Facies and sequence stratigraphy of the tamet formation (middle eocene), eastern sirte basin, Libya

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FACIES AND SEQUENCE STRATIGRAPHY OF THE TAMET FORMATION (MIDDLE EOCENE), EASTERN SIRTE BASIN, LIBYA

By
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DECLARATION

This to certify that the work submitted for the degree of master of science under title of "Facies and sequence stratigraphy of Tamet Formation (Middle Eocene) eastern Sirte Basin, Libya" is the result of original work. No part of this thesis has been accepted in substance for any other degree and is not currently being submitted in candidature for any other degree.

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ABSTRACT

The middle Eocene Tamet Formation on the eastern side of the Sirte Basin, is largely shallow-marine carbonate platform-to-basin transition formed along passive margin. The stratigraphy and deposition of this formation were largely controlled by eustatic processes of superimposed short-term and long-term of relative sea-level fluctuations.

Tamet sediments were mostly deposited in subtidal environments, which ranged from above fairweather- to below storm wave-base. Intrinisic processes such as storm and wave redeposition and reworking may have acted to inhibit aggradation into the zone of peritidal sedimentation. Estimates of water depth during deposition range from a few metres to a few tens metres maximum. There are three major platformal facies associations in the transition from deep subtidal to hypersaline deposits. These facies associations are defined and interpreted on the basis of their constituent microfacies and depend on their palaeogeographic setting on the platform. The spatial distribution of the complete spectrum of the facies associations suggest that deposition took place under low-energy conditions, as a stacked prograding homoclinal ramp.

Ten microfacies types have been distinguished, and their vertical interrelationships reflect metre-scale, shallowing-upward subtidal cycles, which are considered as the basic building blocks of the Tamet ramp. Two different types of subtidal cycle have recognised in the study area. Open-marine subtidal cycles are present along the outer through inner ramp within the transgressive and most of the highstand deposits. They are characterised by relatively deep subtidal microfacies at the base, gradationally overlain by shallow subtidal microfacies. Hypersaline subtidal cycles are present only upon the Cyrenaica Platform and predominated during the late phase of highstand deposition; they are composed of dolomitised shallow subtidal microfacies, capped by anhydrite.

The middle Eocene across this area is not a single carbonate ramp but rather an amalgamation of stacked ramps. Facies associations and cycles within the Tamet Formation have allowed the recognition of three depositional sequences separated by stratigraphic transitional zones. Each sequence represents a prograded ramp. The development of a sequence framework is based on the metre-scale cycle architecture and maintaining microfacies interpretations. Most of the sequences are interpreted as transgressive-highstand deposits. Each transgressive ramp is typically characterised by an aggradational patterns of relatively deep subtidal mud-rich carbonates deposited in a catch-up depositional system and episodically affected by storm events. Away from the ramp-margin, the transgressive facies change and stratigraphically thin into lagoonal facies deposited under keep-up conditions. The subsequent highstand ramp begins with an aggradational geometry but finally shifts into a distinct progradational pattern. The highstand cycles cover a broader area than that occupied by the transgressive sediments and are made up of mud-poor packstones reflecting a keep-up depositional system. The Cyrenaica Platform at this time was occupied by a very shallow and hypersaline sea. Carbonate sedimentation was shut off and replaced by precipitation of shallow-water evaporites; associated with this was dolomitisation, marking the end of sequence.

Several lines of evidence suggest that the magnitude of the middle Eocene sea-level fluctuations on the eastern Sirte Basin were relatively low. First, if the magnitude of the oscillations had been greater, then the sea-level falls would potentially have lead to formation of major sequence boundaries on the Tamet sequence upper surfaces. Second, if the magnitude was larger, then the rapid sea-level rises would have caused drowning of the ramp or domination by "catch-up" style of subtidal deposits. However, the Haq *et al.* (1987) sea-level chart in some circumstances may require modification, at least in terms of magnitudes and eustatic sea-level rises.

Chapter 1

Introduction

1.1 Previous Work:

Middle Eocene deposition in the Sirte Basin in Libya has been the subject of considerable study in recent years because of the importance of sediments of this age as hydrocarbon reservoirs (Hamyuni *et al.*, 1984). The main tectonic evolution of the Sirte Basin has been reviewed by Conant and Goudariz (1967), Burke and Dewey (1974), Van Houten (1983), Anketell and Ghellali (1991) and Gumati and Nairn (1991). The palaeontology and biostratigraphy of the middle Eocene formations are well known only from outcrop studies of Barr and Berggren (1980) and Hammuda (1973).

El-Hawat *et al.* (1986) described the sedimentary and diagenetic characteristics of the cyclic middle Eocene Darnah Formation of Jabal Alkhdar, NE Libya. They proposed a tectono-eustatic mechanism with small amplitude sea-level oscillations as the most likely cause of the repetition of carbonate cycles.

In contrast to the studies of outcrop, the only comprehensive study based on subsurface data from the Nafoora-Augila oil fields in eastern Sirte Basin was carried out by Belazi (1987). He introduced formal formation names based on lithostratigraphic work in the subsurface type section. He focused on the sedimentology, diagenesis, stratigraphy and regional significance of the Tamet Formation and the overlying sedimentary rocks of the Augila Formation as hydrocarbon reservoirs. Also he gave some emphasis to the broader sequence stratigraphic framework in terms of the systematic facies changes within the cycle framework.

1.2 Aim of This Study

The objectives of this study are 1) to describe the microfacies and facies associations of the Tamet Formation, 2) to interpret the spatial and temporal distribution of metre-scale cycles in order to understand the sedimentary dynamics and evolution of the Tamet platform, and 3) to develop a sequence stratigraphic framework of the middle Eocene succession

1.3 Data sources and Techniques

The study area occurs towards the eastern margin of the Upper Cretaceous-Tertiary Sirte Basin, approximately between latitude 29°00′ and 29°50′ and longitude 20°50′ and 21°45′, running generally NW-SE and covering 2500 Km^2. The middle Eocene sediments have been penetrated by numerous petroleum exploration and development wells. Five wells form the basis of this study and constitute a transect across the region. At the present time there are few detailed subsurface studies in this area. For this study, information has also been drawn from published literature and from exposed sections(Jabal Alkhdar, NE Libya).

A total of nearly 200 thin-sections of cuttings were studied from the Tamet wells. The location of the thin sections is shown on logs of the wells in the appendix. Following analytical techniques proposed by Flügel (1982), thin sections of carbonate samples were studied in detail under the petrographic microscope and then were visually subdivided into preliminary microfacies on the basis of their petrographic properties such as the types of biota, texture, proportion of micrite matrix versus sparite cement, etc.

1.4 Regional Geology of The Eastern Sirte Basin

1.4.1 Tectonic framework

The Sirte Basin is located in the north-central part of Libya (Fig. 1.1). It extends from the Gulf of Sirte in the north to about 26° N in the south and stretches for nearly km in an east-west direction from the Hun Graben in the west to the Cyrenaica Platform in the east. The tectonic evolution of the Sirte Basin has been the subject of great debate although in the tectonic basin classification system of Kingston *et al.* (1983), the Sirte Basin has been classified as an interior fracture basin originating from divergence and extension within a continental block. Clifford *et al.* (1986) also termed the Sirte Basin an interior fracture basin, but near the plate margin. Lewis (1990) has commented that such basins are caused by extensional shear forces.

Conant and Goudarzi (1967) and Goudarzi (1980) suggested that the Sirte Basin was formed by large-scale subsidence and block faulting that started during the late Cretaceous. The Sirte was a positive arch during the Palaeozoic and most of the Mesozoic. In order for the arch to collapse, intracratonic rifting would have occurred. Burke and Dewey (1974) suggested that the development of the Sirte Basin was due to widespread extension between two African plates during early Cretaceous time, when the African continent was stationary or slow moving with respect to an underlying mantle plume. According to this situation the Sirte Basin formed the northern end of a fracture zone which extended south through the Chad depression into the Benue trough, separating a Saharan plate to the northwest from an east African plate. Following Burke and Dewey (1973) and their ideas that rifts are initiated as a result of mantle plumes or diapirs (model of mantle-activated rifts), Van Houten (1983) proposed that the Sirte Basin resulted from the passage of the African plate over a fixed hot spot during Early Cretaceous, a time when a significant shift in plate motion took place. The shift in the direction of plate movement led to a change from extensional to compressional stresses within the African plate and to the fragmentation of the thinned lithosphere. Schafer *et* *al.* (1980) in their study of palaeostresses in Libya indicated the role played by differential movements of two African plates. They suggested that during Upper Cretaceous times, Africa drifted northeast towards Europe leading to collision of the Adriatic-African promontory of the Saharan plate with Europe. Within the NE-moving African plate the Saharan sub-plate reduced its velocity relative to the larger east African plate, which continued moving leading to the extension and formation of the Sirte Basin.

Anketell and Ghellali (1991) interpreted fold and fault patterns in the Jifarah region of NW Libya in terms of Riddle mechanics resulting from strike-slip movements of deep-seated basement shears oriented subparallel to the African-European plate margin. They proposed that the basement fracture zone underlying the Jifarah and Tarabulus basins may be extended east and southeast towards the Sirte region with one arm passing to the north of the Cyrenaica Platform and the other dying out in the Western Desert basin. The movement on these faults is responsible for the structural evolution of the Sirte Basin and thus the Sirte Basin rift system should not be interpreted in terms of simple dip-slip extensional tectonics only. However, the idea of the three arm (Sirte Basin) rift was first postulated by Parsons *et al.* (1980). Harding (1984) has pointed out the similarities between the Viking Graben, the Gulf of Suez and the Sirte Basin in terms of their origins, structural styles and hydrocarbon occurrences. He also described the Sirte Basin's arms, including the Sirte Basin deep and Hagfa Trough which represent the northwestern arm, the Tumayn Sub-basin which represents the southern arm, and the Sarir-Hameimat Trough which represents the eastern arm. The three arms define a large triple junction.

More recently, following the idea of an extensional origin of the Sirte Basin, Gumati and Kanes (1985) have discussed the regional history of the northern Sirte Basin and demonstrated the relationship between fault movements and facies patterns. In another paper of Gumati and Nairn (1991), the authors have calculated the rates of basin

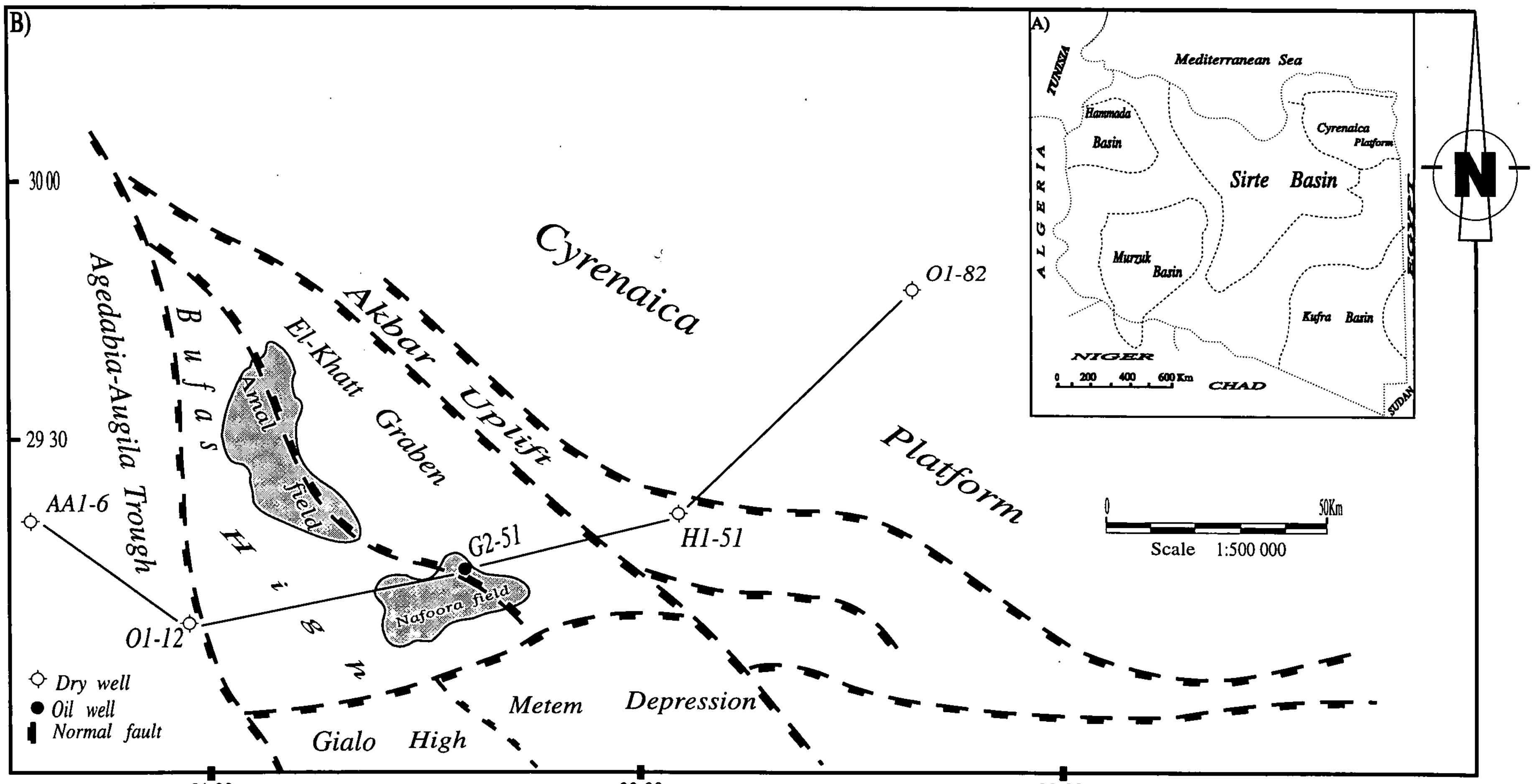

Fig. 1.1 *A) Location map of Libyan sedimentary basins. B) Major tectonic elements of the eastern Sirte Basin and location of wells* used in ths study.

subsidence and sediment accumulation and the facies variations of constructed a model of basin history.

The primary structural pattern of the Sirte Basin is characterised by horst and grabens oriented mainly in the NW-SE direction. However, Parsons *et al.* (1980) indicated that the horsts are often tilted eastwards to form half-grabens in the hanging wall of faults. Interfering with the NW trends, NE, ENE, and E-W clusters of faults and/or fault blocks are also present in the southeastern part, around the southern edge of the Cyrenaica Platform (Fig. 1.1A). The formation of the NW structural trends is believed to follow the pre-existing basement fractures of the Caledonian orogeny, which dominates the whole of Libya, while the occurrence of the Sirte Basin on the NE trending Tibesti-Sirte uplift of Hercynian age may have encouraged the NE trends. However, Kumati (1981), Anketell and Ghellali (1991), and Anketell and Kumati (1991) have all shown evidence of strike-slip or oblique-slip movement on faults in northwest Libya (Jifarah basin) and western Sirte Basin. Movement on the master transcurrent faults across northern Libya, coupled with the reactivation of basement lineaments in a tectonic regime that is basically extensional, may have controlled the development of individual faults in the Sirte Basin.

The average thickness of the basin's sedimentary infill is 2.4 km, with sediments being Cretaceous and Tertiary in age (Parsons *et al.*, 1980). However, Palaeozoic deposits have been documented in fault blocks in some parts of the basin. Sediments vary from shallow-marine to deep-marine deposits with some local reefs and/or restricted deposition of evaporites and dolomite.

The study area (Fig. 1.1B) occurs towards the eastern margin of the Upper Cretaceous-Tertiary Sirte Basin, approximately between latitude 29°00′ and 29°50′ and longitude 20°50′ and 21°45′, running generally NW-SE and covering 2500 Km^2. The El-Khatt Graben, a shallow saddle, separated the area from the Cyrenaica Platform to the east. The Gialo High and Metem Depression separate the area from the south, while the

Agedabia-Augila Trough bounds the Bufas Platform in the west and north. Different opinions and ideas (stated above) have been put forward about the origin of the Sirte Basin. The distinctive geological pattern in this area suggests that this part of the basin has undergone a different deformation regime (Schafer *et al.*,1985). However, the primary structures of the Caledonian and Hercynian orogenies which dominate the entire country would have been rejuvenated and/or locally adjusted during the formation of the Sirte Basin and hence have contributed in the structural developments in this part of the basin. Also more influence from the tectonic events of the Western Desert Basin and the uplift of the Cyrenaica Platform may be expected here. Since the basement in this part of the basin has been intruded by igneous rocks, basement faulting might also have been controlled by the varying characters of these basement rocks. The tectonic influence of granite is well known, with cases of granite buoyancy having been investigated in the North Sea by Scrutton *et al.* (1986) and Donato *et al.* (1990). The east-west trending faults are thought to date from before the Cretaceous and might have originated during the breakdown of the Calinso-Uweinat uplift in the Late Palaeozoic to Mesozoic tectonic phase.

Similar facies of carbonates and shales which dominate the stratigraphic succession of the Sirte Basin as a whole exist in this eastern part of the basin. However, local facies changes do occur in the graben structures. The initial sediments are pre-Lower Cretaceous sandstones and siltstones of non-marine origin, filling the irregular basement topography. This succession is overlain by a thick fluvial sandstone known as the Nubian sandstone which is the major reservoir in this part of the basin. These basal sediments were buried during the late Upper Cretaceous by a thick shale and carbonate sequence. Subsurface stratigraphic data gathered from oil exploration reveal a component of an old fault system reactivated in the Late Cretaceous, which continued through the Miocene and probably into present times (Selley, 1969). The maximum subsidence occurred during the Palaeocene and early Eocene (Gumati, 1984).

In some areas, the basement of the Sirte Basin subsided to depths well in excess of 16,000 ft (5,000m) below present sea-level. Intensive basement block-faulting resulted in the development of northwest-southeast and northeast-southwest oriented subbasins and platforms (Goudarzi, 1979).

1.4.2 Regional Stratigraphy

Precambrian to Tertiary aged strata of the Nafoora Field consist mainly of epeiric platform carbonates, associated with clastics and evaporites, reflecting major cycles of transgression and regression and regional structural features (Hecht *et al.*, 1964; Coltro, 1966; Klitzsch, 1968). A number of major gaps or changes in sedimentation are recognised, and these allow the strata to be subdivided into major rock units ranked as formations. These formations were deposited on the eastern margin of the Sirte Basin, which lay along the southern edge of the Tethyan Ocean during Upper Cretaceous-Tertiary time.

The Nafoora boreholes (all AGOCO) provide a lithostratigraphical and biostratigraphical correlation of Precambrian through Tertiary strata. These boreholes are located on a Pre-Upper Cretaceous structural high, which during this time underwent uplift and erosion. These boreholes illustrate the onlap of earliest upper Cretaceous marine deposits on to the Pre-Upper Cretaceous structural high.

The rock units (formations) that are recognised in the oil-producing areas of interior Libya are described here in ascending stratigraphic order. The descriptions are brief and intended to outline the general sense in which each unit can be discriminated; type sections are not described. The rock unit names used here are the result of oil company work and they are embodied in unpublished oil-company reports.

The stratigraphic succession is over 11,000 ft (3350m) thick, and this was deposited on Precambrian basement. Figure 1.2 gives the generalised scheme of the stratigraphic framework in the study area, which basically is a modified version of that derived from a similar succession encountered in the Nafoora Field (Belazi, 1987).

The succession is dominated by marine carbonates deposited on a tectonically active, broad platform bounded to the north by the central Tethyan Ocean. In addition to the carbonates, there is a significant amount of shale (primarily within the Upper Cretaceous and Oligocene). Under the influence of steadily rising sea-level, basin centre and basin margin facies extended both towards the Bufas high which locally was covered by a thin veneer of deep marine shale. In the basin centre, laminated bituminous shales were deposited under euxinic conditions, and these later became the prolific source rocks for hydrocarbons in basement, Upper Cretaceous and Tertiary reservoirs (Hamyuni *et al.*, 1984). During sea-level lowstands, caused either by a eustatic lowering of sea-level or uplift under an arid climate, evaporites were deposited upon shallower parts of the platform. Periodic incursions of continental sand, were restricted to the far north-eastern corner, probably coming in from the eastern hinterland where the Cyrenaica Platform was an active positive element.

1.4.2a Precambrian (Basement)

Basement is known at outcrops in the south Libyan Mountains (e.g. Jabal Tebesti) and some deep wells.

1400 ft (367m) have been drill in the plutonic and volcanic basement rocks on the Nafoora high, and consist of granophyre, granophyric granite, and rhyolite. Feldspars are generally granulitised along fractures, and also are sericitised and kaolinitised.

The sole age-dating information available in the Nafoora field is from the volcanic rocks penetrated in deep holes drilled on the Bufas structure. Potassium-argon age determinations conducted by Robertson Research indicate these rocks to be early Cambrian to Late Precambrian in age. Williams (1968) reported that the potassium-argon

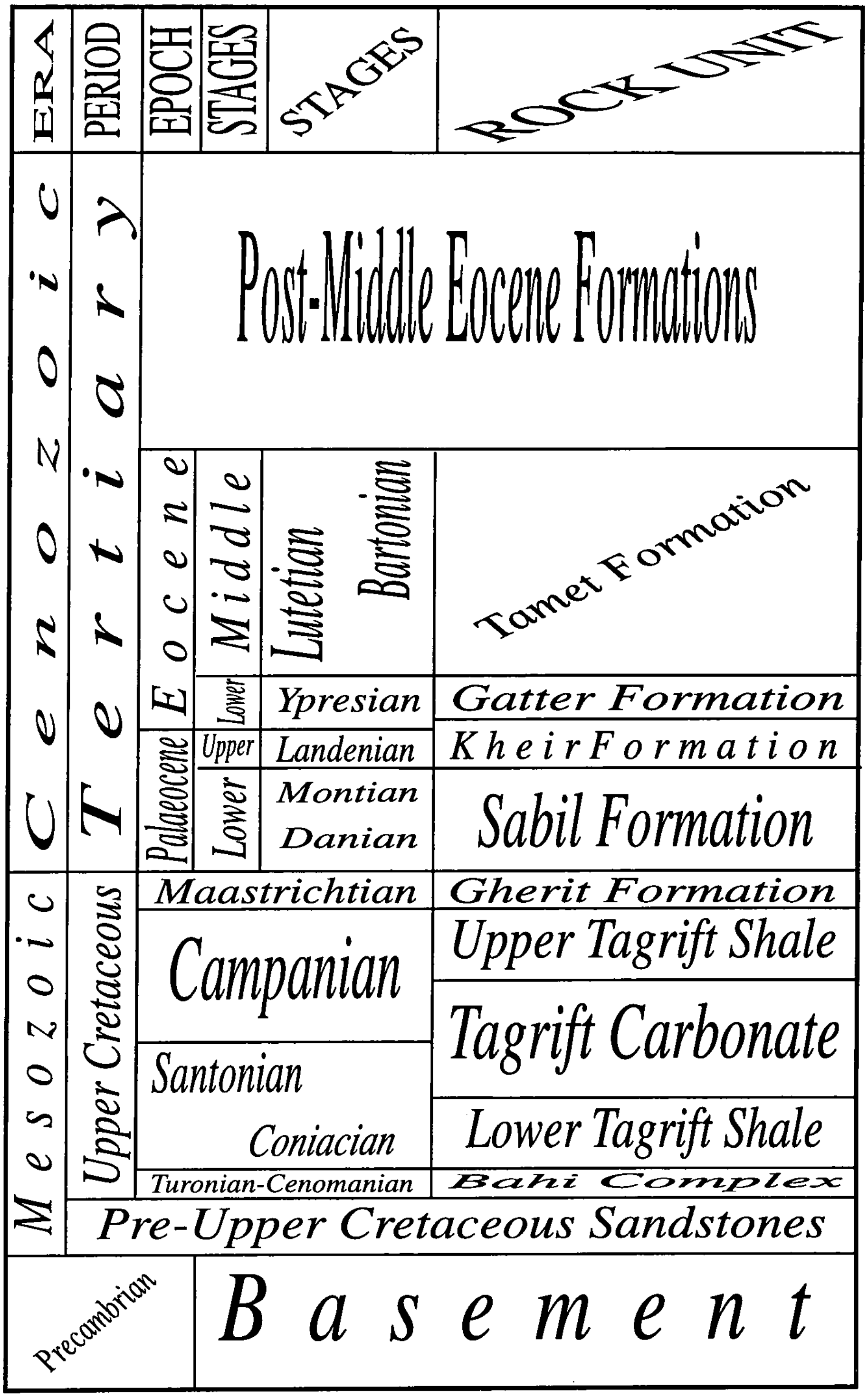

Fig 1.2 *Generalised stratigraphic succession of the Nafoora area, eastern Sirte Basin (modified from Belazi, 1988)*

age determinations of granite samples indicate an age of 402-568 Ma. These ages are considered to be minimum values, because argon losses may have occurred during hydrothermal alteration and mild tectonic deformation after the initial crystallisation; consequently, a Precambrian age is assigned to the entire basement complex in the Nafoora area. The basement complex on the upthrown block of the Nafoora fault is unconformably overlain by the basal sandstones of pre-Upper Cretaceous age or by the shales of lower Tagrifet (Upper Cretaceous).

The basement rocks are the major reservoir in the Nafoora Field. Sections drilled in the basement rocks range from a few tens of meters (e.g. well G2-51) to over a hundred meters (e.g. well G85-51) in thickness and hydrocarbon productivity ranges from nil to several thousand barrels of oil per day. This erratic production is due to the reservoir characteristic of these rocks.

Porosity and permeability within the basement rocks is the result of intense fracturing and weathering and at present is little understood. One explanation is that the basement highs are remnants of the deeply weathered and highly eroded Precambrian land-mass which may have been exposed from early Palaeozoic to late Cretaceous time

1.4.2b Pre-Upper Cretaceous Sandstones

The first sediments deposited in the Sirte Basin and resting directly upon the basement are of continental origin. On the downthrown blocks adjoining the Bufas structure they attain a thickness of at least 3000 ft (900m) (Robert, 1970). The lithology of the initial deposits in the Nafoora area is predominantly fluvial sandstone interfingering with subordinate shales. The sandstones are multi-coloured, and massive to thinly-bedded, with grain size ranging from fine to coarse and sorting generally poor.

There are volcanics intercalated with the fluvial sediments (Williams, 1969). The Bufas High itself was a positive feature during Pre-Upper Cretaceous times, subjected to strong erosion. The lack of marine fossils, together with the unsorted nature of the sandstones as well as the presence of non-marine shales and red beds, suggest that deposition was predominantly non-marine (William, 1969). The granitic and volcanic rocks of the basement landmass of the Nafoora area is considered to be the source of sediments for this formation.

Age determination is based on radiometric analyses of the volcanics and associated dykes and sills in the upper part of this succession, which gives a broad range of Jurassic to Cambrian (Robert, 1970; Barr and Weeger, 1972).

The reservoir qualities of the pre-Upper Cretaceous sandstones are generally poor, except in certain areas, where the sandstones exhibit good to fair reservoir qualities. Fractures within this formation provide the necessary permeability. The huge areal extent and the thickness of the oil column make it one of the major reservoirs in the Nafoora Field, especially when capped with the lower Tagrifet shale, with its better reservoir properties.

1.4.2c Upper Cretaceous Bahi Sandstones

Overlying the basal fluvial clastics is the Bahi sandstones. This is an extensive marginal wedge of fluvial to shoreface facies. The Bahi sandstones vary abruptly in thickness from zero to over 600 ft (180m), with the greater thickness on the basinward side of the Bufas high (William, 1968; Terry and William, 1969).

The lower Bahi sandstones consist of coarse-grained, light brown to white, and subrounded sandstones. The lower part of the unit is composed of minor granite wash, and is overlain by coarse-grained sandstones, probably in a shoreface environment. The upper Bahi sandstones are finer-grained and occur as a veneer of shoreface facies, mainly argillaceous, micaceous and slightly pyritic sandstones. Where the Bahi sandstones rest on the Pre-Upper Cretaceous sandstones, it is usually difficult to define the contact, because there is no sharp break in deposition (Barr and Weeger, 1972).

The basal structureless sandstone facies is thought to have been deposited in a series of extensive fans, dissected by a complex series of braided streams bordering the crystalline highland region. The fans directly fed shorelines, with only narrow alluvial plains with meandering rivers. The overlying marine sediments are believed to have been derived from the erosion of basement ridges, with deposition taking place during a long-term relative sea-level rise, when the shoreline position moved away from the basin on to the inner platform (Williams, 1968). Strata onlap on to the Pre-Upper Cretaceous surface towards the Bufas High, and terminate down dip against the underlying early clastics in the basinward direction. This pattern, however, may be caused by the progradation of the margin over a gently sloping substratum into the open Tethys. The vertical aggradation of shoreface facies continued across the entire platform, which is characterised by normally-graded clastic-shelf deposits. It is overlain by fluvial facies. Progradation developed across the outer platform edge and into a pre-existing basin, when the rate of sea-level fall slowed and then reached a lowstand position. Generally the wedge is characterised by weakly progradational units with a seaward-steepening configuration and rarely a vertical stacking. However, this interpretation is based on log correlation with supporting sedimentological and stratigraphical evidence (Williams, 1968; Belazi, 1989).

1.4.2d Upper Cretaceous Lower Tagrifet Shale

The Upper-Cretaceous (Santonian-Coniacian) marine transgression initiated a sequence of deep-water shales, which are well developed on the fringes of the Bufas High. They were probably not deposited on the crestal areas of the Sirte Basin, because the nature of the basinward margins are uncertain. However, some workers (e.g., Conant and Goudarzi, 1967) believe the former view that shales were also widespread over the present highs. Isopachytes clearly show that they thin towards the Bufas margin where their place is taken by carbonate buildups of the Tagrifet limestones as suggested by Belazi (1989), about 300 ft (90m) on the flank of palaeotopographic highs.

The lower Tagrifet shale reaches a maximum thickness of at least 2000 ft (600m) in the deeper parts of Sirte Basin. These shales with their greater thickness in the troughs around the Nafoora area, are believed to be the source for some of petroleum found in the reservoirs on the high (Hamyuni *et al.*, 1984). This shale is conformably overlain by the upper carbonates, with an apparent transitional contact towards the basin. The lower boundary is conformable with the Bahi sandstones, and in this section, consists of dark-coloured, fissile, calcareous shales interbedded with light grey to light tan, microcrystalline and chalky limestones. Peculiar and highly distinctive benthonic and planktonic foraminiferal assemblages encountered in this deep-water sequence have been dated and confirm a Santonian to Coniacian age (Barr and Weeger, 1972).

A thick succession of shales was deposited across the starved basin deposits during the transgressive Upper Cretaceous seas (Barr and Werggren, 1981). Incomplete drowning resulted from either a rapid large sea-level rise or epeirogenic tilting of the Bufas platform, or both. This Upper Cretaceous incomplete drowning and the large amount of off-bank hemipelagic sediment, resulted in extensive lateral growth of a deep-marine apron which led to closure and infilling of the seaways.

1.4.2e Upper Cretaceous Tagrifet Carbonates

During the Upper-Cretaceous (Santonian/Coniacian to Campanian), the second construction stage of the Bufas platform proceeded, with the sequences showing lateral accretion due to outbuilding of offshore banks. The new margin morphology was sufficient to initiate the first carbonate cycles above the lower Tagrifet shale. They constitute the most important trapping unit of the prolific hydrocarbon reserves in the Nafoora province. This unit is called the Tagrifet carbonates; they have been penetrated by wells in the top of the Bufas structure (Williams, 1969). The thickness of Tagrifet limestones varies from about 200 ft (60m) on the crest to over 900 ft (275m) in the Agedabia-Augila Trough around the Bufas platform. The upper surface is a gradational contact with the Upper Tagrifet shale. They conformably overlie the lower Tagrifet shale, except on the crest of palaeo-highs, where the lower Tagrifet shale was not deposited and it unconformably overlies the Bahi complex.

The Tagrifet carbonate is one of the major reservoirs in the Nafoora Field. Effective winnowing of micrite matrix and/ or faunal activity and dissolution during diagenesis enlarged original porosity to form an excellent reservoir rock on the crest of the high. Towards the offshore areas, the reservoir quality decreases markedly where quiet-water conditions allowed the deposition of muddier tighter limestones.

The sedimentary facies developed along a low-relief, relatively high-energy platform. Three lithofacies occur in the study area, and these reflect, to varying degrees, the subenvironments and organism communities.

The facies distribution of the Tagrifet carbonate succession can be exemplified by the Nafoora area, which, according to Creel (1970), ranges from basin and slope facies distinguished by fine reefal-planktonic foraminifera mud-wackestones, with only a few recognisable fragments of ostracods, echinoderms and calcispheres. This basal part was developed and situated at the top of an open-marine section of lower Tagrifet shale, while the upper part of this facies passing from the toe further up the bank slope, graded upward into fine rudistid-coralline red algal wacke-packstones. There is a predominate crumbly fabric; bivalves and echinoderms are common. Most of the bioclastic grains have been leached out and fine internal sediment was trapped by a sheltering effect.

The second facies consists of poorly-sorted bioclastic packstones/grainstones with comminuted bivalves shells, including rudists, *Inoceramus* and *Exogyra*. The accessory organisms in this facies include echinoderm debris, coralline red algal clasts and gastropods. Many of these clasts have been reduced to rounded sand-size grains. The high energy conditions winnowed out the fine debris and lime mud. This facies grades into the previous facies in a basinward direction.

This facies was deposited in shallow shoals or banks on the order of tens of metres thick and a few hundreds of meters across. They developed for the most part in a relatively low energy, muddy environment. They are superimposed on each other on the open shelf margin, generally as discrete shoals, although in some cases they form more continuous barrier islands.

