present on gentle slope at the edge of the shelf margin where rudists played the most important role. Rudist mudstone is the dominant lithology.

Deeper water deposits are represented by the alternation of coarse calcarenite, fine grained calcarenite-coarse grained calcilutite and pelagic lime mudstones and believed to be deposited by turbidity currents. Proximal and distal environments have been differentiated. The source area was situated in the northeast. Compaction and silicification are the dominant diagenetic processes operating in the calciturbidites.

Eastern part of the investigated area is characterized by alternation of ophiolitic conglomerate, sandstone, siltstone, and mudstones of turbiditic origin. The regional picture for this part is a progradational deltaic complex to the south that fed a turbidite basin located to the north. An Upper Jurassic age for the origin of ophiolites, and Senonian age for the time of their emplacement have been accepted.

In this account "diagenesis" includes all the chemical, physical, and biological changes that sediments have undergone from time of depositon until the present day exclusive of weathering and metamorphism.

The principal diagenetic processes to affect sediments of the Gürün region are biological diagenesis, micritization, cementation neomorphism, compaction, pressure solution, stylolitization, dolomitization, dedolomitization, and silicification.

Biologic activity was contemporaneous with deposition and the first stage of diagenesis. Boring organisms formed micritic envelopes and caused micritzation of the grain supported carbonates. Compaction was not severe and only played an important role in the diagenesis of calciturbidites. Stylolites are common and show more than one mode of formation.

Cementation took place in meteoric, marine, and probably subsurface environments. Cement types encountered are crusts and random calcite mosaic. The criteria for distinguishing sparry calcite formed by cementation and neomorphism have been given. Subaerial exposure and shallow water conditions during deposition of the Gürün sediments favoured early diagenetic cementation with calcite. Two phases of calcite cementation occured in the grainstone facies. The first phase is a crust around grains and the second phase fills the remaining pore spaces. Calcite shell fabrics have also been discussed.

Neomorphism includes inversion (argonite to calcite), recrystallization (calcite to calcite) and is either aggrading or degrading. The origin of the limestones of the Gürün region involved the deposition of large quantities of carbonate mud and convers ion of this carbonate mud to lithified micrite is the most important neomorphic process. It was relatively early diagenetic event. Formation of microspar and pseudospar have also been found important in some samples.

Two stages of dolomitization have been observed. The first stage is characterized by fine grained dolomite and is considered to be of early diagenetic origin. The coarse dolomite rhombs found in the grainstone facies and partly replaced the grains characterize a later stage and may be zoned. Dolomitization of the Gürün sediments was caused by the mixing of seawater and groundwater in the phreatic zone (Dorag dolomitization model). Extensive seawater evaporation and high Mg/Ca ratios in solutions are not likely.

Dedolomitization is a late diagenetic surface or near surface phenomenon and two types of dedolomitization have been observed. In one type the dolomite rhombs are replaced by a mosaic of finer grained calcite, and in the other the dolomite rhombs are replaced by a more coarsely crystalline mosaic of calcite crystals.

Silicification is widespread in the calciturbidites and has been observed both in large scale, as cherts; and in small scale, as void filling cement or replacement of grains and/or matrix. A replacement origin for the cherts of the Gürün region has been revealed. Two types of chert have been differentiated. Granular chert is usually found with packstones and grainstones; vitreous chert is common in mudstones and wackestones. Small scale silica replacement fabrics are common and includes mega quartz, spherulites of length slow chalcedony, microcrystalline quartz and authigenic quartz.

The relative time relationships of various diagenetic processes affecting the Gürün carbonate rocks have been summarized in Table V-3.

The depositional facies of the Gürün region is broadly similar to those of the Golden Lane and Poza Rica Trends in Mexico which are the source of more than 2 billion bbl of oil.

Similarities:

1. Shallow-water facies of the Golden Lane and Poza Rica trend and of the Gürün carbonate platform developed adjacent to a deeper, open marine facies.

2. The rocks overlying the El Abra and Tamabra limestones and those overlying the Gürün limestones are open-shelf or basinal in many areas and are similar in sequence, and lithology. These overlying deposits indicate that, after deposition of the shallow-shelf and organic build-up facies both areas were covered next by deeper water deposits.

3. Formations in the Golden Lane and Poza Rica trend and Gürün platform are predominantly carbonate rocks with rudist pelecypods as an important constituent.

4. Several facies of the Golden Lane and Poza Rica trend are similar to those of the Gurun carbonates (see above p.).

Differences:

1. The carbonate rocks in the Golden Lane and Poza Rica trend are Albian and Cenomanian in age whereas in the Gürün region shallow-water carbonates are Cenomanian-Campanian, and deeper water carbonates are Maestrichtian age.

2. The Golden Lane platform was tilted by an Early Tertiary orogeny which caused the exposure of part of the El Abra Limestone to subaerial weathering and developed considerable vuggy and cavernous porosity. The shelf carbonates of the Gürün region were not exposed to surface weathering for a long geologic time interval.

3. There is convincing evidence that a hot arid climate was prevailing during the Cretaceous time in the Mexico region (see WILSON, 1975, p.327) whereas the present study has shown that the climate was moderate in the Gürün region during that time.

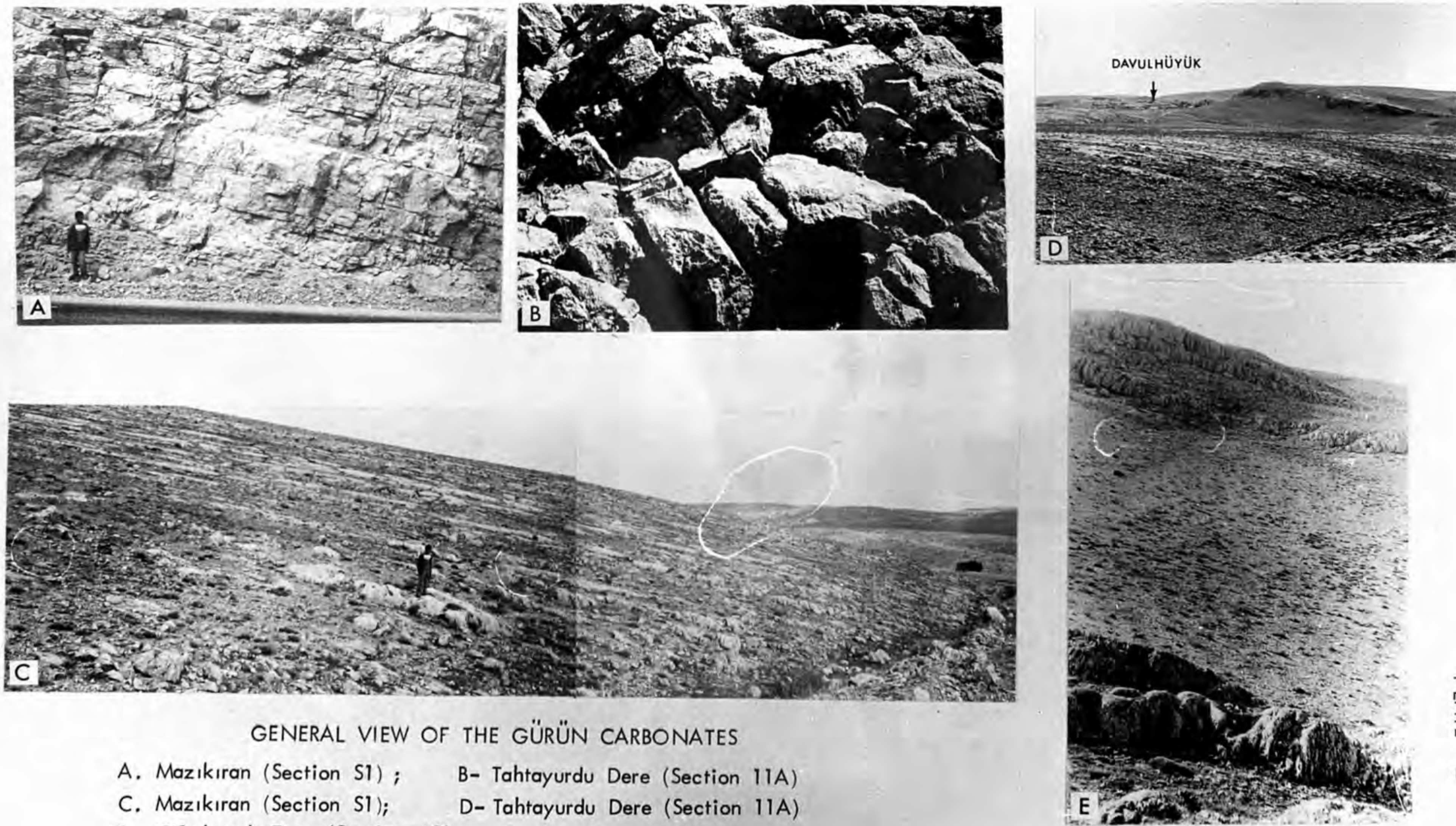
The most important factor controlling the hydrocarbon distribution in the Mexico and Gürün regions seems to be structural. Early exposure of the Golden Lane platform to subaerial weathering caused leaching and development of porosity. On the other hand in the

Gürün carbonates cementation eliminated porosity.

Another important factor is the source rock. The Jurassic shales are thought the source beds of oil in the Mexico region by most of the workers, whereas, in the Gürün region such a shale source is missing.

Finally the Golden Lane platform is overlain by non-porous Upper Cretaceous rocks while the Gürün carbonates have an extensive outcrop in most parts.

PLATE - 2

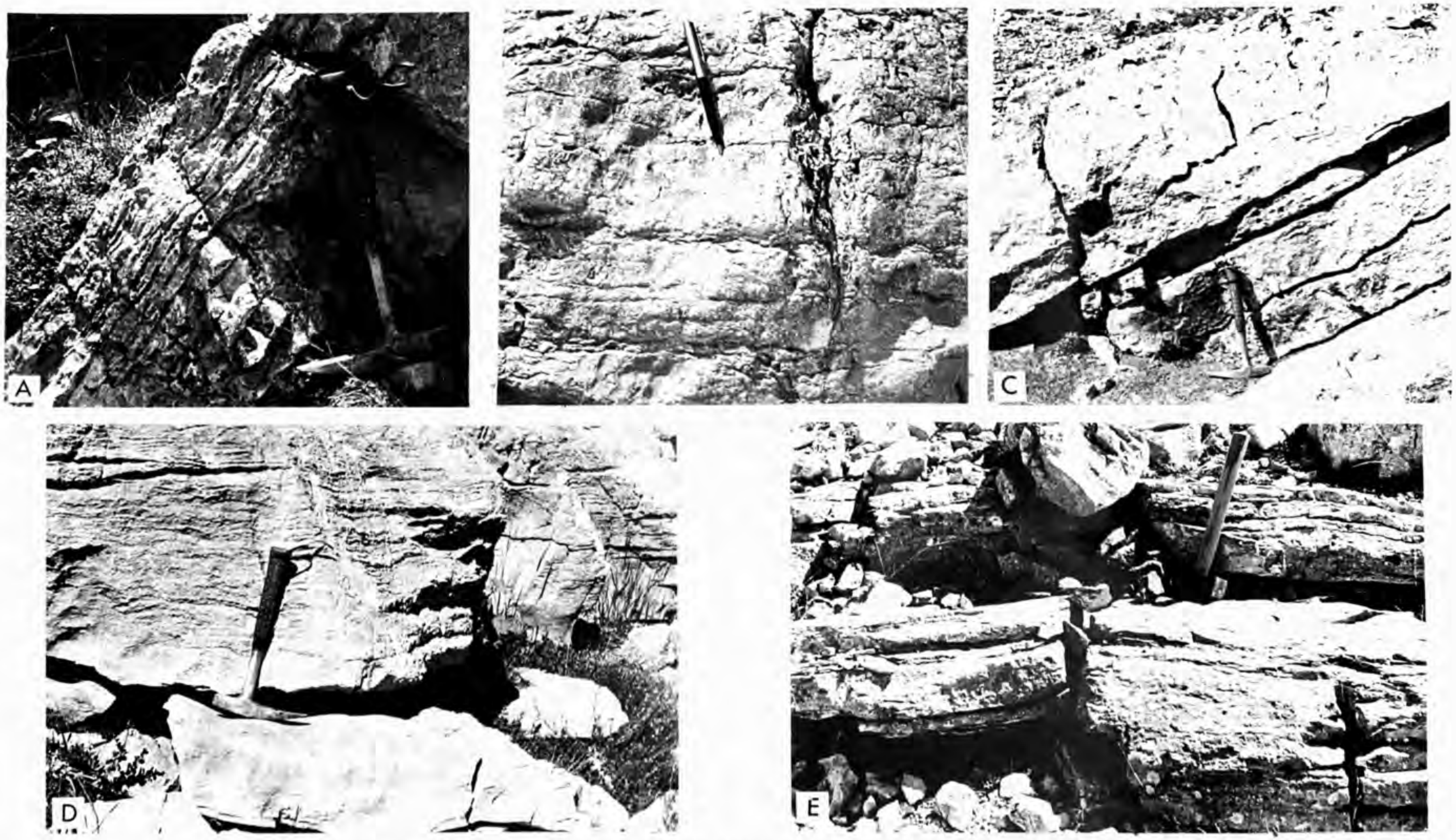


GENERAL VIEW OF THE GÜRÜN CARBONATES

A. Mazıkıran (Section S1) ; B- Tahtayurdu Dere (Section 11A)
C. Mazıkıran (Section S1); D- Tahtayurdu Dere (Section 11A)
E. Ciftekoyak Tepe (Section: 6B)

PLATE - 3

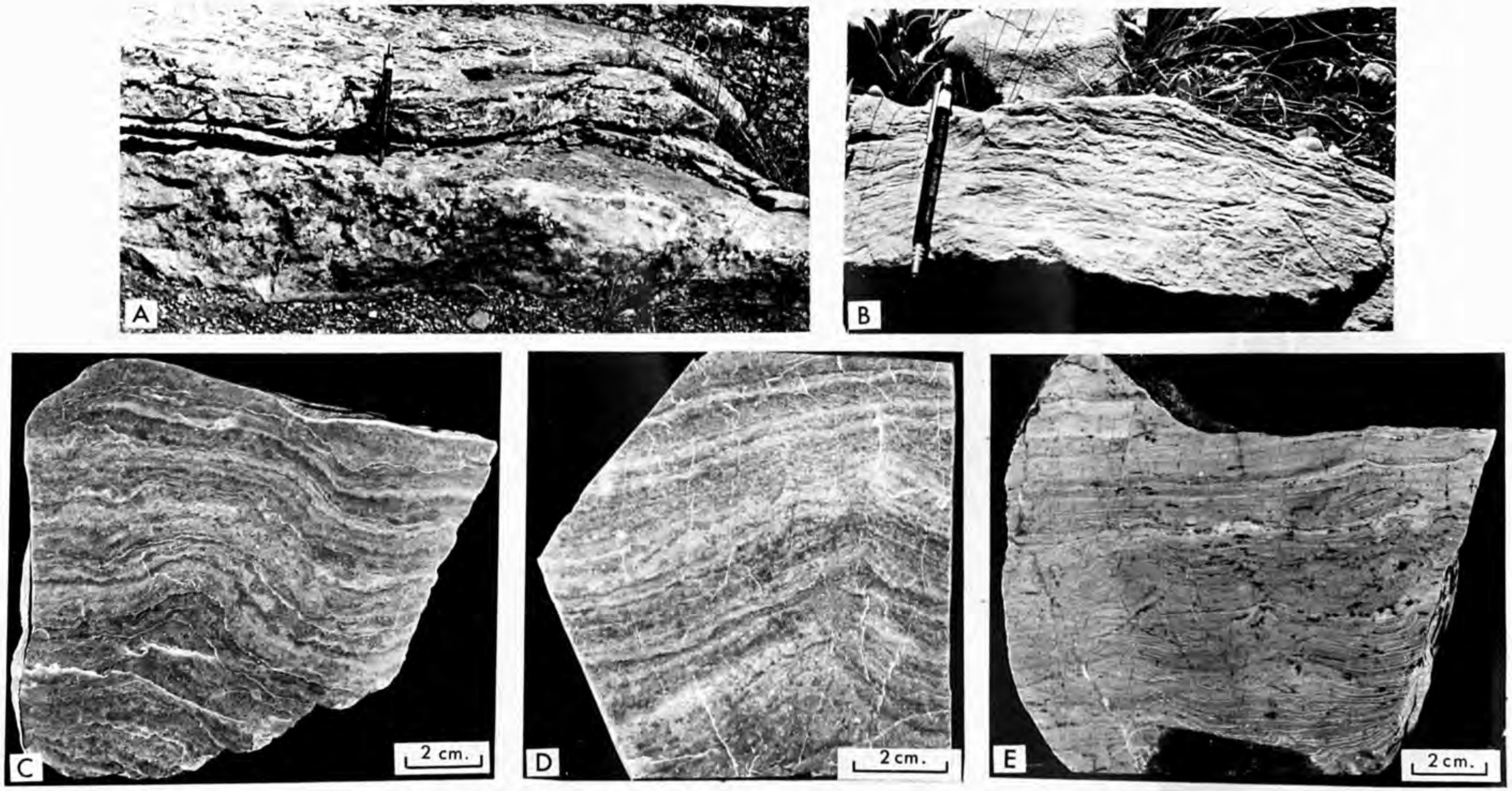
FOLDING IN THE CALCITURBIDITES



FIELD APPEARANCES OF STROMATOLITES AND ALGAL-LAMINATED SEDIMENTS-1

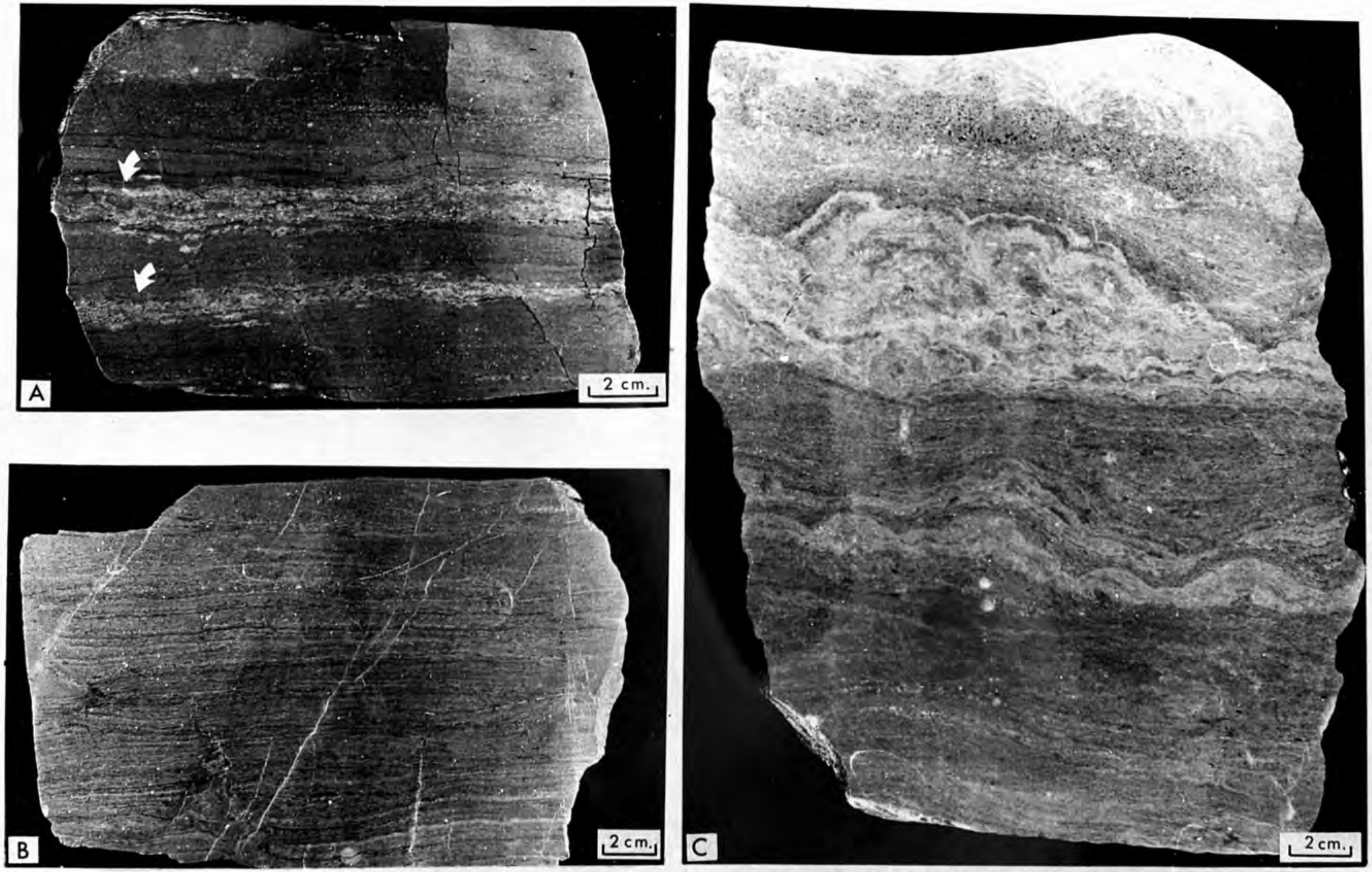
A. Keşlik Dere. Section 10A. Sample No.1031
B. Keşlik Dere. Section 10A. Sample No.1013
C. Keşlik Dere. Section 10A. Sample No.1017
D. Taslıçal Tepe. Section 1. Sample No.106
E. Bölücek Tepe. Section 6A. Sample No. 609

PLATE -5



STROMATOLITES AND ALGAL-LAMINATED SEDIMENTS-2

A. Algal-laminated sediments forming a dome. South of Mazıkıran. Section 8A. Sample No.828
B. A specimen of stromatolitic limestone. Tahtayurdu Dere. Section 11A. Sample No.1122
C. Polished hand specimen. Locality: South of Mazıkıran. Sample No.Mk 1
D. " " : Taşlıcal Tepe. Section 1. Sample No.113A
E. " " : Tahtayurdu Dere. Section 11A. Sample No.1121-1122



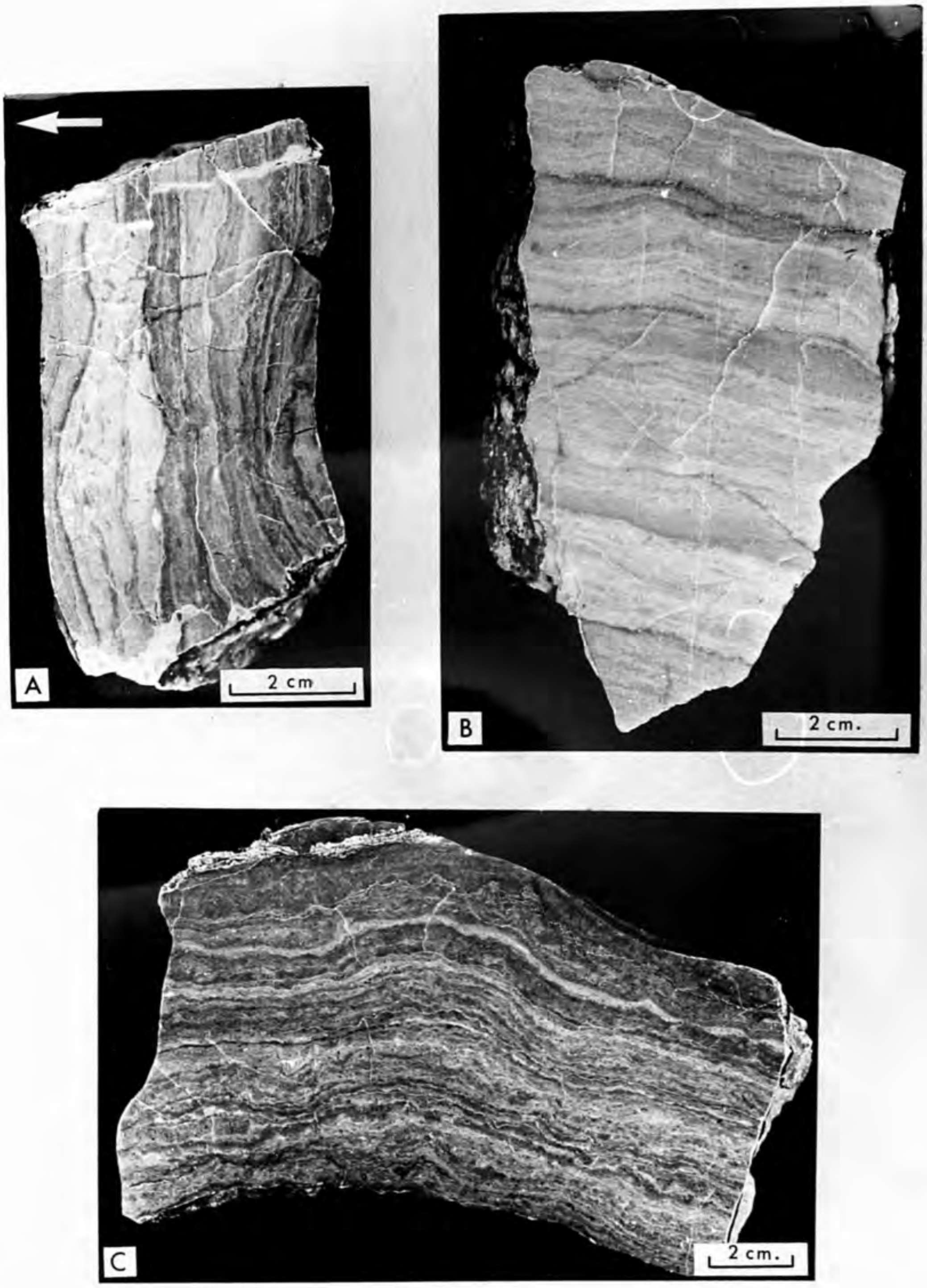
STROMATOLITES AND ALGAL LAMINATED SEDIMENTS-3

A. A polished hand specimen showing alternation of dolomitic (arrows) and calcitic (darker in photograph) laminae. Sample No.74/26; Location: Resadiye (R47).

B. Polished hand specimen of algl-laminated rock. Location: Davulhüyük.

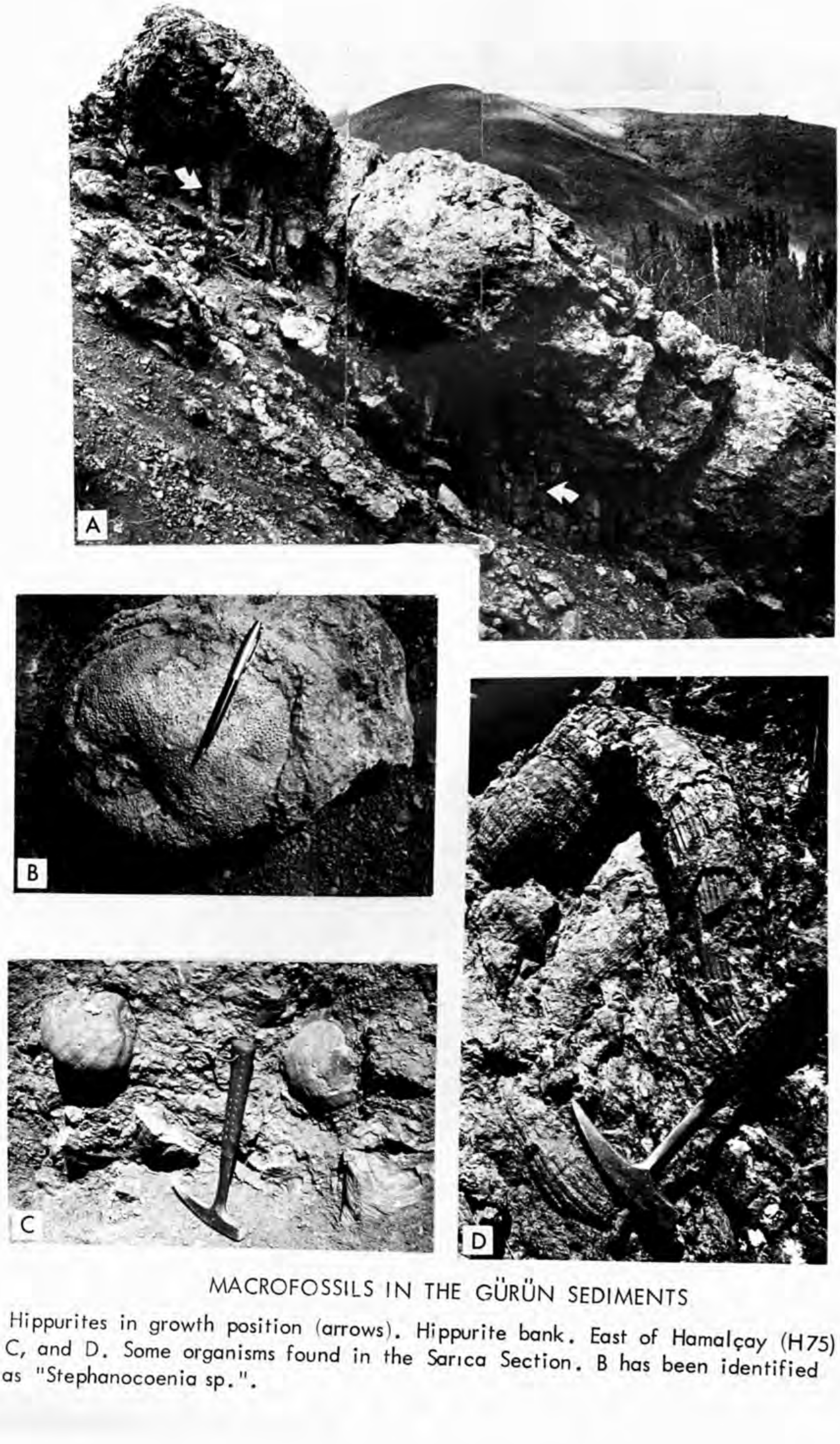
C. Typical apperance of stromatolitic limestone. Polished hand specimen. Location: Mazıkıran. Sample No.72/10

PLATE - 7



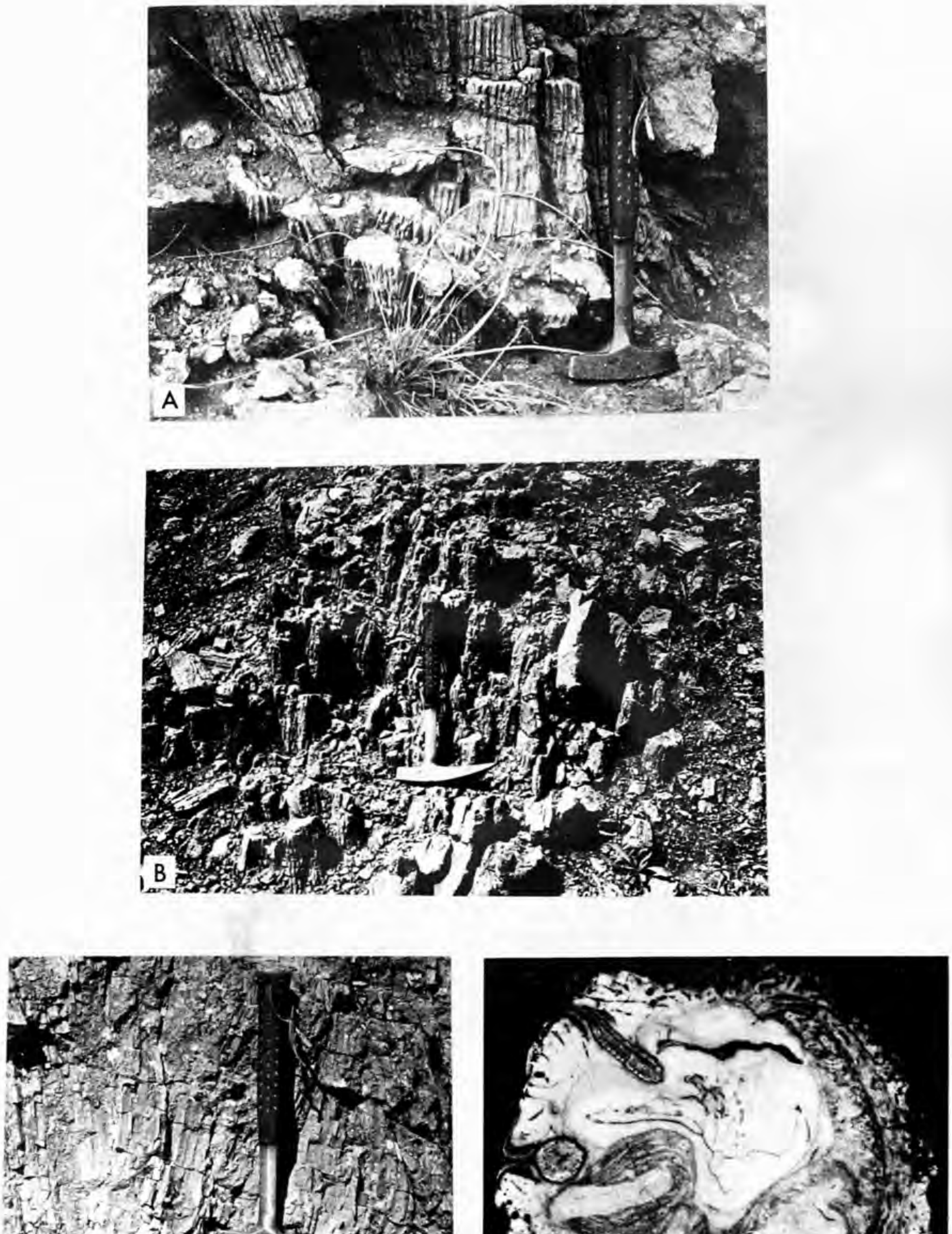
STROMATOLITES AND ALGAL LAMINATED SEDIMENTS-4

A. Polished hand specimen (arrow shows the top). Section 1. Sample No.1/125B
B. Highly dolomitized stromatolitic limestone (For thin section microphotographs see Plate 69A . Section: 7A; Sample No.707
C. Polished hand specimen. Section: 8A; Sample No.Mk.3



MACROFOSSILS IN THE GÜRÜN SEDIMENTS

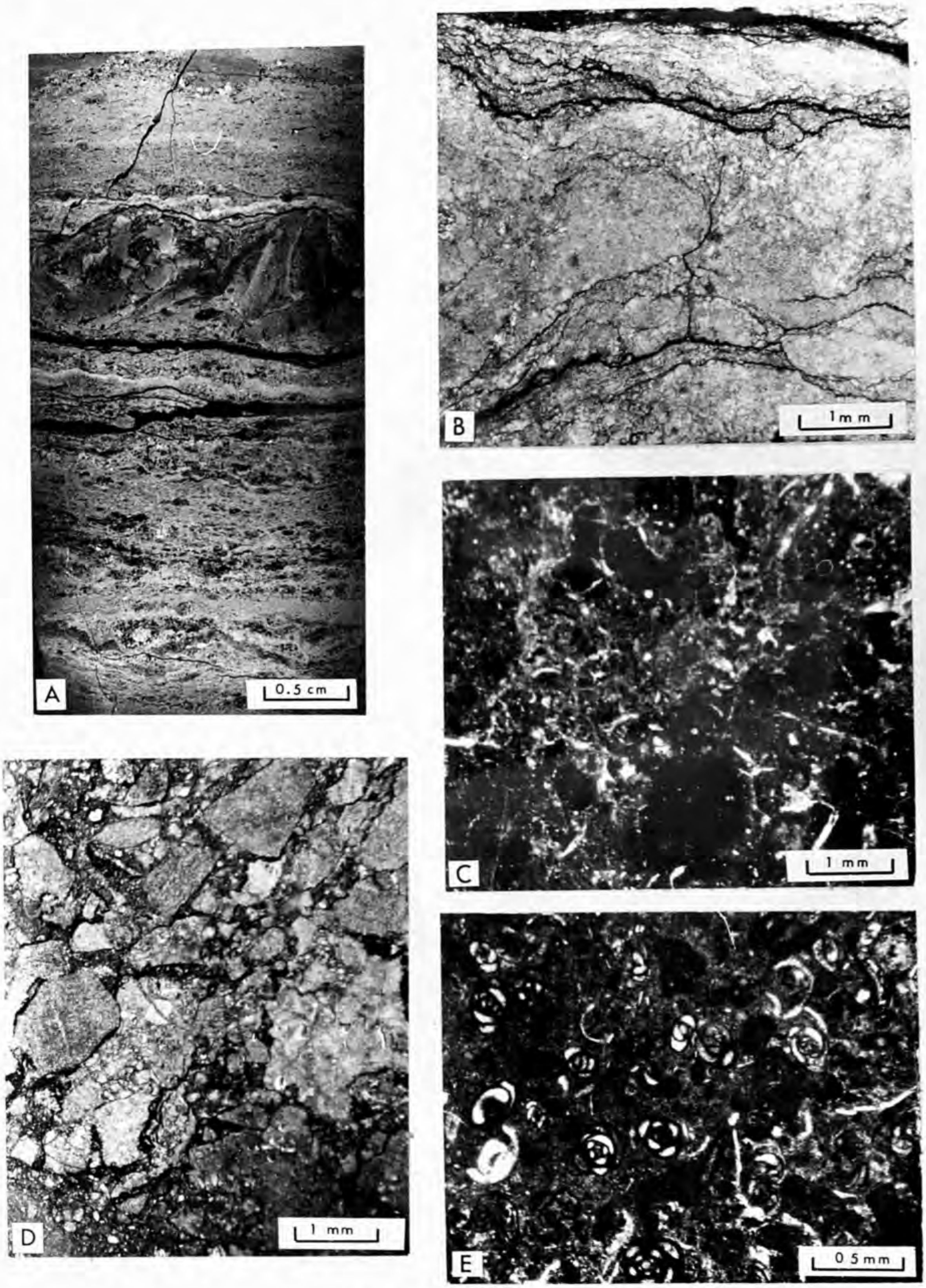
A. Hippurites in growth position (arrows). Hippurite bank. East of Hamalçay (H75)
B, C, and D. Some organisms found in the Sarıca Section. B has been identified as "Stephanocoenia sp.".



MACROFOSSILS IN THE GÜRÜN SEDIMENTS

Hippurites observed in the Hamalçay Dere. A, B, and C show hippurites in growth position. D is a polished hand specimen cut from a hippurite.

