Miocene Basin Evolution of the Isparta Angle, Southern Turkey

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Thesis submitted for the degree of
Doctor of Philosophy

University of Edinburgh
1995
To Lara and Brontë
DECLARATION

I declare that this thesis has been written by myself and is the result of my own research, except where contributions have been stated and duly acknowledged.

Rachel Flecker
Acknowledgements

This project was set up and supervised by Alastair Robertson and John Underhill at Edinburgh University and André Poisson at Orsay (Paris XI) where I spent some part of every year. Alastair and André both came into the field with me and contributed greatly both to the academic content of this thesis and also to my enjoyment of fieldwork in Turkey. All three of my supervisors have read this thesis and I am particularly grateful to them all for their prompt response to my chapters as the deadline for the end of the project drew near.

Perhaps the greatest debt of any fieldwork project is owed to those who brave the heat, the bug-ridden water and the outrageous driving to come and field assist. Toby Harris, James Flecker, Mike Metcalf, Mark Hodson, Elizabeth Pickett and Lawrence Miles have all left an indelible mark on this Ph.D. visible only in my note book and on my sample bags, but permeating unseen throughout.

In Edinburgh my research has been facilitated by the help and guidance of the following people: Pedder Aspen, Mike Shaw, Nicky McEwan, Helena Jack, Diana Batey, Janet Cuthill, Yvonne Cooper, Geof Angell, Shane Voss and Ian Chisholm. I received similar support from the team in Batiment 504 in Orsay and would like to thank Geneviève Roche in particular. Rob Ellam, Anne and Vinney (East Kilbride), Brian Rosen (Natural History Museum) and Carla Müller all provided essential technical expertise.

The Turkish aspect of this project would have been so much more difficult had it not been for the endless help of Mustafer Gerger (Pepsi factory, Antalya) who managed the bureaucracy of exporting rocks and renewing visas with great good humour and patience. I would like to thank Celal Sengör for introducing me to Musafer and providing logistical support from Istanbul Technical University. The hospitality and friendship of those I met in Turkey is unforgettable and I can't wait to go back in a few months time and see them all. Particular mention must be made of the people at Unals Kebab Salonu, Feyvsi, Atilla, Halide, and Timor, Ida and Ece Üstömer.

Useful discussions have been had with the following people: John Dixon, Roger Scrutton, Dick Kroon, Bill Austin, Olivier Monod, Dominique Frizon de Lamotte, Graham Williams, Peter Clift and past and present members of the Tethyan Group. Jon Turner constructed and explained the subsidence curves presented here. I have been particularly fortunate to be writing up at the same time as Anne Payne and Clare Glover. Their knowledge of different regions/times of the Eastern Mediterranean and their support and encouragement throughout the painful business of finishing has been invaluable.

Finally my thanks Phil and Paul for their tolerance; to Flatmate Flem for his yummy meals and enthusiasm and to the KB supper club for their solidarity in adversity: John, Sandra, Jon, Anne , Clare, Jo. As I write, Jerry is checking and formatting my references and La is sticking on page numbers. Thanks too, to Sprout for the flowers.
ABSTRACT

The study of basins developed on top of older suture zones is demonstrably important in the Tethyan region where rifting and convergence have occurred repeatedly from the Mesozoic to the present-day. In southern Turkey, three Miocene basins developed in the Isparta Angle suture zone. Two of these, the Aksu and Köprü basins, are orientated north-south, parallel to pre-Miocene lineaments in the basement (e.g. Antalya Complex). These basins are underlain by a mosaic of deformed Mesozoic carbonate platforms, deep-sea sediments and ophiolitic units. The Manavgat basin, to the east, overlies Permian meta-carbonates (Alanya Massif) deformed by NW-SE trending basement structures. The northern margin of the Manavgat basin, in the south-east of the area, is also orientated NW-SE, but it was open to the Mediterranean Sea to the south.

Prior to Burdigalian-Langhian transgression, a thick (~1.5km) succession of predominantly continental conglomerates and sandstones accumulated in the south of the Aksu and Köprü basins (Kizildag Formation). Coeval thrusting of the Lycian Nappes on the western limb of the Isparta angle is inferred to have induced block faulting of the foreland, exploiting pre-existing basement weaknesses and generating accommodation space.

In the Manavgat basin palaeocurrents and south-to-north diachronity of transgression demonstrate that the Alanya Massif formed a palaeogeographic high to the north of the basin. This was colonised by coral and algal fringing reefs, which shed abundant shallow-water debris, deposited as calcarenites in Langhian times (Oymapinar Limestone). Elsewhere, to the south and west, Late Burdigalian-Langhian patch reefs developed within coastal fan-delta conglomeratic sequences. Detailed study of the interaction between coral growth and clastic influx reveals that relative sea level rise outstripped sediment influx at this time.

Extensional faulting during the Langhian in the Manavgat basin generated numerous micro-faults which trend parallel to NW-SE basement lineaments. This faulting led to the deposition of localised talus (Çakallar Formation) and formation of an asymmetrical horst-graben structure. Reef deposition was abruptly terminated and 3-500m of Serravallian planktic foraminiferal marls (Geceleme Formation) then accumulated as post-rift fill. In the Aksu and Köprü basins, a correlative transition from shallow-water carbonates to deeper-water turbidite deposition (Karpuzçay Formation) is observed.
ÖZ


Burdigaliyan-Langhiyan transgresiyonundan önce Aksu ve Köprü havzalarının güneyinde kümütaşı ve karasal karbonat ağırlıklı kalın (~1.5km) bir çökeme (Kızıldağ Formasyonu) meydana gelmiştir. Lysiyan naplarının eşzamanlı batı doğru Isparta ağısına bindirmeleri, var olan yumuşak temel kayaçlardan yararlanarak ve gerekli sahayi oluşturan ülke blok deformasyonunu oluşturmuştur. Anadolu’nun batıbatı kesiminde yüksek paleo-cografik yapılar meydana getirmiştir. Bu, Langhiyan yaşlı kalkarenit gibi depolanmış bolca sığ-su kırtılı olarak çökemiş (Oymapınar kireç taşı) kenar resifi depolar ve algular tarafından istila edilmişdir. Öte yandan güneye ve batıya doğru, üst Burdigaliyan-Langhiyan parça resifi delta yelpazeli bir konglomeratik sekanslar içerisinde gelişmiştir. Resif depoları ile bolca detritik malzememin arasındaki ilişkinin detaylı incelemesi, göreceli olarak su düzeyinin yükselmeleri sedimentlerin depolanmasına oranından daha fazla olduğunu göstermektedir.

Manavgat havzasındaki paleo-yapılar ve güneyden kuzeye doğru transgresyon diyakroneiti, Alanya masifinin havzanın kuzey kesiminde yüksek paleo-cografik yapıları meydana getirdiğini göstermektedirler. Bu, Langhiyan yaşlı kalkarenit gibi depolanmış bolca sığ-su kırtılı olarak çökemiş (Oymapınar kireç taşı) kenar resifi depolar ve algular tarafından istila edilmişdir. Öte yandan güneye ve batıya doğru, üst Burdigaliyan-Langhiyan parça resifi delta yelpazeli bir konglomeratik sekanslar içerisinde gelişmiştir. Resif depoları ile bolca detritik malzememin arasındaki ilişkinin detaylı incelemesi, göreceli olarak su düzeyinin yükselmeleri sedimentlerin depolanmasına oranından daha fazla olduğunu göstermektedir.

Manavgat havzasında Langhiyan boyunca extensif deformasyon KB-GD doğrultulu temel kayaç çizgiselliklerine paralel olarak pek çok mikro-fay meydana getirmiştir. Bu faylanma, yamaç deposuna (Cakallar formasyonu) ve asimetrik horst-graben yapılarında bir formasyona olanak sağlar. Resif deposu aniden sona ermiş ve 3-500 m kalınılığında Serravalliyan yaşlı planktik foraminiferayı marnlar (Gecelerme formasyonu) rift çökeme sonrası şekilde çökemişdir. Aksu ve Köprü havzalarında sığ denizel karbonatatlardan derin deniz turbidit deposuna (Karpuzçay Formasyonu) doğru bir korelatif geçiş gözlemlenmiştir. Köprü havzasından itibaren paleo-yapılar havzanın kuzeyinden güneyine doğru kanalise
olmuş temel kayaç sedimentler içerisinde K-G yönlü çizgiselliklere işaret ederken güneyde küçük topografik yapılar sahiptir.

Erken Totoniyan’da Manavgat havzasında Geceleme Formasyonu hızlı bir biçimde çökelmiş kiltaşı, kumtaşı ve iri taneli kongromeralara ~10m kalınlığında Alanya masifi meta-karbonatlarından kopan bir blok ihtiva ederek geçmektedir. Kıırıntı-akışkan süreçleri bu sıralı depolanmaya (Karpuz Formasyonu) hakim olmuştur ve foraminiferal etüdlerle birlikte havzanın Tortonoyan-Messiniyan’da yükseldiğini ve depolandığını kanıtlamaktadır. Aksu ve Köprü havzalarının güneyindeki yükselmeler Karpuzçay turbuditlerinin sig denizel delta yelpase kongromelara geçişleri ile birlikte az miktarda parça resiflerini meydana getirmektedir.

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Chapter 1

INTRODUCTION

1.1 Context

The Mediterranean Sea lies along a complex zone of convergence between the Eurasian and Turkish plates to the north and the African, Arabian and Indo-Australian plates to the south. This boundary is marked today by the Alpine-Himalayan orogenic belt (Fig. 1.1). Palaeozoic-Recent fragments preserved within, or on the margins of this suture contain partial records of its history. In the Eastern Mediterranean intensive study of these fragments has been carried out with a view to reconstructing the nature of past and present collision, which in this area is complicated by the presence of a number of micro-plates along the suture zone (Robertson et al., 1991b). Much of this work has focused either on the very young, neotectonic setting of the area, or on its Mesozoic-Early Tertiary history. A number of Miocene basins exposed in south coastal Turkey, document the evolution of the northern margin of the Mediterranean at this time. Three of these, The Aksu, Köprü and Manavgat basins are situated at the junction between two arcuate orogenic belts, the Hellenides and Taurides (Fig. 1.2), in a zone known as the Isparta Angle (Blumenthal, 1963). Study of these Mid-Tertiary basins is essential in trying to bridge the gap in knowledge between Mesozoic-Early Tertiary models and the present day tectonic configuration.

1.2 Plate tectonic setting of the Eastern Mediterranean

The Eastern Mediterranean has a prolonged history of convergence involving the closure of a large Palaeozoic ocean (Palaeotethys) whose suture lies within northern Turkey (e.g. Pickett, 1994, Üstömer, 1993) and the initiation and closure of smaller oceans (Neotethys) during the Mesozoic and Cenozoic (e.g. Robertson and Dixon, 1984). The location of the present-day convergent zone in the Eastern Mediterranean is well established. Continental collision (Bitlis) and strike slip movement is
Figure 1.1 Regional plate setting of the Mediterranean Sea.
Figure 1.2   Map of a) the north-eastern Mediterranean showing the main suture zones and tectonic units; b) Neogene sediments of the Isparta Angle showing the three basins studied in this project.
occurring (North and East Anatolian Faults) in the far east of the Mediterranean. In the west, subduction is taking place on the Hellenic arc while the Aegean Sea undergoes extension. The segment of the Hellenic-Cyprus arc which lies to the south of Isparta Angle is less well delineated (Jackson and McKenzie, 1984), but the general consensus is that some combination of subduction and strike-slip motion are taking place in an arcuate zone to the south of Cyprus.

Pre-Miocene subduction in this area is thought to have been taken place along the Kyrenia Range in Cyprus, to the north of the present location (Robertson and Woodcock 1986). The connection between this subduction system and the Hellenic arc is not at all well defined given the present plate configuration, still less for the Oligocene and earlier. One possibility is that the Hellenic arc system continued onto land up the front of the Lycian Nappes which form the western arm of the Isparta Angle (Poisson, 1984). From here the Hellenic arc might have been connected to the Cyprus system along a suture zone located within the Isparta Angle itself (Robertson and Grasso, in press).

1.3 Structural framework and geological history of the Isparta Angle

The heterogeneous rocks of the Isparta Angle have been studied by many workers over the last three decades. Interest has focused on the complex mosaic of allochthonous and relatively autochthonous Mesozoic successions. These can be divided into six broad groups (Lycian Nappes, Bey Daglari, Antalya Complex, Anamas-Akseki platform, Beysehir-Hoyran-Hadim Nappes and Alanya Massif) located on figure 1.3. The dominant characteristics of each are summarised below:

Lycian Nappes
This extensive nappe system includes Mesozoic carbonate, radiolarites, sandstones and ophiolitic rocks (Graciansky, 1972; Poisson, 1977a). It is thought to have been emplaced during the Mid-Tertiary forming a foreland flexural basin to the south and south-east (Hayward, 1982b). Further thrusting of the Lycian system is thought to have recurred around the Mid-Miocene, deforming the Lower Miocene sediments in the flexural basin (Gutnic et al., 1979; Hayward, 1982) and causing a
Figure 1.3 Map of the Isparta Angle region and the tectonic units described in the text.
widespread remagnetisation event out-board of the nappes (Morris and Robertson, 1993). The Lycian Nappes front and parallel flexural basin strike approximately NE-SW marking the western boundary of the Isparta Angle. Active thrusting of the Lycian system was terminated in the Late Miocene (Hayward, 1982b). The internal structure of the Lycian Nappes and their emplacement style and direction are still controversial issues. These questions are currently being investigated in a Ph.D. project undertaken by A. Collins at Edinburgh University.

**Bey Daglari**

The Bey Daglari-Susuz Dag Cretaceous carbonate platform forms most of the western margin of the Aksu basin. It has generally been viewed as relatively autochthonous (Poisson, 1977; Gutnic et al., 1979; Poisson and Robertson, 1990) and is unmetamorphosed. Palaeomagnetic studies (Kissel and Poisson, 1987; Morris and Robertson, 1993) suggest that it has rotated anticlockwise by 30° since the Langhian. This rotation has been genetically linked to thrusting in the Lycian Nappes (Morris and Robertson, 1993) and the formation of the NNE-SSW trending Bey Daglari anticline (Kissel et al., 1993). Palaeocurrent studies of Miocene sediments in a small basin overlying the eastern flank of the Bey Daglari (Hayward and Robertson, 1982; Fig. 1.4) suggest that sediment was derived from the Antalya Complex to the east. Subsidence of the eastern flank of the Bey Daglari is thought to have been induced by flexural subsidence due to lithospheric loading by the Lycian Nappes (Fig. 8.1; Hayward and Robertson, 1982; Hayward, 1982b). Termination of sedimentation in the Late Miocene resulted from the westward emplacement of the Antalya Complex on top of Mid-Miocene sediments along a basal thrust (Hayward and Robertson, 1982; Hayward, 1982b).

**Antalya Complex**

Perhaps the most contentious debate surrounding the Isparta Angle focuses on the Antalya Complex. Originally named, the Antalya Nappes (Lefevre, 1967), this unit mainly consists of Palaeozoic-Maastrichtian sandstones, shales, carbonates, radiolarites and ophiolitic rocks. The argument centres on the origin and emplacement of the Antalya Complex (e.g. Dumont et al., 1972; Monod, 1977; Poisson, 1977a; Woodcock and Robertson, 1977b; Gutnic et al., 1979; Ricou et al., 1979;
Figure 1.4 A more detailed map of the Isparta Angle showing the north-south structural lineaments within the Antalya Complex and the location of the palaeomagnetic sites of Morris and Robertson (1993). Modified from Waldron (1984).
Woodcock and Robertson, 1979; Hayward, 1984; Robertson and Woodcock, 1980; 1982; Poisson, 1984; Ricou et al., 1984; Waldron, 1984; 1986; Marcoux et al., 1989; Robertson, 1993) the alternative tectonic models for which are summarised below and illustrated in figure 1.5 (Robertson, 1993):

- The Antalya Complex was rooted far to the north and thrust southwards from a single Neotethyan ocean (Ricou et al., 1974; 1975; 1979; Marcoux et al., 1989);
- The Antalya Complex had its origins in a southerly Neotethyan (Pamphylian) basin and was thrust northward by tens to hundreds of kilometres (Dumont et al., 1972; Monod, 1976; 1977);
- The Antalya Complex was routed within the Isparta Angle either in a deep marine basin (Poisson, 1984) or in palaeogeographically complicated ocean basins within which Antalya units were thrust relatively short distances outwards on to neighbouring carbonate platforms (Robertson and Woodcock, 1980; 1982; 1984; Waldron, 1981; 1984a; b; Robertson 1993).

With reference to the present study the emplacement of the Antalya Complex may be of great importance in terms of the underlying lithospheric weaknesses that govern the fabric of the basement. The north-south orientation of the Köprü and Aksu basins for instance parallel structural trends within the Antalya Complex (Fig. 1.4). According to the theory which supports an origin of the Complex within the Isparta Angle (internal), these lineaments are the sites of Triassic rifting and Early Tertiary collision and thrust emplacement. The theories of an external origin of the Antalya Complex either to the north or south, rely on complicated thrusting patterns to produce the dominant north-south structural trend.

Exposure of this unit occurs in the far south west of the Isparta Angle, on the basement promontory dividing the Köprü and Aksu basins, to the south of Lake Egirdir and in a thin strip to north of the Manavgat basin. Poisson et al. (1983) showed that the Antalya Complex was essentially emplaced prior to the Oligocene in the central part of the Isparta Angle, earlier than in the south-western segment, where Hayward and Robertson (1982) suggested that final emplacement occurred in the Late Miocene (see Bey Daglari above). Palaeomagnetic studies carried out by
Figure 1.5 Alternative tectonic interpretations of the Antalya units, based loosely on published concepts: a) thrust from the north from a single Neotethyan ocean basin (Ricou et al., 1974; 1975; 1979; Marcoux et al., 1989); b) thrust from a southerly Neotethyan ocean basin (Dumont et al., 1972; Monod, 1976; 1977); c) thrust from a deep marine basin within the Isparta Angle (Poisson, 1977; 1984); d) an origin as a palaeogeographically complicated small ocean basin within the Isparta Angle; units were thrust relatively short distances outwards onto adjacent carbonate platforms (Robertson and Woodcock, 1980; 1982; 1984; Waldron, 1981; 1984a and b). Diagram taken from Robertson (1993).
Morris and Robertson (1993) suggest that the anticlockwise rotation suffered by the Bey Daglari during the Mid-Late Miocene also affected the south-western segment of the Antalya Complex, causing west-vergent thrusting.

**Anamas-Akseki Platform**

Anamas-Akseki platform, like the Bey Daglari is considered to be a relatively autochthonous unit (Monod, 1977) comprising dominantly Mesozoic carbonate. It forms the eastern border of the Köprü basin and the basement to the Manavgat basin in its far north-western corner. Morris and Robertson (1993) indicated that the Anamas-Akseki Platform in the north of the Isparta Angle was also affected by Mid-Late Miocene anticlockwise rotation.

**Beysehir-Hoyran-Hadim Nappes**

The Beysehir-Hoyran-Hadim Nappes (BHH) consist mainly of Upper Palaeozoic-Lower Tertiary carbonate, sandstone and ophiolitic rocks (Monod, 1977). Arcuate thrust belts trending north-south in the north and NW-SE in the south characterise the Beysehir-Hoyran-Hadim Nappes which mark the eastern arm of the Isparta Angle and border the Anamas-Akseki platform to the west. Narrow exposures of Eocene flysch parallel these thrusts indicating flexural loading at this time. Palaeomagnetic evidence from this area (Kissel et al., 1993) suggests that the BHH nappes underwent 40° of clockwise rotation in the Upper Eocene to Oligocene, but there is no indication that subsequent movement has occurred.

**Alanya Massif**

The Alanya Massif consists mainly of Permian meta-carbonates (Özgül, 1983; Okay and Özgül, 1984). Its structure is complex and still not clearly understood, but it is of great interest as it contains blueschist metamorphic facies and thus is indicative of the high-pressure low-temperature metamorphism associated with subduction zones. It borders the north-eastern part of the Manavgat basin and is exposed at the core of an anticline within the basin itself. It is thought to have been emplaced onto the Antalya Complex to the north during the Early Eocene (Monod, 1977; Okay and Özgül, 1984). Detailed field assessment of the internal
structure of the Alanya Massif has been carried out (Robertson, unpublished data). This suggests that many of the structures within the unit parallel the NW-SE trend of its northern margin.

### 1.4 An overview

The Isparta Angle represents a long-lived suture zone across which episodic compression and extension took place. There is evidence both of rifting and collision throughout the Mesozoic and Early Tertiary successions preserved within the area. Miocene sequences too, on a smaller scale, document the generation of accommodation space and subsequent uplift and erosion. Even today, neotectonic graben systems such as the Kovada graben to the south of Lake Egirdir in the north of the area are concentrated within the Isparta Angle. The Plio-Quaternary evolution within the angle is the subject of a Ph.D. project currently being carried out by C. Glover at Edinburgh University. Significantly, much of the deformational structures within the Isparta Angle, dating from the Early Mesozoic to the present-day, are orientated north-south. Figure 1.6 is a satellite photograph showing the strong north-south grain to the fabric of the region.

### 1.5 General Aims

- To study in detail the sedimentological character of the Miocene succession within the Isparta Angle in order to construct facies models during basin evolution;
- To assess the structural development of the three Miocene basins within the Isparta Angle and compare them;
- To evaluate the role of eustatic sea level change in controlling sedimentation;
- To examine the tectonic implications of these studies with respect to both the evolution of the Isparta Angle as a whole and regional tectonic history of the Eastern Mediterranean during the Miocene.

### 1.6 Techniques used

The major data base for this project comes from field-based studies which
Figure 1.6  Satellite photograph of part of the area showing the strong north-south grain to the fabric of the Isparta Angle.
were carried out during a total of 8 months fieldwork over three years. The geological map of Miocene-Quaternary sediments of the Aksu, Köprü and Manavgat basins produced by Akay and Uysal (1985) as part of his M.T.A. report was used extensively as a guide to the area. Detailed sedimentological logging on various scales was used as a fundamental tool for examining intra- and inter-basin evolution. This was combined with extensive palaeocurrent analysis and provenance studies on the conglomerates where over 100 clasts were recorded for each locality. Where high tectonic dips required, palaeocurrent data were corrected for tilt.

Much of the data base produced by Akay and Uysal (1985) in his study of the area was of a biostratigraphic nature. The current study has expanded his work, using nannoplankton biostratigraphy undertaken with Dr. C. Müller, coral identification carried out with the help of Dr. B. Rosen (Natural History Museum) and strontium isotope stratigraphy. Isotope work was planned in order to assess the age of shallow-water carbonates, which through the paucity or absence of diagnostic foraminifera or nannoplankton, proved difficult to date. It was carried out at the Scottish Universities Research and Reactor Centre in East Kilbride under the guidance of Dr. R. Ellam. The sample preparation and methodology of this technique is described in chapter 6.

Prior to the isotope work a detailed diagenetic study of the biogenic carbonate was required. This involved XRD analysis of coral samples to ascertain the relative quantities of primary aragonite and secondary calcite. Samples of coral were first drilled from a fresh hand specimen and washed in ethanol standing in an ultrasound bath for at least half an hour. After removal, the samples were rinsed and dried and ground to a fine powder using a pestle and mortar. XRD analysis was then carried out and compared to standards containing known compositions of aragonite and low-Mg calcite.

Diagenetic studies of the foraminifera used for Sr isotope analysis were attempted, by first examining washed samples picked at 63µ under a binocular microscope and later using a scanning electron microscope at Stuttgart University, Germany. Each foraminifer was mounted on a stub
with conductive graphite mounting material, or a double sided sticky tab prior to gold-coating.