The lagoonal bio-wackestone and subordinate packstone facies is characterised by miliolid foraminifera and macrofossil fragments. It is partially dolomitised, with wispy laminae, and much stylolitisation. The lagoonal facies extends into the Bufas interior. It occurs immediately behind the coralline-rudistid shoals, and interfingers with back shoal facies. Upper Cretaceous rudist shoals lacked framebuilders and formed low-relief banks. The skeletal communities (bioaccumulations) clearly have not been moved far from their growth site, and are therefore interpreted as essentially autochthonous. They typically developed in sheltered habitats or below-wave base, so that buildups are mud-dominated. The upwards facies transition represents a backstepping of the rudist shoals, and indicates that the rate of relative sea-level rise exceeded the rate of buildup growth. Subsequent buildups were initiated at a new site some distance landward towards the lagoon.

1.4.2f Upper Cretaceous Upper Tagrifet Shale

The "Upper Campanian transgression" may be recognised lithostratigraphically in a number of boreholes in the eastern part of the Sirte Basin. This transgression is a regional event across the whole of north Africa. The thickness of this formation is fairly uniform on the eastern side of the Sirte Basin, averaging about 85 ft (±30m), but increasing on the flanks toward the Agedabia-Augila Trough areas. The Upper Tagrifet shale conformably overlies the Tagrifet carbonate across the entire Bufas High, with a transitional contact. It is conformably overlain by the Gherit Formation.

The Upper Tagrifet Formation in the study area mainly consist of grey, dark brown to black coloured shales. They are subfissile to fissile, firm, hard or brittle. These shales are slightly calcareous, rarely pyritic and locally contain thin interbeds of argillaceous limestones and minor sandstones. On the low areas the Upper Tagrifet shale is frequently interbedded with light to dark grey, microcrystalline, chalky and argillaceous limestone.

1.4.2g Upper Cretaceous Gherit Formation

During Maastrichtian time, a widespread regression occurred and the neritic platform carbonates of the Gherit Formation were deposited over the Bufas High. To the west, in the Agedabia-Augila Trough, thicker Gherit basinal deposits accumulated. The thickness of the Gherit Formation in the Nafoora Field is fairly uniform, averaging about 600 ft (160m). It thickens from the flanks of the Bufas High towards the trough, where it reaches over 1000 ft (600m).

The Gherit Formation conformably overlies the Upper Tagrifet Shale and underlies the Lower Palaeocene Sabil Carbonates with a gradational contact. The Gherit Formation encountered in the Nafoora Field consists predominantly of light-coloured limestone, which is generally soft to medium-hard, micro- to very finely crystalline, argillaceous and chalky. Locally, it contains brown-to-black carbonaceous material and microfaunas. Towards the top of the formation, the limestone becomes dolomitic and rarely interbedded with medium-brown dolomite. This dolomite is cryptocrystalline, dense and anhydritic. Barr and Weeger (1972) reported that the foraminiferal assemblage found in this formation (including planktonic forams) suggests that the Gherit Limestone was deposited in an open-marine environment: they placed the Maastrichtian-Danian contact at the top of this formation. In the Nafoora Field, similar conditions existed during the deposition of the limestone, but shallower conditions were apparent towards the top of the formation as it becomes more dolomitic and anhydritic.

1.4.2h Lower Palaeocene Sabil Formation

The Early Palaeocene rocks subcrop almost continuously along the eastern margin of the Sirte Basin. Rapid changes in the lateral facies occurred between the platform areas and the deeper parts of the basin. The evaporites, carbonates and shales of the early Palaeocene series were deposited on the continental stable platform which extended across the Sirte Basin. This heterogeneous series shows considerable variation in formation thickness which is interpreted as being related to the effects of the Upper Cretaceous structures rather than renewed tectonic activity (Brady *et al.*, 1980).

Based on local and regional information, the early Palaeocene section is made up of two shallowing-upward cycles, where each cycle is underlain-and overlain by deeper, pelagic facies. A gradual shallowing-upward trend from basin to shallow platform is seen in both bio- and lithofacies. Mresah (1993) recognised seven lithofacies in the early Palaeocene, of the northeast Sirte Basin. The environmental parameters of the faunal assemblage apparently reflect very shallow-water, marine deposition under tropical to subtropical conditions. The high recorded content of lime mud suggests low- to moderate wave action and current activity (low energy). However, the presence of some abraded fossil fragments may indicate sporadic, vigorous current activity.

The Sabil carbonate is one of the major reservoirs, containing at least a third of the hydrocarbons in the Sirte Basin (unpublished reports). The reservoirs include pinnacle reefs (Brady *et al.,* 1980), shoals and other associated shallow-marine facies (Bebout and Pendexter, 1975). The Early Palaeocene Sabil Formation described by Gumati (1992) in the type section was divided into lower and upper Sabil formations separated by chalk, argillaceous limestone and shale of the Hagfa Formation. Subdivisions are virtually impossible in the Nafoora area. The Hagfa Formation is very thin on the flanks of the high (Belazi, 1987), and it thickens rapidly westward into the Agedabia-Augila Trough. The Sabil Formation is conformably overlain by the Kheir Formation. The thickness of Sabil section is fairly uniform on the crest of the Nafoora Field, where it averages about 650m. This thickness and geometry of the formation indicates rejuvenation of the faults during the deposition of this formation.

1.4.2i Upper Palaeocene Kheir Formation

A major transgression of the Late Palaeocene sea brought deeper-water conditions over the whole area of the Sirte Basin and basinal shales were deposited. In general, the Kheir shales are thin over early Palaeocene platforms, but they thicken rapidly on the flank into the Agedabia-Augila Trough to the west. This marked the end of the Palaeocene, as the succeeding upper part of Kheir shale is of early Eocene age. In the Nafoora Field, the Kheir Formation is fairly uniform and averages about 60 ft (20m) in thickness in the crestal area, thickness increasing markedly toward the trough with a maximum of 110 ft (35m). The Formation is conformably overlain by the lower Eocene Gatter Formation with a transitional contact.

The Kheir Formation in the Nafoora Field consists predominantly of medium- and dark-brown, hard-to-friable marl, interbedded with highly argillaceous, buff, soft and earthy limestone and minor amounts of dark-grey shale and clay. The benthonic and planktonic forminiferal assemblage in this formation contains some stratigraphically important index fossils (Barr and Weeger, 1972), suggesting that the formation was deposited in an open-marine environment and straddles the Palaeocene-Eocene (Landenian-Ypresian) boundary, the lower part being upper Palaeocene and the upper part being lower Eocene. Kheir sediments are believed to be good source and seal rocks for the hydrocarbons found in the underlying Sabil carbonates (Hamyuni *et al.,* 1984).

1.4.2j Lower Eocene Gattar Formation

Palaeogene Formations were deposited on a wide, broad and shallow platform that formed along the eastern margin of Sirte Basin. The platform formations are characterised by two major sedimentary cycles: the Late Palaeocene-Early Eocene transgressive cycle of Kheir and Gattar Formations followed by a thicker shallow-water regressive cycle of the Tamet and Augila Formations. The termination of the Palaeogene transgression is marked by a widespread break in sedimentation which is widely recognised and is marked by the local occurrence of phosphate, glauconite, dolomicrite and anhydrite. This break between the Ypresian and Lutetian signalled the beginning of the Middle-Upper Eocene regression.

In the Early Eocene, the eastern Sirte Basin platform was un-rimmed and had a gentle slope. This platform was subjected to frequent transgressions and regressions. During the Ypresian, the deposition of the Gattar Formation took place on a substratum characterised by local topographical irregularities of tectonic origin. In general, this platform was covered by wide facies belts with an evaporitic platform in the east, followed westward by a restricted platform and then broad subtidal lagoon with oyster and large-foram banks. In turn these shallow-water facies grade westwards into pelagic and open-marine facies.

The Lower Eocene Gattar Formation extends for about one hundred kilometres around the southeastern part of the Sirte Basin. It has been described by different companies using various formation names and members. The total thickness of Gattar Formation is generally uniform in and around the Nafoora Field. It ranges from just 1150 ft (390m) on the Bufas High to over 1600 ft (550m) in the Agedabia-Augila

Trough; this clearly suggests that it was deposited on a uniformly subsiding region with only gentle bottom slopes. The Gattar Formation generally consists of limestone, dolomite and anhydrite facies. These facies occur in belts that parallel the recent coast of the Gulf of Sirte. The Gattar Formation in this area can be subdivided into three units.

The lower unit is skeletal wackestone, with minor amounts of shale. The limestones are white, cream to light grey, mostly medium hard to hard, and occasionally chalky. They are mainly very fine-grained to micritic; fossils are common and mostly nummulites. The porosity in this unit is mainly of chalky type and ranges from poor to fair. Further towards the east these limestones are interbedded with white to light grey and light brown dolomites. These dolomites are hard, finely crystalline, anhydritic, slightly sandy and cherty and occasionally exhibit good vuggy porosity. The minor amounts of shale in this unit are mainly restricted to the southern area of the Bufas High. These shales are light green to olive green, slightly fissile and moderately soft.

The middle unit consists of interbedded dolomites, limestone, anhydrite and shale. The dolomites are light brown, occasionally dark brown or light tan. They are medium hard, very fine to medium crystalline and occasionally sucrosic with good intercrystalline porosity. To the west, dolomite is locally anhydritic and interbedded with limestone. These limestones are mostly wackestones (biomicrites) and range from light tan to light brown. They are moderately soft, chalky in part, very fine to medium crystalline and occasionally grade into sucrosic dolomite. These limestones are slight to moderately fossiliferous, with nummulites increasing in number in a northward direction. The anhydrites are generally clear, translucent and white. They are finely crystalline, hard and occur as nodules, lenses or beds ranging in thickness from 0.9-1.6m as indicated by the electrical and sonic logs. Associated with these dolomites and anhydrites, minor amounts of shale are restricted to the southern area.

The upper unit consists mainly of limestones, especially in the northwestern parts of the study area, where with the lower unit, they comprise the entire Gattar Formation. The limestone of the upper unit is generally a wackestone, white and light tan to light grey, soft to firm and hard, occasionally chalky to marly. Also these limestones are dolomitic or interbedded with thin beds of dolomites.

The lithological and palaeontological evidence generally suggests that the lower unit of the Gattar Formation has been deposited in a shallow shoal and open-marine environment. The middle unit has been deposited in a very shallow and frequently restricted marine environment. The deposition of the upper unit indicates a return of shallow shoal and open marine environments.

1.4.2k Middle Eocene Tamet Formation

During Middle Eocene (Lutetian-Bartonian) time, major parts of the Nafoora High and its neighbouring regions were covered in a shallow sea. During this time, large parts of the Cyrenaica Platform were under very shallow marine (Tethys) water as is shown by the presence of peritidal and saltern deposits.

Sedimentological studies of the middle Eocene in Libya are sparse. On the basis of faunal and lithological evidence, El-Hawat *et al.* (1986) concluded that the middle Eocene facies in the northeast exposures continue to be a complex of shallow and deep water deposits because of Al-Jabal Alkhdar tectonism and sea-level fluctuations. Rocks are principally carbonates and represent deep to shallow-marine facies; and they are generally interpreted to be open marine, low-energy facies near its base, because of the abundance of plankton and paucity of benthos. Subsequently there appears to have been shallowing upward into facies rich in benthic organisms.

Rocks of the same age encountered in wells drilled in the eastern sectors of the Sirte Basin have a thick succession of carbonate, dolomite and locally evaporite and these cover most of the studied area. They form three major transgressive-regressive units.

These sequences attain a maximum thickness of over 1900 ft (580m), and were deposited on a broad and shallow un-rimmed platform on the eastern flank of the Sirte Basin. This platform had a gentle sloping surface and exhibited different degrees of restriction. In general, this platform was covered by facies exhibiting the characteristics of a saltern environment in the east upon the present-day Cyrenaica Platform, followed by a subtidal platform lagoon where large foram-banks developed. In turn these shallow-water facies grade westwards into pelagic and open marine facies.

Recent detailed descriptions of the Tamet Formation in the eastern Sirte Basin have shown the presence of nine carbonate and evaporite microfacies (see chapter 2). These microfacies show changes in both litho- and bio-facies as well as variations in thickness. It is very probable that the observed variations are the result of factors such as Tethyan eustatic sea-level changes and perhaps geographic position. The whole area was one of long-term tectonic stability at this time (Gumati, 1984).

The ten sedimentary microfacies (A to J) discussed later in this dissertation and can be classified as part of the Tamet and/ or Cyrenaica Platform facies. Middle Eocene rocks on the eastern margin of the Sirte Basin are mainly carbonate and evaporite deposits formed under progressive marine conditions, beginning with moderately deep to shallow subtidal, and followed by tidal flat and saltern environments. The intercalation of these deep, shallow and very shallow-marine facies is a good indication of the Tethyan transgressive-regressive phases, which were dominant during the Tertiary history of the Tethyan Ocean. Said (1990) has concluded that the similarity of the depositional facies of the Middle Eocene of Egypt, Libya and in most of the adjacent countries, suggest that this area was strongly influenced by Tethyan eustatic oscillations. The depositional history of the Tamet platform is different from other middle Eocene platforms on the southern margins of Tethyan Ocean in that 1)- there is a lack of widespread large-foram barriers along the platform margin. 2)- there is an absence of any indication of peritidal deposition.

1.4.2l Post-Middle Eocene Formations

The post-middle Eocene sediments in the studied area consist of three formations. The Nafoora Formation (Upper Eocene) is mainly represented by carbonates and shales. The Nafoora Formation is fairly uniform over this area, averaging about 120ft (37m) in thickness, increasing toward the Agedabia-Augila Trough. The lithological and faunal assemblages of this formation indicate that it was deposited in a shallow-marine platform environment. The end of the Eocene time in this area was marked by shallowing of the Tethyan Sea and introduction of clastics in Oligocene times. Good oil-shows were recorded from the wacke-packstone sections in the inner platform section, especially in the crest of the Nafoora High. Reservoir quality and net pay-section decrease towards the trough.

The Oligocene section in the study area consists two formations, the basal Arida and the top Diba Formations (Barr and Weeger, 1972). The Oligocene formations are represented by fine-grained, glauconitic, poorly-cemented sandstones, which grade into carbonates and shales. The Oligocene section in the Nafoora Field is uniform in thickness, averaging about 670ft (204m). Diagnostic fossils found in these formations indicate that very shallow-water conditions persisted throughout the Oligocene, with minor fluctuation in sea-level. In general, shallowing is upward, with occurrence of open-marine conditions at the middle of the Oligocene section marking the boundary between the Arida and Diba Formations.

Several wells exhibit fair to excellent evidence for hydrocarbon accumulation in limestone, sandy limestone and sandstone of both formations. Due to the gentleness of the structure during the Oligocene time, the potential production area is limited to the crestal part of the Nafoora High.

The latest formations (Miocene-Present) consist of bioclastic limestones interbedded with claystones and minor amounts of anhydrite overlain by fine-grained, calcareous sandstones. The thickness of this section averages about 1100ft (335m). The effects of the faults around the Nafoora High are diminished; the resulting structure is gently with very small closure in the crestal area.

The diversity of macrofossils within the lowermost part suggests an open-marine environment while the sparse remains of molluscan fragments and casts and moulds in the middle of this section indicate a shallow-marine to semi-restricted depositional environment. Finally, the presence of red sandstones and claystones with no fauna strongly indicates that continental conditions prevailed in the study area.

Chapter 2

Microfacies and Depositional Environments of the Tamet Formation

2.1 Introduction:

This chapter introduces the Tamet ramp, facies associations and their constituents microfacies. This includes a petrographic description of the microfacies and their diagenetic phases, followed by an environmental interpretation.

The thickness of the Tamet Formation in the study area, ranges from 1570 ft (480m) in the platform areas to over 2800 ft (870m) in the marginal-platform setting.

Nine distinct microfacies types have been distinguished in the studied wells. These microfacies were alphabetically designated A through J and summarised in Table (1).

2.1.1 Microfacies A: Chalky mudstone

Occurrence: This microfacies occurs in the most basinward stratigraphic sections of the study area and grades landward into deep-ramp planktonic wacke-packstones.

Description: This microfacies consists typically of pure calcitic, fine-grained pelagic mudstone. Determinable components include a few planktonic microorganisms (Fig.2.A) and locally some other nannoplanktonic assemblages. A similar pelagic microfacies has been described by Ahmed (1992), consisting of soft weakly to strongly nodular chalk. By using the SEM, he concluded that this chalk is a relatively monotonous coccolith-rich mudstone. Small crystals can be recognised between and on the coccoliths and coccolith fragments and he believed these to have been either high Mg-calcite or aragonite cements. The most striking feature of this microfacies is the abundance of broken or crushed coccolith shells.

Table (1) Simplified characteristics of facies associations and their microfacies types in Middle-Eocene ramp sequences of eastern Sirte Basin.

Descriptive/ Microfacies	Occurrence	Distinguishing Features	Interpreted Depositional Environment
A). *Chalky mudstones*	This microfacies occurs in the most basinward of the study area in wells AA1-6 and O1-12.	This microfacies consists of fine-grained pelagic mudstone. Determinable components include a few planktonic foraminifera and some other nannoplanktonic assemblages.	This microfacies is interpreted as the deepest deposits probably accumulated by pelagic settling from suspension and represents very distal deposits. This microfacies deposited under straved basin conditions in poorly oxygenated stagnant water below storm wave-base at water depths between 60-100m.
B). *Planktonic foraminifer mud-wackestones*	This microfacies is recognised in the western part of the study area in wells AA1-6 and O1-12.	The most common is planktonic foraminifera wackestone, while mudstone microfacies is rare. This microfacies has a characteristic dark colour and is divided into a lower chalk unit and an upper pelagic micrite.	Pelagic faunas within dark organic-rich micrites characterise the deep water deposition below storm wave-base in more than 50m under marine, occasionally reducing conditions with minimum water turbulence. The assemblage of shallow water biota reflects proximal deposition as a result of episodic increase in sediment flux in response to bioclasts progradation. This microfacies is interpreted to represent suspension deposition in sediment-starved marine environment and is equivalent to the condensed section of Loutit *et al.* (1988).
C). *Microlitho-microbioclast wacke-packstones*	This microfacies occurs in the western and central parts of the study area at different depths.	This microfacies is an alternation of packstones and wackestones and represents deposition in an energetic environment, with fluctuations in storm-wave activity. This microfacies generally consists of mud-supported litho-bioclasts; the fragments display irregularity in their shape and orientation. The matrix often has a neomorphosed texture.	The interbedded nature of this microfacies is most likely due to the input of debris by storms into a generally quiet water environment. However, the influx of undifferentiated bioclasts from nearby carbonate sources, suggests that the depositional environment might have been under the influence of frequent strong and weak storm activity.
D). *Bioclastic mud-wackestones*	This is encountered twice in the central and eastern parts of the study area.	This microfacies is composed of scattered fragments of shallow-marine forminifera, echinoderms and bivalves. The bioclasts almost have a loosely-packed fabric, and range in size from only a few millimetres to greater then 2 cm.	This microfacies was deposited in vast open shallow subtidal ramp, in water depths 20-30m. Bioclasts were probably redeposited during sporadic high-energy events (probably storm-related), during a slow continuous transgression (catch-up condition), the landward retreating surf zone would erode ramp-interior successions and lead to redeposition.
E). *Bio-benthic foraminifera wacke-packstones*	This microfacies is limited to the central part of the study area, in well G2-51. It is repeated twice in the middle section of this well.	This microfacies is composed mainly of wackestones and packstones. Larger foraminifera, especially nummulitids are the dominant constituents locally with echinoderm, bryozoan and molluscan debris. Nummulitids, both large and small, occur as whole and broken fragments, mostly the chambers being filled by peloidal packstone cements.	The larger foraminifera constructed banks during the late transgression of a low-energy, shallow subtidal outer-ramp between normal and storm wave-bases in water depth 25-35m. Ahmed (1992) considered a development of large foram-shoals, caused by drowning by a transgressional pulse. In the beginning this transgression would have caused stabilisation of the banks, and later a drowning of the banks was caused by renewed deepening and increased influx of peloidal micrites.

Continued Table (1)

Descriptive/ Microfacies	Occurrence	Distinguishing Features	Interpreted Depositional Environment
F). *Peloidal-bioclastic wacke-packstones*	This microfacies is extensively developed throughout the study area, particularly in the western and central parts, occurring in thick units. It is repeated twice in well O1-12.	This microfacies is typically mud-supported, and characterised by a robust normal marine fauna of large echinoderm debris, bryozoan, molluscan and minor or small foraminifera. The skeletal material of coarse and sand size varies from whole to fragmented bioclasts. The matrix is compacted muddy peloidal material.	The characteristics of mud-supported and abundant normal marine fauna combined with the lack of sulphides and organic matter indicate deposition under well oxygenated subtidal conditions of normal salinity. Subtidal depths are estimated to be between 20-30m.
G). *Orbitolinid-echinoderm wacke-packstones*	This microfacies is restricted to the central part of the study area, and is encountered only in well G2-51. It is relatively thin.	This microfacies comprises bioclastic material which is highly fragmented. The fauna consists of the small orbitolinids, echinoderm debris and rare miliolids. The matrix consists mainly of silt-sized micrites, which display peloidal fabric.	The depositional environment of this microfacies is interpreted as being shallow marine subtidal in inner-ramp lagoon with open marine circulation. Current activity and wave action is low to moderate on the basis of the high content of lime mud matrix.
H). *Peloidal-miliolid wacke-packstones*	Its distribution is limited to the central part of the study area, in wells O1-12, G2-51 and H1-51.	This microfacies is dominated by miliolids. Other constituents include fine-medium sand sized peloids, with varying admixture of bioclasts. The matrix is organic-rich micrite.	The presence of mud matrix in this microfacies shows that for the most part deposition was in a semi-restricted, shallow subtidal inner-ramp lagoon setting. The miliolids are closely associated with peloids indicating that this lagoon was separated from the open marine by wacke-packstone banks during the deposition of this microfacies, which acted as a sill and restricted lagoon circulation.
I). *Dolomitised peloidal-bioclastic wacke-packstones*	This microfacies is recognised in the central and eastern parts of the study area.	This microfacies is very heavily dolomitised and most of the original depositional textures have been obliterated and altered to anhedral to subhedral dolomite crystals. Relict constituents include peloids and bioclasts. Two distinctive subunits of dolomitisation can be recognised and easily correlated across the region.	Dolomitisation probably took place by seepage reflux and evaporitive pumping. The relationship of the dolomitisation to other diagenetic features and textures indicates that dolomite replacement took place after deposition but before significant compaction.
J). *Anhydrite*	This microfacies is sandwiched between two thick units in the east of the study area, while, the dolomite becomes the dominant mineral westward.	Six stacked cycles of alternating anhydrite and dolomite subunits. Anhydrite subunits are thickening-upward, and have massive or nodular textures at the top of each cycles.	This association is characteristic of sulphate precipitated in a shallow epeiric sea, or a broad hypersaline platform (saltern), it was deposited during periods of restriction or complete isolation of the Cyrenaica Platform, when the platform was cut off from open marine conditions.

Interpretation: This ckalky mudstone is interpreted as the deepest-ramp microfacies that accumulated under partially straved basin conditons in a poorly oxygenated, stagnant water column below storm wave-base at water depths between 70 and a few hundred meters, occasionally affected by storm currents, in an area of the open-Tethyan ocean of normal salinity.

This microfacies was probably deposited by pelagic settling from suspension and represents vey distal deposits. This chalky microfacies is interbedded with a purer carbonate planktonic microfacies and many workers have considered this intercalation to be the result of Milankovitch cyclicity (e.g. Quine and Bosence, 1991).

2.1.2 Microfacies B: Planktonic foraminifera mudstone-wackestone

Occurrence: This microfacies occurs in wells AA1-6 and O1-12, which are the most westerly wells located in a distal position within the studied area in the Sirte Basin.

Description: This microfacies typically is more lime muddy than other microfacies. The basal unit of this microfacies (Fig.2.1A) in well AA1-6 is composed of relatively pure chalk. It is mainly a light-coloured mudstone with sparse wackestone intervals and it contains little organic matter. It is marked by anastomosing swarms of fine dissolution seams and microstylolites. Towards the top, this chalky microfacies passes up into pelagic limestones, typically dark in colour (relatively rich in organic matter), which are chiefly wackestones-packstones (Fig.2.1B). This pelagic unit rests with gradational contact on the chalk, with no clear evidence of omission surfaces.

This microfacies chiefly contains a basinal faunal assemblage. It is biotically diverse and dominated by mostly unbroken Globigerinid foraminifera. Other accessory microfauna near the top include agglutinated foraminifera, which are commonly associated with pelagic benthic forams such as *Textularia* and Nodosiridae. Calcispheres *(Oligostegina)* are present in the uppermost parts in low concentrations. There are also dispersed spicules, probably of sponge origin.

Some of the pelagic horizons seem to be phosphatised locally towards the top of the entire section. There is also a dark coloration, possibly due to the presence of dispersed organic matter, and pyrite is locally abundant as scattered framboids within the pelagic matrix or within the chambers of some planktonic foraminifera.

Interpretation: The depositional environment of this microfacies is interpreted to be deep water on the distal part of outer-ramp to basin setting (c.f. Barnaby and Read, 1990). There is no evidence of currents, indicating that this microfacies was deposited below fair-weather wave base, largely from suspension. Based on characteristic features of this microfacies Ahmed (1992) proposed a marine environment with normal salinity, warm clear water, depth about 50m with a muddy substratum.

In addition, there is no hard evidence for the influence of storm-waves or storm-currents, as inferred from the lack of the skeletal concentrations. However, there is a lack of very deep-water 'basinal' deposits (e.g. laminated claystones), suggesting that the whole system was most likely deposited above the calcite compensation depth (CCD).

The occurrence of phosphatic horizons in the form of discrete grains and as infills of foraminifera tests at the top of this microfacies represents a transgressive event marking the top of the distal boundaries within the Tamet Formation. There is no evidence to suggest that there was a relative fall in sea-level prior to this transgressive event. This rise in sea-level would have decreased the sedimentation rate and changed the circulation patterns on the sea-bed, whilst also possibly being responsible for an increase in organic matter production due to the associated increase in area of shallow sea (Baum *et al.*, 1988). The preserved deep marine organic matter suggests poorly oxygenated to anoxic bottom water conditions. This also explains the absence or limited benthic fauna in these levels. Slow sedimentation rates and probable low circulation rates coupled with the decomposition of organic matter were favourable for phosphogenesis. The phosphate was precipitated in microenvironments present within the voids where oxic, post-oxic and sulphidic microenvironments could occur together close to the sediment-sea-water interface (c.f. Jarvis, 1992).

Pyrite abundances showed no clear relationship to the depositional environment. Increasing abundances of organic matter lowered the Eh of the environment and enabled sulphate reduction to be sustained for longer periods. However, the main controls on sulphurisation are microbial sulphate reduction during organic matter degradation which results in authigenic pyritisation (Hudson, 1982).

2.1.3 Microfacies C: Microbioclastic-lithoclastic wackestone-packstone

Occurrence: This microfacies mostly occurs within wells AA1-6 and O1-12 in the western part of the study area. It occurs at 2 levels within the first well and 3 levels within the second well. There is also a thin unit recognised in well G2-51.

Description: In general this microfacies is characterised by a tightly packed fabric of a variety of bio-lithoclasts (Fig.2.2A) in packstone units which alternate with planktonic-microbioclasts in wackestone units. The contacts are commonly gradational upward, so that boundaries are difficult to define.

There is evidence of an upward decrease in grain size within this microfacies, and in some cases this appears to be repeated several times. These may represent deepening-upward cycles, but unfortunately only cuttings were available in this study, so this cannot be proved.

The packstone layers are made up entirely of bioclasts, which are mostly communited shelf faunal elements, of a highly fragmented and abraded nature. These range from very coarse to fine sand grade (2mm-0.5mm) and they are generally set in a dark, fine mud-

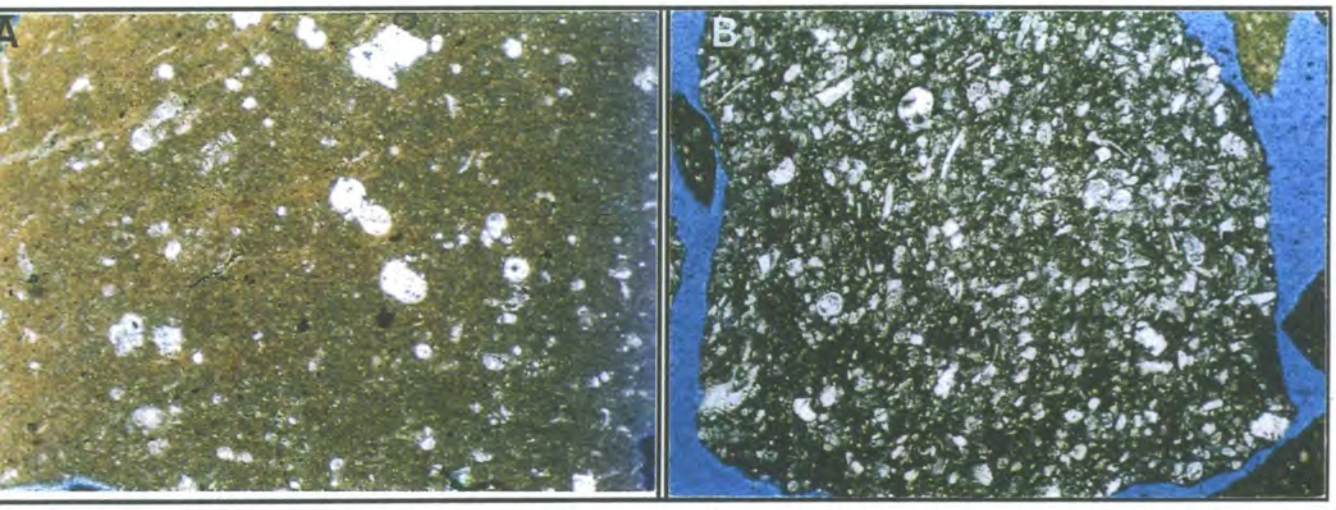

Fig.2.1 *Representative photomicrographs of different microfacies in the Tamet Formation. All photomicrographs were taken under plane polarised light at X 40 magnification. A) Chalky mudstone of the distal outer-ramp environment. It is mainly a light-coloured mudstone with sparse planktonic-forams and it contains little organic matter. It is marked by anastomosing swarms of fine dissolution seams and represents the deepest water sediment preserved in the Tamet Formation. Sample from well O1-12, depth 1887m (6190ft). B) Planktonic mud-wackestone of distal outer ramp environment, composed predominately of globigerinid foraminifera, calcispheres and undifferentiated spicules, which are dispresed in organic matter-rich matrix. The matrix changes gradually upward to phosphatic-bearing. Globigerinid chambers are partially filled with calcite as well as pyrite cement. The prevalence of organic-rich matrix and lack of shallow-marine faunas suggests deposition in a low-energy deep subtidal environment, where sediments were deposited from suspension. Sample from well AA1-6, depth 1867m (6125ft).*

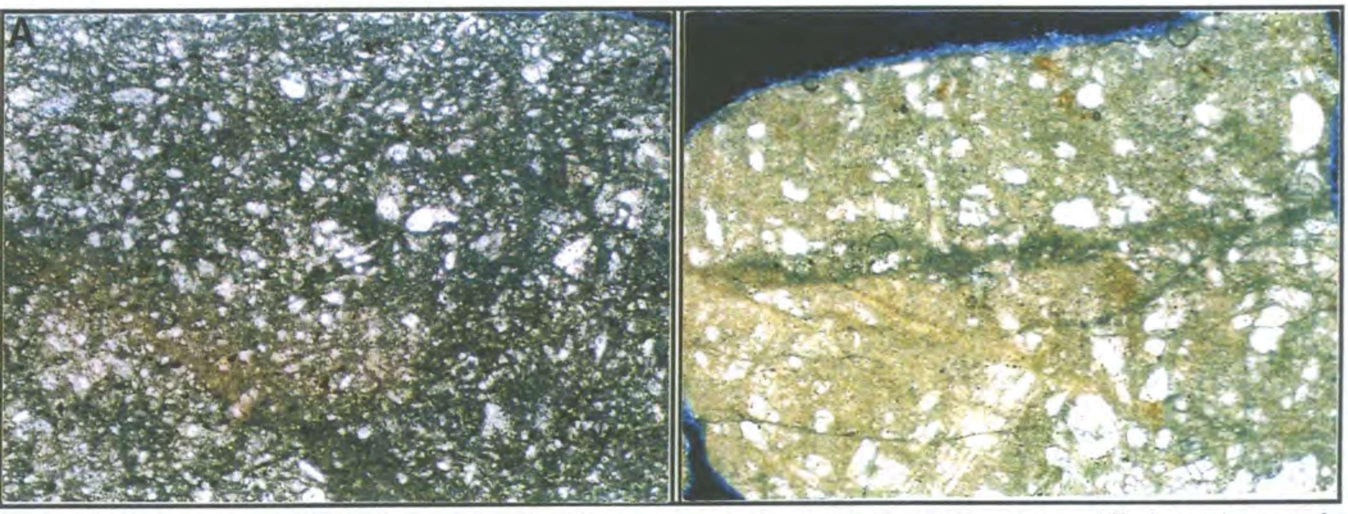

Fig.2.2 *Representative photomicrographs of different microfacies in the Tamet Formation. All photomicrographs were taken under plane polarised light at X 40 magnification A) Microlitho-microbioclastic wacke-packstone of the distal outer-ramp environment, is characterised by relatively packed fabric of silt-fine sand-sized litho-bioclasts. These clasts display irregularity in their shape and orientation. This microfacies is most likely due to input of clasts by storms into a generally quiet water environment. Sample O1-12, depth 2082m (6830ft). B) Bioclastic mud-wackestone of the proximal outer-ramp environment, is composed mixed foraminifera, echinoderm and bivalve bioclasts and range in size from only a few millimetres to greater than 2 cm. The bioclasts almost have a loosely-packed fabric and float in the lime muds. Bioclasts indicate that resedimentation took place in a shallow to moderate subtidal environment during sporadic high-energy events (probably storm-related). Sample from well G2-51, depth 1189m (3900ft).*

wackestone matrix. There is also a variety of other clasts, which may have originally been bioclasts but they have poor preservation, are often very small, and grade into the matrix.

