SEDIMENTATION AND TECTONICS OF THE
PLIO-PLEISTOCENE OF CYPRUS

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DECLARATION

I hereby declare that this thesis has been composed by me, and that the work described in it is my own and was carried out at the University of Edinburgh, except where due acknowledgement is made.
The Troodos Massif, Cyprus, a fragment of Late Cretaceous oceanic crust, has been uplifted since at least Oligocene times. The post-Miocene stages of this uplift are documented by the sediments of the E-W trending Mesaoria basin, located on the north side of the ophiolite, and by Plio-Pleistocene sediments to the south.

The Mesaoria basin initially formed in the Upper Miocene in an extensional setting, when a half-graben began to develop between an emerging island to the south (the Troodos Massif) and a submerged, complex, structural ridge to the north (the Kyrenia lineament). Significant subsidence took place along growth faults in the northern part of the basin in the Pliocene. Relative uplift took place along the Troodos margin to the south, however, in association with antithetic faulting, and is documented by progradation of small, slope fan-deltas. Subsidence declined towards the end of the Pliocene, and the basin shallowed to become a narrow, sandy platform.

In south Cyprus, the dying stages of an important compressional phase, which had begun in the Miocene, are documented by Lower Pliocene silts and fan-delta facies of the Mari basin. Mid-Upper Pliocene sediments are then largely missing from this area.

Regional compression began to affect the whole of Cyprus in the Upper Pliocene-Pleistocene. Former growth faults in the Mesaoria basin were locally reactivated as reverse faults, and fine-grained sediments accumulated in a slightly deepened area ahead of the fault zone. Complimentary movement to the south caused renewed uplift of the Troodos Massif, and a large shelf fan-delta prograded into the basin. This uplift may also be documented by braided fluvial facies in south Cyprus.

Uplift declined once more, however, because fluvial facies overlying fan-delta sediments in the southern Mesaoria basin record a period of hinterland peneplanation. A decline in tectonism is also documented in the still marine northern part of the basin, where earlier deformed rocks were transgressed by shallow marine carbonates. Flooding during this relatively stable period is attributed to eustatic sea level rise.
Severe compression then drastically uplifted the Troodos Massif, and large volumes of very coarse, ophiolite-derived sediment were shed to both north and south. The Kyrenia lineament was also uplifted subaerially, but less sediment was shed from this narrow margin. The Mesaoria basin in between became entirely continental.

Uplift of the Troodos Massif, and ultimately the whole of Cyprus, can be related to underthrusting of the African plate beneath the Eurasian plate south of Cyprus. Subduction has not been steady state, however, but sluggish and episodic at best. Uplift in Cyprus has been pulsed as a result. Miocene continental collision in the convergence zone east of Cyprus (Bitlis zone) is believed to have caused locking of the plate boundary south of Cyprus in the Upper Miocene, resulting in compression south of the Troodos Massif. Extension was induced north of the ophiolite, as the subduction hinge zone subsequently migrated south. Later compression of the whole of Cyprus is attributed to collision of a microcontinental fragment with the subduction zone. Eustatic sea level fluctuations were occurring through much of this Plio-Pleistocene evolution, but their effects are difficult to separate from those of tectonic origin.

Plio-Pleistocene sediments in Cyprus thus document both the infilling of two small sedimentary basins, and the pulsed uplift of their margins, in extensional and compressional, subduction-related environments.
ACKNOWLEDGEMENTS

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Several people outside the department have also contributed to this project. Thanks go to Dr. Harold Reading, who arranged the use of departmental facilities during a stay in Oxford, Dr. John Taylor and colleagues, Drs. David Heppell and Chris Page, and in particular, Dr. Alan Lord, who never gave up on my Cypriot ostracods. Many, many thanks are also due to Rosie Filipiak for typing most of this thesis.

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husband, and Cal the cat, who always kept things in perspective. Last but not least, I must mention wee Niall, whose recent arrival has at last brought this project to a close.
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CHAPTER 1 - INTRODUCTION

1.1 Objectives and Scope

The geology of Cyprus has long been known for its excellently preserved and exposed ophiolite complex, the Troodos Massif. Rather less attention has been paid to the other, no less important, geological units that go to make up the island (Fig. 1.1). In particular, the Upper Cretaceous-Cenozoic sedimentary cover that overlies the Troodos Massif is of considerable importance, because it documents the syn- and post-emplacement history of the ophiolite. Much of the early history of the ophiolite - its generation, initial disruption and a period of rotation - has been established through studies of the older parts of the sedimentary cover, and the ophiolite itself (e.g. Pearce et al., 1984; Clube and Robertson, 1986). Since rotation, the ophiolite has experienced massive uplift, such that its summit is now 2000m above sea level. Uplift of the ophiolite was described in general by Robertson (1977), but relatively little work since has focussed on this latter phase of the evolution of the ophiolite.

On a regional scale, plate tectonic reconstructions for the East Mediterranean have concentrated on the very young neotectonic setting of the area (e.g. Jackson and McKenzie, 1984), or on its Mesozoic-Early Tertiary history (e.g. Robertson and Dixon, 1984). These latter authors noted that poorly understood, local nappé movements, uplift, and subsidence, have prevented post-Mid Tertiary plate interactions in the East Mediterranean from being clearly distinguished. Detailed studies of the Mid Tertiary-Quaternary rocks of the East Mediterranean are therefore required, in order to bridge the gap between existing models.

On sedimentological grounds, domal uplift of the Troodos Massif is of interest because the ophiolite may have acted as an erosional point source. Many large-scale facies models deal with essentially linear environments, e.g. passive continental margin or trench. In this case, a small uplifted block may have shed clastic material
Fig. 1.1 - General geological map of Cyprus, and location of study areas.
radially, and could thus have produced a series of concentric facies belts, traceable around block margins.

The aim of this project was therefore to examine the Plio-Pleistocene part of the Troodos sedimentary cover, in order to evaluate the more recent uplift history of the Troodos Massif, and assess its tectonic implications. The project follows on from earlier studies of the metalliferous sediments directly overlying the ophiolite (Boyle, 1984), and studies of parts of the Tertiary carbonate sequence that lies above (Eaton, 1987). Plio-Pleistocene sediments in Cyprus have received relatively little attention in the past, and specific objectives were:

a) to study in detail the sedimentological and structural character of this part of the Troodos sedimentary cover, in order to construct facies models and assess basin evolution;

b) to examine the tectonic implications of these studies, with respect to both the local evolution of Cyprus and the regional evolution of the East Mediterranean;

c) to compare and contrast the Plio-Pleistocene histories of different margins of the Troodos Massif.

Plio-Pleistocene sediments occur in a number of areas in Cyprus (Fig. 1.1), and investigation of all of these was considered beyond the scope of the project. Furthermore, the northern, Turkish-occupied part of the island (north of the Green Line) is inaccessible for political reasons. Exposure is best in the Mesaoria Plain, and this was chosen as the main field area (Fig. 1.1). Small outcrop areas along the central south coast were also studied, in order to compare the succession on north and south Troodos flanks, and also with a view to correlating onshore geology with offshore seismic data (see below). Plio-Pleistocene sediments of western Cyprus are currently being investigated as part of a separate project, by L. Ward.

In addition to field-based studies, offshore seismic data from south of Cyprus were available for study. These data comprise a shallow penetration, high resolution seismic survey carried out by the (then) Institute of Geological Sciences in 1978, and some deeper seismic data acquired from the Shell International Petroleum
Company, The Hague (see Fig. 1.3 for survey locations). Further seismic data from around Cyprus were also made available by Cambridge University. Seismic data were studied partly with the assistance of an undergraduate student, W. Ferrari.

As the project progressed, it became evident that there was sufficient material within the Pliocene sediments alone to fulfill the requirements of a three-year Ph.D. programme. Thus a relatively small amount of work was focussed on the Pleistocene part of the sedimentary succession. Pleistocene-Recent sediments are now themselves the subject of a separate project, by A. Poole.

1.2 East Mediterranean Plate Tectonic Setting

Cyprus lies within the geologically complex region of the East Mediterranean. The region forms part of the Alpine-Himalayan orogenic belt, which separates the Eurasian and Turkish plates from the African, Arabian and Indo-Australian plates (Fig. 1.2). Plate convergence along this belt has been exceedingly complex, involving the interactions of a myriad of microplates, as well as those of the larger plates themselves. In the East Mediterranean, convergence has incorporated the closure of a large, Palaeozoic ocean (Palaeotethys), whose suture lies in Turkey, and the opening and closing of smaller ocean basins during the Mesozoic and Cenozoic (Neotethys; Robertson and Dixon, 1984).

At present, the location of the convergence zone is well established east of Cyprus, where continental collision (Zagros) and strike-slip movement (Turkey) are taking place. To the west, subduction (Hellenic arc), and extension (Aegean Sea) are occurring. In the Cyprus area itself, however, tectonic activity is seismically poorly defined (Jackson and McKenzie, 1984), and plate boundaries are difficult to delineate bathymetrically (Fig. 1.3). Recent plate reconstructions thus show a variety of plate boundary locations and types (Fig. 1.4).

It is generally agreed, however, that some combination of subduction and strike-slip motion are taking place in the vicinity of Cyprus. Subduction probably began in Oligocene times (see section 1.3.3), but as is argued later (chapter 11), has never developed into
Fig. 1.2 - Regional plate setting of the Mediterranean Sea (double lines represent spreading centres; see Fig. 11.6 for key to other symbols).
Fig. 1.3 - East Mediterranean bathymetry and locations of seismic surveys; contours are at 500m intervals.
Fig. 1.4 - Recent plate reconstructions for the East Mediterranean, a) from Dewey et al. (1986), b) from Rotstein (1984), and c) from Dercourt et al. (1985).
a fully-fledged system with well defined trench, accretionary complex and volcanic arc. Furthermore, palaeomagnetic data from Cyprus (Clube, 1985) suggest that northward drift of Cyprus since the Late Cretaceous has broadly been in unison with that of Africa, relative motion between Africa and Eurasia (Livermore and Smith, 1984) having been largely accommodated by pre-Miocene convergence in Turkey (Robertson, in press). Cyprus has thus lain in a "pre-arc" or "incipient fore-arc" setting at least since the Oligocene, and it is in this regional plate setting that the Plio-Pleistocene evolution of the island has taken place.

1.3 Cyprus Structural Framework and Geological History

On a more local scale, Cyprus itself can be divided into a number of distinct geological provinces or terranes (Fig. 1.5; Robertson, in press), which reflect the complex interactions of a number of microplates that exist, or have existed, in this part of the Africa/Eurasia suture zone. These units are briefly described in order to illustrate the structural framework of Cyprus. The pre-Pliocene geological history of the island is also outlined, and is summarised in Table 1.1. Emphasis is placed on those units which were to play an important role in the Late Tertiary-Early Quaternary evolution of the island.

1.3.1 Troodos and Mamonia terranes

The Troodos Massif (Fig. 1.5, Table 1.1) forms the dominant geological and geomorphological feature in Cyprus. It is an ophiolite complex (Gass and Masson-Smith, 1963) of Late Cretaceous age. It presently forms a NW-SE elongated, dome-shaped mountain range, reaching +2000m at its apex, Mount Olympus. The deepest levels of ophiolite stratigraphy (plutonic complex) are exposed in the Mount Olympus area, while sheeted dykes and pillow lavas outcrop in concentric belts around the plutonic core.

The ophiolite was generated during the opening of a small Neotethyan ocean basin (the Troodos ocean), which lay along the northern margin of Gondwana (Robertson and Woodcock, 1980). Spreading ceased only a short time after its inception, however, and
Fig. 1.5 - Major structural elements and geological units, Cyprus.
<table>
<thead>
<tr>
<th>TROODOS TERRANE</th>
<th>KYRENIA TERRANE</th>
<th>MAMONIA TERRANE</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>U.Mio.</strong> - Evaps. deposited as Messinian salinity crisis affects Med.; tectonism dwindles in S Cyprus, but N of Troodos gypsum is deposited in small basins, localized by active normal faulting.</td>
<td><strong>U.Mio.</strong> - Kyrenia lineament begins to rise along growth faults; clastic input from the north diminishes and mudstone and marl deposited, before gypsum accumulates during Messinian salinity crisis.</td>
<td>Mamonía terrane amalgamated with Troodos terrane.</td>
</tr>
<tr>
<td><strong>Olig.</strong> - Deposition of chalks with increasing marl content. <strong>M.Mio.</strong> (U. Lefkara Fm.), and marls, calciturbidites and conglomerates with reef and igneous clasts (Pakhna Fm.) signal uplift of Troodos ophiolite and emergence, related to onset of subduction S of Cyprus; in M.Mio., tectonic shoaling of basement produces narrow deformation belts in S Cyprus, while extensional faulting starts to affect N Troodos.</td>
<td><strong>Olig.</strong> - Kyrenia lineament subsides drastically in <strong>M.Mio.</strong>. extensional setting; thick turbidite sequence (Kytherea flysch), derived from Turkey, deposited unconformably over lineament; active growth faulting in Mio.</td>
<td></td>
</tr>
<tr>
<td><strong>U.Cret.</strong> - Troodos microplate is detached and rotates 90°; <strong>Eo.</strong> thick sequence of pelagic chalks, now with some replacement chert (L.-M. Lefkara Fm.) deposited as Troodos crust remains in deep carbonate - depositing seas; in S Cyprus, some deformation of chalks along inferred S edge of microplate.</td>
<td><strong>Eo.</strong> - Major S-directed thrusting, associated with continental collision to N, slices Kyrenia area into a series of thrust sheets and creates elongate deformation belt; variety of syntectonic sediments shed.</td>
<td></td>
</tr>
<tr>
<td><strong>U.Cret.</strong> - Thin unit of metalliferous sediments, radiolarites, and mudstone deposited over irregular Troodos crust (Perapedhi Fm.) Crust of Troodos ophiolite generated in small, Neotethyan, ocean basin.</td>
<td><strong>U.Cret.</strong> - Metamorphosed platform subsides, and is covered <strong>L.</strong> with pelagic carbonates; bimodal volcanics extruded and breccias shed from active fault belts as Kyrenia area comes under transtensional stress, associated with Troodos microplate rotation.</td>
<td></td>
</tr>
<tr>
<td><strong>M.-U.</strong> - Carbonates brecciated and interleaved with metamorphic slivers as platform is juxtaposed with Troodos-type crust during subduction.</td>
<td><strong>U.Cret.</strong> - Bentonitic clays, radiolarites and volcanioclastics including olistostrome (Kathikas Fm.) shed from area to NW Cyprus during active strike-slip faulting, associated with rotation of Troodos microplate and juxtaposition with Mamonía rocks.</td>
<td></td>
</tr>
<tr>
<td><strong>Perm.</strong> - Shallow marine limestones and dolomites, <strong>L.</strong> deposited on gently subsiding carbonate platform.</td>
<td><strong>U.Trias.</strong> - Disrupted assemblage of continental margin lithologies (re-deposited mature sandstones and carbonates, lime mudstone and radiolarite) in tectonic contact with oceanic crustal rocks (pillow lavas, metalliferous sediments and re-deposited atoll reef limestone).</td>
<td></td>
</tr>
</tbody>
</table>
subduction and/or continental collision outwith the present Cyprus area are believed to have caused a piece of the new Troodos crust to become detached and to rotate through 90°, as evidenced by palaeomagnetic data (Clube and Robertson, 1986). Rotation about strike-slip lineaments juxtaposed the oceanic microplate against the northern, generally south-facing continental margin of the Troodos ocean. Deformed remnants of this margin were emplaced over and adjacent to the microplate, and a thick series of syn-tectonic sediments with some volcanics was deposited. These sediments and continental margin fragments are now preserved as the Mamonia terrane in south and southwest Cyprus (Fig. 1.5; see Table 1.1 for further details), and as the Kyrenia terrane in north Cyprus (see further, section 1.3.2 and Table 1.1).

In south Cyprus, rotation in part exploited a pre-existing transform fault zone, the Arakapas Fault Belt (Fig. 1.5; Clube and Robertson, 1986), now preserved along the northern margin of the Limassol Forest area (Fig. 1.5). Rocks of the northern and western parts of this block, which include a variety of tectonised plutonics and depleted extrusives, are interpreted as part of the transform zone itself (Murton, 1986). A sequence of more normal extrusives, similar to those of the main Troodos ophiolite body, occur along the southeastern side of the block (MacLeod, 1987).

Rotation and emplacement of continental margin units were completed by the Eocene (Clube and Robertson, 1986). The Mamonia terrane was by then welded to Troodos crust, and the Troodos and Mamonia terranes thereafter acted together as one amalgamated unit (Robertson, in press). The Kyrenia terrane, however, continued to act independently (see following section).

Sometime after rotation was over, probably in the Oligocene (see section 1.3.3), the Troodos ophiolite began to be uplifted. Uplift has continued ever since, and has been a major influence on the latter part of the geological history of Cyprus. The later stages of this uplift are documented in this thesis.

1.3.2 Kyrenia terrane

The Kyrenia Range in northern Cyprus (Fig. 1.5) is a complex structural lineament, comprising a variety of sedimentary,
metamorphic and volcanic rocks, of Permian-Miocene age (Robertson and Woodcock, 1986; for details, see Table 1.1). It forms a narrow, arcuate mountain range, in contrast to the Troodos Massif, up to 1000m high. Magnetic anomaly data (Aubert and Baroz, 1974) suggest that Troodos ophiolitic basement extends north only as far as the lineament, which straddles this basement to the south, and unknown basement to the north (Robertson and Woodcock, op. cit.). The lineament is known from seismic data to extend northeastwards below sea level as a submerged ridge, as far as Turkey (Mulder, 1973).

Although essentially emplaced next to Troodos basement coevally with the Mamonía terrane, the Kyrenia terrane continued to act independently for some considerable time. Following juxtaposition with Troodos crust, it underwent a Late Eocene phase of severe compression and south-directed thrusting, associated with continental collision in the Taurides to the north (Robertson and Woodcock, 1986). The main effect of this in the Kyrenia area was to produce an elongate deformation belt, comprising a series of probably near-horizontal thrust sheets, composed of Mesozoic platform carbonates and younger, deeper water sediments with some volcanics. The lineament then subsided drastically in the Oligocene-Miocene, along with the Cilicia area to the north (Fig. 1.3), and a thick sequence of turbidites (Kythrea flysch) was deposited unconformably over it (Fig. 1.5). These sediments were mainly derived from the Turkish mainland to north and northeast. Subsidence is attributed to crustal extension related to the onset of northward subduction south of Cyprus (Robertson and Woodcock, op. cit.).

Subsidence declined in the Upper Miocene. By this time, there is evidence that an important south-down growth fault became active in the area (the Kythrea fault, Fig. 1.5). The Kyrenia lineament began to rise along it, taking on the form of a submerged ridge. New evidence suggests that other growth faults may also have been active (see sections 3.4.1 and 3.4.2), and subsidence south of these documents the start of evolution of the Mesaoria basin, the main subject of this thesis.

1.3.3 Pre-Pliocene sedimentary cover

Pre-Pliocene sediments in Cyprus (Table 1.1) comprise a
sequence of marine marls and chalks, with some clastics and evaporites, best exposed on the south side of the Troodos Massif (Fig. 1.5). They document the early stages of uplift of the ophiolite. As uplift progressed, these sediments were raised along with Troodos basement, and in part provided source material for Pliocene and younger deposits. They are briefly described in the following.

The oldest part of the Troodos sedimentary cover comprises a thin unit of Late Cretaceous, metalliferous sediments and radiolarites (Perapedhi Formation), which passes up into a thick sequence of Latest Cretaceous-Eocene pelagic chalks with replacement cherts (Lower-Middle Lefkara Formation; Robertson, 1977). These sediments attest to the relatively deep marine setting of the Troodos microplate before and during rotation.

The Upper Lefkara Formation (Oligocene-Mid Miocene) also comprises chalks, but is more marly and bioturbated than its lower part, contains clay bands, and in places passes into benthic foraminifera-rich, calcareous shales (Robertson, op. cit.). These facies changes suggest a period of shallowing, and are believed to signal the onset of uplift of the Troodos Massif.

Lefkara sediments are succeeded by the Mid-Upper Miocene Pakhna Formation, which contains a variety of facies, including marls, calciturbidites with neritic carbonate detritus, reefal material (Koronia Limestone member), and local conglomerates with Troodos-derived clasts. They point to continued shallowing of the Troodos Massif, and for the first time, local emergence above sea level. In south Cyprus, these sediments are partially deformed, together with the underlying Lefkara Formation, along at least two narrow deformation zones, the Yerasa fold and thrust belt and Akrotiri High (Fig. 1.5; Morel, 1960; Eaton, 1987), which involve folding and high-angle thrusting. These thrust lineaments are believed to represent tectonic shoaling of the Troodos Massif in response to underthrusting to the south (Robertson, in press).

In contrast, on the north side of the Troodos Massif, sediments of the Pakhna Formation record the onset of a phase of normal faulting, which controlled thickness and facies variations (Follows and Robertson, in press). Uplifted fault blocks became the site of coral reef development, and large volumes of marl, chalk breccia and
reef talus accumulated in flanking depressions.

Miocene sedimentation culminated in Cyprus with evaporite deposition (Kalavasos Formation), as the well-known Messinian salinity crisis affected the whole of the Mediterranean (Hsu et al., 1978). Deformation largely ended in southern Cyprus, although new evidence (see chapter 9) suggests its waning stages lingered into the very Early Pliocene. North of Troodos, evaporites document deposition in an area that continued to be unstable (Follows and Robertson, in press):

1.3.4 Summary

In summary, prior to the Pliocene, subduction is inferred to have been initiated in the Cyprus area, along a trench to the south of the island. In response, the Troodos and amalgamated Mamonia terrane began to be uplifted. The Troodos Massif shoaled tectonically in the south in the Miocene, while strong crustal extension to the north caused the Kyrenia-Cilicia area to subside. These events declined in the Upper Miocene, however, by which time the Troodos Massif was starting to emerge. At the same time, the Kyrenia lineament began to rise, while extensional faulting affected the northern flank of Troodos. In between, the infant Mesaoria basin began to subside.

1.4 Previous Work

Little detailed work has been carried out on the Plio-Pleistocene succession of Cyprus. Generally rather brief lithological and stratigraphical descriptions, particularly of the Pliocene section, have been given in a number of publications, but little sedimentological or structural detail is available. Gaudry (1862), Russell (1882) and Bellamy and Jukes-Brown (1905) were the first authors to describe Cyprus sediments. Following palaeontological work by Reed (1929) and Ovey (1937), Henson et al. (1949) provided the first comprehensive account of the island's geology.

Between 1959 and 1967, the Cyprus Geological Survey published its eight geological memoirs and accompanying maps, covering the central part of Cyprus. In 1964, the only detailed study of
Plio-Pleistocene sediments was undertaken by Ducloz in the Mesaoria Plain (Ducloz, 1965).

More recent palaeontological studies have been carried out by Mantis (1968, 1970, 1975) and Baroz and Bizon (1974, 1977). Baroz (1979) gives a more detailed account of the previous two references and, together with Robertson (1977), provides palaeogeographic and tectonic reconstructions of Plio-Pleistocene Cyprus geology. Cleintaur et al. (1977) used borehole and geophysical data to study subsurface geology and produced a Neogene structural map of the island. Some Cyprus Geological Survey borehole records are published in Zomenis (1972) and Hadjistavrinou and Constantinou (1977). The Survey also published the most recent geological map of the island (1979). Miocene sediments in southern Cyprus have been described recently by Eaton (1987), while preliminary results of studies of Miocene sediments on the northern flank of Troodos are presented in Follows and Robertson (in press).

The geology of parts of the Kyrenia area is described by Baroz (1979), while a comprehensive review of the area is given by Robertson and Woodcock (1986). Its geomorphology, along with that of the Mesaoria and Troodos, are described by de Vaumas (1959, 1961).

Geophysical studies in Cyprus have included a gravity survey (Gass and Masson-Smith, 1963), aeromagnetic surveys (Vine et al., 1973, and Aubert and Baroz, 1974), and palaeomagnetic studies (Moores and Vine, 1971; Clube, 1985, and others). References to offshore seismic surveys around Cyprus are given in chapter 10.

The tectonic evolution of Cyprus has been discussed in many papers (see references in Clube and Robertson, 1986), and is summarised in Robertson (in press). The neotectonic setting of the island is also the subject of many publications, which are summarised in Kempler and Ben-Avraham (1987).

1.5 Methods Used

Field-based studies, which provide the major data base for this project, were carried out during three field seasons in Cyprus, totalling six months. Mapping was carried out on a reconnaissance
basis initially, using 1:50,000 scale maps. Detailed mapping of several areas was undertaken on a 1:5,000 scale. 1:30,000 geological maps of the Cyprus Geological Survey were used as a guide to mapping, as was the more detailed map of Ducloz (1965).

Sedimentological logging on several scales was carried out, depending on the detail required and the amount of continuous exposure. Standard descriptive terms for lithology, texture, sedimentary structures and other features, as given e.g. in Tucker (1981), were used in logging and are used throughout this thesis. Because of low tectonic dips, palaeocurrent data did not require to be corrected. Lithological and palaeontological samples of selected intervals were collected.

Macropalaeontological determinations were carried out with the help of Dr D. Heppell (Royal Scottish Museum), Dr J. Taylor and colleagues (British Museum), and Dr C. Page (Royal Botanical Gardens, Edinburgh). Dr A. Lord (University College, London) identified both foraminifera and ostracods. These experts also provided palaeoecological information.

Standard laboratory techniques were used to investigate petrographic and diagenetic properties of selected samples. As the project was strongly field-based, these analyses were not intended to be exhaustive, but to focus on particular problems. Standard thin-sectioning of many samples was carried out mainly in Edinburgh, although a small number of sections were made by BP in Aberdeen. Because of their friable nature, all samples were first impregnated with resin, which in most cases was coloured blue. All sections were half-stained to test for carbonate, using the method of Dickson (1966). Point counting of selected sections was undertaken for provenance studies. Over 330 points were counted in each case. Samples were selected to give a reasonable lateral and vertical distribution. A small number of polished thin sections were made for CL (cathodoluminescence) studies. Owing to time constraints, however, only a pilot study could be undertaken, and only general observations made.

XRD analysis of fine-grained fractions was utilised for mineralogical studies. Samples were first crushed with pestle and mortar and sieved to remove sand and coarse silt. Carbonate was
removed by standing in 10% HCl or 30% acetic acid for 24 hours and rinsing several times. Samples were then mounted on glass slides with acetone, and allowed to dry. Glycolation and heating to 500°C for 2 hours were carried out to aid in smectite and kaolinite identification. CuKα radiation was used throughout.

A few fine-grained samples were selected for SEM work. Ostracods were also picked directly from semiconsolidated mudstones, and photographed. In all cases, samples were mounted on stubs with conductive wax or double-sided sellotape and Silverdag applied to sample edges, prior to gold-coating. A Cambridge 2000 SEM was used.

The methods used in the seismic part of the project are outlined in chapter 10.

1.6 Thesis Organisation

This thesis is divided into three parts - Introduction, Plio-Pleistocene sediments and their tectonic implications, and Conclusions. This chapter and the following one on stratigraphy (chapter 2) comprise Part I.

Part II begins with a few introductory comments. Succeeding chapters give detailed descriptions, interpretations and models for each of the formations recognised during the study, and are arranged chronologically (chapters 3-7). Petrography and diagenesis of the sediments from the main field area (Mesaoria Plain) are discussed in chapter 8. Sediments from the minor field area (south Cyprus) and the seismic study are described in chapters 9 and 10.

The findings of the previous chapters, their tectonic implications, and the Miocene-Quaternary tectonic evolution of the Cyprus area are discussed in the first chapter of the final part of the thesis (chapter 11). The major conclusions of the project are then listed in chapter 12.
2.1 Background

Plio-Pleistocene stratigraphy of Cyprus is rather unclear, mainly due to a lack of detailed study of the succession, and a lack of island-wide correlation. A number of conflicting stratigraphies exist (Table 2.1), stemming from studies in different parts of the island (Fig. 2.1). Most of these stratigraphies are based on that of Henson et al. (1949; left-hand column, lower table, Table 2.1), who published the first comprehensive review of Cyprus geology. Henson et al. described, and gave type localities for, three Pliocene formations, the Myrtou Marl, the Nicosia Formation and the Athalassa Formation. Brief comments were also made on Pleistocene-Recent deposits. The later stratigraphies, which were largely the result of mapping by the Cyprus Geological Survey, subdivided the Pleistocene part of the succession into a number of fanglomerate and river terrace deposits, or marine terrace deposits (plus or minus fanglomerate), depending on the location of the study (inland or along the coast, respectively). In addition, not all of the Pliocene formations of Henson et al. (1949) were recognised.

The most detailed stratigraphic study was carried out by Ducloz in 1964, in the Mesaoria Plain (Fig. 2.1), where Plio-Pleistocene facies are best exposed. Ducloz (1965) defined a number of new formations and informal units (upper table, Table 2.1), and gave type localities. His stratigraphy was not adopted in later studies, however, and was not used in the most recent geological map of Cyprus (Cyprus Geological Survey, 1979, lower table, Table 2.1). During field studies by this author, however, Ducloz' Pliocene-Early Pleistocene stratigraphy was found to stand up well, and to provide a sensible framework in which to describe this part of the Troodos sedimentary cover in the Mesaoria Plain. Ducloz' stratigraphy was therefore adopted in this study, although modifications are proposed (described in the following section).

In south Cyprus, the Plio-Pleistocene sequence is much more poorly exposed, and parts of it are missing. No Myrtou Marl or Nicosia Formation was recognised by the two main workers in the
Fig. 2.1 - Locations of previous stratigraphical studies in Cyprus, relevant to this project.
Table 2.1 - Existing Pliocene-Recent stratigraphy of Cyprus

<table>
<thead>
<tr>
<th></th>
<th>NORTH CYPRUS</th>
<th>MESAORIA PLAIN</th>
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<tbody>
<tr>
<td></td>
<td>Moore (1960)</td>
<td>Baroz (1979)</td>
</tr>
<tr>
<td>REC.</td>
<td>River and marine terraces</td>
<td>Recent beaches</td>
</tr>
<tr>
<td>PLEISTO.</td>
<td>Fanglom.</td>
<td></td>
</tr>
<tr>
<td></td>
<td>upper Ath. Fm.</td>
<td>Karka Fm.</td>
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<tr>
<td></td>
<td>lower Ath. Fm.</td>
<td>Kak. Fm.</td>
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<tr>
<td>PLIO.</td>
<td>Nicosia Fm.</td>
<td>Myrtou Marl</td>
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<td></td>
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<td>Potami Fm.</td>
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<td>MIO.</td>
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<tr>
<th></th>
<th>SOUTH CYPRUS</th>
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<tbody>
<tr>
<td>REC.</td>
<td>Alluvium, recent deps.</td>
</tr>
<tr>
<td>PLEISTO.</td>
<td>Fanglom., marine terraces</td>
</tr>
<tr>
<td>PLIO.</td>
<td>Athalassa Fm.</td>
</tr>
<tr>
<td>MIO.</td>
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</table>

* see also Table 2.2
Ath.=Athalassa
Kak.=Kakkaristra
Fanglom.=Fanglomerate
R.B.=Raised beach
= interdigitation
= unconformable contact
area, Bagnall (1960) and Pantazis (1967) (lower table, Table 2.1; Fig. 2.1). With the aid of new micropalaeontological data (section 2.3), however, the Nicosia Formation is now recognised, a modified stratigraphy is proposed, and a tentative correlation with the Pliocene-Early Pleistocene stratigraphy of the Mesaoria Plain made. A full correlation of the marine and fluvial terrace deposits of south Cyprus and river terrace/fanglomerate deposits of the Mesaoria has not been attempted. This is now the subject of a separate study by A. Poole (Edinburgh University).

2.2 Existing Stratigraphy and Proposed Modifications

Proposed modifications to the existing Plio-Pleistocene stratigraphy are based on field and palaeontological investigations undertaken during the present study. In general, existing formations were retained where possible, and new type localities only proposed where old ones are now obscured or inaccessible. One new stratigraphic unit, identified in south Cyprus, is proposed. The complete, revised stratigraphy is summarised in section 2.4 and Table 2.2.

2.2.1 Myrtou Marl, Nicosia Formation and Athalassa Formation

The Myrtou Marl (Henson et al., 1949) is the oldest Pliocene stratigraphic unit in Cyprus. Its main lithology is a thick sequence of green-grey to brown, homogeneous, fossiliferous, calcareous silts. The formation crops out widely in central and northern Cyprus, and has been recognised by most workers. It lies with marked unconformity on Miocene and older rocks. Its relationship with the overlying Nicosia Formation is less clear. The Nicosia Formation is described by most workers as a formation separate from the Myrtou Marl, comprising a series of yellowish, fossiliferous, calcareous sandstones and calcarenites, and finer, silty facies. It was mapped only in the northern Mesaoria by Henson et al. (1949), but was subsequently recognised in the central and southern Mesaoria by Bear (1960), Gass (1960) and Ducloz (1965; Fig. 2.1), and in northern Cyprus by Moore (1960) and other Cyprus Geological Survey workers (notes in Geological Survey annual reports). It was not recognised
Table 2.2 Revised Plio-Pleistocene stratigraphy used in this study

<table>
<thead>
<tr>
<th></th>
<th>N. CYPRUS (KYRENIA RANGE)</th>
<th>MESAORIA PLAIN</th>
<th>SOUTH CYPRUS</th>
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<tbody>
<tr>
<td><strong>PLEISTO.</strong></td>
<td>Fanglom. and marine and river terrace deposits</td>
<td></td>
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</tr>
<tr>
<td></td>
<td>Karka Fm.</td>
<td>Apalos Fm.</td>
<td>Vasilikos Fm.</td>
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<tr>
<td></td>
<td>upper</td>
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<tr>
<td></td>
<td>lower</td>
<td>Athalassa Fm.</td>
<td>Kak. Fm.</td>
</tr>
<tr>
<td><strong>U.PLIO.</strong></td>
<td>Nicosia Fm.</td>
<td>Nicosia Fm.</td>
<td></td>
</tr>
<tr>
<td><strong>M. PLIO.</strong></td>
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<td><strong>L. PLIO.</strong></td>
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<td></td>
<td></td>
</tr>
<tr>
<td><strong>MIO.</strong></td>
<td></td>
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not to scale after Moore (1960), Ducloz (1972), and Baroz (1979) this study, and after Ducloz (1965) this study

Kak. Fm. = Kakkaristra Formation
Fanglom. = Fanglomerate
——— = unconformable contact

* Mesaoria Plain stratigraphy drawn to scale, using maximum formation thicknesses
in the southwest Mesaoria by Wilson (1959) or Carr and Bear (1960; Fig. 2.1). The formation was considered to overlie the Myrtou Marl gradationally (Henson et al., 1949; Moore, 1960), or to be partly laterally equivalent to the Myrtou Marl (Bear, 1960; Gass, 1960; Baroz, 1979). Ducloz (1965) found no clear stratigraphic break between the two formations, and proposed that the Myrtou Marl be considered as the lower part of the Nicosia Formation. The Cyprus Geological Survey adopted this on their 1979 geological map of the island. Baroz (1979) found the Myrtou Marl in the southern Mesaoria to be much sandier than that of the type locality at Myrtou, in northwest Cyprus (Fig. 2.1). He proposed a new formation name, the Potami Formation, for the combined Myrtou Marl and Nicosia Formation in the southern Mesaoria (upper table, Table 2.1).

The upper boundary of the Nicosia Formation is also problematical, particularly in the east central Mesaoria, where its calcarenitic facies chiefly occur, and where it is overlain by the Athalassa Formation. The Athalassa Formation comprises a series of calcitic sandstones, calcarenites and finer facies, notable for their similarity to the calcareous facies of the Nicosia Formation (Henson et al., 1949; Ducloz, 1965; Baroz, 1979). Henson et al. and Ducloz recognised a small angular unconformity between the two calcarenitic series, as did Moore (1960) and Lytras (1963) in northern Cyprus (Fig. 2.1). Baroz (1979), however, recognised no unconformity, and Gass (1960), in the south central Mesaoria, in contrast to Ducloz (1965), could not distinguish the Nicosia and Athalassa Formations, mapping them together as one unit.

In this study, no separation is made between the Myrtou Marl and the Nicosia Formation. Although lithologically different, the sandy Nicosia facies always overlie the silty Myrtou facies, and the contact is always gradational. The two are described together as the Nicosia Formation. An informal distinction is made between the silty, lower Nicosia Formation, which also contains some coarse intervals, the middle Nicosia Formation, which contains sandy intercalations, and the thin, sandy, upper Nicosia Formation (see further, Fig. 3.10 and chapter 3). Although the type locality for the Myrtou Marl could not be visited (it lies to the north of the Green Line), separate formation names for the silts of that locality and
those of the southern Mesaoria seem unwarranted on the basis of only a moderate facies change. The Potami Formation of Baroz (1979) is therefore not adopted here.

The problems of the upper boundary of the Nicosia Formation have not been entirely resolved, because the outcrop area of the angular unconformity between the Nicosia and Athalassa Formations now lies either north of the Green Line, or has been obscured by the expansion of the city of Nicosia. In the area south of Nicosia, in agreement with Gass (1960), no angular unconformity was found between the Nicosia and Athalassa Formations. It was also found that the first occurrence of calcarenitic horizons in this area approximately correlated with the top of the Nicosia Formation to the west (described later). Thus all calcarenitic rocks in the east central Mesaoria are included in the Athalassa Formation in this study, and the base of the Athalassa Formation is taken at the base of the first calcarenite band. It is accepted, however, that in the area of Nicosia itself and further north, towards the northern margin of the Mesaoria basin, a calcarenitic facies of the Nicosia Formation may occur, and that an angular unconformity may be present in this more marginal setting. Such a partial unconformity is present along the southern margin of the basin, where the Nicosia Formation is overlain by a lateral equivalent to the Athalassa Formation, the Kakkaristra Formation (see following section).

Calcarenitic rocks were mapped as Nicosia Formation in the central Mesaoria by Bear (1960; Fig. 2.1). They were later shown as Athalassa Formation on the 1979 Cyprus Geological Survey map of Cyprus. Bear noted that these calcarenites were rather different to those of the Nicosia type locality (now under Nicosia city). The results of this study show that these sediments do not belong to the Nicosia Formation, but to the overlying Kakkaristra and Apalos Formations (not recognised by Bear). The Nicosia Formation does not therefore include any calcarenitic facies at or near its top in this study.

In northwestern Cyprus, the Athalassa Formation unconformably overlies the Nicosia Formation and is apparently thicker than in the eastern Mesaoria (Moore, 1960; left hand column, Table 2.2). It comprises two members: a lower member, composed of fine-grained
silty facies with minor conglomerate, and a thinner, upper member, comprising lithologies typical of the Athalassa Formation of the east central Mesaoria (calcarenites and calcitic sandstones). Soil horizons and brecciated calcarenite horizons are also recorded, however (Moore, op. cit.). Moore’s stratigraphy for northwestern Cyprus is used in this study.

2.2.2 Kakkaristra and Apalos Formations and Fanglomerate

The Athalassa Formation was not recognised by Wilson (1959), Carr and Bear (1960), or Bear (1960) in the central and southwestern Mesaoria (Fig. 2.1). These workers believed the Nicosia Formation to be overlain by a thick Fanglomerate series (Table 2.1). Bear (1960) divided this unit into a marine part, comprising sands and conglomerates with rounded clasts, and a laterally equivalent to slightly younger continental part, comprising sands, silts and coarser, more angular conglomerates. Ducloz (1965) also did not record the Athalassa Formation in the west central Mesaoria, but recognised a lateral equivalent to it, a new formation which he called the Kakkaristra Formation (upper table, Table 2.1). This formation comprised a highly variable sequence of conglomerates and sands, with minor calcareous sands and muds. Its rapid facies changes may have contributed to its previous lack of recognition. This author, in accordance with Ducloz, however, found the Kakkaristra Formation to be a distinct and mappable unit. The formation is therefore included in this study, though some modifications to its top and base are proposed.

Firstly, Ducloz recorded an angular unconformity everywhere at the base of the formation. In this study, the relationship between the Kakkaristra and Nicosia Formations was found to be complex (section 4.2). The contact is unconformable in the southwest, but becomes increasingly conformable to north and east. A partial unconformity only, therefore, is recognised at the base of Kakkaristra Formation in this study. In the central Mesaoria, Ducloz (1965) included two conglomerate units at, and near the top of, the Nicosia Formation. Re-evaluation of the area shows that the two conglomerates are lateral equivalents, and should be included at the base of the Kakkaristra Formation.
Secondly, according to Ducloz (1965), the top of the Kakkaristra Formation is marked by a lacustrine limestone, or where "dull, gray, silty marls" are replaced by "brown silty clays" of the overlying formation. This contact could not be easily traced, and a new top to the formation is proposed. The upper contact is taken at the top of a thin, but prominent, ostracod-bearing mudstone, or the occurrence of the first prominent white caliche horizon in the succession. These two markers are laterally equivalent and are typically overlain by pink- or red-brown, fluvial muds and silts of the Apalos Formation. The new markers occur below Ducloz' top to the Kakkaristra Formation. The Kakkaristra has therefore been reduced in thickness, but its fan-deltaic facies are now uniformly overlain by the fluvial Apalos Formation.

A complication occurs in the south central Mesaoria, where the Kakkaristra and Apalos Formations are only patchily exposed. In this area, the Kakkaristra occurs as a fluvial equivalent to its deltaic facies to the north (Ducloz, 1965; this study; see chapter 4), and is largely indistinguishable from the overlying, fluvial Apalos Formation (see chapter 6). As no lithological means of separating the two formations was identified in this area, and in the absence of palaeontological data, the lower 12m of these fluvial facies are assigned to the Kakkaristra Formation (12m is the average thickness of the formation).

Ducloz (1965) describes the most typical lithology of the Kakkaristra Formation as a fine-grained greywacke. The term greywacke implies poor sorting and a certain matrix content. In this study, typical Kakkaristra sandstones were found to be well sorted and often matrix-free. The Kakkaristra Formation is mapped as part of the Myrtou Marl, Nicosia Formation, continental and marine Fanglomerate on maps accompanying the first four Cyprus Geological Survey memoirs (Wilson, 1959; Carr and Bear, 1960; Bear, 1960, Gass, 1960).

According to Ducloz (1965), the Kakkaristra Formation is conformably overlain by the Apalos Formation. This comprises a series of reddened, fluvial muds and silts, with caliche and some conglomerate. Previous workers mapped the Apalos Formation as continental or marine Fanglomerate. Again in agreement with Ducloz,
the Apalos Formation was found by this author to be a distinctive and mappable unit, and the formation is included in this study. A minor, but important, lithological component of the formation, not mentioned by Ducloz, but present at the type locality, is a skeletal carbonate-bearing sandstone.

The relationship between the Apalos Formation and the lateral equivalent to the Kakkaristra Formation, the Athalassa Formation, is conformable according to Ducloz (1965). The contact is no longer exposed, again due to the expansion of Nicosia city. The greater average thickness of the Athalassa Formation (50m), however, when compared with that of the Kakkaristra Formation (ca. 12m), suggests that the upper part of the Athalassa Formation and lower part of the Apalos Formation are also laterally equivalent (see central column, Table 2.2). This is supported by sedimentary evidence (see chapter 6), although the zone of interdigitation between the two formations is barely exposed.

A very coarse, poorly sorted, angular conglomerate sheet occurs at the top of the Apalos Formation. This conglomerate is typically found as a cap to mesa-type hills in the Mesaoria Plain, and overlies not only the Apalos Formation, but the Nicosia, Athalassa and Kakkaristra Formations as well. In this study, the term Fanglomerate is restricted to this conglomerate, which unconformably overlies all older formations. The Fanglomerate was divided into two units by Ducloz (1965), the Kantara and Kambia gravels, on geomorphological grounds. Later sedimentation events in the Mesaoria Plain cut down through the Fanglomerate and older formations, producing a series of terraced river courses (Laxia gravel, Xeri alluvium and Recent deposits of Ducloz; Table 2.1). The Fanglomerate forms the last unit of Plio-Pleistocene stratigraphy to be included in this study, so these later fluvial deposits (collectively termed Young sediments on Encl. A) have not been studied.

In northern Cyprus (Kyrenia Range), the Kakkaristra and Apalos Formations have not been recognised. A number of units younger than the Athalassa Formation are recorded, but their inter-relationships are uncertain due to very patchy exposure. Moore (1960) reported a Fanglomerate unit and younger river and marine terrace deposits in western Kyrenia. In the central Kyrenia
Range, Ducloz (1972; Fig. 2.1) identified a series of fluvial terraces on the south side of the Range and equivalent marine terraces on the north side, which he correlated with his Fanglomerate and younger series of the Mesaoria Plain. He also recognised an older unit, the Karka, which comprised a series of intramontane fossil talus deposits and breccias, and lacustrine sediments. He correlated the lacustrine facies with the Apalos Formation. Baroz (1979) gave the Karka formation status, though correlated it with the Kakkaristra Formation. As the area could not be visited, the relationship between the Karka, Kakkaristra and Apalos Formations remains unclear.

2.2.3 South Cyprus

In south Cyprus, green-grey silts and overlying sandier facies with distinct conglomerate horizons crop out at a few localities, principally near Mari (Fig. 2.1). These sediments were mapped as Athalassa Formation by Bagnall (1960), and as Pakhna Formation (Miocene) and Athalassa Formation by Pantazis (1967) (lower table, Table 2.1). New micropalaeontological data (section 2.3), however, show the sediments to be Lower Pliocene in age, and therefore correlatable with the Nicosia Formation, which they also resemble lithologically. They are therefore assigned to the Nicosia Formation in this study, and no Athalassa Formation is recognised in the area of southern Cyprus covered in this study (right hand column, Table 2.2). The top of the Nicosia Formation is not exposed and is missing due to erosion.

No Kakkaristra, Apalos or Karka Formation (see previous section) or any lateral equivalents, have been recognised previously in southern Cyprus. A new formation has, however, been identified by this study. The new formation, named the Vasilikos Formation, forms a coherent unit, which unconformably overlies, and cuts down into, green-grey silts of the Nicosia Formation. It is exposed in only a very small area, in a quarry on the coast south of Mari, where the type locality is proposed. The formation is named after the Vasilikos river. The lower contact is concave-up in shape. The formation itself is 20m thick, and comprises a series of coarse, poorly sorted conglomerates, pebbly sands, and aeolian cross-bedded sands. It is
formally described in section 2.4.

Like the Apalos Formation, the Vasilikos Formation is capped by a very coarse, angular conglomerate sheet. This conglomerate correlates on geomorphological grounds with a coarse conglomerate sheet overlying the Nicosia Formation slightly to the north, and these sediments are informally termed "Older River Terrace deposits" in this study. They do not constitute a beach deposit, as suggested by Bagnall (1960), who mapped them as a 120' raised beach. Bagnall correlated this raised beach with a Fanglomeraue unit, which he recognised in the northeast of his mapping area (Fig. 2.1; lower table, Table 2.1). This correlation was difficult to verify due to the lack of continuous exposure, although it is tentatively accepted (see further, section 9.4). Pantazis (1967) recognised two higher raised beaches (120' and 250') and a younger Fanglomeraue unit (Table 2.1), but the Fanglomeraue unit is not marked on his map, and the presence of his two raised beaches was not confirmed.

A lower raised beach (40') is recognised by both Bagnall (1960) and Pantazis (1967), and is confirmed by this study, although it occurs at variable heights above sea level. Some of the facies of the 40' raised beach, mapped by Bagnall (op. cit.) and Pantazis (op. cit.) are fluvial, and are separated in this study from coastal facies, as "Younger River Terrace deposits". They may be both equivalent to, and younger than, the coastal facies. Present-day, terraced river courses, and other areas mapped as recent sediments by Bagnall and Pantazis were not investigated in this study.

The Nicosia Formation is Lower Pliocene in age (see following section). The age of the Vasilikos Formation is unknown, and its correlation with Mesaoria Plain stratigraphy uncertain. It may however correlate with the Kakkaristra and Apalos Formations (Table 2.2). The Kakkaristra Formation, like the Vasilikos, unconformably overlies the Nicosia Formation (at least close to the Troodos Massif), and is coarser-grained than the Nicosia Formation. Both the Kakkaristra and Vasilikos Formations record post-Lower Pliocene uplift of the Troodos Massif (chapters 4 and 9), and it is not unreasonable to assume that they record the same pulse of uplift. Both the Kakkaristra/Apalos and Vasilikos Formations are overlain by the Fanglomeraue, which records a major pulse of uplift of the
Correlation of the remaining units in southern Cyprus is uncertain due to lack of continuous exposure and biostratigraphic data. Their relationships are discussed more fully in chapter 9. Their correlation with the youngest sediments of the Mesaoria Plain was not investigated in this project, as they are now the subject of a separate study, by A. Poole (Edinburgh University).

2.3 Biostratigraphy

The earliest attempts to date the formations of the Plio-Pleistocene part of the Troodos sedimentary cover were made using a combination of foraminiferal and macropalaeontological data (Henson et al., 1949, and references therein; Wilson, 1959; Bear, 1960; Gass, 1960; Ducloz, 1965). Systematic, micropalaeontological, biostratigraphic zonation was not carried out until 1968, by Mantis. He divided Miocene-Recent sediments from borehole samples into eight informal zones, A - H (the upper five zones are shown in Table 2.3), which were later used by the Cyprus Geological Survey for borehole correlation during hydrogeological studies (e.g. Zomenis, 1972). Initially, and when he published his complete biostratigraphic zonation scheme for the entire Troodos sedimentary cover (Mantis, 1970), Mantis believed the base of the Myrtou Marl (now Nicosia Formation) to be Upper Miocene in age. He later revised this to Lower Pliocene (Mantis, 1975), and his modified planktonic foraminiferal zones are shown in Table 2.3. No revision of his 1968, informal eight-fold scheme was made.

Systematic biostratigraphy of the sedimentary cover on the north side of the Troodos Massif has also been investigated by Baroz (Baroz, 1979; Baroz and Bizon, 1974; Table 2.3). He examined material from the Mesaoria Plain and the Kyrenia Range. His zonation scheme was based on that of Bizon and Bizon, (1972; Table 2.3), a scheme previously developed for the western Mediterranean. Pliocene biostratigraphy of the Mediterranean was later revised by Cita (1975; Table 2.3), and this scheme was adopted during the 1978 DSDP cruise in the Mediterranean (Leg 42). Biostratigraphic studies of the Mediterranean are still in a state of flux, however, and several other
Table 2.3 - Plio-Pleistocene foraminiferal biostratigraphy of Cyprus and the Mediterranean

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Ath. - Athalassa Formation  
Nic. - Nicosia Formation  
My. ML - Myrtou Marl  
* - zones not clearly defined  
^ - not included in the table of Mantis (1975)
schemes are also in existence. A recent review of these is given by Iaccarino (1985).

All Plio-Pleistocene biostratigraphic studies of Cyprus share a common feature - the relative ease of dating the fully marine, lower part of the succession, which has a rich foraminiferal content, and the difficulty of dating the shallow marine - continental upper part of the succession, which lacks age-diagnostic fossils. There is thus general agreement that the Nicosia Formation is Lower-Upper Pliocene in age, although Baroz (1979) considered his Nicosia Formation (occurring only in northern Cyprus) to be Lower-Middle Pliocene. Problems are encountered with the Athalassa Formation, however, which was considered to be Upper Pliocene by Henson et al. (1949) and Mantis (1975), but Lower Pleistocene by Ducloz (1965). Baroz (1979) was unable to assign it to a narrower age range than Middle Pliocene-Lower Pleistocene.

The Kakkaristra Formation has only been investigated palaeontologically by Ducloz (1965). He believed the formation to be Lower Pleistocene, based on the identification of mammal bones. New evidence, as a result of this study, confirms a Pleistocene age for the top of the Kakkaristra Formation. In situ ostracods from the mudstone at the top of the formation have been preliminarily identified as the Pleistocene-Quaternary species, Cyprideis torasa, by A. Lord (University College London). The formation is thus assigned an ?Upper Pliocene-Lower Pleistocene age.

Lithostratigraphic field relations, observed during this study, have shown that in the eastern Mesaoria Plain, the Athalassa Formation is equivalent to the Kakkaristra and lower Apalos Formations (sections 2.2.1 and 2.2.2; Table 2.2). In view of the ?Upper Pliocene-Lower Pleistocene age adopted for the Kakkaristra Formation (see above), the Athalassa Formation in the eastern Mesaoria is assigned an ?Upper Pliocene-Pleistocene age. The age of the formation in the northwestern Mesaoria, where it is thicker, is uncertain, however.

It should be noted that the position and absolute age of the Plio-Pleistocene boundary is still highly controversial. Historically, it has been assumed to mark the position of the first occurrence in the geological record of the effects of the last Ice Age. Defining
this in biostratigraphical terms, and choosing a suitable type locality, however, have proved very difficult (Nilsson, 1983). Current absolute ages for the boundary range from 1.6 Ma, based on bio- and magnetostratigraphic methods (Bergerhren et al., 1985), to 2.4 Ma, based on oxygen isotope records (Jenkins et al., 1985). No absolute age data are available in Cyprus. In this study, the more commonly accepted age of 1.6 Ma is adopted, although it is recognised that glacial effects may have begun earlier, i.e. in the Upper Pliocene.

The Apalos Formation, Fanglomerate and river deposits of the Mesaoria Plain are unfossiliferous, as are the Karka Formation and river terrace deposits of the Kyrenia Range. On lithostratigraphic grounds, they are assigned Pliocene to Recent ages. Alpine glacial stratigraphy was applied to these sediments on geomorphological grounds by Ducloz (1965) and Baroz (1979). Mediterranean raised beach stratigraphy (reviewed in Nilsson, 1983) has been applied to the marine terraces of both north and south Cyprus (Pantazis, 1966, 1967; Baroz, 1979), again on geomorphological grounds, but also partly on the recognition of the well-known Mediterranean Strombus fauna. This fauna occurs in the 40' raised beach of southern Cyprus (Pantazis, 1966) and several marine terraces in northern Cyprus (Baroz, 1979), and these have been dated as Tyrrenhian as a result. There is now some dispute, however, as to the correct stratigraphic interpretation of the Strombus fauna (Nilsson, 1983).

No systematic biostratigraphic investigations of the Plio-Pleistocene succession of south central Cyprus have been carried out, except for a very recent study of ostracod assemblages (Pavlakelli, 1987). This work convincingly demonstrates a Lower Pliocene age for the lower part of the succession. Previously, this part had been assigned to the Athalassa Formation of Upper Pliocene age (Bagnall, 1960; Pantazis, 1967). The overlying Fanglomerate, raised beach and river terrace deposits are probably Pleistocene-Recent in age, although detailed biostratigraphic data are lacking. No Middle-Upper Pliocene sediments are conclusively demonstrated in south central Cyprus, therefore, although the new Vasilikos Formation, if correctly correlated with the Kakkaristra and
Apalos Formations, may be of Upper Pliocene-Pleistocene age.

2.4 Revised Stratigraphy

Nicosia Formation

Synonymy: Myrtou Marl and Nicosia Formation of Henson et al. (1949) and Bear (1960); Pliocene marls of Wilson (1959) and Carr and Bear (1960); Athalassa Formation of Bagnall (1960) and Pantazis (1967).

Type locality: old type localities at Myrtou and Nicosia are inaccessible and obscured, respectively; two new type localities are proposed - 1) lower Nicosia Formation, small ridge between Episkopion and Aredhiou, in the southern Mesaoria (grid ref. WD202774); 2) upper Nicosia Formation, lower south flank of Kantara hill, 2km east of Xeri (grid ref. WD313807).

Lower contact: always unconformable; overlies the Pakhna Formation (Miocene) at type locality 1; also overlies the Kalavasos Formation (Messinian) and the Lefkara Formation (Eocene to Miocene) on north and south flanks of Troodos; overlies the Kythrea and rarely the Lapithos Groups (Oligocene to Miocene) in north Cyprus.

Upper contact: unconformable at basin margins, conformable in basin centre; overlain by the Kakkaristra Formation at type locality 2, and in the southwest Mesaoria; overlain by the Athalassa Formation, conformably in the east central Mesaoria and unconformably in the northern Mesaoria and Kyrenia Range; also unconformably overlain by the Fanglomerate in the southern Mesaoria and south Cyprus.

Description of section: type locality 1 - thick, green-grey, homogeneous, fossiliferous, calcareous silts, with channel-shaped conglomerate bodies, which give way to sandier, channelled facies upwards; type locality 2 - thick, green-grey to brown calcareous silts, coarsening up into very fine, orange-yellow sands, with concretionary cemented bands.

Age: Lower-Upper Pliocene.

Remarks: the formation is exposed through much of the southern
Mesaoria, where it weathers to give badlands-type topography.

**Athalassa Formation**


Type locality: Leonardi hill, between Nicosia and Yeri (grid ref. WD 374883; presently occupied by the Greek Cypriot army, therefore not studied by this author).

Lower contact: unconformable at basin margins, conformable in basin centre; overlies the Nicosia Formation, unconformably in north Cyprus and the northern Mesaoria (including the type locality), and conformably in the east central Mesaoria, where its base is taken at the first calcarenite horizon in the succession; upper member unconformably overlies the Kythrea flysch in northwestern Cyprus (Moore, 1960).

Upper contact: generally, no overlying formation present; in the vicinity of Nicosia, conformably overlain by the Apalos Formation (Ducloz, 1965); occasionally unconformably overlain by the Fanglomerate.

Description of section: cross-bedded calcarenites, calcitic sandstones and silty sands, with minor conglomerate lenses.


Remarks: south of the type locality, fine-grained orange-yellow sands (similar to the upper Nicosia Formation) become important, and separate calcarenite horizons; the Athalassa Formation does not occur in the western and west central Mesaoria, where it is replaced by its lateral equivalents, the Kakkaristra Formation and lower Apalos Formation; in northwest Cyprus, the formation comprises a thick, lower, silty member, with minor conglomerate, and a thin, upper, cross-bedded calcarenite member (Moore, 1960).

**Kakkaristra Formation**

Synonymy: Kakkaristra Formation of Ducloz (1965); marine Fanglomerate of Gass (1960); Nicosia Formation and marine Fanglomerate of Bear (1960); part of Myrtou Marl of Carr and Bear (1960).
Type locality: section of the Kakkaristra river, 1km west of Laxia (grid ref. WD334836), as proposed by Ducloz (1965).

Lower contact: unconformable or disconformable at basin margin; conformable in basin centre; overlies the Nicosia Formation, and rarely in the southwest Mesaoria, older formations; contact characterised by a thick conglomerate horizon, or colour change (orange-yellow Nicosia Formation sands pass up into grey-brown Kakkaristra Formation sands) in much of the central Mesaoria.

Upper contact: conformably overlain by the Apalos Formation; contact marked by a thin, ostracod-bearing mudstone or prominent, white caliche horizon.

Description of section: grey-brown sands, with conglomeratic lenses and stringers, overlain by muddy, fossiliferous sediments.


Remarks: the formation is lithologically highly variable; away from the type locality, thick, cross-bedded conglomerates, homogeneous silty facies, thin cream and greenish mudstones and calcitic sandstones occur; a cream-coloured sandstone containing a high proportion of reworked foraminifera occurs in the east, close to the contact with the Athalassa Formation.

Apalos Formation

Synonymy: Apalos Formation of Ducloz (1965); Fanglomerate of Carr and Bear (1960); marine Fanglomerate of Gass (1960); marine and continental Fanglomerate of Bear (1960).

Type locality: Apalos hill, 1/4 km west of Laxia (grid ref. WD330833), as proposed by Ducloz (1965).

Lower contact: always conformable; overlies the Kakkaristra Formation at the type locality; overlies the Athalassa Formation in the vicinity of Nicosia (Ducloz, 1965).

Upper contact: unconformably overlain by the Fanglomerate, where an overlying formation is present.

Description of section: reddened silts and muds, minor sandstones, the lowest of which contain skeletal carbonate grains, and caliche; conglomerates become increasingly important upwards.

Age: ?Pleistocene.

Remarks: laterally equivalent to part of the Athalassa Formation;
possibly laterally equivalent to the Karka Formation in the
Kyrenia Range (Ducloz, 1972).

**Vasilikos Formation**

*Synonymy:* none.

*Type locality:* quarry on the south coast, 1km south of Mari (grid ref. WD276428).

*Lower contact:* unconformably overlies the Nicosia Formation.

*Upper contact:* unconformably overlain by the Fanglomerate.

*Description of section:* coarse, poorly sorted conglomerates, pebbly sands and aeolian cross-bedded sands.

*Age:* ?Upper Pliocene–Pleistocene.

*Remarks:* very limited outcrop area; possible equivalent to the Kakkaristra and Apalos Formations of the Mesaoria Plain.

**Fanglomerate**

*Synonymy:* part of Fanglomerate of Wilson (1959) and Carr and Bear (1960); continental Fanglomerate of Bear (1960) and parts of continental and marine Fanglomerate of Gass (1960); Kantara and Kambia gravels of Ducloz (1965).

*Type locality:* none.

*Lower contact:* always unconformable; unconformably overlies all older formations, and in the western Mesaoria, Troodos basement.

*Upper contact:* no overlying formation is present.

*Description:* thin sheet of coarse, angular, poorly sorted conglomerate, massive to poorly stratified, with occasional sandy intercalations.

*Age:* ?Pleistocene.

*Remarks:* has been divided into two units on geomorphological grounds by Ducloz (1965); typically occurs as caps to mesas in the Measoria Plain; outcrop very limited on the south side of Troodos.
The Plio-Pleistocene sediments investigated in this study crop out in one main area of Cyprus, the Mesaoria Plain (Fig. 1.1, and see Encl. A). They represent a gross shallowing-up sequence, which documents the progressive infilling of a narrow, E-W trending seaway (the Mesaoria basin), which formed in the Late Miocene between an emergent Troodos Massif to the south, and an emerging Kyrenia lineament to the north. The seaway filled initially with thick marine sediments (the Nicosia Formation, Table 2.2, centre column), which were succeeded by much thinner, shallow marine and deltaic facies (Athalassa and Kakkaristra Formations respectively), and finally by continental deposits of the Apalos Formation and the Fanglomerate.

The Mesaoria basin has a gentle synclinal structure, with a WNW - ESE trending axis. It has undergone very little deformation since its formation, except along its margins, where uplift and faulting have tilted bedding up to 70° and generated local unconformities. Folding is virtually absent. In the centre of the basin, the exposed upper part of the basin fill is almost entirely undisturbed, and bedding is either horizontal, or dips gently north or south towards the basin axis.

The present study has focussed on the south side of the basin, as much of its northern part lies within the Turkish-occupied zone of Cyprus. Nevertheless, it has been possible to formulate models for the entire basin, as several published accounts of the geology of the north side of the basin are available. Uplift and erosion along the southern margin of the basin have resulted in exposure of lower parts of the basin succession along this margin, while progressively younger parts crop out towards the basin's centre.

Small outcrops of Plio-Pleistocene sediments also occur along the southern coast of Cyprus (Fig. 1.1). These sediments were investigated in order to compare Plio-Pleistocene successions on both northern and southern flanks of the Troodos Massif, and to correlate onshore geology with offshore seismic data (see chapter 10). This investigation was predictably less extensive than that of the
Mesaoria Plain due to the limited amount of exposure. A small Pliocene basin was identified, however, and is described together with Plio-Pleistocene facies in chapter 9.
3.1 Introduction

The Nicosia Formation forms the oldest part of the Plio-Pleistocene sedimentary succession of the Mesaoria basin. In this study, it includes the informal stratigraphic unit, the Myrtou Marl (see sections 2.2.1 and 2.4 for the formal definition of the formation).