General diagenetic and textural features of the carbonates and sandstones were investigated using standard thin section microscopy and acetate peels. Poorly lithified samples were first impregnated with resin. Point counting of selected sandstones was undertaken as part of provenance investigations. Over 300 points were counted initially and the percentages recorded. A further 100 points were then counted and the total percentages calculated. If the difference for each grain-type between 300 and 400 point counts was less than 5%, then the 400 value was accepted. If however, the difference was greater than 5%, then a further 100 point counts was carried out. This process was repeated until the difference between values over a 100 point count differential was less than 5%.

1.7 Thesis organisation

The introductory chapter is followed by a chapter outlining the stratigraphic framework used in this study. Chapters 3-5 give detailed descriptions, interpretations and models for the sediments seen. They also discuss larger scale issues relevant to the study of similar sediments and processes. Chapter 6 documents the isotopic dating study carried out on biogenic carbonates within the area. Chapter 7 examines the structural evidence for the initiation and evolution of the three basins and discusses the various models put forward. The implications of the work described in previous chapters are summarised in chapter 8. The basins are compared with other Miocene basins within in the Mediterranean and placed in the wider context of the Eastern Mediterranean evolution. The major conclusions are then listed.

Chapters 3-5 describing the sediments of the three Miocene basins are not divided up on the basis of formations. Instead each of the chapters is written from the point of view of examining the fundamental processes involved. Thus, chapter 3 is written from the perspective of the role of biogenic components in carbonate deposition; chapter 4 is written with the mode of deposition of turbiditic sediments foremost in mind and
chapter 5 examines the sediments from the perspective of an interactive fan-delta system. As a result of this method of organisation a certain amount of overlap between chapters occurs. Porites bafflestones described in chapter 3 for instance are crucial in the interpretation of coastal fan-delta sediments discussed in chapter 5. When repetition of this kind is needed, a brief summary of the salient points is given and the reader is referred back to the original description for more detailed information if required. Figure 1.7 is a schematic representation of the overlap of the formations into the three chapters. Information concerning the previous work on the sediments is outlined at the beginning of the relevant chapter.

In justification of this organisational structure, it was felt that to describe the sediments purely on the basis of stratigraphic divisions would have superimposed an arbitrary framework on the study, not valid when considering interactive basinal processes. It is hoped therefore that the approach used will facilitate both the understanding of the integrated depositional and preservational systems active during the Miocene and subsequent publication.
Figure 1.7  Diagram showing the overlap of the formations within the three descriptive chapters. Note that the ornament used here for each formation is the same throughout the thesis.
Chapter 2

STRATIGRAPHY

2.1 Introduction

This chapter describes the time scale used throughout the study and points out the assumptions made in relying on biostratigraphic data. It goes on to introduce information published by previous workers on the Miocene stratigraphy of the area. The most comprehensive of these studies was undertaken by Akay and Uysal (1985) and the stratigraphy used in this thesis is based primarily on the one they produced. Revisions of some of the formation classification have been carried out however and these are described below. Figure 2.1 is a location map showing the position of places mentioned in the text.

2.2 Lithostratigraphy, biostratigraphy and isotopic dates

Figure 2.2 is a chart of the ages of Miocene nannoplankton zones with respect to the record of magnetic polarity from Berggren et al. (1985). Set against this is the time scale used by Haq et al. (1988) when constructing the global eustatic sea level curve. The Berggren time scale is used throughout this study, but as much of the discussion focuses on the role of eustacy in controlling sedimentation, comparison with the eustatic curve and other relative sea level curves which use the Haq time scale must be treated with a little caution. Minor discrepancies as a result of differences between the ages of stage boundaries occur, but these are thought to be insignificant in the context of the resolution of the curve itself.

New biostratigraphic information is listed in Appendix 3. Nannoplankton ages for sections measured during this study were carried out by Dr. C. Müller and can be found in Appendix 3a. This information was combined with data from previous studies of the Miocene sediments of this area (Bizon et al., 1974; Dumont, 1974; Poisson, 1977; Monod, 1977;
Figure 2.1. Location map of type sections mentioned in tables 2.2, 2.3, 2.4 and 2.6.
<table>
<thead>
<tr>
<th>AGE (Ma)</th>
<th>MAGNETIC POLARITY</th>
<th>NANNOPLANKTON ZONES</th>
<th>STAGE</th>
<th>SERIES</th>
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<tr>
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<td>5C</td>
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<td>5D</td>
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<td>5E</td>
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<td>15</td>
<td></td>
<td>6A</td>
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<td>16</td>
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<td>6B</td>
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<tr>
<td>26</td>
<td></td>
<td></td>
<td></td>
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</tr>
</tbody>
</table>

Figures 1.1 and 1.2 Biostratigraphic chart showing the correlation of nannoplankton zones with magnetostratigraphy (after Berggren et al., 1985). The stage boundaries used throughout this thesis are those of Berggren et al. (1985), but Haq et al.'s (1988) stratigraphic framework is also shown because it is against this that the eustatic sea level curve discussed in the text was constructed.
Akay and Uysal, 1985; Akay et al., 1985). Much, though not all of this focuses on foraminiferal biostratigraphy.

Lithostratigraphic correlation without the age framework provided here by nannoplankton and foraminiferal biostratigraphy and strontium isotope dating would have been both a difficult process and hard to justify. Facies vary laterally and it can be demonstrated that completely different depositional processes were active synchronously in different parts of the Isparta Angle. It is only through using well defined ages that secure correlation is possible and basin evolution clarified.

With this in mind it is important to point out that large parts of the successions studied here were deposited in continental environments and are very poorly dated as a result. The base of the Miocene sequence for instance, is only clearly dated in the few places where the basal sediments were deposited in a marine environment and it has been suggested that the thick successions of undated continental conglomerates in parts of the Köprü and Aksu basins may be Oligocene in age (Akay et al., 1985). It was felt that pinning down lithostratigraphic relationships and dating them where possible would be a more effective method of dealing with this gap in the data base than setting out to date the continental sediments directly given the time constraint of three years. Further work on this aspect of continental sediments in the Isparta Angle would enhance palaeogeographic reconstructions of both the Miocene and Plio-Quaternary.

One final note of caution concerning the stratigraphic framework used here concerns redeposition and will be considered again in chapter 4 where this information is most extensively used. The bulk of the samples found to contain nannoplankton or planktic foraminifera and used for dating, came from the Geceleme and Karpuzçay Formations. The possibility that the fine-grained fractions of these formations, sampled for the purpose of age reconstruction were redeposited by low-density turbidity flows cannot be excluded. Indeed, in the case of the Karpuzçay Formation it is extremely unlikely that even the finest material sampled here represents background sedimentation and there is clear evidence that at least parts of the Geceleme Formation are the
products of redeposition. Does this invalidate the biostratigraphic framework outlined below? It would be reasonable to expect a miscorrelation between the biostratigraphic results and the lithostratigraphy observed in the field if it did. In fact this very rarely happens suggesting that although redepositional processes are active, they are not sufficiently effective to significantly reorganise the stratigraphy above the level of biostratigraphic resolution. To obtain an independent view on this problem, Sr isotopes were measured both on in situ reef carbonates (Oymapinar limestone) and on foraminifera from the overlying Geceleme and Karpuzçay Formations. The sequential nature of the results suggests that reliance may be placed on biostratigraphic data.

2.3 Historical development of existing stratigraphic framework

By far the greatest debt is owed to Ergün Akay (see Akay and Uysal, 1985) who, in a team of French and Turkish geologists, first studied the Neogene sediments within the Isparta Angle as a whole, and set up the stratigraphy, which, though refined during this study, remains in place. Prior to his work many authors had dabbled in the Miocene stratigraphy particularly concentrating on the Manavgat basin partly because of its relative geological simplicity, but also due to the ease of access (e.g., Bizon et al., 1974). Blumenthal (1947) was the first to document Miocene stratigraphy in the area. He was followed in the 1970s by a wave of French geologists who mapped the entire Isparta Angle (Dumont, 1974; Marcoux, 1976; Poisson, 1977; Monod, 1977; Ricou 1979) and who triggered the debate over the Antalya Complex in the early eighties which has made the area famous (Robertson, 1993 for review). Although their work concentrated on the Mesozoic successions, all of them studied the overlying Neogene sediments in passing and between them they contributed greatly to the data base of information concerning Miocene sedimentation. Figure 2.3 is an outline map of the Isparta Angle showing the areas covered by the various workers who looked at some aspect of Miocene stratigraphy. Without the detailed knowledge and help of two of these people, Dr. O. Monod and particularly Dr. A. Poisson, this thesis would be much the poorer and less well understood by the author.

Previous work concerning each of the six formations described in this
Figure 2.3  Map of the Isparta Angle region showing the areas covered by previous studies who worked (at least partially) on Miocene sediments.
thesis are outlined at the beginning of the chapter in which they first feature. The minor complication of changing formation names is clarified in tables 2.1, 2.2, 2.3 and 2.4. More difficult to immediately master when perusing the literature are the changing names of the principal sections in the Manavgat basin. These are summarised in table 2.5 along with the names used in this study which relate to the geographical locality of the section and are aimed at reducing confusion.

2.4 Revised Stratigraphy

Figure 2.4 is the chronostratigraphic chart produced by Akay et al., (1985) summarising the relationships between the formations across the Isparta Angle. Much of this stratigraphic framework provided an excellent template on which to base a detailed assessment of the basin evolution through facies analysis. The following modifications have been made however and are illustrated in figure 2.5:

- The Aksu Conglomerate Formation which, in the Akay scheme occurs throughout the Miocene succession in all three basins (Fig. 2.4) has been split up into two main formations, the Aksu Formation and the Kizildag Formation (tables 2.4 and 2.1 respectively). The justification for doing this is based on the increase of biostratigraphic and lithostratigraphic resolution which revealed that there is little evidence for continuous conglomerate deposition throughout the Miocene. Instead, conglomerate successions appear to be concentrated in the Lower Miocene (Kizildag Formation) and Upper Miocene (Aksu Formation).

- These two formations have been further divided into a number of members reflecting the differing environments of deposition (table 2.6).

- One of these members has been drawn from the work of Monod (1972; 1977) on the northern margin of the Manavgat basin. He found and classified lacustrine limestone within Lower Miocene conglomerates (Kizildag Formation) and called it Calcaire de Kepez after a nearby village. This limestone was not documented by Akay, but has been reintroduced into the classification used for this study as a formation group of the Kizildag Formation (table 2.1).

- Akay and Uysal (1985) classified the coarse debris flow-dominated succession in
<table>
<thead>
<tr>
<th>Formation</th>
<th>Kızıldağ Formation</th>
<th>Age</th>
<th>Description</th>
<th>Upper contact</th>
<th>Lower contact</th>
<th>Type locality</th>
<th>Synonymy</th>
<th>Member</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kargı</td>
<td>Member</td>
<td>?</td>
<td>Clast-supported conglomerates and coarse, sometimes laminated, reddened sandstones, interbedded with calcite</td>
<td>Conformably overlain by Oymapınar limestone, Apparently conformably overlain by Karabulut Formation</td>
<td>Unconformable on Mesozoic basement</td>
<td>Kargı, south-west Akku basin</td>
<td>Akku Formation (Akku and Uysal, 1985; Akay et al., 1985)</td>
<td>Akku Member</td>
</tr>
<tr>
<td></td>
<td></td>
<td>?</td>
<td>Laminated, lacustrine limestone</td>
<td>Conformably overlain by both Kargı Group conglomerates and Oymapinar limestone</td>
<td>Unconformable on Mesozoic radiolarites, Conformable on Kargı Group conglomerates</td>
<td>Kepez, northern margin of the Manavgat basin</td>
<td>Calcaire de Calcaire de Kepez (Monod, 1977)</td>
<td>Kepez Member</td>
</tr>
<tr>
<td></td>
<td></td>
<td>?</td>
<td>High angle cross-bedded conglomerates, with interbedded Forites buffy stones or rhodolith-bearing grits</td>
<td>Overlain conformably either by Oymapinar limestone or Kargı Group conglomerates</td>
<td>Conformable on Kargı Group conglomerates</td>
<td>South Kargı, south centralAkku basin and Aspendos, south Kıpöri basin</td>
<td>Tepekli Formation (Monod, 1972, 1977)</td>
<td>Tepekli Member</td>
</tr>
</tbody>
</table>

Table 2.1 Chart of stratigraphic information for the Kızıldağ Formation
Table 2.2  Chart of stratigraphic information for the Oyrapinar limestone, Ĉakallar and Geceleme Formations

<table>
<thead>
<tr>
<th>Formation</th>
<th>Synonymy</th>
<th>Type locality</th>
<th>Lower contact</th>
<th>Upper contact</th>
<th>Description</th>
<th>Age</th>
</tr>
</thead>
<tbody>
<tr>
<td>Geceleme Formation</td>
<td>Geceleme Formation (Akay and Uysal, 1985; Akay et al., 1985); Gecereme Formation (Blumenthal, 1947; Monod, 1977)</td>
<td>Gençler, along the old road north from the coast to Akseki, Manavgat basin</td>
<td>Disconformity with Oyrapinar limestone and Ĉakallar Formation</td>
<td>Conformable and gradual transformation over 20m to Taskesigi Group or more rarely, the Karpuzçay Formation</td>
<td>Planktic foraminiferal marls and rare wedge shape calcirudites</td>
<td>Latest Burdigalian - Langhian</td>
</tr>
<tr>
<td>Ĉakallar Formation</td>
<td>Ĉakallar Formation (Akay and Uysal, 1985; Akay et al., 1985)</td>
<td>Ĉakallar, near Alarahan, south-east Manavgat basin</td>
<td>Conformable with Oyrapinar limestone</td>
<td>Disconformably overlain by Geceleme Formation or Karpuzçay Formation</td>
<td>Calcirudites interbedded with marls and sandstones. Blocky talus sometimes present</td>
<td>Latest Burdigalian - Langhian</td>
</tr>
<tr>
<td>Oyrapinar Limestone</td>
<td>Oyrapinar limestone (Akay and Uysal, 1985; Akay et al., 1985); Calcaire d'Oyrapinar (Monod, 1972)</td>
<td>Oyrapinar, northern margin of the Manavgat basin</td>
<td>Unconformable on Alanya Massif along the north-eastern margin of the Manavgat basin. Conformable on Kizildag Formation elsewhere</td>
<td>Conformably overlain by Ĉakallar Formation locally. Sometimes small angular disconformity below Karpuzçay Formation and Geceleme Formation</td>
<td>Shallow-water biogenic carbonates including coral and algal reefs.</td>
<td>Burdigalian-Langhian. Diachronous from south to north in the Manavgat basin</td>
</tr>
<tr>
<td>Formation</td>
<td>Member</td>
<td>Synonymy</td>
<td>Type locality</td>
<td>Lower contact</td>
<td>Upper contact</td>
<td>Description</td>
</tr>
<tr>
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</tr>
<tr>
<td>Karpuzçay</td>
<td>Beskonak</td>
<td>Karpuzçay Formation (Akay and Uysal, 1985; Akay et al., 1985); Beskonak Formation (Degirmenci, 1992)</td>
<td>North of Beskonak, central Köprü basin</td>
<td>Angular disconformity associated with sections of faulted Oymapinar limestone, e.g. Deniztepesi. Otherwise apparently conformable on Kizildag Formation</td>
<td>?Conformably overlain by Aksu Formation, only rarely seen. Unconformably overlain by Pliocene sediments in the south</td>
<td>Sandstone-siltstone turbidites. Rare conglomerate channels</td>
</tr>
<tr>
<td>Taskesigi</td>
<td>Member</td>
<td>Along the road from the coast to Akseki, near Taskesigi, Manavgat basin</td>
<td>Conformable on Geceleme Formation</td>
<td>Unconformably overlain by Lower Pliocene sediments</td>
<td>Debris flows containing large detached blocks, conglomerates, sandstones and slumped silts</td>
<td>Tortonian-Middle Messinian</td>
</tr>
<tr>
<td>Formation</td>
<td>Member</td>
<td>Synonymy</td>
<td>Type locality</td>
<td>Lower contact</td>
<td>Upper contact</td>
<td>Description</td>
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<td>---------------------------------------------------</td>
<td>--------------------------------------------------</td>
</tr>
<tr>
<td>Aksu</td>
<td>Kapakaya</td>
<td>Aksu Conglomerate (Akay and Uysal, 1985; Akay et al., 1985)</td>
<td>Kapakaya, far north of Aksu basin</td>
<td>Unconformable on Mesozoic basement.</td>
<td>Probably conformable passing up into Kesme Group</td>
<td>Coarse, reddened conglomerates and sandstones. Local evidence of fault generation</td>
</tr>
<tr>
<td>Kesme</td>
<td>Member</td>
<td>Conglomérats de Kesme (Dumont, 1974)</td>
<td>Kesme, far north-east of Köprü basin</td>
<td>Unconformable on Mesozoic basement.</td>
<td>Not seen</td>
<td>Clast-supported conglomerates and fossiliferous, cross-bedded sandstones.</td>
</tr>
</tbody>
</table>
Figure 2.4  Composite stratigraphic framework of the Neogene sediments of the Isparta Angle from Akay et al. (1985). Note that the Karabayır and Karakus Formations outcrop in the flexural basin to the north-west of the Aksu basin. These formations are not discussed here.
Table 2.5  List of sections used by previous workers which have been renamed to avoid confusion in the present study.

<table>
<thead>
<tr>
<th>Name of Author</th>
<th>Section name</th>
<th>Section name in this study</th>
</tr>
</thead>
<tbody>
<tr>
<td>Monod (1977), Bizon et al. (1974)</td>
<td>Manavgat section</td>
<td>Oymapinar section</td>
</tr>
<tr>
<td>Monod (1977)</td>
<td>Alara Çay section</td>
<td>Alarahan section</td>
</tr>
<tr>
<td>Akay and Uysal (1985), Akay et al. (1985)</td>
<td>Manavgat section</td>
<td>Ahmetler section</td>
</tr>
</tbody>
</table>

The Tortonian-Messinian of the Manavgat basin as Karpuzçay Formation. This succession is completely different from any other exposure of his Karpuzçay Formation in which horizons coarser than sandstone are rare. More detailed field study of the debris-flow dominated sequence revealed that it is laterally discontinuous even within the Manavgat basin, although definition of its margins is difficult due to the similarity of the finer-grained components to the "standard" Karpuzçay Formation. The coarse horizons by which this sequence can be recognised appear to be concentrated in a narrow north-south zone ~10km across. This zone has been loosely termed the Akseki corridor in this study because the succession is so well exposed along the new road to Akseki from the coast. The Karpuzçay Formation has been divided into two groups (table 2.3) to distinguish between the debris-flow dominated succession (Taskesigi group), and the finer-grained turbidites (Beskonak group). This re-classification has implications for the origin of the Akseki corridor and these are discussed in chapters 4 and 7.

General definitions, descriptions and ages of the formations used in this study of the Miocene sediment of the Isparta Angle are given in tables 2.1, 2.2, 2.3 and 2.4. A modified stratigraphic framework is shown in figure 2.5 to accommodate the revised formation classification and more detailed biostratigraphic and lithostratigraphic study.
Table 2.6  Re-classification of the Aksu Conglomerate Formation and Karpuzçay Formation (Akay and Uysal, 1985; Akay et al., 1985) and the division into members used in this study.

<table>
<thead>
<tr>
<th>Environment</th>
<th>Sub-environment</th>
<th>Karpuzçay Formation</th>
<th>Aksu Formation</th>
<th>Kizildag Formation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Continental</td>
<td>Alluvial</td>
<td></td>
<td>Kapikaya</td>
<td>Kargi</td>
</tr>
<tr>
<td></td>
<td>Lagoonal and Lacustrine</td>
<td></td>
<td></td>
<td>Calcaire de Kepez</td>
</tr>
<tr>
<td>Marine</td>
<td>Shoreline</td>
<td></td>
<td>Kesme</td>
<td>Tepekli</td>
</tr>
<tr>
<td></td>
<td>Shelf</td>
<td>Taskesigi Beskonak</td>
<td></td>
<td></td>
</tr>
<tr>
<td>AGE (Ma)</td>
<td>SERIES AND STAGES</td>
<td>COMPOSITE STRATIGRAPHIC RELATIONS</td>
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<td></td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>PLEISTOCENE</td>
<td>NW</td>
<td></td>
<td></td>
</tr>
<tr>
<td>2</td>
<td>MESSINIAN</td>
<td>North Aksu basin</td>
<td></td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>TORTONIAN</td>
<td>Central Köprü basin</td>
<td></td>
<td></td>
</tr>
<tr>
<td>4</td>
<td>SERRAVALLIAN</td>
<td>Manavgat basin</td>
<td></td>
<td></td>
</tr>
<tr>
<td>5</td>
<td>LANGHIAN</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>6</td>
<td>BURDIGALIAN</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>7</td>
<td>AQUITANIAN</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

**Legend**

- Aksu Formation
- Karpuzçay Formation
- Beskonak Member
- Taskesigi Member
- Geceleme Formation
- Çakallar Formation
- Oymapinar limestone
- Kizildag Formation

**Figure 2.5** Modified stratigraphic framework (after Akay et al., 1985) used during this study.
Chapter 3

BIOGENIC CARBONATES

3.1 Context

Important information about the early stages of the sedimentary and tectonic evolution of the Aksu, Köprü and Manavgat basins is derived from biogenic carbonates. These occur at two stratigraphic levels in the Miocene: Lower-Mid Miocene and Tortonian. Crucially, the Lower Miocene carbonates occur at, or just above the basal unconformity and mark a basin-wide relative sea level rise.

3.2 Geographical distribution

Biogenic carbonates, primarily coral and algal-rich reefs, calcarenites and planktic foraminiferal marls, are found throughout the study area (Fig. 3.1). The distribution of shallow water carbonates is uneven however, with concentrations of reef and reef related material occurring in the central and eastern parts of the Manavgat basin and more sparsely along the western and northern margins of both the Köprü and Aksu basins.

3.3 Organisation of the chapter

After a brief outline of the information gained from previous work on the area (section 3.4) the assumptions on which the classification used in this thesis for the biogenic carbonates is stated and the rocks divided into facies groups and sub-divided into sub-facies (section 3.5). The facies groups are dealt with in turn (section 3.6) with each sub-facies described and interpreted in terms of its dimensions, content, matrix and cement. A summary chart is given at the end of each facies group and a model of the environments of sub-facies deposition given in section 3.7. Broader topics related to the biogenic carbonates in the study area as a whole, are discussed at the end of the chapter (section 3.8) and the conclusions are listed in section 3.9. Diagenetic alteration of biogenic components is
Figure 3.1. Location map of the biogenic carbonate sections referred to in the text.
touched on here, but is discussed more fully in chapter 6.

### 3.4 Previous work

Table 3.1 displays general information about the principal formations discussed in this chapter: the Oymapinar Limestone; the Çakallar Formation and the Geceleme Formation. The ages of these formations have been deduced from their stratigraphic position and age diagnostic micro-fossil assemblages (Bizon et al., 1974; Dumont, 1974; Poisson, 1977; Monod, 1977; Akay and Uysal, 1985; Akay et al., 1985; Flecker et al., 1995; this study). They occur in the Lower-Mid Miocene (Burdigalian-Langhian) and are predominantly exposed in the Manavgat basin and in the south of the Aksu and Köprü basins. Rare lenses of reef limestone can be found in the north of the Aksu and Köprü basins and these are thought to be Tortonian in age (Akay and Uysal, 1985; Dumont, 1974). Akay classified these Upper Miocene reefs as part of the Aksu Formation. During this study, the Aksu Formation was subdivided and the Tortonian reef limestone lenses are now classified as part of the Kesme Group (Table 2.4).