The mixed assemblage of shallow-water biota is dominated by large foraminifera; there is a minor contribution from echinoderm debris and molluscan fragments. The coarser bioclasts show no preferred orientation, are poorly sorted and offer a striking contrast to the enclosing light brown pelagic matrix. The wackestone intervals are mostly composed of *in situ* basinal assemblages dominated by Globigerinid foraminifera. The remainder consists of comminuted, unidentified light-coloured microbioclasts embedded in a dark-coloured organic-rich micrite. The aggrading recrystallisation is widely recorded in the thin sections, where most of the mud (micrite) tends to aggrade into microspar.

Interpretation: This microfacies is interpreted as mud-rich tempestite deposits on the distal part of outer-ramp. They were probably deposited near the zone of maximum storm-wave reworking; consequently, the skeletal material was derived from the shallow sea-floor of the ramp-interior and redeposited as amalgamated storm deposits. Similar facies were described by Calvet and Tucker (1988) from the Triassic of Spain. This relationship and analogue with modern storm deposits (Aigner, 1985), suggest that deposition was in water depths between 20-40m (assuming a 40m storm-wave base).

The probable occurrence of fining-upward cycles could reflect the upward amalgamation of storm beds, which may be due to decreasing frequency of storms through sea-level fluctuations (c.f. Duke,1985). In addition to sea-level oscillations, climate and sedimentation could also have been affected by short period-high amplitude sea-level changes (Fischer,1991). The deep ramp facies would have been subjected to the effects of strong storm currents generated in response to swells originating in the Tethyan Ocean.

The packstone units, consisting of extremely comminuted and randomly oriented bioclasts, may represent amalgamated storm deposits which formed by reworking during high-energy conditions. In contrast, wackestone units represent either accumulation from low-energy suspensions (low density storm flows), or background sedimentation due to downwelling phases between storms (Lee and Kim, 1992). However, the low-diversity of the pelagic faunas in this interval indicates deposition in a stressed and relatively stagnant environment.

The neomorphism of the micrite may have occurred during the diagenesis of the overlying shale when magnesium ions from this carbonate microfacies were attracted to montmorillonite clays, so facilitating the formation of microspar (Longman, 1980).

2.1.4 Microfacies D: ***Bioclastic mudstone-wackestone***

Occurrence: This microfacies has a laterally continuous nature over the platform interior. This bioclastic microfacies occurs in the basal part of well G2-51 and in the central part of wells H1-51 and O1-82.

Description: This microfacies is mostly composed of scattered coarse sand-sized bioclasts, which are set in a muddy matrix and show repetitive fining-upward layers, that commonly grade up into lime mudstones. The majority of the bioclasts are foraminifera, echinoderms and bivalves, with an average size varying from 50-70mm and the maximum around 180mm. The coarse bioclasts have a loosely-packed fabric and show rare imbrication (Fig.2.2B).

Interpretation: This microfacies contains a heterogeneous mixture of allochthonous bioclasts, and can be interpreted as amalgamated storm deposits (c.f. Kreisa,1979). Thick continuous and lenses of bioclasts at the base of deepening-up successions or cycles, are likely to have been produced by successive reworking events, presumably related to large storm-generated waves and strong currents.

This microfacies is similar to one described by Banerjee and Kidwell (1991), where material was apparently produced by intermittent reworking of ramp-interior facies and deposition took place above normal storm-wave base. The concentration of bioclasts into continuous lenses could have been the result of high-energy storm-reworking due to the initial stage of sea-level rise, but more likely it was the result of reworking during peak and waning storm-surges. This conclusion supports the concept of lowstand shedding derived largely from work around the Bahamian platforms (e.g. Schlager, 1992). Continued storm events caused extensive re-mobilisation of bioclasts and their basinward transport. Deposition took place further offshore.

This proximal bioclastic microfacies shows evidence of multiple episodes of relatively high-energy reworking and lack of grading. This characteristic easily differentiates it from the overlying packstones. Aigner (1985) documented similar coarse-grained bioclastic storm beds and attributed their lack of grading to lateral bed-load transport. The coarseness and slight abrasion of the, open-marine skeletal material suggest reworking and transport of a locally-derived, shallow-ramp biota. It is unlikely that each bioclastic layer represents a single, random storm event interrupting predominantly muddy ramp interior.

The lime mudstone between bioclasts further supports an allochthonous interpretation of the skeletal material. This matrix was formed by deposition of storm-suspended fine sediments (c.f. Markello and Read, 1981).

2.1.5 Microfacies E: Bioclastic-benthic foraminifera wackestone-packstone

Occurrence: This microfacies occurs in the central part of the study area, and in well G2-51. It is interpreted as the result of formation of bioclastic banks. In this case there several banks, which have a low relief. This microfacies is very similar to that forming discrete skeletal banks seen in outcrops in NE Libya, and described by El-Hawat *et al.*

(1986). In addition similar facies occur in the subsurface of this region (unpublished intra-AGOCO reports) where more closely-space wells show a bank geometry.

Description: This microfacies in well G2-51 occurs at two horizons. Basically, it is made up of larger foraminiferal assemblages including nummulitids, operculines, discocyclines and rotaliids. These elements are set in a dark micrite matrix, locally with small unidentifiable benthic debris associated with a poorly defined peloidal texture.

The base of this microfacies is characterised by a densely packed accumulation of large benthic foraminifers (Fig.2.3A & B). Nummulitids (mainly of the species *Nummulites gizehensis*), operculines and discocyclinids are volumetrically the most important constituents. They are enclosed in a mixture of nummuclasts and dark lime muds. The larger-foraminifers have suffered flattening and crushing with the development of low amplitude irregular stylolites. Other skeletal constituents are comminuted open-marine macrofaunas dominated by molluscan bivalve shells and echinoderm and bryozoan debris.

Diagenetically, two generations of cement, were noticed. A rare fringe of fine equant calcite lines the nummulitid chambers. Equant to fine blocky calcite occurs as both intra- and inter-particle cement. Also this microfacies commonly shows abundant fabric and non-fabric selective pyritisation.

Interpretation: This microfacies is interpreted as a hydrodynamic skeletal bank similar to the allochthonous banks of Aigner (1985). Aigner related their origin to storm events in the zone between fair-weather and storm wave-base. He concluded that the top of the banks experienced higher energies than the remainder of the banks or the surrounding sediments. Based on the Libyan offshore analogues (Bernascon *et al.*, 1991) palaeowater depths for nummulitic shoals were probably on order of 25-35m. The nature and thickness of the two foram units suggest that the upper bank was located a little more platform-ward than the lower one.

Skeletal banks and sand shoals are commonly initiated during the transgression of shallow, low-energy ramps following a relative sea-level rise (Sbeta, 1990). The large-foram bodies probably developed as outer-ramp banks and they grew upwards above wave-base. The vertical change in texture would have been in response to deepening conditions. Lateral growth led to the progradation of flanking beds of broken debris on the slopes of the banks. Storms may have been important in the formation of the banks.

Ahmed (1992) concluded that the development of skeletal banks in Palaeocene and early Eocene sections of the Agedabia-Augila Trough, on the eastern edge of the Sirte Basin, Libya, occurred in two stages: bank formation and bank flooding. Two deepening-upward cycles which culminated in low-energy foram shoals are encountered in well G2-51. With the continuation of a short-term relative sea-level rise associated with influx of large foraminiferal assemblages, foram-banks were able to establish themselves on stable substrates under moderate-energy conditions. The banks are interpreted to have aggraded vertically in response to available accommodation space. Stacked and deepening-upward trends within each bank represent repeated adjustment to changes in base level.

The banks are laterally discontinuous, suggesting development as isolated subtidal shoals. There is no evidence that the bank crest was ever above fair-weather wave-base in a high-energy setting during deposition. Probably under the action of wind-induced waves, storms periodically effected the bank-top communities, and the foram tests were reoriented into a more stable state. The crushing and flattening which led to the formation of an overpacked fabric are interpreted as the consequence of compaction resulting from overburden pressure. The sediment of the bank core was supported by internal marine sediments. The precipitation of the fine isopachous fringe cements was probably syndepositional in the active marine phreatic environment (Longman, 1980).

Bernasconi *et al.* (1991) described similar foram banks developed upon topographic irregularities (of tectonic origin) of the underlying strata with very little lateral continuity. They are on the order of tens of meters in thickness and a few hundred meters across in a relatively low-energy, muddy setting. These shell banks are known from the Lower Eocene of Libya and have been studied in the offshore El Bouri field NW of Tripoli, in the Mediterranean Sea. The foram banks are superimposed on each other and generally occur as discrete banks. In some cases they have a high density of occurrence, but they do not form continuous barrier islands.

2.1.6 Microfacies F: Peloidal-bioclastic wackestone-packstone

Occurrence: This microfacies has the greatest volumetric abundance in the overall facies framework and is extensively developed throughout the study area, but particularly in the western and central parts. This microfacies is encountered in wells AA1-6, O1-12, G2-51 and H1-51. In the western part of the area it has the overall shape of a wedge, thinning towards the most westerly and central sectors. In the eastern portion particularly in well H1-51, it capped by thick-massive dolomite microfacies. This microfacies is thickly developed in wells AA1-6 and O1-12.

Description: This microfacies (Fig.2.4A) typically consists of mud and fine skeletal debris in a grain-supported fabric (wackestone -packstone), which is loosely to closely packed. The biota is diverse and dominated by a robust, normal-marine fauna. These assemblages consist of both macro and micro-organisms.

In general, the texture of the matrix varies from grain to mud support fabric, even within a single thin section. In well AA1-6 homogeneous peloidal lime muds are associated with nummuclasts (i.e. broken nummulitic tests). In O1-12, the matrix is partially neomorphosed and farther east the microfacies is more completely neomorphosed to a peloidal grainstone.

Echinoderms dominate the macrofauna and contribute much of the skeletal debris. Some of the fragments show signs of etching and the larger ones commonly have a thin

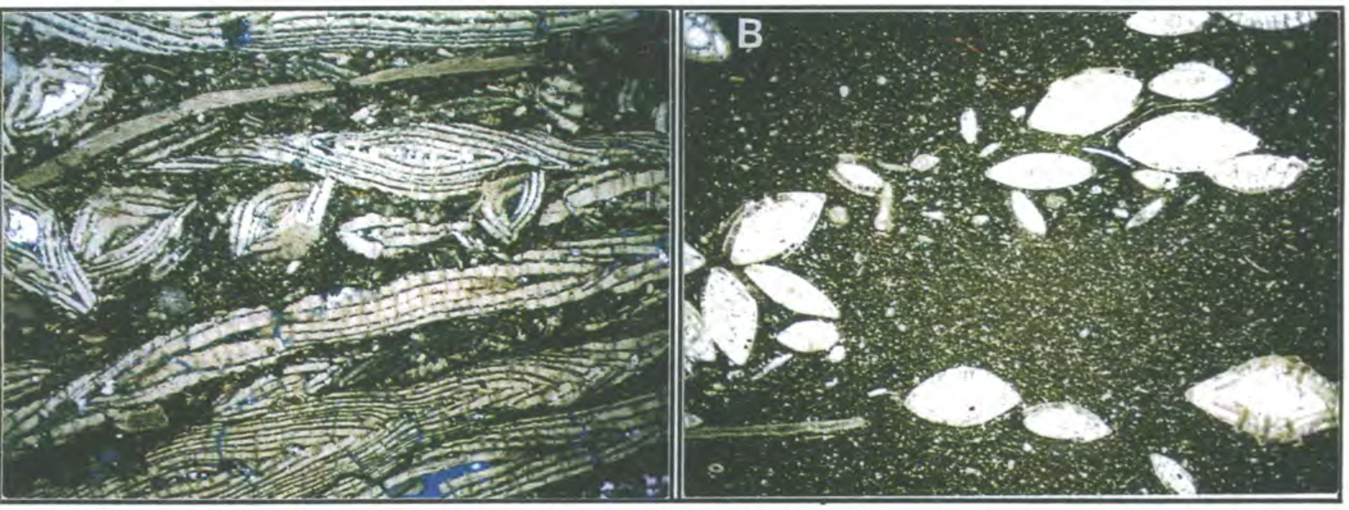

Fig.2.3 *Representative photomicrographs of different microfacies in the Tamet Formation. All photomicrographs were taken under plane polarised light at X 40 magnification. A) Bioclastic-benthic foram wacke-packstone of the proximal outer-ramp environment, this microfacies represents a bank formation and made up of larger foraminiferal assemblages including nummulitids, operculines, discocyclines and rotaliids. These elements are set in a dark micrite matrix, locally with small unidentifiable benthic debris associated with a poorly defined peloidal texture The common occurrence of large foraminifera suggests that the depositional environment might have been under the influence of frequent high-energy conditions, probably storm activity. Sample from well G2-51, depth 936m (3070ft). B) Bioclastic-benthic foram mud-wackestone of the proximal outer-ramp enviroment, this microfacies is considered as the flooding stage of the bank evolution, and consists entirely of small size benthic foraminifera. In this microfacies the water depth became deeper and less energetic than the underlain microfacies. Sample from G2-51, depth 920m (3020ft).*

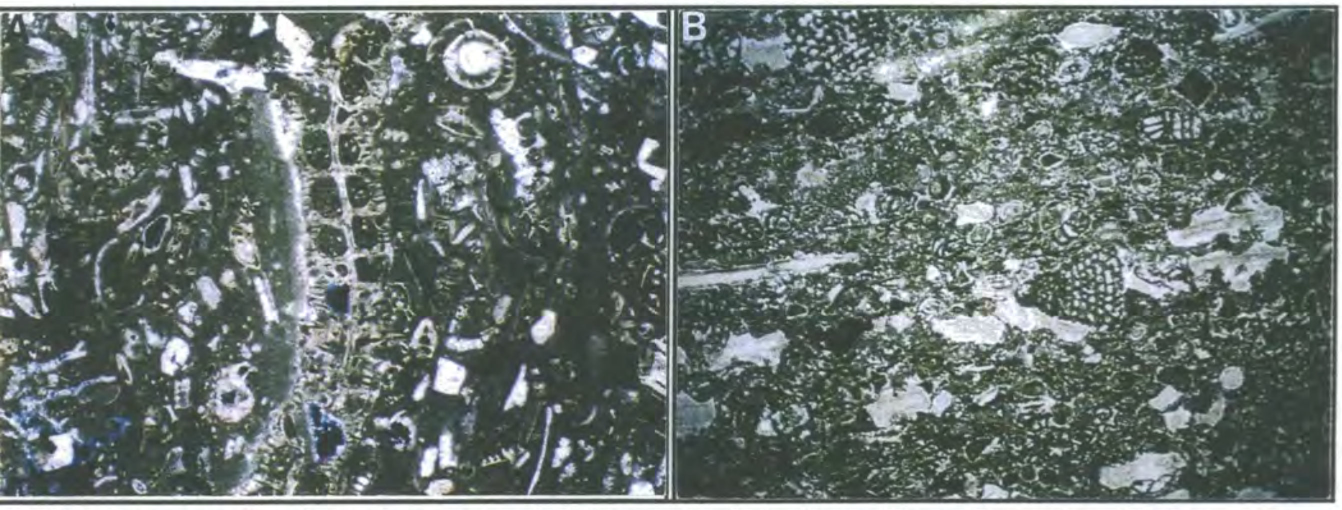

Fig.2.4 *Representative photomicrographs of different microfacies in the Tamet Formation. All photomicrographs were taken under plane polarised light at X 40 magnification. A). Peloidal-bioclastic wacke-packstone. This microfacies is the most abundant in Tamet Formation, showing a diverse open-marine macrofauna. Benthic forams, bryozoans and echinoderm debris are the most common bioclasts. Peloidal lime mud is the usual matrix. This microfacies is interpreted to have been deposited in a low-energy, proximal outer-ramp environment probably below fairweather wave-base. Sample from well O1-12, depth 1750m (5740ft). B) Orbitolinid-echinoderm wacke-packstone of the inner-ramp environment, showing admixture of normal-marine skeletal debris with common fine sand-sized peloidal muds. The faunal community was made up of miliolids, echinoderms, orbitolinids as well as small forams such as rotaliids. This microfacies is interpreted as a shallow-marine deposit with open circulation. Sample from well G2-51, depth 964m (3160ft).*

irregular syntaxial overgrowth; these cements are inclusion-free. Other notable constituents include fragmented shells and fenestellid bryozoans. In some cases, they are relatively complete specimens; they are usually large and well preserved. Most of the chambers of the bryozoans are filled initially by peloidal muds and these are followed by equant to fine blocky cements, giving geopetal fabrics. Farther eastward sandy blankets of molluscan debris become more abundant.

There is also a relatively diverse accessory biota comprising finely comminuted foraminifera distinguished by the presence of small rotaliids, although large foraminifera such as nummulitids (*Nummulites*) and miliolids are also present. The non-skeletal components consist predominantly of micritic grains (*in situ* peloids); they are uniform in size (20-100 μm) and mostly ovoid in shape.

Interpretation: This microfacies consists of two types of sediment, one deposited in a basinal environment and the other derived from a shallow-water environment. Sediment deposition occurred in an open-marine, subtidal setting in waters of moderate to shallow depth (20-40m). Generally the sediment surface was below fair-weather wave-base and possibly above storm wave-base (cf. Elrick and Read, 1991). This interpretation is strengthened by thick-shelled normal-marine fauna and absence of any degree of sorting. The abundance of peloidal muds indicates that wave and tidal current activity was minimal.

However, periodically, there were influxes of the more shallow-water skeletal grains. This skeletal debris may have been derived from scattered carbonate sand shoals and banks, formed on local topographic highs on the ramp interior. These might have been under the influence of frequent, high energy storm conditions. However, an overall modest level of water turbulence may have promoted its development by stabilising the depositional surface over a wide area on the ramp, probably the result of fair and stormy weather alternations.

In many vertical sections of this microfacies, it appears that there is an upward increase in bioclasts and decrease in mud content. This is repeated several times, and the units probably represent shallowing-upward trends. It appears that the sediments are all subtidal; there is no strong indication of intertidal or supratidal features. It is also possible that overall within this microfacies, there is a shallowing-upwards: the two developments of this microfacies in well O1-12 and the single occurrence in well G2-51 both show a general increase in bioclasts/decrease in mud upwards. In addition, it appears that the hypersaline diagenesis effects are better developed toward the top of each microfacies occurrence; these include dissolution enhancement phenomena and interparticle and moldic pores, some of which are now occluded by cement. Neomorphic crystallisation fabrics are probably linked to burial diagensis (cf. Crevello, 1991).

The peloids in the cavities of bryozoan skeletons are generally too small and ill-defined to be microbial precipitates. There is no evidence of micritisation of skeletal grains which would support an origin through alteration of bioclasts (cf. Bridges and Chapman,1988). Aggrading neomorphism of the matrix is particularly well developed in this microfacies. It is commonly interpreted as a typical feature of fresh-water phreatic diagenesis and here it may have been initiated under the action of meteoric influxes through the intraformational unconformity surfaces. The syntaxial overgrowth cements on the echinoderm grains may initially have been precipitated during shallow burial and compaction of the sediments (Moore,1989).

2.1.7 Microfacies G: Orbitolinid-echinoderm wackestone-packstone

Occurrence: This microfacies occurs only within well G2-51. It is the least common of the Tamet microfacies and is surrounded by and closely associated with microfacies G. It apparently occurs as lenses of limited vertical and lateral extent. This microfacies is developed as a unit near the centre of the study area.

Description: This microfacies is characterised by mixed normal-marine skeletal debris with less common fine sand-sized peloids, which commonly are structureless. The biotic constituents are commonly aligned in the mud-rich matrix (Fig.2.4B). They gradually coarsen-upward into peloidal-echinoderm grainstones. The depositional fabric varies from loosely compacted at the bottom, grading up into local concentrations of bioclasts.

The matrix mainly consists of dark silt-sized comminuted shell material which shows evidence of neomorphism to microspar especially in the lower part. The faunal community of this microfacies was made up predominantly of echinoderm bioclasts, in the form of spines, ossicles and plates, commonly without micritised rinds. The upperpart of this microfacies displays an early pre-compactional cement of syntaxial rim overgrowths. The syntaxial rim shielded the echinoderm nuclei and adjacent grains from subsequent intergranular compaction. The additional bioclasts include orbitolinid foraminifera, as well as small foraminifera such as rotaliids. Minor constituents are bryozoan and molluscan fragments; miliolids are less common.

Interpretation: The characteristics of this microfacies: abundant normal marine fauna, a mud-rich framework and overall decrease upward in admixed silt-size bioclasts combined with a lack of sorting, indicate deposition in normal salinity under moderately shallow, oxygenated subtidal conditions. However, the local preferential alignment of allochems and the layers of stacked shell debris are attributed to deposition by bedload processes. This type of microfacies is generally deposited in low-energy, shallow subtidal setting on inner ramp lagoonal environments at water depths below wave base and may form sheet-like stratigraphic belts that are tens to several hundreds of kilometres wide (cf. Elrick and Read, 1991).

The echinoderm debris could have been derived from banks and shoals randomly developed across more distal parts of the inner-ramp by the action of storm and tidal currents. The peloids in this deeper water facies could have been reworked from their original environment of formation in an open-marine, shallow subtidal setting. An alternative explanation is that the peloids were formed *in situ* by microbial alteration of bioclasts in the restricted lagoon.

Comparison with modern peloidal lime muds from south Florida and the Great Bahama Bank suggests that they have undergone significant diagenetic alteration. Microspar-sized crystals have aragonite relics, pits and irregular boundaries. These features could very well have formed in a meteoric diagenetic environment (cf. Lasemi and Sandberg, 1984) or during burial diagenesis.

In the uppermost part, echinoderm debris has usually been cemented before compaction occurred; in other cases a lack of early cement led to a flattening of grains parallel to the stratification and the development of an overpacked fabric. This did not happen in echinoderm grainstones because the framework supplied by the cementation was strong enough to resist the overburden pressure (cf. Peryt, 1987).

2.1.8 Microfacies H: Peloidal-miliolid wackestone-packstone

Occurrence: This microfacies is laterally extensive with a more or less uniform thickness and appears to thin towards the eastern and western parts of the area. This microfacies is encountered in wells G2-51 and H1-51.

Description: This microfacies is very similar to the previously described Orbitolinid-echinoderm microfacies. Texturally, it is dominated by a mud-support fabric (Fig.2.5A), but muddy grainstones are also developed locally. The biotic constituents in this microfacies are scattered within a dark-coloured fine bioclastic matrix. The relatively diverse benthic foraminifera include miliolids (which dominate over other grain types), orbitolinids, alveolinids and textulariids. The recognisable macrofossils include molluscs, ostracods, bryozoans and echinoderms, but they are predominantly found as broken shells. The faunal community in the younger part of the section is more restricted and not very diverse. It is characterised by a greater abundance of peloids. The peloids show variable size and are mostly irregular to spherical in outline and internally structureless.

Interpretation: The fabric and texture of this microfacies reflects a fluctuation from open-to-semirestricted shallow subtidal environments, probably developed in water depths of a few metres (cf. Ahmed, 1992). The basal part of this microfacies was deposited in low-energy, open-marine, shallow-subtidal conditions as suggested by the dominance of a rich fauna and mud-rich matrix. However, the higher part was deposited in less turbulent and more restricted shallow subtidal conditions as suggested by the presence of miliolids and peloids and the occurrence of similar facies in the Palaeocene section in the northeastern edge of the Sirte Basin (Mresah, 1993).

The close association of orbitolinid-echinoderm microfacies and this microfacies is not surprising since the two microfacies probably represent similar depositional settings under the influence of palaeo-water depth oscillations and different degrees of restriction. The peloidal-miliolid microfacies would have been deposited in a more restricted location, with occasional influences of echinoderm and open-marine foraminifera. The nature and origin of peloids are quiet variable; some larger types and irregularly shaped ones are probably micritised fragments of molluscs and algal stems. Shinn and Robbin (1983) pointed out features similar to those of this microfacies in subtidal peloidal muds deposited without the aid of sea grasses in the shallow lagoons of Florida Bay, where storm-generated currents episodically reworked areas of mud.

2.1.9 Microfacies I: Dolomitised peloidal-bioclastic wackestone-packstone

Occurrence: This microfacies is widely recognised in the central and eastern parts of the study area on the Cyrenaica Platform. The distribution of dolomite is shown in Fig. 2.1. The vertical transition between the dolomite body and surrounding limestone varies from well to well and also is noted by Ahmed (1992), who described two different petrographic types of dolomite. However, this microfacies is divided into submicrofacies based upon the characteristics of their texture and fabrics; which dolomite occurs primarily as a replacement mineral and as cement.

2.1.9a Submicrofacies i: Coarse dolomite

Description: Several units of replacive dolomite occur as vertically discontinuous intervals throughout large sections of both wells H1-51 and O1-82. Non-dolomitised microfacies appear to thin out and pass laterally into anhydrite facies in an eastward (landward) direction towards well O1-12. This submicrofacies petrographically ranges from fabric-destructive to fabric-retentive (Fig.2.5B) on a microscopic scale. It consists of a mosaic of tightly interlocking, typically non-planar crystals (Sibley and Gregg, 1987), which increase towards the Cyrenaica platform. The crystal sizes are from 100-700 µm in diameter, with subhedral to anhedral morphologies. Some crystals have intracrystalline microfractures and are associated with solid inclusions of unknown composition which are randomly dispersed. These crystal have faces which are rough and display embayments.

The most common porosity types are fabric-selective moulds, their shapes suggesting precursors which include peloids and possible small benthic foraminifera, while vuggy and intercrystalline areas are well developed regardless of depositional fabric. A baroque dolomite occurs as coarse intercrystalline void-filling cement. This is analogous to the xenotopic-C dolomite of Gregg and Sibley (1984) and is characterised by coarsely-crystalline rhombic crystals (up several mm), with undulose to extremely undulose

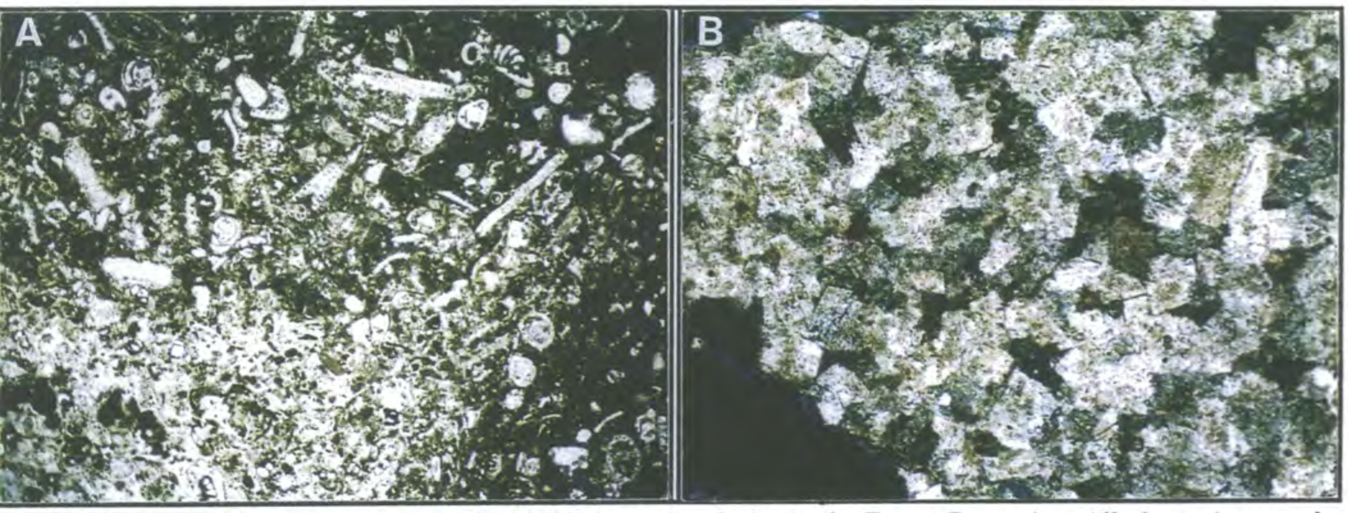

Fig.2.5 *Representative photomicrographs of different microfacies in the Tamet Formation. All photomicrographs were taken under plane polarised light at X 40 magnification A) Peloidal-miliolid wacke-packstone of the inner-ramp environment, the relative diverse benthonic forams include miliolids, alveolinids and textularids. The matrix is mainly fine sand-sized peloids, with admixture of bioclasts. The close association of miliolids and peloids indicates that the deposition was in a semi-restricted lagoonal environment. Sample from well G2-51, depth 960m (3150ft). B) Coarse crystalline dolomite microfacies; displaying a tightly packed mosaic of mostly anhedral to subhedral dolomite with irregular crystal boundaries and slight undulose extinction. Crystal showing cloudy centres and clear rims. Sample from well G1-51, 730m (2394ft).*

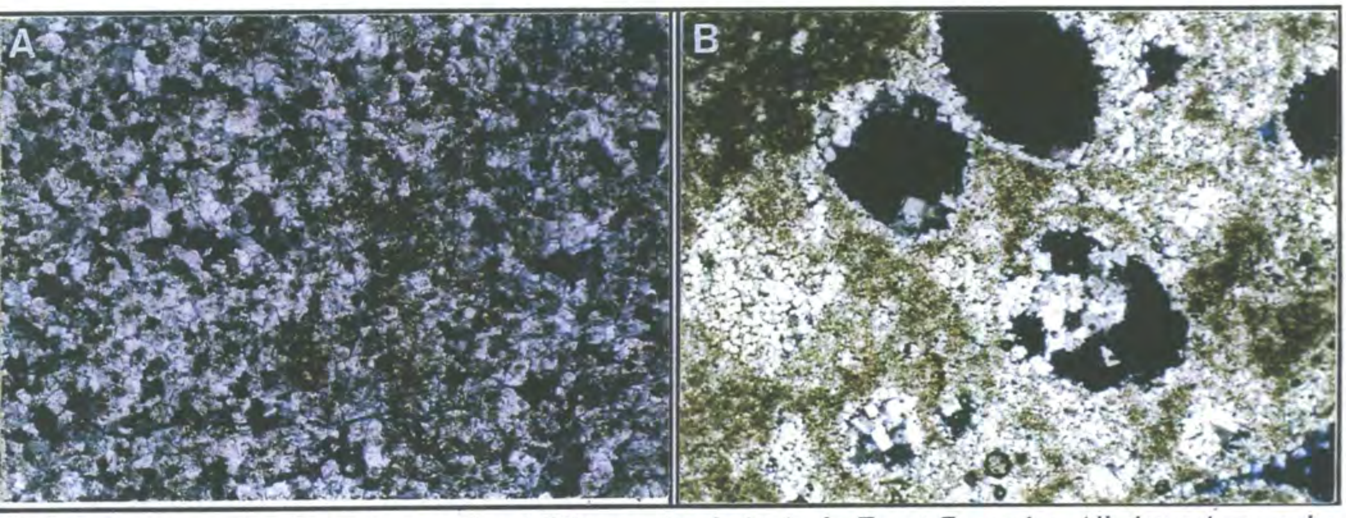

Fig.2.6 *Representative photomicrographs of different microfacies in the Tamet Formation. All photomicrographs were taken under plane polarised light at X 40 magnification A) Fine crystalline dolomite microfacies, composed of an equigranular, anhedral dolomite displying sutured mosaic fabric with a little intercrystalline porosity. Some crystals contain opaque nuclei. Sample from well O1-12, depth 1417m (4650ft). This study indicates that Tamet dolomitisation formed by reflux of hypersaline fluids generated within the Cyrenaica Platform. The various types of dolomite can be explained by conduit systems controlling the flow of the dolomitising fluids downdip. B) Peloidal-foram dolowacke-dolopackstone of the Cyrenaica platform environment has pel-foram moldic porosity (black). Moldic porosity formed through dissolution by the hypersaline fluids. Original shape of peloids can be discerned due to mimetic replacement. Sample from well O1-82, depth 395m (1295ft).*

extinction. Some of the late-stage dolomite precipitation appears to have coincided with hydrocarbon generation. The remaining open space is either filled with bladed anhydrite cement or is open.

2.1.9b Submicrofacies ii:Fine dolomite

Description: This submicrofacies commonly shows a densely-packed mosaic of dolomite crystals (Fig.2.6A), formed by the replacement of pre-existing sediment. The original fabric of this microfacies has been completely obscured by extensive dolomitisation. Depositional textures are mostly lost, but in some cases, faint ghosts of biogenic grains and peloids can be discerned within the dolomite mosaic. This dolomite is homogeneous in appearance, with an overall crystal size of more than 0.1mm in diameter. The crystals are mostly unimodel, equigranular, non-planar (xenotopic), grading into crystals with planar faces, sub- to anhedral shape, and extinction varying from straight to undulose. Intercrystalline boundaries are mostly irregular, serrated and in some places they have a microstylolitic appearance.

