THE PALAEOROTATION OF THE TROODOS MICROPLATE

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DECLARATION

I declare that this thesis has been composed by myself, and that the work is my own except where otherwise indicated and duly acknowledged.
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ABSTRACT

The Troodos ophiolite represents one of the best preserved fragments of ocean floor crust that is exposed on land. As the extrusive series of the ophiolite retains a stable normal polarity magnetisation that is consistently directed westward, the ocean crust is considered to have rotated through 90° since its formation in the Late Cretaceous. Well constrained remanence inclinations from interlava sediment horizons support a northward drift of 15° latitude since crustal genesis. A detailed comparison of structurally well constrained magnetic declinations from numerous sampling sites located within the extrusive succession of the ophiolite show that intra-crustal block rotations in the horizontal plane may have been appreciable (perhaps up to 20-25°) along the northern flanks of the ophiolite, while more dramatic rotations of up to 70° have been recorded in the vicinity of major fracture zones like the Arakapas fault belt. As extrusives on either side of this inferred fossil oceanic transform fault retain a westerly directed remanence, block rotations within the lineament are localised and were probably caused by frictional drag within a strike slip system. Palaeomagnetic data sheds light on the possible geometry of the transform offset. To determine the timing of the rotation event over 2000 orientated samples were collected from the in situ Turonian to Recent sedimentary cover to the ophiolite complex. Pelagic chalks immediately overlying the highest extrusive lava units of the ophiolite complex give reliable magnetic directions indicating that at least 60° of rotation occurred before the lower Eocene and that rotation was completed by the end of the lower Eocene. The major part of the rotation probably took place in the late Campanian-early Maastrichtian interval, only 10-15 Ma. after the end of the spreading ridge volcanism at a time when the Troodos ocean floor crust was known to be stranded close to a passive continental margin. Available palaeomagnetic data from mainland Turkey and Africa support rotation of only a small crustal unit with boundaries potentially located close to the preserved ophiolite complex. A review of field data from earlier Mesozoic basin margin units preserved in south-west (Mamonia Complex), south (Moni Melange) and north (Kyrenia Range) indicates that strike-slip played a critical role in their Late Cretaceous-Early Tertiary deformation. Thus, it is conceivable that segments of the major crustal lineaments around which the ocean crust rotated are preserved in these areas. In the preferred palaeorotation model, the older Mesozoic marginal units of south-west and south Cyprus originated along a transform-fault controlled margin of a microcontinent, possibly the eastern margin of the Alanya massif of southern Turkey, with the Kyrenia Range originating as a separate platform unit to the east. Following rifting to form a narrow Red Sea-type basin, little further spreading occurred during the Mesozoic until mid Cretaceous time, when the Troodos ophiolite was created above
an intra-oceanic subduction zone. After trench-margin collision and ophiolite emplacement to the east of Cyprus, subduction continued to the south of the Troodos, developing an anticlockwise torque on crust stranded in a 'fore-arc' setting. The rotation of a Cyprus-sized microplate then occurred about a local rotation pole. Pre-existing transform lineaments preserved in south-west and south Cyprus were exploited early in the rotation history, juxtaposing contrasting basement blocks along transpressional and transtensional fault strands. Following a phase of pervasive crustal compression in northern Cyprus, Troodos crust was strike-slipped against the basement to the Kyrenia platform. Crustal juxtaposition was accompanied by extension-related volcanism and faulting which persisted into Early Eocene time.
CHAPTER ONE

THE TECTONIC SETTING OF THE TROODOS OPHIOLITE IN THE 
EASTERN MEDITERRANEAN

1.1 INTRODUCTION

Ever since the acceptance of a sea floor spreading hypothesis, the Troodos ophiolite complex has played an almost unique role in advancing our understanding of the complex processes that characterise the evolution of ocean ridge systems. It is therefore somewhat ironic that despite an unrelenting steady accumulation of new geological and geophysical data over the last three decades many of the most fundamental questions regarding its genesis and subsequent tectonic history have been left unanswered. In this respect, one particular outstanding problem centres on the significance of the remarkable $90^\circ$ anticlockwise rotation of the Troodos ophiolite in the complex plate tectonic framework of the Eastern Mediterranean. It is this enigmatic rotation and its manifestation in the surrounding areas that provides the theme for this dissertation.
Most researchers now agree that active deformation in the Eastern Mediterranean area is primarily controlled by the mutual interaction of a number of semi-rigid tectonic plates whose boundaries can be effectively delineated by zones of high seismic activity (Mckenzie 1970, 1972). According to such a model one of these plates, the 'Anatolian Block' is moving westwards away from the locus of major crustal thickening and incipient collision of the Eurasian and Arabian plates in eastern Turkey (see fig.1). Consequently, deformation in the cuspatc Cyprian Arc located at the boundary between the African and Anatolian plates is largely in response to the convergent relative motion of these plates in the area of Cyprus.

Unlike the Hellenic arc to the west, where oceanic crust is clearly being consumed beneath the Aegean Sea (Papazachos and Comninakis 1971), no genuine Benioff zone exists beneath Cyprus where seismicity is notoriously sporadic and generally at a shallow level. This apparent absence of a well defined subducting plate is difficult to reconcile with the result of a recent seismic refraction experiment that indicated that crust beneath the Levantine Sea is either of oceanic or attenuated continental type (Makris et al. 1983). If this crust really is oceanic, why exactly is it apparently not being consumed in the same way as comparable crust under the Hellenic arc? It is possible that major oceanic plateaux (for example, the Anaximander and the Eratosthenes Seamounts, see fig.2) are significantly disrupting normal subduction processes as they approach and eventually collide with the active trench (Ben-Avraham et al., 1981; Rotstein and Ben-Avraham, 1984). Thus, a failure of a buoyant, elevated crustal segment to be subducted can effectively lead to a shift in the Benioff zone; a response that will be accentuated if basement at the trench retains a thick sedimentary cover, as crust beneath the Levantine Sea seems to have (Biju-Duval et al. 1977; Biju-Duval et al. 1977). Certainly sediments on the Florence Rise, close to the inferred trench show deformatonal structures consistent with them having decoupled from the underlying basement which is actively being consumed.

Therefore, in a simplistic plate tectonic scheme, Cyprus is located in a 'fore-arc' setting immediately adjacent to a poorly defined trench which can be traced bathymetrically and seismically along an arcuate lineament into Antalya Bay towards the north west of the island (Rotstein and Kafka 1982; Wong et al. 1974). Reflection profiling to the north and north east of Cyprus reveals two northward tilted asymmetric basins, the Antalyan and Cilician basins, that are infilled by a thick sedimentary cover on their northern sides (Woodside et al. 1977). In a recent paper Jackson and Mckenzie (1984) tentatively suggested that the evolution of these basins has been tectonically controlled; perhaps subsiding as a result of normal faulting and stretching as their south and south western boundaries over-ride the Eastern
Fig. 1 Currently used schematic configuration of principal plates in the Eastern Mediterranean (after McKenzie 1970, 1972), showing how the quasi-rigid Anatolian block, bounded by the Northern (NAF) and Eastern (EAF) Anatolian Faults, is being displaced westward away from the locus of incipient continental collision in the Zagros thrust belt. According to Rotstein (1984) the Anatolian block is not behaving as a single coherent crustal unit and is deforming internally as it rotates anticlockwise in concert with the adjacent Arabian plate.

Fig. 2 Outline map of the Eastern Mediterranean showing the 2000m. bathymetric contour and the principal geological and geomorphological features referred to in the text.

Fig. 3 Geological map of Cyprus showing distribution of Mesozoic and Cenozoic successions around the uplifted core of the Troodos ophiolite complex. Apart from the main Massif, Troodos ophiolitic basement also crops out in the Limassol Forest, on the Akamas Peninsula and in the Troulli inlier. Palaeomagnetic sampling sites of Schoenharting and Abrahamsen (see section 1.4.2) and line of vertical sections (fig. 4 a-c) are also indicated.
1.3 A GEOLOGICAL SUBDIVISION OF CYPRUS

The island of Cyprus can effectively be divided into four distinct geological provinces (see fig.3). These are 1) The Troodos Massif and Limassol Forest areas 2) The Mesara plain 3) The Kyrenia Range and 4) The Mamonia Complex and Moni melange.

A brief synopsis of the geological history of each area is presented below.

1.3.1 THE TROODOS OPHIOLITE COMPLEX AND ITS IN SITU SEDIMENTARY COVER

It is now widely believed that the Troodos Ophiolite Complex is a relic fragment of Late Cretaceous (probably Turonian, Blome and Irwin 1985) ocean crust that retains a complete ocean floor lithostratigraphy, characterised by tectonised peridotites, cumulate plutonic gabbros, a sheeted dyke complex and a series of extrusive lavas overlain by deep sea sediments (Moores and Vine 1971, Gass 1980). Each of these successive oceanic crustal layers now crop out concentrically about the focus of Pleistocene and Recent uplift of the Massif centered around Mount Olympos at heart of the Troodos mountains.

The Massif is dominated by an extensive well developed sheeted dyke complex, whose existence provides unequivocal evidence for crustal formation at linear spreading axis (Wilson 1959, Gass 1968, Moores and Vine 1971). Today, dykes trend predominantly north to south in the western and central parts of the Massif but significant deviations away from this orientation are observed in the east and in the vicinity of a major east-west fracture zone known as the Arakapas Fault Belt (Simonian and Gass 1978). A study of the one way chilling directions at dyke margins in different parts of the complex lead Kidd and Cann (1974) to tentatively propose that the present outcrop of the ophiolite lies to the east of the original spreading ridge axis.

On the northern flanks of the Troodos the sheeted dyke complex grades into a thick (1000m.-1500m.) extrusive succession dominated by units of pillowed and massive lava with subordinate volcanic breccias and hyloclastites (Gass 1960, Schmincke et al. 1983). Geochemically this extrusive series has been subdivided into a lower differentiated andesite to dacitic-andesite series (c.1000m.) that resembles evolved island arc tholeiites and an upper picrite to basaltic andesite series (c.500m.) comparable with
contemporary successions erupted close to island arcs in a forearc setting (Robinson et al. 1983). Although there appears to be a distinct compositional break within this extrusive succession there is no strong evidence for either a major metamorphic discontinuity or a widespread structural disconformity between the 'upper' and 'lower' pillow lava series (Gass and Smewing 1973). This evidence together with a general absence of major sedimentary sequences within the lava succession supports a cogenetic of the lava suites close to a single spreading axis, that probably lay above a subducting oceanic slab (Smewing et al. 1975, Desmet 1976). Detailed field studies on the northern flanks of the Massif demonstrate that crustal accretion occurred episodically, punctuated by phases of extensional normal faulting that resulted in the extrusive sequence being tilted into a series of grabens and half-grabens which were then subsequently infilled by deep sea sediments (Verosub and Moores 1981, Robinson et al. 1983, Boyle and Robertson 1984).

Along the southern flank of the Massif a major east-west depression, the Arakapas Fault, separates the main ophiolite Massif from the complex Limassol Forest area to the south. Although it is clear that the Limassol Forest basement represents a fragment of oceanic lithosphere, the presence of a tectonised harzburgite mantle association intruded at high structural level implies substantial thinning of the crust in this area (Simonian and Gass 1978, Murton in press). As oceanic crust is known to thin toward contemporary transform faults, this provides convincing evidence in support of the Arakapas Fault having behaved as a major transform lineament at the time of crustal accretion. Certainly the expected higher geothermal gradients in this setting might account for the more primitive chemistry of the lavas erupted in the vicinity of the fault (depletion in high field strength elements) and the widespread metamorphic imprint on high level crustal rocks (Simonian and Gass 1978, Murton personal communication 1985). Consistent with it having existed as a deep bathymetric depression, the fault belt is characteristically infilled by a thick succession of pillow lavas intercalated with locally derived clastic and metalliferous sediments that unconformably overlie an intensively fractured and brecciated sheeted diabase basement (Simonian 1975, Robertson 1978).

At specific localities around the periphery of the Massif the stratigraphically highest lavas are directly overlain by a continuous succession of ferromanganiferous oxide sediments (umbers), metal enriched clays, radiolarites (Perapedhi Formation, dated as Campanian) and a sequence of Maastrichtian to Lower Miocene pelagic chalks with subordinate cherts (Lefkara Formation, see Robertson and Hudson 1974 for general review). In south west Cyprus the sedimentary succession includes a thick wedge (c.750m.) of bentonic clays and volcaniclastic sandstones (Kannaviou Formation; Lapierre 1975 and Robertson 1977). In contrast to the northern margin of the ophiolite complex where the Palaeogene pelagic chalk succession is stratigraphically
highly condensed (c.50m.) contemporaneous sequences on the southern and eastern flanks are very thick (c.250m.) and are dominated by calciturbidites (Robertson 1977). Facies variations and palaeocurrent measurements have indicated that during the Palaeogene the Troodos basement formed a submarine palaeoslope dipping toward the south west, allowing downslope reworking of pelagic sediments by turbidity currents.

Although the ophiolite complex was probably partially emergent in Lower Miocene times, a pulse of accelerated uplift centred on the Limassol Forest area in the Middle Miocene lead to the deposition of organic rich carbonates and ophiolite-derived clastics in rapidly subsiding basins in southern Cyprus (Pakhna Formation). Miocene uplift culminated in extensive reef development and evaporite deposition in shallow restricted basins. 'Fanglomerates' were shed radially away from the Troodos Complex in Pleistocene times when rapid uplift was centred on the Mount Olympos area of the Troodos mountains (Robertson 1978).

1.3.2 THE MESAORIA PLAIN

Regional gravity (Gass and Masson Smith 1963) and aeromagnetic anomalies (Vine et al. 1973) over Cyprus together with borehole data (Cleintaur et al. 1977) allow the subsurface extent of the Troodos basement beneath the Mesaoria Plain to be traced. Along the northern margin of the Ophiolite complex the pillow lavas dip at shallow angles (10-15°) beneath the Mesaoria Plain to the north (see fig.4a). This is not compatible with the recorded depth to Troodos basement at Nicosia and Lefkoniko (>2500m.) and it appears that the basement beneath the plain is displaced by a series of north down-throwing faults. Based on magnetic anomaly evidence, Aubert and Baroz (1974) suggest the Troodos slab is abruptly truncated by steep faults that cut obliquely across the Kyrenia Range to the north.

Subsurface cores indicate the Mesaoria Plain is underlain by Upper Cretaceous and Lower Tertiary sediments similar to the succession exposed on the northern flanks of the Troodos Massif, but these are buried beneath a thick pile of Oligocene to Recent sediments (Cleintaur et al. 1977). During the Oligocene and the Miocene, coarse submarine fan sediments flooded in from uplifted areas in southern Turkey across the then submerged Kyrenia area to the north (Kythrea Flysch), and it was only in the Late Miocene and Pliocene that rapid subsidence of the Mesaoria coincided with the pulsed uplift and emplacement of the Kyrenia Range along two major east-west reverse faults that today form the frontal thrusts of this mountain range (Kythrea and Ovgos faults, see Baroz 1975, Robertson and Woodcock in press). Episodes of rapid subsidence and transgression in the Neogene were punctuated by periods of emergence and desiccation (Messinian gypsum) before the whole Mesaoria was dramatically uplifted above sea level during the Pleistocene.
Fig. 4a Simplified cross section from the northern Troodos margin to the Kyrenia Range. Troodos basement has been identified in deep wells beneath the Mesaoria, and magnetic data (see chapter 6 fig.3) suggests that this crust terminates abruptly at a high angle fault near the front of the western Kyrenia Range. The present structure of the Range owes much to Late Eocene (D2) and Plio-Pleistocene deformation (D3).

Fig. 4b According to Lapierre (1975) the Mamonia Complex is entirely allochthonous and consists of two independent nappe sheets (a lower volcano-sedimentary nappe represented by the Dhiarizos Group, and an upper sedimentary nappe sheet represented by the Ayios Photios Group) which have been thrust over the Troodos ophiolitic crust and its in situ Campano-Maastrichtian volcanoclastic sedimentary cover from the north east. All tectonic contacts are considered to be low angle with metamorphic rocks being compared with amphibolite and greenschist 'sole' rocks typically found at the base of a nappe pile.

Fig. 4c Interpretative cross section of Swarbrick (1979) shows how slivers of Dhiarizos Group basement are truncated and separated from identical basement segments along high angle serpentinite screens. Incorporated within this zone of high angle faulting are elongate slivers of Troodos-type ophiolitic basement which like the main Troodos ophiolite to the north east retain a complete and undeformed in situ sedimentary cover of umbers, radiolarite and a thick (>100m.) sequence of Campano-Maastrichtian volcanoclastic sandstones. Greenschist and Amphibolite metamorphic rocks occur in close association with sheared serpentinite that characteristically has been intruded up major strike-slip fault lineaments.
FIG. 4

B) LAPIERRE'S MODEL

Metamorphic sole rocks Ayios Photios Gp. sediments
Dhiarizos Group
Upper Nappe Unit

Low angle thrust Troodos basement Kannaviou Formation

C) SWARBRICKS MODEL

Late-stage serpentinite gravity flows
Greenschist and amphibolite facies slivers
Disrupted sediment sheets, Ayios Photios Gp.

Most folds face and verge W&NW
Troodos-type slivers with Kannaviou Fm. volcanogenic cover

Younger (Troodos) Mesozoic block
Kannaviou Volcanogenic sediments

Sheared serpentinite

Older Mesozoic Block, Dhiarizos Gp.
1.3.3 THE KYRENIA RANGE

Northern Cyprus is dominated by the Kyrenia Range lineament, a convex 160km. long mountain range which comprises four major northwardly dipping thrust arcs (Baroz 1979). Geophysical studies indicate that this lineament can be traced eastwards under the Levant Sea to the Misis Mountains in eastern Turkey and westwards into Antalya Bay (Biju-Duval et al. 1974, 1977; see fig.2).

A tectonostratigraphic synthesis for the area records the existence of four main rock groups each separated by major uncoformities representing important deformational events (D1-D3, of Robertson and Woodcock 1985). A synopsis of the geological history is presented below.

Throughout most of the Mesozoic, a varied succession of shallow water carbonates (Trypa group) were deposited on a basement that is not preserved in Cyprus. These sediments were subsequently locally strongly brecciated and metamorphosed (up to greenschist facies?) in Late Cretaceous time (D1), probably soon after the genesis of Troodos oceanic crust at a spreading ridge to the south. Although often strongly recrystallised, these Mesozoic limestones and dolomites are comparable in many respects to carbonate sediments that accumulated on adjacent gently subsiding Bahama-type platforms now preserved in southern Turkey.

The deformed and deeply indurated Trypa group basement was then unconformably overlain by a Campanian cover of volcaniclastic sandstones (the Kiparisso Vuono Formation of Baroz 1979) that compositionally closely resemble the diverse volcanogenic sediments that lie in situ over Troodos crust in south and southwest Cyprus (Kannaviou Formation). In the Kyrenia Range these exotic sandstones are conformably overlain by a continuous Maastrichtian to lower Eocene sequence of pelagic carbonates intercalated with massive flows of bimodal acid/basic lava (Lapithos group).

Following this period of relative tectonic stability, the Lapithos group together with its associated underlyng basement was imbricated and folded during a Late Eocene south-directed thrust event (D2). Greenschist facies metamorphic rocks probably formed at depth but these grade into pervasively cleaved low grade Lapithos group rocks. At higher structural levels cleaved rocks in turn pass into tectonic breccias and finally into sedimentary breccias that have been deposited by syntectonic processes. These late Eocene sedimentary breccias (Kalograia-Ardana Formation) are dominated by olistostromes that were shed off local fault scarps that dissected the carbonate platform. The Troodos ophiolite was apparently unaffected by this Late Eocene (D2) compression event.

During the Oligocene and Miocene the Kyrenia lineament remained submerged,
as a thick wedge (up to 2200m.) of alluvial conglomerates and turbidites was shed across the area from the north and north east into the Mesaoria basin to the south (Kythrea group). Finally during the latest Miocene and early Pliocene a second major compressive event uplifted the range and again deformed it by southward directed thrusting (D3). Kythrea group sediments were thrusted but remained uncleaved. Rapid deposition of clastics in the Pliocene (Mesaoria group) and Pleistocene record the continued uplift of the Range.

1.3.4 THE MAMONIA COMPLEX

By contrast with the largely undeformed supra-ophiolite sedimentary succession on the northern and eastern flanks of the Troodos Massif, the geology of south west and southern Cyprus has been complicated by the emplacement of a series of Mesozoic continental margin rocks adjacent to the main ophiolite complex soon after its genesis (Lapierre 1975, Ealey and Knox 1975, Robertson 1977, Robertson and Woodcock 1979, Swarbrick 1980). These highly deformed assemblages are now preserved in the Mamonia Complex and the Moni melange of southern Cyprus.

Field mapping in the Mamonia Complex shows that it is dominated by two associated rock assemblages; the Ayios Photios Group comprising a wholly sedimentary succession of Late Triassic to Lower Cretaceous age and the volcano-sedimentary units of the Dhiarizos Group. The Ayios Photios Group can be effectively subdivided into three formations (Swarbrick and Robertson 1980, Robertson and Woodcock 1979), each of which represents successive phases in the evolution of a passive continental margin. Initially, the Upper Triassic Vlambouros Formation records a transition from shallow to deep water terrigenous flysch sedimentation along a rapidly subsiding margin prior to the onset of volcanism associated with sea floor spreading. Locally this is overlain by Halobia-bearing pelagic limestones (Marona Formation) which represent sediments deposited along parts of the margin which were starved of terrigenous clastic input. Finally, the thick Episkopi Formation (>120m.), consists generally of a fining upward sequence of mostly non-calcareous pelagic and hemi-pelagic sediments that represent deposits on a passive margin undergoing continuous subsidence from Jurassic to Early Cretaceous time. The Akamas Member (Mamonia Complex) and Parekklisha Member (Moni melange) record a renewed influx of texturally mature coarse grained orthoquartzites onto the margin in Early Cretaceous time.

By contrast, the Dhiarizos Group is dominated by a thick Late Triassic succession of mafic pillow lavas (Phasoula Formation) and volcanic breccias (Loudra tis Aphrodite Formation) containing intercalations of calciturbidites and cherts (Kholetria Member). The intra-plate alkalic chemistry of these igneous rocks lead Lapierre and Rocci (1976) to suggest they represented volcanism close to a continental margin, a
theory supported by the presence of enormous (>50m.) detached Upper Triassic reefal blocks (Petra tou Romiou Formation) associated with the extrusive succession. The igneous rocks are locally overlain by an in situ condensed pelagic sedimentary succession passing up into distal calciturbidites, probably representing a lateral more distal equivalent of the Episkopi Formation.

Detailed structural studies in the Mamonia Complex indicate that the Ayios Photios Group sediments are stacked as a series of thin sub-horizontal thrust sheets above a largely coherent basement of Dhiarizos Group rocks. As the Ayios Photios Group sediments remain uncleaved and unmetamorphosed it is apparent that there was no significant overburden at the time of their deformation. This is consistent with the deformational style within individual thrust sheets, that support a structural reorganisation of these continental margin sediments by gravity sliding down a continental margin slope onto marginal Dhiarizos Group basement rocks (Robertson and Woodcock 1979). Fold vergence and facing directions imply emplacement of these thrust sheets from the present south west toward the north east. The gravity sliding event must post-date the youngest sediments in the Ayios Photios Group which are dated as Lower Cretaceous (Berrisan).

Although it is clear that the Ayios Photios sediments are allochthonous with respect to the underlying Dhiarizos Group, the tectonic relationship between the main Troodos Massif and the basement to the Mamonia Complex remains unclear. Structural cross sections of Lapierre (1968, 1975), Kluyver (1969) and Turner (1971) show the Mamonia Complex as a grossly allochthonous unit having been thrust onto the Troodos ophiolite complex and its Late Cretaceous sedimentary cover (see fig.4b), whilst Swarbrick (1979), finding no evidence for allochthoneity of the Dhiarizos group igneous rocks envisaged that tectonic juxtaposition of contrasting basement could have taken place by strike-slip along major transcurrent faults (see fig.4c).

In support of this second mode of emplacement Swarbrick (1979) convincingly demonstrated how laterally continuous slivers of Troodos ophiolite basement preserved within the main Mamonia Complex some distance from the main Massif, retain an undeformed sedimentary cover of umbers, radiolarites, Kannaviou volcaniclastic sandstones and pelagic chalks which apparently show no evidence of having been overthrust by the Dhiarizos Group as required by Lapierre's model. This observation combined with the fact that the Dhiarizos Group basement blocks are separated from adjacent Troodos basement blocks by high angle fault contacts would favour the lateral juxtaposition of basement slices in preference to a low angle overthrusting of the Troodos ophiolite by the Dhiarizos Group and its allochthonous sedimentary cover (Ayios Photios Group). More detailed field evidence is presented in chapter 6 in support of this interpretation.

Whatever the mode of basement juxtaposition, emplacement of the Mamonia
Complex occurred within a short time interval subsequent to the deposition of the Kannaviou Formation volcaniclastic sandstones (dated as Campano-Maastrichtian) but prior to the onset of Maastrichtian and Palaeocene pelagic chalk sedimentation.

In contrast to the Mamonía Complex where allochthony of the Dhiarizos Group cannot be confirmed, the Moni melange is undisputably allochthonous with respect to the underlying Troodos ocean crust. Although the matrix to the melange consists of a stratigraphically coherent sequence of hemipelagic clays and siltstones which lie in situ on Troodos basement, a major allochthonous component is represented by a variety of olistoliths that have a provenance identical to that of the Mamonía lithologies (Robertson 1977). All these allochthonous rocks were emplaced by gravity sliding into an in situ host matrix of deep sea hemipelagic sediments at the same time as the Mamonía Complex units were being tectonically juxtaposed against Troodos ocean crust in south west Cyprus. One important feature of the Moni melange is that it contains Early Cretaceous lithologies not represented in the Mamonía Complex (Parekklisha sandstone and Monagroulli siltstone). These have been interpreted to represent proximal continental margin facies, contrasting with the more distal facies of the Ayios Photios Group sediments.

1.4 PREVIOUS PALEOMAGNETIC STUDIES IN CYPRUS

1.4.1 THE TROODOS IGNEOUS COMPLEX

Following the successful application of a working sea floor spreading hypothesis to contemporary mid-ocean ridge systems, the immediate aim of early palaeomagnetic investigations on Cyprus was to identify and delineate areas of normal and reversed polarity within an obducted fragment of ocean crust. This investigation was opportune as, if indeed the Troodos Complex represents a fragment of oceanic lithosphere created at a ridge axis, then the 120km. across strike extent of the sheeted dyke complex should represent between 12Ma. (slow spreading ridge flank, 1cm.a⁻¹) and 1Ma. (fast spreading, 10cm.a⁻¹) of spreading history. A second objective was to obtain typical values for the mean natural remanent magnetisation (\(I_n\)) for each of the different oceanic crustal layers, in order to quantitatively assess the relative contribution of each layer to the marine magnetic anomaly profile recorded over contemporary ocean floor crust.

Thus, with the intention of investigating the interfingering of normal and reversed crustal areas across the complex Vine and Moores (1969) measured the intensity and direction of the natural remanent magnetisation (NRM) of over 900 hand samples collected from over 150 localities with a portable fluxgate magnetometer. This approach was justified, as the mean NRM intensity of Troodos rocks is high.
Fig. 5 Mean westerly primary remanence vector as determined by Moores and Vine (1971) for the extrusives of the Troodos ophiolite complex. A mean inclination of $42^\circ$ for the remanence vector implies an origin for the ophiolite some $15^\circ$ to the south of the present day latitude of Cyprus.

Fig. 6 Remanence directions determined by Shelton and Gass (1980,a) and Schoenharting and Abrahamsen (personal communication) for the autochthonous Turonian to Recent sedimentary cover to the Troodos ophiolite complex. Data of Shelton and Gass are uncleaned, while directions of Schoenharting and Abrahamsen represent directions after cleaning in alternating fields of 50-60mT. Remanence directions for the extrusive series as determined by Shelton and Gass (1980) and Lauer and Barry (1976) are also presented.
MEAN REMANENCE OF TROODOS PILLOW LAVAS

AFTER MOORES AND VINE (1971)

FIG. 5 POLAR STEREOGRAPHIC PROJECTION
and complementary laboratory studies had indicated that the remanence vectors of pillow lava and gabbro samples change very little on progressive demagnetisation. Importantly, at many sites, particularly on the northern margin of the ophiolite complex, pillow lavas only dip at shallow angles, thus the NRM directions obtained in the field could, to a first approximation, be taken to be indicative of the primary remanent magnetisation.

Unfortunately, this preliminary study revealed a marked absence of reversely magnetised crust over the outcrop area of the Massif. Only shallow positive inclinations (mean 42° down) were recorded in the pillow lavas indicating an origin for the Troodos in the northern hemisphere at a palaeolatitude of 21°N consistent with a location intermediate between Africa and Eurasia during the Late Cretaceous. Between-site scatter could be partly attributed to the difficulty in estimating the attitude of the plane of the original horizontal in pillow lava terranes. The inclination was also consistent with the genesis of Troodos crust toward the end of the long Aptian-Santonian normal polarity epoch (118Ma.- 80Ma.) prior to the Campanian when the Perapedhi Formation and Kannaviou Formation sediments were accumulating on the ocean floor (these sediments contain radiolaria diagnostic of the Campanian, Mantis 1972).

Quite unexpectedly, the primary remanent magnetisation retained in the pillow lavas was consistently found to be orientated approximately due west (mean 276°, see fig.5) which, assuming an axial geocentric dipole model for generating the earth's magnetic field, can be interpreted as indicating an anticlockwise rotation of the whole Massif through 90° since its formation (conceivably rotation could be through 270° in a clockwise sense). This implied that the original ridge axis was orientated east-west, more in line with the general Tethyan trend.

The intensity of the NRM for the zeolite facies pillow lavas (1-10Am⁻¹) was typically found to be considerably greater than the greenschist facies rocks of the sheeted dyke complex (0.01-0.5Am⁻¹). As low field susceptibilities for each crustal layer were similar, Koenigsberger ratios (ratio of NRM intensity to intensity of induced magnetisation) for the pillow lavas were typically 10 whilst for the underlying dyke complex they were approximately 0.5. The dramatic decrease of mean intensities in the sheeted dyke complex has been attributed to the metamorphic destruction of titanomagnetite phases into Ti-poor titanomagnetites and paramagnetic Ti-rich ilmenohaematites during hydrothermal circulation of fluids at the spreading ridge crest. The liberated titanium is thought to be accommodated in sphene, whilst iron is incorporated into the crystal lattices of actinolite, chlorite and magnetite.

As the underlying plutonics have NRM intensities similar to those of the sheeted dyke complex (0.1-0.5Am⁻¹), it was concluded that the most potent source of remanent magnetisation in the Troodos ocean crust lay in the pillow lava formations.
This result was confirmed by a high level aeromagnetic survey of southern Cyprus that clearly indicated that the steepest anomaly gradients correlate closely with the outcrop area of the peripheral pillow lavas (Vine et al. 1973). In addition, the systematic association of pillow lava outcrops with positive magnetic anomalies suggested that the pillow lavas were uniformly normally magnetised.

In contrast to the NRM of the Troodos pillow lavas which clearly only include only minor secondary components above the primary thermoremanent magnetisation, samples collected from within the greenschist facies sheeted dyke complex were often dominated by an important secondary viscous component which curiously had a coercive force spectrum that often coincided with that of the primary magnetisation. Therefore, before applying structural corrections, NRM directions within the sheeted dyke complex typically clustered between declinations of the pillow lavas (due west) and the direction of the ambient magnetic field over Cyprus (DEC=003°, INC=54.5°). Dykes sampled at the upper levels of the sheeted complex (Basal Group) consistently record NRM directions coincident with the present day field direction (Lauer and Barry 1976).

The underlying gabbros retain a stable magnetisation that is carried exclusively by magnetite (Kent et al. 1978). Prior to making appropriate structural corrections based on the attitude of cumulate mineral banding, compiled site means were typically well grouped around a mean westerly declination similar to that isolated for the pillow lavas. However, between site scatter increased considerably after re-tilting back to the assumed palaeohorizontal and it was concluded that either mineral layering formed at high angles to the original horizontal or that mineral banding was tilted whilst the ambient temperature was still above the Curie point of magnetite (Moores and Vine 1971).

Unlike the gabbros, the cumulate ultramafics (pyroxenites and dunites) and residual ultramafics (tectonised peridotites) clearly retain an NRM direction close to the present day field. Magnetic overprinting is probably related to pervasive serpentinisation of these rocks during the post-Middle Miocene uplift of the Massif. Associated with these serpentinised ultramafics exposed around Mount Olympos and in the Limassol Forest area are strong negative magnetic anomalies.

1.4.2 THE SEDIMENTARY COVER

Following the discovery of a stable, westerly directed primary magnetisation in the extrusive units of the Troodos ocean crust, palaeomagnetic studies were extended into the continuous Late Cretaceous to Recent sedimentary cover to the ophiolite with the intention of identifying the precise timing of the rotation of the ocean crust. This approach was justified as the sediments lie in situ on the ophiolite and therefore a
change in the direction of a primary magnetic vector within the sedimentary cover must necessarily reflect motion of the underlying basement.

In a preliminary study, Shelton and Gass (1980) measured the natural remanent magnetisation of a small suite of samples collected primarily from the pelagic chalk successions that dominate the sedimentary cover to the ophiolite complex. As the NRM intensity of these sediments was found to often be comparable with the noise level of their measuring instrument (a slow spin balanced fluxgate magnetometer) no absolutely reliable vectors could be defined. Indeed, after even using high spin numbers \(2^7-2^8\) and measuring all specimens in six orientations there was a great deal of variance in resultant vectors. The stability of remanence was not investigated.

In addition, a limited number of samples with measurable NRM intensities were collected from the Perapedhi Formation (ferromanganiferous umbers), the Kannaviou Formation (volcaniclastic sandstones) and from a sequence of Pliocene marls. The declinations of the characteristic remanence vectors for each of these cluster very convincingly in the north west quadrant (see fig.6a), although the limited sample population, the scatter of inclination values and lack of stability tests necessarily makes any interpretation of these vectors controversial. Shelton and Gass were certainly not justified in concluding that rotation was confined to a single Late Miocene event.

To complement this preliminary investigation, a further set of orientated samples were collected from the Troodos extrusive lavas and their in situ sedimentary cover by Schoenharting and Abrahamsen (1982). Although details of their study remain unpublished, they reported remanence directions (after AF cleaning) for the extrusive series (9 sites, 60 samples) that confirm the results of earlier studies (Moores and Vine 1971; Shelton and Gass 1980 and Lauer and Barry 1976)

Remanence directions of weakly magnetised Tertiary carbonate sediments were measured with a cryogenic magnetometer (personal communication, 1984) and their stability investigated by applying alternating field cleaning techniques (in 10mT incremental steps until remanence intensities were comparable with the sensitivity limit of their measuring instrument). The majority of their samples (14 sites, 95 samples) were collected from two successions (see fig.3), one near Pano Lefkara (south east Troodos, Upper Palaeocene-Lower Eocene chalk/chert succession) and a second near Agrokipia (north Troodos, comprising Upper Eocene-Lower Miocene pelagic chalks and Upper Miocene reefal limestones). The biostratigraphy of these sections is well defined (Mantis 1972, Zomenis 1972).

Compiled NRM directions grouped into two populations, each presumably representing normal and reversed polarities in the NW and SE quadrants (see fig.6b). However, although Fisherian statistics somewhat improved upon demagnetisation, remanence directions remained highly scattered, demonstrating that cleaning failed to
remove a significant secondary remanence component. Mean declinations after correction for tectonic tilt clustered between north and north west for the majority of sites.

1.5 OBJECTIVES OF THIS RESEARCH

1) Investigate the significance of variations in the direction of the mean magnetic remanence within the extrusive units of the ophiolite complex. Earlier palaeomagnetic studies recognised the existence of these variations, but failed to comment on their importance (Moores and Vine 1971, Lauer and Barry 1976; see fig.6).

2) To extend the palaeomagnetic sampling coverage through the continuous Late Cretaceous to Recent sedimentary cover to the ophiolite complex with the intention of clearly defining the timing of its 90° anticlockwise rotation.

3) To compare the magnetic properties of pelagic sediments overlying Troodos with comprehensively studied pelagic limestones from the western Tethyan area.

4) Determine the remanence directions of pillow lavas exposed on the Akamas Peninsula and to compare these with corresponding directions from the main Massif to the east; to test for any rotation along the intervening Polis Graben, a major structural feature.

5) To use the palaeomagnetic technique to test whether deviations in dyke trend in the vicinity of the Arakapas Fault Belt are a result of tectonic rotation of large fault blocks subsequent to crustal accretion.

6) Determine the palaeomagnetic declination of Mesozoic sediments and lavas in the Mamonia Complex to investigate whether there has been significant tectonic rotation of these units relative to the Troodos Complex. Corresponding inclinations should constrain the latitudinal position of the Mamonia Complex units in relation to the African plate prior to the closure of the Tethys.

7) To determine whether autochthonous basement elements in southern Turkey have experienced similar 90° anticlockwise rotations as the Troodos ophiolite complex, or have moved independently during the closure of the Tethys ocean.
CHAPTER 2

INTRACRUSTAL BLOCK ROTATIONS ALONG THE NORTHERN MARGIN OF THE TROODOS OPHIOLITE COMPLEX

2.1 INTRODUCTION

The presence of a sheeted dyke complex in the Troodos ophiolite succession provides the key evidence for crustal genesis at a well defined spreading ridge axis. As dykes over the exposed extent of the Massif strike NS, prior to the 90° anticlockwise palaeorotation of the ocean floor crust, dykes were aligned east-west. Thus the Arakapas Fault lineament, interpreted to represent a fossil transform fault, was originally orientated NS, perpendicular to the orientation of the spreading ridge axis.

In detail however, over some areas of the Massif, the structural grain of dykes within the sheeted complex and in the overlying extrusive units deviate considerably away from this general NS trend. Indeed, detailed mapping of dyke orientations along the northern margin of the complex reveals how the oceanic crust can be effectively subdivided into a number of separate areas within which dykes are uniformly orientated (Verosub and Moores 1981). Although on the north and north west margins of the plutonic core to the ophiolite complex dykes are predominantly orientated NS, toward the extreme north west and on the eastern flanks of the Massif a more complex pattern of dyke trends is observed. These are schematically illustrated in fig.1.

In this chapter palaeomagnetism is used to investigate whether discrepancies in dyke trend along the northern flank of the ophiolite complex simply reflect complexities in the original orientation of the extensional stress field at the time of crustal accretion, or whether dyke deviations can be accounted for by relative rotation of intracrustal blocks subsequent to crustal genesis. Deviations in dyke trend in the vicinity of the Arakapas Fault belt and in the Limassol Forest are considered independently in chapter four.

2.2 METHODOLOGY

If all dykes across Troodos were at the time of crustal genesis orientated vertically and in parallel, discrepancies in present day dyke trends can simply be attributed to rotation of faulted blocks in the horizontal plane about vertical axes. As dykes within the sheeted dyke complex provided magmatic feeders to the immediately overlying extrusive units, any systematic rotation within the horizontal plane of the sheeted dyke complex must necessarily be also reflected in the stratigraphically higher extrusive units. Therefore, as the azimuth of a magnetic vector provides a control on
the sense and magnitude of tectonic rotation within the horizontal plane, a comparison of structurally well constrained magnetic vectors retained in lava successions around the periphery of the Massif should reveal whether there have been any significant intracrustal block rotations about vertical axes.

Obviously, a more direct method of investigating the original relationships between dyke orientations across the Massif would involve comparing remanence directions within the sheeted dyke complex itself, but this was considered inappropriate for the following reasons:

a) In sheeted dyke terranes it is not possible to accurately determine the attitude of the plane of the original horizontal, and therefore there is no absolute control on the relative directions of the primary magnetisation vectors. Certainly, if dykes are assumed to have been intruded vertically, they can be restored back to their original attitude by a simple rotation about a line of strike orientated on the dyke margin. However, application of this conventional tilt correction in dyke terranes fails to provide an appropriate reference framework for the re-orientation of the primary magnetisation vector, as any amount of undetectable rotation could have occurred in the plane of the dyke itself. If however, the plane of the original horizontal can be readily identified (as it can at many localities in the extrusive series) and structural tilting occurred about the line of strike, only tectonic rotations about vertical axes in the horizontal plane cannot be corrected for by a conventional tilt method.

b) Early palaeomagnetic studies revealed that remanent magnetisations isolated from the greenschist facies sheeted dyke complex were often dominated by viscous secondary magnetisation components and therefore the remanence acquired at the time of dyke injection could not always be effectively determined (Moores and Vine 1971). Magnetic instability within the sheeted dyke complex can be directly attributed to pervasive greenschist metamorphism associated with the circulation of chemically active hydrothermal fluids at the spreading ridge axis. By contrast, the zeolite facies extrusive units are dominated by primary magnetic components acquired as the lavas cooled rapidly to temperatures below the Curie points of their magnetic carriers. As the metamorphic grade boundary transcends lithological boundaries (Smewing and Gass 1973, Banerjee 1980), but generally occurs at a crustal depth of 1.5km. (corresponding approximately to the depth to the base of the traditional Lower Pillow Lavas) it was considered only appropriate to sample from the extrusive series of the ophiolite complex.
Fig. 1. Schematic map of the Troodos ophiolite complex showing principal orientations of dykes within the sheeted diabase and the overlying extrusive series (after Moores and Vine 1971). Dyke orientations in the Akamas and Troulli inliers are also indicated as is the area covered by fig. 2a.

Fig. 2. Sketch map showing main dyke trends in the Upper and Lower Pillow lava series along the northern (a) and north-eastern (b, after Gass 1960) margins of the Troodos massif. Disposition of Basal group units and principal faults on north east Troodós are also indicated. Sampling sites on northern Troodos (see table 1) were chosen in extrusive successions where intruded dykes clearly deviate markedly away from the predominant north-south trend.

Table 1. Details of sampling procedures at sites along the northern margin of Troodos (see text for details of terminology).

Table 2. Mean remanence directions prior to and subsequent to magnetic cleaning for samples collected at sites along the northern flanks of the Troodos massif. Remanence directions for extrusives sampled at Karavostasi and Apliki (see fig. 2a) are reported in Lauer and Barry (1976). N=Number of subsamples.
Main dyke trends across Troodos

area covered by fig.2

Akamas Inlier

Troulli Inlier

Limassol Forest Block

Arakapas Transform lineament

Gabbro and ultramafic core to Ophiolite

Sheeted dyke complex and extrusive series

FIG. 1
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<td>1756.1</td>
<td>10mT</td>
<td>271.9</td>
<td>41.9</td>
<td>9.7</td>
<td>69.0</td>
</tr>
</tbody>
</table>

**TABLE 2**
2.3 CHOICE OF SITES IN THE LAVA SUCCESSION

The extrusive series on the northern flanks of the ophiolite complex has a total thickness of between 1000m. and 1600m. (Schminke et al. 1983). Detailed mapping along certain well exposed sections through this succession (Gass 1960, Schmincke et al. 1983, Boyle and Robertson 1984) shows that it can effectively be divided into a number (between 10 and 15) of major independent lithologic units, represented by a heterogeneous assemblage of pillowed and massive lava, lava breccia, hyloclastites and minor intrusives. Pillow lavas make up the bulk of the extrusive succession and form lithologic units ranging from several tens of metres to flows more than 100m. thick (Robinson and Gibson 1983). Lateral continuity of lithologic units particularly in the lower parts of the succession would support eruption of the bulk of the lava pile onto a relatively flat sea floor at the axis of a well defined median valley.

Although it is almost impossible to accurately define the plane of the original horizontal in sheeted dyke terranes (see above, section 2.2), numerous potential ‘bedding plane’ criteria can be recognised in the extrusive series. Not all of these are appropriate for the successful application of an accurate structural correction.