The Oymapinar Limestone in the Manavgat basin has an upper age limit of Upper Burdigalian to Mid-Langhian as defined by the nannoplankton (Akay et al., 1985 and Flecker et al. 1995) and foraminiferal (Akay 1985) ages of the overlying marls. Problems with dating the reefs themselves particularly in the Manavgat basin have dogged attempts to define an age for the lower boundary, since no foraminifera have been found within the shallow water carbonates that denotes an age any more precise than Lower Miocene. Blumenthal (1947) first recognised and dated these carbonates as Lower Miocene and they were subsequently correlated with the Aquitanian Karabayir Formation (Poisson and Poignant, 1974) of the Dariören-Kas basin to the west (Fig. 1.3). Monod (1972) first named the limestones calcaire d’Oymapinar. Using the presence of *Praeorbulina glomerosa* found in a marl towards the top of the Oymapinar section, Monod (1977) indicates a Burdigalian to Langhian age, but suggests that the absence of benthic foraminifera characteristic of the Burdigalian such as *Miogypsina* and *Lepidocyclina* indicate that the Oymapinar Limestone is more likely to be Langhian. Bizon *et al.* (1974) date a number of
### 3.1 Chart of stratigraphic information for the Oymapinar limestone, Çakallar and Geceleme Formations

<table>
<thead>
<tr>
<th>Formation</th>
<th>Synonymy</th>
<th>Type locality</th>
<th>Lower contact</th>
<th>Upper contact</th>
<th>Description</th>
<th>Age</th>
</tr>
</thead>
<tbody>
<tr>
<td>Geceleme Formation</td>
<td>Geceleme Formation (Akay &amp; Uysal, 1985; Akay et al., 1985); Geceleme Formation (Blumenthal, 1947; Monod, 1977)</td>
<td>Gençler, along the old road north from the coast to Akseki, Manavgat basin</td>
<td>Disconformity with Oymapinar limestone and Çakallar Formation</td>
<td>Conformable and gradual transformation over 20m to Taskesigi Group or more rarely, the Karpuzçay Formation</td>
<td>Planktic foraminiferal marls and rare wedge shape calcirudites</td>
<td>Latest Burdigalian ?-Langhian</td>
</tr>
<tr>
<td>Çakallar Formation</td>
<td>Çakallar Formation (Akay &amp; Uysal, 1985; Akay et al., 1985)</td>
<td>Çakallar, near Alarahan, south-east Manavgat basin</td>
<td>Conformable with Oymapinar limestone</td>
<td>Disconformably overlain by Geceleme Formation or Karpuzçay Formation</td>
<td>Calcirudites interbedded with marls and sandstones. Blocky talus sometimes present</td>
<td>Latest Burdigalian ?- Langhian</td>
</tr>
<tr>
<td>Oymapinar Limestone</td>
<td>Oymapinar limestone (Akay &amp; Uysal, 1985; Akay et al., 1985); Calcaire d'Oymapinar (Monod, 1972)</td>
<td>Oymapinar, northern margin of the Manavgat basin</td>
<td>Unconformable on Alanya Massif along the north-eastern margin of the Manavgat basin. Conformable on Kizildag Formation elsewhere</td>
<td>Conformably overlain by Çakallar Formation locally. Sometimes small angular disconformity below Karpuzçay Formation and Geceleme Formation</td>
<td>Shallow-water biogenic carbonates including coral and algal reefs.</td>
<td>Burdigalian-Langhian. Diachronous from south to north in the Manavgat basin</td>
</tr>
</tbody>
</table>
Tertiary sections in south coastal Turkey using planktic foraminifera. The basal marls in the Ahmetler section (Table 2.5) are dated as being Mid-Langhian while those of the Alarahan section to the south are Lower Langhian. Akay and Uysal (1985) also proposed a Langhian age for the Oymapinar Limestone, but more recent work on the nannoplankton in the Ahmetler section (Flecker et al. 1995) indicates that the lowest horizons of the overlying marls contain genera specific to the Upper Burdigalian in an assemblage which also contains Praeorbulina glomerosa, the planktic foraminifera diagnostic of the Burdigalian Langhian boundary (Berggren et al., 1985). This indicates that although the Burdigalian fauna must be reworked, their presence suggests that some of the Oymapinar Limestone was deposited in the Burdigalian. In an attempt to resolve the issue of whether Burdigalian limestone is preserved, independent strontium isotopic dates on the reef limestone were obtained. The results of this work are reported in Chapter 6.

The ages of the Çakallar and Geceleme Formations were also defined by Akay and Uysal (1985) using both planktic foraminifera and nannoplankton analysis, as Langhian and Langhian-Serravallian respectively. A summary of the different interpretations of the Lower Miocene carbonate stratigraphy in the area is given in table 3.2. Note that authors prior to Akay and Uysal (1985) did not document the presence of the Çakallar Formation.

3.5 Classification

A rock is classified as a carbonate if it contains more than 50% carbonate minerals. In the Anatalya region, however, classification of many carbonate-rich rocks is complicated by the huge volumes of older (generally Mesozoic) carbonate that are eroded from the surrounding basement hinterland and redeposited during the Miocene. Sticking to the literal definition of a carbonate rock, my entire thesis would have to go into this chapter as nearly every sandstone, siltstone and conglomerate is over 50% carbonate, not including the cement which is ubiquitously carbonate in either micritic or sparite form (Folk, 1959; 1962). From the point of view of looking at the dominant processes involved in the deposition of these rocks it makes little sense to lump the carbonate
Table 3.2  Table showing the various names and inferred ages of the biogenic carbonate succession in the Manavgat basin. For each box, the south of the basin is at the bottom and the north at the top. Thus a south-north diachronous transgression is suggested as part of this study.

<table>
<thead>
<tr>
<th>AGE</th>
<th>Aquitanian</th>
<th>Burdigalian</th>
<th>Langhian</th>
<th>Serravallian</th>
</tr>
</thead>
<tbody>
<tr>
<td>Blumenthal</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(1947)</td>
<td><img src="image1" alt="Diagram" /></td>
<td><img src="image2" alt="Diagram" /></td>
<td></td>
<td><img src="image3" alt="Diagram" /></td>
</tr>
<tr>
<td>Monod</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(1972)</td>
<td><img src="image4" alt="Diagram" /></td>
<td><img src="image5" alt="Diagram" /></td>
<td></td>
<td><img src="image6" alt="Diagram" /></td>
</tr>
<tr>
<td>Monod</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(1977)</td>
<td><img src="image7" alt="Diagram" /></td>
<td><img src="image8" alt="Diagram" /></td>
<td></td>
<td><img src="image9" alt="Diagram" /></td>
</tr>
<tr>
<td>Akay (1985)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>and Akay et al. (1985)</td>
<td><img src="image10" alt="Diagram" /></td>
<td><img src="image11" alt="Diagram" /></td>
<td></td>
<td><img src="image12" alt="Diagram" /></td>
</tr>
<tr>
<td>This study</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td><img src="image13" alt="Diagram" /></td>
<td><img src="image14" alt="Diagram" /></td>
<td><img src="image15" alt="Diagram" /></td>
<td></td>
<td><img src="image16" alt="Diagram" /></td>
</tr>
</tbody>
</table>
grains derived from the erosion of the basement together with those carbonate grains derived from Miocene carbonate-forming environments. Thus, where possible the two have been treated separately and the basement carbonate grains referred to as terrestrial detritus along with quartz and other lithic fragments e.g. mica, chert, pyroxene, opaques, feldspar, chlorite and igneous fragments. In practical terms in thin section this is relatively straightforward since the bulk of the basement carbonate grains contain abundant, small speckely inclusions, appearing dirty in comparison to younger carbonate grains and they frequently display deformation twinning under cross polars (Fig. 3.2).

The formation names and broad definitions of Akay et al. (1985) are retained in this study, but independent of these, the biogenic carbonates are divided into four broad facies groups: reef facies; off-reef facies; reef-associated carbonate facies and shelf carbonate facies (see Table 3.3).

**Reef facies** are distinguished from off-reef facies by the presence of an *in situ* coral (or rarely algal) framestone. In general terms, framework coral or algae bind together both biogenic material and clastic sediment.

**Off-reef facies** include rocks which derive the bulk of their constituent parts from the reef framestone e.g. redeposited coral rudstones. Prolific quantities of off-reef facies are found in a continuous strip parallel to the northern margin of the Manavgat basin.

Rocks containing *in situ* or redeposited bioclasts, which are not derived from the reef framework are classified as **reef-associated carbonate facies** e.g. benthic foraminiferal packstones. The bulk of these rocks are found in close spatial association with reef or off-reef accumulations.

Biogenic carbonate rocks showing little influence of shallow water carbonate production (i.e. containing <50% shallow-water detritus) are classified as **shelf carbonate facies** e.g. planktic foraminiferal wackestones.

In this study these four facies groups have been further subdivided into 14 sub-facies (see Table 3.3). These sub-facies transgress the boundaries of
Figure 3.2  Thin section photo of speckly and deformed carbonate grains in a heterogeneous sandstone.
Table 3.3 Chart showing the overlap between Akay and Uysal’s carbonate Formations and the facies and sub-facies used in this chapter.

<table>
<thead>
<tr>
<th>Akay and Uysal (1985) Terms</th>
<th>Reef Facies</th>
<th>Off-reef Facies</th>
<th>Reef-associated Facies</th>
<th>Shelf Facies</th>
</tr>
</thead>
<tbody>
<tr>
<td>Reef lenses in the Aksu Formation</td>
<td>Domal coral framestone</td>
<td></td>
<td>Calcareous</td>
<td>Echinoid-scaphopod grainstone</td>
</tr>
<tr>
<td></td>
<td>Porites baffiestone</td>
<td></td>
<td>Coral floatstone</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Fan coral framestone</td>
<td></td>
<td>Oyster packstone</td>
<td></td>
</tr>
<tr>
<td>Oymapinar Limestone</td>
<td></td>
<td></td>
<td>Gastropod wackestone</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Echinoid-scaphopod grainstone</td>
<td></td>
</tr>
<tr>
<td>Çakallar Formation</td>
<td></td>
<td></td>
<td>Reef talus</td>
<td></td>
</tr>
<tr>
<td>Geceieme Formation</td>
<td></td>
<td></td>
<td>Planktic foraminiferous marl</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Calcritoids</td>
<td></td>
</tr>
</tbody>
</table>
both formations and facies. Sub-facies have been classified using Dunham's (1962) classification scheme and including the boundstone subdivision suggested by Embry and Klovan (1971).

### 3.6 Facies description

#### 3.6.1 Reef facies

Table 3.4 Table of reef sub-facies, listing their characteristic fauna and the localities at which they can be found (Fig. 3.1).

<table>
<thead>
<tr>
<th>Sub-facies</th>
<th>Characteristic fauna</th>
<th>Localities</th>
</tr>
</thead>
<tbody>
<tr>
<td>Domal coral framestone</td>
<td><em>Porites</em></td>
<td>South Yesilbag</td>
</tr>
<tr>
<td></td>
<td><em>Montastrea</em></td>
<td>Alarahan</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Kasimlar-Kesme</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Saburlar</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Oymapinar</td>
</tr>
<tr>
<td>Porites bafflestone</td>
<td><em>Porites</em></td>
<td>South Kargi</td>
</tr>
<tr>
<td></td>
<td></td>
<td>South Yesilbag</td>
</tr>
<tr>
<td>Fan coral framestone</td>
<td><em>Tarbellastrea</em></td>
<td>South Kargi</td>
</tr>
<tr>
<td></td>
<td><em>Porites</em></td>
<td>North Yesilbag</td>
</tr>
<tr>
<td>Porites rudstone</td>
<td><em>Porites</em></td>
<td>South Kargi</td>
</tr>
<tr>
<td></td>
<td>algae</td>
<td>South Yesilbag</td>
</tr>
<tr>
<td></td>
<td>shell fragments</td>
<td>Altinkaya</td>
</tr>
</tbody>
</table>

The study of modern reefs has revealed that coral type, location and morphology are controlled by a myriad of environmental factors such as salinity, temperature, nutrient supply, light, turbulence, relative sea level change as well as substrate type and sediment flux. In the rock record, preservation of such controls is minimal and most environmental factors can only be tentatively inferred. Substrate and sediment type, because of their high preservation potential, in the past have been credited with controlling coral type, morphology and location in the interpretation of fossil reefs. Although some studies of modern reefs do indicate that such factors can control reef composition, (e.g. Fagerstrom, 1987 in the Caribbean and Acredo *et al.*, 1989) many authors now think that other environmental controls are generally more important as well.
as controls such as species replacement (Barnes and Hughes, 1982). It is therefore extremely difficult to draw any firm conclusions about what controlled the location, species and morphology of fossil corals. Bearing this in mind however, in describing the reef facies, associations between fossils and the sediment have been noted and a few, very tentative conclusions drawn. The implications of the interaction between coarse clastic environments and reef carbonate growth are discussed more fully in chapter 5.

3.6.1.1 Domal coral framestone

*Dimensions and sediment associations*

Domal coral framestones are found all over the study area, but constitute a small percentage of total reef exposure. Framestone horizons are generally 2-3m thick. Laterally, where exposed, units either wedge out on to basement (e.g. Ahmetler) or conglomerate highs over 12 to 20m (e.g. Alarahan) or exhibit a gradual transition to calcarenite (e.g. Kasimlar-Kesme reef). Vertical transitions tend to be abrupt as almost all domal coral framestones are overlain by coarse conglomerates with erosive bases. An example of this can be see just south of Yesilbag. Here, a pronounced channel structure 2m in depth divides two outcrops of domal coral framestone (Fig. 3.3). The corals do not terminate at the channel margins, but appear to have grown out into the main channel conduit which is now filled with siltstones, mudstones and reef debris. The framestone on the Kasimlar-Kesme road by contrast is overlain by a fine grained conglomerate. Parts of the irregular top to the domal coral surface can be seen protruding up into this conglomerate. At Saburlar, marls that overlie the domal coral framestone horizon have been subsequently eroded, exposing an undulating top surface of the harder domal corals beneath.

*Corals*

The domal corals found in this facies can be up to 20cm in diameter and, at Saburlar, they protrude 5-10cm from the matrix. Domal coral species varies between localities, but by far the most dominant coral type is *Porites* *sp.* Poritid morphology is variable. The following main types of morphology occur commonly in reef frameworks:
Figure 3.3  Schematic sketch of the reef at South Yesilbag. Note the increase of stick Porites and correlative decrease in domal corals with increasing clastic content. See also the two different sorts of bafflestone matrix visible at this locality, conglomerate and fine-grained silt. The silt bafflestone appears to be confined within a channel of finer grained material, which from its relationship with the coral framestones and the overlying clastics suggests that the coral growth influence the structure of a coeval channel for fine grained sediments.
- **Domal** - approximating hemispherical shape, 5-10cm across;
- **Tabular** - flattened or dish-shaped, 5-20cm;
- **Sticks** - rods, sometimes branching when in situ, length 2-15cm, diameter 2-5cm.

There are four individual coral-framestone horizons exposed in the Alarahan section (Fig. 3.4). In the lowest of these (reef 1) at the base of the framestone, domal *Porites* corals are interspersed with tabular *Porites* colonies. The framestone passes up over 2m into a reef talus facies containing disorientated fragments of domal *Porites* and rare *Tarbellastrea* and *Porites* stick coral. Stick *Porites* dominate the base of reef 2 with *in situ* domal corals (again dominantly *Porites*) distributed amongst them higher in the section. This reef is abruptly terminated by an erosion surface and overlain by a channelised conglomerate. Reef 3 is similar to reef 2 in that stick *Porites* coral dominates at the base. Towards the top of this reef however domal *Porites* corals become much more prolific and are the dominant morphology. The top of the reef is not exposed. Just below the base of the highest reef (reef 4), a detached block of domal coral framework limestone a metre in length, is seen suspended within cross-bedded conglomerates. The base of the reef itself contains concentrated domal corals. Stubby stick *Porites* are also abundant and become more dominant up section, with domal corals dispersed amongst abundant stick *Porites*.

The Kesme reef displays the greatest diversity of *in situ* coral found within the study area. It is dominated by large, ridged, spreading structures of *Montastrea* (Fig. 3.5), but it also contains tabular and domal *Porites* and *Tarbellastrea*. One sample of the *Montastrea* was submitted for x-ray diffraction analysis and this indicated that all coral in the sample had been completely converted from primary aragonite to low magnesium calcite.

**Matrix**

At Saburlar and Alarahan (Fig. 3.1) the abundant framework interstices are infilled with a coarse, poorly sorted bioclastic matrix rich in echinoid spines and body fossils, abundant large algal fragments, brachiopods and the large benthic foraminifera *Onculina*. Non-bioclastic fragments
Figure 3.4 Alarahan section from the south-east of the Manavgat basin, showing the different vertical stratigaphies in the 4 reefs. Note also that the onset of deposition of planktic foraminiferal marls marks the termination both of reefs and the coarse clastic sedimentation.
Figure 3.5  Photograph of the irregular massive *Montastera* build-up at Kesme.

Figure 3.6  Photograph of spaced *Porites* bafflestone with interstitial micro-conglomerate.
include red chert, sandstone, carbonate grains containing abundant opaques. At Kesme (Fig. 3.1), because of the more massive framework structure, there is little matrix and the colonies of *Montastrea* appear to be largely supported by the close packed nature of neighbouring corals. What matrix there is however is dominantly micrite with miliolids and rare quartz (Dumont, 1974). Dark-grey micro-conglomeratic matrix infills the open framework structure at Yesilbag making up 20-30% of the rock.

**Ages**
The Saburlar and Alarahan sections have a Langhian age inferred from the Langhian nannoplankton and planktic foraminifera found in the overlying marls (Akay 1985; Akay et al., 1985). The Yesilbag reef lies close to the boundary between the Serravallian Karpuzçay Formation (Akay and Uysal, 1985) and the overlying Tortonian Aksu Formation (Dumont, 1974). This reef and the Kasimlar-Kesme reef to the north are considered to be Tortonian because they are interbedded in the Aksu Formation.

**Interpretation**
The Miliolids found in the matrix in the Kasimlar-Kesme reef indicate a low energy environment of deposition (Martin et al. 1989). The structure of this *Montastrea* reef however is very different from that seen elsewhere. Where the abundance and grainsize of the matrix is relatively coarse, this may suggest a relatively high-energy environment of deposition, e.g. strong currents causing break-up and reworking of reef components.

The high terrigenous content of the matrix and the conglomerate channel structures both in and above the domal-coral framestone at Yesilbag suggest that sediment influx, probably from a local fluvial source, was also active. The non-erosive nature of the margins of this channel within the framestone may indicate that coral growth patterns influenced the finer grained sediment pathways. The interaction of coral growth in a clastic environment is discussed more fully in section 5.6.3.2.

Despite displaying the highest diversity of *in situ* coral in the study area, the Kasimlar-Kesme reef is still low diversity in comparison to the coral diversity seen in the Ziqlag Formation in Israel (Esteban 1979) which is
also of Tortonian age. Follows (1990) however, describes a similar
diversity restriction of Tortonian corals (Montastrea, Tarbellastrea and
Porites) in the Koronia Member in Cyprus. The possible reasons for this
restriction are discussed further in section 3.8.2.

3.6.1.2 Porites bafflestone

Dimensions and sediment associations
Porites bafflestone horizons identified in the study area are 2-3m thick.
At South Kargi (Fig. 3.1), differing concentrations of stick Porites occur in
stacked bafflestone beds such that the total thickness of the bafflestones is
over 6.5m. The beds cannot be traced out laterally, but are absent from a
tunnel cutting section 100m to the east (Fig. 3.7). Sediment between coral
sticks is coarse sand to micro-conglomerate grade. At South Kargi the
uppermost bafflestone bed is abruptly terminated by an erosion surface.

There are two distinct bafflestones in the reef just south of Yesilbag: one
with a medium-grained conglomerate matrix and the other with a
muddy matrix. The latter of these is located in the upper part of the
channel structure (Fig. 3.3), wedging out against the channel margin to
the east and passing laterally into a silty coral floatstone to the west.

Coral
The bafflestones studied here are all exclusively Porites. The dominant
Porites morphology is a stick-like form 2-3cm in diameter, up to 10cm in
length and variously spaced from other sticks (Fig. 3.6). The orientation
of these sticks is vertical to sub-vertical in the Kargi and Yesilbag sections
where conglomerate or coarse micro-conglomerate is the associated
matrix. More horizontal sticks are seen in the silt matrix bafflestone at
Yesilbag. Coral concentration in the bafflestones varies from 30-70%.

Matrix
Throughout the bafflestones at Kargi, changes in grainsize and
composition of the matrix are clearly visible as bedding. Coral sticks pass
undisturbed through these boundaries. The matrix here, is made of
angular fragments of red, radiolarian chert and limestone as well as
biogenic fragments including algal debris and echinoid spines. The
Figure 3.7 Tentative correlation of the logged sections at South Kargi, central Aksu basin, and palaeocurrent data measured from the cross-bedded conglomerates. Note the rapid lateral facies changes in reef framestones.
grainsize is dominantly silt to fine sand, but it increases to micro-
conglomerate grade towards the top of the lowermost Porites bafflestone
bed (Fig. 3.6 and 3.7). At Yesilbag both the matrix types (conglomerate and
siltstone) have similar compositions to the matrix at South Kargi. In the
conglomerate the larger limestone clasts are well rounded and algal
debris is abundant. This contrasts with the angularity of the micro-
conglomeratic matrix at South Kargi.

Interpretation

Martin et al. (1989) found a close association of Porites coral with siltstone
in the Tortonian reefs of south east Spain and it is possible that sediment
flux is one of the factors controlling the exclusive colonisation of stick
Porites in these bafflestones. As well as being tolerant to environments
of high terrestrial sedimentation, Porites has also been observed to
colonise areas of low salinity, low temperature and reduced water
circulation in modern reefs (Marshall and Orr 1931, Manton 1935, Wells
1954 and Scoffin and Stoddart 1978). Thus the factor controlling Porites
dominance may be low salinity associated with a high fluvial input
rather than the sediment itself, or a combination of the two. Equally
likely however, is that the matrix infilled the spacious coral framework
after the coral had grown. Evidence supporting this is the observation
that individual stick corals appear to pass through lithological
boundaries. In this case, it would be difficult to argue that sediment
influx and low salinity alone were dominant factors controlling coral
type. The bafflestone located within the channel structure at Yesilbag
may be an exception to this.

The correlation between the orientation of the Porites sticks and the
coarseness of the grainsize may be explained by the stability of the
substrate. Martin et al. (1989) observed in south east Spain, that many of
the Porites colonies embedded in silt appeared to have toppled over
having reached a certain size (see section 3.6.1.3). The coarser packing of
the micro-conglomeratic matrix at Yesilbag and South Kargi may have
better supported the Porites than the finer, more mobile silt observed at
Yesilbag where more horizontally orientated coral sticks were found.
Given that the silt bafflestone occurs within a channel structure however,
it is also possible that some of the horizontal *Porites* sticks may be redeposited.

3.6.1.3 Fan coral framestone

*Dimensions and sediment associations*

The fan coral framestone horizons found at South Kargi (Fig. 3.1) and to the north of Yesilbag are all between 4 and 6m thick. Variations in thickness occur along a single horizon at South Kargi where the lower framestone is at least a metre thicker in the tunnel cutting than it is 100m to the west in the main South Kargi section (Fig. 3.7). The tunnel cutting also exposes another fan coral horizon a few metres higher up the section which is absent to the west. This horizon has a wedge-shaped geometry which passes laterally into a Poritid rudstone containing abundant tabular *Porites*, both in situ and redeposited (Fig. 3.8). In South Kargi all the fan coral framestones overlie *Porites* rudstone horizons. They pass upwards into sand and micro-conglomerate rich *Porites* bafflestones with a rather abrupt transition, or are terminated by an erosion surface.