Fine intercrystalline pores in the generally dense mosaic are rarely developed, but where present they have been enlarged by dissolution and lined by euhedral, saddle dolomite cements with planar, rhombic terminations as the initial phase of cementation (cf. Amthor and Friedman, 1991). In some cases there is residual solid oil. Anhydrite occurs in close association with the late-stage dolomite cement, and has two distinctive modes of occurrence. Most commonly, the anhydrite forms micronodules, is milky-white and composed of fine felted laths less than 100 μm long. The other more common type of anhydrite occurs as a cement. It is composed of fine to medium, bladed and equant crystals up 2mm in size, and partially or totally occludes moldic and vuggy porosity.

Interpretation: The pervasive dolomitisation, leaching of metastable skeletal components and generation of moldic, vuggy and intercrystalline porosity, may have been formed by the lateral migration of dolomitising fluids onto the shallow platform during the rise of brine levels rather than through fresh groundwaters (Sun,1992). It could be controlled by sea-level rises over the Cyrenaica Platform. This interpretation is based on the assumption that dolomitisation developed during deposition decreases off as eustasy peaks accompanied by long-term arid climatic conditions and brine reflux (this discussed below).

Karsts and palaeosols are recorded beneath subaerial exposure surfaces at outcrops in northeast Libya. Most of the subaerial dissolution, karstification and palaeosols and associated calcite cementation affecting these facies in outcrops are considered to have been developed during post-Middle Eocene times (El-Hawat *et al.,* 1986). Dolomite distribution bears no direct relationship to depositional patterns and is formed subtidally in contact with the evaporitic platform. It does not appear to be restricted to a specific microfacies.

2.1.9c Mechanism of dolomitisation

The principal requirements for large-scale replacement dolomitisation are the continued import of magnesium ions, a suitable chemical environment and long-term fluid circulation. Similar textures to those encountered in this study were described by Ahmed (1992) in Palaeocene and early Eocene dolomites of the Agedabia-Augila Trough, on the eastern edge of the Sirte Basin. The geological literature abounds with models to explain the mechanism of dolomitisation, but in this case the most likely model is that involving seepage reflux of high Mg/Ca, hypersaline solutions formed on a broad platform by evaporation. Carbonate-evaporite sediments being deposited today on the hypersaline tidal flats of the Arabian Gulf area are good analogues for the reflux model (Gunatilaka, 1991).

Major changes in the type of deposition took place during time of restriction or complete isolation of the Cyrenaica Platform, when the platform was cut off from open-marine conditions under a dominantly arid climate, as indicated by the absence of meteoric vadose diagenesis, scarcity of palaeosols and pervasive dolomitisation and anhydritisation. In this case a gradual increase in marine restriction could have been due to a sill, barrier or some other topographic high. This lead to the development of a shallowing-up carbonate-evaporite succession infilling a depression in the landscape. This unit formed over the broad regional hinterland that is now called the Cyrenaica platform. This area was an extensive peritidal hypersaline mud flat. It was characterised by long periods of brine concentration, although there were frequent floods by high spring and storm tides (Ahmed, 1992).

As the evaporite sediments were deposited upon the Cyrenaica platform, hypersaline brines (salterns) would have formed, and these could have served as a source of magnesium ions for the dolomitisation. The dolomite is localised along the flanks of this evaporitic region with dolomites passing laterally into limestones and then into more basinal facies along the seaward margin of the Tamet ramp.

The extensive pumping of fluids necessary for pervasive dolomitisation can be caused by the movement of subsurface fluids induced by the descent of hypersaline brine. The large migration of fluids is interpreted to have been driven by uplift of the eastern part of the area and sedimentary loading. The available source is believed to have been three fold: adjoining sea-water (driven onto the saltern during times of supratidal flooding), interstitial aquifers and meteoric water (Migaszewski, 1991).

Under a semi-arid/ arid climate, fluids beneath a saltern are concentrated through evaporation and reach saturation with respect to calcium sulphate, first as gypsum and then as anhydrite. The Ca^{2+} / Mg^{2+} and Ca^{2+}/ CO_3^{2-} ratio drop drastically through sulphate precipitation. During the infrequent rainfall, flooding sea-water (major storms) may move far landward, where both fresh and salt waters pond on the saltern. The infiltration of these waters causes net downward flow.

Refluxing brines continue to flow seaward, and this provides sufficiently high Mg^{2+}/Ca^{2+} ratios (many times greater than normal sea-water). Temperatures are in the range of 30-50°C and there is also an accompanying elevation in alkalinity. Where the efficiency of hydrologic pumping forces the brine level lower and sulphate concentration reaches saturation, the alkalinity of the subsurface fluids may be neutralised through changes in the ionic composition and ionic strength, as a result of mixing with any natural waters. This may have been responsible for dissolution of skeletal aragonite. Consequently, magnesium saturation is probably achieved and high Mg^{2+}/Ca^{2+} ratios appear to be the dominant geochemical factor controlling dolomitisation.

Even if a suitable dolomitising solution is present, to produce large quantities of dolomite there must also be lowering of both the SO_4^{2-} and Ca^{2+} concentrations in the refluxing fluids by extensive gypsum precipitation and a significant flow rate of these fluids within the optimal zone (Gunatilaka *et al.*, 1985).

Dolomitisation becomes more significant and pervasive towards the Cyrenaica Platform, from which the dolomitising fluids were derived, and dies out in a basinward direction. There is no obvious permeability barrier. Perhaps the driving force for fluid movement ceased (probably as a result of differences in porosity and permeability) or the dolomitising potential of the fluid was spent (cf. Given and Wilkinson, 1989).

The selective dissolution which often accompanies dolomitisation can occur on a very local scale, and involve high magnesium calcite or aragonite bioclasts. This dissolution generally was mineralogically selective and formed significant secondary pores, although in some cases post-dolomitisation dissolution enlargement accounts for most of the vuggy porosity. Petrographic observations suggest that aragonitic skeletons, such as echinoderms, and benthonic foraminifera are commonly mimetically replaced by dolomite and some of them were dissolved away. Influx of fresh waters as a result of regional subaerial exposure has been invoked as the cause of aragonite leaching (Armstrong *et al.*, 1980; Coniglio *et al.*, 1988). However, such an explanation is not supported by the lack of prominent karstic features and the paucity of meteoric calcite cementation. Sun (1992) proposed that large-scale dolomitisation, and dolomite cement lining skeletal-moldic pores, together with anhydrite precipitation, suggest that aragonitic skeletons were dissolved by dolomitising fluids (hypersaline brines). The amount of intercrystalline porosity is controlled by two factors: the presence or absence of intercrystalline cementation and the arrangement of dolomite crystals. However, both microfacies of dolomites are dominated by one or the other of these factors and illustrate the relationship between dolomite textures and intercrystalline framework.

Intracrystalline truncation features are present in the dolomite crystals as well as microcavities. These features are probably the result of closed-system burial diagenesis and a stable ratio of brine and residual marine carbonate in the presence of connate fluids at elevated temperature and low water/rock ratios. It is possible that the late dissolution resulted from hydrothermal fluids during burial and compaction of the sediments (Mazzullo and Harris, 1992).

The most dramatic effect of the interaction of connate fluids with the dolomites was the creation of locally significant mesogenetic dissolution vugs. The lack of significant deep burial products such as dedolomites is probably because the system quickly approached chemical equilibrium (mesogenetic stabilisation) with surrounding carbonate rocks (Mazzullo and Harris, 1992).

Hydrocarbon occurrences are commonly present adjacent to late-stage dolomite and anhydrite cement. Furthermore, late dolomitisation coincided with stylolitisation. These striking associations imply late diagenesis and are interpreted as burial cements (cf. Kaufman *et al.*, 1990). This is consistent with the subsidence of early pervasive dolomites through post-Middle Eocene (Gumati, 1984). These conclusions also support earlier notions (e.g. Ahmed, 1992) that saddle dolomite is most likely a vug-filling precipitate rather than a replacement.

The magnesium ions required for burial dolomite cementation probably came from an internal source (e.g. Ahmed, 1992). It is likely that minor amounts of Mg^{2+} ions were derived from the local host dolomite succession, probably liberated by penecontemporaneous stylolitisation. There is also the possibility that dissolution of the initial dolomite framework (as discussed earlier) under the influence of gravity-driven pore-water flow, may have carried a sufficient concentration of magnesium ions.

In the geological literature, there are several hydrological models for burial fluid migration and dolomitisation. One popular model involves topographic-driven fluids generated during continuous subsidence, where the basin is bordered by an uplifted active recharge area. A steep hydraulic gradient for ground water flows away from uplands is established passing downdip into the burial environment. However, the large quantities of fluids with lower degrees of supersaturation may produce only minor amounts of late-stage dolomite. Dolomitising fluids may also be produced by the mixing of near-surface meteoric waters and formation waters. An extensive fluid flow system is required to move and redistribute the Mg^{2+} ions in the fluids.

Two different modes of anhydritisation have been distinguished. The first occurrence is as micronodules of anhydrite in dolomite facies. These nodules probably formed within shallow subtidal sediments and indicate periods of intense evaporation. The modern hypersaline tidal flats of the Arabian Gulf are good analogues (Shearman, 1978). The second occurrence is as minor amounts of cement. Based on its position in dolomite-hosted moulds and vugs, and using timing relations established from the order of vug-filling phases, it post-dates replacement and saddle dolomites. The ultimate source of the anhydrite is not known, but it is probably locally derived from the first-stage anhydritisation.

2.1.10 Microfacies J: Anhydrite

Occurrence: This microfacies is widespread over the Cyrenaica Platform and found throughout the subsurface in AGOCO wells. It becomes thinner and disappears towards the Tamet platform as a result of a change to carbonates. This microfacies is encountered in well O1-82.

Description: The middle Eocene strata occurring farther east in well O1-82, generally consist of repeated shallowing upward carbonate-anhydrite cycles. Carbonate increases in thickness and frequency in a westward (basinward) direction, whereas the amount of anhydrite ranges from zero in the western part of the study area to equal to the carbonate's thickness in the eastern portion. Several stacked cycles occur within the section in well O1-82.

Each cycle can be divided into two microfacies arranged in a shallowing-upward succession. Dolowackestone-dolopackstone makes up the basal part of each cycle, and is very similar to the previously described coarse dolomite microfacies except that it contains skeletal fragments (Fig.2.6B). Other constituents include varying amounts of restricted marine foraminifers such as miliolids. In the basal part of each cycle are anhydrite rip-up clasts. This subunit grades vertically into peloidal dolomites lacking skeletal debris. Porosity in the dolomite is usually either moldic or vuggy, and the pores are now usually filled with either anhydrite or coarsely crystalline dolomite cement.

The anhydrite subunits are predominantly well-bedded. In some cycles, the anhydrite is massive and up to several meters thick. Pseudo-nodular forms pass upward into nodular-bedded anhydrite, in the upper part, consisting of irregularly shaped centimetric nodules interbedded with dolomite.

Interpretation: An extensive shallow epeiric sea formed a large inland seaway under the arid conditions, so that an area of extensive platform evaporites was produced. This type of evaporite deposition was most common during times of long-term tectonic quiescence. Sediments were deposited and stacked as hypersaline platform cycles to form shallowing-upward units to 50m thick. Cyrenaica Platform cycles often had open-marine or restricted marine carbonates at their bases, passing up into saltern evaporites. These associations strongly suggest evaporite accumulation under conditions of isolation and restricted circulation. The flat-topped platform was wide enough to generate the hydrographic restriction necessary to trigger evaporite deposition on the Cyrenaica Platform, which prograded over the adjacent Tamet ramp.

Saltern is a new term to describe extensive shallow-subaqueous evaporites that formed continuous depositional units across hundreds of kilometres (Warren, 1989;1991). This type of evaporite environment does not occur today. In the geological literature, some previous investigators believed that laterally extensive evaporites were deposited subaqueously in lagoon and coastal plain environments. Examples include the basin-wide late Miocene (Messinian) evaporites of the Mediterranean (Cita, 1983), the platform evaporites of the Cretaceous Ferry Lake Anhydrite in the Gulf of Mexico (Pittman, 1985), the platform sulphates of the Permian Zechstein Basin (Taylor, 1984) and the ramp evaporites of Palo Duro Basin (Hovorka, 1987) all of which had been previously interpreted as extensive brine pond deposits.

On the Cyrenaica Platform there were widespread, coastal plain-lagoonal complexes which extended landward for hundreds of kilometres. A topographic barrier probably separated the saltern from the open sea, and this may have been a series of coastal beach-ridges or a tectonic barrier such as a horst block. Roughly equal rates of subsidence and sedimentation led to the development of an aggradational succession, similar to that developing along the coast of the present-day Arabian Gulf, although there basinward progradation has taken place.

The Cyrenaica saltern is an aggrading-prograding succession with a shallowing-upward sequence from subtidal mudstone containing marine biota deposited in a restricted subtidal lagoon, into thick, massive evaporite beds, which are interpreted as hypersaline microfacies probably deposited during the peak of a relative sea-level highstand associated with extremely arid conditions. The cyclic repetition is due to the fluctuating levels of hypersaline water (Warren, 1989). Subsequent restriction either related to very limited accommodation space over the Cyrenaica Platform or the accretion of barrier complexes, leads to the evaporitic stage and the complete isolation of the Cyrenaica Platform from the open-sea. The thickening of middle Eocene anhydrites towards the Cyrenaica Platform matches a significant transgression that occurred at the end of the time period in which the Tamet Formation was deposited and the evaporites are interpreted as saltern deposits. The frequent resupply from storms and seepages keeps the saltern from completely drying up. A phase of extreme aridity necessarily results in a decrease in the surface water of the basin, and this would have led to the development of sabkhas, especially around islands, as suggested by the local occurrence of a thin capping of nodular anhydrites, and these represent the final deposits of the saltern.

2.2 Depositional model and cyclicity

Physiographically, the Tamet platform is part of the southern margin of the Tethyan Ocean but it is not a single platform but rather part of a carbonate megaplatform located in the eastern Sirte Basin region throughout Palaeogene time.

The depositional model proposed for the Tamet Formation is similar to a homoclinal-type ramp of Read (1985), based on microfacies characteristics and stratigraphic position. This suggests that deposition took place on a low-energy, shallow- to deep subtidal ramp subject to episodic high-energy storms. The Tamet ramp differs from the classic model described by other authors in several ways. Most ancient ramps studied to date have an inner terrigenous clastic facies and a continouos reef trends (Ahr, 1973; Tucker and Wright, 1990; Stanton and Flügel, 1995).

Microfacies characteristics are summarised in Table 1. Due to the failure to identify the distinctive boundaries between the microfacies, the large-scale lateral relationships across the Tamet ramp-to-basin transition are constructed on the basis of systematic vertical changes in the characteristics of the metre-scale cyclicity, rather than the internal framework of microfacies. The geometries and stratigraphic position of most cycles which overall show a shallowing-upward trend, are shown in Figure 2.7A. The combination of all these approaches has led to the construction of a 2D-depositional model according to Walther's Law (Fig.2.7B).

Based on microfacies types and their lateral and vertical arrangement in the overall system, the ramp appears to have been partitioned into three major facies associations which in general way, are related to the ramp morphology. These facies associations are broadly mud-dominated reflecting deposition in variety of low-energy, subtidal environments (5-100m depth, Wilson, 1975). Foraminifers, molluscans, bryozoans and echinoderms are all indicative of environment with open-marine circulation (Wilson, 1975). Lack of open-marine macrofauna (molluscans and echinoderms) in the top of the

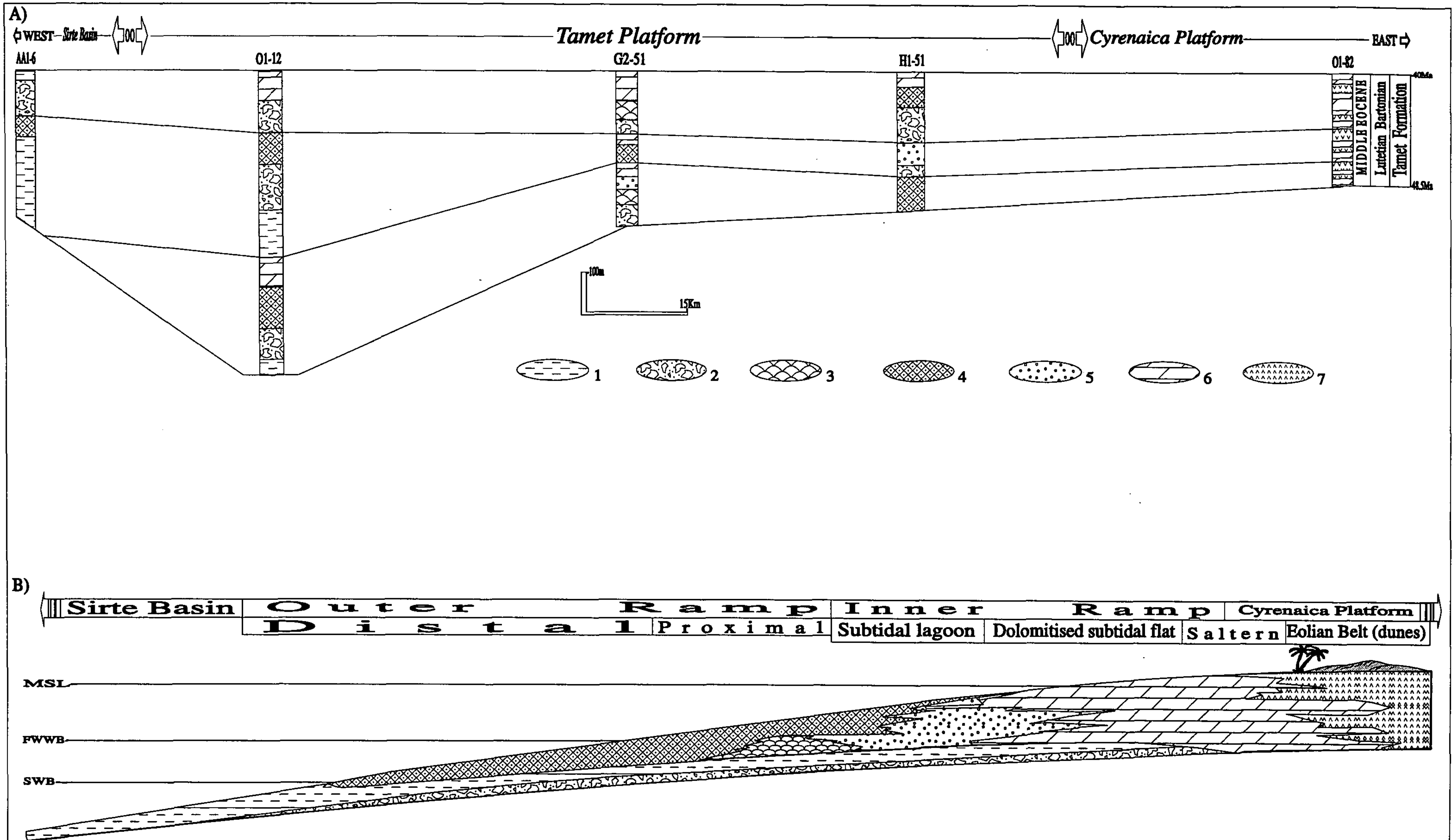

Fig.2.7 *A) Dip-oriented correlation diagram, showing the lateral and vertical relationship of correlative metre-scale cycles within the Tamet Formation. B) Schematic diagram showing the main morphological elements of carbonate ramp in an arid-zone bordered by an evaporitic platform and open Tethyan Ocean. Also shown is the facies architecture for single third-order, depositional supercycles. These facies together make up aggradational-progradational carbonate-evaporite cycle sets.*
Key 1- Deep subtidal cycles, 2- Storm-dominated cycles, 3- Foram-bearing cycles, 4- Open-marine, shallow sutidal cycles, 5- Subtidal lagoonal cycles, 6- Dolomitised subtidal flat cycles and 7- Evaporitic platform cycles.

inner-ramp lagoonal deposits suggests slightly restricted shallow subtidal environments. Based on the dark spiculitic-planktonic mud-wackestones and lack of shallow-marine macrofossils deposition also took place in deep subtidal, moderately oxygenated environments in water depth about 40-60m?.

Three facies associations were identified by refining the textural features and the interpretation of depositional environments of microfacies. The following three facies associations from offshore to onshore were recognised.

2.2.1 Outer-ramp facies association:

This facies association comprises up to two-thirds of the Tamet Formation within the study area. The outer ramp deposits starting from the point where the formation begins, thicken abruptly and are characterised by the accumulation of bioclastic wacke-packstones with a diverse open-marine fauna and thus are referred to as the outer-ramp facies association. The gradual change from mud-dominated subtidal microfacies to relatively grain-dominated microfacies is usually considered as the boundary between the outer and inner-ramp facies association. The outer ramp can be divided into distal and proximal settings based upon the internal microfacies characteristics. The microfacies of the distal outer ramp are characterised by a mud-rich framework, which contains a planktonic assemblage of taxa deposited without evidence of current transport and probably in a low-energy, deep subtidal environment at 40-100m water depth but still in the oxygenated zone. The complete absence of benthic fauna is further supportive of a distal subtidal environment. This type of deposition is also characteristic of several Late Palaeozoic deeper-water formations (e.g. Marquis and Laury, 1989). Also the presence of phosphatisation levels, which are typically more abundant in deeper-water environments between water depths of 50 and 150m, support this deeper water (>100m) interpretation.

The most distal parts of this facies accumulated in a relatively low-energy, deep subtidal environment below storm wave-base as evidenced by micrite-sized carbonate, absence of evidence of shallow water deposition, lack of contrasting rock textures, lack preferential grain orientation and poor sorting of the constituents, which all together indicate relatively deeper water deposition. The proximal outer-ramp setting is characterised by the accumulation of the foraminifers, echinoderms, molluscans and bryozoans. The high faunal diversity are supportive of a stable open-marine, moderate depth, subtidal environment. The foraminiferal wacke-packstones, the dominant rock type of the banks, indicated that the environment was a low-energy setting associated with episodic storm waves that impinged directly on the substrate.

2.2.2 Inner-ramp facies associations:

This association is interpreted as having been deposited in low-energy, shallow to moderately deep subtidal, inner-ramp environments, probably as lagoonal deposits with water depth of between 10-25m, below normal wave base. The low faunal diversity and density, absence of macrofossils such as molluscans, bryozoans and extremely low foraminiferal diversity compared to facies of the outer-ramp environment strongly suggest that the facies accumulated in a shallow subtidal environment and reflect fluctuations in the degree of restriction. in the lagoonal setting. Porosity occlusion by evaporites within the coarse dolomite and the presence of massive anhydrite microfacies defines the boundary between the inner-ramp and Cyrenaica Platform facies. Muddy *Orbitolina* microfacies suggest an open-marine lagoonal stage, whereas, scarce fossil material is characteristic of deposition in restricted subtidal environments where a few marine organisms live (Wilson, 1975). In modern environments, peloidal sediments are often deposited in slightly hypersaline seawater (Bathurst, 1975). Open-marine circulation was inhibited locally by structural highs and buildups.

2.2.3 Cyrenaica platform facies associations:

The cyclic deposition of dolomitised shallow subtidal sediments and anhydrites suggests that deposition took place in an extensive very shallow epeiric sea along a large inland seaway under extreme arid conditions and shallowness of the Cyrenaica Platform (generally <5m). This type of environment is characteristic of Palaeozoic through Cenozoic strata (Sun, 1995). They are well developed in North America (e.g., Upper Ordovician, Upper Devonian and Lower Carboniferous of the Williston basin), North and West Africa (e.g., Middle Cretaceous, Palaeocene and Eocene of the Sirte Basin and Middle Cretaceous of Congo and Cuanza basins), and Middle East (e.g., Upper Permian and Upper Jurassic of the Arabian platform and Miocene of the Mesopotamian basin). Most previous workers believed that this facies association formed primarily in a sabkha environment similar to that which exists today along the coast of the United Arab Emirates in the Arabian Gulf. Ahmed (1992) suggested that most of the Palaeocene and Lower Eocene deposits in the Cyrenaica Platform are characterised by shallowing-upward cycles and deposited subaqueously on a coastal plain which extended landward hundreds kilometres. This environment has been referred to as a saltern by Warren (1991). This facies association occurs in an extensive area behind the Tamet ramp and extended for hundreds of kilometres east to the Cyrenaica Platform. This facies consist predominantly of dolomitised pel-bioclastic wacke-packstones and thick intervals of anhydrite.

As a result of the extended time duration for the saltern development during prolonged periods of relatively stable platform conditions, several inner-ramp microfacies were predominantly influenced by repeated flushing of hypersaline brines which may have played a dominant role in causing extensive dolomitisation

Vertical stacking of Tamet cycles results from increases and decreases in water depth, which is due to cyclic changes of sea-level and/ or sediment supply. Periodically storms can raise water level and allow for re-sedimentation across the platform. Microfacies analyses and interpretations suggest mainly shallow-marine to locally moderate deep-marine (5-100m) environments. These sedimentary microfacies were cyclic, alternating between deep subtidal and hypersaline settings. These cycles were probably controlled directly by eustasy and include give-up, catch-up and keep-up units, different patterns of thickness and microfacies were governed by the interplay of carbonate production and accommodation space. This cyclicity is closely analogous in scale and time-frequency to that recognised in other carbonate-dominated platforms and is attributed to stratigraphic forcing by Milankovich rhythms, or tectonoeustatically driven processes, or autocyclic processes.

The platform-to-basin distribution of small-scale cycles is a function of accommodation space, which includes depositional space generated by sea-level fluctuations and initial water depths due to the depositional physiography. Two different types of subtidal, shallowing-upward cycles are recognised across the ramp-to-basin transition.

2.2.4 Open-marine subtidal cycles:

The subtidal microfacies are cyclical, with deepening- and shallowing-upward facies patterns. The lack of exposure surfaces in the subtidal environments implies that sea-level never fell below the restricted subtidal microfacies, suggesting that the oscillations were on a metre-scale (< 5m). Subtidal cycles typify the central and outer parts of the Tamet Platform where they are volumetrically the most abundant. These were apparently deposited in two different environments, one deeper and more basinal, and the other shallow and more to landward. Upward-shallowing trends in deep subtidal cycles are interpreted from upsection increase in grain content and size. The most common deep subtidal cycles pass downdip into slope facies, and consist of mud-rich tempestite microfacies interpreted as deep-ramp storm deposits. The relationships and analogies with modern storm deposits (Duke *et al.,* 1991) suggest deposition in water depths between 20-40m. These basal cycles are overlain by sub-storm wave-base chalk and pelagic mud-wackestones, and occur in the most basinward stratigraphic position as thick units. The lack of evidence of wave- or current reworking features, indicates deposition below storm-wave base. The concentration of planktonic assemblages within the upper parts of these cycles suggests slower sedimentation rates. These cycles are more common towards outer-ramp localities and water depths were likely greater than 50m. Subtidal *non*-cyclic microfacies that cap some deep subtidal cycles reflect progradation of shallow ramp deposits towards the ramp margin. This would result in shallow subtidal sedimentation rates keeping pace with the generation of accommodation space; therefore, deep subtidal water depths would not be generated along the rest of the ramp during the late transgression of sea-level. Stratigraphic relationships with adjacent cycles suggest deposition in 15-25m water depths. The shallow subtidal cycles located along the broad shallow ramp consist of storm-dominated cycles. The small flooding events associated with storm influence led to the deposition of metre-scale foram-bearing cycles, which pass upward to amalgamated shallow subtidal *non*-cyclic sediments, which are terminated by dolomite-capped cycles. This is similar to the interpretation of subtidal cycles in the Lower Ordovician Elpas Group, in West Texas (Goldhammer *et al.*, 1993). However, it is not always easy to recognise these cycles since the vertical variations are not great. This study shows a gradational change upsection, rather than dramatic changes at the cycle contacts. The controlling mechanism behind incomplete shallowing in these subtidal cycles may be related to their formation during long-term increases in accommodation space when superimposed higher frequency oscillations were modulated by the long-term event. Deposition was influenced by fairweather-wave reworking of the sediments. Diagenetic caps (hypersaline-related features) developed at the top of some cycles suggest slow rises in relative sea-level and then sudden falls in sea-level which did not allow time for peritidal carbonate deposition (Goldhammer *et al.,* 1993). By way of contrast, cycles on the ramp-margin are amalgamated and consist of sub-storm wave-base sediments; these were overlain by the prograding shallow subtidal cycles.

2.2.5 Hypersaline subtidal cycles:

These cycles are widespread on the Cyrenaica Platform and disappear towards the Tamet ramp; internally two microfacies can be recognised. Each cycle contains very shallow subtidal microfacies at the base with evidence of elevated salinity. The hypersaline conditions during deposition are indicated by the low abundance and diversity of the fauna (mainly miliolids). This microfacies passes upwards into massive saltern anhydrite microfacies. Stratigraphic relationships with adjacent microfacies suggest deposition in <5m water depth. Abruptness of this lithological transition and lack of intermediate microfacies reflect the progradation of saltern successions towards the Tamet ramp. This type of response ("keep-up" of Kendall and Schlager, 1981) occurs when the Cyrenaica Platform closely tracks relative sea-level rise. Ahmed (1992) has interpreted the Cyrenaica succession as a sabkha-saline sequence deposited under the influence of processes very similar to those occurring along the southern shore of the modern Arabian Gulf. The origin of cycles of this type has been attributed to allocyclic mechanisms which include eustatic sea-level fluctuations (Goldhammer *et al.,* 1987) and to autocyclic mechanisms, which include processes inherent within the depositional system that govern the production and/ or distribution of carbonate sediments (Pratt and James, 1988).

2.3 Conclusions

* The depositional setting for the middle Eocene strata in this basin is characterised by an areally extensive low-energy platform within a large marine embayment that was in communication with the open Tethyan Ocean at a time of tectonic stability and eustatic oscillations in sea-level.

The Tamet strata were largely deposited in subtidal environments on a homoclinal-type ramp. Based on the microfacies types and their vertical and lateral interrelationships, the Tamet ramp systems include outer ramp and inner ramp and Cyrenaica platform facies. The palaeoslope dip was mainly directed to the west during deposition of the Tamet Formation. As a result, the deeper water facies are generally more abundant to the west, and hypersaline microfacies are more common to the east. There is no evidence for the development of a rimmed platform margin. The slope gradient between the adjacent platform and the basin was small and the transition from shallow to deeper environments was gradual. This is suggested by the lack of slumps and debris flow deposits.

* Microfacies types and diagenetic variabilities are largely dependent on depositional position within this framework. Most microfacies changes are gradational and hence it was often difficult to define boundaries on the basis of textural changes. Microfacies are generally organised into meter-scale, deepening/shallowing-upward cycles. Meter-scale cyclicity in these strata is intermittent, however, and locally breaks into apparently *non*-cyclic intervals.

* Peritidal microfacies are absent in the Tamet ramp; this may be because the tidal flats had shifted several hundred kilometres back onto the Cyrenaica Platform after each sea-level rise and had insufficient time to prograde westward across the Tamet ramp during stillstands and sea-level falls. In addition, there is no evidence of coastal eolianites on the

inner part of the Tamet platform. This reflects either an absence of strong, onshore-directed winds and dominance of offshore storm transport (c.f. Aigner, 1985) or offshore-directed bottom-water counter flows driven by onshore-directed surface waters.

* The regional (lateraly) continuity of most Tamet cycles indicates that the sea-bed there must have been very flat when these cycles were formed. Also the repetition of similar cycles in the succession clearly suggests that they originated by a cyclic process repeated in time. This process is probably eustatic, with the cycles originating by repeated rises and falls of sea-level. These shallowing-upward units are similar to the classic upward-shallowing cycle described by many authors (e.g. Grotzinger, 1986; James, 1979; Read *et al.,* 1986; Tucker and Wright, 1990; Wilson, 1975). Evidence of shallowing-upward cycles includes 1) upward decrease in normal marine fauna, suggesting increased in restriction and 2) presence of hypersaline-related diagenesis at the top.

* Tidal currents and storm waves are capable of both significantly redistributing sediment along a ramp and of supplying sediment to inner ramp depressions (lagoons) and to offshore areas. Changes in constituents and textures with repetitions at different stratigraphic levels, reflect frequency and scale of storms and are discernible from proximal to distal localities on the ramp system. Storm deposits may be dependent on the sea-floor sediments and water depth at the time of storm-wave scouring and redeposition. Thick high-energy tempestite deposits accumulated at shallower depths and pass down dip into low-energy, thin distal storm beds. It is possible that the geometry and orientation of the Sirte Basin, with the open sea (Tethys) towards the west, resulted in periodic intense storm activity in this leeward location.

There are many examples from the stratigraphic record of storm-influenced deposition. Modern ramp margins between 10 and 45 degrees latitude are subject to strong influence by tropical storms (Duke, 1985). The absence of protective buildups on the outer ramp caused both distal and proximal ramp facies to be highly susceptible to the effects of storms.

* Faunally, there is no evidence of offshore (outer ramp) buildups. This can be explained by: either the rates of relative sea-level rise were an order of magnitude lower than the growth potential of platform margins and buildups, or regional transgression during the middle Eocene was interrupted by episodes of sea-level stillstand or slight lowering. However, such long-term environmental stability on the Tamet ramp may be the reason that the microfacies were highly diversified in foraminifera.

Local subtle topographic irregularities controlled the development of isolated banks. These banks are composed of foraminifera and bivalves, which were separated from the restricted evaporitic flats (Cyrenaica Platform) by a lagoon containing miliolids and peloids.