The formation crops out widely throughout the southern Mesaoria Plain (Fig. 3.1, Encl. A). It occupies the major part of the thickness of the sedimentary fill of the Mesaoria basin, ca. 900m out of a total of 1km (see Table 2.2). Because of its great thickness, and generally shallow dip, complete exposed sections through the formation do not exist, and its true thickness is known only from boreholes. Cyprus Geological Survey hydrogeological borehole data (which comprise lithological descriptions and some micropalaeontological information) have thus been used in conjunction with field data in structural and facies analysis of the formation. In general, partial sections of the upper few hundred metres only of the formation are exposed, except along the southern margin of the basin, where the lower part and base of the formation outcrop (see e.g. Fig. 3.4).

The formation is Lower-Upper Pliocene in age (section 2.3). Its lithofacies comprise an entirely marine sequence of sediments, which represent several offshore depositional environments. The formation is dominated by green-grey structureless silts, which characteristically weather to give a badlands type of topography (Plate 3.1a), typical of the scenery of the southern Mesaoria Plain.

3.2 Basal Relations

Before describing lithofacies in detail (section 3.3), sedimentation and structure at the base of the Nicosia Formation are outlined. Important evidence for the structural configuration of the basin is found here, and has implications for facies interpretation.
Fig. 3.1 - Approximate area of outcrop of the Nicosia Formation (stippled), and location of basal unconformity.

Fig. 3.2 - Facies distribution, Nicosia Formation (see Table 3.1 for lithofacies).
3.2.1 Observations

Sediments of the Nicosia Formation unconformably overlie older parts of the Troodos sedimentary cover (Lefkara, Pakhna and Kalavasos Formations, section 1.3.3), and occasionally Troodos basement itself. This relationship is widely recognised on both sides of the Mesaoria basin (Wilson, 1959; Bear, 1960; Carr and Bear, 1960; Gass, 1960; Moore, 1960; Lytras, 1963; Zomenis, 1972; Baroz, 1979), and in south and west Cyprus (Hadjistavrinou and Constantinou, 1977; Hadjistavrinou and Afrodisis, 1977; Ward and Robertson, 1987). In all these areas, the unconformity surface is typically overlain directly by marine silts (facies Al, section 3.3.1; Plate 3.3c), as recorded by both field and borehole data (see e.g. Figs. 3.3 - 3.5).

In the study area, the unconformity is exposed along the northern flank of the Troodos Massif (Fig. 3.1). A number of important observations have been made:

a) east of Kochati (Fig. 3.1), poor exposures of the unconformity show it to have a smooth, planar surface, dipping at about 10° northwards; dips in the underlying Lefkara Formation are uniformly to the north (Gass, 1960); in map form, the unconformity has a rather smooth trace.

b) between Kochati and Pera, the unconformity dips smoothly at about 10°; dips are more variable in the underlying Lefkara Formation, which is also more faulted (Bear, 1960; Gass, 1960).

c) between Pera and Kato Moni, the trace of the unconformity becomes more irregular, with embayments up to a few hundred metres wide; the dip of the unconformity becomes steeper, and is subvertical and probably faulted in places (e.g. WD 156785); between Agrokipia and Kato Moni, an area mapped in detail by E. Follows, structural relationships in the underlying pre-Pliocene sedimentary cover are complex, and indicate the presence of a series of tilted fault blocks (Follows and Robertson, in press); minor normal faults also affect the base of the Pliocene (e.g. WD 188779).

d) in the far west (Potami area, Fig. 3.1), the unconformity is again poorly exposed; significant differences in the topographic elevations of closely-spaced exposures of the top of the Nicosia Formation (and the overlying Kakkaristra Formation) strongly
suggest that young faulting affects this area.
e) where exposed, on close examination, the unconformity surface is smooth and unfissured, and shows no signs of prolonged subaerial exposure.
f) coarse-grained facies occur in the Nicosia Formation at or close to the unconformity, in the centre of the study area (facies B1, Fig. 3.2, section 3.3.1); they occur as a number of discrete lobes, and do not constitute a continuous basal Pliocene unit; these conglomeratic sediments are mainly Troodos-derived, and largely comprise mass flow deposits, including megaconglomeratic debris flows (facies B1c); some conglomerates, however, are rich in angular, pre-Pliocene sedimentary clasts (facies B1d); large, isolated rafts of pre-Pliocene sediment also occur.
g) in the Kato Moni-Agrokipia area, basal Pliocene silts are pale and very calcareous, and grade up over a few tens of metres into more normal brown or grey-green sediments; debris flow conglomerates rich in pre-Pliocene sedimentary clasts are also common, and slumped units of selenitic gypsum occur (Follows and Robertson, in press).

3.2.2 Interpretation

Several important conclusions can be drawn from the previous observations.

Major unconformity surface

A major unconformity separates Pliocene sediments from Miocene and older rocks in the Mesaoria basin, a relationship recorded in much of Cyprus. It is also observed throughout the Mediterranean. This prominent unconformity was largely generated during the well known Messinian salinity crisis, when sea level fell in the region and evaporite deposition occurred (Hsu et al., 1978; but see also following subsection). Fully marine silts (facies A1) directly overlie the unconformity surface in the Mesaoria basin, and elsewhere in Cyprus. Again, this is documented throughout much of the Mediterranean, and represents the rapid return of open marine conditions in the Early Pliocene, following the ending of the salinity crisis. Swift transgression is marked in the Mesaoria basin by the
lack of development of transitional facies, and the absence of signs of significant oxidation of the unconformity.

**Extensional faulting**

Extensional faulting is known to have been affecting the northern margin of the Troodos Massif, prior to the Pliocene (see section 1.3.3). This faulting began in the ?Mid Miocene, and by the end of the Miocene, a number of fault blocks were in evidence (Follows and Robertson, in press). There is abundant evidence that faulting continued into the Pliocene. This includes (see previous observations): local faulting of the unconformity itself; the presence of sedimentary clast-rich mass flows, and isolated blocks of the same material, both of which were probably derived very locally from active fault scarps, comprising pre-Pliocene sediments (see facies B1d); the presence of megaconglomerates, deposition of which was probably triggered by earthquakes (see facies B1c); and the presence of highly channelised, coarse-grained facies, interpreted as marine fan-delta lobes, cut across a steep, narrow, fault-controlled, basin margin (see facies B1a and b). The importance of this faulting with respect to the structural setting of the basin is discussed later (section 3.4.2).

The irregular, embayed morphology of the unconformity surface, in the centre of the study area, may be an expression of fault block topography. Coarse-grained fan-delta sediments (facies B1) are found to be at least partly localised within embayments (see Fig. 3.11a). Although fault geometry was not delineated in detail, it is possible that embayments may represent areas of fault block intersection, which were exploited by fan-deltas as easy routes to the sea (see Fig. 3.11b). Erosion during the Messinian (see previous subsection) was thus not the only control on the morphology of the unconformity surface.

**Reduced faulting to the east**

Also apparent from the previous observations is an easterly decrease in fault intensity, and in the occurrence of coarse basal clastics in Pliocene sediments. The inferred reduction in faulting is also apparent on cross-sections across the Mesaoria Plain,
constructed from borehole data by Cleintaur et al. (1977). The unconformity surface itself also dips less steeply in the east, and has a smoother, less embayed trace (Fig. 3.1).

These observations may in part be due simply to poorer exposure in the east. Furthermore, locations of exposures of the unconformity, which lie along a NW-SE trending line between Kato Moni and Politiko (Fig. 3.1), shift northwards east of Politiko. It is possible that faulting of the unconformity continued further east, in a narrow zone along the same SE trend, but evidence for it is now lost, because the sedimentary cover has been eroded off the ophiolite in this area.

Facies trends in the Nicosia Formation reveal, however, that the reduction in fault intensity to the east may be real. In this area, intervals of bioclastic sands (facies C1) are intercalated with ubiquitous facies A1 silts in the lower Nicosia Formation (Fig. 3.2; section 3.3.1). The presence of bioclastic material, not observed elsewhere, may suggest that a shelly strandline lay shoreward of the locations of this facies (see facies C1, section 3.3.1), and not a series of fan-deltas, as is the case further west (Fig. 3.11a and b). Fan-deltas are believed to have formed in response to faulting and uplift along the south central basin margin (see facies B1, section 3.3.1). Their absence to the east may thus be indicative of a less intensely faulted coastline.

Emergence of the Troodos Massif

Another important feature of early Pliocene sediments in the Mesaoria basin is that they document significant subaerial emergence of the Troodos Massif for the first time (see also section 8.3). Uplift of the ophiolite from sea floor depths began in the Oligocene (section 1.3.3), but by the end of the Miocene, the Troodos Massif was still largely below sea level. Initial signs of emergence are documented by the presence of Troodos-derived conglomerates in the Upper Miocene Pakhna Formation, on both flanks of Troodos (Eaton, 1987; Follows and Robertson, in press), but these are of very local occurrence.

A striking change occurs, however, at the beginning of the Pliocene, where the cream or pale brown marls of the Pakhna
Formation are overlain by the dark brown to green-grey calcareous silts of the Nicosia Formation. Furthermore, Troodos-derived conglomerates and sands (facies B1 and B3) become more widespread. These facies amply document that for the first time, the Troodos Massif was significantly uplifted above sea level.

**Summary**

In summary, uplift of the Troodos Massif, which began approximately in the Oligocene, continued into the Pliocene. The ophiolite was significantly uplifted above sea level for the first time, and became a major detrital sediment source. Normal faulting, which had been affecting the north flank of the ophiolite since the Miocene, continued, and a variety of fault-related sediments were shed into the basin. Faulting was apparently less intense in the east than in the west. Elsewhere, fully marine silts were deposited over the unconformity at the base of the Pliocene, as seas flooded back into the basin following the Messinian salinity crisis.

3.3 Facies and Facies Relationships

3.3.1 Lithofacies description and interpretation

Seven major lithofacies have been recognised in the Nicosia Formation from field studies (Table 3.1), one of these having been divided into several subfacies. Because of the lack of complete exposed sections through the formation, borehole data have been used to augment field data where possible. Borehole data come from a number of wells drilled in the Mesaoria Plain by the Cyprus Geological Survey, for hydrogeological purposes (see Fig. 3.14 for well locations). The data comprise lithological descriptions of rock cuttings recovered during drilling, and some micropalaeontological data (e.g. Mantis, 1968). No core data or detailed sedimentological descriptions are available. In view of this, only the major lithofacies recognised from field studies can be identified in boreholes (facies A1, A2, B1 and B2). The data have been useful, however, in determining sediment body geometry, large-scale facies relations and sediment thickness trends.
Table 3.1 - Lithofacies, Nicosia Formation

<table>
<thead>
<tr>
<th>Code</th>
<th>Lithofacies</th>
<th>Depositional Environment</th>
</tr>
</thead>
<tbody>
<tr>
<td>A1</td>
<td>Calcareous silts</td>
<td></td>
</tr>
<tr>
<td>A2</td>
<td>Fine sands</td>
<td>Moderately deep to shallow sea</td>
</tr>
<tr>
<td>A3</td>
<td>Thin conglomerates</td>
<td></td>
</tr>
<tr>
<td>B1</td>
<td>Conglomerate-sand channels</td>
<td></td>
</tr>
<tr>
<td>B1a</td>
<td>Conglomerates</td>
<td>Steep, faulted, marine fronts of slope</td>
</tr>
<tr>
<td>B1b</td>
<td>Sand-silts</td>
<td>deltas</td>
</tr>
<tr>
<td>B1c</td>
<td>Megaconglomerate</td>
<td></td>
</tr>
<tr>
<td>B1d</td>
<td>Sedimentary clast-rich conglomerates</td>
<td></td>
</tr>
<tr>
<td>B2</td>
<td>Sand-silt channels</td>
<td></td>
</tr>
<tr>
<td>B3</td>
<td>Sheet sands</td>
<td></td>
</tr>
<tr>
<td>C1</td>
<td>Bioclastic sands</td>
<td>Reworked shoreline sands</td>
</tr>
</tbody>
</table>
Lithofacies descriptions and interpretations are given in the following, before sedimentological models are discussed in section 3.3.2.

**Facies A1 - Calcareous silts**

**Description of exposed sections:** this facies comprises homogeneous, mainly structureless, mid to dark brown or green-grey, calcareous, clayey silts, usually poorly consolidated and weathering badly (Plate 3.1a); sedimentary structures are lacking due to intense bioturbation, or are obscured by weathering; where cementation is best, horizontal, thin to thick lamination is evident; the facies is sparsely to moderately macrofossiliferous, with both whole shells and dispersed broken fragments; plant debris is sometimes concentrated along laminae.

**Borehole data:** cuttings descriptions of the unexposed, lower to middle parts of the Nicosia Formation from the centre of the Mesaoria basin, show that, like the upper, exposed part, these lower sections are dominated by very fine-grained sediments (see Figs. 3.3 - 3.7), usually referred to as marl or sandy marl; these terms have also been applied previously to exposed parts of the formation (e.g. Bear, 1960), although Ducloz (1965) noted that calcareous silt was a more accurate description; based on micropalaeontological data, the silts from these boreholes fall into zones C - E of the informal biostratigraphic scheme of Mantis (1968; section 2.3); brief palaeoecological comments by Mantis (1968) indicate a broad shallowing-up through the silts from zone E to zone C, from "moderately deep sea" depths to "shallow marine conditions" (no precise depths given).

**Occurrence:** these sediments form the dominant facies of the Nicosia Formation; the Xeri deep borehole (Fig. 3.4) recorded shelly silts continuously for over 800m; the facies occurs throughout the study area (Fig. 3.2) and throughout the Nicosia Formation, except for the top 20-100m.

**Lateral and vertical facies relationships:** the facies encases all other Nicosia Formation facies, except A2 and A3, and except B1 where B1 occurs directly over the Miocene-Pliocene unconformity; contacts with other facies are sharp, except for A2, which nearly
Fig. 3.3 - Sedimentological logs and interpreted cross-section A–A', constructed from field and borehole data (see Fig. 3.4 for key to section, Fig. 3.9 for section location, and Encl. B for key to logs).
Fig. 3.4 – Sedimentological logs and interpreted cross-section B-B’, constructed from field and borehole data (see Fig. 3.9 for section location and Encl. B for key to logs).
Fig. 3.5 - Interpreted cross-sections C-C' and D-D', constructed from field and borehole data (see Fig. 3.4 for key to sections and Fig. 3.9 for section locations).
Fig. 3.6 - Sedimentological logs, south end of cross-section C-C' (see Fig. 3.5 for log locations and Encl. B for key to logs).

Fig. 3.7 - Interpreted cross-sections E-E' and F-F', constructed from borehole data (see Fig. 3.4 for key to sections and Fig. 3.9 for section locations).
always overlies facies A1 gradationally at the top of the formation.

**Palaeontology:** macrofauna comprise a large variety of molluscs (mainly bivalves and gastropods; see Appendix 1 for a complete faunal list, and see also Wilson, 1959, Bear, 1960 and Gass, 1960); a single, intact, though not upright, head of the branching coral *Cladocera caespitosa* is also recorded; data on macrofauna from the deeper, unexposed parts of the Nicosia Formation are not available, except for the occurrence of oyster beds in the Xeri deep borehole (Fig. 3.4; Gass, 1960); macrofauna in general represent a shallow marine assemblage; microfauna have been studied by Mantis (1968) and Baroz (1979); see Borehole data for comments.

**Interpretation:** this facies represents a sequence of fully marine sediments, mainly structureless due to intense bioturbation; microfaunal studies indicate a general shallowing up through the main silt body of the Nicosia Formation, from ??500m to shallow shelf depths; in the upper, exposed part of the formation, lack of sedimentary structures and fine grain size suggest deposition below storm wave base, or deposition in a relatively low energy environment, where infrequent storm events were obliterated by bioturbation; preservation of an intact head of a delicate branching coral supports a relatively low energy setting; a further possibility is that storm-transported sand was largely trapped in channel-shaped depressions on the sea floor (see facies B2); precise palaeobathymetry is uncertain; grain size is not diagnostic of water depth, even on equilibrium shelves, free from complications caused by relict sediment, because mud deposition depends on the concentration of suspended sediment being supplied to the shelf, and the shelf hydraulic regime (McCave, 1971, 1985); oyster beds in the Xeri deep borehole suggest fairly shallow depths, but it is not known if these beds represent *in situ* fauna or transported accumulations; the gastropod *Turritella communis*, common in the exposed part of the facies, is characteristic of the "outer faunal community" of part of the modern day Mediterranean shelf (Gulf of Gaeta; Dörjes, 1971), and lives in depths of 15-50m; sediments from this zone of the relatively low energy Gulf of Gaeta are also intensely bioturbated (Reinick and Singh, 1973); the gastropod is also typical of the circumlittoral zone of the Mediterranean in
general (Pérés, 1967; water depths of 30-150m).

**Facies A2 - Fine sands**

**Description:** the facies comprises very fine-grained, moderately to well sorted, slightly muddy, orange- or yellow-brown sands, which are generally poorly cemented, structureless and bioturbated; parallel lamination is sometimes evident; in some areas, thin, parallel, cemented layers, separated by 30-60cm intervals of unconsolidated sand, are present (e.g. log 12/5, Fig. 3.3; Plate 3.1b; see also Bear 1960, and Gass, 1960); these layers are either very thin (1-2cm) and completely cemented, or are thicker (3-20cm) and contain discontinuous calcareous concretions; concretions do not distort any visible lamination, are not commonly concentrated around shell fragments, but sometimes contain wood fragments; up to 4 x 10cm in cross-section; bedding, where discernible, is almost always horizontal; large-scale, low-angle, trough-shaped cross-bedding occurs at one locality, along the Pedeios River (WD 257835; Plate 3.2a, Fig. 3.8); at this same locality, two channels are present at the base of the facies; one channel is filled with stratified, rounded conglomerate, the other with partially slumped, slightly pebbly, stratified sand; shells and shell fragments are quite common, both as scattered debris or occasionally concentrated in thin layers or lenses with diffuse contacts; at one locality at the top of the facies (WD 335838), oysters are concentrated in a number of horizontal layers, typically with articulated shells aligned parallel to bedding (Plate 3.1c); bioturbation is usually intense, and distinct burrows are uncommon; *Chondrites, Planolites* and occasional *Skolithos* have been recognised, however.

**Occurrence:** field data show the facies to form a continuous layer at the top of the Nicosia Formation, above facies A1, 10-50m thick; borehole data also support the presence of a laterally continuous sandy layer at the top of the formation (Figs. 3.3 - 3.5), which is up to 100m thick.

**Lateral and vertical facies relations:** the facies always overlies facies A1 silts, usually with a gradational contact over a metre or two (e.g. log 12/5, Fig. 3.3); a sharp, channeled contact is present at one locality only (Fig. 3.8); facies A3, with gradational contacts, and
Fig. 3.8 - Sketch of Pedeios River cliff section, showing the upper part of the Nicosia Formation. Dashed vertical lines show the location of Plate 3.2b.

Fig. 3.9 - Palaeocurrent data, Nicosia Formation, and location of sedimentological cross-sections. Some individual channel trends are shown on the map with small arrows. Outcrop area of the formation is stippled.
facies B2, with sharp contacts, are occasionally present within facies A2.

**Palaeontology:** macrofauna are very similar to facies A1 (see Appendix 1 for species list); the occasional occurrence of additional fauna include whole, possibly in situ, specimens of *Pinna* sp. and *Glossus hummanus*, and the echinoid *Schizaster* (Ora) *canaliferous* (Lamark), one specimen of which still had spines attached; facies A2 probably belongs to micropalaeontological biostratigraphic zone B of Mantis (1968; section 2.3); zone B is typically sandy, not silty, contains few pelagic foraminifera, and represents "very shallow marine conditions" (Mantis, 1968).

**Interpretation:** these sediments represent fully marine, shelf facies, probably deposited in shallower water than the finer-grained facies A1; intense bioturbation again suggests deposition below storm wave base, or obliteration of occasional storm events in a relatively low energy environment by bioturbation; preservation of probably in situ *Pinna*, *Glossus* and an echinoid with attached spines, suggests relatively quiet conditions; the ichnofauna recognised are characteristic of the *Cruziana* ichnofacies, which typifies moderate or relatively low energy shelf areas below fairweather wave base (Seilacher, 1967; Frey and Pemberton, 1984); a similar sequence of bioturbated sands overlying bioturbated silts is recorded in the Gulf of Gaeta (Rienick and Singh, 1973); the sands occur in water depths of 6-15m; bioturbation in the sands is largely due to the heart urchin, *Echinus cordatum*, which is not present in this facies, although another burrowing spatangoid, *Schizaster*, is; some evidence for strong current activity is present in the facies, e.g. the channels at the base of the facies, one of which is filled with gravel (see facies A3 for discussion of gravel transport on the shelf); the low-angle trough cross-beds may be a form of swaley cross-stratification (Leckie and Walker, 1982), or similar to the structures described by Nøttvedt and Kreisa (1987); both types of structure are related to hummocky cross-stratification, which is believed by many workers to result from the combination of oscillatory flow with storm-generated unidirectional currents (see discussions in DeCelles, 1987, and Nøttvedt and Kreisa, 1987); the oyster beds at the top of the facies, with horizontally aligned shells,
were probably reworked by strong currents into discrete beds; the intercalation of facies A3 and B2 also implies storm currents; on the whole, however, storm-related deposits are not common, and a relatively low energy shelf setting is envisaged, with much biological reworking.

**Facies A3 - Thin conglomerates**

**Description:** the facies comprises pebble to cobble conglomerates in thin to thick lenses (10cm-1.5m), up to tens of metres in lateral extent (Plate 3.3a); clasts are mainly subrounded, sometimes flattened, poorly sorted except for small, better sorted pods (Plate 3.3b); conglomerate fabric is clast-supported, with a sandy, sometimes shelly matrix; the matrix lithology is similar to surrounding facies A2 sands; beds have horizontal, or sometimes highly loaded, bases (e.g. log 25/10/1, Fig. 3.4), and planar to wavy tops; clast orientation is very variable, and may be vertical in beds with distorted bases; shelly material is present and comprises broken bivalves, with recognisable *Ostrea* and *Pecten* fragments.

**Occurrence, lateral and vertical facies relations:** the facies is only found at a few localities at the top of the Nicosia Formation, in the central Mesaria (Fig. 3.2); it is always encased in facies A2 (Figs. 3.4 and 3.8); a thin conglomerate recorded in a private borehole (Fig. 3.5) probably belongs to this facies.

**Interpretation:** enclosure in facies A2 implies a similar depositional environment to facies A2, i.e. shallow shelf; conglomerate is not commonly reported from shallow marine settings, and processes of transport and deposition in this environment are still poorly understood (Bourgeois and Leithold, 1984); because unusually strong currents are required for gravel transportation, storm effects are usually invoked (e.g. Wright and Walker, 1981; DeCelles, 1987); poor sorting, loaded bases, lack of grading and vertical clast arrangement suggest very rapid deposition, perhaps from storm-generated, coarse-grained turbidity currents, as suggested by Walker (1983); wavy bed tops may be due to partial, post-storm, reworking of gravel lens tops (cf. Wright and Walker, 1981; Leckie and Walker, 1982); unlike other reported shallow marine conglomerates, these are neither stratified (either horizontal or
cross) nor graded; the facies is not common in the basin; this, in combination with the highly bioturbated nature of the enclosing facies, implies relatively infrequent storm events (see also facies A2).

**Facies B1 - Conglomerate-sand channels**

This facies comprises four subfacies, which occur together, or close to each other, in the lower part of the Nicosia Formation. Each subfacies is described separately, then occurrence, lateral and vertical facies relations and interpretation are discussed together.

**Facies B1a - conglomerates:** this subfacies contains pebble-cobble conglomerates and pebbly coarse sands, poorly to moderately sorted, with subangular to mainly subrounded clasts; coarser beds have clasts up to 10cm, and are either massive, graded (inverse and occasional normal; Plate 3.4a), or crudely horizontally-layered; fabric is mostly clast-supported, with occasional matrix support; matrix is of muddy fine or medium sand; occasional up-dip imbrication is developed (Plate 3.4a); pebbly sands are thinner, often horizontally stratified (Plate 3.4b), internally massive, and occasionally contain huge, outsize clasts, up to 60cm across (Plate 3.4a); beds are 10cm-2m thick, lenticular (i.e. laterally extensive over only a few tens of metres; Plates 3.2b and 3.4c), and have flat, non-erosive, to highly erosive, concave-up bases; clast composition is usually ca. 90% Troodos-derived; largest clasts are typically of pre-Pliocene reef limestone; the sediments may be very shelly, sometimes with intact, thick-shelled molluscs preserved e.g. *Strombus* or *Ostrea*.

**Facies B1b - sand-silts:** this subfacies comprises very thin to thin, parallel-bedded, fine- to occasionally medium-grained sands, with very thin interbedded, very fine sands or silts; thicker beds are massive or faintly parallel-laminated; rarely, very well normally-graded beds, with loaded bases and sometimes flamed tops, are present; siltstone intraclasts are sometimes abundant at the bases of sand beds; beds are laterally persistent over tens of metres at most, between conglomerate lenses or low-angle, concave-up truncation surfaces (Plate 3.4c); distorted bedding and microfaults are sometimes developed beneath conglomerate lenses (e.g. log 19/12/3, Fig. 3.6); occasionally beds are distinctly shelly, with broken molluscan debris; wood fragments, plant debris and
subhorizontal burrows are sometimes present; this facies always occurs interdigitating with facies B1a.

**Facies B1c - megaconglomerate:** occurring at one locality only (WD 218758), this subfacies comprises a 10m thick bed of disorganized, massive, poorly sorted, matrix-supported conglomerate, with clasts from 1cm to 8m in diameter (the latter composed of reef limestone); matrix is pale sandy mud; the base of the bed is sharp, planar, and non-erosive (Plate 3.4b); no internal organisation is apparent; the conglomerate is non-fossiliferous and is interbedded with facies B1a and B1b, only a few metres above the Miocene-Pliocene unconformity surface.

**Facies B1d - sedimentary clast-rich conglomerates:** this subfacies contains a range of mainly conglomeratic sediments, comprising: 20-30cm thick, clast-supported, pebble conglomerate with numerous, up-dip imbricated, tabular chalk clasts in a fine, chalky matrix; 1-2m thick, poorly sorted, massive, pebble to large cobble, clast-supported conglomerate, largely composed of reef limestone debris; 5-20cm thick, cream-coloured, pebbly, chalky silts, slightly burrowed; and large, isolated rafts of chalk, up to 60cm in length, embedded in facies A1 silts; bedding is horizontal, planar, and generally non-erosive; thin layers of facies A1 silts are sometimes interbedded; these sediments interdigitate to form thin wedge-shaped units, up to 2 or 3m in thickness, and are laterally impersistent (Plate 3.3c).

**Occurrence:** the four subfacies occur at the base of the Nicosia Formation, sometimes directly overlying the Miocene-Pliocene unconformity (e.g. log 3/12/1, Fig. 3.4); they are exposed in the south central part of the study area, and are characteristically located in embayments within the unconformity surface (Fig. 3.2); borehole data indicate the facies may extend several kms northwards, towards the centre of the basin (e.g. Fig. 3.5); this is assuming that conglomerate layers in boreholes correlate with facies B1, and do not form isolated bodies, unattached to the shoreline (see also facies B2); facies B1d is particularly common in the Kato Moni-Agrokipia area (Fig. 3.1).

**Geometry, lateral and vertical facies relations:** facies B1a, b and c occur together in channel-shaped bodies (Plate 3.2b), which cut
down into facies A1 silts; channel shapes appear variable, and may be narrow, with steep, subvertical margins, to much shallower features, with gentle margins (Plate 3.2b is intermediate in shape); channel dimensions are also variable, with depths varying from 10-20m (from exposed sections) to 70-80m (from borehole data); widths are less easily estimated, but are up to at least 500m; channels show internal fining- and coarsening-up, but overall, fine up; fining down-dip is also apparent; facies B1d nearly always occurs separately from facies B1a, b, and c, and is under- and overlain by facies A1 (Plate 3.3c).

**Palaeocurrents:** palaeocurrent data (from channel trends and clast imbrication) are rather variable (Fig. 3.9), but are predominantly between NW and E.

**Interpretation:** a marine origin for these subfacies is indicated by the presence of marine fauna, lateral equivalence and encasement in marine facies A1, and absence of any features pointing to subaerial deposition (caliche, rootlets etc); macro- and micropalaeontological data from the surrounding facies A1 silts suggest deposition at shelf depths; channeling cannot therefore be attributed to an outer shelf/slope setting, but may be due to several other factors: climatic effects resulting in unusually erosive processes, eustatic sea level fall resulting in lowered base level, or tectonic uplift of the hinterland (see Ori and Roveri, 1987); changes in climate or sea level are not recorded from the Lower Pliocene in other parts of the Mediterranean; there is much evidence, however, that the southern margin of the Mesaoria basin was an active fault lineament in the Lower Pliocene (section 3.2.2); channeling is thus interpreted as the result of progradation of a number of sediment lobes across the steep, narrow, fault-controlled, southern margin of the basin; clastic influx occurred as relative uplift of the margin took place; sediment lobes probably formed at the fronts of a number of small fan-deltas; localisation of fan-delta lobes within embayments in the Miocene-Pliocene unconformity surface, and variable palaeocurrent data (Fig. 3.9), suggest that fault block morphology of the unconformity surface (section 3.2.2) may have controlled fan-delta location and direction of progradation; sediment was largely reworked from the subaerial and coastal parts of
fan-deltas, as evidenced by the rounding of clasts and presence of subtidal to shallow neritic fauna; in the absence of evidence for traction current processes (e.g. cross-lamination, rippling), sediment transport is considered to have been largely the result of mass flow; matrix-supported conglomerates represent cohesive debris flows; crudely stratified conglomerates may be due to surging debris flows (Nemec and Steel, 1984); graded, clast-supported conglomerates (similar to facies classes A2.3 and A2.4 of Pickering et al., 1986), and horizontally stratified, massive pebbly sands (similar to facies class A2.5) were deposited by a variety of coarse-grained, high density turbidity currents; sandy facies typically do not show Bouma sequences, but resemble sands of facies classes B1.1 and B2.1 of Pickering et al. (1986); these are interpreted as the product of rapid deposition from high density turbidity currents, in this case carrying already-sorted, beach-derived sand; minor suspension sedimentation (silts) and burrowing occurred between flows; mass flow was probably initiated during periods of high sedimentation (during storms), slumping off fan-delta fronts, and seismic shock associated with faulting (facies B1c); facies B1d, also the product of mass flow, indicates rather short-lived (hence thin), very locally-derived bursts of sedimentation, perhaps directly from submarine fault scarps, cut into the pre-Pliocene sedimentary cover of the Troodos Massif; this subfacies is typically located along the most intensely faulted part of the north Troodos margin (section 3.2.2; see also Follows and Robertson, in press); isolated chalk rafts probably spalled of fault scarps (a type of rock fall); channelised, mass-emplaced, conglomeratic sediments, associated with shelf facies, have only rarely been described (e.g. Lewis et al., 1980; Massari, 1984); those described by Massari (op. cit.) are very similar to this facies, and are interpreted as part of a marine fan-delta complex, building into a small, rapidly subsiding basin.

**Facies B2 - Sand-silt channels**

**Description:** this facies comprises large channel-shaped sediment bodies, containing thin, parallel-bedded, fine- to occasionally medium- or coarse-grained sands, with very thin, interbedded, very fine sands or silts (Plate 3.5a); beds are planar, horizontal to
occasionally gently dipping, with sharp, non-erosive tops and bases (Plate 3.5b); internally, beds are massive, faintly parallel-laminated, or sometimes normally-graded; rare wavy lamination is observed; coarser beds are sometimes pebbly (e.g. log 19/12/2, Fig. 3.6), and/or very shelly, with whole shells as well as fragmented material (Plate 3.5b); shells are often aligned parallel to bedding, or occasionally imbricated up-dip; pods of pebbles occur at the base or within channel fills sometimes, rarely with siltstone intraclasts; low-angle, slightly concave-up truncation surfaces cut through the sands (e.g. log 24/4, Fig. 3.4, and log 19/12/2, Fig. 3.6); bedding above these surfaces is either parallel or slightly inclined to the surfaces; at one locality, a small slump fold is developed immediately above a truncation surface, and is probably orientated along it; bioturbation is moderate to rare (Planolites-type burrows have been identified); occasionally, this facies becomes thicker-bedded and more massive; plant debris is sometimes common; at one locality (WD 202785), a series of stacked channels are filled entirely with silts; these silts are identical to the silts into which the channels are cut (facies A1), and are largely structureless; channel shapes are only identified from thin, cemented, cross-cutting horizons.

**Occurrence:** field data show that this facies occurs mainly in the mid to upper parts of the Nicosia Formation (Figs 3.4 and 3.5); a number of sand intervals are also recorded by borehole data; only simple lithological descriptions of these sands are available, so they may or may not be equivalent to facies B2; correlation of sand layers between boreholes and outcrop data is possible, however, (e.g. section C-C', Fig. 3.5), and suggests that at least some of these sands may belong to facies B2, and may be attached up-dip to the palaeoshoreline; sand layers do not correlate well along strike, suggesting ribbon-like geometries; this is also implied by field data; some sand layers appear isolated (e.g. Fig. 3.3), but this may be due to the line of section not being coincident with the axis of the sand lobe in question.

**Dimensions, lateral and vertical facies relations:** the facies occurs in large, channel-shaped bodies, cut into facies A1 silts, and less commonly into facies A2 sands; channels are broad, with gently to moderately-dipping margins; channel dimensions are not known
precisely, but most are at least 20m deep, and 100-200m wide; occasional smaller channels also occur; channel margins are sharp and well defined; channel fills fine distally, from fine to medium, occasionally pebbly sands to very fine sands, which are also more massive and less shelly; nested channel fills are observed; borehole data indicate individual channels may be several kms long.

**Palaeocurrents:** channel orientations in this facies are predominantly N-trending (Fig. 3.9).

**Interpretation:** like facies B1, this facies is interpreted as a number of elongate sediment lobes, cut across a marine shelf, seaward of coastal depocentres; deposition is again believed to have been largely from high density currents (see interpretation of facies B1b for more details); the facies is very similar to the sandy, turbidite-filled, shallow marine channel of Walker (1985a); low-angle truncation surfaces are interpreted as re-incision of filled or partially filled channels, followed by refilling; silt-filled channels suggest that occasionally, channels filled only very slowly with sediment from suspension, and that channel cutting and filling were largely separate events (cf. silt-filled channels of Dott and Bird, 1979); channel incision, as in facies B1, is attributed (at least initially) to the erosive powers of heavily-laden currents, crossing a narrow, steep, basin margin; channel incision is not common on modern shelves (Shephard and Dill, 1967; Walker, 1985a) and ancient, channel-shaped shelf sand bodies are only rarely documented (e.g. Dott and Bird, 1979; Lewis et al., 1980; Squires, 1981; Walker, op. cit.); incision in these examples is related to the presence of a narrow, steepish shelf, except for Walker (op. cit.), who invokes a sea level fall; it is possible that sea level fluctuations may also account for some of the facies B2 channels, particularly those in higher parts of the formation, which are of ?Upper Pliocene age; there is evidence that faulting had declined in the Mesaoria basin by this time (section 3.3.2, model 2), and the basin margin may no longer have been very steep; it is now well documented that eustatic sea level changes, associated with the Quaternary Ice Ages, began before the Pleistocene, in the the Upper Pliocene (e.g. from oxygen isotope data, Shackelton and Opdyke, 1977; see further discussion under section 3.3.2, model 2).
**Facies B3 - Sheet sands**

**Description:** the facies contains single, horizontal beds, 20cm-1m thick, of poorly sorted, muddy, shelly, coarse sand; bases and tops are sharp, planar to slightly wavy; beds are internally massive, or rarely, faintly parallel-laminated; the facies contain a large variety of marine fauna, with thick-shelled molluscs, up to 12cm across, sometimes preserved.

**Occurrence, lateral and vertical facies relations:** this minor facies is found near the base of the Nicosia Formation in the centre of the study area; it occurs as a number of beds separated by facies A1 silts, below sequences of facies B1 sediments; beds are laterally persistent for a few hundred metres.

**Interpretation:** the sediments are interpreted as unconfined, thin, high density turbidity flows; this is not a common facies, suggesting that once facies B1 channels were cut, mass flows tended to be confined to them.

**Facies C1 - Bioclastic sands**

**Description:** this facies comprises 1-3m thick units of parallel, very thin- to thin-, or occasionally medium-bedded, medium to coarse-grained sand, rich in bioclastic material; beds have sharp tops and bases, and are internally massive or parallel-laminated; some larger bioclastic fragments are aligned parallel to bedding; bases of units are mildly erosive, to, in one case, highly erosive and contain siltstone intraclasts; rare low-angle, large-scale truncation surfaces cut the units; slight slumping of beds is visible at the base of one unit, while at another, a fully developed slump fold is developed (log 6/12/6, Fig. 3.3; Plate 3.5c); this slump fold is directed northwards, down the inferred palaeoslope (Fig. 3.9); gross geometry of units is lenticular, or distinctly channel-shaped; channel dimensions are up to 3m thick, ca. 100m wide, and at least a few hundred metres long.

**Occurrence, lateral and vertical facies relations:** the facies is found at only two localities in the eastern part of the study area (Fig. 3.2), in the lower part of the Nicosia Formation (Fig. 3.3); sharp upper and lower contacts occur with the encasing facies A1 silts; the facies is located in a part of the basin where facies B1 does not
Palaeontology: bioclastic debris includes bivalve, red algal, echinoderm, and pelagic and benthic foraminiferal fragments; only small shells and foraminifera remain whole; the fauna represent a littoral to neritic assemblage.

Interpretation: thin, parallel-bedded sands, with occasional truncation surfaces, in channelised bodies, are features similar to facies B1b and B2 sands; a similar depositional process is inferred, i.e. shallow channel cutting on the steepish southern margin of the Mesaoria basin, and infill with sediment from high density currents (see further, facies B1b); slumping suggests rapid deposition on a steep slope, perhaps during storms; the difference in sand composition between this facies and facies B1b suggests that no clastic fan-delta lay shorewards of this facies, but that a shelly strandline may have been developed; the facies occurs just north of the area where the Miocene-Pliocene unconformity surface is believed to be less faulted, and which consequently may have been a less favourable site for fan-delta development (see section 3.2.2, reduced faulting to east subsection).

3.3.2 Sedimentological models

Facies analysis of the Nicosia Formation shows it to comprise an entirely marine sequence of sediments, deposited in the narrow Mesaoria basin. Depositional environments present, and progressive infilling of the basin, can be summarised in three evolutionary models, which represent the lower, middle and upper parts of the formation (note the large time span inferred for the middle Nicosia Formation, Fig. 3.10). These models are described below.

Model 1 - lower Nicosia Formation fan-deltas

Facies of the lower Nicosia Formation (A1, B1a-d, B3 and C1) document sedimentation in a small, elongate marine basin. This basin was bounded to the south by the faulted northern margin of the emerging Troodos Massif (Figs. 3.11a and b; section 3.2.2), and to the north, by a submerged ridge, the Kyrenia lineament (section 1.3.2).

Much of the basin filled with fossiliferous, calcareous silts
Fig. 3.10 - Summary chart for the Nicosia Formation, showing vertical facies variations (col. 1; see Encl. B for key), possible onset of eustatic effects (col. 2), relative decline in faulting intensity along the south (Troodos) basin margin (col. 3), changes in possible causes of channel incision (col. 4), and approximate trend in basin shallowing (col. 5; specific water depths unknown).
(facies A1), structureless due to intense bioturbation, and deposited largely below storm wave base. The maximum depth of deposition in the centre of the basin is uncertain, but was unlikely to have been below 500m. Facies A1 constitutes background suspension sedimentation in the basin.

In the south central part of the basin, facies B1d debris flow conglomerates document continued faulting of the northern margin of the Troodos Massif, which had begun earlier in the Miocene (section 3.2.2). This narrow, steep, fault-controlled margin also provided the location for the development of a number of small fan-deltas, which built out northwards into the basin (facies B1a-c and B3, Fig. 3.11b). The marine toes only of these fan-deltas are preserved. A narrow coastal plain and shoreline probably lay to the south, but evidence for these has now been eroded away. Shelly material present in facies B1a and b may have accumulated in coastal areas initially before being reworked into marine fan-delta lobes. Conglomerate clasts (facies B1a), which are both subrounded and subangular, may have been rounded to some extent in a shoreline setting, prior to reworking.

Sedimentation was dominated by mass flow processes, consistent with the location of fan-deltas along a narrow, steep margin. Some mass flows were capable of incision into underlying sediments, and facies B1a-c are now typically preserved in channel-shaped bodies (facies B1, section 3.3.1). Sea level fluctuations, which were to influence facies geometry later, were probably not operating at this time (Lower Pliocene), and are not believed to have affected fan-delta deposition.

The presence of discrete fan-delta lobes, not coalesced into a more continuous sheet, suggests that fan-deltas were small, perhaps only 1-2km across (Fig. 3.11a). This size, however, compares quite well with some modern fan-deltas, associated with steep, faulted margins e.g. the Red Sea (Hayward, 1985). In the Mesaoria basin, fan-deltas apparently exploited gaps in the fault block topography of the basin margin, and were localised along possible fault block intersections (section 3.2.2).

In the east of the study area, faulting of the basin margin was not so intense (section 3.2.2), relief was probably lower, and
Fig. 3.11a - Schematic palaeogeographic reconstruction for the lower Nicosia Formation, south side of the Mesaoria basin.

Fig. 3.11b - Schematic block diagram reconstruction for the lower Nicosia Formation, south side of the Mesaoria basin.
coarse-grained sediment apparently did not accumulate along the coastline. A shelly strandline may have developed instead (Figs. 3.11a and b). Periodic reworking of shelly material offshore, perhaps during storms, resulted in rapid deposition of a bioclastic-rich facies (facies C1). Once again, currents were able to erode into the shelf, and deposit at least some of facies C1 in channels.

The Nicosia Formation fan-deltas are difficult to classify because of lack of preservation of shoreline and subaerial facies (compare with the Kakkaristra Formation fan-delta, chapter 4). They probably fit best into the slope fan-delta category of Ethridge and Wescott (1984). Slope fan-deltas are typical of small basins with narrow margins, and contain subaqueous mass flow deposits. Submarine fans are often recognised seaward of slope fan-deltas. No such facies are identified in the Nicosia Formation, either from field or borehole data. Presumably, sediment supply from the small Troodos island, perhaps 100 x 30km in size, was insufficient to feed both fan-delta and submarine fan systems.

No direct modern analogue to the Nicosia Formation fan-deltas is known. Red Sea fan-deltas are not fronted by channels, although gravel is transported across the shelf (Hayward, 1985). Channels do occur in front of the Yallahs fan-delta (Wescott and Ethridge, 1980). This is a larger system, however, than the fan-deltas of the Nicosia Formation.

Model 2 - middle Nicosia Formation

Conglomeratic channel facies (facies B1) give way upwards to finer-grained channelised facies (facies B2; Fig. 3.10). This change may be due to several factors:

1) transgression, resulting in a general shift in depositional environment to a sandier, more distal channel setting, perhaps due to a eustatic sea level change; there is no evidence elsewhere in the Mediterranean, however, for sea level changes at this time (ca. Mid Pliocene).

2) climatic variations, or changes in source rock lithology, which might alter the type of sediment being supplied to the basin; again, there is no evidence in other parts of the Mediterranean, however, for major climatic changes in the Lower-Middle
Pliocene, and there is no evidence for a change in source rock.  
3) relative decline in uplift of the Troodos Massif, associated decrease in relief, and consequent supply of finer sediment to the shelf; coupled with lack of other evidence for continued faulting (e.g. the presence of syntectonic sediments similar to facies B1d), this is inferred to be the cause of the observed change.

Towards the middle of the Nicosia Formation, therefore, fault activity along the northern margin of the Troodos Massif, and uplift, are inferred to have declined, finer sediment was supplied to the Mesaoria basin, and a number of small, sandy deltas probably developed along the coast. Channels continued to be cut across the shelf, seaward of these deltas, implying continued existence of a rather steep margin. This can be related to continued subsidence of the centre of the basin, along large faults (see section 3.4.2), in spite of a decline in tectonic activity to the south.

A significant feature of the middle Nicosia Formation is the vertical stacking of sand-filled channel bodies (e.g. Fig. 3.5). Correlation of sand bodies laterally, from cross-section to cross-section (i.e. Figs. 3.3 to 3.7) is not particularly good, but three or perhaps four main episodes of channeling appear to have taken place. The reasons for this repetition are unclear, but could relate to repeated minor bursts of tectonic activity in the hinterland, generating pulses of sediment that were driven into the basin. Some channeling may relate to simple sedimentological control, i.e. abandonment and switching of small coastal depocentres, with consequent shift in the locus of sediment supply for channels.

A final possibility is that incision of the youngest channels was promoted by the start of glacioeustatic effects. This seems possible for two reasons. Firstly, it is now well established from oxygen isotope data (Shackleton and Opdyke, 1977), palynological studies (Suc, 1984), and the first occurrence of ice-rafting (Shackleton et al., 1984) that glaciation began to affect the northern hemisphere ca. 2.5 million years ago (i.e. Upper Pliocene; see section 2.3). The age of the sediments of the upper parts of the Nicosia Formation, including the upper middle Nicosia Formation of Fig. 3.10, is not known precisely, but may be Upper Pliocene (section 2.3). These sediments
may thus have been subject to the onset of glacioeustatic sea level changes. Secondly, the Mesaoria basin eventually shallowed in Upper Nicosia Formation times (see model 3), implying declining gradients along the basin margin. Development of channel bodies towards the top of the Nicosia Formation cannot thus be related to incision into steep basin slopes, as it was earlier, and eustatic fluctuations may now have become the dominant control (see Fig. 3.10; cf. shallow marine channel of Walker, 1985a).

In general, preservation of sand bodies implies that little reworking of shallow marine sediment, by storms or waves, took place. This is probably partly due to channel incision to depths below storm wave base, but may also reflect the relatively low energy basin setting, already implied by intense bioturbation of background silts (see facies A1, section 3.3.1).

**Model 3 - upper Nicosia Formation**

Facies A1 silts coarsen up towards the top of the Nicosia Formation, and the upper 20-100m of the formation comprise fine sands (facies A2). This documents the final shallowing-up of the basin (Fig. 3.10), as subsidence waned and was exceeded by sedimentation. Water depth was probably less than 50m. Intense bioturbation of facies A2 implies continuation of a relatively low energy setting. Signs of strong current activity are present however, e.g. thin conglomerates (facies A3), and swaley cross-stratification and reworked shell horizons in facies A2. These imply that much of the basin was now above storm wave base.

Channelised facies continue to be present in the upper part of the formation. Incision is inferred to have been caused by the onset of glacioeustatic effects towards the top of the middle Nicosia Formation (model 2), and these effects would be expected to continue into the upper part. The rather rapid coarsening of facies A1 into facies A2, over only a few metres is perhaps due to enhancement of basin shallowing by falling sea level.

By upper Nicosia Formation times, thus, the Mesaoria basin had shallowed to above storm wave base, and had evolved into a sandy, shallow marine platform. Sea level fluctuations were very likely affecting it.
3.4 Basin Structure

Facies trends, thickness distribution and structural data from the Pliocene Nicosia Formation, the oldest and thickest formation studied in the Mesaoria basin, contain important information with respect to the early structural configuration of the basin, and its subsequent structural evolution. These aspects are now discussed. As inception of the basin occurred in the Upper Miocene, the pre-Pliocene setting of the area is first outlined.

3.4.1 Pre-Pliocene setting

The Troodos Massif began to be uplifted in approximately Oligocene times (section 1.3.3). By the Upper Miocene, it formed a barely emerging island (Fig. 3.13), and its northern flank had begun to be affected by a phase of normal faulting (section 1.3.3; Follows and Robertson, in press). On the northern side of the Mesaoria Plain area, strong crustal extension had been affecting the Kyrenia lineament and Cilicia basin since the Oligocene, and the region had subsided dramatically (section 1.3.2; Robertson and Woodcock, 1986). A thick turbidite sequence (Kythrea flysch) was deposited as a result. By the Upper Miocene, however, facies changes and thickness trends indicate that an important extensional fault (the Kythrea fault; Fig. 3.12) became active in the Kyrenia area. As a result, the Kyrenia lineament began to subside less rapidly (Baroz, 1979; Robertson and Woodcock, 1986), and began to take on the form of a submerged ridge (Fig. 3.13). Later, in the Pleistocene, the Kythrea fault underwent reactivation under compression, and now crops out as a high-angle thrust.

There is also evidence that another important fault, now buried beneath the Mesaoria Plain and here termed the Mesaoria fault, was also active at the end of the Miocene. This fault has been identified from mapping of the base of the Pliocene, across the Mesaoria Plain, using borehole data. Although the data base is rather scattered, the base Pliocene map (Fig. 3.14b) clearly shows an important break north of the Xeri deep well. Two exploration wells (KL-1 and LEF-1, Cleintaur et al., 1977) penetrated an important, but reversed, fault in
Fig. 3.12 – Major structural elements and geological units, Cyprus.
S

TROODOS MASSIF
ophiolite basement forming locally emerging island

N. TROODOS FLANK
E. Tert-Mio pelagic chalk & marl, and evaps.; affected by normal faulting in Mio., which controlled facies and thickness trends

KYRENIA LINEAMENT
narrow Perm-Eo thrust belt; deeply subsided in Olig-Mio, covered with Kythrea flysch; began to rise in U Mio, as growth faults became active

Fig. 3.13 - Pre-Pliocene setting of the Troodos Massif: schematic N-S structural cross-section (based on data from this study, Cyprus Geological Survey geological map of Cyprus, Cleintaur et al., 1977, Robertson and Woodcock, 1986, and Eaton, 1987; see Fig. 3.12 for section location and Encl. B for key).
Fig. 3.14a - Thickness map of the Nicosia Formation, constructed from Cyprus Geological Survey borehole data; thicknesses are in metres.

Fig. 3.14b - Map of the base of the Nicosia Formation, constructed from Cyprus Geological Survey borehole data and outcrop data (crosses); depths are in metres.
the area of this break. Upper Miocene evaporites are thicker on the southern side of this fault, and also contain halite in addition to gypsum (Cleintaur et al., op. cit.; Fig. 3.15). These thicker evaporite accumulations suggest that, like the Kythrea fault, this fault may have been a south-down growth fault during the Messinian, with deposition of halite on its southern, deeper water side. Like the Kythrea fault, it was later reactivated under compression as a high-angle thrust.

Another important thrust, the Ovgos fault, is exposed in the Mesaoria Plain between the Mesaoria and Kythrea faults (Fig. 3.12). It may also have been a growth fault during the Upper Miocene, although there is no direct evidence for this.

At least two, and possibly three, important south-down faults thus bordered the southern side of the Kyrenia lineament in the Upper Miocene, while smaller, north-down faults bordered the northern side of the Troodos Massif. Together, these faults delineated a half-graben structure in the Mesaoria Plain area (Fig. 3.13), defining the geometry of the infant Mesaoria basin. Extensional faulting was also affecting western Cyprus at this time (Paphos-Polis graben, Ward and Robertson, 1987).

It must be noted that the Kythrea and Ovgos faults now crop out as north-dipping, high-angle thrusts (Fig. 3.12) and Cleintaur et al. (1977) depict the Mesaoria fault as a north-dipping thrust on their cross-section (Fig. 3.15), whereas the proposed earlier growth faults dipped south (Fig. 3.13). The faults are believed to have been reactivated by compressive forces in the Pleistocene (see chapter 5). In an idealised orthogonal stress system, however, in which faults cut through uniform, isotropic material, reactivation of normal faults under compression would not be expected to cause a reversal in the dip directions of fault planes.

The true shapes of the Kythrea and Ovgos faults at depth are unknown, however, and the configuration of the Mesaoria fault in Fig. 3.15 is questionable. Furthermore, the northern part of the Mesaoria basin does not conform to the idealised setting outlined above. An important basement lineament (the probable northern margin of the Troodos microplate) underlies the area (Fig. 3.13; see sections 1.3.2 and 11.2.1, influence of pre-existing crustal structure
Fig. 3.15 - Structural, N-S cross-section across the Mesaoria Plain, incorporating exploration well KL-1 (from Cleintaur et al., 1977).

Fig. 3.16 - Diagrammatic N-S cross-section across the Mesaoria half-graben. B,C,D,E - biostratigraphic zones of Mantis (1968; schematic only).
This exerted a significant control on deformation during an earlier phase of compression, when the structurally weak Kyrenia lineament was first thrust against, and partially over, rigid Troodos basement in the Eocene (Robertson and Woodcock, 1986). It is likely again to have had an influence on deformation during Pleistocene compression, very possibly leading to distortion of existing fault planes as they were reactivated.

In addition, stresses during reactivation were not necessarily purely orthogonal and may have involved elements of strike-slip (see further, section 11.2.2, Mid-Upper Miocene subsection). Horizontal and vertical movements along faults may have further contributed to fault plane distortion. Finally, sedimentary evidence indicates that fault reactivation took place in at least two phases (during Athalassa Formation and Fanglomerate times; sections 5.4.2 and 7.4). Complex interactions between faults reactivated at different times may also have resulted in buckling and disruption to fault planes. Curved and distorted fault plane geometries are apparent on seismic sections from other tectonically reactivated areas. An example of this from a former extensional setting, reactivated during compression, is the Rhône graben (Zeigler, 1983).

As the deep structure of the Kyrenia lineament is unknown, the interpretation outlined must remain tentative. Good evidence that the Mesaoria fault at least was a south-down growth fault during its early history comes from the Nicosia Formation, however, and is described in the following section.

In summary, it is apparent that while strong crustal extension had affected the Kyrenia-Cilicia basin area during the Oligocene-Mid Miocene, more localised extension began to affect the more southerly Troodos-Mesaoria-Kyrenia area by the Upper Miocene, and a half-graben began to form. Whilst some deepening had taken place by the end of the Miocene, dramatic subsidence was to follow in the Pliocene.

3.4.2 Pliocene Extension

Following the inception of the basin in the Upper Miocene, several lines of evidence reveal that extension continued into the Pliocene, and was in operation during much of the deposition of the
Nicosia Formation. Syntectonic sediments and fan-delta facies in the lower part of the formation from the southern margin of the basin attest to continued faulting in the south (sections 3.2.2 and 3.3.2, model 1), while thickness and facies trends elsewhere document subsidence in the north. These trends are now described.

The thickness distribution of the formation has been mapped from borehole data (Fig. 3.14a), and shows thickening of the formation away from the southern margin of the basin, thinning across the Mesaoria fault, and finally thickening again. Micropalaeontological data from the south side of the basin (Mantis, 1968; Baroz, 1979) show that a condensed section is present in this area, and that thinning of the formation is not due to erosion. These trends delineate a wedge-shaped basin fill, depicted schematically in Fig. 3.16. This is typical of half-graben basin geometry, where maximum subsidence, and thickest sediment accumulation, occur along half-graben bounding faults.

It is thus apparent that subsidence continued in the Mesaoria basin in the Pliocene, along the Mesaoria and Ovgos half-graben bounding faults, while antithetic faulting continued in the south (Figs. 3.16 and 3.17). It is uncertain if growth continued along the Kythrea fault to the north.

Footwall uplift is often an important source of sediment in extensional half-grabens (Leeder and Gawthorpe, 1987). In the Mesaoria basin, however, the footwall area, i.e. Kyrenia allochthon (Fig. 3.16), remained largely submerged, as witnessed by the apparent negligible clastic input from that side of the basin (although it is possible that small volumes of such sediments are present, but have not been penetrated by boreholes). Instead, large quantities of silty sediment (facies A1) accumulated (Fig. 3.17), reaching over 800m in thickness (e.g. Xeri deep well, Figs. 3.4 and 3.14a). This lack of coarse-grained facies along half-graben master faults has been documented by other workers, e.g. by Frostick et al. (1988) from modern and ancient continental rifts (in their model, back-tilting of the footwall block results in diversion of sediment away from the developing half-graben).

In contrast, significant quantities of coarse-grained clastics were shed from the Troodos Massif on the south side of the basin (facies
AYIA MAVRI LINEAMENT
minor thrust zone, still active in L. Plio.

TROODOS MASSIF
low-lying, vegetated island; main sediment source

MESAORIA BASIN
sunk along growth faults in northern part of basin; relative uplift of southern margin, with influx of coarse clastics, associated with antithetic faulting

KYRENIA LINEAMENT
uplifted along growth faults, but still submarine

Akrotiri High
no longer active

MARI BASIN
slight deepening ahead of thrust zone

Fig. 3.17 – L.-M. Pliocene setting of the Troodos Massif: schematic N–S structural cross-section (based on data from this study, Cyprus Geological Survey geologic map of Cyprus, Cleintaur et al., 1977, Robertson and Woodcock, 1986, and Eaton, 1987; see Fig. 3.12 for section location and Encl. B for key).
B1, B2 and B3), largely in response to antithetic faulting and associated uplift (Fig. 3.17). In the continental rift example quoted above, coarse clastics are also found in this setting, in facies overlying the hanging wall area.

The initial depth of the deepest part of the basin is uncertain, but was probably less than 500m (see facies A1). Basin subsidence and sediment loading thus presumably combined to allow in excess of 900m of sediment to accumulate in the Pliocene. As no deepening-upwards trends are documented in these sediments, sedimentation rates must have largely kept pace with subsidence. Mantis (1968) subdivided Pliocene borehole successions into five informal biostratigraphic zones (B - E). Cross-sections across the eastern Mesaoria Plain, incorporating Mantis' scheme (Zomenis, 1972), indicate thickening mainly of units C and D across the Mesaoria fault (Fig. 3.16), suggesting the main phase of extension may have been during deposition of these units. Zonation data from Mantis (1968) and Baroz (1979) also show that a number of zones are missing along the southern margin of the basin, further evidence for uplift and erosion, or non-deposition, along an active fault-controlled margin.

Antithetic faulting and uplift of the Troodos Massif began to decline towards the middle of the Nicosia Formation as conglomeratic facies (facies B1) were replaced with finer, sandy sediments (facies B2) along the south margin of the basin (see model 2, section 3.3.2). Basin subsidence continued briefly, but eventually ceased in upper Nicosia Formation times, as the whole basin shallowed, and filled with sandy facies (facies A2; see model 3, section 3.3.2). By the end of the Pliocene, a broad sandy platform extended between the Troodos Massif island and the still largely submerged Kyrenia lineament, and the Mesaoria area represented a relatively inactive, largely infilled basin, occupied by a shallow sea.

3.5 Summary - Basin Evolution during Nicosia Formation Times

1. Prior to the Pliocene, a narrow basin began to form in the area of the Mesaoria Plain. Along its southern margin, the rising Troodos Massif was being affected by normal faulting. To the north, the Kyrenia lineament, which had previously been deeply
submerged, began to rise along important south-down growth faults.

2. Together, the south-down growth faults and north-dipping antithetic faults delineated a half-graben structure, defining the geometry of the infant Mesaoria basin.

3. Following the Messinian salinity crisis, seas rapidly transgressed the area, as elsewhere in the Mediterranean, and marine silts (facies A1) were deposited in the basin, directly over the Miocene-Pliocene unconformity surface.

4. Faulting and subsidence continued into the Pliocene, and over 900m of silty facies accumulated in the deepest, northern half of the basin, next to growth faults.

5. Little clastic input is evident from the north side of the basin during the Pliocene, suggesting that the Kyrenia lineament remained largely submerged.

6. In contrast, antithetic faulting along the Troodos margin to the south prompted the progradation of a number of small fan-deltas into the basin. Only the marine toes of these fan-deltas are now preserved. Their facies comprise a variety of coarse, channelised, mass flow-dominated clastics (facies B1), cut across the steep, faulted, southern margin of the basin.

7. Conglomeratic facies in the southern part of the basin are replaced towards the middle of the Nicosia Formation by finer, sandier sediments (facies B2), reflecting declining fault activity and uplift along this margin of the basin, and associated lowering of relief. These sediments continued to be confined to channels. Channeling reflects the continued presence of steepish sea floor gradients, as subsidence of the narrow Mesaoria basin was sustained along half-graben bounding faults to the north. The onset of glacioeustatic sea level fluctuations cannot, however, be ruled out as a further control on channel incision.

8. Towards the end of Nicosia Formation times (Upper Pliocene), subsidence declined, the whole basin shallowed, and fine sandy facies (facies A2) replaced silty sediments. Evidence from this facies, and the presence of marine conglomerates (facies A3), indicate that some storm-reworking was taking place, implying
the basin floor was now largely above storm wave base. The continued presence of channelised facies in the now shallow basin is attributed to channel incision during periods of lowered sea level.

9. By the end of the Nicosia Formation, basin subsidence had ceased, and the Mesaoria area had evolved into a relatively inactive, largely infilled basin, occupied by a shallow sea.
Plate 3.1

a) Facies A1 silts, exposed along the sides of the Argaki tou Perati valley, east of Pera; weathering gives typical badlands type topography; view looking east

b) Facies A2 sands, showing thin, discontinuous, bed-parallel, cemented layers, alternating with uncemented, structureless intervals; staff is 110cm long

c) Oyster beds in facies A2 sands; oyster valves are often horizontally aligned; cliff is ca. 3m high
Plate 3.2

a) Photo montage of cliff face at Khrysospilliotissa chapel (WD 257835), showing channelised contact (dashed line) between facies A1 silts and A2 sands; refer to Fig. 3.8 for further description.

b) Photo montage of facies B1 conglomerate- and sand-filled channel, southeast of Politiko (WD 218758); base of channel is marked by the dashed line; channel is filled with lenticular units of facies B1a conglomerates and facies B1b sand-silts, and is cut into facies A1 silts; cliff is ca. 25m high.
Plate 3.3

a) Thin horizon of facies A3 conglomerates (2/3 of the way up cliff face), embedded in structureless facies A2 sands; Kakkaristra Formation overlies Nicosia Formation at cliff top; cliff is ca. 20m high

b) Close-up of facies A3 conglomerate, showing subrounded, moderately sorted clasts and lack of consistent imbrication; hammer is 30 cm long

c) Bedded Lefkara Formation (pale) unconformably overlain by facies A1 silts of the Nicosia Formation (dark); thin wedge of Lefkara-derived facies B1d conglomerate (pale) is interbedded with silts towards top of valley side
Plate 3.4

a) Upper half: graded, partially imbricated, conglomerate bed of facies Bla. Lower half: horizontally- to lenticularly-stratified, pebbly coarse sands of facies Bla, with huge, outsize clast of reef limestone; field of view is ca. 3m vertically

b) Upper half: facies Blc megaconglomerate, showing huge, reef limestone clast, and flat base to bed. Lower half: horizontally-stratified facies Bla pebbly coarse sands; hammer on left in middle is 30cm long

c) Lenticularly-shaped units of facies Bla conglomerates and facies Blb sands; conglomerates have poorly sorted, subangular to subrounded, unimbricated clasts; bases to wedges in sands are formed either by conglomerate tops (e.g. right hand side) or shallow, truncation surfaces (e.g. middle, left hand side); bedding within sands parallels wedge bases; hammer in lower middle of photo is 30cm long
Plate 3.5

a) Facies B2 sand-silt-filled channel, cut into facies A1 silts; channel fill is largely parallel-bedded; cliff is ca. 15m high

b) Close-up of facies B2 sand-silts, showing parallel-stratified, sometimes very shelly, internally structureless or parallel-laminated, sand beds; lens cap is 6cm across

c) Slump fold in thin-bedded facies C1 bioclastic sands; staff is 110cm long
CHAPTER 4 - KAKKARISTRA FORMATION

4.1 Introduction

The Kakkaristra Formation (sections 2.2.2 and 2.4) comprises a 10-15m thick mappable unit, located between the underlying Nicosia Formation and the overlying Apalos Formation (see Table 2.2). It outcrops in an E-W trending belt, across the centre of the study area (Fig. 4.1, Encl. A). Exposure is most extensive in the east. The formation does not occur in the extreme east of the study area, or on the northern (Kyrenia) side of the Mesaoria basin, where its lateral equivalent, the Athalassa Formation (chapter 5), is found. The age of the formation is uncertain, but from stratigraphic and limited palaeontological data, is ?Upper Pliocene-Lower Pleistocene (section 2.3).