Fan coral framestones along the track north of Yesilbag (Fig. 3.1) are interbedded with sandstones, conglomerates, and siltstones. The framestones are occasionally observed passing into a more spaced *Porites* bafflestone. Poor exposure does not enable lateral relationships to be observed.

*Corals*

There are two distinct types of fan coral framestone: one composed entirely of *Tarbellastrea* as can be seen at South Kargi and the other exclusively *Porites*. The latter is observed along the track to Kesme, north of Yesilbag (Fig. 3.1).

The *Tarbellastrea* colonies at South Kargi are cone shaped. They are composed of bunches of sticks spreading upwards and outwards and at the same time thickening so that there is virtually no space between individual sticks. Colonies can reach over 3m in height and appear to compete for space with neighbours (Fig. 3.9). In the South Kargi section, well exposed *Tarbellastrea* colonies all appear to grow from a single level.
Figure 3.8 Photomontage and interpretive sketch of the patch-reef and Kargi Group conglomerates in the tunnel section at South Kargi, central Aksu basin. Figure 3.7 shows a log of this section correlated to the main South Kargi section to the west.
Figure 3.9  Photograph of Tarbellastrea fan coral framestone.

Figure 3.10  Photograph of stick Porites framestone at Yesilbag
at the top of the rudstone horizon (Fig. 3.7). 100m to the east however, the tunnel cutting through the same section reveals a series of *Tarbellastrea* cones on top of one another.

The *Porites* fan corals are similar in morphology to the *Tarbellastrea* colonies, but they are less closely packed and much smaller, reaching a maximum vertical height of only 15-20cm. In contrast to the *Tarbellastrea* framework, where there is little displaced coral, *in situ* *Porites* colonies are in the minority. They sit within a dense mêlée of disoriented colonies and single *Porites* sticks (Fig. 3.10).

**Matrix**

As a result of the close packed structure of the *Tarbellastrea* framework, the amount of sediment contained within it is minimal. The *Porites* colonies have more space between individual sticks varying between 5-15% silt to sand grade matrix.

**Interpretation**

According to Martin *et al.* (1989), the degree to which the *Tarbellastrea* cone structures open out is related to water depth i.e. the deeper the water the tighter and narrower the cone structure. Braga and Martin (1988) note micro atoll shapes covered in iron oxides preserved at the top of colonies which reach sea level. These features are not developed at South Kargi suggesting that relative sea level rise was rapid enough to prevent corals from reaching sea level. Furthermore, the *Tarbellastrea* cones are relatively narrow, supporting a deeper water interpretation. The *Porites* fan coral framestones by contrast have a much wider cone geometry and although there is no clear evidence of micro atoll structure development they may have developed much closer to sea level.

The average growth rate estimated for modern reef corals is 7-8m/1000 yrs (Geister, 1983 and Johnson *et al.*, 1986). Thus, even the individual Tarbellastrid colonies in the main South Kargi section which are over 3m in height may well have grown in under half a century. This growth rate has implication for the rate of relative sea level rise which is discussed in section 3.8 and further in chapter 5.
The cause of the total domination of one species of coral in these framestones may well be biotic factors. Martin et al. (1989) suggests that interspecific aggression, susceptibility to predation and colony overgrowth may have caused the *Tarbellastrea* domination observed in south east Spain. *Porites* domination of the Yesilbag reefs may be related to the Late Miocene age of the reefs (see section 3.6.1.1) and will be discussed further in section 3.8.2.

The close packed nature of the *Tarbellastrea* framestone, results in little sediment being retained within the structure. This might initially suggest that the sediment input at this time was low. However, structures as tall as these colonies are require support in order to prevent their top heavy structures from falling over (Scoffin 1987). The outcrop evidence suggests that rather than being supported in sediment, neighbouring colonies, by virtue of the competition for space, supported each other in a close-packed structure with few inter-colony interstices. This is similar to the *Tarbellastrea* frameworks studied by Martin et al. (1989). It is difficult to draw any firm conclusions concerning sediment flux during growth of *Tarbellastrea* fan-coral framestones as their topography may well have affected sediment dispersal patterns. This is discussed further in chapter 5.

Conical shaped stick colonies of *Porites* have been documented by Martin et al. (1989) in the Tortonian reefs of south east Spain. These corals anchor themselves directly into the silt substrate. The authors indicate that the silt was not lithified at the time of coral growth allowing colonies of *Porites* to topple over and become embedded in silt without suffering mechanical abrasion characteristic of firmly cemented corals which have been dislodged by wave or current action. This may well be what has caused the mixture of *in situ* and displaced *Porites* colonies visible along the track north of Yesilbag. A relatively high sediment influx is envisaged for this environment.

3.6.1.4 Porites rudstone

*Dimensions and sediment associations*

*Porites* rudstones occur at two horizons within the South Kargi section.
Towards the bottom of the section, rudstone overlies the interbedded red conglomerates, sandstones, siltstones and calcrete horizons of the Kargi Group (Kizildag Formation; chapter 5). The coarse sandstone to micro-conglomerate substrate below the rudstone displays a weak palaeotopography, but there is no evidence of boring or encrustation. The rudstone itself varies in thickness from 50cm to 2m over a distance of 100m. To the east, the tunnel cutting reveals a wedge shaped Tarbellastrea fan coral framework passing laterally into Porites rudstone (Figs. 3.7 and 3.8). A higher rudstone horizon further up the main South Kargi section occurs above Porites bafflestone. The transition from the underlying bafflestone is abrupt and marked by the termination of the stick Porites and a change in the nature of the matrix.

The rudstone horizon at Altinkaya (Fig. 3.1) is only 50cm thick and is sandwiched between a conglomerate and a coral floatstone. Lateral variations in bed thickness were not observed.

Coral
The in situ coral at South Kargi is exclusively Porites with flat or dish-shaped forms. The concentration of coral increases up the rudstone section and corals generally become more domal in shape. Some redeposited stick-shaped coral fragments are found lying horizontally at the base of the rudstone horizon. Many of these are encrusted with coralline red algae. In the tunnel cutting, the majority of the coral in the rudstone horizon, though larger, appears to be redeposited.

At Altinkaya, the morphology of the Porites rudstone is stick-like. The clasts are much smaller than those at South Kargi with the largest coral clasts being up to 3 cm in length. Once again the long axes of the clasts lie approximately along the plane of bedding, but this bed presents a much more chaotic appearance than the rudstones in the South Kargi section. None of the Porites at Altinkaya appears to be in life position.

Matrix and diagenesis
There is a change in the matrix associated with the bottom of the lower most rudstone horizon at South Kargi. Here, poly lithic micro-conglomerate (radiolarite, carbonate and igneous fragments as well as
abundant shallow-water carbonate debris) gradually grades into the white silt matrix of the rudstone. This silt contains as clasts, small angular chert, bioclastic detritus including abundant algal fragments, and carbonate grains. The transition to a more bioclastic dominated matrix is mirrored in the upper rudstone although the grainsize is here rather coarser (fine sand). The percentage of matrix varies within and between rudstone horizons, but rarely constitutes more than 40%. In the tunnel cutting however, the base of the rudstone horizon has only 30% matrix, but it increases upwards until the coral clasts are only just in contact with each other and the rock resembles a floatstone rather than a rudstone.

The percentage of matrix found at Altinkaya is 50-60% with the coral clasts only just in contact with each other in the plane of exposure. Silt intraclasts of comparable size are also found. The silt matrix is dominated by bioclastic detritus including algal fragments, echinoid spines, benthic foraminifera with agglutinated tests, mollusc and brachiopod shell fragments. The subsidiary terrigenous component consists of rare angular quartz grains, opaques and carbonate grains which show deformation lamellae under cross-polars. Micrite envelopes on the coral clasts and on the other biogenic components are locally coated in microspar and clay minerals.

Interpretation

Particular morphologies and species of coral are characteristic of early colonising assemblages. Flat and dish-shaped coral structures have large surface areas of attachment providing better stability on sandy substrates and in high energy environments (Scoffin 1972). The presence of some Porites sticks lying horizontally at the base of the rudstone horizon in the Kargi section, may indicate either that stick Porites colonies were forming here as in SE Spain (Martin et al., 1989) on an unlithified substrate or that a nearby established reef was shedding stick Porites material. Two pieces of evidence mitigate against the second hypothesis:

- The transition from continental to marine sedimentation is just below the rudstone horizon;
- Palaeocurrent data from the cross-bedded conglomerates indicate that fluvial input was from the W-SW.
The evidence concerning the hardness of the substrate prior to coral colonisation is mixed. The gentle palaeotopography preserved on the surface of the underlying micro-conglomerate may suggest that there had been some lithification of the substrate before coral colonisation. The absence of borings on the top of the micro-conglomerate however, indicates that the upper surface was not hard. The presence of redeposited horizontal sticks of *Porites* within the rudstone may also indicate a soft substrate such that nearby stick-Poritids toppled over due to substrate instability (Martin *et al.* 1989). Mechanical erosion and redeposition of stick *Porites* colonies could also be responsible for these however, and the encrusted nature of the corals argues in favour of this as Martin *et al.* (1989) indicate that immediate burial in silt due to the toppling of a stick *Porites* colony prevents extensive encrustation and abrasion from occurring.

Although coralline red algae are found encrusting individual coral clasts and *in situ* coral and as the main constituent of the bioclastic component in the matrix, algae only very rarely occurs as *in situ* laminated mats. It can therefore be interpreted as a secondary framework builder (Scoffin, 1987).

### 3.6.1.5 Summary of reef sub-facies

A summary table of the information discussed and interpreted above is given in table 3.5.

### 3.6.2 Off-Reef facies

Table 3.6 below displays the off-reef sub-facies, listing their characteristic fauna and the localities at which they can be found (Fig. 3.1).

Generally, transitions from reef to off-reef facies are abrupt as preserved in a vertical section. Where exposed however, lateral transitions are gradational over 5 to 100 metres. This is a result of the reefs exerting a control on the surrounding facies development (e.g. Scoffin 1987.)
<table>
<thead>
<tr>
<th>Sub-facies</th>
<th>Coral type</th>
<th>Matrix</th>
<th>Energy level</th>
<th>Sediment dispersal pattern</th>
<th>Water depth/sea level change</th>
</tr>
</thead>
<tbody>
<tr>
<td>Domal coral framestone</td>
<td><em>Porites</em></td>
<td>20-30% chert, sandstone, carbonate, bioclasts</td>
<td>Moderate-high</td>
<td>found in muddy channel</td>
<td>Photic zone, but no evidence of corals reaching sea level</td>
</tr>
<tr>
<td></td>
<td><em>Montastrea</em></td>
<td>Very little</td>
<td>Low</td>
<td>may have formed in areas of low sediment influx</td>
<td>Photic zone, but no evidence of corals reaching sea level</td>
</tr>
<tr>
<td>Porites bafflestone</td>
<td><em>Porites</em></td>
<td>Sandstone/micro-conglomerate</td>
<td>High</td>
<td>Traps sediment (see chapter 5 for more information)</td>
<td>Relative sea level rise &gt; sediment influx. No evidence of sea surface</td>
</tr>
<tr>
<td>Fan coral framestone</td>
<td><em>Porites</em></td>
<td>Siltstone</td>
<td>High</td>
<td>May trap some sediment</td>
<td>Relatively shallow</td>
</tr>
<tr>
<td></td>
<td><em>Tartbellastraea</em></td>
<td>Very little</td>
<td>?</td>
<td>Topography may have influenced sediment dispersal patterns</td>
<td>Relatively deep? Instantaneous sea level rise &gt; 8m/1000yrs</td>
</tr>
<tr>
<td>Porites rudstone</td>
<td><em>Porites</em></td>
<td>30-60% chert, sandstone, carbonate, high concentration of bioclastic components</td>
<td>High</td>
<td>Colonising colony. May suggest period of low sediment influx</td>
<td>?</td>
</tr>
</tbody>
</table>
Table 3.6  Table of off-reef sub-facies, listing their characteristic fauna and the localities at which they can be found (Fig. 3.1).

<table>
<thead>
<tr>
<th>Sub-facies</th>
<th>Characteristic fauna</th>
<th>Localities</th>
</tr>
</thead>
<tbody>
<tr>
<td>Coral floatstone</td>
<td>Various coral types</td>
<td>Dumanli</td>
</tr>
<tr>
<td></td>
<td>Gastropods</td>
<td>Altinkaya</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Karapinar</td>
</tr>
<tr>
<td>Rhodolithic calcarenite</td>
<td>Algal rhodoliths various shallow-water biogenic fragments</td>
<td>Yaylaalan</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Ahmetler</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Alarahan</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Kepezbelenli</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Akseki Road</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Deniztepesi</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Bucakköyü</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Saburlar</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Oymapinar</td>
</tr>
<tr>
<td>Reef talus</td>
<td>Coral</td>
<td>Oymapinar</td>
</tr>
<tr>
<td></td>
<td>Molluscs</td>
<td>Deniztepesi</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Akseki Road</td>
</tr>
</tbody>
</table>

3.6.2.1  Coral floatstone

*Dimensions and sediment associations*

At Altinkaya (Fig. 3.1), a coral floatstone overlies the Porites rudstone described above. The floatstone horizon is 50cm thick and contains large coral clasts up to 10cm across, oysters and algae, all chaotically orientated. The floatstones at Dumanli (Fig. 3.1) are interbedded with sandstones, conglomerates and rare, thin *in situ* coral framestones. The floatstones wedge out passing into coral and gastropod packstones towards the west and into macro-fossil poor marls and sandstones to the east. The section at Karapinar is just over 18m long and contains 4 individual floatstone horizons.

*Corals*

Two coral samples from the Dumanli section was submitted by Dr. J. P. Cuif for Sr measurements using an energy dispersive spectroscopic
The mean Sr values for these two samples were 0.7% and 0.65%. These values are far closer to the percentages expected in living Scleractinid aragonite (approximately 0.8%) than they are for diagenetic low-Mg calcite (0.02%; Scoffin, 1987). SEM photographs taken of these same samples revealed good preservation of skeletal fibres (Fig. 3.11). There has been some preferential dissolution along organic-rich growth lines, but there is no evidence of general calcitization, (C. Cuif pers. comm. 1994).

The Karapinar floatstones contain a diverse coral assemblage. Table 3.7 lists the species of coral found and their stratigraphical ranges as determined by Chevalier (1961). The coral assemblage found here indicates a Lower Miocene (Aquitanian-Burdigalian) age for the reef. This has important implications for the age of the previously undated underlying conglomerates (see chapters 5 and 6).

Many of the corals collected from the Karapinar section show good preservation of the detailed skeletal structure in hand specimen (Fig. 3.12a), (B. Rosen pers. com., 1993). Of these coral specimens two samples (26J.217.4a and 26J.217.4c) were ground up for x-ray diffraction analysis to determine whether primary aragonite was still present. Sample 26J.217.4a (*Heliastrea oligophylla*) contained approximately 55% aragonite (see Appendix 1). Sample 26J.217.4c (*Favites neglecta*; Fig. 3.12b) was found to have been completely transformed to low-Mg calcite. In hand specimen the Favid appears smoothed and abraded (Fig. 3.12b) and in thin section the skeletal structure can be seen to have been completely destroyed and replaced by irregular dirty calcite crystals. The Heliastrid, by contrast, retains some of the prismatic crystal structure within the coral skeleton.

**Matrix and Diagenesis**

At Altinkaya coral clasts are suspended in a fine to medium-grained sandstone matrix. Smaller bioclasts include echinoid spines, brachiopod fragments showing punctate like structures, benthic foraminifera with agglutinated tests, gastropods and organic fragments. Terrigenous quartz and deformed carbonate grains are scattered liberally through out, constituting 10-20% of the matrix. The coral clasts have a thin, well defined micrite envelope around the skeletal structure which has been...
<table>
<thead>
<tr>
<th>NAME</th>
<th>Aquitanian</th>
<th>Burdigalian</th>
<th>Langhian</th>
<th>Serravallian</th>
<th>Tortonian</th>
</tr>
</thead>
<tbody>
<tr>
<td>Heliastrea oligophylla</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Favites neglecta</td>
<td></td>
<td></td>
<td></td>
<td></td>
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</tr>
<tr>
<td>Tarbellastrea carryensis</td>
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<tr>
<td>Tarbellastrea ellisiana</td>
<td></td>
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<td></td>
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<tr>
<td>Caulastria matheroni</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Porites sp.</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Table 3.7. List of the coral species found in the patch reefs on the western margin of the Köprü basin and their age ranges. The species found in these reefs allowed them to be dated as Early Miocene in age (Chevalier, 1961; B. Rosen pers. com. 1993).
Figure 3.11  SEM photograph of skeletal fibres of aragonite in a Dumanli coral, north Köprü basin. Note the preferential dissolution along organic growth lines and the good morphological preservation of aragonite fibres (Photographs: C. Cuif, 1994).
Figure 3.12 Photograph of a) abraded *Favites neglecta* and well preserved and b) less abraded *Heliastrea oligophylla* (Chevallier, 1961)
replaced by large, equant crystals of neomorphic spar. Smaller stubby, prismatic crystals of sparry calcite coat the micritic envelopes, but the bulk of both matrix and cement is micrite. The matrix at Dumanli is uniformly fine-grained and argillaceous and has yielded a Serravallian age from nannoplankton analysis (C. Müller pers. com., 1994.)

At Karapinar by contrast, the matrix of the floatstones varies from coarse-grained sandstone to argillaceous marls and corals. Gastropods and brachiopods are also found within the floatstones, but are a subsidiary component. Planktic foraminifera are extremely rare. Wood and plant fragments in various states of preservation are found abundantly in some of the floatstone horizons. Thin sections cut perpendicular and parallel to growth direction reveal that the pore spaces in the coral have been variously filled with micrite and drusy sparite, with opaques concentrated round the edges of the spar grains.

Karapinar corals showing sharply defined calice and septal structure including sample 26J.217.4a, are almost entirely found within a silt to mud grade matrix. These horizons are argillaceous, but the grains identified at the size range 63-200µ are almost entirely carbonate, (see Appendix 4.)

Interpretation
Two possible depositional processes may have been active in forming the coral floatstones found in the study area.
1) Mechanical abrasion of a raised reef structure leading to periodic shedding of coral debris down slope into a quieter water area accumulating finer grained sediment. This may have occurred at Dumanli, where the correlative reef is still preserved up slope.
2) Redeposition of fragmented coral by debris flows generated by storms or faulting events. This seems particularly likely when shallow-water material is found interbedded with deeper-water components which appear to have been deposited by turbiditic currents (e.g. chapter 4). At Altinkaya and Karapinar, no in situ reef structure now remains and although this does not preclude the possibility of deposition having occurred as described above, debris flow processes may also have been active.
There is an apparent relationship between the grainsize of the matrix and the preservation of primary structure and chemistry in coral. All the coral samples which have retained primary aragonite were sampled from fine-grained clay-rich rocks and it seems likely that the argillaceous nature of the matrix restricted diagenetic fluid flux. Matrix with coarser grainsizes, have a higher quantity of chert and quartz and are likely not only to have a smaller quantity of clay minerals due to the higher energy of deposition required, but also a greater volume of pore network allowing post-depositional flux of fluids and aiding diagenetic transformation. Coarser grainsizes also may account for the increased abrasion on coral samples found associated with them.

On a larger scale, Scoffin (1987) indicates that rock composition may be a function of whether shallow marine limestones undergo burial after a regression or transgression. Exposure caused by a regression leads to diagenetic change in an open system resulting in flushing out of Sr$^{2+}$ and Mg$^{2+}$ ions. Burial caused by a transgression is more likely to result in a closed system i.e. shallow water carbonates become covered in argillaceous marls and clays which prevent high fluid flux. In this case the isotopic concentrations of the constituents are more likely to be preserved.

3.6.2.2 Rhodolithic calcarenites

*Dimensions*
Rhodolithic calcarenites are the most abundant shallow water carbonate facies found in the study area making up approximately 60% of all shallow water carbonate exposure. Technically, the majority of the rocks classified are having a dominant grainsize >2mm, but there is an infinite range from micro-conglomeratic calcirudites with clasts of rhodolithic algae, coral and shells, to calcilutites with a dominant grainsize <63µ. Sorting and maturity of the carbonate grains are also infinitely varied. The calcarenites are often well bedded on a scale of 0.5-2m (Fig. 3.14) and occasionally show cross bedding.
Figure 3.14 Bedded calcarenites banking up against Alanya Massif basement with marked palaeotopography, Ahmetler section, Manavgat basin.

Figure 3.15 Rhodolithic algae forming round a flat *Porites* nucleus, Ahmetler section Manavgat, basin.
Composition and diagenesis

Calcarenite composition varies from bioclastic sandstones to calcarenites with a negligible terrigenous component. There is apparently no systematic variation of terrigenous input, (individual sections contain a range of terrigenous percentages) except in the north of the Manavgat basin. Here, calcarenites found on top of the Alanya Massif contain some Alanya Massif limestone and schist fragments, but no other detrital clasts.

Coralline red algae of the *Amphiroa* and *Lithothamnium* types, is almost always present in large quantities making up 10-70% of calcarenites point counted. Occasionally whole rhodoliths are preserved as at Ahmetler. Here, fragmented strands of algae are found with rhodoliths several centimetres in diameter which have generally nucleated on fragments of Poritid coral (Fig. 3.15). Whole rhodoliths are rare in calcarenite facies (see echinoid-scaphopod facies, 3.6.3.1 below). The vast bulk of algal material appears as variously micritized fragments. At Halitigalar algal fragments show a complete gradation between fragments with well preserved structure to structureless peloids. Where well preserved, the cells within the algal structure are filled, either with micrite or spar. Conceptacles are also generally spar filled.

Other fauna, including molluscs, coral, brachiopods, echinoids, and benthic foraminifera, found in horizons containing whole rhodoliths are almost always in a fragmentary state. Locally however, 4-8 entire individual echinoids (Clypeasterids) are found in close proximity to each other (e.g. Ahmetler; Ballibucak; Fig. 3.1). Calcarenites found interbedded with planktic foraminiferal marls also contain planktic foraminifera. These tend to be far less fragmented than the shallow water components.

Cements are generally dominated by micrite, but spar is common within the larger pore spaces and particularly in the upper chambers of benthic foraminifera and gastropods. Variously developed micrite envelopes are common on all calcarenite clasts.

Interpretation

Algal rhodoliths in Bermuda form today at depths of 3-5m (Orzag-Sperber et al., 1977). The environment in which they are found is one of high-
energy and with constantly agitated conditions. It is probable fauna other than these rhodoliths have been transported and this helps to explain the advanced state of abrasion and fragmentation that they have suffered.

Algal-rich calcarenites in which there are no whole rhodoliths are also interpreted as having occurred in high energy environments. The common association of coral, and the rich diversity of biota indicates that redeposited reef material is the probable source. The lack of any terrigenous component other than Alanya Massif material in the Ahmetler and Akseki road sections is indicative of the localised source area in this region. The implications of this are discussed in section 5.7. The sorting and maturity of grains is probably a function of the energy of the environments and the length of time the sediment is exposed.