PLATE - 10



MICROFACIES TYPES - I

A. Lithoclastic packstone. Microfacies C. Negative print from a thin section. Sample No.471-472

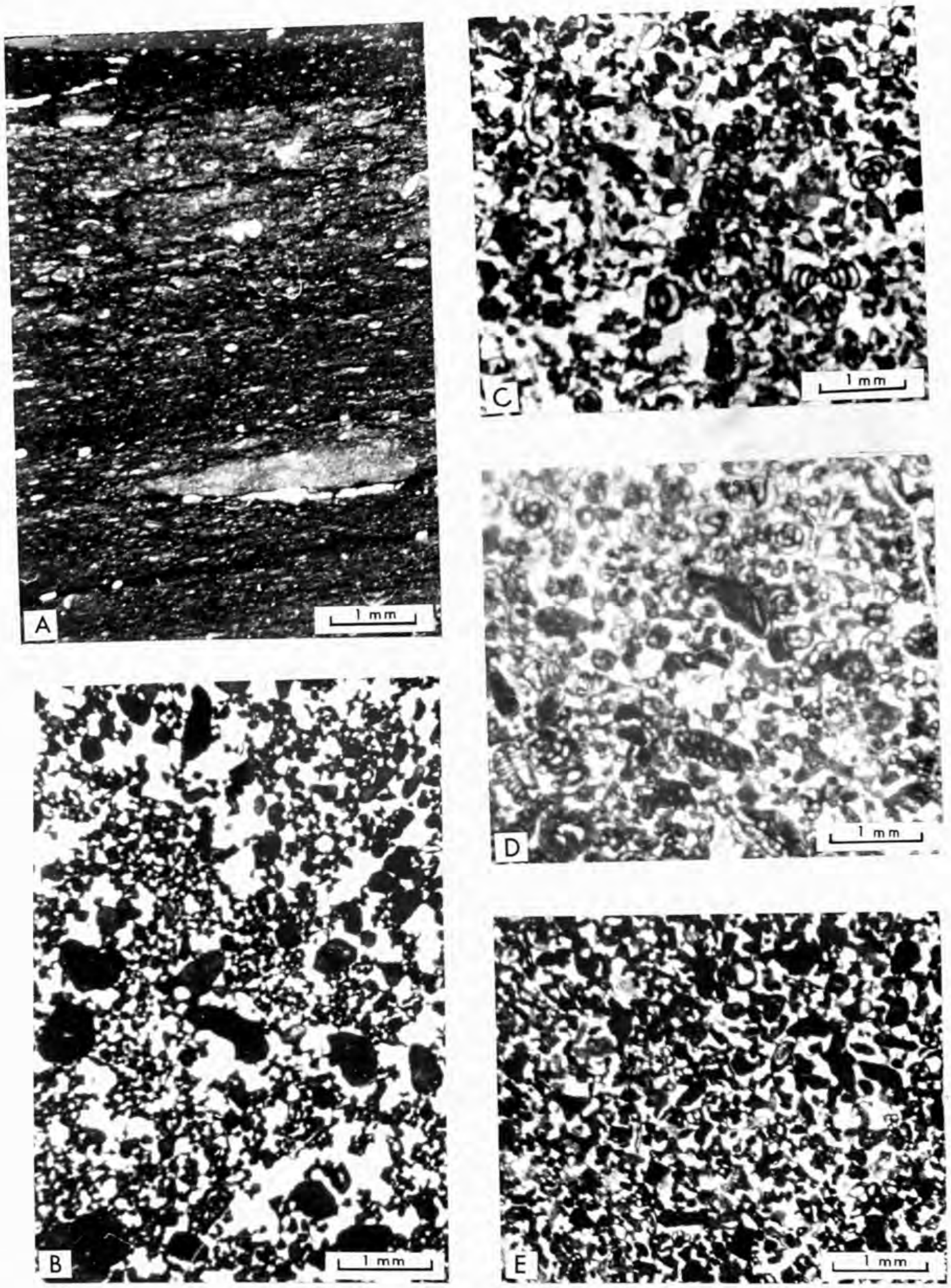
B. Algl-mud lime mudstone. Microfacies B. Peel. Sample No.471-472

C. Lithoclastic packstone. Microfacies C. Sample No.237

D. " . " . " 1176

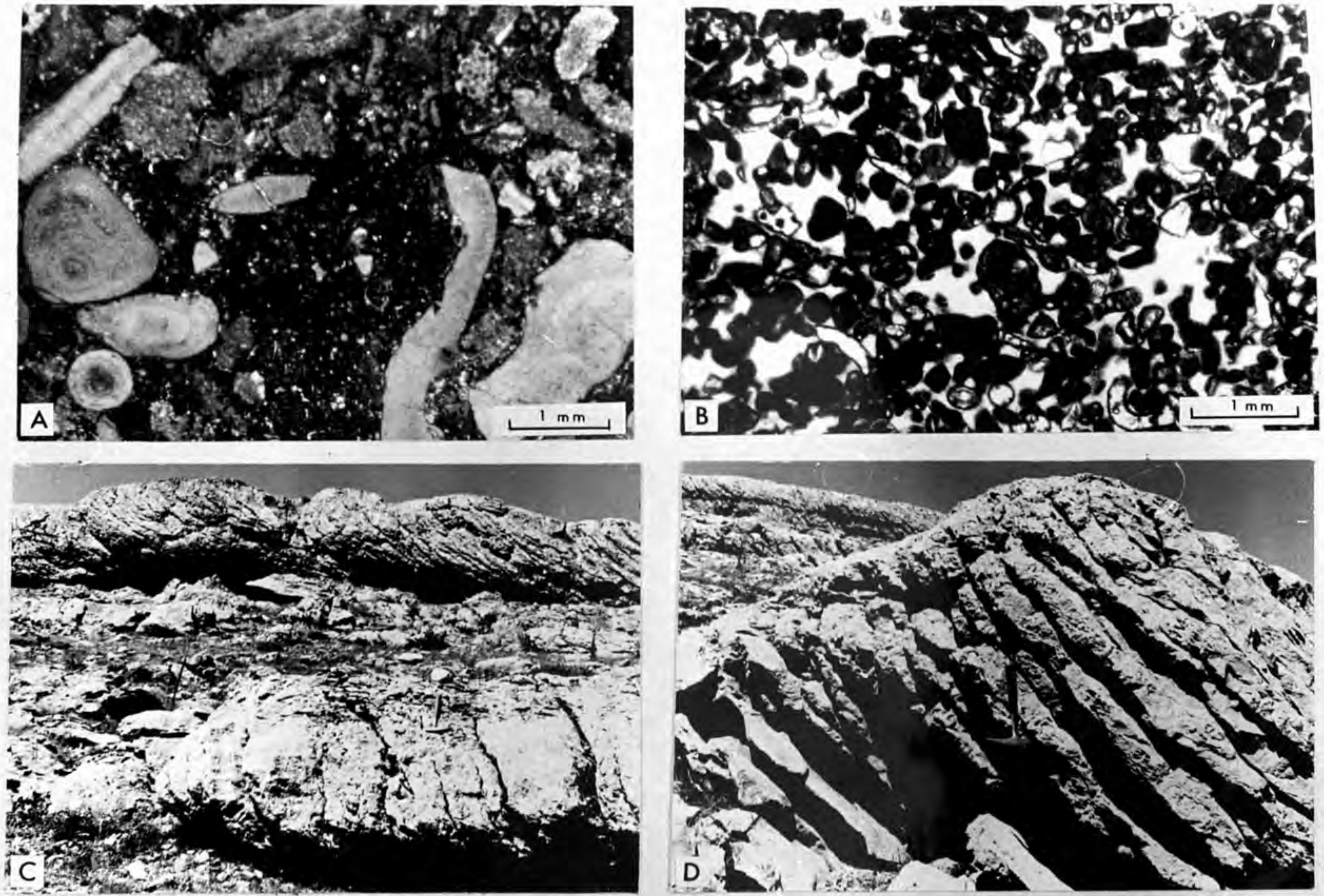
E. Lime wackestone with abundant miliolids. Microfacies E. Sample No.157

PLATE - 11



MICROFACIES TYPES - 2

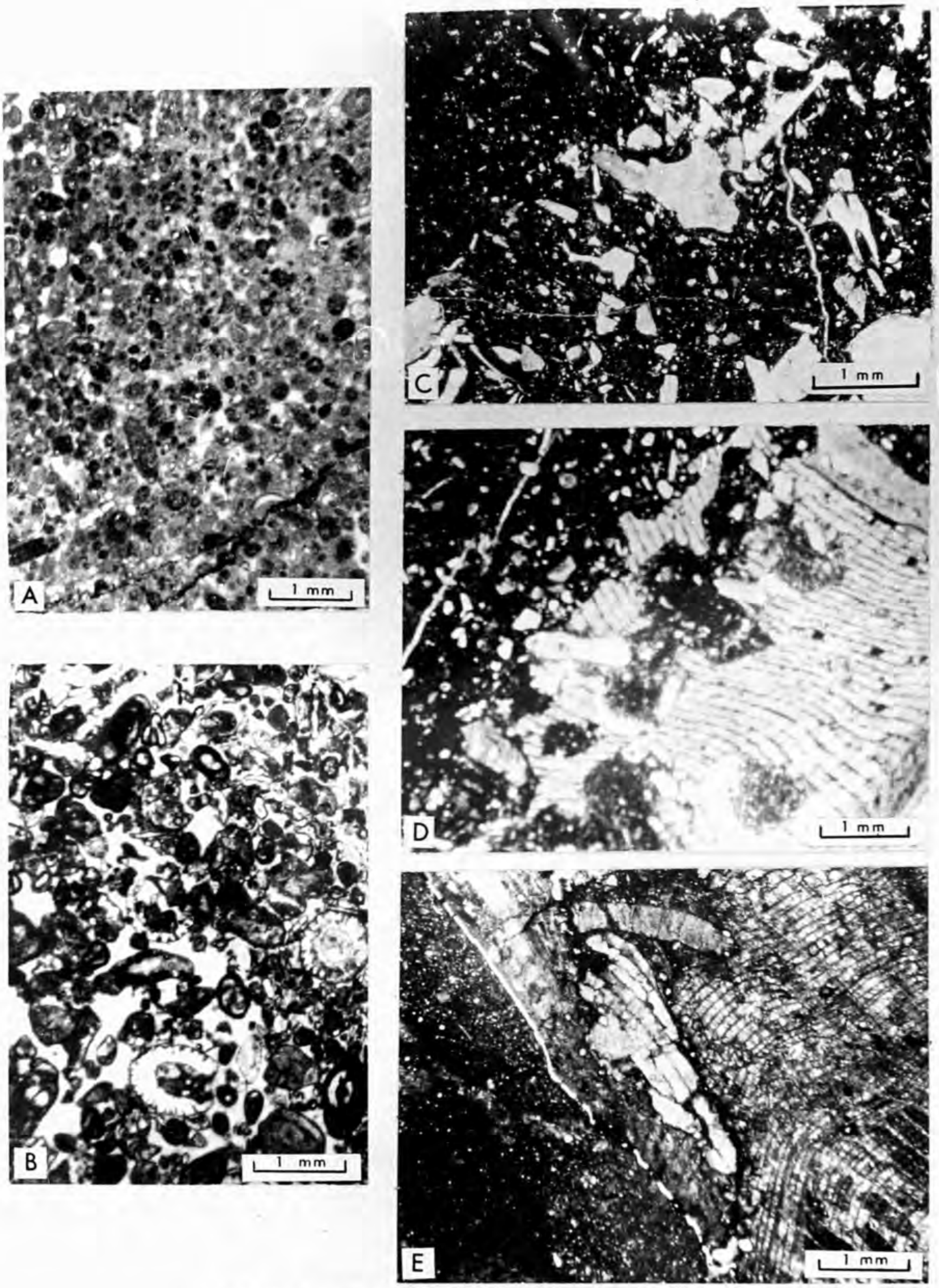
A. Compacted Microfacies E. Sample No.1151
B. Pelletoidal lime grainstone. Microfacies G. Sample No.237
C. Miliolid grainstone. Microfacies F. Sample No.609C
D. Miliolid/pelletoidal grainstone. Microfacies F/G. Sample No.220
E. Pelletoidal grainstone. Microfacies G. Sample No.145



MICROFACIES TYPES - 3

A. Lime wackestone with normal marine fauna. Microfacies H. Sample No.10/108/2
B. Coated grainstone. Microfacies I. Sample No.1210-1211. C. Large-scale cross stratification in the carbonate sand bank lithofacies. Seation: 8B ; Location: Kaz Tepe(D35) D. Detail of C.

PLATE - 13



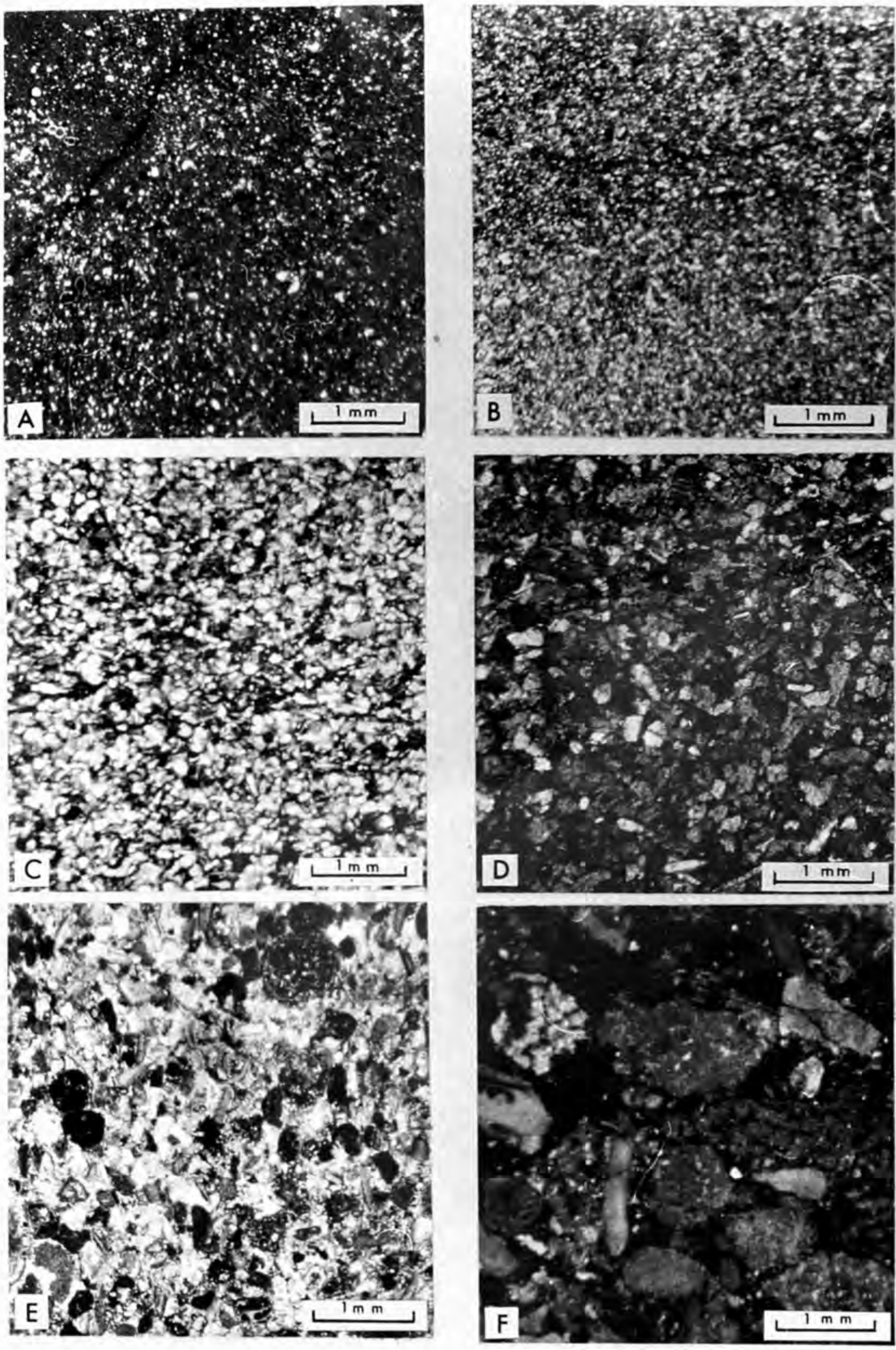
MICROFACIES TYPES - 4

A. Negative print of a thin section showing micritization of grains. Coated grainstone. Microfacies I. Sample No.73/36.

B. Coated grainstone. Microfacies I. Sample No.833

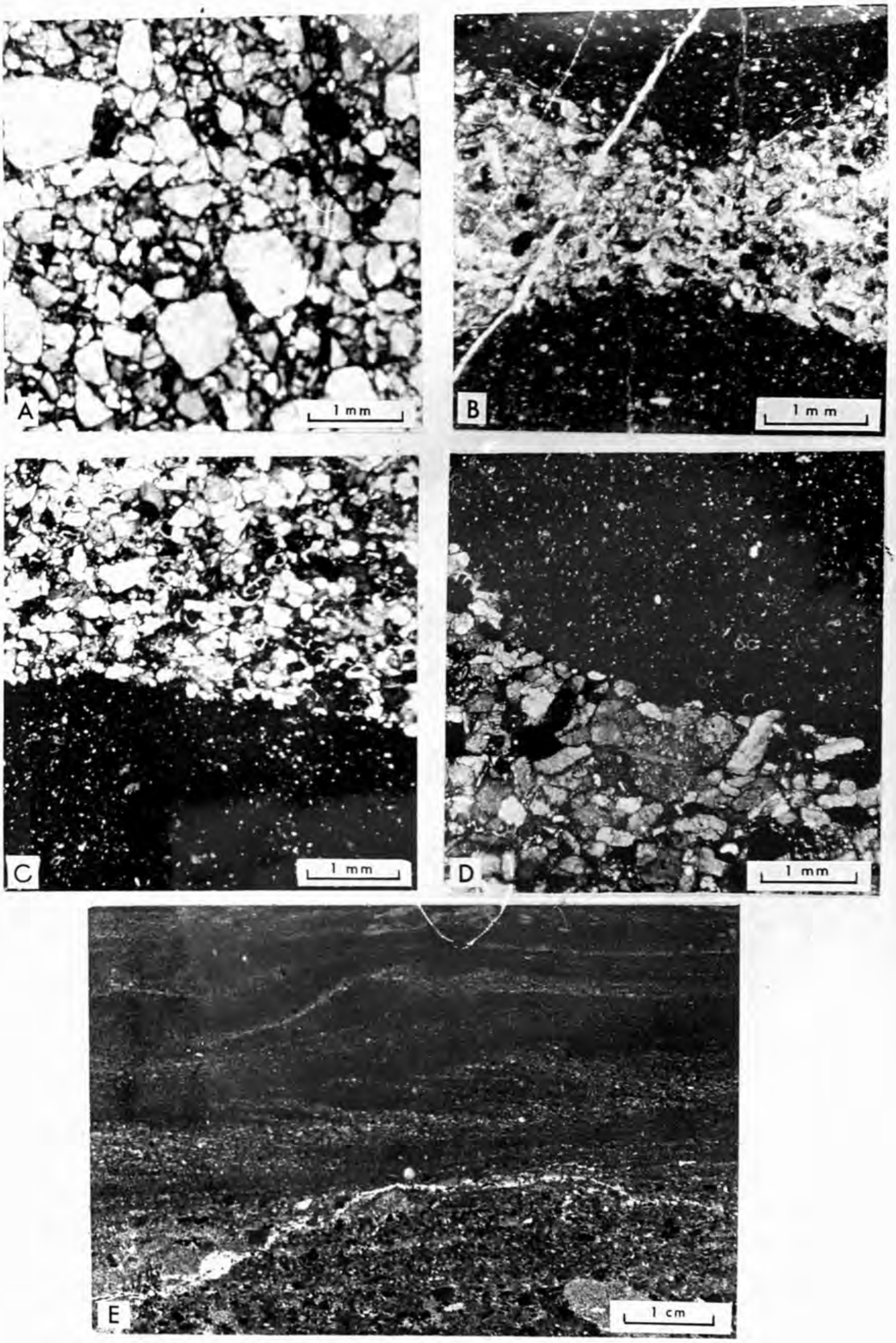
C, D, and E. Rudist lime mudstone. Microfacies J. Sample Nos. C-358C ; D - 514 E - 2

PLATE - 14



MICROFACIES TYPES - 5

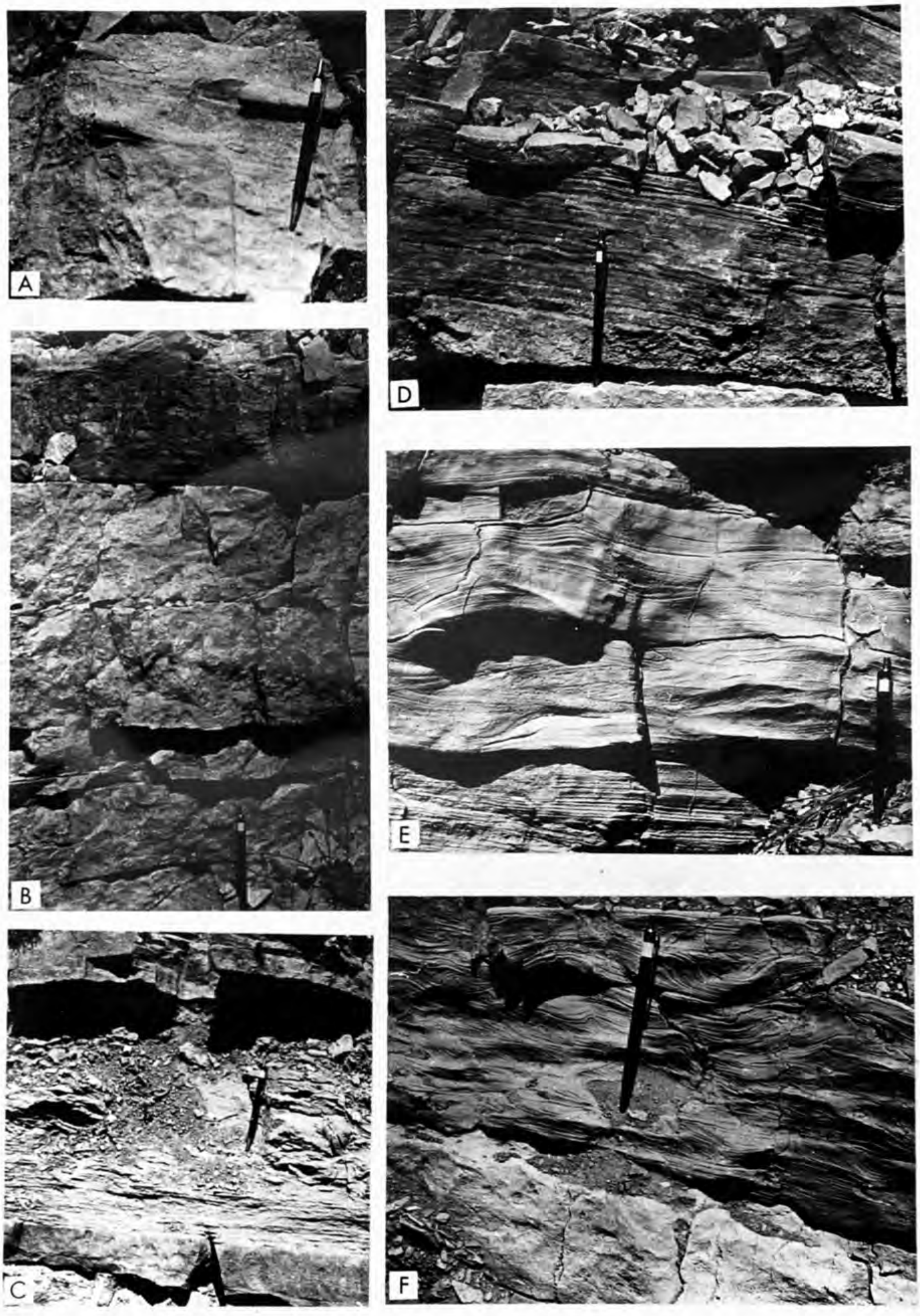
A. Pelagic lime mudstone admixed with calcilutite. Microfacies M. Sample No.901
B. Very fine-grained calcilutite. Microfacies N. Sample No.925/1
C, and D. Fine-grained calcilutite. Sample No.1092/3.Peel. D - 502.
E. Coarse-grained calcilutite. Sample No.911.
F. Bioclastic packstone. Microfacies L. Sample No.1065.



MICROFACIES TYPES - 6

A. Lithoclastic packstone. Microfacies K. Sample No.74/59
B. Pelagic lime mudstone-skeletal debris alternation. Sample No.350
C. " . " .914/2. Peel.
D. " . " .73/2 . Peel.
E. Cross bedding in calciturbidites. Negative print from a peel. Sample No. 1049.


A
B
C
D
E
F
G
H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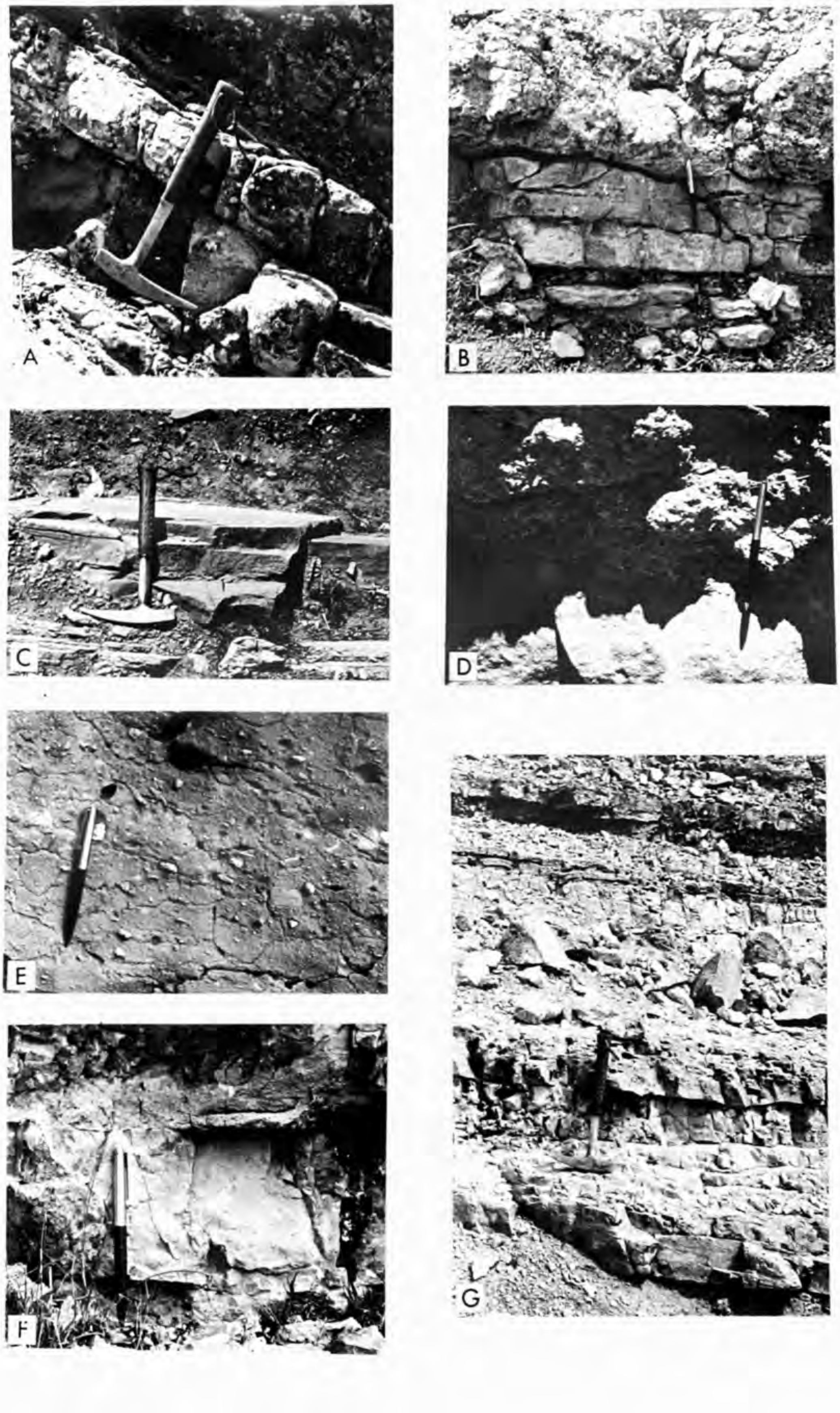
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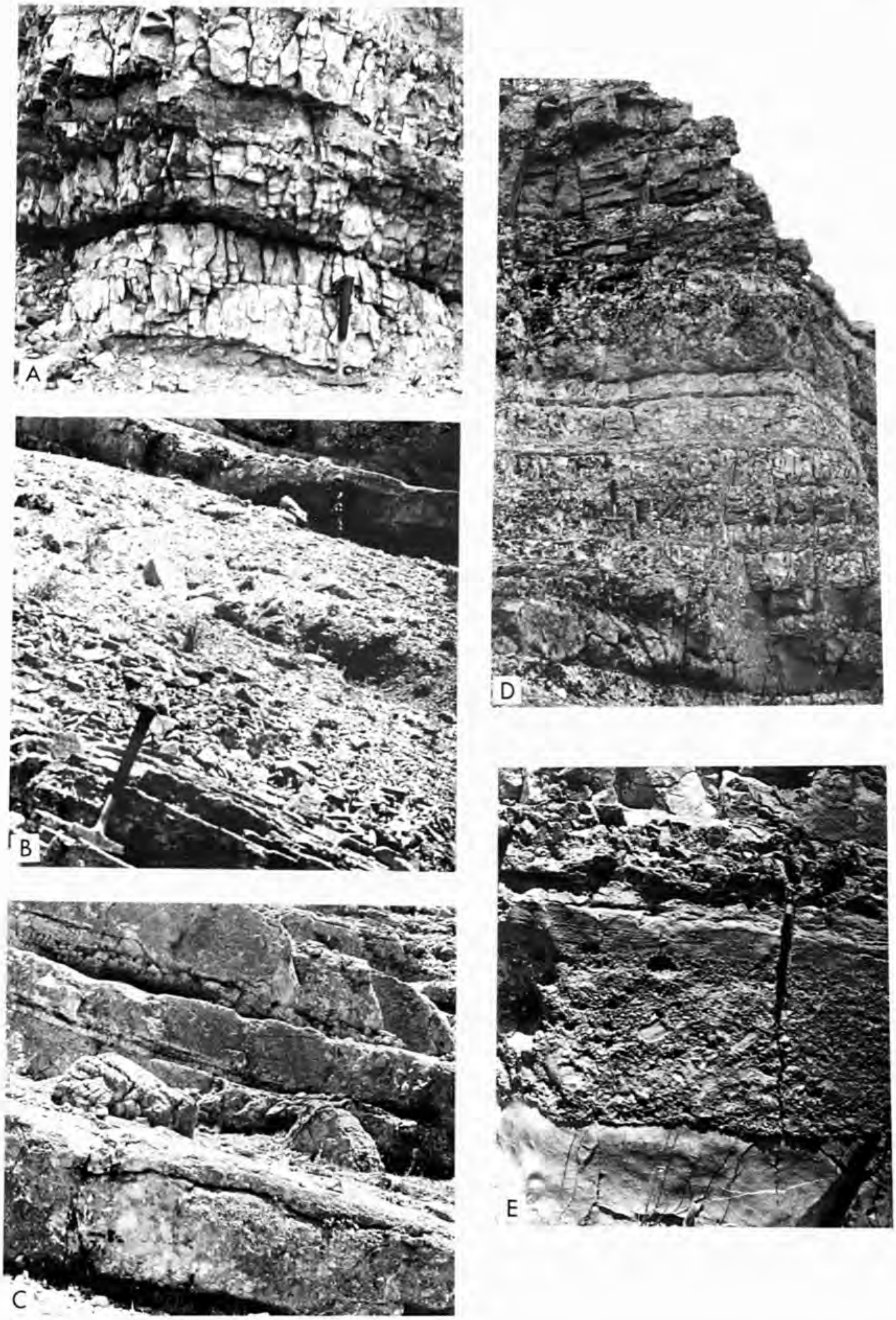
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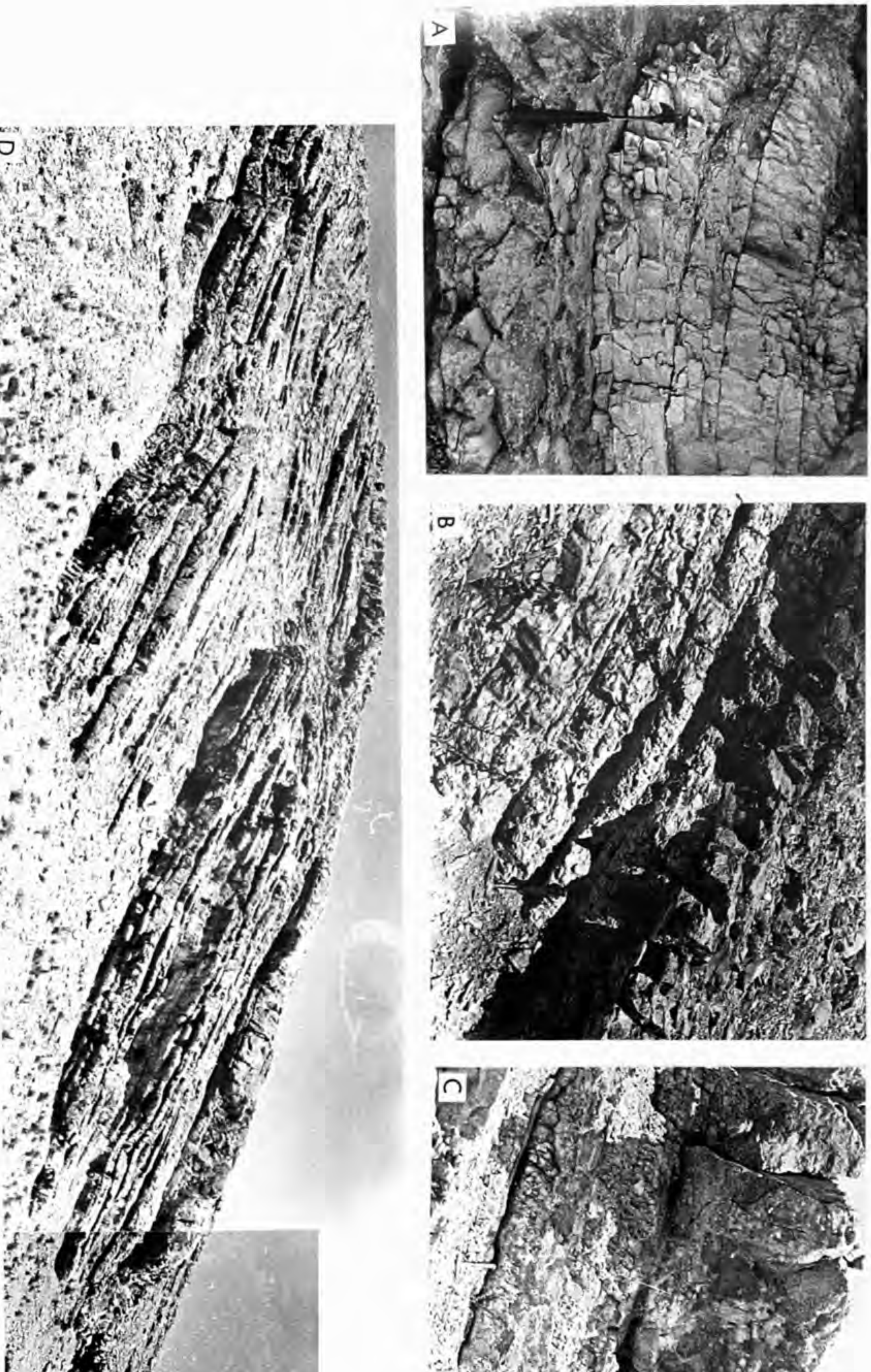
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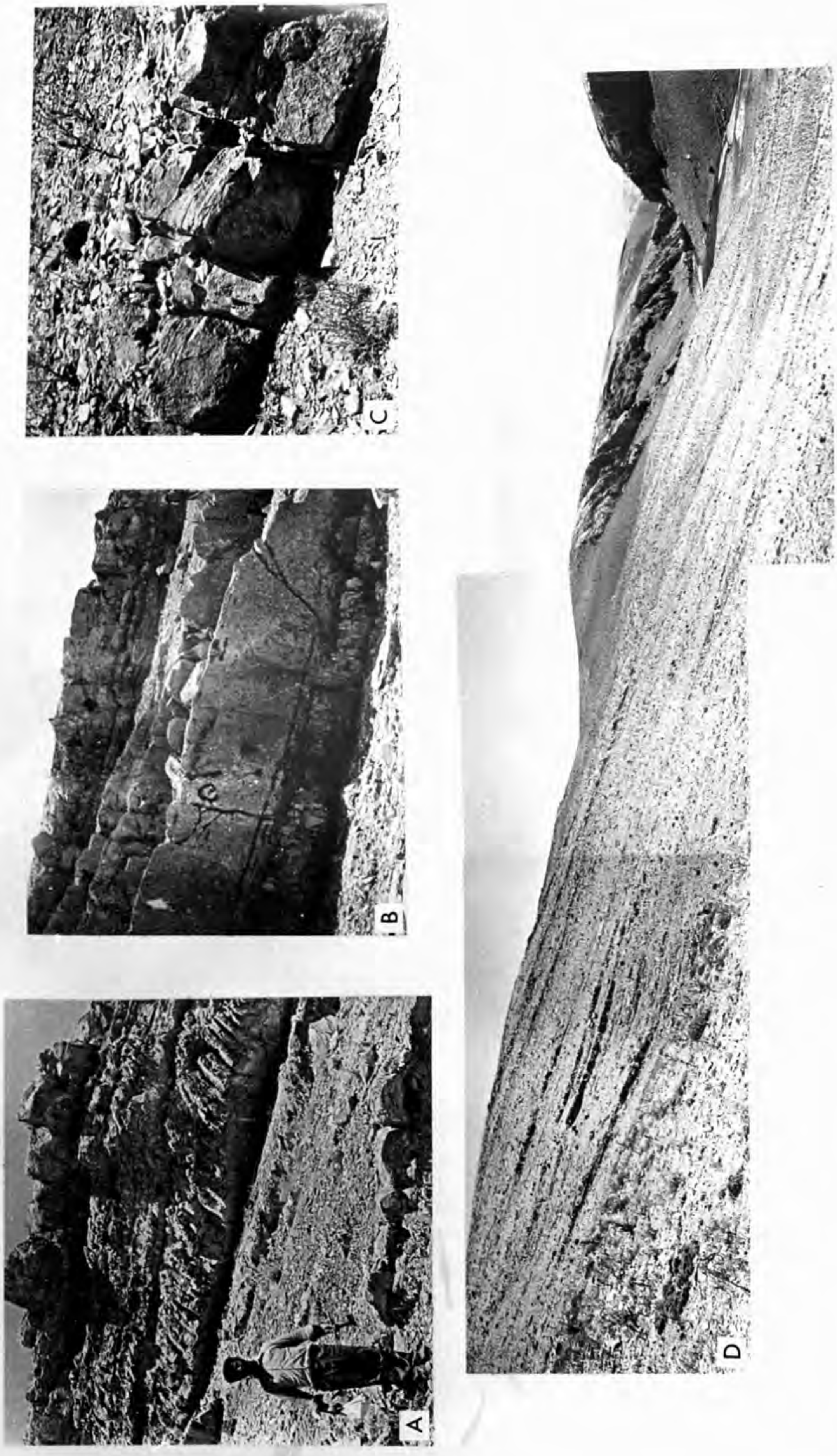
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A
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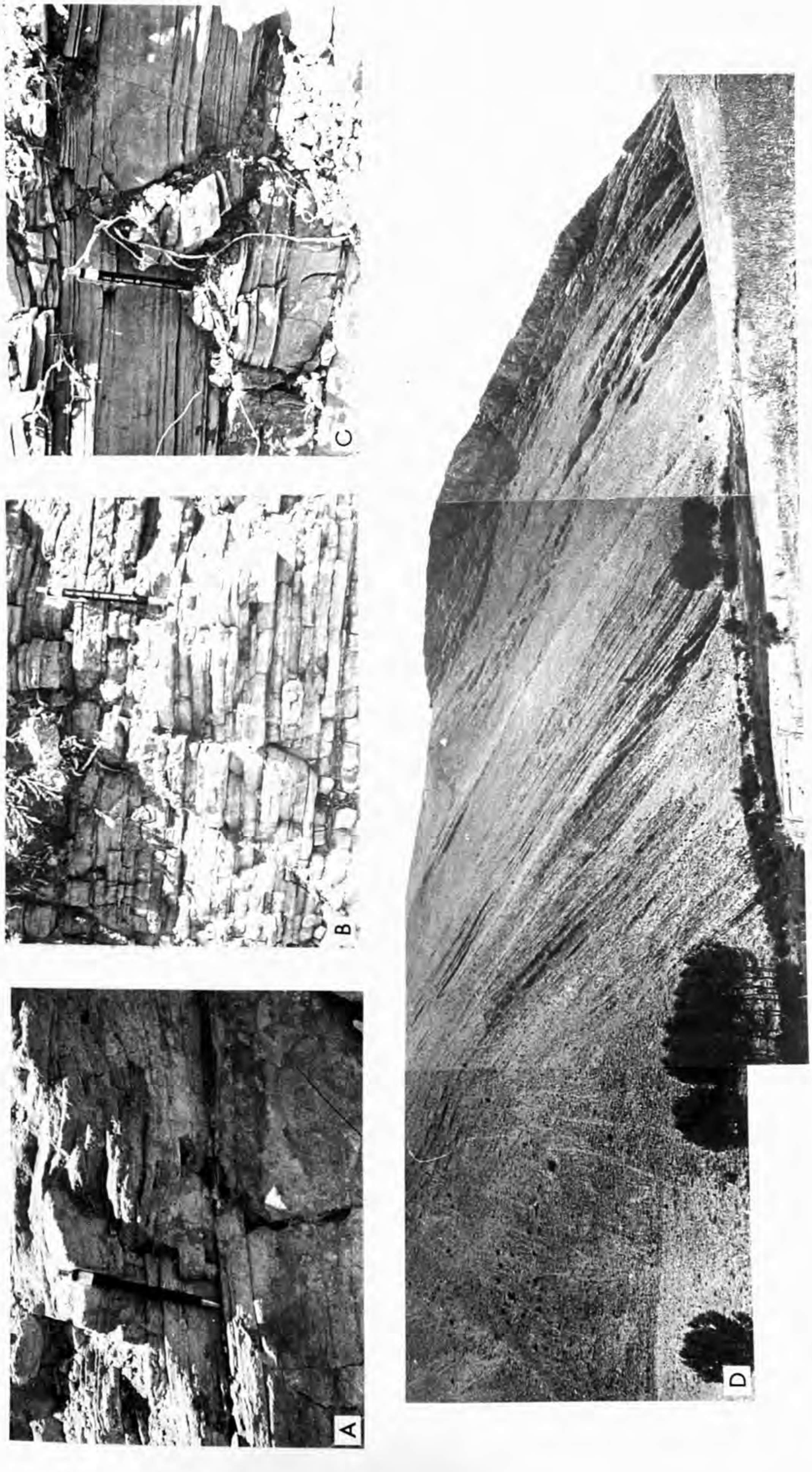
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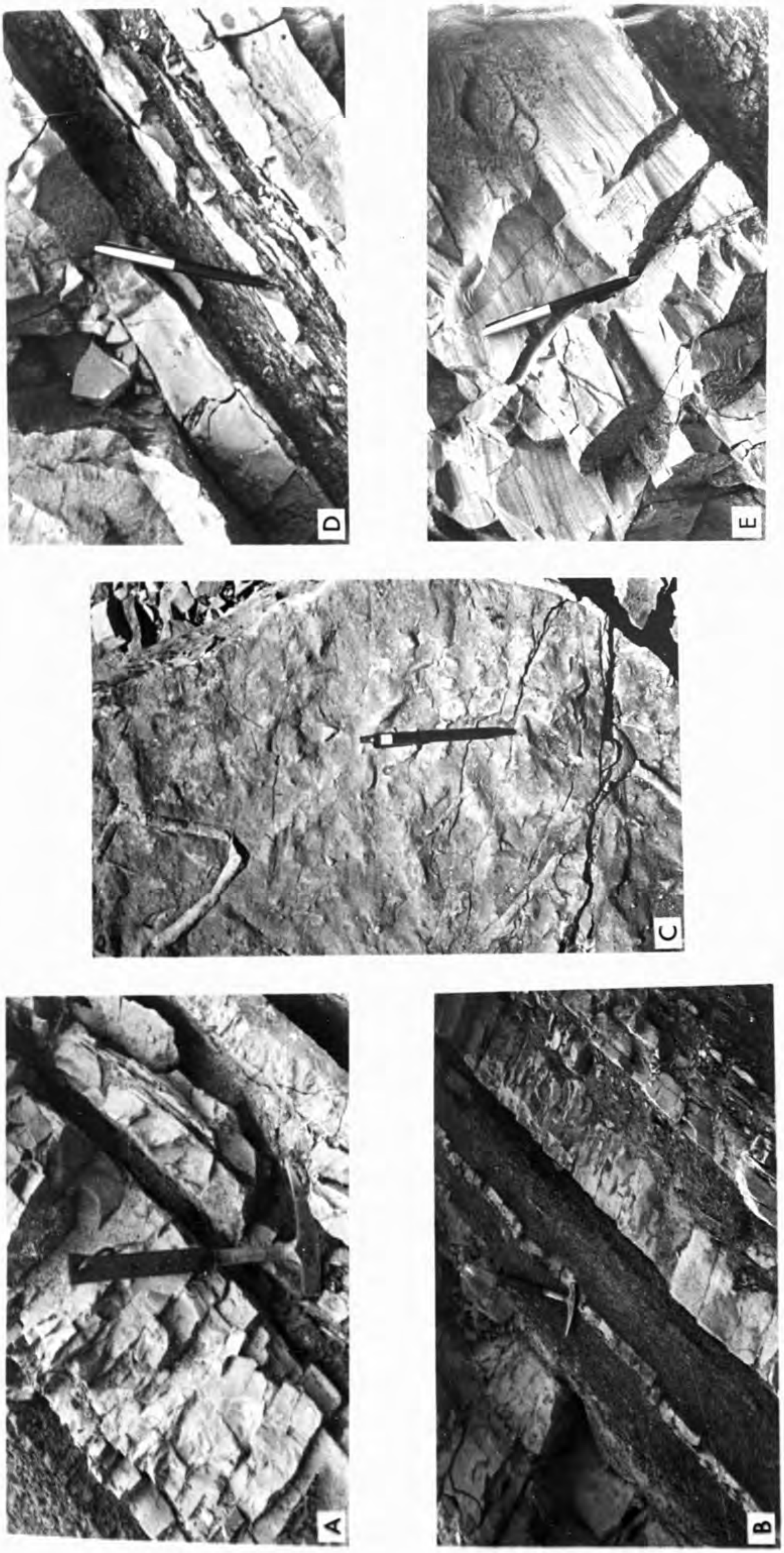
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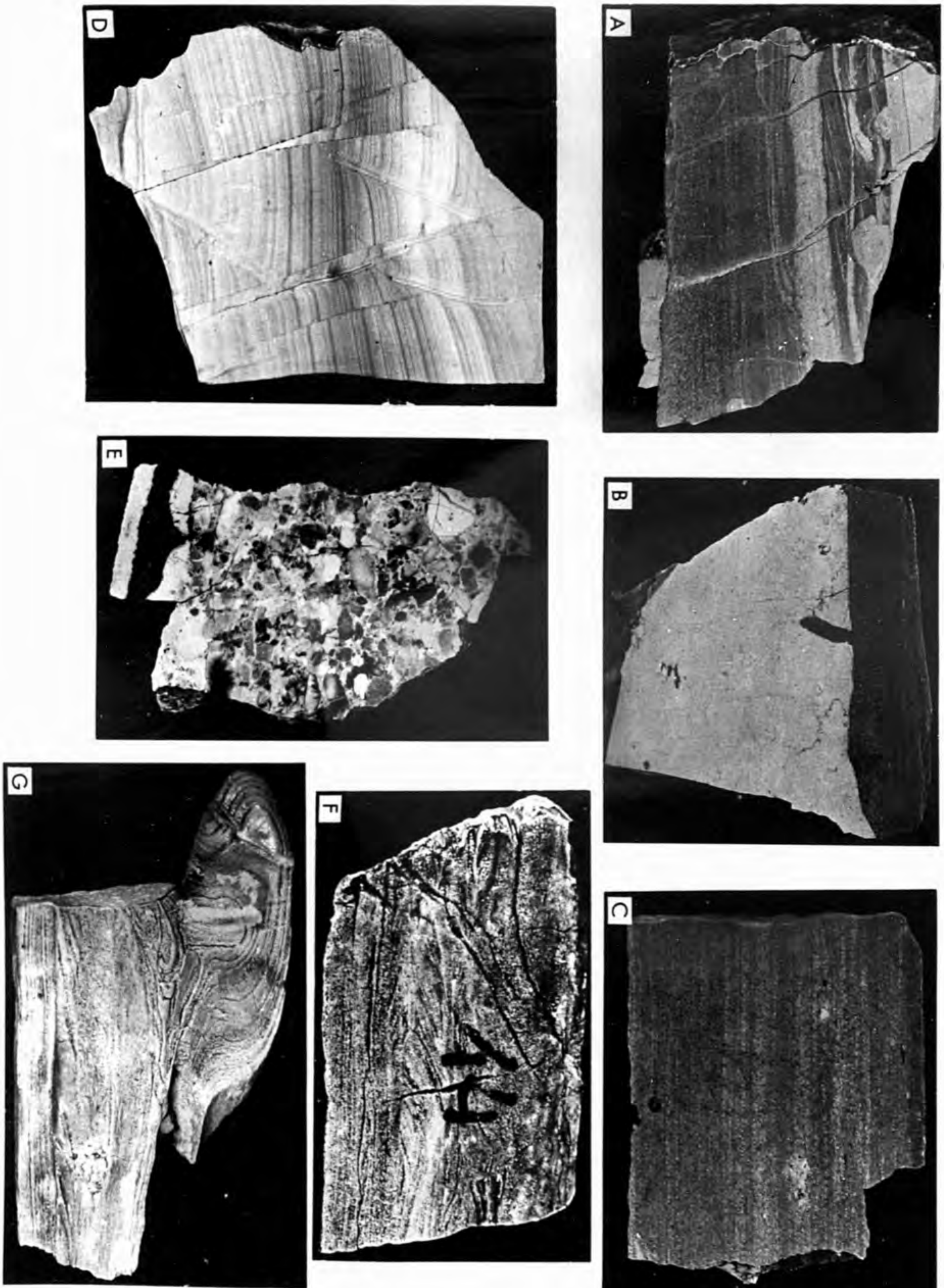
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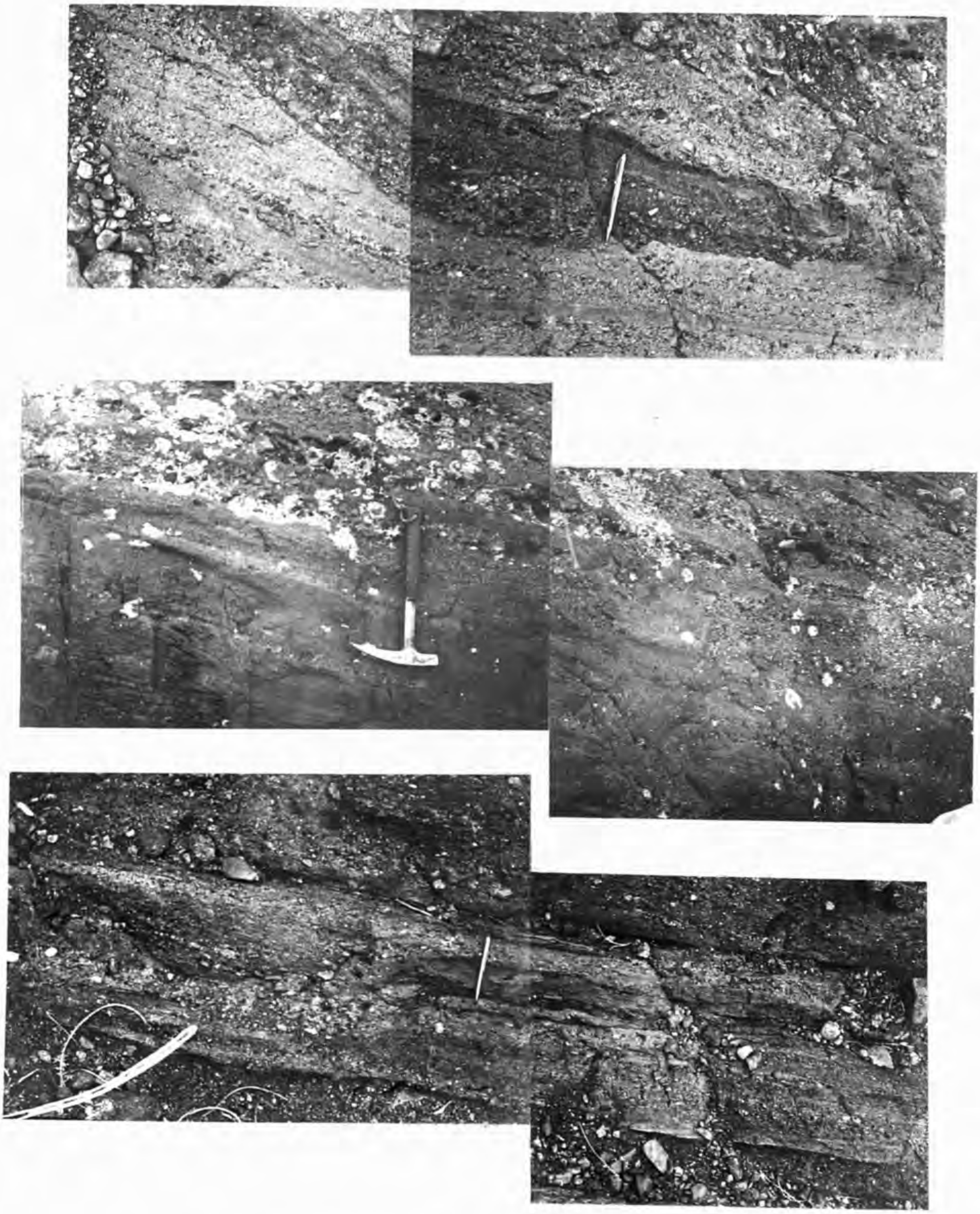
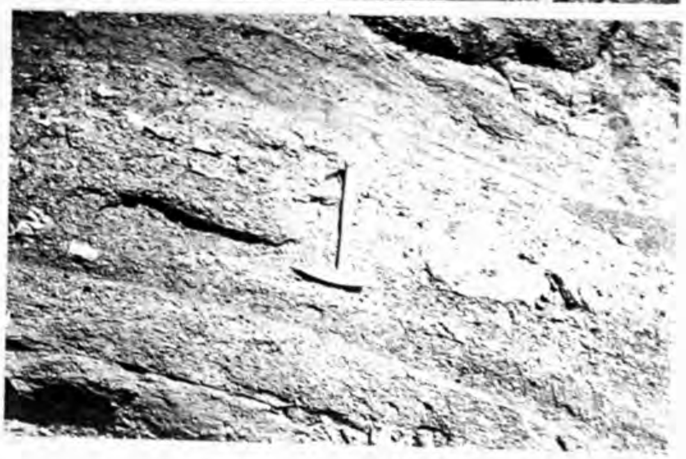
PLATE - 26