1) Pillow lava morphology. In outcrop most lava pillows are uniform in size and shape, with typical lobated lateral and upper contacts. Pillow ‘toes’ provide an effective ‘way up’ criterion. The morphology of the pillow lavas in well exposed sections gives a general impression of the structural dip of the unit, with lines drawn through the long axes of the pillows indicating the inclination of the palaeoslope down which the pillows were extruded. However as lava slopes can be inclined at a considerable angle to the horizontal (Ballard et al. 1981, 1982), gross pillow lava morphology alone cannot be used to determine the plane of the palaeohorizontal.

2) Mineral banding and vesicle layering. Although the stratigraphically lower andesite-dacite-rhyolite lava suite is largely aphyric or sparsely phytic, the upper lava suite of picrite-mafic basalts and basaltic andesites are commonly enriched in phenocrystic olivine and pyroxene (Robinson et al. 1983, Schmincke et al. 1983). In individual pillowed and massive flows these phenocrysts have settled under the influence of gravity into distinct mineral bands which closely parallel the plane of the original horizontal. By contrast, volatiles, which have a density lower than that of the surrounding magmatic fluid, tend to rise and congregate as vesicular-rich layers at higher levels in the lava unit. Again, a consistent attitude in the characteristic internal layering over a number of individual flow units can provide a reliable palaeohorizontal indicator.
3) Bedding lamination in interlava sediments. Iron-rich oxide sediments of hydrothermal origin are dispersed throughout the entire extrusive succession but are particularly abundant in the stratigraphically highest lava units (Boyle 1984). In the majority of occurrences, these sediments fill the cuspatc spaces between adjacent pillows, and any visible bedding lamination is often distorted by over-riding lava flows. Rarely in these settings can a reliable palaeohorizontal surface be established. Occasionally however, thicker laterally continuous horizons of interlava ferromanganiferous mudstones and epiclastic sediments have accumulated during a pause in ridge volcanism. Although the thickness of these units rarely exceeds 1m., individual beds can often be traced along strike over extensive distances (occasionally 200m. or more) and thus probably record deposition on a relatively flat sea floor.

4) Massive lava flow morphology. 25%-40% of the extrusive succession consists of massive flow units (Robinson and Gibson 1983). Individual flows vary in thickness from 2m. to 5m. and often grade into pillowed and pillowed brecciated flows, a transition which can be attributed to a change in the inclination of the palaeoslope. If however, flows can be traced for long distances along strike (>100m.), with little evidence of lateral facies variations or ponding in topographic hollows, the extrusion of lavas probably occurred on a relatively flat basement relief. Thus, the top surfaces of these flows will in most cases adequately record the plane of the original horizontal.

Although the main part of the lava pile (equivalent to the traditional Lower Pillow Lavas) was erupted on a relatively flat sea floor, the extrusion of the higher lava units (Upper Pillow Lavas) was accompanied by normal faulting associated with the formation of an axial rift valley. Thus, as lava extrusion was contemporaneous with structural tilting of the underlying lava units (i.e. syntectonic), structural corrections at sites in the Upper Pillow Lavas series are only locally specific and cannot necessarily be applied to lava units at other levels in the succession. This same rule applies when attempting to relate the bedding of supra-lava sediments to that of the underlying lava units. Although locally lavas pass conformably into ferromanganiferous oxide sediments (umberS), radiolarites and chalks along the major part of the northern periphery of the ophiolite complex, Tertiary pelagic sediments unconformably overlie the stratigraphically highest lava units and therefore the bedding of these younger sediments is unrelated to the tectonic tilt of the underlying extrusive series.

Fig.2 shows the main dyke trends within the extrusive series and the underlying sheeted complex along the northern periphery of the ophiolite complex. The map clearly illustrates how dykes within the sheeted complex closely closely parallel dykes in the overlying extrusive series (although the boundary between the Pillow Lava Series and the sheeted dyke complex is represented by a line, in reality the boundary is a
gradational one, and approximates to a crustal level beneath which 80% of the crust consists of sheeted diabase). On the north east margin of the Massif, dykes deviate markedly away from the NS trend of the main complex. The ocean crust in this area is cut into a series of independent basement blocks by major fault lineaments that extend through the sheeted dyke complex and the extrusive series. Within individual fault blocks there is a general agreement in the strike direction of dykes, although deviations occur especially in the vicinity of major faults.

In this study, the extrusive series was sampled at 7 localities along the northern periphery of the ophiolite complex. Individual localities were chosen both in crustal areas where dykes were aligned NS and in areas where dykes deviate away from this orientation (see table 1 and fig.2). Importantly, at each site an accurate structural correction could be applied to measured remanence directions (see above and table 2), although sites with only minor tectonic tilt were preferred (usually <35°), as this minimised any undetectable 'declination errors' that resulted from rotations about axes inclined at an angle to the strike of an inferred palaeohorizontal surface (Macdonald 1980, see section 2.9.2).

2.4 SAMPLING PROCEDURES

At the majority of sites, pillowed and massive lavas were drilled in situ with an adapted petrol driven 'Stihl' chain saw motor. During coring, the diamond tipped stainless steel drill bit was cooled and lubricated by water pumped from a modified portable crop spraying canister. Occasionally, 15-20cm. long cores could be drilled before internal fracturing occurred, but the usual core length averaged between 5 and 10cm. At an individual site (single pillow or massive lava flow unit) between 5 and 10 independently orientated cores were collected, with preferably 50-70 cores collected per locality (comprising 3-10 sites, usually over an area of less than 1km²). Where possible, care was taken to sample both the exteriors and the interiors of pillows which ranged in size from <0.5m. to >2m. in linear dimensions. At massive flows both the finely crystallised flow margins and the more coarsely crystallised central portion of the flow were sampled.

Drilled cores were orientated using a standard orientation table device attached to a hollow non-magnetic copper tube. The pivotable surface of the orientation table allowed the direction of the drilling to be measured relative to true north and the angle of drilling relative to the horizontal to be determined. Thus, the orientation of a fiducial line marked on each core through a slit in the copper tube (applied with a thin black indelible marker pen) was then measured with both a magnetic and sun compass (Creer and Sanver 1967), which were attached to the upper surface of the orientation table (see fig.3a). Orientated cores were then extracted from the rock volume and
Fig. 3a Orientation of drilled cores using standard sun compass device. Fiducial line of cylindrical sample is marked through the slit in the hollow shaft. This direction is diametrically opposite to the direction of maximum dip of the drilled core. The pivotable surface of the sun compass allows the direction of drilling and the angle of drilling from the horizontal to be determined. From each individual core a number of independent subsamples were prepared by cutting perpendicularly to the long axis of the core.

Fig. 3b Sample orientation system, where the magnetic remanence vector is measured relative to the x axis of the core which is aligned parallel to the direction of drilling (see fig. 3a). Declination is measured clockwise around from the x axis in the horizontal plane, and inclination represents that the remanence vector makes with the xy plane.

Fig. 4 Field and bedding correction conventions. (A) shows direction of remanence vector (in this example DEC=150°, INC=50°) relative to the sample coordinate system illustrated in the previous figure. Relative to due north, the fiducial line is aligned toward 160° (i.e. the top surface of the drilled core is dipping toward 250°). Thus after restoring the subsample back to its original orientation in the field (involving a rotation about a line of strike on the sample, see B) the mean remanence vector moves along a great circle on the stereographic projection towards a more easterly declination. The remanence vector can then be referred back to the palaeohorizontal by rotating the sample about a line of strike on an inclined bedding surface (C).
FIG. 3a
SUBSAMPLE COORDINATE SYSTEM

FIG.3b

Direction of slit on orientation table

Direction of magnetic remanence vector
toward Jr, 350' at about 40'
top surface of core dipping toward 250° at 20°

beds dipping toward 350° at about 40°

FIELD CORRECTION
BEDDING CORRECTION

FIG. 4

FIELD AND BEDDING CORRECTION
clearly labelled with an indelible marker pen. Where possible, broken cores were reconstructed by sticking individual fragments together with a non-magnetic epoxy resin adhesive.

At inaccessible sites, or in lava units which were strongly fragmented or in well bedded sediments, hand sampling was preferred. Samples were orientated by finding a smooth surface (a bedding surface in sediments) and marking on it a horizontal strike line. Where no flat surface could be located, a surface could be artificially attached by gluing a plastic disc onto the sample (see chapter 5, section 5.7). The orientation of this horizontal line relative to true north and the dip of the surface perpendicular to the strike were then measured. Providing the site was not located in the shade, the orientation of this strike line could measured with a sun compass constructed on a specially adapted orientation table. If, however, orientation was limited to the use of a magnetic compass, care was taken to maximise the distance between the compass needle and the adjacent strongly magnetic igneous rock. At each site at least four independent measurements of the strike and dip of an inferred palaeohorizontal surface were taken using an identical technique to that described above. Mean values for bedding plane strike and dip were then calculated.

2.5 PREPARATION OF SAMPLES AND MEASUREMENT OF NRM

In the laboratory between 3 and 6 cores were drilled from each hand sample using a pedestal drill with its bit aligned perpendicularly to the orientated surface. These cores, together with cores collected in the field were then cut into either one or more 2cm. long cylinders, depending on the length of the original sample. Any visibly altered sections of a core were discarded during this subsampling process. The fiducial line along the margin of each subsample was then extended onto the top surface of each cylinder. All remanence directions were then measured with respect to this line (arbitrarily designated the ‘x’ axis, see fig.3b).

After storage in zero magnetic field for a few weeks, the NRM of all these samples was then measured with a portable ‘Molspin’ fluxgate magnetometer. The usual procedure for NRM measurement involved spinning each sample in 6 mutually orthogonal orientations such that any two of the x, y and z axes of the cylindrical samples were in turn perpendicular to the spin axis. As two orthogonal components of magnetisation were obtained from each spin, orientations were chosen to measure each of the x, y and z components 4 times (two of each sign for each component). This procedure tends to average out any random noise in the signals, any effects of inhomogeneity in the NRM and any stable remanent magnetisation associated with the sample holder attached to the rotating spinner shaft. Rapid computation of each of the x, y and z remanence components was achieved by interfacing the ‘Molspin’
magnetometer with an 'Epson HX-20' microcomputer. Declination (azimuth of the remanence vector relative to the x axis, see fig.3b), inclination (angle between the remanence vector and the horizontal xy plane) and NRM intensity were calculated in the same programming routine. All these igneous rocks and interlava sediments recorded NRM intensities at least two orders of magnitude above the sensitivity limit of the fluxgate magnetometer (c. 0.2mAm⁻¹).

The declination, inclination and NRM intensity values for each sample were then transferred by magnetic tape via a 'Sirius' microcomputer onto the ERCC mainframe computer. Conventional field and bedding corrections were then applied to the raw data (i.e. first relating the direction of the NRM to present day geographical coordinates, followed by a correction for tectonic tilt of the units that refers the direction of the remanence vector to the plane of the palaeohorizontal, see fig.4). Field corrected and bedding corrected data for individual sites were then studied on stereographic projection plots.

2.6 STABILITY OF THE REMANENCE

The stability of the NRM retained by these igneous rocks and interlava sediments was investigated by standard AF demagnetisation of at least 3 pilot samples from each locality (usually at 2.5mT incremental steps up to 10mT, then at 5mT steps until the remanence was dominated by a spurious ARM components). This technique offers the most effective method of analysing the coercivity spectra of predominantly titanomagnetite and titanomaghemite bearing rocks, and does not result in the chemical creation and destruction of magnetic mineral phases as may happen with thermal demagnetisation.

The standard demagnetisation coil was located inside a mu-metal shield, to cancel the present day geomagnetic field over the sample during demagnetisation. The procedure involved inserting the sample into the coil in three mutually perpendicular (parallel to x,y,z) orientations at each incremental demagnetisation step. At higher fields (>50mT), ARM acquisition was minimised by demagnetising the samples in a further three mutually perpendicular (parallel to -x,-y,-z) orientations at half the field strength. The intensity and direction of remanence were then measured with directional changes being studied on pairs of two component Zijderveld plots. Up to 8 samples could be demagnetised at the same time.

The decay of a magnetic remanence vector along a straight line toward the origin on a Zijderveld plot indicates the presence of a stable remanence component.

2.7 AF DEMAGNETISATION OF TROODOS LAVAS
Fig. 5 AF demagnetisation of extrusive sample. Following bedding correction, the stereographic projection plot (A) and the normalised intensity plot (B) indicates that a reduction in the remanence intensity during progressive demagnetisation is associated with only minor directional changes in the remanence vector. The NRM direction (identified by a + symbol) is clearly more northerly than the stable remanence vector; the latter being isolated in fields of less than 5mT. The MDF is comparatively low (approximately 10mT) and only 1% of the original remanence remains in fields of >40mT. Pairs of two component plots (C) show how the removal of a steep northerly dipping low coercivity component results in a drift of the stable remanence vector towards shallower inclinations (i.e. z decreases) and more southerly declinations (i.e. x decreases).

Fig. 6 Bedding corrected mean remanence directions (x) together with their associated $\alpha_{95}$ cone of confidence for the 7 sampling localities in the extrusive succession along the northern margin of the Troodos massif. Representative Zijderveld plots are also presented (see appendix 1 for axis convention). N=number of subsamples, *=direction of present geomagnetic field. Boundaries of Upper and Lower Pillow Lava suites are drawn on the map.
DEMAGNETISATION OF PILLOW LAVA SAMPLE

A) CPP 27R
NRM = 1183.68 mAm$^{-1}$

B) CPP 27R
HOF = 98.5"e

C) CPP 27R
Y VS Z
X VS Y
REMANENCE DIRECTIONS AFTER CORRECTING FOR STRUCTURAL TILT

FIG. 6

TROODOS MASSIF
Representative AF demagnetisation curves of samples collected from the Troodos pillow lavas are illustrated in fig.5. These curves qualitatively reveal how the total remanent magnetisation is distributed among magnetic mineral grains with a range of coercivity values. Thus the demagnetising field at which the remanent magnetisation is reduced to half of its initial NRM intensity (median destructive field, MDF) is an important defining parameter of the coercive force spectrum of the magnetic grain carriers.

Nearly all pilot samples responded well to AF cleaning, with intensities dropping to one half of their NRM values in fields of less than 45mT. Values of MDF strengths of Troodos extrusive lavas are similar to those recorded for DSDP basalts (Lowrie et al. 1973) and indicate a good stability of the natural remanence. For all Troodos extrusives, the NRM is dominated by a stable remanence with a westerly declination and a shallow positive inclination. Remanence directions are in agreement with previous palaeomagnetic studies within the extrusive succession on northern Troodos (Vine and Moores 1969, Lauer and Barry 1976, Shelton and Gass 1980). A minor, low coercivity component was removed in low alternating fields (typically less than 10mT). Generally this component was northwardly dipping (positive inclination), and it was assumed to represent a secondary viscous component acquired in the present day field at the site. Removal of this component often resulted in an initial rapid decline in the NRM intensity accompanied sometimes with a directional change in which the NRM direction moves away from the contemporary axial dipole direction (see fig.5a). Occasionally however, removal of a southward directed component with a negative inclination resulted in an initial increase in the remanence intensity. This component was probably acquired during a Quaternary reversed polarity interval and could be related to a chemical remanent magnetisation (CRM) associated with the crystallisation of a magnetic phase during an uplift episode of the Troodos Complex.

The alternating field at which the viscous components were removed from the total remanence was identified from orthogonal projection Zijderveld plots by recognising the minimum field above which the remanence intensity decreased smoothly without a corresponding change in the remanence direction. Thus, on the basis of these pilot demagnetisation runs, a peak alternating field slightly above that of this critical field strength could be applied to the remainder of the samples collected at an individual site. The field strength rarely exceeded 15mT.

A comparison of remanence data before and after demagnetisation treatment (table 2) shows only a small decrease in the scatter of remanence directions subsequent to magnetic cleaning. At localities where the NRM of lavas includes a soft viscous component orientated toward the direction of the earth's present day field, a characteristic reduction in $\alpha_{95}$ is accompanied by a slight movement of the site mean away from the axial dipole direction. Importantly, the consistent grouping of the
remanence vectors away from the present day geomagnetic field direction over Cyprus (DEC=003°, INC=54.5°) both before and after application of a structural correction presents convincing evidence for an NRM dominated by a component with a direction quite unrelated to the direction of the present axial dipole field. Indeed, upon application of the bedding correction to the site mean at each locality there is a dramatic reduction in between-site scatter, demonstrating that this component was acquired prior to the structural tilting of each lava unit. As rotation of individual lava units occurred about intracrustal faults that are characteristically overlain by undisturbed sediments of the supra-ophiolite cover, the remanence was necessarily acquired either at the time of crustal genesis or soon after.

Irreversible thermo-magnetic curves of Beske-Diehl and Banerjee (1979) clearly show that the stable remanence is invariably retained by titanomaghaemite with a Curie temperature ranging between 350°c and 450°c. As the titanomaghaemite in these rocks has not inverted to magnetite and an associated Ti-rich mineral these lavas have not undergone heating greater than 200-250°c, supporting a primary thermoremanent origin for the primary magnetisation. Thus, the the thermomagnetic behavior of these rocks is compatible with them having acquired their primary remanence at the time of crustal genesis.

2.8 DIRECTION OF PRIMARY REMANENCE

Following the standard demagnetisation procedures outlined above, mean remanence directions were computed for each site, together with corresponding Fischerian statistics, $\alpha_{95}$ and $\kappa$ ($\alpha_{95}$ is the semi-angle of the 95% cone of confidence and $\kappa$ is the precision parameter). A standard bedding correction was applied to each site, so that the remanence vector could be related to the plane of the original horizontal. The characteristic mean remanence direction together with their associated 95% circle of confidence were plotted on individual stereographic projection plots (fig.6). Representative orthogonal Zijderveld plots of pilot sample demagnetisation runs are presented adjacent to each stereonet.

Although minor discrepancies exist in the mean remanence direction between localities, the extrusive series on the northern margin of the ophiolite complex clearly retains a stable magnetisation of normal polarity which is consistently orientated toward the west. Importantly, locality mean directions coincide almost exactly with the mean vector of Moores and Vine (1971) for all Troodos pillow lavas. In detail though, mean locality remanence declinations vary between 255° (Mitsero) and 308° (Analiondas) with inclinations also ranging from 46° (Malounda) to 20° (Margi). Within-locality scatter of remanence directions is represented by the size of the circle of confidence, with $\alpha_{95}$ generally varying to a first approximation in proportion to the
number of samples collected at a given locality, from small values of less than $10^0$ to rather greater values at sites where the sampling collection has been more substantial (e.g. Sha, 103 samples).

2.9 DISPERSION OF REMANENCE VECTORS

Clearly, at each individual locality there is a dispersion of remanence vectors about a mean direction, as the circle of confidence has a finite radius. In this section the likely causes for this scatter are investigated both at the site and locality sampling levels. This discussion provides a basis for a consideration of the significance and the implications of between locality variations in the mean remanent magnetization direction along the northern margin of the ophiolite complex.

2.9.1 WITHIN SITE SCATTER

A site comprises 2-10 subsamples collected from either a single pillow or a single sheet lava flow unit. Following magnetic cleaning, remanence directions show a small but finite dispersion, with $a_{95}$ rarely exceeding $5^0$. As an identical bedding correction is applied to each sample at an individual site, dispersion of remanence vectors can be attributed to:

a) Imprecision in sample orientation. If the drilled core is greater than 6cm. in length orientation errors should not exceed the width of the fiducial mark on each sample ($2-4^0$). If however, the drilled core hole is shallow (i.e. <4cm.), the attitude of the core in the rock outcrop cannot be determined accurately with a standard sun compass device. Potential inclination errors of as much as $5^0$ can be introduced at this stage. Absolute determination of strike is impossible in highly magnetic rock terranes, if orientation of either cored or hand samples is based on a conventional magnetic compass reading.

b) Failure to effectively determine the true remanence direction. Remanence inhomogeneity, an anisotropy of magnetic susceptibility, measuring instrument noise levels and ease of acquisition of spurious low coercivity viscous components all contribute towards inaccuracies in assessing the true remanence direction in any one sample. In addition to these factors, if after partial demagnetisation, secondary components are still contributing to the total remanence within some samples at a particular site, remanence directions will necessarily be 'strung-out' between the direction of the primary component and that of the secondary component. As secondary components in the extrusive series are effectively removed at low peak
demagnetising fields, only minor within-site scatter of primary remanence directions can be attributed to inefficient magnetic cleaning.

2.9.2 WITHIN LOCALITY SCATTER

Obviously, within-site scatter is also reflected in the dispersion of remanence directions at a given locality. However, as site mean vectors are never identical, the computation of a locality mean (comprising data from 2-10 sites) will inevitably lead to an increase in directional scatter. This may be attributed to:

a) Palaeo-secular variation. If sites are distributed through an extrusive sequence of lava units that were extruded over a time interval comparable with the period of secular variation cycles, then between-site scatter may be partly related to the drift of the geomagnetic field direction during the course of crustal accretion. As studies on recent terrestrial igneous rocks demonstrate that spot readings of the earth's magnetic field can deviate away from the direction of the axial dipole field by as much as 20° (Coupland and Van der Voo 1980), a reliable locality pole position can only be determined by averaging a series of virtual pole positions compiled from a number of sites along a continuous lava succession.

b) Inappropriate bedding corrections. At specific sites, particularly in pillow lava terranes, it is difficult to accurately determine the plane of the palaeohorizontal. As all site means are related to the palaeohorizontal by rotating the primary remanence vector about a line of strike along an assumed bedding plane, errors are necessarily introduced if this plane cannot be appropriately defined. Thus, the most reliable remanence directions are those recorded at sites where the structural correction has been based upon a reliable palaeohorizontal criterion (bedding planes in interlava sediments).

c) Relative rotation of lava units about inclined axes. A conventional structural correction in palaeomagnetic studies assumes that tilt took place about a line of strike to bedding. However, this basic assumption is incorrect where tectonic rotations have occurred about non-horizontal, inclined axes. Thus, if at any given locality, adjacent sites have undergone tectonic rotations about axes that are not aligned in parallelism, any conventional tilt correction will necessarily introduce both an inclination and a declination error in the site mean. However, at any given locality the attitude of the palaeohorizontal plane is invariably consistent, and only minor within-locality scatter can be attributed to relative rotations about inclined axes.
2.10 MARGI - A CASE STUDY

In the Margi-Mathiati area on the north east margin of the Massif, dykes trend consistently NW-SE, deviating markedly away from the NS orientation observed further to the west (see fig.2). Within the extrusive series, individual lava flow units consistently dip to the ENE at shallow angles (rarely greater than 40°). Regionally low dips and simple fault repetitions allow individual lava units to be traced substantial distances laterally. Typically NNW-SSE striking normal faults down-throw lava units to the WSW with distinctive ferromanganiferous oxide sediments occurring in faulted hollows along the contact between the stratigraphically highest lava units and the overlying pelagic chalk sedimentary cover (see fig.7).

At Margi, orientated samples were collected from six sites, five of these being located in lava successions and a remaining single site in an interlava sediment horizon (see fig.7 and table 3). Four of the Five sites in the extrusive succession were located in a continuous sequence of vesicular-rich aphyric lava flow units at the very top of the Upper Pillow Lavas. As each individual flow was of an even thickness and could be traced laterally along strike for over 200m., the even top surface of each flow was assumed to adequately represent a palaeohorizontal plane. The strike and dip of each flow unit through this succession was virtually identical. In an attempt to average out the stratigraphic secular variation cycles, sites were chosen through this succession at 15m. stratigraphic intervals.

Whereas all lava samples were collected by drilling in situ, the interlava sediments were most effectively sampled by hand, by orientating each sample on a plane bedding surface. These sediments are fine to medium grained epiclastics which are composed predominantly of altered lava fragments in a muddy matrix, rich in iron and manganese (Boyle 1984). These sediments have been interpreted to represent the products of submarine erosion of local topographic highs during a hiatus in volcanism. These sediments are laterally continuous and dip consistently toward the north east at shallow angles (25° to 40°).

A final lava site was located in an olivine-phyric pillow lava flow unit some 30m. stratigraphically above the interlava sediment horizon. Six individual pillows were sampled by hand along a road section approximately 50m. long. This corresponds to a stratigraphic interval of approximately 15m. The bedding correction was based upon the mean attitude of phenocrystic-rich layers in each pillow. The plane of the palaeohorizontal was almost identical to that of the underlying sediments (see table 4).

Both lavas and sediments alike responded well to AF magnetic cleaning with the primary remanence vector being isolated in fields of less than 15mT (see fig.8). Magnetic cleaning resulted in a small reduction in the $\alpha_{95}$ at each site, although prior to cleaning, remanence vectors were already tightly clustered. As between-site scatter
Fig. 7 Distribution of major lava flow units at Margi, NE Troodos (see inset and fig. 2b for details of location). Based on detailed mapping by Boyle (1984). The location of individual sampling sites through the lava succession (see section AB) are shown.

Fig. 8 Representative AF demagnetisation plots of interlava sediment samples (ABM67B) and a pillow lava sample (PPM18E). Both clearly show an NRM dominated by a shallowly inclined westerly directed remanence component (Zijderveld plot conventions are described in appendix 1).

Fig. 9 Bedding corrected mean remanence directions (x) together with their associated $\alpha_{95}$ cone of confidence for six sampling sites in the extrusive succession at Margi. Representative Zijderveld plots are also presented. N=number of subsamples, *direction of prevailing geomagnetic field. Details of site location are shown in fig. 7. Lava flow units are not differentiated on map, although the disposition of umbers (black) and pelagic carbonates (offset boxes) are indicated.

Fig. 10 Declination error attributed to rotation about non-horizontal axes. Notice how declination error increases with bedding dip and plunge of rotation axis.

Table 3 Details of sampling procedures at sites around the village of Margi.

Table 4 Mean remanence directions prior to and subsequent to magnetic cleaning, for samples collected at sites adjacent to Margi village. N=number of subsamples.
STRUCTURAL RELATIONSHIPS AT MARGI

200m. supra-ophiolite sedimentary cover

Aphyric lava unit (C)

Olivine phyric lava unit (B)

Olivine and Pyroxene lava unit (A)

Interlava sediment horizon

Location of map area

1-6 Sampling sites

Dip of massive lava units and pelagic sediments

Lithological boundary

Normal faults

Fault dykes in A lava unit

Line of section

FIG. 7
G.8 DEMAGNETISATION OF INTERLAVA SEDIMENT (A) AND PILLOW LAVA (B) SAMPLE
FIG. 9 REMANENCE DIRECTIONS AT MARGI AFTER BEDDING CORRECTION
FIG. 10  

Axial plunge  

Declination error  

Bedding dip  

i.e. angle inclined axis makes with horizontal
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<th>SITE</th>
<th>SAMPLING METHOD</th>
<th>NO. OF SPECIMENS</th>
<th>NO. OF SUB-SAMPLES</th>
<th>PALAEOHORIZONTAL CRITERION</th>
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TABLE 3
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<th>MINC</th>
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<th>$K$</th>
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<td>15mT</td>
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<td>48.4</td>
<td>10.3</td>
<td>61.7</td>
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TABLE 4
in the massive flow units was negligible, the computed $\alpha_{95}$ for the locality mean was only fractionally greater than that recorded at each site. The mean remanence declination for the interlava sediment site (DEC=270.8, INC=48.4, $\alpha_{95}=10.3$) deviated away from that of the massive flow units (DEC=290.8, INC=26.7, $\alpha_{95}=12.3$). As the interlava sediment site was located in a fault block having a bedding strike closer to NW–SE than the stratigraphically higher massive flow units, this discrepancy in declination might be attributed to relative fault block rotation about non-horizontal axes.

2.11 A COMPARISON OF MEAN REMANENCE DIRECTIONS ALONG THE NORTHERN MARGIN OF TROODOS

2.11.1 SCATTER OF MEAN INCLINATION VALUES

As the Troodos ophiolitic crust was created over a short time interval (probably less than 10 Ma.) in the Late Cretaceous, intuitively one would expect to find no dramatic change in magnetic inclination across the strike of the complex, as this would imply a significant palaeolatitudinal drift of the Troodos oceanic spreading centre during the time of crustal genesis. Indeed, when a remanence vector can be accurately constrained relative to the palaeohorizontal (e.g. at Mitsero and Margi), a consistent inclination value of between 25° and 35° is obtained, corresponding to genesis of the Troodos crust at a latitude of between 20° and 25°N of the equator. If palaeosecular variation has been averaged out at each locality, anomalously high inclination values recorded at certain sites must necessarily be attributed to an incorrect estimate of the palaeohorizontal surface. Thus at Malounda, the apparent tilt of lava units does probably not correspond to the original horizontal surface, but rather the palaeoslope down which the lavas were extruded.

2.11.2 SCATTER OF MEAN DECLINATION VALUES

Like inclination values, there is no recorded swing in the mean remanence declination across the strike of the sheeted dyke complex on the northern margin of Troodos. In this study westerly declinations were found in Troodos oceanic crust exposed on the Akamas Peninsula to the west of the main Massif (see Chapter 4 section 4.5) and at the extreme east of the Massif at Ayia Anna (see table 2 and fig.6). Consistent westerly declinations have also been reported by Lauer and Barry (1976) at Apliki and Karavostas, and in this study at Malounda and Mitsero (see table 2 and fig.2). This implies that no bulk rotation of Troodos ocean crust was occurring at the time of active ridge volcanism and any discrepancies in the mean primary remanence
declination must necessarily be attributed to either real or apparent tectonic rotation of intracrustal blocks subsequent to crustal genesis.

If locality means along the northern margin of the Troodos ophiolite complex represent time averaged palaeomagnetic directions, it is valid to assume that originally, prior to structural tilting all remanence directions were co-linear. However, if structural tilting of a series of lava units at a locality occurred about an axis inclined to the strike of the palaeohorizontal surface, an application of a conventional bedding correction will inevitably lead to a declination error which can be misinterpreted as representing real tectonic rotation about a vertical axis. Although potential declination errors are virtually insignificant for gently dipping lava units (i.e. between 0 and 30°), significant errors (between 5 and 15°) can be introduced if successions dip more steeply, and tectonic tilting occurred about steeply inclined axes (see fig.10). As only gently dipping lava successions were sampled along the northern margin of the complex, only minor discrepancies in declinations can be accredited to application of an inappropriate bedding correction. Failure to allow for possible tectonic rotations about an inclined axis could account for differences in mean remanence declinations at a number of sites (Malounda, Mitsero, Margi, Sha and Ayia Anna), although these could equally well be interpreted to represent relative tectonic rotations about vertical axes.

An anomalous declination value at Analiondas is more difficult to interpret. Although it could conceivably represent a considerable rotation of a small intracrustal block in the horizontal plane, a more likely interpretation is that the palaeohorizontal surface was not correctly determined. At this particular locality olivine phyric pillow lavas dip steeply and radially away from what was probably a small pillow lava volcano. Original palaeohorizontals are notoriously difficult to determine in steep sided pillow flows that form topographic highs around eruptive centres.

2.12 CONCLUSIONS

The primary remanent magnetisation of the extrusive series along the northern margin of the Troodos ophiolite complex has a consistent westerly declination. Inclinations are both shallow and positive, implying that subsequent to formation during a Late Cretaceous normal polarity epoch, the ophiolite complex as a whole has suffered bulk rotation by 90° in an anticlockwise sense. Following a conventional structural correction, small but finite variations in the orientation of the mean declination are recorded at individual localities. These local discrepancies may be attributed to:

a) A declination error, accountable to tectonic rotations about inclined axes.
b) A failure to accurately determine the palaeohorizonatal at any one site.
c) Minor rotations of intracrustal blocks about vertical axes.
Importantly, as remanence declinations in crustal areas where dykes trend significantly away from the predominant NS orientation (e.g. Sha, see fig. 2) are similar to those recorded in areas of the crust where dykes are orientated approximately ns (e.g. Malounda, see fig.2), major discontinuities in dyke orientation cannot be attributed to post-intrusion tectonic rotations about vertical axes. Thus, the general absence of large scale fault block rotations in the horizontal plane along the northern part of the Massif would support the general complexity of dyke trends in this part of the Massif as being a primary feature of the original ocean floor crust.

This complexity in the spreading fabric, as defined by the orientation of sheeted dykes across the complex, would imply that the geometry of the zone of the crustal accretion at times during the spreading history of the ocean floor could not always have existed as a simple NS trending symmetrically spreading ridge, and crust now preserved on the east side of the Massif must have been created at spreading centres that were necessarily orientated at a considerable angle to the principal ridge crest. Furthermore, if crust in this part of the Massif can be subdivided into a number of domains within which dykes are orientated in parallel (Verosub and Moores 1981), it should now be possible to document in detail the migration history of the spreading ridge crest knowing that changes in dyke trend are a primary feature of the ocean floor crust. Changes in the configuration of the Troodos spreading ridge axis during crustal accretion is currently being studied by Robert Varga and Eldridge Moores (in press).
CHAPTER 3

THE PALAEOROTATION OF THE TROODOS OPHIOLITE COMPLEX

3.1 INTRODUCTION

As there is no systematic change in the declination of the magnetic remanence vector retained in the extrusive series across the strike of the sheeted dyke complex on the northern margin of the Troodos Massif, the tectonic rotation of the ocean crust must have been initiated at some time after crustal genesis. Obviously the age of the ocean crust itself (dated as Turonian, Blome and Irwin 1985) provides a lower limit on the timing of the rotation, but in order to constrain the event more precisely, the palaeomagnetic study was extended into the in situ Late Cretaceous to Recent sedimentary cover of the ophiolite complex. The aim of this investigation was to identify any systematic swing in the magnetic declination within the autochthonous supra-ophiolite sedimentary succession, as this must necessarily reflect any component of rotational motion within the horizontal plane of the underlying basement.

3.2 THE PRE-MIOCENE CIRCUM TROODOS SEDIMENTARY SUCCESSION

The pelagic sedimentary cover of the ophiolite complex has been described by Robertson and Hudson (1974). A composite sequence of Upper Cretaceous and Tertiary sediments that overlies the Massif is shown in fig.1. The first deposits to accumulate above the topographically irregular Troodos igneous basement were ferromanganiferous umbers, the precipitates of trace metal enriched hydrothermal fluids that circulated around the active spreading ridge axis. Locally on the northern margin of the ophiolite, major accumulations of these sediments (up to 35m.) occur in generally NS trending grabens and half-graben hollows; inherited structural features of the original extensional tectonic regime. Upwards, umbers are overlain by radiolarites, radiolarian mudstones and bentonitic clays, all of which are non-calcareous and contain well preserved radiolaria of Campanian age (Mantis 1970). All these sediments are ponded in scattered hollows above the stratigraphically highest lava units and show no marked systematic variation on a regional scale.

By contrast, later sediments vary laterally in thickness and composition across the island (see fig.2). On the northern, eastern and south eastern flanks of the Massif, the radiolarian sediments pass into scattered discontinuous deposits of bentonitic (illite-montmorillonitic) clays which fill the upper parts of hollows floored by umbers and radiolarian-bearing rocks. Immediately to the west of the Limassol Forest area these clays thicken and are overlain by the Moni melange (see chapter 5, section 5.8).
Fig. 1. Composite sequence through the Turonian-Miocene sedimentary cover to the Troodos ophiolite (after Robertson and Hudson 1974).

Fig. 2. Comparative sections through the Late Cretaceous-Tertiary sedimentary cover (after Robertson and Hudson 1974).

Fig. 3. Geological map of area south of Margi village to show relationships of metalliferous and pelagic sediments to underlying extrusives (after Boyle 1984). Cross section through umber pit shows how the oldest sediments tend to 'bow' into lava hollows (see inset). Sampling sites are also indicated.

Fig. 4. 'Orientated chip' method of sample collection.

Table 1 and 2. Sampling procedures and details of mean remanence directions prior to, and subsequent to, magnetic cleaning for umber and radiolarite samples collected at Margi. N=Number of subsamples. Radiolarites record both normal (N) and reversed (R) polarity directions.
Bioturbated organic rich marls

Miocene Pakhna Formation

Marls and chert free chalks

Massive chert free chalks (0-200m.)

Brilliant white chalks with cherts (0-250m.)

Pink / white finely bedded marls

Bentonitic clays (0-200m.)

Pink radiolarian cherts

Umber (0-30m.)

Pillow lavas

FIG. 1
COMPARATIVE SECTIONS OF UPPER CRETACEOUS - LOWER TERTIARY SUCCESSIONS

FIG. 2

Maastrichtian-Miocene pelagic chalks
Pink radiolarites and bentonitic clays
umberas
Lavas
1-7 Sampling sites

location of map area
Troodos Ophiolite
50 km.

For detailed inset see fig 2

Line of section
West
East

FIG. 3 SAMPLING SITES IN SEDIMENTARY COVER AT MARGI
Fig. 4

1cm diameter plastic disc

rock chip

plastic sample box
### Table 1

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<th>SAMPLING METHOD</th>
<th>NO. OF SITES</th>
<th>NO. OF SUBSAMPLES</th>
<th>NO. OF PILOT SAMPLES DEMAGNETISED</th>
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Table 2
unit, but are then concealed underneath Tertiary pelagic sediments that overlap directly onto the main Massif between Trimiklini and Pano Panayia (see chapter 1, fig.3). They reappear again in the Paphos District where they are interbedded with a varied succession of volcanioclastic sandstones and siltstones. In this area the succession can reach over 600m. in total thickness. The most convenient stratigraphical system divides the umbers and the radiolarian bearing rocks into the Peraphedi Formation, whilst the Kannaviou Formation comprises the bentonitic clay and volcanogenic sandstone units found in particular abundance in south west Cyprus.

Conformably overlying the Peraphedi Formation and Kannaviou Formation are the Lefkara Group pelagic sediments, which can be divided into a Lower, Middle and Upper units on the basis of combined lithological and micropalaeontological criteria (Wilson 1959, Gass 1960 and Mantis 1970). Biostratigraphical dating and correlation is based on fossil foraminifera associations (after Henson et al. 1949 and Mantis 1970).

The Maastrichtian aged Lower Lefkara marls and chalks rarely exceed 25m. in total thickness and are generally unsilicified. They outcrop discontinuously around the Massif either above Campanian aged sediments that have been ponded in topographic hollows, or directly above the extrusives in broad shallow troughs in the pillow lava surface. Locally to the south of the Limassol Forest block these Maastrichtian marl-chalks reach very much greater thicknesses and contain a high proportion of argillaceous detritus, presumably derived from the subjacent Moni melange unit (see chapter 5, section 5.8).

Immediately overlying Maastrichtian chalks are pelagic sediments of Upper Palaeocene age. Although there is apparently no evidence for a major hiatus in deposition, no Lower Palaeocene (Danian) microfossils have been recognised in the circum-Troodos succession. Conformably overlying the Lower Lefkara on the northern margin of the ophiolite complex, the Upper Palaeocene-Lower Eocene Middle Lefkara is a highly condensed sequence of chalks with bedded and nodular cherts. To the south of the ophiolite the sequence is represented by a thick deep water succession (>300m.) containing both pelagic and turbiditic components. Replacement cherts occurring at the base of the calciturbiditic units can often be traced for many hundreds of metres laterally. Although the succession is characteristically coloured brilliant white, on the northern and eastern margins of the ophiolite condensed Palaeocene chalks often display a strong pink pigmentation, presumably related to an enrichment in haematite.

The stratigraphically lower parts of the Upper Lefkara succession comprises massive, generally chert-free chalks of Middle to Late Eocene age. However, younger chalks of Oligocene to Early Miocene age are less well consolidated and lithologically more variable. Locally these chalks toward the top of the Upper Lefkara are chaotically slumped and it appears that at this time the Troodos Massif underwent differential uplift resulting in folding, slumping and erosion of the overlying sediments.
3.3 CHOICE OF SAMPLING LOCALITIES

As the volcanic basement to the Troodos Massif was not completely buried by sediments until Mid-Tertiary times, only a limited number of complete Campanian to Miocene sections exist around the periphery of the ophiolite complex. Many successions on the western and southern flanks of the Massif are deformed having been subjected to Tertiary tectonism and recent landslipping on Moni and Kannaviou clays and so cannot always be assumed to rest autochthonously on the underlying igneous basement. By contrast, Late Cretaceous and Tertiary sediments on the northern and eastern flanks are generally less deformed and dip only at shallow angles away from the uplifted core of the ophiolite. However, at many localities Middle and Upper Lefkara chalks overlap directly onto the Troodos pillow lavas, 'cutting-out' critical successions through the Upper Cretaceous and Lower Tertiary. Therefore, it is only at a limited number of localities that umbers and radiolarites of the Perapedhi Formation conformably pass into Maastrichtian chalks of the Lower Lefkara Formation. At these sites any early (i.e. Campanian-Maastrichtian) systematic swing in the direction of the primary remanence vector away from that retained in the underlying Troodos basement could be investigated.

3.4 STRUCTURAL RELATIONSHIP OF UMBERS AND RADIOLARITIES TO UPPER PILLOW LAVAS AT MARGI, NE TROODOS

At Margi, on the north east margin of the Massif, the stratigraphically highest extrusive units retain a stable magnetisation with a mean remanence direction of DEC=290.4, INC=20.4, α95=14.2. The umbers and radiolarites commonly occur in faulted depressions in the original lava surface, with the best exposed of these deposits located in a faulted outlier beneath bentonitic clays and Maastrichtian chalks, where mining activities have excavated a number of across-strike sections (see fig.3, section AB). The umber crops out over a distance of 600m. in an eastward facing half-graben in which the controlling NS trending normal fault can be traced southward over a distance of 4km. The umber accumulated in situ on an uneven lava surface that was inclined at 40° to the east, with its general ENE dip preventing umber from being exposed all around the outlier.

In section, the umbiferous sediments 'bow' into the half graben trough, with compaction of these sediments being manifested in inwardly dipping bedding laminations, ptygmatically folded mineral veins, compacted radiolaria and numerous slickensides close to irregular pillow bases (Boyle 1984). Vertical shortening by pure shear can be by as much as 50% (calculations based on tectonic flattening of radiolaria and folding of originally vertical veins). On account of the compaction of umbers,
sediment laminations close to the lava/umber contact parallel the top of the lava surface, but higher in the hollows bedding planes of the radiolarites and chalks dip progressively less and less steeply. As these sediments have been ponded into an elongate hollow, only bedding planes toward the centre of the deposit have a strike that parallels the axis of the half graben. A section across strike (running east-west) at the centre of this half graben deposit was considered appropriate for palaeomagnetic sampling.

3.5 SAMPLING PROCEDURES

The umber is a very homogeneous and fine grained rock which is distinctively coloured dark brown or black. Discontinuous fine bedding laminations are often well preserved. It is impractical to sample umbers using a standard diamond tipped drill bit as umbers become soft and clayey on contact with water, and coherent cores longer than 3cm. are virtually impossible to collect. Bits cooled by compressed air are more effective in cutting through the rock, but vibrations associated with the rotating shaft of the bit tend to cause multiple fracturing of each core.

The overlying pale pink coloured radiolarites are massively bedded (5-20cm.) and are pervasively cut by large scale joints. Like the umbers, drilling in situ is impractical, as sections in the umber pit are structurally unstable and the merest vibration can cause dangerous rock falls. Brittle fracturing of these siliceous rocks is also a common problem that hinders efficient sample collection.

In preference to sampling umbers and radiolarites by conventional drilling methods, small independently orientated 'rock-chips' were collected from the Margi umber pit. The sampling method involved sticking 1cm. diameter plastic discs onto a flat lying rock surface with a fast setting epoxy resin adhesive. These thin discs then provided an artificial flat surface which could then be orientated with both a sun compass and a magnetic compass (see fig.4). The attitude of each disc was then measured with an inclinometer. To remove each sample from the rock surface, material around each disc was carved away with non magnetic chisels to form an in situ pillar which could then be clipped off at the base. Brittle fracture of many pillars close to the disc surface slowed the sampling process down considerably. However, having retrieved a large proportion (>50%) of subsamples from the outcrop, the flat surface of each individual disc was then attached to the inside of a plastic box with the strike line on each disc located parallel to the side of the box. The size of each subsample was trimmed down to the appropriate dimensions of each box.