3.6.2.3 Reef talus

The most widespread and prominent reef talus horizon occurs at the boundary between shallow water carbonates (Oymapinar Limestone) and shelf carbonate facies (Geceleme Formation). Good examples of these are found in the Akseki road section, Oymapinar and at Deniztepesi (Fig. 3.1). In these localities talus horizons between 2 and 6m thick overlie in situ reef framestones and wedge out over 500m to a kilometre. The talus is always overlain by planktic foraminiferal marls, (Geceleme Formation) or turbidites (Karpuzçay Formation). At Deniztepesi, where the talus forms a prominent cliff, these marls overlie it with an angular discordance. The Oymapinar section (Fig. 3.16) contains three separate talus horizons interbedded with in situ reef framestones, calcarenites and bioclastic sandstones. The upper most of these is also overlain by deeper water marls with an angular disconformity.

Composition

The Akseki road section is perhaps the most spectacular of the talus outcrops as it contains huge blocks of domal coral framestone up to 3.5m in diameter. The poorly sorted nature of the clasts means that many of the larger ones protrude from the bed and stick up into overlying marls which have passively filled around them. This section is associated with abundant normal faulting which is discussed in section 7.5.1.2.
Figure 3.16 Log of the type section locality of the Oymapinar Limestone at Oymapinar, northern margin of the Manavgat basin. The dominant in situ facies is domal coral framestone, but the section also contains multiple talus horizons often comprising disorientated blocks of coral-rich reef limestone. At the top of the section a talus horizon is overlain by the planktic foraminiferal marls of the Geceleme Formation with a clear disconformity.
Oymapinar talus horizons as well as containing disorientated blocks of domal coral framestone also contain abundant oysters and gastropods. Deniztepesi by contrast is a talus composed of coral-poor limestone, dominated by algae.

Matrix and diagenesis
The Akseki road talus is well exposed in a stream bed and has a carbonate cemented top up to half a metre thick (Fig. 3.17). Beneath this the talus is a loose rubble, except where faults dissect it. In contrast to this the lower talus horizons at Oymapinar have abundant white silt micritic matrix. The uppermost talus at Oymapinar resembles that at the Akseki road section.

Interpretation
The process of deposition in the lower talus horizons at Oymapinar may be similar to those of the coral floatstones i.e. mechanical and bio-erosion of reef framework depositing into a deeper, quieter water silt dominated environment. It is possible that the talus horizons which occur at the interface between shallow carbonate deposition and the planktic foraminiferal marls were fault generated. The following evidence supports this interpretation:

- The presence of an angular disconformity marking the transition from shallow to deeper water;
- The absence of matrix;
- The very angular nature of the blocks;
- The spatial correlation of talus and Çakallar Formation with areas containing numerous faults (section 7.5.1.2);
- Thickness variations across the Manavgat basin (section 7.5.1.2).

Note however that much talus can be generated by the fast growth of coral due to rapid relative sea level rise. As rapid relative sea level rise would be predicted in association with extension generated faulting, the talus described here may have been formed by both processes and it would be difficult to distinguish between them.
Figure 3.17 Photograph of the cemented top of the Akseki road talus horizon and the loose rubble beneath, Manavgat basin.

Figure 3.18 Echinoid-Scaphopod facies with algal rhodolith encrusting Mesozoic limestone pebble, Aspendos section, west Köprü basin.
<table>
<thead>
<tr>
<th>Sub-facies</th>
<th>Components</th>
<th>Matrix</th>
<th>Cement</th>
<th>Preservation</th>
<th>Water depth</th>
<th>Mode of deposition</th>
</tr>
</thead>
<tbody>
<tr>
<td>Coral floatstone</td>
<td>Various corals</td>
<td>clay</td>
<td>poorly lithified</td>
<td>Primary aragonite preserved in corals</td>
<td>?</td>
<td>Rock fall (and/or debris flow)</td>
</tr>
<tr>
<td></td>
<td>Coral and other shallow water components</td>
<td>sandstone</td>
<td>Poorly lithified</td>
<td>Corals replaced with secondary Low-Mg calcite</td>
<td>?</td>
<td>Debris flow</td>
</tr>
<tr>
<td>Rhodolithic calcarenite</td>
<td>Biogenic reef components Generally little terrigenous material</td>
<td>micrite</td>
<td>Micrite and spar</td>
<td>Secondary replacement Shallow water, shelf environment</td>
<td>?</td>
<td>Current reworking</td>
</tr>
<tr>
<td>Reef talus</td>
<td>Reef framestone blocks Local basement in north of Manavgat basin</td>
<td>none</td>
<td>surficial carbonate cement</td>
<td>Coral aragonite replaced by secondary calcite</td>
<td>?</td>
<td>Rock fall/fault generated talus</td>
</tr>
</tbody>
</table>
3.6.2.4 Summary of off-reef sub-facies

A summary table of the information discussed and interpreted above is given in table 3.8.

3.6.3 Reef-associated facies in coarse clastic environments.

Table 3.9 Table of reef-associated sub-facies, listing their characteristic fauna and the localities at which they can be found (Fig. 3.1).

<table>
<thead>
<tr>
<th>Sub-facies</th>
<th>Characteristic fauna</th>
<th>Localities</th>
</tr>
</thead>
<tbody>
<tr>
<td>Echinoid-scaphopod grainstone</td>
<td>Echinoids, scaphopods, algal rhodoliths, Operculina</td>
<td>Aspendos, South Yesilbag</td>
</tr>
<tr>
<td>Oyster packstone</td>
<td>Oysters</td>
<td>Kesme, North Yesilbag</td>
</tr>
<tr>
<td>Gastropod wackestone</td>
<td>Gastropods</td>
<td>North Yesilbag, Bozburun Dag</td>
</tr>
<tr>
<td>Oyster bafflestone</td>
<td>Oysters</td>
<td>Kargi Baraj</td>
</tr>
</tbody>
</table>

3.6.3.1 Echinoid-scaphopod grainstone

Dimensions and localities
The largest outcrop of this sub-facies occurs to the north of the ruins of Aspendos (Fig. 3.1). Here an 11m interval of echinoid-scaphopod grainstone persist for several hundred metres, passing up into algal limestone (Fig. 5.6). The exposure south of Yesilbag is associated with the channel structure sketched in figure 3.3 and is limited laterally by it. The thickness of the horizon here is only 2m and it passes up in to domal coral framestone.

Fauna and flora
This very distinctive facies consists of a coarse micro-conglomerate with abundant whole and fragmented scaphopods, echinoids, large benthic foraminifera including Operculina and, more rarely, algal rhodoliths up to 5cm in diameter.
These rhodoliths encrust rounded pebbles most commonly veined grey Mesozoic limestone (Fig. 3.18). Individual layers of algae can be clearly seen encrusting first the clast and then other algal layers and included matrix. They are irregular in shape and individual layers do not encircle the clast, but can be traced out to their terminations, which are often abrupt and ragged. Although algal encrustation is visible on many of the micro-conglomerate clasts, the biggest rhodoliths are developed on clasts significantly larger than average.

The orientations of scaphopods and echinoids are for the most part, random. In just a few places however, millimetre scale, graded laminations can be seen and scaphopods have been deposited parallel to these, (Fig. 3.19) The scaphopod most commonly found here is *Dentalium* (Upper Cretaceous-Recent) with its prominent longitudinal ribs. When entire, specimens can reach over 10cm in length. The majority of the scaphopods occur here as large fragments. Entire clypeastrids (minus their spines) are fairly common. Their distribution tends to be rather clustered. These are infaunal or semi-infaunal detritus feeders.

*Terrigenous components*

The proportions of detrital components of the echinoid-scaphopod facies varies from place to place, but the components themselves remain fairly constant. In order of volumetric importance they include grey and fawn crystalline limestone sometimes veined, red and green chert, grey sandstone and yellow silt. The chert is generally subangular to angular, whilst the other grains are rounded to well rounded.

*Interpretation*

No columnar rhodoliths are found here, just laminar ones which form in Bermuda today by overturning on a sandy substrate (Orzag-Sperber *et al.* 1977). The preservation of these large and delicate structures in such a coarse clastic environment shows that they have not been transported. Rhodoliths of a comparable size and structure have been documented from the Miocene of the Meso-Hellenic Trough in northern Greece (Wilson, 1993). The large number of entire echinoid and scaphopod fossils found in this facies indicates that these creatures also inhabited
Figure 3.19 Scaphopods parallel to grit-silt laminations. Randomly orientated scaphopods can be seen in the overlying sediment, Aspendos section, south-west Köprü basin.

Figure 3.20 Pteropod moulds in planktic foraminiferal marls, Akseki road section, Manavgat basin.
this coarse grained environment. Scaphopods are deposit feeders which live with the wider end of the shell embedded in the sediment. Their presence is in accordance with the interpretation of a shallow marine environment subject to periodic sweeping by currents.

3.6.3.2 Oyster packstone

Dimensions
This sub-facies was first described by Dumont (1974). He noted that although individual horizons are of limited lateral extent, several similar beds occur at different localities at about the same stratigraphic level in the Kesme area (Fig. 3.1). The morphology of the beds is often channelised and interbedded with fossil poor sandstones, siltstones and rare conglomerates. Beds generally are less than a metre thick and consist of silts to medium grained sandstone with abundant, large oysters.

Fauna and flora
The two valves of the oysters are generally separated although little abrasion of the shell is visible. The abundance of oysters varies dramatically from being so concentrated that they constitute over 50% of the rock and are in contact with each other, as can be seen along the Kesme-Yesilbag road, to being more sparsely distributed so that the clasts are matrix supported as can be seen between Kasimlar Kesme. In this latter case other fauna particularly gastropods and burrowing bivalves are common. Wood is also associated and on the Kasimlar-Kesme road and some of these fragments reach 20cm in length. The oysters show good alignment sub-parallel to bedding irrespective of whether they are in contact with each other. Some of the gastropods show a similar alignment, but the burrowing bivalves cut across this trend and appear to be in situ.

Matrix
The matrix to this sub-facies is yellow and is dominated by angular chert and limestone fragments. It is generally poorly lithified and argillaceous.
Interpretation
The dislocation and alignment of the oyster valves indicates that they have been reworked. The concentration of redeposited fauna presumably reflects the concentration in the source area and the volume of sediment and post-depositional current winnowing. This source area is likely to have been a marginal, shallow-marine setting, allowing the inclusion of fairly large fragments of wood. The channelised nature of these bodies and the size of some of the larger oysters which were transported as clasts, indicates that the current was probably a high energy one. Subsequent burrowing may have occurred in periods of quiescence.

3.6.3.3 Gastropod grainstone/wackestone

Dimensions and age
This sub-facies resembles the oyster packstone in terms of its matrix composition, bed morphology and texture. It is best exposed along the Kesme-Yesilbag road, but it is also found a few metres above a coal horizon to the north of the Bozburun Dag col, near Pinargözü (Fig. 5.10). At Kesme, gastropod wackestones are interbedded with sandy marls the nannoplankton which have yielded an Upper Miocene age (C. Müller, pers. com., 1994). This confirms the Tortonian age proposed by Dumont (1974) from the planktic foraminifera from interbedded horizons.

Fauna
The dominant fauna are gastropods identified by Dumont (1974) as Cerithidae sp. Other fauna found within these fossil rich beds include bivalves, rare oysters and brachiopods and the benthic foraminifera Alveolina.

Matrix
At Kesme cross-laminated micro-conglomeratic sandstones and laminated silts showing water escape structures are also associated with gastropod wackestones. The gastropods here occur either in winnowed horizons with a fine sand matrix or in argillaceous marls where they show no preferred orientation and are well spaced. At Bozburun Dag however, the matrix is siltstone-micro-conglomerate.
Interpretation
The interpretation of this sub-facies is similar to that for the oyster packstone facies. Deposition by high-energy currents and subsequent winnowing of finer sediments is likely to have caused the grainstones, whilst debris flow processes may have deposited the wackestones.

3.6.3.4 Oyster bafflestone

Dimensions
This sub-facies has only been identified in one locality. A hundred metres north of the Kargi baraj, oyster-rich limestones are exposed overlying Jurassic limestones with a marked palaeotopography. The sequence has a very limited lateral extent. 50m along strike to the south a debris flow with clasts up to a metre in diameter is exposed banked up against the basement. Half a kilometre to the north a small coral patch reef is exposed at the same horizon.

Fauna
The basement limestone is heavily bored, by small, Lithophaga-like borings. Planar-bedded, bored limestones, variably rich in oysters onlap on to the basement. Half a metre above the base of the section and rooting partly on the underlying oyster-rich limestones and partly on the basement, an oyster dome, 50cm in height and 20cm in diameter is located. The oysters are rigidly cemented on to each other producing a framework in which interstitial sediment is trapped. Laterally, the dome passes abruptly into fossiliferous limestone with abundant tubular structures. The sequence is overlain by interbedded marls and sandstones.

Interpretation
This unusual facies has been interpreted as having grown in situ. The bored nature of both the basement and underlying oyster-rich limestones indicates exposure and rapid lithification. The small quantity of terrigenous material found in the oyster-rich limestones suggests efficient channelling was active in the area, leading to rapid lateral changes in facies visible at this locality.
<table>
<thead>
<tr>
<th>Sub-facies</th>
<th>Components</th>
<th>Matrix</th>
<th>Cement</th>
<th>Preservation</th>
<th>Water depth</th>
<th>Mode of deposition</th>
</tr>
</thead>
<tbody>
<tr>
<td>Echinoid-scaphopod grainstone</td>
<td>Echinoids, scaphopods algal rhodoliths benthic</td>
<td>Micrite</td>
<td>Micrite and rare spar in algal</td>
<td>Micrite envelopes developed</td>
<td>Shallow, probably within wave</td>
<td>Periodic influx of sediment reworked by currents (?)waves?)</td>
</tr>
<tr>
<td></td>
<td>foraminiferas angular chert and rounded limestone</td>
<td></td>
<td>conceptules</td>
<td></td>
<td>base</td>
<td></td>
</tr>
<tr>
<td>Oyster packstone</td>
<td>Oysters</td>
<td>Clay-rich</td>
<td>Poorly lithified Micrite</td>
<td>Altered oyster chemistry (see</td>
<td>Relatively shallow and near</td>
<td>Debris flow? subsequently winnowed</td>
</tr>
<tr>
<td></td>
<td>Gastropods</td>
<td></td>
<td></td>
<td>chapter 6)</td>
<td>shore</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Wood</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Chert and limestone</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Gastropod grainstone/wackestone</td>
<td>Gastropods benthic foraminifera bivalves Oysters</td>
<td>micro-</td>
<td>Micrite and spar</td>
<td>Micrite envelopes developed</td>
<td>relatively shallow</td>
<td>Winnowing by currents Some current deposition</td>
</tr>
<tr>
<td></td>
<td>terrigenous carbonate</td>
<td>conglomerate to clay</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Oyster bafflestone</td>
<td>Oysters Shallow-water debris Lithophaga-like</td>
<td>Micrite</td>
<td>Spar</td>
<td>Bored</td>
<td>Shallow</td>
<td>Cementation</td>
</tr>
<tr>
<td></td>
<td>borings</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
3.6.3.5 Summary of reef-associated sub-facies

A summary table of the information discussed and interpreted above is given in table 3.10.

3.6.4 Carbonate shelf facies

Table 3.11 Table of shelf sub-facies, listing their characteristic fauna and the localities at which they can be found (Fig. 3.1).

<table>
<thead>
<tr>
<th>Sub-facies</th>
<th>Characteristic fauna</th>
<th>Localities</th>
</tr>
</thead>
<tbody>
<tr>
<td>Planktic foraminiferal marl</td>
<td>Planktic foraminifera</td>
<td>Alarahan</td>
</tr>
<tr>
<td></td>
<td>pteropods</td>
<td>Ahmetler</td>
</tr>
<tr>
<td></td>
<td>rare benthic foraminifera</td>
<td>Saburlar</td>
</tr>
<tr>
<td></td>
<td>and rare echinoid spines</td>
<td>Akseki Road</td>
</tr>
<tr>
<td>Operculina packstone/grainstone</td>
<td>Operculina echinoids</td>
<td>Halitigalar</td>
</tr>
<tr>
<td>Calcirudites</td>
<td>Algae</td>
<td>Akseki Road</td>
</tr>
<tr>
<td></td>
<td>foraminifera</td>
<td>Alarahan</td>
</tr>
<tr>
<td></td>
<td>echinoid spines</td>
<td>Ahmetler</td>
</tr>
</tbody>
</table>

3.6.4.1 Planktic foraminiferal marl

Dimensions

A thick succession (>330m) of this facies occurs throughout the Manavgat basin. Although the majority of the marls appear to be laterally extensive, rare, better lithified horizons show a wedge-shaped geometry. The Ahmetler succession (Fig. 3.1) is interspersed with rare calcarenite and calcirudite horizons some of which contain large (several metres in diameter) detached blocks of Alanya Massif limestone.

Fauna

The facies consists of a fine grained carbonate-rich marl with an abundance of planktic foraminifera and more sparse benthic foraminifera. Occasionally horizons contain abundant pteropod moulds (Fig. 3.20) e.g. in the Ahmetler section and just above the talus at Akseki road. The abundance of the planktic foraminifera in these marls ranges from less than 1% to over 20% as calculated by point counting of washed,
disaggregated samples. (These figures do not include the clay mineral content; see Appendix 4). Where the foraminifera are most abundant i.e. along the Alarahan-Çakallar road and at the junction of the Ahmetler road with the road to Akseki, the dominant species is *Orbulina sp* and this can constitute over 60% of the planktic foraminifera. Other planktic foraminifera commonly found include *Globorotalia* and *Globigerinoides*. Monod (1977) reports the following species:

- *Globigerinoides bisphericus*
- *Globigerinoides subquadratus*
- *Globigerinoides trilobus*
- *Globoquadrina altisiora*
- *Globoquadrina dehiscens*
- *Globoquadrina langhiana*
- *Globorotalia obesa*
- *Globorotalia scitula praescitula*
- *Praeorbulina glomerosa glomerosa*
- *Praeorbulina transitoria*
- *Praeorbulina glomerosa curva*

Table 3.12 below lists the benthic foraminifera found in these horizons and the water depths at which they are commonly found. The preservational detail of many of these foraminifera, both benthic and planktic, is good (W. Austin pers. com. 1994). Some benthic foraminifera however appear to have suffered abrasion and part of the test ornamentation may be obscured making identification difficult. Where large quantities of benthic foraminifera occur this is most likely to be the case. Shallow water debris, such as echinoid spines and gastropods are also often found in greater concentrations in these horizons which also tend to be those displaying better lithification and wedge-shaped geometries.

**Detrital components**

The terrigenous components in these marls are overwhelmingly dominated by carbonate grains and clay minerals. The carbonate can constitute 100% of grains point counted after washing, although rare quartz and other lithic fragments are more common. In the Ahmetler
section, the lower part of the marl sequence contains virtually no terrigenous material other than carbonate and clay minerals.

**Table 3.12 Table of depth diagnostic benthic foraminifera found in the planktic foraminiferal marls. Depths derived from living foraminifera according to Murray (1973); Boltorskoy and Wright (1976); Murray (1991).**

<table>
<thead>
<tr>
<th>Name</th>
<th>Water depth in meters</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ammonia</td>
<td>0-50</td>
</tr>
<tr>
<td>Amphistegina</td>
<td>5-20</td>
</tr>
<tr>
<td>Bolivina</td>
<td>0-3000</td>
</tr>
<tr>
<td>Bulimina</td>
<td>0-3000</td>
</tr>
<tr>
<td>Cancri</td>
<td>50-150</td>
</tr>
<tr>
<td>Elphidium</td>
<td>0-50</td>
</tr>
<tr>
<td>Melonis</td>
<td>3-1000</td>
</tr>
<tr>
<td>Nonion</td>
<td>0-180</td>
</tr>
<tr>
<td>Operculina</td>
<td>0-70</td>
</tr>
<tr>
<td>Planorbulina</td>
<td>0-50</td>
</tr>
<tr>
<td>Rosalina</td>
<td>0-100</td>
</tr>
<tr>
<td>Ulvigerina</td>
<td>100-&gt;4500</td>
</tr>
</tbody>
</table>

**Interpretation**

The abundance of planktic foraminifera and relative scarcity of benthic foraminifera and shallow-water dwellers indicates that the environment of deposition was deeper water relative to the reef, off-reef and reef-associated facies described above. The fine-grained sediment and planktic fauna may well have collected out of suspension in the water column. The benthic foraminifera listed in table 3.12 which are thought to be indicative of water depth suggest that the environment of deposition was still not particularly deep, corresponding to inner shelf (0-70m; e.g. Murray, 1993). Most of the benthic foraminifera listed however occur in horizons which also contain shallow-water debris (e.g. echinoid spines, algae). It is suggested that these horizons represent redeposition of shallow-water material (probably by low-density turbidity currents, see chapter 4) and the benthic foraminifera within them can therefore only
act as a minimum water depth indicator. The difference in lithification of these horizons may also be related to the different mode of deposition i.e. by a current rather than from suspension.

3.6.4.2 Operculina packstones/grainstones

Dimensions
Horizons of operculina packstone are up to 2m thick and laterally inextensive. They are well exposed at the base of the calcarenites in the Alarahan section and at Halitigalar.

Fauna
The large benthic foraminifera Operculina sp. can constitute from 10-70% of the rock. Its large flattened disc-shaped tests which are in contact with each other often show a parallel to sub-parallel alignment. Horizons are often algal-rich and contain whole, well preserved echinoids, as well as other benthic foraminifera with agglutinated tests.

Matrix and diagenesis
The matrix to these packstones is always a fine grained micrite, its percentage being dependent on the quantity of Operculina present. Micrite envelopes are visible on foraminifera tests and most of the cement appears to be micritic. Rarely however, syntaxial overgrowths on echinoid spines are visible.

Interpretation
These Operculina packstones have been interpreted as foraminiferal shoals similar to those identified by Samuel (1994) in Indonesia. They form in shallow marine conditions where currents winnow and sort material leading to the packstone or grainstone texture. Micritic envelopes on the foraminifera indicate that they were at the sediment-water interface long enough for boring to occur. Similar foraminiferal shoals have been recognised from the Miocene of Cyprus (Eaton, 1987).
3.6.4.3 Calcirudites

**Dimensions**
These calcirudites occur within the planktic marl successions as spaced, wedge-shaped horizons, laterally discontinuous on a scale of several kilometres. They vary in thickness between 50cm and up to 5 m although some of the detached blocks within them may be considerably larger than the thickness of the horizon in which it sits.

**Components**
The calcirudites are dominated by shallow water debris including coral, coralline limestone, algal-rich calcarenite, echinoid spines and benthic foraminifera. Intraformational marl is also a common constituent. Terrigenous components in the lower part of the Ahmetler sequence are restricted to Alanya Massif limestone and schist. These form the large detached blocks which are entrained within some of the calcirudite horizons. Much of the debris within these horizons is angular.

**Matrix**
The matrix is very similar to the planktic foraminiferal marls in which these calcirudites are interbedded. It constitutes micrite with abundant planktic foraminifera. The percentage of matrix varies, but it can make up to 40%.

**Interpretation**
The marked textural immaturity of these calcirudites, their poor sorting and abundant matrix of a different composition suggests that they were deposited either by rock-fall or debris flow processes. The presence of huge detached blocks of Alanya Massif basement several kilometres from the present day outcrop suggests that deposition by debris flow processes is more likely. The nature of the Miocene carbonate material within them indicates that the source area was an area of shallow-water carbonate production.

3.6.4.4 Summary of shelf sub-facies

A summary table of the information discussed and interpreted above is
Table 3.13  Summary of information and interpretation of shelf sub-facies.