* Several phases of diagenesis are recognised in Tamet microfacies, which took place in submarine, meteoric, hypersaline and subsurface environments. Mud-rich platformal microfacies suggests that deposition on the platform was restricted and low-energy (alternating between catch-up and keep-up phases), mainly formed below fair-weather wave-base (Wilson,1975). Meteoric diagenesis is represented only by syntaxial overgrowth cements. This stage of diagenesis has only played a minor role in influencing the Tamet's microfacies. This was the result of a relatively arid climatic regime in which there would have been relatively little recharge of fresh groundwater. However, the most notable diagenetic characteristic of near surface sediments are anhydrite nodularisation and pervasive dolomitisation, the latter affecting most Tamet facies. Another peculiar diagenetic feature is the extensive leaching of skeletal aragonite, such as benthic foraminifera, bivalves, and also leaching of echinoderm debris. The paucity of pre-dolomite calcite cementation in the subsurface, together with pervasive dolomitisation and

dolomite cements lining skeletal-moldic pores, suggest that skeletal aragonite was leached by early hypersaline fluids (Sun, 1992). The last diagenetic event before hydrocarbon emplacement was subsurface diagenesis. It is characterised by stylolitisation and associated saddle dolomites filling partially collapsed dissolution cavities. Other diagenetic features include hydrothermal dissolution and anhydrite cements.

* Based on the large-scale stratigraphic distribution and the petrographic features of the dolomite microfacies, two main types were distinguished. Many Tamet dolomites probably formed when the hypersaline fluids originated within the Cyrenaica Platform and refluxed downward through the ramp microfacies.

As indicated by the very limited fauna (mainly thick-shelled miliolids), the hypersaline conditions over the Cyrenaica Platform seem to have been high enough to generate widespread evaporite deposits. The middle Eocene fine dolomites and massive anhydrites form an excellent seal to the most prolific reservoirs on the Tamet ramp-Cyrenaica Platform transition.

* Porosity distribution in the Tamet microfacies is a function of original character and subsequent diagenetic modifications. The pervasively dolomitised microfacies contain two types of pore systems: 1) fabric-selective dissolution of grains creating moldic porosity, and 2) non-fabric-selective dissolution creating vugs. Pore types in the study area are mainly skeletal-moldic, solution-enlarged moldic, vug and intercrystalline. Some of these pore types are still open, but others have been filled.

Primary inter-intraskeletal porosity is restricted to local mud-free foraminifer wacke-packstone facies.

* Carbonate platforms of middle Eocene age play an important role as reservoirs and as current targets for hydrocarbon exploration in Libya.

The deposits of carbonate ramps may form reservoiring systems and offer a range of subtle stratigraphic play types and lateral facies variations. Diagenetic processes, however, may affect the reservoir facies and so lead to local changes in their petrophysical properties.

In general, the best reservoir quality in the middle Eocene carbonates of the southeastern Sirte Basin occurs in the following facies.

1) Mud-free large foram-bearing cycles. This facies is characterised by a pore system largely composed of intergranular and moldic pores, which are related to dissolution (e.g. Gialo Formation in Concession 59).

2) The saltern deposits of the Cyrenaica Platform consist of thick anhydrite interbedded with dolomitised shallow subtidal microfacies. These deposits form updip stratigraphic traps and result in a vertical stacking of reservoir units.

Shallowing-upward units of carbonates associated with evaporites are extremely common as hydrocarbon traps, where carbonate facies interfinger with updip evaporite facies. Examples of this association include grainstones of the Arab Formation of the eastern Arabian peninsula and Hith Anhydrite.

Chapter 3

Sequence Stratigraphic Concepts

3.1 Introduction

Much of the current research in carbonate depositional systems has been oriented towards the application of sequence stratigraphic techniques to the carbonate rock record (e.g., Sarg, 1988; Franseen *et al.*, 1989; Calvet *et al.*, 1990; Sonnenfeld, 1991; Tucker, 1993; Tucker *et al.*, 1993). One major problem that has been encountered, however, is the translation of sequence stratigraphic concepts defined for siliciclastic systems into the carbonate realm. A primary difference is that carbonate sediment supply is not dependent upon allochthonous sources but is generated directly on the platform and redistributed back onto tidal flats or out onto the deeper platform. Because carbonate production and consequent stratal geometries are so sensitive to changes in the depth of the photic zone, shallow-water carbonates are probably better recorders of accommodation change than siliciclastics.

Seismic stratigraphy was developed by Exxon EPR, Texas with 1970's, and there have been several major publications discussing the concepts, ideas, techniques and examples.

This chapter presents a brief review of seismic stratigraphy and then introduces the newer concepts of sequence stratigraphy.

Memoir 26 of the American Association of Petroleum Geologists published in (1977) contains most of concepts in seismic stratigraphy, now widely used by all oil companies and academic throughout the world. Memoir 26 introduced the new terminology and concepts that are used to interpret the gross depositional framework of sequences and recognise potential exploration plays. There are three aspects to the descriptive terminology: reflection termination, configuration and external form and geometry of seismic facies. The concepts have been refined over the years to describe and account for the major shifts and changes in deposition which have occurred during a basin's evolution and to provide a geologically meaningful basis for the subdivision of strata into genetically significant depositional units (i.e. megasequences, sequences and parasequences).

Sedimentary basins are infilled by depositional sequences, which are commonly arranged in cycles of onlapping and offlaping marine strata. Variations in relative sea-level controlled by the interplay of eustasy, tectonics and sediment supply are inferred to control cyclic deposition. A sequence represents a period of essentially continuous sedimentation and as such may be interpreted as a single depositional episode in the history of the basin. A depositional sequence has been defined by Mitchum *et al.* (1977) as a stratigraphic unit composed of a succession of relatively conformable and genetically related strata deposited by one or more contemporaneous depositional systems. Sequences are bounded at their top and base by erosional or non-depositional unconformities (Mitchum *et al.,* 1977).

Sequence recognition and relative sea-level interpretation were proposed in the mid-1970's (Vail *et al.* 1977). This new approach not only integrated the pre-conditioning of Sloss's (1963) ideas on sequences, earlier seismic stratigraphic and depositional systems concepts, but offered some very powerful methods to analyse sedimentary successions.

By 1983, stratigraphic analysis within Exxon had evolved beyond sequence analysis to the documentation of various stratal expressions within siliciclastic sequences and systems tracts in well logs, cores and outcrops. This represented a major step beyond the concepts of seismic stratigraphy. Using well logs and cores, a high resolution chronostratigraphic framework of sequence boundaries, defined only by the relationship of strata, could be constructed to analyse stratigraphy and facies at the reservoir scale. Integration of data from siliciclastic sequences, similar advances in carbonate facies (Sarg, 1988) and sequence-keyed biostratigraphy (Loutit *et al.*, 1988) with methodology of seismic stratigraphy, produced the framework and methodology for stratigraphic and facies analysis now known as "Sequence Stratigraphy".

3.2 Sequence Stratigraphy

This new approach to stratigraphy provides a method for subdividing and describing a succession, which is useful for interpretation and correlation. Sequence stratigraphy defines units that are the result of changes in accommodation space. These units are bounded by stratal discontinuity surfaces on seismic profiles and geological cross-sections, and are specific surfaces at vertical changes in facies stacking patterns on well logs and outcrops. Where the sediments are thick enough, these surfaces can be identified on seismic profiles, and from outcrop sections and well material they can be dated biostratigraphically. Sequence stratigraphy utilises physical criteria to define chronostratigraphic intervals and uses biostratigraphy to determine their age. These intervals are genetic in the sense that the rocks within the interval are related by facies and bounded by physical surfaces that are discontinuities. In addition, the units are believed to be global by some workers, suggesting that they should be recognisable in any basin around the world with a marine base level.

Sequence and systems tract boundaries are always present in the rock record, although at times in certain situations they may be subdued or one boundary may be a composite of several. In general, sequence boundaries are regional onlap surfaces. In deep-water basins they are characterised by onlap of turbidites and debris flows, or apparent onlap of prograding deltas. In shallow-water or non-marine settings they are characterised by onlap of strata deposited in delta, coastal, or fluvial environments. Subaerial and submarine erosional truncation is commonly present below a sequence boundary

Vail and others proposed that cyclic sequences, which are bounded by unconformities, are typically composed of three subsequences deposited at specific times during a cycle of relative sea-level change. Furthermore, they noted within the unconformity-bounded sequences that these subsequences are separated by marine-condensed sections, minor discontinuities, and /or lapout surfaces. In a series of papers in 1987, Exxon geoscientists named these subsequences depositional systems tracts. The term depositional systems tract was introduced by Brown and Fisher (1977) who defined it as a lithogenetic unit composed of one or more contemporaneous depositional systems. Exxon researchers recognised four systems tracts, three of which ideally compose a depositional sequence.

The tracts were named by Vail (1987) and Van Wagoner *et al.* (1987) for their respective sea-level position when the tract was deposited: lowstand, transgressive (retrograding highstand), highstand (prograding highstand) and shelf margin wedge.
The physical boundary between the lowstand and transgressive systems tract is defined by the first flooding surface. The top lowstand surface merges with the basal unconformity landward of the point where the lowstand or shelf margin systems tract pinches out. The physical boundary between the transgressive and highstand system tract is called the maximum flooding surface. It is a submarine condensed section characterised by downlap above and apparent truncation below. The boundary between the highstand and shelf margin systems tract is a type 2 sequence boundary.

Sequence stratigraphic analysis involves the recognition, mapping and interpretation of depositional systems tracts. It permits greater precision in interpreting the depositional systems and component lithofacies and identifying potential reservoirs, seals, sources and stratigraphic traps. In fact, some petroleum geologists estimate that 85% of the world's hydrocarbons are trapped in reservoirs either deposited or enhanced during lowstands of relative sea-level. It is probably true, however, that the lowstand sandstone reservoirs and carbonate reservoirs that experienced porosity/ permeability enhancement during lowstand of sea-level constitute the majority of the world's reservoirs. For these reasons alone, explorationists and exploitationists should be knowledgeable about systems tracts.

3.3 Factors Controlling on the Internal Architecture of Depositional Sequences

There has been a lot of discussion in recent years over the factors controlling the internal architecture and repetition of depositional sequences. Many workers (e.g. Sloss and Krumbein, 1963; Hardenbol *et al.,* 1981; Wilgus *et al.*, 1988; Van Wagoner *et al.*, 1990) have emphasised that depositional patterns reflect the dynamic interplay of four principal factors, namely eustasy, tectonics, sediment supply and climate. Tectonic and eustatic processes combine to cause relative changes of sea-level which control the space available for sediments (accommodation space). Tectonism and climate are major controls on the amount and types of sediment deposited.

3.3.1 Eustatic effects

Relative sea-level, which is controlled by the interplay of basin subsidence and direction and magnitude of eustatic sea-level changes, is measured relative to an initial underlying depositional surface and provides the principle control of stratal geometries and distribution of lithofacies. The magnitude of relative sea-level at any point is the algebraic sum of two factors, and the rate of relative sea-level change equals the rate of eustatic change minus the rate of subsidence. The space added or subtracted by changes in relative sea-level (i.e., eustasy ± subsidence) is the accommodation space. The interplay of rates of accommodation space added or subtracted and the rate of sediment supply determine water depths and shoreline positions. In general, accommodation space may be filled by two mechanisms: aggradation or upbuilding producing onlap, and progradation or/ outbuilding producing downlap and perhaps toplap.

Eustasy controls the rate of relative sea-level change and is the major controlling factor on the timing of stratigraphic discontinuities; it is the main factor creating the boundaries between sequences and systems tracts. The stratigraphic signatures of eustasy include major continental flooding cycles and depositional sequence cycles (Fig.3.1). Major continental flooding cycles are believed to be caused by tectonoeustasy (changes in ocean basin volume), while depositional sequence cycles are thought to be caused by glacioeustasy (changes in water volume).

3.3.1a. Continental flooding cycles

These are defined on the basis of major periods of encroachment and restriction of sediments on to the cratons. They represent the first-order eustatic cycles. Their stratigraphic signature is a megasequence. There are two Phanerozoic continental flooding cycles. The youngest starts at the base of the Triassic and extends to the present. The Triassic represents a time of gradual encroachment of sediments on to the craton, and a great thickness of non-marine sediments was deposited in grabens and bordering marine basins. This general pattern is believed to be caused by a slow relative rise of sea-level due to long-term rise in eustasy, resulting from continental breakup.

The Jurassic and lower Cretaceous represent times of extensive encroachment of sediments onto the continental margins. During this time period the average relative sea-level rose more rapidly due to an increase in the rate of rise of long-term eustasy. The upper Cretaceous and Cenozoic times are characterised by an overall gradual restriction of sediments to the continental margins and basinal areas. This pattern is believed to be caused by a gradual long-term fall of eustasy, causing a regression or a relative fall and exposure. The older, first order eustatic cycle starts in the uppermost Proterozoic and extends to the end of the Permian. The latest Proterozoic represents the time of slow encroachment with regression, the Cambrian represents the time of extensive encroachment with transgression, the Ordovician represents the eustatic high, and the

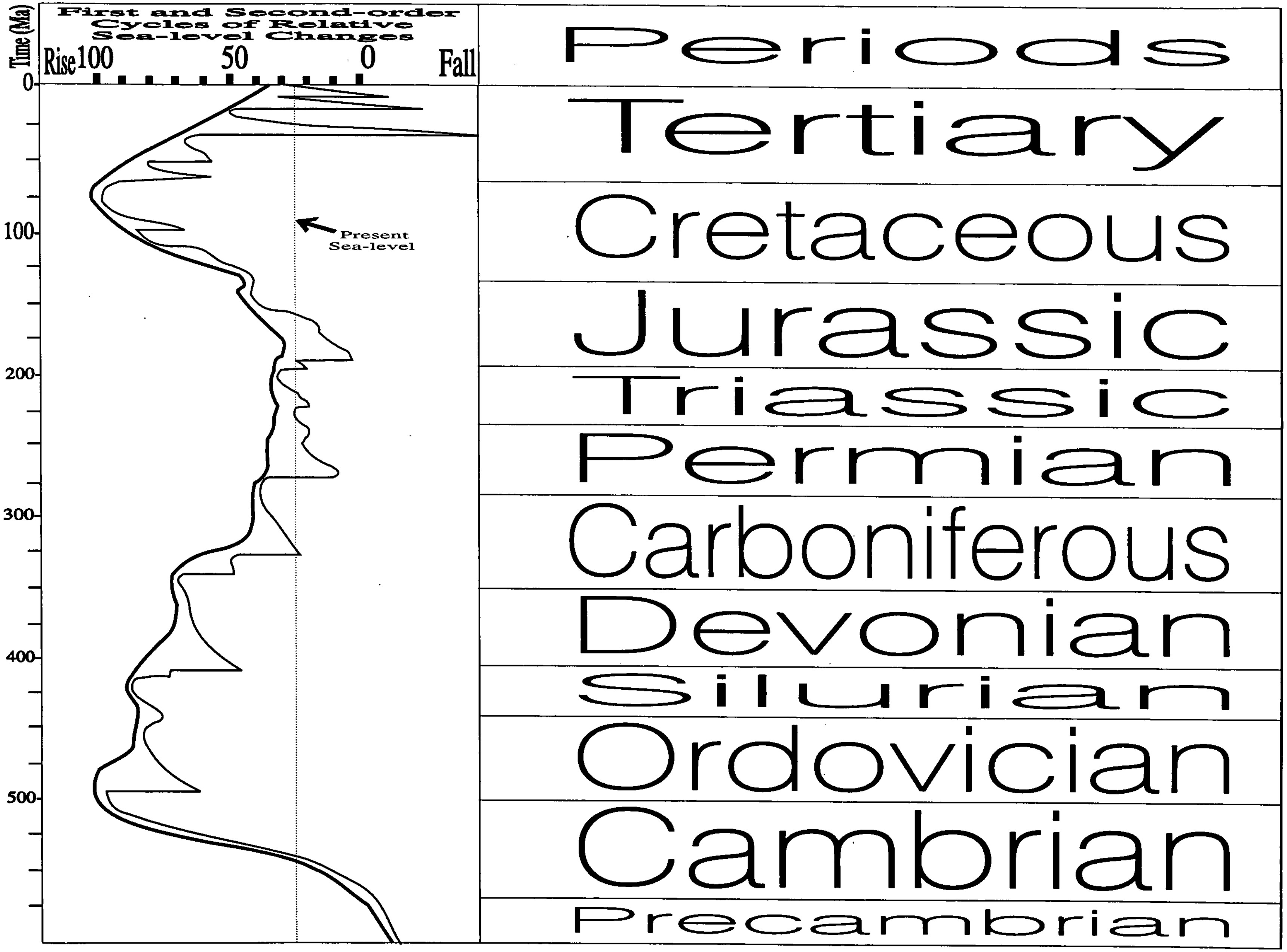

Fig.3.1 *First and second-order global cycles of relative sea-level fluctuations during Phanerozoic time (from Vail et al., 1977)*

Silurian to Permian represents the time of gradual restriction. The two continental flooding cycles are recognisable on all continents and are believed to be global.

Second-to fifth-order eustatic cycles are recorded by sequence cycles, systems tracts and periodic parasequences. They are believed to be glacio-eustatic cycles (Vail *et al.*, 1977; Bartek *et al.*, 1990) with smaller magnitude, but higher frequency than tectonically induced transgressive-regressive facies cycles. Glacioeustatic variations produce high frequency variations similar to those on the sea-level curve of Haq *et al.* (1988).

3.3.1b Depositional sequence cycles

Second-order eustatic cycles consist of sets of third-order cycles (Haq *et al.*, 1987; 1988) bounded by major unconformities. In general, a set of five to seven third-order cycles form a second-order cycle with average duration in range 5-10 Ma. The boundaries of second-order eustatic cycles are characterised by especially large eustatic falls (< 50 m). The stratigraphic signature of a second-order eustatic cycle is a supersequence.

Depositional sequences formed during third-order cycles are the fundamental stratigraphic signature of sequence stratigraphy having a duration of approximately 0.5-5.0 Ma. In general sequence boundaries are regional onlap surfaces. In deep water basins they are characterised by onlap of turbidites and debris flows. In shallow-water settings they are characterised by onlap of strata deposited in deltaic, coastal or fluvial environments. Subaerial and submarine erosional truncation is commonly present below a sequence boundary.

Parasequences and high-frequency sequences are fourth-and fifth-order cycles. They may be episodic or periodic. Episodic parasequences are caused, for example, by delta lobe shifts. They are limited in distribution and of very short duration (usually less than 10000 years). Periodic parasequences are characterised by regional continuity and by systematic changes in thickness within a stratigraphic section. These periodic parasequences are believed to be caused by climatic fluctuations associated with

Milankovitch-scale orbital cycles (less than 500 Ka). These orbital cycles influence the amount of solar energy received on the Earth's surface and thus affect climate. It is believed that these climatic variations induce changes in continental ice volume, which cause eustatic changes and consequently small relative changes of sea-levels.

"Are the boundaries of depositional sequences largely controlled by global eustatic changes in sea-level (e.g. Vail et al. 1977) or created by episodes of tectonism and lasting of millions of years ?".

Vail *et al.* (1984) stated "In general ... the tectonic subsidence along most passive margins is long-term and gradually decreases in rate, because it is related to a thermal decay curve. It does not change rapidly enough to cause regional unconformities. Tectonic subsidence patterns differ from region to region, and are not globally synchronous".

Peter Vail's ideas concerning sequence stratigraphy were seeded during his time as a graduate student supervised by Larry Sloss. Over forty years ago, Sloss recognised major unconformity-bound sequences developed over the North American Craton, and named them after Indian tribes. Sloss (1963) pointed out that each transgressive-regressive episode alternated with phases of orogenic activity on the eastern and western margins of craton.

One of the major problems confronting sequence stratigraphers is to distinguish between relative sea-level changes that are a consequence of fluctuations in eustatic sea-level from those due to variations in the rate of tectonic subsidence.

Some brief examples are given on the controversy underlying the proposition that global sea-level changes (eustasy) are the fundamental control on the sequence boundaries.

On a global scale, many correlations have been established between major plate movements, orogenic episodes and large-scale changes in sea-level. The purpose of this controversy is to developed some simple ideas about the origin of unconformity-bounded depositional sequences, to show how conditions for the development of an unconformity can be quantified in terms of rates of tectonic subsidence, sediment accumulation, eustatic change and change in elevation with respect to the sea surface.

For example, the Triassic rocks in the North Viking graben and the Moray Firth can be subdivided into three unconformity-bounded sequences (Vail & Todd, 1981). Except at the very top of the succession, unconformities developed entirely within non-marine deposits (fluvial, lacustrine) can only be the result of local tectonic activity, so the method of documenting coastal onlap and determining sea-level change cannot be applied here. The reason is that the latter are brought about by volume changes in oceanic spreading centres at times of change in global average spreading rates or during the initiation of new spreading centres.

The majority of proposed sequence boundaries in the mid-Palaeozoic Welsh Basin, U.K., are caused by processes other than pure eustatic sea-level change. Of the 19 boundaries, four are purely tectonic or volcanotectonic and a further seven have an important tectonic or volcanotectonic component. Only eight are plausibly solely eustatic (Woodcock, 1990). This is a useful study in the controversy over the extent of eustatic control on sequence development.

Stratigraphic sequences in foreland basins are clearly controlled by regional tectonism. Recently attempts have been made to correlate specific stratigraphic sequences, such as major molasse pulses, with tectonic events such as terrain collisions. Kauffman (1984) working in the Cretaceous Western Interior Seaway of the Rocky Mountains regions, demonstrated a correlation between transgressive events in the Seaway, major thrusting episodes in Wyoming-Utah fold-thrust belt and periods of volcanism. He attributed this correlation to changes in rates of subduction along the Pacific borderlands, resulting from changes in rates of sea-floor spreading.

In the Karoo Basin of southern Africa, which is also a foreland basin, the Beaufort Group comprises three basin-wide upward-fining cycles ranging from 160 to 500m in thickness (Visser and Dukas, 1979), whereas the overlying Molteno Formation consists of six upward-fining cycles that reach maximum thickness of 140m (Turner, 1983). Tectonic control of these foreland basin cycles is indicated by sedimentological evidence for shifting source terranes. These cycles are third-order in the classification of Vail *et al.* (1977). Foreland basins also contain sequences of shorter, fourth and fifth order, duration (see review by Miall, 1990).

Embry (1991) described the Mesozoic stratigraphic record of the Sverdrup Basin, an extensional basin in Arctic Canada. The 9 km thick succession contains thirty third-order unconformity-bounded fluvial-deltaic sequences, some up to 250m thick. These are now interpreted as the product of fluvial rejuvenation and progradation resulting from regional interplate tilting. He suggested that changes in the interplate stress regime were the cause. Eventually, it may be possible to correlate specific cycles in the Sverdrup Basin with specific tectonic events. This type of correlation has already been attempted for the Cenozoic record of the Gulf Coast (Galloway, 1989a, b). He described the major Cenozoic clastic wedges along the Gulf Coast, where it is apparent that the source areas changed with time, indicating a strong tectonic control. Galloway was able to relate many of the specific third-order depositional cycles, some resulting in up to 2 km of sediment, to tectonic events in the southwest United States. Few of the depositional pulses can be correlated with the supposed eustatic sea-level changes proposed by Haq *et al.* (1988).

There is general agreement that stratigraphic discontinuities (unconformities) are expressed by breaks in facies or biota, and that successions are controlled largely by the interaction of subsidence, eustasy and sediment supply, and less directly or to a lesser extent by such factors as topography climate (Posamentier *et al.,* 1988; Galloway, 1989b). However, serious questions remain about precisely how sediment supply controls the formation of unconformities.

3.3.2 Tectonism

This has the greatest influence on increasing or reducing accommodation space. Also, when coupled with climate, it controls the type and amount of sediment filling that space. The corresponding signatures of each tectonic process can be distinguished on the basis of rates and duration in time and regional distribution. The interaction of eustasy and tectonics causes the observed relative changes of sea-level. In general, tectonic subsidence is of a high magnitude, but it changes slowly with time. Thus tectonic subsidence creates most space (accommodation) for the sediments. The stratigraphic signature of tectonism results from a wide range of processes and has the most profound effect on accommodation. Its imprint on the sedimentary record can be divided into three hierarchical groups.

First-order tectonic events result from thermodynamic processes in the Earth's crust and upper mantle, and are long-term events. Their stratigraphic signature is the sedimentary basin. Second-order tectonic events occur during the evolution of a sedimentary basin, and are characterised by a period of relatively high subsidence rate followed by a period of relatively low subsidence rate. Its stratigraphic signature is a major transgressive-regressive facies cycle. Third-order tectonic events are folding, faulting and diapirism. The stratigraphic signatures of these events are tilted, folded and disrupted strata. They may occur during deposition, and lead to events such as slides, slumps and local facies variations and thickness changes.

3.3.3 Sedimentological effects

Sediments fill the space created by the relative rise of sea-level. Sediments are deposited episodically and are local in distribution. The stratigraphic signatures of the sedimentological effects are the sedimentary structures related to currents or settling processes within each depositional setting which are caused by wind, waves , mass flow, tides, floods, marine and fluvial currents and precipitation. These episodic packages accumulate to form depositional systems composed of lithofacies tracts. They may occur within one sequence or include several sequences. Stratigraphic marker beds may be created by certain unique depositional events. Parasequences, bounded by flooding surfaces, package the beds and bed sets into characteristic upward-shallowing cycles that stack together to form the systems tracts and depositional sequences.

3.3.4. Climatic Effects

Climate, which is a measure of air temperature, precipitation, atmospheric humidity, and wind, helps determine water conditions(salinity, water temperature and circulation) and, hence, the nature of the sedimentation (tropical or temperate) and the types of sediments produced. Shallow-marine tropical waters have a higher degree of $CaCO_3$ supersaturation than the temperate seas of the mid-latitudes. This difference affects the production, stability, and early lithification potential of sediments (Scoffin, 1987). Climate helps determine the types of sediment, that will be deposited within a depositional sequence. Under arid conditions and restricted circulation, evaporite deposition may occur. Differences in climate will affect the style of siliciclastic sediment delivery. Humid climates favour fluvial-deltaic deposition of siliciclastic sediment and arid climates foster aeolian siliciclastic deposition. The presence of these sediments in a carbonate-dominated stratigraphic succession is a clue not only to climatic conditions, but they may signal relative sea-level changes.

3.4 *Depositional Sequence*

The basic unit of sequence stratigraphy has been described by (Van Wagoner *et al.*, 1988; Wilson, 1991). Application of sequence stratigraphic analysis depends on the recognition of a hierarchy of stratal units that range in thickness from millimetres (lamina) to kilometres (megasequence).

Geologists have long recognised that unconformities subdivide the stratigraphic succession in a basin into discrete lithogenetic packages or units that represent the sedimentary products of major depositional episodes. During the early 1970's, Exxon geologists named these units depositional sequences and Mitchum *et al.* (1977) defined the sequence as "... a stratigraphic unit" composed of a relatively conformable succession of genetically-related strata. Brown and Fisher (1977) recognised similar units and called them " seismic stratigraphic units" defined as lithogenetic units composed of one or more contemporaneous depositional systems. Exxon geoscientists (Vail *et al.*, 1984) later redefined sequences as relatively conformable successions of strata deposited between erosional unconformities or their equivalent concordant surfaces produced by a cycle of eustatic sea level. A relative change of sea-level is " an apparent rise or fall of sea-level with respect to the land surface". Movement of sea-level, the land surface, or both may produce the relative change, which can be local, regional, or global. According to Vail *et al.* (1977), indicators of relative sea-level change can be grouped into three basic lines of evidence: i) coastal onlap indicates a relative rise; ii) coastal toplap indicates relative stillstand; and iii) downward shift of coastal onlap reflects a relative (rapid) fall of sea-level.

3.4.1 *Sequence*

This is the fundament unit of sequence stratigraphy. It is defined as " a relatively conformable succession of genetically related strata bounded by unconformities and their correlative conformities" (Mitchum, 1977). The interpretation of sequences, systems tracts and parasequences is based on the identification of lithology, facies, stratal patterns and the discontinuity surfaces that mark the boundaries of sequence stratigraphy units. Recognition and dating of unconformities require careful interpretation with local (well) or regional (seismic) and adequate palaeontological controls. Surfaces of erosion or non-deposition constitute unconformities and they represent typically variable time gaps of short or long duration. Unconformities are not chronostratigraphic surfaces because they do not bound equal time intervals, but they do have chronostratigraphic significance because the rocks above the unconformity are everywhere younger that those below the surface.

Two types of erosional unconformity-bound depositional sequences can be recognised by extensive erosional surfaces and downward shifts of coastal onlap.

3.4.1a Type 1 surfaces

These are initiated when falling eustatic sea-level approaches and passes its maximum rate on the falling limb of the eustatic curve and initiates a rapid basinward shift beyond the pre-existing highstand depositional offlap break. These discontinuities are subaerial and submarine unconformities, and their concordant equivalent surfaces, which develop when eustatic sea-level falls more rapidly (>10 cm/ 1000 yr.) than subsidence at the offlap break. Exposure of the shelf (e.g., palaeosols), incision of fluvial valleys and erosion of submarine canyons provide sediment to basin-floor fans. Slope-front infills and slope fans and subsequent aggradational lowstand deltaic and coastal systems (or evaporite/ carbonate systems) coastally onlap the type 1 surface and fill the canyons and incised valleys during the subsequent relative sea-level rise. Type 1 erosion surfaces in relatively stable, subsiding basins are typified by well developed entrenched valleys and submarine canyons near the relict shelf edge, but erosional evidence diminishes landward. Type 1 erosion of a basin during tectonic uplift and folding produces maximum truncational erosion in the landward direction, and it tends to diminishes toward the shelf edge.

On gently sloping ramps with a lower gradient than the fluvial profiles, relative sea-level falls basinward of the depositional shoreline break, but type 1 unconformities are less erosive and there is limited submarine erosion and deposition.

Recognition criteria for the Type 1 sequence boundary include 1) subaerial erosional truncation, commonly as incised valleys (Van Wagoner *et al.*, 1990), subaerial exposure, and laterally equivalent submarine erosion; 2) downward shift in coastal onlap and basinward shift in facies overlying the sequence boundary; 3) onlap of overlying strata onto the sequence boundary, as either coastal onlap or onlap onto the margins of incised valleys; 4) presence of contemporaneous lowstand deposits beyond the depositional edge, and 5) demonstration of one or more of the criteria on a regional basis to ensure that a sequence boundary and not a local distributary channel is being marked.

3.4.1b Type 2 surfaces

These discontinuities are principally subaerial exposure (e.g., palaeosols) unconformities and their concordant equivalent surfaces, which develop when eustatic sea-level falls slowly to move basinward to a point landward of the shelf edge where rates of eustatic fall and subsidence are equal. Limited deep water sedimentation occurs, and progradational and aggradational shelf-margin systems develop. Coastally onlapping Type 2 surfaces develop during the subsequent relative sea-level rise. Type 2 unconformities display limited fluvial incision, and they are principally regional subaerial exposure surfaces.

Recognition criteria for the Type 2 sequence boundary include 1) a downward shift in coastal onlap; 2) minor subaerial erosion and exposure (no incised valleys), and 3) a vertical change in parasequence stacking patterns from prograding below the boundary to aggradational and retrogradational above. These criteria should occur on the shelf above the depositional shoreline break.

3.4.2 *Cyclic Frequencies of Sequence*

As previously mentioned, the criteria for sequence recognition and definition involve stratal configuration and facies relationships and do not depend upon cyclicity or frequency for sequence definition. After sequence boundaries are recognised, they can be dated using biostratigraphy tied to standard time scales. Third-order sequence boundaries observed on a global basis have been plotted on cycle charts of Haq *et al.* (1988).

High-frequency sequences occur most commonly with fourth-order cyclicity (0.1-0.2 my.) and some with fifth-order cyclicity (0.01-0.02 my.). Conceptually, high-frequency sequences should be best-developed in basins where tectonic subsidence is low and sediment is deposited very rapidly, where apparently the facies have a higher sensitivity to minor high-frequency relative fluctuations of sea-level. These frequencies are about the same as those of parasequences in third-order sequences. Therefore, it appears that fourth- and fifth-order cyclicity may be expressed either as parasequences or high-frequency sequences. Both parasequences and high-frequency sequences may occur in prograding, aggrading, or retrograding stacking patterns.

3.4.3 *Parasequences*

The fundamental building blocks of sequences and systems tracts are parasequences and parasequence sets which are themselves made up of beds and bed sets (Van Wagoner *et al.*, 1990; Campbell, 1969). A parasequence is defined as a relatively conformable succession of genetically-related beds or bed sets bounded by marine flooding surfaces and their correlative surfaces. The general depositional characteristics of a parasequence (generally a coarsening-upward vertical facies association) are interpreted to record a gradual decrease in water depth. This gradual shallowing is followed by an abrupt deepening which produces the marine flooding surface forming the parasequence boundary.

Parasequences commonly occur in sets which are successions of genetically-related parasequences with distinctive stacking patterns. Parasequence sets are progradational, retrogradational or backstepping and aggradational, depending on whether successively younger parasequences in the set build farther basinward, landward, or vertically, respectively (Van Wagoner *et al.*, 1990).

3.5 Depositional Systems Tracts

The objective of sequence stratigraphy is to identify and correlate the genetic chronostratigraphic sequences, systems tracts and parasequences and then relate them to the depositional systems and lithofacies tracts. Each depositional sequence is composed of systems tracts. A systems tract is defined as a linked contemporaneous depositional system. A depositional system is defined as a three dimensional assemblage of lithofacies.

In sequence stratigraphy usage, each systems tract is bounded by a physical surface that is, in part, a discontinuity. Depositional systems within each systems tract are linked by changes in sedimentary facies. Depositional systems tracts may be recognised and interpreted on the basis of stratal geometries and terminations, as well as their relationships to erosional and non-depositional discontinuities. These characteristics are generally less than ideally exhibited by a sequence, but with sufficient seismic profiles, it is normally possible to recognise the diagnostic criteria needed to delineate the systems tract. These include 1) their vertical position within the sequence; 2) the stacking pattern of prograding or retrograding parasequence sets in the systems tract and 3) their lateral position within the sequence.