A total of 58 samples were collected from 5 sites located at 1m. stratigraphic intervals from the the base of the umbers to the top of the radiolarite succession. The tilt and strike of bedding planes at each site was measured by conventional methods.
3.6 ISOLATION OF THE PRIMARY REMANENCE DIRECTION.

The NRM of each sample was then measured with a 'Molspin' magnetometer, by spinning each box in six mutually perpendicular orientations. As the total magnetic remanence of each plastic box and disc (0.1mAm$^{-1}$) is negligible compared with that of either the umber (200mAm$^{-1}$) or radiolarite (120mAm$^{-1}$, see table 1 and table 2) lithologies, the intensity measured can be assumed to represent that of the specimen enclosed within the box. The sample was located centrally inside the plastic box so the fluxgate heads were consistently positioned at the same distance from the specimen during each spin.

Either one or two samples from each site were subjected to alternating field demagnetisation in progressively increasing peak fields. Typically, the procedure involved demagnetisation at 2mT incremental steps up to 22mT and thereafter at 4mT steps until the remanence became dominated by spurious ARM components or the intensity reached a level comparable with the sensitivity level of the magnetometer (0.2mAm$^{-1}$). Each NRM contained a significant soft component that was easily removed by low alternating fields (rarely greater than 15mT). Although variable in direction, this soft component was usually north directed and steep, and probably represented a viscous component acquired by low coercivity magnetic grains in the ambient geomagnetic field. The median destructive field (MDF) for umbers varied between 2.5mT and 5.2mT, and for radiolarites between 8.5mT and 20.4mT. Thus, in most cases the MDF showed that at least half of the NRM was carried by magnetic grains with coercivities of only a few mT or less. On average, 60% of the NRM of umbers was erased by 10mT alternating fields, and 80% of the NRM was erased by 20mT. This compares with the radiolarites, which on average lose only 40% their NRM in fields of 10mT, but have lost 70% of their original remanence in fields of 20mT.

After removal of the initial unstable component, each remanence vector was normally quite stable against further AF demagnetisation, although Zijderveld plots for umber pilot samples rarely show a perfect linear trace toward the origin. However, it is clear that the stability of the NRM is such that it can be cleaned to give a stable remanence that persists for a few tens of mT against AF cleaning. The inclination and declination of the stable remanent vector were usually clearly defined for each pilot specimen (fig.5). It was on the basis of the demagnetisation behaviour of these pilot specimens that remaining samples at each site were cleaned in an optimum field. Generally this field lay in the range 10-20mT.
Fig.5 Representative AF demagnetisation plots for umber (a) and radiolarite (b, c) samples. Both normal (b) and reversed (c) remanence directions are retained by radiolarites. Plots clearly show a remanence dominated by a shallowly inclined stable remanence vector (Zijderveld plot conventions are described in appendix 1).

Fig.6 Stereographic projection plot of umber NRM directions corrected to field (a) and cleaned remanence directions corrected to field and bedding (b). Mean declinations (MDEC), mean inclinations (MINC) and \( \alpha_{95} \) radius (A95) around mean remanence vector (X) are also given. Triangles represent remanence vectors that intersect the lower hemisphere (i.e. positive inclinations) while circles represent remanence vectors that intersect the upper hemisphere (i.e. negative inclinations).

Fig.7 Representative AF demagnetisation plots for youngest radiolarites. In both cases the NRM is clearly dominated by a stable remanence with a more northerly declination compared with that of the underlying lavas, umbers and radiolarites (see fig.5a). Zijderveld conventions are described in appendix 2.

Fig.8 IRM acquisition curves for umber (BM53A) and radiolarite (BM05A, BM41A and BM24A) samples. \( J_0 \) is induced magnetisation at 1000mT, while J represents induced magnetisation in external field. Curves show likely presence of magnetite and specular haematite in radiolarite samples with magnetite being the sole carrier in umber samples.
DEMAGNETISATION OF UMBER AND RADIOLARITE SAMPLES

FIG. 5
UMBER REMANENCE DIRECTIONS

FIG. 6

MARGI UMBERS NRM corrected to field

DEC = 319.9
INC = 37.2
A95 = 27.9

MARGI UMBERS DE corrected to field & bedding

DEC = 285.8
INC = 23.6
A95 = 26.6
FIG. 7 DEMAGNETISATION OF RADIOLARITE SAMPLES
FIG. 8

IRM ACQUISITION CURVES FOR UMBER

AND RADIOLARITE SAMPLES
3.6.1 COMPARISON OF NRM AND CLEANED REMANENCE DIRECTIONS

Field corrected NRM directions for umber samples cluster consistently in the north west quadrant on the stereographic projection, away from the ambient geomagnetic field direction (see fig.6a). Inclinations are both fairly steep and positive. After correcting for bedding tilt at each site the radius of the circle of confidence was fractionally reduced from 27.9 to 27.2. This supports an acquisition of a component of the NRM prior to the compaction and differential tilting of the umber sequence. After cleaning, the mean remanence direction moved sharply away from the present day field direction to a more westerly declination and a lower inclination. However, upon demagnetisation, scattering of declinations is only marginally improved with $\alpha_{95}$ being reduced from 27.2 to 26.6 (both after bedding correction, see table 2 and fig.6b). Curiously, recorded inclinations are surprisingly low, although declinations are comparable with those of the underlying lavas (see section 2.10).

By contrast, field corrected NRM directions for the radiolarites can be separated into two distinctive groups; those with northerly declinations and those with easterly declinations (see table 1). Both groups have mean remanence inclinations that are low and positive. After cleaning though, both mean vector directions move away from the present geomagnetic field toward diametrically opposite groups that are interpreted to represent antipodal normal and reversed polarity directions. Although inclinations are rather low, normal polarity remanence vectors (positive inclination, assuming an origin in the northern hemisphere) are characteristically orientated toward the WNW whilst reversed polarity vectors (negative inclination) are directed toward the ESE. Again, scatter of remanence directions is improved upon application of the bedding correction with $\alpha_{95}$ being reduced from 25.4 to 21.0 (after cleaning).

In conclusion, it appears that the NRM of both umbers and radiolarites lithologies at Margi includes an important stable component with an inclination that is significantly lower than that of the underlying lava units. The umbers are of consistent normal polarity, while the radiolarites preserve both normal and reversed remanent magnetisation directions (of 5 sites, 2 are normal, 3 are reversed). Declinations of the umbers and stratigraphically older radiolarites of normal polarity are comparable to those of the extrusive series. However, samples collected toward the very top of the succession (for example both BBM39A and BBM41A in fig.7 were collected from within the top 3m of the unit), apparently preserve more northerly declinations, although within-site scatter and a low sampling population makes any interpretation inconclusive.
3.7 MAGNETIC MINERALOGY

To identify the nature of the principal carriers of the remanence, isothermal remanent magnetisation (IRM) studies, Curie point determinations and thermal demagnetisation experiments were performed.

3.7.1 IRM STUDIES

The rate of acquisition of an IRM can provide an effective measure of the coercivity spectrum of the magnetic mineralogy in a rock sample (Dunlop 1972). These investigations are particularly versatile in that they permit an examination of the high coercivity magnetic fraction in any rock, a fraction which remains largely unaffected by alternating fields during standard demagnetisation studies. Although a high coercivity fraction may often also be investigated in thermal demagnetisation studies, high temperatures may cause irreversible changes in the magnetic mineralogy.

To investigate the magnetic acquisition properties of umbers and radiolarites, six samples (three of each lithology) were given isothermal remanent magnetisations in successively higher direct fields up to 1000mT. The IRM at this peak field is usually referred to as the saturation IRM \( J_{SAT} \), although samples containing haematite may not be completely saturated at this field. Single and multi-domained magnetite grains typically saturate in fields between 50mT and 200mT although single domain needle-shaped magnetite grains have coercivities of up to 300mT. Remanence components harder than 500mT may safely be attributed to haematite or goethite.

Following Dunlop (1972), coercivity spectra can be analysed rather arbitrarily into their 'soft' (0-100mT), 'intermediate' (100-300mT) and 'hard' (>300mT) coercivity fractions, which represent the approximate IRM contributions of magnetite, specular haematite and pigmentary haematite. IRM acquisition curves of umber samples are invariably characterised by a steep and steady initial increase followed by a drastic flattening-off of the slope as the curve reaches a saturation field of around 150mT or less (e.g. sample BM53A in fig.8). As the coercivity spectra of these samples is dominated by a soft coercivity fraction, the primary remanence carrier is interpreted to be magnetite. Curiously, although the iron hydroxide goethite is known to be abundant in these sediments (Boyle 1984), it apparently does not contribute significantly toward the remanence (typically, the coercivity spectrum of goethite-containing limestones typically covers a broad range up to and in excess of 5000mT in some cases. Typical IRM acquisition curves for goethite bearing limestones are described in Chapter 7, section 7.4.3.2).

Radiolarite samples however, display isothermal remanent magnetisation curves that characteristically increase rapidly up to 100mT but then approach a saturation
Fig. 9 Continuous heating of magnetite-bearing umber sample. Notice remanence is completely eliminated at a temperature of c. 550°C.

Fig. 10 Stepwise thermal demagnetisation of reversed polarity radiolarite sample (AMI03B), showing possible presence of three independent components. The primary remanence direction in this sample appears to be isolated in temperatures of between 350°C and 400°C.

Fig. 11 (a) shows stepwise thermal demagnetisation of saturated IRM acquired in a 1000mT field ($J_0$). In this sample the saturated IRM is reduced to less than 1% of its initial value at a temperature of 570°C. (b) shows normalised plots of $IRM_{SAT}$ (ratio of $IRM_{SAT}$ at $x^\circ C$ to $IRM_{SAT}$ prior to heating) and susceptibility (ratio of susceptibility at $x^\circ C$ to susceptibility prior to heating). These plots indicate that only a minor decrease in these parameters occurs between room temperature and 650°C, confirming that no major new magnetic phase is crystallising during heating.

Table 3 and 4. Sampling procedures and details of mean remanence directions prior to and subsequent to magnetic cleaning for pelagic chalk samples collected at Margi. N=Number of subsamples. Maastrichtian chalks clearly record both normal (N) and reversed (R) remanence directions.
FIG. 9 CONTINUOUS HEATING OF UMBER SAMPLE
FIG. 10 THERMAL DEMAGNETISATION OF CAMPANIAN RADIOLARITE SAMPLE
FIG. 11 STEPWISE THERMAL DEMAGNETISATION OF SATURATION IRM
<table>
<thead>
<tr>
<th>AGE</th>
<th>N</th>
<th>STRIKE</th>
<th>DIP</th>
<th>MDEC</th>
<th>MINC</th>
<th>$\alpha_{95}$</th>
<th>K</th>
<th>MEAN NRM INTENSITY mAm$^{-1}$</th>
<th>CLEANING FIELD</th>
<th>MDEC</th>
<th>MINC</th>
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<td>346.0</td>
<td>16.0</td>
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<td>41.2</td>
<td>10.3</td>
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<tr>
<td>MAAS.</td>
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<td>18.0</td>
<td>126.6</td>
<td>31.9</td>
<td>14.7</td>
<td>31.6</td>
<td>6.7</td>
<td>10mT</td>
<td>139.4</td>
<td>-1.1</td>
<td>11.7</td>
<td>47.5</td>
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<td>38.0</td>
<td>117.2</td>
<td>0.9</td>
<td>13.8</td>
<td>34.3</td>
<td>1.2</td>
<td>10mT</td>
<td>142.6</td>
<td>-18.6</td>
<td>8.7</td>
<td>87.0</td>
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<tr>
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<td>12</td>
<td>352.0</td>
<td>5.0</td>
<td>002.3</td>
<td>45.6</td>
<td>16.3</td>
<td>51.3</td>
<td>0.8</td>
<td>12mT</td>
<td>000.9</td>
<td>39.7</td>
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**TABLE 4**

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<th>SAMPLING METHOD</th>
<th>NO. OF SITES</th>
<th>NO. OF SAMPLES</th>
<th>POLARITY</th>
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<td>THINLY BEDDED</td>
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<td>13</td>
<td>N</td>
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<td>MAASTRICHTIAN</td>
<td>OFF-WHITE MARLS</td>
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<td>PALE PINK CHALCS</td>
<td>HAND</td>
<td>5</td>
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<td>R</td>
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<tr>
<td>UPPER LEFKARA</td>
<td>LOWER EOCENE</td>
<td>MASSIVE CHALCS</td>
<td>HAND</td>
<td>8</td>
<td>12</td>
<td>N</td>
</tr>
</tbody>
</table>

**TABLE 3**
value on a rather shallower path (samples BBM05A, BBM41A and BM24A in fig.8). These samples all saturated in fields of less than 300 mT. According to Dunlop (1972), this type of curve indicates that magnetite is the main carrier of the IRM, with a small additional contribution by a higher coercivity mineral which is usually interpreted to be haematite. It seems possible that this haematite contributes to the distinctive pink colouration of these rocks.

It is important to note that the IRM acquisition curves represent the total spectrum of coercivities in a sample, but do not necessarily indicate the relative contribution of each coercivity fraction to the NRM. Thus, it is possible that haematite is the major carrier of the remanence in radiolarite samples. However, considering the ease with which the bulk of the NRM was erased in fields of less than 200 mT, most of the NRM did not appear to be carried by single domain haematite grains which have coercivities of up to 5000 mT. Thus, both umbers and radiolarites have magnetite as the principal carrier of their remanence, with possibly only a minor contribution being associated with haematite grains in the radiolarite samples.

It is probable that magnetite occurs as detrital grains which have aligned themselves with the ambient magnetic field either during or shortly after deposition. If however, haematite does indeed contribute together with magnetite toward the NRM of the radiolarite samples, the fact that the remanence vector achieves a stable end point direction during demagnetisation, would strongly suggest that if haematite is secondary in origin it was acquired soon after the primary magnetisation. Therefore, it is assumed that if haematite is not of primary origin, it was created during diagenesis either by the oxidation of primary magnetic minerals like magnetite or by the precipitation in situ by percolating groundwater. If either is the case, the remanence of radiolarite samples retains an early secondary magnetisation that was superposed on a depositional remanence carried by detrital magnetic minerals. Consequently, the direction of magnetisation measured in the radiolarites may be a contribution of both primary and secondary magnetisation vectors which cannot be practically distinguished apart.

3.7.2 CURIE POINT ANALYSIS

Plots of saturation magnetisation against temperature were made on crushed radiolarite and umber samples in order to obtain Curie points of the principal magnetic minerals. Thermomagnetic analysis was carried out by heating the samples in air to 700°C at a rate of about 15°C/min. in a maximum field gradient of a 500 mT electromagnet. Using a standard magnetic translation balance, the induced magnetisation of a sample was measured by recording the magnitude of a current required to maintain a sample in steady position relative to a field with a steady
gradient. By monitoring the feedback current, a plot was made of the decay and growth of an induced magnetisation against temperature on an X-Y recorder.

Thermomagnetic curves for both umbers and radiolarites show well defined Curie points between 480° and 540°, corresponding to a magnetite carrier perhaps with minor amounts of titanium in the lattice (see fig.9). Although no haematite Curie point was discernible in the curves from these sites, this does not exclude its presence in smaller amounts than could be detectable. Also visible in the curves is an irreversible mineralogical change, resulting in a reduction of up to 50% in the room temperature value of the induced magnetisation after cooling. Such a change could be attributed to the inversion of maghaemite to haematite at 350°C (at higher temperatures if impure).

3.7.3 THERMAL DEMAGNETISATION OF NRM AND SATURATION IRM

As umber and radiolarite samples were orientated by applying a plastic disc onto the rock surface, this had to be removed before they could be effectively used for thermal demagnetisation purposes. On a limited number of samples (2 umber samples and 2 radiolarite samples) the orientation arrow marked on the disc could be accurately transferred to the rock surface and then the disc removed. Specimens were then systematically heated to successively higher temperatures to a maximum temperature of 650°C. At each 50°C incremental temperature stage, the specimens were cooled in field-free space within a set of Helmholtz coils. In order to further enhance the identification of the magnetic minerals present, stepwise thermal demagnetisation of a saturated IRM was also performed.

All samples showed a pronounced change in intensity on first heating (fig.10) which characteristically consisted of an initial decrease for reversely magnetised radiolarite samples and an increase for the remaining two normally magnetised samples. The intensity and direction of the remanence stabilised at 200-400°C after the initial removal of this soft component of magnetisation. The remanence decayed uniformly from this temperature to 500-600°C after which the magnetisation was rarely more than 1% of its initial NRM intensity. A decay in remanence intensity was associated with only minor directional changes, although after heating to above 600°C the directional scatter increased, giving random magnetisation directions in all four specimens. These directional changes were attributed to the acquisition of spurious thermoremanent magnetisation components (TRM) acquired during cooling.

On a qualitative basis, it appears that stable primary remanence directions isolated by thermal demagnetisation are indistinguishable from those defined in AF demagnetisation runs. Thus, if haematite carriers are significantly contributing to the NRM, then there does not appear to be a large angular discrepancy between the
remanence components carried by magnetite carriers and those carried by haematite grains. The implication of the thermal demagnetisation results is that if haematite grains retain a secondary CRM, this component is essentially in the same direction as the stable primary remanence component carried dominantly by magnetite. Thus, if haematisation did occur, it took place soon after detrital magnetite grains became aligned with the prevailing geomagnetic field.

Continuous thermal demagnetisation of a saturation magnetisation of saturation IRM in radiolarite and umber samples clearly show a smooth and gradual reduction in the magnetisation up to a maximum unblocking temperature of between $520^\circ c$ and $570^\circ c$ (see fig.11a). These curves are characteristic of Ti-poor magnetite bearing sediments and closely resemble curves obtained by Lowrie and Heller (1982) from Scaglia Rossa pelagic limestones. At each incremental heating step the specimens were once again exposed to a saturation field (1000mT) and their IRM and susceptibility measured (fig.11b). This latter technique was used to monitor any mineralogical and/or chemical changes that might have taken place during heating. As the saturation magnetisation and susceptibility values remained approximately constant up to $600^\circ c$, it was assumed that there was no major change in the magnetic mineralogy during heating.

3.8 SUMMARY OF MAGNETIC PROPERTIES OF UMBERS AND RADIOLARITES.

The NRM of umber samples from the Margi area include a stable remanence component which is interpreted to be carried exclusively by magnetite of detrital origin. By contrast, the radiolarites have a stable remanence that is probably carried by a combination of both haematite and magnetite. Considering the ease with which the NRM was erased by AF demagnetisation at each site, any haematite grains were probably of the multi-domain variety. It is clear also, that any diagenetic growth of haematite occurred soon after the acquisition of the primary remanence.

Umber samples are without exception of normal polarity, while both normal and reversed directions have been isolated from the overlying radiolarite lithologies. Mean primary remanence declinations for both umbers and radiolarites are, within the limits of resolution in the data, indistinguishable from those of the underlying lava flow units. Although the youngest radiolarites yield more northerly declinations. Further detailed sampling is required to establish whether these discrepancies are real. Importantly though, both umber and radiolarite lithologies record inclinations that are anomalously low when compared with reliable inclinations from the underlying lava flows. Possible causes for these discrepancies are considered in section 3.16.

3.9 REMANENCE OF PELAGIC CHALKS, MARGI
At Margiumber pit, gently east dipping radiolarites of the Perapedhi Formation pass conformably up into 1-2 metres of finely bedded pale green bentonitic clays. Both radiolarites and bentonites retain a diverse fauna of radiolaria including Dictyomitra multicostata, Crucella espartoensis and Patalibracchium lawsoni, all regarded as being diagnostic of the Campanian (Mantis 1970).

The bentonitic clays are directly overlain by a thin (<10m.) succession of marls, which show a prominent colour gradation from salmon pink at the base of the sequence, to an off-white colour toward the top. These chert-free marls are considered to represent Globotruncana-bearing Lower Lefkara chalks of Maastrichtian age. Elsewhere in the Margi area, these Maastrichtian sediments outcrop discontinuously either above radiolarites in down-faulted half grabens or immediately overlying the highest lava units in broad topographic hollows in the lava surface. More typically, the igneous rocks are directly overlain by either the chalk-cherts of the Middle Lefkara or the more massive chalks of the Upper Lefkara. In the immediate vicinity of the igneous rocks the strata have dips between 30° and 45°, while away from the contact the regional dip is between 5° and 15°. Thus, small structural basins in the extrusive lava topography provide appropriate sites for palaeomagnetic "fold test" investigations.

Sampling sites through the Maastrichtian to Lower Miocene pelagic succession are located on a sketch map of the Margi area in fig.3. Only gently inclined, undeformed sequences away from major fault scarps were sampled. The ages of the sediments were determined on lithological criteria (see section 3.2). At each site, three or four independently orientated handsamples were collected, and from each sample, 2-6 subsamples were obtained (see tables 3 and 4). To avoid unnecessary abrasion of drill bits, massive cherts were preferentially not sampled.

As the NRM intensity of these pelagic sediments was low (between 10 and 0.1mA m⁻¹), and the noise level of the spinner magnetometer is comparatively high (0.1mA m⁻¹), remanence measurements were most effectively made with a cryogenic magnetometer (noise level 5 x 10⁻² mA m² total moment). The cryogenic magnetometer consists of three independent S.Q.I.D. systems (Superconducting Quantum Interference Device), which are arranged so as to provide an output voltage proportional to the three orthogonal magnetisation components of a sample. Theoretically therefore, the remanence intensity and orientation can thus be measured in one operation, but usually either two or three separate measurements were made with the sample in different orientations to cancel out any remanence carried by the sample holder. On a number of occasions one of the SQIDs was not operational, and when this happened it was possible to measure only two of the three orthogonal components in one reading. Usually 2 or 4 separate readings were then used to determine the remanent magnetisation of the sample. However, as the vertical SQID was always operational, each specimen was always inserted into the magnetometer with
Fig.12 Design of sample holder for cryogenic magnetometer.

Fig.13 Representative AF demagnetisation plots for reversed (a) and normal (b) chalks of the Lefkara Formation. Spurious ARM components acquired in higher fields (i.e. >60mT) prevented further AF treatment. Zijderveld plot conventions are described in appendix 1.

Fig.14 Stepwise thermal demagnetisation of NRM; (a) is normal polarity sample while (b) is reversed. After the initial removal of a low blocking temperature component, both samples show marked stability in remanence between 200°C and 400°C, with the stable remanence component only being removed in temperatures of >600°C for (a) and >400°C for (b).

Fig.15 (a) Partially and (b) fully overlapping coercivity spectra for two reversed polarity chalk samples. Both samples show a characteristic initial increase in remanence intensity during cleaning due to the rapid removal of a steep northwardly directed remanence component in low alternating fields.
FIG. 12

2.5 cm.

FIG. 12

sample rotates clockwise
in magnetometer

x direction of sample
(see chapter 2 fig. 3b)

SAMPLE HOLDER DESIGN
FIG. 13 AF DEMAGNETISATION OF LEFKARA CHALKS
THERMAL DEMAGNETISATION OF LEFKARA CHALKS

FIG. 14 a)

FIG. 14 b)
FIG. 15 OVERLAPPING COERCIVITY SPECTRA
its long axis in a vertical orientation. The design of the sample holder is illustrated in fig. 12.

The sample holder magnetisation varied in intensity with use, and it was clear that it was susceptible to acquiring a viscous magnetisation; perhaps when the holder was located close to the entrance of the mu-metal shielding around the cryogenic system. The sample holder was therefore demagnetised at regular intervals and left in a zero field when not in use. Periodically the sample holder was cleaned in a ultrasonic cleaner, to remove any magnetic particles that might have been scraped off specimens as they were inserted into the holder. Taking these precautions, the remanence of the sample holder could be kept down to less than $5 \times 10^{-2} \text{mAm}^{-2}$, which after three measurements averaged out to around $12.5 \times 10^{-6} \text{mAm}^{-2}$. Samples with intensities of less than $0.1 \text{mAm}^{-1}$, were measured twice and the average direction found. In the majority of cases the two independent results were very close (i.e. to within $\pm 2^\circ$), demonstrating the precision of the measuring procedure.

### 3.10 STABILITY OF THE REMANENCE

The intensities of the NRM were generally weak, ranging from $30.0\text{mAm}^{-1}$ for Maastrichtian aged Lower Lefkara chalks with a strong pink pigmentation to $0.1\text{mAm}^{-1}$ for younger brilliant white chalks of the Middle and Upper Lefkara. All samples retained an initial remanence intensity of at least an order of magnitude above the noise level of the cryogenic magnetometer. Thus a remanence of intensity $0.1\text{mAm}^{-1}$ could be effectively measured with less than 2% standard error in intensity and with directional repeatability within $1^\circ$.

Following storage of samples in field free space for a period of a few months, the stability of the NRM of each specimen was investigated using standard AF demagnetisation techniques (see chapter 2, section 2.6). At least two pilot samples from each hand sample were subjected to demagnetisation fields at $2\text{mT}$ incremental steps up to $20\text{mT}$ (occasionally only up to $10\text{mT}$) and thereafter at $4\text{mT}$ steps up to $60\text{mT}$. Most samples showed a pronounced change in intensity on initial demagnetisation, which was represented by an increase for reversely magnetised samples (see fig.13a) and a complementary decrease for normally magnetised samples (see fig.13b), presumably reflecting the removal of a soft normal component of magnetisation. For the majority of samples the intensity and direction of magnetic remanence stabilised after demagnetisation in fields greater than $12.5\text{mT}$. The remanence then decayed uniformly from there to $60\text{mT}$ with only slight directional scatter in most cases.

In peak fields of more than $60\text{mT}$, anhysteritic remanent magnetisations (ARM) became an increasingly serious factor and this made it impractical to continue
Fig.16 Representative AF demagnetisation plots for pink Maastrichtian chalk sample (BBM56A). Notice low median destructive field (MDF) of NRM and the failure of the remanence vector to reach a stable end-point configuration.

Fig.17 Bedding corrected mean remanence direction (X) together with their associated $\alpha_{95}$ cone of confidence for five sampling sites within the pelagic succession at Margi. Representative Zijderveld plots are also presented; N=Number of subsamples, *=direction of the prevailing geomagnetic field. Details of site location are shown in Fig.3. Lava flow units are not differentiated, although disposition of umbers (in black) and pelagic carbonates (offset boxes) are indicated. Reversed polarity mean remanence vectors are represented by circles while normal polarity directions are represented by crosses.

Fig.18 Fold test at site in Upper Lefkara chalks, showing how field corrected remanence vectors for individual subsamples are scattered (a), but grouping improves upon application of a bedding coorection (b). Notice the reduction in the radius of the cone of confidence ($A_{95}$).
FIG. 16  AF DEMAGNETISATION OF BASAL PELAGIC CHALK SAMPLE
FIG. 17 REMANENCE DIRECTIONS OF PELAGIC CHALKS AT MARGI
FIG. 18
FOLD TEST ON UPPER LEFKARA CHALKS
demagnetising in higher fields. In preference, a representative selection of ten samples were subjected to thermal demagnetisation treatment in 50°C incremental steps up to 550°C and then at 25°C steps up to 700°C. In all cases, after the removal of a minor secondary component, the remanence continued to decay without perceptible change in the remanence direction, with many samples retaining a measurable remanence even at temperatures of over 600°C (see fig.14). Thus, the inclination and the declination of the stable remanence vector could be clearly defined for each pilot sample specimen by studying representative pairs of two component Zijderveld plots. A primary direction was identified as a line segment moving toward the origin at successively higher field strengths or temperatures.

Remaining subsamples from each handsample were then demagnetised at a field strength 5mT above that at which the primary remanence component was recognised as singularly contributing toward the remanence. As the coercivity spectra of the secondary and primary remanence components typically overlapped either partially (see fig.15a) or fully (see fig.15b), this field was often greater than 25mT, but typically ranged between 15 and 20mT. No stable remanence vector could be defined for the bright pink chalk specimens collected from the Maastrichtian succession that immediately overlies the Kannaviou Formation bentonitic clays (see fig.16). Compared with other chalks which typically have a median destructive field of between 20mT and 50mT, these chalks have a low MDF (typically <10mT) with at least 90% of the original NRM being erased in alternating fields of less than 12.5mT. NRM intensities are also characteristically rather higher (between 6.5 and 26.7 mAm⁻¹) than stratigraphically higher off-white Maastrichtian chalks. Field corrected NRM directions cluster about the axial geocentric dipole direction, moving westward and toward lower inclinations upon progressive demagnetisation. Thus, it is clear that the NRM is dominated by a lower coercivity component acquired in the recent geomagnetic field. As the remanence vector never achieves a stable end point configuration, the coercivity spectrum of the viscous component presumably overlaps with that of a harder, probably primary component, and it consequently always contributes to the remanence. The primary remanence is interpreted to include a reversed polarity component as the bedding corrected remanence vector isolated in higher alternating fields characteristically intersects the upper hemisphere on the stereographic projection and therefore records a negative inclination. As the direction of the primary remanence component for these samples could not be readily defined, data from two sites (15 subsamples) were rejected.

With the exception of these basal Maastrichtian chalks, reliable mean primary remanence directions were defined for each of the remaining sites in the Margi area (see fig.17 and table 3). Off-white and pale pink Maastrichtian and Palaeocene aged chalks convincingly record diametrically opposite normal and reversed polarity directions that
consistently group away from the present day field direction. By contrast, sampled Eocene and Oligocene chalks are without exception of normal polarity and record declinations close to north, with inclinations being characteristically lower than that of the prevailing magnetic field.

Magnetic cleaning improved the grouping of the characteristic magnetic remanence directions, with Maastrichtian and Palaeocene chalks giving a mean direction distinctively away from the present day field direction even before application of a tilt direction. This implies that the isolated stable remanence direction is quite unrelated to the orientation of the present earth's field. Application of a classical fold test (Graham 1949) to samples collected at two sites in the Upper Lefkara (see fig.18) establishes that the characteristic magnetisation was acquired prior to tectonic or syn-sedimentary folding. In the case of one of these sites to the WNW of Margi village, gentle flexuring of the Upper Lefkara massive chalks can be attributed to differential compaction of sediments within broad topographic hollow in the lava surface, and thus a positive fold test implies that the remanent magnetisation is dominated by a component acquired very soon after the sediment was deposited. The origin of the magnetisation in these chalks is discussed in section 3.13.

On the basis of straigraphical age, mean remanence vectors were then computed for Maastrichtian (Lower Lefkara), Palaeocene and Eocene (Middle Lefkara) and Oligocene (Upper Lefkara) chalks. By inverting all reversed polarity directions back through the origin all palaeovectors could be referred to the westerly declination of the normal polarity Troodos extrusive series. Mean remanence directions for each epoch are summarised in table 4.

Table 4 indicates that stratigraphically younger sediments in the supra-ophiolite succession systematically retain more northerly declination values. Assuming an axial geocentric dipole model for the generation of the earth's magnetic field and a primary time-averaged origin for the remanent magnetisation, magnetic declinations in the autochthonous sedimentary cover must necessarily reflect rotational motions in the horizontal plane of the underlying basement. Thus, this systematic swing in the remanence declination through the pelagic succession can be interpreted as representing an anticlockwise rotation of the basement at the time of sediment deposition.

3.11 ADDITIONAL SAMPLING SITES IN THE CIRCUM-TROODOS PELAGIC SUCCESSION

It is not appropriate to draw conclusions on the palaeorotation history of the Troodos ocean floor crust from the palaeomagnetic data compiled from a single locality. Thus, to complement this detailed palaeomagnetic study at Margi, Late Cretaceous and Tertiary pelagic sediments were sampled at localities around the periphery of the
ophiolite complex (see fig.19 and table 5). At each locality, sites were located in shallowly inclined succession away from zones of local tectonic disturbance. This was a particularly important consideration on the SE, S and SW flanks of the ophiolite complex where the circum-Troodos sedimentary cover has frequently been subjected to intense late Tertiary and Quaternary deformation. Areas of recent landslipping were avoided.

Samples were collected and prepared for measurement following standard palaeomagnetic procedures outlined above (see section 3.9). At least two subsamples from each site were then subjected to detailed AF demagnetisation, with the remaining subsamples at a given site being demagnetised at a critical field, if a stable primary remanence vector could be effectively recognised on pairs of two component Zijderveld plots (see section 3.10). Locality means were then computed (table 6) from compiled site means and remanence directions were plotted on stereographic projections. Individual remanence directions together with their associated \( \alpha_{95} \) circle of confidence, after correcting for bedding are illustrated in fig.19. Representative Zijderveld plots are also presented.

Typical of the behaviour of bright pink chalks at Margi, similar sediments located immediately above either the radiolarites or the stratigraphically highest lava flow units failed to reach a stable end point vector during AF cleaning. Data from these sites (e.g. Analiondas) could not be included in the determination of a locality mean. In addition to these data, selected data from other sites were rejected:

1/ Ayia Anna and Kalavassos (see fig.23).

Massive, brilliant white Upper Lefkara chalks at these localities retain a negligible remanence (between 0.1 and 0.05mAm\(^{-1}\)) which is only marginally greater than the noise level of the ScT cryogenic magnetometer (5 x 10\(^{-2}\)Am\(^{-2}\)). As instrument noise contributes significantly toward the SQID voltage output during measurement, NRM directions are highly scattered with an associated large \( \alpha_{95} \). Although a series of pilot demagnetisation runs were attempted, no stable remanence vector could be isolated before the remanence intensity was reduced to below the sensitivity limit of the magnetometer, and it was considered impractical to attempt a 'blanket' demagnetisation of the remaining sample population at each of these sites.

2/ Pano Lefkara succession (see fig.23).

The 400m. thick Pano Lefkara-Skarinou road section on the SE margin of the ophiolite complex is one of the best exposed sequences through the Middle Lefkara chalk-chert succession. Pale pink Upper Palaeocene chalks rest directly on pillow lavas with a surface dipping unconformably to the SE at 5-15\(^{\circ}\). On the basis of foraminiferal ages, the stratigraphic position along the extensive road section has been determined accurately by Mantis (1972). To complement this study and a preliminary palaeomagnetic study by Schoenharting and Abrahamsen (1982, see chapter 1 section 48.
Fig.19 Bedding corrected mean remanence direction (X) together with their associated $\alpha_{95}$ cone of confidence for eight sampling localities in the supra-ophiolitic pelagic succession. Representative Zijderveld plots are also presented. N=Number of subsamples, *=direction of prevailing geomagnetic field. If cone of confidence intersects upper hemisphere it is dotted (reversed polarity mean remanence vectors are represented by circles, while normal polarity directions are represented by crosses).

Fig.20 Representative AF demagnetisation plot of Middle Lefkara brilliant white chalk sample. Although the characteristic NRM intensity and MDF are both relatively high, at fields of greater than 20mT the remanence becomes dominated by spurious ARM components and no stable remanence component can be defined.

Table 5 and 6 Sampling procedures and details of mean remanence directions prior to, and subsequent to magnetic cleaning for Lefkara pelagic chalks collected around the periphery of the ophiolite complex. MAAS.=Maastrichtian, PAL.=Palaeocene, N=Normal polarity, R=Reversed polarity.

Fig.21 IRM acquisition curves for 12 Lefkara chalk samples. $J_0$ is induced magnetisation at 1000mT, while J represents induced magnetisation in external field. Curves show likely presence of both magnetite and specular haematite in all samples although Lower Lefkara chalks (BM56A, BM57B, MI01C and BM56D) have a rather lower coercivity spectra range. Many samples remain unsaturated even in fields of 1000mT (e.g. HL16E and HL192A).
DEMAGNETISATION OF UNSTABLE MIDDLE LEFKARA CHALK SAMPLE

Y VS Z

X VS Y

NRM = 0.67 mAtm$^{-1}$

HDF = 215.1 oe

FIELD OE.
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TABLE 5
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TABLE 6
FIG. 21
IRM ACQUISITION CURVES FOR LEFKARA CHALK SAMPLES
1.3.2), 3-5 handsamples were collected at approximately 40m. stratigraphic intervals from the very bottom of the sequence WNW of Pano Lefkara to the top near Skarinou.

After cleaning, pink Palaeocene chalks collected from the lower 80-100 m. of the sequence record both normal and reversed polarity directions (see fig.19), while brilliant white massive chert free chalks from the top of the Middle Lefkara (uppermost 100m. of the succession) are represented by only normal polarity directions. After correcting for bedding, inclinations for these Palaeocene chalks (MINC=-12.4 for reversed and MINC=29.3 for normal polarity sites) are consistently lower than those determined for the underlying Troodos extrusives (MINC=36) and the younger chalks of the Middle Lefkara (MINC=32.8). All of these mean inclinations are significantly lower than the inclination of the present day magnetic field over Cyprus which is inclined at 54.5° to the horizontal.

By contrast, the brilliant white chalks of the Palaeocene and Lower Eocene chalk-chert unit that represent the central portion of the Middle Lefkara succession, characteristically often failed to reach a stable remanence direction during progressive AF demagnetisation, and thus could not be included in the computation of a Formation mean. Typically, the number of demagnetisation steps was limited by the original NRM intensity of each pilot sample. Samples with low NRM intensities (<0.2mAm⁻¹) showed poor stability and this reflects the lack of precision in measurement rather than instability of the remanence. Other pilot samples with higher NRM intensities (>10mAm⁻¹) were dominated by significant secondary viscous components which contributed to the remanent magnetisation even after demagnetisation in fields of over 50mT. Representative orthogonal projection plots indicated that remanent magnetisation vectors tracked toward diametrically opposite normal and reversed directions, but failed to reach a stable end point (see fig.20). Thus, in these cases the primary remanent direction could not be isolated before the remanence was dominated by spurious ARM components. On the basis of 18 pilot demagnetisation curves, 42 subsamples (representing 15 handsamples collected from 7 sites) were considered to be unsuitable in the calculation of the Formation mean direction. These samples were rejected from the sample population, and the remaining site mean directions from 6 sites were considered significant and therefore could be included in the determination of the Formation mean.

3.12 THE MAGNETIC MINERALOGY OF THE CIRCUM-TROODOS PELAGIC CHALKS

The colour of the Lefkara chalks varies from dark red (basal Maastrichtian) to delicate pink (Upper Palaeocene) to off-white (Maastrichtian) and brilliant white (Palaeocene, Eocene and Oligocene chalks). It is assumed that the colour of these sediments can be attributed to the relative abundance of pigmentary haematite. If a
component of the NRM is carried by haematite, it is considered unlikely that the haematite will retain a primary remanence, as haematite probably forms either 1/ by oxidation of primary magnetic minerals such as magnetite or 2/ by in situ chemical precipitation from percolating groundwater. In both cases a secondary CRM may be superposed on a primary depositional remanence carried by detrital magnetite carriers, and thus it is important to establish whether the stable magnetisation retained by the Lefkara chalks is primarily carried by haematite grains of secondary origin. If however, haematite represents a hydrogenous component in these sediments, it is possible that haematite does retain a primary remanence.

3.12.1 RATE OF IRM ACQUISITION

In order to determine the coercivity spectrum of the magnetic minerals in the Lefkara chalks, 20 subsamples were subjected to successively higher direct fields between the poles of an electromagnet. The shape of the IRM acquisition curve and the field at which saturation is reached define the total coercivity spectrum of the sample (see section 3.7.1).

IRM acquisition curves for all the limestone samples investigated are characterised by a steep initial increase in the induced magnetisation followed by a drastic flattening-off of the slope at fields of 0.3T or less (see fig.21). The initial part of the coercivity spectrum is interpreted to be due to magnetite and is invariably present even when magnetic minerals with harder coercivity are also present. As the coercivity of single domain magnetite grains is controlled by their shape and reaches an extreme value of 300mT for needle shaped grains, only samples BM56A, BM56D, BM57B and MI01C retain a magnetic mineralogy dominated by magnetite. Surprisingly, these samples were all dark pink Maastrichtian chalks from the Lower Lefkara and thus despite their strong red pigmentation, high coercivity (>300mT) haematite grains do not apparently contribute significantly toward the remanence of these chalks. The presence of magnetite as the principal carrier in these sediments is compatible with their characteristic low MDF (rarely exceeding 5mT) during AF demagnetisation and their susceptibility toward acquiring low coercivity secondary components (see section 3.10).

The remaining samples have grain coercivities that are distributed more widely between 0 and 1000mT, with the spectrum approaching zero as the saturation remanence is reached in fields above 500mT. Although the initial steady increase in IRM acquisition (represented by a high value of $\delta J_r / \delta H$) can be directly attributed to the presence of magnetite carriers, it is very likely that the harder coercivity magnetic mineral in the remaining samples is specular haematite. The hardest part of the remanence (>500mT) may be attributed to pigmentary haematite or goethite. Thus,
the character of the majority of many of the IRM curves indicates that the magnetic mineralogy consists of both magnetite and haematite.

IRM acquisition curves indicate that the total spectrum of coercivities in a sample but not necessarily the fraction carrying the NRM. Therefore, although IRM acquisition curves for many chalk samples show an abundance of haematite in these rocks, but this does not necessarily mean that it contributes significantly toward the NRM, which could theoretically be largely carried by a small original magnetite fraction. It is also important to note that the saturation magnetisation of haematite is less than 1% of that of magnetite, so that an induced IRM will be two orders of magnitude greater for magnetite than it is for haematite.

The most direct way to identify the ferromagnetic content of a rock is probably by investigating the thermomagnetic behaviour of strong field induced magnetisations (saturation IRM) and the determination of Curie points by extraction of magnetic minerals from the rock. However, as the concentration of magnetic carriers in chalk samples is so low, large volumes of rock are required to extract a suitable quantity of magnetic grains for Curie point analysis. As earlier extraction experiments on limestones with relatively higher magnetic intensities than the Cyprus chalk samples have been performed with only limited success (Lowrie and Alvarez 1975, Lowrie and Heller 1982), similar extractions were not attempted on the Cyprus chalk samples.

### 3.12.2 THERMOMAGNETIC BEHAVIOR OF SATURATED IRM

In order to enhance the identification of the magnetic carriers of the Lefkara chalks, the saturation IRM was subjected to progressive thermal demagnetisation treatment (see fig.22). Eight specimens, four from the Maastrichtian Lower Lefkara and four from the Palaeocene-Eocene Middle Lefkara were exposed to a field of 1000mT only once and stepwise thermally demagnetised at 100°C incremental steps up to 500°C and thereafter at 50°C steps up to 700°C. In order to monitor any mineralogical changes that might have taken place during heating the samples were once again exposed to a 1000mT field after each demagnetisation step and their IRM measured.

The samples all display a regular decline in IRM to 300-400°C above which many show a minor increase or change in slope before declining again rapidly to 650-700°C. The rapid decline in IRM between room temperature and 400-500°C is attributed to the presence of magnetite, whilst the slower decline above this temperature to 650°C indicates the presence of haematite. If the remanence components retained by magnetite and haematite grains are anti-parallel, thermal demagnetisation of the IRM reduces the combined moment to a minimum when the two components are equal and then rises to a maximum when the maximum unblocking temperature of magnetite is reached, above which only the haematite component remains. Thus a
minor increase in the intensity between 500°C and 580°C (maximum unblocking temperature of magnetite) can be directly attributed to the presence of magnetite. In sample CMK04A the magnetite IRM is completely lost at c.580°C, the temperature representing the maximum unblocking temperature of magnetite carriers. A Curie point of 580°C corresponds to that of pure magnetite while lower Curie points of between 400°C and 500°C for other samples (estimated by extrapolating the steeper segments of the IRM acquisition curve back to the temperature axis) corresponds to the presence of polydomainal titanomagnetites and/or titanomaghaemites. This is compatible with the rather low values of the MDF for many pelagic chalk samples (see section 3.10).