<table>
<thead>
<tr>
<th>Sub-facies</th>
<th>Components</th>
<th>Matrix</th>
<th>Cement</th>
<th>Preservation</th>
<th>Water depth</th>
<th>Mode of deposition</th>
</tr>
</thead>
<tbody>
<tr>
<td>Planktic foraminiferal marls</td>
<td>Foraminifera pteropods shallow-water debris terrigenous carbonate</td>
<td>micrite and clay</td>
<td>Poorly lithified</td>
<td>Some diagenetic alteration of foraminifera (see chapter 6)</td>
<td>Minimum = inner shelf (0-70m), probably more than this</td>
<td>Partly out of suspension, partly by turbiditic redeposition</td>
</tr>
<tr>
<td>Operculina packstone/wackestone</td>
<td>Operculina echinoids terrigenous carbonate grains</td>
<td>micrite</td>
<td>Poorly lithified Micrite and some spar</td>
<td>Micrite envelopes on foraminifera</td>
<td>Shallow, forming foraminiferal shoals</td>
<td>Current sweeping and winnowing</td>
</tr>
<tr>
<td>Calcirudites</td>
<td>Shallow-water components Large blocks of locally derived basement</td>
<td>clay-rich marl with planktic forams</td>
<td>Micrite and spar</td>
<td>Micrite envelopes developed</td>
<td>Minimum = inner shelf (0-70m), probably deeper</td>
<td>Deposition by debris flow processes</td>
</tr>
</tbody>
</table>
Figure 3.21 Depositional models for the deposition of the various biogenic carbonates discussed in the text.
given in table 3.13.

3.7 Model of sub-facies location.

Figure 3.21 is a model indicating the likely locations of the formations of the sub-facies discussed above.

3.8 Discussion

3.8.1 The Mid-Miocene hiatus in shallow-water carbonate production

The shallow-water carbonates in the study area are restricted to two distinct periods of time: the Early Miocene (Burdigalian-Langhian) and the Late Miocene (Tortonian). There is no evidence of shallow-water carbonate build-ups in the intervening period. This hiatus poses two questions. First, what caused the termination of the Lower Miocene shallow-water carbonates? And second, what allowed shallow-water carbonate production to start again in the Tortonian?

In Cyprus the Miocene reefs have a similar chrono-stratigraphic relationship (Follows 1990) to those in the study area and it was tentatively suggested that the Mid-Miocene hiatus in reef growth (Chevalier 1961) was a result of eustatic sea level rise (Follows 1990).

In the Antalya region however, it is difficult to assess the role of eustacy as no section contains both Lower and Upper Miocene reefs with a complete succession between. The Manavgat basin for instance, does comprise a complete succession from Upper Burdigalian to Messinian age, but it does not contain any Tortonian shallow-water carbonates. At first glance however, a eustatic sea level rise might appear to account for the termination of the Lower Miocene shallow-water carbonates. According to the sea level curve in Haq et al. (1988) the most rapid sea level rise occurring during the Langhian is approximately 75m over about 0.5 million years (Fig. 4.25). This of course represents average, not instantaneous sea level rise. If eustacy did drown the reefs then according to Geister (1983) and Johnson et al. (1986) instantaneous sea level rise must have been >7-8m/1000 yrs. The resolution of the eustatic curve (Haq et al., 1988) is not sufficient to conclusively decide whether or not
Post-hiatus colonisation in the Tortonian is also problematic. Tortonian reefs are concentrated in the north of the study area. Particularly in the far north they overlie thin sequences of continental conglomerates covering the basement. The shallow-water carbonates here, such as those seen at North Yesilbag and Kesme, can be clearly interpreted as a result of relative sea level rise. To the south however, the continuous sequence exposed in the Manavgat basin is entirely marine from the Langhian onwards. The benthic/planktic foraminiferal ratios here (Fig. 4.21) indicate overall shallowing water depths from Early Langhian times. Thus, any concrete evidence of a relative rise in sea level is localised even within the study area. The structural implications of this intra-basinal difference are discussed more fully in chapter 8.

If not eustatic sea level change, then what? Control over modern reef location is now thought to be shared amongst many factors most of which are not preserved in the fossil record e.g. salinity, temperature, nutrient supply, light, turbulence, substrate type and sediment flux. A combination of these factors is likely to have controlled the Mid-Miocene shallow-water carbonate hiatus. However, relative sea level change has not been ruled out as a possible controlling factor. A rapid tectonic subsidence might result in sufficient relative sea level rise to drown shallow-water carbonate production and this possibility is discussed further in section 7.5.2.1. What can to a certain extent be discounted is the control substrate and sediment flux have over the location of shallow-water carbonate deposition. For instance, it is difficult to argue that the high rate of muddy sedimentation in the Manavgat basin in the Tortonian precluded coral colonisation when areas such as South Yesilbag indicate that *Porites* colonisation was possible at the same time in apparently far more adverse conditions i.e. within a mud dominated channel.

### 3.8.2 Coral zonation

The general paucity of coral species in the Mediterranean during the entire Miocene, but particularly in the Late Miocene was first discussed by
Chevalier (1961). Lydendyk et al. (1972) proposed that tectonic movement in the Far East blocked the eastward flowing current from the Indo-Pacific to the Mediterranean at the end of the Oligocene. This resulted in gradual cooling of the Mediterranean and caused selective coral extinctions. As a result the Eastern Mediterranean corals lack the diversity of the Indo-Pacific province (Chevalier 1977).

The dominant coral in the study area is *Porites*. Previous authors working on Miocene Mediterranean corals have also observed a *Porites* domination (e.g. Follows 1990). Frost (1981) indicates that the reason for *Porites* being the dominant "pioneer" coral in the Miocene of the Mediterranean is due to the ability of its planula larvae to settle on mobile wave swept sandy substrates and the ability of the corallites to free themselves from the constant rain of sedimentary particles. In the South Kargi section, subordinate early colonisers of unstable substrates, such as *Acropora* and *Goniopora* (Frost 1981) are absent from the pioneering colonies. This may indicate that other environmental factors were adverse to coral colonisation as well as an unstable substrate. *Porites* has long been observed for being especially tolerant to adverse environmental conditions such as low salinity, low temperature, high terrigenous sedimentation and reduced water circulation (Marshall and Orr, 1931; Manton, 1935; Wells, 1954; Scoffin and Stoddart, 1978). In the South Kargi region, where the coral rudstones directly overlie coarse fluvial clastics and interbedded calcretes (see chapter 5) it seems likely that the fluvial influence was great and that therefore as well as being a high terrigenous sedimentation environment, the salinity was at least periodically lower than normal sea water.

However the South Kargi section is unusual in displaying vertical zonation in coral type. Most of the reefs studied had no systematic pioneering community that was significantly different from overlying framework horizons. A similar observation was made by Hayward (1982b) concerning Lower Miocene reefs of the Kas basin to the west of the study area. He suggested that the reason for this was due to the firm, though unlithified substrate of cobble and pebble gravel.

Hayward (1982a and b) did observe a vertical zonation in coral
Figure 3.22 Progressive change in coral morphology upwards through central reef framework (Hayward, 1982a and b).
morphology from flat, tabulate-dish forms, through branching, reticulate structures, to massive-domal forms (Fig. 3.22). He interpreted this zonation as being consistent with an increase in hydrodynamic stress. In the Antalya region, no systematic morphological variation in coral structure has been observed and in no reef is the whole of Hayward’s morphology zonation exposed. However all of his morphological zones can be identified. For instance, many of the lower coral horizons are rudstones with flat or tabulate forms. The other dominant morphology of basal coral horizons not documented by Hayward, is one of cone-shaped stick Porites, most commonly seen north of Yesilbag. A similar colonising Porites morphology is documented by Martin et al. (1989). The reticulate morphology documented by Hayward (1982a) is similar to the large Tarbellastrea fan framework seen at South Kargi, and the massive domal forms resemble those Montastrea frameworks exposed at Kesme. The tectonic setting of the Kas basins is that of a flexural foredeep with continual subsidence throughout the period of reef deposition (Hayward 1982a). Hayward (1982a and b) showed that reefs developed the complete morphology in areas of locally low clastic input. The possible reasons for the incomplete development of a similar coral zonation in the Antalya area are:

- Different tectonic setting, e.g. not such continual subsidence in the Antalya area;
- Higher clastic input in the Antalya area;
- Preservational differences.

### 3.8.3 Spatial morphology of reefs

Shallow-water carbonates in the Aksu and Köprü basins are generally exposed for less than a kilometre. Even when linear bodies of Miocene carbonates crop out over several kilometres, such as those seen north of Deniztepesi, in detail these lineaments consist of lenses of shallow-water carbonates which are separated from each other. In almost all cases they are interbedded with coarse clastics, often dominated by conglomerate and are interpreted as patch-reefs. Classically, reef generated carbonates were not thought to be associated with areas of high terrigenous sedimentation. Studies of modern reefs in Jamaica (Wescott and Etheridge 1980) and the Red Sea (Gwirtzman and Buchbinder 1978,
Hayward 1982a and 1982b) however have shown that this is not necessarily the case.

Hayward (1982b) studied Miocene coral reefs in the Kas basin to the west of the study area using the Red Sea reefs as a modern analogue. He suggested that coarse gravel sediments of alluvial fans provided ideal substrates for coral planulae to settle. He noted that material finer than 60mm was not directly colonised by coral in the Red Sea, and that Miocene reefs were not found in association with well developed claystone suggesting that they were subject to a similar control. He identified periodic run-off as a critical factor in reef development, particularly stressing that the influence of the one or two flash-floods a year experienced in the Red Sea would be restricted to the active portion of the fan. He concluded that lateral confinement of the fluvial system precludes reef development within channels due to the high sedimentation rates. Although this can be generally applied to the Miocene reefs of the Antalya area, Yesilbag, with its Porites within a fine grained channel is a notable exception.

In contrast to those of the Aksu and Köprü basins, the shallow-water carbonates in the north of the Manavgat basin occur in a continuous NW-SE striking, linear outcrop (Fig. 3.23). They directly overlie Alanya Massif basement here and very little associated terrigenous material. Further south however, coarse conglomerates underlie and are interbedded with the shallow-water carbonates. Very little reef framework is preserved anywhere in the Manavgat basin, but it is particularly sparse along the northern margin. The evidence for it having existed along this margin is strong however. Displaced blocks in talus horizons such as those exposed at the Akseki road section bear witness to its framework structure, but equally convincing is the large quantities of reef-derived carbonate material, particularly calcarenite, that line the basin margin. The Alanya Massif has been uplifted since the Lower Miocene and the shallow-water carbonates now dip at about 50 degrees to the south. The absence of fluvial conglomerates on top of the Alanya Massif where it forms the northern margin to the Manavgat basin indicates that it was a palaeotopographic high during the Lower Miocene and this is borne out by palaeocurrent measurements from the
Figure 3.23. Simplified geological map of the Manavgat basin showing the open folding in the south-east of the basin and the location of the principal localities mentioned in the text.
lowermost conglomerates in the south, none of which indicate flow from the Massif to the south (Fig. 5.33). It seems possible therefore that the reefs directly colonising the Alanya Massif may well have been fringing reefs. Elsewhere in the study area reefs are always interbedded with conglomerate and have limited lateral extent. These have been interpreted as patch-reefs. It is possible to draw the tentative conclusion that high sediment influx and/or processes associated with high sediment influx, may have been some of the factors controlling patch-reef development.

3.9 Conclusions

♦ Most of the shallow-water carbonates are interpreted as having formed as patch reefs interbedded with coarse terrigenous clastics. Reefs, no longer preserved, but deduced to have colonised the south facing margin of the Alanya Massif in the Manavgat basin however, may have been fringing reefs.

♦ Colonisation generally occurs on coarse conglomerates, but is also occasionally associated with silt to mud grade grainsizes.

♦ No distinct coral colonising community exists, but particular coral morphologies are associated with the basal colonies, e.g. flat dish-shaped forms and squat reticulate cones.

♦ Porites is the dominant coral throughout the Miocene of the study area. Minor Tarbellastrea and Montastrea are important in particular reefs.

♦ The full coral zonation of Hayward (1982a and b) is not observed.

♦ Sediment influx and associated processes may have controlled the development of patch reefs.

♦ Reefs topography have had some influence on sediment dispersal patterns.
Porites bafflestones development suggests that relative sea level rise > sediment influx.

A hiatus in shallow-water carbonate production exists spanning the Mid-Langhian to Lower Tortonian.

The eustatic sea level curve of Haq et al. (1988) is not sufficiently high resolution to explain termination of the Lower Miocene reefs in terms of an instantaneous global sea level rise.
Chapter 4

SEDIMENT GRAVITY FLOWS AND THE KARPUZÇAY FORMATION

4.1 Context

Sediment gravity flows deposited units of coarse conglomerate to siltstone grade material (the Karpuzçay Formation) between Langhian and Messinian times. These stratigraphically overlie the shallow-water carbonates described in chapter 3 and indicate a basinward shift in facies. Palaeocurrents, and grain-size changes within the Karpuzçay Formation indicate the geometry of sedimentation patterns and basin morphology in each of the three basins. This has implications for the ultimate controls of the processes of basin formation and evolution.

4.2 Organisation of this chapter

Following a brief outline of the spatial and temporal distribution of the Karpuzçay Formation, a summary of the previous work on these sediments is given in section 4.4. The texture and composition of the Karpuzçay Formation is described and interpreted for each of the three basins taking palaeocurrent directions into account (section 4.5). In the light of this information a comparison of the sediments exposed in the three basins leads to the subdivision of the Karpuzçay Formation into the Beskonak and Taskesigi Members (section 4.6). The discussion is then extended to include sedimentation trends resulting from the correlation of Miocene sequences across the Isparta Angle are noted in section 4.7, prior to the conclusion of the chapter (section 4.8).

4.3 Temporal and Spatial distribution

The majority of the Karpuzçay Formation sediments are exposed on the eastern side of the Köprü basin and in the central part of the Aksu basin. These sediments are mainly Serravallian in age although rare older (Langhian and Upper Burdigalian) and younger (Lower Tortonian)
nannoplankton dates have been derived by previous authors. Exposure of the Karpuzçay Formation is abundant throughout the Manavgat basin, but has a longer age span (Langhian/Lower Serravallian to Upper Messinian). The place names used in the text are indicated on figure 4.1.

4.4 Previous work

Blumenthal (1951) divided the Mio-Pliocene succession found between Alanya and the Köprü Çay into five formations: Conglomérats de base (chapter 5); Calcaire récifal "bordier" (chapter 3); Marnes de Gecereme (Blumenthal 1947; section 3.6.4.1); Molasses gréso-marneuse and lying discordantly above these the Pliocene Conglomerats de Belkis. Monod (1977), studying the same area and using Blumenthal’s 1951 classification improved the dating of these formations and as a result modified the classification so that, Les Marnes de Gecereme, and a new formation, La Molasse de Manavgat, were both grouped under the heading of La Molasse Miocene de Manavgat. Monod noted that the pure pelagic sediments (e.g. Marnes de Gecereme) found near the village of Gençler close to the Akseki road section in this study, (table 2.5) were a very restricted facies evolving rapidly into the sediments of the "Molasse". He describes the "Molasse" as being a thick (1000-2000m?) succession of rhythmically-bedded sandstones and marls with lenses of conglomerate. He also indicated a general trend in the grain-size from north to south with the coarsest beds being found to the north close to Beskonak (Fig. 4.1). Using the Oymapinar and Alarahan sections in the Manavgat basin (Fig. 4.1), Monod identified the following foraminiferal zones in the "Molasse". (N.B. zone stages identified using Berggren et al., 1985):

- *Orbulina suturalis* (Upper Langhian);
- *Globorotalia mayeri* (Serravallian);
- *G. menardii* (Serravallian-Tortonian boundary);
- *G. dutertrei* (Messinian).

In his thesis, Monod (1977) notes the need for further study of the Molasse in the Köprü basin, but leaves the Miocene sediments found there undifferentiated.
Figure 4.1. Location map of the Karpuzçay Formation exposures referred to in the text.
The work published by Bizon et al. (1974) concentrated on the south of the Isparta Angle. They dated the Alarahan and Oymapinar sections (table 2.5) in the Manavgat basin, and the base of the Miocene succession near Aspendos in the south of the Köprü basin (Fig. 4.1).

Poisson (1977) used planktic foraminifera to date a semi-continuous section through the Karpuzçay Formation in the Aksu basin near the Kargi baraj (Fig. 4.1). This section overlies Lower Miocene shallow-water carbonates and spans the Serravallian to Lower Tortonian.

Akay & Uysal (1985) and Akay et al. (1985) classified "shale-sandstone-conglomerate alternation with occasional volcanic tuff interbeds" as the Karpuzçay Formation. The type section Akay & Uysal (1985) used for the Karpuzçay Formation is in the Manavgat basin, located along the road from the coast towards Akseki (Ahmetler section; Fig 4.1; table 2.5; sections 4.6 and 2.4; table 2.3). The formation is named after the river Karpuzçay, above which the road runs and means literally the "watermelon" river formation.

Sections within the Aksu and Köprü basins are often structurally complex such that the resultant dating of this formation in these two basins is rather patchy. The Manavgat basin sections are simpler and generally well dated. Figure 4.2 shows the locations of the sections dated by Akay & Uysal (1985), Akay et al. (1985), Bizon et al. (1974) and Poisson (1977), and the dates that they derived from the study of planktic foraminifera. The salient features of this diagram are listed below:

- The Karpuzçay Formation in the Aksu basin spans Burdigalian to Messinian times. Exposures are generally younger in the north of the basin, although rare Messinian dates are derived from the far south, juxtaposed against the oldest turbidite exposures in this structurally complex area.
- All dated sections in the Köprü basin are Latest Burdigalian-Langhian (NN4-NN5).
- The Karpuzçay Formation in the Manavgat basin spans the Langhian-Messinian (NN5-NN11b).
Figure 4.2  Location and dates of the Karpuzçay Formation sections and samples from Poisson (1974), Bizon et al. (1974), Akay (1985) and Akay et al. (1985). The ages of the nannoplankton zones are taken from Berggren et al. (1985).
Flecker et al. (1995) date the Ahmetler section using nannoplankton. Over 700m of interbedded conglomerates, sandstones, siltstones and marls show progressive younging from the north (Burdigalian/Langhian; section 3.3) to south (Messinian).

4.5 Facies description

4.5.1 Aksu Basin

The Karpuzçay Formation outcrop is concentrated towards the centre of the Aksu basin, to the north of the Kargi baraj (Fig. 4.1). The best exposure occurs in recent road cuttings along the new road north of the Kargi baraj towards Isparta, but due to the steep slopes of these cuttings and the rather crumbly nature of the rock, colluvium formation is rapid and the superb clarity of the turbidite exposure here will probably not be visible in a couple of years time.

Where the base of the Karpuzçay Formation is exposed, it overlies thin, laterally discontinuous Langhian shallow-water carbonates (Oymapinar Limestone; chapter 3) or passively onlaps Mesozoic basement limestone. The relationship of the lower contact of this formation is only observed in the west of the Aksu basin. The upper contact of the Karpuzçay Formation appears to be gradational into Tortonian conglomerates and sandstones with interbedded, small patch reefs (Aksu Formation; chapter 5).

4.5.1.1 Texture and composition

Thickness:
The thickness of the individual beds in the Karpuzçay Formation of the Aksu basin vary from a few centimetres to nearly a metre. Most of the thickest beds are either coarse sandstones and conglomerates or massive siltstones.

Bed shape:
Parallel sided geometries dominate the bed-shapes of the Aksu basin Karpuzçay Formation (Figs. 4.3, 4.4, 4.5 and 4.6). Some parallel bedded
Figure 4.3 Photograph of overturned sandstone and siltstone beds of the Karpuzçay Formation, 1.7km north of the Kargi baraj site, central Aksu basin. (Younging direction to the right).

Figure 4.4 Photograph of Karpuzçay formation turbidites 2.7km north of the Kargi baraj, central Aksu basin. The exposure here indicates a transition from silt dominated turbidites to more sand rich turbidites and back to silt dominated ones again. (The car is pointing in the younging direction.)
Figure 4.5  Log of the part of the succession shown in figure 4.3, Kargi baraj, central Aksu basin.
Figure 4.6 Log of the succession shown in figure 4.4, Kargi baraj, central Aksu basin. For key see figure 4.5.
sand units can be traced laterally over a hundred metres where the exposure permits. Gradual changes in bed thickness were observed in association with beds displaying basal scour. However, complete wedging out of beds was not observed on the scale of 100m.

*Grain size:*
The grain-size varies from relatively rare occurrences of fine-grained conglomerates with pebbles up to a few centimetres in diameter, to rare mudstones. Most Karpuzçay Formation exposures in the Aksu basin are siltstone dominated, but examples of equal sandstone and siltstone domination are exposed just to the north of the Kargi baraj site (Figs. 4.4 and 4.6).

*Composition:*
Terrigenous clasts within the conglomerates include: red and green chert; basement limestone (fawn, white and grey) and rather rare igneous fragments which are generally serpentinized basalts. The sandstones contain similar material (Appendix 4). Biogenic components are dominated by plant debris, but concentrations of shell fragments are also common. In thin section these are overwhelmingly dominated by mollusc debris including oyster fragments although brachiopod fragments have also been identified. Rare entire gastropods are found particularly within the coarser sands and conglomerates. The micropaleontological content of the coarser-grained units is low, but planktic foraminifera and nannoplankton have been identified in the siltstones and mudstones.

*Sorting:*
The terrigenous components of individual units tend to be moderately to well sorted except in the particularly structureless sandstone-conglomerate beds. Biogenic debris generally forms larger clasts than the terrigenous components and this is particularly obvious towards to top of fine sand units where plant debris is concentrated. Concentrations of shell debris appear both as coarse clasts in the basal part of massive sandstones and as winnowed lags associated with laminated and rippled sandstones. Very occasionally, shell fragments are found scattered rather randomly through more massive siltstones.
Grading:
Most of the sand-grade beds display normal grading. Inverse grading is occasionally observed in the lower parts of sandy conglomerates. The thicker bedded siltstone to mudstone units, by contrast, tend not to show grading. However, some of the thickest siltstone units, on detailed inspection are found to be made of millimetre scale silt-mud couplets with which subtle normal grading is associated.

Clast alignment:
Abundant plant debris, generally found towards the top of sandstone units shows distinct bedding parallel alignment, generally with an orientation within that plane. Imbrication within the conglomerates is also relatively common and rose diagrams from measurements on imbricated pebbles are discussed in section 4.5.1.3.

Structures:
Parallel laminations occur commonly in both sandstones and siltstones.

Cross laminations are also common. Most of the cross lamination is seen in section and appears to have a single flow direction. Where rippled surfaces are exposed, as at the junction between the Kargi-Isparta road and the road east from Bucak (11km north of the Kargi baraj site; Fig. 4.1) however, the evidence is equivocal and some of the ripples may be wave rather than current generated. Palaeocurrent measurements taken from the ripples in this locality are discussed in section 4.5.1.3. Climbing ripples are also found, but are much more rare.

Both the laminations and ripples are picked out generally by grain-size differences although occasionally compositional variations in the abundance of plant fragments pick out parallel laminations.

Water escape structures are found associated with sandstones passing rapidly into siltstones. They are generally small and rare.

Burrows, both horizontal and vertical occur throughout the Karpuzçay Formation although concentrations are more obvious both at the very
tops and bottoms of beds. Thick beds where burrows are visible on the bottom of the unit often completely lack structures further up. This may indicate that intensive bioturbation has completely destroyed all structures, save its own traces at the very bottom, where the burial level was potentially too deep for extensive biogenic activity. Concentrations at the top of beds may indicate that sufficient time elapsed between the deposition of one unit and the next for burrowing to occur, but not for it to have caused the destruction of all structures including its own traces. Horizons with well preserved sedimentary structures, but no burrows may indicate periods of low oxygen content both in the upper centimetres of the sediment and in the water just overlying it (Chamberlain 1978; Plaziat, 1984).