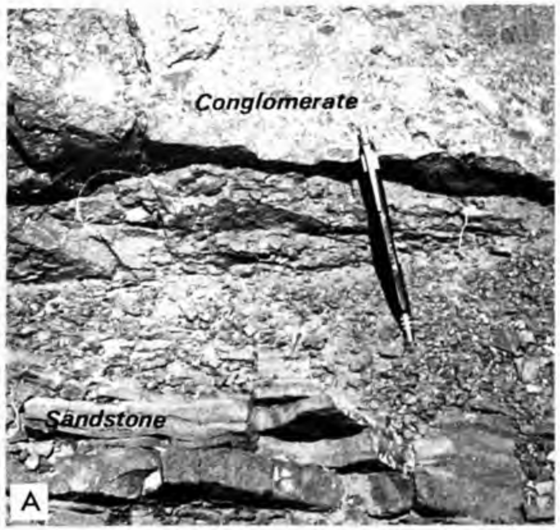


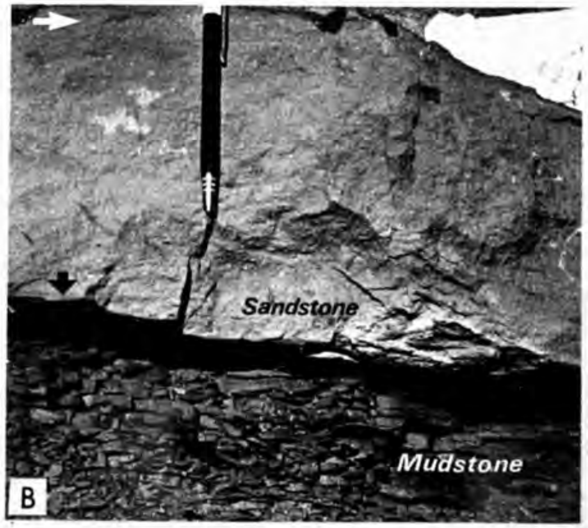
DELTAIC SEDIMENTS IN THE HAMALCAY REGION

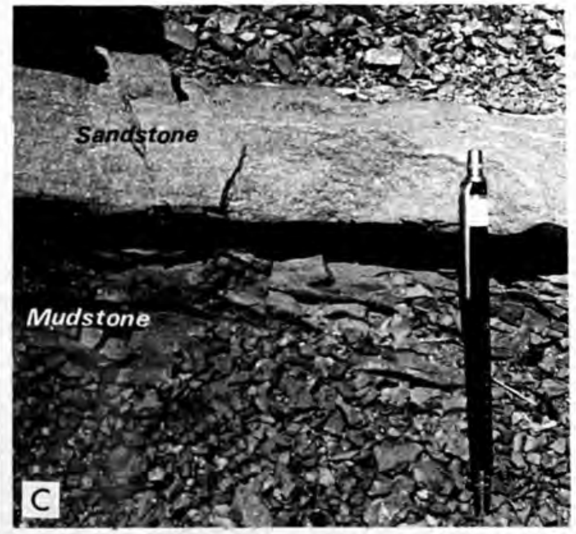
PLATE - 27



DELTAIC SEDIMENTS IN THE HAMALCAY REGION







FIELD APPEARANCES OF TURBIDITES-1

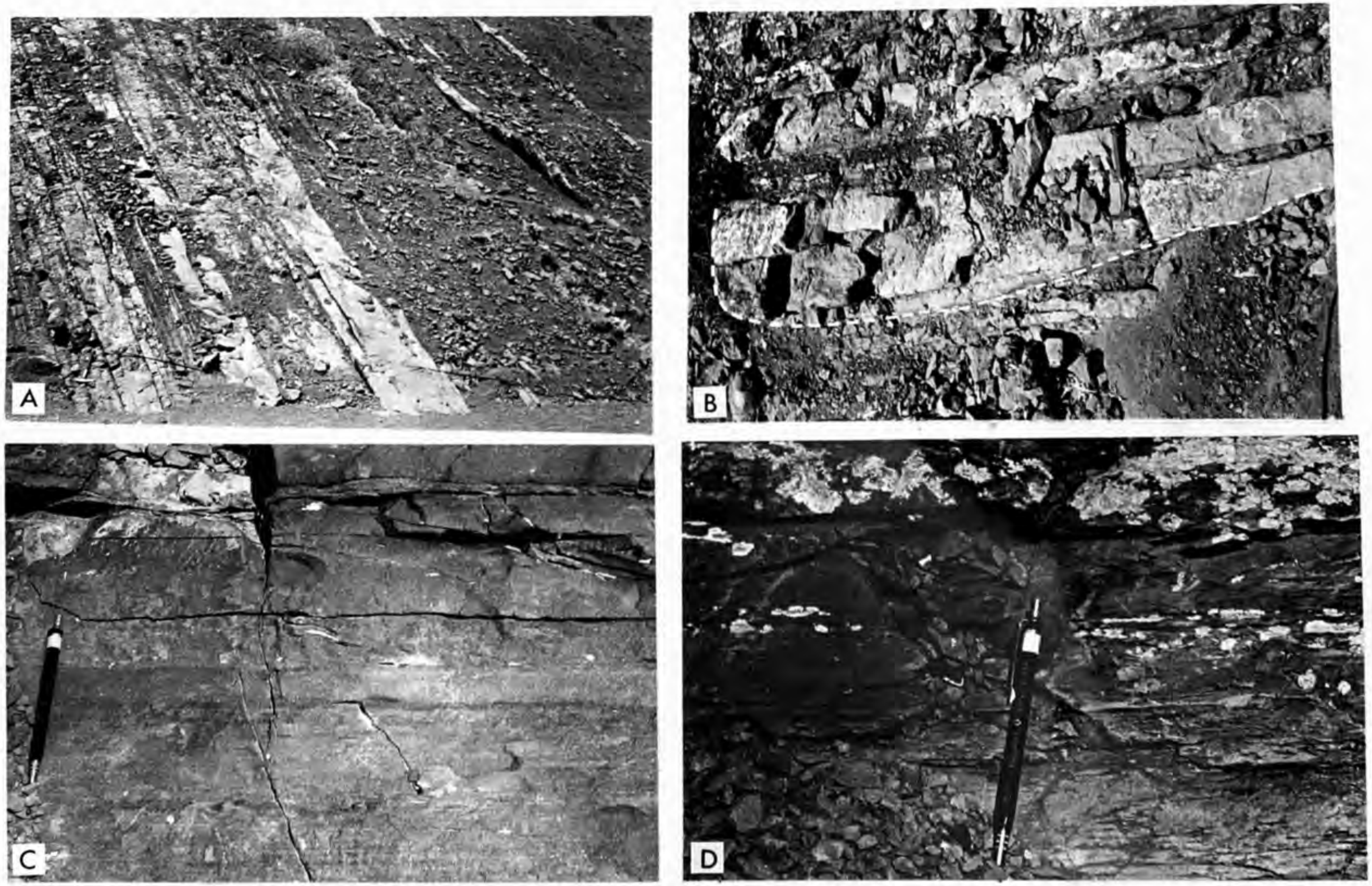
A. Channel facies. Thick conglomerate beds are overlain by coarse-grained sandstones. Beds are overturned. S 2.
B. Although it is not clear in the photo, sandstone bed is coarse-grained at the base (white arrow) and shows fining upwards, followed by a very thin siltstone (black arrow). Hemipelagic mudstone overlies the sandstone bed. Beds are overturned. S 5
C. Broadly channelized sandstone bed encircled in open fan mudstone. Sandstone bed shows upward fining. Beds are overturned. S 6
D. Thin- to thick-bedded sandstone/mudstone interbedding, Facies D of RICCI-LUCCHI (1975). S 7
E. Thin, bioturbated sandstone beds. S 7

FIELD APPEARANCES OF TURBIDITES - 2

A. Thin-bedded sandstone/mudstone interbeddings. Facies E of RICCI-LUCCHI(1975). S 8.

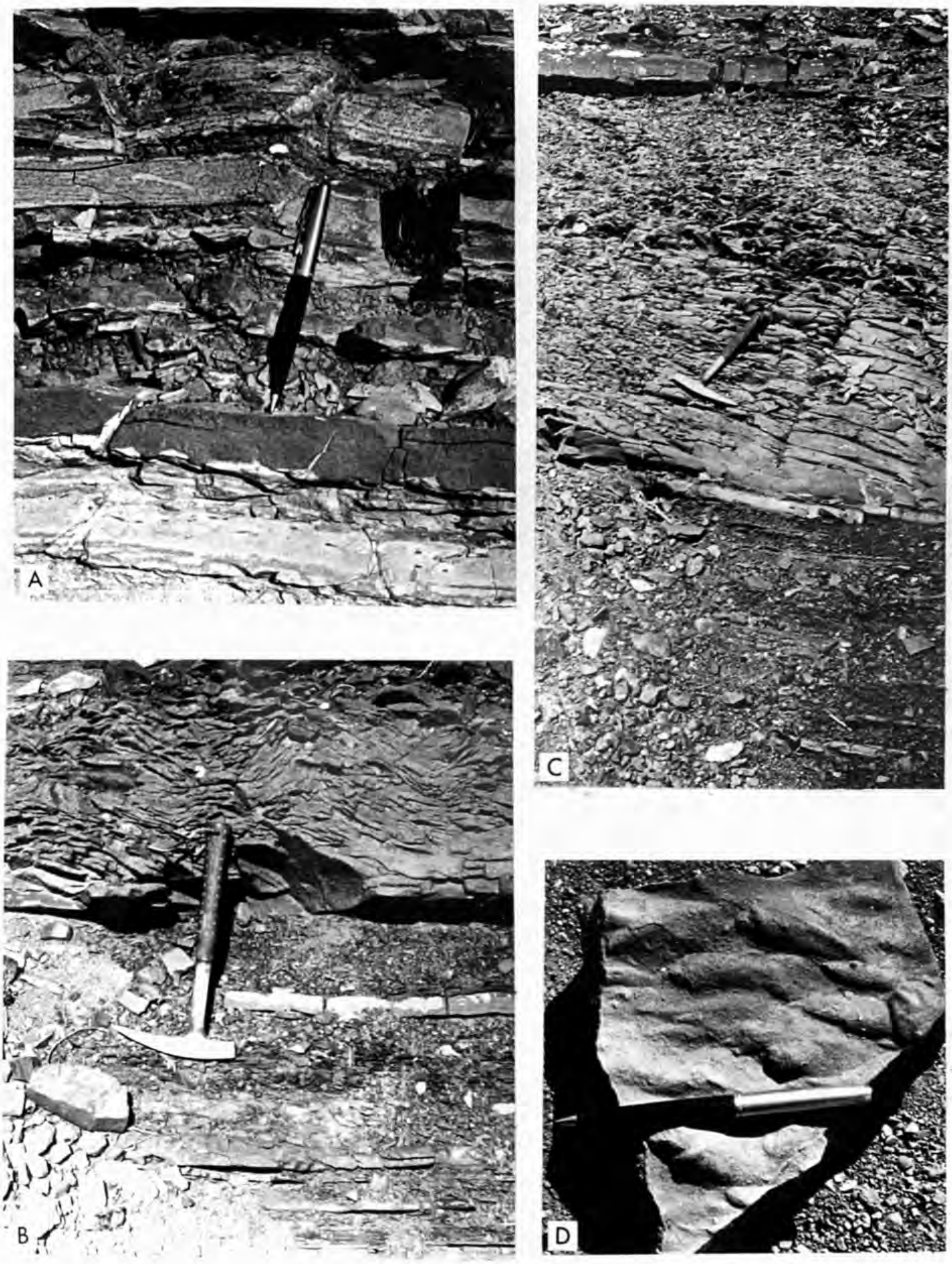
B. Channelized sandstone bed showing BOUMA (1962) A and B Divisions. See also Plate 33D and E. S 8

C. General view of section trend. Arrow shows the rudist bank.



FIELD APPEARANCES OF TURBIDITES-3

A. Alternation of sandstone and mudstone beds as thick as a few metres overlain by open-fan mudstones. S 9-10.
B. A channel cut into thin bedded deposits. S 10
C and D. Basin plain facies. Hemipelagic and turbiditic mudstone. S 14 and S 16. Also see Plate33 F.



FIELD APPEARANCES OF TURBIDITES-4

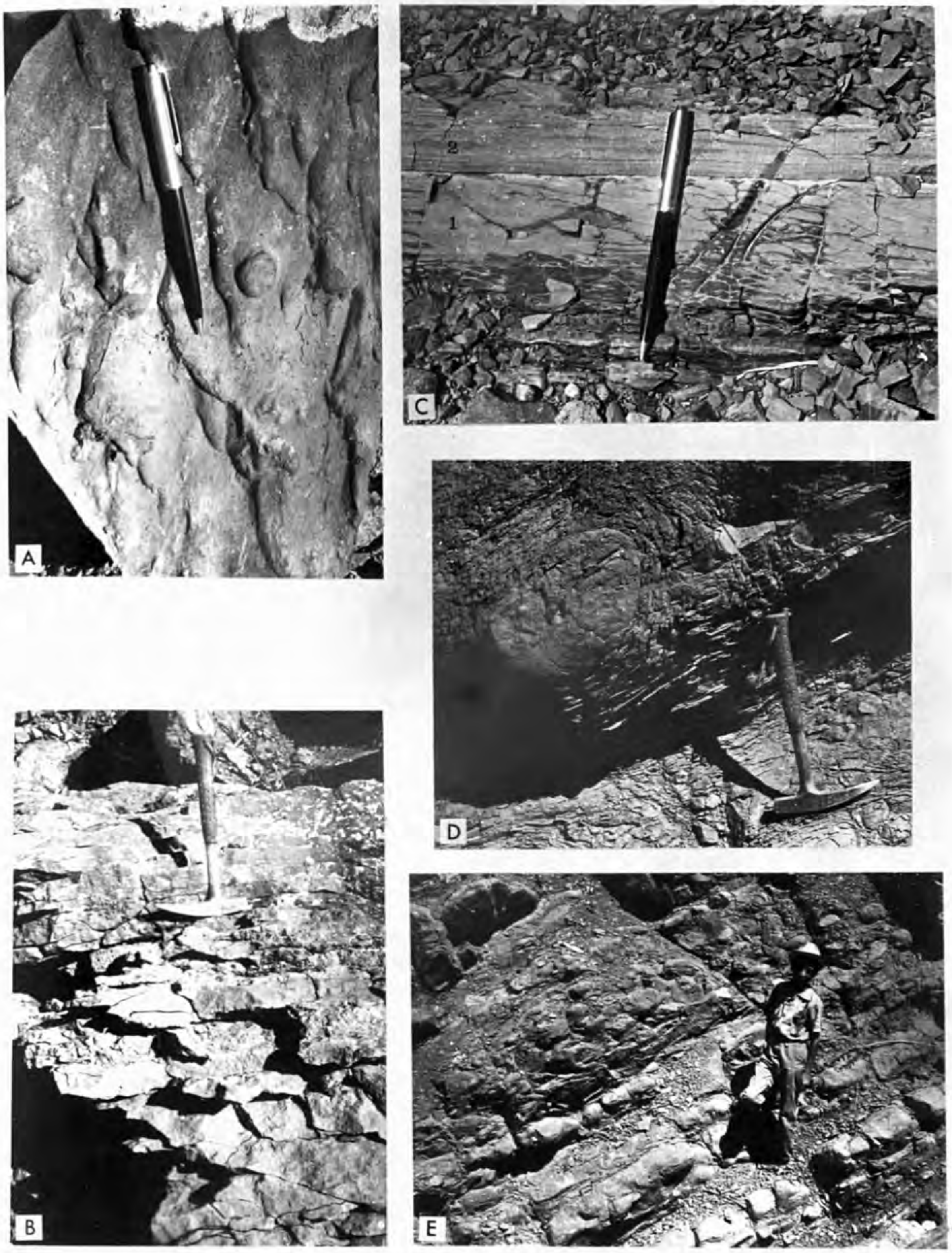
A. BOUMA (1962) T_{d-e} and T_{c-e} units showing thin, parallel and low angle cross laminae. The beds are interpreted as lobe-fringe facies. S 17

B. Convolute bedding in sandstone. S 18

C. Thin bedded turbidites are overlain by hemipelagic mudstones, followed by sandstone (Top of photo). S 19

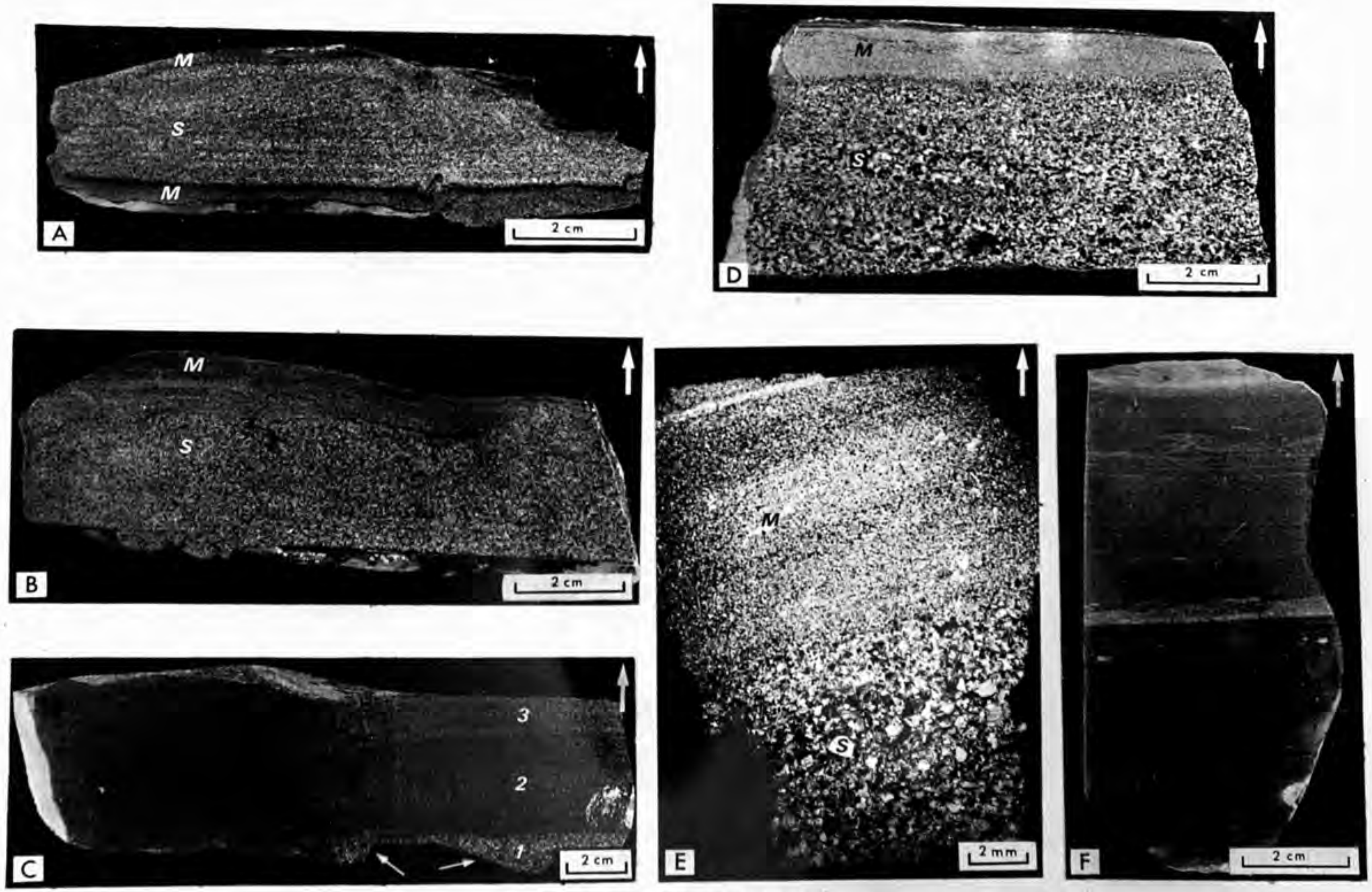
D. Underside of sandstone specimen showing flute-casts. S 19/1. See also Plate 33C.

PLATE - 32



FIELD APPEARANCES OF TURBIDITES-5

A. Sole marks at the base of a sandstone bed.
B. Amalgamation of thin sandstone beds. S 32



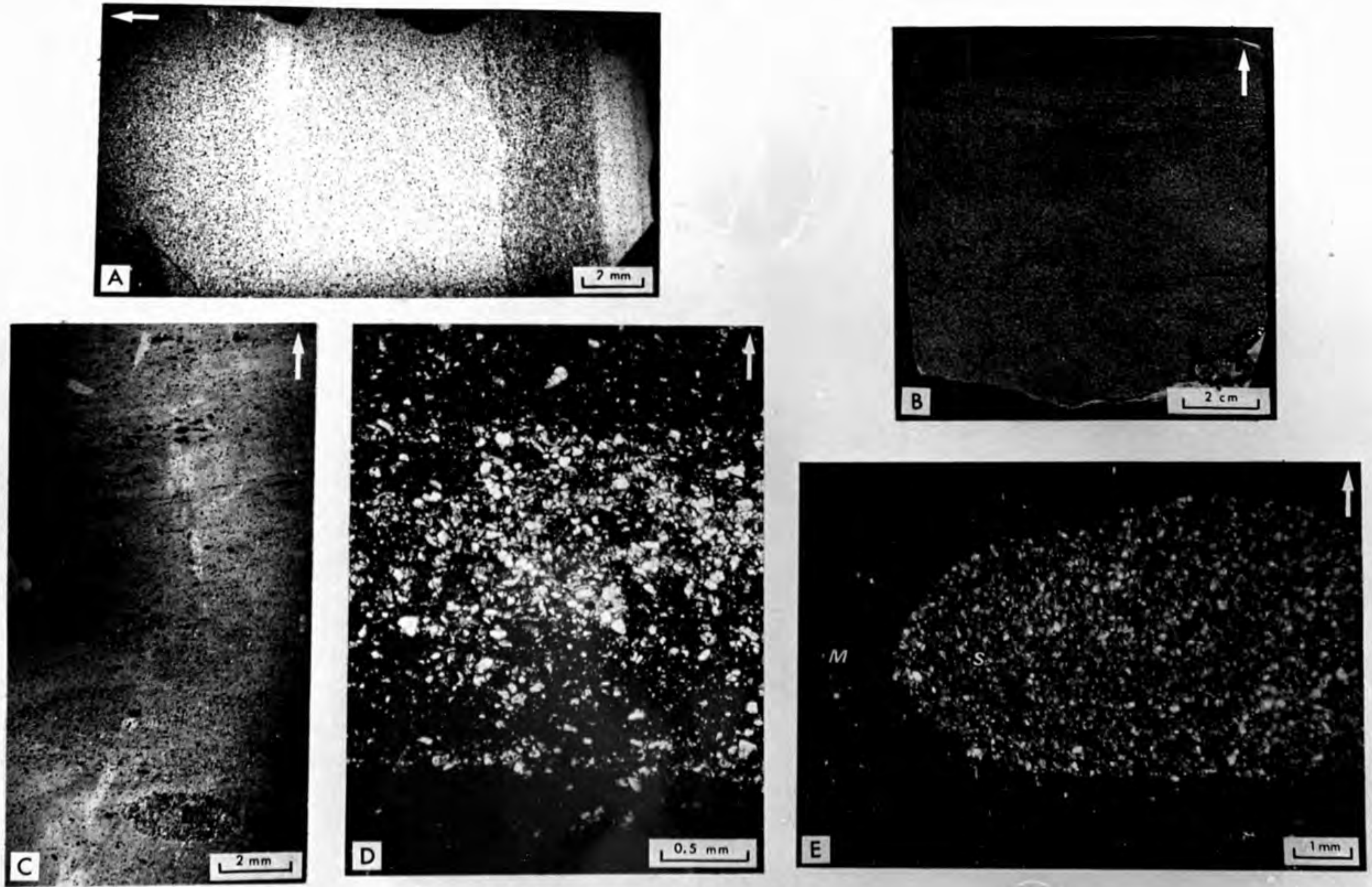
FEATURES OF TURBIDITES-1

A and B. Alternation of sandstone (S) and mudstone/siltstone (M). A. S 4 ; B. S 8'. Polished jand specimens.

C. Polished surface photograph of sample shown in Plate 31 D. Small arrows correspond to flute casts. Sample shows BOUMA's (1962) graded interval (1), lower interval of parallel lamination (2), and ripple lamination divisions (3). S 19/1

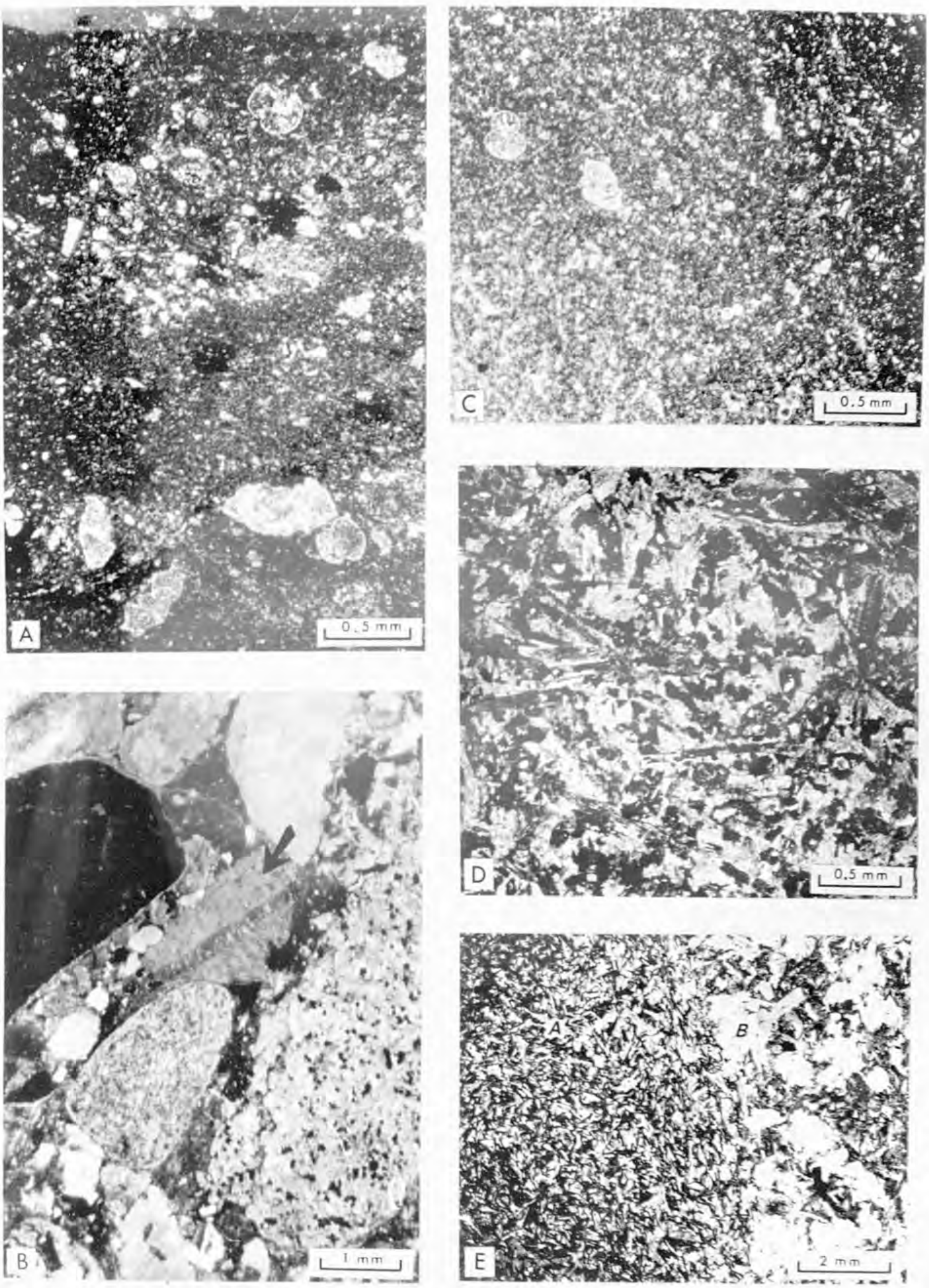
D and E. BOUMA (1962) divisions "a" and "b". A. Polished hand specimen ; B. Negative print from a thin section. Sample S 8.

F. Alternating hemipelagic (light coloured) and turbiditic (dark coloured) mudstones. S 16. Also see Plate 34A. Arrows show top.



FEATURES OF TURBIDITES-2

A. Negative print from a thin section of sample shown in Plate 33F. Dark part is hemipelagic interval. S 16
B. Bioturbated laminated sandstone. S 5
C. Hemipelagic mudstone. Negative print from a thin section. S 20/2
D. Microphotograph of a hemipelagic mudstone. Some pelagic foraminiferas are visible. S 21/2
E. Sandstone burrow in hemipelagic mudstone. S 17. Thin section.



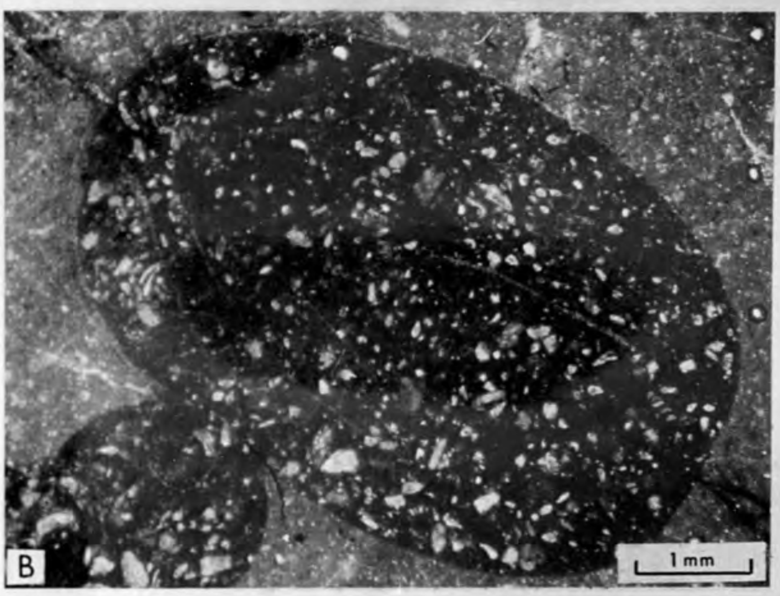
THIN SECTION PHOTOMICROGRAPHS OF SOME TURBIDITE SAMPLES

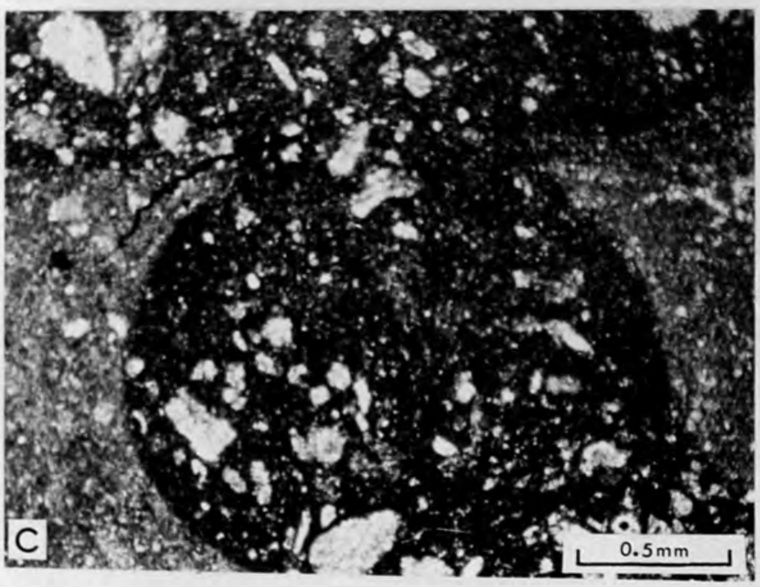
A. Hemipelagic mudstone. Pelagic forams (mostly Globigerina and Globotruncana) are abundant. S 24/1

B. General view of a sandstone thin section. Arrows show transported fossil fragment. All the other grains are intrusive and extrusive in origin. S 21

C. Same as A. S 20/2 D - Microdiorite (?) S 20/1

E. Gabbro/diorite (B) with trachitic (?) margin (A). Negative print from a thin section. S 25.



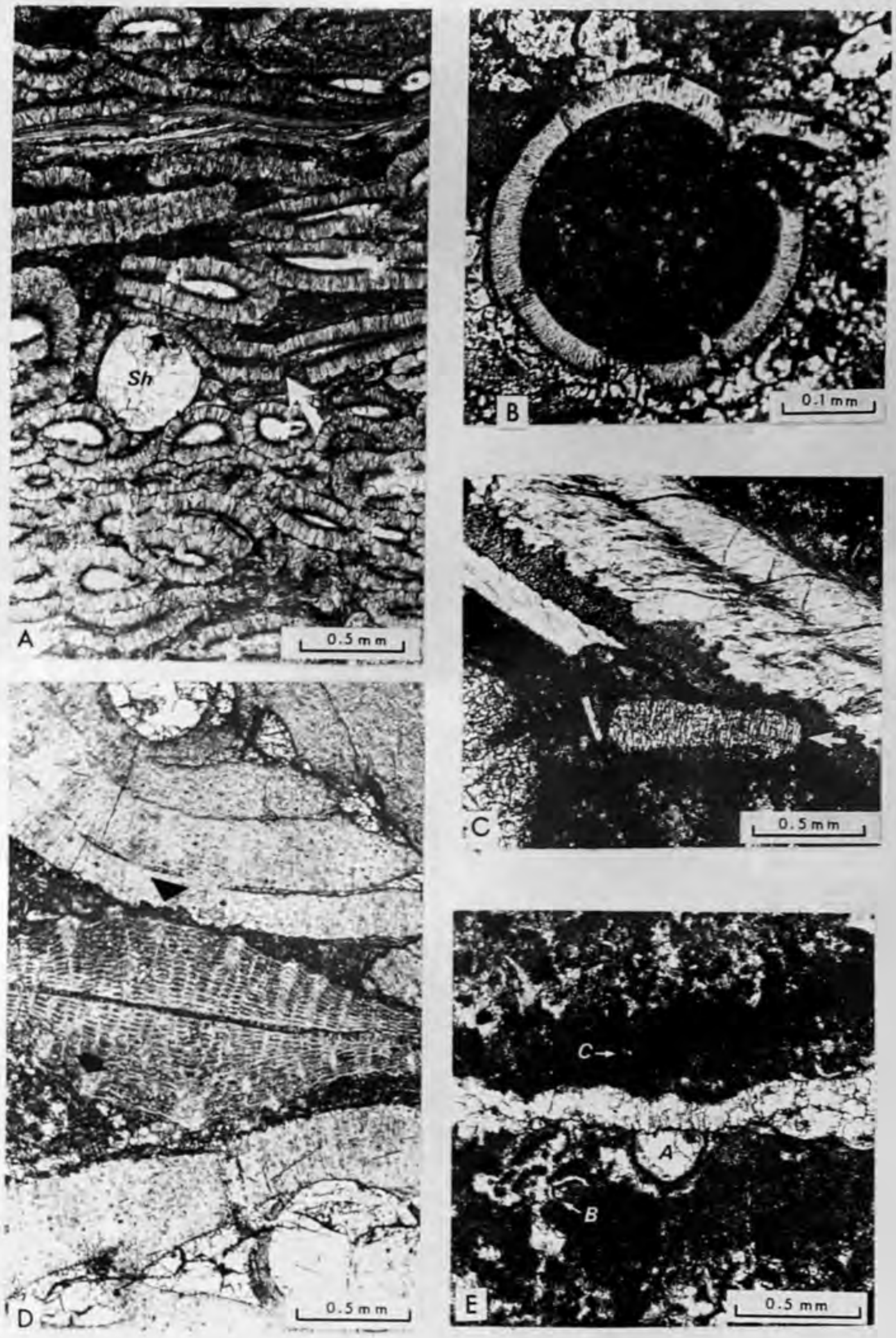


BIOLOGICAL DIAGENESIS

A. Polished hand specimen showing bioturbation. Sample No.8; Locatio: Mazıkıran (139)

B. Burrows are filled with different sediment syrrounding them. Peel. Sample No.74/40 Location: Arpaçukuru (K8)

C. Detail of B.



COMPACTION EFFECTS IN GURUN SEDIMENTS

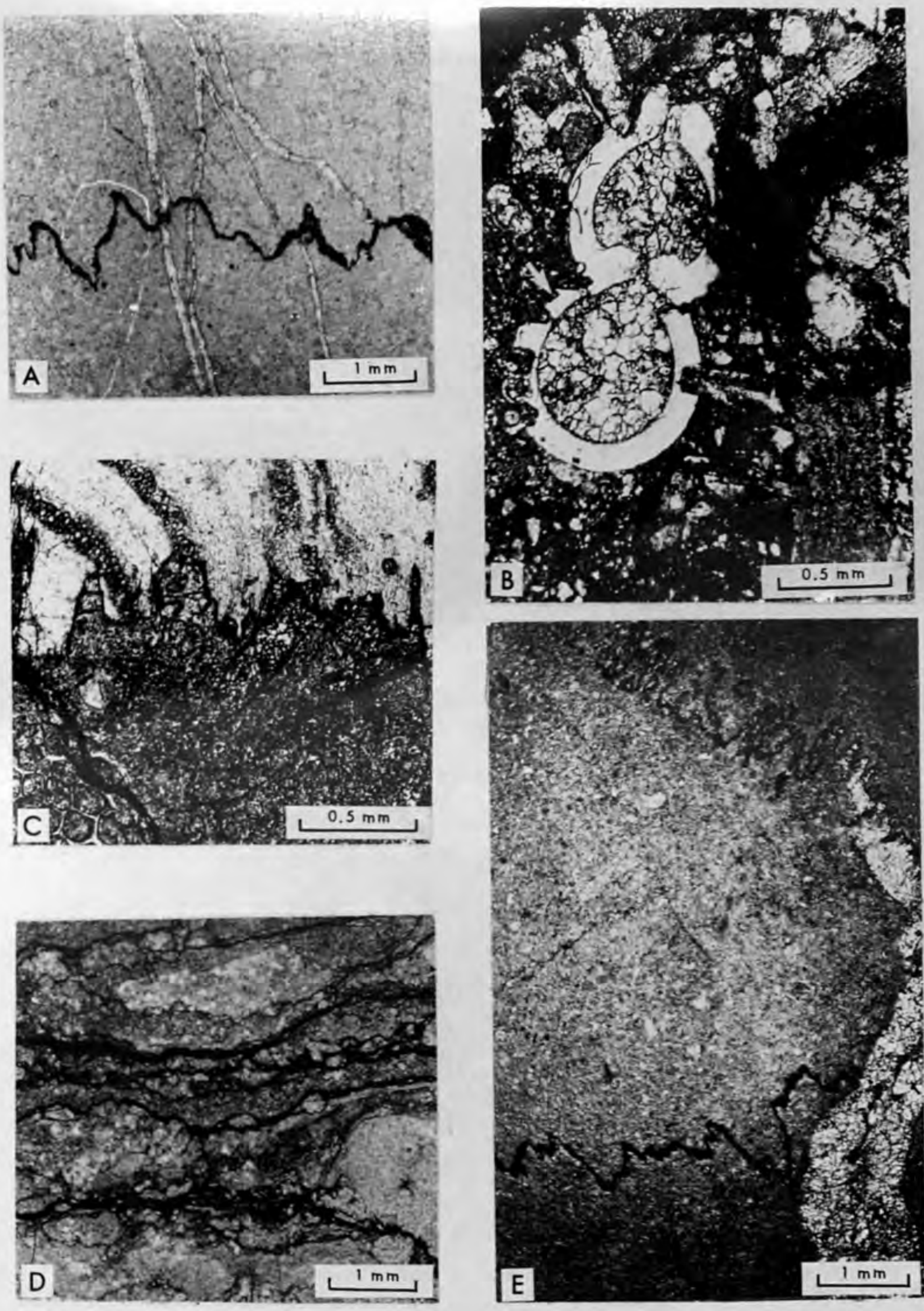
A. Bryzoa show strong evidence of compaction. The sparry calcite infill of the shell (Sh) is responsible from bending and breaking of the bryozoa fragment (white arrow). Note the pressure solution contact between bryozoa and sparry calcite infill (black arrow). Stained thin section. Sample No.1012.