Although occupying only a minor part of the sedimentary fill of the Mesaoria basin (see Table 2.2), the Kakkaristra Formation comprises a wide range of facies, representing deposition in a complex fan-deltaic setting (section 4.4.1). Lateral facies changes are rapid, and correlation between even closely spaced exposures can be difficult. As will be discussed later (sections 4.4.1 and 4.4.2), facies of the subaerial part of the fan-delta are poorly preserved (four localities only). Apart from at the base of the formation, these subaerial facies are similar to those of the overlying Apalos Formation. No consistent lithological means of separating the Kakkaristra and Apalos Formations at these localities was identified. With no palaeontological data available, the lower ca. 12m of these subaerial facies are arbitrarily assigned to the Kakkaristra Formation (12m is the average thickness of the Kakkaristra Formation), and are described in this chapter (see also section 2.2.2).

4.2 Basal Relations

4.2.1 Observations

Several different types of contact have been recognised between the Kakkaristra Formation and the underlying Nicosia Formation (Fig. 4.1):
Fig. 4.1 - Main areas of outcrop of the Kakkaristra Formation, and lower contact types. Outcrop of the Athalassa Formation (stippled) is from the Cyprus Geological Survey geological map of Cyprus.
an unconformity (type not specified) was recognised by Wilson (1959) in the far west of the Mesaoria Plain (outwith the present study area), between his Fanglomerate and Pliocene marls; his description of the Fanglomerate suggests that the lower part of it is equivalent to the Kakkaristra Formation (see section 2.2.2).

b) in the Potami area (Fig. 4.1), the contact is not exposed; differences in topographic elevation of exposed sections of the Kakkaristra Formation, and Nicosia and Apalos Formations, suggest that young faulting affects this area.

c) a distinct angular unconformity of $5^\circ - 10^\circ$ is recognised at three localities in the west and central parts of the study area (Fig. 4.1).

d) fluvial facies of the Kakkaristra Formation lie in erosive contact with open marine sediments of the Nicosia Formation at localities in the west and central parts of the study area (Fig. 4.1); the facies changes at these localities are not simply due to facies transitions between adjacent depositional environments, and the contacts between them are disconformities (Dunbar and Rodgers, 1957; Collinson and Thompson, 1982; cf. Clifton, 1981; Kleinspehn et al., 1984).

e) in the east central and eastern parts of the study area, coastal Kakkaristra Formation facies have channeled erosively into marine Nicosia Formation facies; facies changes here may simply be due to the rapid progradation of the Kakkaristra depositional system over the Nicosia Formation; the close proximity of two of these contacts to an angular unconformity (Fig. 4.1) suggests, however, a possible time gap may be present, and these contacts may also constitute disconformities.

f) non-erosive contacts, but significant facies changes, are recorded between the Kakkaristra and Nicosia Formations predominantly in the east part of the study area; no time gap is necessarily involved at these localities.

g) in the far east of the study area (Fig. 4.1), continuous sedimentation is recorded between the Nicosia Formation and the lateral equivalent to the Kakkaristra Formation, the Athalasssa Formation.
4.2.2 Interpretation

The distribution of contact types between the Kakkaristra and Nicosia Formations (Fig. 4.1) indicates that a partial unconformity exists between the two formations. This unconformity is of angular type (5° - 10°) at three localities in the south (Fig. 4.1). East and west of these localities, disconformities are identified. To the north, and in the far east of the study area, however, the unconformity disappears, and is replaced by only a facies change (Fig. 4.1).

The presence of an angular unconformity is important, because it strongly suggests that a period of tectonism separates the Nicosia and Kakkaristra Formations. Other contact types (e.g. incision of coastal Kakkaristra facies into marine Nicosia silts; observation e) of previous section) could simply be due to intrabasinal, sedimentological effects, or climatic or eustatic fluctuations. Climatic and eustatic variations were very likely occurring (see interpretation of the underlying, upper Nicosia Formation, section 3.3.3), but could not, on their own, generate an angular unconformity. They served to augment, or perhaps retard, tectonic effects, depending on whether sea level was rising or falling at the time of onset of tectonism.

A pulse of uplift is thus believed to have affected the Troodos Massif at the end of deposition of the Nicosia Formation. The sedimentary cover overlying the northern flank of the ophiolite was also uplifted, and its sediments, including those of the Nicosia Formation, tilted. A major influx of detrital sediment was generated, and deposition of the Kakkaristra fan-delta initiated.

Tectonism was very localised, however, and its effects diminished rapidly away from the focus of uplift in the Troodos Massif. Thus, the Kakkaristra Formation was deposited with angular unconformity over tilted Nicosia strata along the southern margin of the basin, then with an erosive contact only further north, where no tilting is recorded (disconformities), and finally, with only a marked facies change. If changes in contact type between the Nicosia and Kakkaristra Formations are assumed to have been distributed radially around the centre of uplift, then this centre probably lay to the southwest of the present study area (see Fig. 4.1).

The type of contact relationship recorded between the
Kakkaristra and Nicosia Formations is similar to the progressive, or intrabasinal, unconformities, described from the unstable margins of other sedimentary basins, e.g. Tertiary Ebro basin, Spain (Riba, 1976) or Meso-Hellenic basin, Greece (Ori and Roveri, 1987). Both these examples come from compressional settings, for which this type of unconformity is characteristic according to Miall (1978). A compressional origin for tectonism is consistent with evidence from the north side of the Mesaoria basin (discussed in the following chapter; section 5.4.2). There, reactivation of former growth faults is recorded by the Athalassa Formation (the lateral equivalent to the Kakkaristra), as the Kyrenia lineament came under compression. It is suggested that complimentary movement to the south caused uplift of the Troodos Massif and southern margin of the basin, and generation of the Kakkaristra fan-delta complex.

Little evidence for faulting associated with uplift is present. Faulting is likely to have been concentrated within, and along the flanks of, the uplifted Troodos area. Proximal facies of the Kakkaristra Formation, deposited along the northern flank of the Troodos Massif, would be most likely to show fault-related effects. They are very poorly exposed, however (Fig. 4.1), and have largely been removed by erosion.

A further tentative piece of evidence in support of uplift of the Troodos Massif during Kakkaristra Formation times, or at least of the presence of a mountainous hinterland, comes from plant material recovered from the formation (pine cone and well preserved leaf fragment, facies C2). Possible living equivalents to the trees from which this material came (see Appendix 1), imply that a rugged, ridge and valley-type terrain existed in the hinterland during deposition of the Kakkaristra Formation.

Renewed uplift of the Troodos Massif is thus inferred to have taken place at the end of the Pliocene. This pulse of tectonism was sufficient to generate a partial or progressive unconformity and a significant change in the depositional nature of the basin, but did not result in large-scale restructuring.
4.3 Facies and Facies Relationships

A large number of lithofacies has been recognised in the Kakkaristra Formation (Table 4.1), some of which appear commonly, and others of which are of very limited occurrence. Broadly, the facies are interpreted to have been deposited in a fan-delataic environment, and are grouped together according to gross overall position on the fan-delta. This has led to an alluvial fan / shoreline / delta front (including bays) subdivision, with a separate group for minor, mud-dominated facies. Lithofacies are described below.

Facies A1 - Matrix-supported conglomerate

Description: the facies comprises 1-5m thick beds of poorly sorted, pebble-cobble conglomerate, which are massive and ungraded, or crudely horizontally-layered; fabric is mainly matrix-supported and unimbricated, though clast-supported horizons do occur in layered exposures; thin sandy layers may also be present; bed bases are flat to rarely slightly scoured; a sandy, fining-up unit, with pebbly cross strata, caps one massive bed; matrix is reddish brown, poorly sorted, muddy sand; at one locality, intraformational clasts of the underlying, upper Nicosia Formation sandstone are present; the facies is unfossiliferous.

Occurrence, lateral and vertical facies relations: the facies is of very limited occurrence; it is found at the base of the Kakkaristra Formation, where it unconformably overlies the Nicosia Formation in the southwest part of the study area (see Figs. 4.2, 4.7 and 4.8); it is overlain by another fluvial Kakkaristra Formation facies (A2); the facies has a probable sheet-like geometry, up to 5m thick, and of unknown lateral extent.

Interpretation: from association with facies A2, lack of fossils, and reddish brown colour, the facies is interpreted as fluvial in origin; general lack of internal organisation and non-erosive bases suggest it is the product of subaerial debris flow processes; debris flows were probably non-cohesive, because of the lack of mud matrix, and occasionally slightly turbulent, leading to minor scouring (Nemec and Steel, 1984); crude layering may be the result of surging debris flows (Nemec and Steel, op. cit.); sandy caps reflect
Table 4.1 - Lithofacies, Kakkaristra Formation

<table>
<thead>
<tr>
<th>Code</th>
<th>Lithofacies</th>
<th>Subenvironment</th>
<th>Environment</th>
</tr>
</thead>
<tbody>
<tr>
<td>A1</td>
<td>Matrix-supported conglomerate</td>
<td>Debris flow</td>
<td>?Alluvial fan</td>
</tr>
<tr>
<td>A2</td>
<td>Massive, reddened, sandy silts and minor</td>
<td>Floodplain fines and minor sheet flows</td>
<td></td>
</tr>
<tr>
<td></td>
<td>conglomerate</td>
<td>Distal braided channels</td>
<td></td>
</tr>
<tr>
<td>A3</td>
<td>Channelised conglomerates and sands</td>
<td></td>
<td></td>
</tr>
<tr>
<td>B1</td>
<td>Clean sands</td>
<td>Shoreface/foreshore</td>
<td></td>
</tr>
<tr>
<td>B2</td>
<td>Conglomerate layers</td>
<td>Fluvial shoreline flooding</td>
<td>Fan-delta shoreline</td>
</tr>
<tr>
<td>B3</td>
<td>Trough cross-laminated sands</td>
<td>Fluvial shoreline flooding</td>
<td></td>
</tr>
<tr>
<td>B4</td>
<td>High-angle, cross-bedded conglomerates</td>
<td>Gilbert-type river mouth bars</td>
<td></td>
</tr>
<tr>
<td>B5</td>
<td>Low-angle, cross-bedded conglomerates</td>
<td>Well washed, gravely foreshore</td>
<td></td>
</tr>
<tr>
<td>B6</td>
<td>Sorted, poorly stratified conglomerates</td>
<td>Poorly washed, gravely foreshore</td>
<td></td>
</tr>
<tr>
<td>C1</td>
<td>Massive, muddy sands</td>
<td>Bay fill</td>
<td>Fan-delta front + Bays</td>
</tr>
<tr>
<td>C2</td>
<td>Rootlet-bearing sands</td>
<td>Bay margin</td>
<td></td>
</tr>
<tr>
<td>C3a</td>
<td>Lenticular-bedded sands</td>
<td>?Storm-affected delta front (?HCS)</td>
<td></td>
</tr>
<tr>
<td>C3b</td>
<td>Poorly bedded sands</td>
<td>Bay fill/delta front</td>
<td></td>
</tr>
<tr>
<td>C4</td>
<td>Cross-bedded, bar sands</td>
<td>Subtidal bars</td>
<td></td>
</tr>
<tr>
<td>C5</td>
<td>Conglomeratic, shelly sands</td>
<td>Distal mouth bars</td>
<td></td>
</tr>
<tr>
<td>D1</td>
<td>Thin, white muds</td>
<td>Protected foreshore</td>
<td>(Muddy facies)</td>
</tr>
<tr>
<td>D2</td>
<td>Ostracodal, green and brown muds</td>
<td>Backshore ponds</td>
<td></td>
</tr>
<tr>
<td>D3</td>
<td>Muddy, calichied sediments</td>
<td>Lagoon/exposed foreshore</td>
<td></td>
</tr>
</tbody>
</table>
Fig. 4.2 - Facies distribution, Kakkariatra Formation, and fan-delta shoreline reconstruction (see Fig. 4.1 for map location and Table 4.1 for lithofacies).

Fig. 4.3 - Palaeocurrent data, Kakkariatra Formation; some individual readings are shown on the map with small arrows (see Fig. 4.1 for map location).
Fig. 4.8 - Sedimentological logs, cross-section E-E' (see Fig. 4.2 for section location and Encl. B for key to logs).

Fig. 4.9 - Sedimentological logs, cross-section F-F' (see Fig. 4.2 for section location and Encl. B for key to logs).
Figs. 4.4 & 4.5—Sedimentological logs and interpreted cross-sections A-A' & B-B' (see Fig. 4.2 for section location, Fig. 4.6 for section abbreviations and Encl. B for key to logs).
Figs. 4.6 & 4.7 - Sedimentological logs and interpreted cross-sections C-C' & D-D' (see Fig. 4.2 for section location and Encl. B for key to logs).
deposition during waning flow; the occurrence of this facies at the base of the Kakkaristra Formation may reflect initial high-energy events, associated with the pulse of uplift that occurred at the beginning of Kakkaristra Formation times (section 4.2.2).

**Facies A2 - Reddened, massive sandy silts**

Description: the facies contains pink or reddish, pale to mid brown gritty silts or occasionally very fine sands, in thick, massive units; pale horizons have a weak, calcareous cementation (?immature caliche); tiny rootlets are common; occasional thin (1-10cm) layers of coarse, sometimes pebbly, sand occur; rarely, these form small, shallow, channel-shaped lenses, less than 50cm thick, and a few meters wide; laterally more persistent, thin, channelised bodies with very irregular, scoured bases and filled with poorly sorted, unstratified, ungraded, clast-supported, sandy pebble conglomerate, occur at one locality (log 1, Fig. 4.7).

Occurrence, lateral and vertical facies relations: the facies is found at the same few localities as facies A1 (Fig. 4.2), overlying facies A1, in units up to 9m thick (although this limit is due to arbitrarily picking the top of the Kakkaristra Formation ca. 12m above the base of the formation, at localities where the formation comprises fluvial facies; see section 4.1).

Interpretation: the association with facies A1, red colouration, presence of rootlet holes, and lack of fossils point to a subaerial environment; the predominantly fine-grained, massive nature of the facies suggests mainly overbank deposition; thin intercalations of coarser sands may be minor crevassse splay-type deposits or sheet flows, also deposited in interchannel areas; the thin, unstratified conglomerate bodies may represent the distal ends of coarse-grained, erosive sheet flows; in the absence of associated, laterally continuous, cross-beded conglomerates and sands, typical of braided fluvial deposits, a medial or distal alluvial fan setting is indicated for this facies; lack of exposure of the facies, however, makes this interpretation rather tentative.

**Facies A3 - Channelised conglomerates and sands**

Description: the facies comprises a heterogeneous group of
interbedded conglomerates, pebbly sands and sands, in cross-bedded and channelised units; a range of structures are present (refer to Fig. 4.11): large-scale, concave-up, cross-bedded, clast-supported pebble conglomerates and pebbly sands, which are occasionally shelly and may show up-dip imbrication; convex-up, moderately bedded, moderately sorted, clast-supported, pebble conglomerates interdigitating with poorly stratified sands; normally graded, sandy, clast-supported pebble conglomerates fining up into massive sands, in channel-shaped bodies; lenses of sorted, matrix-free, small pebble conglomerate; and thick lenses of trough cross-laminated sands; these subfacies interdigitate laterally and vertically (Fig. 4.11).

Occurrence, lateral and vertical facies relations: this facies is found only at three localities (e.g. log 5, Fig. 4.7; Fig. 4.2); it overlies facies B1, B4 and B5, and is overlain by facies B4, or by nothing; a channel-shaped geometry can be inferred for this facies at one locality; channel dimensions are 4-5m deep, ?several hundreds of metres wide.

Interpretation: the general clast-supported fabric, channel- and bar-shaped geometries for subfacies, and possible overall channel geometry, suggest in-channel, braided fluvial deposits; they resemble, for example, part of the upper Cannes de Roche Formation (Rust, 1978), interpreted as distal braided river deposits; association with facies B2, B4 and B5, and occasional shelly nature, imply a fluvial-dominated coastal setting; shells were presumably washed in by large waves, but wave reworking is not apparent, and sedimentation probably took place in fluvial distributaries just landward of the shoreline (supratidal setting); deposition was predominantly by traction current processes, associated with sandy and gravelly braid bar or megaripple migration in channels; fining-up sequences represent filling of small channels between braid bars; fine, matrix-free conglomerates may represent winnowing and reworking of older fluvial deposits (cf. Kleinspehn et al., 1984).

Facies B1 - Clean sands

Description: this facies contains fine, or very fine, matrix-free, well sorted sands, which are often massive in appearance, but this is partly due to weathering; the main sedimentary structure in the
Fig. 4.10 - Sedimentological logs, cross-section G-G' (see Fig. 4.2 for section location and End. B for key to logs).

Fig. 4.11 - Composite field sketch of facies A3.
lower part of the facies is medium-scale, trough cross-lamination (troughs up to 70cm wide and 12cm thick); palaeocurrents are variable, but generally SE-SW (i.e. shoreward-directed; Fig. 4.3); in the upper part of the facies, structures are more variable, and include both on- and off-shore directed shallow-dipping, planar cross-lamination, high-angle, onshore-directed tabular cross-lamination, parallel lamination and heavy mineral lamination (Plate 4.1a); these structures interdigitate, with no discernible pattern; the upper part of the facies is also distinctly pebbly, with granules and small pebbles concentrated along bedding planes, or occasionally along foresets, or scattered throughout (Plate 4.1b); occasional very fine sand to silt interbeds display diffuse rippling of probable wave origin, contain sandy flasers and may be bioturbated; rarely ripple cross-lamination is preserved, and dips landwards; the facies is generally unfossiliferous, although rare oyster and other bivalve fragments occur; no in situ fauna were observed; bioturbation is uncommon, and comprises unidentified subhorizontal burrows, or at one locality, possible Thalassinoides; occasionally, pebbles occur in massive sands in a vertical arrangement (Plate 4.1b), which may be due to burrowing (cf. Phillips, 1984).

Occurrence, lateral and vertical facies relations: this is a common facies, particularly in the east part of the study area (Fig. 4.2), where it can comprise units up to 12m thick and occupy most of the thickness of the Kakkaristra Formation (e.g. log 2, Fig. 4.6); the facies interdigitates with a large number of other facies (see Figs. 4.4-4.10, and facies relationship diagram, Fig. 4.16); it is most characteristically associated with facies B2, B4 and D1, and is typically gradationally overlain by facies B5; it sometimes overlies the Nicosia Formation directly.

Interpretation: the clean, well sorted nature of the facies, occasional presence of shell debris, occasional wave-rippling and association with other fossiliferous facies, suggest a shoreline setting; the lower part of the facies, which is dominated by landward-dipping, trough cross-laminated sediments, represents a typical shoreface zone, which is covered by lunate megaripples, migrating shorewards, under the influence of shoaling waves (Clifton et al., 1971; Elliot, 1986a); the upper part of the facies is pebbly,
and contains mud layers (facies D1), which would not be expected to survive in a subtidal, shoreface setting; a foreshore (or even backshore) environment is implied, although shallow, seaward-dipping lamination, which is typical of the foreshore (Elliot, op. cit.), is not abundant; instead, interdigitation of various sedimentary structures suggests the presence of a variety of bedforms, which mainly migrated landwards, such as is found in relatively low energy foreshore settings, which develop ridge and runnel-type topography (cf. van den Berg, 1977); landward-dipping cross-lamination is partly the result of migration of ridges, which may be covered with superimposed, smaller bedforms; occasionally, offshore-directed lamination, generated by storm or shoaling waves, is preserved (van den Berg, 1977); a low energy environment is also supported by the presence of mud (facies D1), which can accumulate in the lee of ridges and small spits (Wünderlich, 1972; van den Berg, op. cit.); ridge and runnel topography is normally associated with mesotidal settings (Elliot, 1986a), but is also known from microtidal beaches (e.g. beaches in the Netherlands with tidal range of 1.5m, van den Berg, op. cit.).

**Facies B2 - Thin conglomerate layers and lenses**

**Description:** a variety of sedimentary structures and bed geometries are present in this facies (refer to Fig. 4.12): a) thin stringers, one to a few clasts thick, of sorted, pebble conglomerate, supported by well sorted, facies B1 sands (Plate 4.1a); flat sides of clasts are often aligned parallel with bedding; occasionally, these sand-supported conglomerates are up to 1m thick, and flat clasts are sometimes imbricated down-dip; b) 5-15cm thick, pebbly sands, with pebbles dispersed fairly evenly through the sandy matrix; beds have sharp, rather wavy, contacts; c) thick (up to 1m), beds of unstratified, unimbricated, poorly or moderately sorted, sandy, pebble to occasional cobble conglomerate, probably clast-supported, with rather irregular, sharp bases, and gradational tops, and rare, intercalated, sand lenses (Plate 4.1b); d) channelised, sandy, moderately sorted, pebble conglomerates, up to 70cm thick, which are unstratified and ungraded, and matrix- to clast-supported; sometimes, thin, laterally extensive, better sorted conglomerate layers
Fig. 4.12 - Field sketches, subfacies of facies B2 (see text for descriptions and Encl. B for key to symbols).

Fig. 4.13 - Field sketches, subfacies of facies C4 (see text for descriptions and Encl. B for key to symbols).
are developed at the top of channel units; e) thin, poorly sorted, sandy pebble conglomerate lenses, clast-supported, with sharp, concave-up bases and convex-up tops (Plate 4.1b); f) 5-30cm thick, tabular sandy conglomerates, inversely graded and sand-supported except at the top; beds have flat, sharp tops and bases (Plate 4.1c); and g) 10-20cm thick beds of tabular, cross-bedded, pebbly sands, with graded foresets and sharp, flat tops and bases (Plate 4.1c); cross beds dip offshore; none of the above is fossiliferous.

Occurrence, lateral and vertical facies relations: these subfacies are always found interbedded with the upper part of facies B1; type a is the most common, and occurs at several localities throughout the study area; types b - g are less common, and are found in the east of the study area, (e.g. log 4, Fig. 4.5).

Interpretation: the close association of these subfacies with the upper part of facies B1 implies a similar depositional environment, i.e. a foreshore setting; wave- or storm-reworked, shoreline conglomerates are typically well sorted, clast-supported, stratified and often imbricated (Bourgeois and Leithold, 1984; Nemec and Steel, 1984); the poorly sorted, variable nature of this facies, however, suggests fluvial flooding of the shoreline, with little or no wave reworking in the relatively low energy setting (see interpretation of facies B1); several subfacies may be the product of mass flow; type b may be the result of deposition from rather sandy debris flows (cf. channel mouth bar conglomerate type E of Kleinspehn et al., 1984); types c and d represent minor sheet flows, type d flows being perhaps more fluidal and erosive; the thin, extensive, better sorted tops of type d may reflect minor wave reworking; type e sediments are rather similar to type d, except that wave reworking has been able to partially remould gravelly lens tops into megaripples, perhaps during storms (cf. Bourgeois and Leithold, 1984); type f sediments are similar to channel mouth bar conglomerate type F of Kleinspehn et al., (1984), which are interpreted as mass flows in which dispersive pressure played an important role in supporting clasts; type g sediments indicate traction current processes, and migration of pebbly megaripples across the foreshore; type a conglomerates are the best sorted of all the subfacies, and show seaward-dipping imbrication, suggestive of wave reworking; beach
conglomerates are often clast-supported, but if sand is plentiful, wave energy can incorporate both sand and gravel to give bimodal textures (Kleinspehn et al., 1984); in summary, these subfacies are the products of fluvial processes, which sometimes dominated the shoreline, with only minimal wave reworking.

**Facies B3 - Trough cross-laminated sands**

**Description:** this facies comprises two groups of trough cross-laminated sands: a) fine-grained, sorted sands in broad shallow channels, 2-4m wide, up to 50cm thick, cut into finer, silty massive sands; trough cross-lamination is very low-angle, largely parallel to trough margins, though occasionally discordant; one to several sets fill each channel (Plate 4.2a); intraformational mudclasts are sometimes common in the bases of cross-bed sets, and lie parallel to lamination; convolute lamination is recorded at the base of one channel; bioturbation of both the underlying sediment and the channel fill occurs in the form of vertical to subvertical, non-branching, cylindrical burrows (?partly Skolithos); plant debris is sometimes concentrated along laminae; palaeocurrent directions are NE (i.e. probably offshore); b) coarser (medium-grained), slightly pebbly sands, rich in bioclastic debris, in a 3m-thick unit, erosively overlying fine sands; trough cross-lamination occurs in sets up to 20cm thick, 60cm wide; intraformational mudclasts of facies D1 mud and rare convolute lamination are present; pebbles are aligned down foresets; palaeocurrents are directed to the NW (i.e. probably offshore).

**Occurrence, lateral and vertical facies relations:** this is an uncommon facies; type a occurs at only two localities in the west of the study area (Fig. 4.2), where it overlies ?facies B1, and occurs close to exposures of facies B5; type b occurs at one locality at the top of the Kakkaristra Formation, in the centre of the study area, where it overlies facies B1, and is intercalated with facies B2.

**Interpretation:** the association with facies B1, B2 and B5 implies a shoreline setting; channel-shaped geometry of type a, offshore-directed dips, presence of mudclasts, and interdigitation of type b with facies B2 may again suggest occasional fluvial reworking of the shoreline, as is documented by facies B2; in this case, fluvial
currents at least in part cut channels across the backshore - foreshore, and low to moderate relief lunate bedforms migrated along them; type b sediments may occur in channel-shaped bodies, but exposure is insufficient to confirm this; fluvial currents were able to rework beach sand, which was sometimes very shelly, and mud; shelly sands are very uncommon in the Kakkaristra Formation, but do occur close to the exposure of type b (facies C4, Fig. 4.2); deposition was sometimes rapid, giving rise to convolute lamination, but between flows, some bioturbation took place; *Skolithos* ichnofauna are consistent with a shoreline setting (Frey and Pemberton, 1984).

**Facies B4 - High-angle, cross-bedded conglomerates**

*Description:* the facies comprises single units, 70cm - 4m thick, of moderately to sometimes well rounded, granule to small cobble conglomerate; conglomerates form wedge- or bar-shaped units, up to 100m in length and of unknown width (Plate 4.2b); cross beds are 5-30cm thick, and range from poorly to moderately sorted and stratified, with mostly clast-supported conglomerate (most common type; Plates 4.2c and 4.3a), to thinner, finer, better sorted and stratified conglomerates, which are clast-supported and sometimes matrix-free (Plates 4.4a and b); up-dip a-axis imbrication is developed, especially in the better sorted conglomerates (Plates 4.3b and 4.4b); finer conglomerates may have isolated larger clasts, lying parallel to bedding; cross beds are rarely normally graded, planar to concave-up in shape, and have sharp to gradational tops and bases, which may be erosive; matrix-free conglomerates occur in lenses, which are laterally less persistent than other cross beds; cross beds dip up to 30°, but noticeably shallow down-dip (Plate 4.2b); lower contacts are sharp (for high-angle cross beds) to tangential (at bar toes); the conglomerates are rarely fossiliferous, sometimes containing large amounts of oyster debris, and occasional clasts with oyster spats attached; rare cross-laminated sand interbeds occur, which may contain plant remains; occasional exposures transverse to dip direction demonstrate concavity of cross beds in this direction too, and frequent erosive, cross-cutting relationships; tops of conglomerate units may be abrupt, or may grade up quickly into a rather poorly sorted, graded, unstratified, conglomeratic to sandy
layer (Plates 4.3c and 4.4a), which may be rich in oyster debris (including small oyster clumps) and, at one locality (WD 358822), contains small heads of the coral Cladocora caespitosa (Plate 4.3c); bases of conglomerate units are very sharp, and occasionally loaded and deformed; cross strata above these deformed bases are typically also deformed, showing wavy to rotated bedding and vertical clast orientation; palaeocurrent data for this facies are very variable (Fig. 4.3).

Occurrence, lateral and vertical facies relations: these conglomerates are found predominantly in the central and eastern part of the study area (Fig. 4.2), in lower, middle and upper parts of the Kakkaristra Formation; they are typically under- and overlain by facies B1 and B2 (e.g. log 5, Fig. 4.4 and log 4, Fig. 4.5); they also occur in association with other facies (see Fig. 4.16); conglomerates with deformed bases are found only at the base of the formation, cutting erosively into the underlying Nicosia Formation (e.g. log 2, Fig. 4.7).

Interpretation: the presence of oyster debris, close association with coral, and intercalation with facies B1 and B2 imply a shoreline to subtidal setting; much thicker, cross-bedded conglomerates, similar to this facies, have been described from coastal environments as Gilbert-type deltas (e.g. Postma and Roep, 1985; Colella et al., 1987; Ori and Roveri, 1987), and also as huge, migrating bedforms within channels (this interpretation is a result of recognition of channel-shaped geometries; Ori and Ricci-Lucchi, 1981; Ori and Roveri, 1987); sediments of this facies are not restricted to channels, but form small, isolated, bar-shaped units; they are interpreted as a form of small-scale, Gilbert-type deltaic lobe, deposited at the mouths of rivers entering the Mesaoria seaway; in this setting, as fluvial, heavily-laden currents reached the basin, they rapidly mixed, expanded, decelerated and deposited their loads; rapid mixing is favoured if the density of the waters of the receiving basin are similar to that of the fluvial currents (Wright, 1977; homopycnal flow of Bates, 1953); if they are more dense, suspended sediment may be carried away as a buoyant plume above basin waters; if they are less dense, fluvial outflow will tend to pass beneath basin waters, transporting sediment beyond the delta front (Elliot, 1986b); the
favoured conditions are most readily achieved in lacustrine settings (many examples of Gilbert-type deltas come from lacustrine environments), but sufficiently high fluvial output in marine settings can cause dilution of the waters of the receiving basin (Elliot, op. cit.; Colella et al., 1987); once mouth bar sedimentation is initiated, subsequent flows cascade down the face of the incipient mouth bar by avalanching or gravity sliding (Postma and Roep, 1985; Colella et al., op. cit.); the poorly sorted, disorganised nature of much of this facies, lack of grading or imbrication, absence of fining of foresets down-dip and tangential bases, suggest rapid freezing of subaqueous mass flows (Nemec and Steel, 1984; Postma and Roep, op. cit.); well sorted, matrix-free, lens shaped, cross strata may indicate sediment remobilisation on steep, rather unstable, bar fronts, removal of fine sediment in suspension, and resedimentation of coarse material further down the bar front, in a type of grain flow process (cf. flow slide deposits of Colella et al., op. cit.); other evidence of slope instability is uncommon, except for deformed conglomerates at the base of the Kakkaristra Formation; these may have been influenced by particularly high rates of sedimentation, following uplift at the beginning of Kakkaristra Formation times (see section 4.2.2), and/or increased slopes as a result of uplift; once mouth bars were abandoned, tops were variably reworked by wave action; they also formed a firm substrate, which was colonised by oysters, or rarely, coral; palaeocurrent data indicate that mouth bar progradation was not simply perpendicular to the inferred WNW-ESE basin margin, but was in a variety of directions, depending on the direction of entry into the basin of individual river channels; rounding of clasts is sometimes surprisingly good in this facies, and may reflect reworking of partially rounded, older conglomerates of the Nicosia Formation (see further, section 8.2, conglomerates subsection).

**Facies B5 - Low-angle, cross-bedded conglomerates**

*Description:* the facies comprises cross-bedded, sorted, granule to pebble conglomerates and pebbly coarse sands, occurring in wedge-shaped units, up to 5m thick, which thin down-dip; cross beds dip at moderate to low angles (less than 15°), are planar, 5-15cm thick, ungraded, clast-supported and sometimes matrix-free.
(Plate 4.5a); they dip offshore, and predominantly to the NW (Fig. 4.3); matrix is of sorted, fine to coarse sand; intervals of sand-supported pebble conglomerates occur, as do pebbly sands, which often have clasts aligned parallel to bedding; conglomerate clasts are typically well rounded, often flattened, and seaward-dipping imbrication is often well developed (Plate 4.5b); units of this facies comprise one to several cross bed sets, which are 50-150cm in thickness; cross bed sets may cut down at shallow angle into each other (Plate 4.5a); rare lenses of planar cross-laminated medium or coarse sand are sometimes intercalated, and may show shoreward-dipping lamination; cross bed units have gradational bases; tops are typically gradationally overlain by 50-200cm of well sorted, fine, medium or coarse sand, which is massive apart from horizontal layering defined by scattered isolated pebbles (Plate 4.5a); pebbles are often aligned parallel to layering; the facies is unfossiliferous.

**Occurrence, lateral and vertical facies relations:** this facies occurs throughout the study area (Fig. 4.2), and always at the top of the formation (e.g. logs 1 and 2, Fig. 4.6), where it is overlain by the Apalos Formation; the upper part of the facies, close to the contact with the Apalos Formation, quite commonly contains white, powdery layers of facies D3; the base of the facies typically grades down into facies B1 or B4; the lateral extent of individual units of the facies is unknown.

**Interpretation:** the good size and shape sorting and down-dip imbrication of this facies are typical of foreshore conglomerates (Nemec and Steel, 1984); a shoreline setting is supported by association with facies B1 and B4; gradational occurrence above facies B4 may indicate more or less *in situ* reworking of mouth bar conglomerate; offshore palaeocurrent directions are consistent with wave reworking; the typical occurrence of this facies at the top of the Kakkaristra Formation may indicate a broad decline in fluvial influence on the fan-delta at the end of Kakkaristra Formation times, and a corresponding increase in the importance of wave energy; although prevailing wave energy in the Mesaoria basin is inferred to have been relatively low (see facies B2), in the absence of significant fluvial input, waves were able to rework, to some extent,
the fan-delta shoreline; beach conglomerate was apparently deposited first, then as coarse-grained material was used up, the sandy upper part of the facies was deposited; occasional, relatively prolonged, emergence of beach facies is indicated by the development of facies D3 caliche.

**Facies B6 - Sorted, poorly stratified conglomerates**

*Description:* this facies contains moderately sorted, sand- to clast-supported, pebble conglomerates, in ungraded, massive to poorly horizontally stratified units, generally 30-80cm thick, though rarely up to 4m thick; clasts are mainly subrounded, occasionally horizontally aligned or imbricated down-dip, and set in a sorted, fine sandy matrix; conglomerates are often shelly, containing identifiable *Ostrea, Pecten* and other bivalve fragments; bed bases and tops are abrupt and flat; conglomerates form wedge-shaped units, which thin seawards.

*Occurrence, lateral and vertical facies relations:* the facies occurs across the study area, typically at the base of the formation (e.g. log 1, Fig. 4.4, logs 2, 3 and 4, Fig. 4.9); most often, it is overlain by facies C1 or C2; the facies usually forms a single wedge-shaped unit, but where it becomes thick, several beds are present, and other facies (B1 and D2) may be interbedded.

*Interpretation:* the presence of bivalves and association with facies B1, C2 and D2 suggest a shoreline setting; sorting and occasional seaward imbrication suggest the sediments may be beach-type conglomerates (they resemble the wave-reworked conglomerates of Kleinspehn *et al.* (1984), except for poorer stratification in this case), but stratification and imbrication are not nearly as well developed as in facies B5; this facies may be less well washed than facies B5, as a result of lower wave energy or more rapid burial; both these factors can be related to the position of the facies typically at the base of the formation, a time when sedimentation rates were probably high, and fluvial output very significant (hence decreasing the influence of waves), as the Kakkaristra fan-delta first prograded into the basin following uplift in the hinterland.
**Facies Cl - Massive, muddy sands and sandy silts**

**Description:** this facies comprises yellow-brown to greenish-grey, muddy sands, or occasionally, brown sandy silts, all of which are poorly consolidated, massive and bioturbated, and occur in units 4-10m thick (Plate 4.5c); sandier facies sometimes contain thin (1-2cm), horizontal, cemented bands or less common calcite concretions; no structures are visible; plant debris and wood fragments may be common, and include a complete pine cone (Plate 4.6b, log 4, Fig. 4.9; see Appendix 1); the facies occurs in three distinct areas, in the east, central and western parts of the study area (Fig. 4.2), each of which has slightly different palaeontological aspects: in the west, scattered, broken and occasionally whole bivalves, and rare echinoderms, are found (logs 2 and 4, Fig. 4.9); in the centre, shell coquinas, comprising one bivalve species (*Venus* sp.), and with shells often horizontally aligned, form lenses and impersistent layers (Plate 4.6a; log 4, Fig. 4.7); and in the east, moulds of whole bivalves, occasional scattered bivalve debris, and an oyster bed containing intact, articulated oysters and oyster clumps, are found; a thin horizon of tiny, wispy rootlets occurs in the east (log 2, Fig. 4.4); occasional, thin (5-15cm), interbedded, tabular cross-laminated, fine to medium sands, with southerly directed palaeocurrents, are also found in the east, along with a thin unit, at the base of the facies, of partly wave-rippled and partly parallel-laminated, very fine sand and silt (log 2, Fig. 4.5); at one locality (log 1, Fig. 4.4), the sediments of this facies become enriched in the tests of reworked pelagic foraminifera (see further, facies C4, and also section 8.2, subsection on source areas).

**Lateral and vertical facies relations:** these sediments are associated with facies C2, C3 and B6, which under- or overlie them, or less commonly are interbedded with them (e.g. Figs. 4.4, 4.5 and 4.9).

**Interpretation:** the fine grain size, intense bioturbation, presence of marine faunas, and preservation of a pine cone suggest a relatively low energy, subtidal setting, along the Kakkaristra fan-delta front; reconstruction of the fan-delta shoreline (see section 4.4.1) indicates that this facies occurs in three large embayments or bays, situated in the east, central, and western parts of the study area.
area (Fig. 4.2); the most enclosed of these (in the east) contains the finest sediment (silts), and contains a rootlet horizon which may represent the root system of a type of sea grass e.g. *Posidonia*, which grows in some protected, shallow subtidal areas of the present Mediterranean (Pérès, 1967); oyster colonies are also consistent with a shallow subtidal setting (although oysters can live in a wide range of environments; Stenzel, 1971); thin, cross-laminated sands and rippled silts in the eastern bay area, indicate some current activity in the bay; sands may have been introduced fluvially, or by marine reworking; the coquinas of the central bay area indicate periodic, higher energy conditions, and may relate to enhancement of currents as they flowed between the delta front and a detached shelly sand bar (facies C4; see Fig. 4.2); the western bay area is the most open (Fig. 4.2), and contains the most diverse fauna, including echinoderms, which do not readily inhabit partially enclosed areas which may have abnormal salinities.

**Facies C2 - Rootlet-bearing sands**

*Description and facies relations:* this facies contains grey-brown, sorted, clean to slightly muddy sands, which are thin- to medium-bedded (5-20cm); beds are planar, apparently with no internal structure (though they are often badly weathered); scattered pebbles and shell fragments occur infrequently; long, thin, subvertical rootlets are typically present (Plate 4.6c); the facies forms intervals 2-3m thick, which typically occur at the base of the Kakkaristra Formation, overlying facies B6 (e.g. logs 2 and 4, Fig. 4.9); the facies is overlain by other delta front facies (C1, C4 or C5).

*Interpretation:* lack of structure makes interpretation tentative; rootlets and occasional shell fragments suggest a marginal marine setting, e.g. marshy fringe of an interdistributary bay; location of the facies, with respect to the shoreline reconstruction of the fan-delta (Fig. 4.2; see section 4.4.1), is consistent with this interpretation; slightly muddy sand is envisaged to have been intermittently deposited during flooding, while plant colonisation took place between floods; a modern analogue might be the interdistributary bay areas of modern, fluvially-dominated deltas, e.g. the Mississippi (Gould, 1970), though much finer-grained
sediment accumulates in that deltaic system.

Facies C3 - Poorly bedded sands

Description: this facies comprises rather poorly preserved sediments, with indistinct structures; two groups are recognised: a) greenish, fine-grained, sorted, thin to medium bedded (5-15cm) sands; bedding is laterally impersistent and lenticular, with broad pinching and swelling, over a wavelength of 1 or 2m; bed bases are wavy and possibly partly erosive; at one locality only, the facies is well preserved, and reveals low-angle, trough cross-lamination; internal lamination is concordant with bed bases, and also shows pinching and swelling; diffuse, rippled lenses interfinger; elsewhere, shell debris (especially oyster) occurs as scattered material, or is concentrated along bed bases; moderate burrowing is found; b) slightly muddy, fine, sorted, brown sands, which are poorly bedded; indistinct parallel and cross-lamination (probably trough type) occurs (no discernible palaeocurrent trends are evident from sparse data); scattered pebbles and shell debris may be present; thick oyster beds, with broken and whole, sometimes articulated, valves, are also occasionally present (Plate 4.7a); at one locality (WD 367827), a stratified, 2m-thick, oyster bed, containing oyster clumps with attached barnacles, occurs (log 7, Fig. 4.4); stratification of this oyster bank is planar, and dips eastwards.

Occurrence, lateral and vertical facies relations: the facies forms units 1 to a few metres thick, in the eastern and central parts of the study area (e.g. logs 6 & 7, Fig. 4.4; Fig. 4.2); type a is restricted to the east, and occurs within the zone of interdigitation with the Athalassa Formation (Fig. 4.2); it is associated with facies C4 and C5; type b is associated with facies C2, C4, C5 and B6.

Interpretation: in type b, the location of the subfacies with respect to the fan-delta shoreline reconstruction (see section 4.4.1 and Fig. 4.2), association with other facies of group C, the presence of oyster beds, which have undergone relatively little transportation, and general lack of structure suggest a low energy, probably subtidal, bay fill or delta front environment; the presence of minor current activity is indicated by poorly developed parallel and cross-lamination; subfacies C3a is enigmatic; interdigitation with
facies of the Athalassa Formation indicates a shallow, subtidal setting; lenticular bedding may be a mild form of hummocky cross-stratification, or a structure similar to the HCS-related, low-angle trough cross-stratification of Nøttvedt and Kreisa (1987); the restriction of this facies to the east may be significant, because in this area, unidirectional currents are believed to have been flowing through the narrow zone separating the Kakkaristra fan-delta and the Athalassa Formation sand body field (Fig. 4.2; see section 4.4.2, subaqueous subsection); the combination of unidirectional currents with oscillatory flow is believed to generate HCS (Elliot, 1986a; Nøttvedt and Kreisa, 1987); concentration of coarse debris along bed bases and bioturbation are recorded in other HCS facies e.g. DeCelles, 1987; the interpretation is very tentative, however.

**Facies C4 - Cross-bedded, bar sands**

**Description:** this facies contains six sand intervals, all of which are cross-bedded or sometimes cross-laminated, have flat, non-erosive, sometimes gradational bases, and broadly tabular geometries; slightly different features characterise each unit (see Fig. 4.13; refer to Fig. 4.2 for locations and Fig. 4.3 for palaeocurrent data): a) fine-grained, well sorted, sand, largely made up of reworked pelagic foraminifera, in 5-20cm thick, horizontal beds, which are internally structureless or planar cross-laminated (Plate 4.7b), or, in the lower part of the facies, ripple cross-laminated; burrows, including *Skolithos*, are moderately common; cross strata dip southwest, as does asymmetric wave ripple lamination; b) similar sediment type to a), but in a 3m thick, cross-bedded unit; shallower, laterally more extensive, lower cross beds dip west, while steeper, upper cross beds dip east, and partly truncate those below; occasional burrows and scattered pebbles are present; c) coarsening-up, fine- to medium-grained bioclastic sand, in a 2m thick, cross-bedded unit (Plate 4.5c), exposed laterally over ca. 200m; at its north end, cross beds dip at moderate angles northwards; towards the south, dips shallow, the sediment becomes more laminated, and lenses of cross-laminated sand dip mainly north, but occasionally south; ripple cross-lamination and burrows are
sometimes present at the base of the unit; d) greenish, fine, sorted, sand in a 90cm thick unit, with a thick shell coquina at the base; shells are nearly all horizontally aligned bivalves of the genus Glycymeris; the upper part of the unit comprises very shallow, southeastward-dipping, thin, cross beds, with shell debris (principally oyster) aligned parallel to bedding; e) coarsening-up, fine- to medium-grained, slightly muddy, partly bioclastic sand, in a 1.5m-thick unit (log 6, Fig. 4.4), with a thin horizontally laminated lower part, and a trough cross-laminated upper part; cross-lamination dips south; the upper part also contains an un laminated, highly fossiliferous horizon, containing molluscs, serpulids, occasional coral and bryozoan fragments, and algal balls; burrows, including Thalassinioides, are moderately common; f) fine, muddy, bioclastic sand, in a laterally extensive, but badly weathered, unit, 1m thick; this is vaguely cross-laminated, but structure is largely destroyed by caliche formation (caliche fabric is visible in thin section); rootlets are present at the top.

Occurrence, lateral and vertical facies relations: type c occurs in the central bay area of the fan-delta; the others all occur in the east; all are associated with facies C1 or C2, except for type b, whose facies relations are uncertain; type f overlies both facies C2 and B1, (logs 5 and 6, Fig. 4.5).

Interpretation: the association with facies C1 and C2, and location of subfacies with respect to the Kakkaristra fan-delta shoreline reconstruction (Fig. 4.2; see section 4.4.1), imply a delta front/bay environment; the general cross-bedded nature, occasional coarsening-up, unchannelised bases and probable tabular nature suggest a migrating sand bar type of setting; type a represents a sandy swell, in a very fine-grained, protected area (eastern bay area), which was periodically subject to moderate current activity (see facies C1); bar type b is located at the mouth of the eastern bay area; its palaeocurrent directions document fluctuating currents, flowing both into and out of the bay (Fig. 4.3); the bioclastic nature of the sediments of bar type c is anomolous, since most nearby facies are clastic-rich; there is some indication that shell accumulation was taking place in this part of the fan-delta, from the presence of shell layers in the underlying facies C1, and the nearby
presence of shelly facies B3; northward palaeocurrent directions are also odd, in that they reflect current flow not directly through the mouth of the central bay (Fig. 4.3); this may reflect a complex pattern of currents around the shoaling bar and through the mouth of the bay; bar types d and e occur in the area between the Kakkaristra fan-delta and sediments of the Athalassa Formation; they are the most macrofossiliferous bar types, perhaps due to washing in of shell debris from the edge of the Athalassa sand body field; palaeocurrent directions indicate that flow through this area was predominantly from NW to SE (Fig. 4.3); interpretation of facies type f is hampered by poor preservation; caliche fabrics and rootlets indicate emergence; the underlying relationship with both facies B1 (shoreline) and C2 (delta front) may indicate rapid progradation of a shoreline-attached ?spit; lagoonal facies D3 occur above the subfacies; spits are normally associated with wave-dominated settings; a spit is recorded, however, at one location away from channel mouths along the Mississippi delta (Gould, 1970, his Fig. 17), although no description of it is given; the sediment composition of bar types a and b is attributed to derivation largely from the chalky, marly, pre-Pliocene sedimentary cover of the Troodos Massif in the east (see further, section 8.2, subsection on source areas).

**Facies C5 - Conglomeratic, shelly sands**

**Description:** the facies comprises sandy, small pebble conglomerate and interbedded pebbly sands, in units 70-150cm thick; these are moderately to well stratified, and composed of 10-40cm thick, massive, or parallel or cross-laminated beds or lenses; lenses grade into each other, and cross-lamination dips seawards; bases of units are planar or erosive and channelised; channelised units have poorly stratified, conglomeratic bases, which may contain much shell debris, especially oyster fragments; massive conglomerates are sand-supported, ungraded, and sometimes show minor up-dip pebble imbrication; the facies is not well exposed, so lateral dimensions are unknown.

**Occurrence, lateral and vertical facies relations:** this uncommon facies is recorded at three localities across the study area; it is typically interbedded with facies C2 and C3a (e.g. log 6, Fig. 4.5).
Interpretation: a shallow, subtidal setting is indicated by association with facies C2, and occurrence of the facies north of the inferred shoreline of the Kakkaristra fan-delta (Fig. 4.2; see section 4.4.1); palaeocurrent directions and pebble imbrication indicate offshore-directed processes, which could represent deposition from a) distal, fluvially-influenced currents, or b) storm-influenced, marine currents; the general lack of conglomerate-grade material in shoreface sands (lower part of facies B1) and other fan-delta front facies (facies C1 and C3) suggests that marine processes were not readily capable of transporting gravel offshore; storm-related deposits might also have been expected to be more sheet-like (cf. shallow marine conglomerates of Wright and Walker, 1981, and Leckie and Walker, 1982); the facies is thus tentatively interpreted as a fluvially-influenced, distal mouth bar facies.

Facies D1 - Thin, white muds

Description: these sediments comprise pure to slightly silty, white claystone, in thin beds, 0.5-8cm thick (Plate 4.8a); bases and tops are usually flat or wavy, but bases are occasionally beautifully wave-rippled, and tops are occasionally strikingly flamed, with flame structures as thick as beds; internally, beds are massive, or parallel-laminated; rippled sand and silt lenses to wavy beds are intercalated; tiny, vertical cracks are sometimes present; beds are laterally persistent over a few tens of metres at most; occasionally, beds are abruptly terminated, broken and reworked as mudclasts in the surrounding sands; the muds are unfossiliferous, and contain scattered plant debris; burrowed tops are common; SEM studies suggest that most of the sediment comprises detrital clay and carbonate.

Occurrence, lateral and vertical facies relations: this uncommon facies is recorded at two localities only (Fig. 4.2; logs 1 and 2, Fig. 4.6), though similar white mudstones also occur in the Athalassa Formation; the facies is typically interbedded over 1 or 2m with the upper part of facies B1; white mudstone clasts occur in facies B3.

Interpretation: structures in the associated facies B1 together with preservation of this facies imply a relatively low energy, foreshore setting (see also facies B1); mud is inferred to have
accumulated in the lee of small ridges and spits present along the foreshore; these lee areas sometimes had wave-rippled surfaces; mud and thin sands were probably introduced by aeolian processes and large waves; deposition was largely from suspension, with minor ripple formation; strong, storm-generated winds occasionally caused flaming of mud layer tops, and breaching of beach ridges caused some layers to be partially broken and reworked; tiny, vertical cracks are probably evidence of desiccation; burrowing was fairly common; the striking white colour of this facies is due to its highly calcareous nature, and suggests derivation largely from the chalky and marly pre-Pliocene sedimentary cover of the Troodos Massif (see further, section 8.2, clays subsection); muds are not common in modern beach settings, but are occasionally documented (Wünderlich, 1972; van den Berg, 1977).

Facies D2 - Green and brown, ostracodal muds

Description: this facies contains green or brown, pure to sometimes sandy or granuly, claystone, and minor pale, muddy fine sands, in units 30-150cm thick; internally, these units appear massive or show wispy lamination, and sometimes contain very thin, white, horizontal, calcareous partings (Plate 4.8b); sandy and granuly mud forms layers, less than 20cm thick, which have sharp or diffuse contacts; the facies is often fossiliferous, containing bivalves and rarely gastropods (usually as complete shells), sometimes concentrated in layers; oyster clumps were found at one locality; in purer clays, numerous ostracods may be present (see Appendix 1), and occasionally benthic foraminifera; bases of mud units are usually planar and non-erosive, although at one locality (WD 237841), the facies has a concave-up base, lined with pale conglomeratic mud; an especially thick development of this facies (8m; log 4, Fig. 4.10) occurs in the far west of the study area (Fig. 4.2); at this location, green claystones are interbedded with thin, horizontal brown, mudstones, which are sometimes broken (?mud-cracked); SEM studies show the clays to comprise predominantly detrital clay flakes, with rare, unidentified, authigenic material (Plate 4.8c).

Occurrence, lateral and vertical facies relations: this facies is found throughout the study area, typically at the top of the
Kakkaristra Formation, overlying facies B1 or B5 (e.g. log 5, Fig. 4.7 and log 5, Fig. 4.8); less commonly, the facies occurs in lower parts of the formation, interbedded with facies B1, B4 and B6 (e.g. log 1, Fig. 4.5).

**Interpretation:** like facies D1, these sediments were deposited in quiet, protected areas; the thicker development, lack of sandy interbeds, lack of evidence of disruption, and fossiliferous nature imply a more protected, temporally more persistent environment than facies D1, free from wave influence, but still close to the shoreline e.g. backshore; ostracod assemblages suggest brackish to littoral conditions (A. Lord, pers. comm.); the facies probably accumulated largely by suspension sedimentation in backshore ponds, which were colonised by molluscs and ostracods; desiccation was uncommon; very minor influxes of coarser sediment occurred; the common occurrence of the facies at the top of the Kakkaristra Formation is related to the final stages of progradation of the Kakkaristra fan-delta, withdrawal of marine waters from the basin, and subsequent emergence.

**Facies D3 - Muddy, calichied sediments**

**Description:** three types are recognised: a) loose, white, powdery, very fine-grained, chalky material, in one or two 3-10cm thick, horizontal layers, intercalated at the top of facies B5 at the top of the Kakkaristra Formation (e.g. log 1, Fig. 4.6, caliche symbol); white powdery material encases grains or small lenses of facies B5 sand; b) thin, planar-bedded, brown, muddy, fine sands and silt, with occasional mud laminae, and moulds of whole bivalves, interbedded with pale grey to white, hard, cemented beds, which are grainy and massive; beds are 5-12cm thick, laterally continuous over only a metre or two, and have sharp, flat to wavy contacts; c) 1m-thick, hard, massive, pale grey, very fine-grained calcareous sediment, which is mottled and vuggy; the bed is laterally extensive over a few hundred metres; in thin section, types b and c have a pelleted micritic matrix, calcite-filled veins and patches, floating silt grains and dense nodules, and occasional tubules filled with coarse spar; these are typical caliche fabrics (e.g. Wright and Wilson, 1987).

**Occurrence, lateral and vertical facies relations:** the three
subtypes are found at or near the top of the Kakkaristra Formation; type a is found at several localities, type b is found at only one locality in the east, overlying facies C4 (log 5, Fig. 4.5), and type c at only one locality in the centre of the study area, overlying facies C1 (Fig. 4.2); types a and b are overlain by sediments of the Apalos Formation; overlying facies have been eroded from above type c.

**Interpretation:** the occurrence of the facies at the top of the formation, and the caliche fabrics in all three subtypes, indicates emergence, related to the final stages of deposition of the Kakkaristra Formation, immediately prior to the transition into the entirely subaerial Apalos Formation; type a falls into the chalky caliche category of Esteban and Klappa (1983), which is a fairly immature type, and suggests relatively short periods of exposure of the foreshore facies, with which the subfacies is intercalated; types b and c are much better consolidated, and may represent more prolonged exposure of sediments that were already fine-grained, in a backshore or lagoonal environment; type b may have developed in a lagoon behind the bar facies of facies C4, which it overlies; type c represents exposure at the end of Kakkaristra Formation times of part of the fine-grained, central bay area (Fig. 4.2); caliche is more typically described from continental settings, e.g. the Apalos Formation (see facies D2, section 6.2), than from those of deltaic origin.

4.4 Sedimentological Model

4.4.1 Fan-delta interpretation

Several factors lead to an overall deltaic interpretation for the facies of the Kakkaristra Formation: close association of wave- and fluvially-influenced facies, interbedding of fossiliferous and unfossiliferous sediments, rapid lateral and vertical transitions between the large number of facies recognised, and the general stratigraphic position of the Kakkaristra Formation, above the shallow marine, upper Nicosia Formation and below the subaerial Apalos Formation.

The palaeoshoreline of the delta has been reconstructed using the mapped lateral distribution of facies (Fig. 4.2). Known exposures
of shoreline facies (facies B1, B2, B4, B5 and B6) were joined together, and the resulting reconstruction (Figs. 4.2 and 4.3) shows a highly indented delta margin, with fluvially-built lobes, separated by interdistributary bays. Inaccuracies will be present in this reconstruction, because facies transitions occur vertically as well as laterally, and a reconstruction for one particular time interval should ideally be made e.g. lower, middle or upper Kakkaristra Formation. This is not possible, however, because sometimes only part of the formation is exposed at one locality, while a different part is exposed at another locality. In the period of time represented by only 10-15m of sediment (the thickness of the formation), however, it is considered unlikely that the shoreline fluctuated greatly, and the outline shown in Fig. 4.2, which in general represents lower Kakkaristra Formation times, will largely be correct.

The lobate character of the delta is typical of a delta that is fluvially-influenced, similar to, for example, the modern Mississippi delta. The Kakkaristra system was also wave-influenced, however, as recorded by facies B1 and B5. The Kakkaristra Formation delta can further be classified on account of its coarse-grained nature. Coarse-grained deltas have been referred to as fan-deltas in the past (Holmes, 1965; references in McPherson et al., 1987). Recent attempts at classifying coarse-grained deltas have led to the recognition of a second major, coarse-grained delta type, the braid delta (Nemec and Steel, 1987; McPherson et al., op. cit.). The distinction between the two delta types is clear according to McPherson et al.: a fan-delta comprises an alluvial fan that progrades from an adjacent highland directly into a standing body of water, is generally fan-shaped, and mass flow-dominated, while a braid delta is a braided fluvial system that progrades into standing water, is generally elongate and sheet-like, and stream flow-dominated. Nemec and Steel (1987), however, suggest that the term fan-delta should be restricted to the coastal prism of sediment deposited as an alluvial fan progrades into standing water, regardless of whether an intermediate braided system is present. They distinguish between mass flow-dominated and stream flow-dominated fan-deltas. Fan-deltas have also been classified according to the morphology of the receiving basin (shelf and slope
fan-deltas of Ethridge and Wescott, 1984), and it has been suggested that the dominant processes operating in the receiving basin be used for classification (wave-, tidally-, or fluvially-influenced fan-deltas; Kleinspehn et al., 1984).

In the case of the Kakkaristra Formation, classification according to the subaerial part of the system is difficult, because of the poor preservation of fluvial facies. The limited data available suggest an alluvial fan setting (facies A1 and A2), but with a distal braided fringe (facies A3; see also discussion in the following subsection). The delta shoreline reconstruction also suggests a number of discrete fluvial distributaries, building lobes out into the basin, and not a broad-fronted, mass flow-dominated system.

The Kakkaristra Formation is thus probably a braid delta according to McPherson et al. (1987), a fan-delta according to Nemec and Steel (1987), a shelf fan-delta according to Ethridge and Wescott (1984), and could also be termed a wave- and fluvially-influenced fan-delta. No classification takes into account both subaerial and subaqueous parts of the coarse-grained delta system. Combining all aspects, the Kakkaristra Formation could be described as a mass flow- to stream-dominated, wave- and fluvially-influenced, shelf fan-delta. In the following discussion, it is referred to more simply as a fan-delta.

4.4.2 Fan-delta components

A fan-delta is basically composed of three units: subaerial, coastal (transitional), and subaqueous. Controls on each of these three components are now considered.

Subaerial part: as already stated, the subaerial part of the Kakkaristra fan-delta, which occurs in the southern part of the study area, is poorly preserved, and only broad generalisations about the processes operating on it can be made. An alluvial fan setting is generally implied by the lack of laterally continuous, stratified, cross-bedded sands and conglomerates, which are typical of braided environments (e.g. Tucker, 1981). The tectonic setting of the Kakkaristra Formation, in which a nearby hinterland (Troodos Massif) was undergoing uplift (section 4.2.2), is also consistent with this interpretation.
Subaerial debris flow conglomerates (facies A1) are most typical of proximal alluvial fan settings (Collinson, 1986). They are overlain by fine-grained overbank deposits, however, with minor conglomerate and sand (facies A2), suggestive of a more medial to distal fan location. Overbank facies lack preserved organic material, contain incipient caliche, and are partly reddened, suggesting a somewhat arid climate.

A kilometre or two to the north, close to the inferred shoreline of the fan-delta, channelised conglomerates and braided bar deposits occur (facies A3). These sediments suggest that the fringes of the alluvial fans passed laterally into a narrow, partially braided, fringe (Fig. 4.14). The lobate outline of the fan-delta (Fig. 4.2) also points to fan progradation along a number of discrete distributary channels. A similar fan-delta system, containing alluvial fans with braided margins, and of similar size to the Kakkaristra fan-delta (5-7km in radius), is reported from the Miocene of Turkey (Hayward, 1983).

A schematic reconstruction of the inferred Kakkaristra depositional system (Fig. 4.14) shows two small fan-deltas. Proximal parts (now eroded) are very small, and were probably dominated by coarse-grained, mass flow deposits. Minor mass flow deposits reached the medial/distal parts of the system (Fig. 4.15), which probably comprised a few channels and large overbank areas. The hinterland is inferred to have been only 5-7km away from the shoreline.

**Coastal (transitional) part:** this area of a fan-delta system is the most complex, because it is the site of interaction between marine and fluvial processes. In the Kakkaristra fan-delta, fluvial processes were clearly very important, and built a number of protruding, deltaic lobes, out from the shoreline into the basin (Figs. 4.14 and 4.15). Deltaic lobes in general built northwards, although individual river channels entered the basin in a variety of directions (cf. the modern delta of the Mississippi; Gould, 1970), resulting in a wide range of palaeocurrents in mouth bar facies (facies B4; Fig. 4.3). Deposition at the shoreline was both a result of normal stream flow (facies B3 and some of B2) and mass flow processes (facies B4 and some of B2). Mass flow deposits have not been widely documented
Fig. 4.14 – Schematic block diagram reconstruction of the Kakkarista Formation fan–delta system

Fig. 4.15 – Palaeogeographic reconstruction and facies summary, Kakkarista Formation
along the shorelines of fan-deltas (Kleinspehn et al., 1984; Nemec and Steel, 1984).

Between protruding deltaic lobes, interdistributary bays were the site of rather low energy conditions. Bay fringes were at least partly vegetated (facies C2; Fig. 4.15), and muddy sands were deposited during flood inundation.

Marine processes were also important along the fan-delta shoreline, and well sorted sands and conglomerates (facies B1 and B5) were deposited, under the influence of shoaling waves. The prevailing wave energy of the basin was not particularly great, however, as documented by the presence of muds in the foreshore zone (facies D1) and the lack of typical, abundant, shallow, seaward-dipping lamination in foreshore facies. Wave energy was capable of reworking sediment away from direct fluvial input, but was insufficient to smooth the delta shoreline, and generate a cuspate delta margin. Low energy might be expected in the narrow Mesaoria basin (<20km), with limited fetch (Clifton, 1986, however, documents a high energy, conglomeratic, wave-dominated, shoreline sequence from a narrow ancient seaway).

