"SHALLOW MARINE, SANDY CARBONATE CLASTIC
SEDIMENTATION IN A FOREARC BASIN (KALAMAVKA
FORMATION, LATE SERRAVALLIAN-EARLY TORTONIAN,
E. CRETE)".

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ABSTRACT

A literature study of the Neogene of the Aegean region
revealed that the Neogene tectono-sedimentary evolution of
Crete has been largely controlled by the northward subduction
of the African plate beneath the Aegean lithosphere. The
Aegean region has largely responded to this relative motion of
the two plates by extensional tectonics and subsequent
subsidence. Geological data showed that a compressional regime
dominated Crete at least from Late Serravallian-Early
Tortonian.

The Late Serravallian is characterized by the onset of the
break-up stage of the Aegean Landmass. During the same
timespan, a compressional phase prevailed in the Ierapetra
region (E. Crete) which was punctuated by the deposition of
immature carbonate clastics (Prina Complex) and the rhythmical
alternations of calcareous mudstone and sandstone (Kalamavka
Formation).

A field study of the Kalamavka Formation was undertaken to
investigate this rhythmically deposited sedimentary succession
and to interpret the depositional system and the depositional
sequence. The results can be summarized as follows:

The Kalamavka Formation constitutes a (crude) coarsening-
upward sequence which is composed of mainly bioturbated, mud-
dominated shelf deposits which pass transitionally upward into

lower shoreface deposits. In the lower and middle part of the
sequence, debris flow lobes are found.

The Kalamavka deposits are characterized by the presence of
low-angle cross laminations (hummocky-cross-stratification)
and wave ripples found mostly in the proximal (most landward)
part of the sequence and by bioturbated, even laminated thin
sandstone beds found in the distal (more basinward) part.
These beds are thought to have been deposited from storm
induced "ebb" surges. After their deposition and when the
storm assumes normal level, waves produce ripples on the
already deposited sandy surface.

The sedimentation of the Kalamavka Formation seems to have
been influenced by contemporaneous tectonism with differential
subsidence of the basin floor and superimposed eustatic sea
level changes. The observed stacking of minor prograding and
retrograding "cycles" can be attributed to gradual basin
subsidence due to isostatic loading and to eustatic sea level
fluctuations.

Faunal data and ichnofacies also suggest that the Kalamavka
Formation was deposited in a shallow marine environment
(shelf). The combination of storm-induced currents, gradual
basin subsidence and sea level fluctuations can explain the
facies associations and their evolution in time.

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CHAPTER 1

INTRODUCTION

1.1 INTRODUCTION

The island of Crete is situated in the South Aegean region (Greece) (Fig. 1). It is bordered to the west by the island of Kythira and to the east by the islands of Kassos, Karpathos and Rhodes. These five islands constitute an island arc, the Hellenic Arc, which is characterized by an elongated horst-like structure (Angeller, 1976) and connects the structural trends of the Peloponnisos with those of the Taurides in Southern Turkey. Narrow straits are separating Crete from the adjacent islands of Kythira and Kassos (e.g. Peters, 1985).

To the north, Crete is separated from the islands of Cyclades by the Cretan Sea, a roughly E-W oriented bathymetric low, which becomes shallower eastwards (from 2500 m deep to 1300 m deep). To the south, the south Cretan margin is delimited by the Hellenic Trench system.

Crete represents the southernmost exposed part of the Hellenides, which constitute a series of narrow facies belts (called "isopic zones"), formed during the Late Mesozoic-Early Tertiary period. Each one of these isopic zones is

FIG. 1: General map of the Aegean region and adjacent areas of
the eastern Mediterranean, outlining the geography
and the main physiographic domains.

characterized by a specific depositional history and can be traced over large distances throughout the Aegean region.

Furthermore, on Crete, these isopic zones constitute a thick pile of Alpine nappes which are exposed in uplifted blocks and are separated by normal faults from adjacent Neogene and Quaternary sedimentary basins (Drooger and Meulenkamp, 1973; Fortuin, 1977; Meulenkamp, 1979; Peters, 1985).

Thus, the Cretan geology can be divided into two themes:
1. The "pre-Neogene rocks" or the "Alpine basement" (Peters, 1985) which is composed of the southern outliers of the Hellenides and
2. The Neogene and Quaternary deposits which are separated from the "pre-Neogene rocks" by an angular unconformity (Fortuin, 1977; Fortuin and Peters, 1984; Peters, 1985).

In eastern Crete, the Neogene sedimentary evolution of the Ierapetra basin has been studied in detail (Fortuin, 1977, 1978). The stratigraphy of the Ierapetra region is rather complicated by tectonic movements during the period of sedimentation, especially during the Late Serravallian when the break-up stage of Crete from the European landmass started. During that period, Neogene basins were formed due to strong impact of crustal movements which acted on that region. Fortuin (1977,1978), in his contribution to the interpretation of the structural and depositional settings of the Neogene

sedimentary basin fills, studied thoroughly the Ierapetra
region. On the basis of detailed mapping and faunal datings,
he subdivided the basin fill of the Neogene Ierapetra basin
into nine formal rock units: eight formations and one complex.

1.2 SCOPE OF THE PRESENT STUDY

In this thesis, I have restricted myself to the detailed
study of one of the eight formations of Fortuin (1977): the
Kalamavka Formation (Late Serravallian- Early Tortonian). This
formational unit has been selected for the reason that its
depositional timing (Late Serravallian- Early Tortonian)
coincides with the initiation of the break-up stage of the
Aegean crust due to intense extensional tectonics (Angelier,
1979), and therefore, with the onset of the "modern history"
of the Aegean region (Meulenkamp, 1985). Consequently, we
expect that many lithostratigraphical and paleogeographical
variations must have taken place during that timespan. The
Kalamavka Formation is thought to be a representative rock
unit, which depicts the prevailing sedimentary pattern during
that period.

Thus, the purpose of this work is to interpret the
depositional environment of the sediments belonging to the
Kalamavka Formation and its changes through time. The
interpretation is mainly based on the prevailing sedimentation
pattern during deposition.

1.3 METHODOLOGY

In order to accomplish the aim of the present study, the following steps have been taken:

1. Sedimentological field description of the sediments belonging to the Kalamavka Formation. The fieldwork was carried out during the summer of 1988 and the spring of 1989. During that time, detailed sedimentary logs of an appropriate scale were made and about 70 lithological samples were collected.

2. Petrographic description of approximately 30 thin sections, which cover the stratigraphy of the Kalamavka Formation.

3. Finally, in order to interpret the depositional environment and the processes which took place during the sedimentation of the Kalamavka Formation and to illustrate the geological setting of it, a relevant literature study has been made and utilized in the present work.

1.4 OUTLINE OF THE THESIS

The Kalamavka Formation, as already mentioned, has been affected by the tectonics of the broader area known as "Aegean region". During the Late Serravallian- Early Tortonian (depositional age of the Kalamavka Formation), very important tectonic changes took place resulting in the formation of a subduction zone south of Crete and the break-up of Crete from

the European landmass.

In *Chapter 2*, which is entitled "Geological setting of Crete", attention is paid to important tectonic events, which acted on the eastern Mediterranean, the Aegean region and finally, on Crete. In addition to this, some main points of the pre-Neogene and the Neogene history of Crete are discussed.

The Ierapetra region constitutes a basin, which is composed of many lithostratigraphic units, each one reflecting a specific tectono-sedimentary pattern, related to the prevailing tectonic regime of the Aegean region. The overall stratigraphy and structural evolution of the Ierapetra region are discussed in *Chapter 3*. In *Chapter 4*, which is the main chapter, a detailed description of the Kalamavka Formation is presented. This Chapter also comprises a detailed facies analysis of the sediments in space and time. Some results which are concerned with the petrography and biostratigraphy of the studied area are also given.

After the detailed description of the facies associations of the Kalamavka Formation, some comments are made in *Chapter 5* in discussion form and finally, in *Chapter 6*, I have reached some conclusions with respect to the depositional environment and the processes which have controlled it.

CHAPTER 2

GEOLOGICAL SETTING OF CRETE

2.1 NEOGENE TECTONIC EVOLUTION OF THE AEGEAN REGION

In order to understand the sedimentation and the tectonic
evolution of Crete, during the Late Cenozoic Era, it is
important to take into consideration its plate tectonic
setting. For that purpose, many geological investigations have
been done, not only in Crete but also in its surrounding
areas. Some of the results of these investigations are briefly
discussed below.

The Late Cenozoic evolution of Crete has been controlled by
the northward subduction of the African plate beneath the
Aegean lithosphere (Makris, 1977; McKenzie, 1978a; Papazachos
and Comninakis, 1978; Le Pichon and Angelier, 1979). As the
African plate moves northwards, strike slip displacements
along the Dead Sea fault zone causes compression between the
Arabian and the Eurasian plates (Molnar and Tapponier, 1975).
This process results in a crustal thickening of E. Turkey. As
a result, the Aegean and Anatolia plates are being pushed
westwards causing the extension of the Aegean region towards
the eastern Mediterranean (gravitational spreading, Fig. 2),
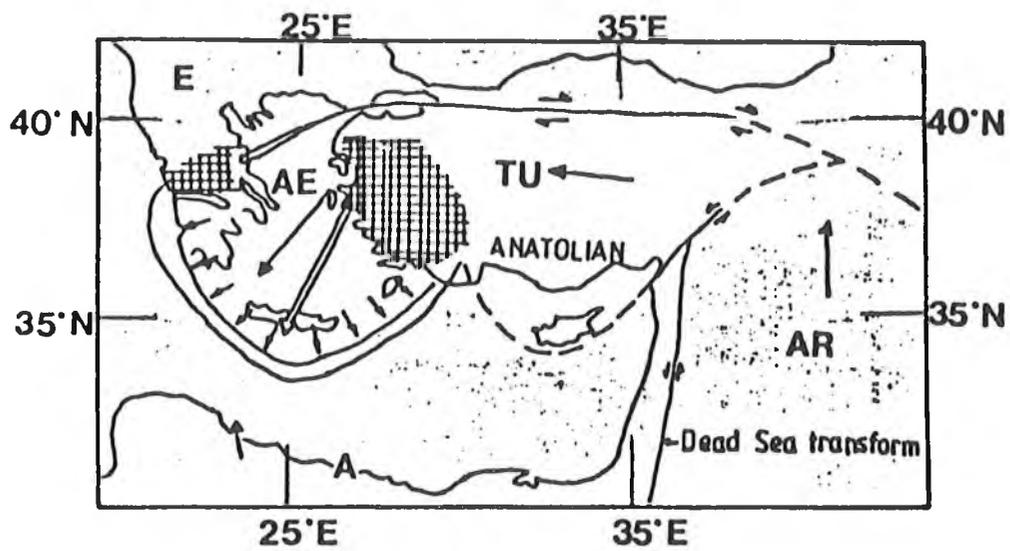
FIG.2: Schematic representation of the plate tectonic configuration in the eastern Mediterranean (after Mckenzie, 1972). The thin black arrows point at the strikes of the movements in relation to the Eurasian plate. The white arrow points at the movement of the Aegean plate towards the African one.

(McKenzie, 1972; McKenzie, 1978a; Le Pichon and Angelier, 1979; Angelier et al., 1981). The gravitational spreading of the Aegean region towards the eastern Mediterranean is evidenced by the presence of a dense pattern of normal faults (Upper Miocene) and horst and graben structures (McKenzie, 1972; McKenzie, 1978a; Le Pichon and Angelier, 1979; Angelier et al., 1981).

Due to the extensional faulting, the crustal thickness of the Aegean plate is reduced, reflecting crustal attenuation (Makris, 1978; Makris, 1985). For instance, the thickness of the Aegean crust in the Cretan Sea is not more than 20 km, which means that the crustal stretching must have been considerable there. According to Makris (1978, 1985), the crustal thickness of the Aegean region is not constant, expressing the variability of the amount of stretching of the Aegean crust. The same author believes that this fact is due to spatial variations in the mode of tectonics in the Aegean plate.

Indeed, apart from normal faulting, the Aegean region was also affected by rotational deformation, since the Middle Tertiary. Le Pichon and Angelier (1979) described a 28° rotation of the Aegean area with respect to Eurasia, around a pole situated in the southern Adriatic Sea. This rotation caused the progressive extension of an inner landmass and the continuous readjustment to this extension of the Hellenic Trench (see below) (Angelier et al.,1982). Since the Mio-Pliocene boundary (5 Ma), another clockwise rotation of 26°

occurred in the western and northwestern Greece (Jamet, 1982; Laj et al., 1982). Palaeomagnetic studies show that this rotation did not affect the southern and southeastern part of the Hellenic Arc (Crete and Rhodes), so that a structural discontinuity between the western and the eastern part of the Hellenic Arc must be assumed (Valente et al., 1982). Because of this discontinuity, the western segment of the Hellenic Arc was marked by a compressional phase since the Mio-Pliocene boundary, may be related to the continental collision of the Aegean domain with the Apulian continental margin (found west of the Ionian Sea), (Mercier et al., 1976; Sorel, 1976), whereas the southern segment remained under extensional conditions with the exception of short intervals of compression (Paquin et al., 1984).

Although the Aegean region is characterized by extensional tectonics resulting in a steady subsidence, the Hellenic Arc, and consequently the island of Crete, has an elevated position relative to the Cretan Sea in the north (**Fig. 3**). Angelier et Le Pichon (1980), Angelier (1981), Le Pichon and Angelier (1981), Angelier et al. (1982) attributed this uplift of the Hellenic Arc to a mechanism of crustal underplating, at least since the Middle Miocene. This would mean that sediments were removed from the subducting plate (African) to form the new basement of the Hellenic Arc (Barton et al., 1983).

The extensionally deformed Aegean region is delimited to the southwest and southeast by the **Hellenic Trench**, a complex
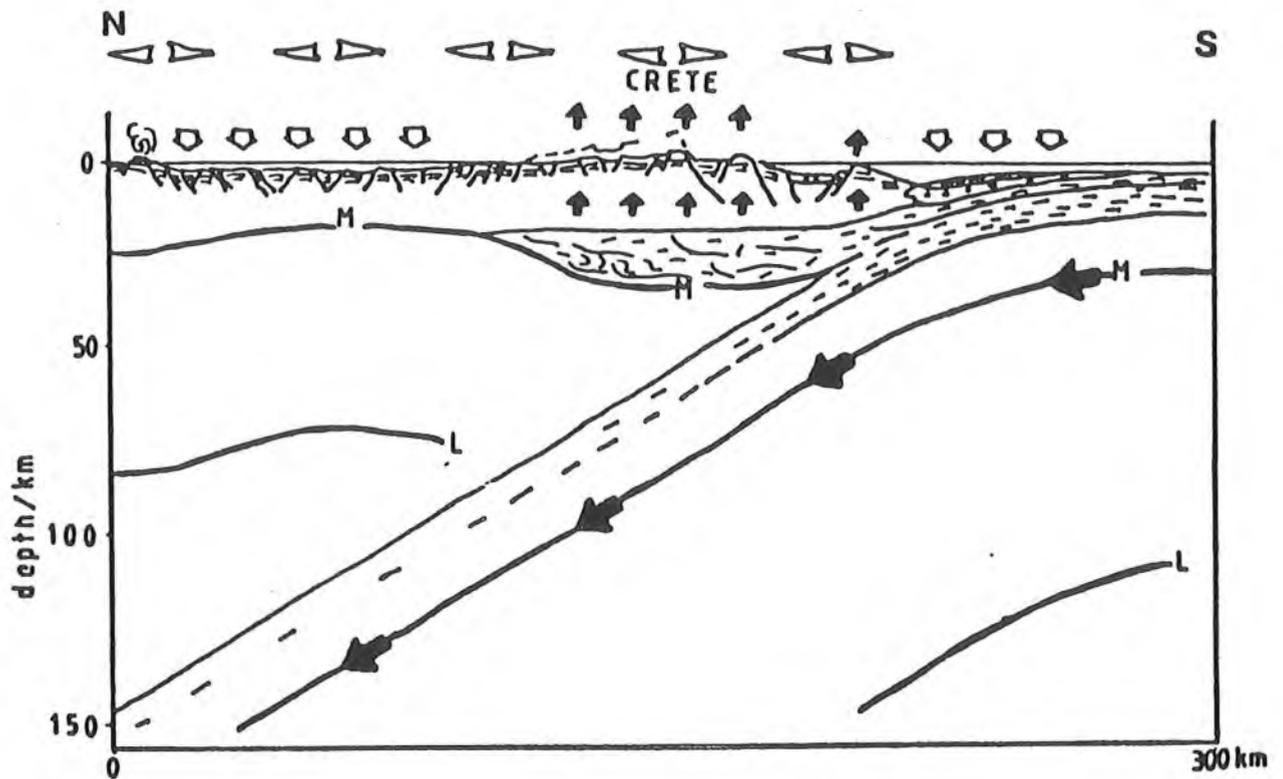
FIG.3: Generalized schematic profile of the South Aegean region
(after Le Pichon and Angelier, 1981). Crete can be
distinguished at the uplifted Hellenic Arc. The big black
arrows are pointing at the submergence of the African
plate, the small black arrows at the uplifting of the arc,
the white ones at the horizontal expansion of the Aegean
area.

system of topographic deeps which is found external to the
Hellenic Arc and marks the site of a subduction zone (Fig.
4),(Angelier, 1979; Le Pichon and Angelier, 1981; Huchon et
al., 1982).

The Hellenic Trench comprises two portions (Fig. 4):
-The Ionian Trench (west) which trends NW-SE and consists of
the Matapan and the Gotrys segments.
-The Levantine branch (east) which consists of a complex
system of subparallel, NE-SW trending narrow troughs: the
South Cretan Trough, the Pliny trench and the Strabo trench
(Jongsma, 1977).