B. Rapture of a shell fragment due to compaction. Peel. Sample No.911.

C. Penetration of two shell fragments. Peel. Sample No.1078.

D. Pressure-solution contact between two foraminifera (triangle). The lower part of the one in the middle has been deformed due to compaction (aarow). Peel. Sample No. Kö 1. Location: Kızılören (L46).

E. Compaction of pelletoidal limestone. The sparry calcite infill of the shell at (A) has been compressed against a calcite vein (Note the sutured contact). The micritic envelope of the shell at (B) has been collapsed due to compaction. The dark zone at (C) is the result of the squashing of the pellets which are barely visible in the microphotograph. Thin section. Sample No.232.



PRESSURE-SOLUTION AND STYLOLITES IN THE GÜRÜN SEDIMENTS

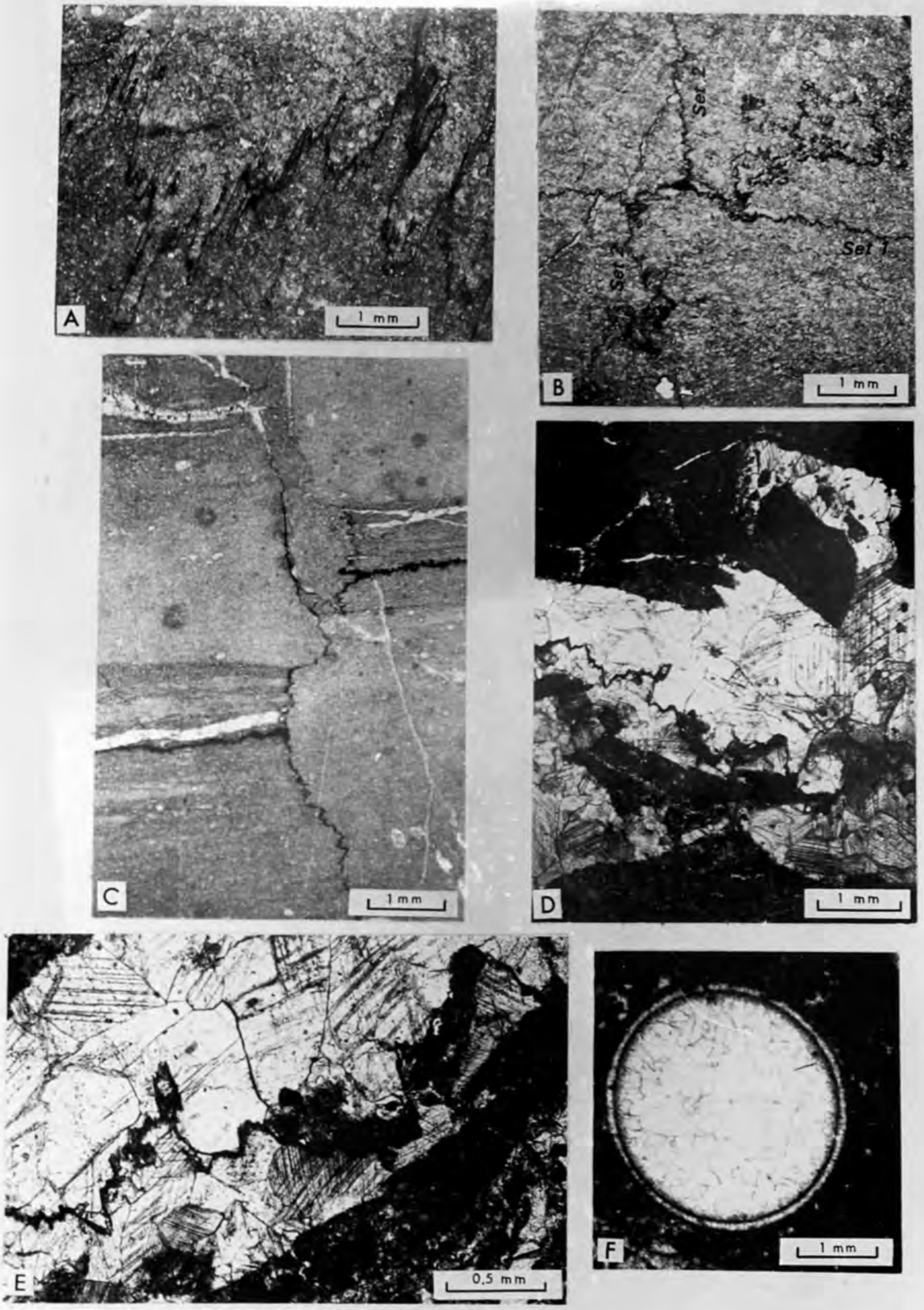
A. Calcite veins are cutting the stylolite seam which indicates that stylolization took place before the formation of the veins. Peel. Sample No.1024

B. Gastropod shell has been penetrated from several points. The pressure-solution contact on the left is clear (arrow). Peel. Sample No.908

C. Stylolitic contact between a coralline algae and bryozoa (?). Peel. Sample No.A4 Location: Arpaçukuru (K8).

D. Multiple stylolites seen in argillaceous limestones. Peel. Sample No.471-472

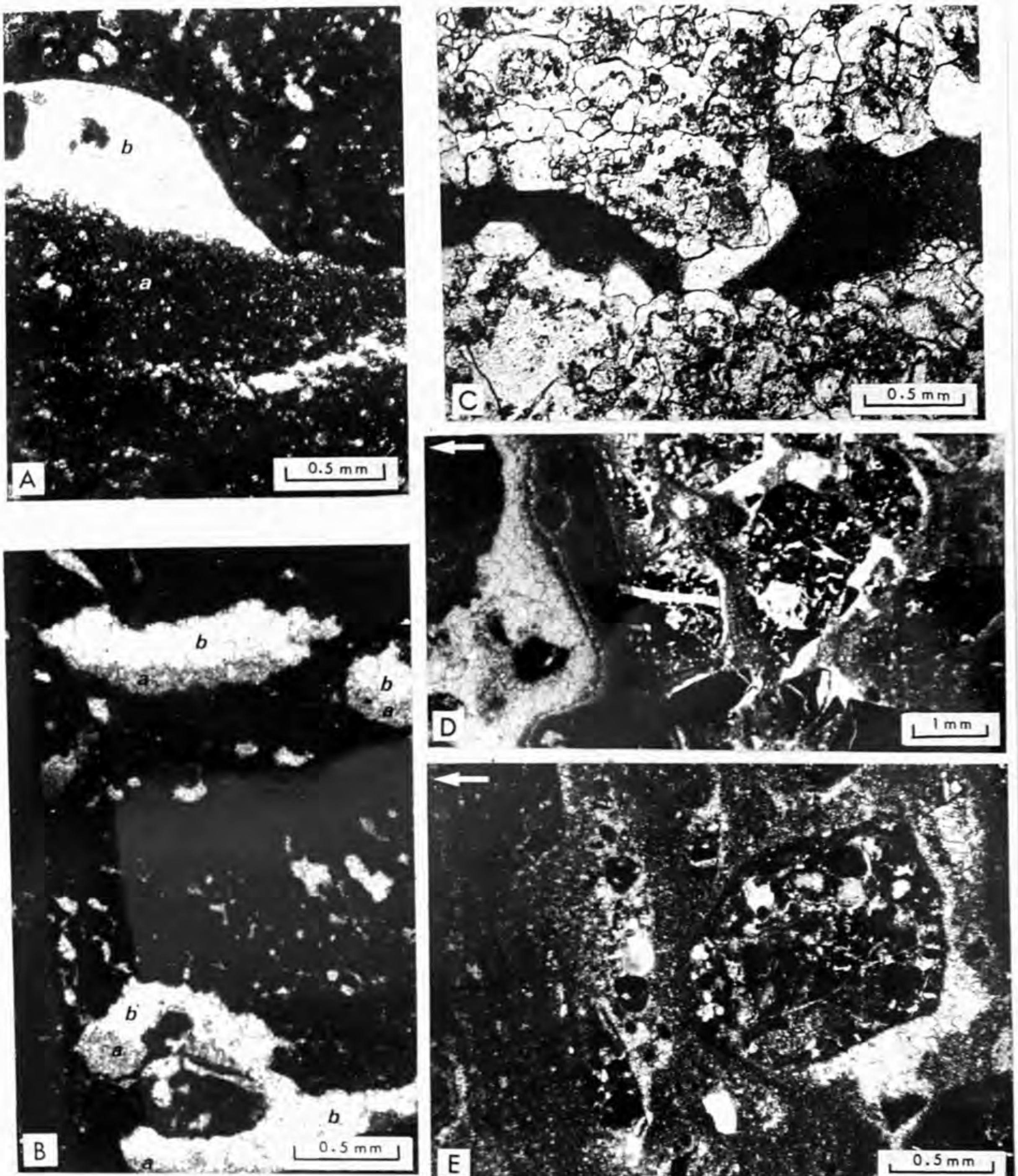
E. Two different types of stylolites seen in the same thin section. Sample No.224



FEATURES OF STYLOLITES

A. a stylolite with high amplitude. Peel. Sample No.1124
B. Two sets of stylolites perpendicular to each other. The vertical one (Set 1) is older than the horizontal one (Set 2). Peel. Sample No.
C. Two sets of stylolites perpendicular to each other. The vertical one is younger than the horizontal one. Peel. Sample No.1/20. Location: Mazikiran.
D. A stylolite cutting a calcite vein. Stained thin section. Sample No.1163-1164
E. Detail of a part of D.
F. Gastropod (?) shell shows the lack of compaction. Peel. Sample No.74/6 Location: Arpacukuru.

PLATE -41



VADOSE ZONE CEMENTATION

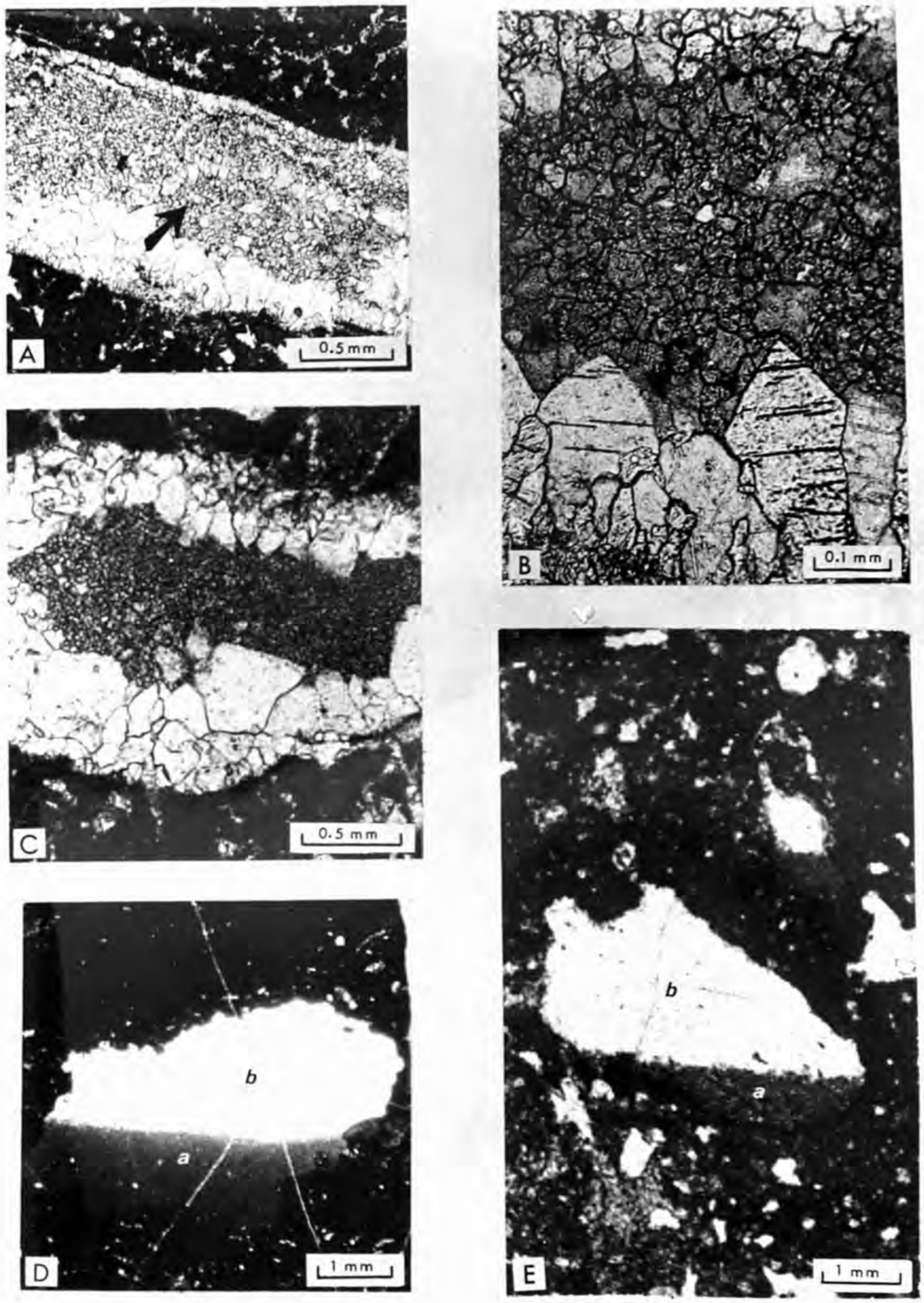
A, and B. Stromatactis filled by vadose silt (a) and random sparry calcite mosaic (b). Thin section. Sample No.: A- 1203/2; B- 760B.

C. Quartz silt. Replacement origin is indicated by relics of undigested calcite in the larger quartz crystals (See DUNHAM, 1969a; p.156, Fig.15F). Sample No.35

D. Gravitational cement (MÜLLER, 1971) has developed at the lower surfaces of grains facing along with the direction of gravity vector and in the vicinity of grain contacts. Arrow shows the top. Thin section. Sample No.414.

E. Detail of D.

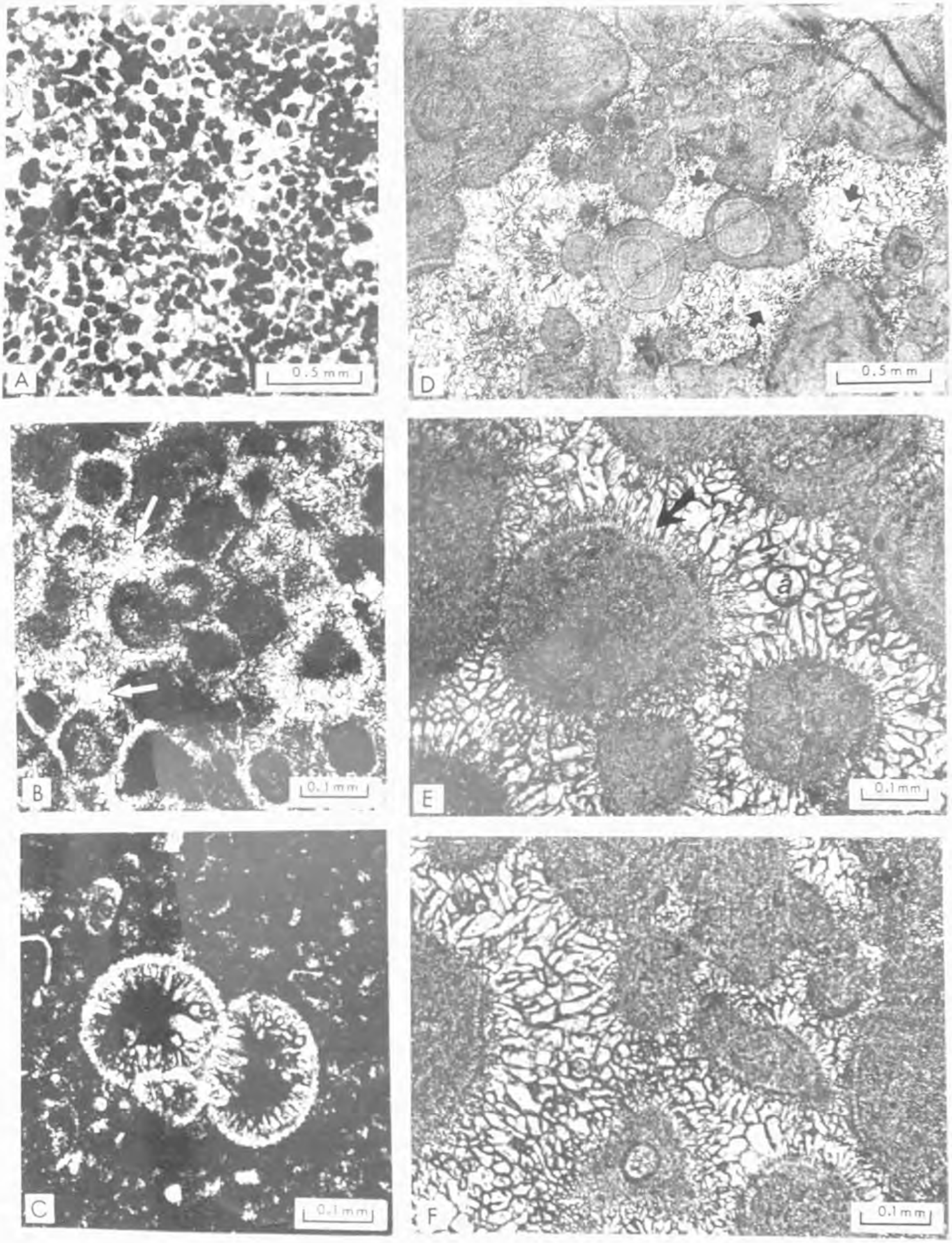
PLATE - 42



CHARACTERISTICS OF VADOSE ZONE

A, B, and C. Vadose silt completely fills the voids and is easily distinguished from original sediment because of the underlying calcite cement is clearly visible. A. Sample No.S1/13; Location: Mazıkıran(139). B. Detail of A.; C. Sample No. 1/16; Locality: Kaz Tepe (D35). All stained thin sections.

D and E. Vadose silt (a) underlies the stromatactis opennings and large calcite crystals (b) fill the remaining pore space. D. Thin section. Sample No.1/28 Locality: Kuz (N66).; E. Thin section. Sample No.74/28 Locality: Kırmızıdere.



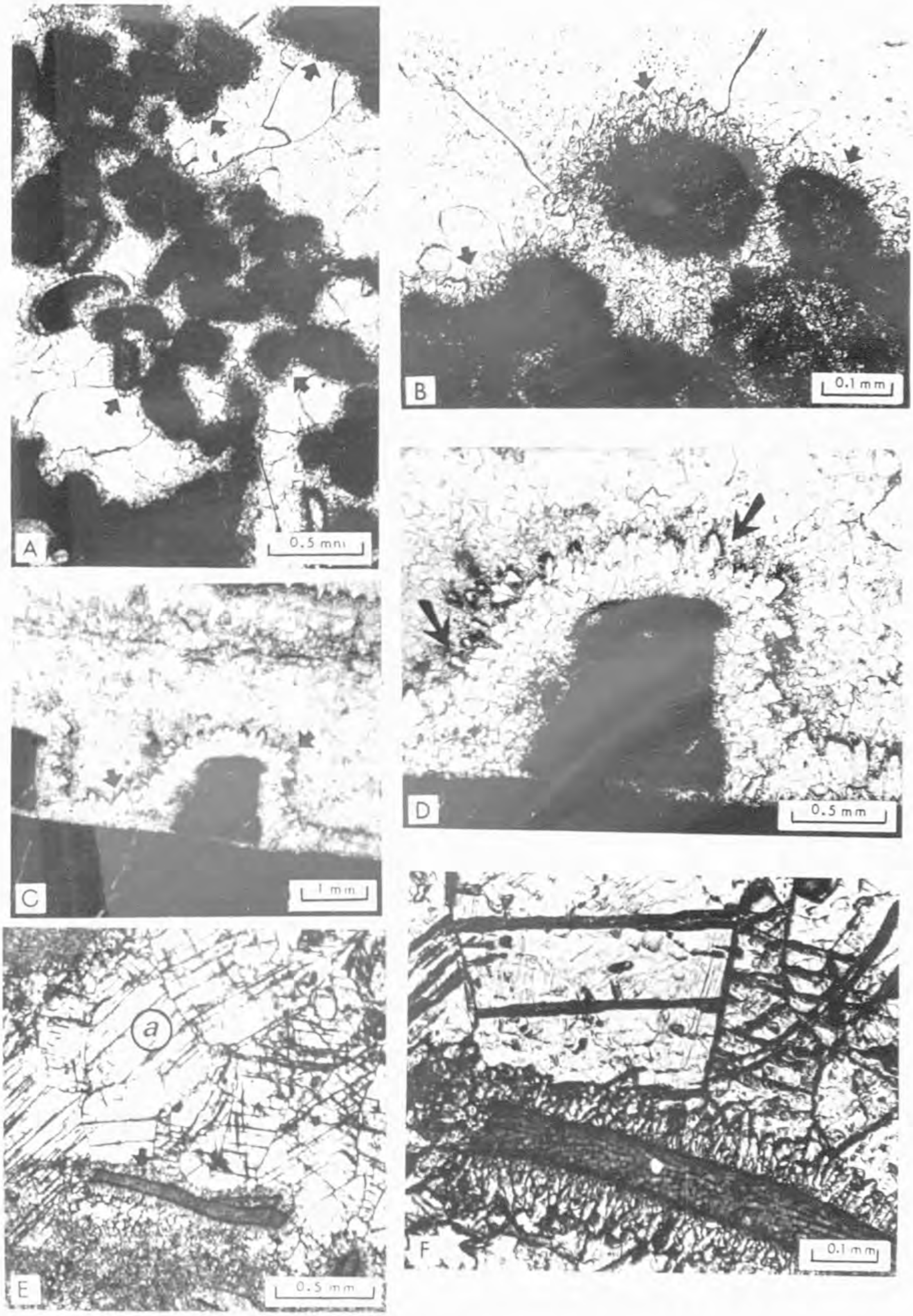
CEMENT FORMING CRUSTS ON GRAINS-1

A and B. Bladed calcite crystals form crusts on pelletoids, Sparry calcite fills the remaining pore space (arrows). A. Sample No.136; B. Detail of A.

C. Fibrous calcite forms a crust in a pelagic foraminifera chamber. Stained thin section. Sample No.1063.

D, E, and F. Microphotographs of an oolitic grainstone. Stained peel. Sample No.637 D. Fibrous calcite fringes the grains (small arrows). The remaining pore space has been filled by sparry calcite cement (larger arrows). E and F are details of D. Arrow shows the fibrous calcite cement. (a) is pore filling cement.

PLATE -44

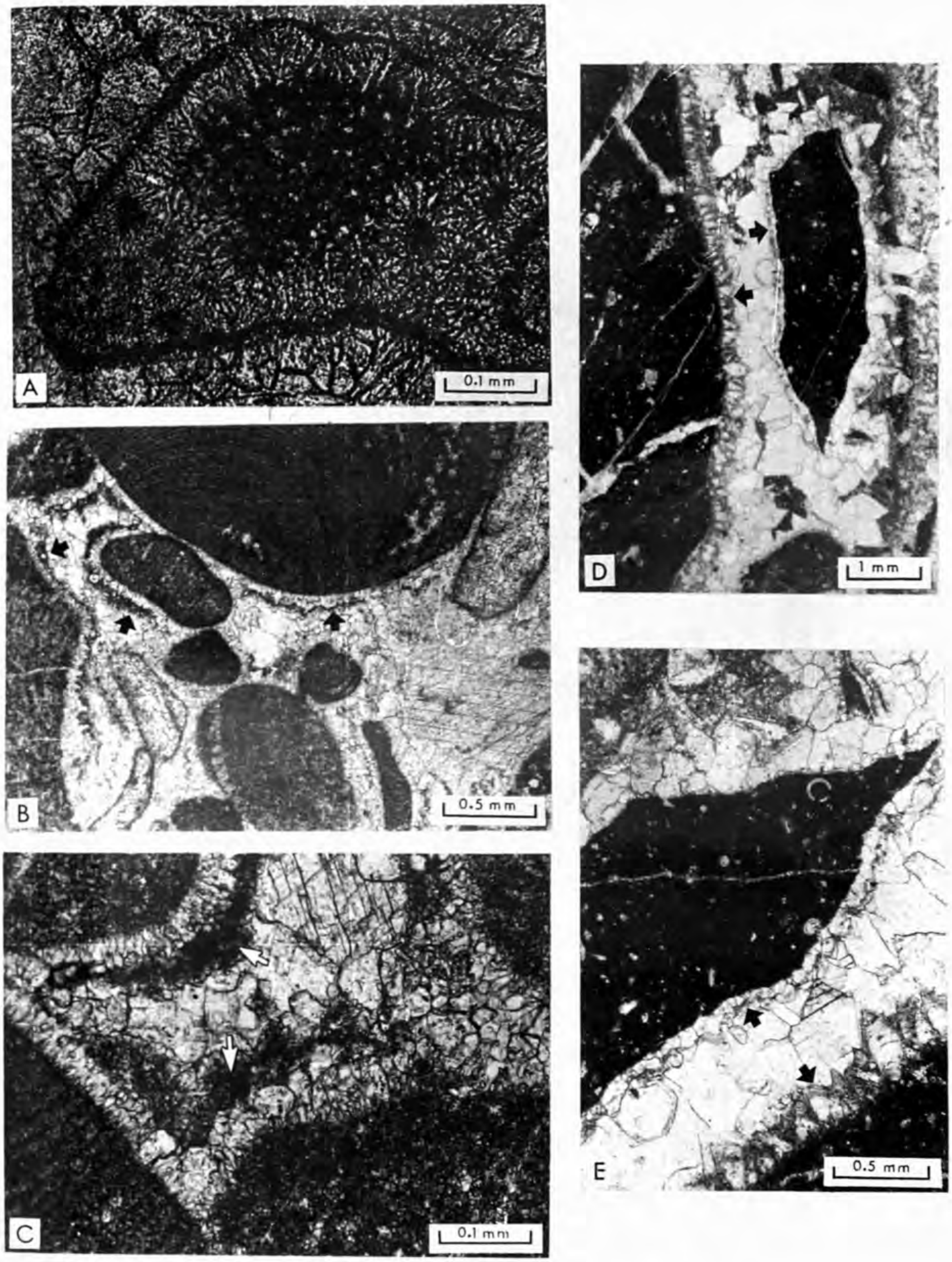


CEMENT FORMING CRUSTS ON GRAINS-2

A and B. Equant calcite crystals form a fringing crust around pelletoids (arrows). Sparry calcite occupies all the available pore space. Thin section. Sample No. S3/32; Location: Kaz Tepe (D35). B- Detail of A.

C and D. Micritic cement (arrows) has been developed on equant calcite crystals forming crust on brecciated lime mudstone, Sample No. S1/30 Thin Section. Locatio: Mazıkıran (139). D- Detail of C.

PLATE -45



CEMENT TYPES IN THE GÜRÜN SEDIMENTS

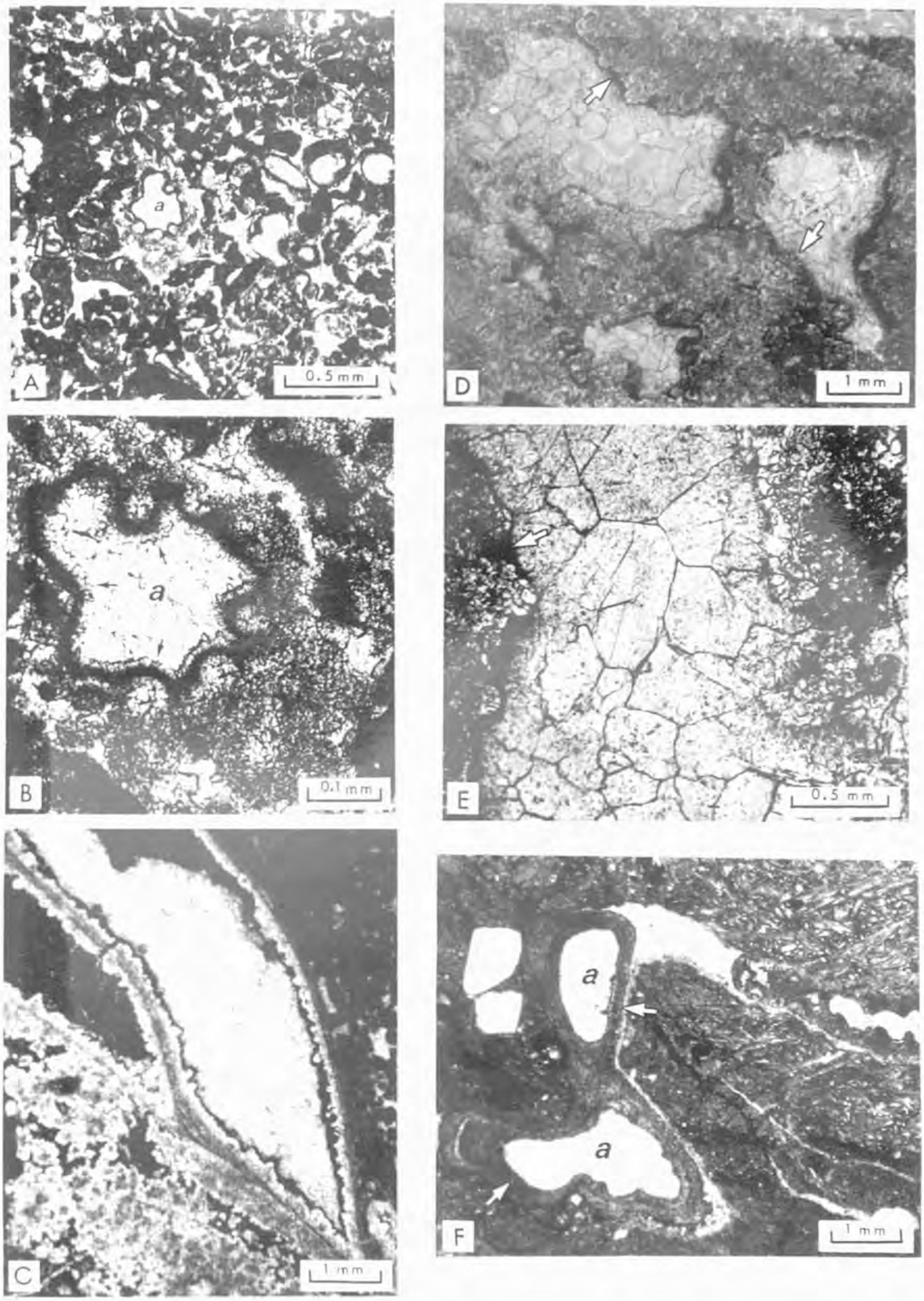
A. Fibrous cement. Peel. Sample No.1/28. Locality: Kuz (N66)

B. Irregular micritic cement (arrows) has been developed on equant calcite crystals which form crust on skeletal grains. Some of the former aragonitic skeletal particles are represented by micritic envelopes. Stained thin section. Sample No.1062C

C. Detail of B.

D. Micritic cement form a rim on first formed calcite crystals in a wackestone, Stained thin section. Sample No.308B.

E. Detail of D.



MICRITIC CEMENT

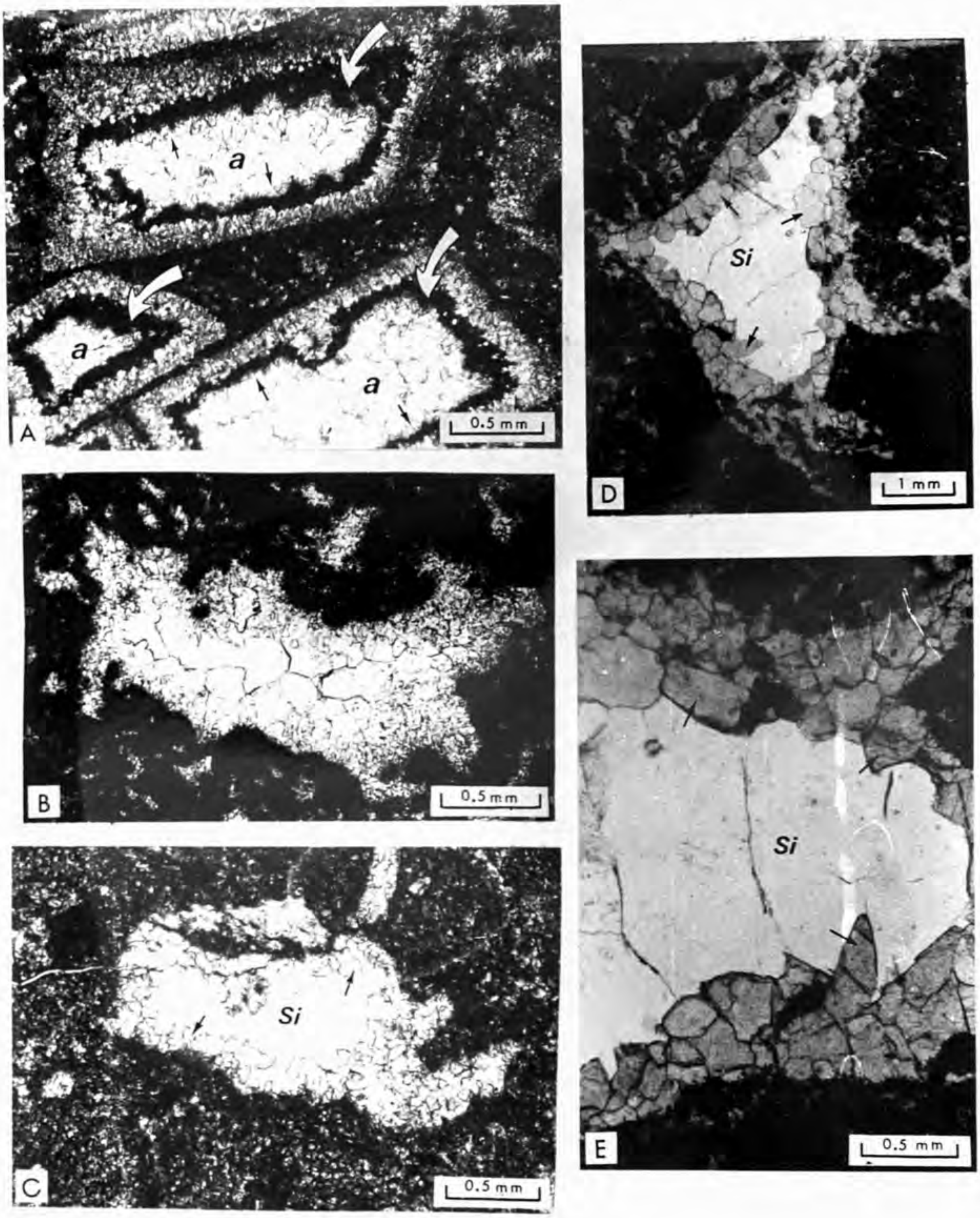
A. Micritic cement forms a crust around pore spaces followed by a crust of equant calcite crystals. Large calcite crystals (a) fill the remaining pore space, Stained thin section. Sample No.1/154

B. Detail of A. Arrows show equant calcite crust.

C, D, and E. Micritic cement lining the pores. Random calcite mosaic fill the rest of the pore spaces. All staned thin sections. C- Sample No.74/40; Location: Arpaçukuru (K8); D- Sample No.74/33; Location: Keşlik Dere; E- Detail of D.

F. Micritic cement forms a thick crust (arrows) followed by equant calcite cement (a). Sample No.620B

PLATE -47

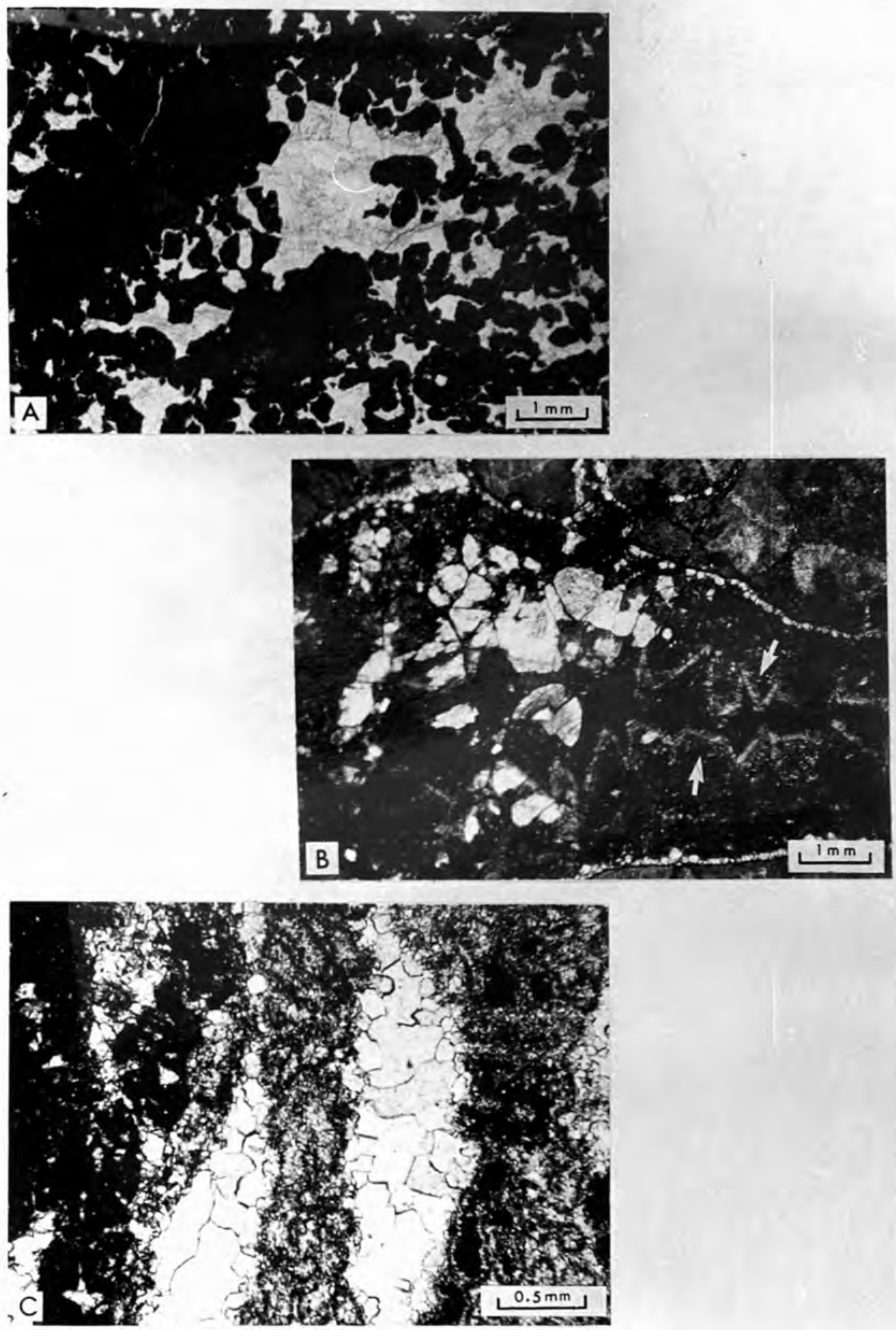


CEMENT TYPES

A. Micritic cement (white arrows) lines the interparticle pore space and is followed by bladed calcite crystals (small arrows). Random calcite crystals fill the whole pore space. Stained thin section. Sample No.1062D.

B. Intergranular pore space has been occupied by a calcite crust and random calcite mosaic. Thin section. Sample No.2/12. South of Yılanocağı.

C, D, and E. Silica cement. Equant calcite crystals line the pore spaces (arrows). The remaining pore spaces are filled by silica cement (Si). C- Thin section; Sample No. 853B. D- Thin section. Sample No.S3/6 Location: Kaz Tepe (D35); E- Detail of D.

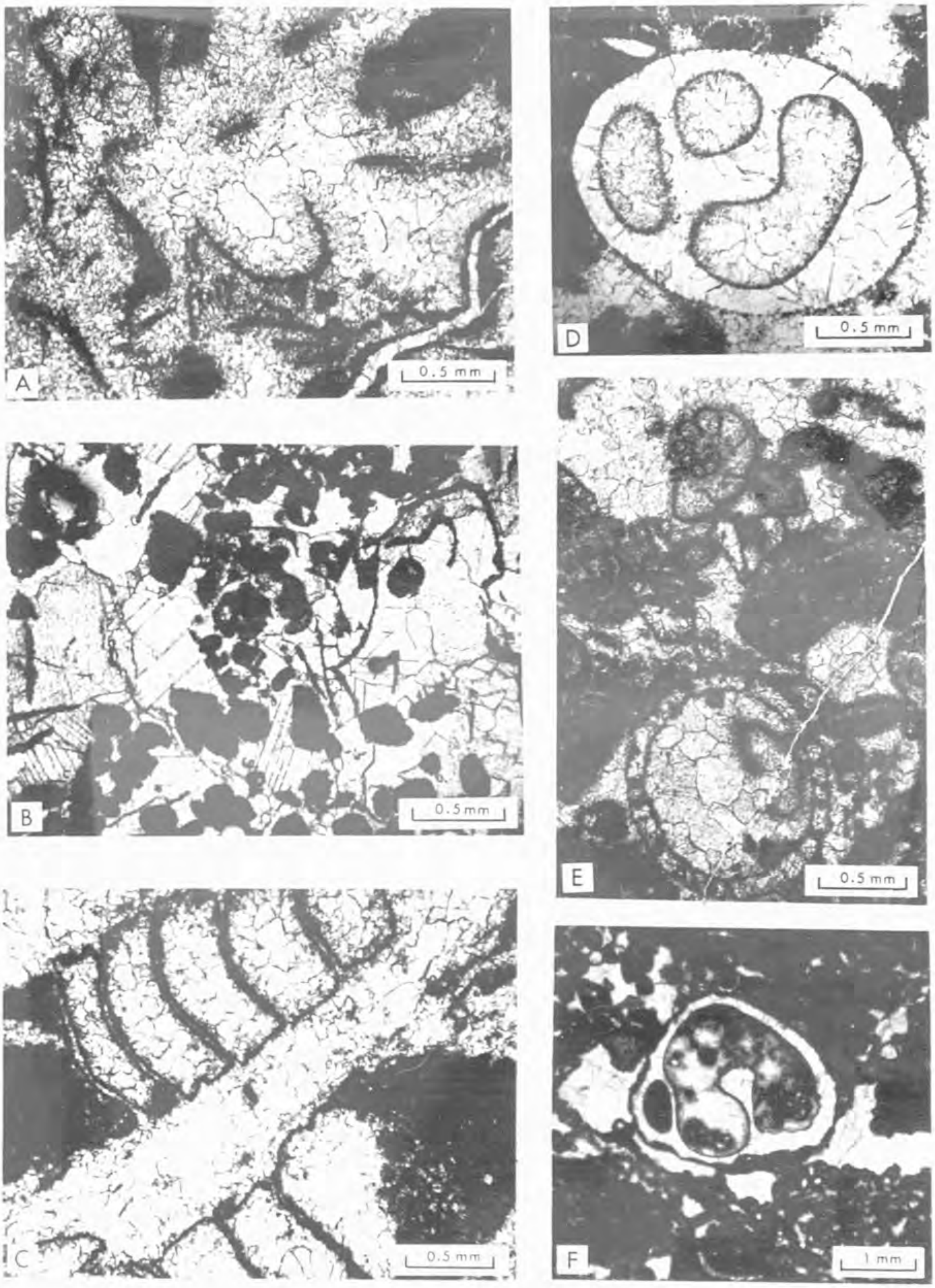


CEMENT TYPES

A. Interparticle pore space has been filled by a large calcite crystal which is in optical continouty in most part of the thin section. Sample No.1002.