Application of this procedure to many different types of basin has shown a close relationship between the systems tracts and this has been used for identification and labelling of systems tract boundaries within different depositional settings. This relationship is discussed below with reference to systems tracts in carbonate successions.

There are four types of systems tracts: lowstand, transgressive, highstand and shelf-margin. Each systems tract is interpreted to be deposited during a specific phase or portion of one complete cycle of relative fall or rise of sea-level.

3.5.1 Lowstand Systems Tracts (LST)

These systems tracts are deposited on Type 1 unconformities when relative sea-level rapidly falls from a highstand to below the pre-existing depositional offlap break. Depending upon the magnitude of the fall and the bathymetry of the basin, three lowstand scenarios may occur: (1) where a shelf/ slope break exists, sea-level may fall below the pre-existing shelf edge to some position on the upper slope; (2) where no distinct shelf/ break exists, sea-level may fall basinward of the shoreline break onto the shelf or ramp; (3) where growth faults occur along the shelf/slope break (most commonly in siliciclastic basins), sea-level may fall basinward on the downthrow side of the fault (Vail, 1987; Posamentier and Vail, 1988).

Variations in the nature and composition of lowstand tracts (Fig. 3.2A) may occur along a siliciclastic basin margin in response to the presence or absence of significant fluvial sediment supply. In fact, if appropriate climatic conditions exist, the lowstand siliciclastic tract may grade into carbonate or evaporite tracts away from the fluvial/ deltaic depocenters.

3.5.1a Carbonate Lowstand Systems Tract

A Type 1 fall of relative sea-level in a carbonate basin exposes the previously deposited highstand carbonate to meteoric waters, and the resulting lenses of fresh water move across the shelf with falling sea-level (Sarg, 1988). Dissolution and various processes of meteoric diagenesis result which may enhance the porosity and permeability of highstand facies.

In general, thick lowstand systems tracts composed of basin floor fans, slope fans and thick lowstand prograding complexes are poorly developed in carbonate environments. Carbonate basin floor fans consisting of megabreccias are common; grainstone basin floor fans may also form at lowstand stage (Fig. 3.2B). Slope fans consisting of turbidites channel overbank deposits and slumps are generally poorly developed in carbonate systems.

Lowstand prograding complexes consist of thick prograding sediments with that facies change from coarse shallow-water grainstones to slope mudstones. Sequence boundaries may be tectonically enhanced during this phase and show widespread erosion, especially on the slope. The top of a lowstand prograding complex at the end of latest regressive phase is the point of maximum progradation into the basin. The lowstand wedges onlap the exposed shelf and their geometries indicate how sedimentation accommodates the new space added by rising sea-level. For example, in restricted basins biogenic productivity is lower and, therefore, deep water, lower energy and finer grained (catch-up) sedimentation (Kendall and Schlager, 1981) does not keep-up with new space added and the wedges exhibit sigmoidal geometries (aggradational parasequence sets). In open basins, the biogenic productivity is normally higher and, therefore, shallower water, higher energy and coarse grained (keep-up) deposition efficiently exceeds new space added by relative sea-level rise.

In the ramp setting, lowstand systems tracts have three parts: lower prograding complex, upper prograding complex and incised valley fill (Posamentier *et al.*, 1988 ; Posamentier and Vail, 1988). The lower prograding complex is characterised by offlapping stratal patterns with downlap at the base and erosional truncation at the top.

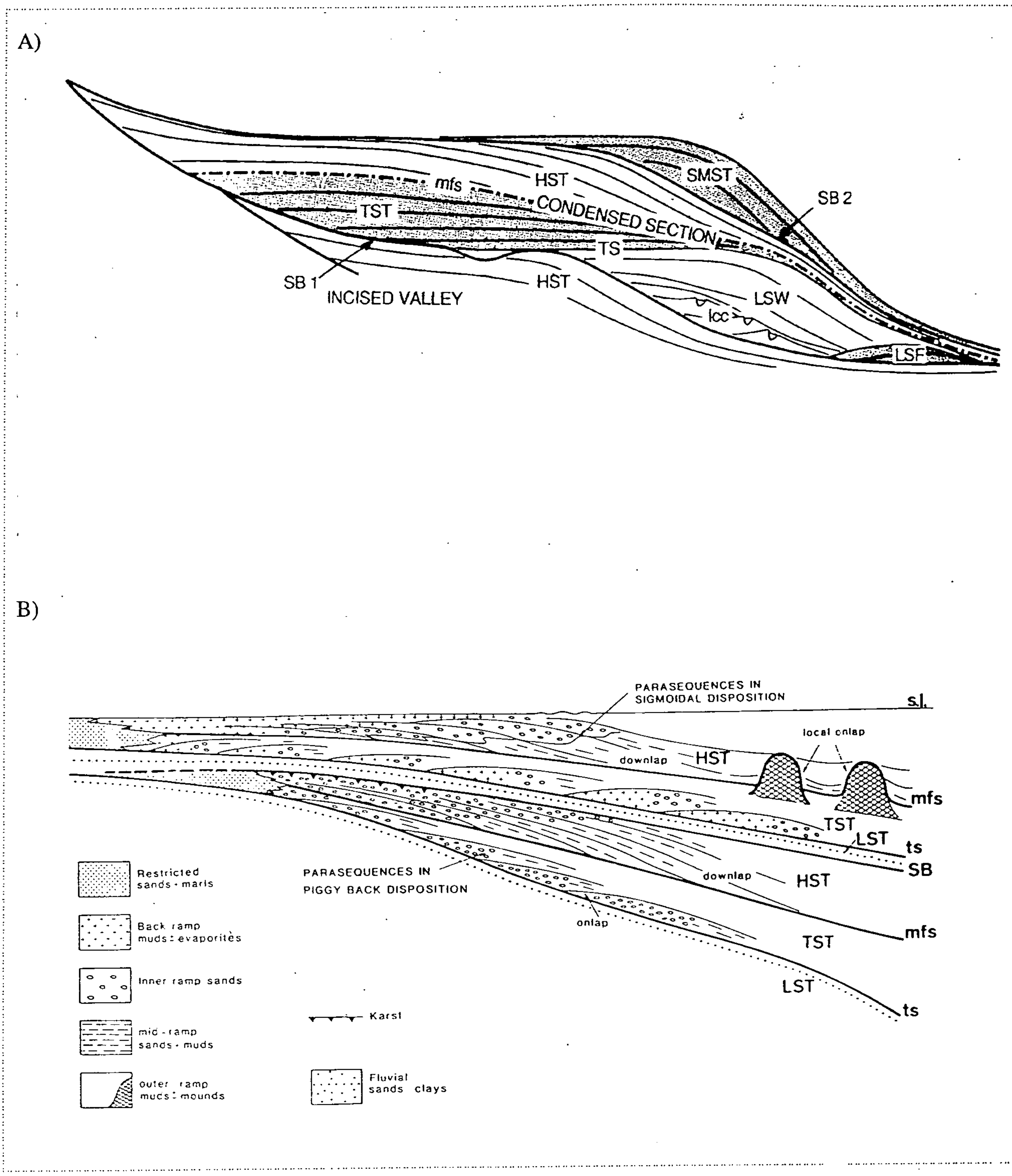

Fig. 3.2 *A)The depositional model of Exxon was generally developed from studies of silicicclastic sequences. Abbrviations are as follows. Surfaces: SB= sequence boundary; mfs= maximum floodin surface; Ts= transgressive surface. Systems Tracts: HST= highstand systems tract; TST= transgressive systems tract; LWS= lowstand wedge systems tract; LSF= lowstand fan systems tract; SMST= shelf-margin systems tract. B) Sequence stratigraphic model for a carbonate ramp and its systems tracts. Abbreviation: SB, sequence boundary; ts, transgressive surface; mfs, maximum flooding surface; LST, lowstand systems tract; TST, transgressive systems tract; HST, highstand systems tract; S.l Sea-level.*

The upper lowstand prograding complex has an erosional base that in places may cut through the lower prograding complex to form incised valleys, and is formed during the slow relative rise of sea-level, following the relative fall (lowstand prograding complex time). The incised valleys are filled either by braided stream deposits, or by estuarine sands during the transgressive systems tract time.

3.5.2 Shelf-Margin Wedge Systems Tracts (SMST)

These systems tracts are deposited upon type 2 unconformities and lap out on the shelf landward of the underlying offlap break. The system tract is characterised by both aggradation and progradation (Fig. 3.2A). Its lower boundary is a conformable sequence boundary and its upper boundary is a top lowstand surface. An unconformity exists landward of where it pinches out. The landward portion of a shelf-margin wedge systems tract is commonly a non-marine wedge that thickens seaward. Some meteoric diagenesis may develop in the landward parts of the subaerially exposed platform, while the marine portion is similar to the lowstand prograding complex and may move into the basin by growth faulting or gravity creep.

3.5.2a Carbonate Shelf-Margin Wedge Systems Tracts

These typically consist of platform-bank margin wedges (Fig. 3.2B) deposited initially at or slightly below the pre-existing highstand platform margin (Sarg, 1988). In basins with good circulation, biogenic production typically maintains shallow, higher energy deposition, while in restricted basins deposition lags behind and is, therefore, deeper water and finer-grained. Depending upon climate and restriction, evaporite deposition may occur within the shelf-margin carbonate wedge.

3.5.3 Transgressive Systems Tracts (TST)

These systems tracts coastally onlap transgressive or first marine flooding surfaces on top lowstand or shelf-margin tracts. These are typically local, diachronous ravinement surfaces produced by marine erosion (Fig. 3.2A). Up-dip these onlapping surfaces coalesce landward and erosionally enhance the subjacent type 1 or 2 surfaces. Retrogradational deposition occurs in response to a generally diminishing fluvial sediment input under the influence of accelerating but periodic paracyclic rises and stillstands of relative sea-level produced by combined subsidence and rising eustatic sea-level. The transgressive tracts are progressively upward thinning and accelerating landward-shifting retrogradational parasequence sets deposited under periodically deepening water, retreating shorelines and diminishing sediment supply.

2.5.3a Carbonate Transgressive Systems Tracts

In the early phase of a transgressive period, thick widespread carbonate banks (Fig. 3.2B) may be developed. These carbonates commonly consist of lagoonal and barrier facies. It is during this stage that a carbonate platform generally has good marine circulation, retrogradational, keep-up, shallow-water, shoaling-upward parasequences are developed. Where there is a degree of restriction, however, carbonate systems may exhibit catch-up deposition composed of lower-energy mudstones which evolve into keep-up deposition near the top of the tract only when relative sea-level rise slows sufficiently to permit shallow-water deposition.

During the latest stage of the transgressive phase, the carbonates cannot generate fast enough to fill all the space being created, thus the facies/parasequences backstep. Backstepping facies thin upward to a surface of drowning (Drowning unconformity), which corresponds to the maximum flooding surface at the top of the late transgressive systems tract. Above the drowning surface, a condensed interval commonly develops where several facies may be merged by sediment starvation. This condensed section is the peak transgression of the 2ndorder transgressive/ regressive succession. Backstepping facies thin basinward because of the processes of sediment starvation that occurs when depositional environments backstep. Basinal transgressive systems tracts are very condensed and organic-matter rich.

3.5.4 Marine Condensed Sections (MCS)

These widespread but thin hemipelagic deposits signal sediment starvation (Loutit *et al.,* 1988). When rising relative sea-level approaches its maximum rate under the influence of accelerating eustatic rise and subsidence, the transgressive tracts are progressively encroached by an extensive blanket of hemipelagic sediments which extends onto the shelves from deep in the basin. Siliciclastic transgressive tracts are normally overlain by black phosphatic, glauconitic high-gamma/ low resistivity. Carbonate transgressive tracts (Sarg, 1988) typically consist of black, phosphatic, micritic high-gamma limestones and shales with hardgrounds and/ or burrowed surfaces. The marine-condensed sections contain rich palaeofaunas and floras which may grade upward with increasing water depths from shallow-water benthic assemblages to deep-water pelagic assemblages.

The marine-condensed sections and its contained maximum flooding surface are widespread and distinctive, and may be recognised from palaeontology, composition, well-log patterns and seismic reflectivity. Low sedimentation rate typically results in marine-condensed sections. In up-dip areas, where transgressive facies are thin, the condensed section or its up-dip equivalent may rest upon, or be separated by only a few meters from the underlying type 1 or 2 unconformity. In deep-basin areas, the condensed sections may be stacked, separated by only a few metres of distal highstand /lowstand shales. Consequently, the marine-condensed section may be recognised even where the bounding unconformities can not, making it a critical element of the depositional sequence.

3.5.5 Highstand Systems Tracts (HST)

These systems tracts are deposited during progradation and they downlap on to the top of subjacent marine-condensed sections when rising relative sea-level slows, permitting an adequate sediment supply to initiate regressive deposition (Vail, 1987). Highstand systems tracts are made up of three parts (Fig. 3.2A): early highstand and late highstand prograding complexes and highstand subaerial complex. The early highstand is characterised by an upward-and outward-building sigmoidal prograding stratal pattern. The late highstand prograding complex is characterised by outward-building oblique prograding stratal patterns. The late highstand subaerial complex is characterised by fluvial sediments deposited during a relative sea-level stillstand.

The late highstand prograding complex and the late subaerial complex are deposited contemporaneously. The boundary at the base of the highstand systems tract is a downlap surface (maximum flooding surface) that is typically associated with a condensed section. The early highstand is very similar to the late transgressive systems tract. The most important difference is that the parasequences retrograde during the transgressive systems tract and prograde during the highstand system tract. The late highstand prograding complex is typically made up of deltaic and interdeltaic or beach and storm deposits; as result of the decreasing relative sea-level rise in the late highstand, the coastal and delta plain sediments are thin-bedded. The late highstand subaerial complex builds above sea-level, enabling the fluvial systems to maintain their optimum equilibrium gradient as the highstand systems tract progrades seaward (Posamentier and Vail, 1988).

3.5.5a Carbonate Highstand System Tracts

These may exhibit widespread (Fig. 3.2B) early aggradational (sigmoidal) and later mounded to progradational (oblique) parasequence geometries on shelves and ramps (Sarg, 1988). The systems tracts downlap on to marine-condensed sections (or hardground / burrowed surfaces), and they are terminated by either type 1 or 2 erosion surfaces. During early moderate rates of relative sea-level rise, increased rate, of added accommodation space result in deeper, more restricted marine conditions on the shelves (i.e., lower oxygen content, lower temperature, poor nutrients, and/or high salinity). The result is catch-up carbonate deposition (Kendall and Schlager, 1981) characterised by slower sediment production and depositional rates, deeper water and lower energy micritic textures and significant early submarine cementation during extended exposure to marine waters. As relative sea-level rise later diminishes, there is generally increased circulation, nutrient production, oxygenation and higher temperatures on the shelves, leading to keep-up carbonate deposition. This phase is characterised by rapid sediment production and deposition, shallow environments, grain-rich/mud-poor sediments, and limited marine cementation. Evaporite deposition may occur in proximal platform areas within shallow lagoonal and sabkha environments. Subaerial exposure during subsequent type 1 or 2 falls of relative sea-level provides the framework for dissolution, fracturing and eventual evaporite replacement/dolomitisation, factors that either enhance or diminish porosity and permeability.

3.6 Depositional Systems Tracts & Hydrocarbon Occurrences

In this section, the hydrocarbon-play potential of the various systems tracts and their component depositional elements will be considered in terms of *reservoir, trap, seal(s) and source* by integrating sequence and depositional systems concepts. Reservoirs are component facies within depositional systems which have primary and/or secondary-enhanced porosity and/or diagenetic processes influenced by relative sea-level cycles. Traps herein are (1) primary stratigraphic pinchouts related to depositional processes (e.g., lateral facies changes and lapouts on to discontinuity surfaces) and/or secondary porosity/ permeability pinchouts caused by diagenetic barriers (e.g., dissolution, karstification and cementation) related to subaerial exposure or deep-marine diagenesis, both influenced by relative sea-level cycles or (2) non-tectonic structural closures (e.g., growth faults, compactional drape) resulting from syndepositional or early post-depositional processes. *Seals* include flooding surfaces, transgressive, marine-condensed sections and downlapping shales, toplapping subaerial mudstones and diagenetic barriers. *Source beds* are principally marine-condensed sections, distal highstand mudstones and delta-plain organic facies which underwent appropriate burial/ thermal/temporal maturity and which have available pathways for vertical and lateral migration to reservoir.

The concept permits extrapolation of the critical association of play elements once a depositional systems/sequence framework has been constructed. Consequently, extrapolation of plays based on the occurrence of the key play elements provides the basis for establishing a play trend, sometimes called a play fairway. The fairway is a geologically delineated area where the 3-dimensional play elements are inferred to occur and, hence, constitute prospective areas. Unassessable until after further drilling and analysis are factors which may affect the quality and quantity of the plays (e.g., limiting diagenesis, reservoir volumes/quality, adequate seals and migration pathways). If these factors are favourable, the plays may provide an idea of the prospectivity of the basin.

3.6.1 Lowstand Systems Tracts

This systems tract probably contains the greatest potential for stratigraphic hydrocarbon plays because of its several distinctive depositional elements associated with type 1 erosion, localised lowstand deposition, common juxtaposition to seal and source beds, deep burial and in some basins, the occurrence of contemporaneous growth faults near shelf breaks. Two potential reservoirs may exist in carbonate successions, allochthonous debris-fan wedges and autochthonous wedges (Sarg, 1988), as well as secondary porosity and permeability generated in pre-existing highstand facies during subaerial exposure. Autochthonous lowstand carbonate wedges may contain shallow-water, high energy keep-up grainstone and packstone reservoirs, including some reefal buildups. Traps may be stratigraphic and possess marine-condensed seal and source beds. During lowstand, subaerially exposed relict highstand facies typically undergo dissolution and karstification to produce secondary plays.

3.6.2 Transgressive Systems Tracts

These systems tracts, which are deposited during accelerating rise of relative sea-level, may contain carbonate plays. If well developed, the tracts display a retrogradational geometry and a series of diachronous transgressive surfaces which, up dip of the lowstand or shelf-margin tracts, coincide with type 1 or 2 unconformities. In carbonate transgressive tracts, if marine circulation and other factors favour biogenic production in balance with rising relative sea-level, thick retrogradational keep-up grainstones, oolites and packstones with limited marine diagenesis may characterise the tract (Sarg, 1988). Because of typically slow subsidence rates in such basins, the reservoir quality may also be affected by many higher orders of eustatic cycles and paracycles which can induce short-term exposure and, hence, dissolution that enhances porosity and permeability. Stratigraphic traps, seals and source beds are typically distal transgressive and marine-condensed micritic limestones and marls.

3.6.3 Highstand and Shelf-Margin Systems Tracts

These two systems tracts exhibit very similar depositional systems. Highstand systems are deposited during late decelerating relative sea-level rise and shelf-margin systems are deposited during early accelerating relative sea-level rise (following a slow relative fall that did not drop below the depositional shoreline break).

In carbonate basins, highstand and shelf-margin systems tracts have excellent primary hydrocarbon-play potential if the biogenic productivity and rising relative sea-level are balanced, permitting extensive deposition of keep-up grainstones, oolites and packstone facies (Sarg, 1988). Classic reservoirs are thick shelf-edge buildups which may occur under optimum rates of sea-level rise. During late highstand, hypersaline dolomitisation associated with restricted evaporite deposition and meteoric dissolution associated with early falling sea-level may continue during subsequent type 1 falls of sea-level to enhance porosity and permeability. Traps may be structural, but stratigraphic traps may result from complex diagenesis of shelf and platform-edge facies and buried biogenic buildups. Deep-water micritic catch-up sediments, deep-basinal muds and marls and marine-condensed sections are typical source beds. The marine-condensed section also may provide a seal for both highstand and shelf-margin tracts.

Chapter 4

Sequence Stratigraphic Framework of the Tamet Formation

4.1 Problems of sequence boundary recognition

No evidence is recognised anywhere on the Tamet ramp for major erosional truncation that may have been the result of long-term subaerial exposure and basinward shifts of microfacies. Also in the absence of biostratigraphic calibration with mid-Eocene strata in the eastern Sirte Basin, it is extremely difficult, and perhaps impossible, to locate precisely the genetic surfaces of these unconformities and the extent of their hiatuses, particularly since there are no detailed chronostratigraphic subdivisions. Therefore, sequence boundaries bracketing the three depositional sequences in the Tamet Formation are interpreted to be type 2 sequence boundaries generated when the rate of third-order eustatic sea-level fall was not sufficient to drop below the platform top across the passive margin.

The biostratigraphic subdivision used in this study area corresponds to the zonation of the middle Eocene by Abul-Nasr and Thunell (1987) and Baum and Vail (1988). This zonation provides correlation with planktonic foraminifera and calcareous nannoplankton for Lutetian and Bartonian time. The apparent good correlation of the Tamet Formation with the middle Eocene of Egypt could simply be the result of the whole of the southern passive margin of the Tethyan Ocean being subject to the same tectonic regime which would also have existed in Southern Europe and the Middle East.

Aubry (1991) suggested that there is ample evidence of widespread unconformities within middle Eocene strata. However, it cannot be assumed that these correlate (in the stratigraphic sense) and so result from single eustatic lowering events because there is no well-established mechanism yet known which could have initiated their simultaneous development on the platforms and slopes in the middle Eocene.

The boundaries between the sequences are not classic sequence boundaries described by Van Wagoner *et al.* (1988) because there is no evidence of erosional truncation, and/or subaerial exposure. In the absence of such features it becomes much more arbitrary and difficult to detect the sequence boundaries. Under these circumstances, these "transitional boundaries" do not meet the original definition of sequence boundaries by Van Wagoner *et al.* (1988).

Failure to identify the sequence boundaries reported in the Exxon global cycle chart (Haq *et al.,* 1988) in Lutetian and Bartonian time may reflect a combination of unrelated causes.

1. Sea-level cycles affect flat-topped platforms with a steady background of tectonic subsidence during greenhouse times are probably characterised by low-amplitude oscillations (Tucker, 1993). This could keep the platform tops near to sea-level during the apparent eustatic falls. Therefore, the boundaries between the depositional sequences are defined by *zones* of renewed phases of deposition, because the sea-level may not drop below the platform top between deposition of these sequences.

2. The sequence boundaries may have formed during short-lived periods (about 1 Ma. or less) with insufficient time to cause large-scale subaerial features. However, such short-lived subaerial exposure cannot be identified biostratigraphically or by other features.

3. The erosional effects of a subsequent rise of relative sea-level might also have removed some of the early diagenetic features developed at the top of the sequences (Walker and Eyles, 1991).

Similar to the conclusion of Schlager (1991), middle Eocene sequences in this area require a more flexible definition of sequence boundary than that given in Van Wagoner *et al.* (1988). The "sequence boundary" of Van Wagoner *et al.* (1988) requires subaerial exposure with a significant hiatus and /or erosional truncation (even with platform-margin systems tracts and a type 2 sequence boundary). As discussed above, during the long-term sea-level fluctuations (third-order), thick carbonate successions should be deposited, and boundaries between the sequences are stratigraphically *transitional*, without an overlying basinward microfacies shift, and are located on the basis of cycle stacking. Long-term changes in accommodation linked to low-amplitude third-order global eustatic cycles on passive margins are the main control on this low-order cyclicity. The background subsidence is assumed to be constant (Gumati and Kanes, 1985), or at least changing at a very slow rate over the time span of Tamet deposition. Therefore, Schlager's (1989; 1991) definition that "a sequence boundary represents a geometrically manifest change in the pattern of sediment input and dispersal" in a basin is more acceptable.

As demonstrated in this study, third-order sequence boundaries of the mid-Eocene platform in Libya are stratigraphically transitional rather than sharp unconformities. At these transitional boundaries, there is no evidence of a downward shift in the overlying facies, no lowstand wedge on the downdip ramp, and also no indication of an erosional hiatus. Goldhammer *et al.* (1993) have suggested that this transitional surface is a type 2 sequence boundary (in the *sense* of Vail, 1987 and Sarg, 1988). Such conformity-bounded sequences would not be identified in seismic sections because of the lack of seismic discontinuities at their boundaries (i.e., lack of erosion or karstification in a pure carbonate system), and the lack of pronounced stratal geometries. These transitional zones are recognised only by cycle stacking patterns without an overlying basinward facies shift, and are identified on the basis of anomalous facies changes where Walther's Law of adjacent versus vertical facies relations appears to be violated. Goldhammer *et al.* (1990) suggested that this transitional type of sequence boundary will be a particular attribute of shallow-water carbonates because of the characteristic flat-topped geometry of carbonate platforms. Its reasoned that with such flat-topped platforms any slow third-order sea-level fall will have an essentially equal effect across the almost entire platform.

This in turn, would allow the deposition of "conformity-bounded" sequences across almost the entire platform. Such conformity-bounded sequences would not be identified in seismic sections because of the lack of seismic discontinuities at their boundaries and lack of pronounced geometries.

4.2 Identification of transitional zones

Detecting transitional boundaries is not always straightforward. Detailed analysis of the lithology, biostratigraphy, comparison with global eustatic sea-level curves, and cycle stacking patterns are generally the best means of detecting such boundaries. In the Mid-Eocene of the eastern Sirte Basin all transitional boundaries are type 2 (*sensu* Van Wagoner *et al.*, 1988), and three third-order sequences can be recognised. The planktonic foraminifera and calcareous nannoplankton assemblages give an age ranging from Lutetian to Bartonian (about 8.5 Ma. duration according to Haq *et al.*, 1988). In the study area, two transitional boundaries occur between two major unconformities. These can be correlated with the eustatic sequence boundaries at 48.5 and 40 Ma respectively on the Haq *et al.* (1988) chart. These transitional zones and unconformities are present on the shallow-water platform, which was characterised morphologically by a flat top.

It is clear from the evidence presented in this study later that the diagenetic pathways in the middle Eocene carbonates were strongly controlled by an arid climate against a background of eustatic peaks, with saltern progradation resulting from shoreline migration across the platform. Hypersaline brines originated within the Cyrenaica Platform refluxed downward and seaward through the cyclic section causing widespread regional dolomitisation (Ahmed, 1992); the dolomitisation is often associated with leaching of the metastable skeletal grains and generation of moldic, vuggy and intercrystalline porosity. The association of dolomitisation with sea-level fluctuations might be a clue to the mechanism for the formation of thick dolomites in the Tamet ramp sediments, as noted by the increasing ratio of dolomite/ limestone near the transitional boundaries. This apparent association led Sun and Esteban (1994) to propose that transitional boundaries are the common cause of updip high-frequency diagenetic alteration.

Above the transitional zones, the frequency of storm activity was high during the period following the initial rise of relative sea-level; there was extensive re-mobilisation of bioclasts. Subsequent repetitive processes of reworking of bioclasts and amalgamation of beds affected the sediments immediately upon the transitional boundaries across the ramp and Cyrenaica Platform. The successions are characterised by rapidly deepening-upward deposits, leading to catch-up with the new space created after the relative sea-level rise.

This widely-correlative interval of storm beds serves as an effective depositional timeline and demonstrates the use of the individual storm-event cycles as chronological markers in ramp settings (Aigner, 1985).

The physical expressions of the third-order transitional boundaries (termed Tbs) are quite similar. For example, the Tb1 at the top of the lower sequence is marked mainly by significant hypersaline features. This is the time when dolomitisation and other diagenetic products were developed on the flat-top of the Tamet platform; these probably reflect late highstand-early lowstand of relative sea-level or widespread subaerial exposure of the Cyrenaica Platform because relative sea-level was actually falling.

Further eastward and across the Cyrenaica Platform, the stacked carbonate-evaporite cycles reaching thickness 185 ft (57m) provide a record of brine-level fluctuations. The stacking of these hypersaline cycles indicates that during periods of third-order eustatic rise, sea-level was subjected to small-scale episodic rises and stillstands, followed by coastal progradation. In this case, evaporite microfacies represent periods of non-deposition and could be related to complete isolation of the Cyrenaica Platform associated with long-term arid conditions. Evaporite signatures represent regionally correlatable events. However, the use of brine-levels as markers to define the transitional boundaries is more applicable. Stacks of hypersaline cycles with the eastern Sirte Basin are not unique to the Tertiary. Similar carbonate-evaporite stacks have been documented within the late Miocene (Messinian) on Sicily and are classically separated into cycles by a subaerial unconformity (Butler *et al.,* 1994).

The transitional zones are overlain by bioclastic cycles, which may be interpreted as storm-derived material transported into a pre-existing depression in the shallow ramp setting. An alternative is that following the initial rise of relative sea-level new topographic lows, may have formed and been involved with lithoclasts produced under the relatively high-energy conditions. This indicates that during the early transgressive phase, limited accommodation space was available in the ramp interior region, and so that large amounts of a variety of reworked bioclasts was transported basinwards.

4.3 Depositional framework

The middle Eocene palaeogeography of the eastern Sirte Basin was largely inherited from Upper Cretaceous times, and the basin infill displays a very low-order aggradational to progradational stacking pattern. Cyclic carbonates were deposited on a flat-topped, fully-aggraded platform that extended approximately 200 to 250 km across the middle Eocene passive margin of the southern Tethyan Ocean. During long-term (third-order) sea-level rises, times of carbonate deposition should be long, cycles should be thick, and boundaries between sequences and their components should be thin and *transitional zones.* These depositional sequences commonly range from 1 to 3 Ma. in duration, and reflect long-term changes in accommodation linked to low-amplitude, third-order global eustatic cycles on passive margins. The main control on this low-order cyclicity may be due to the decrease in the rate of subsidence (Gumati, 1986).

Two scales or orders of cyclicity are apparent in the strata studied here. Using the hierarchical classification of Goldhammer *et al.* (1990) and Mitchum and Van Wagoner (1991), fourth-order cycles (parasequences) can be recognised within the long-term third-order sequences. In the absence of detailed age constraints for these stratal units in the study region, this hierarchy can only be applied in a genetic or relative sense.

The Tamet Formation has been divided into three stratigraphic sequences (termed DEP. SEQ.-T1, T2 and T3); two of them are continuous from the platform to the basin transition, with a tabular shaped and tilted base. The Lower-Middle Eocene boundary lies at the base of DEP. SEQ.-T1 whereas DEP. SEQ.-T3 extends up the boundary between the Middle and Upper Eocene.

Tamet sequences are regionally traceable in the middle Eocene of the Sirte Basin with a thickness about 80-420m. In the western part of the study area, deep subtidal to shallow subtidal sediments were deposited (average approximately 200-300m thick); third-order amalgamated sequences accumulated. During this phase the subtidal deposits display a catch-up succession on the marginal-platform. Towards the platform, the progressive slowing of the third-order sea-level rise resulted in upward-thinning cycles (parasequences) from the base to the top. Subtidal sedimentation was able to keep-up with net sea-level change, leading to third-order rhythmic sequences, in some cases with diagenetic caps representing periods of late highstand of relative sea-level. The boundaries separating these third-order sequences are stratigraphically *transitional zones*, without an overlying basinward facies shift, and are located by the vertical stacking patterns (systematic variation in the thickness and microfacies reflecting the long-term accommodation changes).

High-frequency (metre-scale) shallowing-upward cycles are the building blocks of the large-scale, lower-frequency (third-order) depositional sequences on the Tamet Platform. These high-frequency cycles are equivalent to parasequences in siliciclastic deposits (Van Wagoner *et al.,* 1988). Cycle stacking patterns are most likely controlled by long-term third-order changes in accommodation space and provide the crucial link between the individual (meter-scale) cycles and the large-scale depositional sequences (Goldhammer *et al.,* 1990).

Establishment of a sequence stratigraphic framework is based on an integration of microfacies identification and the nature of their shifting and stacking within the sequences. The sequences and systems tracts identified here in the middle Eocene were compared with those obtained in other sedimentary basins. The Tamet sequences, their internal architecture, and stratal geometries can be matched with the third-order sequences of Baum and Vail (1988) to pick candidate stratal surfaces (e.g. sequence boundaries, transgressive surfaces and maximum condensation surfaces), and to evaluate the duration times of each sequence, regardless of whether carbonate or siliciclastic sediments are involved.

For the purposes of this chapter, only one sequence will be described. The remaining sequences also comprise these same systems tracts, though in some cases not as well expressed.

During DEP. SEQ.-T1, sedimentation was widespread in the eastern part of the study area. Its more dramatic expression is on the platform margin, where well AA1-6 is located on a structural high (intra-basinal dome). The absence or incomplete sequence in this well indicates strong truncation of sequence 1. This could represents an intra-DEP. SEQ-T1 unconformity, which took place prior to the deposition of sequence 2 (DEP. SEQ.-T2). As a result, the whole sequence 1 in well AA1-6 was eroded. The eroded sediments were probably deposited in neighbouring depressions, producing a local thickening of sequence 2.

The transgressive and highstand systems tracts of DEP. SEQ.-T2 represent an important onlapping from the depressions to the surrounding high. After this onlap the ramp became morphologically more uniform and sedimentation during DEP. SEQ.-T2 and T3 was more widespread due to relatively uniform platform topography in most of the studied wells.