The Kakkaristra fan-delta was thus both wave- and fluvially-influenced. Although basin energy was not particularly great, it was sufficient to modify fluvial processes. In comparison with fine-grained deltas, the Kakkaristra fan-delta is less lobate than the Mississippi delta, but more lobate than the Ebro and Nile deltas (note these latter deltas are much larger than the Kakkaristra fan-delta).

Subaqueous part: this constitutes all subtidal parts of the fan-delta, and includes the fan-delta front (slope and toe) and bay areas. Facies of the fan-delta front are not particularly well exposed. They mainly comprise rather structureless, bioturbated, fossiliferous, slightly muddy sediments (facies C1 and C3b), and probably reflect deposition below fairweather wave base. Minor conglomeratic facies (C5) represent distal mouth bars, and reflect dwindling fluvial influence beyond the delta shoreline. Further north, marine silts and clays (Bear, 1960; Ducloz, 1965; see Fig. 4.2) probably reflect prodelta sedimentation.

Oyster banks are sometimes present in bay areas (Fig. 4.15).
Some signs of current activity are also present, e.g. cross-laminated sands, and reworked shell layers in the eastern and central bay areas respectively (facies C1). Larger-scale sand bar facies (C4) also occur in bay areas, and imply some moderately persistent currents. Bars migrated in a number of directions (Fig. 4.3), relating to individual bar location, direction of opening of bay mouths, and local circulation patterns within bays and around bars.

A special part of the subaqueous fan-delta is the area that lies between the eastern fan-delta shoreline and the nearby field of sand bars, belonging to the Athalassa Formation (Figs. 4.14 and 4.15; chapter 5). Interdigitation of the two formations occurs only over a narrow zone (Fig. 4.2), and the largely bioclastic sediments of the Athalassa Formation, derived from the northern side of the basin (section 5.3.3), did not apparently mix very much with the clastic sediments of the Kakkaristra Formation. Mixing might have been expected, in view of the predominantly south-directed palaeocurrents recorded by the sand bars, as they migrated under the influence of prevailing northerly winds (section 5.3.3). It is possible, however, that additional, more regional currents, generated by circulation in the Mediterranean, may also have flowed through the Mesaoria basin, which was open at both ends. These currents were probably strongest around margins of the Athalassa sand bar field, where resistance to flow was least. Evidence that such currents existed, and were capable of transporting sediment, comes from two facies C4 sand bars (types d and e; section 4.3), located in the zone between the fan-delta and sand bar field. These bars migrated under the influence of currents flowing from NW to SE through this zone (Figs. 4.3 and 4.15). It is suggested that these currents partly inhibited interdigitation of the Kakkaristra and Athalassa Formations by transporting sediment eastwards out of the transition zone.

Further, though rather speculative, evidence in support of unidirectional currents operating close to the eastern part of the fan-delta, comes from facies C3a, which may be a form of hummocky cross-stratification (HCS). It is postulated that enhanced unidirectional flow (?storm-amplified) may have combined with oscillatory motion to generate a mild form of HCS. The lack of detailed structure and palaeocurrents in facies C3a makes this,
4.5 Controls on Sedimentation

Lateral and vertical facies sequences in a sedimentary succession document the response of sedimentation to both extrabasinal controls, e.g. tectonic, eustatic or climatic changes, and to intrabasinal, sedimentological controls, e.g. channel avulsion, river mouth bar aggradation or abandonment, or variable wave-reworking (e.g. Clifton, 1981; Kleinsphön et al., 1984). The influence exerted by these controls on the Kakkaristra Formation is now addressed.

4.5.1 Extrabasinal controls

Initiation of fan-delta sedimentation at the beginning of Kakkaristra Formation times has already been attributed to tectonic effects (section 4.2.2). Detailed facies observations at the base of the formation reflect this tectonism. Deformed sediments of facies B4 (cross-bedded, conglomeratic, Gilbert-type mouth bars) are attributed to high initial sedimentation rates, slope steepening and instability, as the Kakkaristra fan-delta first prograded into the Mesaoria basin, following uplift. Laterally equivalent to these sediments are gravelly beach sediments of facies B6. These conglomerates are relatively poorly washed when compared with beach conglomerates occurring stratigraphically higher in the Kakkaristra Formation (facies B5). This is again attributed to initial high sedimentation rates, lack of time for extensive wave-reworking, and rapid burial.

Sea level fluctuations were also undoubtedly occurring at this time as well, as they are documented in the Athalassa Formation (section 5.5), which is laterally equivalent to the Kakkaristra. Eustatic effects are difficult to separate from those of tectonic origin, however, in tectonically active areas (Miall, 1984), and they have not been specifically identified in the Kakkaristra Formation. They may have hindered fan-delta progradation, if sea level was rising, or enhanced it, if sea level was falling.

A change in climatic conditions might have occurred from Nicosia to Kakkaristra Formation times. Reddening and lack of preserved organic material in subaerial overbank facies suggest somewhat arid
conditions during deposition of the Kakkaristra Formation. Climatic conditions during deposition of the Nicosia Formation are not known, however. An arid climate during deposition of the Kakkarista Formation could have implications for eustatic effects. Aridity in mid latitudes, however, cannot be confidently equated with glacial or interglacial periods during the Quaternary. For example, during the glacial maximum of ca. 18,000 years ago, the north side of the Mediterranean was apparently rather dry, yet high lake levels from North Africa suggest the south side was moist (Goudie, 1984).

4.5.2 Intrabasinal controls

A wide variety of facies (Table 4.1) and facies transitions occur in the Kakkaristra Formation, in spite of its thinness. A facies relationship diagram (Fig. 4.16), and cross-sections showing lateral and vertical facies changes (Figs. 4.4-4.7) were constructed, to see if distinctive facies patterns were discernible. (N.B. Markov chain analysis was not attempted because sophisticated statistical methods are required to achieve meaningful results in this type of process (Walker, 1984a), and a large number of facies transitions need to be recorded).

Fig. 4.16 was constructed to show delta front/bay fill facies at the base, shoreline facies in the centre, and delta top pond and fluvial channel facies at the top. 20 transitions can be traced in an upward direction, and only 13 downwards. This largely reflects the gross prodelta to delta top sequence documented by the Kakkaristra Formation, which in turn forms part of the shallowing-up cycle of the Mesaoria basin, progressively recorded by the Nicosia, Kakkaristra and Apalos Formations. Downward and lateral transitions, however, reveal intimate facies interdigitation. No laterally continuous, widespread trends were identified within these transitions.

Cross-sectional reconstructions amply reflect the highly variable facies patterns. For example, the logs and interpreted section of Fig. 4.4 show a bay environment existed at the southern end of the section (log 1) during early Kakkaristra Formation times, with bay bar developed in the middle of the section (log 3). The shoreline encroached on the bay area in mid-upper Kakkaristra Formation
Fig. 4.16 - Facies relationship diagram, Kakkaristra Formation.
times, and prograded across part of it. Shoreline conditions prevailed throughout the formation in the vicinity of log 5, though with variable facies, i.e. mouth bar progradation, abandonment and limited colonisation by oysters and coral, establishment of sandy shoreline facies, and finally further mouth bar progradation. Delta front facies are found further north (log 6).

Similar reconstructions can be made for other cross-sections. As no widespread patterns were identified, detailed facies changes within the Kakkaristra Formation are attributed to local, sedimentological, intrabasinal processes, and not to tectonic or eustatic controls. The vertical facies pattern in log 5 (Fig. 4.4), just described, for example, is probably the result of channel switching events.

In total, intrabasinal processes have resulted in an intimate stacking of fluvial, shoreline and subaqueous facies, with the most commonly preserved transitions occurring between shoreline facies (B1, B4 and B5; Fig. 4.16).

In summary, extrabasinal controls influenced the initiation of deposition of the Kakkaristra Formation, while intrabasinal processes controlled detailed facies changes.

4.6 Summary - Basin Evolution during Kakkaristra Formation Times

1. A pulse of uplift of the Troodos Massif initiated deposition of the Kakkaristra Formation. A partial unconformity was generated and a large fan-delta system prograded into the shallow Mesaoria basin. Sea level fluctuations may also have had an influence, but are inseparable from tectonic effects.

2. Initial high sedimentation rates and influx of fluvially-transported sediment resulted in deposition of thick, oversteep, river mouth bars and poorly washed beach conglomerates along the fan-delta front.

3. Fluvially-influenced deltaic lobes built out into a relatively low energy basin. Wave reworking took place away from direct fluvial input, and fine-grained sediments accumulated in bay areas and along the subaqueous fan-delta front. Subaerial alluvial fan facies are poorly preserved.
4. Local intrabasinal processes controlled detailed facies changes, while the formation in general records a gross prodelta to delta top sequence.

5. Complete emergence of the basin is heralded at the end of the Kakkaristra Formation by the incoming of caliche horizons.
Plate 4.1

a) Facies B1 sands, showing parallel lamination (above conglomerate horizon) and trough cross-lamination (below conglomerate horizon), emphasised by dark, heavy mineral concentrations; conglomerate horizon represents facies B2, type a; vertically, field of view is ca. 18cm

b) Facies B2 conglomerates, interbedded with slightly pebbly facies B1 sands; thin conglomerate represents subfacies type e, with concave-up base and convex-up top; upper conglomerate represents thick, unstratified subfacies type c; vertically, field of view is ca. 180cm

c) Interbedded facies B2 conglomerates and facies B1 sands; upper conglomerate represents inversely graded, subfacies type f; middle conglomerate represents cross-bedded, subfacies type g; portion of staff showing is ca. 56cm long
Plate 4.2

a) Facies B3 trough cross-laminated sands, showing broad, shallow troughs filled with laminated, vertically burrowed sands; hammer at top is 30cm long

b) Long distance view of cross-bedded facies B4 conglomerates; photo represents a ca. perpendicular cross-section through a facies B4 mouth bar; cross-bedding shallows to right; car at top left for scale

c) Poorly sorted and stratified cross beds of facies B4 conglomerates; portion of staff showing is ca. 67cm long
Plate 4.3

a) Better sorted and stratified cross beds of facies B4 conglomerates; hammer is 30cm long

b) Close-up of a), showing conglomerate layers which are clast-supported, well sorted, and contain subrounded, sometimes flattened and up-dip imbricated clasts

c) Facies B4 mouth bar top, comprising a thin, lower conglomeratic section containing reworked pebbles, a middle section rich in oyster debris, and coral development (arrowed) in the upper section; cross beds of main body of mouth bar are not visible in dark area at bottom right; vertically, field of view is ca. 1m
Plate 4.4

a) Facies B4 mouth bar conglomerate, showing very well developed cross beds of main mouth bar body, less well sorted, conglomeratic bar top interval, and overlying facies B1 sands; staff is 110cm long

b) Close-up of a), showing a virtually matrix-free, well sorted, imbricated conglomerate cross bed; note lenticular shape
Plate 4.5

a) Gently dipping, very well cross-bedded beach conglomerates of facies B5; white, vein-like material in upper right (opposite the letter a) is caliche development at the top of the facies at the top of the Kakkaristra Formation; reddened facies of the Apalos Formation overlie this horizon; staff is 110cm long

b) Close-up of a), showing well sorted conglomerate with clasts imbricated down-dip (compare with Plate 4.4b); staff is marked off in 10cm intervals

c) Hillside composed of structureless, largely grass-covered, facies C1 muddy sands, overlain by tabular, cross-stratified sands of shallow marine bar facies C4 (cross beds are not visible because section is perpendicular to palaeocurrent direction); hillside is ca. 10m high
Plate 4.6

a) Shelly horizons in facies C1 muddy sands from the central bay area (fig. 4.2); shells belong to one species (Venus sp.) and form impersistent lenses; hammer is 30cm long

b) Fossil pine cone (see Appendix 1), recovered from facies C1 muddy sands, western bay area (Fig. 4.2); specimen is 9cm long

c) Poorly bedded, internally structureless, facies C2 rootlet-bearing sands, showing thin, subvertical, root traces
Plate 4.7

a) Oyster bed developed within facies C3b poorly bedded sands; articulated valves and small oyster clumps are present in the debris; hammer is 30cm long

b) Bedded and cross-laminated facies C4 shallow marine bar sands (type a); hammer is 25cm long
Plate 4.8

a) Thin beds of facies D1 white mudstones, interbedded with facies B1 sands; some layers show partially rippled bases and tops; second layer from top on left shows flamed top; thick layer in middle is broken and ends abruptly; vertically, field of view is ca. 75cm

b) Structureless muds of facies D2, overlying facies B1 sands; thin, horizontal, calcareous partings are present at the top; mud interval is ca. 2m thick

c) Scanning electron photomicrograph of facies D2 muds, showing an abundance of detrital clay flakes
5.1 Introduction

The Athalassa Formation (sections 2.2.1 and 2.4) crops out in the eastern part of the study area (Encl. A), and also in the northern part of the Mesaoria basin, along the southwest and south flanks of the Kyrenia Range (Fig. 5.1). Northern exposures of the formation could not be investigated during this study, because they lie close to, or beyond, the Green Line. Published descriptions of these rocks are available, however (Moore, 1960; Lytras, 1962, 1963; Baroz, 1979), and this literature is reviewed in section 5.4.

Stratigraphic correlation of the Athalassa Formation with other formations in the basin is not entirely clear because of the limits of exposure. Field relations in the southeast part of the study area show that the Athalassa and Kakkaristra Formations are lateral equivalents and interdigitate, though only over a limited zone (see Fig. 5.6). The base of the Athalassa Formation (taken at the first occurrence of a calcarenite interval; see section 2.2.1) is ca. 10m below that of the Kakkaristra Formation, so that the lower part of the formation and the topmost part of the underlying Nicosia Formation are equivalent (see Table 5.1).

The Kakkaristra Formation is only ca. 15m thick, while the Athalassa Formation is ca. 50m thick. The upper part of the Athalassa Formation is thus laterally equivalent to the lower part of the Apalos Formation, which overlies the Kakkaristra (Table 5.1). Interdigation between the Athalassa and Apalos Formations has not been directly confirmed in this study because of poor exposure in the central Mesaoria Plain. Some facies in the continental Apalos Formation show the influence of a nearby shoreline, however (section 6.3.2), suggesting the presence of an adjacent marine environment. This marine environment is documented by the Athalassa Formation.

The thickness of the Athalassa Formation has not been directly measured in this study, because the top of the formation is either unexposed, or exposed only in zones of restricted access. Ducloz (1965) identified Apalos facies overlying Athalassa facies in the vicinity of Nicosia (now no longer exposed), and gave the formation a
Table 5.1 – Relationships between the Athalassa, Nicosia, Kakkaristra and Apalos Formations.
Fig. 5.1 - Outcrop area of the Athalassa Formation (stippled; slightly modified from the Cyprus Geological Survey geological map of Cyprus). Only exposures of the formation in the east were examined during this study.
thickness of ca. 50m. Borehole data from the eastern Mesaoria suggest a thickness of ca. 80m (see Fig. 5.10).

The base of the Athalassa Formation is conformable over, and partly interdigitates with, the underlying Nicosia Formation in the study area. To the northwest, however, the relationship becomes unconformable, and a low-angle unconformity separates the two formations. The formation also thickens to the northwest, and near the Ovgos fault (Fig. 5.1), is about 150m thick (Moore, 1960).

In the study area, the Athalassa Formation mainly comprises a series of shallow marine, cross-stratified sand bodies, separated by finer, muddier facies (see Plate 5.6c). These shallow marine sediments contrast with the fan-delta facies of the Kakkaristra Formation, which crop out to the west (Fig. 5.1). Sands of the Athalassa Formation are characteristically enriched in skeletal carbonate material (always greater than 40%; see section 8.2). In order to distinguish them from the clastic-rich sands of the Kakkaristra Formation, they are referred to as calcarenites, although skeletal carbonate content is not strictly always greater than 50%.

The shallow marine facies of the formation are described in the following section, before sedimentological models and geological implications are considered.

5.2 Facies and Facies Relations

Seven lithofacies, of which one has been subdivided into four subfacies, are recognised in the Athalassa Formation (Table 5.2). Facies are grouped on the basis of broad depositional environment. Facies of group A represent the shallow marine calcarenite bodies so characteristic of the formation. Group B contains a single, finer-grained, largely structureless sand facies, representing interbar sedimentation. Facies of groups C and D belong to nearshore and coastal environments.

**Facies A1 - Planar-stratified calcarenites**

*Description:* this facies occurs in two forms: a) parallel-laminated, fine-grained calcarenite, which occurs in thin, planar beds, 5-20cm thick, or lenses up to 1 or 2m in length (Plates
Table 5.2 - Lithofacies, Athalassa Formation

<table>
<thead>
<tr>
<th>Code</th>
<th>Lithofacies</th>
<th>Depositional environment</th>
</tr>
</thead>
<tbody>
<tr>
<td>A1</td>
<td>Planar-stratified calcarenites</td>
<td></td>
</tr>
<tr>
<td>A2</td>
<td>Cross-stratified calcarenites</td>
<td></td>
</tr>
<tr>
<td>A2a</td>
<td>Cross-laminated calcarenites</td>
<td>Shallow marine sand bodies</td>
</tr>
<tr>
<td>A2b</td>
<td>Cross-bedded calcarenites</td>
<td></td>
</tr>
<tr>
<td>A2c</td>
<td>Large-scale, cross-bedded calcarenites</td>
<td></td>
</tr>
<tr>
<td>A2d</td>
<td>Ripple cross-laminated calcarenites</td>
<td></td>
</tr>
<tr>
<td>B1</td>
<td>Fine, massive sands</td>
<td>Background shelf sands</td>
</tr>
<tr>
<td>C1</td>
<td>Pebbly calcarenites</td>
<td>Beach to shallow subtidal facies of a barrier island</td>
</tr>
<tr>
<td>C1a</td>
<td>Wedge cross-stratified calcarenites</td>
<td></td>
</tr>
<tr>
<td>C1b</td>
<td>Planar to lenticular-bedded calcarenites</td>
<td></td>
</tr>
<tr>
<td>D1</td>
<td>White muds</td>
<td></td>
</tr>
<tr>
<td>D2</td>
<td>Brown muds and associated sands</td>
<td>Lagoons/ponds</td>
</tr>
<tr>
<td>D3</td>
<td>Wackestone</td>
<td></td>
</tr>
</tbody>
</table>
5.1a and 5.5c); bed-aligned macrofossiliferous debris is sometimes present; b) planar-bedded, fine- to coarse-grained calcarenite, which forms beds 5-40cm thick; these beds are apparently internally structureless (Plate 5.1b), and have flat, non-erosive to mildly erosive bases.

Occurrence, lateral and vertical facies relations: the facies forms part of the numerous, thin to thick calcarenite bodies which occur in the Athalassa Formation, separated by finer, muddier facies B1 (e.g. logs 15/11/4 and 15/11/6, Fig. 5.2); planar-laminated facies are typically interbedded with, and pass laterally into, facies A2a; occasionally they form the dominant facies in parts of calcarenite bodies (Plate 5.1a); planar-bedded calcarenites form intervals up to 2m thick, and are often found in thicker calcarenite bodies (e.g. log 26/11/1, Fig. 5.5); the facies occurs throughout the study area (Fig. 5.6).

Interpretation: parallel lamination in fine-grained sediments is either the result of deposition from suspension in a quiet depositional setting, or the result of sedimentation during upper flow regime conditions, when current velocities were high; in view of the interfingering of this facies with cross-laminated calcarenite (facies A2a), the latter interpretation is more likely; planar-bedded calcarenites are probably partly structureless due to weathering, and perhaps also due to the lack of clarity of structures often associated with carbonate sands (Tucker, 1981); much of this subfacies was probably deposited along with parallel-laminated and cross-laminated calcarenites during normal traction current processes.

Facies A2 - Cross-stratified calcarenites

Four subfacies are recognised within this facies. Their sedimentary structures, occurrence and lateral and vertical facies relations are described separately, before interpretations are discussed together at the end.

Facies A2a - cross-laminated calcarenites: this subfacies comprises fine- to occasionally medium- grained, tabular and trough cross-laminated calcarenite, in beds or lenses 20-40cm thick (Plates 5.1b and c); tabular cross-lamination is less common than trough,
Fig. 5.2 - Sedimentological logs, cross-section A-A' (see Fig. 5.6 for section location and Encl. B for key to logs).

Fig. 5.3 - Sedimentological logs, cross-section B-B' (see Fig. 5.6 for section location and Encl. B for key to logs).
Fig. 5.4 - Sedimentological logs, cross-section C-C’ (see Fig. 5.6 for section location and Encl. B for key to logs).

Fig. 5.5 - Sedimentological logs, cross-section D-D’ (see Fig. 5.6 for section location and Encl. B for key to logs).
with laminae dipping up to 20°, and often having tangential bases; rarely, low-angle tabular cross-lamination is preserved in thin wedges; trough cross-lamination occurs in troughs up to 1m wide, and sometimes fills erosional scours at the base of calcarenite bodies (e.g. log 5/12/4, Fig. 5.2; Plate 5.2a); cross-laminated calcarenites are a common facies in the Athalassa Formation (Fig. 5.6); they are typically interbedded with, or pass laterally into, facies A1 calcarenites; they form the dominant sedimentary structure in some thin calcarenite bodies (e.g. log 14/10/3, Fig. 5.2); occasionally, the facies is bioturbated, and thin, curving cylindrical burrows may be present on bedding planes.

**Facies A2b - cross-bedded calcarenites:** this subfacies comprises medium- to coarse-grained, cross-bedded calcarenite, in units 1-3m thick (Plate 5.2b); units are either made up of single, cross-bedded sets, or 2-3 sets, sometimes with varying orientation, which may be separated by low-angle truncation surfaces (Plate 5.2c); cross beds are 3-30cm thick, dip at 10°-30°, and typically have tangential bases; cross bed sets are largely trough-shaped perpendicular to flow, and up to 100m wide; thin, subhorizontally-laminated bottom sets sometimes occur, as do rare cross-bedded lenses with orientation opposite to that of the main cross bed set; cross-bedded calcarenites have sharp or gradational bases over facies B1 sediments and sharp tops, which may be littered with oyster debris (e.g. log 8/1, Fig. 5.5); cross beds sometimes have a 'nobbly' appearance as a result of the presence of *Thalassinoides*-type burrows; at one locality (WD 827367; log 27/11/2, Fig. 5.4), a calcarenite body comprises a series of erosive cross bed sets, some of which are sigmoidal in shape (Plates 5.3 and 5.4a); this unusual calcarenite unit overlies a thick, vaguely bedded oyster and barnacle bank (facies C3b of the Kakkaristra Formation; section 4.3); at this locality and others, cross-bedded calcarenite passes up-current into planar-stratified calcarenite (facies A1), sometimes with facies A2a; this subfacies is less common than facies A1 or A2a, and is found throughout the study area (Fig. 5.6).

**Facies A2c - large-scale, cross-bedded calcarenites:** this subfacies is similar to facies A2b, except that it occurs in much larger units, 5m-25m thick (Plate 5.4b), and 100-several 100m wide;
like facies A2b, cross beds are thin to thick, often tangentially based, and form large, trough-shaped units (Plate 5.4c); basal and lateral relations are uncertain because of lack of exposure; vertically, large-scale, cross-bedded units may overlie each other (e.g. log 27/1, Fig. 5.2; Plate 5.4c); small cross bed sets, with varying orientations, are rarely interbedded in the facies (e.g. log 5/10/2, Fig. 5.4); curved, large-scale cross beds are visible on aerial photographs of the study area; the facies is not common, occurring at only two localites (Fig. 5.6).

**Facies A2d - ripple cross-laminated calcarenites**: this subfacies comprises fine-grained, ripple cross-laminated calcarenite lenses; cross laminae have irregular lower set boundaries, may be form discordant, and are of wave origin; symmetrical ripple bedforms are rarely preserved on bed tops (Plate 5.5a); lenses of this facies are only 10-20cm thick, up to 70cm long, and pass laterally and vertically into facies A2a calcarenites; the subfacies is not very common (Fig. 5.6), and only occurs in the lower parts of thin calcarenite bodies, which otherwise comprise facies A1 and A2a (e.g. log 5/12/4, Fig. 5.2).

**Palaeocurrents**: palaeocurrent directions for cross-stratified subfacies have been plotted together in Fig. 5.7; some individual readings are shown on Fig. 5.6; a very strong predominance of SSE to SW directions is evident.

**Interpretation**: sediments of these subfacies are the product of migration of a variety of sandy bedforms; these bedforms were shallow marine in origin, as witnessed by their fossiliferous content, interbedding with marine facies B1, and the occasional occurrence of the typical shallow marine trace fossil, *Thalassinoides*; cross-laminated subfacies A2a records migration of straight- and sinuous-crested megaripples; this subfacies is found most typically in thin, laterally extensive, sheet-like calcarenite bodies (see also section 5.3.3), across the top of which megaripples migrated; occasionally, sheet tops were partially reworked by waves, perhaps during fairer weather periods, to give facies A2d rippled sediments; subfacies A2b and A2c reflect migration of larger-scale sand bodies, at least up to 5m high (section 5.3.3); these sand bodies were largely asymmetrical, with steep downcurrent faces and much gentler,
Fig. 5.6 - Facies distribution & schematic palaeogeography, Athalassa Formation. Stippled areas represent known locations of sand bodies, which were not all necessarily developed at the same time. Other bodies may also have been present. Arrows represent principal cross bed directions. See Fig. 5.1 for map location and Table 5.1 for lithofacies.

Fig. 5.7 - Palaeocurrent data, Athalassa Formation (see also Fig. 5.6 above).
subhorizontal lee slopes (Fig. 5.8); low-angle truncation surfaces reflect minor changes in migration direction; bedforms were occasionally intensely bioturbated; rarely, abandoned bar tops were colonised by oysters; bedforms largely migrated southwards across the Mesaoria basin (Fig. 5.7), probably under the influence of south-blowing storms, evidence of tidal influence e.g. herring-bone cross-stratification, mud drapes or numerous reactivation surfaces, being absent (see further, section 5.3.3); the calcarenite body with sigmoidal cross beds (log 27/11/2, Fig. 5.4), which dip east, is anomalous; this structure was probably influenced by migration across a large, pre-existing, roughly N-S trending, oyster bank (facies C3b, Kakkaristra Formation); normal, southerly transport of the calcarenite bedform was diverted across the east-facing slope of the bank; deposition over the increased slope resulted in a series of erosively stacked, sigmoidal-shaped cross bed sets; depositional processes associated with facies A2 are further discussed in section 5.3.3, where comparisons with modern and ancient examples are also made.

Facies B1 - Fine, massive sands

Description: the facies comprises very fine to fine, moderately sorted, slightly muddy, orange-yellow sands, which are poorly consolidated, generally structureless and thoroughly bioturbated; occasionally, vague parallel lamination is visible (Plate 5.5c); the sands are fossiliferous, mostly containing broken mollusc fragments; coral fragments are recorded from one locality, as are rare brachiopods; thick accumulations of oyster debris (including some articulated valves) are found at several localities (e.g. log 14/10/3, Fig. 5.3); sediments of this facies are very similar to those of facies A2 of the Nicosia Formation (section 3.3.1); thin cemented layers, however, which are quite common in facies A2, are relatively uncommon in this facies; the facies may be up to 15m thick.

Occurrence, lateral and vertical facies relations: this facies is common throughout the Athalassa Formation, and throughout the study area (Fig. 5.6), especially in the south; it typically occurs in thick intervals between calcarenite bodies containing facies A1 and A2, e.g. logs of Fig. 5.2.
Interpretation: the similarity with facies A2 of the Nicosia Formation suggests a similar depositional environment i.e. shallow marine shelf (see section 3.3.1); the intercalation of calcarenite bodies, deposited largely by storm-generated currents (see section 5.3.3), indicates deposition above storm wave base, but any signs of storm reworking in this facies have largely been destroyed by bioturbation; oyster debris, perhaps transported during storms, accumulated in shell hashes; in the northern part of the study area, the facies becomes rich in plant debris, perhaps reflecting proximity to the northern margin of the Mesaoria basin.

The following facies (groups C and D) are less common than those previously described, and occur only in the northern part of the study area (Fig. 5.6). Facies changes in this area are numerous and complex, and not always clear, because lateral continuity of exposure is poor.

**Facies Cl - Pebby calcarenites**

**Description:** this facies comprises two types of pebbly sediment: type a) sorted, fine- to medium- grained calcarenite, displaying shallow-angle, tabular cross-lamination in wedge-shaped sets, up to 40cm thick and 1m long (Plate 5.5b); some layers are very shelly, and may also be pebbly, with shell debris and pebbles aligned parallel to lamination; lamination is sometimes highlighted by dark mineral layers (Plate 5.5b); wedge-shaped cross bed sets are stacked laterally and vertically, and have variable orientations, mainly between NE and SE (Fig. 5.7); the subfacies passes laterally into rather poorly cross-laminated calcarenite, which is burrowed and contains a horizon with whole, possibly in situ bivalves (log 8/2b, Fig. 5.5); rootlets are present at the top of this interval; type b) weathered, thin-bedded (5-15cm), fine, well sorted, slightly pebbly calcarenite in planar to lenticular beds (log 8/2a, Fig. 5.5); some bed bases are covered with scattered, bed-parallel macrofossil fragments and pebbles; other pebbles are scattered throughout the sediment; internal structure within beds is not apparent.

**Occurrence, lateral and vertical facies relations:** both subfacies types occur only at one locality in the northern part of the study.
area (Fig. 5.5 and 5.6); type a is 4m thick, type b 2m thick; type a overlies facies B1 and contains thin intercalations of facies D1 muds; it is separated from type b above by facies D1 and D2 lagoonal sediments; subfacies b is overlain by facies A2b.

Interpretation: the wedge-shaped, shallow-angle, tabular cross-laminated subfacies a, with heavy mineral lamination, strongly suggests a beach environment (Elliot, 1986a); mudstone intercalations (facies D1) are not common in beach settings, but are occasionally documented (e.g. Wünderlich, 1972; van den Berg, 1977), and are present in beach facies of the Kakkaristra Formation (facies D1, section 4.3); the occurrence of lagoonal facies D2 north of the location of subfacies a, but at the same stratigraphic level, indicates that this subfacies may represent a barrier island, rather than shore-attached beach; palaeocurrents (Fig. 5.7) are rather variable and may reflect a component of longshore drift, resulting in oblique migration of beach ridges; the barrier island may also have been curved, facing S to SSE (Fig. 5.6); the poorly laminated, burrowed and fossiliferous part of this subfacies may represent lower energy deposition on the lee side of the barrier island, which became emergent and was colonised by plants; subfacies b shows no evidence of emergence; interpretation of this subfacies is very uncertain; its lenticular bedding is reminiscent of facies C3b of the Kakkaristra Formation (section 4.3), tentatively interpreted as possible hummocky cross-stratification; this implies a subtidal setting, seaward of the higher energy beach environment; cross-lamination would normally be expected to be preserved, but is probably obscured by weathering; good sorting of the facies may reflect reworking of beach sediment offshore.

**Facies D1 - White muds**

Description: the mudstones of this facies are calcareous, white or cream-coloured, sometimes silty, and either structureless or finely laminated; lamination is usually horizontal, but sometimes wavy and disturbed (Plate 5.6a); thin sand flasers are sometimes present; jagged, vertical, sand-filled desiccation cracks are common (Plate 5.6a); individual beds range from minor lenses, a few millimetres thick, to beds up to 30cm thick; bed bases are sometimes
spectacularly symmetrically rippled (Plate 5.6b); in places, beds are sharply truncated, and have been reworked as rounded mudclasts in the surrounding sediment; an horizon of whole, possibly in situ, bivalves occurs in the lower part of one bed; many of the features of this facies are similar to facies D1 of the Kakkaristra Formation (section 4.3).

**Occurrence, lateral and vertical facies relations:** the facies is found at only one locality in the northern part of the study area (log 8/2a, Fig. 5.5); it occurs as four to five thin intercalations within facies C1a beach sediments.

**Interpretation:** like facies D1 of the Kakkaristra Formation, which is also intercalated with beach sediments, the mudstones of this facies are interpreted to have accumulated in small pools in the lee of beach ridges and spits, developed along a relatively low energy foreshore (cf. Wünderlich, 1972; van den Berg, 1977); some lee areas had well developed wave-rippled surfaces; mud was probably introduced into pools by wind and large waves; lamination is mainly due to suspension fallout; desiccation cracks developed and were filled with sand; strong, ?storm-generated waves occasionally breached beach ridges, ripped up semiconsolidated mud and reworked it as mudclasts; mud ponds were occasionally colonised by bivalves; muds were probably largely derived from the chalky and marly pre-Pliocene sedimentary cover of the Troodos Massif (Lefkara and Pakhna Formations, section 1.3.3), hence their very calcareous nature (see section 8.2, clays subsection).

**Facies D2 - Brown muds and associated sands**

**Description:** this uncommon facies comprises a 1m thick unit of brown, very fine, well sorted sand, which is poorly consolidated and partly parallel-laminated; it is overlain by brown, slightly sandy, mudstone, which is apparently structureless apart from a few sand-filled desiccation cracks (log 8/2a, Fig. 5.5); the sand is partly bioclastic in composition; the mudstone is unfossiliferous.

**Occurrence, lateral and vertical facies relations:** this facies occurs at only one locality on the north side of the study area (Fig. 5.6), where it overlies facies C1a (beach sediments) and is overlain by facies C1b (shallow subtidal sediments).
Interpretation: the rather structureless to parallel-laminated, fine-grained nature of the facies suggests a low energy environment, and the association with barrier island facies C1a indicates a possible lagoonal setting, with sedimentation largely from suspension; lack of exposure prevents a more detailed interpretation.

Facies D3 - Wackestones

Description: this facies comprises very fine- to medium-sized bioclastic and clastic sand grains, usually moderately sorted, set in a pale grey to brown mudstone; large fossil fragments tend to be horizontally aligned, but strong layering is not evident; whole fossils are not found; the facies forms single beds, up to 60cm thick, and of unknown lateral extent.

Occurrence, lateral and vertical facies relations: the facies is found only in the northern part of the study area (Fig. 5.6) where it occurs at several stratigraphic levels; it is usually found in association with facies A2 calcarenites, which under- or overlie it (e.g. log 26/11/1, Fig. 5.5), or may be found overlying facies B1 sands (e.g. log 8/1, Fig. 5.5); facies relations and geometry are not entirely clear because of limited exposure.

Interpretation: wackestones are typical of low energy, mainly mud-accumulating environments; the location of the facies on the north side of the basin, where lagoonal and barrier island facies (C1a, C1b, D2) are also found, suggests that this too may be a lagoonal facies; the association with facies A2 (calcarenite sand bodies) and facies B1 (shelf floor sands) suggests, however, that quiet areas may have formed in the lee of carbonate sand bodies developed close to the northern shoreline of the basin (see Fig. 5.6); these areas were perhaps protected from normal shelf floor processes, and very fine sediment was able to accumulate, with only minor influx of clastic and bioclastic sand.

5.3 Sedimentological Model

5.3.1 General model

Sediments of the Athalassa Formation were deposited in a shallow marine, predominantly subtidal, environment. They fall into three
broad groups. The first contains the ubiquitous facies Bi (slightly muddy, bioturbated shelf sands), which is present throughout the formation (see Figs. 5.2 - 5.6). This facies can be considered to represent background sedimentation during Athalassa Formation times, on a sandy, shallow marine shelf. The second group comprises facies A1 and A2a-d, which go to make up the numerous thin to thick calcarenite bodies present within the formation. These calcarenite sand bodies developed on, and migrated across, the shallow Athalassa shelf. The small third group contains the remaining facies (facies C1a and b, D1, D2 and D3), which occur only in the northern part of the study area. These relatively uncommon facies represent a variety of nearshore to coastal environments, including barrier island, beach and lagoon, and suggest the nearby presence of the northern margin of the Mesaoria basin.

Each of these three groups is discussed in turn below.

5.3.2 Background shelf sediments

Group one sediments are represented by one facies, facies Bi. These sediments are slightly muddy, very fine to fine sands, largely structureless due to extensive bioturbation. They document shallow marine sedimentation on a relatively low energy shelf. They are very similar to facies A2 of the upper Nicosia Formation, which is stratigraphically directly below the Athalassa Formation. The shallow shelf, which evolved towards the end of Nicosia Formation times (see model 3, section 3.3.2), thus continued to be in existence during deposition of the Athalassa Formation. Water depths on the shelf are not known exactly, but were largely above storm wave base. The facies resembles the predominant facies of many other shallow marine shelves, both ancient and modern, which receive or received plentiful supplies of sediment e.g. the bioturbated silty sands of the Bering shelf (Nelson, 1982) or Gulf of Gaeta in the Mediterranean (Reinick and Singh, 1973), bioturbated shelf sandstone facies of the Cretaceous Western Interior Seaway of the U.S.A. (Tillman and Martinsen, 1984), or unbedded fine sandstone of the Miocene of California (Clifton, 1981).
5.3.3 Carbonate sand bodies

Group two sediments comprise facies A1 and A2a-d, which represent a series of cross-stratified, carbonate sand bodies, which migrated across the shallow Athalassa shelf. Sandy bedforms are known from a variety of modern shelf settings, including tidal-, storm- and oceanic current-dominated. Bedform size and geometry are well documented from these areas, but internal structures are still poorly known. Furthermore, descriptions of ancient shallow marine sand bodies are surprisingly few (Walker, 1985b), and their interpretation is often uncertain. Comparison of the Athalassa Formation sand bodies with other examples is thus relatively meagre.

The Athalassa Formation sand bodies display a variety of geometries and sedimentary structures, and four sand body types have been identified (Fig. 5.8). These are described below, before depositional processes are considered in detail.

**Type I sand bodies** are thin (2m or less), sharply to erosively based, may show coarsening-up from fine to medium calcarenite, and are dominated by small-scale, trough and tabular cross-lamination. Planar or very shallow-dipping cross-lamination may also occur, and occasionally ripple cross-lamination is preserved in the lower parts of this sand body type. The sand bodies are sometimes traceable in a direction parallel to flow for at least 2km. Perpendicular to flow, they have only been traced for ca. 4km. This may suggest a ribbon-like geometry, but lateral dimensions normal to flow are uncertain. Type I sand bodies are termed sheets therefore. Sheets on the sea floor were largely covered by megaripples (Fig. 5.9). Accretion took place laterally as megaripples migrated, and vertically as new megaripples trains were initiated, leading to sheets up to 2m thick. Thin sand sheets with erosive bases have been described from the shallow marine sediments of the Upper Devonian of southern Ireland (MacCarthy, 1987).

**Type II sand bodies** are up to 3m thick, coarsen up from fine to medium (or coarse) calcarenite, and have sharp or gradational bases. The down-current parts of these bodies comprise a single set of cross-bedded calcarenite (facies A2b; Fig. 5.8), dipping up to 30°, while up-current parts are gently sloping to subhorizontal, and dominated by thin, planar-bedded calcarenite (facies A1).
Fig. 5.8 - Carbonate sand body types. (see Encl. B for key to symbols).
Fig. 5.9 - Block diagram reconstruction for the NE Mesaoria basin, showing the low-lying Kyrenia landmass to the N (right) and shallow marine sand bodies migrating southwards across the shelf (to the left). Coastal facies document transgressive and regressive events (shown only schematically here).

Fig. 5.10 - Summary logs for the Athalassa Formation (compiled from this study, Moore, 1960, & Lytras, 1962; see Encl. B for key).
Thalassinoides burrows may be common in this sand body type and bar tops are sometimes littered with oysters. One good exposure of this sand body type (WD 342867) is ca. 250m long parallel to flow and 1km normal to flow, suggesting sand body elongation mainly normal to flow direction. Sediment accreted to these sand bodies both laterally as they migrated down-current, but also vertically, as sediment was added to lee slopes. These sand bodies thus stood above the sea floor (Fig. 5.9), and are referred to as ridges. Sand bars of the Cretaceous Western Interior Seaway of the U.S.A., like type II sand bodies, coarsen up, are cross-stratified and have gradational bases. They are much larger than the sand bodies of the Athalassa Formation, however, typically show a well developed, interbedded bar - interbar unit, and may contain hummocky cross-stratification (Walker, 1984b).

Type III sand bodies involve very large sets of cross-bedded calcarenite (facies A2c), from 5 to 25m in thickness (Fig. 5.8). Basal relations, detailed structures and lateral dimensions are unknown because of lack of continuous exposure or lack of access to exposure. It is unknown, therefore, how large the bedforms were that generated these structures. A good example of the facies (WD 385798) has a large, upper trough cross-bedded set, ca. 5m thick and 200m wide, suggesting the presence of crescentic sand waves up to at least 5m high on the sea floor. These very large sand waves were not common. Other ancient shallow marine sand bodies, with very thick cross bed sets, are usually documented from tidal settings e.g. Cretaceous Upper Woburn sands of southeast England (Bridges, 1982). There is no evidence for tidal influence in the Athalassa Formation, however (see further, following subsection).

Type IV sand bodies are the thickest, and comprise several facies (A1, A2a and A2b; Fig. 5.8). They are composite bodies and reflect stacking of several sand body types (ridges and sheets), which migrated across each other. Composite sand bodies are described from other ancient shallow marine settings containing shelf sandstone bodies (e.g. Lower Carboniferous Kinsale Formation of southern Ireland, de Raaf et al., 1977; Cretaceous Western Interior Seaway, Tillman and Martinsen, 1984; Upper Devonian Old Head Sandstone Formation, also of southern Ireland, MacCarthy, 1987).
Sand body classification - tidal- or storm-built

Sandy bedforms occur on modern shelves of three main types: tidal, e.g. the southern North Sea (Belderson et al., 1982), storm-dominated, e.g. the northwest Atlantic shelf (Swift et al., 1973), or oceanic current-dominated, e.g. the southeast African shelf (Flemming, 1980). Ancient shallow marine sand bodies are often classified in terms of these three settings.

Oceanic currents were not important in the case of the Athalassa Formation, because the Mediterranean was, as it is now, a partially enclosed sea, protected from major oceanic influence. Tidal effects are also in general very small in the Mediterranean, usually being less than 0.3m (Rand McNally Atlas of the Oceans, 1979). They can be important in certain narrows and straits, however, e.g. the Messina Strait, whose bottom is also partially covered by sand waves (Colella and d’Alessandro, 1988). The Messina Strait is exceptional, however, because it lies between two seas (the Ionian and Tyrrhenian), whose own tides are very weak, but are in opposite phase. This results in a continuous sea surface gradient in the Messina Strait, and powerful currents flow as a result. In addition, tidal bedforms in straits typically face towards one end of the strait, because strong tidal currents tend to be through-flowing, e.g. Straits of Malacca, Indonesia (Keller and Richards, 1967).

Palaeocurrent directions from the Athalassa Formation sand bodies are almost without exception to the south (Fig. 5.7), i.e. directed across the E-W trending Mesaoria seaway. Although the scale of some Athalassa sand bodies is more akin to those of tidal origin (see type III sand bodies, previous subsection), features typical of tidal deposits, e.g. mud drapes, herring-bone cross-stratification or numerous reactivation surfaces (Johnson and Baldwin, 1986), are not observed.

General circulation in the Mediterranean may have generated currents which flowed east or west through the E-W trending Mesaoria seaway. There is some evidence for such currents from facies of the subaqueous part of the fan-delta of the Kakkaristra Formation (facies C4, section 4.3), which lay to the southwest of the Athalassa sand body field (Fig. 5.6). This facies documents small
subaqueous shoals, two of which lie between the main Athalassa and Kakkaristra depositional areas, and have southeast-oriented cross-stratification. The shoals are believed to have been deposited by currents flowing from NW to SE through this area (see further, section 4.4.2, subaqueous subsection). If such currents were responsible for the Athalassa Formation sand bodies, however, they too would have been expected to have broadly east- (or west-) dipping cross-stratification, instead of the strikingly southerly orientations found. Currents flowing through the Mesaoria seaway were probably strongest in the area between the Kakkaristra fan-delta and the Athalassa sand body field, where they were uninhibited by topography on the sea floor (i.e. sand bodies), and were capable of transporting small quantities of sediment.

Fairweather processes are not capable of transporting large amounts of sand across shelves (Swift and Niedoroda, 1985). A final possibility thus remains for the origin of the Athalassa sand bodies—they were storm built.

Currents flowing across ancient and modern shelves, in particular those associated with storms, have been much discussed (e.g. Swift, 1985; Swift and Niedoroda, 1985; Walker, 1984b, 1985b), and are still not fully understood. Two processes may have generated the south-flowing currents across the Mesaoria basin: 1) currents generated directly by winds blowing from the north; these winds would have been subject to only minor baffling by the narrow low-lying Kyrenia landmass (see sand supply subsection at end of this section), or 2) currents generated by a type of seaward-returning flow, resulting from coastal set-up as winds blew north across the Mesaoria seaway, towards the Kyrenia landmass.

The second type of current, which includes storm surge ebb (relaxation) currents, enhanced rip currents, and seaward-returning storm flows, (which may evolve into geostrophic currents influenced by the Coriolis force; Swift, 1985), are important on modern shelves. Most studied modern shelves are wide, however, and mainly N-S trending (therefore subject to the Coriolis force). The narrow E-W trending Mesaoria seaway was geometrically very different to these shelves. It is uncertain how effective such seaward-returning flows would have been in such a narrow basin.
Furthermore, most of the modern storm systems crossing Cyprus come from the NW, N and NE. As the major geographic features of the modern Mediterranean were in place by the end of the Pliocene (Steinger et al., 1984), and the modern Mediterranean climate had probably begun to evolve (Suc, 1984), it is suggested that storm currents in the Mesaoria basin were generated in direct response to dominant, south-blowing storm winds.

There is also a strong possibility that storm winds were stronger during Athalassa Formation times than previously, or at present. The formation is ?Upper Pliocene-Pleistocene in age, a time of glacioeustatic sea level fluctuations and accompanying climatic change. Arid periods in the Quaternary have been correlated with enhanced wind activity, which generated huge onshore dune fields (Lowe and Walker, 1984). According to Goudie (1984), intensification of climatic patterns should be expected in the Quaternary, particularly during glacial periods, when temperature gradients were increased. It is suggested here that stronger winds would have affected offshore as well as onshore areas, very possibly leading to enhanced shallow marine bedform development. In the Mesaoria basin, this may well have prompted generation of the Athalassa sand body field.

Sand body depositional processes

The initial formation of the bedforms represented by ancient, storm-built shelf sandstone bodies, their subsequent development and final abandonment, are processes still poorly understood (Walker, 1984b). Comparison with modern, nontidal sand bodies is hampered by the global transgression which affects all modern shelves. Sand ridges on the Atlantic margin of North America, for example, were originally shore-attached shoals, which have become progressively detached as sea level has risen (Swift and Field, 1981). They are rooted in a shelf-wide, transgressive sand sheet. Ancient shelf sandstones e.g. those of the Athalassa Formation or Western Interior Seaway form isolated bodies.

Swift and Rice (1984) have shown that sand patches can begin to accumulate on a muddy, storm-influenced shelf, if slight topographic irregularities are present on the sea floor. A system of positive
feedback then operates: as the bedform becomes larger and its relief increases, it extracts more sand from successive storm flows. A variety of storm and fair weather processes may modify bedforms, leading to a considerable variation in size, geometry, sedimentary structures and palaeocurrent orientations. Several different sand body types are present, for example, in the Western Interior Seaway (Walker, 1984b; Johnson and Baldwin, 1986).

In the Mesaoria basin, types II and III sand bodies may have been initiated as sand patches associated with slight topographic highs on the sea floor. Scouring of the shelf bottom was occurring because type I sand sheets have erosive bases (cf. thin sand sheets of MacCarthy, 1987). This scouring may have created a rather undulating sea floor surface. Once established, sand ridges and waves grew and migrated under the influence of storms. Large sand waves (type III) probably only moved during the most severe of these. Depressions were also filled with sand, which overflowed to form thin sand sheets (type I bodies). Smaller bedforms migrated across the tops of these sheets, which were also occasionally wave-rippled, perhaps during fairweather periods. Bedform migration was nearly always south, under the influence of the dominant storm regime.

Reasons for sand body abandonment and re-establishment elsewhere are also unclear. One important factor is likely to have been eustatic sea level fluctuations. There is no evidence that sand bodies actually emerged in the main part of the study area. No caliche development or vadose cements, for example, are recorded, or beach cross-stratification (cf. main bar sandstone of de Raaf et al., 1977). Fluctuating water depths must have influenced current strengths and patterns, however. Accompanying climatic changes are also likely to have been occurring, resulting in shifting wind patterns and intensity (see also previous subsection).

The four types of sand body recognised are developed at all levels within the Athalassa Formation. Laterally, sheets and large sand waves (types I and III) are present in the south and centre of the study area (Fig. 5.6), while type II ridges are apparently best developed closer to the northern margin of the basin. Reasons for this distribution pattern (if it is real, given the limits of exposure)
are uncertain. A variety of sand body structures and geometries are recorded from other ancient shallow marine settings e.g. the Cretaceous Western Interior Seaway of the U.S.A. (Swift and Rice, 1984) and the Upper Devonian of southern Ireland (MacCarthy, 1987). In these examples, reasons for sand body variation are also uncertain. In the Athalassa Formation, fluctuating storm patterns and intensity, changing water depths, distance from the shoreline and local sea floor conditions must all have played a role.

Initiation of sand body deposition

The mechanisms which triggered ancient, shallow marine sand body deposition have also been the subject of debate. Transgressive reworking of abandoned shoreline deltas has been suggested as a method of supplying offshore areas with sand, subsequently deposited as offshore bars e.g. Western Interior Seaway, U.S.A. (Philips and Swift, 1985; Johnson and Baldwin, 1986). Some sand bodies are associated with regressive events, however, and may be former shoreline deposits, drowned during later transgression (Bergman and Walker, 1987). This model is based on the identification of important erosion surfaces at the base of sand bodies, which are often difficult to recognise (Bergman and Walker, op. cit.). Tectonic effects are also proposed as initiators of sand body deposition, e.g. increased relief as a result of thrusting in the Rocky Mountains, adjacent to the Western Interior Seaway, is believed to have generated pulses of sand, which were transported to the coast and out across the shelf to be deposited as offshore bars (Swift and Rice, 1984).

In the case of the Athalassa Formation, eustatic sea level fluctuations had already begun to affect the Mesaoria basin (see section 3.5), and were presumably operating during deposition of the formation. It is possible that during eustatically-initiated transgressions, sediment may have been reworked southwards from the flooded margins of the Kyrenia landmass. In addition, a pulse of tectonism is recorded in the western Kyrenia lineament at the beginning of Athalassa Formation times (section 5.4.2), as the area came under compression. This compression may have triggered a supply of clastic sediment. It was relatively small, however,
because the Athalassa Formation sand bodies are highly bioclastic (see following subsection), despite the proximity (ca. 5km) of the Kyrenia landmass.

It can only be inferred that a combination of shallow marine setting, supply of sand-grade bioclastic, with some clastic, sediment and evolution of a favourable weather regime contributed to the onset of sand body deposition during Athalassa Formation times.

**Sand supply**

Predominantly south-directed palaeocurrents for the sand bodies of the Athalassa Formation imply that sediment was largely supplied from the north side of the basin (Kyrenia lineament). The presence of clastic grains, which could only have been derived from the Kyrenia lineament (e.g. dolostone; see section 8.4), supports the existence of an exposed Kyrenia landmass. The high proportion of skeletal carbonate grains in Athalassa sands, however, suggests that the landmass was small and not very elevated, and consequently supplied relatively little clastic material. Because of the low clastic influx, its margins became favourable sites for skeletal carbonate production. Carbonate and minor clastic sediment was subsequently transported into the basin, where it mixed with sediment derived from the volumetrically much larger Troodos Massif, and was reworked during storms to form sand bodies.

5.3.4 Nearshore to coastal environments, and transgressive events

Sediments of the third group of Athalassa Formation facies comprise facies C1a and D1 (beach, back beach and associated mud pools), C1b (?subtidal, hummocky cross-stratified facies), and D2 and D3 (lagoonal facies). These sediments interfinger on the northern side of the study area (Figs. 5.5 and 5.6) and indicate that the northern margin of the basin was partly fringed by a barrier island and lagoons (Fig. 5.9).

Barrier islands were perhaps a form of shallow marine sand body, emergent because of proximity to the shoreline. Lagoons developed behind these barrier islands. Lagoonal facies (D3) also interfinger with submerged sand body facies (A1 and A2), e.g. log 26/11/1, Fig. 5.5. This suggests that these sand bodies may have
been very shallow, again because of proximity to the shoreline, and that sheltered conditions were able to develop behind them (see Fig. 5.6).

The presence of thin pebble horizons in the calcarenites of log 26/11/1 (Fig. 5.5) is a further indication of a nearby coastline. The rounded pebbles, which are themselves of calcarenite, were probably locally derived from the north, from older shoreline facies of the Athalassa Formation, or from the Nicosia Formation, which may be calcarenitic in the Kyrenia area (Moore, 1960; Baroz, 1979). They were subsequently reworked a little way offshore. Conglomerates in shallow marine settings are typically found not far from the coast (e.g. Bourgeois and Liethold, 1984).

Facies of this group of sediments document shallowing-up (shelf facies B1 overlain by shoreline and lagoonal facies C1a, D1 and D2; log 8/2a, Fig. 5.5.), as is typical of the general trend of the sedimentary fill of the Mesaoria basin. An apparent deepening is then evident, however, because the facies in question are overlain by possible subtidal sands (facies C1b) and finally a shallow marine sand body facies (A2b). This transgressive event may be the result of some tectonic effect, but there is no other evidence for tectonism in this area at this time. Transgression may alternatively have been due to an intrabasinal (autocyclic) change, e.g. flooding of the barrier island as that part of the shoreline was abandoned as a depocentre. Transgression, however, may also have been the result of eustatic effects. Sea level changes were affecting the basin at this time, and are believed to have influenced shallow marine sand body deposition to the south (see previous section). In addition, there is evidence elsewhere in the Athalassa Formation for eustatic transgressive effects (see section 5.4.2). Eustatic effects would be most likely to be documented in this part of the Athalassa Formation, close to the shoreline, an area most susceptible to the consequences of fluctuating sea level.

5.4 North Side of the Basin

The Athalassa Formation is widely exposed on the north central and northwestern sides of the Mesaoria basin (Fig. 5.1). These areas
were not accessible to study during this project. Their geology is described in a number of Cyprus Geological Survey and other publications, however, and is now reviewed.

5.4.1 Description

In northwest Cyprus (western part of the Kyrenia Range), the Athalassa Formation is up to 150m thick, and comprises two members, upper and lower (Moore, 1960). The lower member is best developed south of the Ovgos fault (Figs. 5.1 and 5.10), is up to 120m thick, and comprises a basal conglomerate and a thick sequence of marls. The marls sometimes contain pebbly facies, consisting of interbedded fine conglomerate and sand, or conglomerate in a marly matrix, and occasionally show slump features (Lytras, 1962). Away from the Ovgos fault, pebbles are largely igneous, and probably Troodos-derived. The contact with the underlying Nicosia Formation is unconformable (Fig. 5.10). In some places, the sandy uppermost part of the Nicosia Formation is missing, so that the Athalassa Formation rests directly on the silty lower part of the Nicosia Formation (formerly the Myrtou Marl).

The upper member is thin (10-25m), and found only close to, or north of, the Ovgos fault. North of the Ovgos fault, the lower Athalassa member is not developed, and the upper member rests directly and unconformably on deformed Miocene Kythrea flysch (section 1.3.2; Fig. 5.10). Sediments of the upper member comprise a basal conglomerate, containing clasts now largely derived from the Kyrenia lineament, and pervasively cross-stratified, bioclastic-rich sands. No data on cross bed orientations are available. North of the Ovgos fault, thin soils, containing angular calcarenite clasts, occur within the sands, as well as occasional heavy mineral concentrations (Moore, 1960).

Along the Ovgos fault itself, a thin wedge of cleaved, Lower Miocene chalk (Lapithos Formation) has been thrust along the fault zone. It passes rapidly to the north into folded and deformed Kythrea flysch (A. Robertson, pers. comm.). Chalky sediment has been locally reworked into the Athalassa lower member to form a chalk conglomerate (Lytras, 1962).

The formation dips in general a few degrees south, or west to
northwest in the northwestern corner of the island (Cape Kormakiti, Fig. 5.1). Its true thickness distribution is difficult to ascertain because its upper surface is an unconformity, and it has suffered erosion. The formation infills depressions in the unconformity surface at its base, which are apparently due to folding of the Nicosia Formation (Lytras, 1962).

5.4.2 Interpretation

Several important differences are apparent between the Athalassa Formation of the eastern part of the Mesaoria basin, examined in this study, and that of the northwestern part, reviewed in the previous section. The major differences are (see Fig. 5.10):

1) the base of the formation is an unconformity in the northwest, but is conformable in the east;
2) the formation is up to 150m thick in the northwest, and only ca. 80m thick in the east;
3) two members are present in the northwest, and only one in the east;
4) an important fault (the Ovgos fault) is present in the northwest; it bounds the southwestern side of the Kyrenia lineament (Fig. 5.1), but its extension eastwards beyond Nicosia is uncertain.

The Ovgos fault is an important structural element in the northern part of the Mesaoria basin. It is proposed as one of the original master faults bounding the north side of the early Pliocene Mesaoria basin half-graben (section 3.4.1). Movement on these faults and subsidence of the half-graben came to an end during Nicosia Formation times (section 3.4.2). The faults were subsequently reactivated under compression, because they now crop out as high-angle reverse faults.

Reactivation of the Ovgos fault is believed to have begun at the start of Athalassa Formation times. It accounts for the E-W variations in thickness and facies trends documented by the formation, either because the fault terminates west of Nicosia, or because reactivation was very localised, affecting only its western portion. Reactivation did not apparently affect the Mesaoria fault to the southeast at this time (Fig. 5.10), as this fault appears to
have had no detectable influence on deposition of the Athalassa Formation in the east. Differential reactivation of faults was probably associated with uneven application of stresses, elements of both vertical and horizontal movement, and the influence of an important, underlying basement lineament (see further, section 3.4.1). The effects of initial reactivation are described as follows.

**Initial reactivation of the Kyrenia lineament**

Initial, and localised, reactivation of the Kyrenia lineament began in the west, along the Ovgos fault, at the beginning of Athalassa times. Compressional forces along the fault resulted in folding of the Kythrea flysch along its northern side (the Kythrea flysch may have already been partly deformed by earlier fault movements associated with half-graben development), and a thin slice of Lower Miocene Lapithos Formation was thrust along the fault plane. On the south side of the fault, the Nicosia Formation was tilted and partially folded, although on the whole deformation was not apparently great. The formation suffered some erosion, and an unconformity was generated.

As a result of thrusting, it is suggested that slight deepening took place ahead (south) of the Ovgos fault, and into this subsided area, marine marls of the lower member of the Athalassa Formation were deposited. Some conglomeratic material of Troodos origin was also swept into this area ahead of the advancing front of the Kakkaristra Formation fan-delta to the south (chapter 4). Complimentary movement in the Troodos Massif at this time, associated with the advent of compression of the Kyrenia lineament, is believed to have been responsible for generation of the fan-delta sediments of the Kakkaristra Formation.

**Sedimentation**

As fine-grained marine sediments (marls) began to fill the deepened area south of the Ovgos fault, conglomerate locally derived from the Kyrenia lineament was deposited close to the fault, in a number of small, lobes (Fig. 5.11a), generating pebbly intercalations within the marls. Locally, Lower Miocene chalk, thrust along the fault zone, was reworked as a chalk conglomerate. This setting is
Fig. 5.11 - Palaeogeographic reconstructions of the Mesaoria basin during A) lower Athalassa/Kakkaristra Formation times, and B) upper Athalassa/Apalos Formation times (see Encl. B for key to symbols). Note that fan, lake, dune and coastline positions in B) are shown schematically only (lack of exposure does not permit a detailed reconstruction for the Apalos Formation).
reminiscent of the lower Nicosia Formation, where small, slope fan-deltas were building off the faulted northern margin of the Troodos Massif in the early Pliocene (model 1, section 3.3.2). The pebbly facies in question were probably rather similar to facies B1a of the Nicosia Formation. Slumping in the marls (Lytras, 1962) suggests rapid deposition and/or syntectonic disturbance.

The minor foredeep eventually filled and shallowed. A period of relative quiescence followed, as the undisturbed shallow marine facies of the upper member of the Athalassa Formation were deposited. These facies transgressed northwards over the now inactive Ovgos fault (Fig. 5.11b) and were thus deposited both conformably over lower member marls and unconformably over deformed Kythrea flysch. During this relatively tranquil period, a eustatic rise in sea level is suggested as the cause of transgression. This minor transgression apparently did not extend across the southern margin of the basin, which had completely emerged by this time (Fig. 5.11b), probably because sedimentation rates were sufficient to inhibit flooding (see further, section 6.3.4).

Facies themselves of the upper member of the Athalassa Formation represent shallow marine to coastal enivronments. Detailed interpretations cannot be made in the absence of detailed description. Cross-stratification and heavy mineral lamination, however, suggest some sediments were perhaps beach deposits. The presence of soils suggests back beach environments. The high proportion of skeletal carbonate material indicates that the western end of the narrow Kyrenia lineament, which had earlier been supplying conglomeratic material, was once again partially submerged by the transgression. As in the eastern part of the Mesaoria basin, it now supplied only a limited amount of clastic detritus.

By the end of Athalassa times, tectonic activity had thus come to an end, and as in the northeastern part of the basin, shallow marine to coastal facies were being deposited in a temporarily stable setting (Fig. 5.11b).

5.5 Glacioeustatic Effects

In the preceding sections, reference has been made several
times to the possible effects of eustatic sea level changes on the Athalassa Formation. The formation is ?Upper Pliocene-Pleistocene in age (section 2.3), and was undoubtedly subject to these. It is indeed considered likely that eustatic effects were responsible for the onset of shallow marine sand body deposition which is so characteristic of this formation. Two reasons for this are: a) intensification of climatic patterns associated with glaciation may have resulted in enhanced storm activity, prompting development of storm-built sand bodies (section 5.3.3, subsection on sand body classification), and b) transgressive events may have enhanced sand supply to offshore areas, by enabling reworking of flooded coastal sediments from the fringes of the Kyrenia landmass (section 5.3.3, subsection on initiation of sand body deposition). In addition, eustatic effects are considered a likely cause for transgression of upper Athalassa Formation facies across the western end of the Kyrenia lineament (section 5.4.2; this transgression is not identified in the east in the area investigated in this project, because the upper Athalassa Formation could not be examined).

This large-scale effect of sea level variation (initiation of deposition of the Athalassa Formation) contrasts with the smaller effects documented within the formation itself. These include possible drowning of barrier island facies in the northern part of the study area (section 5.3.4), and the likely influence of small-scale fluctuations on the variety of geometries and sedimentary structures seen in the sand bodies of the formation (section 5.3.3, sand body depositional processes subsection).