Importantly, the iron oxyhydroxide goethite (α-FeOOH), is a common constituent of many sedimentary rocks, formed by the weathering of iron minerals or by the direct precipitation from iron-bearing solutions appears not to be an important carrier in these pelagic sediments. As the maximum blocking temperature of goethite lies in the range 50-90°C, the continuous thermal demagnetisation of an IRM retained by a goethite bearing specimen is represented by a dramatic decrease in the remanence intensity at temperatures well below 100°C (see chapter 7, section 7.4.3.2). This is also compatible with the form of the saturation magnetisation curves during thermal demagnetisation (fig.22). With the exception of a single specimen (CMK07C) which shows a slight progressive increase in the IRM saturation magnetisation up to 550°C, the remaining samples display a notable decrease over the same temperature range. If goethite had been a principal carrier, the saturation magnetisation curve would have demonstrably shown an appreciable increase at temperatures above 300°C, the temperature at which goethite characteristically dehydrates to ferromagnetic haematite.

3.13 SUMMARY OF MAGNETIC PROPERTIES AND DIRECTION OF CIRCUM-TROODOS PELAGIC CHALKS

Like the underlying radiolarites, the remanence of the majority of chalk samples investigated is carried by a combination of both haematite and magnetite. Pink Maastrichtian chalks at the base of the sequence have an NRM dominated by magnetite carriers. As the remanence retained by haematite carriers is similar in direction to that carried by high coercivity magnetite carriers, it is clear that any diagenetic growth of haematite occurred soon after the acquisition of the primary remanence. A positive fold test at a number of localities indicates that the primary remanence was acquired prior to tectonic tilting. Goethite appears not to be an important carrier of remanence in these sediments.

Following blanket demagnetisation at appropriate field strengths, Formation mean directions were computed by compiling data from all localities in the
Fig. 22 (a, lower axes) shows stepwise thermal demagnetisation of saturated IRM acquired in an external 1000mT field ($J_0$). Samples MI16A, CMK04A, AN03C and KA12A are of Maastrichtian age (Lower Lefkara) while HL16G, HL163C, CMK07C and HL42B are of Palaeocene-Eocene age (Middle Lefkara). (b, upper axes) shows normalised plots of $IRM_{SAT}$ (ratio of $IRM_{SAT}$ at $x^0c$ to $IRM_{SAT}$ prior to heating) and susceptibility (ratio of susceptibility at $x^0c$ to susceptibility prior to heating). For all samples only a minor decrease or increase in these ratios occurs between room temperature and 650°C, confirming that no major new magnetic phase is crystallising during stepwise heating.

Fig. 23 Location of sampling sites within the Miocene and Pliocene sedimentary cover to the ophiolite complex. Stippled area marks extent of post-Lefkara sediments on the southern margin of the Troodos Complex.

Fig. 24 Representative AF demagnetisation plots of (a) ophiolite derived clastic sample collected at Tokhni and (b) Pliocene marl collected in the Polis graben. Both clearly show an NRM dominated by a stable steeply dipping northwardly directed remanence component. Complementary IRM acquisition curves show that this remanence is probably carried at least partly by low coercivity magnetite grains.
FIG. 22. STEPWISE THERMAL DEMAGNETISATION OF SATURATION IRM.
DISPOSITION OF POST-LOWER MICOCENE SEDIMENTS TO THE SOUTH OF TROODOS

(S) SAMPLING SITES

FIG. 23
AF DEMAGNETISATION PLOTS AND IRM ACQUISITION CURVES FOR
a) OPHIOLITE DERIVED SEDIMENT AND b) PLIOcene MARL

FIG. 24

CTT 270
NRM = 64.79

CTT 270
NRM = 89.1

CPL 01A
NRM = 33.82

CPL 01A
NRM = 109.9
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TABLE 7
Table 8. Summary of available palaeomagnetic data from the Troodos Complex and its in situ sedimentary cover (all data represent cleaned remanence directions).

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TR = Troodos Pillow Lavas (north Troodos, Akamas and Southern Limassol Forest Areas)

AK = Akamas

UM = Umbers (north Troodos and southern Limassol Forest areas)

RA = Radiolarites (North Troodos)

PE = Pelagic chalks

N = number of localities, n = number of samples, Dec = declination, Inc = inclination, $\alpha_{95}$ = 95% radius of area of confidence, No = normal polarity, Re = reversed polarity

*At this locality mean declination of radiolarites is within $5^\circ$ of mean declination of underlying lavas. Reversed polarity sample directions have been inverted through the origin.
circum-Troodos pelagic succession (see table 8). Cleaned remanence directions for the Maastrichtian-Palaeocene chalks were conveniently divided into two populations, each representing complementary antipodal normal and reversed polarity directions. All Eocene sampling localities were of normal polarity. The declinations and inclinations of the Formation means could then be effectively compared with the mean remanence vectors determined for the main ophiolite complex and the Campanian sedimentary successions immediately overlying the ocean floor crust. Following the analysis of remanence directions retained in the Neogene sedimentary cover, the implications for the systematic changes in declination through the supra-ophiolite cover are discussed in section 3.16.

3.14 NEOGENE SEDIMENTARY COVER TO THE TROODOS OPHIOLITE COMPLEX

Locally, towards the top of the Upper Lefkara succession (Lower to early Middle Miocene), chalks are characteristically chaotically deformed having been subjected to synsedimentary slumping processes. Thus, it appears that during Upper Lefkara times the Troodos Massif underwent differential uplift resulting in arching, slumping and erosion of the earlier Tertiary sediments. Uplift has resulted in an angular unconformity between the Lefkara Formation and the overlying Pakhna Formation in some areas on southern and eastern flanks of the Massif, whilst in the more basinal sequences the contact is conformable and transitional and can only be accurately be defined by using microfossils.

The Pakhna Formation of the Middle and Upper Miocene, consists of a diverse assemblage of ophiolite-derived clastics, shallow water calcarenites, organic rich calcilutites and calcareous marls. The succession varies widely in thickness (from >300m to the south of the Massif to 0m on north west Troodos) and is characterised by dramatic facies variations that faithfully record the variable depositional conditions around the rising and eroding Troodos ophiolite basement (Robertson 1977, Eaton 1986).

To the south of the Massif, gradual uplift until the Middle Miocene is documented by shallowing upward carbonate sequences culminating in the deposition of highly bioturbated chalky marls in oxygen deficient lagoonal basins. Locally reefal limestone debris was shed from nearby patch reefs that lay immediately adjacent to the emergent areas of vegetated and weathered ophiolitic basement. A pulse of rather accelerated Middle Miocene uplift centered on the Limassol Forest area resulted in the intense slumping, erosion and associated deformation of sedimentary successions on the southern flanks of the ophiolite. This uplift was associated with the rapid deposition of both ophiolite-derived clastics (particularly on the SW and SE flanks of the Limassol Forest), coarse shallow water reefal-derived carbonates (Koronia limestone) and deeper
water organic rich hemipelagic carbonates (which are typically strongly bioturbated). Graded bioclastic calciturbidites are particularly abundant in deeper parts of the basin and these are interpreted as the erosion products of shallow shelf sediments which have been reworked downslope.

According to Eaton (1986), the southern limits of the basins on the southern flanks of the Massif were defined by a broad submarine topographic high, a feature which developed during the Miocene period and probably evolved as an upward propagating thrust or series of thrusts, possibly sharing a common sole thrust with faults in the Yerasa fold belt on the south western margin of the Limassol Forest block. Features associated with this zone of deformation on the southern flanks of the Pakhna basins include intense folding of Pakhna sediments, the shedding of talus deposits off south-facing submarine fault scarps (Akrotiri boulder bed) and the back-shedding of olistoliths of shallow water limestones into the adjacent basins to the north.

Miocene uplift culminated in the deposition of monomineralic gypsum deposits in oxygen deficient restricted basins. These evaporitic deposits blanket over both ophiolite-derived shallow water clastics and reefal limestones alike, but appear to be of subaqueous origin. Their monomineralic texture, thickness and facies supports a refluxing barred basin model for their formation (Eaton 1986). During Pliocene time, following penetrative peneplanation of the Troodos ophiolite, the Massif remained low lying while to the north in the Mesaoria (see chapter 1 section 1.3.2) sedimentary basins developed along high angle faults initiated during the Late Miocene emplacement to the adjacent Kyrenia Range. Thick sequences of shallow marine sands and marls accumulated in the neotectonic Polis graben to the west of the main Massif while the area to the south of the present day outcrop of the ophiolite was exposed subaerially and deposition of coarse Troodos derived clastics only occurred during local marine incursions.

3.15 SITE LOCATION AND SAMPLING PROCEDURES

A total of 426 cores were collected from 6 localities in the Neogene sedimentary cover to the ophiolite complex (see fig.23 and table 7). A wide variety of lithologies were sampled including ophiolite-derived clastics (one locality; Tokhni), bioturbated organic rich hemipelagics (three localities; Tsada, Kandou and Ayios Tykonas), shallow water calcarenites (one locality; Ypsonas) and Pliocene grey marls (one locality; Polis graben). Typically, successions were well bedded and only gently inclined, making structural corrections easy to apply.

Cores were taken in groups of four or five at 1m intervals if bedding planes were tilted at a sufficiently large angle. If sediments were more flat lying only limited stratigraphic intervals could be effectively sampled. At all localities orientation was
recorded solely with a magnetic compass. This orientation procedure was considered appropriate at localities sited some distance from the outcrop of the highly magnetic extrusive series of the ophiolite complex. Rarely could cores longer than 3cm be collected from friable marl or hemipelagic chalk successions, although coarser bioclastic limestones and massively bedded ophiolite-derived clastics sometimes yielded coherent cores longer than 15cm.

The NRM of each subsample was then measured on either a cryogenic magnetometer (for samples with intensities less than 1.0mAm⁻¹; i.e. at 5 localities) or a 'Molspin' magnetometer (for ophiolite-derived clastic sediments collected at Tokhni, which have an NRM intensity of >100mAm⁻¹). After application of a field correction the mean NRM directions clearly grouped close to the present day geomagnetic field direction over Cyprus (DEC=003°, INC=54.5°) while following application of the bedding correction between-locality directional scatter is increased. This can be interpreted as an indication of the dominance of soft recent components that were acquired after the structural tilting of the succession.

Three relatively high intensity specimens were then chosen from each locality sampling population and subjected to AF demagnetisation at 2mT steps up to 12mT and thereafter at 4mT steps until the remanence was comparable with that of the noise level of the measuring instrument. Pairs of 3 component plots show that the NRM of coarse ophiolite-derived sediments from Tokhni (see fig.24a) and Pliocene marls (see fig.24b) from the Polis graben is dominated by a stable magnetisation with a direction close to that of the present day field direction. Complementary IRM acquisition curves clearly show that this magnetisation is carried at least partly by magnetite (see fig.24), and is therefore presumably primary in origin.

By contrast, bioclastic limestones and hemipelagic sediments from the remaining localities appear to be dominated by secondary components which are not fully removed during magnetic cleaning. Remanence vectors on the stereographic projection track toward both normal and reversed polarity directions, but rarely reach a stable end point configuration. As the NRM intensities of these sediments are rarely more than an order of magnitude greater than the sensitivity of the magnetometer, instrumental noise contributes significantly toward the measured remanence intensity in fields above that of the MDF (usually between 2.5mT and 7.5mT). On a qualitative basis it is clear that normal polarity remanence declinations close to due north with inclinations comparable with the present day geomagnetic field direction. After blanket demagnetisation at 15mT only remanence directions at 2 localities (Kandou and Tokhni) show any improvement in grouping.

Although there is considerable scope for detailed palaeomagnetic studies on Neogene successions in Cyprus, it is clear that recorded declinations are close to the present day geomagnetic field direction. Miocene sediments record both normal and
reversed polarity directions, compatible with them retaining their remanence at the
time or soon after deposition. Contrary to the published data of Shelton and Gass
(1980), no north westerly declinations were recorded in Pliocene successions (see
chapter 1, section 1.4.2).

3.16 DISCUSSION OF REMANENCE DATA FROM THE SUPRA-OPHIOLITE
SEDIMENTARY COVER: IMPLICATIONS AND CONCLUSIONS

Table 8 summarises the available cleaned remanence data from the Late
Cretaceous to Eocene sedimentary cover. Only data from this study are included,
although directions can be effectively compared with complementary mean remanence
vectors determined by previous workers (see chapter 1, section 1.4.2).

Remanence declinations for the basal Perapedhi Formation (umber and basal
radiolarites) are directly comparable with those determined for the underlying extrusive
series. However, there is a clear systematic increase in the declination of the mean
remanence vector through the Lefkara Formation pelagic chalk succession. After
inversion of reversed polarity directions through the origin, a mean remanence direction
for Maastrichtian and Palaeocene chalks is calculated to be DEC=334; INC=+23;
α95=17.3; N=217 compared with DEC=274; INC=+36; α95=12.3; N=663 for the
underlying extrusives. As magnetic declinations in the sediment cover reflect rotational
motion in the horizontal plane of the underlying ophiolite basement, this implies that c.
60° of the palaeorotation occurred subsequent to the deposition of the basal radiolarites
(i.e. late Campanian) but prior to the end of the Palaeocene. As Lower Eocene chalks
of the Upper Lefkara Formation retain remanence declinations that are close to due
north (DEC=357.0; INC=38.0; α95=10.1; N=136), rotation was certainly complete by
the end of the Lower Eocene. No further rotation occurred during the Late Palaeogene
and Neogene as indicated by the northerly declinations recorded by the Miocene and
Pliocene sediments. This constrains the rate of rotation to a maximum of 3° per Ma.,
although it is possible that rotation occurred in Campanian-Maastrichtian time, a
period of only a few million years (c.20 Ma.), immediately following the end of
spreading ridge volcanism. If more northerly declinations recorded in the
stratigraphically higher levels of the radiolarite succession are considered reliable (see
section 3.6), it can be assumed that rotation was initiated at some time during the late
Campanian.

By contrast, no systematic change in the mean inclination is recorded through
the supra-ophiolite sedimentary cover. Both umber (MINC=+6.0) and radiolarite
(MINC=+13) lithologies preserve anomalously low inclinations which contrast strongly
with the mean magnetic inclination of the underlying extrusive series (MINC=+36).
Rather than appealing to a rapid migration of the ophiolite complex toward the
equator, or a major prolonged excursion of the geomagnetic pole away from the
g eo-centric axis of rotation, low inclinations are more likely to be attributed to the
physical rotation of detrital carriers in the vertical plane about horizontal axes during
sediment compaction. This is compatible with estimates of vertical shortening in these
sediments which indicates that compaction can be by as much as 50% (Boyle 1984).
This effect may be superposed over an original ‘inclination’ error which can result in a
shallowing of the net magnetisation by as much as 20° compared with that of the
ambient field inclination. If deposition of detrital magnetic carriers occurred on a
sloping surface (which is probable for the basal umbers which overlie inclined massive
lava flow units) this inclination error may be as high as 25° (King 1955). Both
processes leave declinations unaffected as realignment of detrital carriers only occurs in
the vertical plane without any significant rotation of particles in the horizontal plane.

Inclinations in the overlying Late Cretaceous and Tertiary pelagic succession are
similarly rather low, but rather more variable, with individual demagnetisation plots
(see fig.13 and fig.15) indicating that inclinations for stable remanence vectors can vary
from a minimum of +05° up to a maximum of +45° for normally polarity samples.
Complementary reversed polarity samples have inclinations that vary from a maximum
of -05° down to a minimum of -43°. Although no single cause can be identified these
anomalous directions may originate partly from the acquisition of remanence during a
transitional excursion of the geomagnetic field between two polarities, or a partial
remagnetisation in an antipodal field direction or a physical reorientation of detrital
carriers during compaction processes. However, as higher coercivity magnetite carriers
appear to retain a remanence vector parallel to that of the haematite carriers (see
section 3.13) the magnetisation is assumed to be of primary origin and any magnetic
overprinting must have occurred relatively soon after the sediment was deposited.

Upper Lefkara chalks show more consistent inclination values (MINC=38.0)
compatible with them having been deposited at a latitude of c.19°, providing that
detrital magnetite grains were aligned in parallelism with the ambient geomagnetic field
at the time of sediment deposition and have not been reorientated subsequently. If this
inclination is considered reliable the gradual northward drift of the ophiolite complex
towards its present latitude at 35°N can be constrained to have been initiated after the
end of the Lower Eocene. Further, more detailed palaeomagnetic studies through the
Lefkara succession are required before the northward drift of the ophiolite complex can
be mapped out more comprehensively.
4.1 INTRODUCTION

Paleomagnetic studies along the northern flanks of the ophiolite complex have conclusively established that there is no systematic change in the declination of the primary remanence vector retained by the extrusive series across the strike of the sheeted dyke complex. This implies that no bulk rotation of the ocean floor crust in the horizontal plane occurred at the time of crustal genesis, a result which is in agreement with primary remanence directions isolated from the supra-ophiolite sedimentary cover, which support an initiation of the palaeorotation no earlier than the late Campanian, some 10-15Ma after the end of spreading ridge volcanism.

As dykes were aligned predominantly NS across the ophiolite complex it can be assumed that if well exposed and undeformed sequences of the extrusive series existed along the southern margin of the main Massif (i.e. to the north of the east-west trending Arakapas Fault), they would retain identical remanence vectors to the pillow lava succession on the northern flanks of the ophiolite. This however, cannot be assumed for the extrusive sequence exposed in the Limassol Forest and Akamas Peninsula crustal blocks, as these fragments of ophiolitic basement are separated from the main Massif by major structural lineaments which may represent transcurrent faults along which the rotation of the main Massif may have taken place. In this chapter palaeomagnetism is used to investigate whether these outlying blocks of ophiolitic basement have rotated together with the main ophiolite complex or whether they have behaved independently through the Late Cretaceous and Early Tertiary. To complement this study, the evolution of the Arakapas and Limassol Forest transform fault domain and its inferred extension into south west Cyprus is discussed. Preliminary palaeomagnetic data from intracrustal fault blocks in the vicinity of the Arakapas lineament provide possible evidence for the sense of offset of the original fracture zone.

4.2 THE LIMASSOL FOREST AND ARAKAPAS TRANSFORM DOMAIN.

4.2.1 THE OPHIOLITIC BASEMENT OF THE LIMASSOL FOREST

Along the southern flank of the main ophiolite complex (see fig.1), a prominent linear east-west depression, the Arakapas Fault, separates the main Massif from the
Limassol Forest, a fragment of oceanic lithosphere which has experienced a more complex magmatic and tectonic history to that of the main complex (Simonian and Gass 1978). In the Limassol Forest, the tectonised harzburgite-dunite mantle sequence which is usually located at the base of the ophiolite succession, occurs at a high structural level and has been differentially serpentinised and brecciated before being intruded by wehrlite and gabbro plutons and several sets of dykes. Important structural features, including east-west trending vertical mylonite shear zones in both gabbros and harzburgites, horizontal lineations and serpentinite melange zones with large phacoidal ophiolitic blocks are all indicative of major deformation in an oblique-slip stress regime. These features, together with pronounced crustal thinning of the oceanic lithosphere and the intrusion of dykes into a conjugate fracture zone system (Bechon and Rocci 1982) lead Murton and Gass (in press) to suggest that this area of crust was generated in a major transform fault zone with a northern boundary delineated by the Arakapas Fault.

4.2.2 THE EXTRUSIVE SUCCESSION OF THE LIMASSOL FOREST AND THE ARAKAPAS FAULT BELT.

On the SW and SE flanks of the triangular shaped Limassol Forest Block, the mafic and ultramafic rocks of the ophiolitic basement are unconformably overlain by the extrusive succession containing numerous interlava sediment horizons. Typically, these sediments resemble interlava epiclastics described from the northern flanks of the main Massif (see chapter 2, section 2.10), being predominantly fine to medium grained, evenly bedded, poorly graded and containing locally derived lava fragments set in a muddy matrix. Coarse grits and conglomerates are common, these immature clastics represent sediments shed off local active fault scarps in the transform domain. The abundance of hyloclastites and lava breccias in the succession supports the extrusion of lava in a tectonically unstable setting.

The Arakapas Fault lineament marks the northern boundary of the Limassol Forest transform domain. The volcanic and sedimentary ‘infill’ of this fault belt includes basaltic lava breccias, volcaniclastic sandstones, pillow lavas, massive lavas, limburgites and intercalated Fe-rich mudstones (Simonian 1975). Like epiclastics interbedded with extrusives in the Limassol Forest block, the volcaniclastic sedimentary rocks of the Arakapas Fault belt were produced by submarine weathering and gravity deposition of basalts within the fault zone. By contrast, the iron-rich mudstones originated as chemical precipitates from hydrothermal solutions generated by leaching of basalt by seawater (Robertson 1978). Pervasive greenschist metamorphism of the underlying sheeted dyke complex can be attributed to the same phenomena.
Fig.1 Detailed map of the Limassol Forest area (after Simonian and Gass 1978). Localities referred to in the text are also marked. Inset shows areas covered by fig.1, fig.8 and fig.9. Notice how the Arakapas Fault is in reality a complex lineament comprising a series of major east-west trending faults that extend for long distances laterally.

Fig.2 Alternative geometries for the Arakapas oceanic transform fault zone. i) Field map of the Limassol Forest block showing systematic swing in dyke trend in sheeted complex to the north of the Arakapas Fault lineament (after Simonian and Gass 1978). The curvature could be related to dyke injection into a sigmoidal stress field associated with a dextrally offset spreading ridge (ii) or, alternatively dykes could be attributed to clockwise fault block rotations close to a sinistrally offset ridge axis (fig.iii).

Fig.3 Representative AF demagnetisation plots for umber sample collected at Asgata (a) and interlava sediment sample collected at Kalavassos (b). In contrast to the umber sample in which the stable remanence vector is not isolated in fields of less than 17.5mT, the interlava sediment sample is clearly dominated by a stable remanence component which is isolated in fields of less than 10mT.

Table 1 and 2 Sampling procedures and details of mean remanence directions prior to, and subsequent to, magnetic cleaning for extrusives and sediments overlying the plutonic core to the Limassol Forest block. N=Number of subsamples.
FIG. 1 MAIN STRUCTURAL ELEMENTS AND ROCK TYPES IN AND ADJACENT TO THE ARAKAPAS FAULT BELT
(AFTER SIMONIAN AND GASS 1978)

LEGEND

- UPPER CRETACEOUS TO RECENT SEDIMENTS
- UPPER PILLOW LAVAS AND ASSOCIATED SEDIMENTS
- PILLOW LAVAS
- SHEETED DYKE COMPLEX
- AXIAL SEQUENCE
- PLUTONIC COMPLEX
- MANTLE ASSOCIATED ROCKS

Inset

FIG. 9

FIG. 8

FIG. 1

10Km

SCREE BRECCIA
FAULT
FOLD AXIS IN YERASA
FOLD BELT
FIG. 3 AF DEMAGNETISATION OF UMBER AND INTERLAVA SEDIMENT SAMPLES

a) NRM = 46.95

b) NRM = 6.65

AKA 12A

AKA 03A

MDF = 295.6

MDF = 120.1

Y VS Z

X VS Y

Y VS Z

X VS Y
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<th>NO. OF SPECIMENS</th>
<th>NO. OF SUBSAMPLES</th>
<th>SAMPLING METHOD</th>
<th>PALAEOHORIZONTAL CRITERION</th>
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<td>SEDIMENT LAMINATION</td>
<td>SILLS/INTERLAVA SEDIMENTS</td>
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<td>CHIPS</td>
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**TABLE 1**
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<th>MDEC</th>
<th>MINC</th>
<th>$\alpha_{95}$</th>
<th>K</th>
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<th>CLEANING FIELD</th>
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<th>MINC</th>
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**TABLE 2**
4.2.3 THE SENSE OF RIDGE OFFSET OF THE ARAKAPAS TRANSFORM DOMAIN

The occurrence of a well developed NS trending sheeted dyke complex in the main Massif provides unequivocal evidence for spreading at right angles to the inferred transform trend, as represented by the Arakapas Fault. However, according to Simonian and Gass (1978), as one approaches the Arakapas Fault Belt from the north, the dykes in the Troodos Massif swing progressively westward into eventual alignment with the transform lineament over a distance of 10-15km (see fig.2i). The swings in dyke trend could be theoretically explained by the intrusion of sheeted dykes into a sigmoidal stress field operating close to an active transform fault or, alternatively, frictional drag along the transform could have caused bulk rotation of the sheeted dykes into parallelism with the transform.

In support of the first model Clube and Robertson (in press) pointed out that a deviation in the tectonic spreading fabric, as represented by a swing in the orientation of normal fault scarps has been recognised close to contemporary transform faults undergoing high slip rates, as in the Gofar and Quebrada fracture zones of the East Pacific Rise (Searle 1983). If ocean floor normal fault configurations close to contemporary transform domains represent complementary swings in dyke trends within the underlying ocean layer 2, the original sense of offset of the Troodos spreading axis may be inferred by comparing the curvature of dyke trends near the Arakapas Fault with representative deviations in normal fault orientations in the vicinity of modern oceanic transform faults. Thus, if deviations in dyke trend can be attributed to dyke injection into a sigmoidal stress field operating close to an active transform with a component of transtension (i.e. a leaky transform), swings in the tectonic spreading fabric are directly analogous to modern day sinistral transform faults that offset the spreading ridge segment in a dextral sense (see fig.2ii). The curvature of the spreading fabric reflects dyke injection into a stress regime that changes continuously from tension along the spreading direction far from transforms to shear parallel to that direction near transforms. The minimum compressive stress ($\sigma_3$ of the stress ellipsoid) rotates continuously from a direction parallel to the spreading direction far from transforms to a direction 45° from the spreading direction close to fracture zones (Fox and Gallo 1984). Close to contemporary fracture zones, some axial faults appear to curve more than 45° as predicted by this hypothesis, and these are assumed to represent Riedel shears that form at low angles to the spreading direction (Wilcox et al. 1973).

Simonian and Gass (1978) were the first to propose that deviations in dyke trend to the north of the Arakapas transform fault could be attributed to crustal genesis in a sigmoidal stress field, perhaps in a zone between two spreading axes not linked by a transform, as is the case in the Afar depression (Barberi et al. 1975). In
their analysis however, they failed to correctly identify $\sigma_3$ axis of the stress ellipsoid, leading them to conclude that the spreading ridge segment was sinistrally offset. The relative orientation of the stress ellipsoid can now be accurately defined as long range sidescan sonograph surveys provide detailed pictures of the seafloor topography in the vicinity of fracture zone systems (Searle and Laughton 1977, Searle 1979, 1981; Tamsett 1983).

Alternatively, deviations in dyke trend might be attributed to physical rotation of intracrustal fault blocks in the horizontal plane about vertical axes. In this model the deviation is due to horizontal drag impressed by continuous lateral movement along the fault zone. By contrast with the first model, the sense of rotation of dykes in the vicinity of the Arakapas is compatible with the transform having operated as a dextral fault system offsetting a sinistrally displaced spreading ridge segment (see fig.2iii).

In choosing between the two models Simonian and Gass (1978) were impressed by the lack of fracturing within the zone of dyke deviations (10km. to the north of the Arakapas) and also by the radius of curvature described by the arcuate dyke pattern (see fig.2i). These factors lead them to prefer the first sigmoidal stress field hypothesis, although they suggested that further dyke deviation could be caused by fault block rotations. It is also important to note that within the Arakapas Fault Belt itself there is a marked difference between the structure of the igneous basement (i.e. the sheeted dyke complex) and that of the overlying interbedded extrusives and sediments. In contrast to the basement, which is pervasively brecciated and metamorphosed, the extrusive succession is largely undeformed and unmetamorphosed, only being cut by a series of northwardly dipping normal faults. To produce the brecciation of the type displayed by the basement there necessarily must have been a major component of compression across the fault zone prior to the onset of lava extrusion and sedimentation in a presumably transtensional setting. Thus, the preferred 'leaky transform' model involving crustal genesis in a dominantly transtensional setting cannot be compatible with field data from the Arakapas Fault zone and the Limassol Forest which indicates that these areas suffered a history of pervasive shear and compression in a transpressional setting during crustal accretion.

### 4.3 OBJECTIVES OF PALAEOMAGNETIC STUDY

The objectives of the palaeomagnetic study were twofold;

1/ To determine the orientation of the mean primary remanence vector retained by extrusive units overlying the plutonic core of the Limassol Forest. The direction of this vector should constrain the sense and magnitude of any relative rotation that might have occurred between the ophiolitic basement exposed in the Limassol Forest and that of the main Massif to the north.
To determine the orientations of primary remanence vectors retained within successions outcropping along the length of the Arakapas Fault lineament. A comparison of these directions with those of the basement both to the north and the south should indicate whether any significant intracrustal block rotations have occurred within the fault lineament. The sense of any relative rotation recorded by intracrustal fault blocks might conceivably throw light on the sense of fault motion within the strike slip zone, thus constraining the offset sense of the original spreading ridge segment. Theoretically, a complementary palaeomagnetic investigation in the sheeted dyke complex to the north of the Arakapas Fault could reveal whether dykes have suffered significant tectonic rotations in the horizontal plane with respect to dykes on the northern margin of the Massif. These results might indicate whether deviations in dyke trend can be attributed to horizontal drag along the fault zone in a transpressional stress regime or if dyke injection occurred in a sigmoidal stress field within a dominantly transtensional stress regime.

4.3.1 PALAEOMAGNETIC INVESTIGATIONS IN THE LIMASSOL FOREST

The extrusive succession overlying the plutonic core of the Limassol Forest basement comprises a thick sequence of pillow lava flows, pillow breccias and minor massive flow units intercalated with ophiolite-derived epiclastics. The presence of numerous sills and well bedded sediments at many localities provides a reliable indication of a palaeohorizontal surface on which an appropriate structural correction can be based. However, on the south western margin of the Limassol Forest block the extrusive sequence has been strongly deformed during late Tertiary compressional tectonics, and individual lava flow units typically dip at steep angles beneath folded and faulted Lefkara and Pakhna sediments. Locally in this 'Yerasa fold and thrust belt' the lavas are overturned and have been thrust over Neogene sediments. As directional errors are always introduced when retilting steeply dipping successions back to the original horizontal only sites on the south eastern margin of the Limassol Forest block were considered appropriate for palaeomagnetic studies, as there lava units rarely dip at angles greater than 40°.

In all, 58 samples were collected at four localities (see fig.1 and table 1); two within the extrusive succession (Parekklisha, pillow lava flow unit and Kalavassos, interlava sediments invaded by numerous sills) and two within the Perapedhi Formation (Monagroulli and Asgata, both umbiferous sediments immediately overlying breccia flow units). Pillow lavas at Parekklisha were drilled in situ in the field, while interlava sediments and regularly jointed hypabyssal rocks were orientated on a planar surface and drilled with a pedestal drill back in the laboratory. Finely laminated umber sediments from the remaining localities were sampled by collecting small
orientated chips (see chapter 3 section 3.5).

Following subsampling and NRM measurement, the stability of the NRM carried by at least two specimens from each locality was investigated using standard AF demagnetisation procedures (see chapter 2 section 2.6 and chapter 3 section 3.6). Igneous and interlava sediment specimens were demagnetised at 2-5mT incremental steps up to 15mT and thereafter at 5mT incremental steps up to 60mT, while umber samples were demagnetised at 2mT incremental steps up to 14mT thereafter at 4mT steps until the remanence was dominated by spurious ARM components. Changes in the orientation of the remanence vector during progressive demagnetisation were investigated on pairs of 2 component Zijderveld plots. The remainder of the sample population from each locality were then demagnetised at a level above (usually 5mT) the field at which the stable remanence vector was isolated. Cleaned remanence directions were then corrected to bedding by rotating the remanence vector about a line of strike on the palaeohorizontal surface.

After demagnetisation there was a slight reduction in the radius of the cone of confidence at each site, together with a complementary migration in the direction of the mean locality remanence vector away from the present day geomagnetic field direction (see table 2). Two component Zijderveld plots clearly show that both umbers (see fig.3a) and interlava sediments (see fig.3b) have an NRM that includes a soft northwardly dipping component that is removed by low alternating fields (typically 17.5mT for umbers and 15mT for interlava sediments). Igneous specimens also retain this secondary magnetisation component, but its effect is removed in low field strengths (<7.5mT) with only a minor reduction in the remanence intensity.

With the exception of umber samples collected at Monagroulli, the direction of the stable magnetisation at all localities is consistently directed toward the west (see fig.4), although there is evidently considerable scatter in the inclination of the mean remanence vector between localities. To account for the range in inclination values which vary from +11.5° for umbers at Askata to +47.1° for pillow lavas at Parekklisha, I appeal to application of inappropriate structural corrections (in extrusive terranes, see chapter 2 section 2.9) and the physical rotation of detrital carriers in the vertical plane (for umber samples, see chapter 3 section 3.16). However, the strong similarity between stable remanence vector directions isolated for interlava sediments and intrusive sills at Kalavassos provides a particularly accurate estimate of the attitude of the palaeofield relative to the palaeohorizontal surface, as defined by the sediment laminations in well bedded interlava sediments. The reliability of the remanent vector orientation is further enhanced, as the origin of the magnetisation in each rock type is different, being of thermoremanent origin in igneous rocks and of probable detrital origin in finely bedded interlava sediments. Thus, the mean inclination of the remanence vector at this site gives the best indication of the latitude
Fig. 4 Bedding corrected mean remanence directions (X) together with their associated α95 cone of confidence for two sampling sites in the extrusive series (Parekklisha and Kalavassos) overlying the Limassol Forest block and two sampling sites in the supra-ophiolitic cover (both umbers at Asgata and Monagroulli). Representative Zijderveld plots are also presented; N=Number of subsamples, *direction of the prevailing geomagnetic field. Legend of rock types is presented in fig. 5.

Fig. 5 Bedding corrected mean remanence direction (X) together with their associated α95 cone of confidence for five sampling sites located in the axis sequence of the Arakapas Fault belt. Representative Zijderveld plots are also presented; N=Number of subsamples, *direction of the prevailing geomagnetic field.

Table 3 and 4. Sampling procedures and details of mean remanence directions prior to and subsequent to magnetic cleaning for extrusives and sediments collected along the Arakapas Fault belt. N=Number of subsamples. See fig. 1 for details of site location.

Fig. 6 Rotation of fault block in the horizontal plane about a vertical axis in a strike-slip fault lineament. Schematic diagram shows how rotation of a semi-rigid fault block can occur between two coherent basement units (The Limassol Forest block and the Troodos massif). Adjacent to the rotating block is a zone of intense cataclastic deformation. A clockwise sense of rotation is consistent with a right lateral sense of fault motion.

Although the limited data support the sinistral (frictional drag) offset model, the Arakapas fault zone may exhibit a long history of both transpressional and transtensional deformation and more data are clearly needed before a firm conclusion can be reached. In this thesis reconstruction of the Troodos ridge system shows the Arakapas and the Limassol Forest areas existed as a dextral transform fault lineament that sinistrally offset the Troodos spreading ridge segment (as in fig. 2ii).
FIG. 4 CLEANED REMANENCE DIRECTIONS AT SITES IN THE LIMASSOL FOREST BLOCK
FIG. 5 CLEANED REMANENCE DIRECTIONS ALONG ARAKAPAS FAULT LINEAMENT
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Table 4
STRIKE-SLIP FAULT (-ARAKAPAS FAULT)

RIGID BLOCK

ZONE OF CATACLASTIC DEFORMATION

COHERENT BASEMENT BLOCK
(TROODOS MASSIF)

COHERENT BASEMENT BLOCK
(-LIMASSOL FOREST)

FIG.6 BLOCK ROTATION WITHIN STRIKE-SLIP SYSTEM
of the Troodos ophiolite in the Late Cretaceous. Applying the axial geocentric dipole formula \(2\tan \lambda = \tan I\) (\(\lambda = \)latitude, \(I = \)inclination), an inclination of 32.9° gives a palaeolatitude of 17.8°, some 17° to the south of the present day latitude of Cyprus.

The mean westerly declination of the stable remanence vector indicates that the basement to the Limassol Forest area has rotated 90° anticlockwise together with the main Troodos Massif to the north. This implies that the Arakapas Fault lineament did not operate as a major crustal transcurrent fault along which the palaeorotation of the Troodos ophiolite complex occurred. This fault is not preserved in southern Cyprus and therefore must lie to the south of Parekklisha where the extrusive units clearly retain a stable remanence with a due west declination (see fig.4).

4.3.2 Palaeomagnetic Investigations Within and Adjacent to the Arakapas Fault Belt

1/ The extrusive and sedimentary infill.

The Arakapas Fault Belt forms a major topographic depression that extends east-west for some 35km before disappearing, at both ends beneath an undeformed cover of Tertiary sedimentary rocks. Topographically the Arakapas Fault Belt varies along its length from a single valley to a series of valleys separated by parallel ridges. To the north, the boundary that marks the contact with the main Massif is taken as the line or narrow zone across which the uniform sheeted structure of the dyke complex is replaced by dominantly brecciated rocks. To the south, it is the intensely sheared contact with the plutonic core of the Limassol Forest complex that delineates the southern boundary of the fault belt. The surface configuration of the basement varies markedly along its length, from a rugged surface defined by linear depressions of intensely brecciated basement adjacent to east-west spines of coherent axis sequence outcrops between Kato Dhrys and Akapnou, to uniformly dipping east-west-running south-facing fault scarps aproned by extrusive pillow breccias and immature epiclastics between Perapedhi and Arakapas (see fig.1).

The palaeomagnetic study was concentrated on the western segment of the fault zone between Arakapas and Trimiklini, as in this section, successions through the sedimentary and igneous infill to the fault are largely coherent and undeformed. In all, 96 samples were collected from 6 localities sited at regular intervals along this section of the fault (see table 3 and fig.5). In addition to these localities, the extrusive series to the SEE of Trimiklini and umber sediments at Perapedhi were also sampled. Sampling procedures at each site have been described in detail elsewhere (orientated chips from finely laminated mudstones and siltstones, hand sampling from massively bedded epiclastics and well jointed sheeted intrusives and coring of pillow lavas and well bedded massive sediments). All samples were orientated using a sun compass, and
the attitude of the palaeohorizontal was assessed by conventional methods. Where aphyric pillow lava flow units are not interbedded with sediments, the palaeohorizontal can only be estimated by measuring the attitude of the plane that holds the long axes of pillows in an individual flow unit. As this represents the inclination of the plane of the palaeoslope, remanence data from Trimiklini can be assumed to be poorly constrained (see table 3). Pillow breccia units were not sampled, as in these fragmental sequences any remanence with a primary thermoremanent origin will be randomly distributed if brecciation occurred subsequent to the cooling of the lava beneath the Curie points of their magnetic carriers.

With the exception of one locality where the bedding corrected remanent magnetisation vectors cluster in the south west quadrant of the stereonet (e.g. Trimiklini, see fig.5), the NRM of all lithologies at the remaining localities is clearly dominated by a stable component which is typically orientated between due west and due north and is inclined downwards (i.e. positive inclination). Even allowing for minor declination errors being introduced in the application of a standard structural correction (see chapter 2 section 2.11.2), the mean remanence vector directions are consistently orientated away from the mean westerly remanence vector determined for the main Troodos Massif to the north (see section 4.1) and that for the Limassol Forest block to the south (see section 4.3.1). Assuming that the stable remanent magnetisation retained by these successions is primary (i.e. was acquired at the time of, or soon after, the formation of the rock), and that the palaeorotation of the ophiolite complex was initiated subsequent to the deposition of the youngest sediments in the Arakapas Fault belt 'infill', the north westerly declinations retained by successions within the lineament necessarily record the clockwise rotation of intracrustal fault blocks relative to ophiolitic basement both to the north and south of the fault lineament. Clockwise rotations of up to 70° relative to the Troodos basement are represented (e.g. at Arakapas where a 12m. long section of gently dipping Fe-rich mudstones retains a stable magnetisation with a mean inclination of 340°. In this section the anomalously low inclinations recorded after application of a structural correction, suggest that the stable remanent magnetisation was acquired subsequent to the structural tilting of the succession (see table 4 and fig.5). As the coercivity spectrum of the remanent magnetisation is diagnostic of a magnetite carrier, it is conceivable that faulting and block rotation occurred shortly after sediment deposition.

Assuming that remanence declinations retained in the Arakapas successions record net tectonic rotation about vertical axes, the clockwise rotation of fault blocks in the fault lineament can be directly attributed to deformation within a right lateral strike slip system as illustrated in fig.2iii. In this system, rotation of rigid or quasi-rigid blocks in the horizontal plane would occur within a diffuse zone between two rigid basement segments represented by the main Troodos Massif to the north and
the Limassol Forest block to the south. These small discrete intracrustal fault blocks would remain coherent at the expense of the surrounding basement which experienced severe cataclastic deformation and brecciation. Importantly, in this tectonic setting zones of localised transpression may exist in the vicinity of the rotating intracrustal fault block while regionally a dominantly transtensional stress regime operates (see fig.6). Further, more detailed palaeomagnetic studies through coherent sequences in the sedimentary and extrusive infill to the fault lineament should indicate whether complementary anticlockwise intracrustal block rotations have occurred (as may have been the case in the Trimiklini segment, see fig.7), as well as constraining the timing of the rotation of individual fault blocks relative to the adjacent ophiolitic basement.

2/ Sheeted dykes within the Arakapas Fault lineament and in the Limassol Forest.

Dykes intruding the extrusive and sedimentary infill of the Arakapas fault belt are few, thin, narrow, sinuous and discontinuous. At high levels these dykes provided magmatic feeders for sheeted intrusions injected along sedimentary bedding planes. The presence of these sills and other irregular intrusions together with the general absence of a well developed sheeted dyke complex implies that the east-west tensional stress field that affected the main Massif did not exist in the Arakapas Fault Belt area during the 'infill' period. The orientation of dykes intruding the higher crustal sections varies widely and probably reflects intrusion into an isotropic stress field, although post-intrusion rotations in the horizontal plane are known to have occurred (see above).

To the south, in the Limassol Forest block, three major dyke trends have been identified (Murton in press). The first is a set of near vertical greenschist facies metabasalts that trend between 100-110°N and invade the harzburgite/wherlite plutonic core to the Limassol Forest basement. A second dyke suite cuts this second set and strikes at between 030-040°N. Again these basalt dykes are usually subvertical, and probably represent a complementary swarm to those striking between 100-110°N, both sets having been intruded along a conjugate fracture zone system. The emplacement of a final set of sinuous, bifurcating, dolerite dykes striking 000-040°N represents intrusion into a low stress regime.

In a preliminary palaeomagnetic sampling investigation two subvertical dykes intruding the sedimentary and extrusive infill to the Arakapas lineament were sampled 2km. to the west of the village of Arakapas (see fig.1). One dyke had a WNW strike while the other was striking due E. In all, a total of 15 cores were drilled in situ with each being orientated with a standard sun compass device. After application of the field correction, NRM directions clustered convincingly close to the present day field direction (DEC=003, INC=54.5), while on application of a structural correction (involving rotating the dykes back to vertical about a strike line on the dyke margin)
there is an increase in the radius of the circle of confidence (see fig. 7). This corresponds to a negative fold test (Graham 1949) and demonstrates that the remanence is dominated by a soft present day field component. AF demagnetisation analysis of two pilot samples indicated that the MDF was less than 10 mT, implying that the remanence was carried by low coercivity grains which are susceptible to acquiring viscous magnetisation components. The mean intensity of the magnetisation was low (<0.01 Am⁻¹) compared with the adjacent pillow lavas (>1.0 Am⁻¹) of the Arakapas 'infill', and it appears that the remanence of the dykes is carried by Ti-poor titanomagnetites like that of the sheeted dyke complex of the main Troodos Massif.

To complement this sampling collection, a number of handsamples were collected through a swarm of east-west trending subvertical basalt dykes in the Limassol Forest block 1 km to the west of Dhierona (see fig. 1). In all 20 subsamples were prepared for analysis from 6 handsamples, each hand sample being collected from a different dyke over a 100 m long road section. The mean NRM intensity is comparable with that determined for dykes within the fault lineament, and it seems likely that the primary titanomagnetite phases have been oxidised during pervasive greenschist metamorphism of the basement. Again, prior to application of a structural tilt correction, uncleaned NRM vectors cluster about the present day field direction, and demagnetisation of representative pilot specimens show no systematic directional change in the magnetisation vector in increasing peak field strengths. The MDF of these samples is also rather low (typically less than 7.5 mT). The remanence of these metabasalt dykes was therefore interpreted to have been acquired in the present earth's field, and no firm conclusions could be made regarding the relative post-intrusion rotation of dykes in this segment of the Limassol Forest block.