B. Iron rich calcite crystals (arrows) line a pore space. This cement has most probably formed in the phreatic zone. The sample has later been silicified (white in micropho - graph). Peel. Sample No.S4/11. Location: Kuz (N66)

C. Sparry calcite fills pore spaces of stromatoporoid (?). Sample No.72/35; Location: Sincan Tepe.

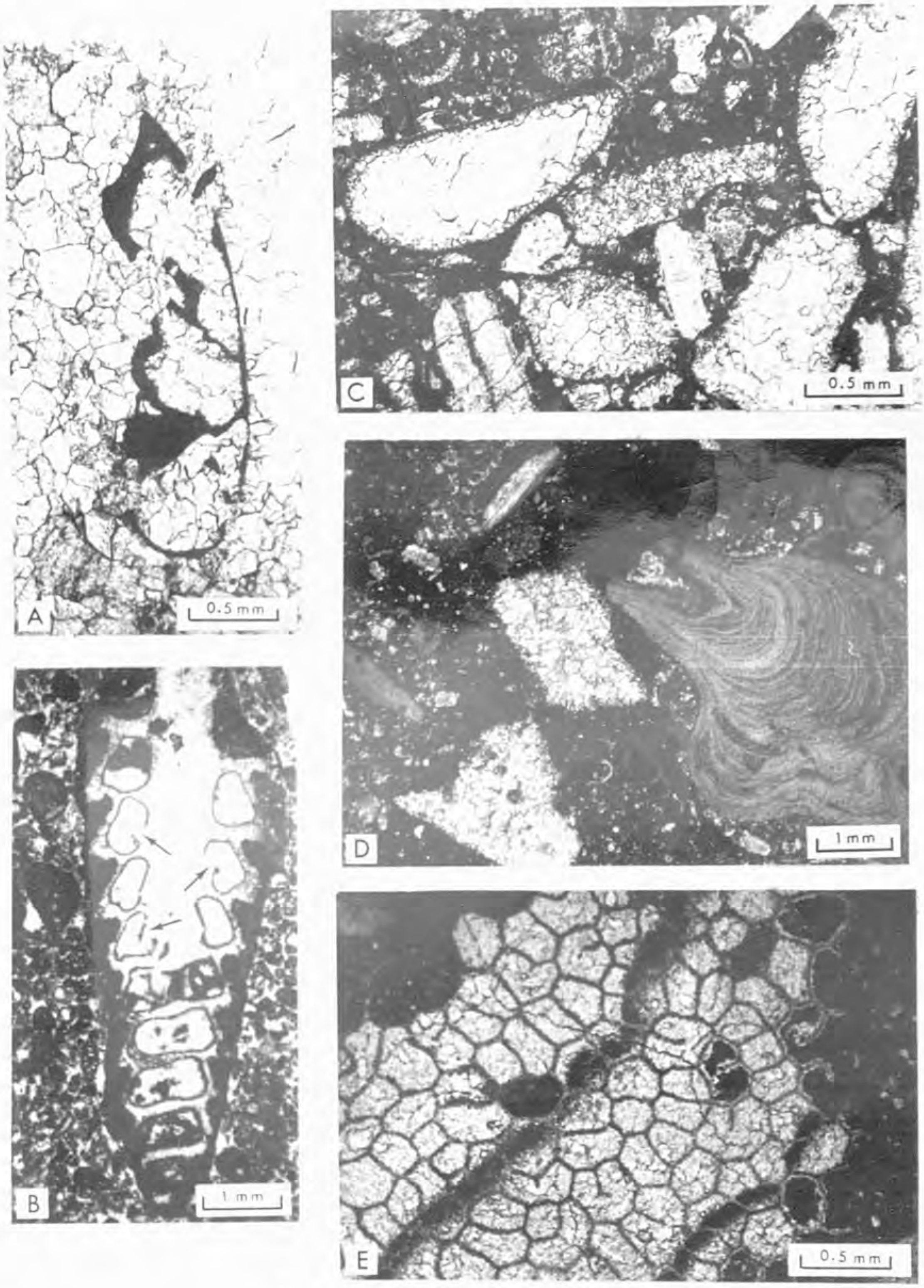


CALCITE SHELL FABRICS - 1

A and B. Sparry calcite containing fragments of micritic envelopes probably crystallized as a secondary infilling of a partially collapsed cavity formed by the solution of aragonitic skeletal fragments. Stained thin sections. A- Sample No.645 B- Sample No. 1 118.

C-F. Original pores of the skeletal fragments had started to fill with sparry calcite cement before the solution of the aragonitic skeletons and filling of the resulting pore with random calcite mosaic. All stained thin section. Sample Nos. C-1/56; Location: Dayakpınar; D-74/6; Locality: North of Arpaçukuru(K8); F-828.

PLATE -50



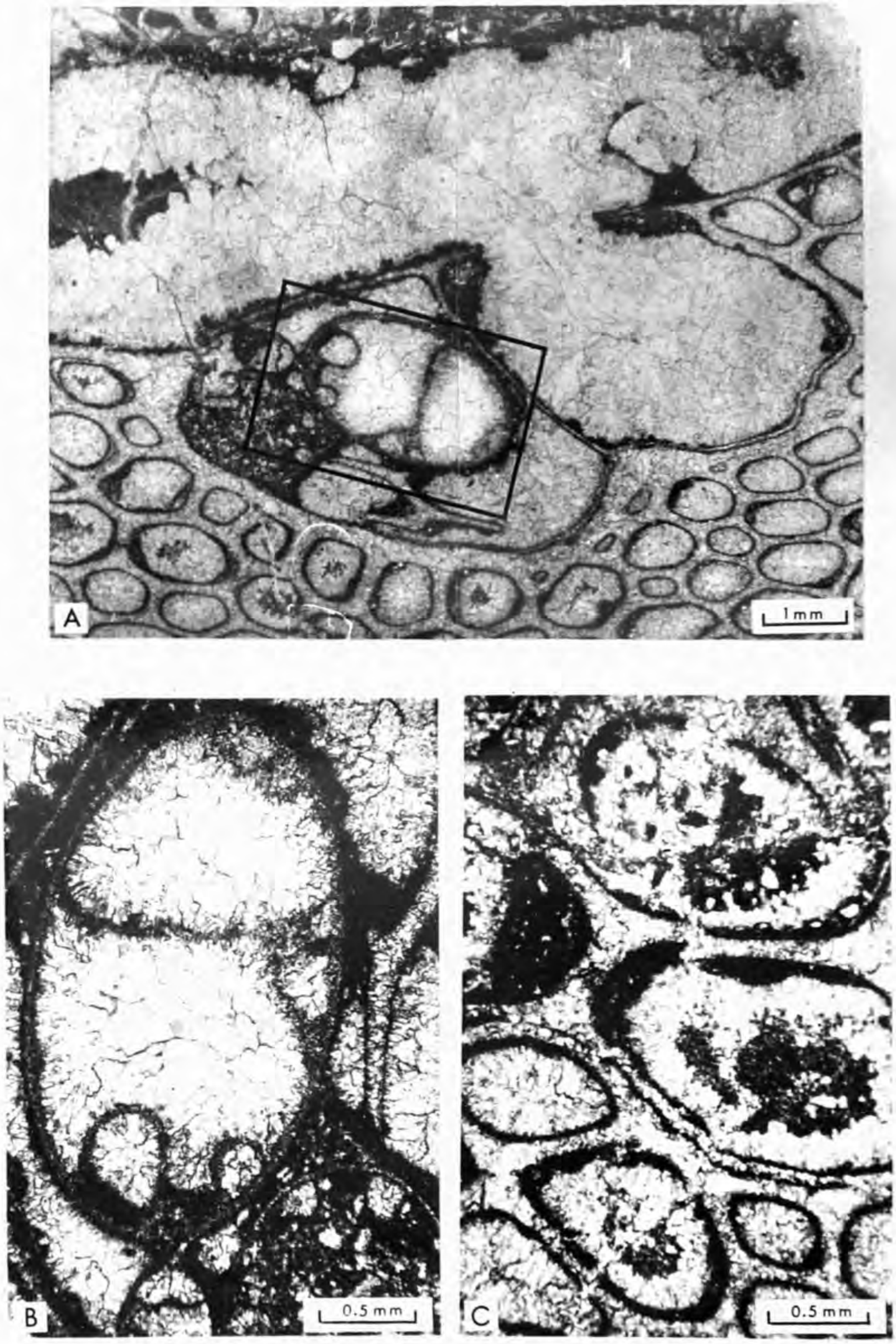
CALCITE SHELL FABRICS

A. Broken micritic envelope fragments in a random calcite mosaic. Sample:1/3B Sarıca.

B. Preservation of a gastropod shell. Solution of aragonite probably took place atleast some time before the infilling of original pores with sparry calcite as shown by broken micritic envelopes (arrows). Thin section. Sample No.248C

C and D. Aragonitic skeletal fragments have been completely dissolved and resulting voids are filled with calcite cement. The original structure of skeletons are completely lost. Stained thin section. Sample No.1/24; Location:Kuz (N66); D- Sample No.KÖ5; Location: Kızılören.

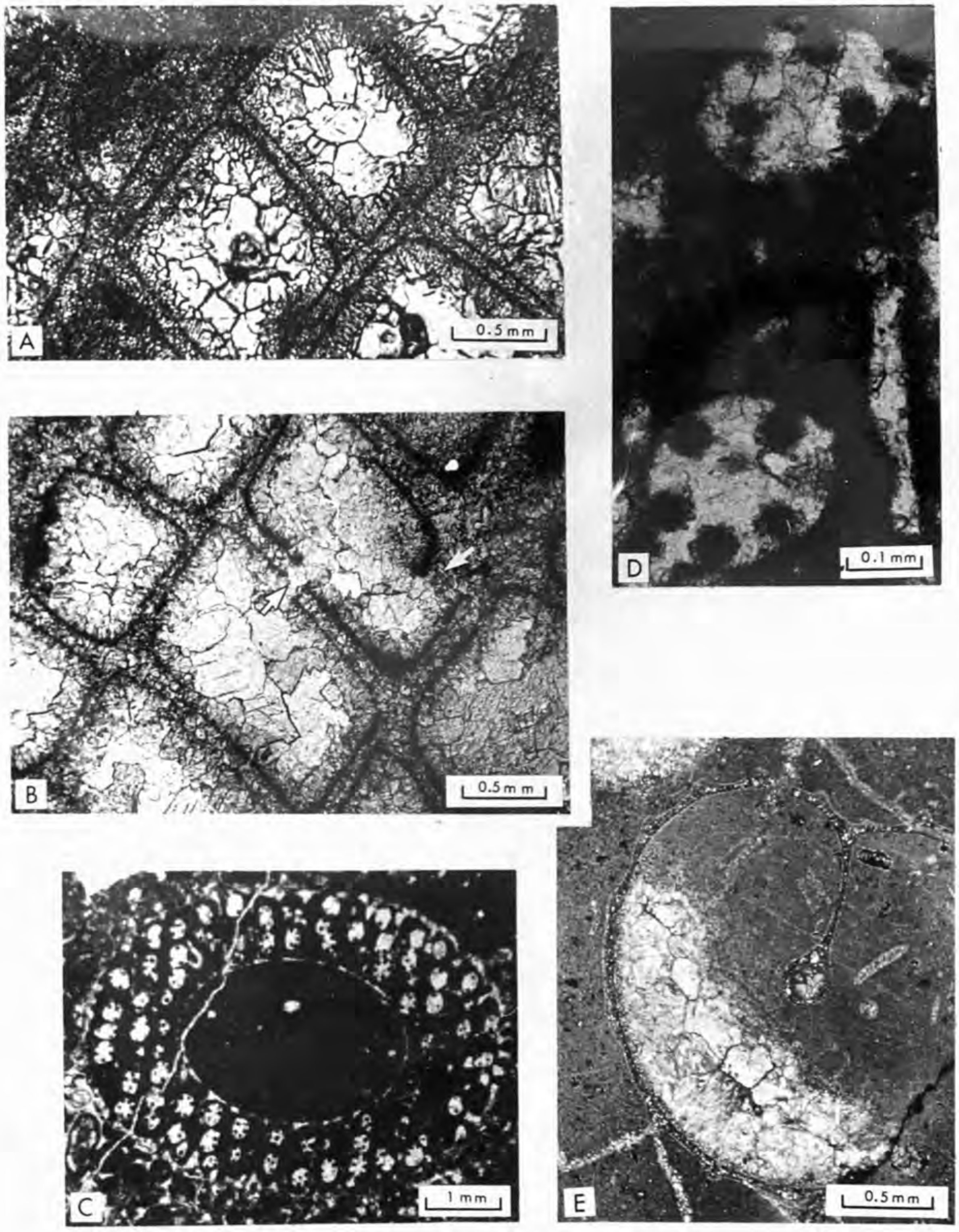
E. Sparry calcite fills most of the skeletal pore spaces which the rest are filled with lime mud. Stained thin section. Sample No.662.



CALCITE SHELL FABRICS OBSERVED IN CORALS

A. Sample No.651. Stained thin section.
B. Detail of area shown in A.
C. Detail of the lower part of A.

PLATE -52



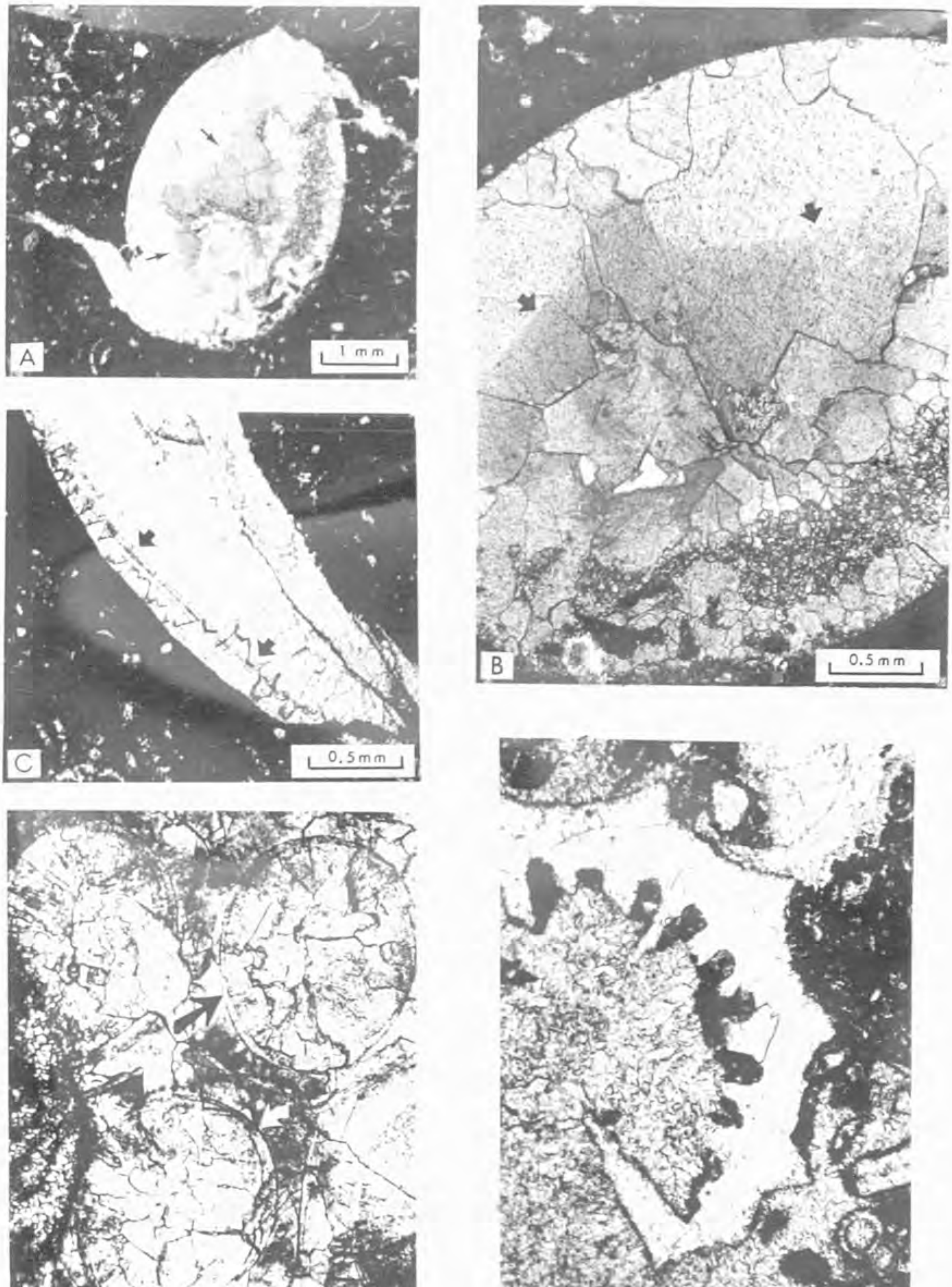
VARIOUS CALCITE SHELL FABRICS

A and B. The solution of aragonite of coral shown in these microphotographs probably took place after the infill of original pores with calcite cement. But broken micritic envelopes seen in B (arrows) indicate that cementation was not complete at the time of solution of aragonite. Sample No.:1/17; Location: Dayakpınar; A- Peel; B- Thin Sec.

C. The preservation of coral septa in the mud free cavity suggests that deposition of sparry calcite in the original cavity had atleast been started before solution occured. Thin section. Sample No.1/11 Locality: Kaz Tepe (D35); D- Detail of C.

E. The skeletal aragonite of the gastropod shell which was dissolved after the filling of the specimen partially with carbonate mud and partially sparry calcite. Peel. Sample No.1/28; Location: Kuz (N66).

PLATE -53

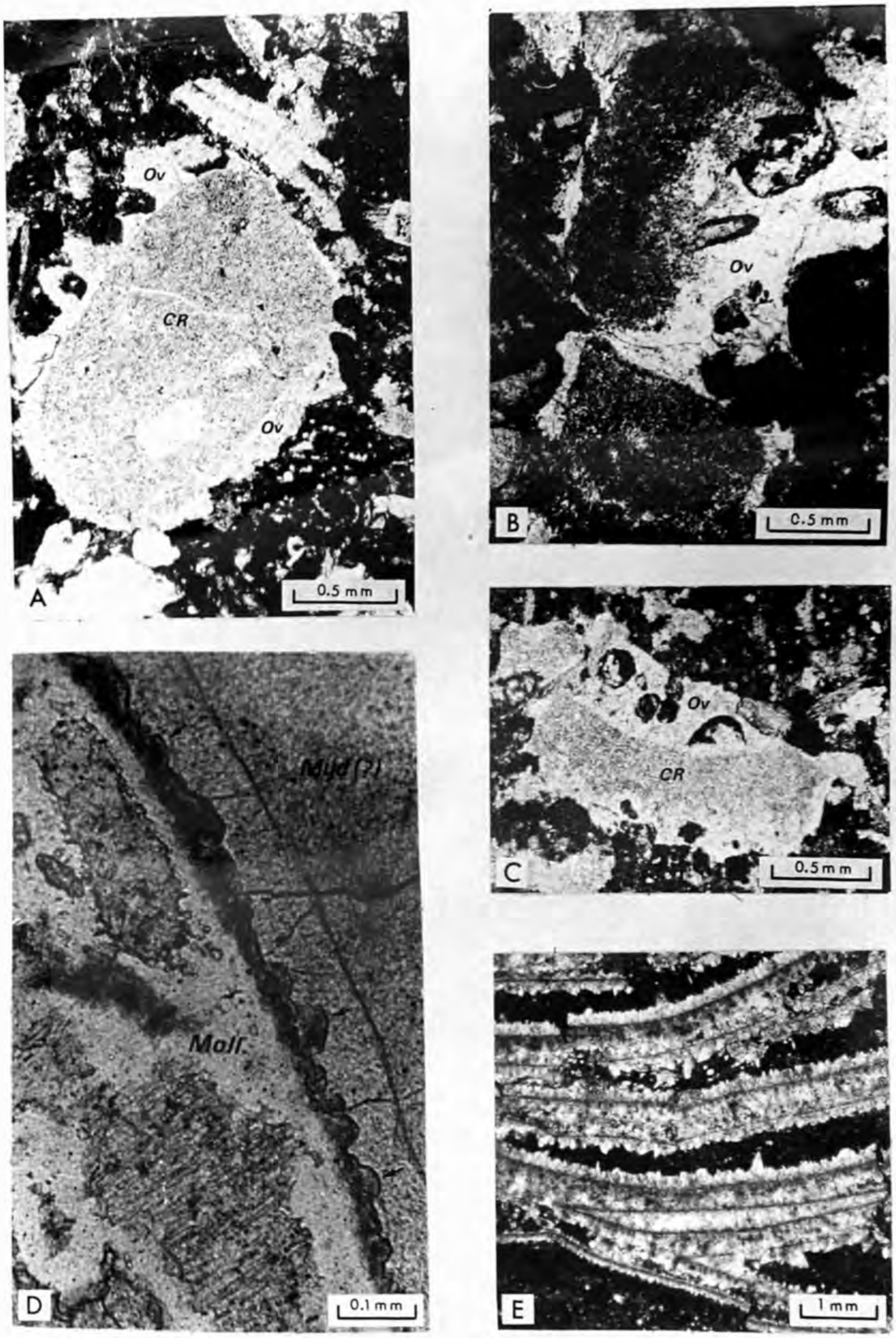


CALCITE SHELL FABRICS OBSERVED IN THE GÜRÜN SEDIMENTS

A. Ostracod (?) has been filled with iron free (light colour in microphotograph) and iron rich (darker) sparry calcite cement. Boundaries of two types have been shown by arrows. Iron rich calcite may represent phreatic zone. Vadose silt fills some of the pore space. Stained thin section. Sample No.837; B - Detail of A.

C. Molluscan shell has been replaced by a neomorphic sparry calcite. Its crystal mosaic is transected by lines (arrows) that are the relics of the original layered structure of the shell. Thin section. Sample No.309

D. In situ inversion of rudist. Relics of original structure are preserved (arrows). Stained thin section. Sample No.2/54; Location: Kaledere.



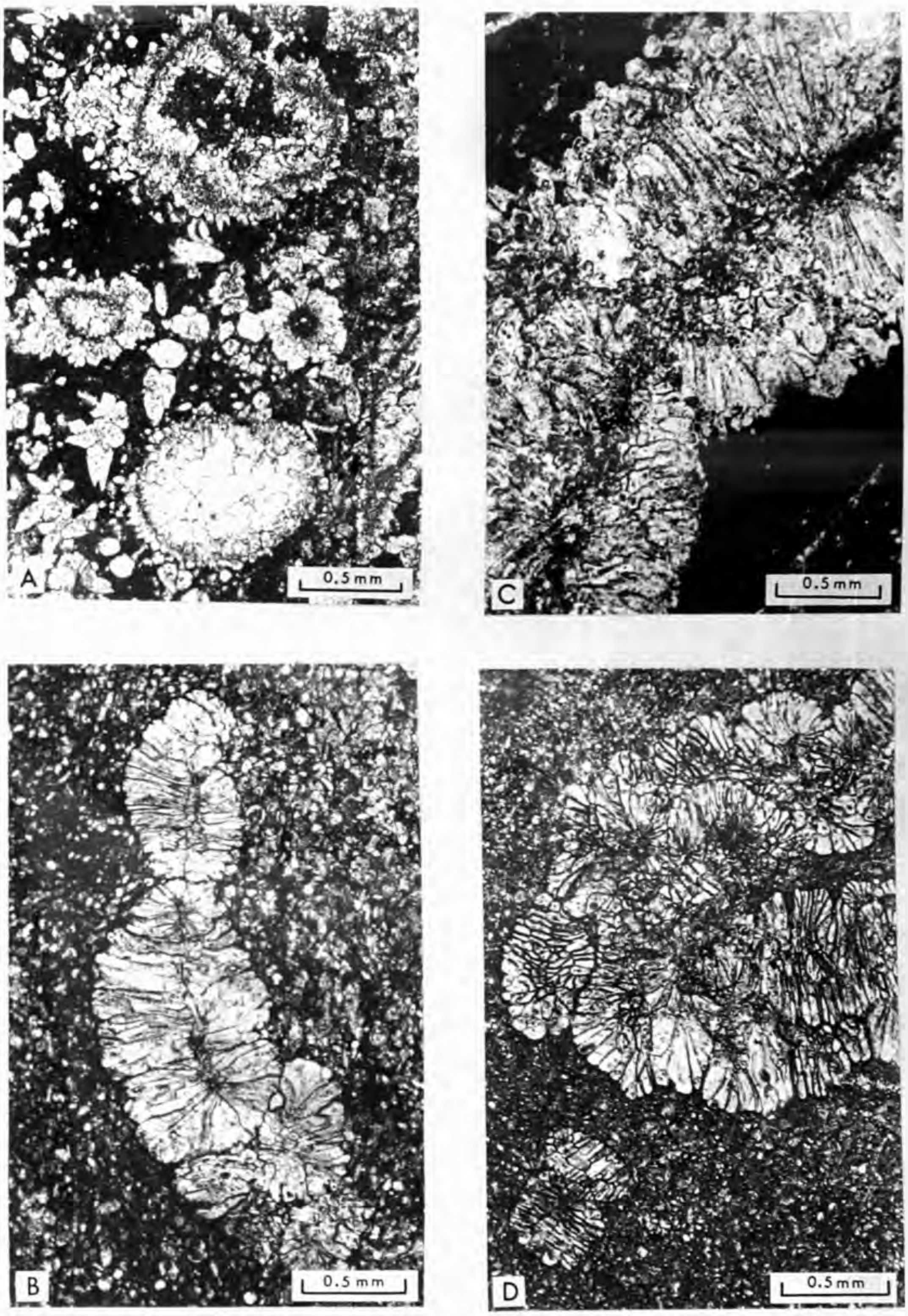
NEOMORPHIC CALCITE FABRIC TYPES

A, B, and C. Neomorphic overgrowth (Ov) on crinoid fragments (CR). Neomorphic origin of the overgrowts is evident from the patches of matrix in the overgrowths. A - Thin section. Sample No.1/48; Location: Reşadiye; B - Stained thin section; Sample No.336; C - Stained thin section; Sample No.1/48; Location: Reşadiye.

D. Neomorphic overgrowth (?) shown by arrows, on a silicified molluscan shell wall extending syntaxially into the adjacing matrix which is now completely silicified. Thin section. Sample No.663

E. Equant neomorphic crust. Stained thin section. Sample No.723B.

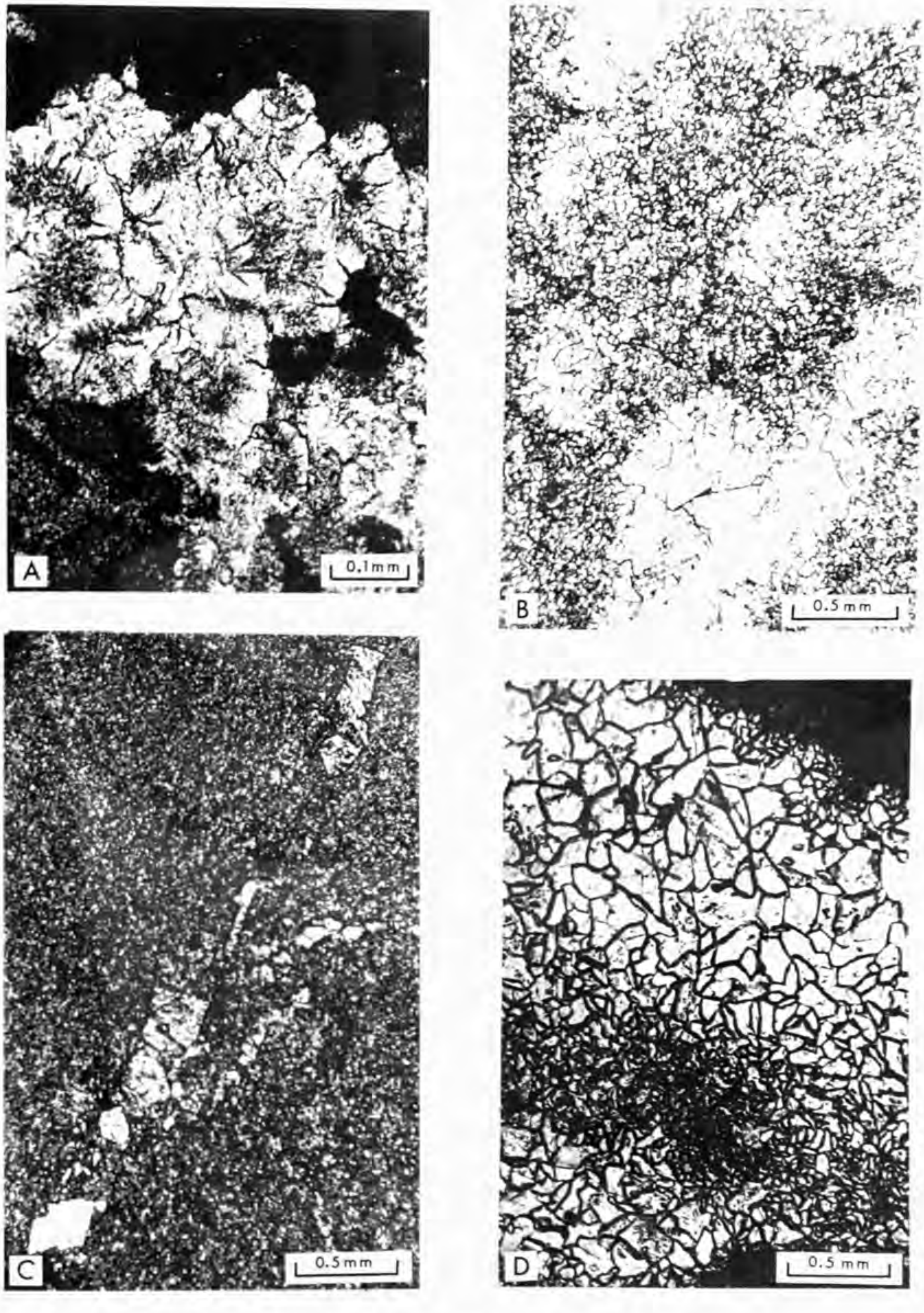
PLATE -55



NEOMORPHIC CALCITE FABRIC TYPES

A. Pseodospar both fringes the grains and grows towards the centre of skeletal fragments. Some of the skeletal fragments are completely neomorphosed by pseudospar. Stained thin section. Sample No.723B

B, C, and D. Neomorphic fibrous calcite. B – Stained thin section; Sample No.74/40 Location: Konakpınar; C – Stained thin section; Sample No.74/33; Location: Keşlik Dere; D – Peel. Sample No.74/40; Location: Konakpınar.



NEOMORPHIC CALCITE FABRIC TYPES

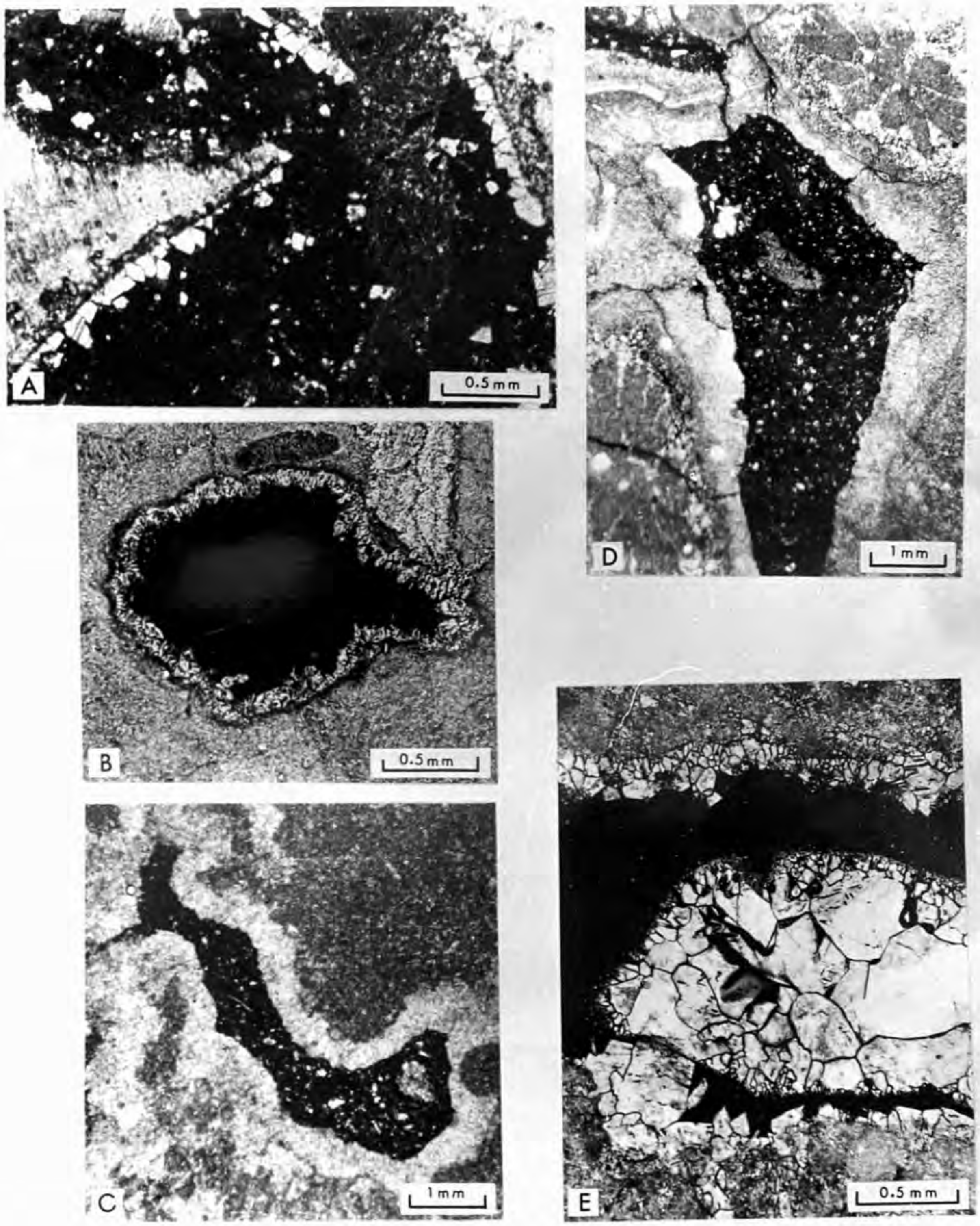
A. Stellate. Neomorphic fibrous calcite with cores of uquigranular microspar (Compare with BATHURST, 1971; Fig.333; p.482). Stained thin section. Sample No.469B.

B. Microspar and pseudospar. Pseudospar crystals are clear whereas small neomorphic crystals are cloudy. Stained thin section. Sample No.1/26; Location: Kuz (N66)

C. Euhedral calcite crystals have been arranged in a line. Origin unknown. They are probably very late diagenetic neomorphic pseudospar formed along a joint. Peel. Sample No.441.

D. Micrite, microspar and pseudospar can be seen in the same microphotograph. Peel. Sample No.1157.

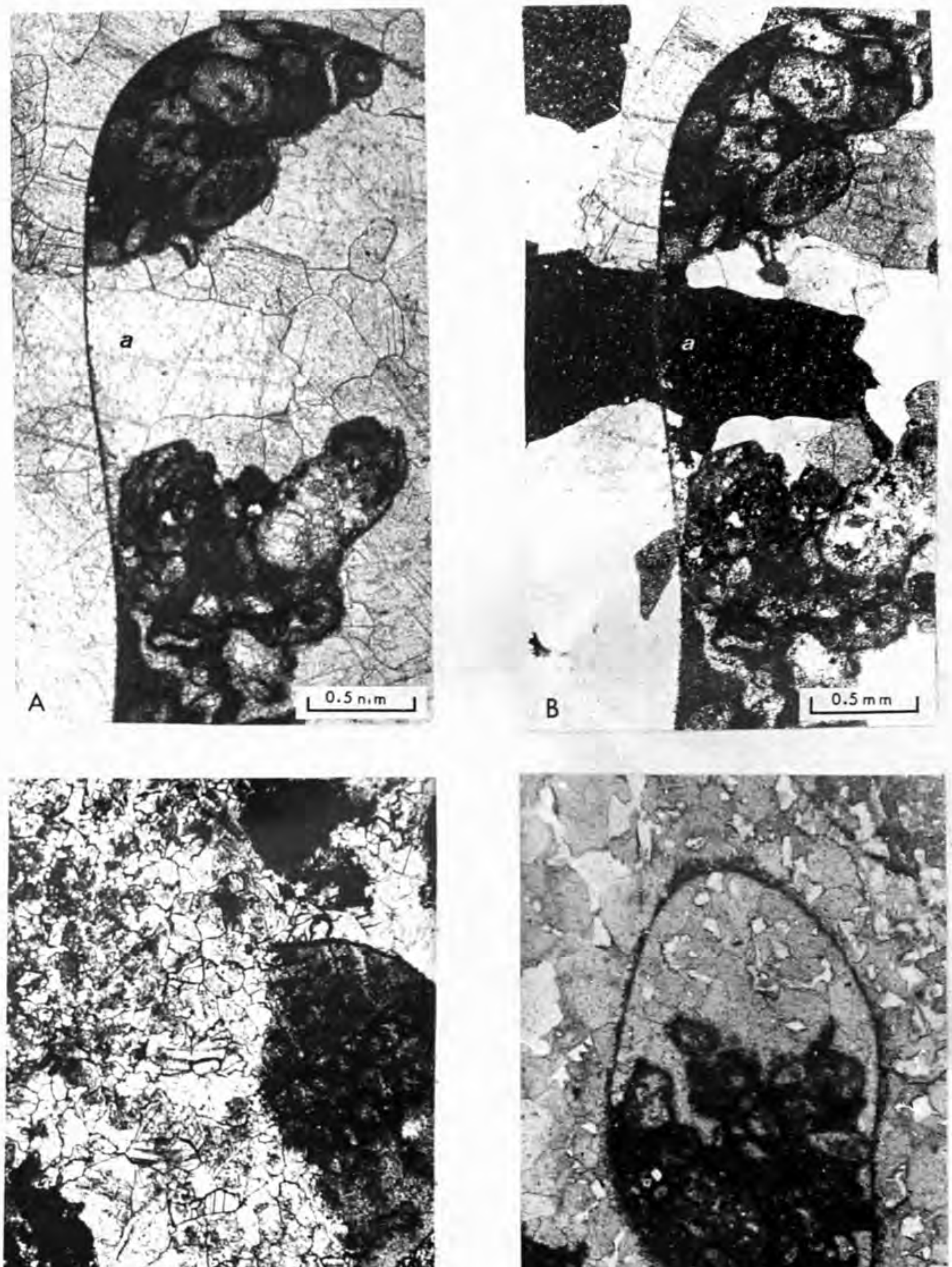
PLATE -57



NEOMORPHIC CALCITE FABRIC TYPES

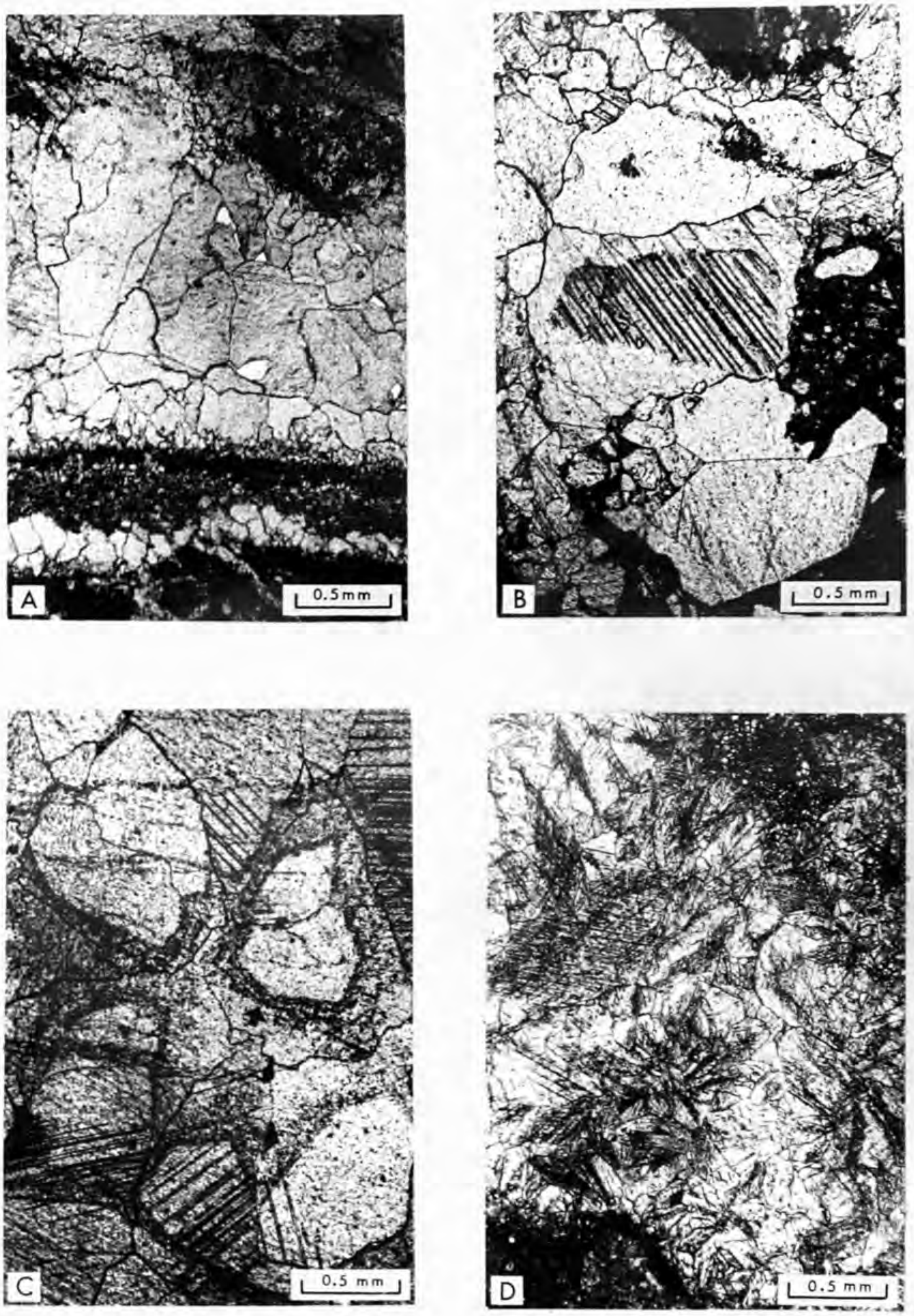
A. Neomorphic equant calcite crystals formed on skeletal grains. Stained thin section. Sample No.1/44; Location: Sincan Tepe.