**Tool marks**, generally grooves or prods are found on the bases of many of the coarser sandstone units particularly where they overlie siltstones. Palaeocurrent directions measured from the tool marks in the Aksu basin are discussed in section 4.5.1.3.

**Slumps** are found throughout the Karpuzçay Formation and indicate an east-dipping palaeoslope to the basin.

### 4.5.1.2 Processes of deposition

The features of the sediments described above can be interpreted terms of the mechanisms of sediment gravity flows (Middleton and Hampton, 1973). Some of them could also be interpreted in terms of storm deposits. Aigner and Reineck (1982) distinguished shelf storm sand layers from turbidites using the following criteria:

- Wave-ripple cross-lamination;
- Wave-rippled top surface;
- *In situ* shallow-marine shelf fauna in interbedded shale layers;
- Marked increase in the bioturbation of storm sand layers from proximal to distal settings;
- Association with shallow-water facies.
Wave-rippled surfaces only occur in one locality in the Aksu basin. Here, they represent rippled top-surfaces of sandy units in a sequence which is significantly coarser than the Karpuzçay Formation elsewhere in the basin. This sequence may indeed represent at least partial deposition by storm-generated currents. Other successions studied in the Aksu basin however, failed to reveal structures characteristic of storm deposits. Hummocky cross-stratification (e.g. Dott and Bourgeois, 1982) for example, was not observed in the basin. Shallow-water fauna are only found as lags in the base of sandstone units, never in situ in the finer-grained layers. The absence of these typical storm structures suggests that although storm processes may have affected some of the deposits in the Aksu basin, there is little evidence that they were the dominant mechanism of deposition.

A brief summary of the depositional processes involved in sediment gravity flows and the formation mechanisms of the characteristic structures observed in Karpuzçay Formation of the Aksu basin, is given below.

Bouma (1962) collected detailed information of the internal characteristics of turbidites from the Maritime Alps in southern France. The Bouma sequence (Fig. 4.7) represents a summary of this information. The sequence was interpreted in relation to the progression of bedforms observed during decelerating flow in flumes (Harms and Fahnestock, 1965; Walker, 1965). Further work suggested that the Bouma sequence is a good model only for medium-grained turbidites deposited by the deceleration of low concentration turbidity currents.

The classification of sediment gravity flows by Lowe (1979) is based on flow rheology (fluid or plastic) and particle support mechanism (turbulence, pore fluid, dispersive pressure and matrix support; Fig. 4.8). In turbidity currents, grains are suspended by the turbulence of the flow and must therefore be treated in terms of several size populations (see Lowe, 1982 for review). Clay to medium-grained sand particles can be suspended by fluid turbulence alone, largely independent of the concentration and hence can be transported by low-density flows (Pantin, 1979). The vast majority of the turbidites observed in the Aksu basin, fall
<table>
<thead>
<tr>
<th>GRAIN SIZE</th>
<th>BOUMA (1962) DIVISIONS</th>
<th>INTERPRETATION</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mud</td>
<td>E Laminated to homogeneous mud</td>
<td>Deposition from low-density tail of turbidity current: settling of pelagic or hemipelagic particles</td>
</tr>
<tr>
<td>Silt</td>
<td>D Upper mud/silt laminae</td>
<td>Shear sorting of grains &amp; flocs</td>
</tr>
<tr>
<td>Sand</td>
<td>C Ripples, climbing ripples, wavy or convolute laminae</td>
<td>Lower part of lower flow regime of Simons <em>et al</em> (1965)</td>
</tr>
<tr>
<td></td>
<td>B Plane laminae</td>
<td>Upper flow regime plane bed</td>
</tr>
<tr>
<td>Coarse Sand</td>
<td>A Structureless or graded sand to granule</td>
<td>Rapid deposition with no traction transport, possible quick (liquefied) bed</td>
</tr>
</tbody>
</table>

Figure 4.7  Idealised sequence of sedimentary structures in a turbidite bed (after Bouma, 1962, with interpretation after Harms and Fahnestock, 1965; Walker, 1965; Walton, 1967; Stow and Bowen, 1980).
<table>
<thead>
<tr>
<th>Flow behaviour</th>
<th>Flow type</th>
<th>Flow character</th>
<th>Sediment support mechanism</th>
</tr>
</thead>
<tbody>
<tr>
<td>Fluid</td>
<td>Fluidal flow</td>
<td>Turbidity current</td>
<td>Fluid Turbulence</td>
</tr>
<tr>
<td></td>
<td>Fluidized flow</td>
<td></td>
<td>Low density turbidity current</td>
</tr>
<tr>
<td></td>
<td>Liquified flow</td>
<td>Escaping pore fluid (full support)</td>
<td>High-density turbidity current</td>
</tr>
<tr>
<td>Plastic</td>
<td>Debris flow</td>
<td>Grain flow</td>
<td>Dispersive pressure</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Mudflow or cohesive debris flow</td>
<td>Matrix strength/matrix density</td>
</tr>
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<td></td>
<td></td>
<td></td>
<td>Mud flow</td>
</tr>
</tbody>
</table>

Figure 4.8. Nomenclature of laminar and turbulent sediment gravity flows based on flow rheology and particle support mechanisms (modified after Lowe, 1979).

Figure 4.9 Origin of traction carpet layers, with sketches from Lowe (1982) and explanation based on Hiscott and Middleton (1979, 1980): a) Grain interaction at the base of a high-concentration turbidity current produces inverse grading of the flow. B) Continued fallout from suspension produces a dense traction carpet that is supported by dispersive pressure and sheared by the main body of the flow above. c) The upper part of the traction carpet freezes (vertical segment in the velocity profile) with shear being concentrated at the base of the carpet. d) The entire sheared layer ceases to move and becomes welded to the bed. Repetition of this process forms a division of inversely-graded stratification.
within this grainsize population and many of the structures with which they are associated can be formed during the deceleration of a low-density turbid flow (see below). Coarse sand to small pebble-sized gravel however, cannot be entirely suspended within a dilute flow. In high density flows where there is a wide range of particle sizes, the coarser grains can be supported by a combination of turbulence; hindered settling due to their own high concentration and buoyant lift caused by the interstitial mixture of water and finer-grained sediment. Pebble to cobble-sized clasts in concentrations >10-15% can also be supported by dispersive pressure from grain collision (Lowe, 1982). Although the quantity of coarse grained material within the Karpuzçay Formation in the Aksu basin is relatively small, none-the-less, different depositional mechanisms can be inferred from their presence and internal structures.

The process of deposition is mainly a function of a) flow concentration or flow density and b) rate of deposition. Generally more than one process (Traction sedimentation, suspension sedimentation, frictional freezing and cohesive freezing) is operative either synchronously or serially during the deposition of an entire unit. Deceleration of a low-density turbid flow, for instance, results in the passage of sediment from suspension to bed load. Sediment is then deposited by traction sedimentation. Bedforms resulting from the combination of these two processes include parallel lamination, ripples and normal grading, features that are characteristic of Bouma division B, C and D (Fig. 4.7) and are commonly found in the Aksu basin turbidites.

High-density turbidity currents with a coarse sand-gravel population, may deposit some load as a sand bed in a slightly unsteady, but fully turbulent flow state. These sediments show traction-related structures, e.g. flat lamination or oblique cross-stratification and high-density turbidity currents are locally erosive, causing scour structures. As flow unsteadiness increases, suspended load becomes increasing concentrated towards the base of the flow, and, particularly in coarser size fractions, a basal layer is formed and maintained by dispersive pressure due to grain collision. This layer is known as a traction carpet (Dzulynski and Sanders, 1962) and leads to the formation of inversely graded coarse-grained layers.
When it freezes, due to continued sediment fallout from the overlying flow, a new carpet forms on top and the process is repeated (Fig. 4.9).

In cohesive mudflows (or debris flows), grains are supported by the cohesiveness and buoyancy of a sediment-water matrix rather than by turbulence. In some of these flows, the clasts are not truly suspended in the matrix, but the buoyancy and lubrication (preventing frictional locking) provided by the sediment-water mix, allows them to bounce and roll down-slope whilst remaining more or less in contact with other clasts. The features produced by such flows are variable and may depend on whether the flow was turbulent at some point during deposition (Enos, 1977). Chaotic units where larger clasts are at least partially suspended in a finer-grained matrix are generally attributed to debris flow processes.

Tables 4.1 and 4.2 indicate the range of structures observed in the Aksu basin. Using the classification scheme of Pickering et al. (1989) which is a modified version of the much earlier work by Mutti and Ricci Lucchi (1972), the Karpuzçay Formation has been divided into 6 sub-facies and the probable depositional processes and current types have been identified.

4.5.1.3 Palaeocurrent analysis

Figure 4.10 displays representative rose diagrams measured in the Kargi area plotted against a generalised section which has been derived, in part from Poisson (1977a). A clear bimodal current distribution can be seen; broadly to the east and to the SSE. On closer examination the east and ESE directed current measurements are derived from pebble imbrication, whilst the south to SSE directed palaeocurrents are derived from tool marks. The limited slump data collected from the area indicate that there was a palaeoslope dipping towards the east. Palaeocurrent directions measured on six rippled surfaces in the turbidites at the Bucak junction with the new Isparta-Kargi baraj road indicate a dominantly east directed flow with a subordinate westerly flow direction. As discussed earlier, cross-sections through these rippled surfaces did not indicate
Table 4.1  Description and classification of medium-sand to clay grade units from the Aksu, Köprü and Manavgat basins, in terms of low- and high-density turbidity currents (after Pickering et al., 1989).

<table>
<thead>
<tr>
<th>Structures</th>
<th>Sub-facies (Pickering et al. 1989)</th>
<th>Bouma divisions (Bouma 1962)</th>
<th>Depositional process</th>
<th>Current type</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tool marks, scour, water-escape, normal grading, parallel laminations, ripples.</td>
<td>C2.1 sand-mud couplets</td>
<td>A-D</td>
<td>Grain-by-grain deposition from suspension. Tractional transport active in Bouma division B and C</td>
<td>High concentration turbidity flow</td>
</tr>
<tr>
<td>Generally parallel bedded, normal grading, parallel laminations, ripples</td>
<td>C2.2 sand-mud couplets</td>
<td>B-D</td>
<td>Grain-by-grain deposition from suspension. Tractional transport active</td>
<td>Intermediate strength turbidity flow</td>
</tr>
<tr>
<td>Generally parallel bedded, normal grading, parallel laminations in silt</td>
<td>C2.3 sand-mud couplets</td>
<td>C-D</td>
<td>Grain-by-grain deposition from suspension. Tractional transport active</td>
<td>Relatively dilute turbidity flow</td>
</tr>
<tr>
<td>Silt-mud laminations, gradational tops and bases</td>
<td>D2.3 Thin silt and mud laminae</td>
<td></td>
<td>Slow uniform deposition from suspension with shear sorting</td>
<td>Low concentration turbidity flow</td>
</tr>
</tbody>
</table>
Table 4.2 Description and classification of coarse sand and conglomerate units from the Aksu, Köprü and Manavgat basins in terms of high-density turbidity currents (after Pickering et al., 1989).

<table>
<thead>
<tr>
<th>Structures</th>
<th>Sub-facies (Pickering et al. 1989)</th>
<th>Bouma divisions (Bouma 1962)</th>
<th>Depositional process</th>
<th>Current type</th>
</tr>
</thead>
<tbody>
<tr>
<td>Inverse grading, imbrication, massive, water escape, rippled silts at top</td>
<td>B2.1 Parallel-stratified sands</td>
<td>Freezing of traction carpets at base. Intense grain interaction producing imbrication and inverse grading</td>
<td>High concentration turbidity flow</td>
<td></td>
</tr>
<tr>
<td>Thicker bedded, flat based, poorly sorted, mudstone intraclasts common</td>
<td>A1.1 Disorganised gravel</td>
<td>freezing on decreasing slopes due to intergranular friction and cohesion</td>
<td>High concentration turbidity currents or debris flows</td>
<td></td>
</tr>
</tbody>
</table>
Figure 4.10  A schematic log of the Kargi baraj section, central Aksu basin with palaeocurrent rose diagrams. The ages are derived from Poisson (1977).
unequivocally what sort of current deposited them. The bimodal ripple direction suggests that they were formed by wave action.

Interpretation of the palaeocurrent data can be listed as follows:

- There is no evidence to suggest that the east margin of the Aksu basin was a major sediment source;
- Tool marks on the base of sandstone-siltstone units suggest that the turbidites flowed axially, from north to south;
- Imbrication in conglomeratic units near the Bucak junction indicate derivation from the western margin;
- Wave ripples are associated with the conglomeratic sequence.

4.5.1.4 Temporal and spatial variations

Dating of this formation in the Aksu basin, due to intense folding and large scale faulting requires detailed sampling and mapping on a scale which was not compatible with the template of the project. Thus, the discussion of temporal relationships within the formation is limited either to broad generalities or relationships visible within a single continuous exposure.

Where redeposition of material is a dominant process in the formation of a sedimentary unit, (e.g. turbidites) biostratigraphic data should be treated with caution. In this case, although the biostratigraphic template is based on nannoplankton which are notoriously easily reworked, the younging to the north trend is likely to be real even if the zones identified in individual samples (particularly the older samples), are inaccurate for the date of deposition of the turbiditic unit. It is therefore assumed that the age of the Karpuzçay Formation in the Aksu basin spans the Late Langhian/Early Serravallian- Lower Tortonian.

The timing of input of the Bucak junction conglomerates from the western margin occurs in the Serravallian. Poisson (1977a) indicates that these conglomerates pass upwards into finer turbidites which are also Serravallian in age, suggesting that the western source of coarse material was both developed and abandoned in a relatively short space of time.
Interpretation of the coarse, often matrix-supported conglomerates in this succession is discussed more fully with reference to the Tortonian and Messinian succession in the Manavgat basin, where this type of deposit is more common (section 4.5.3.2). A tentative interpretation of the episodic nature of the conglomeratic sequence seen at the Bucak junction is that it represents shelf deposition of an ephemeral fan-delta lobe. On a smaller scale (i.e. 100m) a similar pattern of periodic coarse influx can be seen in the finer-grained units near Kargi baraj (Figs. 4.4 and 4.6). This succession may document the formation and abandoning of a sand lobe such as those discussed by Stow (1984).

The Karpuzçay Formation in the north of the Aksu basin is significantly richer in conglomerate than in the south (Fig. 4.11). This may indicate a proximal to distal relationship within the basin from north to south and the north-to-south direction of axial flow inferred from the turbidite palaeocurrents would support this interpretation. However, given the younging trend, the distribution of conglomerate may also represent a shallowing of the basin through time. In this scenario, axial flow from north-to-south would be maintained throughout the Serravallian, but a relative fall in sea level would cause increasingly proximal facies to be deposited near the source area, in the north. The continental and shallow-marine Tortonian facies in the north of the Aksu and Köprü basins discussed in chapter 5 and foraminiferal evidence from the Manavgat basin, support the interpretation of Serravallian shallowing throughout the Isparta Angle.

One further possibility, however, is that the conglomerate-rich Karpuzçay Formation represents a period of active deposition on a fan-lobe, much like that seen in the centre of the basin at the Bucak junction.

4.5.2 Köprü Basin

The outcrop of the Karpuzçay Formation is mainly confined to the east of the basin between the river Köprü and the Kirkkavak fault and to the south of the village of Yesilbag. An important exception to this is the exposure which occurs in the south west of the Köprü basin in the region of the village of Akbas (Fig. 4.1).
Figure 4.11 Logs of typical Karpuzçay Formation sections from the north and south of the Aksu basin.
Where the base of the Karpuzçay Formation is exposed, it overlies laterally discontinuous Langhian shallow-water carbonates (Oymapınar Limestone; chapter 3) and coarse conglomerates and sandstones (Kızıldag Formation; chapter 5) sometimes with a disconformable relationship. The base of the formation is only observed in the south-west of the Köprü basin, near Aspendos and Deniztepesi and in the north where the basin bifurcates, near Ballibucak. The upper contact of the Karpuzçay Formation is exposed along the road north from Beskonak to Kesme near the village of Yeşilbag and appears to be rapidly gradational into Tortonian conglomerates and sandstones with interbedded, small patch reefs (Aksu Formation; chapter 5).

4.5.2.1 Texture and composition

Thickness:
Conglomerate beds in the Köprü basin vary in thickness from a few centimetres at the base of fining upwards sequences, to 50cm thick massive units. Sandstone beds tend to be thinner, generally between 5 and 20cm although much thicker beds are occasionally found. An example of a 75cm thick massive sandstone horizon was observed on the road to Karabucak. Channelised sandstone and conglomerate units such as those located near the junction between the road to Karabucak and the road from the coast to Beskonak are generally in the order of 1-2m thick. Siltstone and mudstone horizons are interbedded on a centimetre scale.

Bed shape:
The vast majority of the Karpuzçay Formation in the Köprü basin consists of parallel bedded units. However, lenticular geometries are far more common in the Köprü basin than in the Aksu basin. Such geometries include: wedging out of coarse horizons (e.g. the micro-conglomerates at Duzagaçköy); cross-sections through channels (e.g. near the junction between the road to Karabucak and the road from the coast to Beskonak, Fig. 4.12) and rare cross-cutting relationships on a scale of a hundred metres, indicating transverse sections through channels. Most of these bedforms are associated with conglomerate and coarse sandstone.
Figure 4.12 Photograph of a channel in Karpuzçay Formation, central Köprü basin.

Figure 4.13 Photograph of rip-up clasts in a sandstone unit from the Karpuzçay Formation, central Köprü basin.
However, at Çayıçi in a relatively sandstone poor succession, fine sandstone is confined within narrow channels approximately 1m across.

Grain size:
A few broad generalisations can be made concerning the overall grain-size of the turbidites in the Köprü basin:

- The Akbas syncline area in the west of the basin contains fine grained turbidites, with very rare conglomerates. These have clasts up to about 4cm in diameter and a mean clast size of 0.5cm;
- The eastern margin of the Köprü basin is also generally conglomerate poor, but the percentage of sand seems to vary unsystematically from 5-90%.
- Conglomerate-rich units are concentrated in the centre of the basin in association with channel geometries. Here maximum clasts size varies from 5-10cm with a mean of 2-4cm. Rare boulders up to 50cm in diameter are found in a poorly sorted conglomerate to the north west of Beskonak.

Composition:
The clast types found in the conglomerates of the Köprü basin are more varied than those found in the Aksu basin. They include: red, green, black and diagenetic grey chert; pale grey limestone; dark grey limestone with gastropods; white fossiliferous limestone; fawn recrystallised limestone; brown sandstone; green silt; calcarenite; igneous fragments including serpentinite and weathered basalt and intraformational rip-up clasts of mud and silt. Abundant, large rip-up clasts of grey-green silt are also found at the base of sandstone units such as those found near the junction of the road to Kepez and Mütülü with the road to Beskonak (Fig. 4.13).

The terrigenous components of the sandstones are broadly similar to those of the conglomerates, but the sandstones also contain individual quartz grains making up about 25% of the total number of grains (Appendix 4b). Micrite is also a fairly abundant component constituting up to 45%.

Biogenic detritus is a significant component of many sandstone and some conglomerate horizons. One of the most fundamental differences
between the composition of the Aksu basin and the Köprü basin is the quantity and preservational state of the wood and plant fragments. In the Aksu basin the turbidites contain brown plant material, generally concentrated in particular horizons. In the Köprü basin the quantity of organic material is much greater and consists of coal fragments. This may reflect differences in both the nature of the organic material (i.e. more wood rather than leaf detritus influxed into the Köprü basin) and in the burial depth between the two basins. Figure 4.14 is located a few kilometres to the north of Beskonak and shows a siltstone horizon containing a large wood clast (now coal) which has been copiously bored. Organic fragments of various sizes are found in horizons of all grain sizes.

Large pieces of coral included as clasts in conglomerate are also relatively common in the Köprü basin. The most common species found is *Porites sp.*, but *Heliastrea* (Vindobonian, Early-Mid Miocene, Chevalier 1961) has also been identified (Appendix 3c). Shells and shell fragments are concentrated in particular horizons the grain-size of which varies from conglomerate to siltstone. At Duzagaçköy, a shell-rich sandstone contains pectens, oysters, bivalves, gastropods and small bored limestone clasts, whilst 7% of clasts in a pebble conglomerate 50m higher up the section are oysters. A further 2% of clasts in this horizon are gastropods (Appendix 4a).

**Sorting and physical maturity:**

The physical maturity of the conglomerate clasts in the Köprü basin is sometimes difficult to assess accurately as it is, to some extent, dependent on the relative quantities of chert and limestone. Angular conglomerates rich in chert clasts are not uncommon, but angular limestone conglomerates such as the 50cm thick horizons seen in the Dedek section are rare. Most conglomerates and sandstones contain clasts and grains which are subangular-subrounded.

Sorting of conglomerates varies, but is on the whole poor to moderate. Matrix supported conglomerates are rare, whilst clast-supported conglomerates with abundant matrix are the norm. Grain sorting in the
Figure 4.14 Photograph of a bored coal fragment in a Karpuzçay Formation siltstone unit a few kilometres north of Beskonak, central Köprü basin.

Figure 4.15 Photograph of a slumped horizon overlain by a channelised sandstone, overlain by a debris flow unit containing a detached block of Alanya Massif metacarbonate from the Karpuzçay Formation, 1km south of the Ahmetler junction, north-eastern Manavgat basin.
sandstones tends to be moderate to good with the exception of larger clasts near the base of beds.

**Grading:**
Almost all the grading seen in the basin is normal. Conglomerates located along the Kepez section contain rare examples of inverse grading at the base overlain by subsequent normal grading.

**Clast alignment:**
As in the Aksu basin, imbrication in the conglomerates is relatively rare. Alignment of organic fragments is less common in the Köprü basin than in the turbidites in the Aksu basin, perhaps due to the larger size of the organic clasts.

**Structures:**
Parallellaminations in both sandstones and siltstones are common and are well exposed in sandstone near Gökceler in the Akbas syncline area;

Cross laminations associated with both straight crested and crescent-shaped ripples are common in sandstones. Mega-ripples with a ripple height of 5cm and a wave length of approximately 20cm are found in one locality, a few kilometres south east of the Köprü bridge on the road to Yesilbag. Cross-sets in sandstones at the top of channelised units such as those along the Karabucak section can be up to 1.5m in height;

Water escape structures were observed infrequently;

Bioturbation is abundant in all but the coarsest conglomerates. Both horizontal and vertical tubes are common.

Tool marks found in the Köprü basin include grooves, prods, more rare flutes, load balls and sludge marks.

Slumps are abundant particularly along the eastern margin of the basin. They vary in thickness from a few centimetres to 10m.
4.5.2.2 Interpretation

The Karpuzçay Formation exposed in the Köprü basin closely resembles the succession in the Aksu basin and can be interpreted in terms of high and low-density turbidity currents in a similar way (section 4.5.1.2; tables 4.1 and 4.2).

4.5.2.3 Palaeocurrent analysis

Figure 4.16 shows the locations of representative rose diagrams from the Karpuzçay Formation in the Köprü basin. Slumps are more common in the Köprü basin than in the Aksu basin, but are concentrated along the eastern margin, marked today by the Kirkkavak fault (Fig. 4.16). In the north of the basin the slump directions indicate a palaeoslope dipping towards the west, whilst the main transport direction of the turbidites in the central part of the basin is indicated by measurements of the tool marks which show north-south movement. Sandstone channels associated with these turbidites are exposed in three dimensions on a scale of 3-5m in central parts of the basin away from the margins and these are also orientated north-south. Towards the south of the basin however a marked change in palaeocurrent direction is indicated. Grooves measured within 2 kilometres west of the Kirkkavak fault close to the village of Kepez have NNW-SSE direction (Fig. 4.16). A few kilometres further south, channels exposed in road cuttings are orientated NW-SE. Slumps measured just to the east of the Kirkkavak fault near the village of Karabucak show a wide variety of directions, but there is a conspicuous absence of slumping towards the north.