The magnetic instability within the sheeted dyke complex of the Limassol Forest and dykes within the Arakapas Fault lineament can be directly attributed to pervasive greenschist metamorphism and oxidation of primary titanomagnetite phases associated with the circulation of chemically active hydrothermal fluids in the vicinity of the transform fault zone. Hydrothermal metamorphism was particularly important in this section of the Troodos ophiolite complex as tectonic brecciation of the basement allowed greater volumes of seawater to penetrate and interact with basalts existing in a steeper geothermal gradient than elsewhere on the Massif. This is compatible with some of the rocks within the transform domain being more primitive (dominantly olivine phyric) than those of the main Massif (Simonian and Gass 1978, Murton and Gass in press, Murton in press). Presumably, the steeper geothermal gradient associated with a thinner lithosphere, allowed higher percentages of partial melt in the underlying peridotite mantle to rise up deep fracture systems and then be extruded both in the bathymetric depression of the Arakapas lineament as well as onto the deformed core of the Limassol Forest basement.
CONCLUSIONS OF PALAEOMAGNETIC INVESTIGATION IN THE LIMASSOL FOREST TRANSFORM DOMAIN

1/ The extrusives of the Limassol Forest block preserve westerly declinations, indistinguishable from the mean remanence declination determined for the main ophiolite complex. This implies that no significant rotation has taken place across the Arakapas Fault lineament.

2/ The limited palaeomagnetic data from unmetamorphosed extrusive and sedimentary successions within the Arakapas Fault lineament support a systematic clockwise rotation of fault blocks relative to ophiolitic basement of the main Troodos Massif to the north and the Limassol Forest block to the south. Although this might conceivably support a sinistral (frictional drag) offset model (see fig.2iii), it is more probable that the Arakapas Fault zone exhibited a long and complex history of both transpressional and transtensional deformation. Thus, dyke injection may have occurred originally within a sigmoidal stress field generated close to a dextrally offset spreading ridge segment, before the tectonic rotation of intracrustal fault blocks within the Arakapas Fault zone.

3/ Dykes within the Arakapas Fault belt and in the Limassol Forest retain a remanence that is dominated by soft viscous components. By contrast to the extrusive and sedimentary 'infill' to the Arakapas Fault, the primary magnetic carriers have been pervasively oxidised and any original thermoremanent remanence has been lost. Further palaeomagnetic studies to the north of the Arakapas Fault should reveal whether the remanence retained by sheeted dykes in the zone of dyke curvature is dominated by similar secondary components.

AN EXTENSION OF THE ARAKAPAS FAULT BELT INTO SOUTH WEST CYPRUS?

The geology of the south west margin of the main Troodos ophiolite complex to the west of Perapedhi has not been studied in detail, having attracted far less attention than the adjacent Mamonia Complex. Reconnaissance mapping by Kluyver (1969) showed that the Lower Pillow Lava Series was erupted on a generally flat lying ocean floor topography, while the overlying Upper Pillow Lava unit was extruded onto a more rugged bathymetry. Volcanic agglomerates, sheet intrusions, fault breccia screes and pillow flows erupted down steep palaeoslopes are abundant in higher stratigraphic levels of the succession. The lava pile has been strongly deformed by a series of major faults (at least three different sets according to Kluyver) and at many localities lavas have been thrust S and SW over the supra-ophiolite sedimentary cover (e.g. at Kannaviou where shallowly E dipping thrusts juxtaposed lava breccias above thick
Fig. 7 Remanence directions for samples collected from two dykes prior to (i) and subsequent to (ii) application of a bedding correction. Notice how the radius of the cone of confidence is increased in (ii).

Fig. 8 Bedding corrected mean remanence directions (X) together with their associated $\alpha_{95}$ cone of confidence for five sampling localities along the south-west margin of the Troodos massif. Representative Zijderveld plots are also presented; N=Number of subsamples, *=direction of the prevailing geomagnetic field.

Fig. 9 a) Bedding corrected mean remanence directions (X) for sites located on Troodos ophiolitic basement exposed on the Akamas Peninsula. b) shows mean direction at southerly site assuming that the palaeohorizontal can be effectively estimated from the attitude of pillow lava flows. Representative Zijderveld plots are also presented. Geological map based on field map of Turner (1971).

Table 5, 6 and 7. Sampling procedures and details of mean remanence directions prior to, and subsequent to, magnetic cleaning for extrusives and sediments collected along the south-west margin of the Troodos and on the Akamas peninsula. In table 7b bedding corrections are based on the attitude of dykes, which are assumed to have been originally intruded vertically.
FIG. 7 FOLD TEST ON ARAKAPAS DYKES
FIG. 8 LOCATION OF SITES ON SOUTH-WEST MARGIN OF TROODOS
FIG. 9 SAMPLING LOCALITIES ON THE AKAMAS PENINSULA (MAP AFTER TURNER 1971)
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<th>MINC</th>
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<th>K</th>
<th>MEAN NRM INTENSITY mAm⁻¹</th>
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**TABLE 7a)**

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volcaniclastic sandstones of the Kannaviou Formation, see fig.8). Preliminary data indicate that the extrusive succession includes primitive lavas similar to those of the Arakapas Fault belt (Murton and Gass, op. cit.). Metalliferous mudstones and siltstones similar to those reported in the Arakapas Fault 'infill' (see above) have also been found interbedded and overlying Troodos lava units near Statos some 20km to the west of Trimiklini, where the Arakapas Fault disappears beneath Tertiary sediments (Robertson 1977).

Available geological data supports an extension of the Arakapas Fault along the southern margin of the Massif toward Asporyia and Pano Panayia. If this is the case, then the lava breccias that dominate the higher levels of the extrusive pile can be interpreted as submarine lava screes derived from a transform fault scarp terrane located not far to the north (Robertson 1977). A history of continuing fault movement and general tectonic instability is implied by the repeated angular discordances both within the lava pile itself and in conglomerate horizons at the base of the overlying Kannaviou Formation. Umbert and radiolarite lithologies occur in scattered hollows at the top of the extrusive succession, and these sediment inliers provide the most convenient successions for determining a reliable declination for the underlying Troodos basement. This is because at many localities along the south west margin of the ophiolite the palaeohorizontal can only be poorly constrained on account of the extrusive succession being dominated by pillow and lava breccia units that were shed down steeply dipping palaeoslopes. In sampling Turonian-Campanian umber and radiolarite successions to determine a complementary remanence vector for the underlying Troodos crust, it was assumed that there was no rotation of the Troodos basement prior to the end of radiolarite deposition in the late Campanian (see chapter 3 section 3.8).

In a preliminary palaeomagnetic study, 92 samples were collected at 5 localities along the south west and western margins of the main ophiolite complex (see fig.8 and table 5). Massive pillow lavas, sheeted intrusives, metalliferous umbers, pink radiolarites and Fe-enriched mudstones intercalated with lavas were sampled using standard procedures described elsewhere (see chapter 2 section 2.4 and chapter 3 section 3.5). The attitude of the palaeohorizontal at each site was measured at least three times and a mean value was then calculated. Details of site location and sampling procedures are presented in table 5. The stability of the NRM was then investigated by standard AF demagnetisation, with directional changes in the remanence being studied on pairs of two component Zijderveld plots. Following isolation of a stable remanence direction in pilot specimens, the remaining samples at any one site were then exposed to an appropriate 'blanket' demagnetising field to remove the effects of any low coercivity secondary remanence components. For each locality, the value of this field, and the mean remanence vector orientation together with its associated cone
of confidence is listed in table 6. Stereonets and representative Zijderveld plots are presented in fig.8.

With the exception of samples collected at one locality (purple epiclastics and finely laminated mudstones at Statos) a reliable stable remanence vector mean was determined for all the sites. After application of an appropriate bedding correction, which incidently rarely involved tilting the field corrected remanence vector through an angle greater than 30°, remanence directions at many localities clearly cluster in the NW quadrant away from the prevailing geomagnetic field direction. Due westerly declinations were recorded at Asporyia and Anadhiou. This convincingly demonstrates that intracrustal fault blocks have locally undergone clockwise rotation along the southern and south western margins of the main Massif, supporting an extension of the major Arakapas Fault lineament into south western Cyprus. Future palaeomagnetic studies along the southern margin of the main Massif will aim at constraining the size and relative displacement of individual fault blocks. This will only be possible when the structural history of this section of the Massif has been investigated in detail.

4.5 A PALAEOMAGNETIC STUDY ON THE AKAMAS PENINSULA

A major inlier of Troodos igneous basement crops out to the west of the neotectonic north south trending Polis graben in western Cyprus (see fig.9). This late Tertiary structural feature essentially isolates the Akamas igneous basement from the main Troodos Massif to the east, although gravity data indicate that the thick Neogene sediments in the graben probably blanket over a major crustal lineament around which the palaeorotation of the main Troodos Complex might have taken place (see chapter 5, section 5.3). The aim of the palaeomagnetic investigation was therefore to determine the declination of any stable remanence retained by the extrusive series on the Akamas Peninsula, in order to resolve whether or not this basement sliver has rotated together with the main Troodos Complex through 90° anticlockwise.

The igneous basement that crops out on the Akamas Peninsula is dominated by extensive flows of olivine phyric pillow basalts and agglomerates which have been intruded by a swarm of NNW–SSE trending dykes. As these dykes typically form at least 30% of the rock volume along the coastal exposures, these igneous rocks formally belong to the Lower Pillow Lava suite. Along the coast individual lava flow units and sheeted intrusives typically dip at 60-70° due east while dykes margins are inclined at shallower angles (rarely greater than 45°). This contrasts with pillow lava units exposed inland in the Akamas Forest, where the Troodos basement together with the sedimentary cover (represented by Perapedhi Formation umbiferous clays and Kannaviou Formation bentonites and volcaniclastic sandstones) are locally overturned toward the west. An inverted contact of Upper Pillow Lavas, umbers and Kannaviou
clays is extremely well exposed in the Akamas Forest where olivine phric layers in pillow basalts dip steeply towards the NE at 60-75°. Reconnaissance mapping by Gass (1960), Turner (1971, 1973) and Lapierre (1975) showed that the Troodos basement exposed on Akamas has been folded about a synclinal axis which plunges NNW at about 15°.

In all, 226 samples were drilled at three localities; two localities being sited on the western flank of the synclinal structure, with the third being located inland in the Akamas Forest (see fig.9). Sampling on the eastern flank of the fold was restricted to two pillowed flow units, while on the coast sheeted intrusives, dykes and pillowed flow units were sampled along 30-50m across-strike sections (see table 7a).

At each locality, demagnetisation studies showed that the NRM was dominated by a stable remanence component which was directed away from the ambient geomagnetic field direction both prior to and subsequent to the application of a bedding correction. The mean intensity of the NRM was comparable with lavas sampled in the UPL succession (between 0.1 and 2.0Am\(^{-1}\)) on the northern flanks of Troodos, and the MDF was typically rather higher (rarely less than 3.5mT). The stable remanence like that of the Troodos extrusives is assumed to be of primary thermoremanent origin. If, at the two coastal localities, cleaned remanence directions are corrected back to the assumed palaeohorizontal on the basis of the attitude of pillow lava flow units, the mean inclination is computed to be 69° (see fig.9a). If however the palaeohorizontal is applied on the basis of the average attitude of dyke margins at each locality (see fig.9b), this gives a mean inclination of 41.4°, closer to that determined for the extrusive series of the main Massif to the east. This implies that the extrusives along the western coastline of the Akamas Peninsula were shed down a considerable palaeoslope that was dipping between 20° and 30° toward the east. Importantly, this result also indicates that dykes were intruded close to vertical and have been subsequently tilted westward after they have acquired their remanence.

After tilting the extrusive succession back to an assumed palaeohorizontal at each locality on either side of the synclinal axis, there was a dramatic decrease in the radius of the cone of confidence around the mean remanence direction for the entire Akamas Peninsula sampling population (see chapter 3 table 7). This corresponds to a positive 'fold test' and convincingly demonstrates that stable, assumed primary, remanence was acquired prior to tectonic tilting. The mean declination of the remanence is clearly directed westward, implying that like the main Troodos ophiolite complex and the Limassol Forest block to the east, the Akamas igneous basement has also undergone a 90° anticlockwise rotation since its genesis. Thus, no significant relative tectonic rotation can have occurred across the Polis graben structural lineament.
CHAPTER 5

THE MAMONIA COMPLEX AND THE MONI MELANGE AND THEIR RELATIONSHIP TO THE TROODOS OPHIOLITE COMPLEX.

5.1 INTRODUCTION

As magnetic remanence directions recorded in the in situ sedimentary cover to the ophiolite complex necessarily reflect complementary motions of the underlying basement, the data presented in chapter three unequivocally demonstrate that the Troodos ophiolite complex rotated relatively rapidly between late Campanian and early Eocene time, a period lasting probably no longer than 20-25Ma. Indeed, magnetic declinations preserved in Maastrichtian sediments would suggest that much of the rotation occurred in Campano-Maastrichtian time, a period of only a few million years duration, immediately following the end of spreading ridge volcanism.

The Campano-Maastrichtian interval also marks a particularly critical period in the tectonic history of the south and south western segments of the preserved Troodos ophiolite complex. In contrast to the present northern and eastern flanks of the Massif where pelagic carbonates were slowly accumulating on a topographically elevated segment of ophiolitic basement, in south west Cyprus quite coarse terrigenous sediments (the Kannaviou Formation) appear interbedded with spreading ridge-related manganiferous radiolarites (Perapedhi Formation, of late Campanian-early Maastrichtian age), only several metres stratigraphically above the youngest lavas of the Troodos extrusive series (Robertson 1977). In south Cyprus, Mesozoic continental margin lithologies are present as large detached blocks (olistoliths) within the in situ sedimentary cover overlying Troodos ophiolitic basement of the Limassol Forest block (Moni melange, Robertson 1977). These observations implicitly require the Troodos oceanic crust to have been located close to a Mesozoic continental margin as early as late Campanian-early Maastrichtian time, probably only 10-15Ma. after its genesis.

As the palaeorotation of the Troodos ocean crust occurred at a time when the present south west and southern segments of the Troodos Massif were located in close proximity to a Mesozoic continental margin, it is conceivable that the rotation was directly related to the tectonic interaction of Troodos oceanic crust with the adjacent continental margin. Remnants of the basement to this continental margin are now preserved in the Mamonia Complex in south west Cyprus. In an attempt to understand the mode of basement juxtaposition in south west Cyprus, in this chapter I extensively review, and where necessary, reinterpret salient features of the Mesozoic and early Tertiary geology of the south western margins of the Troodos Massif, the
Fig. 1 Outline geology of south west Cyprus showing the inferred distribution of contrasting Older (Mamonia) and Younger (Troodos) Mesozoic basement blocks, separated by arcuate serpentinite fault lineaments. Localities referred to in the text and line of cross sections drawn in fig. 2 are also included. Palaeomagnetic sampling localities within the Dhiarizos Group (Phasoula lavas), the Ayios Photios Group (Vlamborous Formation) and Troodos basement exposed on the Akamas peninsula (see chapter 4) are also shown. Map is based on Lapierre (1975) and Swarbrick (1979).

Fig. 2 Interpretative cross sections for south-west Cyprus. a) northern area across Polis graben and b) southern area across the Arkhimandrita-Marathounda serpentinite lineament. Note the juxtaposition of contrasting Older (Mamonia) and Younger (Troodos) Mesozoic basement blocks along the high angle serpentinitic fault lineaments. The Older Mesozoic blocks comprise volcano-sedimentary lithologies (Dhiarizos Group) overthrust along a low-angle contact by an upper structural unit composed of wholly sedimentary lithologies (Ayios Photios Group). In (b) the neotectonic Polis graben is superimposed over the older Late Cretaceous and Early Tertiary structural grain. Both sections are taken from Swarbrick (1979).

Fig. 3 The Bouguer gravity anomaly map of Cyprus (after Gass and Masson Smith 1963). A major positive anomaly lies over the exposed Troodos massif reaching a maximum in the Pamos area on the eastern side of Khrysokhou Bay. Notice prominent negative anomalies over the intrusive serpentinite exposed at the heart of the Troodos mountains and over south-west Cyprus. Densities used for Bouguer reductions: Igneous and volcanic rocks 2.7g/cm³ and other rocks 2.4g/cm³.

Fig. 4 Diagram showing why the Mamonia Complex is unlikely to preserve the remnants of an accretionary complex. Both Pearce (1975) and Moores et al. (1984) considered that the Mamonia Complex formed above a subducting oceanic slab, although the inferred direction of crustal consumption is not considered to be the same.
FIG. 2a Metamorphic rocks
Inferred limits of Old Mesozoic Block basement

FIG. 2b

FIG. 1 GEOLOGICAL MAP OF THE MAMONIA COMPLEX
FIG. 2 TWO SCHEMATIC SECTIONS ACROSS THE MAMONIA COMPLEX

a) Gently dipping seafloor serpentinite; gravity flows

Neotectonic Polls Graben

Maas. & Tertiary sediment cover

Multiple debris-flow Kathikas melange

Akamas Peninsula thin Troodos crust with high Mg lavas

Gravity low-inferred Dhiorizos Gp. basement - Older Mesozoic block

WSW

ENE

b) Late-stage serpentinite gravity flows

Most folds face and verge W&NW

Greenschist and amphibolite facies slivers

Disrupted sediment sheets, Ayios Photios Gp.

Kannaviou Volcanogenic sediments

Younger (Troodos) Mesozoic block

Troodos-type slivers with Kannaviou Fm. volcanogenic cover

Sheared serpentinite

Older Mesozoic Block, Dhiorizos Gp.

SW

NE

0 5km.
BOUGUER ANOMALY

Igneous Rocks 2.7 g/cm
Sediments 2.4 g/cm

FIG. 3 BOUGUER GRAVITY ANOMALY MAP

Mamonia has proximal rift and passive margin sediments not open ocean
Troodos crust not regionally thrust over Mamonia

No trench-flysch intercalated with Mamonia sheets
No arc or pre-arc oceanic crust preserved

FIG. 4 THE MAMONIA COMPLEX AS AN ACCRETIONARY COMPLEX?
Mamonia Complex and the Moni melange. In the light of the contrasting tectonic and deformational histories of these areas, I then consider potential constraints on the original plate tectonic setting of the Troodos ophiolite complex at the time of palaeorotation. To complement this discussion, preliminary palaeomagnetic data from the contrasting crustal units of south west Cyprus are presented and these are compared with available data from the main Massif (see chapter 2), the Limassol Forest and the Akamas Peninsula (see chapter 4).

This chapter was jointly written with A.H.F. Robertson, and forms c.50% of a paper entitled 'The palaeorotation of the Troodos Microplate in the Late Mesozoic-Early Cenozoic plate tectonic framework of the Eastern Mediterranean'. A.H.F.R. was the principal author of sections 5.5 and 5.6.

5.2 THE MAMONIA COMPLEX

In chapter 1 (see section 1.3.4) I described how Swarbrick (1979, 1980) convincingly demonstrated that the Mamonia Complex comprises two fundamentally different types of basement terrane each of which is separated by major arcuate fault lineaments intruded by thick sheets of sheared serpentinite. A SW-NE section through the south west segment of the Mamonia Complex (the Arkhimandrita-Marathounta belt, see fig.1 and fig.2b) shows how discrete, largely intact slivers of Older Mesozoic Block (OMB) basement (comprising Dhiarizos Group volcano-sedimentary units overlain by allochthonous Ayios Photios Group sediments) are characteristically abruptly truncated by and separated from identical basement units along high angle fault lineaments intruded by serpentinite screens. Incorporated within this zone of high angle faults are slivers of Younger Mesozoic Block (YMB) basement, consisting of Troodos-type ophiolitic crust overlain by an in situ sedimentary cover of umbers and radiolarites of the Perapedhi Formation. These slivers of Troodos ophiolitic basement are also overlain by a thick largely undeformed succession of bentonitic clays and volcanioclastic sandstones, indistinguishable from the Kannaviou Formation sediments found overlying YMB basement on the main Massif and on the Akamas Peninsula (see chapter 4, section 4.4).

fig.1 shows that on a larger scale, the two contrasting basement types (OMB and YMB) are interleaved along major fault lineaments that closely parallel the arcuate south western margin of the main Troodos Massif. In the south, a c.15km. wide OMB is sandwiched between two major fault lineaments (Statos to the NE, and Marathounda-Arkhimandrita to the SW, see fig.2b) while to the north the inferred extension (see below) of this OMB is bounded to the west by the YMB cropping out on the Akamas Peninsula and to the east by the main Troodos Massif.
5.3 THE REGIONAL DISTRIBUTION OF CONTRASTING BASEMENT TYPES THROUGHOUT SOUTH WEST CYPRUS.

The detailed regional distribution of contrasting basement types throughout south west Cyprus can be inferred from a combination of field mapping, well and gravity data.

5.3.1 FIELD MAPPING AND WELL DATA

In contrast to the Marathounda-Arkhimandrita area where elongate slivers of YMB and OMB basement are juxtaposed along a complex anastomosing belt of high angle faults, in the northern area, OMB basement that is inferred to underlie the Polis graben area (see section 5.3.2), is separated from adjacent YMB basement on the Akamas Peninsula by a single steep eastwardly dipping sheet of intrusive serpentinite (see fig.2a). An analogous westwardly dipping contact in the east marks the boundary between OMB basement and the main Massif of the Troodos Complex, but exposure of this serpentinite lineament is poor and discontinuous beneath Maastrichtian melange and Tertiary pelagic sediments.

To the west of the Polis graben the contact between Troodos basement exposed on the Akamas Peninsula and the adjacent OMB basement can be traced from Loudratis Aphroditis in the north to Trimithousa in the south. The contact running approximately NS is marked by a single line of serpentinite which is periodically lost beneath Late Cretaceous and Tertiary sediments. YMB basement has been identified at a number of localities to the west of this lineament (e.g. at Lara, Peyia, Kissonerga, Mavrokolymbos and Khlorokas; see fig.1). To the south of Trimithousa the strike of the contact swings into alignment with the east-west structural trend of the Arkhimandrita-Marathounda area. No OMB basement has been identified to the west of this serpentinite lineament, and it therefore seems likely that YMB basement occupies the whole area from the contact westwards to the coast and possibly further offshore. This is in agreement with gravity data which indicates that Troodos-type basement with a higher density crustal signature to that of the OMB basement extends to the south of the present day outcrop on the Akamas Peninsula (see section 5.3.2).

To the east, the line of contact is less clearly defined. It is apparent however that the same relationship exists between YMB and OMB basement. The contact is first seen in the south east near Kithasi, where narrow strongly foliated serpentinite striking NW-SE marks the boundary between Troodos complex breccias overlain by Kannaviou Formation bentonitic clays to the north east and OMB sediments (Ayios Photios Group) to the south west. The contact reappears again 5km. to the north near Ayios Ioannis where once again sub-vertical screens of serpentinite separate Kannaviou
Formation sediments from OMB basement. Serpentinite reappears again further to the north west at Statos where thick south westerly dipping sheets truncate YMB basement against lavas of the Dhiarizos Group.

Borehole data (Aphrodisis personal communication, Swarbrick 1979) indicate that the contact between YMB and OMB basement continues north westwards beneath the Polis basin, the inferred trace being shown in fig.1. Kannaviou Formation clays associated with the YMB basement have been identified from boreholes at Kholi and Khrysokhou. The lineament passes close to mineral baths (sulphur springs, associated with Troodos basement?) at Yiolou before meeting the westerly basement contact to the north of Polis in Khrysokhou Bay. This northward closure, inferred to lie some 10-15km. northwest of Polis is also predicted from the geophysical data of Gass and Masson Smith (1963, see below).

5.3.2 GRAVITY ANOMALY OVER SOUTH WEST CYPRUS

The Bouguer anomaly map of Cyprus shows that in south west Cyprus a major negative anomaly lies superposed over the large positive anomaly associated with the gently northwardly dipping Troodos ultramafic slab (see fig.3). The anomaly pattern for south west Cyprus shows a northward closing structure defined by closely spaced isogals, this trend indicating that a mass of lower density material exists at a high structural level relative to higher density material to the north east (the main Massif) and to the north west (the Akamas Peninsula and its southerly extension). The size of the anomaly (>80 mgal) over the Polis graben would imply that in this area a large mass of lower density material is juxtaposed against YMB basement both to the east and west (Gass and Masson Smith 1963). The depth to the base of the Cenozoic sediments in the Polis basin is approximately 400m below sea level (Cyprus Geological Survey Unpublished report), and assuming a maximum density contrast of 0.5gcc\(^{-1}\) between basement and sedimentary cover, this would lead to an anomaly of 8mgal, corresponding to an anomaly an order of magnitude less than that actually observed. Thus, it is apparent that the anomaly over south west Cyprus cannot be solely attributed to sediments overlying basement in the Polis graben, and it seems more likely that the anomaly is related to an actual mass deficiency in the basement.

Although this mass deficiency could be explained by a high concentration of low density serpentinite intruded at a higher level in the crust, Swarbrick (1979) tentatively suggested that a narrow wedge of relatively lower density OMB type basement juxtaposed to the east and west against Troodos basement along high angle fault contacts, might account for the observed distribution of isogals in south west Cyprus.

Having above attempted to apply spacial constraints to the regional distribution
of contrasting basement types throughout south west Cyprus, I now continue by considering the independent histories of each basement type prior to tectonic juxtaposition. First, the early passive evolutionary phase of the Mamonia continental margin are described, and then its deformation and subsequent juxtaposition against YMB basement of the Troodos Massif are considered. The synopsis that follows is based on the studies of Lapierre (1975), Ealey and Knox (1975), Robertson and Woodcock (1979), Swarbrick and Robertson (1980) and Swarbrick (1979, 1980).

5.4 A GEOLOGICAL HISTORY OF THE OLDER MESozoIC BLOCK: THE CONSTRUCTION DEFORMATION AND SUBSEQUENT EMPLACEMENT OF A PASSIVE CONTINENTAL MARGIN SUCCESSION.

The OMB can be effectively divided into two contrasting lithotectonic units separated by an important regionally low angle thrust (see chapter 1, section 1.3.4). This tectonic contact is characterised by localised shearing and disruption to bedding within several metres of the principal thrust plane, but shows no development of penetrative deformational fabrics.

5.4.1 THE LOWER TECTONIC UNIT: A MAFIC CRUSTAL BASEMENT

The lower tectonic unit (lower nappe of Lapierre 1975) consists mainly of Late Triassic mafic extrusives (Dhiarizos Group) including both pillow lavas (Phasoula Formation) and fragmental extrusives (Loudra tis Aphroditis Formation). Subordinate interlava radiolarites and hemipelagic carbonates (Kholetria Member) also occur in the extrusive succession. According to Swarbrick (1979, 1980) the Dhiarizos Group forms a relatively coherent structural basement and is overlain by a para-autochthonous cover of Late Triassic to Early Cretaceous deep sea sediments (Mavrokolymbos Formation) together with isolated detached blocks (up to 100m in diameter) of Halobia-bearing Late Triassic reefal limestone. Regionally, lithologies are usually only gently inclined, although structural dips and intensity of deformation increase markedly toward serpentinite fault lineaments (see section 5.6.2). No structural base to this lower tectonic unit has been found in south west Cyprus.

The petrography, major (Lapierre and Rocci 1976) and 'immobile' trace (Pearce 1975) element compositions of the mafic extrusives support either a continental rift (e.g. Afar) or ocean island (e.g. Azores) origin. This is consistent with the abundance of volcanic breccias and blocks of reefal limestone in the succession which indicate that lava extrusion occurred along active fault scarps located close to carbonate build-ups or small atolls. Peri-platform ooze deposits, presumably derived from a more extensive carbonate platform unit, accumulated on the igneous basement.
during periods of volcanic inactivity, while the supra-lava sedimentary cover was deposited in a deep sea setting below the carbonate compensation depth away from any significant supply of coarse terrigenous sediment.

5.4.2 THE UPPER TECTONIC UNIT: DISRUPTED CONTINENTAL MARGIN SEDIMENTARY SUCCESSION

The upper tectonic unit (upper nappe unit of Lapierre, 1975; or the Ayios Photios Group of Robertson and Woodcock, 1979) consists of a series of disrupted low angle sedimentary sheets which range from tens of metres to several kilometres in horizontal dimension and up to 200m. in vertical thickness. The fold vergence and facing directions of numerous mesoscopic and rarer megascopic folds indicate relative displacement of the sedimentary sheets towards the east and north east (Robertson and Woodcock 1979). Folding about near horizontal axial planes has lead to the inversion of successions in many areas (e.g. Khapotami river, see fig.1).

A composite sequence through the Ayios Photios Group can be pieced together by comparing successions preserved in individual fragmentary sheets. In the restored succession, the oldest sediments are Late Triassic quartzose sandstones (Vlambouros Formation), derived from a plutonic and metamorphic terrane (up to high grade, as represented by the presence of sillimanite, Henson et al. 1949). These sediments were deposited largely by turbidites in settings ranging from deltaic to slope and base of slope settings. Locally, interbedded calciturbidites were shed downslope from an adjacent carbonate platform area.

In the south (e.g. Khapotami River) the sandstones are overlain by Late Triassic Halobia-bearing limestones (Marona Formation), which closely resemble modern periplatform ooze deposits. Above, the Episkopi Formation consists of up to 120m. of mudstones, siltstones, redeposited limestones, quartzose sandstones, radiolarian cherts and minor metalliferous oxide sediments. Towards the top of the succession pelagic carbonates occur locally. Debris flow conglomerates in the sequence reflect continued local input from a nearby platform. Distinctive, texturally mature orthoquartzites (Akamas Member) were deposited into deep water below the carbonate compensation depth in Early Cretaceous time (see section 5.4.3). The most distal-type successions (e.g. Khapotami river in the south) are lithologically similar in many respects to the sedimentary cover of the Late Triassic mafic lavas in the lower tectonic unit (Mavrokolymbos Formation), but probably represent more distal equivalents.

5.4.3 A RECONSTRUCTION OF THE MESOZOIC PASSIVE MARGIN

It appears therefore that the upper tectonic unit preserves the disrupted
remnants of a Mesozoic passive margin that formed adjacent to a continental terrane fringed by a carbonate platform unit (Robertson and Woodcock 1979, Swarbrick 1980). The Late Triassic Vlambouros Formation records high terrigenous input during initial rifting and the early stages of basin formation, while the Marona Formation micritic limestones were shed from the adjacent platform in a tectonically quiescent setting toward the end of the Triassic. The overlying Episkopi Formation accumulated slowly (c.1-2mm/ thousand years) on a gently subsiding passive margin below the carbonate compensation depth in Jurassic to mid-Cretaceous time. Facies range from near-platform margin (carbonate debris-flows) to channelised base of slope turbidites (e.g. Mavrokolymbos Dam) and non-channelised basin plain and/or inter-fan type successions (e.g. Khapotami River). Marked tectonic disturbance and renewed volcanism in the Early Cretaceous is represented by redeposition of the orthoquartzite (Akamas Member) in proximal successions and by the accumulation of bentonitic clay, local tuff, and hydrothermally-derived metal-oxide sediments in the most distal-type successions.

In conclusion therefore it is clear that the Ayios Photios Group sediments were deposited on a narrow fault controlled passive margin segment that was located close to a metamorphic hinterland bordered by fringing reefs and carbonate banks. A possible modern analogy for the early rifting stage of the passive margin is in the Eilat area of the northern Red Sea (Ben Avraham et al. 1979; Mergner 1971 and Hayward 1985).

In contrast to the Ayios Photios sediments, which record the transition across a proximal to distal Mesozoic continental margin of a landmass now located outside the present area of Cyprus, the lower tectonic unit (Dhiarizos Group, see section 5.4.1) represents crust formed in a tectonically unstable setting close to the margins of an embryonic Mesozoic ocean basin which began to open in Late Triassic time. The distal origin of the Mavrokolymbos Formation would suggest that the marginal oceanic crust (lower tectonic unit) was probably located oceanward of the passive margin successions of the Ayios Photios Group (upper tectonic unit). Importantly though, the intervening crustal basement (i.e. that crust which originally underlay the Ayios Photios Group) is not preserved in Cyprus, but it could have been represented by transitional or marginal oceanic crust with MORB affinities. The passive margin-ocean transition was thus probably originally quite wide (c. tens of kilometres) and it is likely that substantial crustal shortening must have occurred to account for the present structural relationship between the two units.

5.4.4 MODE OF TECTONIC JUXTAPOSITION: ACCRETION OR SLIDING?
The timing of emplacement of the upper tectonic unit over the lower tectonic unit of the OMB must post-date the youngest sediments in the allochthonous Ayios Photios Group (probably of Berrasian/Hauterivian age; Ealey and Knox 1975) but pre-date the onset of deposition of Late Cretaceous/Early Tertiary (probably Maastrichtian, Turner 1971) pelagic chalks, as these sediments unconformably overlie the major structural lineaments along which the YMB and OMB have been juxtaposed (Swarbrick 1979, 1980).

Two alternative models have been proposed for the emplacement of the upper tectonic unit (Robertson and Woodcock 1979). In the first model, the OMB was considered to preserve the remnants of an accretionary complex generated above a subducting plate (Pearce 1975, Moores et al. 1984; see fig.4), while in the second model, the passive margin sediments of the upper tectonic unit were believed to have been emplaced above the marginal oceanic crust by gravity sliding (Robertson and Woodcock 1979, Swarbrick 1980) down a continental margin slope. Detailed mapping however, clearly indicates that the lithostratigraphy and structure of the Mamonia Complex is quite unlike that of a conventional accretionary prism formed by steady state subduction. Although a gravity sliding model is more consistent with the detailed field relations, there do however appear to be problems with this interpretation; a) The Episkopi Formation of the Ayios Photios Group is dominated by deep water base of slope to abyssal plane lithofacies, and thus any major gravity sliding is likely to only have followed major vertical displacement relative to the Dhiarizos group basement. b) If the OMB represents an originally wide passive margin-ocean crust transition, it is difficult to envisage how the thrust sheets of the upper tectonic unit could have been emplaced over the lower tectonic unit basement purely as low volume multiple gravity slides without first invoking significant crustal shortening.

It seems likely therefore, that a component of underthrusting (perhaps associated with limited subduction?), tectonically juxtaposed the upper and lower structural units, and during this process the former passive margin sediments were disrupted and sliced with marginal oceanic crustal units being emplaced beneath (see chapter 6, fig.10c). Following the initial crustal juxtaposition, large scale down-margin gravity sliding of the uplifted former passive margin sheets then occurred on deformed, originally more distal marginal crust.

5.5 THE SUPRA-OPIHOLITIC SEDIMENTARY COVER OF THE YOUNGER MESOZOIC BLOCKS

In addition to the Troodos-type slivers associated with the arcuate serpentinite fault lineament in the Marathounda-Arkhimandrita area, YMB basement also crops out on the western flank of the main Massif as well as on the Akamas Peninsula and its
Although it has not been studied in detail, the ophiolite stratigraphy along the south west flank of the Troodos Massif is very sheared and faulted toward its contact with the OMB basement (e.g. near Kannaviou, see chapter 4, section 4.4). In south west Cyprus this Troodos type crust, including basement exposed on the Akamas Peninsula (see chapter 4, section 4.5) is overlain by up to 750m. of late Campanian-lower Maastrichtian volcanogenic sediments of the Kannaviou Formation (Lapierre 1968, Robertson 1977). These sediments include coarse volcanoclastic sandstones, not exposed elsewhere around the north, east or south east flanks of the Troodos Massif. At its base, the Kannaviou Formation consists of radiolarian siltstones and mudstones, while at higher stratigraphic levels these sediments generally coarsen upwards into thick bentonitic clays interbedded with pumaceous sandstones. Petrographic studies indicate that the clastic component is derived from three independent sources: a) Troodos-type ophiolitic crust (e.g. at the base of the succession to the east of Kannaviou, see fig.1); b) A basalt-andesite-rhyodacite extrusive centre and c) Terrigenous units, including lithologies similar to those of the upper tectonic unit of the Older Mesozoic Blocks (e.g. polycrystalline quartz and radiolarian chert).

The Kannaviou sandstones were deposited by a combination of mass-flow, traction currents and minor turbidity currents from a source area presumed, on the basis of regional sediment distribution to have been located to the NW, W and SW of south west Cyprus. At all localities, with the exception of Petra tou Romiou (Swarbrick 1979), the Kannaviou Formation is absent from the cover of the OMB basement.

Limited microprobe data on basic volcanic glass in the Kannaviou sandstones can be compared with chemical analyses of the Troodos extrusive succession. Along the northern flank, Robinson et al. (1983) identified two extrusive units: a lower one of basalts and basaltic andesites, and an upper one of basalts, basaltic andesites and rhyodacites (see chapter 2, section 2.3). The upper unit is comparable with the magnesian andesites (boninites) of arc and fore-arc areas (e.g. Bonin arc). The characteristic chemistry of the upper unit can be attributed to high degrees of partial melting of an already depleted mantle source.

The basic Kannaviou glass occurs together with abundant very vesicular transparent acid volcanic glass (up to 75% SiO₂ based on refractive index tests, Robertson 1977). Although the acid volcanic glass could not be reliably analysed, plots of the major element oxides (FeO, CaO, Na₂O, SiO₂, TiO₂) relative to MgO plot on the same trend as the Troodos basalt-andesite-rhyodacite suite, but exhibit slightly higher MgO contents in several cases (see fig.5 and table 1). The limited Kannaviou glass data support derivation from an evolved arc-type magma similar to that of the Troodos lower basalt-andesite-rhyodacite suite, but clearly dissimilar to extrusives like
Fig. 5 MgO-FeO variation plot after Jakes and Gill (1970). The limited available geochemical data for the Kannaviou Formation basic volcanic glasses are plotted in relation to the two fields of Troodos glasses (after Robinson et al. 1983) and other igneous units. With one exception the Kannaviou glasses lie on the evolved island-arc tholeiitic trend, and thus it is possible the sediments were derived from similar extrusives as those that overlie the plutonic core to the ophiolite complex.

Fig. 6 Field map of the Marathounda-Arkhimandrita area showing the regional distribution of major rock types. According to Swarbrick (1979), the anastomosing pattern of serpentinite marks the line of high angle fault contacts. The elongate sliver of Troodos Complex and Kannaviou Formation rocks was probably introduced into its present position adjacent to OMB basement units by strike-slip movement.

Fig. 7 (a) Geological map of the area to the north-east of Ayia Varvara, where both greenschist and amphibolite facies metamorphic rocks outcrop. (b) shows simplified cross section, modified after Swarbrick (1979) and Spray and Roddick (1981). Notice how metamorphic grade increases towards sliver of Troodos-type crust exposed to the north of the area.

Fig. 8 Alternative settings for genesis of the Mamonia Complex metamorphic rocks after Swarbrick (1979), Spray and Roddick (1981), Searle and Malpas (1980) and Robertson and Woodcock (1979).
Key

A Andesite - dacite - rhyodacite
B Basalt - basaltic andesite

{ Troodos assemblage

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<table>
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<tr>
<th>MgO wt.%</th>
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fig. 5.
THE STRUCTURAL SETTING OF THE AYIA VARVARA METAMORPHIC ROCKS

- Greenschist and lower grade sedimentary and alkali extrusive protoliths, grade and deformation intensity increases towards 'Troodos' ophiolite slivers
- Amphibolites—MORB alkalic and sedimentary protoliths, max. est. T 600°C, local grade variations, dominant S1 schistosity, F3 kinks imply dextral strike-slip, main fabric parallel to boundary faults
- Steep-dipping sheared low temperature serpentinite intrusions
- Low-angle sheets - Cret. passive margin sediments
- Sharp metamorphic grade changes across thrusts
- Metasomatism at contact
- Troodos-type ophiolitic rocks of transform fault affinities disrupted in strike-slip zone

Phosoulo luvos
K Kholetria limestones cherts
AV Ayia Varvara metamorphics
Ph Phosoulo lavas
S Serpentinite
K Kholetria limestones cherts
TC Troodos Complex
Fault
Foliation, dip & strike
Bedding
Graded contact
Inferred
Pillow lava orientation
Young Troodos crust

Shear heating along strike-slip faults

METAMORPHISM OCCURS AS 'HOT' CRUST PASSES TOPographically LOWER BASEMENT IN TRANSFORM OFFSET

A) SWARBRICK (1979)

ALTERNATIVE MODES OF GENESIS FOR THE MAMONIA COMPLEX METAMORPHICS

B) SPRAY AND RODDICK (1981)

ALTERNATIVE MODES OF GENESIS FOR THE MAMONIA COMPLEX METAMORPHICS

Amphibolites, MORB and sedimentary protoliths, S2 schistosity dominant, partial melts, amphibolite facies metamorphism c. 650-750°C.

Mantle sequence

Sheared peridotite, serpentinised, mylonitic bands

Downward protrusions of sheared serpentinite with inclusions

C) AS AN OPHIOLITIC SOLE, AFTER SEARLE AND MALPAS (1980)

Imbriated Tr. to Lr. Cret., passive margin-basin plain successions

Within-plate extrusives (alkalic), deep water sediments and Triassic neritic limestone blocks

D) ROBERTSON AND WOODCOCK (1979)

E) THE PREFERRED MODEL

MORB OCEAN CRUST (UTRI-LJUR IN AGE?)

LOCALISED TRANSPRESSION
the Late Cretaceous-Early Tertiary bimodal within-plate volcanics of the Kyrenia Range (see chapter 6 section 6.3, Robertson and Woodcock in press).

As quite coarse terrigenous sediments (including polycrystalline quartz) are found only several metres above the stratigraphically highest Troodos extrusives and metalliferous sediments along the western flanks of the main Massif (e.g. at Kinousa, see fig.1) it is clear that soon after its genesis the Troodos spreading ridge axis was located close to a source of continentally derived sediments (late Campanian-early Maastrichtian). Thus, the Kannaviou volcanogenic sediments were derived from localised, predominantly intermediate calc-alkaline volcanic centres, which were spacially closely associated with both the Troodos ridge system and the ‘Mamonia’ continental margin.

5.6 ACTIVE DEFORMATION OF THE MAMONIA CONTINENTAL MARGIN: STRIKE-SLIP JUXTAPOSITION OF OMB AND OMB BASEMENT SLIVERS

5.6.1 ARCULATE FAULT LINEAMENTS

Swarbrick (1979) recognised that OMB and YMB basement blocks have been tectonically juxtaposed along a pair of extensive fault lineaments, along which intruded serpentine occurs, together with elongate slivers of both OMB and YMB lithologies. Small volumes of greenschist and amphibolite facies metamorphic rocks also crop out locally within the fault lineaments.

5.6.2 SHEARED SERPENTINITE STRANDS

In the Arkhimandrita-Marathounda area (see fig.1, fig.2b and fig.6), a 3km. wide fault lineament is defined by numerous thin (typically 5-10m. thick) anastomosing strands of sub-vertical sheared serpentine (e.g. south west of Stavrokono) while in the north (fig.2a) there is a simple, but wider (typically 10-250m. thick) zone of intruded serpentinite (e.g. at Statos or Mavrokolymbos Dam). In all areas harzburgite appears to be the main protolith. Swarbrick (1979) suggested that the serpentinites were generated by low temperature alteration of ultramafic ophiolitic rocks at depth, followed by protrusion up high angle faults within a transtensional strike slip regime. Similar serpentine intrusions have been reported from the San Andreas Fault system (Dickinson 1966).