The nature and the structure of the Hellenic Trench are
controlled by the kinematics of the subduction. Seismic
studies have shown that the African lithosphere is moving
towards the Northeast (McKenzie, 1978a; Le Pichon and
Angelier, 1979). Consequently, the Ionian branch is placed
perpendicular to the plate motion and is characterized by
compressive strain while the Levantine branch exhibits
transform motion (left-lateral strike-slip) (Fig. 4), (Le
Pichon et al., 1979).

The Hellenic Trench system constitutes the boundary between
the extended Aegean region to the north and the Mediterranean
Ridge to the south (Fig. 5). The latter is a 1300 km long,
broad, submarine swell which comprises a pile of heavily
deformed sediments, derived from the collision between the
very thick sedimentary cover of the subducting African plate
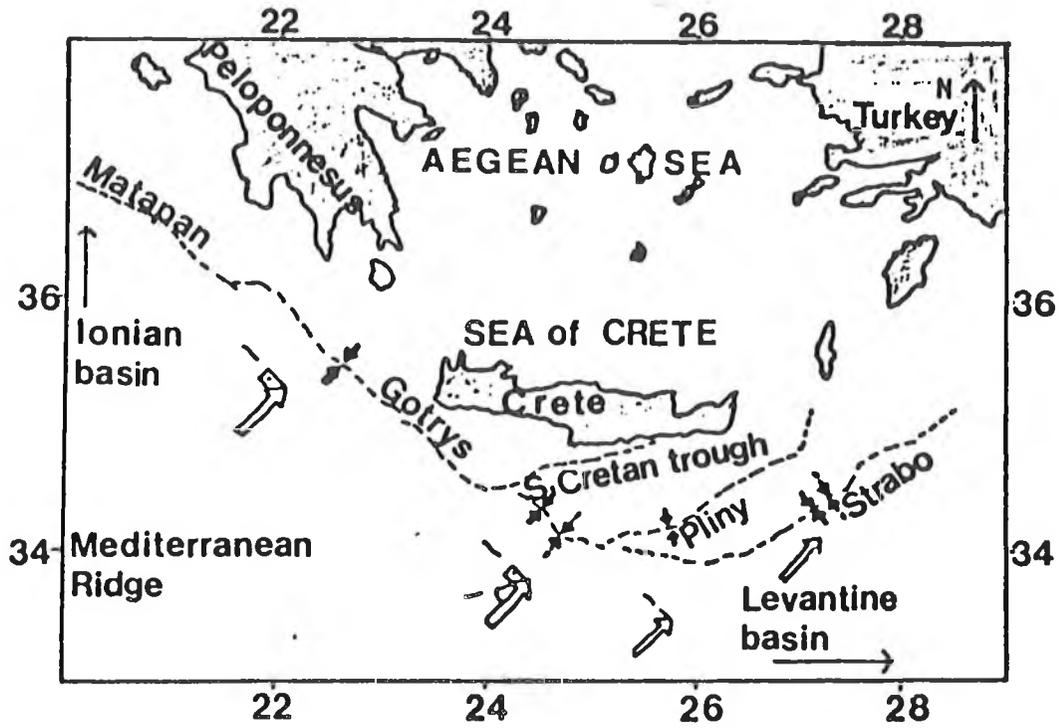
FIG. 4: The Hellenic Trench System (after Le Pichon and Angelier, 1979). Schematic structural and kinematic map showing the angular relationship between slip vectors.
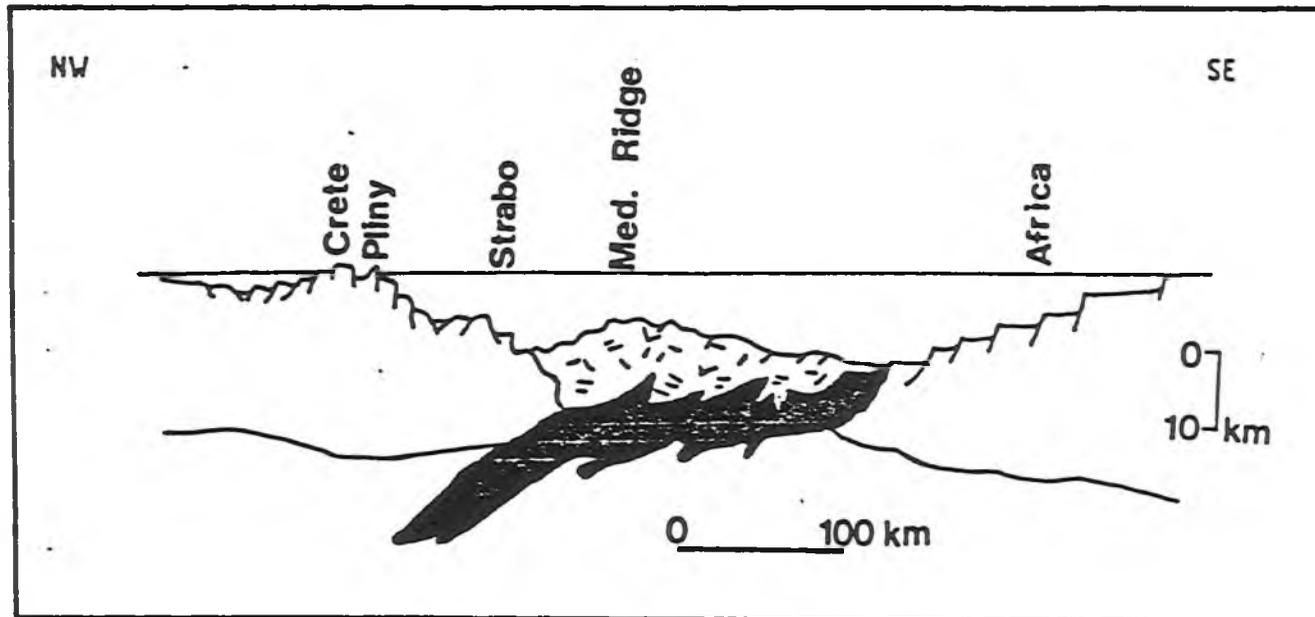
FIG.5: Profile across the south Aegean region, illustrating the collision between the thick sedimentary cover of the subducting African plate and the gravitationally spreading Aegean plate, and the formation of the Mediterranean Ridge System (after Mascle et al., 1986).

and the gravitationally spreading Aegean continental
lithosphere (Ryan et al., 1982; Le Pichon et al., 1982b;
Peters and Huson, 1985). It constitutes an accretionary prism
subject to compressional tectonics (Rabinowitch and Ryan,
1970).

The structural evolution of the ridge is directly related
to the dynamics of the Hellenic convergent zone. This has been
shown by data from DSDP-sites in the eastern Mediterranean,
which showed that the initiation of the compression in the
Mediterranean Ridge (Middle- Late Miocene) coincides with the
onset of the extensional phase in the Aegean region (Montadert
et al., 1978).

Mascle et al. (1986), based on the results of numerous
seismic surveys, presented and discussed the evolution of the
Hellenic subduction. The outline of this subduction history is
presented succinctly below (Fig. 6):

(A). During the Middle-Late Miocene (maybe Serravallian), the
subduction of an oceanic basin enclosed between Europe and
Africa, was taking place (Le Pichon and Angelier, 1979).
Seismic data have shown that during that period the relative
movement of the two plates was roughly N-S, which means that
the subduction was frontal along the Levantine branch and
oblique along the Ionian Trench.
(B). During the Late Miocene, the progressive closure of the
oceanic basin resulted in the slowdown of subduction along the

MIDDLE—UPPER MIOCENE
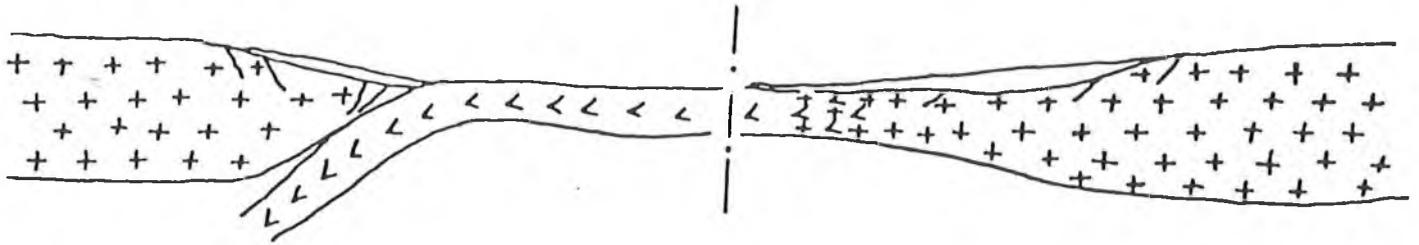
LATE MIOCENE
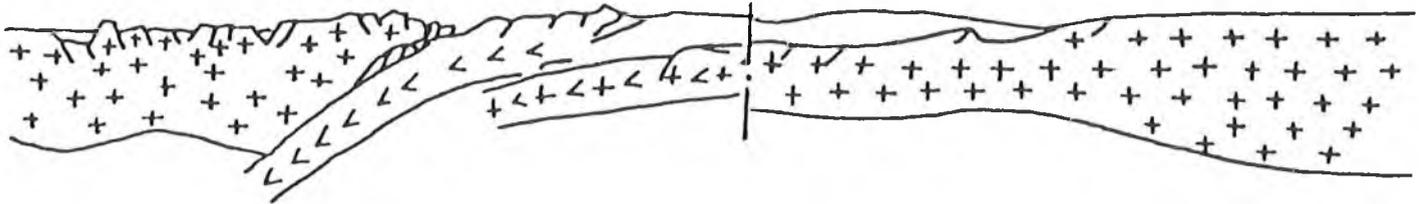
LATE MIOCENE—EARLY PLIOCENE
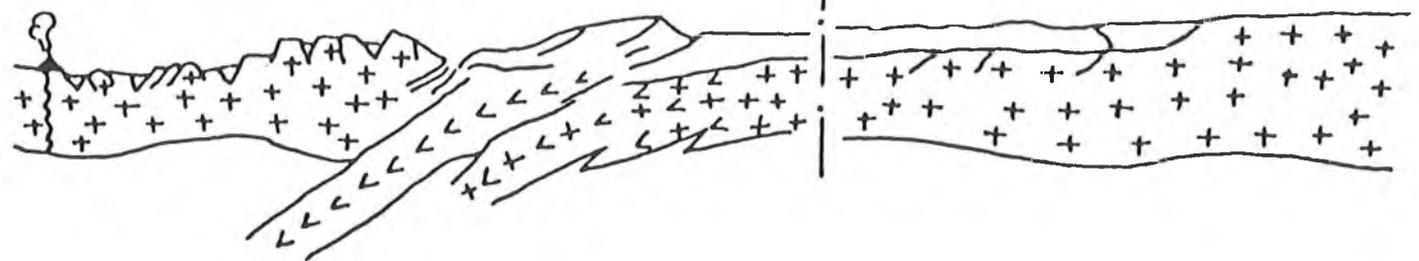
PLIO-QUATERNARY
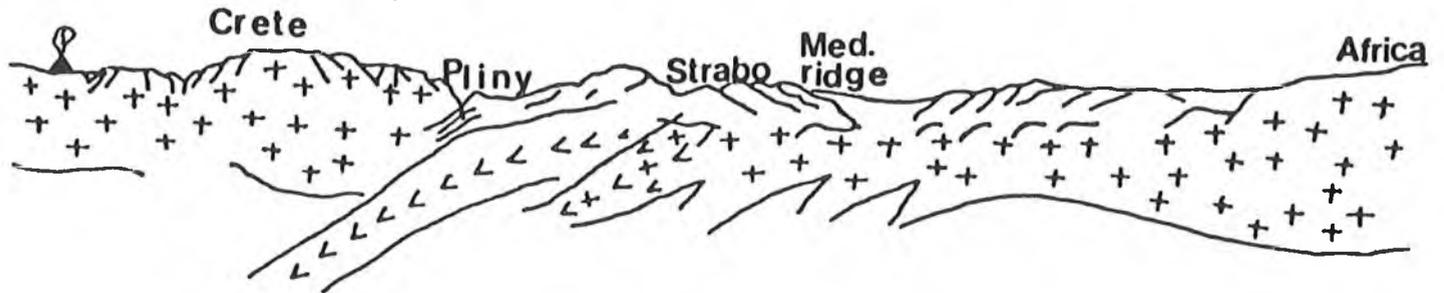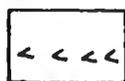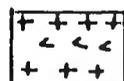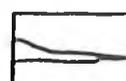
FIG. 6: Evolutionary model of the Hellenic subduction since the Middle-Upper Miocene (after Mascle et al., 1986).
1. continental crust, 2. oceanic crust,
3. intermediate crust, 4. sedimentary cover,
5. Aegean volcanism.

Levantine basin and the consequent migration of the subduction zone towards the SW. Thus, along the Ionian basin, the subduction continued resulting in the S-SW directed extension of the Aegean region.

(C). After the closing of the oceanic basin, in the **Late Miocene - Early Pliocene**, the African continental lithosphere continued to subduct only in the Ionian basin, giving rise to an extensional regime in the back-arc region, which led to general subsidence and formation of the present-day Aegean Sea (Le Pichon and Angelier, 1981). On the other hand, the rate of subduction in the Levantine basin was very low. But as the African plate continued to move towards the north, the Mediterranean Ridge with its compacted sediments started to overthrust the Levantine crust. The break-up of the Levantine crust gave rise to the Pliny and Strabo trenches.

(D). When in the last stage of the continental collision, the subduction came to a standstill (Le Pichon and Angelier, 1981), the Aegean region continued to extend in a predominantly NE-SW direction. Hereby, the subduction zone migrated southwards over the sinking lithosphere (Le Pichon and Angelier, 1981) promoting the subduction of the African Mesozoic lithosphere into the asthenosphere of the Aegean plate. This process is known as underplating and could have been responsible for both the elevation of the South Aegean region and the sinking of the Hellenic Trench in the south (Makris, 1976).

2.2 POSITION OF CRETE IN THE HELLENIC ARC

Crete, which is the southernmost part of the Aegean region, is an emerged part of the Hellenic Arc. Its elevated position seems to be contradicted by the extensional tectonics and the subsequent subsidence which began to act upon the Aegean region in the Late Miocene. However, geological data support the idea that compressional tectonics may have played a greater role in Crete than in the rest of the Aegean region.

During the Late Miocene (maybe Late Serravallian according to Le Pichon and Angelier, 1979), when the extensional regime of the Aegean region started, the island of Crete is thought to have transformed into a mosaic of culminations and depressions (Meulenkamp and Hilgen, 1986). Extensive analysis of the properties of the fault system of Crete (Angelier, 1979; Angelier et al., 1982; Meulenkamp et al., 1988) showed the coexistence of extensional and compressional features. Indeed, folds, reverse faults and oblique-slip transverse faults were found, deforming the pre-Neogene basement and the Neogene sequences, throughout the island. In other words, the Neogene of Crete is characterized by periods of universally directed extension alternating with periods of approximately NE-SW oriented compression (Meulenkamp et al. 1988).

This particular situation in the tectonic regime of Crete can be explained if we consider that the extensional regime in

the Aegean lithosphere gave rise to the generation of south-
dipping, low-angle shear zones (see Lister et al., 1984),
(Fig. 7). Along these zones, a supracrustal slab was formed
and started to slide in a southward direction, driven by
gravity (Simpson and Schmid, 1983; Lister et al., 1984). Above
the "detachment plane" of this supracrustal slab, the
extensional faulting resulted in the crustal thinning and
finally, in subsidence of the crust, leading to the generation
of the Sea of Crete (Lister et al., 1984). On the other hand,
the frontal part of the supracrustal slab may be affected by
compression as dissects Crete and meets the bulge of the
Ionian plate (eastern Mediterranean, a part of the African
plate) (Meulenkamp et al., 1988), especially during the
periods when tectonic transport of the supracrustal slab takes
place. So, NE-SW and NNE-SSW oriented compressional forces
obliged the pre-Neogene basement to be folded and thrusted. In
the meantime, E-W oriented, sinistral oblique-slip faults and
NNE-SSW oriented, dextral oblique-slip faults formed,
controlling the generation of the Neogene basins (Meulenkamp
et al., 1988).