The effects of small-scale, sea level fluctuations are being documented increasingly in the literature. This contrasts with older studies, which typically record more regional variations, related for example to the first, second or third order sea level cycles of Vail et al. (1977). Examples of these smaller effects come from both Holocene studies, e.g. Quaternary of the Texas coastline (Morton and Price, 1987), and from ancient examples, e.g. Silurian of the Appalachian basin, Pennsylvania (Cotter, 1988), where 1-3m thick coarsening-up cycles in mid shelf offshore bar deposits are equated with sea level changes. This development is coupled with evidence from oxygen isotope data that glacioeustatic fluctuations in the
Quaternary have been much more numerous than previously realised (Goudie, 1984).

Recognition of possible eustatic influence during deposition of the Athalassa Formation contrasts with other formations in the Mesaoria basin of similar age, i.e. the Kakkaristra and lower Apalos Formations (Table 5.1). The fan-deltaic facies of the Kakkaristra Formation were generated by a pulse of tectonism in the Troodos Massif (section 4.2.2). No eustatic effects are recognised in the formation. Small-scale eustatic effects are unlikely to be detectable in tectonically active areas, subject to major influxes of detrital sediment (Miall, 1984). The Apalos Formation is entirely continental. In the absence of excellent stratigraphic control and exposure, sea level effects are difficult to detect in fluvial sequences (see section 6.3.4).

Glacioeustatic effects are thus only documented in the Mesaoria basin by shallow marine facies (Athalassa Formation and upper Nicosia Formation). This reflects the susceptibility of shallow marine to coastal sequences to the consequences of fluctuating sea level.

5.6 Summary - Basin Evolution during Athalassa Formation Times

1. At the end of Nicosia Formation times, prior to deposition of the Athalassa Formation, a narrow, sandy, shallow marine platform occupied the Mesaoria basin (section 3.5).

2. In the eastern part of the basin, shallow marine conditions continued, and a number of carbonate sand bodies formed and migrated south across the platform. Their initial formation may have been a response to a changing climatic environment, as glacioeustatic effects began to have a pronounced effect on the basin. In particular, storm activity, which was the major sand body-building agent, may have been enhanced.

3. Several sand body types are recognised, and are developed throughout the Athalassa Formation. Their initiation, varying structures and configuration, and abandonment, reflect fluctuating current patterns and intensity, variable reworking of sediment from the shoreline, and local sea floor conditions. At least some of these factors are believed to relate to small-scale
glacioeustatic and climatic changes.

4. Sediment supply for sand bodies was mainly from the northern Kyrenia margin of the basin. The barely emergent Kyrenia landmass supplied little clastic sediment. Instead, its fringes became the site of skeletal carbonate production, which was subsequently reworked southwards by dominant storm winds, to form the characteristically carbonate-rich sand bodies of the formation.

5. The margins of the Kyrenia landmass in the east were at least partly fringed by barrier islands and lagoons. Preserved coastal to nearshore sequences reveal evidence of barrier island drowning, perhaps the result of a eustatic sea level rise.

6. Evidence for tectonism comes from the northwestern part of the basin (an area not accessible to this study). Here, the Athalassa Formation comprises a thick lower member, comprising up to 120m of marls and silts, with coarse-grained fan-deltaic intercalations. These sediments are believed to have accumulated in a deepened area just south of the Ovgos fault, which was being reactivated as a high angle thrust, as compression began to affect the Kyrenia lineament. This deepening did not occur further east, either because reactivation was very localised, or because the Ovgos fault terminates west of Nicosia. Complimentary movement in the Troodos Massif to the south at this time, resulted in progradation of the Kakkaristra Formation fan-delta into the basin (section 4.6).

7. In upper Athalassa Formation times, movement on the Ovgos fault declined as compression was temporarily halted. The minor foredeep filled, and bioclastic, shallow marine sediments were deposited in the northwest part of the basin (upper member of the Athalassa Formation) as well as in the east.

8. Upper Athalassa sediments in the northwest transgressed across deformed lithologies of the Kyrenia lineament. Transgression and partial resubmersion of the Kyrenia lineament are attributed to a further eustatic sea level rise, during this tectonically quiescent time.
Plate 5.1

a) Planar-bedded and -laminated calcarenite of facies A1a; hammer is 30cm long

b) Trough cross-laminated calcarenite of facies A2a, overlain by an interval of internally structureless, A1b calcarenite; hammer is 30cm long

c) Cross-laminated calcarenite of facies A2a; staff is 110cm long
Plate 5.2

a) Facies B1 sands, erosively overlain by thin calcarenite body, comprising trough cross-laminated facies A2a sediments; staff is 110cm long

b) Trough cross-bedded calcarenite facies A2b; figure at left for scale

c) Facies A2b cross-bedded calcarenite body showing two cross bed sets, separated by a truncation surface (which passes through centre of staff); staff is 110cm long
Plate 5.3

a) Photo montage of unusual calcarenite body, showing several sets of tabular to sigmoidal cross beds, overlying a vaguely bedded oyster bank (visible to left of hut); figure to right of hut for scale
Plate 5.4

a) Upper part: close-up of cross bed sets of Plate 5.3a. Lower part: partly bedded oyster and barnacle bank of facies C3b of the Kakkaristra Formation

b) Large-scale, cross-bedded calcarenite interval, facies A2c

c) Large-scale, trough cross-bedded calcarenite body of facies A2c, ca. 5m thick, overlying other facies A2c cross-bedded calcarenites (cross-bedding not visible)
Plate 5.5

a) Poorly preserved wave ripples, facies A2d

b) Wedge-shaped sets of cross-laminated facies C1a beach sediments; sands vary from very fine- to medium-grained and may be shelly or pebbly; hammer is 25cm long

c) Faintly parallel-stratified facies B1 sands, gradationally overlain by facies A1a planar-stratified calcarenite
Plate 5.6

a) Horizontal to wavy, finely laminated facies D1 mudstone, showing irregular, sand-filled, desiccation cracks; lens cap is 6cm across

b) Facies D1 mudstone, showing a spectacularly symmetrically rippled base, interbedded with cross-laminated, facies C1a beach sediments; hammer is 25cm long

c) General view of the Athalassa Formation, showing two, thin calcarenite bodies (facies A1 and A2), separated by structureless, poorly consolidated, largely grass-covered, sands (facies B1); hillside is ca. 20m high
CHAPTER 6 - APALOS FORMATION

6.1 Introduction

The Apalos Formation (section 2.2.2 and 2.4) is exposed in a number of small areas in the central to northern parts of the study area (Fig. 6.1, Encl. A). Exposure is generally restricted to the flanks of mesa-type hills in the Mesaoria Plain, which have escaped recent erosion. Partly because of erosion, the original thickness of the formation is unknown. Its present maximum thickness is ca. 60m (Ducloz, 1965).

The Apalos Formation is unfossiliferous (other than containing reworked microfossils), and its age is uncertain. On lithostratigraphic grounds, it is assigned a Pleistocene age (section 2.3). It conformably overlies the Kakkaristra Formation in the central and southern parts of the Mesaoria basin. In the east and north, where the Kakkaristra Formation is replaced by its lateral equivalent, the Athalassa Formation, relationships are less clear due to lack of exposure. The Athalassa Formation is thicker than the Kakkaristra, and in the study area, its upper part is considered to be equivalent to the lower part of the Apalos Formation (Fig. 6.2). The Apalos Formation is then conformable over the Athalassa Formation, according to Ducloz (1965). In the northwestern part of the basin (outside the study area), the Apalos Formation has not been recorded. In the Kyrenia Range itself, the Apalos Formation may be partly equivalent to the Karka Formation (Ducloz, 1972). This was not investigated during this study.

The Apalos Formation comprises an entirely continental sequence of sediments, dominated by fine-grained overbank facies. Facies recognised in the formation are described in the following section, before sedimentary models are discussed.

6.2 Facies and Facies Relations

Lithofacies and depositional environments are summarised in Table 6.1
Fig. 6.1 - Main areas of outcrop and facies distribution, Apalos Formation (see Table 6.1 for lithofacies).

Fig. 6.2 - Relationship between the Apalos, Athalassa and Kakkaristra Formations.
### Table 6.1 - Lithofacies, Apalos Formation

<table>
<thead>
<tr>
<th>Code</th>
<th>Lithofacies</th>
<th>Subenvironment</th>
<th>Environment</th>
</tr>
</thead>
<tbody>
<tr>
<td>A1</td>
<td>Colour-banded mud/siltstones</td>
<td>Floodplain fines</td>
<td>Interchannel areas of distal alluvial fans</td>
</tr>
<tr>
<td>A2</td>
<td>Interbedded sands and muds</td>
<td>? Coastal lake</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>B1</td>
<td>Thin sands</td>
<td>Minor sheet flows</td>
<td>(including small lakes)</td>
</tr>
<tr>
<td>B2</td>
<td>Calcarenitic sands</td>
<td>? Coastal lake</td>
<td></td>
</tr>
<tr>
<td>B2a</td>
<td>Trough cross-laminated</td>
<td>Dunes</td>
<td></td>
</tr>
<tr>
<td>B2b</td>
<td>High-angle cross-laminated</td>
<td></td>
<td></td>
</tr>
<tr>
<td>C1</td>
<td>Massive conglomerates</td>
<td>Debris flow/sheet flow</td>
<td>Alluvial fan braided channels</td>
</tr>
<tr>
<td>C2</td>
<td>Conglomerate-sands</td>
<td>Braided river channels</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>and bars</td>
<td></td>
</tr>
<tr>
<td>D1</td>
<td>Pale muds</td>
<td>Shallow ponds</td>
<td>Interchannel areas of distal alluvial fans</td>
</tr>
<tr>
<td>D2</td>
<td>Caliche</td>
<td>Palaeosols</td>
<td></td>
</tr>
</tbody>
</table>
Facies A1 - Colour-banded mud/siltstones

Description: this ubiquitous facies comprises variably coloured (pale to dark, red-, yellow- or occasionally grey-brown), fine-grained sediments, which range from silty claystone (least common) through silty or sandy mudstone (most common) to clayey siltstone and muddy very fine sandstone (moderately common); collectively, they are termed mud/siltstones; they are poorly consolidated, weathered and nearly always massive in appearance, with no visible structure other than prominent, horizontal colour-banding (Plates 6.1a and b); clay-rich layers are often the most reddened; pale layers are usually sandy or silty; very fine parallel lamination is occasionally evident; caliche intercalations (facies D2) are quite common; tiny blackened rootlet holes are common, although root traces are not; small, vertical, cylindrical burrows are sometimes present; mud/siltstones occasionally fill channel-shaped bodies, cut into other mud/siltstones (Plate 6.1c), in areas where other channelised facies (sands and conglomerates) are present.

Occurrence, lateral and vertical facies relations: this facies occurs everywhere within the Apalos Formation (Figs. 6.1 and 6.4–6.7); it encases all other facies except A2; it forms intervals from less than 1m to 20m thick.

Interpretation: the facies comprises typical floodplain sediments, which are fine-grained, rather structureless, brown and red in colour, contain rootlet holes and caliche, and lack fossils; deposition was largely from suspension during flooding; much original structure has been obscured by weathering, bioturbation and root disturbance; colour variations depend on original sediment composition, porosity and permeability of the sediment, length of time prior to burial, and pedogenic processes (Retallack, 1983); there are many examples of similar, structureless, reddened, caliche-bearing overbank deposits, e.g. Eocene of the Temp-Graus basin, Spain (Atkinson, 1986), Old Red Sandstone of Wales (Allen, 1970) and Triassic of Wales (Tucker, 1977).

Facies A2 - Interbedded sands and muds

Description: the facies comprises very thin- to medium-bedded
Fig. 6.3 - Palaeocurrent data, Apalos Formation.
Figs. 6.4 & 6.5 - Sedimentological logs, cross-sections A-A' & B-B' (see Fig. 6.1 for section locations and Encl. B for key to logs).
Fig. 6.6 - Sedimentological logs, cross-section C-C’ (see Fig. 6.1 for section location and Encl. B for key to logs).

Fig. 6.7 - Sedimentological logs, cross-section D-D’ (see Fig. 6.1 for section location and Encl. B for key to logs).
(2-30cm), very fine to fine, sorted, grey-brown, partially cemented sandstones, interbedded with 10-40cm thick intervals of poorly consolidated, grey-brown sandy mudstones (Plate 6.2a; Fig. 6.5); the sands have poorly developed sedimentary structures, including parallel and cross-lamination, and rarely preserved undulatory ripples; lamination is occasionally defined by heavy minerals; sands may be gritty, containing scattered coarse grains; bed tops and bases are sharp, and bases are sometimes concave-up; palaeocurrents are rather variable, but mainly between south and west (Fig. 6.3); the muds are massive or rarely, faintly parallel-laminated; two small bivalves (?Parvicardium) and oyster debris were recovered from one mudstone interval (log 7/12/2a, Fig. 6.5); Parvicardium is a marine bivalve; rootlet holes are present higher in the section.

Occurrence, lateral and vertical facies relations: this facies only occurs in a small area in the north central part of the study area (Fig. 6.1); it occurs in the lower part of the formation (Fig. 6.5), overlying oyster-bearing muds of the Kakkaristra Formation; the upper part of the facies is characteristically interbedded with facies B2a; the facies forms intervals up to 12m thick.

Interpretation: this facies shows a possible marine influence near its base (presence of marine bivalves), but a continental aspect towards its top (rootlet holes and lack of fauna); fine-grained muds indicate a periodically quiet environment; a very shallow embayment in the coastline is suggested; this received marine influence initially, but was then perhaps closed off from the sea to form a small, shallow, coastal lake; mud accumulated in this area, periodically interrupted by small influxes of sand, introduced fluvially; the sands were then subject to reworking by south- and west-flowing currents, giving rise to the observed palaeocurrent pattern; these currents were probably generated in the lake by the dominant southerly storm winds blowing across the Mesaoria basin (see section 5.3.3, sand body classification subsection); facies B2a sands were also reworked by similar currents; this facies is similar to the thin-bedded facies of the Old Red Sandstone of the Anglo-Welsh basin (Allen, 1965), which is also interpreted as a lake deposit.
**Facies B1 - Thin sands**

**Description:** this facies comprises fine- (or sometimes medium-) grained, moderately sorted, brown sands in beds 20-80cm thick; sedimentary structures are poorly preserved, but where present, comprise trough cross-lamination or rarely parallel lamination; beds sometimes have coarse bases, containing granules and small mudstone intraclasts, and fine up; bed contacts are sharp; bases are sometimes erosive and gently concave-up, leading to rather lenticular bedding (Plate 6.2b); lateral continuity is poor, and beds are only ?10's of metres normal to flow; rare rootlet holes are present; only three palaeocurrents were recorded, all with northerly directions.

**Occurrence, lateral and vertical facies relations:** this is an uncommon facies, occurring mainly in the central part of the study area (Figs. 6.1 and 6.5); the sands occur as single beds, surrounded by facies A1 mud/siltstones (Plate 6.2b).

**Interpretation:** fining-up and trough cross-lamination are characteristic of channel-fill sequences in sandy fluvial systems (e.g. Allen, 1964); the thinness of this facies, however, suggests that sedimentary structures may be due to decelerating, sheet flow events (e.g. Leeder, 1973); the restricted lateral extent of the facies suggests that it was probably associated with crevassing events, originating from the main distributaries in the system, and not the product of major, catastrophic flooding (Collinson, 1986).

**Facies B2 - Calcarenitic sands**

**Description:** this facies comprises fine-grained (or occasionally medium-grained), well sorted, pale brown to pink-brown sands, containing substantial amounts (up to 30%) of skeletal carbonate material (echinoid, bivalve, red algal and foraminiferal fragments); this sediment is similar to that found in the calcarenites of the Athalassa Formation (chapter 5); two subfacies exist: a) type a sediments form beds 20-40cm thick, which appear massive, or occasionally reveal shallow trough cross-lamination (troughs up to 50cm wide); the subfacies forms lenticular intervals with flat tops, up to 180cm thick and a few 100 metres wide; they sometimes show a slight coarsening-up; sparse palaeocurrent directions are directed SSW; b) type b sediments form individual beds up to 85cm thick; the
beds are characteristically nobbly in appearance (Plate 6.2c), due to weathering, and possibly pedogenic processes; this obscures sedimentary structures, though occasionally high-angle, tabular cross-lamination is visible; beds have sharp contacts, and are laterally persistent over only a few 10's of metres; they are distinctively lenticular, with both flat tops and concave-up bases, or flat bases and convex-up tops (Plate 6.2c); palaeocurrent data are limited, but show south and west directions (Fig. 6.3).

Occurrence, lateral and vertical facies relations: subfacies a occurs in the centre of the study area (Fig. 6.1), and is characteristically associated with facies A2 (Fig. 6.5); subfacies b occurs in the centre and east of the study area (Fig. 6.1), and is encased in facies A1 (e.g. log 29/10/4, Fig. 6.4).

Interpretation: the location of subfacies a coincides with that of facies A2, interpreted as the deposits of a small, shallow, coastal lake; the skeletal carbonate fraction of this subfacies is believed to have been washed into this lake from the nearby, shallow marine Mesaoria seaway, during high tides, or transported by wind; this material was subsequently reworked with clastic sediment by currents associated with south-blowing winds (see also facies A2); in subfacies b, clastic and bioclastic material is interpreted to have been reworked by aeolian processes to form small dunes, with high-angle cross-stratification; reworking of bioclastic sediment into aeolian facies is not uncommon (e.g. Chan and Kocurek, 1988); beds with convex-up tops may be exhumed dunes, those with concave-up bases reflect aeolian infilling of small depressions in the floodplain; palaeocurrent directions partly reflect dominant southerly storm winds, but westerly directions suggest lesser winds also transported sediment; two possible origins for bioclastic sediment are 1) shelly beaches along the south coast of the basin, if these were developed (the southern shoreline is not exposed), or 2) material reworked from the shallow marine sediments of the Athalassa Formation (chapter 5), which are stratigraphically partly equivalent to the Apalos Formation (section 6.1).

Facies C1 - Massive conglomerates

Description: this facies comprises poorly to sometimes moderately
sorted, pebble to cobble conglomerate, with angular to subrounded clasts, in thin sheets or lenses up to 2m thick (Plates 6.1b and 6.3a); clasts are mainly 1-8cm in diameter, though occasionally up to 25cm; matrix comprises brown, poorly sorted muddy sand; the conglomerates are ungraded and only rarely show a vague horizontal stratification; conglomerate fabric is either matrix-supported (uncommon; Plate 6.3b) or clast-supported (common; Plate 6.3c); clast-supported conglomerate may show poor to moderate, up-dip clast imbrication; conglomerate bodies have flat tops, bases may be slightly to distinctively scoured or concave-up (Plate 6.3a); conglomerates typically form single, isolated bodies, (e.g. log 28/10/3, Fig. 6.7); laterally continuity of conglomerates is of the order of several 100's of metres; conglomerates are unfossiliferous.

Occurrence, lateral and vertical facies relations: the facies occurs throughout the study area, and the formation, but is particularly common in its upper part; conglomerates form isolated bodies, surrounded by facies A1 mud/siltstones; they usually occur close to exposures of facies C2 conglomerates and sands.

Interpretation: much of this facies resembles facies Gm of Miall (1977), usually ascribed to deposition on longitudinal bars in braided river channels, during normal stream flow (e.g. Hayward, 1983; Ramos and Sopćena, 1983; Miall, 1984); their isolated nature, erosive bases and often sheet-like nature, however, is suggestive of sheet flow processes, marginal to the main fluvial channel system represented by facies C2; some sandy matrix-supported facies may represent rather high concentration flows, which are transitional to true, mud-supported mass flows, i.e. similar to the fluidal sediment flows of Nemec and Steel (1984); the facies is thus interpreted as representing the more distal portions of unconfined, gravelly sheet flows of both low and high concentration; the facies is never found far from braided channel facies C2, perhaps suggestive of overflow of sediment-laden currents from channels, during torrential flooding.

Facies C2 - Conglomerate-sands

Description: this facies contains a heterogeneous group of sediments, comprising (Fig. 6.8a and b): a) massive to vaguely stratified, predominantly clast-supported, pebble to large cobble
Fig. 6.8 - Field sketches, facies C2 conglomerate-sands (see Encl. B for key).
conglomerate (Plate 6.4a), in beds 30cm-1m thick, which are mostly
sheet-like; imbrication is sometimes well developed; sorting is usually
poor, b) trough cross-bedded conglomerate in laterally impersistent
wedges, up to 1m thick; conglomerates are pebble to small cobble
grade, moderately sorted, clast-supported, and sometimes almost
matrix-free (Plate 6.4c); rarely, if surrounded by sandy facies,
gravelly megaripples are preserved (Fig. 6.8b); c) trough
cross-stratified, fine, sorted, pale to dark brown sands, in laterally
impersistent wedges, sometimes grading up from conglomerate
wedges; sands may contain pebbly intercalations; tabular
cross-stratified sands are less commonly present; d) horizontally, or
low-angle cross-stratified sands, generally in thin (up to 20cm),
tabular to lenticular units (Plate 6.4b) which are laterally
impersistent; e) massive fine sands in sheet-like or less regularly
shaped bodies up to 1m thick; thin pebbly stringers, or thin clayey
layers are sometimes present, as are rootlet holes; f) thin layers and
lenses of moderately to well sorted granule to pebble conglomerate,
which are up to 20cm thick, sometimes have erosive bases (Plate
6.4b), and are of limited lateral extent; these subfacies interdigitate
intimately (Figs. 6.8a and b), sometimes on quite a small scale (Plate
6.4b).

Occurrence, lateral and vertical facies variations: this facies
occurs primarily at two locations - Kantara hill (WD 313813) and
Vouniotika (WD 258850), in the eastern and central parts of the
study area (Fig. 6.1), though is occasionally present elsewhere; it
forms intervals 1-6m thick, interbedded with facies Al mud/siltstones
(e.g. log 29/10/3, Fig. 6.4); neither fining- nor coarsening-up
characterises these intervals.

Palaeocurrents: palaeocurrent directions for the two main
locations are plotted in Fig. 6.3; both show a strong northerly
component; conglomerate imbrication from all areas is combined in the
lower rose diagram (Fig. 6.3), confirming the predominant northward
direction.

Interpretation: sediments of this facies can be likened to facies
Gm (type a), Gt (type b), St and Sp (type c) and Sh (type d) of
Miall (1977), all typical of braided river or distal alluvial fan
deposits (Miall, 1977; Rust, 1978); type a sediments are largely the
product of bedload deposition on a flat bed e.g. longitudinal bar or channel floor (Collinson, 1986); lack of grading and erosive bases suggest deposition relatively early, during early or peak flood stage; type b sediments are the product of migrating, crescentic braided bars; these may have formed in deeper channel areas (Walker and Cant, 1984); matrix-free textures may reflect post-depositional winnowing (cf. Kleinspehn et al., 1984); subfacies c and d represent deposition of sand during moderate to upper flow regime conditions; fining-up of trough cross-bedded conglomerates into similarly stratified sands is not uncommon in braided settings (e.g. Rust, 1978) and probably reflects waning flow; type e sediments are probably structureless in part due to weathering, but may also represent the nuclei of sandy bedforms, which accreted vertically, rather than by lateral migration (Walker and Cant, op. cit.); such bars can become vegetated if they emerge; moderate to good sorting in type f sediments may reflect minor winnowing and reworking of gravelly bedform tops by late stage watery flows (cf. Gjelberg and Steel, 1983); in total, this facies represents a mix of sandy and gravelly bedforms and flows, deposited in a series of braided river channels; the relatively small thickness of intervals of the facies suggests braided river tracts were rather short-lived, and never very deep.

**Facies D1 - Pale muds**

*Description:* this uncommon facies comprises thinly bedded to laminated, pale brown to cream-coloured, calcareous silty claystone; it is found in intervals up to 2m thick, and may contain very thin beds of sandy silt, or partings of rather pure clay; weathering tends to obscure sedimentary structures, but bedding and lamination are virtually always horizontal; rarely, very fine cross-lamination is evident, as are thin intervals containing brown mudflakes; tiny, blackened rootlet holes are sometimes common; mottling in places is suggestive of bioturbation; the facies is unfossiliferous.

*Occurrence, lateral and vertical facies relations:* the facies occurs in the eastern part of the study area (Figs. 6.1 and 6.4), and forms lenticular units traceable over ca. 10km; it is under- and overlain by facies A1 or C1 (Plate 6.1b).
Interpretation: the fine-grained nature of this facies suggests a quiet depositional environment, e.g. shallow pool or lake, in which only finest sediment accumulated; its pale colour and calcareous content indicates derivation largely from the chalky and marly pre-Pliocene sediments overlying the Troodos Massif (see section 8.2, clays subsection), similar to the pale mudstones of the Kakkaristra and Athalassa Formations (facies D1 of both formations); aeolian processes probably transported a large amount of this very fine material; horizontal lamination is due to settling from suspension; cross-lamination reflects minor rippling and reworking; coarser sediment was introduced occasionally during flooding; ponds were partly vegetated, or occasionally desiccated (mud flakes).

**Facies D2 - Caliche**

Description: sediment of this facies comprises diffuse to well formed blebs, or subvertical veins, of white, powdery, calcareous material, found in discontinuous layers and lenses within facies A1 mud/siltstones (Plate 6.5a); variable proportions of host sediment are disseminated through the powdery sediment; blebs are never more than a few centimetres across, veins are up to 10cm long; occasionally, well formed, discrete, vertically oriented calcareous nodules are developed, and rarely, these are fused to form coalesced horizons (Plate 6.5b); the facies is developed throughout the Apalos Formation, wherever facies A1 occurs.

Interpretation: this facies is a form of caliche; it most closely resembles the chalky caliche of Esteban and Klappa (1983), with minor development of nodular caliche; chalky caliche is rather immature, reflecting insufficient time for the development of nodular and hardpan types; the thinness of caliche zones, and lack of development of profiles exhibiting several caliche types (Esteban and Klappa, op. cit.) also reflects immaturity; the presence of caliche indicates a semi-arid climate with seasonal rainfall (Allen, 1986); it is typical of distal alluvial fan settings, which are well drained, and have large, inactive, overbank areas; examples of caliche horizons within floodplain sequences are described by many workers, e.g. Old Red Sandstone of northwest Scotland (Steel, 1974), Eo-Oligocene of South Dakota (Retallack, 1983).
6.3 Sedimentological Model

6.3.1 General Model

Alluvial fan or braidplain

The general characteristics of Apalos Formation facies, i.e. their brown or red colour, the presence of rootlets and caliche, relative immaturity of coarse-grained intervals, and lack of fossils, are typical of a continental environment. Red colouration and caliche development suggest a semi-arid climatic setting. The high percentage of floodplain fines (up to 75% in exposed sections; see Figs. 6.4-6.7) indicates one of two terrestrial settings: distal alluvial fan (or a more proximal location on an alluvial fan supplied with fine-grained sediment) or an alluvial plain crossed by a few braided river tracts, fed by small alluvial fans (Figs. 6.9a and b). The latter may be considered a type of semi-arid, fine-grained braidplain.

Recently, the importance of distinguishing between alluvial fans and braidplains has been emphasised by McPherson et al. (1987), because of the different palaeogeographic and tectonic implications of the two environments. According to them, the two settings are easily separated: alluvial fans are small, wedge-shaped bodies, usually associated with a fault-bound margin, and containing coarse, poorly sorted sediment; braidplains are extensive, sheet-like, contain better sorted and rounded facies, and do not necessarily have tectonic implications. These authors, however, in the main, compare the proximal parts of typical semi-arid alluvial fans, characterised by massive conglomerates, with typical, humid braidplains, characterised by cross-bedded sands and conglomerates. They do not consider distal facies, or discuss climatic factors and sediment supply variations.

Preferred model

Large-scale facies geometry and lateral facies transitions would distinguish between alluvial fan and braidplain models. In the Apalos Formation, large-scale, three-dimensional geometry is unknown because of lack of exposure. Down-slope variations in maximum grain size, which would provide an indication of former
Fig. 6.9 - Schematic plan views of depositional models for the Apalos Formation.
geomorphological gradients, cannot be measured, and down-slope facies distribution is uncertain.

In spite of the difficulties imposed by a lack of laterally continuous exposure, an alluvial fan interpretation is preferred for the Apalos Formation. Firstly, an alluvial plain flanking a number of alluvial fans would perhaps be more likely to be characterised by sandy, channel-fill sequences and sheet flows, with conglomerates largely restricted to higher gradient fan surfaces (cf. Permian of the Karoo basin, South Africa; Stear, 1983). Conglomerates may occur in the alluvial plain, but would be likely to be restricted to channels (Freidman and Saunders, 1978). In the Apalos Formation, braided channel sequences contain appreciable amounts of conglomerate (facies C2), and conglomerate sheets, partly of mass flow origin (facies C1), occur in interchannel areas.

Secondly, the main braided river tracts of the Apalos Formation (represented by facies C2) coincide with the major depositional lobes of the older Kakkaristra Formation fan-delta (Fig. 6.10), which were fed by alluvial fans (section 4.4.2; Fig. 4.14). This implies that these earlier fans did not shift location, but continued to supply the more central parts of the basin. These fans would have had to shrink to some extent in order for an alluvial plain to have evolved adjacent to them. Although the Apalos Formation is finer-grained than the Kakkaristra Formation, sedimentation rates are not considered to have declined, because the transition from deltaic Kakkaristra facies to continental Apalos sediments records continued shallowing and filling of the Mesaoria basin (see also section 6.3.3). Shrinking of alluvial fans does not therefore seem likely.

The Apalos Formation is thus interpreted as a distal alluvial fan sequence (Fig. 6.11). The one small exposure of the formation close to the Troodos Massif (see Fig. 6.1), is not dominated by very coarse facies (see log 25/1, Fig. 6.7). It is interpreted as occuring in an interfan area.

Several other ancient examples of similar mid-distal alluvial fan deposits include the Old Red Sandstone of the Anglo-Welsh basin (Allen, 1970), the Marginal Triassic of Wales (Tucker, 1977), the Lower Cretaceous Lombriz Formation of Chile (Flint et al., 1986), and the Oligocene Capella Formation of the Temp-Graus basin in the
Pyrenees (Atkinson, 1986). The latter example is most similar to the Apalos Formation in containing a high percentage of floodplain fines, with isolated, intercalated conglomerate-sand units, thin sandstone sheets, caliche intervals and occasional lacustrine facies.

6.3.2 Depositional processes

Alluvial fans, established during Kakkaristra Formation times, continued to build off the Troodos Massif during deposition of the Apalos Formation. Their coarsest, proximal parts (not exposed) were presumably dominated by coarse, conglomerate facies.

Down-slope, these fans passed into fine-grained distal margins, crossed by a few impersistent braided channel tracts. The sediments of this environment are preserved in the currently exposed Apalos Formation. Floodplain facies comprise colour-banded muds and silts (facies A1), mainly deposited from suspension. These facies were sometimes reddened, and contain immature caliche (facies D2). In some infrequently inundated areas, the finest sediment accumulated in shallow pools (facies D1). Overbank areas were vegetated to some extent, as witnessed by numerous tiny rootlet holes. Occasional dune facies developed (facies B2b).

Distal fan distributaries, or braided river tracts, comprised a variety of gravelly longitudinal and crescentic bars, sandy bedforms and channels. Together these resulted in deposition of interfingered horizontally and cross-stratified conglomerates and sands (facies C2). Periodically, overbank flooding led to crevassing and deposition on the floodplain of thin sandstones (facies B1). During torrential flooding, unconfined mass flows and sheet flows were swept far along distributaries and across channel margins, to be deposited as thin sheets in interchannel areas (facies C1).

During early Apalos Formation times, the coastline did not lie far to the north. A shallow coastal lake, probably originally connected to the sea to form an embayment, is recorded by facies A2. Fine-grained muds accumulated in this lake, which periodically received sandy material both from its landward southern side (clastic sediment) and its seaward northern side (bioclastic sediment). This material was reworked largely by currents generated by the dominant south-blowing winds of the Mesaoria basin.
Lake facies lie just northeast of a prominent embayment in the coastline of the older, Kakkaristra Formation fan-delta (Fig. 6.10). This may reflect a gradual migration and filling of this embayment, as the southern margin of the Mesaoria basin was transformed from a coastal and shallow marine environment (Kakkaristra Formation) into a fluvial one (Apalos Formation). Other lakes were perhaps also present, correlating with the presence of other embayments in the Kakkaristra fan-delta shoreline. These lakes filled as the coastline prograded north, as the Mesaoria basin continued to shallow.

6.3.3 Tectonic implications

Two factors are important. Firstly, alluvial fans, established during deposition of the Kakkaristra Formation, continued to exist during Apalos Formation times, and continued to be the main conduits of sediment supply to the Mesaoria basin (section 6.3.1). There was no apparent shift in fan location or generation of new fans. Secondly, a fining-up is recorded between the sands and conglomerates of the Kakkaristra Formation, and the sequence dominated by mud/siltstones in the Apalos Formation.

The type of sedimentation which occurs in tectonically active alluvial basins depends on the relative rates of uplift, erosion, sedimentation and subsidence (Miall, 1978). If sedimentation exceeds subsidence, then fining-up sequences will result when uplift rates are less than rates of erosion (Miall, op. cit.). Sedimentation rates were clearly in excess of basin subsidence in the Mesaoria basin, because the transition from the deltaic Kakkaristra Formation to the continental Apalos Formation records the continued filling and shallowing of the basin. Fining-up between the two formations is thus believed to signal a decline in uplift of the Troodos Massif (which had initially resulted in progradation of the Kakkaristra fan-delta; section 4.6), and the onset of peneplanation of the ophiolite. Other large-scale, fining-up, alluvial sequences document similar decreases in uplift rates, and hinterland peneplanation, e.g. the Old Red Sandstone of the Midland Valley, Scotland (Bluck, 1967; Morton, 1979). The fact that no new sediment transport paths developed during Apalos Formation times is consistent with the interpretation of a period of relative stability.
Fig. 6.10 - Relationships between the main localities of facies A2 lake sediments (circles) and facies C2 braided channel deposits (crosses), and the Kakkaristra fan-delta. Main areas of exposure of the Apalos Formation are stippled.

Fig. 6.11 - Schematic block diagram reconstruction for the Apalos Formation (nature and configuration of the shoreline are uncertain).
Tectonic quiescence is also evident on the northwest side of the Mesaoria basin at this time. There, initial compression of the Kyrenia lineament had resulted in thrusting and folding at the beginning of Athalassa Formation times (section 5.4.2). Tectonism then declined, however, and the undisturbed shallow marine to coastal facies of the uppermost Athalassa Formation were deposited partly over the earlier deformed lineament (see Fig. 5.11b). These sediments are approximately laterally equivalent to the middle Apalos Formation (Fig. 6.2).

### 6.3.4 Glacioeustatic effects

The effects of glacioeustatic sea level fluctuations must also be considered when evaluating the Apalos Formation, as these were undoubtedly occurring during deposition of this Quaternary formation.

One important observation, which might have signaled a sea level fall, would have been the presence of incised river valleys, recording lowered base level. There are several problems with this interpretation, however:

a) incision is an intrinsic fluvial process; a flood event will incise naturally, once the local geomorphic threshold for channel entrenchment is overcome (Maizels, 1987a);

b) detailed lithostratigraphic or biostratigraphic control, which is often lacking in alluvial sequences, is required so that the relative ages of channel fills and surrounding floodplain sediments can be assessed (see Leeder and Gawthorpe, 1987, p. 145);

c) the response of a fluvial system to both internal control, e.g. geomorphology, nature and rates of sediment supply, and external control, e.g. climatic, tectonic or eustatic factors, is complex, and may involve both deposition and incision (Schumm, 1981); a single change in a controlling factor can produce a series of erosional and depositional events; recent examples of this are the generation of several river terraces during Holocene deglaciation (Maizels, 1987a), and arroyo cutting and filling in Colorado (Womack and Schumm, 1977);

d) external effects other than eustatic may cause incision, e.g. 
Sea level changes have occasionally been documented in ancient continental sediments, e.g. the Lower Old Red Sandstone of southwest Wales (Allen and Williams, 1982). Several factors facilitated their recognition in this example, however: excellent lateral exposure, the presence of extensive, correlatable, ash-fall tuffs, and the occurrence of extensive, well-developed caliche horizons. Such features are lacking in the Apalos Formation.

Initial deposition of the Apalos Formation may have been influenced by sea level change. Prior to its deposition, a pulse of uplift of the Troodos Massif generated the fan-delta complex of the Kakkaristra Formation (section 4.6). This formation passes up very rapidly into the Apalos Formation. The swiftness of this transition may suggest that regression was enhanced by falling sea level.

On the north side of the basin, a transgression attributed to sea level rise is recorded in the uppermost Athalassa Formation (section 5.4.2), which is approximately stratigraphically equivalent to the middle of the Apalos Formation (Fig. 6.2). Transgression flooded the western end of the Kyrenia lineament. It did not extend far south, however, because no evidence for transgression is found in the Apalos Formation (see Fig. 5.11b). This probably reflects a continued, plentiful sediment supply from the Troodos Massif, sufficient to enable the rate of sedimentation to keep pace with the rate of transgression.

In summary, sea level fluctuations were occurring during deposition of the Apalos Formation. No direct evidence for them has been found in this entirely fluvial formation, however.

6.4 Summary - Basin Evolution during Apalos Formation Times

1. Immediately prior to deposition of the Apalos Formation, the fan-delta system of the Kakkaristra Formation had prograded northwards into the Mesaoria basin, in response to a pulse of uplift in the Troodos Massif (section 4.6).
2. As plentiful supplies of sediment continued, the subaqueous and coastal fringes of the fan-delta began to emerge, and were transformed into the distal portions of alluvial fans.
3. These distal alluvial fans comprised large floodplain areas, crossed by a few, short-lived, braided channel tracts. This reflects the supply of large quantities of fine-grained sediment, as peneplanation of the Troodos Massif began, and uplift declined in the hinterland.

4. The north side of the basin remained marine, as facies of the upper Athalassa Formation (chapter 5), which is laterally equivalent to the lower to middle Apalos Formation, continued to be deposited. Although shoreline facies are not preserved within the Apalos Formation in the study area, the nearby presence of the Mesaoria seaway is documented by the occurrence of reworked bioclastic material in Apalos facies. These facies were partly deposited in small, coastal lakes, and also by aeolian processes.

5. Eustatic sea level fluctuations were occurring during deposition of the Apalos Formation. Their effects are difficult to identify, however, in the fluvial facies of the formation.

6. Facies of the formation eventually prograded far across the basin, at least in the east, where the upper Apalos Formation overlies the Athalassa Formation. Whether the entire basin became continental at this stage is uncertain, because sediments younger than the Athalassa Formation have largely been eroded from the north side of the basin.
Plate 6.1

a) Horizontally colour-banded, continental facies A1 mud/siltstones, showing characteristic weathering

b) Colour-banded, continental facies A1 mud/siltstones, and a thin horizon of pale, facies D1 muds near top; dark, lenticular layer of facies C1 conglomerate overlies muds

c) Channelised body filled with largely structureless, facies A1 mud/siltstones, cut into other mud/siltstones; staff is 110cm long
Plate 6.2

a) Thin, cemented, horizontal sandstone layers, interbedded with thicker, poorly consolidated, sandy mudstones, facies A2; hammer is 30cm long

b) Channelised, lenticular unit of facies B1 sands, interbedded with grass- and rubble-covered facies A1 mud/siltstones; hammer, which is 30cm long, rests on base of unit

c) Nobbly, calcarenitic sand unit of facies B2b, with convex-up top, which may represent a small, exhumed dune; high-angle cross-lamination, characteristic of the facies, is not visible in this example; hammer is 25cm long
Plate 6.3

a) Channelised and lenticular intervals of facies C1 conglomerates, interbedded with facies A1 mud/siltstones; vertically, field of view is ca. 10m

b) Matrix-rich, channelised interval of facies C1 conglomerate, ca. 70cm thick, interbedded with facies A1 mud/siltstones

c) Facies C1 conglomerate bed, showing clast-supported fabric, poor to moderate sorting and very weakly developed imbrication (clasts dip to left); staff is 110cm long
a) Interbedded conglomerate-sand facies C2, showing two, massive, poorly sorted, type a conglomerate beds, with a thin, intercalated, sand interval; staff is 110cm long

b) Interbedded conglomerate-sand facies C2, showing a lower, poorly sorted, sandy conglomerate bed (type a), and middle and upper, well sorted, fine conglomerates in irregular lenses (type f), all interbedded with massive to faintly parallel laminated sands (type d); portion of staff showing is 95cm long

c) Interbedded conglomerate-sand facies C2, showing a thick set of cross-bedded conglomerate (type b), overlying facies A1 mud/siltstones; one cross bed (upper right) is almost matrix-free; portion of staff showing is 95cm long
Plate 6.5

a) Poorly developed facies D2 caliche veins and blebs, within facies A1 mud/siltstones, overlain by a poorly sorted, channelised, facies C1 conglomerate; staff is 85cm long

b) Vertical facies D2 caliche nodules, fused together in upper part of photo to form a well developed, coalesced, caliche horizon; staff is 110cm long
CHAPTER 7 - FANGLOMERATE

7.1 Introduction

The Fanglomerate is the youngest stratigraphic interval investigated during the present project. It crops out throughout the study area, but is very thin, in general only 2-5m, though occasionally reaching 12m. It typically forms a thin capping to the numerous, flat-topped mesas present in the Mesaoria Plain (Plate 7.1a).

The Fanglomerate unconformably overlies all other Plio-Pleistocene formations in the basin. A remarkable feature is the flatness of its basal contact with the majority of these underlying formations, although close to the Troodos Massif, some broad channeling is evident (see further section 7.2, facies B1 and section 7.3). Indeed, an unconformity is not readily identified in the centre of the basin, where the Fanglomerate overlies the stratigraphic unit occurring immediately below it, i.e. the fluvial and conglomerate-bearing Apalos Formation. The Fanglomerate can be traced southwards, however, where it overlies progressively older formations, and even onlaps on to Troodos basement in places along the southern edge of the basin.

Sedimentologically, the Fanglomerate comprises a very coarse, widespread, alluvial fan deposit, and consists of one principal lithology - coarse, angular, poorly sorted conglomerate.

The Fanglomerate has not been studied in as much detail in this project as the other formations of the Mesaoria basin, because it, and the youngest Quaternary sediments of the basin are now the subject of a separate study (A. Poole, Edinburgh University). These young sediments are morphologically of interest because they form a series of river terrace deposits. When combined with studies of the raised beaches of Cyprus, they may reveal in detail the latest uplift history of the Troodos Massif, and its relation to recent glacioeustatic sea level fluctuations.
Only three lithofacies are recognised in the Fanglomerate, and are described below.

**Facies A1 - Massive conglomerates**

*Description:* this facies comprises coarse, poorly sorted, pebble-cobble conglomerate, with occasional boulders, up to 55cm in diameter; clast size tends to be between 1 and 10cm, however, and clasts are angular to subrounded; conglomerates are mainly massive, or occasionally show crude, horizontal layering, defined by variations in clast size; grading is uncommon although coarsening-up is sometimes observed; clast fabric is matrix-supported (Plate 7.1c) to clast-supported (Plate 7.1b), the latter being more common; matrix comprises poorly sorted silty sand, which is often degraded and calcified to form a caliche-like white powder; up-dip clast imbrication is weakly developed, if at all; the conglomerates generally form single beds, 1-3m thick, with planar bases, most commonly (Plate 7.2a), although erosive, mildly channelised bases also occur; lenses of intercalated, brown, silty sand are rarely present; caliche crusts are sometimes developed, typically on the underside of clasts; the conglomerates are unfossiliferous; clasts are largely of ophiolitic origin (see section 8.2), though some chalk and chert clasts are also present, and are occasionally predominant (e.g. Plate 7.1c).

**Occurrence, lateral and vertical facies relations:** this facies is the most common in the Fanglomerate, occurring across the study area; it is typically the only facies developed in a Fanglomerate exposure; in places, it grades laterally into facies A2, and overlies facies B1.

**Palaeocurrent directions:** palaeocurrent indicators, measured from clast imbrication (Fig. 7.2), are rather scattered, though all lie between WNW and ENE.

**Interpretation:** this facies resembles to some extent facies C1 of the Apalos Formation (section 6.2), although these sediments are coarser, more poorly organised, thicker and of greater lateral extent; facies C1 was interpreted as representing the more distal portions of gravelly sheet flows, of both low and high concentration; this facies
Fig. 7.1 - Facies distribution, Fanglomerate; asterisks indicate localities where the Fanglomerate occurs in valleys incised into pre-Pleistocene sediments or the Troodos Massif itself.

Fig. 7.2 - Palaeocurrent data and maximum clast size distribution for the Fanglomerate (clast sizes in cm). Arrows represent individual clast imbrication directions.
represents a more proximal, conglomerate-dominated setting; it is also believed to have been deposited from a variety of high to low concentration sheet flows, and resembles two types of recently documented, very coarse alluvial deposits: 1) proximal, waterlain, sheet flow conglomerates, which are characteristically clast-supported, poorly- to well-imbricated, crudely layered, sometimes erosively based, contain occasional interbedded sand lenses, but lack cross-bedding or fining-up sequences (e.g. the Tertiary Simmier Formation, California, of Ballance, 1984, or the Cretaceous Upper Coloso Formation, Chile, of Flint et al., 1986), or 2) subaerial, cohesionless, mass flow deposits of modified grain flow type (Lowe, 1976), which are clast- to sandy matrix-supported, non-erosively based, unimbricated and unstratified or crudely layered (e.g. Lombriz Formation, Chile, of Flint et al., 1986, or Devonian sediments of the Hornelen Basin, Norway, of Nemec and Steel, 1984); some conglomerates are likely to represent deposits intermediate between those of mass flow and waterlain origin (cf. fluidal sediment flows of Nemec and Steel, 1984); a spectrum of depositional processes was therefore probably responsible for deposition of the Fanglomerate; further work would be required to map out in detail the areal extent and interdigitation between these sediment types; all are witness, however, to a very coarse-grained, subaerial depositional setting, and suggest that the Fanglomerate was largely deposited in a proximal, alluvial fan environment.

Facies A2 - Cross-bedded conglomerates

Description: this facies comprises coarse, poorly sorted pebble-cobble conglomerate (maximum clast size of 20cm) and finer, better sorted conglomerate (maximum clast size of 5cm), in cross-bedded sets up to 2m thick; cross-bedding is only crudely defined in the coarsest sediments (Plate 7.2b), but it is better defined in finer conglomerates; cross bed sets are mainly of tabular type, and have non-erosive bases; cross bed directions (Fig. 7.2) are in line with clast imbrication orientations for facies A1; conglomerates are clast-supported and contain a poorly to moderately sorted brown sandy matrix; they are unfossiliferous.

Occurrence, lateral and vertical facies relations: this is an
uncommon facies (Fig. 7.1); it is typically found in association with facies A1, with which it is laterally gradational.

**Interpretation:** these cross-bedded conglomerates, similar to facies Gp of Miall (1977), are interpreted as the product of migration of the slipfaces of crude, longitudinal braid bars; these bars were occasionally developed on an alluvial fan surface which was otherwise dominated by sheet flows of various kinds (facies A1); bars were perhaps developed during periods of more stable flow during prolonged flooding.

**Facies B1 - Interbedded sands and minor conglomerates**

**Description:** the facies comprises very friable, poorly to moderately sorted, brown, silty, fine-grained sands, interbedded with thin (5-20cm) conglomerate layers and lenses; the sands, which form intervals up to 70cm thick, are largely structureless, or show faint horizontal lamination; thin, subhorizontal, calcareous veins, which are a type of recent caliche, sometimes permeate the sands; conglomerates form discontinuous layers of clast-supported sediment, which are finer-grained than the conglomerates of facies A1 and A2 (maximum clast size recorded is 8cm), and better sorted (Plate 7.3b); clasts also show better rounding, and are occasionally imbricated.

**Occurrence, lateral and vertical facies relations:** this facies is not common, and is mainly recorded in the east of the study area, close to or at a short distance from, the Troodos Massif (Fig. 7.1); it typically occurs where the Fanglomerate is thickest, and fills broad channel-shaped depressions, a few 100 metres in width (Plate 7.3a); the facies is typically overlain by facies A1, and together with facies A1, may be up to 12m thick.

**Interpretation:** the sediments of this facies are the finest-grained found in the Fanglomerate, although they occur in relatively proximal locations, close to the Troodos Massif; they also typically occur in the lower part of the Fanglomerate, where the Fanglomerate is thickest and fills small valleys cut into underlying sediments or Troodos basement itself (e.g. just north of Kambia; see Fig. 7.1); it is suggested that these valleys were cut during initial Fanglomerate deposition; these valleys filled, not with coarse, angular conglomerate, but with sands and finer, rounded conglomerate; this
may reflect initial reworking of the sedimentary cover of the Troodos Massif, prior to the introduction of large volumes of newly-derived coarse, ophiolitic detritus (facies A1 and A2); in particular, it may reflect reworking of the poorly consolidated, sandy and silty Nicosia, Kakkaristra and Apalos Formations, which occur immediately below the Fanglomerate stratigraphically, and which would have acted as a source of fine-grained sediment and minor conglomerate; when sediment-laden currents reached the Troodos foothills, they experienced a sharp change in gradient, rapidly expanded and decelerated, and deposited their load; this produced a series of horizontally stratified sands and conglomerates, with little evidence of channel or bar development (e.g. channel-fill sequences or cross-bedding); deposition thus mainly involved sheet flow processes.

7.3 Sedimentological Model

7.3.1 General model

The coarse, unfossiliferous sediments of the Fanglomerate form a thin stratigraphic unit, which was deposited in an alluvial fan environment. Two phases of deposition are identified.

Initially, a phase of valley incision occurred along the northern margin of the Troodos Massif (Fig. 7.3a). At least two such valleys are preserved near Kambia and Vyzakia (see asterisks, Fig. 7.1), though others may exist. Following incision, valleys were partially filled with the thickest observed Fanglomerate deposits (up to 12m). In the east, these deposits are sandy and contain conglomerates with rounded clasts (facies B1), even close to the Troodos Massif. This is believed to reflect initial reworking of the sedimentary cover of the ophiolite, in particular poorly consolidated Pliocene–Early Pleistocene silts, sands and conglomerates, prior to flooding with large volumes of coarse, newly-derived sediment. Most sediment was deposited rapidly, largely by sheet flow processes. In the west, near Vyzakia, facies are apparently more conglomeratic (Wilson, 1959), and near Potami, at least two generations of conglomeratic channel fill have been observed.

Following initial valley incision and filling, huge quantities of very coarse, conglomeratic sediment, largely derived from the
Fig. 7.3 - Schematic block diagram reconstructions for the Fanglomerate, showing:

a) initial valley incision, and deposition of a small conglomerate and sand lobe, followed by
b) deposition of a large sheet of coarse conglomerate.
Troodos Massif, were shed off the ophiolite in a number of thin sheets (facies A1), during the second phase of deposition (Fig. 7.3b). These sheets covered the earlier Fanglomerate deposits, and spread far out (up to 15km) into the Mesaoria basin, covering all older sediments. Some renewed valley incision may have occurred in the Troodos Massif, giving rise to new feeder channels for the sheets. Fanglomerate facies, preserved in patches in elevated positions overlying Troodos basement, not studied in this project, may stem from this depositional period. In the basin, however, conglomerates spread out, with very little in the way of down-cutting.

As with earlier Fanglomerate deposits, no organised system of channels or braid bars was developed, and conglomerates were largely transported in a number of huge sheet flows of varying sediment concentration. Deposition was partly from mass flow, normal stream flow and probable intermediate flow types. Variable palaeocurrent directions and maximum clast size (Fig. 7.2) may reflect the presence of a number of different, overlapping sheets, which moved in slightly different directions and carried variable clast size populations. Further, carefully collected data are necessary to verify this, however. Poole (unpublished report, Edinburgh University) records the development of braided river facies in more distal locations.

Following deposition of only 2-5m of these coarse conglomerates, sudden and deep incision into them began. Since this time, incision has continued, and a series of terraced river courses are now cut into the Fanglomerate and older formations of the basin. These young sediments (Young sediments of Encl. A) have not been studied.

7.3.2 Extrabasinal controls and tectonic implications

The transition from rather passive alluvial fan sedimentation, with deposition of significant volumes of fine-grained overbank facies, during Apalos Formation times (chapter 6), to the sudden appearance of very coarse, angular, alluvial conglomerates, marks a period of dramatic change in the depositional character of the Mesaoria basin. Several extrabasinal controls may have been
Firstly, climatic factors must be considered. The Fanglomerate is very similar to the conglomerates of the Fan II deposits of Maizels (1987b) from the Pleistocene of Oman. These latter sediments are also characterised by coarse, massive, conglomerates with little cross-bedding, or evidence of channel and bar development. They are interpreted as the products of broad sheet flows, deposited during an arid climatic period in the Quaternary, when discharge was highly episodic. Cyprus is unlikely to have been as arid as the Oman, however. In addition, the thinness and areal extent of the Fanglomerate (at least 1200 km$^2$ compared with 400 km$^2$ for the Fan II deposits) are remarkable. Climatic effects may have been sufficient to influence depositional processes (and may have been the cause of domination of sheet flow processes). They are not on their own, however, believed to have been capable of introducing such large volumes of coarse sediment into the Mesaoria basin, so suddenly, during Fanglomerate time.

Secondly, a eustatic sea level fall may have been a contributing factor to Fanglomerate deposition, but fluvial incision would have been expected to be an important process within the Mesaoria basin, as well as along its southern margin, if this were the case. Instead, Fanglomerate deposits in the centre of the basin are characterised by remarkably flat bases.

Tectonic effects are therefore considered to have been primarily responsible for generation of the Fanglomerate. A period of renewed uplift in the Troodos Massif initially resulted in valley incision, and subsequent infilling largely with sandy, reworked, older Plio-Pleistocene sediments. This was followed by the vast outpouring of coarse, ophiolitic detritus, which travelled far across the basin. Uplift was drastic and resulted in deposition of the coarsest sediments seen anywhere in the entire sedimentary fill of the basin.

**Serpentinisation of the Troodos Massif**

Fanglomerates in the western Mesaoria, not investigated in this study, contain appreciable numbers of ultramafic clasts (Wilson, 1959), derived from the plutonic core of the Troodos Massif. This is the first recording of significant amounts of ultramafic debris in the
sedimentary cover of the ophiolite, suggesting that unroofing of its plutonic core was a late stage event, only occurring for the first time in the Pleistocene (see section 8.3). The timing of the unroofing correlates with possible large-scale serpentinisation of the plutonic core (see section 11.2.2, Upper Pliocene-Pleistocene subsection), suggesting that uplift of the Troodos Massif may have been enhanced at this time by the release of serpentinising fluids.

Later incision

Incision into the Fanglomerate, and subsequent cycles of renewed incision and deposition, resulting in the generation of the terraced river courses which cross the Mesaoria Plain today, may be the product of continued, pulsed uplift of the Troodos Massif. In addition, the Mesaoria Plain itself may now be experiencing uplift.

Eustatic sea level fluctuations and accompanying climatic changes must also be taken into consideration, however. Furthermore, river terrace sequences may develop in response to purely sedimentological factors. During a period of major flooding, incision will occur in places where local geomorphic thresholds for channel entrenchment are exceeded. Downstream, however, deposition may take place. Successive flood events, of varying strength, then similarly incise at some localities, but deposit at others, and a series of river terraces may develop e.g. arroyo cutting and filling in Colorado (Womack and Schumm, 1977), and glacial outwash terrace formation in Iceland (Maizels, 1987a).

Wilson (1959) believed stream entrenchment in Cyprus was a result of pulsed uplift, and recognised four stages of uplift. Ducloz (1965), on the other hand, correlated the Fanglomerate with the four classical Alpine glacial periods. He also divided the Fanglomerate deposits investigated in this study into two units, on the basis of different topographic elevations. He believed an older Fanglomerate unit (his Kantara gravel) was preserved close to the Troodos Massif and dipped gently towards the basin centre, and that a younger unit (his Kambia gravel) was also preserved in the basin centre, but at higher elevations. This interpretation was not confirmed during the present study.

As river terrace deposits were not the subject of study during
this project, a solution to the problem of their genesis is not presented here. It seems likely, however, that tectonic and eustatic factors would both have played a role, and combined to give the morphological and sedimentological products preserved today.

7.4 Coarse Clastics on the North Side of the Basin

Several Pleistocene units younger than the Athalassa Formation have been recognised in the Kyrenia lineament. These include the Karka Formation (intramontane lacustrine facies and angular scree conglomerates; Ducloz, 1972; Baroz, 1979), Fanglomerate (Moore, 1960), older Pluvial deposits (fossil talus deposits, breccias and lacustrine sediments; Ducloz, 1972), and river terrace and raised beach deposits. Their exposure is patchy, however, and correlation with the Apalos Formation, Fanglomerate and river terrace deposits of the Mesaoria Plain uncertain (see also section 2.2.2).

These Pleistocene sediments do reveal, however, a marked change from the shallow marine conditions prevailing at the end of Athalassa Formation times (section 5.6), and attest to major uplift of the Kyrenia lineament. As a result of uplift, coarse sediment (screes, breccias and fanglomerate) and lacustrine facies were deposited, often in incised, intramontane settings.

Uplift of the Kyrenia lineament is believed to mark the main phase of compression of the lineament, which had begun earlier at the beginning of Athalassa Formation times (section 5.4.2), but then waned. Early compression had first affected the poorly consolidated sedimentary cover on the south side of the lineament (the Kythrea flysch; section 1.3.2). This was folded and faulted as reactivation of the Ovgos fault took place. During renewed compression, deformation was then concentrated at the core of the lineament, where hard, competent, pre-existing thrust sheets (see section 1.3.2) were upended to give their present, near vertical orientations (Robertson and Woodcock, 1986). Many of the Pleistocene deposits described above are found in this core area. Because of its much smaller size, less coarse sediment was shed from the Kyrenia lineament than from the Troodos Massif, and was not transported far beyond the range.
Timing of uplift of the Troodos Massif, documented by the Fanglomerate, and the onset of renewed compression of the Kyrenia lineament cannot be correlated precisely because of the poor correlation between Pleistocene sediments of the Mesaoria basin and Kyrenia area. The two events were broadly coeval, however, and rejuvenation of the Troodos Massif is also believed to have been a response to compression. For the first time, the whole of Cyprus was reacting to tectonism as a single structural entity.

7.5 Summary - Basin Evolution during Fanglomerate Times

1. Immediately prior to deposition of the Fanglomerate, several small alluvial fans were building off the northern margin of the Troodos Massif into the Mesaoria basin (Apalos Formation). The basin was probably entirely continental by now, and filling with alluvial facies.

2. A major pulse of uplift then affected the Troodos Massif. Initially, valleys were incised into the margins of the ophiolite, and at least partly filled with sand-rich facies, probably reworked from older Plio-Pleistocene sediments.

3. As uplift continued, huge quantities of coarse, Troodos-derived sediment were shed from the ophiolite in a number of vast sheets, that reached far across the basin.

4. Deposition was dominated by mass flow and sheet flow processes, with very little evidence of channeling. This may reflect an arid climate associated with unstable, flashy discharge.

5. Uplift of the Troodos Massif did not occur in isolation. Large-scale compression and uplift of the Kyrenia lineament also took place, resulting in deposition of fossil talus, breccias and lacustrine sediments, often in incised mountain valleys.

6. After deposition of only a few metres of coarse conglomerate, deep incision into the Fanglomerate began. Repeated incision, with some deposition, has continued to today, producing a series of terraced river courses. These reflect the interplay between possible continued uplift of Cyprus, and eustatic sea level fluctuations.
Plate 7.1

a) Valley west of Xeri; grass-covered valley sides comprise facies A1 silts of the Nicosia Formation, overlain by a horizontal capping of Fanglomerate; view looking east

b) Poorly sorted, clast-supported, facies A1 conglomerate, showing vague, central coarse layer; staff is 110cm long

c) Clast- to matrix-supported, massive facies A1 conglomerate, predominantly composed of angular to subangular chalk clasts
Plate 7.2

a) Massive, poorly sorted, facies A1 conglomerate, with flat base, overlying facies A2 sands of the Nicosia Formation; staff is 110cm long

b) Coarse, poorly sorted, crudely cross-bedded facies A2 conglomerate; cross beds dip to the right; ca. 100cm of staff are showing
Plate 7.3

a) Interbedded sands and minor conglomerates of facies B1, filling a broad channel cut into facies B2 sands of the Nicosia Formation; cliff face is ca. 10m high

b) Interbedded sands and minor conglomerates of facies B1, showing structureless sands, veined with recent caliche, and thin conglomerates with moderately sorted, subrounded clasts; hammer is 30cm long
8.1 Introduction

Petrographic analysis of selected samples of sediments from the Mesaoria basin has been carried out to determine provenance, and to study diagenetic effects. In addition, it is now well established that detrital mineralogy of sandstones can be an important aid in the evaluation of palaeogeographic and tectonic evolution of source regions (e.g. Dickinson and Valloni, 1980). Petrographic trends in the Mesaoria basin have thus been used to try to shed light on the unroofing history of the main source area for the basin, the Troodos Massif.

Results are based on thin section studies and some X.R.D., S.E.M. and cathodoluminescence work (section 1.5). Thin section work included point counting of sandstone samples from the main sedimentary formations of the basin. As this project was mainly field-based, these analyses were not intended to be exhaustive, but to be sufficient to document general trends. Conglomerates were point counted in the field.

8.2 General Composition

Clays

The finest fractions from all formations were analysed using X.R.D. techniques (section 1.5). Clay mineral assemblages throughout the studied succession were found to be dominated by smectite, with subordinate quantities of illite and kaolinite, and very little chlorite (Fig. 8.1). No vertical changes in composition were observed (though quantitative X.R.D. measurements were not made).

This assemblage of clay minerals is very similar to that found in Plio-Quaternary clays from DSDP holes 375 and 376, which lie 60 km west of Cyprus (Mélières et al., 1978). The clay content of these latter sediments is largely attributed to weathering of the basic igneous Troodos ophiolite. Recent clays in Morphou Bay, on the northwest coast of Cyprus and adjacent to the Troodos Massif, are also rich in smectite (Shaw, 1978). Clay minerals of the
Fig. 8.1 - Typical X.R.D. trace for the fine-grained fraction from Plio-Pleistocene sediments of the Mesoaria basin (Nicosia, Kakkaristra, Athalassa and Apalos Fms.), showing a clay mineral assemblage of smectite and subordinate illite, chlorite and kaolinite. The trace for the heated sample shows collapse of the 14Å smectite and 7Å kaolinite peaks (though a small chlorite peak remains), and development of illite. The glycolated trace shows a typical shift of the 14Å smectite peak.

C/K - chlorite/kaolinite  F - feldspar  I - illite  Q - quartz  S - smectite
Plio-Pleistocene sediments of the Mesaoria basin are clearly predominantly derived from the same source.

A small contribution to clay mineralogy is also likely to have been made by the Kyrenia lineament, on the north side of the basin (see Fig. 8.5). This would be difficult to detect, however, because the lineament is also likely to have contributed smectitic clays, derived from its main lithology, the smectite-bearing Kythrea flysch (Baroz, 1979).

An increase in smectite and decrease in chlorite is apparent across the Miocene-Pliocene boundary in DSDP holes 375 and 376 (Mélières et al., 1978). This may reflect two factors: increased erosion of the Troodos Massif as a consequence of its emergence above sea level at end-Miocene times (section 1.3.3), and subsequent reduction of chlorite, which was perhaps largely derived from other terrigenous sources, e.g. Turkey.