5.6.3 SLIVERS OF OLDER MESOZOIC BASEMENT IN THE ZONE OF HIGH ANGLE FAULTING.
In the Arkhimandrita-Marathounda area, slivers of OMB lithologies (i.e. Dhiarizos Group volcanics and associated sediments) occur as steeply dipping, strongly sheared and flattened screens (up to 1km. wide and 2km. long) between sheared serpentinite strands (e.g. along the Nea Kholetria-Stavrokono road section, see fig.6). Towards the fault lineament, regionally gently inclined Ayios Photios Group sedimentary sheets overlying Dhiarizos basement become more steeply dipping, with axial planes locally dipping south east (e.g. near Mavrokolymbos Dam, Swarbrick 1979). Detailed structural studies indicate that the OMB lithologies were rotated to near vertical within the strike slip lineaments subsequent to the earlier tectonic juxtaposition of the upper tectonic unit (Ayios Photios Group) against the lower tectonic unit (Dhiarizos Group, see Swarbrick 1979, 1980).

5.6.4 YOUNGER MESOZOIC BLOCK SLIVERS: GENESIS CLOSE TO AN ACTIVE TRANSFORM DOMAIN?

Blocks of ophiolite basement enclosed within the sheared serpentinite zones range from metre-sized detached blocks to large 300m. wide x 1200m. long fault bound slivers. Palaeohorizontal surfaces are often only moderately inclined, but in some areas are locally overturned (e.g. in the Marathounda area and on the east margin of the Akamas basement sliver, see fig.2a) with basement slivers being generally less highly sheared than OMB counterparts. Typically, the YMB basement slivers are conformably overlain by Kannaviou Formation volcanogenic sediments, and this implies that they have an origin similar to that of the Troodos crust exposed on the main Massif and on the Akamas Peninsula (Robertson 1977).

Swarbrick (1979) however, recognised important differences in the ophiolite crustal stratigraphy of the basement slivers in the Arkhimandrita-Marathounda area to that of the main ophiolite complex. Field studies indicate that; a) successions through the ophiolite lithostratigraphy in this area are typically anomalously thin (200m. in the Ezousa valley and 35m. near Marathounda, compared with >1000m. on the main Massif); b) Coarse gabbros are commonly intruded by pyroxenite dykes and sills (Ezousa valley) and c) the extrusive breccias of the succession are texturally similar to the clinoenstatite-phyric basalts preserved within the Arakapas Fault Belt.

Importantly, further to the north, along the periphery of the Troodos Massif and the Akamas basement sliver, fault scarp type lava breccias locally dominate the Troodos extrusive succession (see chapter 4, section 4.4 and 4.5) and the Kannaviou Formation is locally underlain by red haematitic volcanlastic sediments typical of the Arakapas Fault belt (e.g. at Statos, Robertson 1977). Thus, it is conceivable that the original Arakapas oceanic transform fault domain extended westward into the Mamonia Complex of south west Cyprus.
5.6.5 GREENSCHIST AND AMPHIBOLITE FACIES METAMORPHICS

5.6.5.1 FIELD RELATIONSHIPS

Locally sandwiched within the fault lineaments are polyphase-deformed slivers of metamorphic rocks up to 1.5km wide x 2km long (Lapierre 1975, Swarbrick 1980, Spray and Roddick 1981). The southern outcrops around Ayia Varvara (see fig.6 and fig.7) are less disrupted than in the north (Loudra tis Aphroditis), where the metamorphic rocks are cut by major NE-SW striking faults. In both areas two metamorphic assemblages are separated by steeply northerly (in the south) and easterly (in the north) dipping thrust faults. The volumetrically larger (c. eight times) assemblage consists of amphibolites and micaschists interstratified with marbles and quartzites. Mineralogical studies and mineral compositions indicate that these rocks have been metamorphosed at an epidote to low amphibolite facies grade at an inferred maximum temperature of 600°C (Spray and Roddick 1981). 'Immobile' trace element analyses suggest protoliths of alkalic basalts similar to volcanic rocks of the Dhiarizos Group, but some compositions are closer to MORB, unlike either the OMB (Dhiarizos Group) or the YMB (Troodos) extrusives (Spray and Roddick op. cit.). Generally the assemblages are representative of relatively uniform PT conditions, although grain size changes and mineral compositions point to northward temperature increases (in the south, see fig.7b). Structural fabrics are dominated by a moderately-to-steeply dipping schistosity (S1) with axial planar isoclinal folds (F1) and co-axial lineations (L1). Formation of this fabric near the metamorphic peak in a semi-plastic state was followed by later more open folds (F2) then by co-axially refolded kinks (F3) which locally are diagnostic of dextral strike slip motions along the fault lineament. Hornblende separates (2 from Ayia Varvara in the south and 2 from Loudra tis Aphroditis in the north) were dated by the 39Ar-40Ar technique. Isochron plots shows that rocks have undergone primary metamorphism at c.83-90 Ma. (Santonian-Campanian; Spray and Roddick 1981), a similar age to that of the main Troodos ophiolite complex as determined by isotopic methods (Delaloye et al. 1980).

By contrast, the volumetrically minor greenschist metamorphic facies assemblage exposed at Loudra tis Aphroditis is represented by metalavas, metacherts, marbles, quartzites, phyllites and micaschists. Field and mineralogical studies suggest protoliths similar to the volcano-sedimentary successions of the Dhiarizos Group (OMB; Robertson and Woodcock 1979, Spray and Roddick 1981 and Swarbrick 1980). Most outcrops show abrupt changes from sheared but unmetamorphosed Dhiarizos Group lithologies, to foliated meta-lavas and recrystallised sediments of the greenschist facies assemblage, then amphibolite metamorphics (fig.7b). At each locality schistosity is parallel to the bounding faults. In the Ayia Varvara area YMB lithologies are
juxtaposed with the highest grade metamorphics. Metastomatism has been extensive where amphibolites and serpentinite are in contact.

5.6.5.2 ALTERNATIVE MODES OF GENESIS FOR MAMONIA COMPLEX

METAMORPHIC ROCKS

Originally, Swarbrick (1979) suggested that metamorphism was associated with shear heating as hot newly formed Troodos-type oceanic crust was tectonically juxtaposed against cooler basement of the Dhiarizos Group along an active strike slip lineament (see fig.8a). In support of this hypothesis he cited the steep metamorphic fabrics and their parallelism with the fault lineaments as being indicative of metamorphism within a strike slip regime. A pure strike-slip mode of genesis is considered unlikely however as; a) The dominant F1 isoclines are diagnostic of dip-slip rather than strike-slip displacements; b) The regionally consistent metamorphic fabric and structural history is unexpected in a complex strike-slip regime and c) The sub-parallel orientation of the palaeohorizontal in unmetamorphosed units (for example the Dhiarizos Group lavas at Ayia Varvara see fig.7b) and the fabrics in metamorphosed units might be equally consistent with a post-metamorphism tectonic rotation for the steep fabrics.

In a second model, Spray and Roddick (1981) favoured metamorphism associated with an offset spreading ridge axis (see fig.8b). In this setting metamorphism was considered to occur when young, topographically elevated crust in the vicinity of the ridge axis passed by older and cooler crust created on the opposite side of the transform lineament. In support of this model Spray and Roddick (1981) described how similar lithologies to those preserved within the Mamonia Complex (amphibolites, metagabbros and sheared serpentinites) have been dredged from Atlantic fracture zones. However, it is clear that these metamorphic rocks cannot be directly compared with those of south west Cyprus as; a) The characteristic MORB composition of amphibolite metamorphic rocks in the Mamonia Complex cannot be consistent with a derivation from Troodos-type crust; b) Substantial volumes of siliceous and calcareous sediments with passive margin affinities are not expected in typical open-ocean ridge transform offsets; c) To create the required geothermal gradients for amphibolite genesis, a significant contrast in topographic relief might be expected to juxtapose 'hot' and 'cold' oceanic crust. This is not consistent with reconstructions of the ocean floor palaeo-topography along the northern margin of the Massif which indicate the extrusive series was not significantly rifted, with detailed mapping in the Margi area (north east Troodos, see chapter 2; Boyle and Robertson 1984) indicating normal fault throws of only a few hundred metres, in marked contrast to the topography of the Mid Atlantic Ridge system; d) Amphibolites like those of the
Mamonia Complex do not occur along the Arakapas transform fault zone to the east.

In a third model, the metamorphics might be compared with sub-ophiolite sole rocks like those of the Semail nappe, Oman (Searle and Malpas 1980) and the Bay of Island Complex, Newfoundland (Jamieson 1980). A comparison of fig.7b with fig.8c indicates that there are a number of lithological and structural similarities between the metamorphic sole rocks and the Mamonia complex rocks but they differ in that; a) a major ophiolitic nappe never overthrust the Older Mesozoic Blocks, with the upper tectonic unit (Ayios Photios Group sediments) only showing evidence of high level deformation; b) folds in the upper tectonic unit (passive margin sediments) face toward, rather than away from the ophiolitic rocks of the main Massif; c) amphibolites are volumetrically more important than greenschists; d) the metamorphic rocks of the Mamonia Complex show a less complex structural history than the Oman metamorphic sole rocks of the Semail ophiolite; e) characteristic metamorphic gradients were demonstrably less steep than in most 'sole' rocks; f) maximum temperatures were lower (c. max. 600°C) than at the base of metamorphic soles where granulite facies rocks are typically preserved; and g) Troodos ophiolitic slivers in the Arkhimandrita-Marathounda area preserve an undeformed in situ sedimentary cover of Campano-Maastrichtian Kannaviou Formation volcanioclastics and Maastrichtian/Palaeocene pelagic carbonates (see fig.4). This precludes any Late Cretaceous overthrusting of an ophiolite nappe onto the continental margin rocks preserved in the Mamonia Complex.

In a fourth model, Robertson and Woodcock (1979) attributed the formation of the Mamonia Complex metamorphic rocks to the 'juxtaposition by transform faulting of hot, near axial oceanic crust with cold continental margin rocks'. Displacement occurred along a major sinistral trench-trench transform fault located along the Antalya margin to the west (see fig.7d). Although this model is attractive in many respects, regional geological considerations now indicate that the Mamonia Complex and the Antalya continental margin were probably not juxtaposed in Late Cretaceous time (see chapter 6, section 6.6.1).

In the light of the regional tectonic setting, it is more probable that the metamorphic rocks formed along a major fracture zone (an extension of the Arakapas Fault belt) in which young, buoyant crust was tectonically overlapped against topographically lower, older marginal MORB-type crust and its sedimentary cover. In this setting the metamorphic rocks would probably be formed along the hanging wall of a steeply dipping fracture zone experiencing localised overthrusting in transpressional stress regime. As the radiometric ages of the amphibolites are coeval with Troodos genesis, this is consistent with formation in a transform-offset spreading setting.

Thus, pervasive strike-slip tectonics juxtaposed marginal oceanic crust (Dhiarizos Group), MORB and transform-related Troodos crust. The metamorphic
rocks were formed in a transform setting and were then disrupted and rotated within the strike-slip lineament in a transpressional stress regime.

5.6.6 HIGH LEVEL SYNTECTONIC AND POST TECTONIC UNITS

5.6.6.1 THE KATHIKAS MELANGE-MULTIPLE DEBRIS FLOWS

Both the OMB and YMB basement units are locally overlain by up to 270m. of matrix-supported conglomerates which have been identified as multiple debris flows (see fig.1). This Kathikas melange (Kathikas Formation) has been described by Swarbrick (1979) and Swarbrick and Naylor (1980). Clasts derived from all the OMB and YMB lithologies are recognised within the melange, although no massive serpentinite clasts have been identified (see below, section 5.6.6.2). The Kathikas melange is regionally undeformed and contains thin interbeds of Maastrichtian pelagic chalks (Swarbrick and Naylor 1980). In stratigraphically higher sections of the melange, local decreases in both debris flow and pelagic chalk interbed thicknesses implying a cyclical steepening of the source areas, which on the basis of regional thickness variations are considered to have been located along the major fault lineaments. Close to the fault lineaments, the typical debris flow units pass laterally and downwards into a matrix-free block melange, which shows evidence of having been penetratively sheared (e.g. near Mamonia, see fig.1). Throughout south west Cyprus, the Kathikas Formation is conformably overlain by Maastrichtian and Early Tertiary pelagic carbonates of the Lefkara Formation (see section 5.6.6.3).

5.6.6.2 MASSIVE SERPENTINITE GRAVITY FLOWS

Locally the Kathikas melange is overlain or interfingers with low angle sheets of massive serpentinite up to tens of metres thick (Mavrokolymsos Dam, Akamas and the Xeros valley; Swarbrick 1979). The serpentinite is predominantly harzburgite-derived and has a massive to blocky fabric cut by occasional thin low angle shear zones. In some areas, individual serpentinite sheets can be traced toward and into fault lineaments, where they progressively steepen to sub-vertical and root downwards (e.g. Akamas, Mavrokolymsos see fig.2a). Local overthrust relationships confirm that the intrusion of the serpentinite sheets necessarily post-dates the tectonic juxtaposition of OMB (e.g. Ayios Photios Group at Mavrokolymsos Dam) and YMB (e.g. Kannaviou-type area). Like the Kathikas melange, the massive serpentinites are locally overlain by Maastrichtian and Palaeocene pelagic carbonates (e.g. Mavrokolymsos and Akamas areas; Turner 1971). Swarbrick (1980) suggested that the massive serpentinites were extruded from the fault lineaments and, under the influence of
gravity flowed onto both OMB and YMB alike (see chapter 6, fig.10e).

The Kathikas melange and the massive serpentine sheets can be interpreted in the context of active strike-slip tectonics. Earlier strike-slip was dominantly transpressional and caused localised uplift, with the shedding of the Kathikas debris flows onto topographically lower Older and Younger Mesozoic blocks away from the zone of active deformation. However, later strike-slip deformation was more transtensional, allowing hydrated ultramafic rocks within the fault lineaments to rise up and flow out onto the surrounding topographically subdued seafloor.

5.6.6.3 THE POST TECTONIC SEDIMENTARY COVER

The sedimentary cover overlying the zones of high-angle faulting documents the end of strike-slip tectonics in south west Cyprus. Although in the north (e.g. Akamas Peninsula) mid-Tertiary erosion has largely removed any older cover (Robertson 1978), in the south (e.g. Arkhimandritita) the strike-slip lineaments disappear beneath undeformed pelagic sediments, correlated regionally with the Upper Palaeocene-Lower Eocene Middle Lefkara Formation (Mantis 1970). Although additional age data on the chalk cover is clearly required, it appears strike-slip ended at some time during the Maastrichtian-Palaeocene interval.

5.7 A PRELIMINARY PALAEOMAGNETIC STUDY IN SOUTH WEST CYPRUS

As the mean primary remanent magnetisation vector retained by the extrusive series on the Akamas Peninsula YMB is indistinguishable from the stable westerly directed remanence vector determined for the main Massif, both basement units have undergone a 90° anticlockwise rotation subsequent to crustal genesis in Turonian times (see chapter 4, section 4.5). Although no palaeomagnetic data exist for the in situ supra-ophiolitic sedimentary cover overlying YMB basement on the Akamas Peninsula, it is likely that the rotation of this detached YMB occurred during the same upper Campanian-Lower Eocene time interval as the main Massif. In an attempt to provide additional constraints on the structural relationship within, and between, OMB and YMB basement in the Mamonia Complex, a limited number of orientated samples were collected from the Lower (Dhiarizos Group) and Upper (Ayios Photios Group) structural units.

In all, 25 handsamples were collected from four localities (see fig.1), and from these 86 subsamples were prepared for palaeomagnetic analysis. Where possible, sampling sites were located as far from the deformed zones close to the high angle fault lineaments. As regular surfaces were rare in rock outcrop, 1" diameter plastic discs were attached to the rock to provide an artificial flat plane with which the
Fig.9 Orientating irregular hand samples by attaching flat plastic disc to upper surface. After extracting the sample from the rock outcrop, the lower surface could be cut with a circular saw to provide a flat base on which the rock could rest during drilling. Ideally, this surface is cut parallel to the upper disc surface so that cores can be drilled perpendicularly to this plane.

Fig.10 Representative thermal demagnetisation plot. Phasoula lava samples. As there is little change in remanence intensity during progressive AF demagnetisation the principal carrier is considered to be haematite. Thus, only thermal demagnetisation is effective in removing secondary unstable components.

Fig.11 Summary of the magnetic directions and properties of the Older Mesozoic Block lithologies, including Dhiarizos Group basalts and Vlamborous Formation sandstones (both are Late Triassic in age). (a) and (b) are stereographic projections of cleaned remanence data from Dhiarizos Group (4 localities, 86 subsamples) and Vlamborous Formation (4 localities, 318 subsamples) respectively. Directions have been corrected for simple tectonic tilt, with structural corrections in Dhiarizos Group lava terranes being based on the attitude of sedimentary laminations in primary micritic limestone interbeds (Kholetria Member). High coercivity haematite is a major remanence carrier in both sediments and lavas (this cannot be assumed to be a primary carrier); and so thermal demagnetisation was preferred as the most effective cleaning method. All samples were demagnetised at 350°C, the temperature at which the stable remanence component was considered to be isolated. (d) and (e) show typical demagnetisation plots for Vlamborous Formation sandstones. In (d) the remanence is dominated by a shallow northerly directed component, while in (e) following the removal of a shallowly inclined northerly directed component, the remanence is clearly defined by a southerly directed component, antiparallel to the lower blocking temperature secondary component.
3cm. diameter plastic disc

Orientation arrow

Irregular bottom
surface planed off

FIG.9 HAND SAMPLING OF DHIARIZOS GROUP LAVAS
FIG. 10 THERMAL DEMAGNETISATION OF PHASOULA LAVA SAMPLE
FIG. 11: MAGNETIC REMANENCE DIRECTIONS AND PROPERTIES OF DHIARIZOS GROUP LAVAS AND VLAMBOROUS FORMATION SANDSTONES
sample could effectively be orientated. Cores were then prepared in the laboratory by drilling perpendicularly to this flat surface (see fig.9). The Dhiarizos Group lavas are typically purple, vesicular and non-porphyritic and are thus easily distinguished from Troodos lavas which are characteristically green, vesicular and porphyritic. Structural corrections were based on the attitude of well bedded hemipelagic sediments that interdigitate with the pillows (Kholetria Member), and as these are typically steeply inclined, significant declination errors were necessarily introduced upon application of the structural correction.

In this preliminary study, following measurement of the NRM with a 'Molspin' magnetometer, the remanence of four pilot samples was investigated by demagnetising two samples in alternating fields (2.5mT steps up to 10mT and thereafter at 5mT steps up to 80mT) and the remaining two samples in the furnace (50°C incremental steps up to 650°C). As the MDF of each pilot sample was not reached in fields of less than 80mT, the remaining samples were demagnetised at 350°C, the temperature at which the stable remanence component was considered to be solely contributing toward the remanence (see fig.10). Following 'blanket' demagnetisation, the 'cleaned' remanent magnetisation directions determined for the four sites in the Dhiarizos Group extrusives show a broadly bimodal distribution (see fig.11a), perhaps reflecting diametrically opposed normal and reversed polarities. As the MDF of these lavas was not reached in fields of less than 80mT, high coercivity haematite is likely to be the principal carrier and cannot be assumed to be of primary origin.

A limited number of palaeomagnetic samples were also collected from the relatively structurally coherent Late Triassic Vlambouros Formation (Ayios Photios Group) that allochthonously overlies the extrusive units of the Dhiarizos Group. In contrast to the structurally underlying extrusives which are often steeply dipping and pervasively sheared close to the serpentinite fault lineaments, extensive successions through the massively bedded Vlambouros Formation sandstones are characteristically only moderately inclined. In all, 288 samples were drilled in situ from four localities, and from these 318 subsamples were prepared for palaeomagnetic measurement (see fig.1). Natural remanent magnetisations were measured with a cryogenic magnetometer as intensities were typically rather low (<2.0mAm⁻¹).

Thermal demagnetisation of the Vlambouros Formation haematitic quartzose sandstones showed that the NRM often includes antiparallel components, each with its own associated blocking temperature spectrum (see fig.11c and d). However, after blanket demagnetisation of all samples at 350°C, the remanence directions clearly record a similar bimodal distribution to the underlying Dhiarizos Group pillow lavas (see fig.11b). Assuming that the stable remanence retained by these haematitic sediments is of primary origin, the low inclination of the mean remanence vector would support a near-equatorial origin, although, unlike the lavas, the steepness of the vector can be
more accurately constrained relative to the horizontal. Vectors with a positive inclination are typically associated with a south directed remanence direction. If, as is currently favoured, the Late Triassic units of the OMB formed close to the north margin of Gondwanaland (Robertson and Woodcock 1980), along with counterparts in southern Turkey (Sengor and Yilmaz 1981), both the Dhiarizos Group crustal basement and its allochthonously overlying sediment cover have been rotated together by c. 180° in the horizontal plane after Early Cretaceous time. However, if as suggested by Lauer (1984, see chapter 7 section 7.2.3), the Triassic units originated south of the equator, the recorded declinations imply no net rotation subsequent to genesis.

By contrast with the coherent thrust sheets of the Vlamborous Formation, the relatively deformed, thinly bedded siltstones and cherts of the overlying Episkopi Formation show little consistency in remanence directions, and even after magnetic cleaning, show a general scattering in all four quadrants. Although no firm conclusions can be drawn from these data, the available remanence directions from the Dhiarizos Group and the overlying Vlamborous Formation would suggest; a) that unlike the Troodos ophiolitic basement exposed on the Akamas Peninsula and on the main Massif, westerly remanence declinations are not retained by either the Dhiarizos Group basement or its allochthonously overlying sedimentary cover, b) generally consistent declinations within the Dhiarizos Group basement would support only minimal relative rotations between the different basement outliers, consistent with the Dhiarizos Group possibly behaving as a single coherent unit during juxtaposition against Troodos-type ophiolitic crust, c) little relative rotation in the horizontal plane has occurred during the emplacement of the lower and upper structural units of the OMB.

5.8 THE MONI MELANGE OF SOUTHERN CYPRUS

Lithologies juxtaposed within strike-slip lineaments in south west Cyprus are also exposed as exotic bodies in the Moni melange of southern Cyprus. Poorly exposed equivalents of the Moni melange also crop out further east beyond Larnaca (Paralimni area, see fig.12). The following description of the Moni melange is based on a paper by Robertson (1977).

Along the southern margin of the Limassol Forest block, the highest ophiolitic extrusives are overlain by in situ umbers and radiolarites of the Perapedhi Formation before passing up into several hundred metres of bentonitic clays (Moni clays), radiolarian mudstones and non-calcareous siltstones which have been correlated both biostratigraphically and lithologically with the Kannaviou Formation of south west Cyprus (although volcaniclastic sandstones are not present). Embedded within these sediments are detached olistolith blocks up to several metres long and tens of metres thick. Lithologies similar to the Late Triassic extrusives and associated sediments
(Dhiarizos Group and Kholetria Member) are recognised, together with detached blocks (sometimes greater than 1km.) of friable orthoquartzites (Parekklisha sandstones) and impure limestones and siltstones (Monagroulli siltstones). These latter lithologies are unknown elsewhere in Cyprus, although texturally, the Parekklisha sandstones resemble the Early Cretaceous Akamas sandstone.

The provenance of the Parekklisha and the Akamas sandstones is enigmatic as no source for these lithologies is preserved in Cyprus. Robertson (1977) and Robertson and Woodcock (1979) suggested that these sediments were derived from the W or NW assuming palaeorotation together with the structurally underlying ophiolitic basement. However, Cleintaur et al. (1977) interpreted the highly mature textures as being indicative of aeolian sandstones, possibly related to the Nubian sandstones of North Africa or Arabia. Similar sandstones of the same age have indeed been identified in the Baer-Bassit area of northern Syria, there being interpreted as preserved remnants of the margin of the Arabian plate to the south (Keeir Formation of Delaune-Mayere 1984). However, Robertson and Woodcock (1980) also report petrographically similar orthoquartzites locally interbedded with radiolarian cherts in the south west segment of the Antalya Complex (see chapter 7, section 7.4.2), which was interpreted as part of the northern margin of the Troodos ocean basin (see chapter 6, section 6.4.3). An African origin is considered to be unlikely as it is clear that a significant width of Mesozoic oceanic crust still probably exists in the Levantine basin to the south of Cyprus (Makris et al. 1983). The existence of this basin would presumably have prevented African Nubian sandstones from reaching Cyprus. It is therefore more probable that the sandstones originated as coastal aeolian dunes located between a narrow carbonate platform and continental areas further inland, perhaps in a setting similar to the hinterland of the fringing reefs of Eilat on the northern Red Sea coastline (Mergner 1971). In the Early Cretaceous the shelf was tectonically destabilised and the dune sand was rapidly shed into shallow (Parekklisha sandstone) and deeper (Akamas sandstone) margin settings.

In addition to these allochthonous components there are also separate low-angle sheets of massive serpentinite up to 5km long and 60m thick, with fabrics similar to the massive serpentinites sheets in south west Cyprus (see fig.13). Sedimentary lithologies similar to those of the Ayios Photios Group in south west Cyprus (e.g. quartzose sandstones, cherts and redeposited limestones) become locally abundant in the 20m thick 'small-olistolith' melange that outcrops towards the top of the succession in the south of the area (e.g. Pedakomo-Monagroulli, see fig.12). The bedding attitude in most of the larger olistoliths is orientated towards the S or SSW contrasting with the shear-fabrics in the serpentinites that dip toward the north.

Originally the Moni melange was interpreted as having formed as a result of down-margin sliding from a Mesozoic passive margin that was, relative to the present
Fig.12 Outline geology of Cyprus showing the main structural units and basement features as delineated by geophysical surveys. Localities referred to in the text are also shown (localities in south-west Cyprus are included in fig.1). The older Mesozoic units described are located in south-west (Mamonia Complex, see section 5.2), south (Moni melange, see section 5.8) and north (Kyrenia Range, see chapter 6 section 6.3). Line of interpretative cross sections and trend of Bouguer anomaly isogals (after Gass and Masson Smith 1963) are also included. Inset shows Moni melange area.

Fig.13 Interpretative cross-section of south Cyprus to show the inferred relationships of basement blocks beneath a thick Tertiary sedimentary cover. For convenience, a composite section (see fig.12) is illustrated, compiled from existing field and bore hole data for the western area in the south (Akrotiri Peninsula) and the eastern area in the north (Monagroulli). This was considered necessary to avoid the complications of the Miocene Yerasa fold-and-thrust belt on the eastern flank of the Limassol Forest area (see chapter 4, section 4.7.1). The Arakapas oceanic fracture zone is bordered to the south by fault lineaments along which the Troodos oceanic basement was juxtaposed against older Mesozoic continental margin units (Moni melange) in Campanian-Maastrichtian time.
FIG. 12 GEOLOGICAL MAP OF CYPRUS WITH GEOPHYSICAL ANOMALIES SUPERIMPOSED
FIG. 13 SECTION ACROSS THE LIMASSOL FOREST AND AKROTIRI PENINSULA
geographical coordinates, located to the south of the ophiolite (Robertson 1977). Swarbrick (1980) suggested emplacement of detached blocks occurred by sliding off active strike-slip fault scarps.

The presence of very proximal deltaic sandstones (Parekklisha sandstones) structurally overlying ophiolitic crust of the Limassol Forest block would support significant crustal shortening prior to emplacement. In this setting it is conceivable that localised uplift, possibly associated with underthrusting, allowed Late Triassic marginal oceanic crustal elements to slide onto undeformed Troodos ophiolitic crust now located to the north. By analogy with the serpentinite gravity flows of south west Cyprus, the massive serpentinite sheets were possibly extruded from transtensional fault lineaments located immediately to the north of the present outcrop of the Moni melange outcrop and then flowed south onto the melange. The upper 'small olistolith' melange is interpreted to represent debris flows, shed from active strike-slip fault lineaments to the south, an origin similar to the Kathikas melange in south west Cyprus is inferred.

It seems highly likely therefore that a zone of major strike-slip fault lineaments extended from south west Cyprus along the south coast, separating Older Mesozoic basement from younger ophiolitic crust to the north. Certainly, this is consistent with the trace of the boundary of the Troodos Bouguer gravity high that closely parallels the southern coastline of the island (Gass and Masson Smith 1963, see 100 mgal isogal in fig.12). The Moni melange has also been discovered underlying Moni power station to the south of the present outcrop (Mantis 1977) while along the southern cliffs (Morel 1960) and underlying (Hadjistravinou and Constantinou 1977) the Akrotiri Peninsula Mesozoic radiolarian cherts are exposed without a sedimentary matrix. Lithologies are similar to the Episkopi Formation of the Ayios Photios Group in south west Cyprus and contain mesoscopic folds indicative of north-directed compression.
CHAPTER 6

THE PALAEORotation OF THE TROODOS MICROPLATE IN THE LATE MESOZOIC-EARLY CENOZOIC PLATE TECTONIC FRAMEWORK OF THE EASTERN MEDITERRANEAN

6.1 INTRODUCTION

The geophysical and geological field evidence presented in chapter 5 would strongly support the existence of a major crustal lineament that extends around the south western and southern margins of the main ophiolite Massif, broadly following the line of the present-day southern coastline of the island. Although the major fault lineament around which the palaeorotation of the Troodos ocean floor crust occurred must necessarily lie to the west of the Akamas Peninsula and to the south of the Limassol Forest Block (see chapter 4), it is clear the major strike-slip faults that juxtaposed OMB and YMB basement slivers in the Mamonía Complex were active in south west Cyprus at the time of tectonic rotation. In this chapter, I argue how the arcuate system of high angle faults that dominates the geology of south west Cyprus probably represents a segment of a major crustal fracture zone around which the Troodos oceanic crust rotated in late Campanian-Early Eocene time. The master fault is not preserved over the area of southern Cyprus, but it is inferred to run close to the western, south western and southern coastlines of the island. In an attempt to apply constraints to the size of the rotated crustal unit, I review the limited available palaeomagnetic data from the Turkish mainland and Africa to assess whether basement elements in these areas have rotated together with the Troodos Complex. It is clear however that existing data support rotation of only a small crustal unit whose delimiting northern boundary is potentially preserved within the present area of Cyprus. Appropriately therefore, I discuss the Late Cretaceous-Early Tertiary history of the Kyrenia Range, and conclude that during this time interval this former passive margin formed part of a major crustal lineament around which the palaeorotation of the Troodos oceanic crust may have taken place.

In the light of the regional plate tectonic framework, I then consider a number of potential settings in which the rotation of a small crustal unit could occur and conclude in favour of a model in which a small Cyprus-sized microplate rotated about a local Eulerian pole. The key control on the palaeorotation was probably the collision of an intraoceanic subduction zone with the Arabian margin to the east of the Cyprus area, while oblique subduction continued to the west, developing an anticlockwise rotational torque on a small crustal element stranded above a subducting plate in the Cyprus area.
This chapter was written jointly with A.H.F. Robertson and forms c.50% of a paper entitled 'The palaeorotation of the Troodos microplate in the Late Mesozoic-Early Cenozoic plate tectonic framework of the Eastern Mediterranean'. A.H.F.R. was the principal author of sections 6.3, 6.4 and 6.5.

6.2 THE SCALE OF THE ROTATED CRUSTAL UNIT

Can any constraints be placed on the size of the rotated crustal unit? Although only the Troodos ophiolitic basement records a reliable westerly directed remanence vector, Shelton and Gass (1980) speculated that the rotated unit may have been quite sizeable, and included basement elements in southern Turkey as well as crust to the south of Cyprus in the Levantine Basin (see fig. 1). Having incorrectly concluded that rotation was a late Tertiary phenomena, they proposed that major neotectonic lineaments such as the Northern Anatolian Fault, the Jordan and the Sinai Rift systems may have provided potential plate boundaries. Intuitively though, they preferred rotating a smaller crustal unit, either on the scale of Cyprus or even smaller; a suggestion supported by Robertson and Woodcock (1980) who considered rotation was limited to a minor platelet no larger than the Troodos ophiolite itself.

Using a combination of apparent polar wander path and sea floor spreading data, Livermore and Smith (1984a,b) have accurately documented the motion of stable Africa relative to Eurasia since the initiation of sea floor spreading in the Atlantic Ocean (c. 173 Ma, see fig. 2). Their reconstructions, and those of Smith and Briden (1971,1977,1981), would support only 15-20° of anticlockwise rotation for stable Africa subsequent to the genesis of Troodos oceanic crust in Turonian time (c. 91-88.5 Ma.). This result implies that the Troodos ophiolite has rotated approximately 70° relative to the northern margin of the African plate since Turonian time. During the same time interval Africa has migrated approximately 15° of latitude to the north.

Existing palaeomagnetic data for mainland Turkey adjacent to Cyprus are sparse and often, not surprisingly contradictory, since the results are largely derived from deformed igneous rocks cropping out within ophiolitic suture zones. According to Lauer (1981, see chapter 7 section 7.2.3) and in agreement with Waldron (1984a,b;see chapter 7 section 7.4.2), the eastern and western Tauride blocks that now make up southern Turkey were separated by a significant Mesozoic ocean basin, the remnants of which is preserved in a NS trending suture zone at the heart of the Isparta Angle in south west Turkey (see chapter 1 fig.2). The most reliable palaeomagnetic results come from the western Tauride block (south west segment of the Antalya Complex, see chapter 7 see section 7.5), where the mean isolated primary magnetic vector is orientated DEC=335, INC=+5, a95=20.6, N=56 within a thick succession of relatively undeformed Late Triassic mafic extrusives (Calbali Dag; Lauer 1981). Results from the
Fig.1. Possible rotation models (after Shelton and Gass 1980). Rotation of Cyprus away from Antalya Bay along a major transform fault. (b) shows the rotated Troodos Complex forming part of a much larger plate, including basement elements of southern Turkey.

Fig.2. Motion of Africa relative to Europe for the past 173Ma. Reproduced from Livermore and Smith (1984a). Diagram shows the track of a present day rectilinear geographic grid rigidly fixed to Africa. Sea floor spreading data give the present relative longitudinal positions of Africa and Eurasia, while the palaeomagnetically determined apparent pole positions give the precise relative latitudinal positions. Numbers on the track corresponding to 50°E, 30°N grid positions are ages in Ma. In the Turkish area the principal motion of Africa relative to Eurasia has been sinistral and convergent since Mid Cretaceous times. Notice periods of accelerated motion around 170Ma. and 100Ma.

Fig.3. Magnetic anomalies over Cyprus area (after Vine et al. 1973, and Aubert and Baroz 1974). Map shows resulting magnetic anomaly over south Cyprus after removal of an appropriate regional field from the total field aerosurvey. Notice prominent positive anomalies over extrusive series of the Troodos ophiolite Complex both on the main Massif and on outlying basement blocks (i.e. the Limassol Forest, the Akamas Peninsula and the Troulli Inlier). Prominent magnetic highs in the Famagusta and Larnaca areas suggest extrusive basement basement probably lies close to the surface here. Importantly, NNE–SSW trending negative anomalies exist over the Polis graben area and to the west of Famagusta. Both these anomalies match corresponding negative Bouguer anomalies (see chapter 5 fig.5). According to Aubert and Baroz (1974), the disposition of isogammas in northern Cyprus is consistent with Troodos basement being cut by a high angle fault that underlies the southern edge of the western and central Range but cuts NE across the eastern Range.
Fig. 4. Discriminant trace element diagrams for Maastrichtian-Lower Eocene igneous rocks of the Kyrenia Range (after Robertson and Woodcock 1985 in press). (a) TiO$_2$ vs. Zr. Samples plotting below the horizontal line are significantly fractionated (these are mostly rhyolites) and cannot be used for tectonic discrimination. Of the remainder, the majority of samples plot near the boundaries of MORB, within-plate and volcanic arc fields. There are no obvious systematic differences in the inferred tectonic setting for either Maastrichtian (Malounda Formation) or the Palaeogene (Ayios Nikalaos Formation) igneous rocks. (b) Cr vs. Y plot for Kyrenia Range basalts. Volcanic arc basalts possess a lower content of immobile elements (e.g. Y) relative to MORB and within-plate basalts. Cr is an index of fractionation (Pearce 1980). The Kyrenia Range basalts fall in the MORB and within-plate fields, and so none should be classified as volcanic arc basalts as determined from major element compositions (Baroz 1980).

Fig. 5. The palaeorotation of the Troodos microplate in the plate tectonic framework of the Eastern Mediterranean area (see also Fig. 10). a, a small Red Sea-type ocean basin opens, offset by major (Levant, Antalya) and minor (Mamonia) transform segments; b, Subduction is initiated in the Turonian near the spreading axis and the Troodos ophiolite forms above a subduction zone in a spreading setting: the Arakapas oceanic fracture zone forms as an extension of the transform-rifted margin offset; c, the intra-oceanic subduction zone impinges on the Arabian margin to the E. Regional compression then initiates subduction of the marginal oceanic crust immediately beneath the Kyrenia platform and reactivates the Arakapas oceanic fracture zone with metamorphism (e.g. Ayia Varvara) and localised volcanism (Kannaviou Formation). Continued compression emplaces ophiolites onto the Arabian margin prior to Late Campanian (Hatay and Baer-Bassit), while subduction continues to the E. An anti-clockwise torque is then developed on the Troodos microplate (which might include the unsubducted buoyant former spreading axis) causing it to be progressively rotated, carrying with it slivers of the already deformed older marginal units. Rotation is over by the end of Early Eocene time.
Eastern Tauride block immediately to the north of Cyprus are less definitive, but our own preliminary palaeomagnetic results from little deformed Mesozoic carbonate platform units located on both sides of the Isparta Angle (e.g. Bey Daglari and Anamas Dag, see chapter 7 section 7.4.3) record anticlockwise rotations of no more than 20°, a similar post-Late Cretaceous rotational displacement to that of the African plate to the south (Van der Voo and French 1974).

Thus, it is clear that the area of rotated crust did not include adjacent areas of southern Turkey and the Afro-Arabian platform and was thus probably quite small. It is therefore quite appropriate to look for preserved remnants of the rotated microplate boundaries in the immediate vicinity of the preserved Troodos ophiolite complex. As one of these microplate boundaries might conceivably lie beneath the Kyrenia Range lineament, where Troodos type basement is abruptly truncated by a major high angle fault (see chapter 1 section 1.3.2), I discuss below the important elements of the Late Cretaceous-Early Tertiary history of this prominent lineament, with the intention of clarifying its relationship to the ophiolite complex, which during this interval was undergoing a major tectonic rotation.

6.3 THE LATE MESOZOIC-EARLY CENOZOIC HISTORY OF THE KYRENIA MARGIN: THE DEFORMATION OF A FORMER PASSIVE CARBONATE PLATFORM

Dominating the north of Cyprus, the Kyrenia Range is a narrow convex c.160km. long lineament comprising four major northwardly dipping thrust arcs (Baroz 1979). Geophysical studies indicate that the lineament can be traced east under the Levant Sea to the Misis Mountains near the Gulf of Iskenderun and westwards into Antalya Bay (Biju-Duval et al. 1974, 1977; see chapter 1 fig. 2). Magnetic (Aubert and Baroz 1974), gravity (Gass and Masson Smith 1963) and deep well (Cleintaur et al. 1977) data indicate that deeply Troodos-type ophiolitic crust underlies the Mesaoria plain to the front of the western part of the range, where it is abruptly truncated by a high angle fault (see chapter 1, section 1.3.2). By contrast, a prominent magnetic high extending under the eastern segment of the range indicates that the Troodos-type basement probably extends further to the north in this area (Aubert and Baroz 1974, see fig. 3). The Kyrenia Range comprises four rock groups separated by major unconformities recording important deformational events (D1-D3 of Robertson and Woodcock 1985). Only events prior to and including the major Late Eocene D2 event are described here.

From Late Triassic to the Mid Cretaceous, shallow water carbonates (Trypa Group) accumulated on a basement not exposed in Cyprus. A largely dolomitic succession (Sykhari Formation) is succeeded by mostly neritic limestones (Hilarion
Formation). After the mid-Cretaceous (but prior to Maastrichtian) the carbonate platform was recrystallised, pervasively brecciated, and interleaved with slivers of greenschist (schists and pelites) and amphibolite facies (schists) grade metamorphic rocks, now particularly well exposed along the south margin of the range. This corresponds to the D1 tectonic event.

The former platform then subsided and was overlain, probably unconformably, by poorly exposed volcanlastic sediments of late Campanian-early Maastrichtian age (Kiparisso Vuono Formation; Baroz 1979). These sandstones contain sedimentary (radiolarite, quartz and limestone), volcanic (basic-acid volcanic glass, alkalic and low-K basalt) and metamorphic (marble, schist and amphibolite) clasts, together with fossils of benthonic (e.g. Orbitolines) and planktonic (e.g. Globotruncana) foraminifera. In view of the heterogeneous petrography, the source area for these sediments must have consisted of igneous extrusives, a diverse assemblage of metamorphic rocks as well as deep and shallow water Mesozoic sediments (radiolarites, micrites and neritic limestones). Robertson and Woodcock (in press) favoured derivation from a source area close to the margin, as the nearest exposed metamorphic basement lies 250km. to the north. No metamorphic basement has been identified in the western Tauride mountains, where unmetamorphosed successions extend down to the Ordovician in the Antalya Complex (e.g. Tahtali Dag, Gutmich et al. 1979; Kesme section in Kemer zone, Robertson and Woodcock 1982, see chapter 7 section 7.4.3) and to the Cambrian in the Hadim nappes (Okay and Ozgul 1984).

The petrological similarities of the upper Campanian-lower Maastrichtian Kiparisso Vuono Formation volcanics and the Kannaviou Formation sediments of south west Cyprus provides a unique link between the two areas in Late Cretaceous time. Both the Kannaviou and Kiparisso Vuono Formations are strikingly similar in that both contain grains of fresh basic and acid volcanics, metamorphics and both shallow and deep water sedimentary rocks. Originally, Robertson (1977) believed the volcanics were derived directly from the Late Cretaceous volcanics of the Kyrenia Range (Moore 1960, see below), but this is now considered unlikely following the demonstration by Baroz (1980) that these are slightly younger (Maastrichtian and Palaeocene). As the acid-basic volcanic clasts of the Kannaviou Formation have been shown to have been probably derived from an evolved Troodos-type magma source (see chapter 5 section 5.5), it is likely that both the Mamonha Complex and the deformed Kyrenia platform were located close to Troodos ocean floor crust. This is compatible with available field evidence from south west Cyprus which indicates that Mamonha continental margin and Troodos ocean floor crust were juxtaposed in upper Campanian-lower Maastrichtian time (see chapter 5 section 5.1). It is considered likely that the Troodos crust was similarly juxtaposed against the Kyrenia margin during this time interval. An alternative interpretation, is that the Troodos ophiolitic crust, now
located to the south, was emplaced during a D2 (Late Eocene) thrusting event related to northward subduction of Troodos crust immediately to the south of the Kyrenia platform.

The pervasively brecciated and metamorphosed basement together with its patchy volcaniclastic cover was then unconformably overlain by Maastrichtian pelagic carbonates (Lapithos Group) which are interstratified with two series of bimodal basic-acid volcanics ranging from Maastrichtian (Melounda Formation) to Late Palaeocene-Early Eocene (Ayios Nikalaos Formation) age. Originally, Baroz (1980) reported that the major element chemistry of this distinctive bimodal suite was diagnostic of calc-alkaline to shoshonitic trend. Their chemistry appeared to be consistent with genesis above a northward dipping subducting plate prior to southward obduction of the Troodos and Kyrenia margin rocks onto the African continent, as proposed by Ricou et al. (1984). Analysis of 'immobile' trace elements in these highly altered extrusives now shows that all the lavas are of similar within-plate type, with apparently no identifiable above-subduction zone component (Robertson and Woodcock 1985, see fig. 4).

Penetrative faulting of the basement accompanied fissural eruptions, with submarine screes being shed off active fault scarps into deeply submerged areas undergoing pelagic carbonate deposition. Breccias are particularly common at the base of the succession, these presumably being derived from fault scarps that cut through the already deformed and metamorphosed Trypa Group basement. In Late Eocene time, the Lapithos Group, together with its basement then underwent major south-directed thrust emplacement (D2, see chapter 1 section 1.3.3).