To conclude, the southward migration of the supracrustal
slab is supposed to have been accompanied by compression in
its frontal part (Meulenkamp et al., 1988). The consequent
folding and thrusting, which acted upon Crete, are thought to
be the processes responsible for the uplift of Crete. In this
way, the compressional regime and the general uplift of Crete

S   C R E T E   Sea of Crete   Santorini   N

M
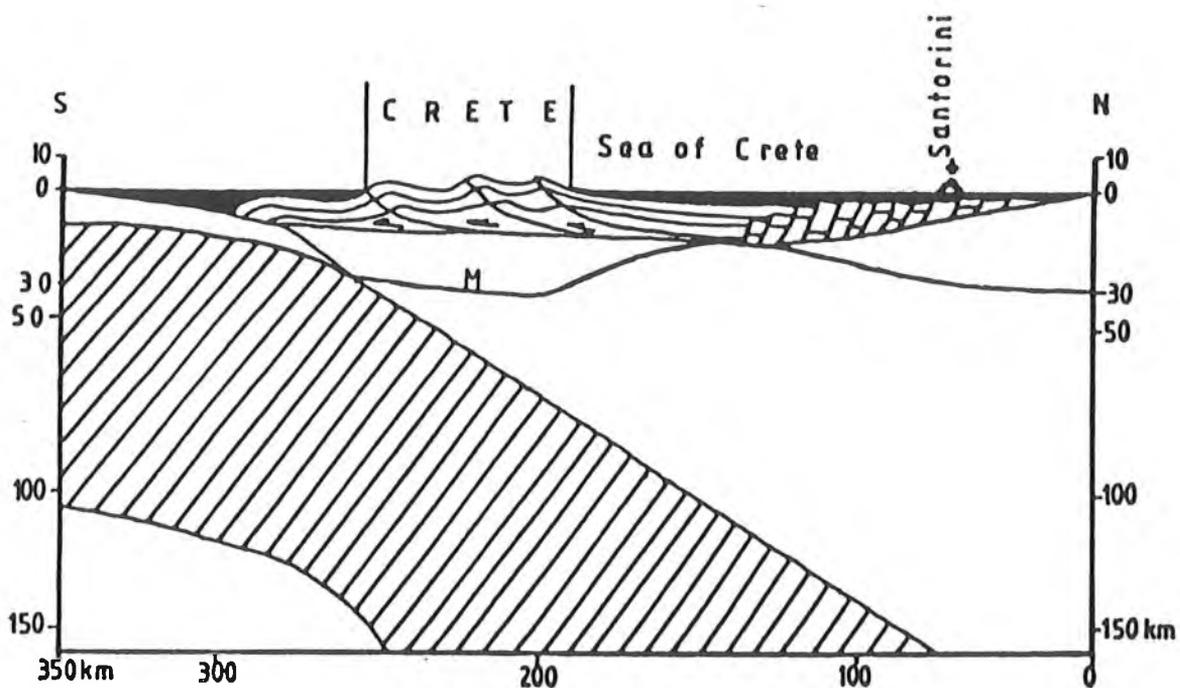
L E G E N D

Subducting lithospheric slab   sedimentary wedge

M : Moho

FIG. 7: Schematic N-S cross section of the South Aegean region illustrating a possible model of the supracrustal slab (after Lister et al., 1984; Meulenkamp et al., 1988). The position of the Moho in the Aegean lithosphere is indicated schematically. For further explanation see text.

is the outcome of the overall tensional stress regime of the
Aegean region.

2.3 TECTONO-SEDIMENTARY HISTORY OF CRETE

2.3.1 Pre-Neogene tectono-sedimentary history of Crete

The "pre-Neogene rocks" of Crete are composed of a stacked
series of highly heterogeneous tectonic nappes which are
exposed in uplifted blocks and are separated by normal faults
from adjacent Neogene and Quaternary basins (Fig. 8). In this
section, a synopsis of these tectonic nappes is given because
their structural position and lithological characteristics are
thought to be of importance for the present study. Some of the
main lithological characteristics of this nappe pile are
discussed below. For details on age and thickness of the
units, deformational structures and metamorphism the reader is
referred to Fig. 9.

The "Plattenkalk Series" (sensu Chaliklopoulos, 1903) or
"Cherty limestones" are the oldest rocks of Crete and
constitute the autochthonous "backbone" of the island (Kuss
and Thorbeck, 1974). They consist of Permian to Palaeogene
pelagic, partly metamorphic limestones, dolomites and fine
clastics (Creutzburg and Seidel, 1975). The whole unit has
undergone low-grade metamorphism and is deformed into folds
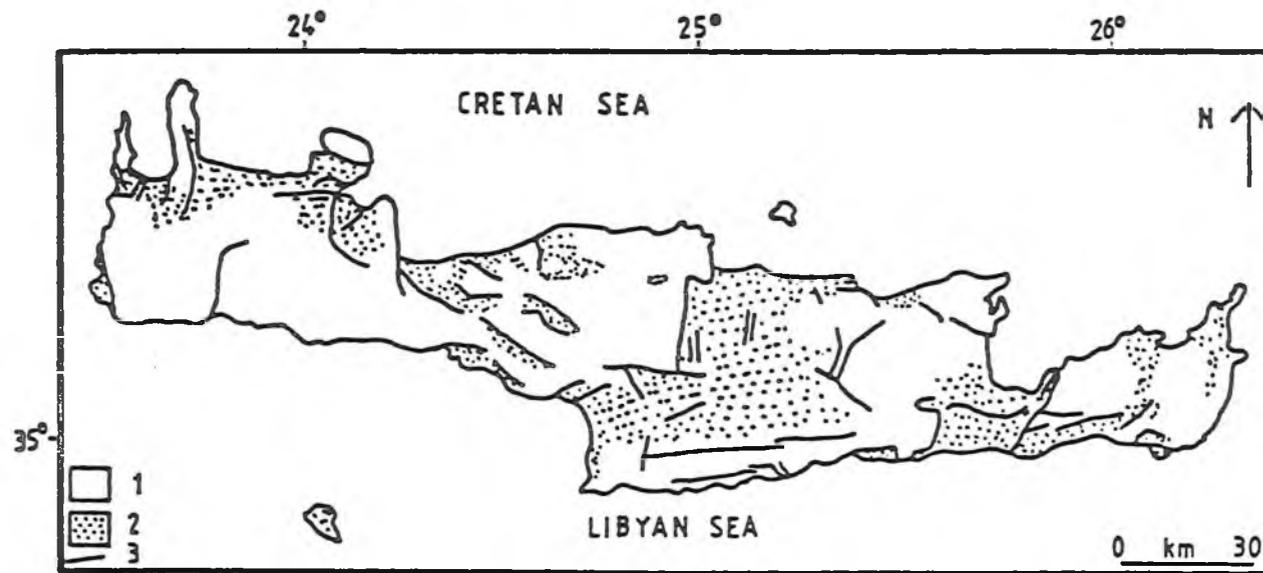
FIG.8: Geological sketch map of Crete. Distribution of the Alpine
basement rocks and the Late Cenozoic sedimentary basins
(after Peters, 1985).

1 . Alpine basement

2. Late Cenozoic sediments

3 Normal fault
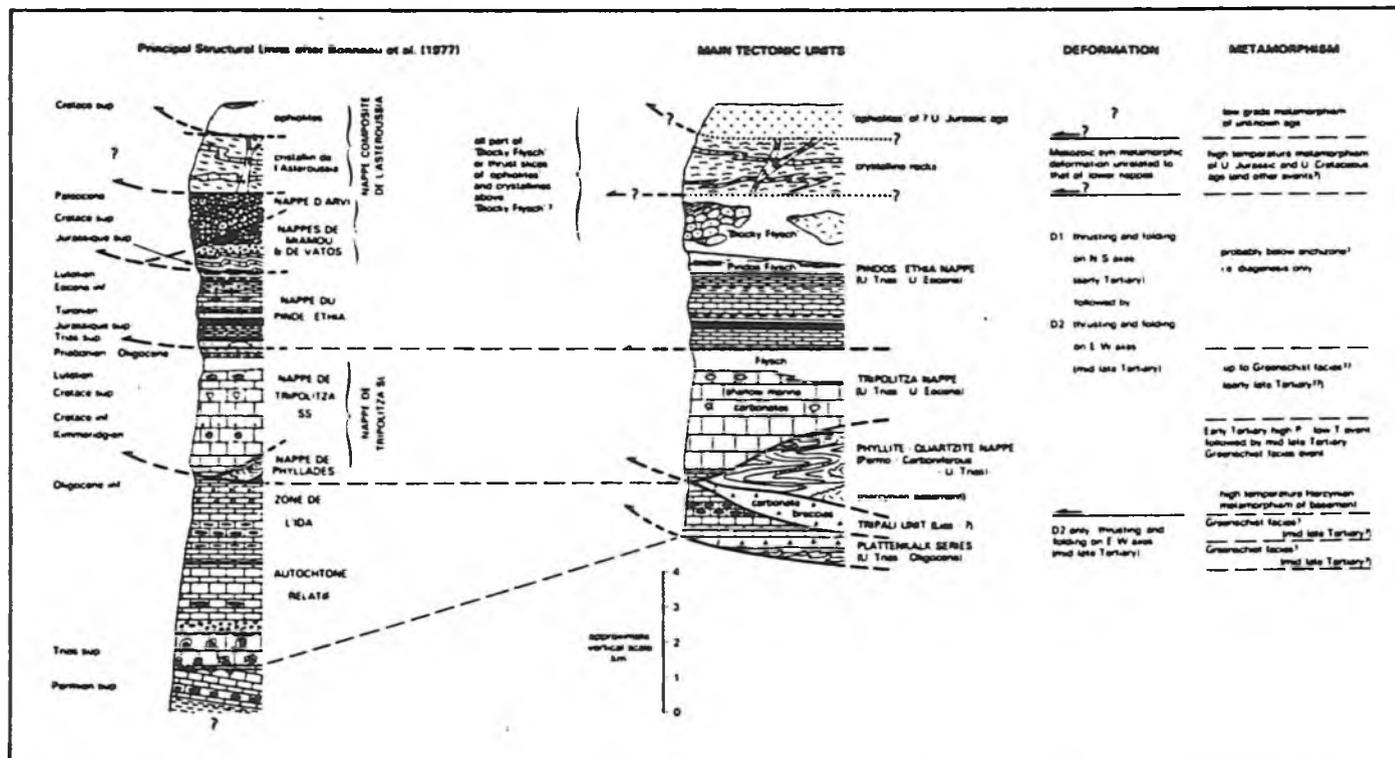
FIG.9: Summary of the main tectonic units referred by Hall et al.
(1984), showing a comparison of their tectonic scheme with
the units of Bonneau et al. (1977).

with wavelengths ranging between meters and kilometers
(Baumann et al., 1976). Over this unit, a series of other
allochthonous units are placed, bound by basal thrust planes
representing décollement levels (Bernoulli et al., 1974),
(Fig. 9 and Fig. 10). These allochthonous units are the
following:

1. Tripali unit: (Upper Triassic - Lower Jurassic; Kopp and
Richter (1983) suggested an Oligocene/Miocene age, though). It
is the lowermost allochthonous unit, consisting of various
types of carbonates, found only in western Crete (Creutzburg
and Seidel, 1975).

2. Phyllite-Quartzite unit: (Permian - Triassic). It is the
lowermost allochthonous unit of Eastern Crete (Wachendorf et
al., 1974; Seidel, 1978; Bonneau and Karakitsios, 1979;
Greiling, 1982; Krahl et al., 1983) which is composed of
phyllites, quartzites, meta-conglomerates, meta-sandstones,
lenticular crystallized limestones, meta-basites and meta-
andesites (Creutzburg and Seidel, 1975). The Phyllite-
Quartzite unit has been interpreted as a melange (sensu
Greenly, 1919) because it is thought to be a tectonic mixture
bound by shear surfaces.

3. Tripolitza Series: (Jurassic to Eocene). It is the most
dominant pre-Neogene rock type in southeastern Crete
(Wachendorf et al., 1975; Baumann et al., 1976; Zambetakis,
1977a). It consists of unbedded to thickly bedded, partly
dolomitized, reefal, dark-grey limestones (the so-called
Tripolitza limestones) overlain by Palaeogene flysch-like
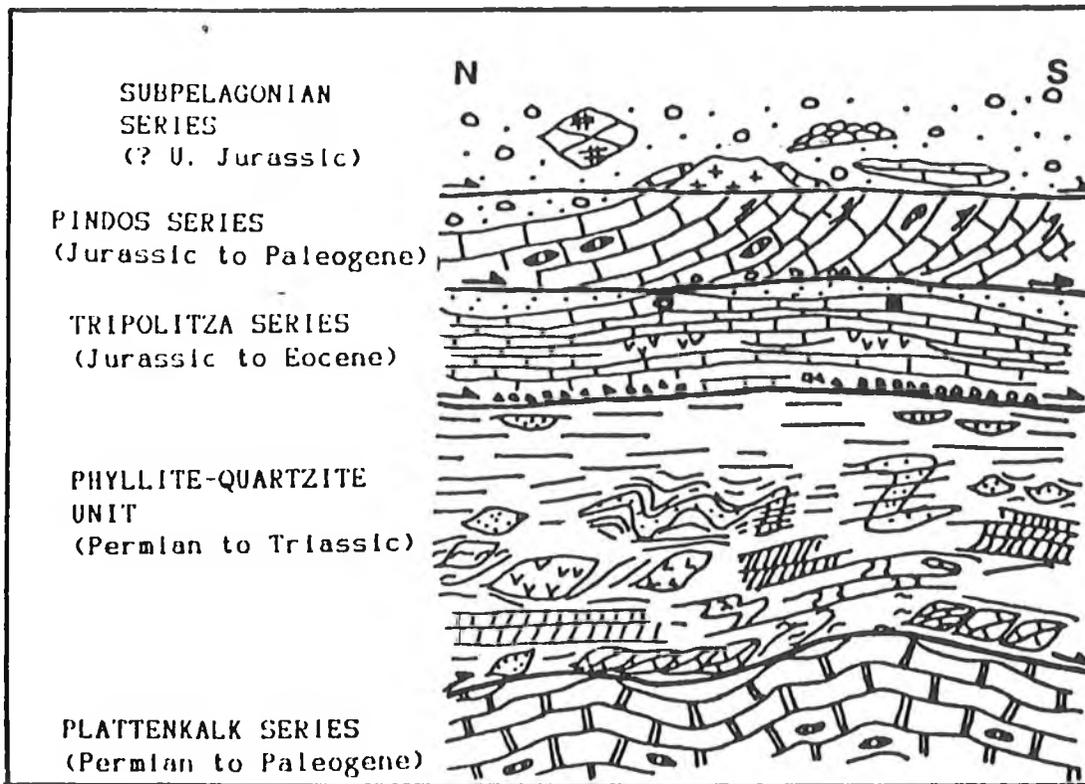
FIG. 10: Schematic section through the nappe pile of eastern Crete
(after Baumann et al., 1976).

deposits (Tripolitza flysch). Locally the base of the
Tripolitza Series is mylonitized.

4. Pindos Series: It consists of pelagic limestones, cherts
(radiolarites) and intercalations of redeposited shallow water
limestones of Jurassic to Palaeogene age (Zambetakis, 1977b).
The Pindos Series is found only fragmented in minor
extensional graben structures.

5. Sub-Pelagonian Series: It is the uppermost allochthonous
unit and consists of isolated bodies of amphibolites,
gneisses, mica schists, calcium silicate rocks and
serpentinite. These relicts are chaotically mixed and
distributed only in the southern part of central and eastern
Crete (Bonneau, 1972; Bonneau et al., 1977; Wachendorf et al.,
1980; Krahl et al., 1982).

    Apart from the previously described five nappes, small,
scattered ophiolite and granodiorite fragments can be found
underlying the Neogene in several places along fault zones,
pointing to a Late Cretaceous age (Baranyi et al., 1975;
Wachendorf et al., 1975). -

    The above mentioned tectonic units (of Mesozoic to
Paleogene age) were originally developed in distinctive
troughs and ridges in the Cycladic area (Aubouin, 1965). The
age of the emplacement of the nappes on Crete is still
uncertain. Angelier (1979) claims that it occurred before the
late Early Miocene. The finding of miogypsinids of Early

Burdigalian age in nonmarine, post-orogenic conglomerate successions is in favour with this assumption. In addition, before the Middle Miocene, the Cycladic area was affected by diapiric uplift (Durr et al., 1978; Lister et al., 1984) which is indicated by the elevation of the Moho discontinuity under the Cyclades and the tectonic denudation of the Cycladic crystalline massif. The uplift of the Cycladic area was accompanied by large-scale gravity sliding in the direction towards the Hellenic Arc (Baumann et al., 1976). The gravity sliding gave rise to the lateral transport of the allochthonous units of Crete. According to Wunderlich (1965), this nappe transport occurred in successive undulatory movements. Now, the whole succession of the allochthonous units rests on the Plattenkalk Series which responded to the overriding pile by deformation into folds (Baumann et al., 1977).

After the emplacement of these tectonic units, Neogene and Quaternary normal faulting broke the nappe pile into large 'rectangular' blocks and locally caused substantial vertical displacements of the originally (sub)horizontal tectonic contacts (Drooger and Meulenkamp, 1973). According to Fortuin (1977), this blockfaulting affected the sedimentation by causing numerous gravity slides at various stratigraphic levels and in different depositional environments.

2.3.2 Neogene tectono-sedimentary history of Crete

The "pre-Neogene rocks" or the "Alpine basement" of Crete
are unconformably overlain by the Neogene and Quaternary
deposits which fill up the graben structures formed during the
post- nappe period of normal faulting.