4.4 The link between metre-scale cycles and sequence stratigraphy

The sequence stratigraphy model developed for the Tamet Formation (Fig. 4.1) is based on the role it plays in the linkage between meter-scale cyclic stratigraphy and seismic-scale sequence stratigraphy. Meter-scale (depositional cycles) stratigraphic packages stack into discrete sets which aggrade or prograde depending on the relative rates of accommodation development and carbonate production. However, little detailed work could be carried out on the large-scale depositional patterns. It may be partly the absence of erosional unconformities that has led to third-order sequences being more difficult to define in certain carbonate successions compared to siliciclastic strata. In this dissertation mid-Eocene ramp carbonates are used to demonstrate that high-order cycles can be identified in the Tamet ramp successions. This is not simply a matter of semantics because the process by which high-order cycles and sequences are generated is different and their correct identification is important in understanding the relative rates at which sea-level is fluctuating within a basin. These stacking patterns potentially allow the division of the Tamet sequences into transgressive (TST) and highstand (HST) systems tracts.

4.4.1 Transgressive systems tract

The transgressive systems tracts of the Tamet sequences are composed of normal marine strata and they appear to have been subtidally deposited because no peritidal microfacies or subaerial exposure features were identified. Most deposition in this episode is characterised by an overall deepening-upward of both lithofacies and biofacies, as well as an increase of early marine diagenesis marking a progressive decline

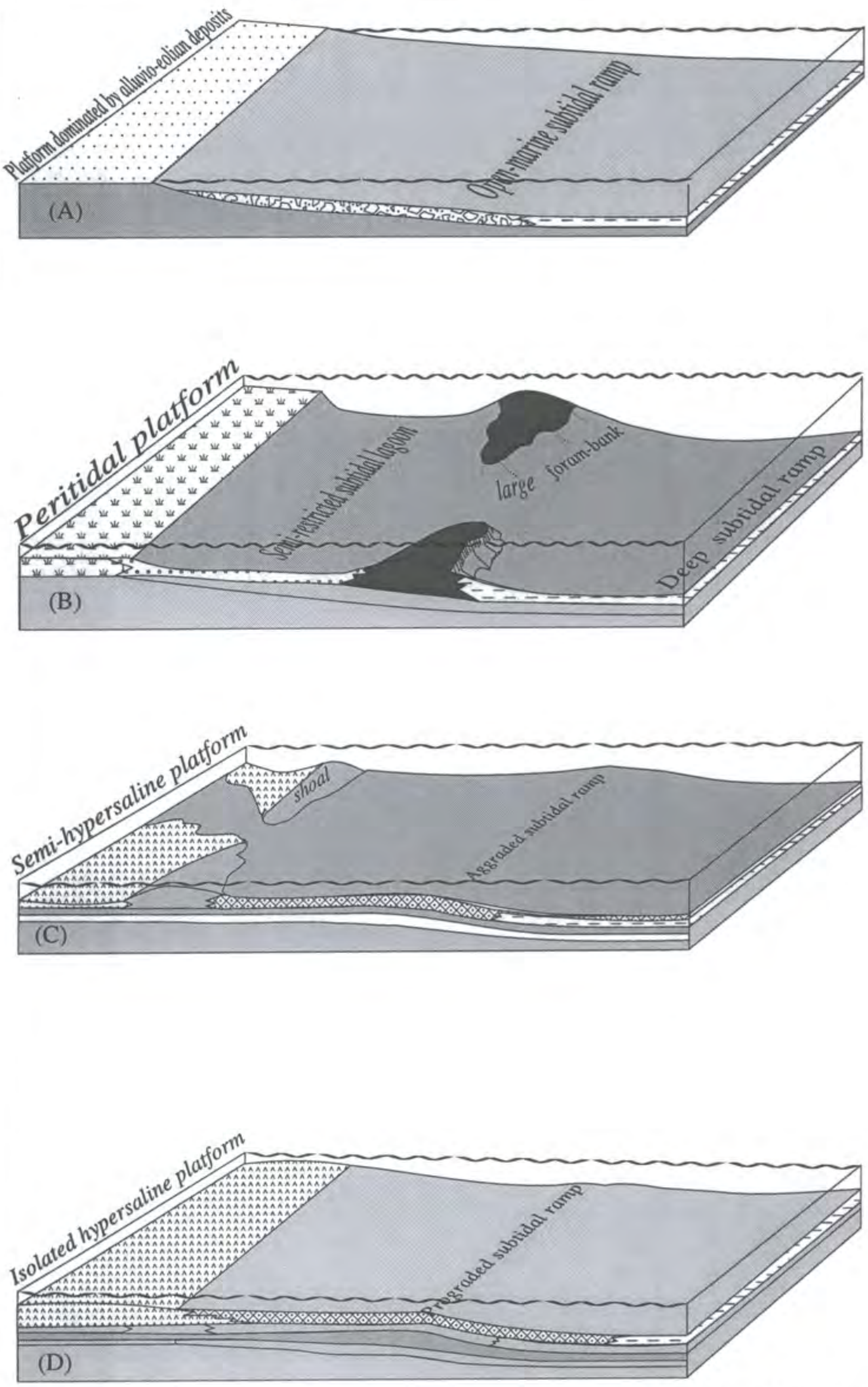

Fig.4.1 *Schematic sequence stratigraphic model showing the middle Eocene depositional style on the eastern flank of the Sirte Basin. Changes are interpreted to reflect fluctuations in sea-level and carbonate productivity. Relative sea-level (A-D) is the sea surface relative to the position of a datum (e.g., the lower Eocene Gatter Formation) at or near the sea-floor. Regional subsidence is assumed to be negligible. The time period covers the deposition of Tamet Formation.*
A) Initial flooding of the ramp, slow rate of creation of accommodation space allowed rampwide depositon of storm-dominated cyles. The Cyrenaica Platform in this stage probably was dominated by alluvo-eolian deposition.
B) Flooding stage, rapid and maximum rate of creation of accommodation space. In most cases, the distal outer-ramp was dominated by a deep subtidal environment. Large foram-banks formed during a few flooding events mainly on palaeohighs and resulted in back-bank subtidal lagoons. The maximum landward extent of sea-level rise permitted open-marine peritidal deposition across the Cyrenaica Platform. C) Slowing rate of creation of accommodation space and lower carbonate accumulation resulted in relatively thick non-cyclic deposits across the ramp; probably the deposition did not approach sea-level. D) slower sea-level rise and high carbonate production rate over the ramp resulted in large-scale progradation into the ramp-margin. Towards the Cyrenaica Platform, slow rate of accommodation space creation allowed the saltern deposits prograde into the lagoon.

in sedimentation rates related to increased palaeowater depth. The lack of evidence of shallow subtidal caps in areas downdip of the basin transition, indicates that sea-level changes were too rapid to allow the microfacies of the ramp interior to prograde seaward.

Deposits representing this episode across the ramp are about 80m thick in the marginal are and not more than 20m on the platform interior. The transgressive systems tracts of the Tamet sequences may be subdivided into an early transgressive phase (the lower lithoclast-dominated cycles) and a late transgressive phase (the upper foraminifera-bearing cycles). The early transgressive systems tract (Fig.4.1A) across the Tamet platform is characterised by storm-dominated cycles, which may be present as storm overwash during the lag period prior to carbonate production and following the initial rise of relative sea-level. This indicates that limited accommodation apace was available in the ramp region so that large amounts of a variety of clasts were reworked basinward. These cycles depict monotonous conditions of subtidal sedimentation. Each cycle is composed of two depth- and energy-controlled microfacies and the vertical variations record deposition below and above a fluctuating zone of storm wave reworking in a shallow ramp setting.

In an updip position, the early transgressive deposition was influenced by fairweather-wave reworking, when relatively thin and bioclastic-dominated microfacies were deposited. The lack of basal shell lags at the base of these cycles is probably due to the "patchy" distribution of benthic foraminifera. These skeletal materials were transported by storm-generated currents and then reworked by oscillatory shear currents (Aigner, 1985). This microfacies is overlain by a thick unit of fine skeletal fragments in a pelagic matrix. Deposition of this microfacies probably occurred during waning storm conditions, combined-flows deposits settled from suspension during fairweather conditions and were draped by a pelagic matrix. These cycles more likely represent deposition from multiple storm events during initial short-term sea-level rises;

comparisons with modern analogues suggest that cycles were deposited in water depths of between 20-30m.

Downdip in the Tamet system, the lateral deep-water equivalents are mud-rich cycles and these are interpreted as deep ramp storm-tempestites. The lack of wave and current reworking of microbio-lithoclasts indicates that these cycles were deposited further away from the sources. Analogies with modern storm deposits (Aigner, 1985; Reading, 1986; Sneeden *et al.*, 1988) suggest deposition in water depth between 20-40m (assuming a 40m storm-wave base)

Storm-influenced platform deposits have been reported by numerous authors from many different depositional environment ranging from deep sea-pelagic to shallow subtidal settings. Duke *et al.* (1991) made the connection between ramp physiography and laterally extensive amalgamated storm deposits; the latter tend to be more common on very gently dipping, low-angle ramps than flat-topped shelves, because open ramps would be subjected to constant high-energy conditions due to their vulnerability from strong storms. Recent equivalents are the offshore area of Lily Bank (north of Little Bahama Bank; Hine, 1977) and Cat Island (eastern Great Bahama Bank; Dominguez, 1988), which have similar deposits from an initial phase of the Holocene sea-level rise.

During continuation of the relative sea-level rise, the ramp margin eventually became the site of deep subtidal cycles, which are characterised by an overall deepening-upward trend. Chalk and pelagic micrites represent the depositional cycles of a transgressive systems tract on the ramp margin and are interpreted as having been deposited in an open marine, relatively deep subtidal environment, that was occasionally influenced by storm-induced waves. The top of these cycles at this site is typically marked by abundant and diverse planktonic faunas in the form of a "condensed section". The greater abundance of planktonic specimens on the platform margin suggests that deposition occurred at depths >50m with minimum water turbulence. During the long-term flooding phase, sea-level oscillations may have occurred too far above the sediment-sea floor interface to cause a distinct change in microfacies types, and therefore the high-frequency sea-level signal is not recorded in this direction.

The matrix of this microfacies in the distal parts of the ramp gradually changes upwards to phosphatic micrite with increased patches of framboidal pyrite; wisps of organic matter are common. The development of a biocondensation horizon marking very slow rates of deposition aids in the identification of a maximum flooding surface. However, it is unclear on a flat-topped, fully aggraded ramp, where no condensed section exists and other criteria for the recognition of maximum flooding (e.g. organic matter-rich, phosphate and hardground) are lacking; in these cases cycle stacking patterns provide the best information for identifying the transition. Similar to sequence boundaries, the transition between the transgressive systems tracts and highstand systems tracts is complicated and is considered to be a *zone* of maximum flooding.

In the interior platform, these pelagic deposits replaced by cyclic bank facies with larger foraminifera especially nummulitids, operculines and discocyclines (Fig.4.1B). The term bank here is used to designate an individual structure having low positive-relief and discernible morphology inferred to have formed by hydrodynamic processes (Aigner, 1983).

In the past, many authors (e.g. Bishop, 1975; Moody, 1987; Keheila and Al-yyat, 1990) have interpreted the larger foraminifera buildups as shallowing-upward cycles. In contrast, Ahmed (1992) concluded that the development of larger foraminifera banks in Palaeocene and early Eocene sections of the Agedabia-Augila Trough, on the eastern portion of the Sirte Basin, occurred in two stages, one of bank formation and then bank flooding. He suggested that foraminiferal accumulation occurs only during storm phases of a eustatic rise, that foraminifera influx decreases as eustasy peaks (i.e. when the rate of sea-level rise reaches a minimum). By using published literature of similar age, the presence of submarine features such as mineralisation (Sbeta, 1991), hardgrounds and borings (Danielli, 1988) further supports Ahmed's interpretation.

Palaeotopographic highs (probably formed by normal faults) within the shallow outer-ramp became the site of foraminifera accumulations. Two cycles are encountered in the Tamet nummulitic strata. Meter-thick packages of larger-foraminifers occur as discontinuous horizons and exhibit a more domal geometry; they can be traced for tens of kilometres across the ramp. Each cycle exhibits a gradual upward decrease in the abundance, diversity and size of forams, as well as a transition from foram-bearing to mud-rich deposition.

Ahmed (1992) described the basal microfacies as a storm-dominated bank and he related its origin to primary active production of larger foraminifers and sporadic storm events. Progressive deepening within individual cycles is indicated by subtly grades allochthonous debris and fine muds deposited under weak traction currents as energy waned and the storm receded. Presumably this microfacies reflects a somewhat deeper water condition than the underlying beds and could be interpreted as overbank microfacies. This vertical change from foram-bearing to a mud-rich bank represents the temporary or prolonged marine inundation into shallow realms. However, the top of foram-rich cycles are interpreted as candidate maximum flooding zones and cannot be traced laterally across the subtidal lagoon cycles.

Further eastward, the subtidal lagoonal deposits are a lateral stratigraphic-equivalent to foram bank cycles. There is not enough evidence to indicate either latest transgressive deposition or early highstand deposition. El-Hawat *et al.* (1986) believed that the stacking of foram-banks around the lagoon, may have a great influence on sedimentation inside the lagoon. An idealised cycle in the lagoon reflects the fluctuation from an open-to-semi restricted shallow-subtidal environment. These deposits accumulated behind foram-banks and infilled the deep inner-ramp lagoon. The open lagoon systems consisted of orbitolinid-echinoderm wacke-packstones deposited in water depths between 10 and 35m, with conditions of normal circulation and salinity. As the foram-banks continued to stack vertically, they must have acted as a sill and restricted circulation within the lagoon.

Zones of maximum flooding in Tamet sequences are characterised by the vertical transition from thicker cycles to successively thinner cycles with a concomitant increase in shallow subtidal microfacies. This shift in stacking patterns may reflect the progressive decrease in accommodation space as the rate of sea-level rise slows towards the eustatic peak. The lack of an abrupt backstepping of the larger foram-banks on the shallow ramp accompanied by the absence of the offshore buildups suggest that the drowning was gradual. This result supports Schlager's (1991) conclusion that the third-order eustatic cycles of Haq *et al.* (1987) did not always lead to carbonate platform drowning.

4.4.2 Highstand systems tract

This depositional tract is more aerially extended than the transgressive systems tracts. Deposition progressed from a mud-dominated ramp, best developed during transgressive catch-up deposition to a grain-dominated ramp, common in the late highstand keep-up deposition. This reflects a long-term decrease of accommodation space, from the deepening phase followed by an overall shallowing of the platform.

This gradational changeover in the microfacies architecture is associated by the dramatic changes in the style of diagenesis. For example, in the shallow ramp sediments, the peloidal fabrics were best developed in large foram-rich microfacies of the transgressive systems tract and early highstand peloidal-bioclastic microfacies (catch-up sea-level rise), whereas the hypersaline features are considered to be a very common process, particularly in the late highstand deposits (keep-up sea-level rise).

Basinward, the highstand systems tract is represented by strong prograding of peloidal-bioclastic wacke-packstones which downlap underlying chalky-pelagic mud-wackestones, thereby documenting a shift in the platform margin. The boundary between the transgressive and highstand deposits is regarded as the maximum flooding zone. This zone marks the change from aggradational to progradational microfacies architecture, corresponding to the change from transgressive to highstand. However, this close juxtaposition of two entirely different microfacies, suggests the shallow subtidal microfacies are easily able to keep pace with relatively slow sea-level rises.

The highstand deposits covered an extensive area of the ramp interior and are very difficult to separate from the transgressive deposits, because both consist of vertically stacked subtidal, *non*-cyclic microfacies (Fig.4.1C). Consequently, it is difficult to explain these microfacies as the result of rises or highstands of sea-level, and so their respective lower boundary is not "drowning" or "suffocation". Tamet strata of the early highstand systems tract vary in depositional texture (mainly between wackestones and packstones) and fauna, but distinct shallowing-upward trends are difficult to identify. Some large foram-rich beds (packstones) may indicate the top of the shallowing-upward cycles, but deposition probably did not approach sea-level in most cycles.

The platform interior parts are dominated by bioclastic wackestones and packstones and marked by abundant and diverse open marine fauna; thus they are referred to as open ramp microfacies in which shallowing-upward cycles were difficult to distinguish. The controlling mechanism behind incomplete shallowing in these units could reflect a subdued sedimentological response to a long-term increase in accommodation space. Sea-level may have been oscillating too far above the sediment surface to cause a distinct sedimentologic change in microfacies.

This microfacies developed on the outer and inner ramp at intermediate water depths above the zone of storm wave reworking, and reflects the response of the Tamet interior to catch-up deposition to fill space created by this episode; this most likely is the reason for the development of subtidal *non*-cyclic (amalgamated microfacies) units within the early highstand depositional tract. This situation is similar to bioclastic packstones which comprise most of the back-reef sediments in the Pleistocene and Holocene of Enewetak Atoll (Goter and Friedman, 1988). Those Pleistocene and Holocene back-reef sediments rarely show systematic vertical changes in the depositional lithologies.

As the rates of eustatic rise slowed down, the water depths decreased; therefore, rapid deposition on the Cyrenaica Platform allowed saltern sequences to aggrade-prograde (Fig.4.1D) across very shallow parts of the Tamet ramp. This pattern of keep-up deposition and outbuilding is a characteristic feature of late highstand systems tracts (Sarg, 1988). The apparent increase in amplitude of sea-level on the adjacent submerged ramp tended to inhibit hypersaline cycles prograding completely across the Tamet ramp at these times (c.f. Wright, 1984), but allowed extensive brines to reflux downdip to dolomitise the tops of subtidal *non*-cyclic units. Other factors, in addition to sea-level oscillations, likely played a role in generating saltern deposits. These factors may include long-term tectonic quiescence and accompanied by more arid conditions, which influenced sedimentation over the Cyrenaica Platform after the connection to the open Tethyan Ocean was lost. This is illustrated by the saltern deposits, which are intimately related to cycles of sea-level change and environments ranging from very restricted, shallow-subtidal to saltern evaporites.

Cyrenaica sequences deposited as part of a basin-wide setting, passed seaward into time-equivalent open-marine, subtidal deposits. The platform was up to several hundred kilometres wide and separated from the open sea by a beach-ridge complex and/or possibly a tectonic sill. The deposition on the Cyrenaica during this episode recorded the combined effects of arid climate and smooth rise of relative sea-level. Several stacked cycles must have resulted from the aggradation-progradation of the Cyrenaica Platform following a relative sea-level rise. There is no evidence at the scale of these cycles for significant lowering of depositional base level to indicate that the deposition was subaqueous. A typical arid cycle consists of restricted, very shallow subtidal microfacies as indicated by the limited diversity of biota. The depositional textures of this microfacies have been overprinted by dolomitisation. This microfacies is overlain by massive anhydrite. The extensive evaporite precipitation would occur in response to increasing salinities upward and reflects the decrease in accommodation space over the Cyrenaica Platform.

The most remarkable diagenetic characteristic during the late highstand systems tract is pervasive dolomitisation affecting most of the Tamet's microfacies. It has been suggested that the bulk of these dolomites may have been formed by the reflux of the hypersaline brines generated within the restricted platform (Ahmed, 1992). Another diagenetic feature is the extensive leaching of skeletal aragonite. The paucity of fresh-water diagenesis together with the observation of dolomite cements lining skeletal-moldic pores, suggest that skeletal aragonite was dissolved away by dolomitising fluids (Sun, 1992).

In contrast, the rest of the entire area of the Tamet platform is characterised by large-scale progradation across the drowned outer ramp, suggesting deposition during relatively slow third-order sea-level highstand, with sedimentation rates equalling or moderately outpacing production of accommodation space. This increased the width of the carbonate-producing area and decreased the width of the outer-ramp that had to be crossed by the sediment before it could reach the basin; this increased the carbonate input into the basinal areas. This progradation is dominated by subtidal *non*-cyclic microfacies overlain by dolomitised tops, which formed after deposition as extensive hypersaline fluids migrated downdip. Since there is no evidence of a marginal escarpment, this progradation can be interpreted as depositional to accretional type according to Read (1985).

4.5 Possible mechanisms of sea-level fluctuations on the eastern Sirte Basin

Fluctuations in sea-level occur on both regional and global scales, and range from glacially induced changes over several tens of thousands of years to longer-term variations in the volume of the ocean basins as the rate of sea-floor spreading changes. Furthermore, short-term cycles are superimposed on long-term cycles, which are in turn superimposed on long-term periodic trends.

Donovan and Jones (1979) reviewed the possible mechanisms for sea-level change, and considered that there are only two plausible candidates to explain third-order cyclicity: the waxing and waning of polar ice-caps (glacio-eustasy), and the change in the volume of ocean basins (tectono-eustasy). As discussed by Goldhammer *et al.* (1993) the origin and control of third-order sequences remains problematic, especially for periods of time lacking evidence for major glaciations, such as the Palaeogene.

In general, the third-order sequences can be correlated around the periphery of Laurasian-Tethyan passive margins, suggesting a eustatic control or some underlining subsidence cycle with coincident and widespread effects. This approach has been used by Read and Goldhammer (1988), Read (1989), Osleger (1991) and Kerans and Lucia (1989). In all cases, a similar interpretation has been invokved of a coincident and widespread third-order tectono-eustatic driving mechanism, controlling the magnitude and periodicity of relative sea-level change and facies architecture during the middle Eocene (Abul-Nasr and Thunell, 1987; Baum and Vail, 1988).

During the Palaeogene, it is probable that there were no large polar ice caps so that a glacio-eustatic mechanism is unlikely to account for the third-order cycles of this age. On the contrary, these eustatic changes may correspond to a global tectonic phenomenon (i.e. tectono-eustatic movements), which could have affected a series of genetically-linked basins (e.g. Miall, 1986). Ultimately, many of the possibilities raised above are coupled to the stress fields in intraplate zones (Cloetingh, 1986), a novel mechanism apparently able to generate non-glacial transgression-regression cycles on a third-order scale *sensu* Vail *et al.* (1990). Cathles and Hallam (1991) observed that the buildup of stress can induce changes in plate density that will propagate across the entire plate in less than 30,000 years, and they further suggest that stress at plate margins can cause a transgression by forcing water up onto the platforms and flooding the continents. Likewise, a decrease in sea-floor spreading rates should cause a regression and may result in plate elevation up to several hundreds of meters. Overall, the anomalous regional differences in sea-level change during the Palaeogene, showing simultaneous transgressions in some regions and regression in others, may reflect the position of different areas with respect to regional geoidal highs and lows, rather than changes in the volume of the ocean basins or glacio-eustatic variations.

During periods of very slow subsidence (as in the mid-Eocene time) regional tectono-eustatic changes would have affected at least several Tethyan basins. However, a more systematic study of those basins with a well-documented subsidence history, located in separate tectonic regions is needed in order to test the eustatic origin of sequences.

The deposits of the Tethyan Ocean contain an extensive record of Palaeogene relative sea-level fluctuations. These records are considered to represent very long-term transgressive-regressive cycles. The origin and development of such cycles as well as the longer and shorter-term cycles resulting mainly the tectonic behaviour (Sengör, 1985; Said, 1990) are probably related to the accretion of the African plate towards the European plate during collisional deformation. This motion was caused by sea-floor spreading during the evolution of Atlantic Ocean. These events were followed by a long-term (second-order) sea-level rise covering the southern Tethyan passive margins, beginning in the Upper Albian and continuing into the Eocene, that is punctuated by numerous third-order cycles of relative sea-level.

Cyclic changes in global sea-level and associated relative sea-level changes in coastal onlap during the late Cretaceous and Palaeogene epochs have been recognised by Baum and Vail (1988) and Haq *et al.* (1988). Baum and Vail (1988) used four global

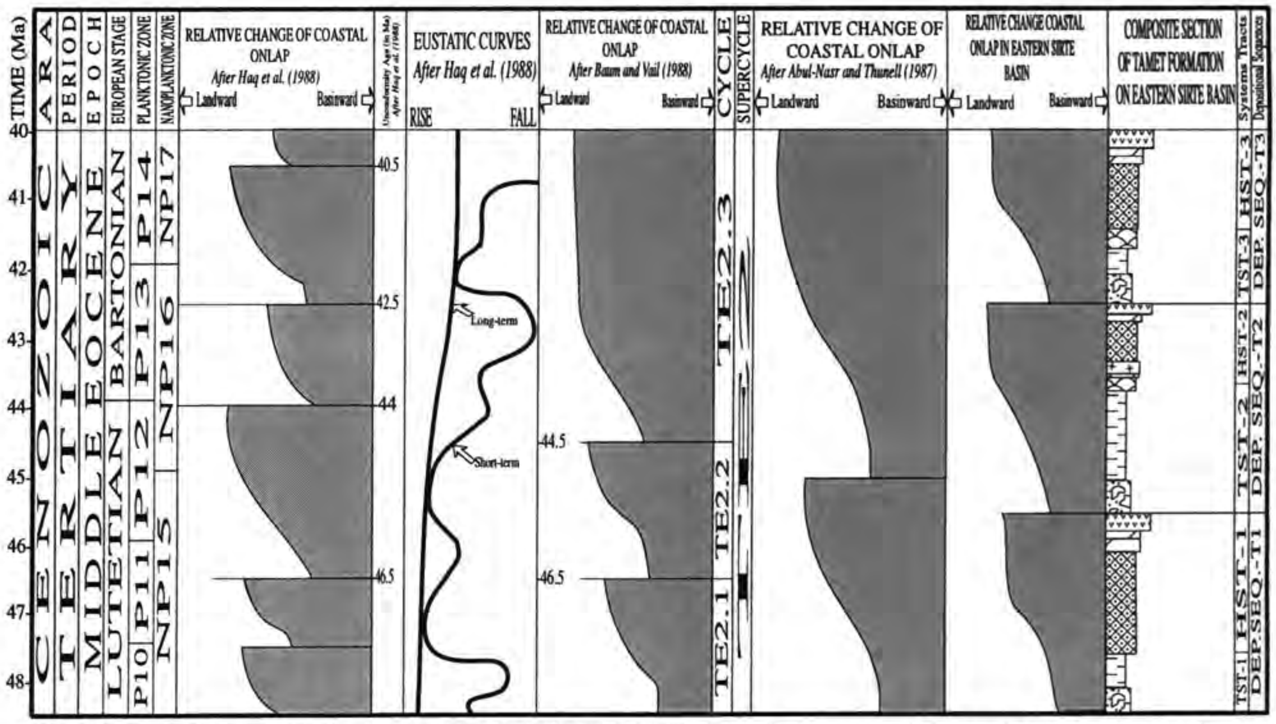

Fig 4.2 *Correlation chart showing ages of stratigraphic sections, relative ages of biostratigraphic zones and comparison of coastal onlap curves with those illustrated on the global chart of Haq et al. (1988) and selected curves of the U.S. Gulf Coast by Baum and Vail (1988); middle Eocene of Sinai, Egypt by Abul-Nasr and Thunell (1987) versus the observed coastal onlap curve of the eastern Sirte Basin, constructed on the basis of depth dependent cycles, which appear as sets of transgressive and highstand deposits.*

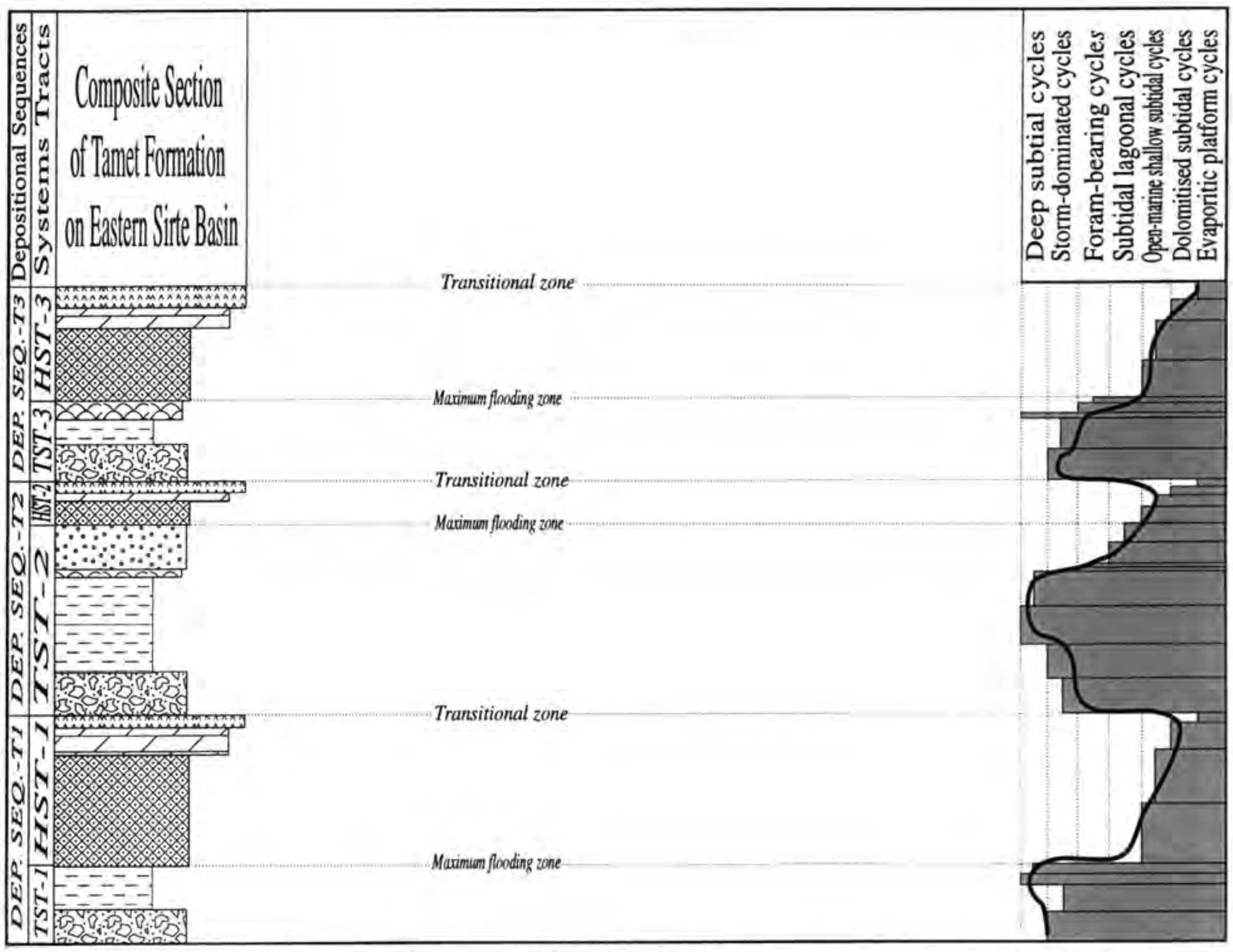

Fig 4.3 *Generalised composite stratigraphic column of the middle Eocene sequences on the eastern Sirte Basin and corr esponding relative sea-level curve. Curve constructed on the basis of metre-scale cycles.*

unconformities to divide the Lutetian-Bartonian strata into three third-order depositional sequences resulting from eustatic sea-level changes. These sequences of Baum and Vail (1988) include TE 2.1 (49-46.5 Ma), the TE 2.2 (46.5-44.5 Ma), and TE 2.3 (44.5-40 Ma).

Middle Eocene strata in the Western Sinai area, Egypt documented by Abul-Nasr and Thunell (1987), can be correlated to the sequences of the Tamet Formation. The apparent correlation of sequences in two widely separated areas supports the idea that individual Middle Eocene sequences were the result of eustatic sea-level fluctuations as proposed by Abul-Nasr and Thunell (1987). Local tectonic movements could have played a significant, although not an easy to read, role in this platform's deposition during the stages of the passive margin development.

Depositional and diagenetic patterns observed in this study area are similar to those described by Ahmed (1992) in the Palaeocene and early Eocene succession of the Agedabia-Augila Trough, on the eastern edge of the Sirte Basin. He also concluded that eustatic fluctuations, together with long-term arid conditions, were largely responsible for the transgressive-regressive depositional sequences and these factors were important in controlling evaporite deposition over the Cyrenaica Platform.

The mid-Eocene strata in the eastern Sirte Basin can be subdivided into three depositional sequences (termed DEP. SEQ.-T1, T2 and T3), which were probably formed under low amplitude, third-order sea-level oscillations (whatever the cause). Most of the sequences defined in this area lack precise biostratigraphic control of the ages of the sequence boundaries. These sequences are regionally correlative, and consist of a lower transgressive part and a regressive highstand part (Fig.4.2). The sequences are traceable across most of the mid-Eocene platform and marginal areas. Based on sedimentological data, the sequences are interpreted as third-order sequences because of the regional nature of the bounding zones and the widespread occurrences of the component systems tracts. The main control on this low-order cyclicity may be a decrease in the rate of subsidence, with time after the main Sirte Basin rifting phase (Late Cretaceous to Early Palaeogene, Gumati, 1985).

Correlation between the global eustatic relative sea-level chart and sea-level history of the eastern Sirte Basin by the classic techniques such as fossil dating is difficult because of the lack of datable fossils in the Tamet section in this area. However, correlation between both curves is facilitated by integrating sedimentological and sequence stratigraphic data. If the locally derived relative curve can eventually be correlated with the global eustatic curve, then this provides a useful method for determining the importance of the global versus local and tectonic versus eustatic mechanisms responsible for the depositional sequence geometries and may aid in age determination of strata.

According to the estimates of the relative depths from the Tamet microfacies within each depositional sequence, relative sea-level elevations for the middle Eocene were in the order of 5m to 100m above the sediment surface; during this time interval, the Tamet shoreline would have shifted back many hundreds of kilometres towards the Cyrenaica Platform during sea-level highstand positions. A. sea-level curve for the Tamet Formation can be constructed (Fig.4.3) based on the vertical changes of meter-scale cycles from distal to proximal areas of the ramp. The curve shows long-term shallowing-upward trend. Relative fluctuations in water depth were determined from interpreted changes in depositional environments of meter-scale cycles. Relative water-depths of microfacies and facies associations in these cycles were estimated from Palaeogene analogies (Ahmed, 1992; Mresah, 1993).