B - E. Neomorphic aurrsoles surrounding clay clast in lime mudstones (For discussion see text p.). B - Peel. Sample No.640B; C - Peel. Sample No.74/6; Location: Yöz Tepe; D - Same as C.; E - Peel. Sample No.1/16; Location: Dayakpınar.



CHARACTERISTICS OF NEOMORPHIC CALCITE

A and B. The crystal at (a) is in optical continuity in both sides of the micritic envelopes. Stained thin section. Sample No.828. B - Crossed Nicols.
C. Stained thin section. Sample No.2/20; Location: Çiçekyurt.
D. Stained thin section. Sample No.S3/8.

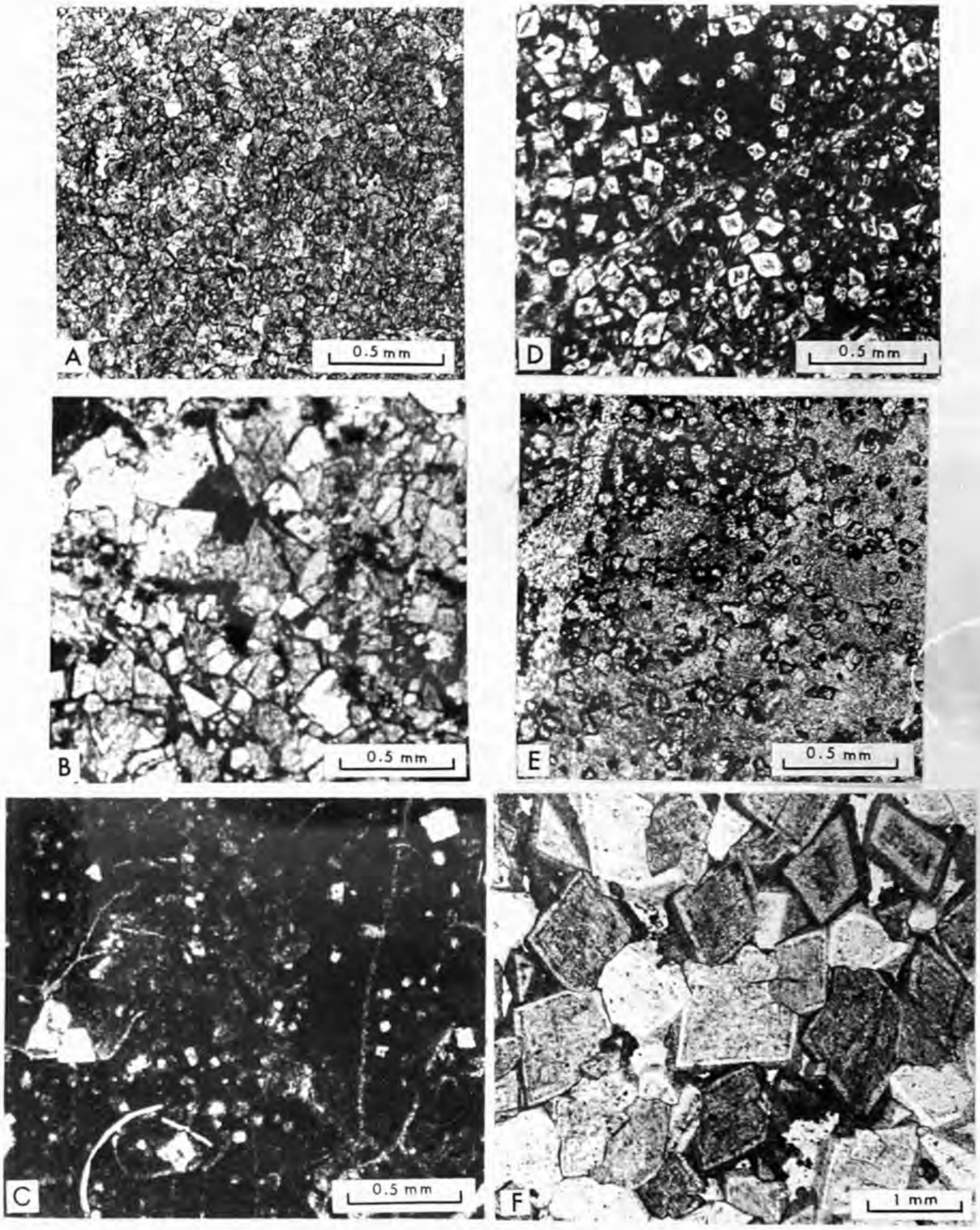


CHARACTERISTICS OF NEOMORPHIC CALCITE

A. Microspar and pseudospar. Stained thin section. Sample No.1/16; Location: Dayakpinar.

B. Pseudospar crystal showing twin laminae. Stained thin section. Sample No.72/53; Location: Kale Dere.

C. Pseudospar crystals show the outlines of former crystals indicating the conversion of small crystals to larger ones. Stained thin section. Sample No.54/11. Location: Kuz (N66)



TYPES OF DOLOMITE

A. Xenotopic dolomite crystals in a dolostone. Stained thin section. Sample No.73/54 Location: Kale Dere (L56)

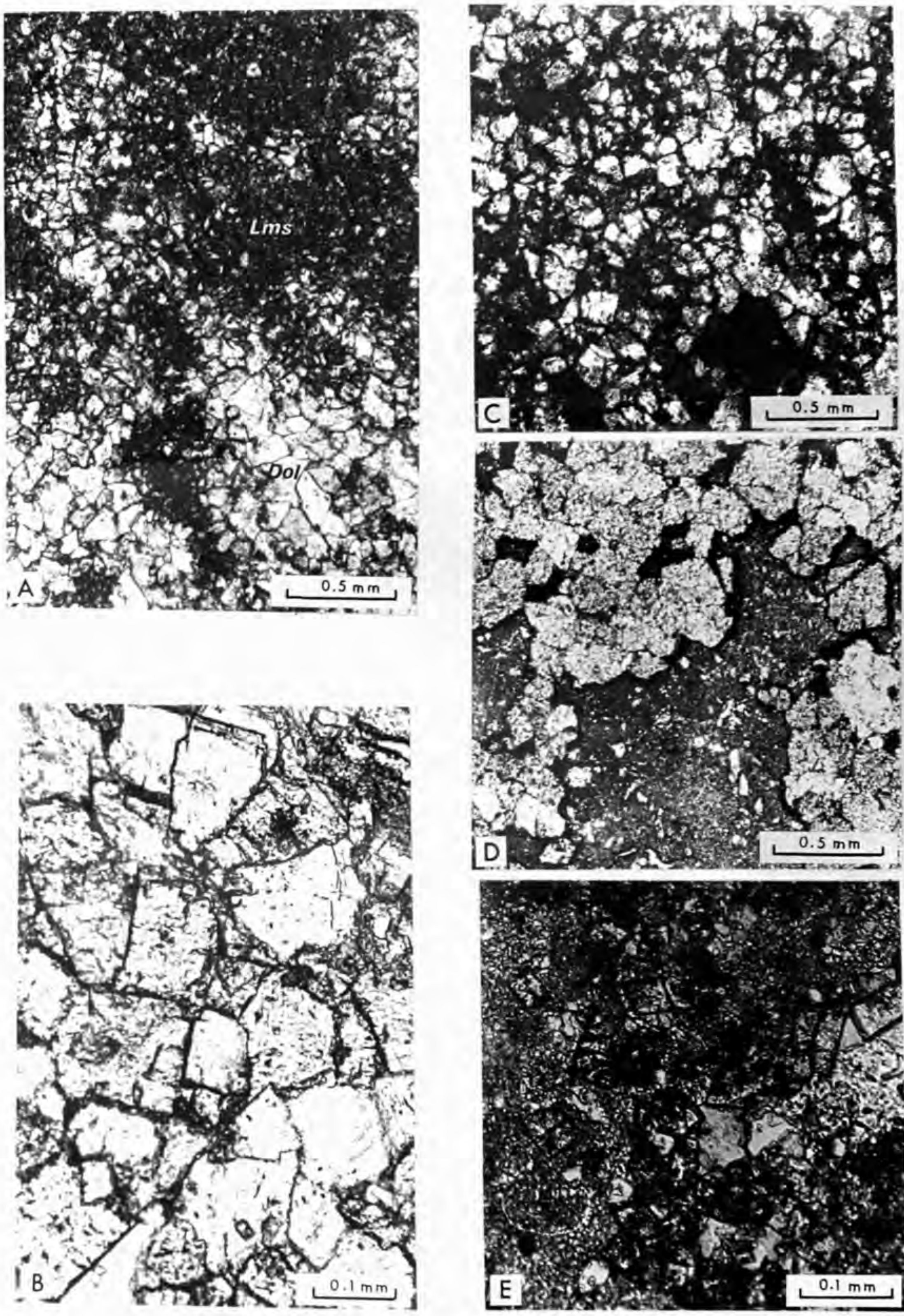
B. Hypidiotopic dolomite crystals. Stained thin section. Sample No.73/25; Location: Kuz

C. Scattered dolomite rhombs in a lime mudstone. Stained thin section. Sample No.72/39 Location: Irmac (North of Behram "X56").

D. Idiotopic dolomite crystals in same sample at C.

E. Microphotograph of a peel showing scattered dolomite rhombohedra in a lime mud matrix. Sample No.456B

F. Hydrotermal dolomite (Courtesy of C.MAHMUT).



FEATURES OF DOLOMITIZATION

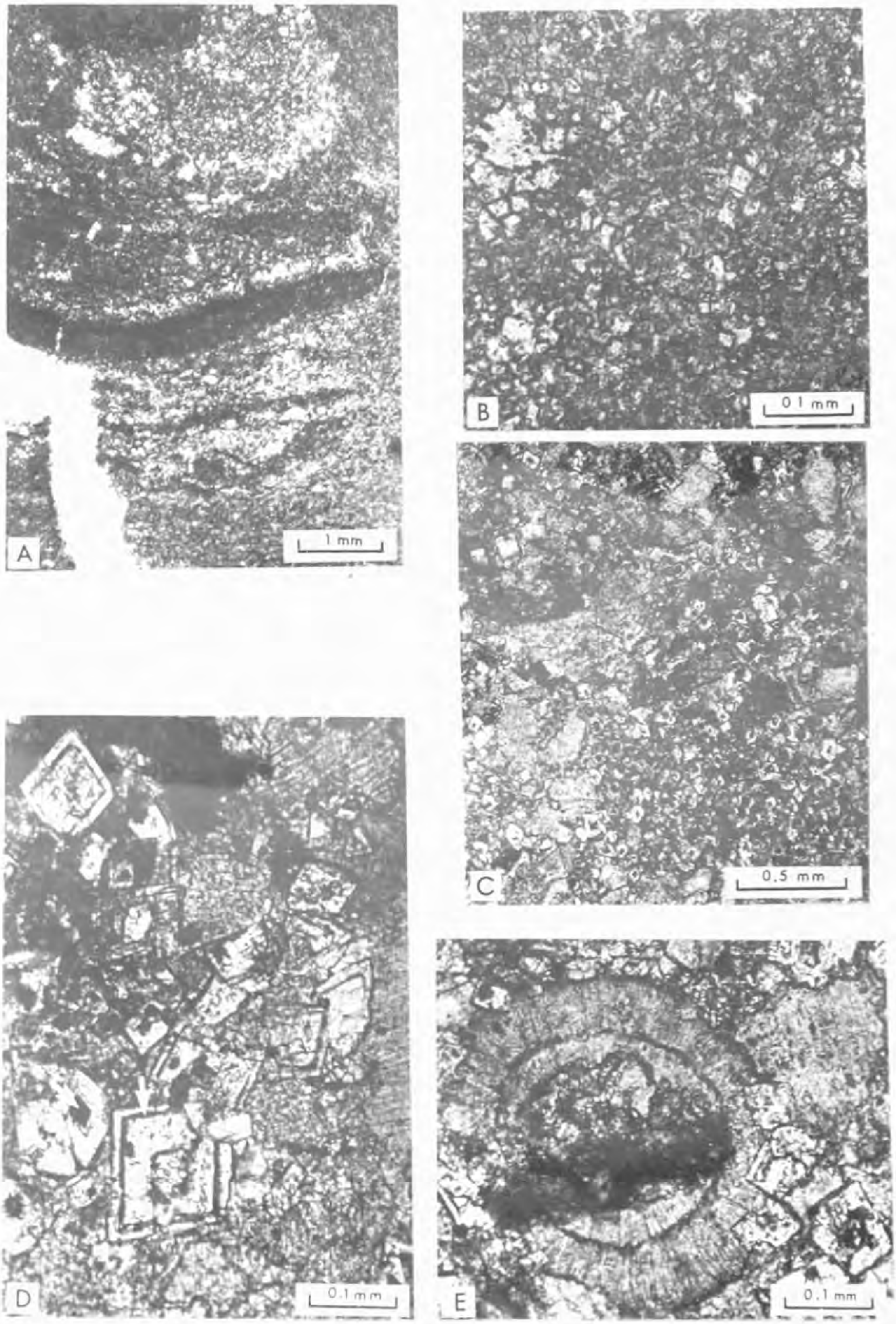
A. Dolomites (Dol) replacing lime mudstone(Lms). Thin section. Sample No.450

B. Detail microphotograph of dolomitized parts of sample at A.

C. As A. Dolomite crystals have a hematatic rim. Stained thin section. Sample No. 1010.

D. Same sample at C. Peel.

E. Microphotograph of a peel showing scattered dolomite rhombohedra in a lime mud matrix. Sample No.456B.



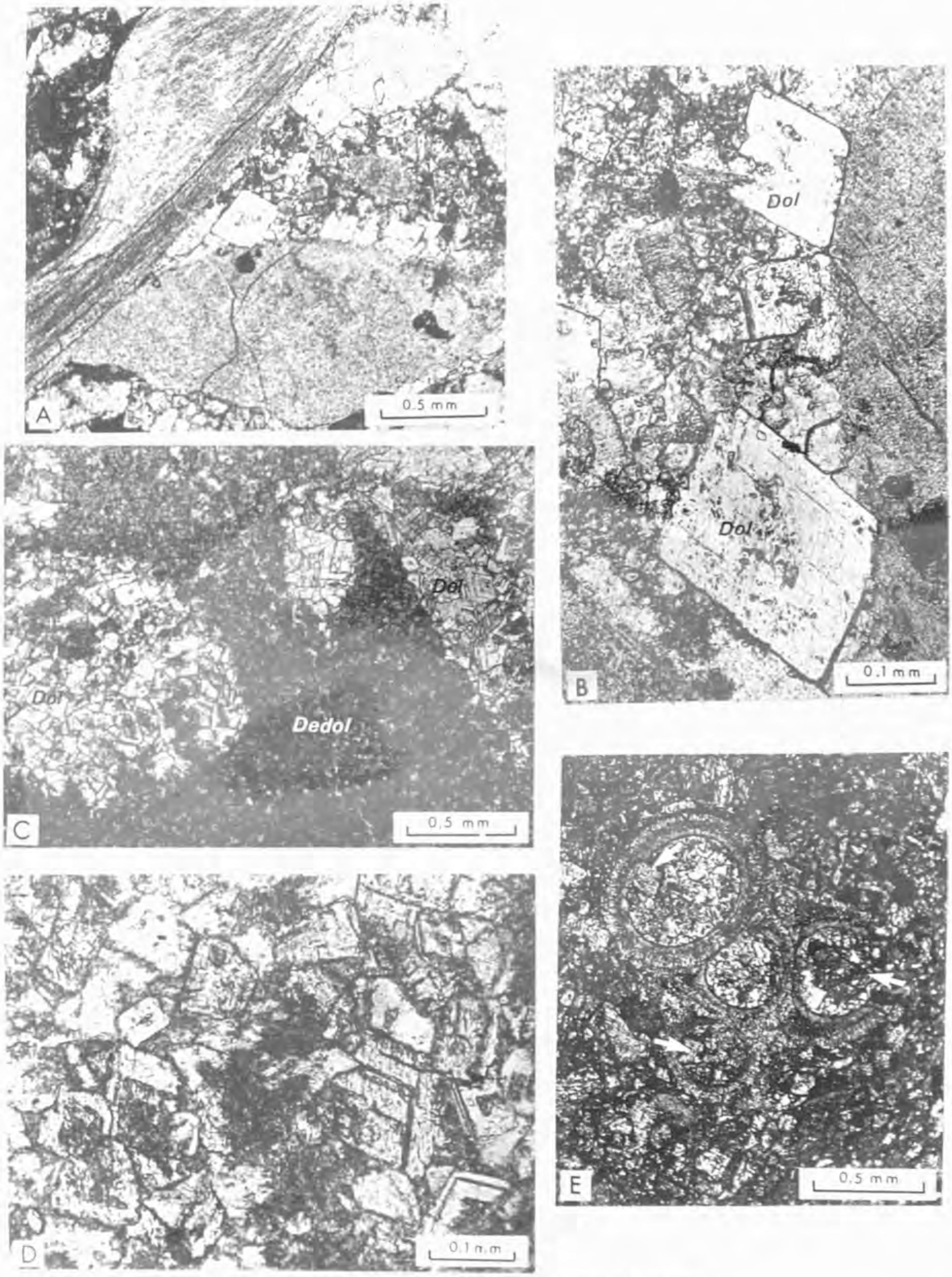
DOLOMITIZATION IN THE GÜRÜN SEDIMENTS

A. Dolomitized stromatolitic limestone. Dolomite crystals are white in the microphotograph. The white feature on the left hand bottom is a calcite vein. Stained thin section. Sample No.74/29; Location: Konakpınar (R60) B- Detail of A.

C. A calciturbidite bed has been dolomitized by zoned dolomite crystals. Most of the centres of dolomite crystals are composed of hematite. Stained thin section. Sample No.KAY 10; Location: Kaynarca.

D. Detail of C. Hematite core and zone have been shown by arrows.

E. Microphotograph of same thin section showing dolomite rhombs replacing shell fragment.

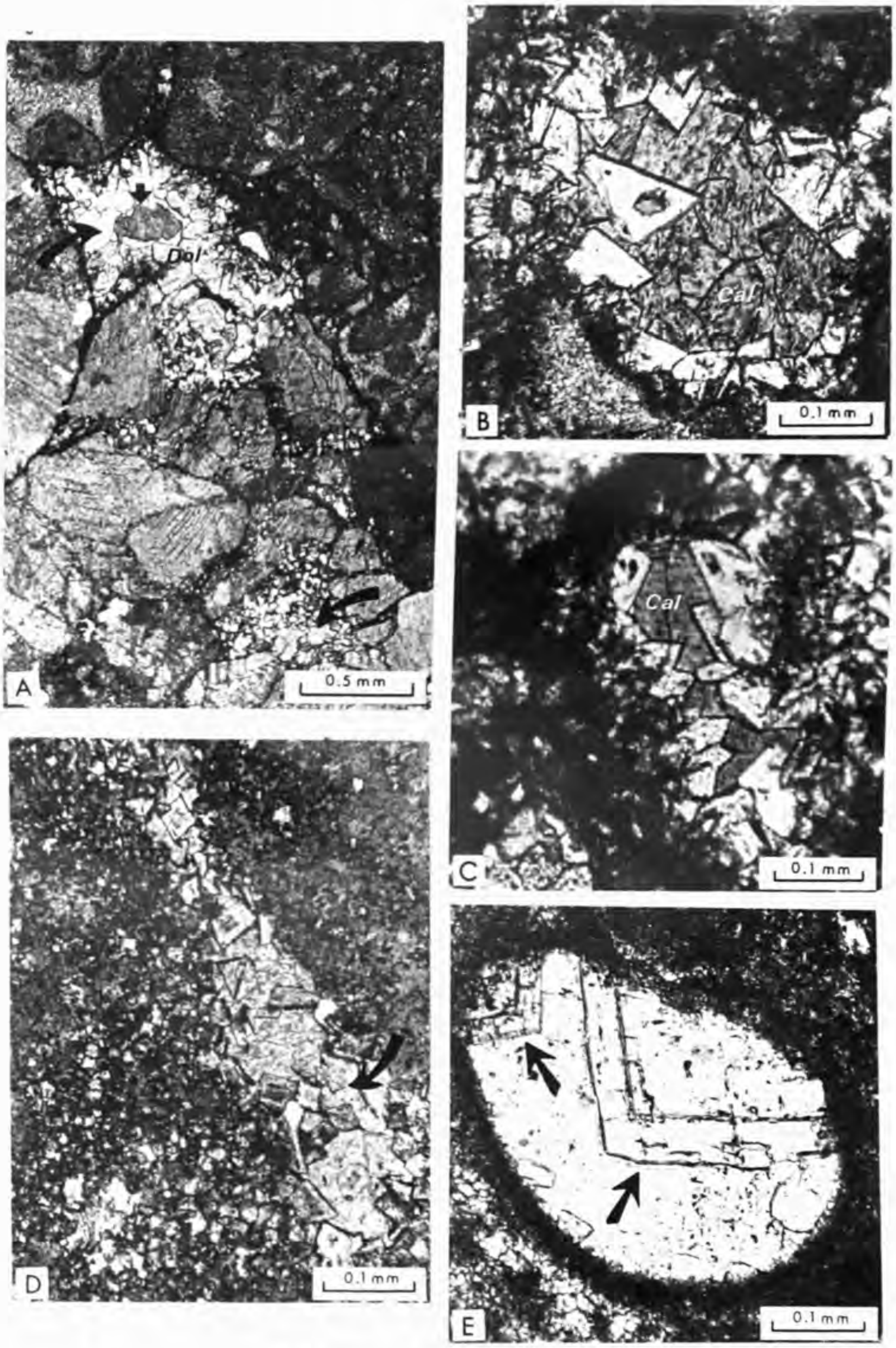


DOLOMITIZATION AND DEDOLOMITIZATION

A. Dolomite rhombs replacing matrix. Shells are not affected from dolomitization. Thin section. Sample No.338-339. B - Detail of A.

C. Dark centered and clear rimmed dolomite crystals have partially been replaced by calcite. Resultant texture is pseudo-micropelletoidal (EVAMY, 1967). Stained thin section. Sample No.451. D - Detail of dolomitized parts of C.

E. Dolomite rhombs replacing matrix and infills of the gastropod chambers. Peel. Sample No. 338-339.



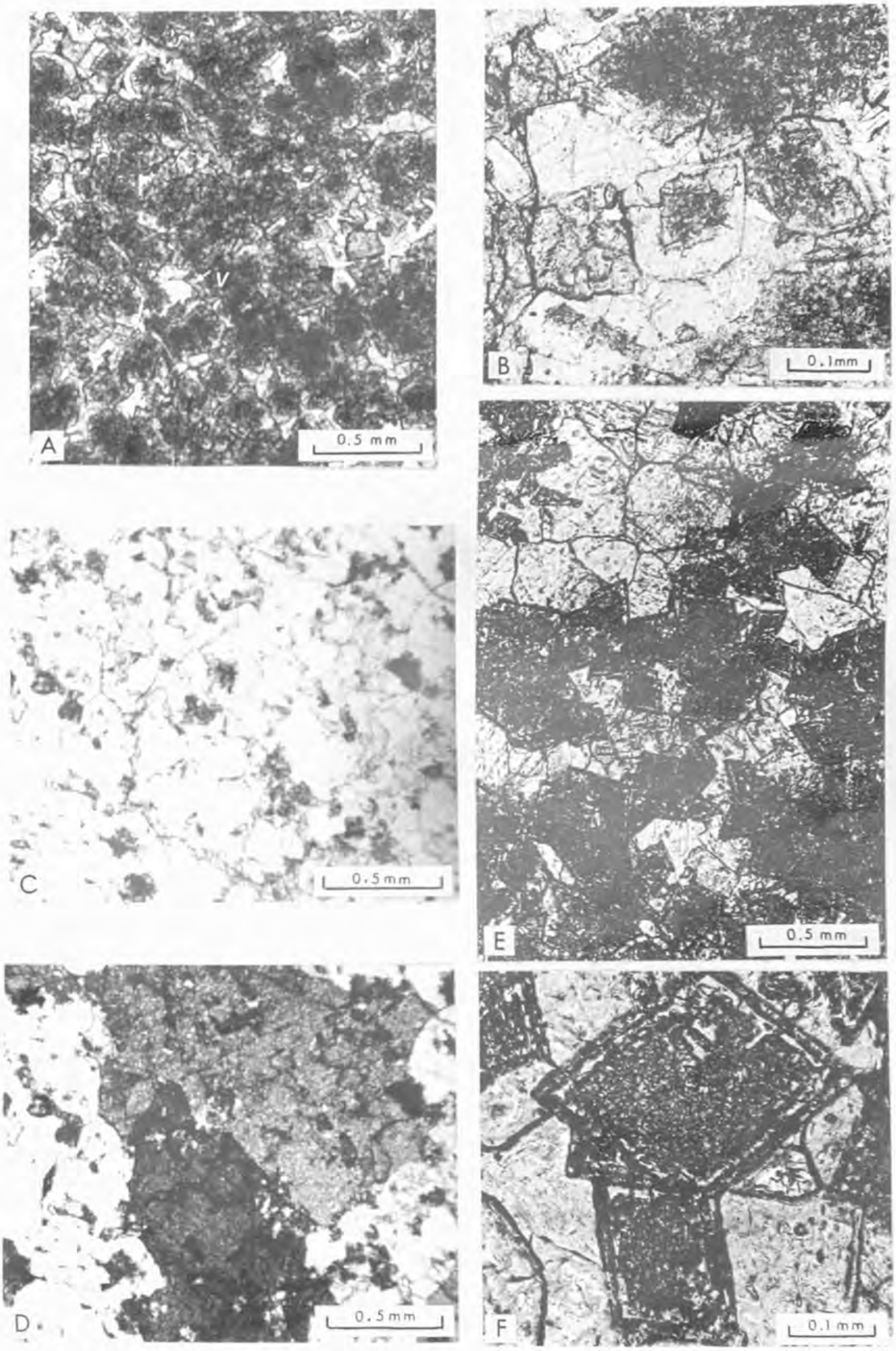
DOLOMITE AS A CEMENT

A. Dolomite cement (large arrows) in a brecciated limestone. The centre of the top one is filled with iron rich calcite (small arrow). Stained thin section. Sample No. Kuz 2

B and C. Details of dolomite cement. Dolomite crystals are seen white in the microphotographs. Iro rich calcite (Cal) fills the remaining pore space.

D. Dolomite crystals filling a crack (arrow) in a dolomitized limestone. Stained thin section. Sample No.121

E. Dolomite cement filling a void formed by dissolution of a shell fragment. The uncomplete perfect dolomite zones (arrows) suggest that shell has been broken after filled by dolomite cement. Stained thin section. Sample No.1010.



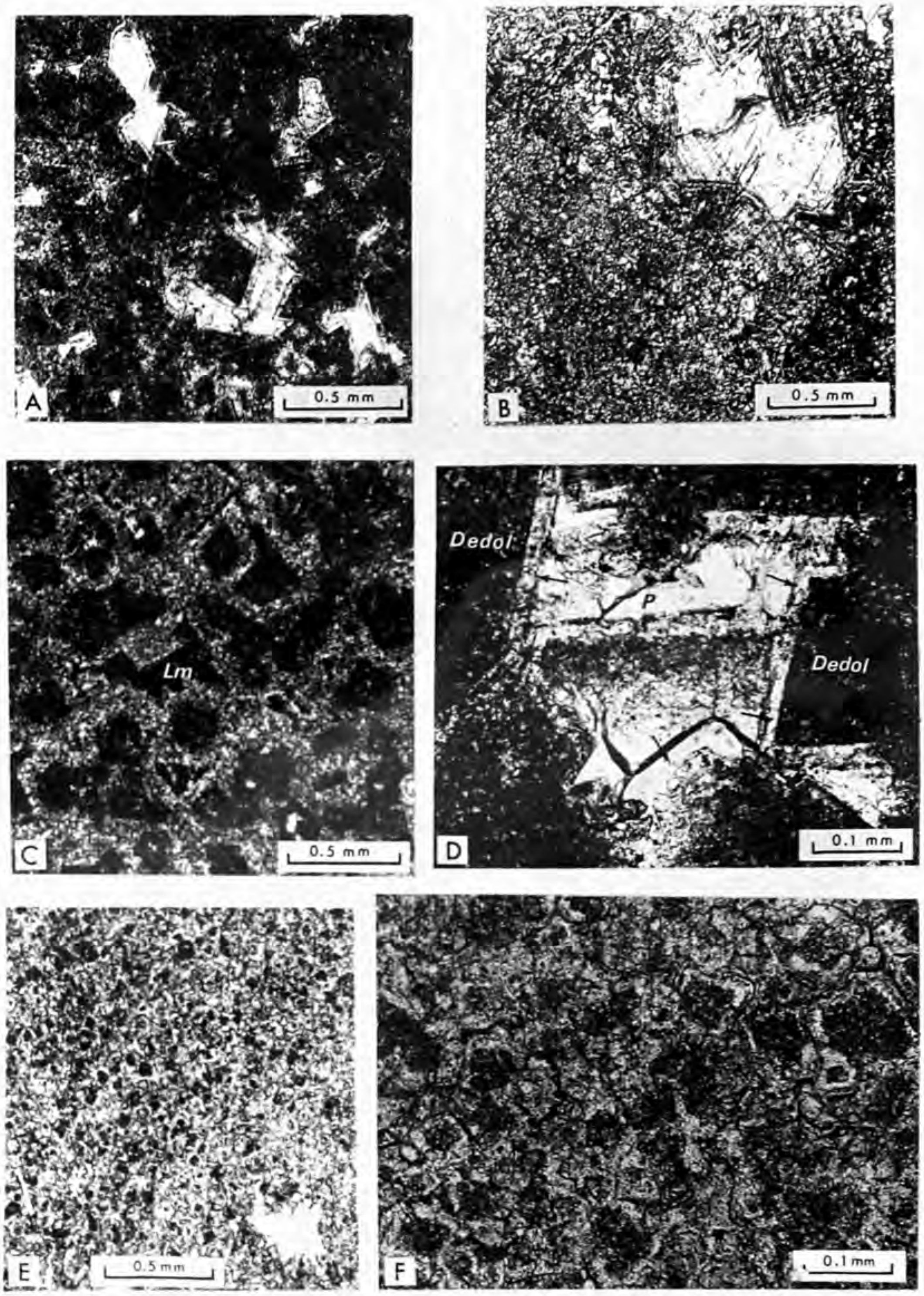
DEDOLOMITIZATION IN THE GÜRÜN SEDIMENTS

A. Partial dedolomitization. Cores of rhombs are composed of calcite (Black in microphotograph). Most of the rims are still dolomite. Some dolomite cement and voids (V) are visible. Stained thin section. Sample No.337; B - Detail of A.

C. Centripetal dedolomitization. The dolomite rhombs are replaced by a more coarsely crystalline mosaic of calcite crystals. Stained thin section. Sample No.359

D. Crossed Nicols. E - Limestone after dolomite. Peel. Sample No.52/53

F. Detail of E.



CENTRIFUGAL DEDOLOMITIZATION

A and B. Zoned dedolomite rhombs and rhombo-mouldic pores, filled with calcite. Peels. Sample Nos. A-1021-1022; B- 72/60; Location: Reşadiye.

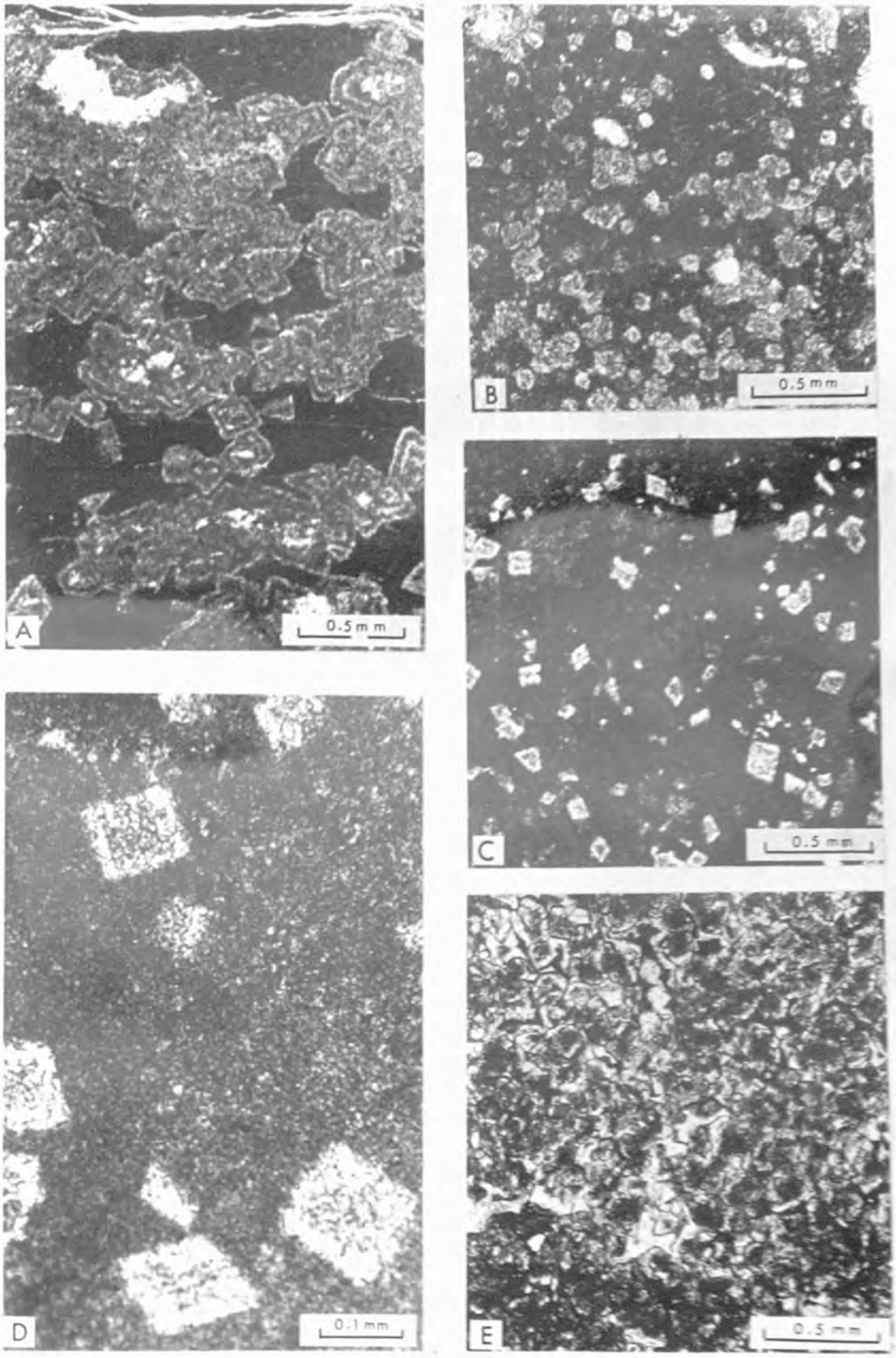
C. Dedolomitization of partially dolomitized limestone. Original lime mud matrix has been preserved in some places (Lm). Dedolomite rhombohedra are composed of dark cetres and lighter and coarser outer zones. Stained thin section. Sample No.72/62; Location: Reşadiye.

D. Rhombo-mouldic pore (P) bordered by zoned (arrows) dedolomite rhombohedra (Dedol) Stained thin section. Sample No.2/53 Locality: Reşadiye.

E. Pseudo-pellotoidal texture. See Fig.V-12. Stained thin section. Sample No.448

F. Detail of E.

PLATE - 67



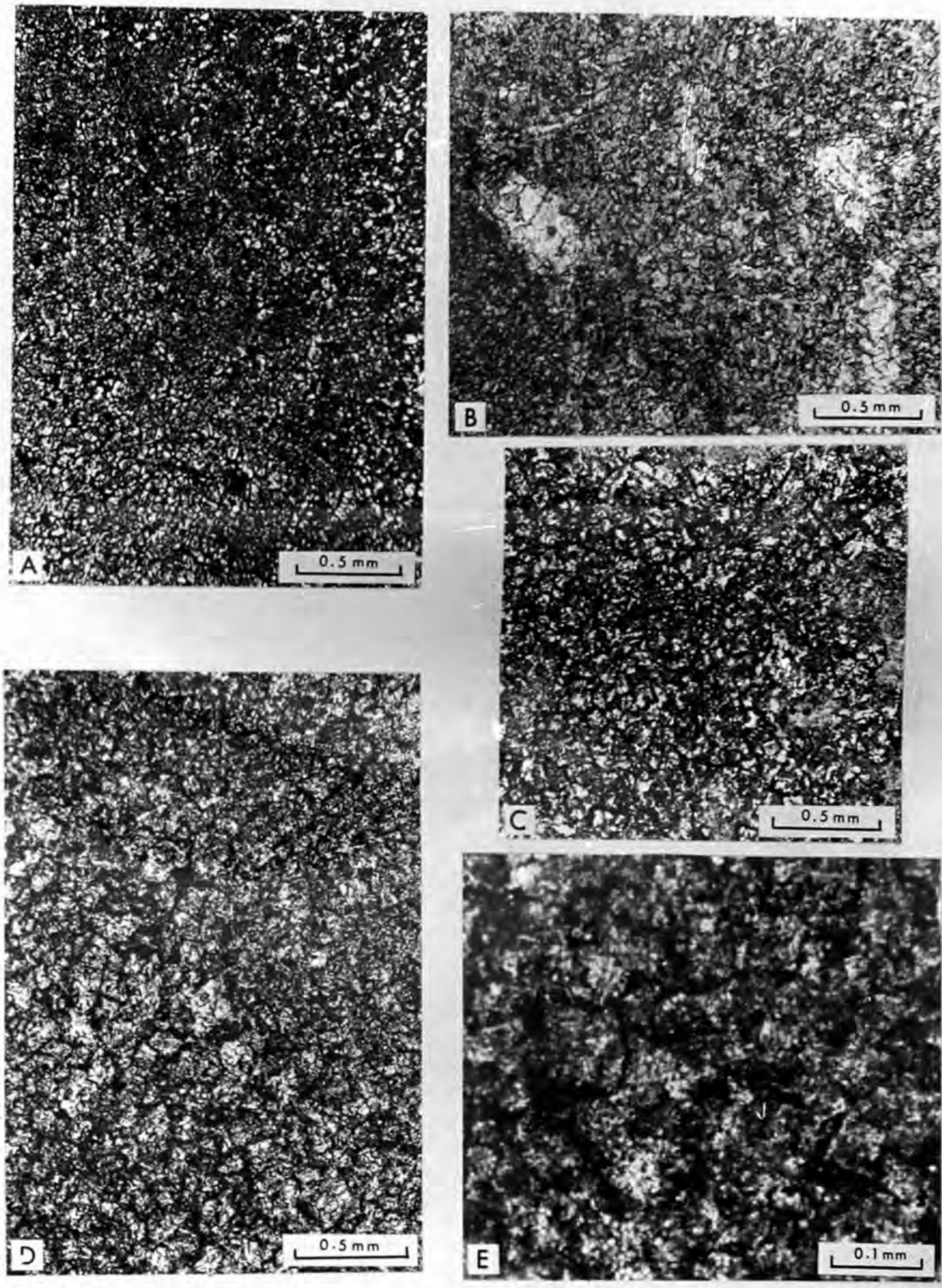
CENTRIFUGAL DEDOLOMITIZATION

A. Lime mudstone is replaced by zoned dolomite crystals has been completely dedolomitized. Stained thin section. Sample No.74/25; Location: Reşadiye.

B and C. Dedolomitization of scattered dolomite crystals, Stained thin section. Sample Nos. B-1166 ; C-749.

D. Detail of composite calcite rhombohedra seen at C.

E. Centrifugal replacement of a dolostone. Cores of rhombohedra are calcite while the outer parts are still dolomite. Sample No.359



CENTRIPETAL DEDOLOMITIZATION

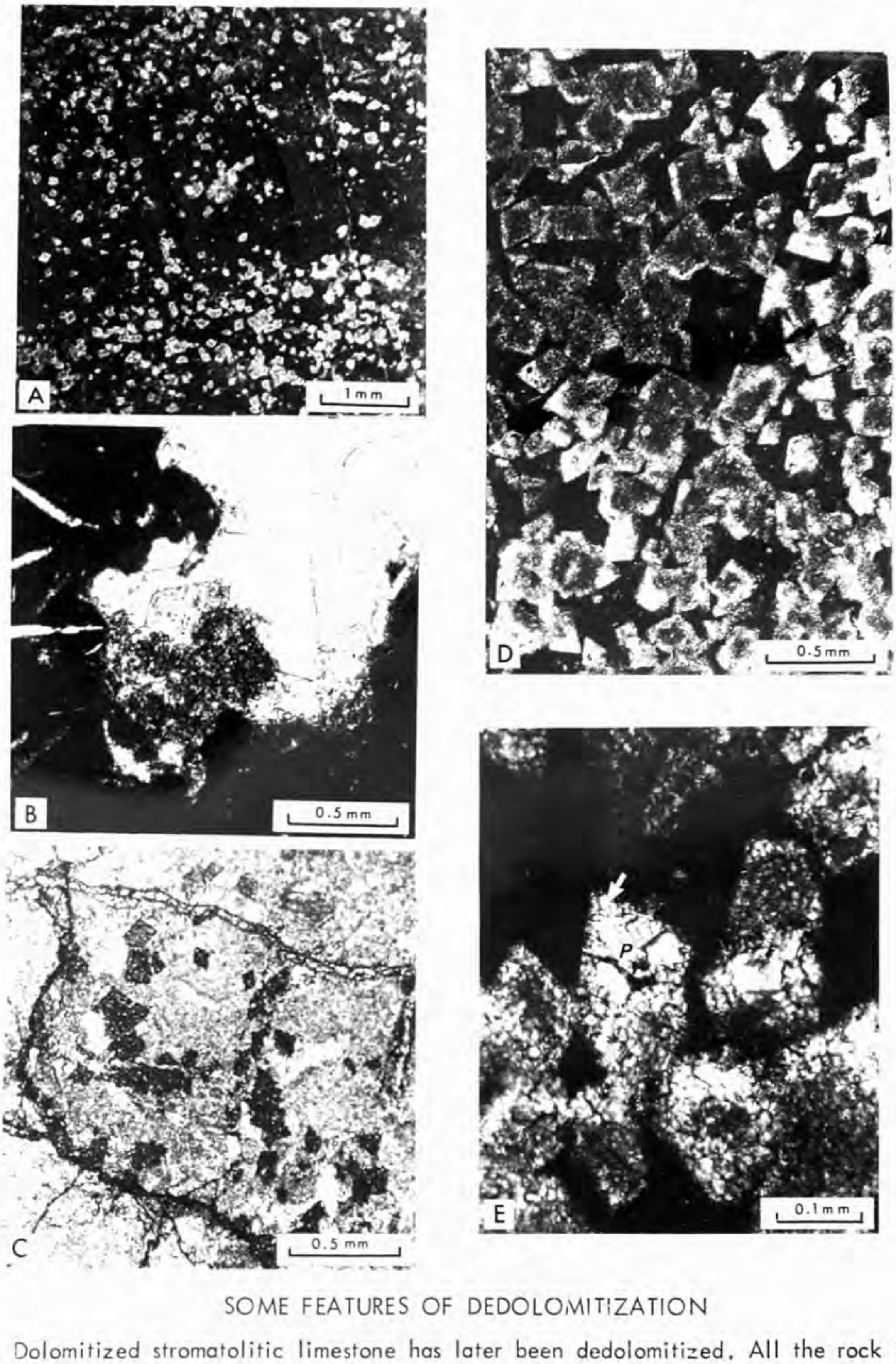
All the microphotographs shown in this plate are limestones after dolostones. It is beleived that they are formed by centripetal replacement (Fig.V-11). C, D, and E show ferric oxide ghosts of earlier dolomite rhombs.