The Neogene record of Crete permits a fair reconstruction
of its palaeogeographic and tectono-sedimentary history
(Drooger and Meulenkamp, 1973; Fortuin, 1977; Angelier, 1979;
Meulenkamp et al., 1979a; Meulenkamp, 1982a,b; Meulenkamp,
1985). Consequently, the Late Cenozoic evolution of Crete has
been divided into six successive intervals from the Oligocene
to Recent. Each interval corresponds to a specific combination
of sedimentary pattern, tectonic movements and
palaeogeographic configuration. The main characteristics of
these six intervals are cited below (Fig. 11):

1. From the Oligocene until a poorly defined time level in the
Miocene.

After the emplacement of the Alpine pile nappes, Crete was
part of an uplifted landmass which was connected with the
European mainland and was extended to the north and to the
south of the present island (Meulenkamp, 1971). This is known
as the "Southern Aegean Landmass" (sensu Meulenkamp, 1971).
During that period, there was hardly any sedimentation or much
of it has been removed because of erosion. The erosion of the
pre-Neogene nappe pile resulted in the local accumulation of
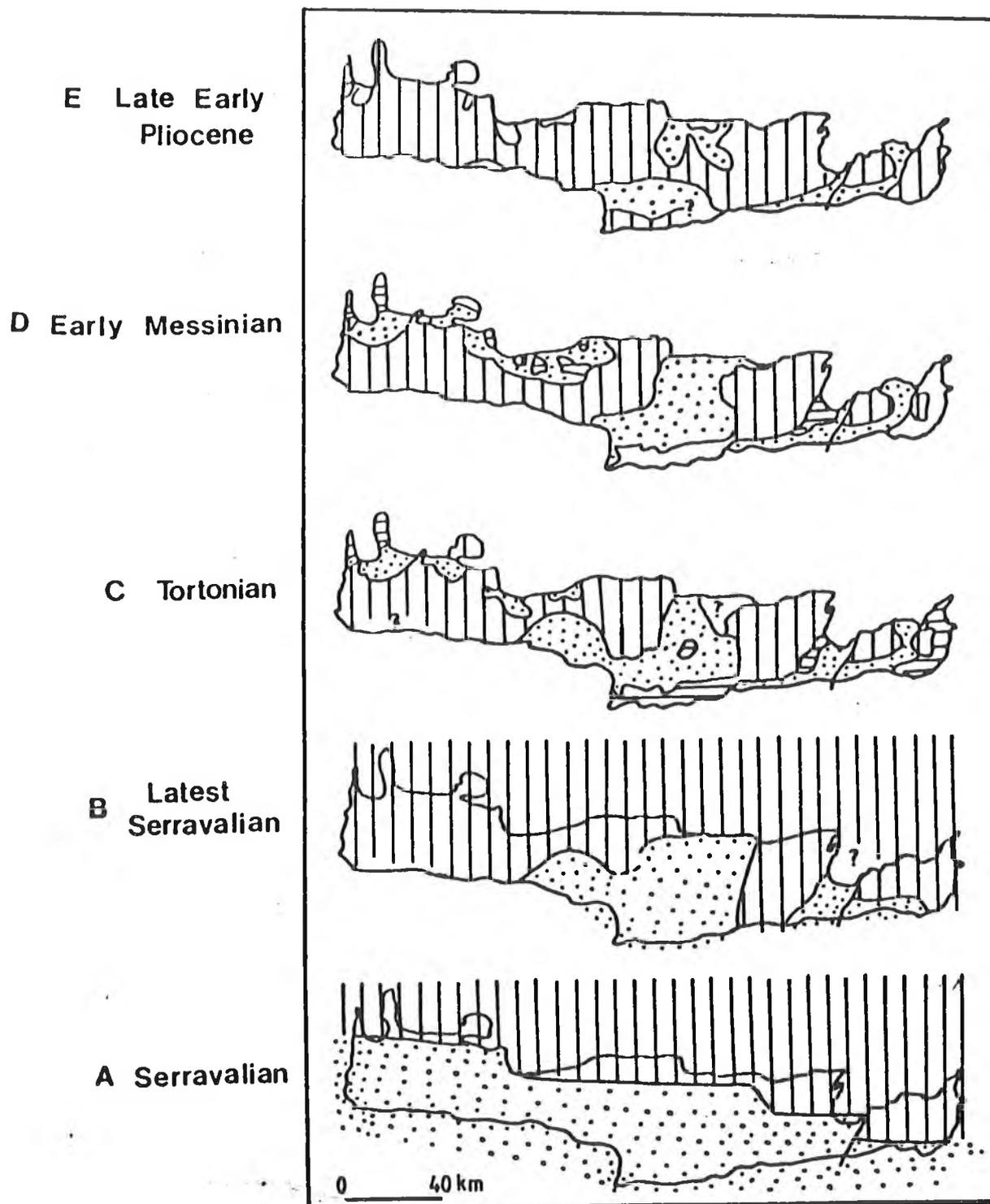
FIG.11: Middle Miocene to Early Pliocene basin configuration on
Crete based on data in Freudenthal, 1969; Meulenkamp,
1969; Gradstein, 1973; Fortuin, 1977; Meulenkamp
et al., 1979; Meulenkamp, 1985 and unpublished reports.
Vertical shading: emerged; horizontal shading: partly
submerged.

coarse, nonmarine clastics (Drooger and Meulenkamp, 1973). In addition, a N-S trending fault system found mainly in the north part of Crete and an E-W trending fault system found mainly in the south part of the area, began to affect the whole region according to Drooger and Meulenkamp (1973).

2. Middle Miocene

During the Serravallian (14-13 Ma ago) (Fig. 11A), Crete became incorporated in a regime of general subsidence, related to large scale tilting to the south (Meulenkamp and Hilgen, 1986). The differential vertical movements of the E-W, N-S trending fault systems resulted in the breaking up of Crete into numerous small blocks and in its "incorporation" into a large basin in which thick successions of fluviatile and lacustrine sediments, consisting of basal Neogene clastics with intervals of finer-grained terrigenous deposits, were laid down (Freudenthal, 1969; Meulenkamp, 1969; Gradstein, 1973; Fortuin, 1977; Meulenkamp, 1985; Peters, 1985).

At the transition from the Serravallian to the Tortonian (10,6 Ma ago), the palaeogeographic configuration of Crete changed completely. In the Late Serravallian (Fig. 11B), a more pronounced subsidence affected Crete (Drooger and Meulenkamp, 1973).The sea slowly invaded Crete and locally covered the older limnic deposits (Zachariasse, 1975; Boger and Willman, 1979). This transgression, which was more pronounced in Eastern Crete, must have been induced tectonically since it cannot be correlated with a global

eustatic sea level rise (see Vail et al., 1977; Vail and
Mitchum, 1979).

Besides, a chain of events, possibly related to the
evolution of the Hellenic subduction, caused the break-up of
the Southern Aegean Landmass and of the basin bordering it to
the south (Drooger and Meulenkamp, 1973; Meulenkamp, 1985). At
the same time, the Cyclades were subject to large scale uplift
(Durr et al., 1978; Lister et al., 1984).

According to Meulenkamp (1979, 1985) and Meulenkamp et al.
(1988), compressional tectonics in the Late Serravallian
resulted in slight folding which became more pronounced in the
Early Tortonian and caused the development of synclinal
depressions and anticlinal culminations. According to Fortuin
(1977) and Meulenkamp and Hilgen (1986), the gravitational
spreading, which affected the Aegean region at the same time,
caused pre-Neogene and Neogene slabs from the culminations to
slide into small basins.

Furthermore, the movements which took place in the Late
Serravallian-Early Tortonian time, were responsible for
shaping the general configuration of the present Cretan region
(Meulenkamp, 1985).

3. Late Miocene (Late Tortonian-Messinian)
During that period, Crete again formed a fairly stable
large block which was for the larger part submerged (Fig.
11C), (Drooger and Meulenkamp, 1973).

The Tortonian-Messinian time interval boundary was

characterized by a tectonic instability which caused changes in basin configuration and sedimentation. This is expressed by tilting and erosion of the older parts of the Neogene sequences (Meulenkamp, 1985). Accumulation of carbonates are found over the previously eroded and uplifted blocks and suggest a transgression (Meulenkamp, 1985).

In the Messinian (Fig. 11D), the partial obstruction of the passage between the Mediterranean Sea and the Atlantic ocean ("Messinian salinity crisis") influenced the tectono-sedimentary pattern of Crete by causing many drastic geodynamical, palaeogeographical and biological changes.

During the Early Messinian, evaporites of different nature and thicknesses were laid down in various parts of Crete (Meulenkamp, 1985). The variation in thicknesses may reflect different subsidence rates of the E-W and N-S oriented sub-basins (Meulenkamp et al., 1979a; Meulenkamp et al., 1979b).

The sedimentation pattern during the Messinian was also controlled by a rejuvenation of the relief, which resulted in the deposition of both coarse conglomeratic, nonmarine successions and fluvio-lacustrine to lagoonal deposits, the latter predominantly in western Crete (Meulenkamp, 1985).

4. Early Pliocene

At the beginning of the Pliocene, the marine connection between the Atlantic Ocean and the Mediterranean Sea was restored, causing a geologically rapid flooding of the Mediterranean. In western Crete, the upper Messinian clastics

6. Pleistocene

During this timespan, Crete obtained its present landscape.
Renewed fragmentation and strong differential vertical
movements were the most prominent tectonic features
controlling the Quaternary landscape evolution. The old N-S,
E-W fault systems are still active and determine the shape of
the island, but the relief and the terrestrial sedimentation
are controlled by a new set of NW-SE and SW-NE faults (Drooger
and Meulenkamp, 1973; Meulenkamp, 1985). It is noted that the
orientation of this new fault system seems to follow the
orientation of the Hellenic Arc, as the NW-SE system is more
important in western Crete and the NE-SW system in eastern
Crete. Both these systems can be followed over great distances
outside Crete (Drooger and Meulenkamp, 1973)

Many displacements along the N-S, E-W and the NW-SE, SW-NE
fault systems occur even today, which result in the continuing
accentuation of the topographic relief. The majority of these
displacements cause a general tilt of the island to the north
(Drooger and Meulenkamp, 1973).

To summarize, the Neogene - Quaternary evolution of the
Cretan area can be divided into six successive intervals, each
one marked by pronounced changes in basin configuration and
sedimentation patterns. The most important changes seem to
have taken place in the Late Serravallian - Early Tortonian
times between about 12 Ma and 11 Ma ago, when the Southern

Aegean Landmass, hitherto connecting Crete with the European mainland, started to break up. At the same time, Crete itself was being fragmented into several basins (Meulenkamp and Hilgen, 1986). The fragmentation of Crete has been attributed by Le Pichon and Angelier (1979) to the initiation of subduction of the African plate beneath the Aegean lithosphere (13 Ma ago).

New data on the structure of the Upper Mantle underneath the Aegean region indicate an age for the Hellenic subduction of at least 26 Ma (Meulenkamp et al., 1988). Taking into consideration this assumption, the important changes which occurred in the Late Serravallian/Early Tortonian boundary interval (12-11 Ma ago) are not associated with the initiation of subduction but with the inception of "back-arc processes" (Meulenkamp et al., 1988) which led to the south-southwestward migration of the Hellenic Trench system. The beginning of these "back-arc processes" seems to have significantly affected the stress field (see Wortel and Cloetingh, 1986) in particular along the arc, and turned it into a tensional regime (for more details see Ch. 2.2). According to Meulenkamp et al. (1988), these "back-arc processes" can be related to the final stages of the collision between Africa/Arabia and Turkey (see Ch. 2.1).

CHAPTER 3

IERAPETRA REGION

3.1 REGIONAL SETTING

The Ierapetra region is located in Eastern Crete and
occupies the area of the Ierapetra and Merabellou districts of
the Prefecture of Lasithi (Fig. 12; Fortuin, 1977).

Its Neogene geological evolution is controlled by the
active continental collision of the African and Eurasian
plates, that is the whole region was affected by extensional
deformation during the Early Miocene and compressional
deformation during its further evolution in the Miocene
(Meulenkamp et al., 1988).

The Neogene rocks of the Ierapetra region extend over an
area of 500 km² and consist of coarse clastic sediments at the
base which pass upwards into marine marls, sands and
limestones (Fortuin, 1977). They are underlain by the pre-
Neogene rocks, which form a complex structure of an
autochthonous basement of Permian-Oligocene age (Plattenkalk
Series) overlain by four allochthonous units (Phyllite-
Quartzite, Tripolitza, Pindos and Sub-Pelagonian) (Baumann et
al., 1976). After the emplacement of this nappe pile in the
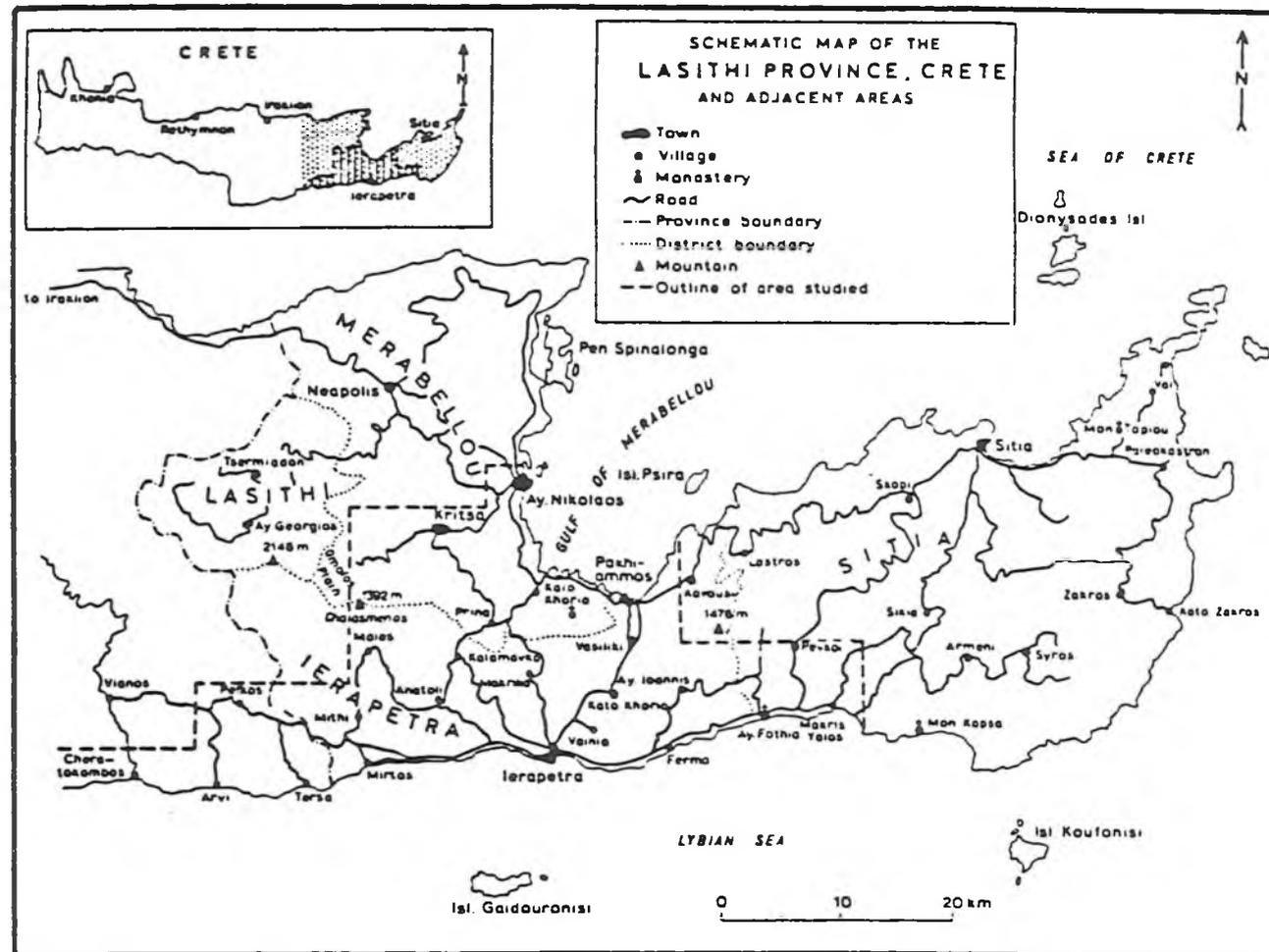
FIG. 4: Schematic map of the Lasithi Province and adjacent areas
(after Fortuin, 1977).

Oligocene-Early Miocene times (Meulenkamp, 1971; Angelier,
1979), strong block-faulting along E-W and NE-SW directions
affected the whole region and formed small sedimentary basins
in which the complex interaction between the faults resulted
in the frequent change of land-sea distribution (Drooger and
Meulenkamp, 1973; Meulenkamp, 1979). This complicated pattern
of land and sea during the Late Cenozoic caused rapid lateral
and vertical changes in the lithology (Fortuin, 1977).
Besides, the stratigraphy of the Ierapetra region became more
complicated in the Messinian when the Messinian salinity
crisis and its consequences had a strong impact upon
sedimentation.

According to Drooger and Meulenkamp (1973), Jongsma (1977)
and Jongsma et al. (1977), there is a strong relationship
between sedimentation, tectonics, climate and erosional
processes. Thus, the pattern of sedimentation in the Ierapetra
basin is controlled by the strong crustal movements which act
upon the area and is characterized by sediments of detrital
origin and gravity slides found at various stratigraphic
levels and in various depositional environments (Fortuin,
1977; Fortuin, 1978; Fortuin and Peters, 1984).

It should be pointed out that the present topography of the
Ierapetra region reflects some important Late Pliocene-
Quaternary movements along faults. The most striking tectonic
feature of the present day morphology of the Ierapetra region
is the central NE-SW depression with its eastern margin
bounded against the Quaternary Ierapetra fault (La fosse d'

Ierapetra: Angelier, 1976; Angelier, 1977). According to many tectonic maps of the South Aegean region, its western margin is a series of active normal faults, (although this structure is doubted by Fortuin and Peters (1984)). The orientation of this graben or half-graben depression conforms to the orientation of the South Cretan Trough (a branch of the eastern Hellenic Trench system), which, according to Leité and Mascle (1982), is the offshore continuation of the Ierapetra region (Fig. 13).