In the eastern Mesaoria basin, some claystones in the Kakkaristra, Athalassa and Apalos Formations are very pale and highly calcareous (up to 60% carbonate by weight; facies D1, D1 and C1 respectively). This is attributed to local derivation of clay largely from the chalky and marly Pakhna and Lefkara Formations (section 1.3.3) of the pre-Pliocene sedimentary cover of the Troodos Massif. These formations crop out most extensively in the east part of the Mesaoria Plain at present (Fig. 8.5), and this may also have been the case in the Plio-Pleistocene. Clay-grade material derived from these formation was probably the finest sediment generated in the Troodos source area. It would have been very susceptible to aeolian transport and was periodically concentrated and deposited as the facies in question.

**Sandstones**

Sandstones of the Mesaoria basin contain three major components: rock fragments, mineral grains and bioclastic material (Table 8.1). Rock fragments are volumetrically the most important component (Fig 8.2a; Plates 8.1b and 8.2a), except for samples from the Athalassa Formation (Plates 8.3a and b) and a few others. These others belong to bioclastic-rich facies of the Kakkaristra Formation (facies B3 and C4) and the Apalos Formation (facies B2).
Fig 8.2 - Composition of sandstones from the Mesaoria basin.
Rock fragments are of igneous, metamorphic and sedimentary origin (Table 8.1), with igneous types being by far the most common (Fig. 8.2b). Sedimentary rock fragments, which are mainly of chalk or foraminiferal marl, are present to varying degrees, and are occasionally predominant. Metamorphic rock fragments are very minor, but of great significance, because they have only one possible source area, the Kyrenia Range (see Table 8.2 and section 8.4).

Mineral grains mainly comprise feldspar, with generally low proportions of quartz and ferromagnesian minerals (Fig. 8.2c). Bioclastic grains are predominantly of skeletal carbonate (Table 8.1), and largely represent carbonate production and deposition contemporaneous with clastic sediment deposition. Some highly corroded grains, which are uncommon, may represent reworking of older limestones. In addition, pelagic foraminifera occur in a variety of unexpected facies, e.g. deltaic and fluvial (Table 8.3). Some of these foraminifera have been identified as of Eocene and Miocene age (Table 8.3), and represent material reworked from older parts of the Troodos sedimentary cover (Pakhna and Lefkara Formations; see also source areas subsection).

In general, sandstones of the Mesaoria basin can be described as mineralogically immature, containing large quantities of igneous rock fragments and feldspar, with relatively little quartz (Tucker, 1981). Texturally, they vary from moderately to well sorted, with subangular to subrounded grains. Lack of well rounded grains, even in beach facies, probably reflects short transport paths from source to depositional sites. Sorting is a product of individual environment e.g. facies B2 shoreline sands of the Kakkaristra Formation are very well sorted, while fluvial sands of the Apalos Formation (facies B1) are only moderately sorted.

Conglomerates

Conglomerates contain two major clast types: igneous and sedimentary. Igneous clasts are of basalt, dolerite or rarely gabbro, whilst sedimentary clasts are of chalk, indurated marl, occasional chert and jasper, and rarely reef limestone. These clasts are all derived from the Troodos ophiolite, and its overlying sedimentary cover. No clasts originating from the Kyrenia Range were identified.
Table 8.1 - Grain types identified in sandstones

| Rock fragments: igneous:       | glassy basalt                           |
|                               | crystalline basalt                      |
|                               | dolerite                                |
|                               | gabbro                                  |
|                               | peridotite                              |
|                               | plagiogranite                           |
| sedimentary:                  | chalk                                   |
|                               | foraminiferal marl.                     |
|                               | chert                                   |
|                               | sandstone                               |
| metamorphic:                  | recrystallised limestone/               |
|                               | dolostone                               |
|                               | schist                                  |
| Minerals:                     | feldspar (mainly plagioclase)           |
|                               | quartz                                  |
|                               | ferromagnesiants (pyroxene              |
|                               | and amphibole)                          |
|                               | epidote                                 |
|                               | opaque minerals                         |
| Bioclastic grains:            | echinoderms                             |
|                               | bivalves                                |
|                               | calcareous algae                        |
|                               | benthic foraminifera                    |
|                               | pelagic foraminifera                    |
|                               | gastropods                              |
|                               | miscellaneous shell                     |
|                               | fragments                               |
|                               | peloids                                 |
Igneous clasts are by far the most important, and of these, basalt and dolerite are predominant (Fig. 8.3a). Gabbroic clasts were only occasionally observed, and very rarely plagiogranite. No ultramafic clasts were identified. Dolerite clasts predominate over basalt in all formations (Fig. 8.3b). This perhaps partly reflects the greater ease of mechanical breakdown of lavas into sand-, silt- and clay-sized sediment, compared with coarser dolerites.

Sedimentary clast content is usually small, but is very occasionally predominant e.g. in conglomerates of the Apalos Formation and Fanglomerate. Sedimentary clast-rich conglomerates are also found in the lower Nicosia Formation (facies B1d). Chalk and indurated marl, derived from the Pakhna and Lefkara Formations, are the most common sedimentary clast type, but locally reef limestone occurs. These latter clasts are only found close to isolated outcrops of Miocene Koronia Limestone (section 1.3.3), from which they are derived. Rarely, reef limestone forms huge megaclasts in conglomerates of the lower Nicosia Formation (facies B1c) e.g. near Politiko (WD 218758).

Texturally, conglomerates are variable. Broad trends are apparent in clast roundness and sorting, however (Fig. 8.4). These largely reflect the shallowing and infilling of the Mesaoria basin. For example, good sorting and rounding occur in the fan-deltaic Kakkaristra Formation, which contains beach conglomerates. Sorting and rounding are poorer in the younger, entirely fluvial Apalos Formation, while textural parameters are poorest of all in the most proximal sediments in the basin, the alluvial fan conglomerates of the Fanglomerate.

A reversal of the trend is seen between the two oldest formations, the Kakkaristra and Nicosia. This is again, however, a product of depositional setting. The conglomerates of the Nicosia Formation belong to the subaqueous parts of small, slope fan-deltas, which rapidly built off a steep basin margin (section 3.3.1). They did not undergo extensive reworking in a beach environment, as did some of the conglomerates of the Kakkarista shelf fan-delta.

Good sorting and rounding is found in some unexpected facies, e.g. mouth bar conglomerates of the Kakkaristra Formation (facies B4). These conglomerates were transported fluvially, and then
Fig. 8.3 - Composition of conglomerate clasts from the Mesaoria basin.

Fig. 8.4 - Conglomerate clast textures

Horizontal bars indicate the range of clast rounding or sorting (qualitative only, based on eyeball estimates of field exposures). Dashed lines indicate general trends in rounding and sorting.
rapidly dumped at river mouths. Well rounded clasts in this facies may reflect reworking of older, partially rounded conglomerates of the poorly consolidated Nicosia Formation, which was uplifted along with the Troodos Massif at the beginning of Kakkaristra Formation times (section 4.2.2). Clast composition cannot be easily used to distinguish source lithologies because nearly all clasts were ultimately derived from a single source, i.e. the Troodos ophiolite.

Rounded conglomerates are also found in facies B1 of the alluvial fan Fanglomerate (section 7.2). These again are thought to reflect reworking of older Plio-Pleistocene conglomerates.

**Source areas**

Two source areas exist for the Plio-Pleistocene sediments of the Mesaoria basin: the Troodos Massif to the south and the Kyrenia Range to the north (Fig. 8.5, Table 8.2). The large quantities of basaltic and doleritic clasts in conglomerates, and the predominance of igneous rock fragments and feldspar in sandstones clearly point to the Troodos Massif as the major source area for the sediments. This is not unexpected in view of:

a) the much larger size of the Troodos Massif available for erosion in comparison to the Kyrenia Range (Fig. 8.5);

b) exposure of the Troodos Massif since end-Miocene times, but of the Kyrenia lineament only since approximately the Upper Pliocene, and

c) the proximity of the study area, located mainly in the southern part of the basin, to the Troodos Massif.

The petrography of sediments from the north side of the basin, not studied in this project, is not well documented, but some Troodos-derived material is reported by Moore (1960) and Baroz (1979). Also, samples from the Athalassa Formation from near Nicosia, investigated in this study, contain a variety of Troodos-derived grains. Ophiolitic sediment was thus transported to, and deposited in, the northern part of the basin. The Kyrenia lineament contributed some clastic sediment, however, as witnessed by the occurrence of metamorphic rock fragments. This is further discussed in section 8.4.

The bulk of the sediment derived from the Troodos source area
Table 8.2 - Main lithologies in source areas

<table>
<thead>
<tr>
<th>Troodos Massif:</th>
<th>basalt</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>dolerite</td>
</tr>
<tr>
<td></td>
<td>gabbro</td>
</tr>
<tr>
<td></td>
<td>plagiogranite</td>
</tr>
<tr>
<td></td>
<td>ultramafic lithologies</td>
</tr>
<tr>
<td></td>
<td>serpentinite</td>
</tr>
<tr>
<td></td>
<td>chalk</td>
</tr>
<tr>
<td></td>
<td>marl</td>
</tr>
<tr>
<td></td>
<td>minor chert</td>
</tr>
<tr>
<td></td>
<td>gypsum</td>
</tr>
<tr>
<td></td>
<td>jasper, umber &amp; ochre</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Kyrenia Range:</th>
<th>lithic sandstone</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>recrystalised limestone/dolostone</td>
</tr>
<tr>
<td></td>
<td>minor chalk</td>
</tr>
<tr>
<td></td>
<td>chert</td>
</tr>
<tr>
<td></td>
<td>gypsum</td>
</tr>
<tr>
<td></td>
<td>acid and basic volcanics</td>
</tr>
<tr>
<td></td>
<td>schist</td>
</tr>
</tbody>
</table>
comprises material of ophiolitic origin. Sediment derived from the pre-Pliocene sedimentary cover of the Troodos ophiolite is also present. Occasionally, it becomes predominant, both in conglomerates (Fig. 8.3a; see conglomerates subsection) and sandstones (Fig. 8.2b). In sandstones, it is found particularly in a small area in the eastern part of the fan-delta system of the Kakkaristra Formation (Fig. 8.6). Here, some samples of coastal and nearshore facies (facies B1, C1 and C4) contain up to 80% reworked pelagic foraminifera and foraminiferal marl fragments (Plate 8.2b). Some of the foraminifera have been identified as reworked from the Pakhna and Lefkara Formations (Table 8.3). It is thus apparent that some tributaries feeding the eastern part of the Kakkaristra fan-delta were draining across a predominantly chalky, marly hinterland. This type of hinterland was also an important source of very fine sediment, which was periodically reworked by aeolian processes (see clays subsection).

Thus, although the Troodos ophiolite was the major sediment source for the Mesaoria basin, locally the overlying sedimentary cover was very important. This reflects very local sourcing in some instances.

8.3 Troodos Unroofing History

The Troodos Massif had been a source of clay-grade sediment since before the Pliocene (Mélières et al., 1978; Baroz, 1979). Only in the Upper Miocene did it begin to shed coarser material, heralding its emergence above sea level (section 1.3.3). Initial deposition of ophiolite-derived sands and conglomerates was only very local, however (Eaton, 1987), and large volumes of biogenic sediment were still being deposited at this time.

A dramatic change ensued at the beginning of the Pliocene, ably demonstrated by the striking change in colour from the cream and pale brown marls of the Pakhna Formation, to the dark brown and grey-green silts of the Nicosia Formation (section 3.2.2). In addition, conglomerates and sands of ophiolitic origin are present in much of the lower Nicosia Formation. Thus, for the first time, the Troodos Massif was significantly exposed, and became the major detrital source for the sediments of the Mesaoria basin.
Table 8.3 - Sandstone samples containing reworked pelagic foraminifera

<table>
<thead>
<tr>
<th>Sample</th>
<th>Fm. and age</th>
<th>Depositional environment</th>
<th>Name and age of reworked foram.</th>
<th>Fm(s). from which forams. are derived</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Nicosia Fm. (Plio.)</td>
<td>Marine toe of fan-delta lobe</td>
<td><em>Orbulina</em> sp. (Miocene)</td>
<td>Pakhna Fm.</td>
</tr>
</tbody>
</table>
| 2      | Kakkaristra Fm. (?)  | Fan-delta shoreline facies | *Gumbeletria* ?aff. *triseriata* (Eo.)  
*Globigerina* sp.       | Lefkara Fm.                                
Lefkara, Pakhna or Nicosia Fms. |
| 3      | Kakkaristra Fm. (?)  | Fan-delta shoreline facies | *Cibicides* ?westi (Eo.)  
*Globigerina* and  
*Globorotalia* sp.     | Lefkara Fm.                                
Lefkara, Pakhna or Nicosia Fms. |
| 4      | Apalos Fm. (?)       | Fluvial facies            | *Globigerina* and  
*Globorotalia* sp.          | Lefkara, Pakhna or Nicosia Fms. |
The Troodos Massif itself comprises a typical ophiolite assemblage of pillow lavas, sheeted dykes and plutonic complex (Gass and Masson-Smith, 1963; Fig. 8.5). By documenting vertical changes in detrital mineralogy, it is possible to shed light on the timing and manner in which the ophiolite was unroofed. Such petrographic trends have been reported from the sedimentary cover of other ophiolites, e.g. the Zambales in the Philippines (Schweller et al., 1983, 1984). Before these trends are discussed, the stratigraphy of the Troodos Massif is briefly outlined.

**Troodos stratigraphy**

In the Troodos Massif, pillow lavas, representing the shallowest levels of ophiolite stratigraphy, are traditionally divided into upper and lower groups (Gass, 1980). These groups are geochemically slightly different, and show differing degrees of hydrothermal alteration (roughly zeolite facies metamorphism in the upper group, and zeolite to greenschist facies in the lower group; Gass, op. cit.). The lavas themselves comprise basaltic, pillowed and sheeted lava flows. The cores of some massive flows may be quite coarse, and dykes become increasingly common in the lower group. The Basal Group (over 50% dykes) separates lower pillow lavas from the sheeted dyke complex, which comprises doleritic to dioritic, and occasional basaltic, lithologies (Panayiotou, 1987). The contact with the plutonic core of the ophiolite is gradational and complex. The core itself comprises gabbroic and plagiogranitic rocks at higher levels, and ultramafic lithologies at deeper levels.

The ophiolite presently forms a huge, dome-shaped, structure. Ultramafic rocks crop out at the apex of this dome, in the Mount Olympus area (Fig. 8.5). Successively shallower levels of the ophiolite's stratigraphy outcrop in roughly concentric belts around this ultramafic core (Fig. 8.5).

**Sandstone and conglomerate petrography**

In this study, the proportions of coarse-, medium- and fine-grained igneous rock fragments in sandstones were used as indicators of derivation from a particular level within the Troodos complex. Fine-grained and glassy rock fragments are derived from
Fig. 8.5 - Major structural elements and geological units, Cyprus.
the shallowest, pillow lava, levels of ophiolite stratigraphy. No attempt was made to distinguish between upper and lower pillow lava groups. Medium-grained igneous rock fragments are largely derived from the sheeted dyke complex, although some may come from the cores of massive lava flows. Coarse-grained rock fragments (with average grain size greater than $\frac{2}{3}$ mm) are considered to have been derived from coarser dolerites within the sheeted dykes, and from finer-grained gabbros and plagiogranites from the plutonic core of the ophiolite. A marked increase in the numbers of feldspar and ferromagnesian mineral grains would indicate derivation largely from gabbroic rocks, while a majority of ferromagnesian grains would reveal erosion of the ultramafic central core of the ophiolite (cf. Scheweller et al., 1983).

Results show that sandstones of the oldest Pliocene Formation of the Mesaoria basin, the Nicosia Formation, are richest in fine-grained igneous rock fragments (Fig. 8.2d). The succeeding stratigraphic unit, the Kakkaristra Formation is generally richer in medium-grained rock fragments, while the youngest Apalos Formation shows an enrichment in coarse-grained igneous rock fragments. Very coarse-grained rock fragments, or ultramafic igneous rock fragments, have not been observed, nor are any clear variations in feldspar or ferromagnesian content apparent (Fig. 8.2c). These trends reflect exposure and erosion of successively deeper levels of ophiolite stratigraphy, but only as far as the outer margins of the plutonic core of the Troodos complex. Sediments with a large content of ultramafic or gabbroic detritus, predominantly derived from the plutonic core of the Troodos Massif, are absent.

Samples from the Athalassa Formation, which is stratigraphically equivalent to the Kakkaristra and lower Apalos Formations, show increased amounts of coarse-grained igneous material relative to their lateral equivalents (Fig. 8.2d). The Athalassa Formation is found in the northeast corner of the study area, and this trend may reflect input from the Kyrenia lineament, igneous material being reworked from the lithic sandstones of the Kythrea flysch (A. Robertson, pers. comm.; see further, section 8.4). Other evidence from the Athalassa Formation also points to sediment contribution from this source (see section 8.4). Kakkaristra sandstone samples
from within, or near, the zone of interdigitation with the Athalassa Formation (marked with b on Fig. 8.2d) also show this enrichment.

Trends in conglomerate clast composition (Fig. 8.3a) show broadly the same features as sandstone petrography. Trends in dolerite:basalt clast ratios are not very clear (Fig. 8.3b), but show the Nicosia Formation (oldest formation) to contain the highest proportions of basalt, and the Fanglomerate (youngest formation) the most dolerite. Gabbroic clasts are very uncommon, and are not found in the oldest, Nicosia Formation. No ultramafic clasts were recorded at all. These changes indicate progressive erosion of first pillow lavas, with some sheeted dykes, then sheeted dykes with lesser amounts of pillow lava clasts, and finally the occasional contribution of sediment from the outer parts of the plutonic core of the Troodos Massif.

**Plutonic core**

As already outlined, the plutonic core of the Troodos Massif is not well represented in the sediments studied. It must be noted, however, that the eastern part of the Mesaoria basin, which is the area most extensively investigated in this study, was perhaps largely outwith the influence of rivers draining the plutonic complex (Fig. 8.5), as it is today. A small number of sandstone samples were available from the west, but none showed significant amounts of gabbroic detritus, or any ultramafic debris. Fanglomerates from this area were not studied. Wilson (1959), however, records the presence of peridotite clasts in Fanglomerate facies in the west, and this has been confirmed by A. Poole (pers. comm.).

It is concluded that exposure of the plutonic core of the Troodos Massif, presently found at the apex of the dome-shaped Troodos mountains (Mount Olympus area, Fig. 8.5) is a very late stage event. Ultramafic rocks were apparently only exposed for the first time during Pleistocene uplift associated with Fanglomerate deposition (section 7.3.2). Recent conglomerates would therefore be expected to show the greatest enrichment in ultramafic and gabbroic clasts.
8.4 Sediment Contribution from the Kyrenia Range

Clastic sediment

The contribution of material derived from the Kyrenia lineament to the sediments of the Mesaoria basin is difficult to evaluate for two reasons. Firstly, petrographic studies of sediments from the north side of the basin, closest to the lineament, have not been extensive. Secondly, material derived from the lineament is to some extent indistinguishable from that derived from the Troodos Massif. The predominant lithology in the lineament is the Kythrea flysch (Fig. 8.5), which comprises poorly consolidated lithic sandstones, containing large amounts of igneous rock fragments, and mudstones. Reworking of these sediments would be likely to produce quartz, feldspar and igneous rock fragments, all of which may also be derived from the Troodos Massif (though quartz in minor quantities). Minor lithologies, such as chalk, chert, gypsum and basic volcanics (see Table 8.2) are also present in the Troodos Massif.

Three lithologies are, however, diagnostic of the Kyrenia lineament: recrystallised limestone/dolostone from thrust sheets at the core of the lineament (Fig. 8.5), and acidic volcanics and schist, which are of limited areal extent in the lineament. None of the three diagnostic lithologies has been identified in sediments (sandstones and conglomerates) from close to the Troodos Massif. Recrystallised dolostone grains, however, have been identified in this study from the north side of the basin, in samples from the Athalassa Formation (Plate 8.3a; Fig. 8.6). These grains luminesce bright red under CL. Two occurrences of schistose fragments have also been observed in the same formation.

Other petrographic evidence suggests that the Kyrenia lineament was a sediment source for the Athalassa Formation. The Athalassa Formation contains a higher percentage of coarse-grained igneous rock fragments than its lateral equivalent to the south, the Kakkaristra Formation (Fig. 8.2d). These rock fragments may be reworked from lithic sandstones of the Kythrea flysch. The Kythrea flysch itself is largely derived from the northeast, where Tauride ophiolite terrain was exposed in Turkey during its deposition (Weiler, 1965; Robertson and Woodcock, 1986).
Fig. 8.6 - Schematic palaeogeographic setting during Kakkaristra/lower Athalassa Fm. times.

- Athalassa Fm. sandstones containing dolostone grains
- Kakkaristra Fm. sandstones containing dolostone grains
- Kakkaristra Fm. sandstones along the eastern fan-delta front containing increased proportions of quartz & coarse-grained, igneous rock fragments
- Areal extent of Kakkaristra Fm. sandstones rich in reworked pelagic forams.
Samples from the Athalassa Formation also show higher proportions of quartz than most other sandstones (Fig. 8.2c). Quartz is present locally in the Troodos Massif in plagiogranites and some hydrothermally metamorphosed parts of the ophiolite, but it is more common in sandstones of the Kythrea flysch. Quartz grains in the Athalassa Formation tend to be better rounded than those in the Kakkaristra Formation, perhaps reflecting derivation from Kythrea sandstones and a second cycle of reworking. In addition, Gass (1960) reported heavy mineral assemblages from the Athalassa Formation containing kyanite, which occurs in the Kythrea flysch according to Henson et al., (1949).

A few samples from the Kakkaristra Formation show trends similar to the Athalassa Formation i.e higher percentages of coarse-grained igneous rock fragments and quartz (Figs. 8.2c and d). One sample contains a dolostone grain (Fig. 8.6). All these samples occur in, or close to, the area of interdigitation between the Athalassa and Kakkaristra Formations, along the eastern front of the Kakkarista fan-delta (Fig. 8.6).

Bioclastic sediment

The Kyrenia lineament supplied limited amounts of clastic sediment to the north side of the Mesaoria basin, during Athalassa Formation times (see previous subsection). The remainder of Athalassa sediment largely comprises bioclastic material (Fig. 8.2a). That this material was also derived from the north side of the basin is implied by the predominantly south-directed palaeocurrents in the formation (chapter 5). As argued in section 5.3.3 (sand supply subsection), these observations imply that the Kyrenia lineament formed a small, low-lying landmass during Athalassa Formation times. As clastic influx was limited, its margins became favourable sites for skeletal carbonate production. During storms, skeletal and clastic sediment was swept south across the Mesaoria basin, and deposited as a number of carbonate sand bodies.

History of the lineament as a sediment source

The role of the Kyrenia lineament as a sediment source prior to deposition of the Athalassa Formation is uncertain. Marls of the
early Pliocene lower Nicosia Formation from the Myrtou area (western Kyrenia Range) are very fine and free of arenaceous material (Baroz, 1979), suggesting the lineament was largely submerged. Evidence from the upper Nicosia Formation is conflicting. Baroz (1979) reported conglomerates containing recrystallized limestone clasts, though the exact location and extent of these conglomerates is not specified. Moore (1960) makes no mention of conglomerate in the Nicosia Formation in the northwestern part of the Kyrenia Range. If the lineament was emergent at this time, it was probably only locally, and so provided only a very limited sediment source.

Conglomerates with Kyrenia-derived clasts are clearly present in the Athalassa Formation in the northwestern part of the Mesaoria basin (Moore, 1960; Baroz, 1979). Coupled with evidence from the sandstones of the formation from the east, investigated in this project, it is clear that the lineament was beginning to be significantly eroded during Athalassa Formation times, although only relatively small volumes of clastic sediment were being shed.

The Kyrenia lineament only became a major sediment source after deposition of the Athalassa Formation. Following this, a variety of continental deposits, often very coarse, were deposited within the lineament and on its south flank (section 7.4). These sediments were apparently derived exclusively from the Kyrenia lineament, and imply that for the first time, the lineament was extensively uplifted.

8.5 Diagenesis

8.5.1 Introduction

Diagenetic studies of Mesaoria basin sediments have involved the use of several techniques, including staining of thin sections, and limited SEM and cathodoluminescence work (section 1.5). Detailed investigation of cement compositions using microprobe analyses and oxygen and carbon isotopes, and also of cement stratigraphies, involving extensive cathodoluminescence studies, were considered beyond the scope of this project. Diagenetic interpretations are thus necessarily limited, but several important observations have been made. Before discussing diagenesis on a formation by formation basis, some general points, applying to all sediments in the
basin, can be made.

Firstly, much of the succession is poorly consolidated or uncemented. This reflects the young age of the sediments, and their relatively limited burial. From tectonic considerations, only the lower part of the succession (i.e. lower to middle parts of the Nicosia Formation) in the middle of the basin is likely to have suffered any significant burial. These sediments are not exposed, however. As a consequence of lack of burial, compaction features, e.g. sutured grain boundaries, distorted grains or stylolites, are virtually absent from the sediments studied.

The only cement identified in the succession is calcite (ferroan and non-ferroan). Part of the source of CaCO_3 for cementation has come from the dissolution of aragonitic shell fragments. This is found in the Kakkaristra and Athalassa Formations, but not in the Nicosia and Apalos Formations. Cement fabrics are nearly always of a drusy spar type.

In summary, diagenetic effects are not pronounced in the Plio–Pleistocene sediments of the Mesaoria basin. They have produced partial calcite cementation and some aragonitic shell dissolution. Near-surface to shallow burial processes have been dominant.

8.5.2 Nicosia Formation

The Nicosia Formation comprises two main lithologies: silts (facies A1) and muddy sands (facies A2). Minor lithologies include interbedded conglomerates and sands. These occur mainly along the uplifted southern margin of the basin, and have suffered very little diagenetic alteration.

The silts and muddy sands of the bulk of the exposed part of the formation are generally poorly consolidated and very friable. Thin cemented horizons are present, however, in both (e.g. Plate 3.1b). In thin section, interstitial material in these cemented layers stains blue and comprises a very fine microspar or micrite, sometimes too fine to be detected with a microscope. SEM studies show this material to comprise an interlocking fabric of micritic grains (Plate 8.1a), with only rare, stubby authigenic calcite. This fabric may represent neomorphosed primary lime mud, an
interpretation given to similar micritic fabrics by Miller (1986, Lower Carboniferous mud mounds from Ireland) and Lasemi and Sandberg (1984, Plio-Pleistocene micrites from Florida). Neomorphism does not require compaction and burial (Lasemi and Sandberg, op. cit.), but probably took place in reducing conditions where Fe²⁺ ions were available, because the micrite is ferroan. A shallow burial environment is implied. Restriction of neomorphism to certain horizons may reflect the occurrence of clay-poor, carbonate-rich muddy zones within the sediment.

Coarser, sandy facies also contain ferroan interstitial material, but this time as a drusy calcite spar cement (Plate 8.1b). Cementing fluids were clearly able to move through these more permeable sediments, again in a reducing environment. Ferroan spar cements are often associated with burial diagenesis (e.g. Purser and Schroeder, 1986). There is no other evidence, however, that these sediments from the upper Nicosia Formation were ever deeply buried (previous section) and cementation probably took place again in a shallow burial setting. Restriction of cement to certain horizons most likely reflects variations in porosity and permeability in the sediment. Creation of favourable porosity and permeability may have been the result of periodic winnowing of the sea floor, as these sediments were deposited.

The source of CaCO₃ for cementation is unknown. Very little leaching of aragonitic shells has occurred in the Nicosia Formation, unlike the overlying Kakkaristra and Athalassa Formations. This suggests that either the marine sediments of the formation were beyond the influence of corrosive, meteoric diagenetic fluids, permeating into the basin from the shoreline, or that such fluids were inhibited from flowing through much of the silty, relatively impermeable formation.

Minor pyrite cement is sometimes present in Nicosia Formation sediments. Pyrite formation would have been a very early diagenetic event, occurring not far beneath the sea floor in the sulphate-reducing zone. Iron was presumably supplied from detrital ferromagnesian grains and sulphate from sea water (Bjørlykke, 1983).

In summary (Fig. 8.7), following deposition and early pyrite cementation, the studied Nicosia Formation sediments were buried,
DEPOSITION

MINOR PYRITE CEMENT (IN SILTS)

MINOR COMPACATION

FERROAN CALCITE SPAR IN COARSER SEDIMENTS

NEOMORPHISM OF CARBONATE-RICH LAYERS TO FERROAN CALCITE MICROSPAR IN MUDDY SEDIMENTS

OXIDATION

Fig. 8.7 - Diagenetic pathways, Nicosia Formation

DEPOSITION

PARTIAL VOID-FILLING, NON-FERROAN, DRUSY CALCITE SPAR

DISSOLUTION OF ARAGONITIC SHELLS IN MUDDY FOSSIL-INFEROUS FACIES

MINOR OXIDATION

Fig. 8.8 - Diagenetic pathways, Kakkaristra Formation
though not to great depths. Neomorphism of lime mud to ferroan micrite or microspar took place along discrete horizons in silty sediments. Additional carbonate was also introduced and porous horizons in sandier facies were cemented with ferroan calcite. Sediments were not exposed to corrosive, diagenetic fluids. As a result, little leaching of aragonite took place.

8.5.3 Kakkaristra Formation

Sediments of the Kakkaristra Formation are also poorly consolidated. They were deposited in a deltaic environment (see chapter 4), and have been buried to ca. 50m at most. Only sandy facies show any significant cementation. They are generally better sorted and more matrix-free than those of the Nicosia Formation. The most extensive cement development is found in well sorted, shoreline facies (facies B1).

Lithified sandstones are characteristically cemented with non-ferroan, drusy calcite spar, which partially to totally fills voids between grains. The cement is very coarse and poikilotopic in some samples (Plate 8.2a). No fringe cement fabrics were observed. Occasionally, floating grain textures are recorded as the result of displacive calcite growth.

From palaeogeographic considerations, it is unlikely that the Kakkaristra Formation ever underwent any diagenesis in the marine realm. Marine conditions withdrew from the south side of the Mesaoria basin as the formation was deposited, and subsequently fluvial facies were deposited (Apalos Formation) as the area completely emerged (section 6.4). Diagenesis is thus inferred to have occurred in a meteoric setting. The presence of drusy to poikilotopic calcite spar cement is consistent with this. The non-ferroan nature of cements and lack of pendant or meniscus fringe fabrics suggests cementation in phreatic, oxidising conditions.

Fossiliferous sediments in the Kakkaristra Formation are found in muddy, subaqueous bay facies (e.g. facies C1). Although fine-grained, lenses of fresh, corrosive diagenetic fluid were able to permeate these facies and cause leaching of aragonitic shells, leaving behind bivalve moulds and casts. This may have provided a source of carbonate, although of limited extent, for cementation of
In summary, the Kakkaristra Formation has undergone little diagenetic alteration. Minor dissolution of unstable shell fragments has occurred. Calcite cementation took place in the shallow, meteoric realm, but is very patchy, suggesting only local flushing of cementing fluids, largely through the most porous and permeable facies. Diagenetic pathways are summarised in Fig. 8.8.

8.5.4 Athalassa Formation

The Athalassa Formation comprises two main facies: very fine, muddy sands (facies B1), which are poorly consolidated and very friable, and calcarenitic sands (facies A1 and A2), which are poorly to well cemented. Diagenesis of these latter sediments is considered here.

Calcarenites of the Athalassa Formation exhibit four main diagenetic characteristics - early echinoderm overgrowths (Plate 8.3b), non-ferroan, calcite fringe cements, dissolution of unstable shell fragments, and void-filling, non-ferroan, drusy calcite spar cement. Fringe cements are equant and isopachous to nonisopachous (Plate 8.3a). Degree of development varies from sample to sample, and is typically best around bioclastic grains (Plate 8.3a). Dissolution of aragonitic shell fragments occurred mainly after fringe cements formed, as witnessed by moulds supported by thin cement crusts (Plate 8.3b). Later, inter- and intra-particle voids were filled with drusy spar. Again, this cement varies from poorly developed (Plate 8.3b) to completely void-filling (Plate 8.3a).

The interpretation of carbonate cements on the basis of morphology alone must be treated cautiously (Purser and Shroeder, 1986; Hird and Tucker, 1988). Equant fringing spar cements, for example, have been interpreted as both meteoric and marine in origin (see Hird and Tucker, op. cit.). A meteoric origin is favoured here, however, because there is no other evidence for marine cementation e.g. fibrous fringing cements or algal borings. Lack of meniscus or pendant cements further suggests a phreatic environment.

That cementation took place early on is evident from the occasional occurrence of reworked cemented calcarenites in some Athalassa marine sand bodies. Reworking is thought to have taken...
place when old, abandoned, partially buried, sand bodies were ripped up during violent storms, and reworked as calcarenite clasts into active sand bodies. Lithification of these sand bodies was not the result of exposure and beach rock formation, as beach rock cements are not observed in these reworked clasts.

Calcarenites are thus inferred to have experienced cementation soon after deposition. The earliest diagenetic changes involved echinoid overgrowth and fringe cement development, which formed in a shallow, phreatic, meteoric environment. This implies that fresh waters were circulating beneath much of the Mesaoria basin, during deposition of these marine sediments. The region of freshwater diagenesis typically protrudes offshore beneath the surface (e.g. Galloway and Hobday, 1983, Fig. 8.11a). As the Mesaoria basin was very narrow at this time, it is not difficult to envisage the meteoric realm extending virtually the entire width of the basin (Fig. 8.11b). Freshwater recharge took place through the Troodos Massif to the south, and also through the emerging Kyrenia lineament to the north. Lowering of sea level can also induce extension of the meteoric lens offshore (Bjørlykke, 1983). Sea level fluctuations were taking place during Athalassa Formation times (section 5.5), and may well have had some influence on diagenetic changes.

Following early cementation, concentrations of calcium and carbonate ions in diagenetic fluids were lowered, until eventually they were sufficiently undersaturated for dissolution of aragonite to take place (Fig. 8.9). Very minor compaction accompanied this, because cement fringes around moulds are occasionally broken. Finally, remaining pore space was partially to totally occluded in a later phase of cementation, by non-ferroan, drusy calcite spar. A shallow, oxidising, meteoric environment is again implied.

8.5.5. Apalos Formation

Sediments of the Apalos Formation are generally very poorly consolidated. Rare cemented horizons in most facies are associated with caliche development, described in section 6.2 (facies D2).

The only other semi-consolidated sediments occur in calcarenitic facies (facies B2). These sediments contain patchy, non-ferroan, drusy calcite spar, similar to the Kakkaristra Formation. Cementation
DEPOSITION

NON-FERROAN CALCITE ECHINOID OVERGROWTHS AND EQUANT FRINGE CEMENT

DISSOLUTION OF ARAGONITE

PARTIAL TO TOTAL, VOID-FILLING, NON-FERROAN, DRUSY CALCITE SPAR

MINOR OXIDATION

Fig. 8.9 - Diagenetic pathways, Athalassa Formation

DEPOSITION

MINOR, NON-FERROAN, DRUSY CALCITE SPAR IN CALCARENITE FACIES

CALICHE DEVELOPMENT

MINOR OXIDATION

Fig. 8.10 - Diagenetic pathways, Apalos Formation
Fig. 8.11a - Principal diagenetic realms in a sedimentary basin (from Galloway and Hobday, 1983).

Fig. 8.11b - Possible diagenetic realms in the Mesaoria basin during Kakkanstra/Athalassa Formation times.
is again considered to have taken place in an oxidising, meteoric environment (Fig. 8.10). One of the controls on the development of carbonate cements is substrate type (Bathurst, 1975). Preferential cementation of calcarenitic facies in the Apalos Formation may have been due to the presence of calcitic bioclastic grains, which acted as favourable nucleation sites for calcite cement growth.

8.6 Summary

1. Plio-Pleistocene sediments of the Mesaoria basin are mineralogically immature. They are dominated by igneous rock fragments and clasts, and feldspar grains. This reflects derivation largely from the Troodos Massif. Textural immaturity generally increases upwards through the succession, reflecting shallowing and infilling of the basin.

2. Prior to the Pliocene, the Troodos Massif was a source of clay-sized sediment. At the beginning of the Pliocene, it emerged significantly above sea level for the first time, and became a major detrital source area.

3. The unroofing history of the Troodos Massif can be documented from vertical changes in the mineralogy of the Mesaoria basin sediments. Proportions of coarse-grained igneous rock fragments increase upwards, reflecting uplift, exposure and erosion of first pillow lava and then sheeted dyke levels of the ophiolite stratigraphy. Ultramafic clasts or grains, significant quantities of gabbroic material or large numbers of ferromagnesian mineral grains were not observed, indicating that unroofing of the plutonic core of the ophiolite is a relatively recent event.

4. Some reworking of the pre-Pliocene sedimentary cover of the Troodos Massif has occurred. Reworked material generally forms only a small component of the sediments in the basin, but is occasionally predominant. This reflects occasional, very localised sourcing.

5. The Kyrenia lineament contributed very little sediment to the basin in the Early-Mid Pliocene, because it was largely submerged. By upper Pliocene-Pleistocene times, it began to
emerge, and shed limited quantities of clastic sediment into the north side of the basin. At the same time, the fringes of the emerging lineament became favourable sites for the production of skeletal carbonate, which was subsequently transported into the basin. Significant volumes of Kyrenia-derived sediment are only observed in Pleistocene-Recent sediments, exposed within or close to the Kyrenia Range.

6. Most sediments in the basin are poorly consolidated, and diagenetic effects are not pronounced. Where cementation has occurred, it has largely been under shallow, oxidising, meteoric conditions. These observations support other sedimentological and tectonic inferences that the upper part of the succession has never been significantly buried.
Plate 8.1

a) Scanning electron micrograph of interstitial material from a cemented band from a facies A1 siltstone, Nicosia Formation, showing micritic and microsparitic grains with interlocking fabric. Magnification x1370

b) Sandstone from the Nicosia Formation, comprising a variety of igneous rock fragments (mainly basalt, B, and altered glassy basalt, AB), plagioclase laths and grains (P) and minor quartz (Q), set in a calcite spar cement (C; dark in photograph because of staining). Magnification x32
Plate 8.2

a) Sandstone from the Kakkaristra Formation, largely comprising altered igneous rock fragments (I) and plagioclase grains (P), with minor bioclastic material (B). Grains are set in a calcite cement (C; light grey colour), which is coarse and poikilotopic. Magnification x32

b) Sandstone from the Kakkaristra Formation, largely comprising reworked pelagic foraminifera. Minor clastic grains are also present. Interstitial material is mainly epoxy resin. Magnification x32
Plate 8.3

a) Calcarenite from the Athalassa Formation, comprising a variety of bioclastic grains (mainly foraminifera, F, echinoderms, E, calcareous algae, A, and bivalves, B) and clastic material (including a dolostone fragment, D; note also some holes, H). Grains are set in calcite cement, which shows an early fringing phase (f), commonly around bioclastic grains, and a later, well developed, void-filling, drusy calcite spar (s). Magnification x32

b) Calcarenite from the Athalassa Formation, showing an echinoderm fragment with early overgrowth (e), fringing calcite cement (f), and later dissolution of unstable carbonate grains, leaving moulds supported by cement crusts (m). Later, void-filling cement is very poorly developed, and most pore space is filled with epoxy resin (r; light grey colour). Magnification x32.
CHAPTER 9 - PLIO-PLEISTOCENE SEDIMENTS OF SOUTH CENTRAL CYPRUS

9.1 Introduction

Plio-Pleistocene sediments in south central Cyprus were also studied in this project, with a view to comparing Plio-Pleistocene sedimentary successions from north and south of the Troodos Massif, and to try to correlate onshore geology with offshore seismic data (see chapter 10). Pliocene sediments outcrop in only a small area of this region, around Mari and Maroni (Fig. 9.1), while Pleistocene facies occur mainly in a narrow strip along the coast.

Several Plio-Pleistocene sedimentary units have been identified in the study area. Because of limited exposure and lack of biostratigraphic data, however, their stratigraphic relationships are not always clear, and parts of the Plio-Pleistocene stratigraphy are missing. A preliminary stratigraphic correlation is presented in Fig. 9.2 (see also section 2.2.3), but this may be modified by future work. The units identified are as follows:

a) the Lower Pliocene Nicosia Formation;

b) the Vasilikos Formation, a newly recognised formation (see section 2.2.3 and 2.4), of ?Upper Pliocene-Pleistocene age;

c) the Fanglomerate (as mapped by Bagnall, 1960), found only in the eastern part of the study area;

d) fluvial sediments, informally termed Older River Terrace deposits here, found in the western part of the study area, and probably laterally equivalent to the Fanglomerate;

e) Raised Beach deposits, which outcrop along the coast, and correlate with the 40' raised beach of Bagnall (1960) and Pantazis (1967);

f) and fluvial sediments, informally termed Younger River Terrace deposits here, which also outcrop along the coast and were previously mapped as raised beach deposits.

Raised beach and river terrace deposits have not been exhaustively studied, as they are now the subject of a separate study, by A. Poole (Edinburgh University). In addition, modern terraced river courses (e.g. the Vasilikos river) and recent
Fig. 9.1 – General geological map, south central Cyprus (modified from the Cyprus Geological Survey geological map of Cyprus, 1979).

Fig. 9.2 – Preliminary stratigraphic column, Pliocene–Recent sediments, south central Cyprus.
sediments, as mapped by Bagnall (1960) and Pantazis (1967), were not investigated, although it is possible that some of these sediments may include older facies, correlatable with units b–e.

The sedimentary units are now described, in chronological order.

9.2 Nicosia Formation

9.2.1 Introduction

The Nicosia Formation crops out in south central Cyprus only in a small area, around Mari and Maroni (Fig. 9.1). It was formerly mapped as the Upper Pliocene Athalassa Formation (Bagnall, 1960; Pantazis, 1967), but new biostratigraphic data (Paviakelli, 1987) demonstrate it to have a Lower Pliocene age (see section 2.3). Furthermore, lithologically, its facies are very similar to the Nicosia Formation of the Mesaoria Plain (chapter 3).

As in the Mesaoria Plain, the formation unconformably overlies Miocene and older sediments. Its true thickness is unknown, the maximum thickness exposed in the field being only 30m. In addition, it is unconformably overlain by the Vasilikos Formation, and its upper part is missing.

The formation comprises an assemblage of open marine silts, and shallow marine sands, silts and conglomerates, of fan-deltaic origin.

9.2.2 Basal relations

Observations

The basal relations of the formation, like those of the Nicosia Formation in the Mesaoria basin, are important because they reveal information about the structure of the basin into which the formation was deposited. As elsewhere in Cyprus (see section 3.2.1), an unconformity is present at the base of the formation, and fully marine, Pliocene silts overlie Miocene evaporites and marls (Kalavasos and Pakhna Formations, respectively; Fig. 9.3). Coarse-grained intercalations occur within the silts at locality A (Fig. 9.3), close to the unconformity, and are of probable shallow marine, fan-deltaic origin (facies N1b; following section). At locality B to the south, however (Fig. 9.3), micropalaeontological analysis of silts from the
Fig. 9.3 - Geological map of the Mari-Maroni area (see Fig. 9.1 for location). Note stratigraphy is only preliminary, and reconnaissance mapping only was undertaken of sediments younger than the Vasilikos Formation. Based on the geological maps of Bagnall (1960) & Pantazis (1967).
base of the exposed section at this locality, suggest a rather deeper marine depositional environment in this part of the basin (? >200m; see facies N1a, following section).

The unconformity is not well exposed, and can largely be traced only between isolated outcrops (Fig. 9.3). In general terms, its trace forms a large embayment open to the southeast. East of Maroni, field relations are complex, and the unconformity is very likely faulted. Just south of Maroni, gentle folding of Pliocene sediments is visible close to the base of the formation. In the west, the unconformity dips west or south at up to 10°. The surface is only scrappily exposed, and could not be examined in detail.

In the west, Miocene sediments immediately below the unconformity, which have not been described in detail before, contain breccias and conglomerates. At locality C (Fig. 9.3), a 60cm bed of chalk conglomerate is interbedded with more normal chalky marls of the Pakhna Formation. The conglomerate contains angular chert and more rounded chalk clasts, up to 4cm in diameter. No Troodos-derived sediment is present. At locality D (Fig. 9.3), a 20-40cm thick, sucrosic gypsum layer overlies chalky Pakhna marls. This layer is brecciated and folded in a large, gentle, syncline, about a southeast-plunging axis. Above lies a ca. 60m interval of chalky marls, which contains gypsum and chalk conglomerates in its lower section. Near its top, however, is a chalk and chert conglomerate, which contains no gypsum clasts.

These conglomeratic sediments near the top of the Miocene sequence, and folding of the gypsum surface, are suggestive of syndepositional, late Miocene tectonic activity in this area. At locality D, gypsum appears to have been locally reworked into conglomerates after folding and brecciation. Erosion may then have cut deeper to produce the chalk and chert conglomerate, which was probably derived from the underlying Lefkara Formation (the Pakhna Formation contains no chert). Complex field relations in the east may also imply tectonism, but this area was not mapped in detail, and faulting may be post-Pliocene.
Interpretation

A well established phase of compression began to affect southern Cyprus in the Mid Miocene (Robertson, 1977; Eaton, 1987). A narrow deformation zone, containing folded and thrust rocks, developed along the southwestern margin of the Limassol Forest block of the Troodos Massif (the Yerasa fold and thrust belt; Fig. 9.1; sections 1.3.1 and 1.3.3). Eaton (op. cit.) suggested that several thrust-controlled lineaments developed at this time, and sediments of the Pakhna Formation were deposited in basins between them. These lineaments included the Akrotiri High to the south (inset, Fig. 9.1), and a possible smaller lineament, the Ayia Mavri, to the northeast (Fig. 9.1). A further lineament may run through Petounda Point (Fig. 9.1; Cleintaur et al., 1977). Its presence is supported by folding of the Lefkara and Pakhna Formations, visible on offshore seismic data along its possible offshore extension (section 10.4.2).

Eaton (1987) showed that tectonic activity ceased along the Yerasa fold and thrust belt in the Tortonian, because undeformed, uppermost Miocene, Koronia reef limestone facies of the Pakhna Formation transgress across the fault zone. It is now postulated, however, that deformation continued into the Messinian along the Ayia Mavri lineament to the northeast, a time when evaporites were being deposited around the margins of the Troodos Massif in response to the Mediterranean-wide Messinian salinity crisis (Hsü et al., 1978; section 1.3.3). As a result of tectonism, some folding of gypsum occurred, and syntectonic gypsum conglomerates were shed.

The unconformity separating the Miocene and Pliocene is present throughout Cyprus, and much of the Mediterranean. It was largely generated during periods of lowered sea level during the salinity crisis. Seas rapidly flooded back into the Mediterranean, following the ending of the salinity crisis, and in southern Cyprus, as in the Mesaoria basin (section 3.2.2), fully marine sediments were deposited over the unconformity at the beginning of the Pliocene. Micropalaeontological data (see previous subsection) imply that at least part of the basin into which these Pliocene sediments were deposited, was initially quite deep. This is equated with local subsidence associated with continued tectonism along the Ayia Mavri lineament (Fig. 9.4a).
Shallow marine, fan-deltaic facies are recorded further north (locality A, Fig. 9.3), implying the basin shallowed in this direction (Fig. 9.4a). Shallowing to the northeast is supported by offshore seismic data, which show Pliocene sediments to thin in this direction, and to pinch out west of Petounda Point (Figs. 9.1 and 9.4b; section 10.4.2). This suggests that tectonic activity had ceased along the Petounda lineament to the northeast (as it had along the Yerasa fold and thrust belt to the south).

The Pliocene basin, here termed the Marl basin, presumably also shallowed to the northwest, towards the uplifted Limassol Forest block of the Troodos Massif. The setting was somewhat analogous to earlier, Upper Miocene times, when sediments of the Pakhna Formation were fed, via fan-deltas, across the margins of the thrust-controlled Maroni basin, and deposited partly in deep, slope apron facies (Eaton, 1987). In the Pliocene, subsidence was probably less dramatic, coarse-grained sediment was trapped closer to shore on less steep slopes (marine fan-delta facies), and slope apron-type facies are not recorded. Significant subaerial emergence of the Troodos Massif had occurred by this time, and both coarse- and fine-grained Pliocene facies are richer in Troodos-derived material than Miocene sediments.

The Marl basin probably did not deepen much offshore, beyond Zyyi (Fig. 9.1). No deformation to pre-Pliocene sediments is apparent on offshore seismic data, along the trend of the Ayia Mavri lineament (Fig. 9.4b), suggesting it terminates onshore. Furthermore, seismic data do not show thickening of Pliocene sediments offshore along a deepening Miocene/Pliocene unconformity surface (see chapter 10). The Marl basin thus apparently developed a very local depocentre, in the vicinity of Mari, associated with very localised subsidence along the Ayia Mavri lineament.

In summary, compression in southern Cyprus, which had begun in the Mid Miocene, continued into the Messinian and very Early Pliocene along the Ayia Mavri lineament. Local deepening took place, and the small Marl basin evolved. Other tectonic lineaments had become inactive by this time, and the basin shallowed to northeast and northwest. Small fan-deltas at least partly fringed the margins of the basin, while fine-grained marine sediments were deposited
Fig. 9.4a - Schematic cross-section showing the structure of the south Troodos flank in the L. Pliocene. Only the Ayla Mavri lineament was active, and local subsidence ahead of it formed the depocentre of the Mari basin (see Fig. 9.1 for section location and Encl. B for key to symbols).

Fig. 9.4b - Schematic cross-section, based on seismic data, showing the NE pinch-out of Pliocene sediments on the shelf adjacent to the south central Cyprus coast (see Fig. 9.1 for section location).
over the top Miocene unconformity surface in its deeper parts. Continued, but very localised, tectonic activity, along only the Ayia Mavri lineament, can be interpreted as recording the dying stages of this important compressional phase in southern Cyprus.

9.2.3 Facies and facies relations

Facies of the Nicosia Formation of south central Cyprus are described in three groups: silts (facies N1a), which also include coarse-grained intercalations (facies N1b), sands and sand/silts (facies N2a-N2c), and conglomerates (facies N3). These sediments were deposited in deepish to shallow marine environments.

Facies N1 - silts with coarse-grained intercalations

Subfacies N1a silts comprise grey-green, calcareous, fossiliferous, clayey silts, very similar to facies A1 of the Nicosia Formation of the Mesaoria Plain (section 3.3.1). Very little structure is apparent in these silts because of biological reworking, and weathering of these poorly consolidated sediments. Rare calcite concretions occur, and are sometimes subhorizontally elongated, probably parallel to bedding. Macrofossil content comprises broken or occasionally intact mollusc shells, scattered throughout the facies. The fauna are of shallow, neritic origin, and include species of Ostrea, Pecten, Meretrix, Trochus, Turritella, the scaphopod Dentalium, and worm tubes (probably Ditrupa sp.). Comminuted plant debris is sometimes visible. The facies is volumetrically the most abundant in the Nicosia Formation (as it is in the formation of the Mesaoria basin), and occurs throughout the Pliocene outcrop area (Figs. 9.5-9.7).

Palaeontological analysis of ostracod assemblages from these silts, in the south of the study area (locality A, Fig. 9.3; WD 255428), suggests that the lower part of the exposed section was deposited in a rather deep marine setting (Pavlakelli, 1987). The section contains in situ specimens of Oblitacythereis mediterranea, which imply depositional depths of over 200m (Benson, 1978). A second section, further north, from a quarry west of Mari (WD 268444), contains a shallower assemblage (Pavlakelli, op. cit.). Both sections contain mixtures of autochthonous deeper water fauna and displaced,
Fig. 9.5 - Sedimentological logs, cross-section a-a' (see Fig. 9.3 for section location and Endl. B for key to logs).

Fig. 9.6 - Sedimentological logs, cross-section b-b' (see Fig. 9.3 for section location and Endl. B for key to logs).

Fig. 9.7 - Sedimentological logs, cross-section c-c' (see Fig. 9.3 for section location and Endl. B for key to logs).
shallower water species, and also show rapid upward shallowing, to shallow shelf depths at the top of each section (Pavlakelli, op. cit.).

This facies, like facies A1 of the Nicosia Formation of the Mesaoria basin, represents deposition of fine-grained sediment in open marine conditions. Lack of structure is attributed to bioturbation, and masking by weathering. Micropalaeontological data suggest the basin into which the silts were deposited, initially deepened quite steeply to the south. Shallow water fauna (ostracods and molluscs) were transported into this area.

Subfacies N1b coarse-grained intercalations are interbedded in facies N1b silts in their most northerly exposures (log 14/10/4, Fig. 9.6). These facies include: clast-supported, pebbly conglomerates, with igneous and chalk clasts, set in a shelly, sandy matrix, in beds up to 20cm thick with erosional bases; graded sands up to 60cm thick, with erosive, pebbly, shelly sand bases and fine sand tops; and fine- to occasionally medium-grained sands, in 5-60cm thick, planar beds, which are massive, rarely parallel-laminated, or graded, sometimes with bioturbated silty tops, and loaded bases, and which occasionally contain scattered siltstone intraclasts. Thin siltstone bands (1-2cm) are sometimes intercalated with these sediments.

The facies forms intervals up to 5m thick, which thin down-dip, and pass into facies N1a silts. They are also of probable limited lateral extent, although poor exposure prevents reasonable estimates of lateral dimensions from being made.

Intercalation with marine silts (facies N1a) and shelly content imply a marine depositional environment for these sediments. They are similar to the channelised conglomerate and sand of the Nicosia Formation of the Mesaoria basin (facies B1 and B2, section 3.3.1), interpreted as marine fan-delta deposits, although restriction to channel-shaped bodies was not established in this case. Nevertheless, the sediments are also interpreted as the marine toes of small fan-deltas building into the Mari basin. Sediment supply to the offshore area was probably not as prolific as in the Mesaoria basin, so these lobes are thinner, less channelised and less conglomeratic than those of the Mesaoria.

Sedimentary structures in the facies (e.g. normal grading, parallel lamination, planar to erosive bases, presence of intraclasts)
are again attributed to deposition of high concentration sediment flows, issuing directly from the shoreline during periods of substantial flooding (see facies B1 and B2, section 3.3.1). Some reworking of sediment during storms cannot be ruled out, however.

Facies N2 - sands and sand/silts

This facies comprises a variety of very fine- to fine-grained, moderately to well sorted, yellow-brown sands, which are only poorly consolidated, and intercalated with thin, pale yellow-brown silts. The sediments are largely unfossiliferous, although comminuted shell debris is sometimes present. Bioturbation is rare.

Three subfacies are recognised. Subfacies N2a comprises thin-bedded (1-5cm) sands with occasional intercalated silt bands. Beds have flat tops and bases, and are internally structureless or faintly parallel-laminated, or rarely in thicker beds, cross-laminated. Occasionally, beds contain siltstone intraclasts. The subfacies characteristically occurs in wedge-shaped units, 0.5m-a few metres thick, bound by erosion surfaces (Plate 9.1b). Bedding above each erosion surface is parallel to that surface. The subfacies is characteristically interbedded with facies N3 conglomerates, and the less common sand subfacies b (Plate 9.1b). Subfacies N2b comprises massive sand, which lacks any structure other than occasional faint parallel lamination. It forms wedge-shaped intervals, 5cm-1m thick, similar to subfacies a, confined between erosion surfaces.

These two subfacies are found only in the westernmost outcrop of the Nicosia Formation (see Figs. 9.3 and 9.5), where they overlie facies N1a silts. They pass laterally down-dip into facies N1a silts.

Subfacies N2c is further restricted in occurrence, being found only at the northern end of the westernmost outcrop area of the Nicosia Formation (see Figs. 9.3 and 9.5). It comprises very thin- to thin-bedded sands and silts (Plate,9.1a). Bedding is plane to wavy, or occasionally flaser or lenticular bedding is developed. Symmetrical to mildly asymmetric ripples are present. Preserved ripple cross-lamination is unidirectional and dips east. The subfacies forms units up to 5m thick between intervals of facies N3 conglomerates (Fig. 9.5), and grades laterally and down-dip into subfacies a and b.
Facies N2 is affected by synsedimentary deformation at the northern end of its outcrop area. Synsedimentary normal and reverse faults have offsets of up to km (Plate 9.1b), and the facies is sometimes folded beneath the loaded bases of conglomerate intervals. Dewatering structures occur at one locality (WD 266443).

**Interpretation:** the association with marine facies N1a silts and occasional shelly nature of the facies tentatively point to a coastal to marine depositional environment. Intercalation with facies N3 conglomerates, lack of bioturbation, the presence of synsedimentary deformation and characteristic erosion surfaces suggest a high energy environment. The sediments do not resemble typical high energy coastal/marine deposits, e.g. foreshore/shoreface or shelf storm facies (as described for example by Elliot, 1986a, and Johnson and Baldwin, 1986). High energy is therefore believed to have been associated with a fluvially-dominated fan-delta margin. During periods of substantial flooding, erosion surfaces were generated, and silt layers were ripped up to be reworked as siltstone intraclasts. Parallel-laminated sands were later deposited over erosion surfaces during slightly less energetic events (subfacies a). Only occasionally did bedforms develop and migrate, giving rise to cross-laminated sands. Massive subfacies b sands may represent rapid dumping of sediment, perhaps from rather high concentration flows. Synsedimentary deformation was induced in the sands as facies N3 conglomerates were rapidly dumped on top of them (see following facies).

Subfacies c sediments are interpreted to have been deposited in a shallower setting than subfacies a or b, either in a more shoreward location (e.g. Fig. 9.10), or in a locally shoaled area. This area received fine-grained sediment during flooding, and rippling of the sediment surface occurred. This probably largely took place after flooding, under the influence of east-blowing winds, which generated symmetric to mildly asymmetric ripples. Silt was deposited from suspension following fluvial incursions, draping rippled surfaces, and forming thin beds.

Lack of evidence for wave-reeking in the facies may be a result of the orientation of the Mari basin with respect to the main direction of palaeowave approach. The basin was probably
**Fig. 9.8** - Palaeocurrent data, Nicosia and Vasilikos Formations.

**Fig. 9.9** - Conglomerate clast composition, Nicosia and Vasilikos Formations.

**Fig. 9.10** - Schematic palaeogeographic reconstruction during deposition of the upper Nicosia Formation fan-delta.
semi-circular, and open to the southeast (see section 9.2.2 and Fig. 9.10). Wind and wave approach is largely from the southwest at present, and may also have been from this direction in the past. Ripple cross-stratification in subfacies c, and cross-stratification from younger (?Pleistocene) dune facies of the Vasilikos Formation (facies V1b, section 9.3.2) both support the existence of northeast-to east-blowing winds. If this were the case, the southeast-facing Mari basin would not have received waves with very long range fetch. In the absence of wave-reworking, and strong tidal effects (the Mediterranean is microtidal), the fringes of the fan-delta, represented by this facies, became fluvially-dominated.

**Facies N3 - conglomerates**

This facies comprises 10cm-3m thick units of massive to rarely horizontally-stratified, moderately to well sorted, ungraded or occasionally normally graded, pebble-cobble conglomerate (Plate 9.2a), which typically contains rounded, though not flattened, clasts set in a sandy matrix. Conglomerates are usually clast-supported (although sand-supported beds also occur), and show weak to rarely moderate up-dip clast imbrication. Imbrication directions (Fig. 9.8) suggest conglomerate derivation broadly from the northwest (see also Fig. 9.10). Rarely, shelly debris is present in the matrix. Bedding varies from laterally rather continuous over hundreds of metres, with tabular to mildly channelised geometries (Plate 9.1c), to highly lenticular (Plate 9.1b). Beds sometimes have loaded and deformed bases. Clasts are predominantly chalk in composition, although less well rounded Troodos-derived clasts are also always present (Plate 9.2a; Fig. 9.9). These include a small proportion of gabbroic, sometimes pegmatitic, clasts.

The facies is found at the same locations as facies N2, i.e. the westernmost outcrop of the Nicosia Formation (Figs. 9.3 and 9.5). They are interbedded with facies N2 (Fig. 9.5), and pass laterally down-dip over 1-1.5km into facies N2.

**Interpretation:** like facies N2, this facies is interpreted as the product of fluvial flooding of the coastal to subaqueous fringes of a fluvially-dominated fan-delta.

Conglomerates lack evidence of traction current deposition, e.g.
well developed horizontal or cross-stratification, and mass flow processes may have been largely responsible for their deposition. It is envisaged that heavily sediment-laden currents flowed out across the shoreline during periods of maximum flooding, mixed with waters of the receiving basin, and rapidly dumped their loads. Rapid deposition is suggested by general lack of grading, and stratification, loaded bed bases and synsedimentary deformation to underlying sands.

The conglomerates form thin, well defined, generally tabular bodies, separated by sand intervals. There is no development of intermediate facies, e.g. pebbly sands, or evidence of discrete, river channel mouth bars, commonly associated with fluvially-dominated deltas (e.g. Kleinspehn et al., 1984; cf. the Kakkaristra Formation fan-delta, chapter 4). This may imply that conglomerate deposition normally took place onshore, but that during very severe flooding, or perhaps in response to seismic shocks, pulses of conglomerate were swept across the fan-delta front and rapidly deposited.

Good clast sorting and rounding, and the relatively high percentage of sedimentary clasts in this facies, suggest local derivation from the northwest, where the pre-Pliocene sedimentary cover of the Troodos Massif is exposed (Fig. 9.1). This cover would have included pre-existing Upper Miocene fan-delta deposits, developed around the margins of the earlier Maroni basin (Eaton, 1987). Rounded and sorted clasts may be a result of cannibalisation of these older sediments.

9.2.4 Sedimentological model and basin evolution

Facies close to the base of the Nicosia Formation document the presence of a small, but steep-margined basin, whose depocentre lay in the vicinity of Mari. The basin developed during the dying stages of a phase of compression, which had affected southern Cyprus since the Miocene. The marine toes of fan-deltas are recorded close to the probable northern margin of the basin, while silty facies accumulated to the south.

Ostracod assemblages from the silts suggest the basin shallowed rapidly through the Lower Pliocene, presumably as local deepening and tectonism declined.
Subsequently, another fan-delta prograded into the basin from the northwest (Fig. 9.10). The subaqueous fringes of this fan-delta are preserved. The geographic setting of the southeast-facing basin, protected from probable southwest wave approach, allowed fluvial, not wave, processes to dominate this fringe. Fan-delta sediments were largely derived from the pre-Pliocene sedimentary cover of the Troodos Massif.

Progradation of the fan-delta may reflect final readjustment in the hinterland as compression eventually ceased in southern Cyprus, or may record a separate tectonic pulse.

9.3 Vasilikos Formation

9.3.1 Introduction

The Vasilikos Formation is a newly recognised formation (section 2.2.3), cropping out at only one main locality, south of Mari (Figs. 9.1 and 9.3). It rests unconformably on facies N1a silts of the Nicosia Formation (Plate 9.2c), into which it gently cuts. The base of the formation is thus broadly concave-up.

The formation comprises a 20m thick sequence of unfossiliferous, fluvial sediments with minor dune facies. Its true thickness is not known, because overlying formations and probably its top have been eroded. Its age is unknown, as is its precise correlation with other formations and stratigraphic units in southern Cyprus. It is much thicker than the Fanglomerate and Older River Terrace deposits (section 9.1 and see section 9.4), is cut erosively into the Nicosia Formation (unlike the latter units), and occurs at a lower elevation than these units in the Mari area. It is thus believed to represent an older depositional unit, deposited after the Nicosia Formation, but prior to the Fanglomerate and associated sediments.