A. Peel. Sample No.442 B. Stained thin section. Sample No.741

C. Stained thin section. Sample No.1006 D. Peel. Sample No.444

E. Stained thin section. Sample No.72/40; Location: Irmaç (North of Behram).

PLATE -69



SOME FEATURES OF DEDOLOMITIZATION

A. Dolomitized stromatolitic limestone has later been dedolomitized. All the rock is composed of calcite. Stained thin section. Sample No.707

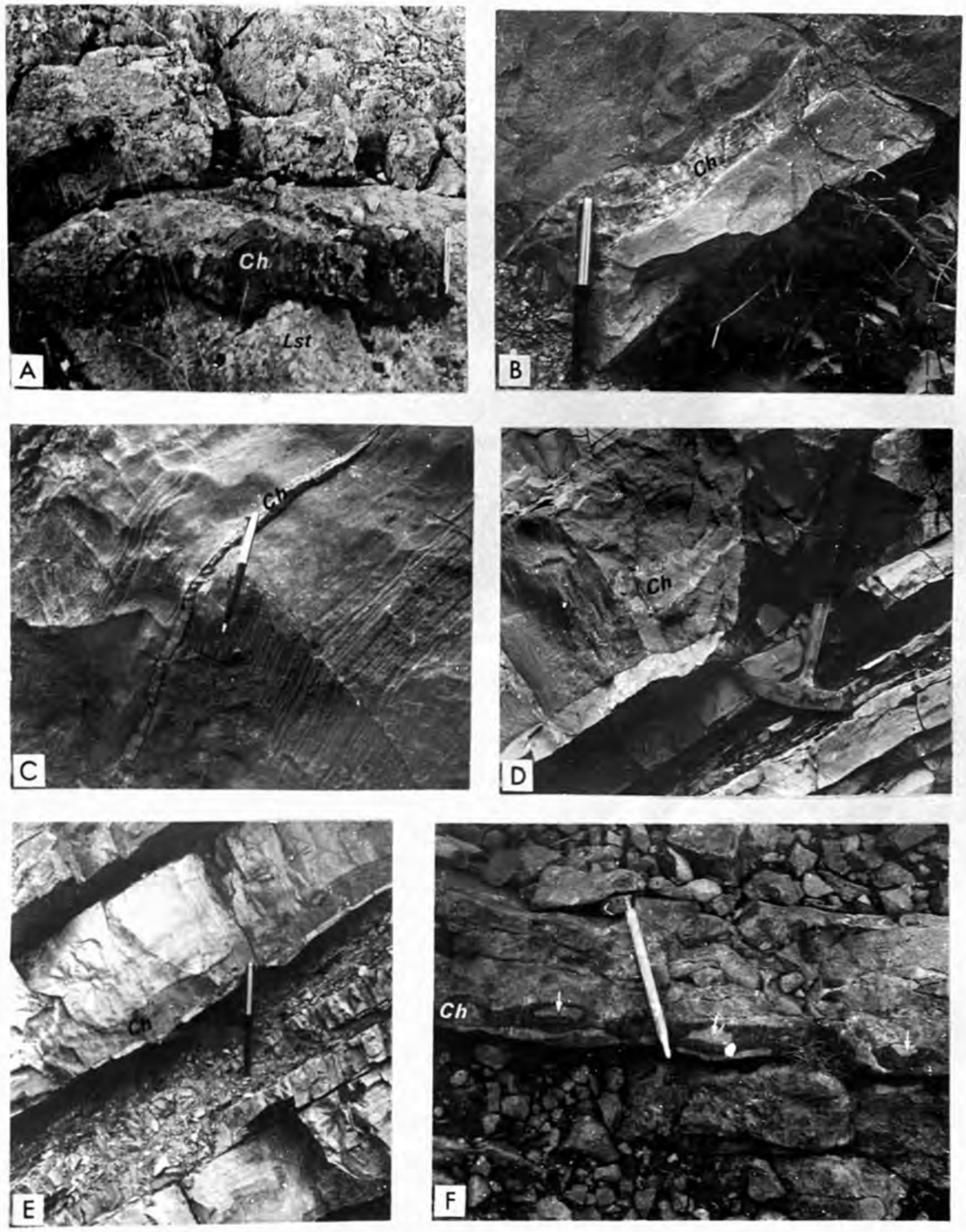
B. Replacement of dolomite rhombohedra replacing calcite cement. Stained thin section. Sample No.1152

C. Rhombohedra seen in this microphotograph are composed of iron rich calcite. Most probably they were formed by dedolomitization of iron rich dolomites. Peel. Sample No.4/11; Location: Kuz (N66).

D. Partially leached dedolomite rhombohedra. Later calcite cementation (white in micro-photograph has filled the voids. Stained thin section. Sample No.357B

E. Detail of E. Calcite cement (arrow) and pore space (P) are visible.

PLATE - 70

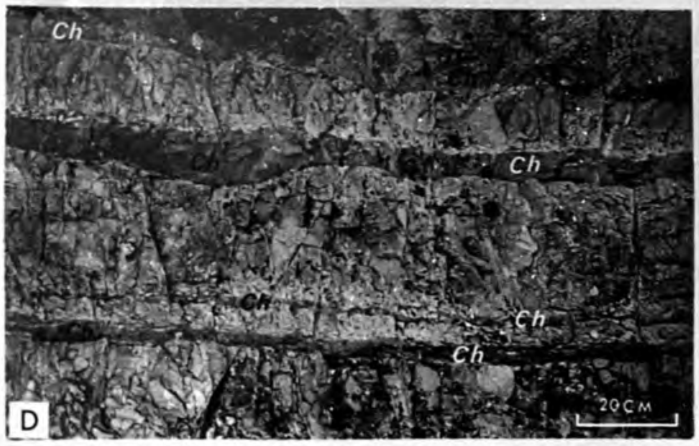


OCCURENCE OF CHERT- 1

A. Chert associated with shelf carbonates. Sample No.663

B, C, D, and E. Bands and lenticular bodies of granular chert (Ch) in calciturbidite beds. Locality: Kaynarca (T29)

F. Band of vitreous chert (Ch). Note the occurence of unchertified areas within the bands (arrows). Sample No.510



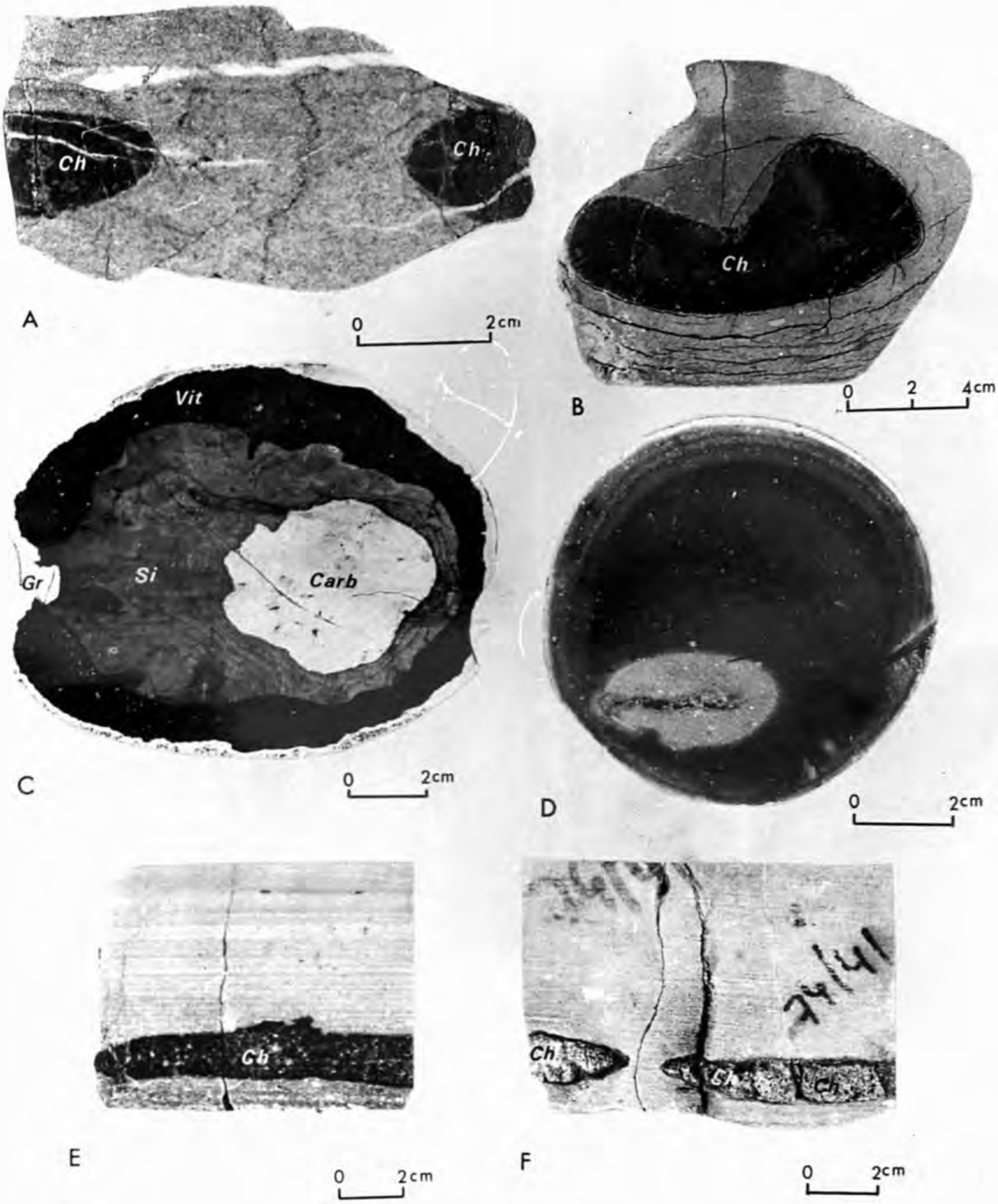
CCCURENCE OF CHERT-2

A and B. Bands of granular chert (Ch). Locality: Kaynarca (T29)

C. Chert nodule. Vitreous core (Vit) and granular rim (Gr) are visible. Note the sharp contact (arrow) between nodule and host rock. Sample No.912

D. Lateral continous "beds" of granular chert (Ch). Sample No.1071

PLATE - 72



MEGASCOPIC CHARACTERISTICS OF CHERT

A. Discontinous bands of granular chert (Ch). Polished surface. Sample No.1078

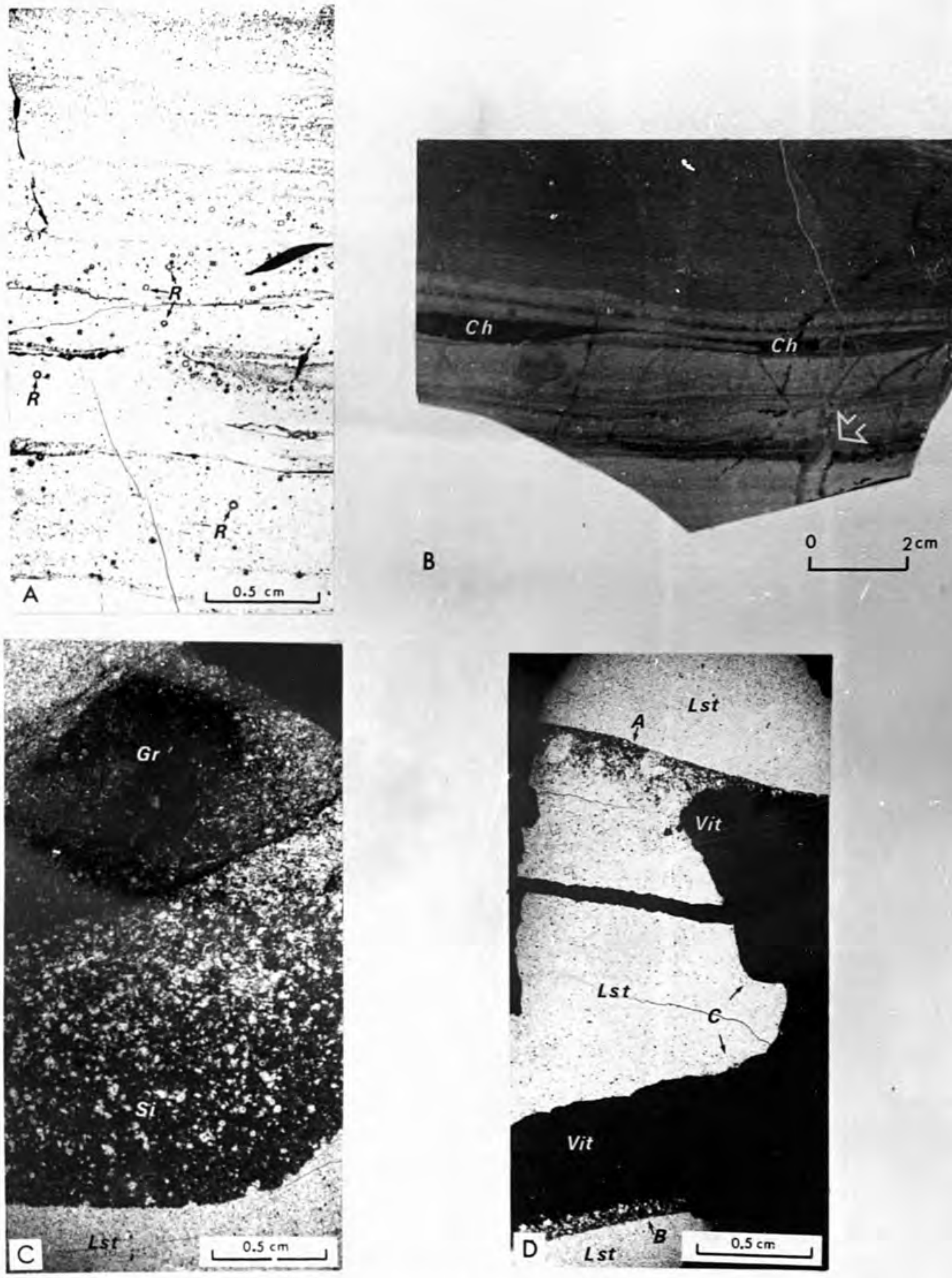
B. Lenticular granular chert (Ch) in a turbiditic layer. Polished surface. Location: Kaynarca.

C. Polished surface photograph of a "nodule" found in a river bed. Carb- Carbonate Core; Si- Silicified carbonate rock; Vit- Vitreous chert; Gr- Granular chert rim. Note the conformity of bands with core and cross-cutting with vitreous chert. This "nodule" is thought to be formed in a similar way to band of Plate 70 F. For thin section microphotograph see Plate 74 D. Sample No.74/9 Location: Abdalpinar Dere (Section 1).

D. Chert nodule showing parallel banding (Liesegang rings). The whitish part is due to a crack in the chert. Location: North of Camiiliyurt (P12).

E. Chert band in the base of a thin laminated calciturbidite bed. Polished surface. Sample No. 74/41; Location: Arpacukuru (K8).

F. Other surface of the sample at E. As seen on this photograph the chert band is not continous in three dimensions. The thickness of the sample is 5cm. Hand Specimen.



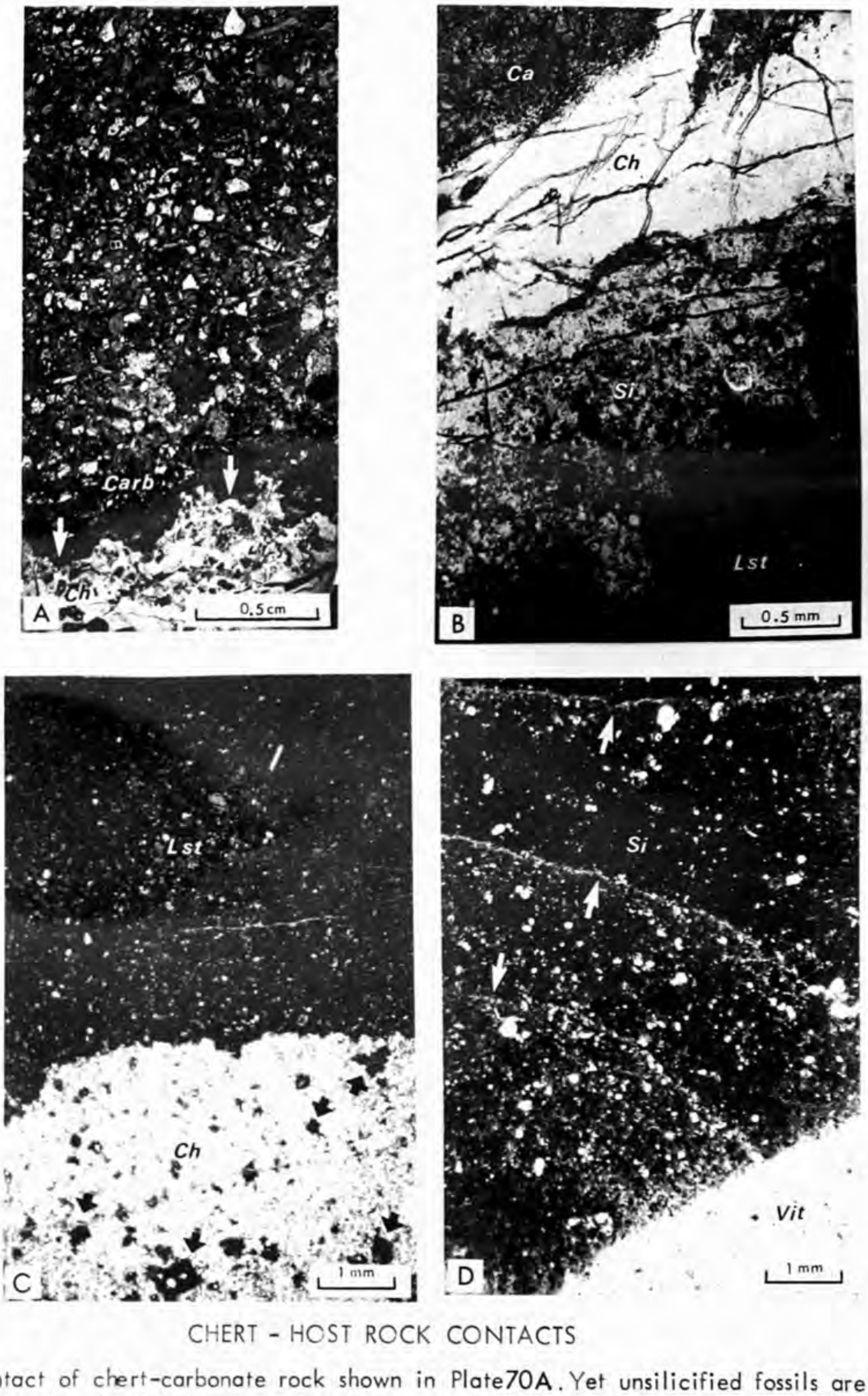
CHARACTERISTICS OF CHERT

A. Negative print of a jasper (ferruginous chert) thin section. It shows current activity. Radiolaria (R) abundant. Sample No.74/32; Location: Unknown.

B. Vitreous chert (Ch) in laminated calciturbidite layer. The feature at the right hand bottom (arrow) is due to biologic activity. Polished surface. Sample No.504.

C. Negative print of a stained thin section showing granular chert (Gr)-Limestone (Lst) contact. Transition zone is partially silicified limestone (si). Sample No.73/25; Location: Kızılören (L46)

D. Vitreous chert (Vit)-Limestone (Lst) contact. Negative print from a stained thin section. Note the transitional contact at C. C resembles silicification front. Sample No.509.



CHERT - HOST ROCK CONTACTS

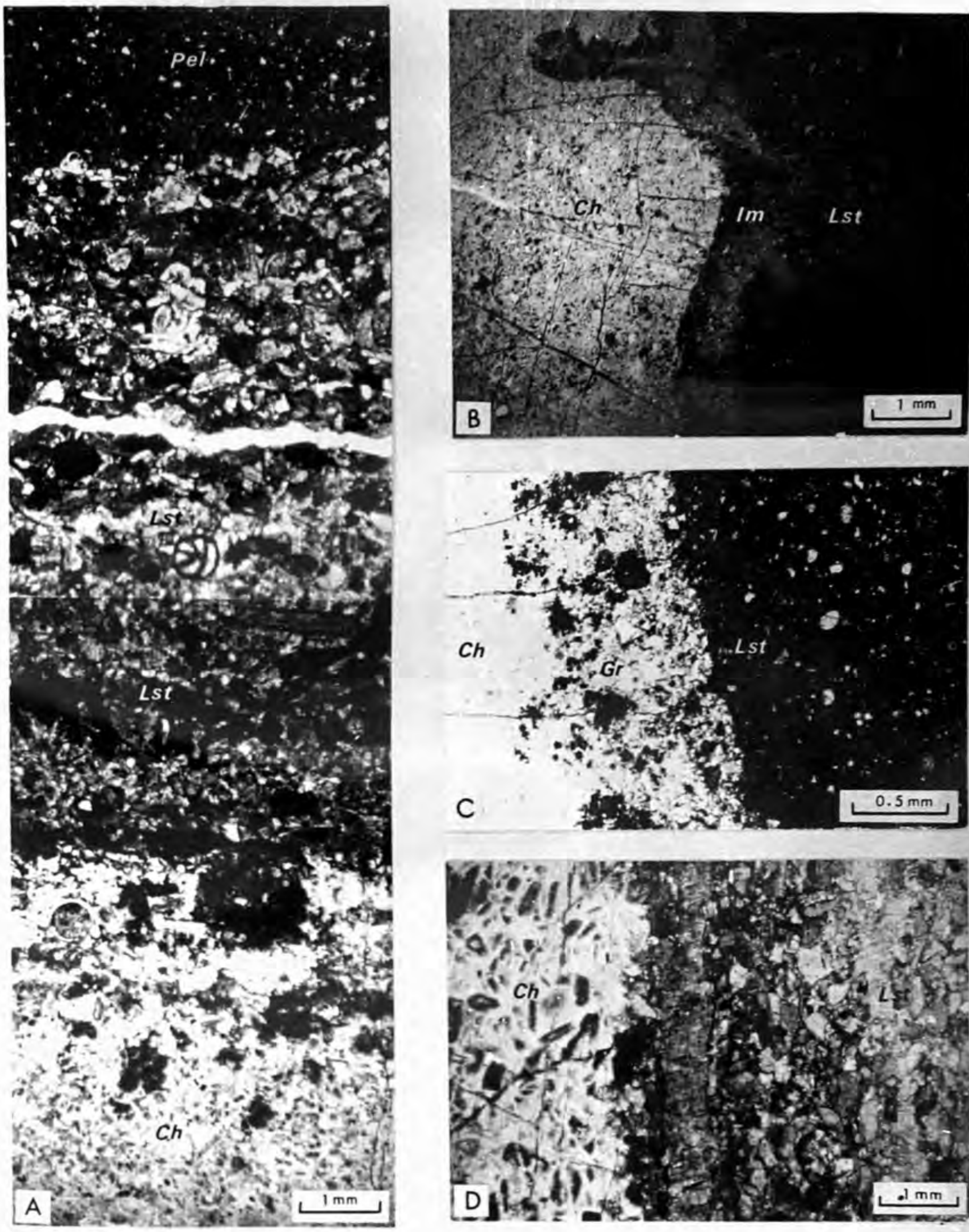
A. Contact of chert-carbonate rock shown in Plate70A. Yet unsilicified fossils are seen within the chertified part (Ch). Negative print from a thin section. Sample No.663

B. Partially chertified limestone; Ch-Chert; Ca-Calcite vein. Stained thin section. Sample No.72/22; Locality: Kuz (N66)

C. Some patches of pelagic limestone are preserved in the chert (arrows). Stained thin section. Sample No.73/25; Locality: Kızılören (L47)

D. Microphotograph of "nodule" in Plate72C. Vit-Vitreous chert; Si-Partially silicified carbonate rock. Cross-cutting relations of the Liesegang rings (arrows) are well seen. Stained thin section. Sample No.74/9 Locality: Abdalpınar Dere (Section 9).

PLATE -75



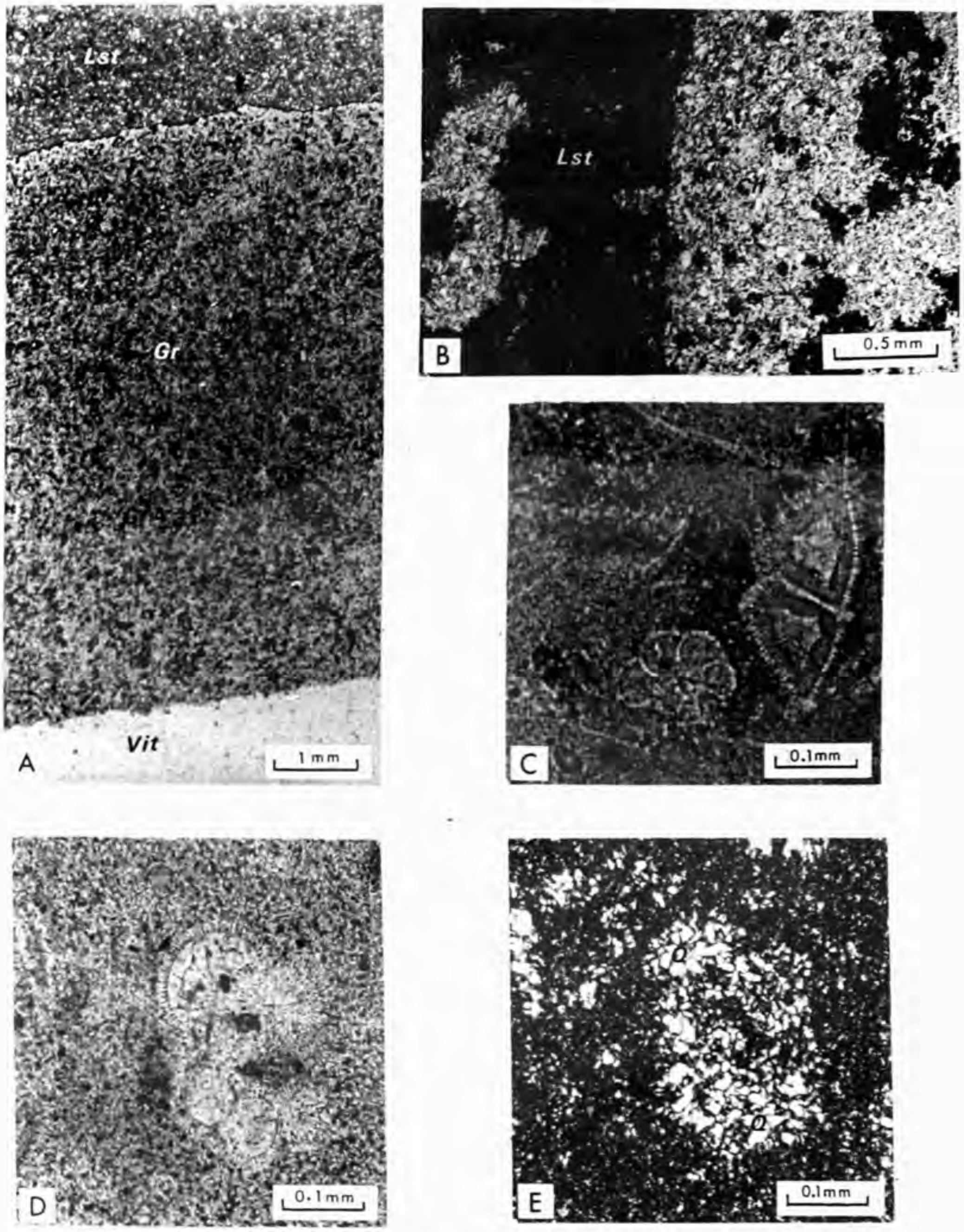
CHERT - HOST ROCK RELATIONS

A. Contact between granular chert (Ch), fossiliferous limestone (Lst) and pelagic limestone (Pel). Sample No.925/3

B. Zone of impurities (Im) marks the transition between chert (Ch) and host rock (Lst). Stained thin section. Sample No.904/4

C. Boundary between rim and calcareous sediment at the periphery of a vitreous chert nodule. Stained thin section. Sample No.509

D. Granular chert (on the left) and fossiliferous lomestone (on the right) contact. Some remnants of calcitic skeletons are preserved within the chert. Stained thin section. Sample No.1077



MICROSCOPIC CHARACTERISTICS OF CHERT

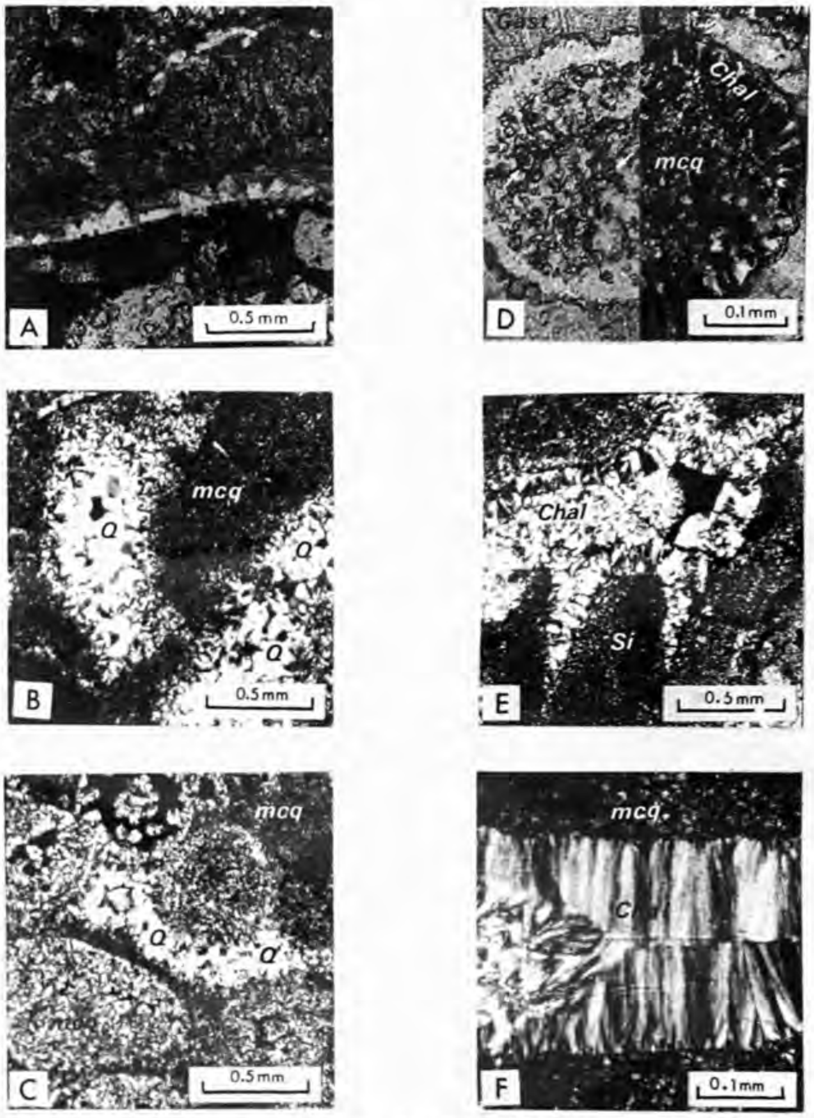
A. Contact between vitreous chert (Vit) and limestone (Lst). Transition zone is composed of granular chert (Gr). Stained thin section. Sample No.73/2; Locality: Tersakan (N7).

B. Some patches of host rock (dark colour) are partially silicified (light colour) while some patches are preserved within the chert. Stained thin section. Sample No.504D

C. Some remnants of calcitic foraminifera tests in vitreous chert. Chalcedony is restricted to tests. Chambers are filled with megaquartz. Detail structure is still barely visible in plane polarized light photographs. Only megaquartz indicates the presence of foraminifera ghosts in crossed-nicols photograph (See Plate79A). Sample No.354

D. Same as C. Sample No.354

E. Same area in D. Crossed-Nicols.



CAVITY FILLING SILICA FABRICS

A. Quartz crystals seem to grow from the wall towards the inside. Process may be due to solution of calcite and precipitation of silica in pH controlled micro-environment. Stained thin section. Sample No.72/5; Location: Hamalcay (H75)

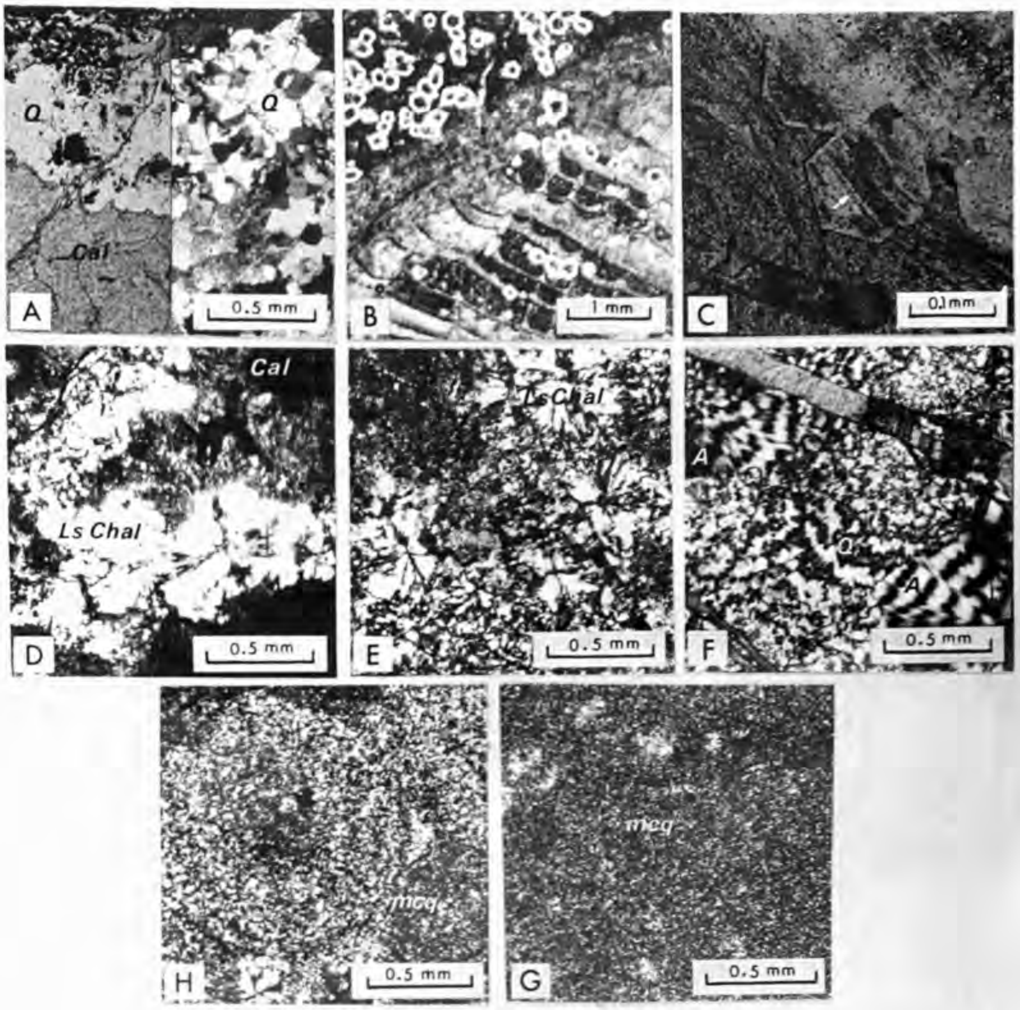
B. Megaquartz filling pore spaces between silicified allochems. Q - Megaquartz; mcq-Microcrystalline quartz. Stained thin section. Crossed Nicols. Sample No.73/6 Location: Arpacukuru (K8).

C. Same as B. Sample No..54/20; Location: Kuz (N66)

D. Chalcedonic crust (Chal) lining gastropod (Gast) chamber. Rest of the chamber is composed of microcrystalline quartz (mcq) and calcite inclusions (arrows). Thin section. Left: Plain polarized light; Right: Crossed Nicols. Sample No.663

E. Chalcedonic crust (Chal) lining the pore space between silicified allochems (Si). Stained thin section. Crossed Nicols. Sample No.906.

F. Chalcedony (Chal) filling a vein in microcrystalline quartz (mcq) matrix. Thin section. Crossed Nicols. Sample No.357B.



REPLACEMENT SILICA FABRICS

A. Megaquartz (Q) replacing calcitic shell. Cross-cutting relations and calcite inclusions in megaquartz are strong evidence of replacement. Thin section. Left: Plain polarized light; Right: Crossed Nicols. Sample No.Kuz 3; Location: Kuz (N66)

B. Quartz euhedrons replacing a rudist fragment and matrix. Stained thin section. Sample No.73/55; Location: Kale Dere.

C. Detail of qurtz euhedron replacing shell fragment. Hexagonal outlines of euhedron is well developed. Note the shell structure preserved in quartz euhedron as inclusions (arrow). Stained thin section. Sample No.347

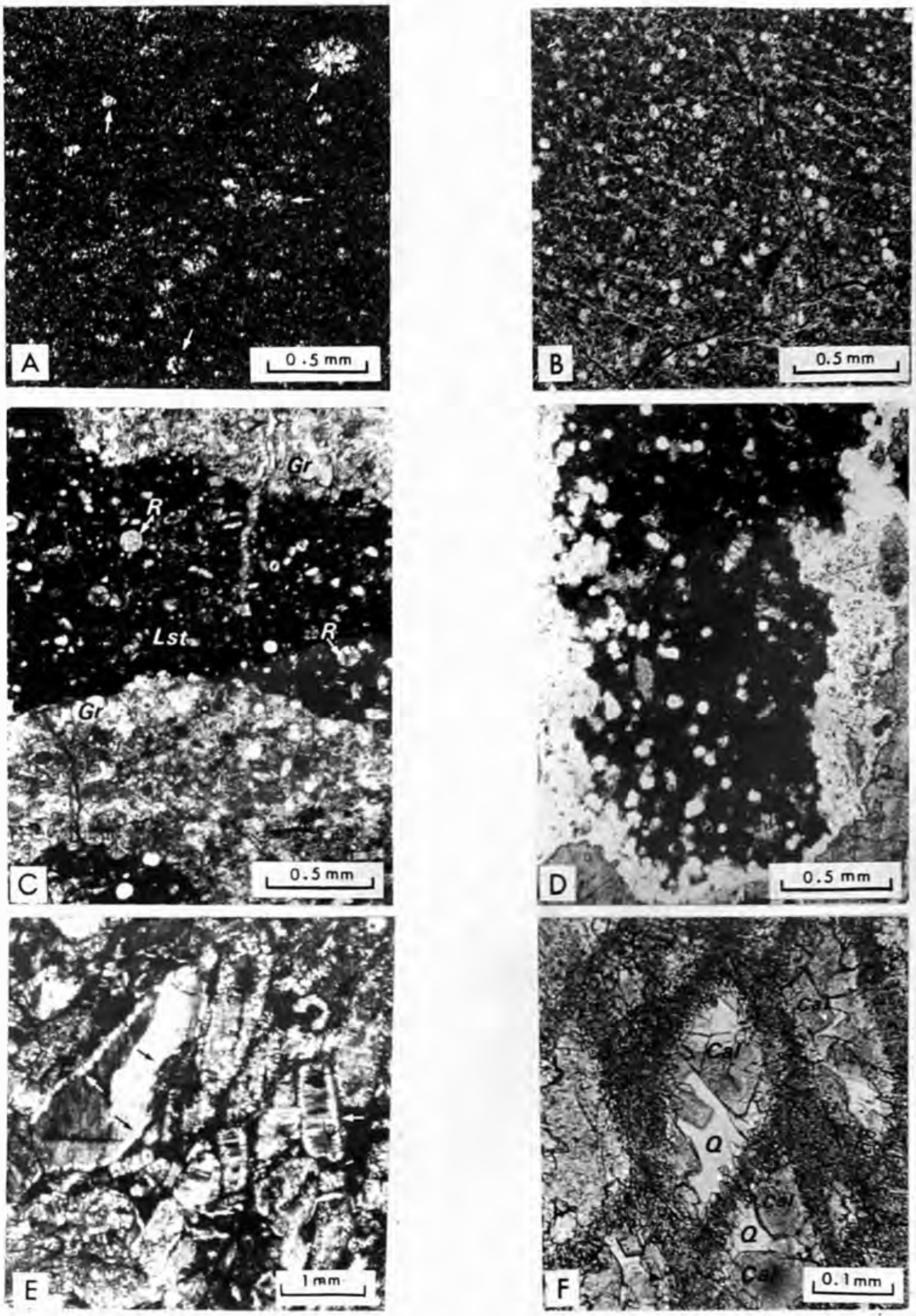
D. Spheroidal length-slow chalcedony (Ls Chal) replacing a shell. Stained thin section. Crossed Nicols. Sample No.501

E. Spherulites of length-slow chalcedony forming a replacement of micritic limestone. Stained thin section. Crossed Nicols. Sample No.Kö 1; Location: Kızılören (L47)

F. Spherulites of length-slow chalcedony (A) are grown at the expense of megaquartz (Q). Note the well developed outlines of spherulites. Thin section. Crossed Nicols. Sample No.72/32; Location: Irmaç (North of Behram).

G. Microcrystalline quartz (mcq) replaces both fossils and matrix. Larger crystals indicate the presence of former fossils. Thin section. Crossed Nicols. Sample No.73/8; Location: Bozhüyük.

H. Silicified oölite. All the sample is replaced by microcrystalline quartz (mcq). Nuclei and consantric structure of oölite is barely visible (arrow). Thin section. Crossed Nicols. Sample No.73/29; Location: Böğrüdelik.



GENERAL FEATURES OF SILICIFICATION

A. Microphotograph of a vitreous chert. All the texture of host rock has been obliterated. Only former presence of fossils is evident (arrows). Thin section. Crossed Nicols. Sample No.338B

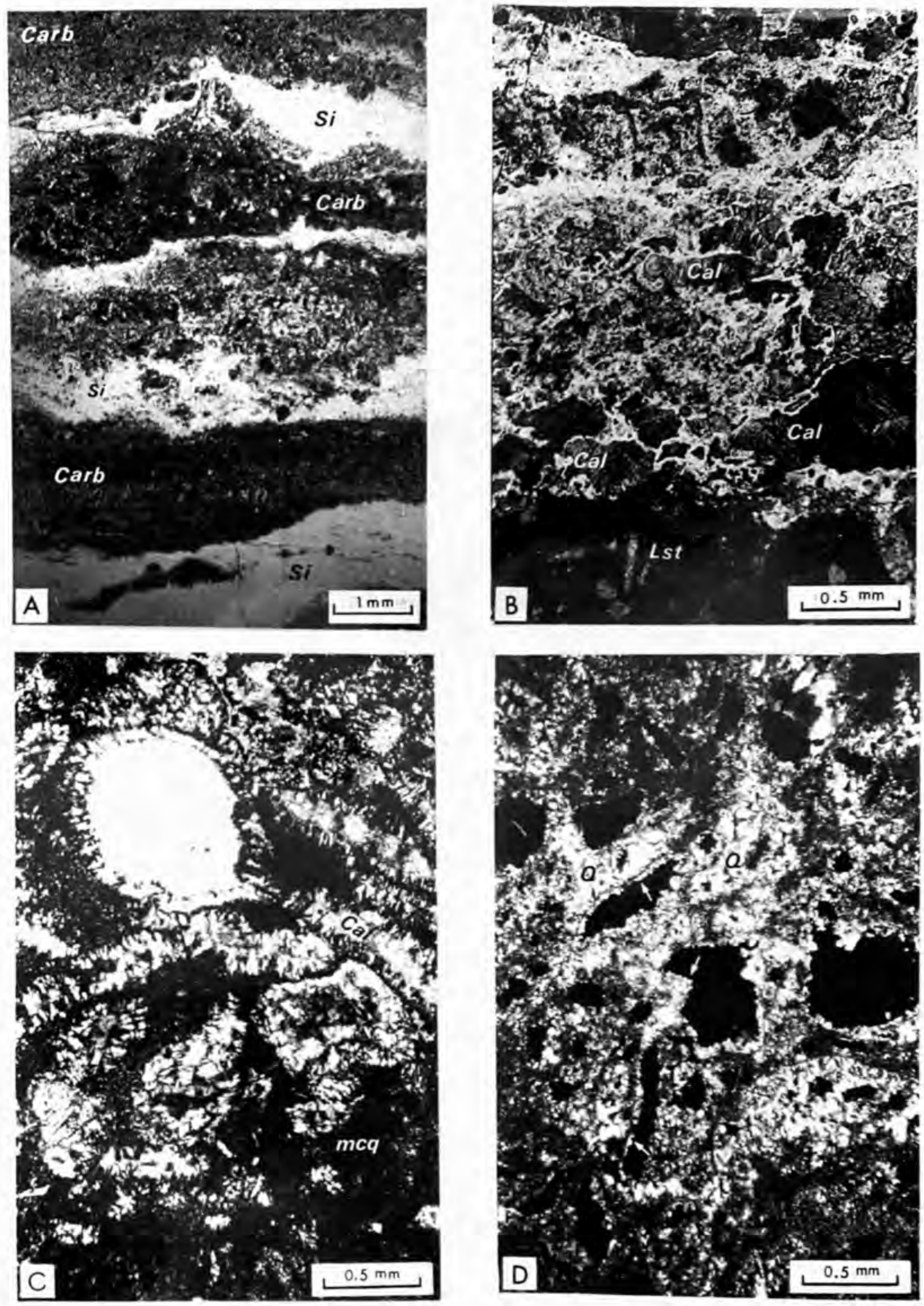
B. Chert with abundant radiolaria. Thin section. Crossed Nicols. Sample No.1078

C. Granular chert (Gr) and host rock (Lst) contact. Probably dissolution of some radiolaria (R) seen in host rock, now filled with calcite, is source of silica. Stained thin section. Sample No.72/22; Location: Kuz (N66).