3.2 STRATIGRAPHY OF THE IERAPETRA REGION

Fortuin (1977) subdivided the Neogene of the Ierapetra region into nine formal rock units: eight formations and one complex, each one corresponding to different environmental conditions. The stratigraphy is complicated by tectonic movements during the sedimentation, especially during the Late Serravallian when the break-up stage of the Aegean landmass started. The Neogene sediment distribution of the Ierapetra basin is shown in Fig. 14 and Fig. 15. The main characteristics of these formational units are cited below (from bottom to top) (Fig. 16):

1. Mithi Formation (150 m thick)

Conglomerates of poorly-sorted pebbles with a reddish or greyish weathering colour. The components consist
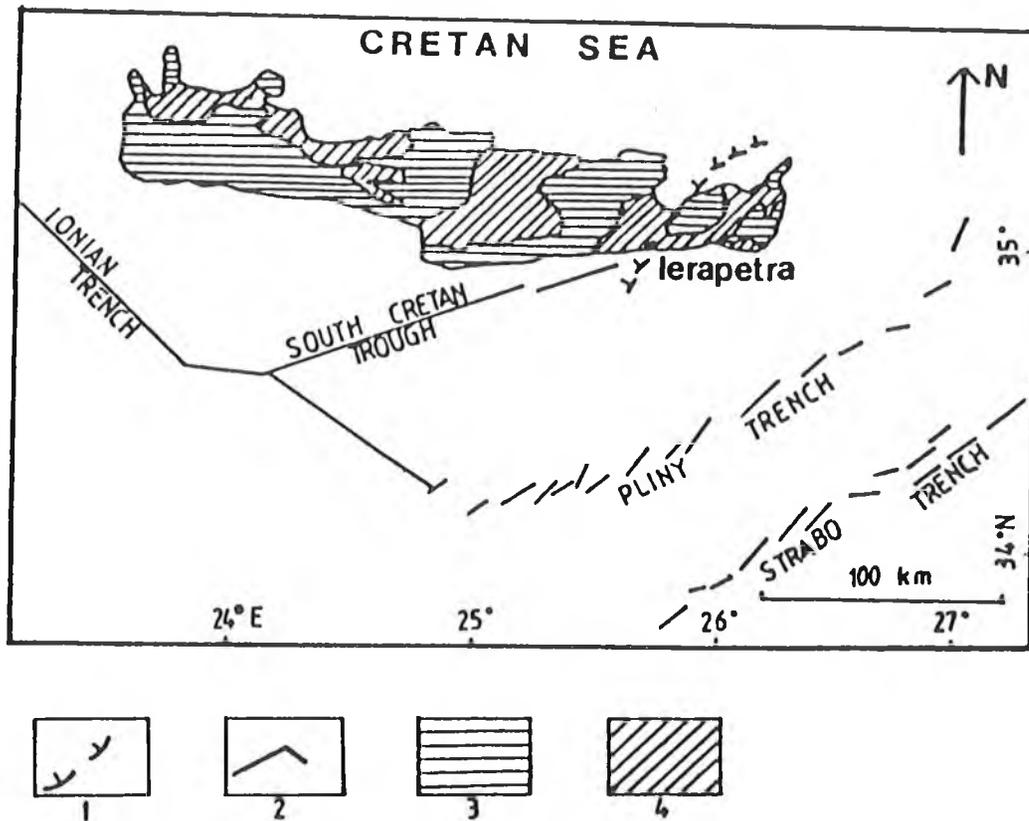
FIG. 13: Geological sketch map of the south Hellenic arc showing the Ierapetra fault (after Angelier et al., 1982 and Mascle et al., 1982b).
1. normal fault, 2. trench axis, 3. pre-Neogene nappe pile, 4. Neogene and Quaternary sediments.

LEGEND

Alluvium

Pliocene

Allochthonous Neogene

Autochthonous Neogene

Subpelagonian Series

Phyllite-Quartzite, Tripolitza, Pindos Series

Plattenkalk Series

thrust

blind thrust

anticline

normal fault

tectonic boundary

CRETE

Ay. Nikolaos

Ierapetra

0   km   15
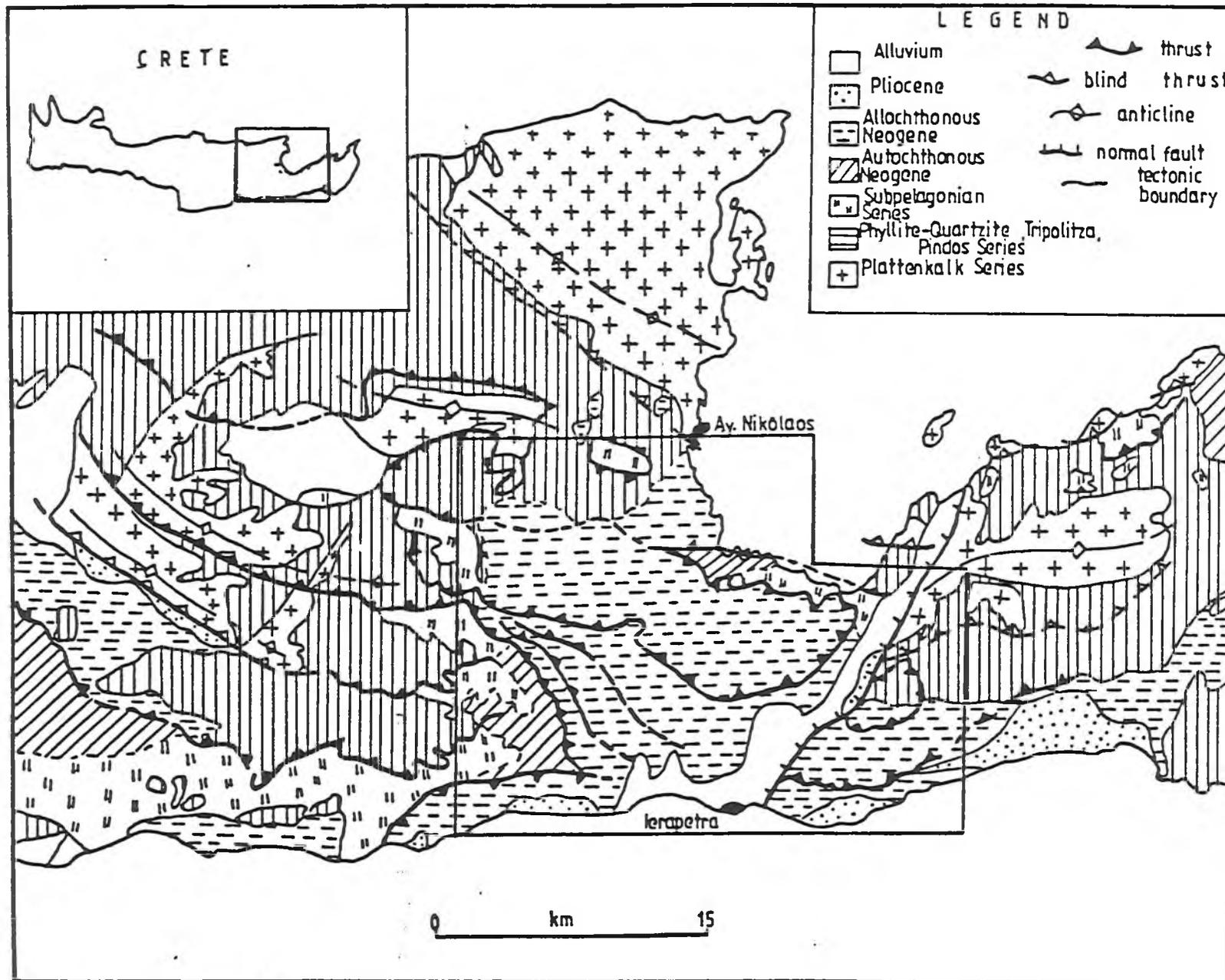
FIG. 14: Simplified geological map of the Ierapetra region (after Postma, pers. comm.), indicating the outline of the central part of the region (FIG. 15).
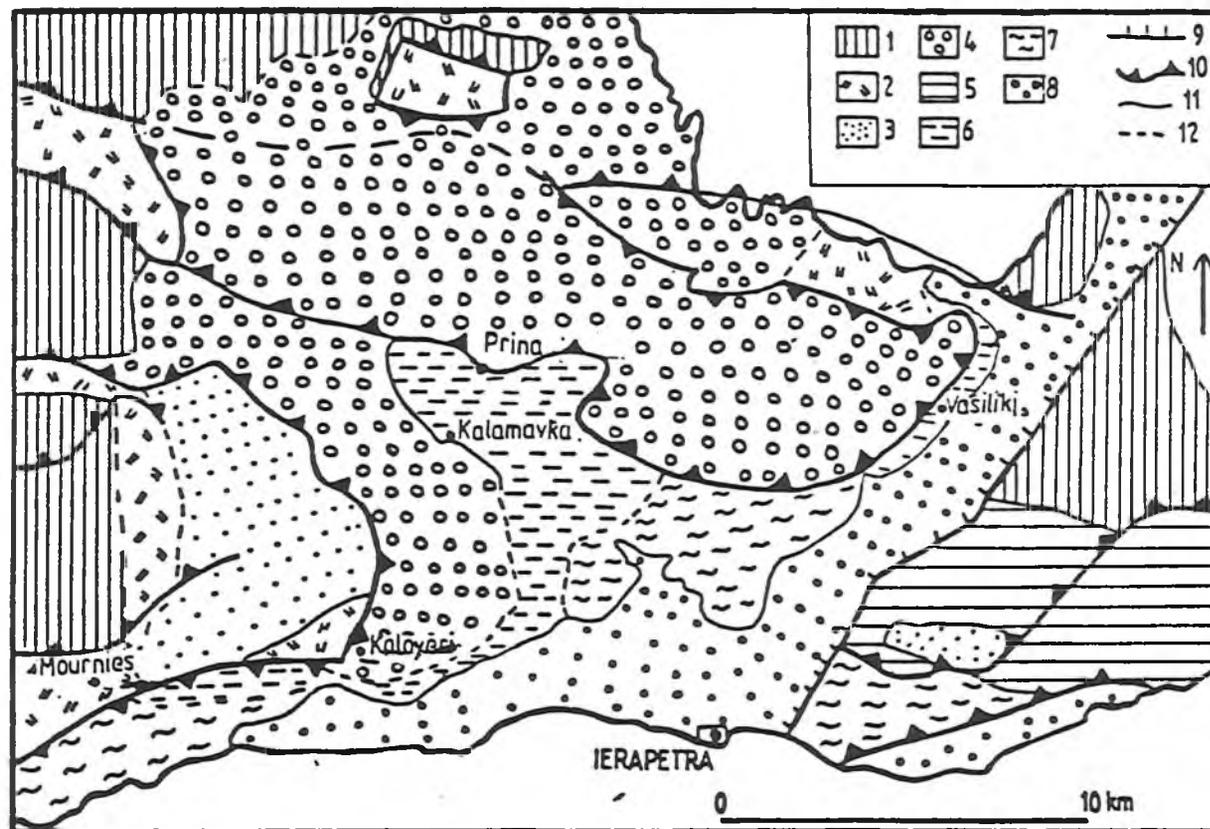
FIG. 15: Simplified geological map of the central part of the
Ierapetra region (after Fortuin, 1977; modified by Postma,
pers. comm.).
1. Tripolitza Series; 2. Subpelagonian Series;
3. Mithi and Males Formations; 4. Prina Complex;
5. Fothia Formation; 6. Kalamavka Formation;
7. Makrilia Formation; 8. post-Tortonian sediments;
9. normal faults; 10. thrust faults;
11. tectonic contact; 12. stratigraphic contact.

FIG. 16: Generalized stratigraphic column of the Ierapetra basin.

M.a	epoch	age	FORMATION	LITHOLOGY
	PLEISTOCENE	Tyrrenian		Bioclastic limestones
		Milazzian		
		Sicilian		
1		Emilian		
		Calabrian		
2	PLIOCENE	Piacenzian	PAKHI-AMMOS	
3				
4		Zanclean	MIRTOS	
5				
6	LATE MIOCENE	Messinian	AMMOU-DHARES	
7				
8		Tortonian	MAKRILIA	
9				
10			KAVALA PRINA / FOTHIA	
11				
12	MIDDLE MIOCENE	Sarravallian	MALES	
13				
14		Langhian	MITHI	
15			? ? ? ? ?	