The sudden appearance of the Vasilikos Formation, cut into the Nicosia Formation, is attributed to tectonic factors (see section 9.3.3), and may correlate with the appearance of the Kakkaristra Formation in the Mesaoria basin, also attributed to tectonism (section 4.2.2). If this is so, the formation may be ?Upper Pliocene-Lower Pleistocene in age (section 2.3).
9.3.2 Facies and facies relations

Two major lithofacies are recognised in the formation (conglomerates and sands), each of which is split into two subfacies.

Facies V1 - sands

Subfacies V1a, fluvial sands: this subfacies comprises fine, moderately sorted, slightly muddy, yellow-brown, unfossiliferous, poorly consolidated sands. Sedimentary structures have largely been obscured by weathering (e.g. Plate 9.3a). In the lower part of the facies, pebbly sand intercalations form lenses up to 25cm thick, with channelised bases, and are trough cross-laminated (log 3/11/2, Fig. 9.5), or form thinner, irregular lenses. Cross-stratification dips east (Fig. 9.8). In the upper part of the facies, tiny black rootlet holes are sometimes visible. Parallel lamination is occasionally evident throughout the facies, which also contains occasional thin, wavy silt bands.

The subfacies is interbedded with other facies of the Vasilikos Formation (logs 3/11/1 and 3/11/2, Fig. 9.5; Plate 9.2c), and is more common in the eastern part of the outcrop area. It is interpreted, along with the formation's other facies (except V1b), as being of fluvial origin, and was deposited in a sandy and gravelly braided river or alluvial fan setting. Pebby, trough cross-bedded sands are typical of these environments (e.g. Rust, 1978; Collinson, 1986), and represent migration of lunate bedforms along braided channel floors. Horizontally-laminated sands represent deposition across braided bar tops during upper flow regime conditions. Some bar tops were exposed and later vegetated. Thin silts were deposited between flood events, and only occasionally preserved.

Subfacies V1b, dune sands: a 4m thick unit of pervasively high-angle (up to 33°), cross-laminated sands occurs within the Vasilikos Formation (log 3/11/2, Fig. 9.5). These sands are fine-grained, well sorted, and contain minor sand-sized shell debris. Also present are numerous, very thin, calcareous, vertical tubules, up to 10cm long, which may be a form of rhizoconcretion. Cross-stratification dips northeast (Fig. 9.8).

The subfacies is interbedded with, and laterally equivalent to, other facies of the Vasilikos Formation, which are of fluvial origin.
It is interpreted as an aeolian dune facies, deposited on an inactive part of the fluvial system which deposited the rest of the Vasilikos Formation. Winds were from the southwest. Dominant winds along the south central Cyprus coast are from this direction at present. Bioclastic debris in the subfacies suggests the coastline was not far away, and that shelly sediment was reworked along with fluvial sand from exposed, braided bar tops into small dunes. Dunes were apparently vegetated after they became inactive.

**Facies V2 - conglomerates**

This facies comprises pebble-small cobble conglomerates, which are poorly or occasionally moderately sorted, unfossiliferous and contain angular and subangular, or rarely subrounded, clasts, set in a poorly sorted, muddy sand matrix. Clasts are predominantly of igneous composition (Fig. 9.9), a small proportion of which are gabbroic, although sedimentary clasts are also always present. Two subfacies are recognised.

*Subfacies V2a* conglomerates are massive, or contain diffuse, horizontal layers of better sorted sediment. Clast size is mainly up to 15cm in diameter, though outsize clasts, up to 70cm in diameter, occasionally occur (Plate 9.2b). Conglomerate fabric is predominantly clast-supported, imbrication and grading are not present. Conglomerates form lenses and beds (Plate 9.2c), up to 4m in thickness, with mildly to heavily scoured bases. The subfacies directly overlies marine silts of the Nicosia Formation and is interbedded with facies V1a sands (Plate 9.2c; Fig. 9.5).

*Subfacies V2b* conglomerates are characteristically trough cross-bedded, in units up to 1m thick, and tend to be less coarse, and better sorted than subfacies a. Cross-bedding dips east. The subfacies forms a unit ca. 3m thick, interbedded between fluvial and dune sand facies (log 3/11/2, Fig. 9.5; Plate 9.3a). Laterally, it interdigitates with horizons of subfacies a.

These unfossiliferous, rather poorly sorted, conglomerates are interpreted as fluvial deposits. They are similar to type a and b sediments of facies C2 of the Apalos Formation (section 6.2), interpreted as longitudinal and transverse braided bar deposits.
Some massive conglomerates may also have been deposited from sheet flows, some of which were erosive, and capable of carrying very large outsize clasts. These flows moved across and between gravelly, braided bars. The facies resemble other typical braided river deposits, e.g. facies Gm and Gt, described by Miall (1977) and Rust (1978). Gravelly tracts were apparently more persistent in the west of the outcrop area of the Vasilikos Formation (log 3/11/1, Fig. 9.5), and migrated laterally over a sandier area to the east (log 3/11/2, Fig. 9.5).

9.3.3 Sedimentological model and implications

Model

Facies of the Vasilikos Formation were deposited in a fluvial environment. Lack of overbank muds suggests a relatively proximal, braided river or alluvial fan setting. The incised nature of the formation (section 9.3.1) perhaps favours the former environment. Sediments were derived from the west, and transported east by stream and sheet flows. A series of braided channels and gravelly and sandy bars developed. Inactive fluvial tracts were sometimes covered by small aeolian dunes, which later became vegetated, along with exposed sandy bar tops. A fluvial setting not far from the coast is suggested by the presence of bioclastic material in dune facies.

Tectonic and eustatic implications

Incision and deposition of the formation can be attributed to tectonic and/or eustatic factors. Possible correlation with sediments of the Mesaoria basin suggests that tectonic factors were at least partly responsible. The Vasilikos Formation is younger than the Nicosia Formation, but older than the Fanglomerate (Fig. 9.2). The Kakkaristra and Apalos Formations occupy this same stratigraphic position in the Mesaoria basin (see Table 2.2). Deposition of the Kakkaristra Formation is attributed to a pulse of uplift in the Troodos Massif, which generated a partial unconformity and led to progradation of the Kakkaristra Formation fan-delta into the Mesaoria basin (section 4.2.2). It seems reasonable to infer that this pulse of
uplift, which marked the start of a compressional phase in Cyprus recorded by both the Troodos Massif and the Kyrenia lineament (section 5.4.2), may also be documented in southern Cyprus, by the Vasilikos Formation. As in the Mesaoria basin, eustatic factors may have enhanced or hindered the effects of uplift, depending on whether sea level was falling or rising. If the correlation is correct, it implies that the Vasilikos Formation is Upper Pliocene–Pleistocene in age.

The extent of the Vasilikos depositional system is unknown. Although only one braided fluvial tract is apparently preserved, it seems likely that others would have developed in response to uplift, but have since been removed by erosion. The formation is similar in this respect to the Fanglomerate (next section), which is poorly preserved in southern Cyprus when compared with the Fanglomerate of the Mesaoria basin (see further, section 9.4.3).

**Composition**

Conglomerates of both the Vasilikos and Nicosia Formations contain gabbroic clasts, unlike equivalent formations in the Mesaoria basin (Nicosia and Kakkaristra), which contain none, or very few (see Fig. 8.3a). This clearly reflects derivation of the igneous component of the Nicosia and Vasilikos formations in southern Cyprus from the Limassol Forest block of the Troodos Massif, which contains a large proportion of plutonic lithologies (see section 1.3.1 and Fig. 1.5). The Vasilikos Formation is richer in Troodos-derived sediment than the older fan-delta at the top of the Nicosia Formation (Fig. 9.9), exposed to the north. This probably reflects derivation of the Nicosia fan-delta largely from pre-Pliocene sediments to the northwest of its location (section 9.2.3). The Vasilikos Formation, however, was derived from the west. An eastward-extending "arm" of the Limassol Forest block is present in this area (Fig. 9.1). It was presumably uplifted along the Yerasa fold and thrust belt in the Upper Miocene, and may well have been exposed by the Upper Pliocene.

9.4 Fanglomerate and Older River Terrace Deposits
9.4.1 Introduction

Coarse, alluvial deposits (the Fanglomerate) overlie the Kakkaristra and Apalos Formations in the Mesaoria basin north of Troodos (chapter 7). These sediments represent a major phase of uplift of both the Troodos Massif and the Kyrenia lineament in northern Cyprus (see section 7.3.2). In view of the relative magnitude of this uplift pulse, Fanglomerate facies might be expected to have been shed to the south of the Troodos Massif as well as to the north. Fanglomerate has been mapped in the south, by Bagnall (1960) and Pantazis (1967), but at only a few isolated localities. Work by this author supports the general lack of preservation of the Fanglomerate in southern Cyprus. Some areas around Mari and Maroni, however, previously mapped as a 120' raised beach, are covered by fluvial facies, more akin to the Fanglomerate than beach sediments. Like the Fanglomerate, they occur in topographically elevated positions compared with other Quaternary fluvial sediments, and may well be laterally equivalent to the Fanglomerate. As this correlation is only tentative, however, these sediments are termed Older River Terrace deposits here (Figs. 9.2 and 9.3).

9.4.2 Facies and facies relations

Fanglomerate mapped in the east of the study area by Bagnall (1960), north of Tersephanou (Fig. 9.1), comprises up to 3m of poorly sorted, non-stratified, clast-supported, pebble to cobble conglomerate. Clasts are angular and predominantly of chalk. Matrix is pinkish, sandy silt, probably also largely of non-Troodos origin. Conglomerates pass laterally into pinkish, sandy silts with occasional thin conglomerate stringers. These sediments are partly altered, probably as a result of recent caliche processes. Fanglomerate facies unconformably overlie Lefkara Formation chalks (of pre-Pliocene age).

Older River Terrace deposits in the Mari and Maroni area (Fig. 9.3) are more igneous-rich. They mainly comprise poorly stratified, clast-supported, pebble-cobble conglomerate, with angular to occasionally subrounded clasts, set in a poorly sorted sandy matrix. Rarely, matrix-supported horizons are present. Imbrication is occasionally developed (Plate 9.3b), particularly in better stratified
intervals. The few imbrication directions recorded are scattered, but suggest currents from between west and north. Clast size is generally less than 12cm, but outsize clasts, up to 50cm in diameter, sometimes occur. Rarely, trough cross-bedding is developed. Conglomerates are up to 2m thick, with flat or only very mildly erosive bases (Fig. 9.6). Sandy intervals, up to 50cm thick, are intercalated with conglomeratic sediments. Structures within these sands are destroyed by recent caliche development. All facies are unfossiliferous.

Interpretation: the Fanglomerate and Older River Terrace deposits of south Cyprus are fluvial sediments, deposited unconformably, largely as a sheet, over older sediments. They are similar to the bulk of Fanglomerate sediments in the Mesaoria basin in this respect (chapter 7). They differ from the latter in containing less matrix-supported conglomerate, and more sand intercalations, perhaps indicative of a slightly more distal alluvial fan setting. General lack of cross-bedded facies, also characteristic of Mesaoria basin Fanglomerate, again suggests dominance of sheet flow processes (see section 7.3.1), perhaps associated with flashy, episodic discharge (see section 7.3.2). As in the Mesaoria basin, some Fanglomeratic sediments are rich in sedimentary clasts, reflecting local derivation from the pre-Pliocene sedimentary cover of the Troodos Massif.

9.4.3 Summary and implications

The Fanglomerate and Older River Terrace deposits are possible lateral equivalents, and both are correlated with the Fanglomerate of the Mesaoria basin. These sediments record a major pulse of uplift of the Troodos Massif, associated with Quaternary compression of the whole of Cyprus (section 7.3.2). Uplift resulted in the outpouring of large volumes of coarse sediment in huge sheets on both sides of the Troodos Massif. Sediments in south central Cyprus apparently represent more distal facies than those examined in the Mesaoria basin. This may reflect a greater distance from the Mount Olympus area of the Troodos Massif (see inset, Fig. 9.1). Uplift of the Troodos Massif at this time may have been focussed on Mount Olympus as a result of large-scale serpentinisation of the plutonic core.
of the ophiolite (see section 7.3, subsection on serpentinisation of the Troodos Massif), and Fanglomerate sediments may have been shed more or less radially from this centre.

The Fanglomerate and associated sediments are much more poorly preserved in south central Cyprus than in the Mesaoria basin. This very likely reflects the differences in geomorphology to north and south of the Troodos Massif. To the north is the flat-lying Mesaoria Plain. Fanglomerate facies have largely been trapped here, and subsequent reworking has only occurred along river courses incised into the Fanglomerate. The southern flank of the Troodos Massif, with its pre-Pliocene sedimentary cover, slopes down virtually to the coast, where only a very narrow coastal plain is developed. Fanglomerate sediments were deposited on this plain, and probably in the foothills, but large quantities may have been transported further, and deposited offshore. Subsequently, as uplift of the ophiolite continued, any Fanglomerate deposits were probably substantially reworked off the steep southern flank of the ophiolite, redeposited in incised valleys, and also again transported offshore. As a result, only minor patches of Fanglomerate are now preserved onshore.

A unit of young sediment, up to 50m thick, is identified from seismic data on the narrow Cyprus shelf, adjacent to the coast (unit D; section 10.4.2). This may contain Fanglomerate and reworked Fanglomerate facies. Quaternary, and also Pliocene, sediments are in general rather thin across the shelf, however, although they are thicker across the adjacent, seaward, irregular slope area, and also on the deeper sea floor (section 10.4.3, sediment package subsection). Both seismic and geological data thus suggest large-scale sediment bypassing of the southern margin of the Troodos Massif and shallow shelf area, with deposition at greater depths. This further reflects the lack of sediment trapping on the south side of the ophiolite, associated with steep gradients as the ophiolite was progressively uplifted.

9.5 Raised Beach and Younger River Terrace Deposits

9.5.1 Introduction
Raised Beach deposits, as mapped by Bagnall (1960) and Pantazis (1967) along the south coast of the study area, were examined during this project, but not studied in great detail. True beach facies were identified in these deposits. In addition, fluvial sediments were recognised, here termed the Younger River Terrace deposits to distinguish them from the Older River Terrace deposits, which occur inland in topographically elevated positions, and which may correlate with the Fanglomerate (see previous section). Younger River Terrace deposits may be in part lateral equivalents to Raised Beach facies, but at several localities, clearly cut into the latter, and therefore post-date them (e.g. log 7, Fig. 9.11).

Fluvial sediments also occur inland at topographically low elevations (e.g. log 2/11/1, Fig. 9.6), but were not investigated in detail. They mainly represent the products of rivers incised into older sediments in post-Fanglomerate times. Their relationship with Younger River Terrace deposits is uncertain. Bagnall (1960) and Pantazis (1967) mapped these facies as Recent sediments.

Two series of Raised Beach deposits, belonging to 40' and 120' raised beaches, were recognised by Bagnall (1960). His 120' raised beach is covered by fluvial facies, however, here assigned to the Older River Terrace deposits (see section 9.4.1). Pantazis (1967) recognised a 40' raised beach, and small outcrops of a 125' and a 250' raised beach. The presence of these latter raised beaches was not confirmed in this study.

Facies of the 40' raised beach occur at a variety of heights above sea level along the coast, ranging from 0' to 40' (see Fig. 9.11). Between the Moni power station and Cape Dolos, for example, the raised beach is clearly warped into an anticlinal surface (logs 1-6, Fig. 9.11). To the east, westward tilting is evident (logs 12-14, Fig. 9.11), an observation also applicable to a drowned wave-cut platform, identified offshore from seismic data (section 10.4.2, unit B subsection). This deformation suggests recent movement along the coast, probably related to young faulting, a topic under investigation by A. Poole (Edinburgh University).

Raised Beach and Younger River Terrace deposits rest unconformably on older sedimentary rocks, chiefly the chalks and marls of the pre-Pliocene Pakhna and Lefkara Formations. East of
Fig. 9.11 - Sedimentological logs, Raised Beach and Younger River Terrace deposits (see Encl. B for key to logs).
Zyyi, however, they overlie yellowish silty sands, which may belong to the Nicosia Formation (logs 9-11, Fig. 9.11). Occurrence of the Nicosia Formation at the coast here correlates with its outcrop area inland, and with the eastern limit of the formation offshore, as identified from seismic data (Fig. 9.1).

The 40' raised beach of Bagnall and Pantazis has been dated as Tyrrhenian (Pantazis, 1966). This age is questionable, however (see section 2.3).

9.5.2 Facies and facies relations

Raised Beach deposits comprise beach conglomerates and sands (facies RB1), and some sediments indicating fluvial inundation of the coastline (facies RB2). Fossil trottoir (facies RB3) has also been identified. These sediments, and the Younger River Terrace deposits, are now described.

*Facies RB1* comprises well sorted, rounded, stratified, shelly granule-pebble conglomerate (sometimes with less well sorted, coarser layers containing small cobbles), and well sorted, fine to medium sands, with pebbly and shelly horizons. The main sedimentary structure in both lithologies is gently seaward-dipping parallel lamination or stratification (Plate 9.4a). Sands are occasionally bioturbated, and sometimes very bioclastic-rich, becoming sandy calcarenites. The sediments overlie the inferred raised beach which sometimes has a very irregular surface. Clefts and crannies are often filled with cobbles and coarse shell debris.

The sediments are interpreted as typical foreshore deposits, which are well sorted and characterised by seaward-dipping stratification (e.g. Elliot, 1986a).

*Facies RB2* comprises rather poorly sorted, occasionally shelly, poorly horizontally stratified, pebble-cobble conglomerates, with subangular to subrounded clasts. Conglomerates may pass up into sands or pebbly sands, with indistinct parallel lamination, which then commonly pass up into pinkish, muddy silts (e.g. log 1, Fig. 9.11). Silts are structureless, include thin layers and lenses of angular conglomerate, and occasionally contain fossil land snails (e.g.
Planorbis sp.).

Although conglomerates contain marine shells, their poor sorting and stratification, when compared with facies RB1, is believed to reflect fluvial inundation of the shoreline. Shelly material and sand, as well as newly introduced sediment, were mixed together during fluvial incursions, and deposited in the shoreline area. Lack of subsequent wave-rewrorking may be due to deposition during a regressive cycle, an inference supported by fining-up of these fluvially-influenced shoreline deposits into silts of continental origin. Similar cycles are documented from raised beaches along the north coast of Cyprus (Baroz, 1979).

**Facies RB3** comprises an unusual rock type, a very fine-grained, pale grey, highly calcareous sediment. It is typically massive and nobby in appearance (Plate 9.4b), and contains sandy patches, occasional pebbles and cobbles, and whitish blebs and tubes, which comprise calcareous algal and serpulid remains. Small heads of *in situ* coral (*Cladocera caespitosa*) are occasionally present. The rock forms layers a few 10s of cm thick or thicker units, comprising the bulk of the Raised Beach deposit (e.g. log 9, Fig. 9.11). In thin section, calcitic tubules and cell-like structures are visible.

The rock is believed to represent fossil trottoir. Trottoir is an organically-generated, calcareous crust, that forms in warm climates on rocky shores in the intertidal to uppermost subtidal zones, as a result of colonisation by vermetid gastropods (Pérès, 1967). Vermetid gastropods have long, uncoiled, calcareous shells, and live attached to a hard substrate. In warmer parts of the Mediterranean (e.g. Israel, Libya and Sicily), these gastropods (e.g. *Vermetus triqueter* and *V. cristatus*, also called *Dendropoma petraeum*), live with calcareous algae and serpulid worms, and cement together to form platform-like areas, up to 10m wide and 10-20cm thick (e.g. Fevret and Sanlaville, 1966; Safriel, 1966). Sand and gravel, washed on to platforms, are incorporated into their fabric. Fossil trottoirs, up to 90cm thick, are recorded from raised beach deposits in Libya (Fevret and Sanlaville, *op. cit.*).

Fossil trottoir deposits are good indicators of former mean sea level, because they are only found in the intertidal to uppermost
subtidal zone. They may also indicate that climatic conditions during their formation were similar to the present eastern and southern Mediterranean (Fevret and Sanlaville, 1966).

Younger River Terrace Deposits were only cursorily examined, and comprise a variety of unfossiliferous, poorly to moderately sorted and stratified, conglomerates (Plate 9.4c), laminated or massive, occasionally current-rippled sands, and pink, continental, muddy silts. Pottery fragments were found in these sediments at one locality (log 11, Fig. 9.11). The facies forms either the entire thickness of the Raised Beach deposit (e.g. log 10, Fig. 9.11), or channels down into beach facies (e.g. logs 7 and 11, Fig. 9.11). The sediments are interpreted as a variety of fluvial facies, deposited either as lateral equivalents to shoreline deposits, or perhaps more commonly, incised at a later date into shoreline sediments. The relationship between these facies and young, fluvial deposits inland is unknown.

9.5.3 Summary and implications

At least one raised beach and associated sediments are present along the south central coast of Cyprus. This raised beach is no longer horizontal, but has been warped and tilted, along with a drowned wave-cut terrace, recognised from seismic data. Raised beach facies include beach conglomerates and sands, fossil trottoir, and fluvially-influenced shoreline sediments. These are interrupted in places by fluvial facies of the Younger River Terrace deposits, which may be partly lateral equivalent to, and partly younger than Raised Beach deposits.

Raised Beach and Younger River Terrace deposits are younger than the Fanglomerate and Older River Terrace deposits, and may have been deposited during the last interglacial (Tyrrenian age), although this is questionable. Along with the adjacent, drowned, wave-cut terrace, they have tectonic, eustatic and climatic implications for the recent geology of Cyprus. A detailed analysis of this was outwith the scope of this project. Baroz (1979) recognised a series of five raised beaches in northern Cyprus, which he equated with continuous positive movement of the Kyrenia Range
through the Quaternary. The interplay between tectonics and eustatic sea level changes was not discussed, however, and the relationship between these two factors during the Quaternary remains to be unravelled.

9.6 Comparison of the North and South Troodos Flanks

In the introduction to this thesis (section 1.1), it was suggested that the Troodos Massif may have acted as an erosional point source during uplift. The Troodos crustal block is believed to be relatively small, perhaps only 75 x 150 km² in areal extent (see section 11.2.1, influence of pre-existing crustal structure subsection). Similar sedimentary facies could thus perhaps have been expected to have been deposited around the margins of the ophiolite block, if clastic sediment had been shed radially.

This has not been the case, however, until relatively recently, because the structural evolution of different margins of the Troodos Massif has been very variable. Robertson (1977) documented differing early evolutions of various segments of the ophiolite. Follows and Robertson (in press) show variations in carbonate and evaporite development along the north Troodos flank and in southeast Cyprus, in Mid-Upper Miocene times. And this study, and that of Ward in western Cyprus (Ward and Robertson, 1987), record the variable Pliocene evolution of the northern, southern and western margins of the Troodos Massif (summarised in chapter 11).

Some similarities are present in the Plio-Pleistocene sediments to north and south of the Troodos Massif, investigated in this study. The lower Nicosia Formation in both the Mesaoria and Mari basins largely comprises fossiliferous, calcareous, grey-green silts. These facies record the region-wide return of marine conditions to the Mediterranean following the Messinian salinity crisis (sections 3.2.2 and 9.2.4). From tectonic considerations, however, these sediments were deposited in different structural settings. On the north side of the Troodos Massif, a half-graben was subsiding in the Lower Pliocene (section 3.4.2). To the south, however, a phase of compression was coming to an end (section 9.2.2).

Only in the Upper Pliocene is it likely that the structural and
depositional histories of the north and south Troodos flanks began to merge. The fan-deltaic Kakkaristra Formation (Mesaoria basin) and the braided river deposits of the Vasilikos Formation (Mari basin) may record the same pulse of uplift of the ophiolite (section 9.2.4). Their facies are different, but only because the Vasilikos Formation probably represents a more proximal environment. Fan-delta (or braid delta) deposits may have accumulated in more distal areas in southern Cyprus, but have not been preserved.

Pleistocene Fanglomerate facies are recorded throughout Cyprus, both north and south of the Troodos Massif (chapter 7 and section 9.4), to the west (Cyprus Geological Survey Department geological map of Cyprus), and in northern Cyprus in the Kyrenia Range (section 7.4). Poor preservation of Fanglomerate sediments in south central Cyprus prevents a detailed comparison with those of the Mesaoria basin from being made. Both sets of facies, however, apparently record the outpouring of large quantities of very coarse, angular conglomerate in the Pleistocene in huge alluvial sheets, with little braided bar or channel development (sections 7.3 and 9.4.3). They document drastic uplift of the Troodos Massif, and imply that the ophiolite was by this time acting as an effective, erosional point source, shedding sediment north and south, and giving rise to similar depositional facies to both north and south. An ongoing study of later Quaternary sediments in Cyprus may establish if younger river terrace and raised beach deposits are also correlatable around the margins of the Troodos Massif.

It is thus apparent that the geological evolutions of different margins of the relatively small, Troodos crustal block have been very variable. In this case, this further increases the geological complexity of the East Mediterranean, an area which is already known to be the highly complex.

9.7 Summary - Plio-Pleistocene Evolution in South Central Cyprus

1. Prior to the Pliocene, a number of WNW-ESE trending, thrust-controlled lineaments were active in south central Cyprus. Sediments were deposited in small basins between these structural highs.
Movement along the lineaments largely ended by the Tortonian. Activity along the Ayia Mavri lineament, however, lingered into the very Early Pliocene, during the dying stages of compression.

As a result, a small basin, with very localised depocentre located ahead of this lineament, evolved in the Early Pliocene. It was bordered by land to the northwest (Limassol Forest) and north (Troodos Massif), and was open to the southeast.

Marine silts of the Nicosia Formation were deposited in the basin in the Early Pliocene, as open seas returned to the Mediterranean following the Messinian salinity crisis. Faunal evidence suggests depths were in excess of 200m in deeper parts of the basin. Small fan-deltas built southwards across the margin of the basin in the north.

The basin shallowed rapidly, however, as deepening associated with movement along the Ayia Mavri lineament swiftly came to an end.

Towards the Mid Pliocene, a fan-delta prograded into the basin from the northwest, recording either final adjustment of the Limassol Forest block of the Troodos Massif in response to compression, or a separate tectonic pulse. Because of the morphology of the basin in relation to the main direction of palaeowave approach, the subaqueous fringes of the fan-delta were protected from wave-reworking, and show considerable fluvial influence.

The later Pliocene evolution of the area is unknown, because Mid-Upper Pliocene sediments are largely missing in southern Cyprus.

The Nicosia Formation is unconformably overlain by the Vasilikos Formation, of possible Upper Pliocene-Pleistocene age. The formation comprises a series of braided river deposits, incised into the underlying Nicosia Formation. Their deposition is attributed to rejuvenation of the Limassol Forest block, which may correlate with the pulse of uplift of the Troodos Massif, which generated the Kakkaristra Formation in the Mesaoria basin.

Very major uplift affected the Troodos Massif and Kyrenia lineament in the Pleistocene, resulting in deposition of the very
coarse, alluvial Fanglomerate in the Mesaoria basin. Patches of Fanglomerate are also mapped on the southern flank of the ophiolite, and the Older River Terrace deposits recognised in this study probably correlate with this Fanglomerate.

10. Fanglomerate and Older River Terrace deposits are very poorly preserved in southern Cyprus (as are the Vasilikos and Nicosia Formations). This reflects reworking and removal of sediment from the steep, sloping, south flank of the Troodos Massif, with probable redeposition offshore. In contrast, north of Troodos, Plio-Quaternary facies are much better preserved in the flat-lying Mesaoria Plain, which acted as a sediment trap.

11. A raised beach and associated sediments crop out along the south central Cyprus coast. The raised beach surface has been tilted and warped. Development of the raised beach and an adjacent, drowned, wave-cut terrace offshore, deposition of overlying sediments, and subsequent distortion, reflect the interplay between recent eustatic and tectonic effects in Cyprus. Analysis of these was outwith the scope of this project.
a) Facies N2c sand/silts, showing thin, parallel - bedded, parallel-laminated sands with some wavy bedding and ripple cross-lamination; vertically, field of view is ca. 15cm

b) Facies N2a, thin, parallel-bedded sands, and thin horizons of intercalated facies N3 conglomerate; conglomerates and sands occur in wedge-shaped units; in centre left, a small synsedimentary fault distorts and offsets strata; massive facies N2b sands are exposed at top left; vertically, field of view is ca. 5m

c) Interbedded facies N2a thin-bedded sands, and facies N3 conglomerates; conglomerates form tabular to mildly channelised units; cliff face is ca. 10m high
Plate 9.2

a) Facies N3 conglomerate, showing horizontal stratification and a high proportion of subrounded, pale chalk clasts; the conglomerate is moderately to well sorted, and clast-supported; staff is 110cm long.

b) Facies V2a massive conglomerate, showing poor sorting, presence of angular to subrounded igneous (dark) and sedimentary (pale) clasts, and a huge, outsize clast; staff is 110cm long.

c) General view of cliff face, ca. 30cm high, at the Vasilikos cement works, showing structureless facies N1a silts of the Nicosia Formation, unconformably overlain by interbedded facies V2a massive conglomerates and facies V1a fluvial sands of the Vasilikos Formation.
Plate 9.3

a) Slightly pebbly, facies V1a fluvial sands, overlain by facies V2b cross-bedded conglomerates, facies V2a massive conglomerates, and further facies V1a fluvial sands; facies V2b dune sands are exposed at the top; cliff face is ca. 10m high

b) Massive, poorly sorted, imbricated conglomerate of the Older River Terrace deposits; staff is 110cm long

c) Raised Beach deposits, overlying slumped chalks of the Lefkara Formation, near Petounda Point
Plate 9.4

a) Parallel-stratified, coarse sands and fine conglomerates of facies RB1, overlying an irregular surface cut into chalk of the Lefkara Formation; the coarsest conglomerate directly overlies this surface; vertically, field of view is ca. 1.5m

b) Typical, nobbly, massive appearance of trottoir facies RB3; lens filter is 5cm in diameter

c) Poorly sorted, sandy conglomerates and pebbly sands of the Younger River Terrace deposits exposed in shoreline cliff; modern beach gravel is present at the foot of the cliff; staff is 110cm long
CHAPTER 10 - SEISMIC STUDY

10.1 Introduction

Evaluation of the tectonic setting of Cyprus is of prime importance when assessing the evolution of the island. The island presently lies within the complex Alpine-Himalayan orogenic belt (section 1.2, Fig. 1.2), and a major plate boundary (Cyprus arc) is located to the south of it. Convergence has probably been occurring along this boundary since Mid Tertiary times, but there is limited agreement on the exact location of the plate boundary, or on the type of motion taking place (see Fig. 1.4).

A wealth of seismic data exists in the East Mediterranean. In order to shed further light on the nature of the Cyprus arc, and its influence on the geology of Cyprus, some of these seismic data (to the south of the island) were reviewed. In addition, two sets of previously unpublished seismic data, located close to the south central coast of Cyprus, were made available to this project. They comprise several profiles of a multichannel seismic survey carried out by the Shell Internationale Petroleum Matschappij in 1971, covering the irregular slope area south of south central Cyprus (Fig. 10.1), and a shallow penetration, high resolution, single channel survey, carried out by the (then) Institute of Geological Sciences in 1978 (Roberts et al., 1978), covering the very narrow shelf adjacent to south central Cyprus (Fig. 10.1).

It was possible to use the high resolution IGS seismic survey for direct correlation with the onshore geology of southern Cyprus, and then to correlate this with the Shell data (although there is not direct overlap of the two surveys). In this way, the structure of the south Cyprus continental margin and the sedimentary processes operating on it were examined. Correlation with regional seismic data was then undertaken and some important conclusions were reached regarding the location, structure and evolution of the Cyprus arc.

The methods used and results of the seismic study are now presented. Their implications for the Plio-Pleistocene evolution of Cyprus, which is the main concern of this thesis, are also outlined,
Fig. 10.1 - East Mediterranean bathymetry and locations of seismic surveys; contours are at 500m intervals.
although they are more fully discussed in chapter 11 (basin synthesis and regional tectonic implications), along with the other major results of the project.

10.2 Data base

The IGS, shallow penetration, high resolution seismic data comprise a series of 32 dip and 5 strike seismic lines (Fig. 10.2b). They were recorded using single channel receiving equipment (Huntec hydrophone) and a 300J EG and G boomer sound source. Maximum depth of penetration is approximately 60ms below the sea floor. No processing of data was carried out, other than bandpass filtering. The IGS also carried out limited sampling of the sea bottom using a vibrocorer.

7 dip lines and 1 strike line were used from the Shell seismic survey (Fig. 10.2a). All data were acquired using 24-fold multichannel seismic profiling equipment, with 3 x 300cu.in. air gun sound sources, and a 52m shot interval. The data were subject to standard processing, though no migration was carried out.

Regional seismic data from around Cyprus (Fig. 10.4) were obtained from the Geophysical Survey Report on the East Mediterranean (Cambridge University, 1976), and other published work (Stride et al., 1977; Woodside, 1977; Biju-Duval et al., 1978; Kempler and Ben-Avraham, 1987).

Bathymetric charts for the East Mediterranean were available from the Cambridge University report, and from the Israeli Geological Survey.

10.3 Methods

All seismic lines were examined to identify consistent, mappable reflectors and seismostratigraphic units. A seismic basement and three overlying sedimentary units were identified from the IGS data. Depth contour maps of the tops of each unit were constructed by converting two way travel times into depths, using sonic velocities of 1500m/s for water and 1700m/s for shallow, uncompacted sediments. Isopach maps for the three sedimentary units were
Fig. 10.2a – Location of the Shell seismic survey. Geology from the Cyprus Geological Survey geological map of Cyprus. Stratigraphy is shown in inset.

Fig. 10.2b – Location of the IGS seismic survey. Geology from Bagnall (1960) & Pantazis (1967), tectonic lineaments from Eaton (1987) & Cleintaur et al. (1977).
contoured in two way time.

Where dipping reflectors were present (in seismic units A and D), apparent dips were calculated by correcting for vertical exaggeration on seismic profiles. Where apparent dips were available at, or close to, the intersection of crossing dip and strike lines, approximate true dips were determined using stereographic projections.

One consistent, regional reflector was identified from the Shell data. This reflector, and the thickness of sediment overlying it, were mapped.

Once mapping was complete, cross-sections were constructed from both sets of seismically-derived maps, and extended onshore using south Cyprus topographic and geological maps. Cross-sections were then used to facilitate correlation between on- and offshore data.

10.4 Results and Interpretation

10.4.1 Bathymetry

The bathymetric map (Fig. 10.1) reveals a steep and irregular south margin to Cyprus. In the study area, this margin can be divided into two parts.

Firstly, there is a very narrow, shallow portion, not clearly visible on Fig. 10.1, but identified from seismic data (see Encl. 10.3, and cross-section on Encl. 10.9). This area, termed the shelf for simplicity, extends for 7 or 8km offshore, and down to a depth of ca. 120m. It is covered by the IGS seismic survey.

Secondly, there is an adjacent, wider, irregular, outer slope area, largely covered by the Shell seismic survey (Fig. 10.1). This outer slope extends down to depths greater than 2000m, which are the deepest found in the East Mediterranean. The slope exhibits a prominent ridge, which extends eastwards from the south end of the Akrotiri Peninsula (Fig. 10.2a). The Hecataeus Plateau lies at the eastern end of this ridge (Fig. 10.1).

A major trough, up to 2600m deep, is present to the south of the outer slope. It is bathymetrically rather irregular, however, and exhibits small highs and lows, particularly in the west (see Fig.
The trench curves around a prominent seamount to the south, the Eratosthenes Seamount, and then rapidly shallows to the east. West of Cyprus, it becomes much less clearly defined.

Bathymetric features further away from Cyprus are also important, but are described later, in section 10.4.5 (regional data).

10.4.2 IGS data

Several seismic units were recognised from the IGS survey. They are summarised in Table 10.1 and illustrated in Encls. 10.1 and 10.2. The IGS interpretation of the data (Roberts et al., 1978) is very much in line with the results presented here. The main differences are that seismic unit B was not recognised by Roberts et al., and that seismic unit C was correlated with the Miocene Pakhna Formation, not Pliocene sediments.

Unit A

The first unit identified, unit A, represents seismic basement. It is often not well penetrated seismically, but where it is, is seen to comprise mainly parallel reflectors, which are sometimes deformed, and which are overlain by an unconformity surface and undeformed strata (Table 10.1, Encls. 10.1 and 10.2; see Fig. 10.2b for seismic line locations). Basement reflectors mainly dip south, though dip north on Encl. 10.1 due to folding. Seismic basement thus constitutes a well bedded, sometimes deformed, succession, capped by an unconformity.

It is correlated with the Lefkara and Pakhna Formations (section 1.3.3; inset in Fig. 10.2a), which crop out along the south Cyprus coast (Fig. 10.2b). They comprise bedded chalk, chert, marl and calciturbidites, which typically dip south between 10° and 20° (Bagnall, 1960; Pantazis, 1967; Eaton, 1987), and are capped by the Miocene-Pliocene unconformity surface. The interpretation is supported by the one vibrocorer sample to penetrate seismic unit A, which recovered chalk (Roberts et al., 1978). The thin and impersistent evaporitic Kalavasos Formation was not specifically identified.

Folding of the Pakhna and Lefkara Formations occurs onshore, and is concentrated along WNW-ESE trending lineaments (Eaton,
Table 10.1 - Description of seismic units identified from IGS seismic data

<table>
<thead>
<tr>
<th>Unit</th>
<th>Lower contact</th>
<th>Upper contact</th>
<th>Internal character</th>
<th>Distribution &amp; correlation</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>N/A</td>
<td>generally a strong, smooth reflector, with considerable relief; weakens where overlain by C; structures within A truncated at this surface</td>
<td>often little visible due to lack of penetration; dipping parallel, sometimes folded, reflectors clearly present on some lines, however</td>
<td>present throughout survey area; correlated with Pakhna and Lefkara Formations</td>
</tr>
<tr>
<td>B</td>
<td>strong to weak, smooth to wavy reflector</td>
<td>generally a strong, smooth to slightly hummocky reflector</td>
<td>mainly moderate amplitude, discontinuous, subhorizontal reflections; minor continuous and hummocky reflections too</td>
<td>thin, discontinuous unit, present only in centre and east; correlated with raised beach deposits</td>
</tr>
<tr>
<td>C</td>
<td>rather weak reflector between high amplitude unit D and low amplitude unit A</td>
<td>strong, smooth reflector, when overlain by unit D</td>
<td>high amplitude, subhorizontal discontinuous to semi-continuous, occasionally hummocky, reflections</td>
<td>present only in west of survey area, correlated with Plio. Nicosia Formation</td>
</tr>
<tr>
<td>D</td>
<td>moderate to strong, smooth to irregular reflection over unit A</td>
<td>N/A</td>
<td>low to moderate amplitude, moderate to widely spaced, parallel, subhoriz. to dipping reflections; excellent to moderate continuity; downlap and onlap on to underlying units</td>
<td>present only in east of survey area, in 3 lobes; correlated with recent sediments</td>
</tr>
</tbody>
</table>
1987). The most important of these, the Yerasa fold and thrust belt, plunges southeastwards below the surface, to the west of the IGS survey area, near Cape Dolos (Figs. 10.2a and b). Folding on IGS line 3 (Fig. 10.2b and Encl. 10.1) may be associated with the subsurface extension of this lineament. The wavelength of the seismically observed folds (350-500m) compares well with the wavelength of folds associated with the lineament onshore (Fig. 7.2 of Eaton, 1987). The most marked folding is seen on IGS line 18. This line is located just south of Petounda Point, where another structural lineament may run (Cleintaur et al., 1977; Fig. 10.2b).

The top of unit A (Encl. 10.4) dips gently over 5-6km before dipping more steeply offshore. The surface is rather smooth in the west, but in the centre and east of the study area, exhibits a number of small highs and lows. Flattish areas with few contours are also evident along this eastern part of the shelf. Cross-sections reveal that these areas are narrow and backed by steeper slopes closer to shore (Fig. 10.3, cross-sections 1 and 2), and they resemble wave-cut terraces. Fossil wave-cut terraces, which are familiarly elevated to form raised beaches, are common along Mediterranean coasts, and reflect Quaternary sea level fluctuations and/or neotectonic movements. Small exposures of at least one raised beach occur along the south Cyprus coast (Bagnall, 1960; Pantazis, 1967; section 9.5). Drowned wave-cut terraces are also recorded from the Mediterranean, however, e.g. Crete (Peters et al., 1985), and from elsewhere, e.g. the western United States (Mullins et al., 1985). These flat areas are thus interpreted as a drowned, wave-cut terrace.

Flattish areas are also present in the west. These, however, are not bordered by steeper zones on their landward sides, and extend as gentle slopes across much of the shelf (cross-sections 3-5, Fig. 10.3). They form an integral part of the gently dipping, top basement surface, and are not believed to represent wave-cut terraces.

Unit C

Seismic unit C (Table 10.1) overlies seismic basement in the west of the study area (Encl. 10.1), but in the centre and east, basement
Fig. 10.3 - Cross-sections across the Cyprus shelf, illustrating the morphology of the top of seismic basement (unit A), without sediment cover (left column) and with sediment cover (right column). Landwards end of sections is to the left. Sections were constructed from IGS seismic data. See Encl. 10.4 for section locations.
is overlain by unit B and the bulk of unit D (Encl. 10.2, Fig. 10.3). A facies change was initially considered as a possible explanation for this relationship. Careful inspection of the seismic profiles in the central area where the units overlap, however, revealed that unit D overlies, and is therefore younger than, unit C. Also the occurrence of unit C corresponds closely with the onshore outcrop area of the Pliocene Nicosia Formation (Encl. 10.5). This formation is composed of bioturbated, structureless silts at the coast (see section 9.2.3). It is suggested that unit C correlates with the Nicosia Formation and the dense, rather structureless seismic character of the unit reflects the character of the formation exposed onshore.

The absence of unit C to the east may be the result of erosion or non-deposition. Evidence from onshore indicates that a small basin evolved in the Lower Pliocene in south central Cyprus, with depocentre in the west of the area of currently exposed Pliocene sediments (shown in Encl. 10.5; see section 9.2.2). This depocentre was very local in extent, and the Pliocene basin shallowed rapidly to north and east. Thin Pliocene sediments were thus probably deposited in the east, and are believed to have since been completely removed by later erosion. Hence, unit C is no longer present in the east of the offshore study area, and Pliocene sediments are no longer found in the corresponding, adjacent onshore area.

Unit B

Seismic unit B (Encl. 10.2, Table 10.1) is found in only two small areas (Encl. 10.6). It typically overlies the drowned, wave-cut terrace cut into the top of seismic basement (Encl. 10.2 and Fig. 10.3). It is thus interpreted as a thin veneer of beach sediment, deposited soon after the wave-cut terrace was formed. Low-angle cross-bedding, observed in raised beach deposits onshore, may account for its rather incoherent seismic character. Some relief is evident on the terrace surface (Encl. 10.2), which is also characteristic of the raised beaches onshore (section 9.5.2).

The terrace area dips gently seawards. It also dips westward, because unit B occurs at 40–60m below sea level in the east, but at 60–75m below sea level in the centre of the study area. This is also
reflected by raised beach deposits onshore, which are ca. 12m above sea level at Petounda Point, but at about sea level further west towards Zyyi (Encl. 10.6; see further, section 9.5.1).

Unit D

Seismic unit D (Table 10.1, Encl. 10.2) comprises a series of parallel reflectors, draped over the unconformity surface at the top of seismic basement, and also over minor marine terrace deposits (unit B). It is thickest in the east, with only a very thin veneer developed in the west (Encl. 10.7). Several depositional packages are probably present within unit D (e.g. lower part of unit D on line 30, Encl. 10.2), though they were not mapped in detail.

The unit is interpreted as a recent wedge of sediment prograding across the shelf, largely fed by modern rivers draining off the south coast of the island. It is correlatable with the most recent deposits onshore (marine, terrace and floodplain alluvium of Bagnall, 1960), although may perhaps also contain older Quaternary sediments, e.g. Fanglomerate facies (see further, section 9.4.3). Individual sediment packets may reflect the presence of these units, but detailed correlation was not undertaken.

The isochore map of unit D shows two main area of sediment deposition, one in the centre of the study area (comprising two sublobes), and one in the east (Encl. 10.7). These lobes do not correspond with the locations of the principal rivers along the coast (Encl. 10.7). True depositional dips in unit D (Encl. 10.7) reveal ESE to SE transport directions for the central depositional sublobe closest to shore (no data available for the other sublobe), and easterly to southerly directions for the eastern depositional lobe. The sediments of this latter lobe have been deposited in an embayment in the coastline, marked by Cape Kiti at its eastern end (Encl. 10.7). This embayment is also evident in the top of seismic basement (Encl. 10.4). To a lesser extent, an embayment exists west of Petounda Point, and this is also weakly reflected in the top of seismic basement.

It is believed that longshore reworking of relatively small volumes of sediment feeding on to the south Cyprus shelf via intermittent rivers, may explain the sediment distribution and
depositional dips in unit D. Sediments have been reworked eastwards and southwards, and trapped in two embayments on the shelf (eastward longshore drift is occurring at present along the south Cyprus coast; A. Poole, pers. comm.). Thinning in the central depositional lobe, to form the two sublobes, occurs across the edge of the proposed, drowned, wave-cut terrace (see Fig. 10.3, section 2).

Seismic unit D has largely infilled topographic irregularities in the shelf, smoothing its surface, so that this surface now dips rather evenly seaward (see Encl. 10.3, which largely represents the top of unit D, and sections 1-3 of Fig. 10.3). Young sediments on the slope beyond show a similar feature (see further, section 10.4.3, sediment package subsection).

**Erosion surfaces**

In the light of the preceding descriptions, it can be seen that the erosion surface at the top of seismic basement (unit A) is of end Miocene age in the west, where it is overlain by Pliocene unit C. In the centre and east, however, a wave-cut terrace has been cut into the top of seismic basement, which is of probable Quaternary age (see unit A subsection).

These erosive events can be explained as follows. Erosion at the end of the Miocene was associated with the Mediterranean-wide Messinian salinity crisis, a time of sea level fall and evaporite deposition (Hsü et al., 1978; section 1.3.3). An end Miocene unconformity was generated widely through the Mediterranean at this time. Deposition of marine Pliocene sediments then followed, as seas flooded back into the Mediterranean. Sedimentation was largely confined to the western part of the south central Cyprus shelf, however, where a basin with very local depocentre developed (see unit C subsection and section 9.2.2). Only a thin sequence of Pliocene sediment is likely to have been deposited further east, away from this depocentre.

Quaternary sea level fluctuations subsequently modified the shelf. During a period of lowered sea level, the thin Pliocene sediments in the east were eroded, and a wave-cut terrace was cut into the hard, underlying Miocene bedrock. In the west, however,
thick Pliocene sediments were only partially removed. No marine terrace was cut, perhaps because the soft, poorly consolidated character of these sediments made them unsuitable for terrace development. The wave-cut terrace, and a thin veneer of associated sediment, were then drowned, tilted and covered by more recent deposits (unit D).

10.4.3 Shell data

The Shell seismic survey covers the wide slope area beyond the narrow Cyprus shelf (Fig. 10.1). This area extends for ca. 20km down to depths greater than 2000m. Resolution is not as good as on the IGS sections, and only one regional reflector of consistent character was identified.

**Top Messinian reflector**

Only one consistent reflector was picked on all eight Shell seismic lines. By correlation with published seismic profiles south of Cyprus (Woodside, 1977; Biju-Duval et al., 1978) and DSDP holes 375 and 376 west of Cyprus, this reflector is interpreted as the Miocene-Pliocene boundary or top of Messinian evaporites (A reflector of Woodside, 1977; M reflector of Ryan, 1979). The reflector has a markedly hummocky character along the outer part of the continental slope, but is smoother both closer to shore, and also beyond the edge of the slope (Encl. 10.8; see Fig. 10.2a for the location of the line).

The depth map of the top Messinian reflector (Encl. 10.9) exhibits the same ridge (termed the Akrotiri High after Eaton, 1987) as the bathymetric map (Fig. 10.1). A number of other highs and lows occur between this ridge and the coast. Although smaller than the Akrotiri High, they still exhibit up to 500m of relief. Their origin is uncertain. Anticlinal structures are also mapped in the top Messinian surface in the Cilicia basin north of Cyprus (Smith, 1977). These features are smaller than those south of Cyprus, however, are clearly diapiric in character, and are attributed to salt diapirism. In the southeast Mediterranean, Ryan (1979) attributed significant relief in the upper Messinian surface beneath the Nile cone (up to 1km), to erosion during lowered sea level in the
Messinian. As Cyprus lies in the centre of the eastern Mediterranean, at some distance from significant fluvial input, it is questionable as to whether such large structures would be generated in this manner.

A final possibility is that the structures are fault- or fold-controlled. Because of the poor seismic resolution of the data, this is difficult to verify. There is some evidence, however, on line 2811 (Encl. 10.8) that the north side of one of these features may be bound by a high-angle fault.

**Structure beneath the top Messinian reflector**

Very little structure is visible below the top Messinian reflector on the majority of the Shell seismic lines. This is commonly found in the Mediterranean, because of the absorption of seismic energy by deformed Messinian evaporites (e.g. Woodside, 1977). On the southern ends of lines 2811 (Encl. 10.8) and 2812, however, some gently dipping reflectors are present. These reflectors disappear landwards at the same point at which the hummocky nature to the top Messinian reflector begins. The disappearance may be due to a facies change, but may also be due to absorption of seismic signal by mildly deformed evaporites (hence the hummocky surface). By comparison with other seismic profiles from the eastern Mediterranean, the topmost of the package of dipping reflectors may represent the base of Messinian evaporites.

On Shell line 2811 (Encl. 10.8), a faint reflection horizon has been identified beneath the reflectors just described. A complication at the level of this horizon is the presence of multiples from shallower structures. An irregular, probably faulted, surface has been recognised, however, which rises upwards to be shallowest beneath the Akrotiri High. The structure predates much of the Messinian interval, which pinches out against it. The Akrotiri High has previously been interpreted as a thrust-controlled, basement lineament, one of several WNW-ESE trending lineaments in southern Cyprus (Figs. 10.2a and b), active during the Mid-Upper Miocene (Eaton, 1987). Basement rocks, forming part of this lineament, outcrop at the southwestern tip of the Akrotiri Peninsula (Encl. 10.9), and are unconformably overlain by the Pakhna and Nicosia
Formations (Morel, 1960; see inset in Fig. 10.2a for stratigraphy). It is possible that the structure on line 2811 may correlate with the Akrotiri High basement lineament, and the faulted reflection horizon may represent the top of basement rocks. The structure very likely underlies the whole of the Akrotiri High, although is not visible on other seismic sections because of poor data resolution.

**Sediment package**

Pliocene-Recent sediments across the south Cyprus slope area are significantly thicker than on the narrow shelf (up to 1 sec compared with only ca. 50 msec; compare Encl. 10.10 with Encls. 10.5, 6 and 7; see also the cross-section on Encl. 10.9). Their thickness is rather irregular over much of the mapped area, although is more uniform in the south (Encl. 10.10). This distribution reflects infilling of end Miocene topography and subsequent spreading out of sediment to form a thin, continuous unit across the slope area, and on to the deep sea floor below (cross-section, Encl. 10.9). This suggests that relatively little Pliocene-Recent subsidence has occurred along the south Cyprus margin, and Miocene structures have been buried beneath a blanket of underformed, younger sediment. This supports evidence from onshore that movement largely ceased along the Akrotiri High and other tectonic lineaments after the Miocene (see section 9.2.2).

It is thus apparent that the south Cyprus margin has been relatively inactive since the Miocene. Pliocene-Recent sediments have largely bypassed the narrow shelf area, and been deposited as a blanket at greater depths.

**10.4.4 Implications of the Shell and IGS seismic data**

The implications of the Shell and IGS seismic survey data can be summarised under a number of headings as follows.

**Tectonic lineaments**: both Shell and IGS seismic surveys support the presence of a number of WNW-ESE trending, tectonic lineaments in south central Cyprus, which were first identified onshore (Morel, 1960; Cleintaur et al., 1977; Eaton, 1987), and which can now be traced offshore. Deformation along these lineaments is mainly of Mid-Upper Miocene age (Eaton, 1987; although see further, section
9.2.2) and has been attributed to tectonic shoaling of Cyprus in response to subduction (Robertson, in press). Seismic data support a cessation of tectonic activity at the end of the Miocene, as witnessed by the blanketing of Messinian structures with undeformed, younger sediments.

*Top seismic basement erosion surface:* an irregular unconformity is present at the top of seismic basement (unit A or top Messinian reflector). The morphology of the surface is partly attributable to erosion during the Messinian salinity crisis, but is also due to tectonic activity e.g. formation of the Akrotiri High, and on the Cyprus shelf, to later erosion during the Quaternary.

*Pliocene–Quaternary sedimentation:* evidence from IGS seismic data supports the development of a small basin in south central Cyprus in the Pliocene, whose depocentre lay in the west of the area of currently exposed Pliocene sediments. This depocentre was very local, and the absence of Pliocene sediments to the east is attributed to the complete erosion of only a thin Pliocene sequence, deposited in shallower parts of the basin. Shell and IGS survey data show that much Pliocene–Quaternary sediment has bypassed the Cyprus shelf, and has been deposited downslope as a blanket over older structures. This reflects relative inactivity along the south Cyprus margin, following tectonism in the Miocene.

*Wave-cut terraces:* a drowned, wave-cut terrace is recognised along the eastern part of the south central Cyprus shelf. Its age in relation to the raised beach onshore is uncertain. Both dip gently westwards, suggesting very recent tilting along the coast. They reflect the interplay between Quaternary sea level fluctuations and recent tectonics, a subject now under investigation by A. Poole (Edinburgh University). No evidence for sea level fluctuations was recognised on the slope area, beyond the shelf.

10.4.5 *Regional seismic data*

As outline in the Introduction (section 10.1), existing seismic data from the south of Cyprus were reviewed, firstly to shed further light on the nature of the major plate boundary located south of Cyprus, and secondly to relate this to the onshore geology of Cyprus via information gained from the Shell and IGS seismic
surveys. The results of this review are now presented.

The data base for the regional seismic data review was outlined in section 10.2, and is illustrated in Fig. 10.4. The data can be divided into 3 geographical groups: southwest of Cyprus (i.e. southwest of the west coast of Cyprus), south of Cyprus (i.e. south of Cyprus, as far east as the Hecataeus Plateau; Fig. 10.5), and southeast of Cyprus (i.e. southeast and east of the Hecataeus Plateau). The seismic character of each area is described below, before tectonic implications are addressed. Bathymetry is also outlined, as this has a bearing on structural interpretation.

Southwest area

The area southwest and west of Cyprus is bathymetrically rather irregular. A broad swell (the Florence Rise, Fig. 10.5) separates the deep Antalya basin to the north from a low area to the south, which contains a chain of depressions and small highs. This poorly defined trench (Pytheus trench of Anastasakis and Kelling, in press) eventually connects to the west with the Strabo trench. In the east, the trench splits, and has a small northern fork, which separates the most elevated part of the Florence Rise, just west of Cyprus, from a second, lower ridge (Fig. 10.5; outer Cyprus ridge of Stride et al., 1977). The south fork of the Pytheus trench connects eastwards with the well defined Cyprus trench, which passes round the northern side of the Eratothsenes Seamount.

The south fork of the Pytheus trench is associated with a zone of deformation on seismic (Fig. 10.6a; Giermann fault of Kempler and Ben-Avraham, 1987). Multiple diffractions within this zone make structural interpretation very difficult, though Kempler and Ben-Avraham believe the zone to be compressional, and associated with subduction. Earthquake data west of Cyprus (Rotstein and Kafka, 1982; Kempler and Ben-Avraham, op. cit.) support the presence of a subduction zone, although bathmetrically it is not expressed by a well developed trench. The direction of subduction is disputed, being northeast according to some workers (Dewey and Sengör, 1979; Jackson and McKenzie, 1984; Kempler and Ben-Avraham, op. cit.) or northwards according to others (Rotstein and Kafka, op. cit.). In addition, Anastasakis and Kelling (in press) suggest the
Fig. 10.4 - Location map for the regional seismic data examined in this study.
Fig. 10.5 – Detailed bathymetric map (from the Geophysical Survey Report on the East Mediterranean, Cambridge University, 1976; contours are in metres).
Fig. 10.6a - Line drawing of a seismic section from west of the Eratosthenes Seamount (see Fig. 10.8 for location), showing a deformation zone associated with diffractions and disturbed bathymetry. This zone is interpreted as the actual plate boundary in this area by Kempler and Ben-Avraham (1987; their Giermann fault), and correlates with the Pytheus trench of Anastasakis and Kelling (in press). Drawing is from Kempler and Ben-Avraham (op. cit.).

Fig. 10.6b - Seismic section showing the complex northern wall area of the Cyprus arc southwest of Cyprus (see Fig. 10.8 for location). The section shows several ridges with small, intervening lows. The thick sediment cover on the subducting plate is relatively undeformed, and underthrusting is not visible beneath it. M is the probable top of Messinian evaporites. Section is from Stride et al. (1977).
Fig. 10.7a - Seismic section from the steep, faulted northern wall of the Cyprus arc southeast of Cyprus (see Fig. 10.8 for location). The thick sediment cover on the subducting plate is again relatively undeformed (see also Fig. 10.6b), and underthrusting is not easily discernible beneath it. Section is from Woodside (1977).

Fig. 10.7b - Seismic section across the northern margin of the Eratosthenes Seamount (see Fig. 10.8 for location), showing possible underthrusting of the seamount and overlying sediments, ?including the Messinian (M). Ridges and intervening lows are again visible in the steep, northern wall of the trench (see also Fig. 10.6b). Section is from Woodside (1977).
Pytheus trench is a zone of transtensional right lateral strike-slip, with more transpressional regimes at eastern and western ends, marked by the uplifted Florence Rise and Anaximander Mountains respectively.

South area

The only area in the eastern Mediterranean where a trench is bathymetrically well defined is south of Cyprus (Fig. 10.5). It lies between the elevated blocks of Cyprus to the north and the Eratosthenes Seamount to the south. The axis of the trench clearly curves around the north margin of the seamount, and then broadens and shallows further east.

The north wall of the trench is partly defined bathymetrically by the steep continental slope south of Cyprus, which extends eastward along the south side of the Akrotiri High and the Hecataeus Plateau (Fig. 10.5). To the southwest of Cyprus, however, the wall area contains additional ridges, including the outer Cyprus ridge, and a second ridge just to the south, marked by a chain of small highs.

The transition from trench to wall is marked by a steep, faulted zone, comprising a series of narrow highs and small, intervening, empty to partially filled lows (Figs. 10.6b and 10.7a). Some tilting of basin sediments is evident. Sediments on the subducting plate are relatively undeformed. These features are rather similar to some other subduction zone plate margins, e.g. the Makran (see White and Louden, 1982, their Fig. 8).

The faulted zone with small perched basins does not follow the steep, continuous, south Cyprus continental slope. Instead, it is only located on this slope south of the Hecataeus Plateau and Akrotiri High (e.g. Fig. 10.7a). Westwards it passes south to the additional ridges described above (Figs. 10.6b and 10.7b), and the Cyprus trench connects with the Pytheus trench along the southern side of these ridges (Fig. 10.5). It is thus evident that the wall area, immediately north of the Cyprus trench, is rather irregular and characterised by several discontinuous zones of perched sedimentary basins and intervening highs.
Actual underthrusting along the trench is generally not visible because of the thick sedimentary fill in the trench. An exception to this is in the vicinity of the Eratosthenes Seamount, where northward underthrusting of the seamount appears to be occurring (Fig. 10.7b; Anastasakis and Kelling, in press). Sediments in the trench are also noticeably tilted here and have been interpreted to be of Pliocene-Quaternary age (Anastasakis and Kelling, op. cit.).

The location of the transition from trench to north wall, as identified from available seismic data, has been plotted on Fig. 10.8. This line is clearly kinked around the Eratosthenes Seamount. This kinking, together with the general curvature of the trench and the evidence for underthrusting, suggest that the Eratosthenes Seamount is colliding with the Cyprus arc, and distorting it. Block faulting of the seamount is occurring, perhaps to enable it to be subducted as a number of smaller blocks (Kempler and Ben-Avraham, 1987).

The origin of the Eratosthenes Seamount is uncertain. It has been interpreted as an oceanic plateau (Rotstein, 1985; Ben-Avraham and Nur, 1986), and as a microcontinental block (Kempler and Ben-Avraham, 1987). It is associated with a zone of magnetic anomalies, which suggest that its subsurface extent may be greater than its bathymetric expression (Ben-Avraham et al., 1976). Cyprus appears to be underlain by a slab of continental crust at present (Makris et al., 1983). This may represent a subducted portion of a continental Eratosthenes block. If so, a sizeable continental fragment (Olympus microcontinent of Robertson, in press) may be in the process of being subducted.

Southeast area

This area in general shallows eastwards towards the Levantine coast (Fig. 10.5). Two NE-SW trending escarpments cross it. The first connects to the west with the north wall of the Cyprus trench, and becomes increasingly less well defined eastwards. The second is parallel to, and north of, the first. It is marked by the deepish Cyprus basin at its southwestern end, and by another basin just west of Syria at its northeastern end. The escarpments connect onshore to the east with the Baer Bassit overthrust zone, and then to the Dead Sea transform fault and East Anatolian fault via the
Fig. 10.8 - Possible location of the Cyprus arc, constructed from the seismic and bathymetric data examined during this study. Structural elements in Cyprus are also shown. The extension of the plate boundary onshore to the east, and the nature of the lineament extending northeastwards from northern Cyprus are uncertain.
Hatay graben (Sengör and Yilmaz, 1981; Sengör et al., 1985).

The two escarpments are associated with narrow deformation zones, interpreted to include thrusts by Biju-Duval et al. (1978; Fig. 10.9a). They separate blocks covered with relatively undeformed sediments. Disturbance increases to the west, however, as connection with the Cyprus trench is approached (Fig. 10.9b).

A strike-slip plate boundary has been interpreted to separate the African and Anatolian (Turkish) plates in this area by many workers (e.g. Dewey and Sengör, 1979; Jackson and McKenzie, 1984; Dewey et al., 1986). Scarce and scattered seismicity (Rotstein and Kafka, 1982) supports a lack of active subduction, although these authors and Rotstein (1984) still prefer a subduction zone model. Anastasakis and Kelling (in press) regard the strike-slip zone as predominantly transtensional, on the basis of extensional structures onshore in Syria and Turkey, and lack of deformation of blocks between fault zones offshore.

The strike-slip zone east of Cyprus is often shown to swing northwards, and to connect onshore with the south side of the Gulf of Iskenderun (Fig. 10.5; e.g. Sengör et al., 1985; Dewey et al., 1986; Fig. 1.4). Bathymetrically, such a zone has no consistent expression, and must cross the basin west of the Baer Bassit complex.