6.3.1 AN OUTLINE TECTONIC HISTORY

In a recent synthesis Robertson and Woodcock (in press) considered that prior to deformation in the Late Cretaceous, the Kyrenia margin evolved as a stable gradually subsiding carbonate platform, similar to other marginal units in the Neotethyan area. During the Late Cretaceous, but prior to the deposition of the Kiparisso Vuono Formation, the original external passive margin successions were removed (either by underthrusting and/or strike-slip) and Troodos-type oceanic crust was juxtaposed adjacent to the now active margin by strike-slip tectonics. Tentatively, they proposed that active deformation of the Trypa Group basement could have been related to the initiation of northward subduction beneath the platform in pre-Maastrichtian times. This event corresponds to the early 'tangential tectonics' of Baroz (1979), during which the platform edge was sheared and compressed, leading to the pervasive brecciation of carbonates and the genesis of low-grade psammitic and pelitic rocks along active fault strands. Substantial dip-slip components along these
faults allowed amphibolite facies meta-sedimentary and meta-igneous rocks, presumably formed by shear heating, to be tectonically transported upward along the faults.

Here a two-stage model is adopted to account for the complex Late Cretaceous-Early Tertiary deformation along the Kyrenia lineament. In this scheme, earlier northward-directed underthrusting (related to Troodos crustal consumption) beneath the south-facing Kyrenia platform removed the external passive margin succession and pervasively deformed the adjacent carbonate basement. Later, following the initiation of the palaeorotation of the Troodos ocean floor crust to the south, the lineament became dominated by strike-slip tectonics as the ophiolite was tectonically juxtaposed adjacent to the former passive margin along a major dextral transcurrent fault (see section 6. 7). Although initially the lineament was dominated by transpressional tectonics, during the Maastrichtian and Palaeocene, intraplate volcanics were extruded onto the subsided platform from fissures in a presumably transtensional stress regime. The localised acidic volcanism of the western range could conceivably be related to genesis in a transpressional stress regime, causing greater fractionation and crustal assimilation during magma ascent.

Although the master fault around which the rotation of the Troodos ocean floor crust occurred must necessarily run to the west of the Akamas basement sliver, available geological and geophysical data is consistent with an extention of the prominent crustal lineament preserved in south west Cyprus towards the southern front of the Kyrenia Range. Certainly, complementary strike-slip tectonics in the Mamonia Complex of south west Cyprus can be closely correlated with contemporaneous deformation along the inferred southern margin of the Kyrenia platform to the north. Although major lateral displacements of fault-bound slivers occurred in pre-Tertiary time, strike-slip deformation in the Mamonia Complex, like that along the Kyrenia lineament, was demonstrably characterised by two independent phases. During an earlier transpressional phase, OMB and YMB were tectonically juxtaposed along faults with considerable dip-slip components. Basement juxtaposition was associated with metamorphism, pervasive shearing, local overturning and the shedding of multiple debris flows into fault-bound depressions. By contrast, later deformation was dominantly transtensional, allowing both massive serpentinites to flow laterally onto adjacent ocean floor crust and Kannaviou Formation volcanioclastic sediments to accumulate along the length of the lineament to the west of the elevated Troodos oceanic crust of the main Massif. Importantly, the thick bentonitic clays (lateral equivalents of the Kannaviou Formation) that occur around the uplifted Massif (e.g. beneath the Mesaoria, Cleintaur 1977; in the Cape Greco area, and beneath the Akrotiri Peninsula, Mantis 1977) imply that during upper Campanian and early
Maastrichtian time the preserved Troodos ocean crust was topographically elevated and so does not preserve an extensive volcanogenic sedimentary cover. It is possible therefore that the former spreading axis is preserved within the present exposed area of the Troodos ocean floor crust.

6.4 A REGIONAL TECTONIC SETTING FOR THE PALAEOROTATION

According to Robertson and Woodcock (1980), at the time of the Late Cretaceous-Early Tertiary palaeorotation, the preserved Troodos ophiolite complex formed part of a narrow Neotethyan ocean basin that was located immediately to the north of the stable Africa and to the south of the complex mosaic of Tauride microcontinental units preserved in southern Turkey (see fig. 5a). This 'Troodos ocean' was located to the south of the main Tethys ocean (or Palaeozoic Palaeotethys sensu stricto) which was probably consumed further to the north beneath the Pontide block of northern Turkey (see chapter 7, section 7.2).

6.4.1 THE NORTHERN MARGIN OF THE TROODOS OCEAN BASIN

Accepting that the Troodos ophiolite complex rests in an essentially in situ position relative to the surrounding autochthonous basement elements in the Eastern Mediterranean area, and has not been tectonically transported southward across the Tauride carbonate platform units, as envisaged by Ricou (1975, see chapter 7 section 7.2.1), the Alanya Massif, cropping out on the east side of Antalya Bay (see chapter 1 fig. 2), probably originally lay on the northern margin of the Troodos ocean basin close to the Kyrenia platform unit. In a preliminary field study, Okay and Ozgul (1984) showed that the Alanya Massif comprises a complex polyphase-deformed greenschist and blueschist facies metamorphic nappe pile, composed predominantly of Palaeozoic carbonate and siliciclastic rocks, that was emplaced northward over the adjacent Akseki carbonate platform unit (see chapter 1, fig. 2) in Late Eocene time. Contrasting with other autochthonous basement elements in the Tauride belt, no thick Mesozoic platform successions overlie Triassic rift-related sediments on the Alanya Massif, and this implies that this area was probably subaerially exposed during much of Mesozoic time. As passive margin successions now structurally underlie the northern margin of the Alanya Massif (Okay and Ozgul 1984, Robertson unpublished data) it is likely that a minor ocean basin once separated the Alanya Massif from other Tauride platform units to the north in Mesozoic time. It is quite possible therefore that the Alanya Massif originated as an off-margin microcontinental unit that lay close to the Kyrenia carbonate platform. This is consistent with the petrography of lithologies preserved within the Mamonía Complex and the Moni melange which have been shown to contain
a major terrigenous component (e.g. Early Cretaceous Parekklisha sandstones, Akamas sandstones and other arenaceous rocks of the Episkopi Formation), unlike some other preserved Mesozoic passive margin units in the Eastern Mediterranean area (e.g. south west Antalya Complex, Robertson and Woodcock 1982; north east Antalya Complex, Waldron 1984 a,b; and Baer-Bassit, Delaune-Mayere 1984). Thus, it is conceivable that passive margin units preserved in the Mamonia Complex and in the Moni melange originated from a southward, or eastward facing Alanya margin that was bordered by a narrow fringing reef. This allowed carbonate sediments to be rapidly redeposited into deep water areas close to the margin (e.g. debris flow carbonate conglomerates of the Episkopi Formation, see chapter 5 section 5.4.3), while mature terrigenous clastics, derived from a subaerially exposed microcontinent, accumulated on a narrow shelf behind the platform. Although the adjacent Kyrenia platform might have developed on an easterly extension of the Alanya microcontinent, the preferred model shows how the platform developed along the southern margin of a separate Tauride platform unit (possibly an eastward extension of the Akseki platform unit, see fig.5).

Further to the east, the Tauride platform units were probably offset northward by the Ecemis fault, while beyond the northern basin margin runs into the Late Cretaceous-Early Tertiary active margin terrane of Eastern Anatolia (Aktas and Robertson 1984). To the north west of the Kyrenia platform, a significant NS trending ocean basin was bounded to the east (Karacahisar and Anamas Dag) and west (Bey Daglari) by Bahaman-type carbonate platforms (Waldron 1984 a,b; see chapter 7 section 7.4.2). Closure of this basin occurred in Upper Palaeocene-Lower Eocene time, leading to the complementary thrusting of off-margin successions onto the stable platforms on either side of the Isparta angle. If closure of this ocean basin involved the westward translation of the Akseki platform relative to Africa, the Kyrenia platform may well have originally been situated further to the east during the Mesozoic.

6.4.2 THE SOUTHERN MARGIN OF THE TROODOS OCEAN BASIN

The southern margin of the Troodos ocean basin is well preserved to the east of Cyprus in Baer Bassit, Syria (see chapter 1, fig.2). Here, characteristic sedimentary and volcanic passive margin successions are believed to preserve the remnants of an originally northward facing continental margin (Delaune-Mayere 1984). The structurally overlying Baer Bassit ophiolite and less deformed Hatay ophiolite exposed to the north (see chapter 1, fig.2) are considered to have originally formed above a subduction zone, and were then emplaced southward onto the northward facing Arabian platform edge following trench-margin collision. Obduction was complete by upper Campanian times as the emplaced ophiolite was unconformably overlain by shallow marine sediments (Selcuk 1981). Importantly, the sheeted dyke complex preserved in
the Hatay ophiolite is orientated east-west, and appears not to have been significantly rotated like the Troodos oceanic crust.

During the Mesozoic the southern margin of the Troodos ocean basin probably closely followed the present NNE-SSW Levant margin before turning again toward an EW orientation in the vicinity of the Nile cone (Garfunkel and Derin 1984). This is consistent with seismic reflection profiles from Sinai to the Levantine Sea which indicate a marked thinning from continental to oceanic crust over the inferred line of the Mesozoic continental margin (Ginzburg and Gvirtzman 1979). Seismic refraction data of Makris et al. (1983) confirm that oceanic crust in the Levantine basin to the south of Cyprus probably extends right up to the coastal areas of Sinai and Israel and to the west beneath the Eratosthenes Seamount. According to Makris et al. (1983), crust to the north of this prominent topographic feature appears to thicken, and lower inferred crustal densities support a basement with more continental affinities. As the Troodos ophiolite is currently located in a 'fore-arc' setting it is possible that this wedge of thickened crust represents a 'microcontinental block' that has collided with the active trench in the Neogene and is currently underthrusting the 10km. thick ophiolitic slab. Certainly, Rotstein and Ben Avraham (1984) agree that numerous oceanic plateaux that may once have existed along the southern margin of the Troodos ocean basin have significantly disrupted normal subduction processes as they approached and eventually collided with the trench (see chapter 1, section 1.2).

Thus, during the constructive phase of the Troodos ocean both the northern and the southern margins of the basin were fringed by a complex assemblage of small platform units (e.g. the Tauride platforms), subaerial metamorphic and igneous terranes (e.g. the Alanya Massif) and prominent seamounts (e.g. the Eratosthenes seamount). The relative spatial distribution of these basement elements relative to the main African continent is schematically illustrated in fig.5a.

6.4.3 THE CONSTRUCTIVE MARGIN HISTORY

Following widespread Late Triassic rifting of the northern margin of Gondwanaland, the Troodos ocean basin formed a narrow passage between the African continent and the Tauride microcontinental blocks. The Kyrenia lineament either formed part of the northern margin to this basin, or existed as an off-margin platform unit. The passive continental margin units preserved in the Mamonia Complex are considered to have formed along the eastern margin of the Alanya microcontinental unit, close to a significant transform fault offset. The initiation of ocean floor spreading in the Late Triassic (Dhiarizos Group of the Mamonia Complex) led to the development of a narrow ocean basin. Basalt breccias were shed from active fault scarps (Loudra tis Aphroditis breccias), while close by
small atolls (Petra tou Romiou patch reefs) formed on local seamounts (Phasoula lavas). As no Jurassic oceanic crust is preserved in the area, it is assumed that sea floor spreading had effectively ended by earliest Jurassic times, with only negligible generation of any MORB-type crust. Following the end of sea floor spreading, the Mamonia passive continental margin continued to subside, while carbonate sediments accumulated on adjacent platform areas (e.g. Hilarion and Sikhari Formations of the Kyrenia Range).

Active spreading probably began again in Early to Mid Cretaceous time, with renewed volcanism close to the basin margins. Bentonitic clays, tuffs and hydrothermal sediments were deposited onto marginal Upper Triassic crust (Mavrokolymbos Formation overlying the Phasoula lavas) and the adjacent passive margin (distal successions of the Episkopi Formation). During this renewed phase of ocean floor spreading, the passive margin was tectonically destabilised and mature orthoquartzites were redeposited into coastal (Parekklisha sandstone) and deeper water (Akamas sandstone) environments, while the Kyrenia platform continued to gently subside to the north.

6.5 GENESIS OF THE TROODOS OPHIOLITE COMPLEX

Any model for the palaeorotation of the Troodos ophiolite complex must be consistent with the genesis of the ocean crust above a subducting oceanic slab in Late Cretaceous (Turonian time, 91-88.5Ma). Most geologists working in the Eastern Mediterranean and Middle Eastern areas (e.g. Aktas and Robertson 1984, in the Eastern Taurides; Berberian et al. 1982 in Iran; Glennie et al. 1974 and Pearce et al. 1980 in the Oman Mountains) now agree that the predominant Late Cretaceous and Tertiary subduction direction was generally northward like that beneath the present-day Cretan arc (Fytikas et al. 1984, see chapter 1 section 1.2). A history of prolonged southward subduction along the southern margin of the Troodos ocean and its easterly extension is considered unlikely, as there is a general absence of any arc-type volcanism along the Arabian continental margin (for instance in the Hatay and Baer Bassit areas of northern Syria). Thus, if indeed the Troodos ophiolite was created above a subducting plate in the Late Cretaceous, a model that incorporates the microplate rotation into a northward closing subduction system would seem the most appropriate.

In a recent paper Moores et al. (1984) considered that the Troodos ophiolite was created in a back arc basin similar to that of the Andaman Sea. Unfortunately, in a regional tectonic framework this setting is considered inappropriate as: a) One would expect to find the products of island arc-related volcanoes overlying ocean crust formed in a marginal basin. No volcanic arc of the same age or older than the ophiolite
appears to be preserved in the area (including south west Turkey, Hatay and Baer Bassit). b) Although oceanic crust underlying marginal basins is typically variable in composition, no crust similar to the Troodos ophiolite has been dredged from contemporary basins like the Andaman Sea (Saunders and Tarney 1984). c) The deformed passive continental margin rocks preserved in the Mamonia Complex cannot have been emplaced in a conventional accretionary prism normally associated with steady state subduction. The general absence of trench-fore arc flysch sediments and a regular stacking order of thrust sheets would support an alternative emplacement origin (see chapter 5 section 5.4.4).

Pearce et al. (1984) postulated that many eastern Tethyan ophiolites, including the Troodos, formed by 'pre-arc' spreading in small Mesozoic ocean basins. In their model intra-oceanic subduction was initiated away from the continent-ocean boundary, and was followed immediately by extension above the subduction zone allowing ophiolitic crust to be generated in a spreading environment close to the active trench. The ophiolite was then obducted, or spreading stopped before any substantial arc carapace had time to form. Although this model is more compatible with many features of Tethyan ophiolites, there are problems in its application: a) A conventional marginal basin appears to form by the 'splitting' of a pre-established arc (Karig 1982). This 'arc' is not apparently preserved close to the Troodos Complex or in other associated ophiolite complexes in the area (e.g. Antalya, Hatay or Baer Bassit). b) In most conventional intra-oceanic arc-trench gaps, an area of crust older than 'oceanic' crust in the marginal basin is typically preserved. However, with the possible exception of minor occurrences of MORB-type amphibolite facies metamorphics in 'sole' rocks, no crust of this type is associated with obducted Neotethyan ophiolites.

In an attempt to resolve inherent problems with both these models Robertson and Dixon (1984; see also Dixon and Robertson 1985, Smith and Spray 1984) postulated that subduction could be initiated at, or near the spreading ridge axis when a small ocean basin was in a state of regional compression. Future obducted ophiolites were then generated above the an embryonic subduction zone as the down going plate 'rolled back'. The already depleted aesthenosphere, enriched in large-ion lithophile elements that overlies the subducting plate, then upwelled into the area created in front of the rolling back plate.

In this model the ophiolite, created in a spreading-related setting, becomes in a sense the 'arc' itself, and no true conventional arc carapace need form. The most obvious problem with this model is that young crust may be too buoyant to be effectively subducted in such a system. However, if only little Jurassic and Lower Cretaceous crust was created subsequent to initial spreading in the Upper Triassic, slabs of more thermally mature denser crust (presumably including seamounts and
microcontinental blocks) may thus have been located close to the spreading axis, thus favouring a transition from initial ridge collapse to steady-state intra-oceanic subduction. Although there are inherent problems with this model (such as the conspicuous absence of modern analogues), it does explain many of the characteristic features of Tethyan ophiolite geology (such as the near synchronicity of spreading and metamorphic sole ages, the general absence of arc edifices in the area and the genesis of 'remelted' high Mg boninitic magmas so typical of Mid-eastern ophiolites).

6.6 A PLATE SETTING FOR THE PALAEOROTATION

6.6.1 ALTERNATIVE SETTINGS FOR THE PALAEOROTATION

In this section I discuss a number of potential settings in which the palaeorotation of a small oceanic microplate within a narrow Neotethyan ocean basin undergoing regional compression might be accommodated. In each of the four possible plate settings that are schematically illustrated in fig.6, genesis of the Troodos ophiolite occurred above a northwardly dipping subduction system that may represent the locus of 'roll-back' referred to in section 6.5.

In (a) the Troodos ocean crust is rotated by non-symmetrical spreading along an east-west ridge system about a pole located close to the south eastern margin of Antalya Bay. If the Kyrenia Range has not rotated through 90° like the adjacent ophiolite complex, this margin must have been strike-slipped into its present position subsequent to the palaeorotation of the Troodos ocean floor crust. This setting is broadly analogous to the Neogene opening of the Ligurian Sea marginal basin, where both Corsica and Sardinia rotated anticlockwise away from the stable European craton (De Jong et al. 1969). Alternatively in (b) the palaeorotation is driven by a collision between a seamount or microcontinent with a subduction zone. In this setting a stranded crustal element rotates essentially in situ during incipient underthrusting to the south. Moores et al. (1984) attributed the Troodos rotation to the collision of a microcontinent with a NW-SE trending subduction zone. In their model, Troodos oceanic crust was generated in a fore-arc or back-arc setting similar to crust underlying the Andaman Sea. In (c) the ophiolitic crust is expelled from the closing Isparta Angle ocean basin as the Akseki platform is translated westwards away from the zone of active continental collision to the east. Robertson and Woodcock (1980) attributed rotation to a Miocene expulsion of ocean crust away from Antalya Bay along a major arcuate transform fault. Finally in (d), Troodos crust, originally sited adjacent to the Antalya platform or its southerly extension was rotated away to the east together with marginal remnants preserved in the Mamonia Complex. Swarbrick (1979) argued that the Mamonia Complex and the Troodos formed adjacent to a NS trending transform
Fig.6. Theoretical, but unsatisfactory models for the Troodos palaeorotation. Above-subduction zone genesis of the Troodos is assumed in each case. a, Opening of a small marginal basin about a local rotation pole; b, collision of a microcontinent with the subduction zone; c, expulsion of a crustal wedge from the closing Isparta angle seaway to the NW; d, origin along the Antalya transform passive margin followed by rotation away. Associated space problems (a, c, d) and the lack of internal Troodos deformation (b, c) are considered to be major handicaps of these models.

Fig.7. Favoured model of palaeorotation of Troodos ophiolitic crust about a local pole. A possible driving mechanism which could be related to oblique NE motion of Africa relative to Eurasia, imparting a rotational torque on Troodos crust stranded adjacent to the subducting plate. Continental margin slivers become attached to the rotating microplate. See Fig.10 for the regional tectonic setting.

Fig.8. Oceanic microplate rotations at active margins (triangles - trench, ..... - sea floor magnetic anomalies, = spreading ridge, arrows indicate direction of convergence). (b) Sunda Arc (SA) plate tectonics. Oblique consumption of oceanic crust (Indian Plate, IP) under the western Sunda Arc has lead to the displacement of a major crustal wedge along a pair of transcurrent faults in the over-riding plate. Displacement is in the direction of the transverse component of oblique underthrusting. SU, Sumatra. (a) Sea floor tectonics off Central America. The East Pacific Rise (EPR) forms an impediment to subduction at the locus of ridge-trench collision (X), causing the Cocos Plate (CO) to 'pivot' about a pole centred in the Gulf of California. Rotation is implied by the pattern of arcuate transforms and the fanning of magnetic anomalies away from the pole of rotation.

Fig.9. Sketches to show alternative subduction modes. (a) Subduction is initiated near the continent-ocean boundary followed by creation of the Troodos ophiolite near the N basin margin. As the Troodos ridge is already located close to the Kyrenia platform, only one subduction zone is required to juxtapose Troodos with a terrigenous source (Kannaviou Formation) in late Campanian-early Maastrichtian. (b) the preferred model, in which intra-oceanic subduction begins near the spreading axis in a wider ocean. A second subduction sequence then juxtaposes Troodos ophiolitic crust with the margin to the N.
TO = TROODOS
K = KYRENIA RANGE
MAM = ORIGIN OF MAMONIA
To DIRECTION OF CRUSTAL CONSUMPTION
Pie-Upper Campanian

a i) Cr Tr +

Kc bi) Kp Ka

U.Cr. Ka

Active margin

Kc Collapsed

Pre-Upper Campanian

b i) R U.Cr.

Kp

ii) U.Cr. Ka

'Sroll-back' spreading

Kyrenia margin

Kp Passive

Ka Active

fig.9
margin, and were rotated to the south east along a major arcuate transform lineament.

Each of the models (a),(c) and (d) involve translating crustal elements considerable distances laterally, a process that must be necessarily complex as it involves the creation and/or destruction of oceanic crust in the path of a rotating microplate. Model (a) is also inconsistent with field observations in the Antalya Complex which suggest that this margin was an active strike-slip and compressional zone in the Late Cretaceous and Early Tertiary, and not a passive margin lying adjacent to a spreading ridge system. Although available geophysical data is sparse, there is no evidence for Late Cretaceous and Early Tertiary oceanic crust in Antalya Bay as required by models (a) and (c). Both models (b) and (c) imply rotation was driven by a forceful collision, a mechanism that is inconsistent with the mainly undeformed state of the Troodos ophiolitic basement and its in situ sedimentary cover. It is unlikely that the wedge of thickened crust currently underthrusting the ophiolite complex collided with the Troodos ocean crust as early as the Late Cretaceous as the dramatic uplift of Massif only began in Middle Miocene times. An early collision would also necessitate a long history of Tertiary crustal consumption to the south of Cyprus to be consistent with the probable width of crust subducted beneath the Hellenic arc as inferred from the Africa-Eurasia convergence paths (Livermore and Smith 1984a,b).

6.6.2 THE PREFERRED ROTATION MODEL

The favoured model for the ophiolite palaeorotation involves the oblique northward subduction of Troodos ocean crust beneath the Kyrenia platform to the north (see fig.7), leading to impingement of the spreading ridge on the margin to the north. A young oceanic crustal unit was then decoupled and started to rotate about a pole centred on or close to the present exposed area of the Troodos Massif. Former slivers of the north basin margin then became attached to the rotating microplate and were carried southward.

If the Troodos crust was sited in a fore-arc setting close to the principal Late Cretaceous Neotethyan subduction zone, oblique crustal consumption beneath the Troodos oceanic crust might have imparted an anticlockwise torque onto the over-riding plate. Certainly, in the case of the Sunda arc, oblique convergence of the Indian plate relative to south east Asia has lead to the northwest displacement of a (>10^5 km^2) forearc crustal wedge relative to the adjacent island arc (see fig.8a). On a broader scale the complex rotations (sometimes >70°) and displacement of allochthonous terranes along the western borderland of the USA and Canada have been attributed to the long-term effects of oblique crustal consumption and the northward drift of the Pacific oceanic plate relative to the stable American continent (Coney 1980). Importantly, in this case both oceanic and continental crustal elements are
involved and there are no limits, a priori, on the extent of the rotations involved (Beck 1983). Such a driving mechanism is consistent with correlations of Atlantic magnetic anomaly and polar wander path data that indicate Africa was moving north eastwards relative to Eurasia in the Eastern Mediterranean area during Late Cretaceous and Early Tertiary time (Livermore and Smith 1984a,b, see fig.2).

As the tectonic rotation involved the interaction of a small crustal segment with a Mesozoic continental margin, it is appropriate to identify analogous settings where a similar tectonic juxtaposition exists. One area of immediate interest lies of the western margin of the North American plate. There, subduction carried the Farallon plate obliquely against the consuming margin of the Californian borderland (Atwater 1970). As a result of transform-trench intersections, the original plate has disintegrated into a number of small fragments (sometimes <10km$^2$) showing considerable independent relative rotations (for example, the Gorda and Explorer plates ). The record of earlier independently moving, but now consumed fragments is left in the Pacific plate anomaly patterns. In this setting, a possible key control on the rotation of larger lithospheric plates like the northern Cocos plate (see fig.8b) appears to be related to the failure of young, hot buoyant oceanic crust to be subducted, resulting in the spreading ridge crust pivoting toward the continental margin at the locus of ridge-trench collision (Menard 1978). Although this may well explain the rotation of larger lithospheric fragments, small oceanic fragments bounded by transform segments may rotate in the opposite sense (e.g. the Gorda plate ). Rotations up to 30-40° are recorded, considerably less than the 90° of the Troodos ophiolite. Field observations (see chapter 5, section 5.1) require the Troodos ophiolite to have been located close to the 'Mamonia' continental margin as early as late Campanian-early Maastrichtian time (only 5-10 Ma after its genesis), a possible additional component of rotation could have been provided by pivoting about a ridge-trench-transform triple junction.

It is conceivable however that the palaeorotation was initiated by some form of collisional event between Troodos oceanic crust and another crustal unit located outside the present area of Cyprus. Importantly, only c.80 km to the east of Cyprus the Hatay and Baer Bassit ophiolites were emplaced onto the Arabian margin immediately prior to the initiation of the tectonic rotation in the upper Campanian. As ophiolite obduction was probably related to the collision of a northward facing continental margin, the favoured model for the palaeorotation involves the collision of an active trench with the Arabian margin to the east of Cyprus, while subduction continued further toward the west. As oceanic crust attached to the leading margin of the African plate continued to migrate north east toward the intra-oceanic subduction zone to the south of Cyprus, an anticlockwise torque developed on the Troodos crust, driving the progressive rotation of the Troodos without significant internal deformation.
As the ring of dextral fracture zones around the rotating crustal unit might not have necessarily defined a perfect small circle, any 'space problems' during rotation are considered to be accommodated by continued local underthrusting toward the south.

Before attempting to incorporate the details of Cyprus geology into this model, one other important aspect of the Late Cretaceous-Early Tertiary history of the Troodos Complex is considered: the siting of subduction zones relative to the preserved segment of the Troodos oceanic crust.

6.6.3 RELATIONSHIP OF SUBDUCTION ZONES TO THE ROTATING MICROPLATE

The preferred model for the Troodos microplate rotation requires the existence of two subduction zones, one to generate the Troodos Complex and a second to bring the ophiolite up against the Kyrenia platform (see fig.9b). However, as it is unclear whether the more southerly subduction zone associated with Troodos genesis was intra-oceanic, I also consider an alternative configuration. In this second model (see fig.9a), subduction was initiated below the Kyrenia margin in Mid to Late Cretaceous time, followed soon after by 'roll-back' and genesis of the Troodos ocean crust close to the margin (a setting favoured by Smith and Spray 1984 for the Hellenic-Dinaric ophiolites). In this model, little relative northward translation of the Troodos ocean floor crust is required to juxtapose it against the continental marginal units to the north soon after genesis.

A number of circumstantial arguments tend to favour the genesis of Troodos oceanic crust in an intraoceanic setting like that illustrated in fig.9b. a) There is a general absence of terrigenous sediments interstratified with the Troodos extrusive series, implying that the ocean floor crust was generated some distance away from a source of continental margin sediments. It was only in late Campanian-early Maastrichtian time that quite coarse terrigenous sediments are found interdigitating with the supra-ophiolitic sedimentary cover in south west Cyprus (see chapter 5, section 5.1). b) Available field data from the Mamonia Complex supports a period of Late Cretaceous margin directed crustal shortening during which the more proximal Ayios Photios sediments were thrust over the Dhiarizos Group basement (see chapter 5, section 5.4.4). c) As the Hatay and Baer Bassit ophiolites were formed in a Late Cretaceous supra-subduction zone setting, and were then emplaced southward over the northward facing Arabian continental margin, a single subduction zone model would require 'roll-back' across the whole width of the Troodos ocean basin. South east Turkey also experienced important Late Cretaceous and Early Tertiary arc-related volcanism, implying that during this period a well developed subduction zone existed along the northern margin of the Troodos ocean basin.

In the comprehensive model presented below, an intra-oceanic
'above-subduction-zone' genesis for the Troodos is adopted, followed by the activation of a more northerly subduction system, which lead to the preserved ophiolite being juxtaposed against the southern margin of the Kyrenia platform.

6.7 THE DESTRUCTIVE MARGIN HISTORY

In Mid-Cretaceous time subduction was initiated at, or near the spreading axis, and following progressive 'roll-back' the Troodos ophiolite was generated in a supra-subduction zone spreading setting. An original passive margin offset to the east of the Alanya Massif was propagated southward in the Arakapas oceanic transform fault zone (see fig.5b and fig.10a,b).

Soon after the Troodos oceanic crust was created, the Hatay and Baer Bassit ophiolites were emplaced onto the Arabian continental margin to the east. Crustal compression was then taken up along the continent-ocean boundary to the north of the Troodos oceanic crust and underthrusting of crust was initiated there (see fig.10c). During this phase of crustal shortening, the former Kyrenia Range passive margin was removed by underthrusting beneath the former gently subsiding carbonate platform. At the same time the basement to the platform was pervasively brecciated and metamorphosed. Further to the west, dextral strike-slip began to translate Troodos oceanic crust toward the Mesozoic margin offset and localised transpression along the northerly extension of the Arakapas transform fault lead to the overlapping of hot Troodos crust with older and cooler MORB crust created adjacent to the Dhiarizos Group extrusives earlier in the Mesozoic. The Ayia Varvara amphibolite and greenschist metamorphics were created along the extension of the transform lineament while in the same interval localised transtension allowed minor calc-alkaline eruptions from the dying Troodos spreading axis to shed pumaceous Kannaviou Formation volcanioclastic sediments onto the adjacent Troodos oceanic crust. Further to the north, older marginal crust (Dhiarizos Group) obliquely underthrust the former passive margin (Ayios Photios Group), leading to uplift and down-margin sliding of sedimentary thrust sheets. Detached blocks from the adjacent former passive margin were able to slide directly onto Troodos-type crust.

6.7.1 'TURNTABLE TECTONICS' - THE ROTATION HISTORY

As Africa continued to migrate north eastwards with respect to the Tauride microcontinental blocks, the principal zone of underthrusting then returned to the south margin (see fig.5c). The Baer Bassit ophiolite, together with the more intact Hatay ophiolite were then thrust southward over the Arabian margin as trench-margin collision occurred. As ophiolite slices were being obducted onto the Arabian margin to
Fig.10. Sketch maps and cross-sections to show the tectonic evolution of SW Cyprus (Mamonia Complex), S Cyprus (Moni melange) and N Cyprus (Kyrenia Range) in relation to palaeorotation of the Troodos microplate. For regional model see Fig.5).

The older Mesozoic rocks of the Mamonia Complex and the Moni Melange originate as a southward offset in a rift basin margin to the N. The Kyrenia platform is located further to the E. With Late Triassic spreading a narrow Red Sea-type ocean basin zone opens. b, ii). Following only minor sea-floor spreading in the Jurassic, spreading restarts in Early-Mid Cretaceous time, followed by initiation of a N-dipping intra-oceanic subduction zone near the spreading axis. The Arakapas oceanic fracture zone forms an extension of older marginal strike-slip passive margin. With progressive 'roll-back' of the subducting slab, the Troodos ophiolite forms above a subduction zone in a spreading setting. c, iii) Northward consumption of marginal oceanic crust in a second subduction zone to the N begins; the Kyrenia margin is removed by underthrusting with the platform being brecciated and metamorphosed. The amphibolites and Kannaviou volcaniclastics form along respectively transpressional and transtensional transform strands. The former Mamonia passive margin is uplifted, with gravity-sliding of olistoliths locally onto Troodos ophiolitic crust (Moni melange); d, iv). Troodos microplate rotation begins and Troodos and Mamonia crust elements are interleaved along major strike-slip faults. Debris-flows are shed from transpressional fault lineaments (Kathikas melange) and serpentinite gravity-flows are extruded from transtensional fissures. Bimodal within-plate volcanics and carbonate fault-scarp carbonate breccias were generated in a transtensional setting on the deformed subsided Kyrenia platform to the N. Rotation continued into Early Eocene time.
<table>
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<tr>
<th>LITHOLOGIES</th>
<th>SYMBOLS</th>
<th>LETTERS</th>
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<tbody>
<tr>
<td>'Troodos' oceanic crust (Turonian)</td>
<td>Transform fault</td>
<td>T Troodos ophiolite</td>
</tr>
<tr>
<td>Serpentinite</td>
<td>Spreading axis</td>
<td>S Serpentinite</td>
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<tr>
<td>Marginal oceanic crust (U. Tr./Ju.-Lr. Cret.)</td>
<td>Thrust fault</td>
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<td>Debris-flows (Maast)</td>
<td>Subduction zone</td>
<td>M Metamorphics</td>
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<td>Volcaniclastic sediments (U. Camp-Lr. Maast)</td>
<td>Uplift</td>
<td>Km Kathikas melange</td>
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<td>Passive margin sediments (U. Tr.-Lr. Cret.)</td>
<td>Subsidence</td>
<td>Kv Kannaviou volcanics</td>
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<td>Neritic limestone blocks (U. Tr.)</td>
<td>Gravity sliding</td>
<td>K Kyrenia Range</td>
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<td>Shelf Carbonate (Tr.-Cr.)</td>
<td>Sediment dispersal</td>
<td>D Dhiarizos Gp.</td>
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<td>Continental crust (not exposed)</td>
<td>Strike-slip</td>
<td>AP Ayios Photios Gp.</td>
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<td></td>
<td>Mega-rotation</td>
<td>AF Arakapas fault belt</td>
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<tr>
<td></td>
<td>Detached blocks</td>
<td>A African plate</td>
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U. Trias.

U. Cret. (Turonian)

U. Jurassic - L. Cretaceous

Passive margin
Seamounts and collapsed atolls
Carbonate platform Marginal crust 'MORB' crust Marginal crust

Above-subduction zone Troodos spreading

U. Cretaceous (Turonian)

L. Campanian

Uplift and sliding
Amphibolites along transpressional transform to S
Oblique underthrusting

U. Campanian-L. Maastrichtian

Sliding off fault scarps Calc-alkaline eruptions

U. Maastrichtian - L. Eocene

Serpentinite debris flow
Kathikas debris-flow-transpressional
Arakapas transform extension

Serpentinite protrusions into strike-slip lineaments
the east, ocean crust continued to be subducted to the south of Cyprus, thus developing an anticlockwise torque on Troodos crust stranded in a 'fore-arc' or 'pre-arc' setting. The Troodos microplate then became detached and began to rotate around an inter-connected 'ring' of major dextral fractures (see fig.10d). One of these fractures is inferred to run to the west of the Troodos ophiolitic basement exposed on the Akamas Peninsula and to the south of the Limassol Forest block.

Former oceanic fracture zone crust in south west Cyprus became part of the rotating microplate, while pervasive dextral strike slip deformation continued to dominate the segment of the lineament preserved in south west Cyprus. Previously deformed former passive margin units became attached to the rotating microplate and were carried southwards to be interleaved with Troodos crustal wedges (e.g. the Akamas Peninsula block and the elongate YMB preserved in the Arkhimandrita-Marathounda area). In south west Cyprus the earlier stages of rotation were dominated by transpressional tectonics as both YMB and OMB were strongly sheared and serpentinites were intruded into braided strike-slip zones. Locally, uplift and overturning (for instance the 'flower structure' deformation of Troodos basement on either side of the Polis graben, see chapter 5 fig.2a) resulted in the shedding of multiple debris flows into nearby topographic hollows (the Kathikas and 'small olistolith' melange of south west and southern Cyprus respectively). By contrast, in northern Cyprus, Troodos ophiolitic crust, perhaps including the former spreading axis was rotated into place adjacent to the deformed Kyrenia platform in a generally transtensional setting. Within-plate bimodal volcanics were extruded from fissures along the subsided platform margin while carbonate breccias were shed from active fault scarps into adjacent basins floored by pelagic sediments (see fig.10e).

Following a period of more rapid plate rotation in the late Campanian and early Maastrichtian the Troodos ophiolite palaeorotation continued more slowly until the Early Eocene, when the northward convergent motion of Africa relative to Eurasia became less rapid (Livermore and Smith 1984a,b). Pelagic carbonates gradually buried the former spreading axis crust while fault-activity continued near the microplate boundaries, especially in southern Cyprus where calciturbidites filled a deep trough (Robertson 1976). Continued seafloor fault activity along the inferred southern microplate boundary is implied by the abundant evidence for intra-formational slumping in the Upper Palaeocene and Lower Eocene Middle Lefkara to the south of the main Massif (Eaton, personal communication 1985). Intraplate volcanics continued to be erupted onto the Kyrenia platform into Early Eocene time.

In the Late Eocene the Africa-Eurasia convergence rate increased (Livermore and Smith 1984a,b) and the Kyrenia margin was deformed again by southward-directed thrusting (D2 of Robertson and Woodcock 1985). The Troodos ophiolite complex remained largely unaffected by this event. Throughout the Tertiary oceanic crust
continued to be consumed beneath the ophiolite complex that remained stranded adjacent to the Kyrenia margin in a 'fore-arc' setting. It was probably only in Middle Miocene time that continued crustal consumption to the south of Cyprus brought a wedge of thickened crust against the ophiolite complex. Phases of dramatic uplift in the Middle Miocene and in the Plio-Pleistocene may well record periods of accelerated underthrusting of this block beneath the preserved ophiolite complex. During the Neogene, the Kyrenia Range could have been translated westward relative to the ophiolite along the Kythrea Fault as the Anatolian Block was displaced west away from the 'jaws' of the Arabian and Eurasian collision further to the east.
CHAPTER SEVEN

A PRELIMINARY PALAEOMAGNETIC STUDY IN THE ISPARTA ANGLE, SW TURKEY

7.1 INTRODUCTION

In Chapter One I described how contemporary active deformation in the Greece-Turkey-Syria-Iran segment of the Alpine-Himalayan orogenic belt is primarily controlled by the mutual interaction of the semi-rigid 'Anatolian Block' with the surrounding Aegean, African, Eurasian and Arabian plates. In this plate tectonic framework, Turkey is itself cut by a series of major plate boundaries, some of which are active today, whilst others preserved within ophiolitic suture zones were active in the past. Although there is widespread disagreement as to the detailed relative configuration of individual microplates in any one time window through geological time, there is a popular view among geologists working in the area that Turkey, and indeed much of the central segment of the Alpine-Himalayan orogenic belt, comprises a complex mosaic of basement elements that have been brought together at different times between the converging African and Eurasian plates. By imposing constraints on the relative positions of Africa and Eurasia throughout the closure of the Palaeotethys ocean, Livermore and Smith (1984a,b, see chapter 6 fig.2) have effectively provided a framework within which geologists can analyse the geodynamic evolution of the Turkish mosaic, as if parts of Turkey have in the past 250Ma belonged to either stable Africa or Eurasia, then their motions and polar wander paths must necessarily correspond to those calculated for the adjacent major converging plates. Thus, it is only palaeomagnetism that provides geologists with a technique with which they can quantitatively apply physical constraints to the relative spatial distribution of elemental component basement units during the long closure history of the complex branching ocean basin network that characterises the Neotethyan orogenic belt.

In this chapter, I first review the current contradictory plate tectonic hypotheses that have been proposed for the geodynamic evolution of the Turkish area with the intention of clarifying the scope for palaeomagnetic studies in a region that has experienced a complex history of intra-continental rifting, sea-floor spreading, ocean closure and intra-continental crustal compression. I then complement this discussion by presenting preliminary palaeomagnetic data from the critical 'Isparta Angle' segment (see chapter 1 fig.2) of the Taurus mountains 'Axe Calcaire' and show how autochthonous and para-autochthonous basement elements in this part of southern Turkey have unequivocally experienced a different palaeorotation history to that of the adjacent Troodos ophiolite complex. I critically compare these data with existing data,
and consider whether palaeomagnetism can be feasibly used as a tool for testing the relative merits of contradictory tectonic and sedimentary models that have been developed in recent years by geologists working in the area.

7.2 THE ORIGIN OF NEOTETHYAN OPHIOLITES AND THEIR RELATIONSHIP TO THE AFRO-ARABIAN PLATE:
THE SINGLE VS. MULTIPLE SUTURE ZONE MODEL

One of the diagnostic features of Turkish geology is unquestionably the abundance of Late Mesozoic ophiolitic rocks which occur in variable states of preservation along a series of anastomosing suture zones, often in close spatial association with rift-related and passive continental margin sediments (see fig.1a). As ophiolites are the only preserved remnants of ocean crust that originally floored the Neotethyan ocean basins, their geodynamic evolution from initial intra-continental rifting to eventual closure can only be effectively pieced together by investigating the successions deposited at their margins. It was the general similarity of many marginal sedimentary successions preserved within separate ophiolite belts that lead Ricou (1971) to propose that all Turkish ophiolites, including the Troodos Massif, originated from a unique Tethyan ocean basin that originally lay to the south of the Pontide and Sakarya blocks in northern Anatolia (see fig.1a, b and fig.2). This model strongly contrasts with other reconstructions which invoke the existence of at least a single independent Neotethyan ocean basin (the 'Troodos ocean basin' described in chapter 6; Dumont et al. 1977, Biju-Duval et al. 1977, Robertson and Woodcock 1980) and those which infer a complex palaeogeography characterised by a braided pattern of interconnecting ocean strands separating adjacent microcontinental slivers (Sengor and Yilmaz 1981, Robertson and Dixon 1984; see fig.1c).

7.2.1 DUPLICATION OF OPHIOLITE BELTS BY LARGE SCALE THRUSTING

Although the single basin hypothesis is attractive in that it attempts to incorporate many of the complexities of Turkish geology into a unified framework, it appears to have a number of fundamental flaws. In this model, developed over the past decade to a high degree of sophistication by Ricou and others (1979, 1984), Triassic rifting occurred along the northern margin of Gondwanaland to the north of the Tauride-Anatolide platform unit (see fig.1b). This resulted in the formation of a single Mesozoic (Tethyan) ocean basin that may, or may not, have formed part of the original much wider Palaeotethys. Following the initiation of northward subduction beneath the Pontide arc in the Mid Cretaceous, platforms on the northern margin of Gondwanaland (for example the Tauride 'Axe Calcaire') rapidly collapsed to pelagic
Fig.1 The origin of Neotethyan ophiolites in the Turkish area (after Robertson and Dixon 1984). (A) shows location of principal ophiolite belts in relation to major 'autochthons'. Postulated suture zones for 'single' and 'multiple' root zone models are also indicated as is approximate line of section represented by (B) and (C). (B) shows the main features of the single (Izmir-Ankara-Erzican) root zone model of Ricou et al. (1979, 1984). In this model the northern margin of the Afro-Arabian plate extends as far north as the Sakarya and Pontide units of northern Turkey. A major ophiolite nappe emplaced initially onto the northern margin of the Tauride-Anatolide platform in the Upper Cretaceous is differentially disrupted and transported southward in episodic Tertiary thrusting events, to create three independent ophiolite belts stranded on the northern margin of the African Plate. (C) shows multiple (see A) root zone model favoured by Sengor and Yilmaz (1981) and Robertson and Dixon (1984). Schematic diagram shows the Sakarya, Kirsehir, Tauride-Anatolide and Cimmerian blocks being rifted-off the northern margin of Gondwanaland as a result of southward subduction of the Palaeotethys (Sengor and Yilmaz 1981). Each rift acted as an ophiolite source once an individual basin experienced compression. Following incipient collision, a complex braided network of four suture zones were preserved in Turkey (Intrapontide, Izmir-Ankara-Erzican, Inner Tauride and Peri-Arabic sutures). Multiple suture zone model of Robertson and Dixon (1984) again adopts a 'separate block' configuration, but maintains a northward-subducting Palaeotethys between the Sakarya-Kirsehir blocks and the Pontides until the Mid-Tertiary.