LEGEND

~~~ Marl

.... Sandstone

Limestone

>>> Gypsum

Conglomerate

Breccia

Pre-Neogene rocks

predominantly of multicoloured igneous and metamorphic rocks, most of which appear to have originated from the underlying pre-Neogene. This formational unit which is thought to be the oldest continental Neogene of Crete, has not be found east of the Ierapetra fault (Fig. 15) (Fortuin, 1978). It has not been dated because of the lack of datable material in these deposits.

2. **Males Formation** (350 m thick, ? - Late Serravallian)

The Males Formation consists of alternations of mature conglomerates, sandstones and clayey marls. The conglomerates are well-sorted and well-rounded of rather constant composition. Most of them were derived from the allochthonous pre-Neogene Pindos unit. On top of this succession, marly deposits predominate forming the **Parathiri Member**. The Males Formation is usually found on top of the Mithi Formation or the pre-Neogene rocks and is overlain by either the Prina Complex, the Kalamavka Formation, the Fothia Formation or the Makrilia Formation (east of the Ierapetra fault; **Fig. 16**).

3. **Prina Complex** (350 m thick, Late Serravallian - Early Tortonian).

This formational unit is a widely distributed rock unit which covers the northern part of the Ierapetra region (Fig. 15). It consists of stratified breccias and breccio-conglomerates and conglomerates with fine-grained interbeds. Most of the clastics were derived from the pre-Neogene

Tripolitza unit. In the NW part of the region, marine boulder conglomerates with regular intercalations of well-sorted calcareous sandstones predominate. Slabs and very large masses of dark, brecciated Tripolitza limestones usually prevail in the basal parts of the complex.

4. Fothia Formation (450 m thick, Late Serravallian - Early Tortonian).

Irregularly stratified, polymict conglomerates with ill-sorted and poorly rounded pebbles, alternating upwards with sands and marly deposits. This formational unit is found only east of the Ierapetra fault (Fig. 15) and its composition is related to the pre-Neogene rock units which are found northwards of its exposure. The Fothia Formation overlies the Males Formation or the pre-Neogene rocks and is overlain by the marine marls of the Makrilia Formation (Fig. 16).

5. Kalamavka Formation (300 m thick, Late Serravallian - Early Tortonian).

Rhythmically alternating marls and calcareous sandstones. Pebbly mudstone intercalations as well as other conglomeratic bodies are common in different levels. The Kalamavka Formation overlies the Parathiri Member of the Males Formation or the Prina Complex and in part is a lateral equivalent of the latter. It is overlain by the marine marls of the Makrilia Formation (Fig. 16). The Kalamavka Formation constitutes the

core of this study and will be dealt later with in much more detail.

6. Makrilia Formation (450 m thick, Early Tortonian).

Marine bluish marls alternating with brownish, graded sands (turbidites). It differs from the Kalamavka Formation by the more regular bedding, the brownish colour and the non-lithified nature of the sands. The Makrilia Formation overlies or is, for a part, laterally equivalent with the Kalamavka Formation. In some places, it directly overlies the marine top of the Males Formation (Parathiri Member). East of the Ierapetra fault , the formation is found overlying the Fothia Formation (Fig. 15). The Makrilia Formation is overlain unconformably by the Ammoudhares Formation (Fig. 16).

7. Ammoudhares Formation (50-100 m thick, Late Tortonian - Messinian).

Yellow-grey, highly calcareous, homogeneous or laminated marls, alternating with sandy bioclastic limestones. Sponge spicules are abundant in the laminated marls. Pebbly mudstones or slumped strata with limestone olistoliths are also very common. The whole formation has been disturbed by intensive small-scale faulting due to post-depositional slumping or sliding. The Ammoudhares Formation overlies the Makrilia Formation, apart from the surroundings of Prina where it overlies the Kalamavka Formation. Itself, it is overlain by the Mirtos Formation (Fig. 16).

8. Mirtos Formation (60 m thick, Early Pliocene).

Heterogeneous deposits of gypsum, white and greyish marls, marl breccias and sands. Usually, the basal part of the formation is disturbed by discontinuous, displaced slabs or bodies of gypsum (Fig. 16). The Mirtos Formation is only found in the southern part of the Ierapetra region and is overlain by Quaternary deposits.

9. Pakhiammos Formation (30 m thick, Middle Pliocene).

This formation is only found in the north coast of the Ierapetra region. It consists of cavernous, micritic limestones, marl breccias and whitish marls. It can be found overlying pre-Neogene rocks or conglomerates of the Prina Complex. Wherever it overlies pre-Neogene rocks, terrigenous clastic deposits are found at its base.

In some parts of the Ierapetra region, especially along the south coast, post-Neogene rocks are found to form large terraces at various heights (Fortuin, 1977). The Quaternary deposits can be divided into three main characteristic groups: 1) bioclastic limestones which are whitish, massive packstones consisting of Miliolids and algae, 2) siliciclastic terraces which are composed of conglomerates, sands and marls and 3) rockfall and landslides.

comm.; **Fig. 17**). Thrusting caused the uplift of the northern part of the region and exotic blocks and coarse clastics of continental to littoral origin started to accumulate south of the thrust fault (**Prina Complex**). The continuous accumulation of sediment there and the imposed weight of the thrust mass (isostatic loading) caused flexure and gradual subsidence of the crust and the formation of a graben-like depression in the central part of the region, which had an open marine connection in the south (Fortuin, 1978). Due to the main thrust fault, other secondary thrust faults were formed (**Fig. 17**), which deformed the already continental to littoral sediments of the Prina Complex and caused the deposition of fluvio/marine conglomerates (still **Prina Complex**) and calcareous mudstones and sandstones (**Kalamavka Formation**).

East of the Ierapetra fault, the lithostratigraphic succession is different (Fortuin, 1977). Instead of the Prina Complex, the **Fothia Formation** is deposited next to the fault (**Fig. 15**). According to Fortuin (1978), the sedimentation of the Fothia Formation was initially controlled by rapid accumulation of coarse terrigenous clastics in a piedmont area along pronounced relief found in the north. In a later stage, the relief levelled-off and most of the clastics were supplied by braided streams over a floodplain and transported towards the SW. Continuous subsidence finally resulted in a gradual shift towards marine sedimentation conditions.

In the Early Tortonian, in response to further subsidence, a deeper basin was formed which was filled with marl and a few

3.3 TECTONO-SEDIMENTARY HISTORY OF THE IERAPETRA REGION

According to Meulenkamp (1979) and Drooger and Meulenkamp
(1973), the Neogene sedimentary history of the Ierapetra
region started with the deposition of the **Mithi Formation** (the
timing of which has not been defined), which comprises coarse
clastic alluvial fan deposits originated from earlier tectonic
movements or from the erosion of a pronounced relief. During
that time, the region was part of a continental borderland,
situated at the southern margin of the South Aegean Landmass
(Drooger and Meulenkamp, 1973).

On top of this formation, conglomerates, sandstones and
clays of the **Males Formation** were laid down into a large, east
to west flowing, fluvial drainage system. In the Late
Serravallian, the fluvial sedimentation ceased because of the
gradual submergence of the area which was caused by strong
blockfaulting due to crustal tension (Drooger and Meulenkamp,
1973). As a consequence, fossiliferous, shallow marine marls
(**Parathiri Member**) were deposited on top of the Males
Formation.

During the same period, the break-up stage of the Southern
Aegean Landmass started (Drooger and Meulenkamp, 1973;
Meulenkamp, 1985; Meulenkamp et al., 1988). As a consequence a
N-S shortening affected the Ierapetra region. During the Late
Serravallian-Early Tortonian period, compressional tectonics
affected the Ierapetra region resulting in the formation of a
major, approximately E-W oriented, thrust fault (Postma, pers.

FIG.17: Palaeogeographical reconstruction of the Ierapetra region
during the Late Serravallian-Early Tortonian.
The activation of the major thrust fault caused the uplift
and denudation of the northern part and subsidence (due
to isostatic loading) of the southern part. In the
beginning, mainly continental to littoral coarse clastics
were accumulated next to the fault (Prina Complex).
Further to the south, more pronounced subsidence (due to
the activation of secondary thrust faults) resulted in the
deposition of fluvio/marine conglomerates (still Prina
Complex) and marine marls and sandstones of the Kalamavka
Formation.

turbiditic sandstones and extended over the south coast areas.
This basin was filled with sediments derived from relatively
distant sources in the west (Makrilia Formation) (Fortuin,
1977). During this period, the gradual submergence of the
northern parts of the Ierapetra region started (Drooger and
Meulenkamp, 1973).

The Messinian is characterized by the gradual emergence of
the area, which resulted in the "Messinian facies" and is
attributed to the Late Miocene salinity crisis of the
Mediterranean sea (Hsü et al., 1976). Limestone sedimentation
became more pronounced all over the island of Crete. Before
the end of the Messinian, marine conditions returned again,
causing the submarine sliding and slumping of the Late
Tortonian - Messinian calcareous deposits, in the south coast
areas (Fortuin, 1977).

Finally, during the Early Pliocene, marl breccias were
deposited, indicating submarine mass transport (Fortuin,
1977). On top of the marl breccias, marls were deposited in an
open and quiet marine environment. In the Middle Pliocene, a
renewed uplift started and by the end of the Pliocene, the
whole area had emerged (Drooger and Meulenkamp, 1973).

From the previously discussed structural evolution of the
Ierapetra region, it is evidenced that a big change in its
tectonic regime occurred in the late Serravallian times: The
N-S extensional phase which acted on Crete and was evidenced

by the deposition of mature, gravelly braided river facies
(Males Formation) switched to a N-S compressional phase which
caused the deposition of the Prina Complex (immature carbonate
clastics) and the Kalamavka Formation (rhythmical alternations
of calcareous mudstone and sandstone). According to Meulenkamp
et al. (1988), the timing of this change coincides with the
onset of the south-southwestward migration of the Hellenic
Trench system and the inception of the "back-arc processes"
and is postulated to be directly related to it.

CHAPTER 4

KALAMAVKA FORMATION

4.1 STRATIGRAPHY AND GENERAL CHARACTERISTICS

The Kalamavka Formation is a rock unit with a limited geographical extension (20 km E-W and 10 km N-S). It mainly covers the area between the villages of Kalamavka-Prina-Kaloyeri (Fortuin, 1977) (Fig. 15), and is also found in a narrow E-W strip between Kaloyeri and Mournies and around the village of Vasiliki, NE of Ierapetra (Fortuin, 1977). The Kalamavka Formation is well developed near the Kalamavka village (over 350 m thick), (type section of the Formation). The type section (which has been studied in detail in this work) is exposed along the road from Kalamavka to Prina, 12 km NW of Ierapetra (Fortuin, 1977) (Fig. 18).

The Kalamavka Formation developed during the Late Serravallian - Early Tortonian times, when the extensional phase which acted on Crete switched to a N-S compressional phase. It is found southwards of the thrust fault which delimits the northern uplifted deposits from the southern subsiding areas (Fig. 17). It usually overlies the marly deposits of the Parathiri Member (Males Formation), although its contact is poorly exposed (Fig. 16). As far as the Prina

LEGEND

ROCK UNITS

PRINA COMPLEX

KALAMAVKA FORMATION

TRIPOLITZA LIMESTONE BODIES

GEOLOGICAL SYMBOLS

sandstone beds

conglomeratic bodies

THRUST FAULT

NORMAL FAULT

STRIKE-SLIP FAULT

FIG. 18: Geological map of the studied region, indicating
the studied cross-sections AB and CD (scale 1:50.000).

Complex is concerned, Fortuin (1977, 1978) suggested that it

can either underlie or interfinger with the Kalamavka

Formation. (However,) field data shows that the boundary between

the Prina Complex and the Kalamavka Formation is uncertain.

The Kalamavka Formation is overlain by, or it is a lateral

equivalent of the brownish graded sands and the marine marls

of the Makrilia Formation (Fig. 16). East of the Ierapetra

fault, the only place the Kalamavka Formation may be

represented is in the lower 30 m of the Makrilia Formation

where it overlies the Fothia Formation (Fortuin, 1977).

The contact between the Prina Complex and the Kalamavka

Formation can be either of a tectonic or sedimentary nature

(Fig. 15). Where there is a tectonic contact, it constitutes a

very gently dipping thrust fault, found at the northern margin

of the basin. Many other faults secondary to the main thrust

fault were formed dipping to the north and causing the

displacement of the Prina Complex and the deformation of the

Kalamavka sediments. This is clear from the fact that near the

thrust fault, Prina Complex deposits are found with

intercalations of deformed Kalamavka-type sediments.

At the western margin of the basin, Kalamavka sediments are

overlying a fining-upward sequence of immature breccio-

conglomerates, belonging to the Prina Complex (stratigraphic

contact). Although indications for a clear, gradual transition

have not been found, there is still some evidence of normal

faulting or thrusting. The dips of the Prina deposits are

quite similar to the dips of the Kalamavka sediments (towards

NE), which means that both formational units might have undergone the same tectonic history. Biostratigraphic data showed that where a stratigraphic contact exists between the Kalamavka Formation and the Prina Complex, the Prina sediments have a slightly different age (younger) from the Prina sediments which have a tectonic contact with the Kalamavka Formation (see also explanation of **Fig. 17**).

The Kalamavka Formation is characterized by predominantly marine fine-grained clastics (Fortuin, 1977). Its submature sediments possibly originated from both the uplifted basement (the Mesozoic Tripolitza limestone) found NW of the major thrust fault and partly from the adjacent coarse-grained Prina Complex. Paleocurrent measurements indicate a SE directed transport of sediments (Fortuin, 1977).

The studied section (type section of the Kalamavka Formation) is exposed along the road to Prina (Fig. 18, pl. 1A, pl. 1B). It consists of alternating, bluish-grey fossiliferous marls and dark grey, strongly lithified, calcareous sandstones. Its basal, predominantly marly beds (section AB Fig. 18), which conformably overlie breccio-conglomerates of the Prina Complex, are overlain by thin to thick, partly amalgamated sandstones (section CD Fig. 18). In the lower and middle part of the Formation, there are some conglomeratic bodies similar to those found in the Prina

PLATE 1A: The Kalamavka Formation along the road from Kalamavka
to Prina

PLATE 1B: General view of the Kalamavka Formation

Complex. The whole section has an average dip 30°NE and forms
a crude coarsening-upward trend. A detailed vertical profile
through the studied type section of the Kalamavka Formation is
found in Appendix B.

4.2 PETROGRAPHICAL AND BIOSTRATIGRAPHICAL COMMENTS

Before entering a description of the Kalamavka sediments,
it will be worthwhile making some comments about their
petrography and biostratigraphy.

Selected thin sections, which cover the sequence of the
Kalamavka Formation in its type-section, have been examined in
order to establish the petrographic and biostratigraphic
framework of the Kalamavka sediments. Detailed results are
found in Appendix A.

Generally, the Kalamavka sandstones are lithic sandstones,
relatively low in quartz and feldspars, and contain a variety
of rock fragments. These sandstones are, mostly, very fine to
fine and calcareous. The rock fragments are usually of
sedimentary or secondary diagenetic origin, (although there
may be some transported metamorphic fragments), chiefly
limestone and chert. Micritic limestones are the most common
types of sedimentary rock fragments. These calcareous
sandstones are usually fossiliferous with the most common

fossils being foraminifera. Other fossils such as fragments of gastropods and ostracod valves also occur.

The finer grained sandstones tend to contain more calcareous mud or more sparry calcite than the coarser sandstones.

The fossils that have been identified in the thin sections studied are the uniserial *Uvigerina* and some planktonic foraminifera, mainly Globigerinidae. The identified fossils are in complete agreement with those that Fortuin (1977) found, which are the following:

- *Uvigerina melitensis* or *U.* ex. interc. *pappi melitensis* (usually found in the lower part of the studied section)
- *Uvigerina praeselliana* which together with *Uvigerina gaulensis* are found in the higher levels of the type section.
- And finally, the primitive *Uvigerina selliana.*