D. Silicification of lime mud. Stained thin section. Sample No.663.

E. Peripheral replacement (arrows) of fossil fragments. Matrix is microcrystalline quartz. Thin section. Crossed Nicols. Sample No.663

F. Possibly secondary calcite (Cal) recrystallization in silica filled rudist cell. Secondary calcite is iron rich and seems to grow from the walls the inside and may partially replace megaquartz (Q) filling. Stained thin section. Sample No.1078.



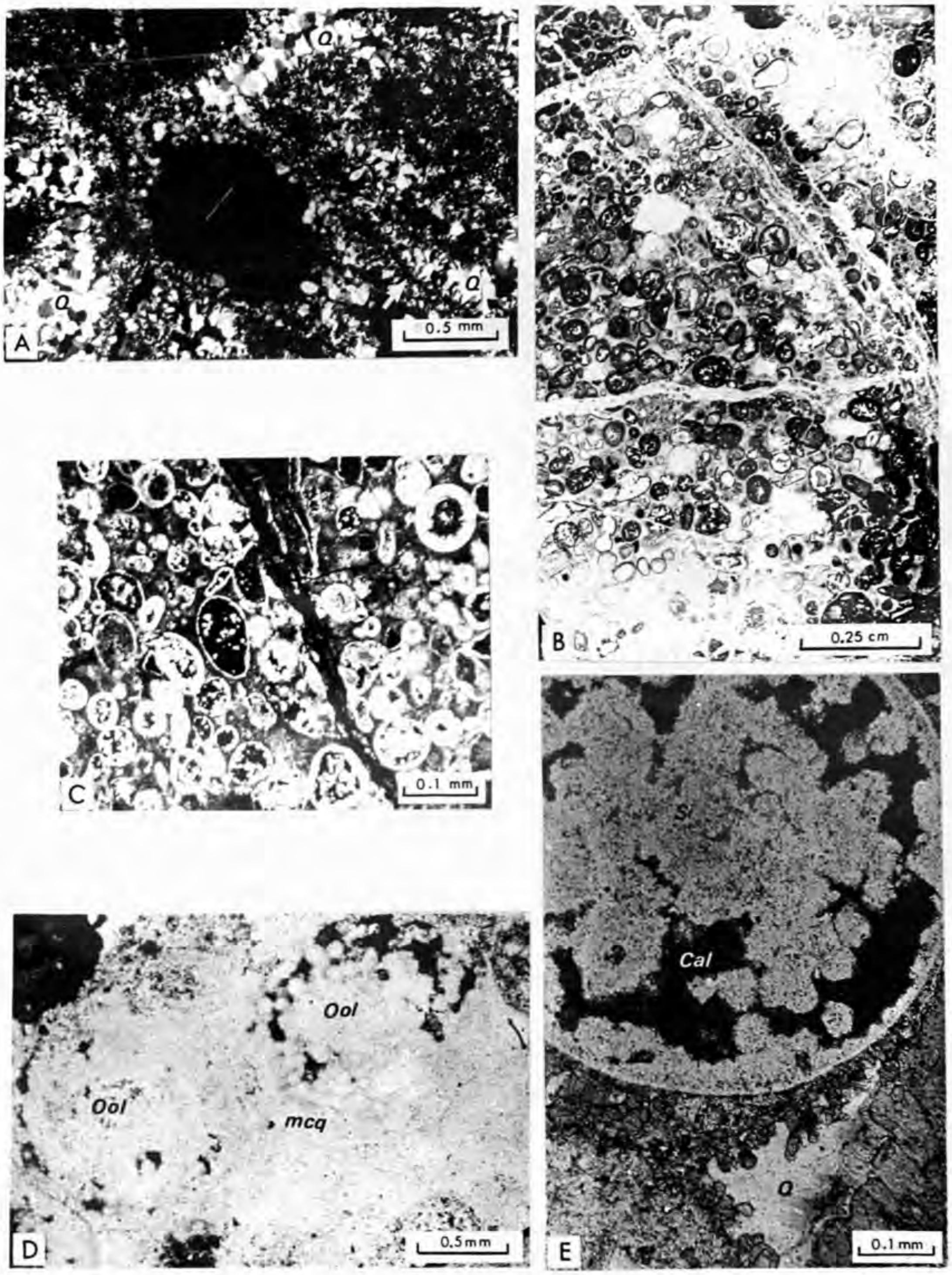
EXAMPLES OF SMALL SCALE SILICA REPLACEMENT-1

A. Partially silicified (Si) stromatolitic structure (Carb). Stained thin section. Sample No.74/I (Section: 1)

B. Silica replacement of strained calcite (cal) by penetration along the crystal boundaries. Stained thin section. Sample No.101

C. Advanced peripheral replacement. A small part of the shells (Cal) has been escaped from silicification. Matrix is microcrystalline quartz (mcq). Thin section. Crossed Nicols. Sample No.601C

D. Leaching of calcite has caused the voids which are lined by chalcedonic crust (arrows). Megaquartz is probably cement in origin. Thin section. Crossed Nicols. Sample No.642C



EXAMPLES OF SMALL-SCALE SILICA REPLACEMENT-2

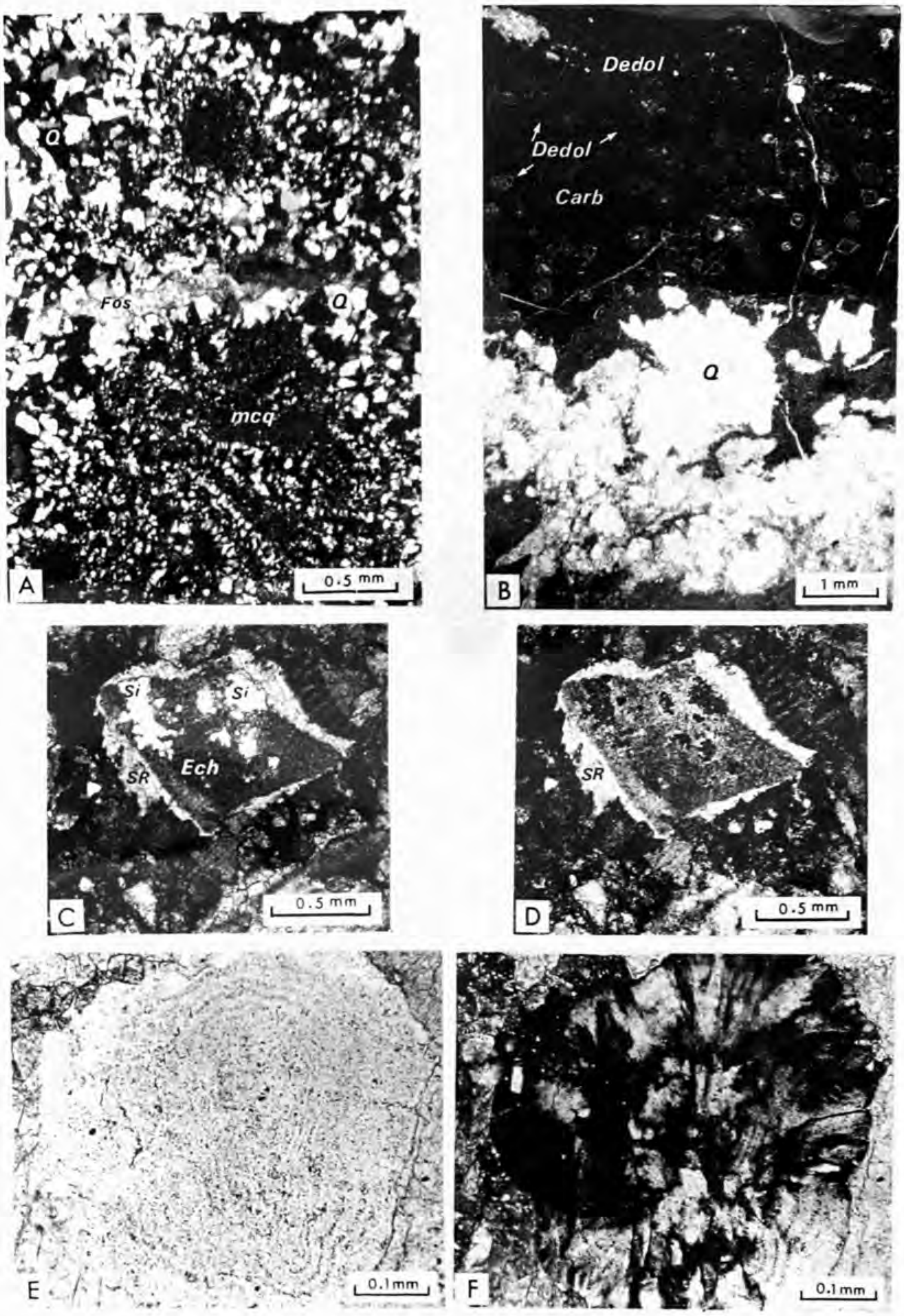
A. Same as Plate 80D. Some traces of former micritic envelope can be seen (arrow). Thin section. Crossed Nicols. Sample No.642C

B. Negative print of an oolitic pebble thin section. Sample No.S4/20; Locality: Kuz.

C. Detail of part of B. Thin section. Plain polarized light.

D. Advanced silicification almost entirely obliterates internal structure of oölite. Thin section. Sample No.73/29; Location: Böğrüdelik.

E. Detail of partially silicified oölite. Si-Silica; Ca-Calcite. Megaquartz (Q) rims the allochems, and intergranular pore space. Thin section. Sample No.S4/20; Kuz (N66)



EXAMPLES OF SMALL-SCALE SILICA REPLACEMENT-3

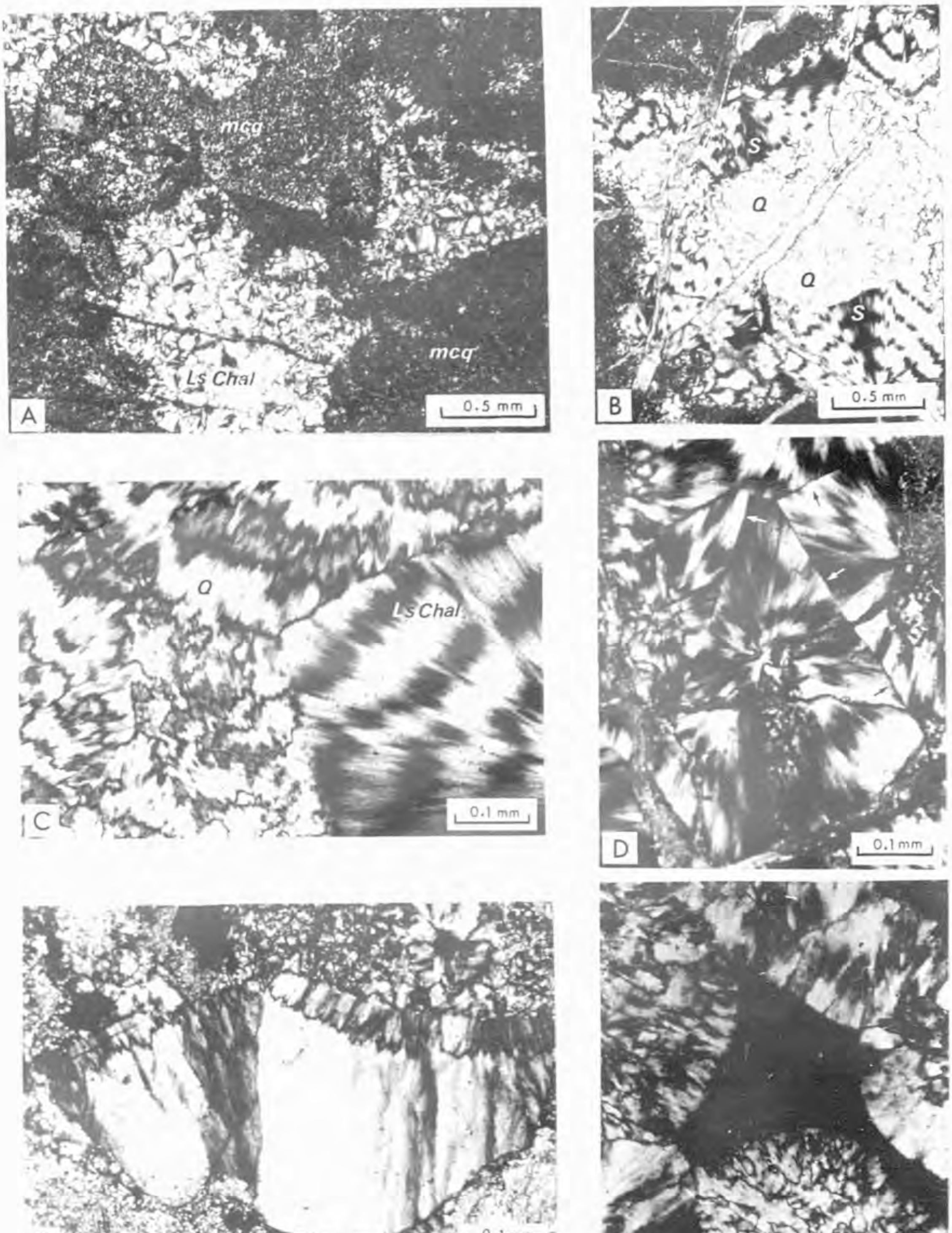
A. Replacement of a fossiliferous micrite by megaquartz (Q) and microcrstalline quartz (mcq). Excluding the one in the centre (Fos) of photo all fossils have been replaced. Thin section. Crossed Nicols. Sample No.Kuz 3; Location: Kuz (N66).

B. Megaqurtz (Q) replacing a dedolomitized limestone. Thin Section. Sample No.74/25 Location: Tahtayurdu Dere (Section 11A).

C. Replacement of an echinoid plate and its syntaxial rim (SR) by silica (Si). Stained thin section. Sample No.908. D - Same as C. Crossed Nicols.

E. Colloform banding seen a spherulitic chalcedony. Thin Section. Sample No.355

F. Same as E. Crossed Nicols.



SOME CHARACTERISTICS OF CHALCEDONY

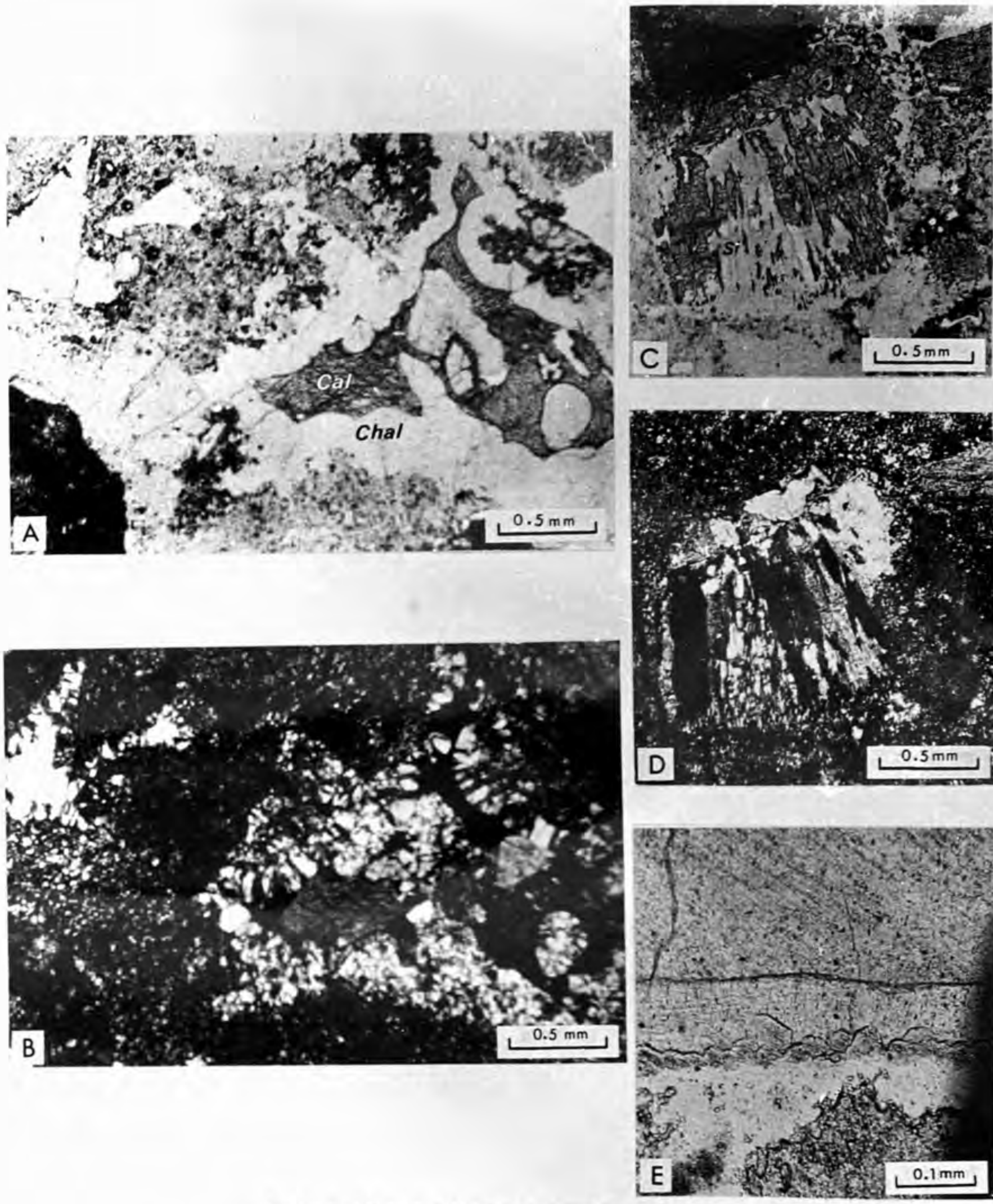
A. Allochems are completely silicified (mcq) and matrix has been replaced by spherulites of length-slow chalcedony (Ls Chal). Thin section. Crossed Nicols. Sample No.906

B. Spherulites of length-slow chalcedony (S) have grown at the expense of megaquartz (Q). Thin section. Crossed Nicols. Sample No.72/32; Location: Irmaç.

C. Detail of some part of B. Zoned extinction is due to outward spiral twisting of the spherulite fibers (SIEDLECKA, 1972).

D. Straght boundaries (arrows) between spherulites. Thin section. Crossed Nicols. Sample No.72/32; Location: Irmaç (North of Behram).

F. Chalcedonic crust lining allochems (Detail of Plate 77E). Intergranular pore space is still preserved. Stained thin section. Crossed Nicols. Sample No.906



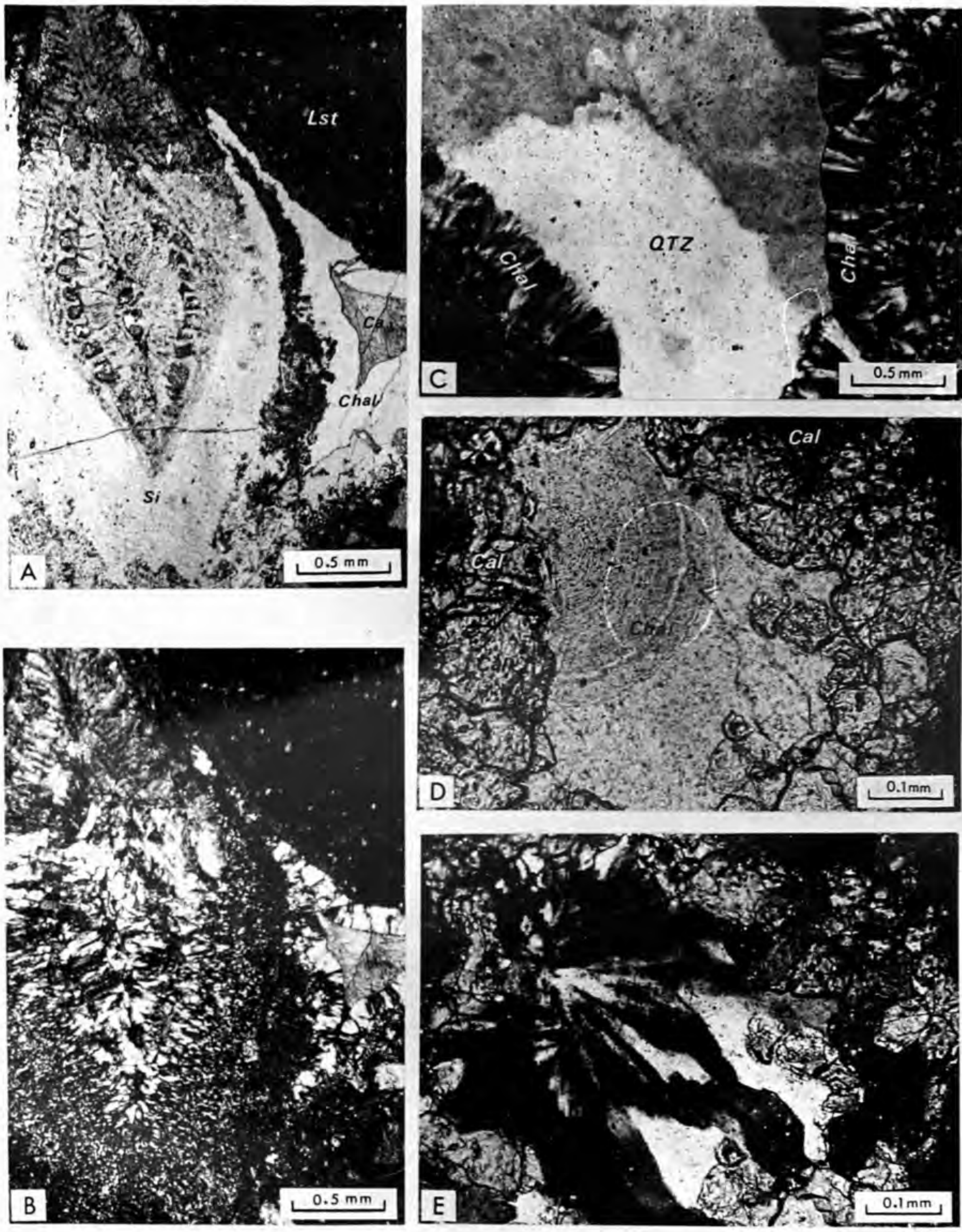
SMALL-SCALE SILICA FABRICS

A. Chalcedonic crust (Chal) lining and partially filling pore spaces. Secondary iron rich calcite (cal) fills the rest of the void. Stained thin section. Sample No.906

B. Same as A. Crossed Nicols.

C. A prismatic molluscan shell has been partially replaced by silica. Stained thin section. Sample No.905/2   D - Same as C. Crossed Nicols.

E. Silicified molluscan shell. Sample No.663.



VARIOUS SMALL-SCALE SILICA FABRICS

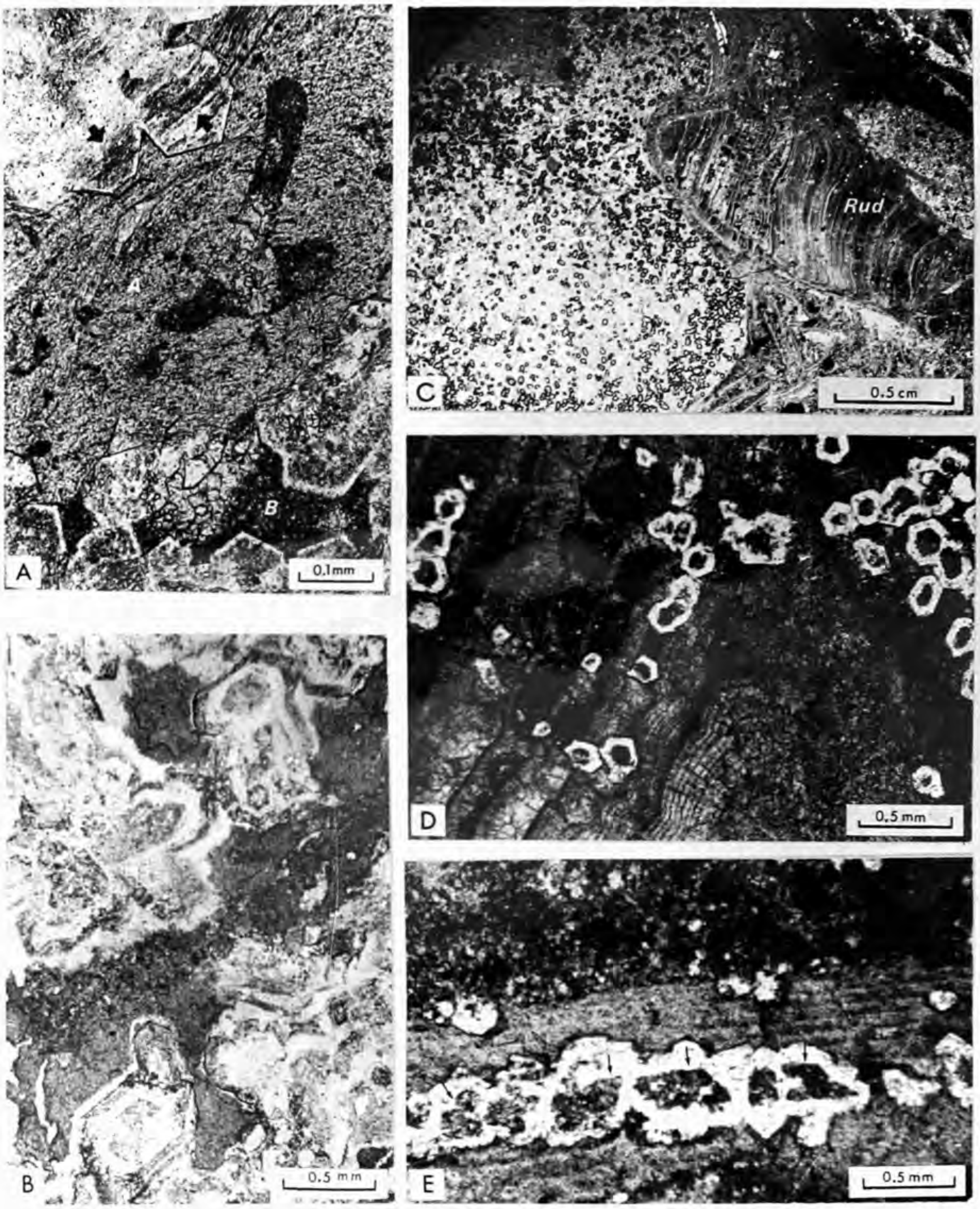
A. Partial silicification of a large foram. Arrows show boundary between silicified and unsilicified parts. Chalcedonic crust (Chal) lines the pore space. Secondary iron rich calcite obliterates the porosity. Stained thin section. Sample No.906

B. Same as A. Crossed Nicols.

C. A pore space is lined by chalcedonic crust (Chal) which has later been filled with megaquartz (Q). Stained thin section. Sample No.906

D. Chalcedonic quartz (Chal) showing colloform banding fills the cavity left by cementation (Cal). Stained thin section. Sample No.853B

E. Same as D. Crossed Nicols.



QUARTZ EUHEDRON

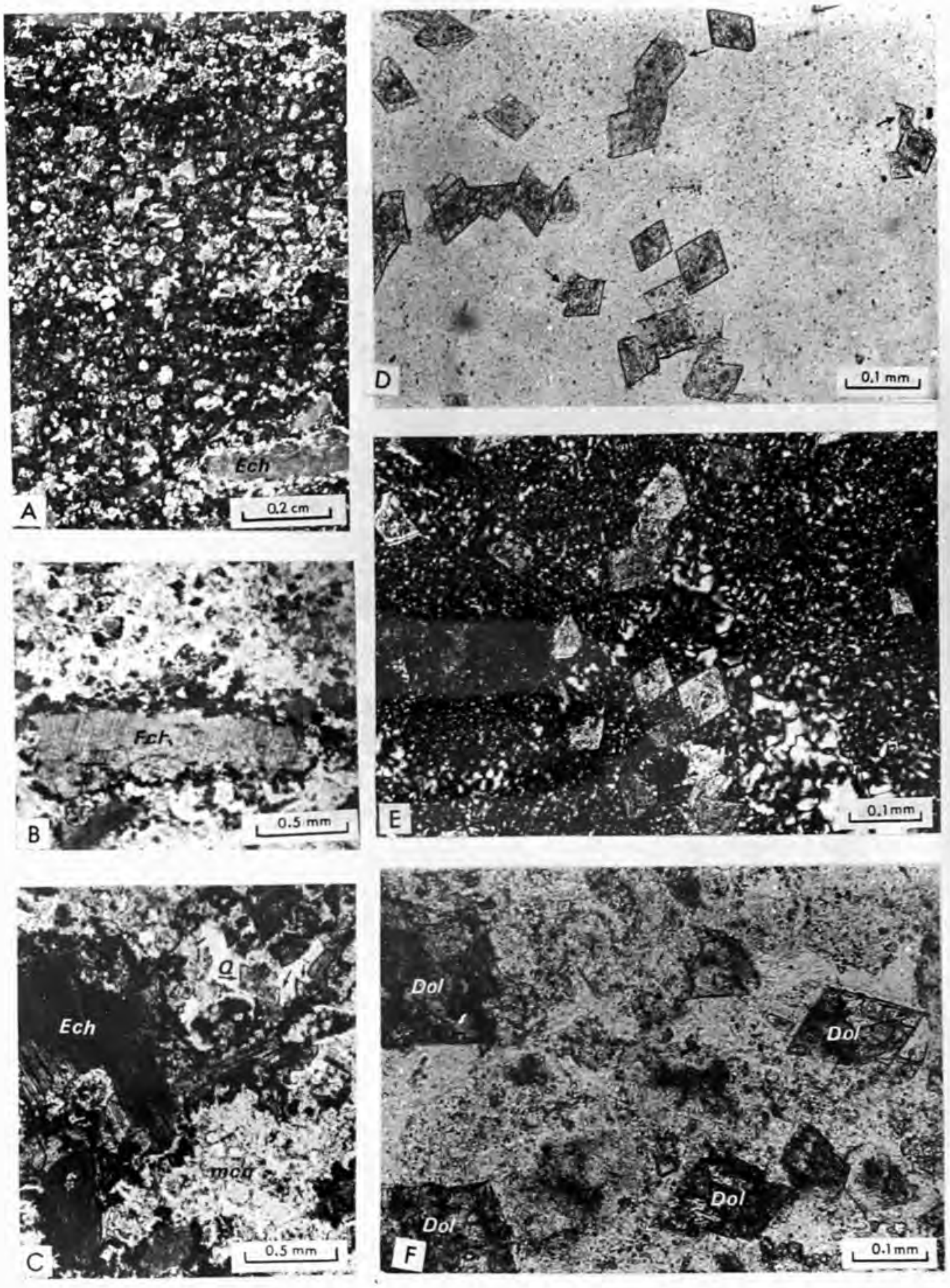
A. Quartz euhedrons have replaced both shell (A) and micritic matrix (B). Structure of shell is preserved as inclusions in quartz euhedrons. Thin section. Sample No.347

B. Perfect crystal shape and zonation of quartz euhedrons. Stained thin section. Sample No.73/55 Location: Kale Dere.

C. Negative print of a thin section showing replacement of a rudist (Rud) fragment and matrix by quartz euhedrons. Stained thin section. Sample No.73/55; Location: Kale Dere. D - Detail of C. Thin section.

E. Replacement of a shell by quartz euhedrons. Original shell structure is preserved as centers of euhedrons (arrows). Stained thin section. Sample No.73/55; Location: Kale Dere.

PLATE - 87



DOLOMITE IN CHERT

A. Negative print of an echonoid rich chert thin section. Dolomites (white in photo) are usually associated with echinoid plates (Ech). Stained thin section. Sample No.72/32; Location: Erikli.

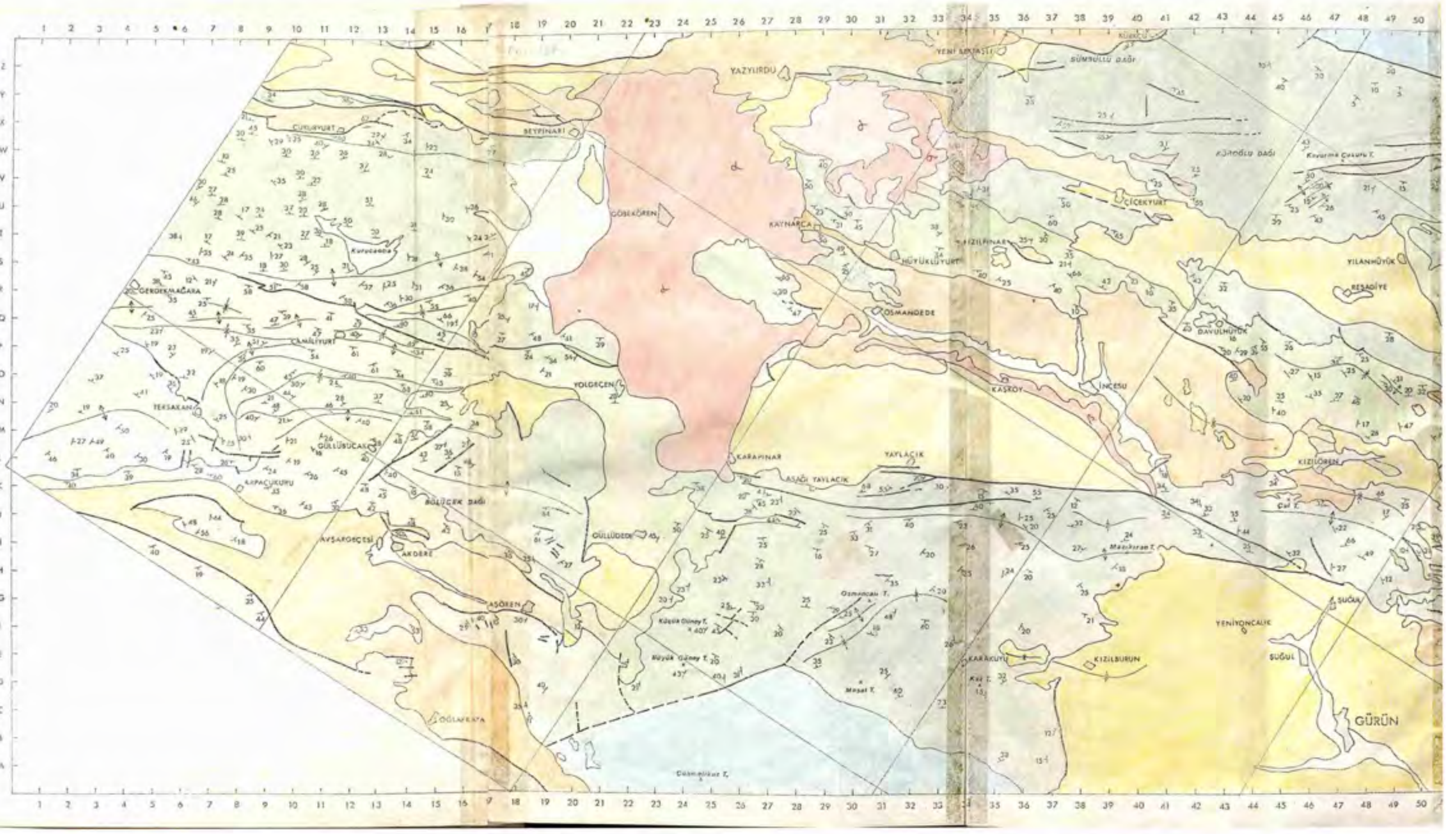
B. Thin section microphotograph of the large echinoid plate at the right hand bottom of A. Dolomites are black. Matrix is microcrystalline quartz. Stained thin section.

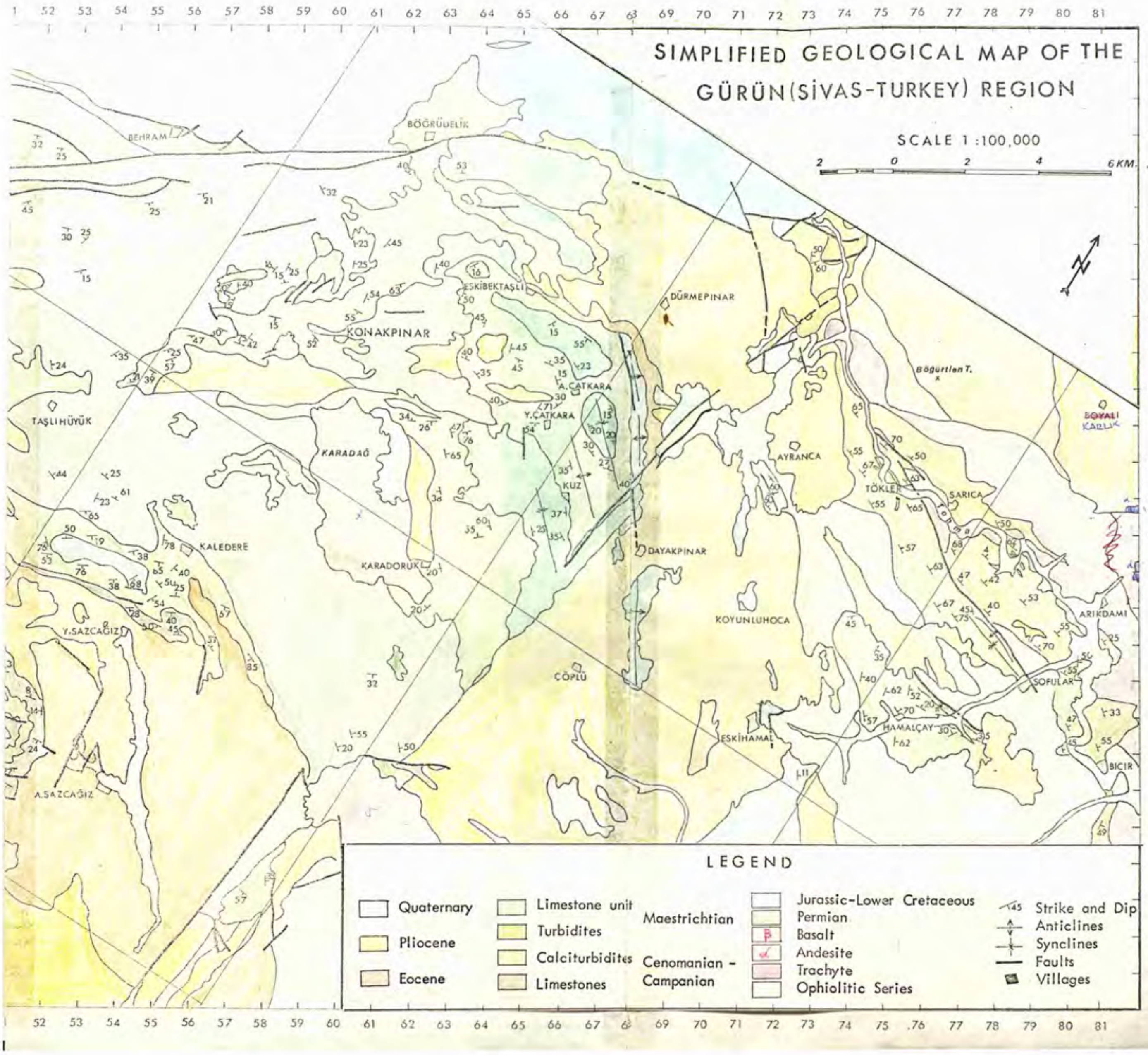
C. Detail of a part of the same thin section. Ech- EEchinoid plate; mcq- Microcrystalline quartz; Q-Megaquartz. Dolomites are black except the ones in the megaquartz which are shown by arrows.

D. Dolomite rhombs in a vitreous chert. Note the presence of some perfect rhombs and some corroded ones (arrows). Stained thin section. Sample No.73/6; Location: Arpacukuru (K8)

E. Same as D. Crossed Nicols.

F. Enlarged view of some dolomite rhombs (Dol) in echinoid rich chert. Stained thin section. Sample No.72/32; Locatio: Erikli.


YAZYURDU
YENİ BEKTAŞLI
SÜMBÜLLÜ DAĞI
KÖROĞLU DAĞI
ÇİÇEKYURT
ÇUKURYURT
BEYPINARI
GÖBEKÖREN
KAYNARCA
KIZILPINAR
HÜYÜKLÜYURT
OSMANDEDE
YILANHÜYÜK
REŞADİYE
DAVULHÜYÜK
KAŞKÖY
İNCESU
Kurucaova
GERDEKMAĞARA
ÇAMLIYURT
YOLGEÇEN
TEKSAKAN
GÜLLÜBUCAK
KAPAÇUKURU
BÖLÜCEK DAĞI
KARAPINAR
YAYLACIK
AŞAĞI TAYLACIK
KIZILÖREN
Çal T.
GÜLLÜDERE
AVŞAROBÇESİ
AKDERE
AŞÖREN
Mazıkıran T.
Osmancalı T.
Küçük Güney T.
Büyük Güney T.
SUÇAT
YENİYONCALIK
KARAKUYU
KIZILBURUN
SUĞUL
Mezar T.
Kel T.
OĞLAKKAYA
GÜRÜN


SIMPLIFIED GEOLOGICAL MAP OF THE GÜRÜN(SİVAS-TURKEY) REGION
SCALE 1 :100,000
2 0 2 4 6 KM.
LEGEND
Quaternary
Pliocene
Eocene
Limestone unit
Turbidites
Maestrichtian
Calciturbidites
Limestones
Cenomanian - Campanian
Jurassic-Lower Cretaceous
Permian
Basalt
Andesite
Trachyte
Ophiolitic Series
Strike and Dip
Anticlines
Synclines
Faults
Villages
KONAKPINAR
TAŞLIHÜYÜK
KARADAĞ
KALEDERE
Y.SAZCAĞIZ
A.SAZCAĞIZ
KARADORUK
BÖĞRÜDELİK
BEHRAM
ESKİBEKTAŞLI
A.ÇATKARA
Y.ÇATKARA
KUZ
DAYAKPINAR
DÜRMEPINAR
ÇÖPLÜ
AYRANCA
KOYUNLUHOCA
ESKİHAMAL
HAMALÇAY
TÖKLER
SARICA
SOFULAR
ARIKDAMI
BICIR
Böğürtlen T.