Some workers recognise two tectonic lineaments east of Cyprus. The first is that just described, i.e. connecting the Cyprus trench with the south side of the Gulf of Iskenderun. The second connects the Kyrenia Range in northern Cyprus with the Misis Mountains of Turkey (Fig. 10.5; e.g. Dewey et al., 1986; Anastasakis and Kelling, in press). The Kyrenia Range can be seen to extend bathymetrically eastwards as a submerged ridge. Thrusting is apparently visible on seismic sections crossing the submerged ridge (Mulder, 1973). The Adana and Iskenderun basins either side of the lineament are believed to be extensional features, with thick Pliocene-Quaternary sedimentary fills (Mulder, op. cit.; Sengör et al., 1985; Dewey et al., 1986). Anastasakis and Kelling (in press) interpret the lineament as a second strike-slip fault zone, with alternating zones of transpression (uplifted Kyrenia Range and Misis Mountains) and transtension (Adana and Iskenderun basins). Furthermore, they extend the lineament west of the Kyrenia Range, through the Antalya
Fig. 10.9a - Seismic section across the two escarpments southeast of Cyprus (see Fig. 10.8 for location). Thrust faulting is interpreted by Biju-Duval et al. (1978). Sediments between the two fault zones are undeformed. Section is from Biju-Duval et al. (op cit).

Fig. 10.9b - Seismic section showing increased deformation between the two escarpments southeast of Cyprus, close to the junction with the Cyprus trench proper (see Fig. 10.8 for location). Section is from Stride et al. (1977).
basin to southern Turkey, thus defining a separate Cyprus block between the African and Anatolian plates in the eastern Mediterranean.

Exposed Miocene-Recent sediments in the Kyrenia lineament, however, show little evidence of strike-slip influence (Robertson and Woodcock, 1986; this study), although possible distortion to Pliocene fault planes during later compression may suggest some lateral as well as vertical movement (section 3.4.1). Furthermore, the post-Mid Miocene histories of the Kyrenia and Misis areas are apparently rather different (compare Robertson and Woodcock, 1986, with Kelling el al., 1987; see also section 11.2.2), and it seems unlikely that a major plate boundary joins them. In addition, palaeomagnetic data do not support rotation of a separate Cyprus block (Clube and Robertson, 1986). The lineament may represent a secondary zone of weakness, however, and the Kyrenia area has perhaps been affected by minor strike-slip, as more important (? although still small) lateral movements were accommodated along structures to the south.

10.4.6 Implications of regional seismic data

West and southwest of Cyprus, the boundary between the African and Turkish plates is believed to be marked by a poorly defined trough, the Pytheus trench. The type of motion occurring along this plate boundary is disputed, although seismicity patterns give a strong indication of north or northeastwards subduction (Fig. 10.8). Subduction is poorly defined bathymetrically and on seismic sections, perhaps because the crust in the East Mediterranean is old and thickened (Makris et al., 1983), thus not easily subducted, and perhaps also because of a slow, sluggish rate of subduction (see further, section 11.2.1).

South of Cyprus, subduction is better defined bathymetrically, and a clearly visible trench is developed. In addition, perched sedimentary basins and intervening structural highs, typical of some subduction plate boundaries, are present along the north wall of the inferred subduction zone, although they form a series of discontinuous ridges. Underthrusting of the Eratosthenes Seamount is also apparently evident.

Evidence from onshore Cyprus and from Shell seismic data
(sections 10.4.2 and 10.4.3) reveal the presence of a number of thrust-controlled lineaments north of the Cyprus trench, in or close to southern Cyprus (Fig. 10.8). Formation of these lineaments is attributed to tectonic shoaling of southern Cyprus in response to subduction during the Miocene (Robertson, in press). Tectonic activity along these lineaments largely ceased at the end of the Miocene. The important implication here is that the hinge of the subduction zone may have migrated south, from close to Cyprus to its present location at the base of the slope area, in approximately Upper Miocene-Lower Pliocene times. Several other, important, tectonic events also took place in Cyprus at this time, including half-graben formation north of Troodos (the Mesaoria basin; section 3.5), and collapse of the Paphos-Polis graben west of Troodos (Ward and Robertson, 1987; Fig. 10.8). In addition, major tectonic readjustment was occurring to the east of Cyprus, in the wake of continental collision in the Bitlis zone (Dewey et al., 1986). The implications of these events are further discussed in section 11.2.2 (Mid-Upper Miocene subsection).

Collision in the Bitlis area undoubtedly influenced plate interactions east of Cyprus, in the area that broadly connects the Bitlis zone with the Cyprus trench. Strike-slip motion probably began in this area at this time, as westward expulsion of Turkey between the Northern and Eastern Anatolian faults commenced (Sengör et al., 1985). A well defined strike-slip plate boundary does not occur in this area at present, however. Instead, it is possible that strike-slip motion is exploiting a series of narrow zones (the two escarpments described in section 10.4.5, southeast area subsection, and the Kyrenia-Misis lineament; Fig. 10.8), which may partly represent the locations of older crustal lineaments. Little evidence of movement is recorded in the area, perhaps because African/Eurasian convergence is now largely being accommodated along other plate boundaries, e.g. the Anatolian and Dead Sea faults and Bitlis zone (see also section 11.2.2, Mid-Upper Miocene subsection).
10.5 Summary of Seismic Data and their Implications

1. A correlation has been made between the onshore geology of south central Cyprus, and offshore seismic data, which cover the narrow, adjacent, shallow shelf and a wider, irregular, slope area.

2. The correlation supports the presence of a number of WNW-ESE trending, tectonic lineaments, in, and close to, southern Cyprus. These structures were active in the Miocene. This activity is attributed to tectonic shoaling of southern Cyprus in response to subduction.

3. The structures have since been buried beneath a blanket of Pliocene-Quaternary sediments. These sediments have infilled topographic irregularities on the south Cyprus continental margin, and then bypassed it to be deposited on the deep sea floor below. Very little Pliocene-Quaternary subsidence has thus occurred along the margin.

4. Tectonic activity associated with subduction is now concentrated south of Cyprus, along the Cyprus trench. It is thus possible that southward migration of the hinge of the subduction zone may have taken place in ca. Upper Miocene-Lower Pliocene times. Other important tectonic events also occurred at this time in, and near, Cyprus.

5. Earthquake data suggest that subduction is occurring west of Cyprus at present. It is not well defined bathymetrically, however, or on seismic sections. East of the island, strike-slip may be taking place. It is possible that small amounts of movement are being taken up along several lineaments.

6. The shallower portions of the south Cyprus continental margin have been modified by Quaternary eustatic effects and/or recent movements of the Cyprus block. A drowned wave-cut terrace has been identified, which, together with a raised beach onshore, has been tilted.
PART III - CONCLUSIONS

CHAPTER 11 - BASIN EVOLUTION AND REGIONAL TECTONIC IMPLICATIONS

Plio-Pleistocene sediments of the Mesaoria and Mari basins have been described in the preceding chapters, and their tectonic implications discussed. The evolution of these basins is now summarised (section 11.1), and placed in a regional context by first outlining the regional tectonic setting of Cyprus (section 11.2.1), and then addressing the Miocene-Quaternary tectonic evolution of the area (section 11.2.2). By considering the implications of this part of the geological history of Cyprus, the younger evolution of the East Mediterranean, in particular the Cyprus arc segment of the Alpine-Himalayan orogenic belt, can be further refined, and new light shed on the understanding of this geologically complex area.

11.1 Plio-Pleistocene Basin Evolution in Cyprus

The evolution of the Mesaoria and Mari basins is summarised in the following as a series of points. Points are referred back to the chapter or section from which they come, so that if clarification is required, reference can be made to the appropriate part of the thesis.

11.1.1 Evolution of the Mesaoria basin

Pre-Pliocene setting

1. Prior to the Pliocene, the Troodos Massif had been uplifted from sea floor depths to form a locally emergent island (Fig. 11.1; section 1.3.3). It supplied only limited clastic material, however, because Upper Miocene sediments around the ophiolite are mainly calcareous and marly. During the Miocene, the north flank of the ophiolite began to be affected by a phase of normal faulting (Fig. 11.1; Follows and Robertson, in press), which divided the flank into a number of fault blocks, each several kilometres wide. Faulting controlled thickness and facies variations in the
Fig. 11.1 - Pre-Pliocene setting of the Troodos Massif: schematic N–S structural cross-section (based on data from this study, Cyprus Geological Survey geological map of Cyprus, Cleintaur et al., 1977, Robertson and Woodcock, 1986, and Eaton, 1987; see Fig. 11.5 for section location and Encl. B for key).
Miocene Pakhna Formation, and continued to influence depositional processes in the Messinian. At this time, evaporites were deposited around the Troodos Massif in response to the Mediterranean-wide Messinian salinity crisis (Hsü et al., 1978; section 1.3.3).

2. The Kyrenia lineament, to the north, had been deeply submerged in the Oligocene-Mid Miocene, along with the Cilicia area further north, as the region was affected by strong crustal extension. A thick turbidite sequence (the Kythrea flysch), derived from the Turkish mainland to the north, was deposited over the lineament as a result (Baroz, 1979; Robertson and Woodcock, 1986; section 1.3.2). By the Upper Miocene, however, the lineament began to rise as growth faults became active along its southern flank (Fig. 11.1). These faults influenced thickness and facies variations in younger parts of the Kythrea flysch and Messinian evaporites.

3. Thus, by the end of the Miocene, growth faulting was affecting the northern margin of the Mesaoria Plain area in Cyprus, while normal faulting, antithetic to growth faults, affected its southern margin. Together, these faults delineated a half-graben structure (Fig. 11.1), defining the geometry of the infant Mesaoria basin. Some deepening had occurred by this stage, but significant subsidence had yet to take place (section 3.4.1).

**Nicosia Formation**

4. At the beginning of the Pliocene, following the ending of the Messinian salinity crisis, seas flooded back into the Mediterranean, and open marine conditions were re-established. In the Mesaoria basin, fully marine silts of the Nicosia Formation were deposited over the end Miocene unconformity surface, which was partially generated during periods of lowered sea level during the earlier salinity crisis (section 3.2.2).

5. Over 900m of silts then accumulated in the northern part of the basin during the Pliocene, attesting to continued subsidence along growth faults in the north, while much thinner sediments were deposited to the south (Fig. 11.2). Little direct clastic input is evident from the north side of the basin, suggesting
S

AYIA MAVRI LINEAMENT
minor thrust zone, still active in L. Plio.

TROODOS MASSIF
low-lying, vegetated island; main sediment source

MESAORIA BASIN
sudsided along growth faults in northern part of basin; relative uplift of southern margin, with influx of coarse clastics, associated with antithetic faulting

KYRENIAS LINEAMENT
uplifted along growth faults, but still submarine

Fig. 11.2 - L.-M. Pliocene setting of the Troodos Massif: schematic N-S structural cross-section (based on data from this study, Cyprus Geological Survey geological map of Cyprus, Cleintaur et al., 1977, Robertson and Woodcock, 1986, and Eaton, 1987; see Fig. 11.5 for section location and Encl. B for key).
the Kyrenia lineament remained largely submerged during Nicosia Formation times.

6. In contrast, coarse-grained sediments are present in the lower Nicosia Formation along the southern side of the basin. Fault-generated breccias and conglomerates document continued faulting along this margin, and channelised, mass flow-dominated, slope fan-delta facies are witness to its steep, narrow character (Fig. 11.2; sections 3.3.2, model 1, and 3.4.2, extensional faulting subsection). Antithetic faulting and relative uplift of the Troodos Massif, which had begun in the Miocene, thus continued along the Troodos margin of the basin in the Lower Pliocene.

7. Fan-delta sediments are largely Troodos-derived, and along with the switch from mainly marl accumulation in Upper Miocene Pakhna Formation times to predominantly silt deposition in the Nicosia Formation, record significant subaerial emergence of the Troodos Massif for the first time (section 3.2.2, emergence of the Troodos Massif subsection).

8. Coarse-grained fan-delta facies give way upwards to sandier facies. Locally-derived, fault-generated sediments are no longer recorded, and together these observations document a decline in fault activity along the south basin margin, in mid Nicosia Formation times (section 3.3.2, model 2). Sandy facies are still channelised, however, suggesting basin margin gradients continued to be steep, perhaps as a result of continued subsidence along growth faults to the north. Glacioeustatic effects may have begun to affect the basin at this time, however, and some incision may have been prompted during periods of lowered sea level.

9. In the upper Nicosia Formation (approximately Upper Pliocene times), the ubiquitous silty facies of the formation pass up into very fine-grained, bioturbated sands. This reflects shallowing of the basin as subsidence finally ceased (section 3.3.2, model 3). By the end of Nicosia Formation times, the Mesaoria basin had largely filled, and evolved into a relatively stable, sandy, shallow marine platform.
Following deposition of the Nicosia Formation, a marked change in sedimentation is recorded in the southern Mesaoria basin, by the incoming of the Kakkaristra Formation. This formation comprises a series of coarse- to fine-grained clastics, which document the progradation of a large fan-delta system (section 4.4.1).

An angular unconformity separates the Nicosia and Kakkaristra Formations along the south margin of the basin. Renewed uplift of the Troodos Massif is thus inferred to have taken place at the beginning of Kakkaristra Formation times, resulting in tilting of the Nicosia Formation, generation of an unconformity and subsequent fan-delta progradation. The effects of uplift were localised, however, because the unconformity disappears towards the basin centre (Fig. 11.3; section 4.2.2).

The Kakkaristra fan-delta does not include major channelised facies, like the slope fan-deltas of the lower Nicosia Formation. This reflects progradation of a shelf-type fan-delta across the gentle margin of the now largely filled Mesaoria basin (section 4.4.1).

The south side of the basin soon emerged, because the thin Kakkaristra Formation is rapidly succeeded by fluvial facies of the Apalos Formation (chapter 6). These facies are characterised by large volumes of overbank mud/siltstones, with little conglomerate or sand. This is believed to reflect a decline in uplift of the Troodos hinterland to the south, peneplanation and subsequent supply of fine-grained sediment (section 6.3.3).

Eustatic sea level fluctuations were undoubtedly occurring during deposition of the ?Upper Pliocene-Pleistocene Kakkaristra and Apalos Formations. Their effects are difficult to separate form those of tectonic origin, in an area receiving a major detrital influx as a result of uplift, however (Miall, 1984; section 4.5.1), and are also difficult to detect in entirely fluvial sequences e.g. the Apalos Formation (section 6.3.4).
Fig. 11.3 - U. Pliocene–L. Pleistocene setting of the Meeaoira basin: schematic N–S structural cross-section (based on data from this study, Cyprus Geological Survey geological map of Cyprus, and Robertson and Woodcock, 1986; see Fig. 11.5 for section location and End B for key). Thicknesses of the Kakkaristra, Apalos and Athalassa Formations are exaggerated.
Athalassa Formation

15. On the north side of the basin, the Nicosia Formation is overlain by the lateral equivalent to the Kakkaristra and Apalos Formations, the Athalassa Formation. In the northeast, this formation documents the southerly migration of a series of shallow marine, bioclastic-rich, sand waves across the shallow Mesaoria basin, under the influence of dominant south-blowing storms (section 5.3.3). Development of shallow marine sand bodies at this time may have been due to enhanced wind activity, associated with climatic variations during Upper Pliocene–Pleistocene glacial fluctuations (sections 5.3.3 and 5.5).

16. Kyrenia-derived clastic material is present in Athalassa facies, suggesting the lineament was beginning to emerge (Fig. 11.3). The high percentage of bioclastic sediment in the formation, however, implies that clastic input was still limited (section 5.3.3, sand supply subsection).

17. In the northwestern part of the basin, in contrast to the northeast, the Athalassa Formation is unconformable over the Nicosia Formation and comprises a thick, silty, occasionally slumped lower member, and a thinner, cross-bedded, bioclastic upper member (section 5.4.1).

18. These facies are believed to record local reactivation of the Ovgos fault, under compression. Thrusting took place along this former growth fault, poorly consolidated Miocene sediments on its northern side were deformed (Kythrea flysch), and the silty, partially slumped, lower Athalassa member was deposited as slight deepening took place ahead of the fault zone to the south (Fig. 11.3; section 5.4.2, initial reactivation of the Kyrenia lineament). Thin conglomerate intervals in the silts may represent minor fan-delta facies, implying local emergence of the western end of the Kyrenia lineament. Complimentary movement to the south at this time gave rise to renewed uplift of the Troodos Massif and progradation of the Kakkaristra fan-delta.

19. Fault reactivation was short-lived, however, as was uplift to the south. The minor foredeep filled, and shallow marine to coastal carbonate sands of the upper Athalassa Formation were deposited. These facies transgressed northwards across the
Ovgos fault zone to be deposited unconformably on deformed Kythrea flysch (Fig. 11.3). Flooding of the west end of the Kyrenia lineament during this relatively stable period is attributed to a eustatic sea level rise (section 5.4.2, sedimentation subsection). The whole Mesaoria basin was not resubmerged as a result of transgression, presumably because sediment supply from the larger Troodos Massif allowed sedimentation to keep pace with the rate of sea level rise.

20. Fault reactivation apparently resulted in a reversal of fault plane dips because former south-dipping growth faults (Figs. 11.1 and 11.2) now crop out at the surface as north-dipping reverse faults (Figs. 11.3 and 11.4). True fault plane geometries are unknown at depth, however, and faults may have been distorted during reactivation. This may have arisen from a variety of factors (see section 3.4.1 and also 11.2.2).

**Fanglomerate and related sediments**

21. The final stage in the evolution of the Mesaoria basin is marked by the sudden and widespread appearance of very coarse, angular, Troodos-derived conglomerate (the Pleistocene Fanglomerate). These conglomerates spread far across the basin (Fig. 11.4), and record a dramatic pulse of uplift of the Troodos Massif (section 7.3.2). Eustatic sea level fall may have also been a contributing factor.

22. On the north side of the basin, sediments younger than the Athalassa Formation crop out within the core of the Kyrenia lineament, and include breccias, screes and lacustrine muds. Their age equivalence with the Fanglomerate is not certain, but the switch from shallow marine Athalassa facies to often coarse continental sediments attests to major uplift of the Kyrenia Range at this time (Fig. 11.4; section 7.4), as well as of the Troodos Massif.

23. Uplift represents the major stage of a phase of compression, which had begun earlier with deposition of the Athalassa and Kakkaristra Formations (point 18), but had then waned (point 19). In the Troodos Massif, uplift may have been enhanced by large-scale serpentinisation of the core of the ophiolite (section
Fig. 11.4 - Pleistocene setting of the Mesaoria basin: schematic N–S structural cross-section (based on data from this study, Cyprus Geological Survey geological map of Cyprus, and Robertson and Woodcock, 1986; see Fig. 11.5 for section location and Encl. B for key). Thicknesses of the Kakkaristra, Apalos and Athalassa Formations, and Fanglomerate, are exaggerated.
7.3.2, serpentinisation of the Troodos Massif subsection; see also section 11.2.2, Upper Pliocene-Pleistocene subsection). In the Kyrenia lineament, deformation, which had first affected the poorly consolidated Kythrea flysch along its southern flank (Fig. 11.3), was now concentrated at its core, where pre-existing, subhorizontal thrust sheets were rotated and uplifted along high-angle faults, to give their present, near vertical, orientations (Fig. 11.4; Robertson and Woodcock, 1986; section 7.4).

24. Only 5-10m of Fanglomerate accumulated before fluvial incision became a major process in the basin. No further sediments were deposited topographically above the Fanglomerate, which is now cut by a series of terraced river courses (section 7.3.1). These terraces and their sediments, along with river terrace and raised beach deposits in the Kyrenia Range, document the interplay between possible continued uplift of Cyprus (including the Mesaoria basin) and eustatic sea level fluctuations (section 7.3.2, later incision subsection). These events were not studied during this project.

Summary

In broad terms, the Plio-Pleistocene sediments of the Mesaoria area record the infilling of a narrow basin and episodic uplift of its margins, in extensional, then compressional settings (Table 11.1). Extension began in the Mid Miocene. By the Upper Miocene, a half-graben began to form, but significant subsidence along growth faults in the north did not occur until the Pliocene. Antithetic faulting and relative uplift took place to the south. Extension declined in the Upper Pliocene, and relative stability ensued.

Shortly afterwards, compression began to affect the basin. This was localised, and short-lived initially, however, and a period of temporary stability returned. Severe compression then uplifted both margins of the basin in the Pleistocene and eventually the basin itself. This uplift may be continuing today.
<table>
<thead>
<tr>
<th>AGE</th>
<th>MESAORIA BASIN</th>
<th>MARI BASIN</th>
<th>SEA LEVEL FLUCTUATIONS</th>
<th>HINTERLAND UPLIFT/PENEPLANATION</th>
<th>EVENTS IN CYPRUS</th>
<th>COMMENTS</th>
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<tr>
<td></td>
<td>FORMATION</td>
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<td>FACIES</td>
<td>FORMATION</td>
<td>SUMMARY LOG</td>
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<tr>
<td></td>
<td>FANGLOMERATE</td>
<td></td>
<td>Coarse, angular</td>
<td>FANGLOMERATE,</td>
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<td>Coarse, angular conglomerate sheets</td>
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<td>conglomerate sheets</td>
<td>OLDER RIVER TERR. DEPOSITS</td>
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<tr>
<td>PLEISTO-RECENT</td>
<td>APALOS/UPPER</td>
<td>Mud-rich alluv. fans, minor conglomerate and sand/shallow marine to coastal carbonate sands</td>
<td></td>
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<td>Period of relative stability as compression and uplift wane.</td>
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<td>ATHALASSA FM.</td>
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<td></td>
<td>KAKKARISTA</td>
<td>Shelf fan-delta/shallow marine sand bodies/minor fan-delta intercalations</td>
<td>?VASILIKOS FM.</td>
<td>Braided fluvial deposits</td>
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<td>LOWER ATHALASSA</td>
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<tr>
<td>U. PLIO.-L.</td>
<td>NICOSIA FM.</td>
<td>Marine silts; conglomeratic slope fan-delta intervals along south basin margin, passing up into sandy intercalations</td>
<td>Nicosia FM.</td>
<td>Marine silts overlain by fan-delta facies</td>
<td>Period of relative stability</td>
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<tr>
<td>PLEISTO</td>
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<td>L.-U. PLIO.</td>
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<td>MIO.</td>
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<td></td>
<td>PAKHNA/KALAVASOS</td>
<td>Marl, chalky debris, evaps.</td>
<td>KALAVASOS FM.</td>
<td>Evaporites</td>
<td></td>
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</table>

**NOT TO SCALE**

1. Solid line - sea level fluctuations definitely affecting basins. Dashed line - fluctuations possibly affecting basins
2. Schematic depiction of phases of hinterland uplift (widening column) and hinterland peneplanation (narrowing column).
3. C - compression  E - extension
4. TM - Troodos Massif
5. KL - Kyrenia lineament
11.1.2 Evolution of the Mari basin

*Pre-Pliocene setting*

1. Prior to the Pliocene, compression had been affecting south central Cyprus. This compression resulted in the formation of a series of narrow, ESE-WNW trending, thrust-controlled lineaments (Fig. 11.5; section 9.2.2), which are traceable offshore on seismic data (sections 10.4.2, unit A subsection and 10.4.3, subsection on structure beneath the top Messinian reflector). Marly sediments of the Pakhna Formation and Messinian evaporites were deposited in basins between these lineaments (Fig. 11.1; Eaton, 1987; section 9.2.2).

2. Deformation ended along the most prominent of these lineaments, the Yerasa fold and thrust belt and the Akrotiri High, in the Upper Miocene, but continued into the topmost Miocene (Messinian) along the smaller Ayia Mavri lineament, as witnessed by local disruption and syntectonic sedimentation (section 9.2.2).

*Nicosia and Vasilikos Formations*

3. In south central Cyprus, as elsewhere in Cyprus, fully marine, Lower Pliocene silts of the Nicosia Formation overlie an unconformity surface at the top of the Miocene. The unconformity was largely generated during the Mediterranean-wide salinity crisis (Hsü et al., 1978; section 9.2.2).

4. These silts locally record moderately deep basin conditions (>200m), in the Mari area. This is attributed to subsidence associated with continued tectonic activity along the Ayia Mavri lineament (Fig. 11.2). Absence of Pliocene sediments further east, a feature also evident along the narrow Cyprus continental shelf, from offshore seismic data (section 10.4.2, unit C subsection), is believed to reflect shallowing of the basin, deposition of only thin Pliocene sediments and subsequent complete erosion. A basin with a very local depocentre (the Mari basin) is thus inferred to have formed (section 9.2.2).

5. Offshore seismic data support the notion that activity along the majority of thrust-controlled lineaments ceased by the end of the Miocene, because they are covered by undeformed
Pliocene-Quaternary sediments. These sediments form a blanket over the top Miocene unconformity surface, implying that very little Pliocene-Quaternary subsidence has occurred along the south Cyprus continental margin (section 10.4.4, Pliocene-Quaternary sedimentation subsection).

6. Thus, very local subsidence in the Lower Pliocene, along only the Ayia Mavri lineament, documents the final stage of compression in southern Cyprus, which had begun in the Miocene. This subsidence itself was short-lived, because silt facies close to the base of the Nicosia Formation show shallowing-upward trends (section 9.2.2).

7. Silt facies pass up into coarse-grained, fan-deltaic sediments, still of probable Lower Pliocene age. Fan-delta facies prograded southeastwards into the basin, and either document final adjustment of the Troodos Massif, in association with the end of compression, or record a separate tectonic pulse (section 9.2.4).

8. Mid-Upper Pliocene sediments are probably missing in south central Cyprus. The Nicosia Formation is unconformably succeeded by the Vasilikos Formation (at one locality). This formation comprises braided fluvial facies, derived from the west, and records further uplift of the Troodos Massif (Limassol Forest block; section 9.3.3). It is possible that this pulse of tectonism correlates with the phase of uplift of the Troodos Massif, which generated the Kakkaristra Formation in the Mesaoria basin. This would imply that the Vasilikos Formation is Upper Pliocene-Pleistocene in age. Interpretation of the formation is uncertain, however, because of its very restricted exposure, and unfossiliferous nature.

Fanglomerate and later sediments

9. Coarse, alluvial conglomerates (the Older River Terrace deposits) overlie the Nicosia Formation at other localities in south central Cyprus. These sediments are very similar to the Fanglomerate of the Mesaoria basin, and are correlated with it, and with the few, isolated exposures of Fanglomerate itself mapped in southern Cyprus (section 9.4.1).

10. These sediments record the same drastic pulse of uplift of the
Troodos Massif as the Fanglomerate of the Mesaoria basin (section 9.4.3). Their poor preservation in southern Cyprus is attributed to their location on the steepish southeast flank of the Troodos Massif, probable reworking off this flank, and subsequent deposition offshore.

11. Raised beach deposits crop out above at least one wave-cut terrace along the coast of south central Cyprus (section 9.5.1). They are cut in places by fluvial sediments (Younger River Terrace deposits; section 9.5). Another, drowned, wave-cut terrace is identified offshore from seismic data (section 10.4.2, unit B subsection). These terraces and sediments record the interplay between possible further tectonic uplift of Cyprus, and eustatic sea level fluctuations (section 9.5.3; cf. point 25, previous section).

Summary

Lower Pliocene sediments in the Mari basin record the dying stages of an important compressional phase in southern Cyprus, which began in the Miocene. Local subsidence was short-lived, and the basin shallowed rapidly.

The later evolution of the area (post Lower Pliocene) is difficult to evaluate because of lack of or poor exposure. Two pulses of uplift of the Troodos Massif are recorded by the Vasilikos Formation and the Fanglomerate and associated facies. These may correlate with uplift recorded by the Kakkaristra Formation and the Fanglomerate in the Mesaoria basin (Table 11.1), associated with Upper Pliocene-Pleistocene compression of both the Troodos Massif and Kyrenia lineament.

11.2 Regional Tectonic Implications

Having summarised the evolution of the Plio-Pleistocene Mesaoria and Mari basins in the previous sections, their geological histories are now used to further elucidate the Miocene-Quaternary tectonic evolution of the Cyprus area. The regional tectonic framework is first outlined (see also section 1.2).
11.2.1 Regional tectonic framework - subduction-related tectonics

Subduction setting

Uplift of the Troodos Massif, and ultimately of the whole of Cyprus, have been the major features of the Pliocene-Quaternary geological history of the island. Studies of the older parts of the sedimentary cover of the ophiolite indicate that significant uplift did not begin until the Oligocene (Robertson, 1977; section 1.3.3). This followed the ending of the previous, complex, rotational history of the Troodos microplate, and its amalgamation and juxtaposition with continental margin units (Mamonia and Kyrenia terranes; Robertson, in press; sections 1.3.1 and 1.3.2). The switch to uplift is believed to relate to the start of underthrusting of Cyprus by the African plate, which has continued to the present day, at least to the west of Cyprus (Jackson and McKenzie, 1984). Furthermore, strong crustal extension began north of Cyprus at this time (Kyrenia-Cilicia area), and has also been related to underthrusting to the south (Robertson and Woodcock, 1986). On a regional scale, therefore, uplift in Cyprus has been driven by subduction.

Poorly defined subduction zone

Although subduction may have begun in the Oligocene, several arguments suggest that it has never evolved into a fully-fledged, steady-state system:

a) there is no well defined subduction zone at present near Cyprus (see review of the bathymetric and seismic character of the plate margin, section 10.4.5); northward subduction is generally agreed to be taking place south of Cyprus, while northeast subduction (Jackson and McKenzie, 1984; Kempler and Ben-Avraham, 1987), or northward subduction (Rotstein, 1984), or strike-slip motion (Anastasakis and Kelling, in press) are inferred west of Cyprus; east of Cyprus, strike-slip motion is taking place according to most workers (except Rotstein, 1984, who prefers northward subduction), but the location of the plate boundary, particularly how it connects to the east with the East Anatolian fault and Bitlis zone, is uncertain.

b) there is no volcanic arc associated with subduction; Neogene
volcanics do exist in Turkey (Innocenti et al., 1982), but they do not form a chain that parallels the inferred trench, and are over 300km away.

c) palaeomagnetic data from Cyprus (Clube, 1985) indicate that no detectable convergence has taken place between Africa and Cyprus since the Miocene (section 1.2).

d) seismic refraction data (Makris et al., 1983) suggest that the crust south of Cyprus is not easily subductable, normal oceanic crust, but may be of transitional or thinned continental type, with a thick sediment cover; the refraction data also suggest that Cyprus itself may be underlain by continental crust.

These points are most easily reconciled with subduction that has, at best, been sluggish and/or episodic, and Cyprus must be considered to lie in a "prearc" or "incipient forearc" subduction setting (Robertson, in press).

An additional complication is that the Cyprus arc forms part of the complex Alpine-Himalayan orogenic belt, along which several types of plate motion occur (section 1.2). In some areas, one type of motion has evolved into another. Changes of this nature have the potential to induce tectonic readjustment in adjacent parts of the convergence zone. Such effects are documented in the Pliocene-Quaternary evolution of Cyprus.

Influence of pre-existing crustal structure

A strong influence on the tectonic evolution of the area has been the control exerted by pre-existing crustal structure. Compartmentalisation of the crust began soon after Troodos oceanic crust was generated in the Late Cretaceous, when a fragment of this crust (Troodos microplate) became detached, and was rotated (Clube and Robertson, 1986; section 1.3.1). The boundaries of the microplate are believed to lie partly within the confines of the present island of Cyprus. The northern margin is perhaps located beneath the Kyrenia lineament, while the western and southern margins are documented by the complex geology of the Mamonia complex and Moni melange (Fig. 11.5; Clube and Robertson, op. cit.; sections 1.3.1 and 1.3.2). Reactivation of these boundaries has had an important influence on the later evolution of Cyprus.
Fig. 11.5 - Major structural elements and geological units, Cyprus.
The other important crustal unit in Cyprus is the Kyrenia Range (Fig. 11.5), a complex structural lineament (see section 1.3.2). The lineament extends east of Cyprus as a submerged ridge and curves north, apparently to join the Misis complex in southeast Turkey (see Fig. 10.1). The geological history of the Misis complex (Kelling et al., 1987) is apparently rather different to that of the Kyrenia lineament, however, so that although they may form a continuous morphological feature, the two areas have undergone different Neogene-Quaternary structural evolutions.

Other crustal blocks in the Cyprus area include the Eratosthenes Seamount (see Fig. 10.1), interpreted as an oceanic plateau (Rotstein and Ben-Avraham, 1985), or a microcontinental fragment (Kempler and Ben-Avraham, 1987; Robertson, in press; section 10.4.5). The Hecataeus Plateau has also been interpreted as an accreted oceanic block (Ben-Avraham and Nur, 1986).

11.2.2 Miocene-Quaternary evolution of the Cyprus area

Having outlined the subduction-related tectonic setting of Cyprus and its structural framework, the Plio-Pleistocene tectonic evolution of the island and surrounding area is now considered. As events prior to the Pliocene had an influence on this, Miocene tectonics must also be described.

Lower-Mid Miocene

In Lower-Mid Miocene times, an oceanic area still lay between Arabia and Eurasia to the east of Cyprus. Continental collision had not yet occurred in the Bitlis zone, and a continuous, arcuate subduction zone is inferred to have extended through this area, into the East Mediterranean and west of Cyprus (Fig. 11.6a; Dewey and Sengör, 1979; Sengör and Yilmaz, 1981). The Troodos Massif had begun to be uplifted, but was still submerged (Fig. 11.6a). Subsidence was occurring north of it, in the Kyrenia-Cilicia area, where the turbiditic Kythrea flysch was being deposited (point 2, section 11.1.1). This subsidence is attributed to strong crustal extension behind the subduction zone, perhaps linked in some way to southward migration of the subduction hinge ("roll back"; Robertson, in press), a model also proposed for extension in the Aegean and
Figs. 11.6a and b - Plate tectonic reconstructions, Cyprus region.

A) L−M MIO

1. Extension in Kyrenia-Cilicia-Misis area
2. Uplift of Troodos Massif (though ophiolite still submerged)

B) M−U MIO

1. Tectonic shoaling in S Cyprus (S Troodos)
2. ?Evolution of strike-slip between Cyprus and Turkey (single line shown for simplicity)
3. Start of extension in Mesaoria area (N Troodos)
4. Thrusting in Misis area
1. Pronounced extension in Mesaoria area; significant emergence of the Troodos Massif
2. Pronounced extension in Paphos–Polis area
3. Southward migration of subduction zone (shown as a continuous line for simplicity)

1. Severe uplift of Troodos Massif
2. Severe uplift of Kyrenia Lineament along reactivated faults
3. Collision of Eratothenes Seamount with subduction zone
4. Extension in Adana–Iskenderun area
Tyrrhenian Seas (Le Pichon and Angelier, 1979; Malinverno and Ryan, 1988). The zone of subsidence apparently extended eastwards into southeast Turkey, where turbidites of the Karatas Formation, outcropping in the Misis complex, are equated with the Kythrea flysch (Kelling et al., 1987).

Mid–Upper Miocene

Collision occurred in the Bitlis zone in Mid–Upper Miocene times (Sengör and Yilmaz, 1981; Dewey et al., 1986). As a result, the Anatolian plate (Turkey) began to be expelled westwards along the newly formed Anatolian faults (Fig. 11.6b). Dewey and Sengör (1979) also inferred a change in convergence direction from north to northeast west of Cyprus, while subduction probably ended at this time immediately east of Cyprus (Fig. 11.6b). This latter segment of the plate boundary probably began to evolve into a strike-slip zone.

Tectonic shoaling began in southern Cyprus, as the area came under compression (point 1, section 11.1.2). Compression may relate to locking of the subduction zone, as tectonic readjustments were occurring in the collision zone to the east. Shear stresses generated by coupling of the underthrusting and overriding plates locally exploited pre-existing lines of weakness, i.e. the southern margin of the Troodos microplate, and a series of thrust lineaments developed (Fig. 11.6b). These lineaments paralleled the inferred trench, which may have lain closer to southern Cyprus than at present. Deformation was limited to a small area, coinciding with the point of maximum curvature of the plate margin, presumably where shear stresses were concentrated. This area also broadly marks the zone where transition from subduction to strike-slip motion began to take place.

The effects of strike-slip motion between Cyprus and southeast Turkey are not well documented, and even today, activity along this part of the Cyprus arc is of uncertain type and location (section 10.4.5, southeast area subsection). Two fault escarpments connect the Cyprus trench with Syria, and it is suggested that small amounts of movement may be occurring along these (section 10.4.6). Minor strike-slip activity may have affected a third lineament, which
connects the Kyrenia Range in northern Cyprus with the Misis mountains in Turkey (section 10.4.5, southern area subsection). Connection of the strike-slip segment of the Cyprus arc with the Dead Sea and East Anatolian faults to the east is uncertain, hence is shown as questionable on Figs. 11.6b-d.

While the southern margin of the Troodos Massif shoaled, extension began to effect its northern and western flanks. Previous, more regional extension in the Kyrenia-Cilica-MISIS area came to an end as subduction faltered (section 3.4.1), and thrusting, relating to collision further east, is recorded in the Misis complex (Fig. 11.6b; Kelling et al., 1987). Local extension in Cyprus was not a major process yet, however, as it was later to become (following subsection).

Upper Miocene–Pliocene

Tectonic shoaling of south Cyprus began to decline in the Upper Miocene, and by the Messinian–Lower Pliocene, only one minor thrust lineament apparently remained active (point 2, section 11.1.2). Meanwhile, local extension to the west and north of the Troodos Massif became major events. First, the Paphos–Polis graben west of Troodos subsided rapidly in the Messinian (Fig. 11.6c; Ward and Robertson, 1987), and then the Mesaoria half-graben subsided more gently through the Pliocene (point 7, section 11.1.1).

Seismic data also suggest that an important reorganisation of the plate boundary south of Cyprus took place at this time. Miocene structures, associated with earlier tectonic shoaling along the south Cyprus coast, are blanketed by undeformed Pliocene–Quaternary sediments, suggesting very little post–Miocene activity along the south Cyprus continental margin (section 10.4.4, Pliocene–Quaternary sedimentation subsection). Tectonic activity is now located further south, beyond the base of the continental slope, where perched sedimentary basins, with tilted sediment fills, features typical of the inner walls of subduction zones, occur (section 10.4.5, southeast area subsection). The inference is that southward migration of the subduction hinge zone may have taken place in Upper Miocene–Lower Pliocene times (Fig. 11.6c), perhaps as the effects of collision to the east became more pronounced. Readjustment may have partly
fragmented the plate boundary, because perched sedimentary basins do not define a continuous, inner subduction zone wall, but instead occur in a series of discontinuous sections.

Migration of the subduction hinge zone may account for the apparent southward migration in the zone of extension north of the Cyprus arc, from its initial location in the Kyrenia-Cilicia area to the flanks of the Troodos Massif. A type of "roll back" mechanism (Dewey, 1980; Carlson and Melia, 1984) may again have been responsible. In addition, with subduction no longer active between Cyprus and southeast Turkey, extension might also have been induced by sinking of the previously subducted slab along this part of the Cyprus arc (cf. Jackson and McKenzie, 1984, their Fig. 38, and Channel, 1986, his Fig. 7). The localisation and apparent directional divergence in extension (the Paphos-Polis graben trends NNW-SSE, the Mesaoria basin E-W; Fig. 11.6c) can be attributed to control by pre-existing structural trends. The Paphos-Polis graben subsided over the western margin of the Troodos microplate, while the Mesaoria basin formed adjacent to the inferred northern margin of the microplate.

Extension also apparently resumed at this time in southeast Turkey, as the Gulf of Iskenderun and Adana basins began to subside (Sengör and Yilmaz, 1981). Extension was not pronounced initially, however, and some thrust activity continued in the Misis complex in the Pliocene (Mulder, 1973; Kelling et al., 1987).

**Upper Pliocene-Pleistocene**

A major tectonic event occurred in Cyprus in the Upper Pliocene-Pleistocene: drastic uplift of the whole island (points 21 and 22, section 11.1.1, and point 10, section 11.1.2). In contrast, the region to the northeast (Adana-Iskenderun) subsided substantially (Fig. 11.6d) and 2-3km of continental to deltaic, Pliocene-Quaternary sediments now occur below sea level (Mulder, 1973).

These two events are not believed to be linked, and have separate origins. Extension in the Adana-Gulf of Iskenderun has been explained as the result of space problems at the junction of the East Anatolian fault, Dead Sea fault, and the fault which connects to the Cyprus trench to the west. The Arabian plate moves
faster than the African plate, while the Anatolian plate is moving westward away from the others (Fig. 11.6d). A compatibility gap results at the strike-slip fault triple junction, and the Adana-Iskenderun area is extending in response (Sengör et al., 1985; Dewey et al., 1986).

The extensional zone extends towards Cyprus, but drastic uplift of the island is believed to relate to another factor: subduction of a continental fragment south of Cyprus. Underthrusting of the Eratosthenes Seamount along the Cyprus trench is evident on seismic data (section 10.4.5, south area subsection), and deflection of the plate boundary around the seamount is further evidence of collision between the seamount and the subduction zone (Fig. 11.6d; section 10.4.5, south area subsection). The Eratosthenes Seamount is of uncertain origin (oceanic or continental; section 10.4.5). Seismic refraction data (Makris et al., 1983), however, which suggest that Cyprus may be underlain by continental crust, support the notion that the seamount forms a sizeable, partially subducted microcontinent. This subduction may have begun in the Pliocene (section 10.4.5, south area subsection). Uplift in Cyprus was delayed, however, perhaps because subducted continental crust can initially sink with associated oceanic basement, before buoyancy forces cause the continental material to rise (model of Powell, 1986).

Uplift resulted in contrasting styles of deformation in Cyprus. Rigid Troodos basement was apparently uplifted more or less vertically, with little internal disruption, while the Kyrenia lineament to the north suffered extensive deformation. This may be explained in terms of further exploitation of an old crustal boundary during compression, i.e. the northern margin of the Troodos microplate. Concentration of stresses along this boundary deformed the structurally weak Kyrenia lineament, initially by folding and disrupting its poorly consolidated sediment cover (Kythrea flysch; point 18, section 11.1.1), and then by rotating and uplifting the hard, competent, lithologies of the thrust sheets at its core along high-angle reverse faults (point 23, section 11.1.1).

Pleistocene uplift of the Troodos Massif was severe, as witnessed by the deposition of the very coarse Fanglomerate (point 21, section 11.1.1), and an additional factor may have contributed to uplift at
this time - serpentinisation of the plutonic core of the ophiolite. Serpentinisation was originally proposed as a general mechanism for Troodos uplift (Moores and Vine, 1971; Robertson, 1977). It is a process which may occur during or soon after oceanic crust has been formed (Bonatti, 1976). Oxygen isotope data from the serpentinites of exposed ophiolites, however, often differ from oceanic serpentinites, suggesting later stage serpentinisation (Wenner and Taylor, 1971). Oxygen isotope ratios from Troodos serpentinites show $^{18}O$ enrichment, which has been attributed to the involvement of evaporitic fluids during the serpentinisation process (Magaritz and Taylor, 1974). These fluids may have been derived from Messinian evaporites. Messinian evaporites were very likely deposited over the Eratothsenes Seamount. It is possible that they may have been subducted along with the seamount in the Plio-Pleistocene, thus generating a potential source of evaporitic waters for later serpentinisation (Robertson, in press).

In addition, the present dome-shaped morphology of the Troodos mountains may relate to the serpentinisation of the plutonic core of the ophiolite. The core crops out at the apex of the dome (Mount Olympus area, Fig. 11). Petrographic data from the Mesaoria basin show that ultramafic material, derived from the core, is only recorded for the first time in the basin, in the Fanglomerate, implying unroofing of the core for only the first time in the Pleistocene (section 8.3, plutonic core subsection). Although not proven, it is possible that serpentinisation of the Troodos Massif at this time caused uplift of its plutonic core, doming of the massif, unroofing of the core and subsequent erosion and deposition of ultramafic material in Pleistocene facies. This uplift was superimposed on the more general uplift of the ophiolite, associated with microcontinental subduction.

The location of serpentinisation in the Troodos Massif may relate to its original structure, which incorporates three fossil graben spreading centres (Varga and Moores, 1985). Normal faulting associated with oceanic spreading centres can penetrate into ultramafic rocks beneath sheeted dykes. Such faults are mapped in Cyprus (Varga and Moores, op. cit.). They may have acted as local conduits for rising mantle diapirs, which were hydrated by
upward-migrating evaporitic fluids being driven off the subducting slab beneath. A fossilised oceanic transform fault zone (Arakapas fault belt, Fig. 11.1, section 1.3.2) also lies to the south of the plutonic core of the Troodos Massif. It may further have had some influence on localisation of serpentinisation.

**Pleistocene-Recent**

Collision between the Eratosthenes Seamount and the Cyprus arc, which is deflected around the seamount (Fig. 10.8, section 10.4.5, south area subsection), is probably currently continuing. The difficulties of subducting continental material (of which the seamount may be composed), however, may be "choking" the subduction zone, and impeding movement along much of the Cyprus arc, which is relatively inactive seismically at present.

Uplift may be continuing in Cyprus. Difficulties in identifying this arise, however, because the effects of Quaternary sea level fluctuations are superimposed on those of tectonic origin (point 24, section 11.1.1, and point 11, section 11.1.2). This final phase in the evolution of Cyprus has not been investigated in detail, as it is now the subject of a separate study by A. Poole (Edinburgh University).

**Summary**

Evaluation of the Plio-Pleistocene sediments of the Mesaoria and Mari basins, and of accompanying offshore seismic data, have been used to further elucidate the Miocene-Quaternary tectonic evolution of the Cyprus area. This evolution has been dominated by subduction along the Cyprus arc south of Cyprus. Subduction has been sluggish and episodic, however. Together with the knock-on effects of continental collision to the east (Bitlis zone), this has given rise to extensional and compressional tectonic phases in Cyprus, resulting in pulsed uplift of the island (Table 11.1).

**11.2.3 Influence of eustatic sea level effects**

The possible effects of eustatic sea level fluctuations must not be ignored when evaluating sediments of Plio-Pleistocene age. Such effects may have begun to affect basinal processes in Cyprus in
Mid-Upper Pliocene times (see section 3.3.2, model 2). They are difficult to separate from those of tectonic origin, however, in tectonically active areas. Nonetheless, they are believed to have been detected in the Mesaoria basin by:

a) channel incision in a shallow marine setting in the Upper Pliocene upper Nicosia Formation (point 9, section 11.1.1);
b) the onset of shallow marine, sand body deposition in the Mesaoria basin during Athalassa Formation times (point 15, section 11.1.1), and the sedimentary features within these sand bodies (section 5.5);
c) transgression across the western end of the Kyrenia lineament following waning of the initial stages of Upper Pliocene-Pleistocene compression (point 19, section 11.1.1).

Two important factors are reflected by these events. Firstly, they partly occurred during temporary periods of relative tectonic stability (Table 11.1). Secondly, they are recorded by sediments from shallow marine to coastal settings - environments which are most likely to manifest the effects of a fluctuating shoreline.

Pleistocene-Recent eustatic effects were not studied in this project, as they are the subject of a separate investigation.

11.3 Comparable Tectonic Settings

Examples of ancient or modern geological analogues to Cyprus are difficult to locate. Lack of subduction-related volcanics in the East Mediterranean implies that Cyprus lies in a unique "prearc" or "incipient forearc" setting (section 11.2.1). Furthermore, the subducting crust south of Cyprus is not normal oceanic, but is of old, thickened oceanic or thinned continental type, and may include at least one microcontinental block. In addition, unlike many other ophiolites, the Troodos Massif in Cyprus has been uplifted essentially in situ, with little apparent internal deformation. Its sedimentary cover, also largely undeformed, reveals that uplift has been pulsed in nature, as extensional and compressional phases have affected the island.

The Hellenic and Calabrian arcs are other, subduction-related segments of the African/Eurasian convergence zone in the
Mediterranean. The Hellenic arc is morphologically very similar to the Cyprus arc in that it is arcuate, comprises a western section associated with subduction and an eastern section associated with probable strike-slip, and includes an uplifting island at its point of maximum curvature (Crete). Crust south of the arc is not typical oceanic, and extension occurs behind the arc to the north.

Subduction, however, is associated with a well developed volcanic arc, and extension has been taking place continuously since the Upper Miocene in the Aegean, in a back arc setting (Le Pichon and Angelier, 1979). Furthermore, evolution of the Hellenic arc has involved significant rotations of its eastern and western portions (leading to its present curvature; Kissel and Laj, 1988), as well as thickening and imbrication of the sedimentary cover overlying the subducting plate (Mediterranean Ridge; Mascle et al., 1986). Underthrusting of this deformed sedimentary wedge is responsible for the uplift of Crete (Le Pichon, 1982). Further west, subduction along the Calabrian arc has resulted not merely in extension behind the subduction zone, but fully-fledged, active back arc spreading, and generation of new oceanic crust (Kastens et al., 1988). Thus, both the Hellenic and Calabrian arcs are in a more evolved state than the Cyprus arc.

An example of a largely undeformed, uplifted ophiolite occurs in the Costa Rican forearc, landward of the Middle America trench (the Jurassic-Cretaceous Nicoya complex). It is interpreted as an autochthonous, essentially in situ ophiolite, disrupted along only a few discrete zones (Lundberg, 1983). The lower part of the sedimentary cover of the ophiolite comprises thin-bedded trench slope turbidites and mudstone on its seaward side, while thick turbidites are found on its arcward side. Shelf deposits overlying these turbidites document rapid uplift of the ophiolite. Like the present Cyprus arc, no sedimentary accretionary prism is developed along this part of the Middle America trench, and uplift of the Nicoya complex has been attributed to accretion of incoming bathymetric highs (?cf. the Eratosthenes Seamount), although assumed to be of oceanic rather that continental origin (Lundberg, op. cit.).
The Nicoya complex lies in an intraoceanic setting, however, where oceanic crust of the Cocos plate is being subducted beneath the essentially oceanic Carribean plate. In addition, it lies in a true forearc setting, between a trench and well developed volcanic arc. Furthermore, interpretation of the complex as an in situ, uplifted ophiolite has been questioned, and it may alternatively represent slices of oceanic crust, accreted from the subducting plate, and uplifted during accretion of further oceanic slices (Karig, 1982).

The Zambales ophiolite in the Philippines is another relatively undeformed, uplifted ophiolite. Its sedimentary cover also documents the infilling of a basin along its arcward flank, as the ophiolite was progressively uplifted and tilted (Schweller et al., 1984). This ophiolite has a complex tectonic history, however, involving rotation and northward drift along strike-slip faults, resulting in its accretion along with a number of other allochthonous terranes, in the Philippines archipelago (Karig, 1983).

The Plio-Pleistocene geology of Cyprus thus has few, if any, geological counterparts, attesting to its unique "prearc" or "incipient forearc" setting, in a long-lived and complex convergence zone.
CHAPTER 12 - CONCLUSIONS

1. The Neogene-Quaternary geological evolution of Cyprus has been dominated by uplift of the Troodos Massif. Uplift has not been uniform, however, but pulsed in nature as extensional and compressional tectonic phases have affected the island. Furthermore, although the Troodos block is relatively small, the structural and depositional histories of different margins of the uplifting block have been variable. The post-Miocene stages of uplift are well documented by the sediments of two, small, Plio-Pleistocene sedimentary basins, the Mesaoria basin to north of the Troodos Massif, and the Mari basin to south.

2. The E-W trending Mesaoria basin initially formed in the Upper Miocene, in an extensional setting, when a half-graben began to develop. Significant subsidence then occurred in the Pliocene along growth faults in the northern part of the basin, while relative uplift of the Troodos margin to the south took place. Small, slope fan-deltas prograded into the basin as a result, as the remainder of the basin filled with fine-grained, marine facies.

3. In south Cyprus, Lower Pliocene sediments in the Mari basin record the dying stages of a phase of compression, which had begun in the Mid Miocene. Marine silts accumulated in a small deepened area, ahead of a thrust-controlled tectonic lineament. Subsidence was short-lived, however, and silts rapidly shallow up into fan-delta facies.

4. Uplift of the Troodos Massif and extension in the Mesaoria basin declined towards the Upper Pliocene, and the basin shallowed to become a sandy platform. The Mid-Upper Pliocene evolution of the Mari basin is unknown, because sediments of this age are missing.

5. Renewed uplift is then recorded as both the Troodos Massif and the Kyrenia lineament, on the north side of the Mesaoria basin, were raised, this time in a compressional setting. Compression initially caused local reactivation of former growth faults as high-angle thrusts along the northern margin of the basin, and slight deepening took place ahead of the fault zone. Silts and
minor fan-delta facies accumulated in the deepened area. To the south, complimentary movement uplifted the Troodos Massif, and a large shelf fan-delta prograded into the Mesaoria basin, while braided fluvial facies in the Mari basin may record this pulse of uplift south of Troodos.

6. Uplift declined again, however, as mud-rich fluvial facies overlying fan-delta sediments in the southern Mesaoria basin document peneplanation of the Troodos hinterland, and consequent supply of fine-grained sediment. Tectonic quiescence is also recorded on the north side of the basin, where undeformed, shallow marine facies transgressed across earlier deformed rocks.

7. Major compression in the Pleistocene then drastically uplifted both the Troodos Massif and Kyrenia lineament, and the coarsest facies present in Neogene-Quaternary sediments in Cyprus were shed in a number of huge, alluvial, fanglomerate sheets. Uplift may be continuing today.

8. Uplift can in general be related to convergence of Africa and Eurasia along a plate boundary (Cyprus arc) south of Cyprus. Subduction has not been steady-state, however, but has varied in rate and direction, and evolved into strike-slip motion in places. This is largely related to a) the presence of not easily subducted, old, thickened oceanic or thinned continental crust in the subducting plate, and b) the influence of continental collision in the convergence zone east of Cyprus (Bitlis zone).

9. Miocene—very Early Pliocene compression in southern Cyprus is attributed to locking of the subduction zone, which at that time probably lay closer to Cyprus, as collision occurred in the Bitlis zone to the east. At the same time, strike-slip motion probably began to replace subduction immediately east of Cyprus.

10. As the effects of collision became more pronounced, further tectonic readjustment took place and the Cyprus arc apparently migrated south to its present location, in Upper Miocene—Lower Pliocene times. This migration may have been the cause of extension behind the arc, which resulted in the formation of the Mesaoria basin half-graben, in a type of "roll back" process.

11. Later compression of the whole of Cyprus is attributed to
collision with the subduction zone of a microcontinental block, the Eratosthenes Seamount. Uplift of the Troodos Massif at this time may also have been enhanced by serpentinisation of its plutonic core, triggered by underthrusting of the seamount.

12. Localisation of tectonic events in Cyprus has been greatly influenced by pre-existing, underlying crustal structures, in particular the boundaries of the former Troodos microplate. Miocene tectonic shoaling in southern Cyprus exploited parts of its southern margin, Pliocene extension took place adjacent to its northern margin (and also across its western margin, where the Paphos-Polis graben is located), while compression in the Upper Pliocene-Pleistocene again exploited the northern boundary of the microplate, where the adjacent, structurally weak, Kyrenia lineament was severely deformed.

13. Eustatic sea level fluctuations were occurring through much of the Plio-Pleistocene evolution of Cyprus. Their effects are only detectable, however, when they can be separated from those of tectonic origin. They are, therefore, documented only during temporary periods of relative stability, where they are recorded by shallow marine to coastal sediments. Facies from these environments are the most susceptible to the consequences of fluctuating sea level.

14. From a sedimentological point of view, the Mesaoria basin contains a number of facies which have been little described in the past. These include a) the channelised, mass flow-dominated, shallow marine toes of the slope fan-deltas of the lower Nicosia Formation; b) the conglomeratic, cross-bedded, river mouth bars and associated sediments of the wave- and fluvially-influenced, shelf fan-delta of the Kakkaristra Formation; and c) the shallow marine sand bodies of the Athalassa Formation, whose initiation, geometry and sedimentary features may have been influenced by glacioeustatic effects.

15. Evaluation of the Plio-Pleistocene sediments of the Mesaoria and Mari basins in Cyprus, and of accompanying offshore seismic data, have further refined the understanding of the geologically complex East Mediterranean region around Cyprus. The Neogene-Quaternary evolution of the island has few, if any,
geological counterparts, attesting to its unique "prearc" or "incipient forearc" setting, in a long-lived and complicated convergence zone.
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APPENDIX 1 - PALAEONTOLOGY

MACROFAUNA

The following invertebrate macrofauna have been identified from Plio-Pleistocene sediments in Cyprus. The list is not exhaustive because detailed faunal analysis of these sediments was outside the scope of this project. Letters to the right of each name indicate the formation(s) from which each species was recorded (Nicosia, N, Kakkaristra, K, or Apalos, A, Formations). Material from the highly bioclastic Athalassa Formation was not studied, because of its fragmented nature. Additional faunal lists are given in the memoirs of the Cyprus Geological Survey Department (Wilson, 1959; Bear, 1960; Gass, 1960; Bagnall, 1960; Pantazis, 1967).

Bivalves

Acanthocardia echinata K
   " tuberculata N
Amusium cristatum N
Anadara ?diluvii K
Aquipecten opercularis N
Arca tetragona N
Arctica islandica N
Barbatia barbatia N
?Callista sp. N
?Cerastoderma edule/glaucum K
Chama Gryphoides N
   " sp. N,K
Chamalea gallina N,K
Chlamys varia N
Corbula sp. N
?Crassostrea sp. N
Digitaria digitaria N
Diluvarca corbuloides N
Divaricella divaricata N
Donax sp. N
Glycymeris glycymerisa N,K
Glycymeris violacescens N
Glossus humanus N
?Laevicardium oblongum K
Lucinoma borealis N
Lutraria lutraria N,K
   " angustior N
Macoma sp. N
Modiolus sp. N
Nucula nucleus N
Ostrea edulis N,K
Parvicardium sp. N,K,A
Pecten jacobus N,K
Pinna sp. N
Solen marginatus N
Spisula elliptica N
Tellina (Fabula) fabulina N
   " sp. N,?K
?Thyasira sp. N
Venecardia antiquata N
   " sp. N,K
Venerupis sp. N
Venus casina N,K
Gastropods
Aporrhais pespelicani N
Aspa cf. marginata N
Bathymoma sp. N
Bolma sp. N
Cassidaria echinophora
Cerithium ?alucastrum N
Cypraeccasis pseudocrumena N
Fusinus (Murex) longiroster N
Gibbula sp. N
N
Conus ventricosus
(mediterraneus) N

Scaphopods
Antalis sp. N,K
Laevidentalium sp. N

Echinoderms
?Echinocyanthus pusulum N
Schizaster (Ora) canaliferous N

Brachiopods
Terebratalid sp. N

Arthropods
Balanus sp. N,K

Annelids
Ditrupus sp. N,K
Filograna N

Corals
Cladocora caespitosa N,K

Gyrineum marginatum N
Hinia ventriculatum K
" sp. N
Naticarius ?dillwyni N
Neverita ?josephina N
Strombus ?coronatus N
Turritella communis N
" mediterranea N
Truncularia (Murex) trunculus
Detalium -sexigarium N
Seylocidaris affinis N
MICROFAUNA

Microfauna were not extensively studied. Foraminifera were investigated from a limited number of samples, mainly to establish if reworked species were present. Ostracods were only studied from the Kakkaristra Formation (facies D2 muds), to assist with dating and palaeoecological interpretations. The following in situ microfauna were identified. Further micropalaeontological information is given in section 2.3.

Foraminifera

Ammonia beccarii Ath.
" tepida K
" sp. K
Cibicides sp. N
Elphidium crispum Ath,K
" sp. N,K
Globigerina sp. N
Globorotalia sp. N
Patellina sp. N

Ostracods

Cyprideis seminulum
" torasa
Loxoconcha sp.
Aurila sp.

FLORA

Two very well preserved samples of plant material were recovered from the Kakkaristra Formation (facies C1 bay sands and silts). Although not identified specifically, possible close living relatives were suggested by Dr. C. Page (Royal Botanical Gardens, Edinburgh).

Pine cone (Plate 4.6b): similar to cone of Pinus armandii
Leaf fragment : similar to leaves of broad leaf trees, e.g. oak or tulip oak

Pinus armandii is a 5-needle pine, currently found living in rugged, ridge and valley-type terrain in China.
The combination of pine and broadleaf trees, which prefer damper, valley bottom-type conditions, suggests a mountainous, ridge and valley topography in the hinterland, during deposition of the Kakkaristra Formation (C. Page, pers. comm.).
GEOLOGICAL MAP OF THE STUDY AREA

constructed from this study and Duclos, 1985

SCALE 1:50,000

KEY

- YOUNG SEDIMENTS
- PLEISTOCENE-RECENT
- CLAY COMMERATE
- APALOS FM.
- ATHALASSA FM.
- KAKARISTRA FM.
- NICOSIA FM.
- PRE-PLEISTOCENE & TROODOS MASSIF
- PRE-PLIO-PIEISTOCENE SEDS. & TROODOS MASSIF
- MAIN ROADS
- MAIN RIVERS
- GEOLOGICAL BOUNDARY
- FAULTS

note that in some areas, not mapped in detail, undifferentiated Young sediments may be present.
Encl. 10.3
BATHYMETRIC MAP
constructed from BGS data
Contour interval: 10m

5km
Encl. 10.5
ISOCHORE MAP
of
SEISMIC UNIT C
constructed from BGS data
Contour interval: 2.5 millisec.
Encl. 10.6

ISOCHORE MAP
of
SEISMIC UNIT B
constructed from BGS data
Contour interval : 2.5 millisec.
Encl. 10.7

ISOCHORE MAP
of
SEISMIC UNIT D
Constructed from BGS data
Contour interval: 10 millisecc.


direction of sediment dip
areas where unit D is absent

5km
Encl. 10.10
ISOCHORE MAP
of
PLIOCENE-RECENT SEDIMENTS
constructed from Shell data
Contour interval : 200 millisec.

25km
DEPTH CONTOUR MAP
of
TOP OF MESSINIAN
constructed from Shell data
Contour interval: 200m

Encl. 10.9

Surface shown in map

MESSINIAN AKROTIRI HIGH

Surface shown in map
KEY TO SEDIMENTOLOGICAL LOGS, CROSS-SECTIONS AND MAPS

LITHOLOGY & GRAIN SIZE

- Mud
- Silt
- Very fine sand
- Silty fine sand
- muddy medium sand
- Coarse sand
- Poorly sorted, angular/ moderately sorted, rounded conglomerate

OTHER LITHOLOGIES

- Calcarenite (sand with <40% bioclasts)
- Foraminifera-rich sand
- Troodos basement
- Chalk
- Marl
- Evaporites
- Calcarenite

SYMBOLS

- Planar bedding
- Cross-bedding
- Indistinct bedding
- Lenticular bedding
- Parallel lamination
- Tabular cross-lamination
- Trough cross-lamination
- Wavy lamination
- Ripple cross-lamination
- Contorted lamination
- Up-dip conglomerate imbrication
- Down-dip conglomerate imbrication
- Horizontally-aligned conglomerate clasts
- Concretionary layers in poorly cemented sediments
- Microfaults
- Water escape structures
- Caliche horizons
- Desiccation cracks
- Heavy mineral layers
- Vegetated
- Silt layers
- Rootlets
- Burrows
- Bioturbated
- Oyster beds
- Oyster debris
- Shells, broken & unbroken
- Shell moulds, orientated
- Echinoderms
- Ostracods
- Foraminifera
- Pine cone
- Plant debris
- Wood fragments
- Mud clasts
- Pebbly
- Outsize clast
- Barnacles
- Coral debris
- Lens of e.g. conglomerate
- Lens of e.g. mud, with rippled base