These *Uvigerina* specimens suggest a **Late Serravallian – Early Tortonian** depositional age for the Kalamavka Formation (Fortuin, 1977).

This assumption is supported by the fact that *Glob̶i̶g̶e̶r̶i̶n̶a̶* Orotalia *menardii* has also been detected (Fortuin, 1977).

The previously mentioned faunal data indicate deposition in an open marine environment at an approximate depth of 100 to 200m (Murray, 1973).

This conclusion is also supported by the marine burrows which have been found in the Kalamavka sediments. Burrows are of the *Scolithos* and *Cruziana* Ichnofacies, that belong to the marine realm.

According to Ekdale et al. (1984), the *Scolithos* Ichnofacies exemplifies intertidal settings in which shifting and fluctuating water conditions favour preservation of primary sedimentary structures. The *Cruziana* Ichnofacies occurs in shallow marine settings below the low tideline but usually above storm wave base (Ekdale et al., 1984).

The above palaeontological data indicates that the sediments belonging to the Kalamavka Formation were probably deposited in a shallow marine shelf environment with depths not exceeding 200 m.

4.3 SEDIMENTARY FACIES: DESCRIPTION AND INTERPRETATION

The studied section of the Kalamavka Formation has been divided into five Facies, each one identified by its field aspect, physical and biological sedimentary structures, grain size and lithology. These facies are as follows (Fig. 19, Appendix B):

Facies 1: Calcareous mudstone.

Facies 2: Alternations of very thin-bedded, laminated sandstone and mudstone.

FA-CIES	DEPOSITIONAL PROCESS	DEPOSITIONAL SETTING
4	Storm-generated currents Wave reworking	LOWER SHOREFACE
5	Flood-generated subaqueous debris flow Wave reworked at the top	DEBRIS FLOW LOBE
2	Storm-generated turbidity currents marine reworking	OFFSHORE
3	Storm-generated currents	OFFSHORE-TRANS
1	Suspension punctuated by waning traction currents	OFFSHORE (shelf)
5	Flood-generated debris flow	DEBRIS FLOW LOBES
1	Suspension	OFFSHORE

m

500

100

0

m (sst) cong

5m

m (sst) cong

40 cm

0

m (sst) cong

5m

m (sst) cong

50 cm

0

m (sst) cong

5m

m (sst) cong

50 cm

0

m (sst) cong

5m

m (sst) cong

50 cm

0

m (sst) cong

FIG. 19: Vertical profile through the Kalamavka Formation (type section). The legend is found on the next page.

LEGEND

▢	: Sandstone
▢	: Marls
——— f	: Minor fault
✕	: Not exposed
⊔	: Burrows
⌒	: Current ripples
‿‿‿	: Small-scale cross stratification
△	: Hummocky cross stratification
~	: Drape of mud
∫	: Slumping

▤	: Very thin alternations of sandstone and marls
▨	: Conglomerate
~	: Undulatory boundary- erosive contact
⋀	: Wave ripples
=	: Parallel lamination
⌒	: Oyster shell
⊚	: Fossil
●	: Clast of mud
ℓ	: Intraclast

Legend of Fig.19, Fig.20, Fig.21, Fig.22, Fig.23, Fig.24 and Appendix B.

Facies 3: Fine to medium-grained, thin-bedded
          sandstones.

Facies 4: Medium to coarse-grained, thin to thick-
          bedded, hummocky-cross-stratified sandstones.

Facies 5: Debris flow lobes.

In this part of the thesis an initial interpretation of
each facies is given after each description. An integrated
interpretation will be suggested later in Chapter 5
(Discussion). The facies distribution of the studied sequence
is shown in Appendix B.

FACIES 1

Description

Facies 1 (Fig. 20, pl. 2, Appendix B, p.B1, B2) comprises
bluish, fossiliferous, calcareous mudstones with occasional
lenses or layers of fine to very fine grained sandstone (pl.
2A). The mudstone is generally massive but in places thin
laminations are visible (pl. 2B). The sandstone beds are very
thin and increase in abundance and thickness as Facies 1
grades into Facies 2 (interbedded mudstone and sandstone).
The sandstone layers are usually massive but parallel-
lamination, ripple cross-lamination, lenticular and wavy
bedding also occur.

Due to the strong bioturbation most commonly of the
Scolithos Ichnofacies (pl. 2C) which affected the sediments of
Facies 1, the lower and upper surfaces of the thin sandstone

Facies 1

FIG.20: Selected vertical log through Facies 1 (LOC 2, Appendix B, p.B1). The facies distriburion and its relationship with the other facies is shown in Appendix B. For further explanation see text.

PLATE 2C: Bioturbated Facies 1 (*Scolithos* Ichnofacies).

layers are frequently indistinct.

Facies 1 dominates mainly the basal part of the studied section and is largely exposed in LOC 1 and LOC 2 (Fig. 18, Appendix B.

Interpretation

The mud-dominated Facies 1 is thought to represent open marine (shelf) environment. The fossiliferous and burrowed mudstones indicate deposition from suspension in a quiet environment which has hardly been influenced by waves or other currents. The deposition of mud by suspension was occasionally interrupted by waning sand-laden traction currents (deposition of the sandstone beds).

FACIES 2

Description

Facies 2 (Fig. 21, pl. 3, Appendix B, p.B3, B8, B9, B10, B11) consists of very thin to thin-bedded and very fine to fine-grained sandstones found embedded in shelf mud.

All these sandstone beds begin with a sharp base and mostly grade upwards into overlying mud layers. The sandstone:mudstone ratio is usually 1:1 and the average thickness of the beds is 1 to 2 cm. The sandstone beds are either evenly laminated or are developed as laminated rhythmites (sensu Reineck and Singh, 1972), (pl. 3A), in which the lower laminae are thicker and coarser grained and grade upwards into thinner and finer-grained laminae. The upper

Facies 2

FIG. 21: Selected vertical log through Facies 2 (LOC 5,
Appendix B, p. B5). For further details see text and
Appendix B.

contact of each bed with the intercalated mudstone is sharp
and smooth, yet there are some sandstone beds with wave-
rippled tops.

On the surface of the very thin-bedded sandstones, there
are some curvilinear, randomly distributed imprints which are
possibly traces of crab (pl. 3B) (Fortuin, 1977). Primary
sedimentary structures throughout Facies 2 are reworked by
strong, mainly vertical, bioturbation (pl. 3C).

Facies 2 is dominating in the southern part of the basin
(LOC 3 and LOC 4) and usually overlies the massive mudstone of
Facies 1. Locally, it is characterized by a crude fining-
upward trend.

Interpretation

According to Walker (1979), the mechanism for the transport
and deposition of sand offshore is by density driven turbidity
currents which evolve from storm-generated currents and can
transport and deposit sediment well below storm wave-base.

Hamblin and Walker (1979) and Leckie and Walker (1982)
worked on offshore/shelf facies which include sandstone beds
1-2 cm thick, characterized by parallel-lamination grading
upwards into wave ripple lamination and interpreted them as
originating from storm-generated turbidity currents which
escaped beyond storm wave base and deposited sediment in the
offshore-shelf area.

The crude fining-upward trend of such deposits is thought
to be due to the declining energy of successive, episodical

PLATE 3: FACIES 2

PLATE 3A:
Alternations of very thin-
bedded sandstone and mudstone,
parallel-bedded and varve-
like in appearance.

PLATE 3B: Curvilinear imprints on the
surface of very thin bedded
sandstones.

PLATE 3C: Reworking of Facies 2 by strong, mainly vertical bioturbation (indicated with the arrow). .

currents. That is, in the beginning, when the energy level of
the storm-generated current is high, only sand in the form of
parallel laminae is deposited, so that laminated sandstone
beds are produced. Later, with a decrease in wave energy,
finer-grained, wave cross-stratified sediments and finally mud
with rather ill-defined laminae are laid down, forming the
observed fining upward, small-scale sequences. This kind of
mechanism of deposition has been studied by Reineck and Singh
(1972) and proposed by Nøttvedt and Kreisa (1987).

FACIES 3

Description

Facies 3 is characterized by fine to medium grained, thin
to medium-bedded, moderately sorted sandstones (Fig. 22, pl.
4, Appendix B, p.B4, B5, B8, B10, B11). It usually grades into
Facies 4 and is laterally persistent.

Generally, the base and the top of each sandstone bed of
Facies 3 are sharp and smooth surfaces but they can also be
disrupted by moulds or burrows.

The base of many sandstone beds of this Facies usually show
a distinct laminated interval with some scattered outsized
pebbles. This laminated interval is covered by wave rippled
sand (pl. 4A and pl. 4B). The sandstone beds are separated by
thin marly horizons or by very thin, strongly bioturbated
alternations of mudstone and wave-worked, very fine-grained
sandstones (pl. 4C). Otherwise, they are amalgamated.

In some beds, granules and pebbles are either scattered

Facies 3

LOC 5

FIG. 22: Selected vertical log through Facies 3 (LOC 5, Appendix B, p. B5). For further details see text and Appendix B.

PLATE 4: FACIES 3

PLATE 4A:
Medium to fine-grained
laminated sandstone bed.

PLATE 4B: Wave rippled top of a medium-grained sandstone

PLATE 4C: Medium-bedded sandstones separated by very thin, strongly
bioturbated alternations of mudstone and sandstone.

throughout the entire thickness of the bed or are concentrated as discontinuous streaks along laminae and basal erosive surfaces.

Interpretation

The type of stratification present in these sandstone beds is thought to be indicative of sedimentation in the offshore-transition zone: the flat lamination being the product of storm produced ebb-surge currents (Elliott, 1978) and the abundance of wave rippled cross lamination that occur between the storm generated beds being indicative of a return from storm to fair weather conditions. In other words, high energy conditions ("storms") alternated with periods of lower energy ("fair weather" conditions). The wave rippled sandstone cappings and the bioturbated alternations of mudstone and very fine sandstone possibly represent sand reworking by waning storm currents and fairweather sedimentation punctuated by minor storms respectively (Maejima, 1988). The accumulation of pebbles into the sandstones suggests effective segregation of larger clasts from sands and is interpreted to be the result of wave processes related to storms (Maejima, 1988).

FACIES 4

Description

Facies 4 (Fig. 23, pl. 5, Appendix B, p.B12) follows Facies 3 gradationally in a lateral and vertical sense, and consists of medium to coarse grained, medium to thick-bedded sandstones

Facies 4

LOC 8
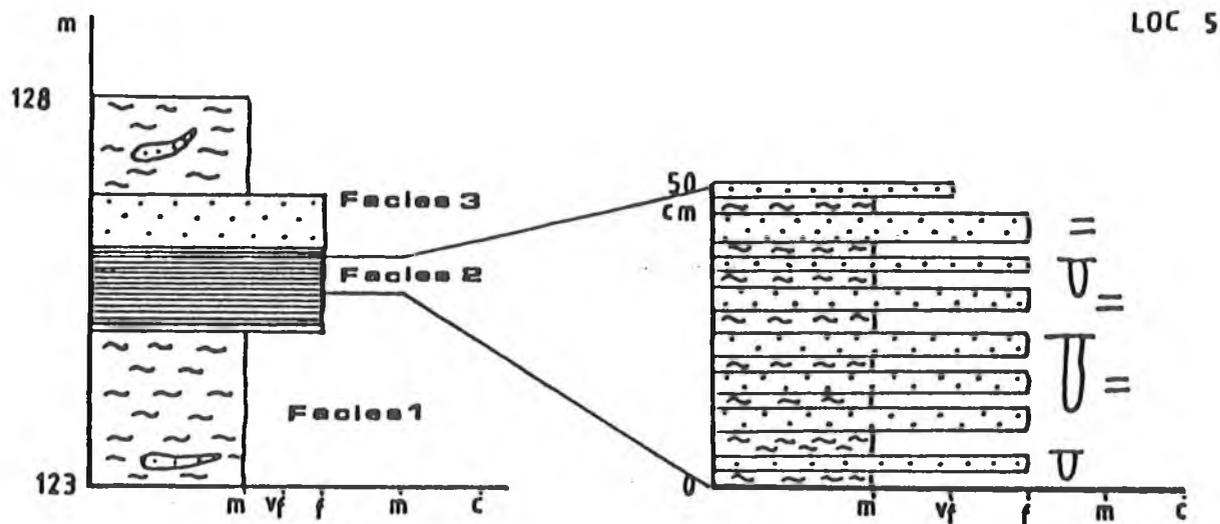

FIG. 23: Selected vertical log through Facies 4 (LOC 8,
Appendix B, p. B12). For further details see text and
Appendix B.

and bioturbated mudstones (up to 5 cm thick) (pl. 5A). These deposits are elongate, lenticular units, ungraded and poorly sorted. The sandstones contain hummocky-cross-stratification. This structure consists of low-angle, gently undulating laminations. The base of these HCS beds is usually sharp whereas the top commonly show small-scale symmetrical ripples. The HCS beds become amalgamated upward in the sequence, forming a sand body of 2-3 m thick. The top of this amalgamated sand body is very undulatory and partly reworked (wave ripple cross-stratification), whereas the base is characterized by load structures (pl. 5B).

Outsized pebbles are scattered throughout the amalgamated, sandy beds or occur concentrated as discontinuous stringers along the internal scour surfaces, forming traction carpets (pl. 5C). Some of these clasts show imbrication indicating current direction to the SE.

The mudstone intercalations which have developed between the sandstone beds are homogenized due to bioturbation (*Rhizocorallium*). Bioturbation is almost absent where there is predominance of coarse sand.

Facies 4 is common in the upper part of the type section (LOC 8) and has never been found southwards of Kalamavka (Fortuin, 1977).

Interpretation

Facies 4 is characterized by the presence of hummocky-cross-stratification. According to Harms et al. (1975, p. 87-

PLATE 5: FACIES 4

PLATE 5A: Medium to coarse-grained sandstone beds, generally
amalgamated with crudely-developed internal tractive
structures.

PLATE 5B: Load structure in Facies 4.

PLATE 5C: Outsized pebbles in a sandy bed of Facies 4.

88), the long, low, undulating hummocky-cross-stratification is produced by storm waves, below fair weather wave base. Hamblin and Walker (1979) and Wright and Walker (1981) suggested for the origin of this structure that a storm of hurricane proportion can entrain sand in very shallow water creating a density current. This density current, as originally proposed by Hayes (1967a), transport sands offshore. After the deposition of these sands under the influence of strong oscillatory flows (i.e. above storm wave base but below fair weather wave base), storm waves of the same storm which created the density current, subsequently reworked the sands creating hummocky-cross-stratification to the depth of storm wave base. As the storm abates, normal fair weather deposition of mudstone would resume (deposition of the bioturbated mudstone intercalations).

Upwards in the sequence, the HCS sandstone beds become amalgamated by the erosion of the interbedded mudstones. According to Leckie and Walker (1982), the amalgamation suggests slightly shallower water where more frequent storms were able to generate and to rework the bottom, preventing the preservation of fair weather mudstones.

The isolated pebbles which are occasionally found floating in the sandy beds may indicate that the storm currents were very erosive at some stages, deriving the clasts of the nearshore region through bed erosion and transporting them basinwards.

FACIES 5

Description

Facies 5 (Fig. 24, pl. 6, Appendix B, p.B2, B8) consists of matrix- to clast-supported, cobble to boulder conglomerates. Usually, these conglomerate beds are stacked one at the top of the other, to form composite conglomerate units up to 10 m thick. Individual conglomerate bodies show flat bases and convex-upward tops, surrounded by sandstones (lobe-shaped) (pl. 6A).

The gravel size-distribution ranges from bimodal to polymodal. Clasts are subrounded to subangular and are poorly cemented (pl. 6B). The matrix is marly to sandy and is more abundant in beds belonging to the lower part of the conglomerate units. Irregular patches of muddy matrix which are locally associated with surrounding slumping also occur (see pl. 6C).

The base of the conglomerate body consists of marls with boulders (clasts of Tripolitza limestone, laminated sandstone and conglomeratic clasts) floating in it. Marls are also disturbed by important load structures. Gastropod shells and burrows are indicative of a marine environment.

At the top, there is a wavy, parallel-laminated, wave ripple cross-laminated, sandy bed with intercalated very thin gravel layers.
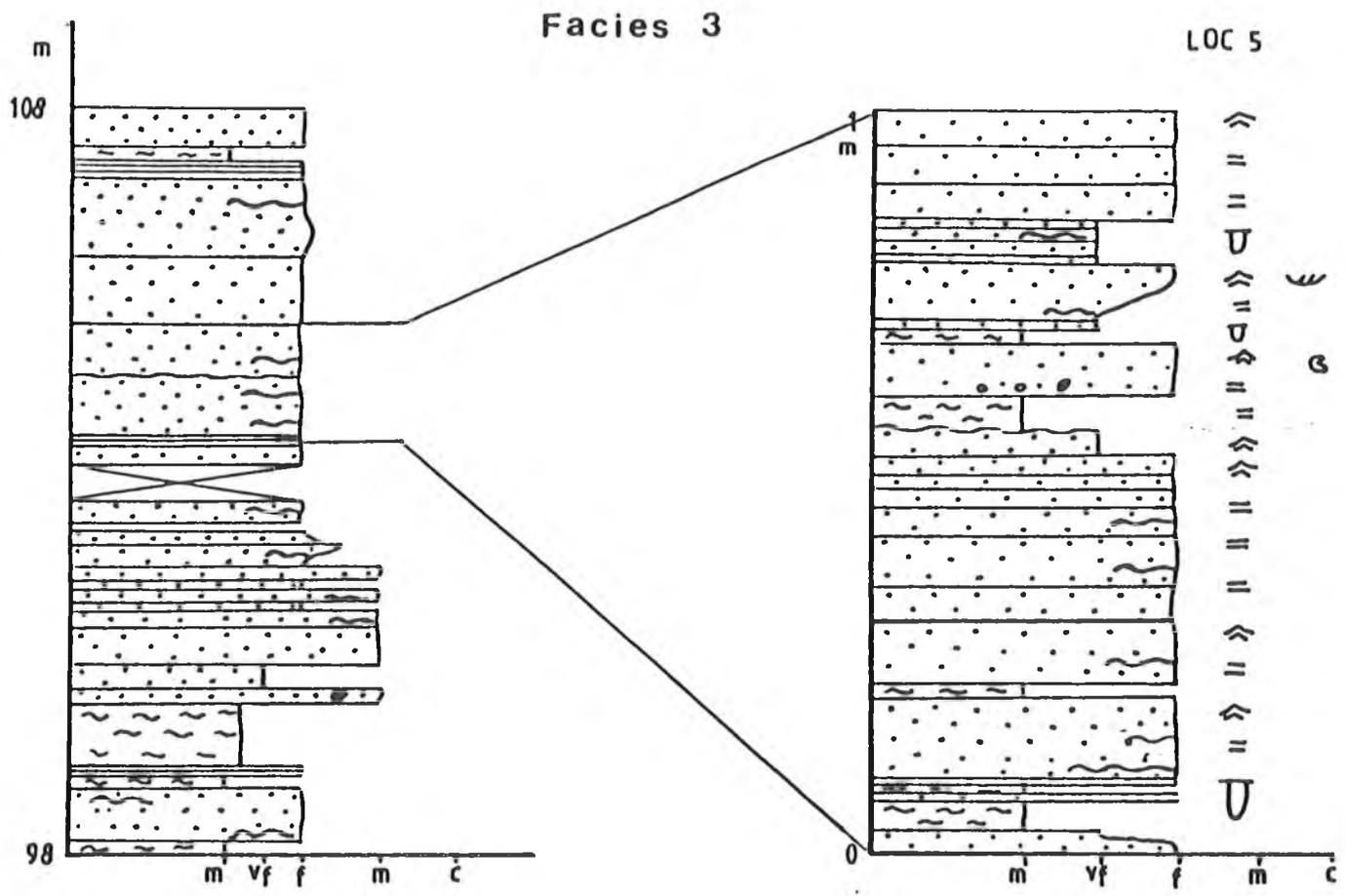
Facies 5 is only found in LOC 2 and LOC 7.

Facies 5

FIG.24: Selected vertical log through Facies 5 (LOC 7,
Appendix B, p.B8). For further details see text and
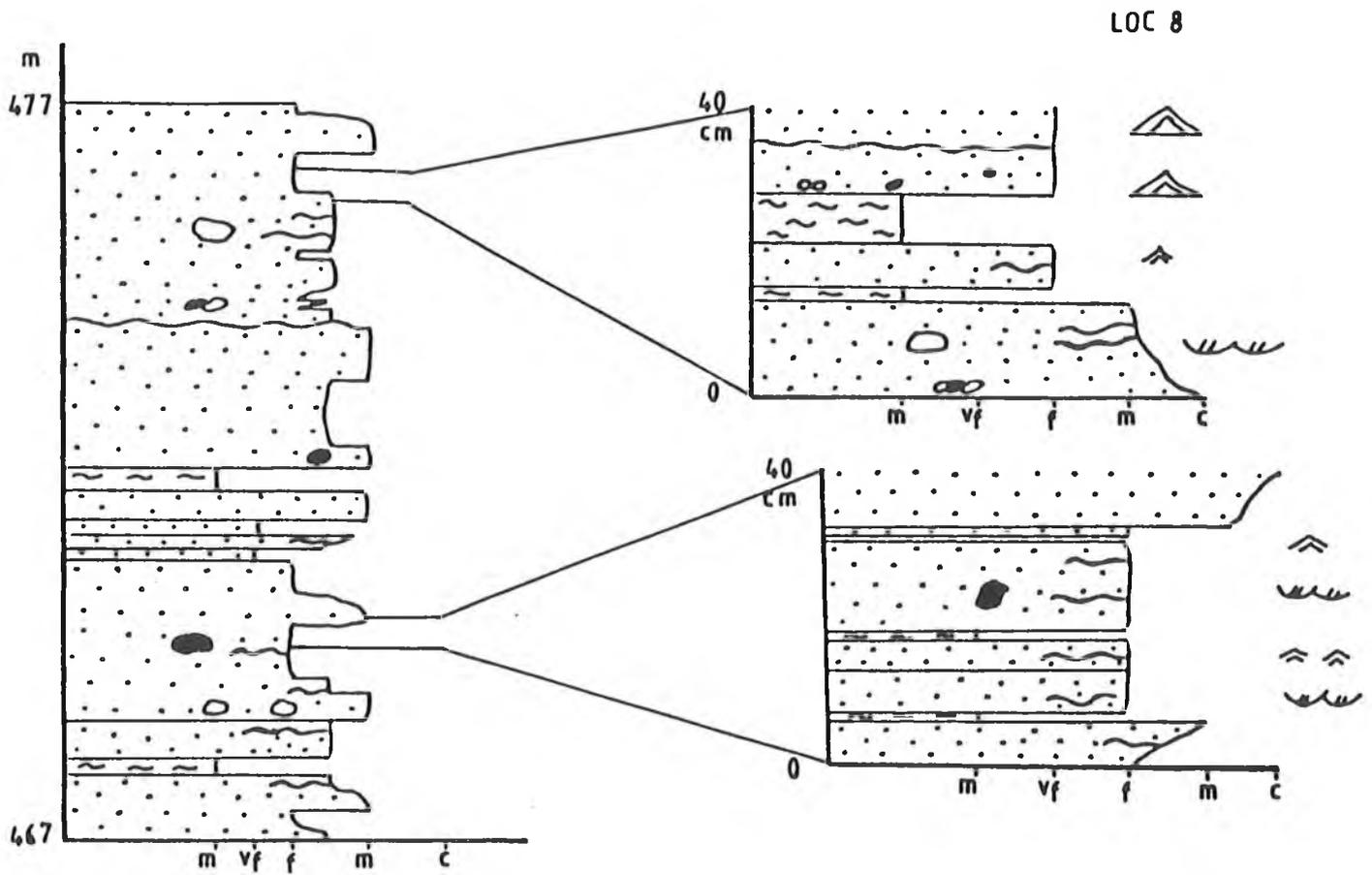Appendix B.

PLATE 6: FACIES 5

PLATE 6A: Matrix- to clast-supported conglomerate body of Facies 5.

PLATE 6B: Subrounded to angular clasts, poorly cemented.

PLATE 6C: Irregular muddy patches, associated with local slumping of the muddy sea floor.

Interpretation

The conglomerates of Facies 5 are thought to represent mass flow deposits, emplaced as cohesionless or cohesive debris flows (Nemec et al., 1980; Nemec and Steel, 1984). These mass flow deposits may have originated either from slope instabillity or flood-generated currents. The first possibility is rather unlikely as indications for a steep slope have nowhere been found. Flood-generated river currents may have been responsible for the cobble and boulder transport basinwards.

The tops of many conglomeratic beds record brief reworking which is an indication for deposition above storm-wave base.

To summarize, the vertical sequence of the Kalamavka Formation (Fig. 19, Appendix B) in its type section comprises bioturbated, mud-dominated shelf deposits (*Facies 1*, LOC 1 and LOC 2) which pass transitionally upward into shoreface deposits (*Facies 4*, LOC 8). The transition consists of couplets of offshore, very thin, flat laminated sands and mud (*Facies 2*) which resemble the storm-sand layers or rhythmites of Reineck and Singh (1972). The offshore-transition Facies is represented by up to 20 cm thick-bedded sandstones with intercalated mud layers. The sandstones are usually horizontally laminated beds which are often succeeded by wave rippled cross-laminated cappings (*Facies 3*) and grade into

amalgamated sandy hummocky-cross-stratified beds (*Facies 4*).

The whole sequence has been affected by strong bioturbation which obscures the primary sedimentary structures. Nevertheless, relicts of hummocky-cross-stratification are still preserved in the upper part of the sequence (in *Facies 4*).

Furthermore, it is clear from the vertical profile of the studied area (Fig. 19, Appendix B), that rhythmic deposition of calcareous sandstone and mudstone prevailed during the deposition of the Kalamavka Formation, yet cyclicity is not well-developed. An overall crude coarsening-upward trend can be inferred for the Kalamavka Formation based on a predominantly marly lower part (see Appendix B: LOC 1 and LOC 2) and a predominantly sandy upper part (Appendix B: LOC 8). In addition, smaller scale "cycles" (up to 10 m thick), stacked one at the top of the other, can be distinguished which comprise sequences of a marly base and a sandy top. Proceeding upwards in the section, the thickness of the marls decreases and the thickness and the grain size of the sandstone increases.