Comparison of the Tamet water depth profile with the global sea-level cycles for the middle Eocene time as published by Haq *et al.* (1987), suggests that at least the large-scale depositional sequences are partly recognisable in the Tamet Formation, which is part of the southern margin of the Tethyan Ocean. However, the geometric expression of the middle Eocene sequences is different from the idealised sequence stratigraphic models (e.g. sequence boundary and maximum flooding surface). The difference arise from the low-amplitude of the third-order eustatic oscillations and due the slow subsidence rates (Gumati, 1985) on the eastern flank of the Sirte Basin during the middle Eocene time. This further suggests a link between plate latitudinal motion and changes in climate as major forces in carbonate platform demise.

4.5 Conclusions

* The middle Eocene Tamet Formation on the eastern limb of the Sirte basin records an extensive history of accommodation changes which resulted from relative sea-level fluctuations of different frequencies and amplitudes. The Tamet shallow-marine platform carbonates contain three, third-order sequences. The sequences display aggradational-progradational deposition during transgressive and relative highstands of sea-level. Sequences and systems tracts within the Tamet Formation are identified solely on the basis of the vertical stacking pattern of depositional metre-scale cycles. The stacking trends of these cycles and gradational shift potentially allow the division of sequences into transgressive (TST) and highsand (HST) systems tracts.

Deposits of the transgressive systems tracts are composed of storm-dominated deposits in the lower part of the transgressive section passing upward into deep-ramp cycles, foram-bearing and lagoonal deposits. The highstand deposits show a general aggrading to prograding character and are dominated by bioclastic wacke-packstone with diverse of open-marine fauna.

* In the Tamet carbonate ramp system, the concept of utilising a downward shift in microfacies to identify the sequence boundaries is difficult. In this case, the sequence boundaries and transitions between systems tracts are gradational in nature and are defined as *zones*. The definition of stratal geometries requires more flexibility than that given in Van Wagoner *et al.* (1988), because during times of low-amplitude sea-level fluctuations, sea-level may not drop below the platform, and subaerial unconformities and/or significant depositional gaps will not develop on the ramp to separate the constructional phases in vertical sequences and to denote the end of a phase of sedimentation.

* The depositional sequences in the mid-Eocene carbonate-evaporite system of the eastern edge of the Sirte Basin are different in some respects from the carbonate models described by Sarg (1988). This study shows that these sequences consist of gradual and rather symmetrical shifts of microfacies, probably due to longer-term (third-order) fluctuations in relative sea-level. Thus many of the stratigraphic features observed in this study are similar to those described by Calvet *et. al.* (1990); Goldhammer *et al.* (1990); Schlager (1989, 1991) and Tucker *et. al.* (1993)

* The Tamet ramp-Cyrenaica Platform transitional grainstone (oolitic or bioclastic) is similar to other inner ramp shoreline sand bodies and major shoal complexes in the geological record. These grainstones form reservoiring systems, and offering a range of subtle stratigraphic play types.

It is important in this context to determine whether grainstone packages were deposited during the transgressive or highstand systems tracts, since porosity-related late highstand hypersaline leaching will be of greatest magnitude in highstand grainstones.
To approach grainstone reservoirs in middle Eocene ramp systems, the use of detailed sequence stratigraphic and diagenetic models is highly recommended.

Chapter 5

General Conclusions

In the eastern Sirte Basin, the middle Eocene (Lutetian-Bartonian) carbonates were deposited in generally shallow-marine environment contained within a vast platform system, which developed on the south Tethyan passive margin. The study of the Tamet strata provide an opportunity to examine deposition and diagenesis of a carbonate platform. These observations led to the following conclusions.

* The Middle Eocene Tamet strata accumulated as genetically related units of carbonate and evaporites corresponding to three major transgressive and relative highstands of sea-level. The abundance of lime mudstones/packstones, diverse open-marine biota, storm deposition and absence of any preserved slope breaks, reef margin or slope and basin sediments corresponds to Read's (1985) definition of a homoclinal ramp. Analysis of ten microfacies suggests that deposition occurred on a low-energy, muddy carbonate ramp, 5-100m deep. Two main depositional processes were involved: deposition during fairweather periods and high-energy physical sedimentation from flow and from suspension during episodic storms. The ramp appears to have been partitioned into three major facies associations deposited in environmentally-related provinces including evaporitic platform, inner ramp and outer ramp facies.

* The Tamet cyclicity is the product of small-scale fluctuations in relative sea-level and on the basis of the symmetrical arrangement of microfacies in vertical section, a gradual drowning and shallowing are indicated as accommodation space was being created and then infilled. Each cycle reflects the interplay of sedimentation and eustatic sea-level changes. Some of these changes may reflect local conditions of storm energy, sea-floor

topography and subtle tectonic effects. Thick subtidal cycles probably formed under higher-amplitude fluctuations of at least 20-25m, which caused short-term drowning of the ramp and inhibited development of peritidal microfacies on the submerged ramp interior. By way of contrast, the section on the Cyrenaica Platform is dominated by cyclic subtidal dolomite capped by evaporites; these hypersaline cycles prograded out onto the Tamet ramp indicating that a limited amount of accommodation space developed upon the Cyrenaica Platform.

* Pervasive dolomitisation is closely associated with evaporites of the Cyrenaica Platform, but it decreases progressively towards the more normal marine portions of the Tamet ramp, suggesting that reflux of hypersaline fluids generated within the Cyrenaica Platform was the main cause of dolomitisation. Two dolomite-rock textures are recognised and classified according to the crystal-size distribution. The explanation might lie with factors such as flow direction and volumes of dolomitising fluids. The most common pore types include (1) intercrystalline and moldic porosity generated during hypersaline dolomitisation and (2) primary intergranular porosity in mud-free large foraminifera microfacies.

* The study of the Tamet strata suggests that superimposed different orders of eustasy controlled the development of large-scale depositional sequences and the component meter-scale cycles that comprise them. platform-to-basin transitions are gradational and small-scale cyclicity alternated between catch-up and keep-up modes. Systematic changes in the patterns of these cycles define three third-order depositional sequences which reflect third-order sea-level fluctuations superimposed on long-term tectonic stabilities.

* The middle Eocene across this area is not a single carbonate ramp but rather an amalgamation of ramps. Based on facies associations and cycles within the Tamet Formation three depositional sequences are recognised, separated by stratigraphic transitional zones. Each sequence represents a prograded ramp. The sequence framework was developed from the metre-scale cycle architecture and microfacies interpretations that based on fabrics. Most of the sequences are interpreted as transgressive-highstand deposits. Each transgressive ramp is typically characterised by an aggradational pattern of relatively deep subtidal mud-rich carbonates deposited in a catch-up depositional system and episodically affected by storm events. Away from the ramp-margin, the transgressive facies change and stratigraphic thinning into lagoonal facies deposited under keep-up conditions. The subsequent highstand ramp begins with an aggradational geometry pattern but finally shift into a distinct progradational pattern. The highstand cycles covers abroader areas than that occupied by the transgressive deposition and made up of a mud-poor framework reflect a keep-up depositional system. The Cyrenaica Platform at this time was occupied by a very shallow and hypersaline sea. Carbonate sedimentation was shut off and replaced by deposition of shallow-water evaporites which become the main cause of dolomitisation and marking the end of sequence.

* A few limitations emanate when translating sequence stratigraphic concepts originally defined for siliciclastic systems to carbonate deposits. There is no indication of discernible large-amplitude third-order eustatic movements in the Tamet sequences compared to those produced by Cretaceous eustatic fluctuations. This may be due to the depositional setting of the Tamet platform which was not sensitive enough to record sea-level signals that may have been too low. Under this hypothesis, the sequence boundaries and transition between systems tracts should be recognised as *zones* rather than distinct surfaces. Therefore, the Haq *et al.* (1987) sea-level curve may require some modifications.

The geometric expression of the middle Eocene sequences is different from the idealised sequence stratigraphic models of Vail (1987) and Posamentier *et al.* (1988). These differences arise from the very limited amount of accommodation space created on the eastern flank of the Sirte Basin due to slow subsidence (Gumati, 1985) and slow Palaeogene second-order eustatic oscillations (Haq *et al.*, 1987). Low accommodation potential resulted in limited aggradation of the ramp interior strata and extensive accretion of the ramp-margin beyond the former margin.

* There is potentially a large undrilled petroleum resource remaining in stratigraphic traps (probably concentrated at relatively shallow depths) in the Palaeogene sequences. Shallowing-upward sequences of carbonates and evaporites are extremely common as hydrocarbon traps. By using high-resolution sequence stratigraphic and diagenetic approaches, we hope to provide direction and to reduce risk for future hydrocarbon exploration.

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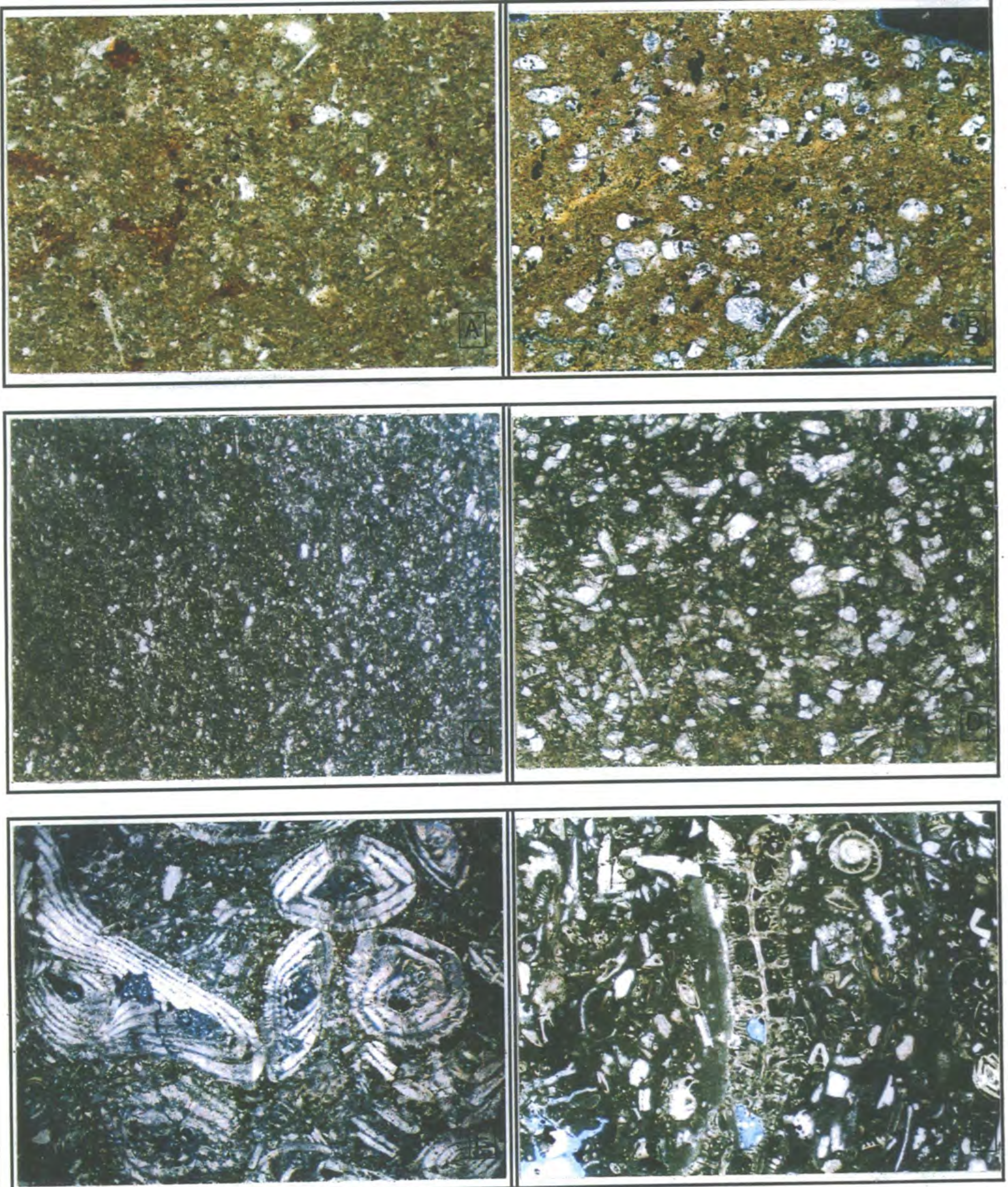
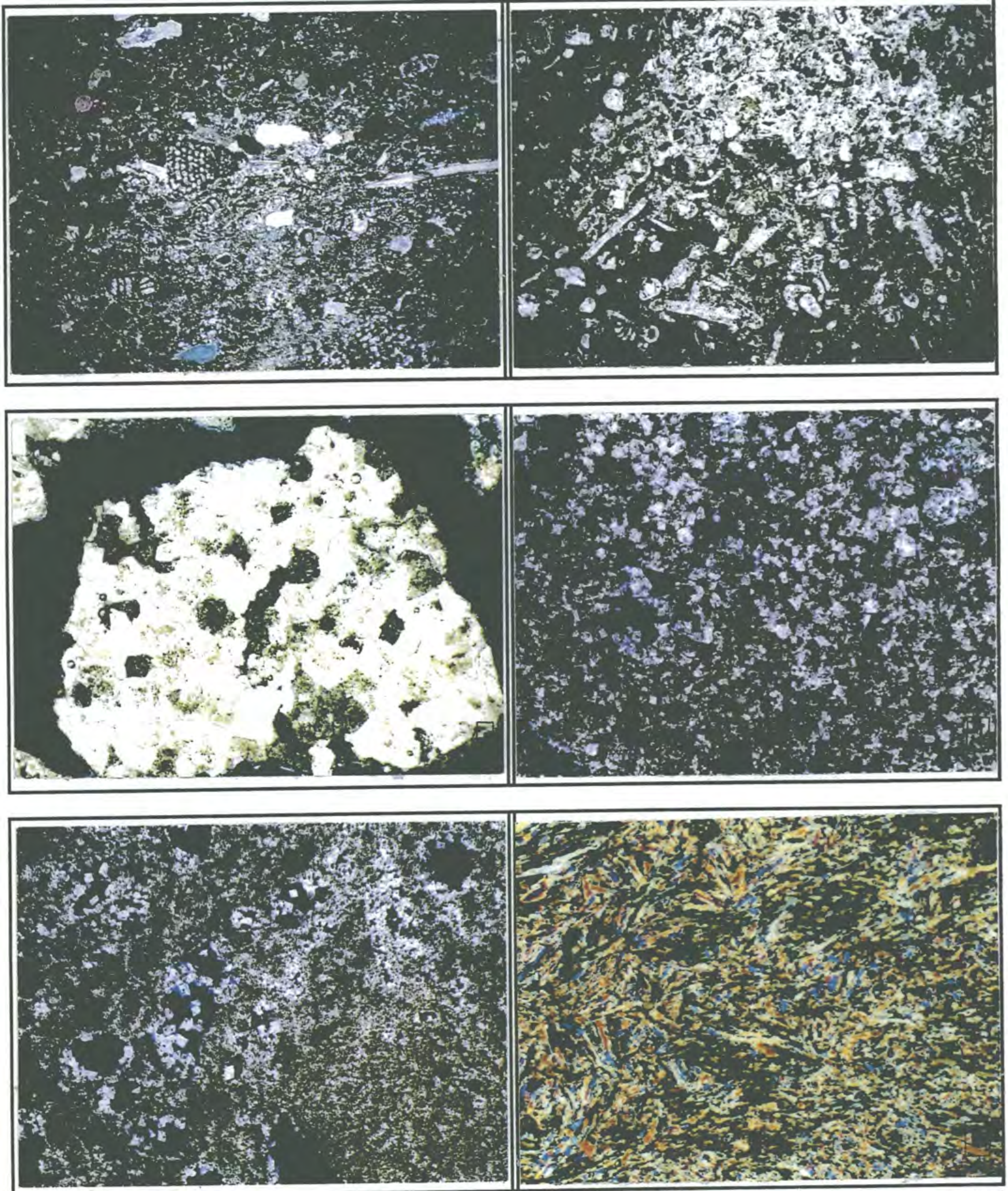
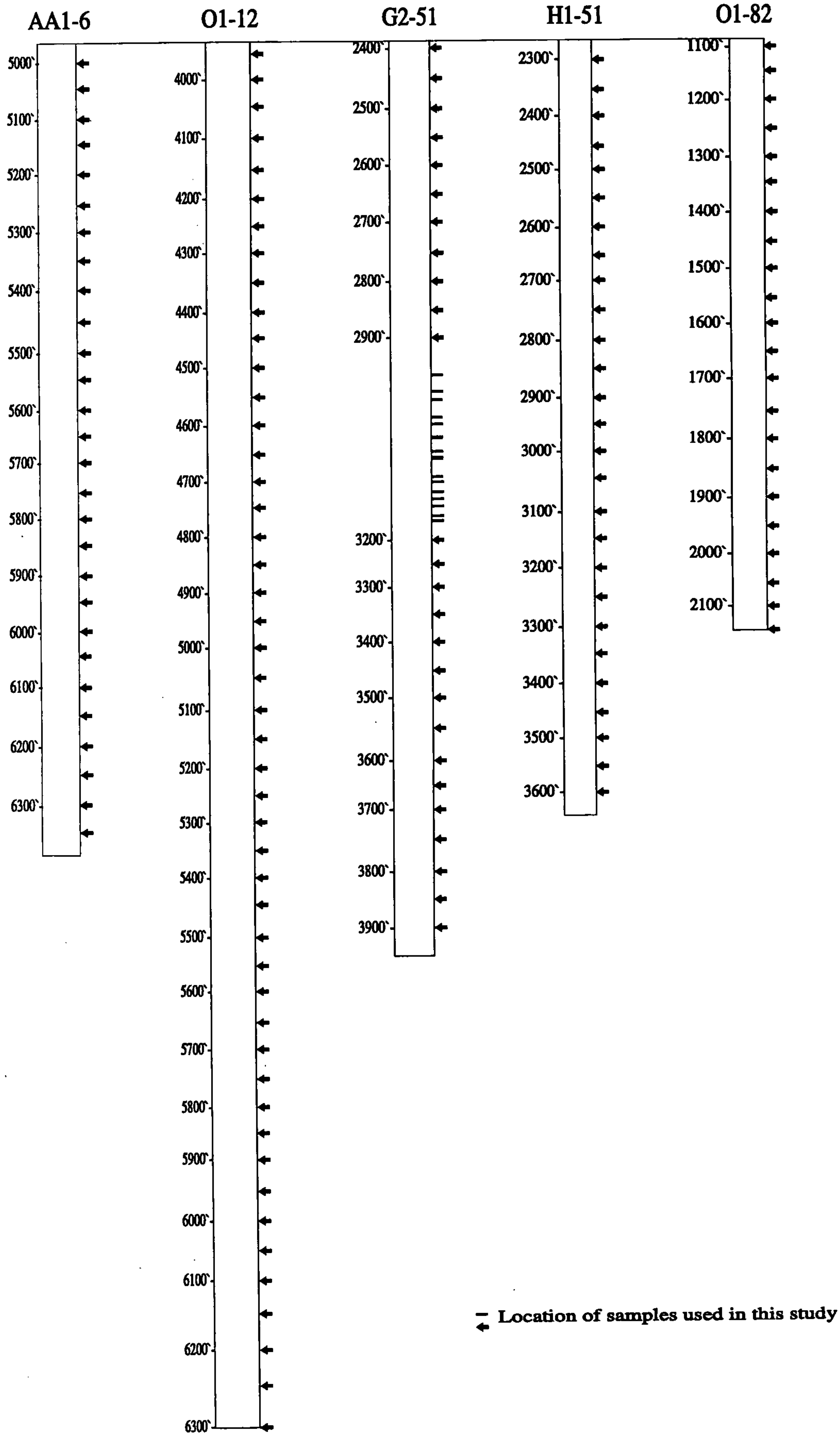


Appendix 1: *Typical microfacies photomicrographs of the Tamet Formation on eastern Sirte Basin. All photomicrographs were taken under plane polarised light and 40X magnification. A) Chalky mudstone microfacies, which consists sparse planktonic foraminifers and little organic matter. Sample from well O1-12, depth 1887m (6190ft). B) Planktonic-foram mud-wackestone microfacies, which composed predominately globigerinid foraminifers and undifferentiated spicules. Sample from well AA1-6, depth 1867m (6125ft). C. Microlitho-microbioclastic wacke-packstone microfacies, is characterised by relatively packed fabric silt-fine sand-sized bioclasts. Sample from well O1-12, depth 2082m (6830ft). D) Bioclastic mud-wackestone microfacies, is composed mixed foraminifer, echinoderm and bivalve debris. Sample from well G2-51, depth 1189m (3900ft). E) Bioclastic-benthic foram wacke-packstone microfacies, is composed of mixed large benthic foraminifers including nummulitids, operculines, discocycines and rotaliids. Sample from well G2-51, depth 936m (3070ft). F) Peloidal-bioclastic wacke-packstone microfacies, showing a diverse of open-marine macrofauna such as benthic foraminifers, bryozoans, bivalves and echinoderm. Sample from well O1-12, depth 1750m (5740ft).*



Appendix 1: (continued). *G) Orbitolinid-echinoderm wacke-packstone microfacies, showing admixture of open-marine skeletal debris with common fine sand-sized peloidal muds. Sample from well G2-51, depth 964m (3160ft). H) Peloidal-miliolid wacke-packstone microfacies, consists relative diverse benthic foraminifers include miliolids, alveolnids and textularids, which are dispersed in peloid-rich matrix. Sample from well G2-51, depth 960m (3150ft). I) Coarse crystalline dolomite submicrofacies, displaying a tightly packed mosaic of mostly anhedral to subhedral dolomite with irregular crystal boundaries. Sample from well G2-51, depth 730m (2394ft). J) Fine crystalline dolomite submicrofacies, consists of an equigranular, anhedral dolomite crystals. Some crystals contain opaque nuclei. Sample from well O1-12, depth 1417m (4650ft). K) Pel-foram dolowacke-dolopackstone microfacies, has moldic porosity. Locally these pores are solution-enlarged and resulted in a more vuggy appearance. Most larger pores are partially or totally occluded by dolomite cements. Sample from well O1-82, depth 395m (1295ft). K) Anhydrite microfacies, forming pseudomorphs to a lath-shaped crystals. Sample from well O1-82, depth 1578m (5175ft).*

Appendix 2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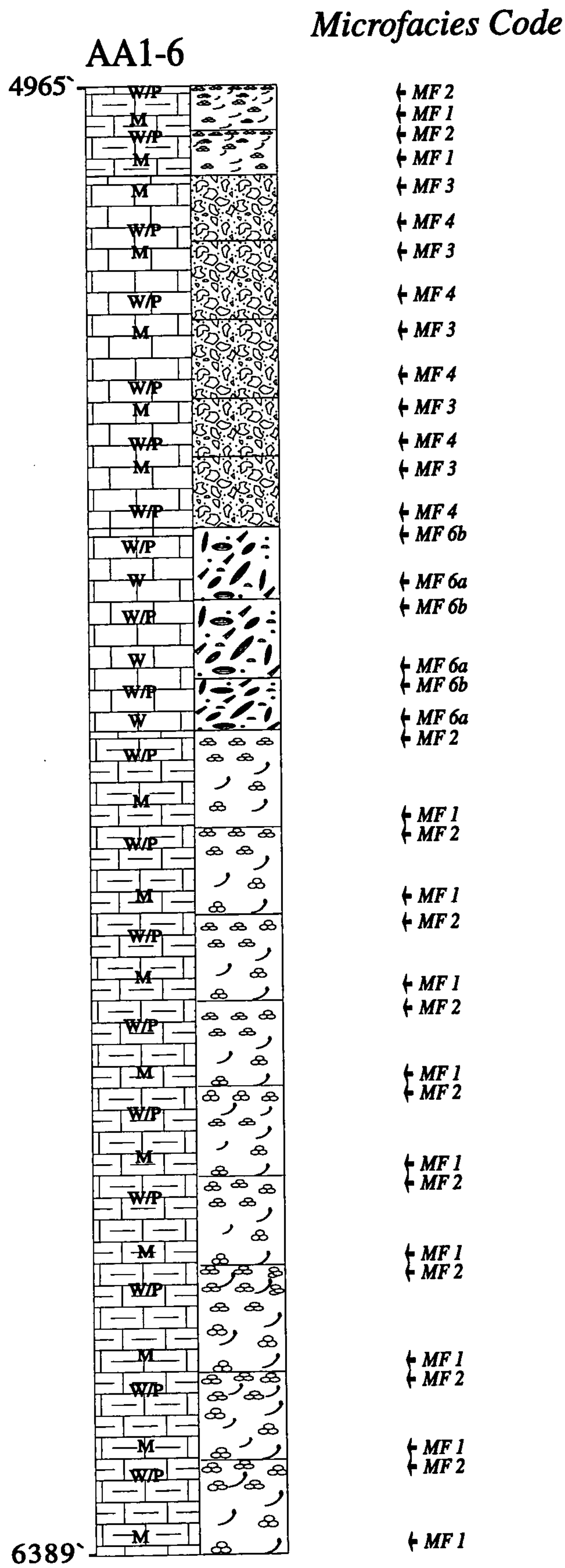
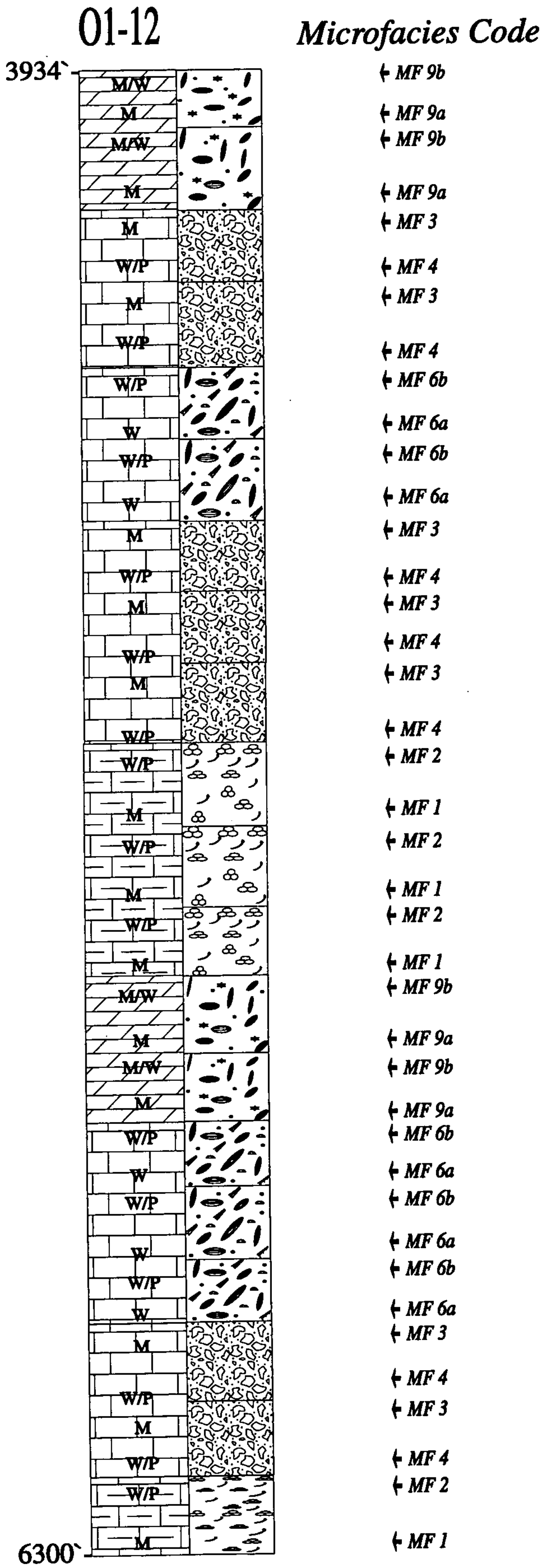
Location of samples used in this study

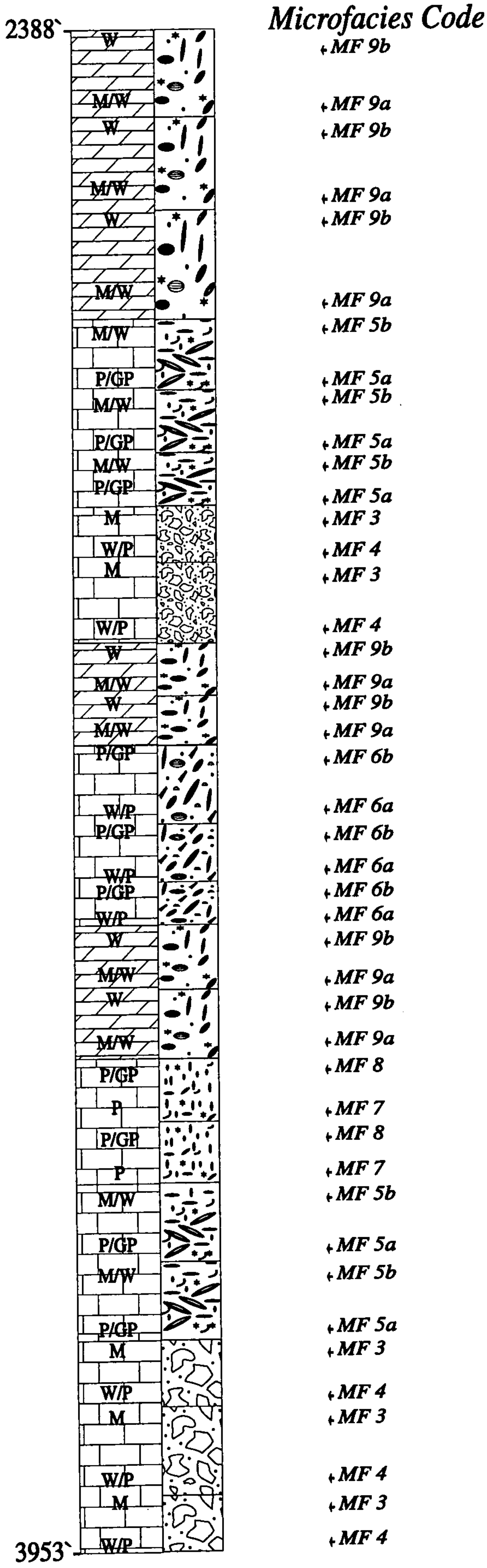


Appendix 3: *Columnar sections of AA1-6, O1-12, G2-51, H1-51 and O1-82 wells; showing the main microfacies sequences through the middle Eocene of the eastern Sirte Basin. These microfacies were probably deposited during relatively uniform conditions (sedimentation rates and subsidence rates) and should give a uniform response to any cyclic signal.*

Appendix 3: (continued)



G2-51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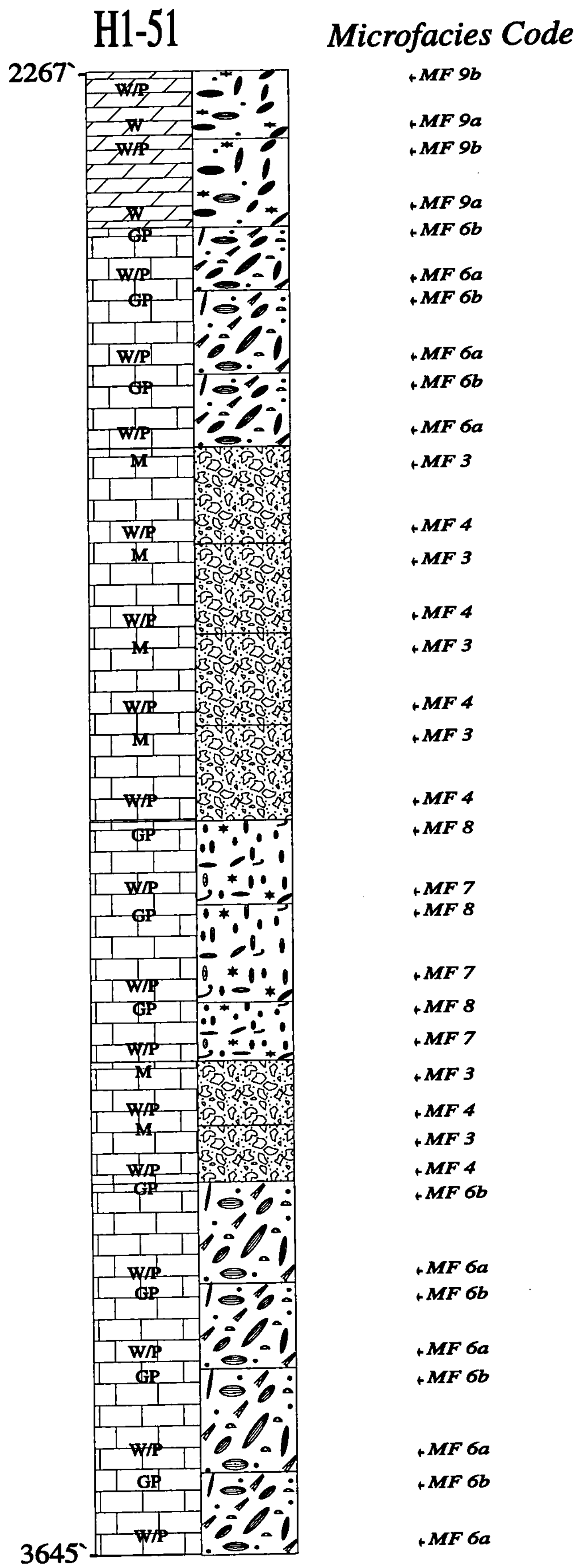
Appendix 3: (continued)



Appendix 3: (continued)