Fig.2 Important tectonic units of Turkey (after Robertson and Dixon 1984). Palaeomagnetic sampling sites of Evans et al. (1982, Bilecik), Van der Voo (1968, Bayburt) and Lauer (1981, Balya and Gorede) in Pontide 'block' are shown, as is area covered by fig.5.
FIG. 1 ALTERNATIVE MODELS FOR ORIGIN OF OPHIOLITES
FIG. 2 TECTONIC ELEMENTS OF TURKEY
depths. After a period of deep water pelagic sedimentation ophiolitic olistostromal deposition heralded the arrival of major ophiolite nappes that moved southward in the Upper Cretaceous by 'gravity sliding' from the principal Tethyan suture zone in northern Turkey. After successive stages of southward tectonic transport over the leading edge of the 'Arabian' platform, the nappes reached their final position by Late Eocene to Late Miocene. Subsequently, their original connection to the 'root zone', located some hundreds of kilometres to the north, has been eroded leaving them isolated as grossly allochthonous units stranded above the Tauride-Anatolide platform autochthon. In this model the whole of the Eastern Mediterranean would always have been underlain by continental crust, essentially the northern extension of the African and Arabian plates. An alternative, and generally preferred model, shows almost every ophiolite body 'rooted' in an almost in situ position in a separate basin, and it is only following a long history of Alpine deformation that the terranes have been thrust into an allochthonous position relative to the underlying relatively autochthonous platform units (see fig.1c).

Although it is conceivable that similar basin margin facies could be diagnostic of a single basin origin, the available evidence certainly does not preclude a derivation from different geographical locations in a Tethyan region represented by a complex pattern of ocean basins and microcontinental blocks. Certainly, many of the rift-related and passive margin sedimentary facies typical of ophiolite-associated allochthonous terranes are ubiquitous throughout the Mesozoic Tethyan region and were thus not necessarily formed within a single connected ocean basin. In apparent support of a single basin model, it is clear that ophiolitic nappes were emplaced southward in the latest Cretaceous over a wide area along the northern edge of the Tauride carbonate platforms (for example the Bozkir nappes of Sengor and Yilmaz 1981, see fig.2). Extensive nappe transport however, is not compatible with a growing body of evidence from the Bey Daglari platform unit in the Antalya Complex (see fig.2 and section 7.4.2), that supports continuous Early Cretaceous to Recent sedimentation, thus precluding any simple overthrusting of ophiolitic nappes from the north west and north east. Sengor and Yilmaz (1981; see also Monod 1977) also list an additional five localities where successions on the Tauride carbonate platforms extend intact into Palaeocene-Early Eocene time. Having accepted that there are segments of the Tauride-Anatolide platform that have not been overthrust by a nappe pile until at least Late Eocene time, Ricou et al. (1979) now consider that the Troodos and Antalya ophiolites together with associated carbonate platform (e.g. the Eastern Bey Daglari, see fig.3) and off-margin pelagic sediments were originally located to the north west of the Menderes metamorphic Massif and have been transported into their present position by a major dextral shear, followed by anticlockwise rotation through the Isparta angle. In their preferred model, nappe piles (for example the Mamonia
nappes of Lapierre 1975, see chapter 1 fig.4b; and the Antalya nappes of Brunn et al. 1974, see section 7.4.2) were assembled in the Late Cretaceous on the northern margin of the Tauride-Anatolide platform and were only translated to their more southerly position in the Tertiary by a combination of strike-slip faulting and major dextral shear. Similarly, the Hatay, Baer Bassit and Pozanti-Karsanti ophiolite allochthons are considered to have been initially assembled to the north of the 'Tauride axis' in the Late Cretaceous and then thrust southward onto the leading edge of the Arabian platform in an upper Eocene phase of regional compression. Other units remained largely unaffected by late Cretaceous tectonics and experienced mainly Late Eocene transport to the south (for example, the Beysehir-Hoyran nappes, see fig.2).

If the Antalya Complex allochthon originated on the northern side of the Tauride calcareous axis, then one would expect off-margin sediments to have been imbricated against a generally northward facing margin during the Late Cretaceous deformational phase. This however, is inconsistent with the palinspastic reconstructions of Waldron (1984a, b see section 7.4.2) which clearly show that in the north east segment of the Antalya Complex (east of Lake Egridir) off-margin sediments were thrust north east against the Anamas Dag relative autochthon in the Late Cretaceous, prior to a later south east directed phase of compression in the Late Eocene.

If indeed the circum-Mediterranean ophiolitic rocks, together with the associated carbonate margin and pelagic units, are considered to represent the remnants of far-travelled nappes, it is difficult to believe how allochthonous units such as the Eastern Bey Daglari platform of Ricou (1979, see fig.3) could have remained largely undeformed, accumulating continuous sedimentary successions throughout much of the Tertiary with little or no evidence of internal deformation. Similarly, it seems particularly unlikely that the Troodos ophiolite complex can have been transported more than 400km in at least two independent phases without disturbing the passive accumulation of pelagic sediments (Robertson and Woodcock 1980). This history contrasts strongly with the emplacement of the Semail ophiolite which is known to have been obducted over the Arabian foreland in the Late Cretaceous. In this case, pelagic sedimentation on the ocean floor crust ended abruptly before the ophiolite was incorporated within an imbricated thrust stack and finally emplaced onto the leading edge of the Arabian continental margin.

7.2.2 DUPLICATION OF OPHIOLITE BELTS BY STRIKE-SLIP DEFORMATION

Although the available outcrop evidence suggests that the ophiolite Massifs were not thrust substantial distances southward over the the Tauride-Anatolide platforms, could a single basin origin for the circum-Mediterranean ophiolites still be feasible if the juxtaposition of sub-parallel ophiolite belts was achieved by major strike-slip motion
Fig.3 Alternative models for the origin and the emplacement of the Antalya Complex marginal units in the Isparta Angle. a) Derivation from a unique northern Mesozoic ocean basin as postulated by Brunn et al. (1974) and Ricou et al. (1975). b) Ricou et al. (1979) modified 'internalist' model, showing Antalya Complex allochthon being emplaced over the northward-facing margin of the Menderes Massif (which includes the Eastern Bey Daglari relative autochthon) to the north in the Upper Cretaceous. After being translated southward by dextral shear into the Isparta Angle in the Eocene, the relative autochthon together with the overlying Antalya Complex marginal units were emplaced westward (by the Salir thrust) onto the Western Bey Daglari platform in the Miocene. In the 'externalist' model (c), the Antalya allochthon is thrust from a southerly ocean basin onto the main Bey Daglari platform in Upper Cretaceous time (Dumont et al. 1972). Woodcock and Robertson (1982) argue that a strike-slip component along the eastern margin of the Bey Daglari platform strongly influenced the mode of emplacement.
Fig. 4 Lauer's hypothesis, showing division of Turkey into three separate blocks (A-Pontide Block, B-Western Tauride Block and C-Eastern Tauride Block). Northward drift of each block relative to the northern African margin is constrained on the basis of palaeomagnetic inclination data (b) compiled from both autochthonous and allochthonous elements from each block.

Fig. 5 Tectonic map of the Isparta Angle (after Brunn et al. 1971, with modifications by Woodcock and Robertson 1982, Gutnic et al. 1979 and Waldron 1981). Location of palaeomagnetic sampling sites are indicated. Inset shows schematic Upper Cretaceous (Campanian-Senonian) palinspastic map for the main platform units in the Lake Egridir region, assuming no rotation or relative horizontal displacement prior to emplacement (i.e. units have been unstacked without rotation and assuming minimal translation for consistency with outcrop). The actual size of banks and/or basins may have been much larger.
FIG. 3 ALTERNATIVE ORIGINS FOR THE ANTALYA COMPLEX

orament on allochthonous units only
- derived internally from NW
- " " " NE
- " externally from SE or SW
- postulated allochthonous platform
- " derivation uncertain

FIG. 4 LAUER'S HYPOTHESIS
FIG. 5 THE ISPARTA ANGLE

- Post-nappe sediments & volcanics
- TERTIARY ALLOCOTHONS
  - Lycian nappe
  - Beysehir-Hoyran-Hadim nappe
- ANATOLA COMPLEX
  - Ophiolites, serpentinite
  - Units with basinal Mesozoic sequences
  - Units with shallow-water Mesozoic
  - Anamas Dağ margin (sheet A)
  - Structurally high (NE area sheets 1, 3)
  - Structurally low (NE area sheet 6)
  - Undifferentiated (SW area)
- ALANYA MASSIF
- TAUER DE PLATFORM UNITS
  - Bey Dağları
  - Anamas Dağ - Akseki
  - Tertiary flysch & molasse
  - Aksu thrust (U Miocone)
  - Earlier Tertiary thrusts

Oceanic zone of unknown width

Scale for minimum shortening reconstruction
along the northern margin of the Afro-Arabian plate? If major strike-slip displacements did occur, they must necessarily have taken place during a short interval in the Upper Cretaceous, after the formation of the youngest ophiolites (probably during the Turonian to Santonian interval) but prior to their emplacement on to the adjacent autochthonous carbonate margins. A later juxtaposition would require displacement to have occurred along major intra-continental lineaments like the Northern and Eastern Anatolian transform faults (Robertson and Dixon 1984). With the exception of these contemporary intra-continental faults no major Mesozoic and Early Cenozoic strike-slip lineaments of this kind have been identified. This observation is consistent with the convergence path of Africa relative to Eurasia in the Turkish area from the Mid Cretaceous, which would support only a limited history of oblique crustal consumption beneath either margin prior to the onset of more orthogonal NS convergence in the Early Tertiary (Livermore and Smith 1984a,b). In this framework no conceivable driving mechanism for major 'Pacific margin' strike-slip displacements exists (Coney 1980).

Duplication of ophiolitic suture zones by major lateral displacements would also imply that the southern margin of the Tauride carbonate platforms was located close to a major intra-continental transform fault at the time of lateral juxtaposition in the Late Cretaceous, yet it is clear that in many cases, deposition of continental margin sediments continued undisturbed throughout this time interval (for example, on the Alanya Massif; Okay and Ozgul 1984). With the exception of the arcuate Arakapas transform fault lineament in Cyprus (see chapter 6, fig.10), other intra-oceanic transform faults in the circum-Mediterranean ophiolite belt (e.g. in the Tekirova ophiolite in the Antalya Complex, Reuber 1984) were regionally orientated NS allowing little opportunity for significant east-west displacements (Robertson and Dixon 1984).

On a local scale however there is strong evidence that strike-slip faulting did indeed play an important role along some oblique convergent margins during the long history of crustal consumption in the Neotethyan belt. Aktas and Robertson (1984) for example, recognised the importance of strike-slip generated pull-apart basins in the formation of the Maden complex in south east Anatolia, while Norman (1984) interprets the Ankara Melange as having formed during a long history of oblique northward subduction to the south of the ‘Sakarya microcontinental block’ in Jurassic and Cretaceous time. Similarly, both the formation (Reuber 1984) and the subsequent tectonic emplacement (Robertson and Woodcock 1984) of the Antalya Complex are now interpreted in terms of transform faulting.

Despite this evidence of localised strike-slip deformation, it is apparent that the present distribution of sub-parallel ophiolite belts through Turkey cannot be easily explained by large scale lateral shearing along the leading margin of the Afro-Arabian plate. However, before any hypothesis that supports a derivation of all the Turkish
basement elements from a proto-Indian ocean by major north west directed shear can be safely disregarded, the available palaeomagnetic data of Lauer (1981, 1984; see below) has to be scrutinised, to determine whether Turkey did indeed originate from low latitudes to the south of the inferred northern margin of the Afro-Arabian plate.

7.2.3 LAUER'S HYPOTHESIS: DERIVATION FROM A PROTO-INDIAN OCEAN

According to Lauer (1981, 1984), the available palaeomagnetic data allow Turkey to be sub-divided into three blocks (A-PONTIDE BLOCK, B-WESTERN TAURIDE BLOCK and C-EASTERN TAURIDE BLOCK, see fig.4a), each of which has moved independently until continental collision and eventual accretion against the leading edge of the African plate in the Neogene. Although he suggests that the spatial extent of each block has been defined solely on palaeomagnetic criteria, the specific boundaries to each unit were in fact chosen so that they corresponded to important suture zones (i.e. the southern boundary of BLOCK A runs east-west along the northern Anatolian suture, while the boundary between BLOCK B and BLOCK C coincides with the axis of the Isparta angle lineament). Discrepancies in remanence declinations between rock units of the same age in any one of these three 'rigid' blocks were attributed to localised internal 'block rotations' or large scale plastic deformation.

Although no palaeomagnetic data exists for autochthonous units in BLOCK B, the low inclinations recorded in Triassic and Jurassic successions in both BLOCKS A and C requires these blocks to have been originally located at equatorial latitudes to the east of the Arabian plate, whose northern margin can be independently constrained to have lain between 10° and 20° to the north of the equator. Not unexpectedly, inclination data for allochthonous units in BLOCK A and BLOCK C do not differ considerably from contemporaneous autochthonous successions in the same block, implying that given the poor resolution in accurately assessing palaeolatitudes from palaeomagnetic data, there has not been a significant relative displacement of nappe sheets relative to underlying basement units. Similarly, the low inclinations recorded in allochthonous units above BLOCK B are assumed to imply that, like the other blocks, the western Taurides also has an equatorial origin.

If the sparse pre-Cretaceous palaeo-inclination data are considered reliable (see fig.4b), Lauer's hypothesis would require each of the three separate blocks to have been rifted-off adjacent areas of the eastern Arabian margin in the Triassic, possibly with BLOCK A originating from the lowest latitudes. As ammonite faunas of the Tauride carbonate platforms have a distinctive 'south Tethys' provenance (Hirsch 1984), the three Turkish blocks migrated northward but probably never drifted far from the Arabian margin. Each of the three independent units were then juxtaposed close to
their present position, prior to the emplacement of ophiolites in Late Cretaceous time.

In evaluating the merits of Lauer's hypothesis it is first necessary to appreciate how accurately one can realistically constrain the palaeolatitudinal positions of each of the three independent blocks relative to the Arabian plate. Importantly, the inferred equatorial palaeolatitudes for BLOCK B and C were primarily based on remanence directions isolated from pillow lava formations where appropriate structural corrections are notoriously difficult to apply (see chapter 2 section 2.3). For example, even in an allochthonous 250m thick coherent succession on BLOCK B (Calbali Dag succession), remanence inclinations measured for over 50 samples were found to vary between -32.6° and +15.1°, leading to wide error margins in the weighted mean. Similarly, the equatorial origin for BLOCK A relies heavily on the Permian inclination data of Gregor and Zijderveld (1964) and the Jurassic data of Van der Voo (1968), supplemented by new Upper Triassic (one locality at Bayla, see fig.2) and Upper Jurassic (one locality at Gorede) data of Lauer (1981). Without a complete polar wander path the low Permian inclination data are open to alternative interpretations, while the Lower Jurassic result from Bayburt was not considered reliable by Van der Voo as the stable remanence was carried by haematite of possible secondary origin. Only a thorough analysis of 191 samples from Upper Jurassic limestones in the western Pontides (Bilecik limestones, Evans et al. 1983) has provided a reliable remanence direction, implying that the area has drifted northward by 21° and has been rotated through more than 90° clockwise since the sediments were deposited. Although at least a proportion of the remanence was carried by magnetite, curiously no reversed polarity sites were identified. Thus, although the fossil evidence points to a Eurasian affinity for the Pontides, further palaeomagnetic data is clearly needed to assess whether the Palaeozoic and Mesozoic polar wander paths for BLOCK A correspond to complementary ones of either Africa or Eurasia.

The problem of assessing the position of the three Turkish blocks during the Mesozoic is further compounded by the poor definition of the African apparent polar wander (APW) path during this interval. The best available palaeomagnetic data shows that during the latest Carboniferous and Early Permian there was little APW, while during the Lower Triassic the palaeopole migrated rapidly southward (i.e. corresponding to a northward drift of Africa) to a new stable position before moving northward again in the Jurassic (i.e. southward drift of Africa). After a period of southward drift, Africa started to move steadily northward from Mid Cretaceous time (see chapter 6 fig.2). The sparsity of reliable data do not allow us to map out the exact position of the northern margin of the Afro-Arabian plate during the critical Late Triassic-Early Jurassic period, and it is still conceivable that the Turkish blocks originated from an 'equatorial' location close to the northern margin of the African Plate. According to Smith et al. (1981) the northern margin of the African Plate may
have been located at a latitude of little more than 10°N in this interval.

If it is assumed that the Turkish basement elements were derived from the eastern Arabian margin, then the model implies that the independent blocks should be bordered on at least one edge by passive continental margin successions with a long Palaeozoic history, and that during the Mesozoic they were translated by several thousand kilometres along major transforms bounding the leading margin of Arabia (Robertson and Dixon 1984). Unfortunately there is little definitive evidence for either of these prerequisites, and it seems likely that until a more reliable data base is available, the apparent 10-20° difference in inclinations between the southern Tauride blocks and stable Africa does not preclude these areas from having originally been located against North Africa rather than south east Arabia.

### 7.3 OBJECTIVES OF FUTURE PALAEOMAGNETIC STUDIES IN TURKEY

Clearly, there is now an urgent need for a more comprehensive and integrated palaeomagnetic study of the elemental basement component units of Turkey. As existing data for the Turkish area are often not surprisingly contradictory, having been confined to relatively deformed igneous successions exposed in the vicinity of ophiolite suture zones, future palaeomagnetic studies will concentrate sampling primarily within largely undeformed successions that overlie basement autochthons. Specific aims include:

(a) Checking whether all the autochthonous Massifs of southern Turkey (e.g. the Menderes, Tauride and Kirsehir blocks) represent tectonic windows into the main passive Afro-Arabian platform or a number of microcontinental units that were originally separated by wide ocean basins. Necessarily this will test whether ophiolitic 'nappes' in southern Turkey, northern Syria and Cyprus are grossly allochthonous or relatively in situ with respect to adjacent autochthons.

(b) Providing a more comprehensive set of inclination data from both allochthonous and autochthonous units to determine whether or not the various tectonic elements of Turkey originated at low latitudes from a Proto-Indian ocean (as proposed by Lauer, see section 7.2.3). Reliable Mesozoic palaeo-inclination data from the Pontide and Sakarya blocks should reveal if these units were derived from Eurasia, as preferred by Robertson and Dixon (1984) or Gondwanaland as preferred by Sengor (1979).

(c) Checking whether the Tauride 'Calcareous Axis' is split by a major suture zone now preserved within the important NS trending 'Isparta Angle' lineament (this corresponds to the boundary between BLOCKS B and C of Lauer 1981). This will
involve recognising the rotational behaviour within the horizontal plane of relatively undeformed Mesozoic carbonate platform blocks both during initial rifting (in the Permo-Triassic) and subsequent tectonic emplacement (Late Cretaceous and Tertiary).

(d) Determining whether the 90° anticlockwise palaeorotation of the 'Troodos microplate' in Cyprus has influenced plate rotation in southern Turkey.

7.4 PALAEOMAGNETIC STUDIES IN THE ISPARTA ANGLE, SW TURKEY

7.4.1 INTRODUCTION

In the final section of this thesis I discuss in detail the tectonic evolution of the 'Isparta Angle' lineament in south west Turkey, to determine whether the available geological field evidence supports the existence of a substantial Mesozoic oceanic tract between the western and eastern Tauride blocks, as proposed by Lauer (see above). To complement this discussion, the published palaeomagnetic data for the area are critically reviewed and compared with our own preliminary results from autochthonous Late Mesozoic-Early Cenozoic platform successions on either side of this prominent lineament.

7.4.2 THE STRUCTURAL HISTORY OF THE ISPARTA ANGLE

The geology of the Tauride mountain belt in the vicinity of Antalya Bay is dominated by the occurrence of major Mesozoic platform limestone Massifs that outcrop as a series of tectonic inliers beneath a variety of allochthonous units, characterised by basinal Mesozoic sediments (turbidites, shales and cherts), together with ophiolite fragments and some clearly allochthonous platform limestone sheets (see fig.5). The two largest platform units, the Bey Daglari and the Anamas-Akseki Massifs, together form the 'backbone' of the Taurus mountains on either side of the prominent north pointing cusp of the Isparta Angle, while the Davras Dag, Barla Dag, Dulup Dag and Karacahisar limestone Massifs occur as separate smaller platform units in the Lake Egridir region at the heart of the Isparta Angle.

Both major platform units were deformed by major Tertiary thrust events recognised throughout the Hellenides and Taurides. While movement on the Tauride thrusts to the east of the 'Isparta Angle' ended in the Eocene (associated with SW-directed emplacement of the Hadim nappes, Monod 1977) to the west thrusting continued until Miocene times (associated with SE-directed emplacement of the Lycian nappes over the Bey Daglari; Poisson 1977, Hayward and Robertson 1982). Finally, the Late Miocene Aksu thrust at the centre of the Isparta Angle transported the
western edge of the Taurides an unknown but probably short distance over the Bey Daglari platform and its adjacent units (Poisson 1977, Akbulut 1977).

These Tertiary structures are superimposed over a Late Cretaceous complex of thrust sheets, tectonic slices, melanges and ophiolitic slivers that generally outcrop in the areas immediately adjacent to individual platform units. Traditionally, these complexly deformed successions have been assigned to the Antalya nappes (Lefevre 1967, Brunn et al. 1971 and Juteau 1975) or the Antalya Complex (Woodcock and Robertson 1977). This group of allochthonous units is confined to the Isparta Angle, although similar rock associations are also found in the Mamonia Complex of south west Cyprus (see chapter 5). To the west of Isparta Angle along the eastern flank of the Bey Daglari, the Antalya Complex units have been thrust westward during the Upper Cretaceous onto the Jurassic and Cretaceous shallow water carbonates of the relative autochthon while on the north west of the Massif the Lycian nappes were emplaced south eastwards over the platform in the Tertiary.

According to Waldron (1981, 1984a,b), the structure of the north east Antalya Complex between the Davras Dag and Anamas Dag platforms on the eastern limb of the Isparta Angle is consistent with a two phase history of thrusting. During an earlier Maastrichtian-Early Tertiary deformational phase \( t_1 \) the basinal off-marginal sedimentary successions of the Antalya Complex were imbricated against the westward facing platform margin of the Anamas Dag, while in a later Eocene phase \( t_2 \), related to nappe emplacement from the north east, both the Antalya Complex and the adjacent platform were thrust westwards. The ‘Aksu’ phase of Late Miocene thrusting oblique to \( t_1 \) and \( t_2 \) has further complicated the structural pattern of the area.

On the eastern limb of the Isparta Angle, erosion of the main Anamas Dag-Akseki autochthonous Massifs has reached a deeper level than in the Bey Daglari, revealing Triassic sandstones and shales resting unconformably on a Palaeozoic basement (Monod 1977). In reality, the autochthon here consists of a series of major thrust slices emplaced toward the south west. Resting upon these slices are the Beyschir-Hoyran-Hadim nappes (Brunn et al. 1971, Monod 1977) which, like the Lycian nappes contain an ophiolitic unit together with a variety of Upper Palaeozoic to Late Cretaceous sediments. These units are well preserved within a major klippe along the axis of the Akseki platform (see fig.5).

Some authors, notably Ricou et al. (1979, see fig.3) consider that each of the distinctive platform units exposed around the Isparta Angle should not be regarded as separate palaeogeographic entities, but as fragments of a single contiguous platform unit. They propose that the Antalya Complex and certain adjacent platform units (for example the Eastern Bey Daglari), are the remnants of far-travelled nappes that were transported southward over the Tauride platform during the latest Cretaceous, and
were then translated into the Isparta Angle by dextral shear in the Eocene. Finally, the Antalya Complex together with the adjacent platform successions were emplaced over another platform unit (the Western Bey Daglari) in Miocene time. This model sharply contrasts with that of Waldron (1984 a,b; see also Woodcock and Robertson 1977) who interprets the Antalya Complex not as a series of far travelled nappes from a single Tethyan ocean basin to the north, but as fragments of a number of smaller carbonate banks and adjacent basins that were originally located between the main Bey Daglari and Anamas Dag platforms. In support of this hypothesis, Waldron (1981) cites the following field evidence:

a) The Maastrichtian-Early Tertiary tₕ phase of north eastward directed thrusting of the Antalya Complex over the Anamas Dag (see above) is incompatible with major southward nappe emplacement during the Late Cretaceous unless the platform has been subsequently rotated through a large angle in the horizontal plane. Similarly, structural studies in the south western segment of the Antalya Complex (Woodcock and Robertson 1977, 1982) suggest off-margin sediments were emplaced as an imbricate thrust stack against the margin of the Bey Daglari platform. The polarity of facies in the imbricated sediments (Robertson and Woodcock 1981, 1982) confirms an origin on an east-facing platform margin.

b) Distinctive sedimentary facies preserved in situ along the south western edge of the Anamas-Akseki platform are similar in many respects to other proximal margin units of the Antalya Complex that now structurally overlie the autochthon. These units are not found in any other platform or basin successions in the area. Also, Monod (1977) reports that marginal successions to the Anamas-Akseki platform contain detritus derived from Antalya Complex units in the Late Cretaceous to Palaeocene, while carbonate sedimentation continued in central parts of the platform unit in Eocene time.

c) According to Hayward and Robertson (1982) Cretaceous to Oligocene platform limestones of the Bey Daglari pass conformably up into Miocene clastic sediments derived from the adjacent Antalya Complex. Westward-directed palaeocurrents in these sediments indicates that the Antalya Complex was emplaced onto the eastern edge of the Bey Daglari. As these sequences are continuous through to the Upper Miocene, the Antalya Complex could not have been transported over the Bey Daglari in the Middle Miocene as necessitated by the allochthonous hypothesis of Ricou et al. (1974,1979). Poisson (1977) has clearly demonstrated that the south west segment of the Antalya Complex was not finally emplaced by easterly directed thrusting until the Late Miocene, yet to the north at the heart of the Isparta angle, the Antalya Complex is unconformably overlain by flysch sediments of Upper Oligocene and Early Miocene.
age. Thus, southward translation of ophiolite-related allochthonous units must necessarily have been completed prior to the end of the mid Oligocene, well before the final emplacement of the Antalya Complex in the Miocene further to the south.

This evidence lead Waldron (1981) to favour an origin for the Antalya Complex units in the Isparta Angle as a complex mosaic of small carbonate banks and basins that lay between the principal Bey Daglari and Anamas-Akseki platform units. A similar configuration of marginal carbonate platform units on the eastern edge of the Bey Daglari in the SW Antalya Complex was proposed by Robertson and Woodcock (1982). The general distribution of carbonate platform units was broadly analagous to the present-day Bahama Banks. By unstacking the numerous thrust sheet complexes that have been emplaced above the relatively autochthonous platform units, Waldron (1981) was able to reconstruct the original palaeogeographic disposition of off-margin platform units relative to the adjacent major carbonate platforms (see fig.5 inset). It must be emphasised however, that the actual dimensions of the platforms and basins were probably much greater than in his reconstructions, as balanced cross sections only allow an estimate of the minimum shortening between adjacent platform units. Also, the structural data only place a lower limit on the distance that thrust sheets have been transported and do not indicate the nature of the original basement on which the platform and basinal successions were deposited. Importantly, in his reconstructions all platform units were arranged in their present-day orientations as no constraints could be placed on the rotations that these units might have undergone during crustal shortening.

7.4.3 PALAEOMAGNETIC STUDY

7.4.3.1 CHOICE OF SAMPLING SITE AND SAMPLING STRATEGY

In Chapter 3 I described how a systematic swing in the declination of the mean primary remanence vector retained within a sedimentary succession lying in situ over lithospheric basement provides a particularly versatile method of determining the sense, the magnitude and the timing of a rotational displacement about a vertical axis in the horizontal plane. If it can be assumed that the extensive Mesozoic shallow water limestone sequences that dominate the geology of the Isparta Angle lie either autochthonously or para-autochthonously on lithospheric basement, it should in principle be possible to palaeomagnetically determine whether each ‘independent’ platform unit has experienced an identical or contrasting rotation history relative to the stable African plate to the south.

Detailed sedimentological and stratigraphical studies (Waldron 1984a, Poisson
1977, Robertson and Woodcock 1980a,b) have demonstrated that the Mesozoic carbonate successions deposited on individual platform units record a similar sedimentary history, represented by a progressive change from low energy restricted environments in the Triassic and Jurassic to more open marine conditions in the Cretaceous. Typically, Upper Triassic and Lower Jurassic dolomites, algal laminated and reefal limestones, diagnostic of low energy inter-tidal and supra-tidal environments, pass up into thick (e.g. 500m on the Karacashihar Massif, Dumont 1976 and 2500m on the Bey Daglari, Poisson 1977) neritic and stable platform limestones characterised by a variety of biosparites, stromatolites, coraline limestones, ooli;es and pelsparites. With the important exception of the Barla Dag succession which from the Lower Jurassic time onwards the succession is dominated by deeper water pelagic sediments, individual platforms subsided at differing rates while remaining in the neritic carbonate deposition zone. Following a period of emergence, brecciation and fissuring of the platforms in the early Late Cretaceous, the Upper Cretaceous was characterised by a phase of rapid subsidence during which pelagic Globotruncana-bearing limestones were deposited on both large and small platforms alike. Fine clastics, derived from ophiolites and deep water sedimentary units of the Antalya Complex were shed onto the subsided platform in the latest Cretaceous-early Palaeocene interval before outboard Mesozoic passive margin successions were thrust onto the relatively autochthonous platform units. On some platforms, unbroken pelagic sediment deposition continued into the Palaeocene (e.g. south west Bey Daglari, Poisson 1977).

In a preliminary palaeomagnetic study over 2000 orientated samples have been collected from sites located in thick, structurally coherent platform and pelagic limestone successions on each of the main 'relative autochthons' in the Isparta Angle area (i.e. the Bey Daglari, Anamas-Akseki, Davras Dag, Dulup Dag, Barla Dag and Karacasihar Massifs, see fig. 5). The ages of sampled successions were based on combined lithological and palaeontological criteria (see table 1).

Although a few sites were located in Triassic, Jurassic and Lower Cretaceous carbonate platform successions, sampling was concentrated in the Late Cretaceous-Early Tertiary stratigraphic interval for the following reasons:

a) pelagic chalks deposited onto the platform during this interval typically contain fine clastic material derived from deeper water sedimentary and ophiolite units. As limestones are known to be only weakly magnetic, an 'exotic' clastic component in the sediments should improve the chances of collecting samples with a measurable remanence. These 'impure' carbonate sediments are typically coloured pale or dark grey, contrasting with the lighter coloured platform limestones.

b) Late Cretaceous and Early Tertiary pelagic successions are thinly bedded and fine
grained while many earlier Mesozoic platform limestones are massively bedded, coarse grained and recrystallised. Finer grained sediments are generally more appropriate for palaeomagnetic studies as their sedimentation rate is low enough to average out secular variations of the geomagnetic field over the thickness of the sample (Kissel et al. 1984). By contrast with the the Triassic and Jurassic neritic limestones which are commonly irregularly bedded, the more regular bedding planes of pelagic sediments allows an accurate determination of the structural dip.

c) Intuitively, one might expect important tectonic rotations to occur between independent basement blocks during the Late Cretaceous and Tertiary phase of crustal convergence and collisional tectonics rather than in the Triassic to Mid Cretaceous 'passive margin' phase.

Where possible, each independent platform unit was sampled at a number of geographically separated localities to test for internal block rotations, and in order to minimise any declination error that might be attributed to tectonic tilting about an inclined rotation axis, sampling sites were preferably located only in shallowly dipping successions. Sections close to faults, landslides and zones of penetrative veining were avoided.

All carbonates were drilled in situ and orientated with both a sun and magnetic compass. Where appropriate, samples were usually collected in groups of 2-3 at 2-5m. stratigraphic intervals. Over long sections it was hoped that this sampling procedure might result in the identification of both normal and reversed polarity intervals. The optimum core length was 4-5cm., although at many sections it was only possible to drill 3-4cm. into the rock surface before the core fractured internally. The weathered surface of each core was then trimmed-off with a circular saw leaving a standard cylindrical subsample 2cm. high.

7.4.3.2 LABORATORY PROCEDURES AND RESULTS

The natural remanent magnetisation of each sample was then measured with a two component cryogenic magnetometer at the University of Edinburgh. A second independent measurement of the NRM was made on the two component cryogenic magnetometer at the University of Newcastle. The vector difference between these two independent estimates and/or the noise level of each magnetometer (c. 5x10⁻³ mAm⁻¹) provided a measure of the reliability of the magnetisation obtained.

NRM intensities are generally very low, with values ranging from below the noise level of the magnetometer to greater than 1.0mAm⁻¹. The majority of samples have an NRM intensity of between 0.01mAm⁻¹ and 0.1mAm⁻¹, less than an order of
Fig. 6 NRM directions at each site (a, shows sites on main Bey Daglari platform, Davras Dag and Kemer Zone off-margin limestones; b, shows sites on other platform units). All mean NRM directions (indicated as crosses) prior to and subsequent to application of bedding correction have a positive inclination. Mean vectors characteristically group in the NW and NE quadrants close to the direction of the present day geomagnetic field (*). Typically, the radius of the $\alpha_{95}$ cone of confidence increases on application of a structural correction (full circle shows cone of confidence after application of structural correction while dotted circle indicates cone of confidence after application of field correction), indicating that the NRM is probably dominated by a viscous present-day field component. Representative pairs of two-component Zijderveld plots are also presented. N=Number of subsamples.

Fig. 7 IRM acquisition curves for two limestone samples (TPB126 from Pembeli, Davras Dag platform; TFD17 collected at Finike, Bey Daglari platform). $J_0$ is induced magnetisation at 1000mT, while J represents induced magnetisation in external field. Curves show likely presence of magnetite and haematite (or goethite) in both samples. These limestones remain unsaturated in a maximum external field of 1000mT.

Fig. 8 Representative demagnetisation data from a variety of carbonate platform successions in the Isparta Angle. See table 1 for list of site abbreviations and fig. 6 for details of site location. Zijderveld plot conventions are summarised in appendix 1.

Table 1 Summary of NRM data from sites in Isparta Angle area. Mean remanence vectors have been corrected for structural tilt. Site abbreviations are also included. N=Number of subsamples, MB C=Mean bedding correction (strike measured 90° anticlockwise away from the direction of maximum dip).
FIG. 7 IRM ACQUISITION CURVES

- TP8126
- TFD17
FIG. 8 AF AND THERMAL DEMAGNETISATION OF PILOT SAMPLES
<table>
<thead>
<tr>
<th>SITE</th>
<th>PLATFORM</th>
<th>AGE</th>
<th>LENGTH OF</th>
<th>N</th>
<th>MDEC</th>
<th>MINC</th>
<th>A95</th>
<th>K</th>
<th>M.B.C.</th>
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</thead>
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<tr>
<td>FINIKE</td>
<td>BEY</td>
<td>Pal.</td>
<td>c.20m</td>
<td>29</td>
<td>001.44</td>
<td>34.57</td>
<td>33.95</td>
<td>5.69</td>
<td>038/43</td>
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<tr>
<td>(FI)</td>
<td>DAGLARI</td>
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<td></td>
<td></td>
<td></td>
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<td></td>
<td></td>
</tr>
<tr>
<td>SAKLIKENT</td>
<td>BEY</td>
<td>Pal.</td>
<td>c.40m</td>
<td>44</td>
<td>026.54</td>
<td>67.54</td>
<td>39.38</td>
<td>4.23</td>
<td>332/47</td>
</tr>
<tr>
<td>(DK)</td>
<td>DAGLARI</td>
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<td></td>
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</tr>
<tr>
<td>CHAMLEDERE</td>
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<td>U.Cre.</td>
<td>c.150m</td>
<td>130</td>
<td>007.00</td>
<td>41.28</td>
<td>29.40</td>
<td>7.59</td>
<td>086/33</td>
</tr>
<tr>
<td>(CD)</td>
<td>DAGLARI</td>
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<tr>
<td>BAYATBADEMLISI</td>
<td>BEY</td>
<td>U.Cre.</td>
<td>c.30m</td>
<td>40</td>
<td>335.34</td>
<td>50.47</td>
<td>23.05</td>
<td>12.35</td>
<td>145/12</td>
</tr>
<tr>
<td>(YA)</td>
<td>DAGLARI</td>
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<td></td>
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<td></td>
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</tr>
<tr>
<td>PEMBELLI</td>
<td>DAVRAS</td>
<td>U.Cre.</td>
<td>c.100m</td>
<td>126</td>
<td>012.25</td>
<td>41.26</td>
<td>33.21</td>
<td>5.94</td>
<td>052/37</td>
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<tr>
<td>(PB)</td>
<td>DAG.</td>
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<td></td>
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<td></td>
<td></td>
</tr>
<tr>
<td>FINDOS COL</td>
<td>DAVRAS</td>
<td>U.Cre.</td>
<td>c.30m</td>
<td>49</td>
<td>349.70</td>
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<td>47.89</td>
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<td>(FD)</td>
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<td>KESME BRIDGE</td>
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<td>Jur.</td>
<td>c.100m</td>
<td>110</td>
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<td>59.05</td>
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<td>c.80m</td>
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<td>KARADICKEN</td>
<td>KARACAHISAR</td>
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<td>c.25m</td>
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<td>058.87</td>
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<td>45.68</td>
<td>3.14</td>
<td>359/20</td>
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<td>BARLA DAG</td>
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_table_1_
magnitude above the remanence intensity of the sample holder (usually <0.01mAm⁻¹). Samples with an NRM intensity greater than 0.025mAm⁻¹ showed directional repeatability to within ±5°. Prior to magnetic cleaning all samples with intensities less than 0.025mAm⁻¹ were rejected from the sample population. At many sites over half the samples were discarded at this stage.

Stereonets of field corrected NRM directions at each locality show a broad concentration of vectors in the north east and north west quadrants (see fig.6a,b). Without exception, mean remanence inclinations at each site are positive (down) and do not distinctively group away from the present day geomagnetic field direction. There is no observable improvement in the grouping of the mean remanence vectors at each locality after correcting for the dip of the beds. On the contrary, at Adada on the Anamas Dag Massif, a negative fold test implies that the NRM is dominated by a remanence component acquired subsequent to tectonic tilting and thus cannot be assumed to be primary in origin.

In order to identify the magnetic minerals present, isothermal remanent magnetisation (IRM) studies were carried out. Four samples were exposed to magnetising field from 50mT to 500mT in incremental steps of 50mT and then at 100mT steps up to 1000mT (see fig.7). Each sample displayed a similar behavior, rapidly acquiring an IRM in fields of 50 to 500mT. This was followed by a short plateau in the magnetisation curve prior to acquiring a further IRM, this time more slowly up to 500mT. None of the samples studied is saturated at 500mT. The initial IRM is believed to be due to magnetite whose maximum coercivity is 300mT (see chapter 3 section 3.7) while the IRM acquired above this field could be due to either goethite or haematite both of which have saturation fields well above 500mT.

To determine if the NRM includes a stable remanence component, samples were magnetically 'cleaned' using both stepwise thermal and alternating field demagnetisation techniques. A minimum of four samples (sometimes as many as ten) from each locality were exposed to AF demagnetising fields in 2mT incremental steps up to 20mT and then at 4mT steps until either the remanence was comparable with the noise level of the measuring instrument, or the magnetic intensity stabilised. The NRM of a further two or three samples from each site were then thermally demagnetised at 50° incremental steps up to 500°C or 600°C. If alternating field demagnetisation failed to reduce the remanence intensity effectively in high fields (i.e. >40–50mT), samples were then demagnetised thermally at 100°C incremental steps up to 500°C.

Samples collected from the main Bey Daglari platform near Saklikent (see fig.8f), Bayatbademlisi (see fig.8a,b) and Finike (see fig.8e) responded well to both thermal and alternating field demagnetisation, with a stable remanence vector being isolated in fields of less than 15mT and temperatures of less than 200°C. Typically,
any low coercivity or low blocking temperature components were northerly directed and probably represented a secondary component acquired in the present day geomagnetic field. Similarly, samples collected from Upper Cretaceous platform successions on the Barla Dag (by Lake Egridir, see fig.8d), Anamas Dag (at Cay Koy gorge, see fig.8c) and Davras Dag (at Findos Col, see fig.8o) are dominated by a stable remanence direction that is typically orientated between north and north west, with an inclination between $30^\circ$ and $45^\circ$ after correcting for bedding tilt. The isolated stable remanence direction is quite unrelated to the orientation of the present day field direction. The MDF of samples collected at all these sites is typically very low (almost always $<20$ mT) implying that the remanence is carried predominantly by low coercivity components. Magnetite is probably the principal carrier, although low coercivity haematite might also contribute to the remanence. Complementary pairs of two component plots show that for these normal polarity samples a stable end point vector can be clearly defined by fitting a straight line through the last 5-8 points. By contrast, it appears that at some sites (e.g. at Finike Dam, see fig.8k,m; and at Bucak Gorge, see fig.8j) remanence vectors move toward reversed polarity directions but fail to reach a stable end-point. At these sites, the remanence is clearly dominated by a viscous present-day field component which is not completely removed before the intensity reaches a value comparable with the noise level of the measuring instrument. Although further pilot demagnetisation studies clearly need to be performed, it seems likely that the remanence vector for these reversed polarity samples is moving toward a direction diametrically opposite to that retained by the normal polarity samples.

A proportion of samples at many sites remained largely unaffected by AF demagnetisation treatment (e.g.fig.8p), and it was assumed that in these cases the remanence was carried by either goethite or haematite of probable secondary origin. Samples with larger concentrations of goethite (blocking temperature 60-100$^\circ$) decreased in intensity rapidly at low temperatures (e.g.see fig.8q), while samples containing haematite as the principal remanence carrier show a steady decrease in intensity to higher temperatures. As vector plots of haematite and magnetite bearing samples at many sites are virtually indistinguishable (e.g. compare fig.8a with fig.8b), it can be safely assumed that the haematite and magnetite components in many samples carry a similar direction of magnetisation. This strongly suggests that the remanence was acquired at, or shortly after deposition.

Contrasting with the NW to N directed stable remanence of the main platform units, some off-margin platform successions typically give stable remanence directions that group in the NE quadrant. For example, steeply dipping Jurassic and Lower Cretaceous platform limestones in the SW Antalya Complex (Kemer Zone of Robertson and Woodcock 1982) to the east of the main Bey Daglari platform (two sites, see fig.6) have an NRM that includes at least two components with overlapping coercivity and
I draw three main conclusions from this palaeomagnetic study. First, the remarkable rotation of the Troodos ophiolite occurred relatively rapidly in the Late Cretaceous-Early Tertiary interval, quite soon after its genesis. Second, the crustal area involved was relatively small, with at least one of its boundaries possibly being preserved within the present area of Cyprus beneath the Kyrenia Range lineament. A second boundary probably lies close to the south and south-west margins of the preserved ophiolite complex. Third, although the exact mechanism of the rotation cannot yet be specified in detail, it seems likely that a major driving force resulted from the oblique subduction of oceanic crust beneath the Troodos ophiolite which was then probably sited in a 'fore-arc' setting immediately adjacent to the Kyrenia continental margin to the north. Active deformation in the Mamonia Complex (SW Cyprus) and Kyrenia Range (N Cyprus) during the Late Cretaceous-Early Tertiary interval is consistent with these areas having been located close to a major crustal lineament during the critical period of microplate rotation.
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All pairs of two component plots have axis conventions as illustrated overleaf. Upper graph shows remanence vector projected onto vertical (yz) plane, while lower plot shows remanence vector projected onto horizontal (xy) plane. Declination can be measured directly from the lower graph, while true inclination can only be measured from the upper graph if component out of the plane of projection is zero (i.e. the x component).