These stacked, usually coarsening-upward "cycles" are not obvious in the parts of the section where the stratigraphy is poorly exposed or shows intercalated debris flow deposits.

4.4 LATERAL CHANGES IN THE KALAMAVKA FORMATION (type section).

Generally, the Kalamavka deposits show lateral consistency (a feature of a wave dominated environment), yet some variations in the facies distribution exist in the areas adjacent to the type locality. The Kalamavka sediments of the study locality have been examined along their strike and normal to this direction (NW-SE) in order to determine any lateral variability.

Southeast of the village Kalamavka, another section has been studied which distinguishes itself from the type section by a thicker basal marly succession. The thick-bedded, coarse grained sandstone beds present North of the village Kalamavka are almost lacking here. Instead, there are only some fine grained, up to 10 cm thick sandstone beds intercalated in marly sediments. In other words, the deposits SE of the village Kalamavka are characterized by a predominantly uniform lithology which obscures any possible cyclic pattern of the sediments and apparently accounts for a thickness reduction of the calcareous sandstone sequences and an increase in the thickness of the marl intercalations. This has also been observed by Fortuin (1977).

Northwest of the village Kalamavka, the opposite situation prevails: the thick-bedded calcareous sandstone deposits are more pronounced, comprising a major part of the Kalamavka sediments. Tracing these beds laterally is quite difficult as that is hampered by faulting.

Taking into consideration the above observations, it is inferred that from NW to SE (along the strike), the average grain size of the sediments and the thickness of the sandstone beds decrease rapidly, which means that there is a proximal-distal facies transition southeastwards.

The greatest thicknesses of the sandstone beds of the Kalamavka basin are more proximal to the major E-W thrust fault (source area), whereas the thinner deposits, south of Kalamavka, have distal properties.

CHAPTER 5

DISCUSSION

5.1 DEPOSITIONAL MODEL AND IMPLICATIONS

The purpose of this section is to discuss the depositional environment and the processes which took place during the sedimentation of the Kalamavka Formation, and to compare them with those interpreted by Fortuin (1977).

According to Fortuin (1977), the Kalamavka Formation was deposited in a submarine fan environment. This interpretation was mainly supported by the depositional depth of the Kalamavka sediments (more than 100 m) and by the sedimentary structures of the thin-bedded sandstones which resemble those of turbidites.

The depositional depth of the Kalamavka Formation probably did not exceed the 200 m suggested by benthonic foraminiferal fauna (Fortuin, 1977) and ichnofacies. Besides, the observed sedimentary structures suggest deposition in rather shallow water, at least above wave base and on this basis, a submarine fan origin as suggested by Fortuin (1977), can be ruled out.

Furthermore, this thesis shows that the Kalamavka Formation sediments are usually characterized by rhythmical, turbidite-like deposits. However, they differ from real turbidites by

the presence of low-angle cross laminations (hummocky-cross-
stratification) and wave ripples, resulting from storm wave
oscillation effects as discussed earlier (Ch. 4.3). These
structures are found mostly in the proximal part of the
sequence (in *Facies 4*), and not in the distal part (*Facies 3*
and *Facies 2*), where a predominance of bioturbated, even
laminated thin sandstone beds (storm-layers) occur. ·

In other words, the rhythmical deposition of the sediments
of the Kalamavka Formation is attributed to the repeated
storm-wave action: during stormy weather, the storm situation
raises the sea level and causes erosion in the nearshore sand
deposits (Schumacher and Tripp, 1979; Cacchione and Drake,
1982; Nelson, 1982). This mechanism is characterized by a two-
layer circulatory system in which the wind-driven surface
waters move landwards whilst the bottom waters move offshore
(Fig. 25; Morton, 1981; Swift, Figueiredo et al., 1983). The
bottom currents achieve maximum velocities during the storm
rather than after, and transport sediment entrained by the
storm waves. With the decreasing energy of the storm the
transported sediment is laid down forming parallel sand
layers. When the sea level drops after the storm to assume
normal level, waves can produce ripples on the previously
deposited sandy surface (Masters, 1967; Häntzschel and
Reineck, 1968; Reineck et al., 1968; Reineck and Singh, 1971).
When the deposition of sand prevails, energy levels will be
too high for the deposition of marls. Thus, marls can be
expected basinwards, which is indeed the case (Ch. 4.4). After

FIG. 25: Illustration of the proposed mechanism for the generation of storm deposits (after Walker, 1979; Morton, 1981; Swift, Figueiredo et al., 1983)

the storm, when quiet conditions prevail, deposition of marls predominates.

To conclude, storm-related density flows seem to have been produced on a shallow gradient slope, (there are no indications of sedimentation on a steep slope such as characterice real turbidite deposits), and provided a mechanism for delivering sediment offshore (Hamblin and Walker, 1979). This sediment was deposited under the influence of strong oscillatory flows (i.e above storm wave base but below fair weather base).

5.2 TECTONIC AND SEDIMENTARY CONTROLS

The development of the Kalamavka Formation was closely related to the onset of the break-up stage of the Aegean region in the Late Serravallian - Early Tortonian times. During that period, the Ierapetra region was probably affected by compressional tectonics, which resulted in thrust faulting (formation of the E-W oriented main thrust fault, Postma, pers. comm.) and associated folding in the study area (Fortuin and Peters, 1984; Fig. 17). Due to thrusting, the northern part of the region was uplifted and denudated whereas the area in front of the thrust was gradually subsiding (probably as a result of isostatic loading), (Fig. 26).

According to Fortuin (1977; 1978), at the onset of this

crustal deformation, a thick series of breccias and
conglomerates of continental to littoral origin with clastics
supplied from nearby Tripolitza limestone to the North were
laid down (Fortuin, 1977). Further to the south, more rapid
subsidence (due to the activation of secondary thrust faults)
resulted in the deposition of 300 m of fluvio/marine
conglomerates of the Prina Complex and calcareous sandstones
and mudstones of the Kalamavka Formation.

The deposition of the Kalamavka Formation was largely
controlled by this tectonic setting. During the Late
Serravallian - Early Tortonian, thrust plate loading with
attendant flexural subsidence, plus eustatic sea-level changes
of low amplitude resulted in relative sea level rise along the
Ierapetra basin.

The subsidence which led to the preservation of the sheet-
like sandstone deposits of the Kalamavka Formation combined
with an assumed low-amplitude eustatic signal caused the
relative sea-level curve to consist of periods of (slow) relative
relative sea-level rise separated by periods of rapid rise
(Fig. 27). This variation in the rate of relative sea-level
rise without associated periods of sea-level fall can be
attributed to the fact that the eustatic fall may fail to
lower sea level significantly and it merely suffices to slow
the relative rise. The vertical stacking of the sandstone
bodies and the lack of major unconformities between them
support this idea.

FIG. 26: Diagrammatical portrayal of the development of the
depositional basin of the Kalamavka Formation in Late
Serravallian-Early Tortonian times. Faults are schematic,
wavy arrows indicate transport of eroded debris.

FIG. 27: Schematic diagram showing the relative sea-level curve as
a function of the overall tectonic subsidence due to
isostatic loading and low amplitude of absolute (eustatic)
sea-level fluctuations (after Nummedal and Swift, 1987).
The relative sea-level curve consists of periods of slow
rise separated by periods of rapid rise.

Fluctuations in the rate and extent of transgressions and regressions in the shoreline, which are themselves controlled by the rate of sediment supply and the rate of sea-level change, have significantly influenced the spatial and temporal distribution of facies on the shelf.

According to Nummedal and Swift (1987), during a fast relative sea-level rise the shoreline translates upward and landward through a process of shoreface retreat. Nearshore sediment eroded by storm currents is commonly redeposited farther seaward, e.g. as in the Kalamavka basin. The slow relative sea level rise only induces shoreline progradation by transporting coarser sand over the shelf by rivers.

These processes are repeated multiple times during the overall transgression and a series of stacked sand bodies formed. Even during the short-term regressions included in this model, the seaward limit of the prograding clastic wedge will be fairly close to the shore, resulting in a stair-step progression toward the basin margin.

In other words, during the overall transgression, high-frequency oscillations of the shoreline during the retreat caused intermittent progradation, shoreface retreat and the formation of multiple sand bodies separated by a series of non-erosive surfaces.

The above described processes, which may account for the Kalamavka Formation, would predict an overall fining-upward trend of the Formation. However, an overall (crude)

coarsening-upward trend has been observed. This coarsening-
upward trend must be attributed to shoreline progradation.
Shoreline progradation in a relative sea-level rise regime
might then mean that the local sediment supply was high enough
to cause progradation.

The local sediment supply can be mainly controlled by the
climate. Although indications for the prevailing climatic
conditions during the deposition of the Kalamavka shelf
deposits have not been found, periods of heavy rainfalls can
be inferred by the presence of the flood-generated debris flow
lobes into the sequence. During heavy rainfalls, an excess of
sediment supply is transported by the flooded rivers
basinwards. In this case the sedimentation rate exceeds the
rate of the relative sea-level rise resulting in the (crude)
coarsening-upward trend of the Kalamavka Formation.

Excess of local sediment supply can be also attributed to
periods of sea-level stillstands or minor falls. This
possibility seems more unlikely as disconformities indicative
for a punctuated transgression have been found nowhere along
the study area.

CHAPTER 6

CONCLUSIONS

A generalized depositional model for the Kalamavka Formation is illustrated in Fig. 28. The tectonic setting facies and the interpreted sedimentary pattern of the Kalamavka Formation lead to the following conclusions about the Late Serravallian-Early Tortonian sedimentation in the Kalamavka basin.

(i). The sedimentary succession of the Kalamavka Formation was deposited in front of an E-W oriented thrust fault which separated a northern uplifted part from a southern subsiding sedimentary basin.

(ii). The Kalamavka Formation deposits consist of fine-grained sediments, derived mainly from the uplifted Tripolitza limestone and partly from the resedimented conglomerates of the Prina Complex.

(iii). The presence of low-angle laminations (hummocky-cross-stratification) and wave ripples found mostly in the proximal part of the sequence and of bioturbated, even laminated thin sandstone beds found in the distal part,

LEGEND

⊞	Tripolitza Unit
⟨△⟩	Prina Complex
⟨∴⟩	Kalamavka Fm
☰	Males Fm

FIG. 28: Schematic representation of sediment facies distribution of the Kalamavka shelf depositional environment. Numbers 1 to 5 refer to facies associations described in text.

Indicates deposition of these fine grained sediments in
a storm-dominated (wind- and wave-driven) shelf
environment. After deposition, the whole succession was
intensely reworked by waves which led to the formation
of wave rippled structures.

(iv). The sedimentation of the Kalamavka Formation has
been influenced by contemporaneous tectonism with gradual
subsidence of the basin floor and superimposed eustatic sea
level changes. The observed vertical stacking of the
sandstone bodies can be attributed to gradual subsidence
due to isostatic loading and due to eustatic sea-level
fluctuations of low amplitude.

(v). The Kalamavka sedimentary succession itself
constitutes a (crude) coarsening-upward sequence with
offshore facies grading upwards into lower shoreface
sandstones. This coarsening-upward trend is thought to
indicate progradation of the coast line which can be
attributed to the excess of local sediment supply maybe
due to favourable climatic conditions.

(vi). Finally, foraminiferal data as well as ichnofacies
indicate that the depositional depth of the Kalamavka
Formation did not exceed the 200 m, that is the deposition
of the Kalamavka sediments took place in a shallow marine
(shelf) environment.

Therefore, taking into consideration (a) the sedimentary facies, (b) the facies associations and (c) the depositional environment (shelf) of the Kalamavka Formation, the "submarine fan" hypothesis introduced by Fortuin (1977) is opposed. According to Fortuin (1977, p. 86), "the Kalamavka Formation is interpreted as part of a submarine fan, which developed in the proximity of the shore after crustal downwarping of a marginal sea area. The fan accumulated on a relatively steep slope, related to a fault or flexure, which remained active throughout deposition of the sediments".

In this thesis, it has been suggested that the combination of storm-induced currents, gradual basin subsidence and sea level fluctuations can explain the facies associations of the Kalamavka Formation and their evolution in time.

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APPENDICES

APPENDIX A

BIOSTRATIGRAPHICAL AND PETROGRAPHICAL RESULTS

Schematic representation of some identified fossils.
1, 2, 3, primitive *Uvigerina selliana*
4, *Uvigerina* with micrite-filled chambers
5, *Globigerina* with micrite-filled chambers
6, thin-walled *Globigerina Foraminifera* with spar-filled
   chambers
7, *Gastropod*. The aragonite walls have been replaced by
   micrite. The voids are filled with sparite.
8, a pair of *ostracod valves* or micritic envelope.
9, *Peneroplid Foraminifera* illustrating early diagenetic
   changes within depositional environment.
10, *Gastropod*. Original aragonite wall has been inverted to
   calcite with loss of internal details. Yet original
   internal and external outlines are preserved by micritic
   sediment.

LOC	sample	m	COMPOSITION						FOSSILS		ROUNDNESS						SORTING						GRAIN SIZE		CEMENT	NOTES
			carbonate rock	chert	feldspar	quartz	Mica M	a	fossils	identification	Angular	Angular	sub-angular	sub-rounded	rounded	well rounded	very sorted	well sorted	moderately sorted	poorly sorted	v. poorly sorted	matrix inequiv.				
2	33 0 17	bd	•			•	•			?	Gastropod ?			▾						▾			Supratidal	▾ ▾		Calcite	Burrows
	35	50.6	•	/	••	••							▾						▾			••	▾ ▾		"	Compaction Calcite replacement
3	38	55.2	••			•			/		Gastropod Forams												••	▾		Carbonate mud	
	18	3.5			•				•		Bivalvement												••	▾		"	
	19	15.7	•			•	•?		/		Ligcering Gastropod		▾						▾				••	▾		"	
3	55	6.3	•	••	•••	••			?		▾							▾				••	▾		Calcite	Compaction Replacement
	8	11.9	•	/		••	•?				▾								▾			••	▾ ▾		Carbonate mud	
	9	17.1	••	•			•?		••		Gastropod vaive											••	▾			Burrows
4	1	21.3	•		••	•	•?					▾						▾				••	▾ ▾		Calcite	
	47	25.5	•	/		••	•?							▾			▾					••	▾		"	Compaction
	48	33.1	/	/		••								▾			▾					••	▾		"	Replacement / Compaction
	49	39.6	•	••	••	•			••		Gastropoda				▾			▾				••	▾ ▾		Carb. mud	
	50	50.1	•				•?							▾			▾				••	▾ ▾		Carb. mud	
4	51	5.9	•	/	••	•	•?					▾								▾		••	▾ ▾		"	"
	3	7.5	•	••		•							▾				▾					••	▾		Calcite	Replacement
	4	11.1																				••			Carb. mud	
5	6	1.0	/	••		•	•?															••	▾		Carb. mud	
	7	2.2	/	••		•			••	Forams		▾					▾					••	▾		Calcite	
7	12	3.7	••			•	•?		••	"			▾				▾					••	▾			Burrows
	13	5.1	•		•••	•				"			▾				▾					••	▾		"	Compaction
	16	9.1	•			•							▾				▾					••	▾		"	Compaction / Burrows
	45	1.0							•?	"				•						▾		••	▾		Carb. mud	
8	20	1.9	•						•?	"		▾										••	▾		Carb. mud	Dissolution surface Deformation
	41	2.63	/	••			•?											▾				••	▾		Calcite	
	32	5.2	/	••					•?								▾					••	▾		"	
	33	9.1	•	••		•	•?		•?			▾						▾				••	▾		Carb. mud	
	34	11.3	•			•	•?					▾						▾				••	▾		Carb. mud	Deformation

LEGEND	
•	abundant
/	many
••	few
•	very few

APPENDIX B

DETAILED LOGGING OF THE STUDIED TYPE SECTION

m
55

→ 88D38

48

30

20

10

1
A 0

m vf f m c
(sst)
cong.

	FACIES				
Data	1	2	3	4	5

→ 88D35

→ 88D17

FACIES

Data	1	2	3	4	5

→ 88D55

→ 88D53

→ 88D19

→ 88D18

LOC 3

Dir. 40°NE / Dip 36°

LOC 2

Dir. 40°NE / Dip 25°

LOC 1

Dir. 37° / Dip 22°

m (vf. f. m c) cong
    sst

FACIES

Data	1	2	3	4	5

Dir. 33°/Dip

88D48

88D47

88D1

88D9

88D8

m (vf f m c) cong
   ss t

FACIES

Data | 1 | 2 | 3 | 4 | 5

LOC 4
Dir N⁄Dip30°

m
88
→ 88D6
78
6 m
70
→ 88D4
→ 88D3
60
→ 88D51
→ 88D50
50
m vf f m c   cong

B5

LOC 5

Dir. 20°NE/Dip 30°

FACIES

Data	1	2	3	4	5

m

183

173

158

148

m (vf f m c) cong
     sst

FACIES

Data	1	2	3	4	5

m

227

f

f

217

f

f

→ 88D13

→ 88D12

207

10 m

193

m (vf f m c) cong
ss t

LOC 6
Dir 28°NE / Dip 30°

LOC 7
Dir 30°NE/Dip 30°

FACIES

DATA	1	2	3	4	5

LOC 8

Dir 40°NE / Dip 38°