Although the slump data in the north of the basin indicate that there was a west facing slope close to the present day eastern margin of the Köprü basin, there is no related change in the grain-size of the turbidites from east to west. Thus, as in the Aksu basin the major source of turbidites seems to have been to the north.

The change in direction from axial flow documented by the groove marks in the north of the basin to flow directed to the south-east, towards and over the present day basin margin (Kirkkavak fault) in the south
Figure 4.16 A map of the Köprü and Manvagat basins showing the location and orientation of some directional structures in the Karpuzçay Formation units.
indicates either that the Kirkkavak fault did not exist at the time of deposition or, if it did, that it had no marked topographic expression. At any rate the south of the Köprü basin appears to have been overfilled during the Early Miocene allowing turbidites to spill out towards the south east.

There is little evidence to suggest that the Kirkkavak fault was overstepped by turbidites along its entire length, although it is possible that due to subsequent uplift at the end of the Miocene (the Aksu Phase; chapter 7) much of what was deposited to the east of the Kirkkavak fault has been removed as a result of footwall uplift. Miocene outcrop east of the Kirkkavak fault occurs in only two places (Fig. 4.16), both in the far north close to a palaeo-incised valley which breaches the Kirkkavak Fault (chapter 5; Fig. 5.16). The exposures at Dumanli, though of similar age (Serravallian; Appendix 3.A.2) constitute shallow-water carbonates and conglomerates interbedded with marls that are rich in benthic foraminifera. The other exposure located between the villages of Çetmi and Gencek consists of a single small bank outcrop of marls containing no nannoplankton and abundant large oysters. Both these localities are discussed further in chapters 3, 5 and 6.

4.5.2.4 Spatial and temporal variations

Akay’s M.T.A. report (1985) indicates that the Karpuzçay Formation in the Köprü basin is generally older and represents a much shorter time span (uppermost Burdigalian- Langhian) than that found in the Aksu basin (Langhian to Tortonian, Fig. 4.2). Nannoplankton analysis of sections in the south west of the basin (e.g. Aspendos and Deniztepesi; Fig. 4.1; C. Müller pers. com., 1994) suggest that here, Langhian shallow-water carbonates pass up into Early Serravallian turbidites (Appendix, 3.A.2; Fig. 4.32) thus extending the time period over which Karpuzçay turbidites are thought to have been deposited in the Köprü basin. Once again, because of the redeposited nature of the material, these dates must be regarded as providing an estimate of the maximum age of deposition (section 4.5.1.3). Potentially however, if the older samples do reflect Burdigalian-Langhian deposition of the Karpuzçay Formation in the east of the basin coeval
with shallower-water carbonates deposition in the west, this indicates an asymmetric geometry to the Köprü basin at this time (Fig 4.17).

Monod (1977) suggested that there was a north to south decrease in the grain-size of the Karpuzçay Formation in the Köprü basin. This trend was not observed, although local coarsening is associated with the transition to the conglomeratic Aksu Formation near Yesilbag in the north-eastern part of the basin.

4.5.3 Manavgat basin

Outcrop of the Karpuzçay Formation is abundant in the Manavgat basin, but particularly spectacular exposures are found along the new road from the coast towards Akseki. Every road cutting exposure of Karpuzçay Formation along this road was logged and is summarised in the upper half of the Ahmetler section (Fig. 4.18). References to sites and samples in the following text can be located in Appendix 5 where a detailed centimetre-meter scale log of the entire Ahmetler section is presented. Elsewhere in the basin, the outcrop is more patchy, but good road cutting sections can be found at Alarahan, Akdam, Oynaminar and on the road to Yaylaalan (Fig. 4.1).

The Karpuzçay Formation in the Manavgat basin is different from that exposed in the other two basins in the following ways:

- Nannoplankton analysis of samples from the Ahmetler section reveals that the Karpuzçay Formation in the south-east of the Manavgat basin is Tortonian to Upper Messinian in age (Fig. 4.18; Appendix 3a) much younger than Aksu and Köprü basin exposures which are mainly Serravallian.
- In the Manavgat basin, the Karpuzçay Formation is characterised by much coarser deposits than in either the Köprü or Aksu basins.

The Karpuzçay Formation in the Manavgat basin overlies conglomerates and sandstones of the Kizildag Formation in the north-west of the basin (Fig. 3.23). In this area, nannoplankton analyses (Akay & Uysal, 1985) suggest that deposition occurred during the Late Burdigalian-Early Langhian or at some time there after. In the south-east of the basin, the
Figure 4.17  Schematic palaeogeographic reconstruction of the Köprü basin based on the apparent coeval deposition of turbidites in the east of the basin and shallow-water carbonates in the west in the Burdigalian. Note however that there is little evidence to suggest that the carbonates were shedding material east into the basin at this time and it is possible that the ages of the turbidites reflect reworking of older material rather than the age of deposition of the Karpuzçay Formation.
Figure 4.18 Summary log of the Ahmetler section showing the lithology, formation and the ages derived from nannoplankton analysis (modified after Flecker et al., 1995).
Karpuzçay Formation overlies Serravallian planktic foraminiferal marls (Geceleme Formation). This contact is gradational over ~20m. The upper contact of Karpuzçay Formation is poorly exposed, but can be seen in a few places in the south of the Manavgat basin along the coast road where there is a subtle, shallow angular discordance with the overlying Pliocene succession. The implications of this unconformity are discussed in chapter 7.

4.5.3.1 Texture and composition

**Thickness:**
Bed thickness generally increases up section in the Manavgat basin. Siltstones and mudstones are rarely more than 10cm thick and have an average thickness of less than half that. At the base of the Karpuzçay Formation in the Ahmetler section (exposed at the junction between the road to Akseki and the track to Ahmetler), sandstone horizons 5cm thick or less are interbedded with 20-40cms of thinly bedded siltstones and mudstones (Appendix 5). Occasionally, coarser sandstone horizons up to 40cm thick are observed, e.g. just south of Alarahan near the village of Ulugüney. Higher up in the succession however, the variability of sandstone thickness increases and some massive sandstones are several metres thick. Near the base of the Karpuzçay exposure in the Ahmetler section, rare pebble horizons are thin, generally only a few centimetres thick. Conglomerates and debris flows further up the section, however, vary in thickness, but can be up to 8m thick. Channelised beds vary in thickness from 10cm to 2-3m as can be seen in the central region of the Ahmetler section (near sample 185.344.7, Appendix 5). The thickness of slumped beds varies from 0.5-10m within the Ahmetler section.

**Bed shape:**
Siltstones and mudstone beds are dominated by parallel-sided geometry, as are the sandstones in the lower parts of the succession. Occasionally, sandstone lenses in mudstone are seen as at site 308 (Appendix 5). Further up the succession however, sandstones become increasingly erosively bedded and channelised (Fig. 4.15). The superimposition of conglomerate and sandstone channels produces abundant wedging out geometries in the central and upper parts of the succession (Fig. 4.12).
Some of these channelised conglomerates pass laterally into sandstones. Large scale slumps, several meters thick (Fig. 4.15), often have undulating exterior bedding surfaces.

Grain size:
The Karpuzçay Formation in the Ahmetler section contains a full range of grain-sizes from mud-grade to conglomerate with detached blocks 7-10m in diameter (Fig. 4.15). Siltstones and mudstones dominate the lower half of the section and over the transitional interval from the Geceleme Formation (planktic foraminiferal marls). Higher up the section sandstones and conglomerates become increasingly dominant.

Composition:
Figure 4.19 is a triangular diagram plotting the relative proportions of various groups of clast types found in the conglomerates and debris flows in the Ahmetler section. The analyses plotted include measurements from the Çakallar and Geceleme formations and these can be clearly distinguished from the Karpuzçay Formation measurements. What this indicates is that there was a change in the composition of coarse clastic deposits with time. Early calcirudites in the Gecereme and Çakallar Formations contain only Alanya Massif metacarbonates and Miocene shallow-water carbonate detritus. The later clast-supported and matrix-supported conglomerates of the Karpuzçay Formation by contrast, contain much more varied clast-types. In detail this change occurs abruptly over about 20m of stratigraphic thickness and coincides with both the appearance of sandstone in the section for the first time and quartz grains in the siltstones. Subsequently the quantity of Miocene shallow-water carbonate detritus gradually decreases whilst the percentage of Alanya Massif metacarbonate remains relatively constant. Almost without exception, the mega-detached blocks in the debris flows are made of Alanya Massif metacarbonate.

Rip-up clasts of marl make up a significant part of many of the clast-supported and matrix-supported conglomerates, particularly at the base of beds. They are also found at the base of many of the coarser sandstone units. Marl and marly siltstone is the most common matrix of the
Figure 4.19   Triangular diagram plotting the relative proportions of Alanya Massif metacarbonate, Miocene shallow-water carbonate and lithic clasts in conglomerates from the Çakallar, Geceleme and Karpuzçay Formation in the Ahmetler section, east Manavgat basin.
matrix-supported conglomerates in the Manavgat basin, often making up 40% or more by volume of the rock.

In the Messinian, extremely rare, thin horizons of concentrated, fresh-looking, detrital Biotite are found. It is not clear what this represents and it is only seen in one other locality, in Middle Miocene marls (C. Müller pers. com., 1994) to the south east of Gebiz. Akay et al. (1985) report occasional volcanic tuff interbeds in the Karpuzçay Formation, but no such beds were identified during this study. The freshness of the biotite suggests that it has not been transported far and may have a volcanic origin. Lower Pliocene-Quaternary volcanism in the area has been reported by Lefevre et al. (1983) and K-Ar studies of Pliocene ignimbrites are currently being undertaken by C. Glover as part of a Ph.D. project. Miocene aged tuffs were noted in the Kithrea Flysch in northern Cyprus (A. Robertson pers. com., 1994) suggesting that Miocene volcanism is not entirely unknown in the area.

Sandstones, particularly massive sandstones, often contain shell fragments entrained along the base of beds. The fauna identified from these beds includes, bivalves, coral, rhodoliths, bryozoa and the large benthic foraminifera, Operculina sp. Slumps can also be rich in shell fragments. Brown plant material is common throughout the Karpuzçay succession in the Manavgat basin, but organic detritus is dominated by coal fragments of varying sizes which occurs in marls, siltstones and sandstones alike. Although ubiquitous on the large scale, millimetre thick horizons of concentrated coal fragments are the most common form of occurrence. One 7cm thick coal detritus layer which occurs towards the top of the section (sample 20S.352.3, Appendix 5), thins laterally, having a channel-shaped geometry. The biogenic components of the marls and siltstones resemble those of the Geceleme and Çakallar Formations and are described in section 3.6.4.1.

**Sorting and physical maturity:**
One of the most major differences between the clast-supported and matrix-supported conglomerates in the Manavgat basin, is that the clast-supported conglomerates tend to be moderately sorted, while the matrix-supported conglomerates are generally very poorly sorted. In the latter,
detached blocks several metres in diameter and often wider than the thickness of the debris flow horizon in which they have been transported, are surrounded by clasts with an average grain-size of a few tens of centimetres and sometimes less. The sorting of sandstones varies from moderately well sorted to poorly sorted. The latter are sometimes associated with bioturbation traces.

The physical maturity of clasts in the Manavgat basin, as in the Köprü basin, depends to some extent on the nature of the clasts (section 4.5.2.1). Miocene shallow-water carbonate clasts tend to be angular in the lower part of the section and more rounded towards the top. Alanya Massif metacarbonate schist clasts are also angular. The dark grey recrystallised form of Alanya Massif limestone occurs as both rounded and angular clasts in the lower part of the succession, but becomes increasingly rounded at the top.

Grading:
Normal-graded beds of sandstone to siltstone or marl are found with increasing abundance towards the top of the Ahmetler section. Clast-supported and matrix-supported conglomerates also sometimes show normal grading as do some of the debris flow deposits. Figure 4.20 is an example of a matrix-supported conglomerate unit with a normally-graded, imbricated, clast-supported basal layer.

Clast alignment and grain interaction:
Both Pteropod moulds and organic fragments show bedding parallel alignment. However, alignment within the plane of the bed was rarely observed. Imbrication in both clast- and matrix-supported conglomerates is relatively common particularly in the upper parts of the succession (Fig. 4.20).

Structures:
Parallel laminations were observed relatively frequently in sandstones and siltstones. Some mudstones at the base of the Ahmetler section are laminated to the point of appearing fissile.
Figure 4.20  Photograph of a unit with a clast-supported, imbricated basal layer and an upper layer of matrix-supported conglomerate from the Karpuzçay Formation, Ahmetler section, east Manavgat basin.

Figure 4.21  Photograph of distorted beds interpreted as a water escape structure intruding into an imbricated conglomerate, Karpuzçay Formation, Ahmetler section, east Manavgat basin.
In comparison with both the Köprü and Aksu basins the Manavgat basin has a much lower frequency of cross-laminations and ripples. They occur in both sandstones and siltstones with increasing frequency at the top of the succession and are often capped with a mud drape, a few millimetres to several centimetres in thickness. Many coal-rich layers are also rippled. No wave ripples have been observed in this section. Cross-stratification was not observed in the Ahmetler section, but outcrops of cross-bedded sandstones and conglomeratic sandstones occur in a channel to the south of Akdam (the palaeocurrent direction indicated by these cross-sets is discussed in section 4.5.3.4) and in the Messinian section to the south of Alarahan.

3) Several spectacular examples of different types of distorted beds interpreted as water escape structures can be seen at the top of the Ahmetler section. These include the vertical structure 1.5m in height, shown in figure 4.21 (site 348, Appendix 5); a large load structure shown in figure 4.22 (site 356, Appendix 5) and a series of 6 asymmetrical flame structures (0.5-1m in height) along a single bed.

The Geceleme and Cakallar formations are largely free of burrowing traces whilst the Karpuzçay Formation in contrast is extensively burrowed. Burrowing, generally Chondrites, was first observed fairly low down the Ahmetler section (site 316, Appendix 5). The diversity of trace fossils increases upwards reaching a maximum diversity in the Tortonian (site 327, Appendix 5) where extensive networks of the branching echinoid burrow, Thallasinoides are found with Chondrites, Zoophycos, Lophocstenium, Arenicolites and other unidentified traces. Upper parts of the Ahmetler section contain a lower diversity of burrowing traces, but Chondrites is fairly ubiquitous. According to Chamberlain (1978) and Pemberton (1992) Chondrites, Zoophycos, Lophocstenium, Arenicolites and Arenicolites are common as an assemblage associated with turbidites in off-shore to bathyal environments, indicating an approximate water depth range of 150-3000m (Fig. 4.24). Thallasinoides however is characteristic only of off-shore environments. Thus, the ichnofacies assemblage found at site 327 indicates that the water depth of the basin was in the region of 150m or less. Chondrites outside this assemblage can be found in bathyal to nearshore environments (Chamberlain 1978,
Figure 4.22 Photograph of a large load structure located in the Messinian of the Karpuzçay Formation, Ahmetler section, south eastern Manavgat basin.

Figure 4.23 Photograph of layered bioturbation at site 327, Karpuzçay Formation, in the Ahmetler section, north-eastern Manavgat basin.
Figure 4.24 Chart of ichnofossils plotted against depth (after Chamberlain, 1978; Pemberton, 1992). Numbers refer to references in the original text.
Figure 4.23 shows a more common exposure of burrowing, where contrasting layers of marl and sandstone allow clear trace identification.

**Tool Marks** are far less abundant than in the Köprü and Aksu basins, but they are common enough from site 344 up-section (Appendix 5) to give convincing palaeocurrent directions which are discussed in section 4.5.3.4.

**Slumps** are common from site 344 onwards (Appendix 5; Fig. 4.15), but vary greatly in thickness and grain-size. Where the exposure permits towards the top of the sequence a slump has been observed, passing laterally into a matrix-supported conglomerate with no discernible slump structures. This is in agreement with the common consideration of debris flows as being part of a continuum between slumps and turbidity currents (Rupke 1978). The orientation of the palaeoslope deduced from these slumps is discussed below (section 4.5.3.4).

**Syn-sedimentary deformation structures**

Syn-sedimentary faulting was occasionally observed within the Karpuzçay Formation in the Manavgat basin, but it is not common. Figure 4.25 shows a locality where approximately 1.5m of marls interbedded with siltstones and sandstones have been fractured and forced up into the overlying matrix-supported conglomerate. Both brittle and ductile deformation have taken place in the processes.

4.5.3.2 Interpretation

Many of the finer-grained units (medium-grained sandstone to clay grain-size populations) exposed in the Manavgat basin can be interpreted in a similar way to the Karpuzçay Formation in the Köprü and Aksu basin (section 4.5.1.2). Tables 4.1 and 4.2 summarise the turbidite sub-facies identified using a classification scheme after Pickering *et al.*, (1989). Exposure of the coarser grain-size population (coarse sandstone to detached blocks several meters in diameter) is restricted to the Manavgat basin and the small locality near the Bucak junction in the Aksu basin (section 4.5.1.3 and 4) and the processes of deposition of these facies are briefly described below.
Figure 4.25 Photograph of the "snapped beds" at site 345, where both ductile and brittle deformation have occurred during deposition of the overlying conglomerate unit, Karpuzçay Formation, Ahmetler section, south-eastern Manavgat basin.
The coarsest grain-size population can be divided into 3 facies: clast-supported conglomerates; matrix-supported conglomerates and sandstones. These have been subdivided into sub-facies as shown in Tables 4.3 and 4.4.

The internal structures of matrix-supported conglomerates can be explained in terms of debris flow processes. The flow rheology and particle-support mechanism have been briefly described in section 4.5.1.2. Chaotic structureless units with clasts dispersed fairly uniformly through the matrix suggest the existence of weak intergranular dispersive pressure (Lowe, 1976; 1979) or flow turbulence (Enos, 1977) and are thought to be preserved by freezing of the flow due to intergranular friction and cohesion. In some units, an upper cap of structureless matrix-supported conglomerate overlies a clast-supported conglomerate base (e.g. Fig 4.20). Typically, this basal layer is coarser and the grainsize distribution reflects the particle size limit that the matrix was capable of supporting during flow, such that clasts larger than this were able to settle through the matrix (Lowe, 1982). Imbrication of such layers is common and indicates intense grain interaction. Much of the variety of structures observed in debris flows is dependent on the percentage of matrix and whether the flow was turbulent at any time during its depositional history. Enos (1977) suggested that a fully turbulent flow could support much larger clasts than matrix cohesiveness and buoyancy alone. The size of some of the detached blocks (i.e. up to ~10m across) suggests that even a fully turbulent flow is unlikely to have been able to support them. Instead, they may have slid and bounced down slope, lubricated by the water-sediment mix. Turbulent flow is likely to have been active however, in the deposition of units with a relatively coarse matrix (e.g. sand-grade).

Debris flows also occur in subaerial environments and are discussed within the context of alluvial sedimentation in chapter 5 (section 5.6.1.3). Nemec and Steel (1984) suggested a number of internal structural criteria that could be used to distinguish the two environments of deposition (Table 5.5). The structures generally associated with subaqueous debris flow units can be seen in figure 4.26 (after Nemec and Steel, 1984).
<table>
<thead>
<tr>
<th>Structures</th>
<th>Sub-facies (Pickering et al. 1989)</th>
<th>Depositional process</th>
<th>Current type</th>
</tr>
</thead>
<tbody>
<tr>
<td>Clast supported, thickly bedded, flat bedded-deeply scoured, poorly sorted clasts, plastically deformed mud clasts common</td>
<td>A1.1 Disorganised gravel</td>
<td>Freezing on decreasing bottom slopes due to intergranular friction and cohesion</td>
<td>High concentration turbidity flow or debris flow</td>
</tr>
<tr>
<td>Matrix supported, 10-50% mud-grade material, medium-thick bedded, often little erosion into underlying beds but upper surface hummocky, large boulders dispersed throughout, enormous detached blocks</td>
<td>A1.2 Disorganised muddy gravel</td>
<td>Freezing on decreasing bottom slopes due to intergranular friction and cohesion</td>
<td>Cohesive debris flow. Detached blocks may slide into place on a cushion of over-pressured or liquefied mud (Labaune et al. 1983)</td>
</tr>
<tr>
<td>Similar to A1.2, but contains 50-95% mud, irregular shape over short distance</td>
<td>A1.3 Disorganised gravelly mud</td>
<td>Freezing on decreasing bottom slopes as the shear stress at the base of the flow becomes less than the cohesive strength</td>
<td>Cohesive mud flows (debris flows)</td>
</tr>
<tr>
<td>Similar to A1.1, but clasts are dispersed through a sand matrix</td>
<td>A1.4 Disorganised pebbly sand</td>
<td>Rapid collective grain deposition of a pebble sand mixture due to increased intergranular friction as the flow decelerates</td>
<td>High concentration turbidity flow</td>
</tr>
</tbody>
</table>
Table 4.4. Description and classification of clast-supported conglomerates and sandstones in the Manavgat basin.

<table>
<thead>
<tr>
<th>Structures</th>
<th>Sub-facies (Pickering et al. 1989)</th>
<th>Depositional process</th>
<th>Current type</th>
</tr>
</thead>
<tbody>
<tr>
<td>Imbrication, cross-stratification, lenticular to wedge-shaped bodies</td>
<td>A2.1 Stratified gravel</td>
<td>Grain-by-grain deposition from suspension and then traction transport as bed-load</td>
<td>High concentration turbidity flow</td>
</tr>
<tr>
<td>Erosive, laterally thinning, poorly sorted, imbricated, inverse grading</td>
<td>A2.2 Inversely graded gravel</td>
<td>Rapid deposition of a concentrated traction carpet due to increased intergranular friction. Inverse grading and Imbrication are caused by intense grain interaction</td>
<td>High concentration turbidity currents</td>
</tr>
<tr>
<td>with or without overlying normal grading</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Normal grading, scour, clast-supported</td>
<td>A2.3 Normally graded gravel</td>
<td>Grain-by-grain deposition from suspension with little subsequent tractional transport</td>
<td>High concentration turbidity currents</td>
</tr>
<tr>
<td>Normal grading, scour, clast-supported</td>
<td>A2.7 Normally graded pebbly sand</td>
<td>Grain-by-grain deposition from suspension with little subsequent tractional transport</td>
<td>High concentration turbidity currents</td>
</tr>
<tr>
<td>Laterally continuous, grading absent or poorly developed, water escape</td>
<td>B1.1 Thick-medium bedded disorganised sands</td>
<td>Rapid mass deposition due to intergranular friction in a concentrated dispersion near the bed. Resultant open grain packing may collapse during or after deposition resulting in the escape of poor fluids and formation of fluid escape structures</td>
<td>High concentration turbidity currents</td>
</tr>
</tbody>
</table>
Figure 4.26 Characteristics of subaqueous debris flows (Nemec and Steel, 1984).
The disorganised character of some clast-supported conglomerates and coarse sandstones allows comparison with debris flow deposits. Where the amount of medium sand to mud grade matrix is <5% by volume, this suggests a non-cohesive transport mechanism (Ineson, 1989). In the field this was difficult to assess accurately, but it is thought that units with <5% matrix were relatively uncommon so that some buoyant lift from the matrix would have been active in reducing the effective weight of clasts (Rodine and Johnson, 1976). This may help to explain the generally coarse nature of the clast-supported conglomerates, although fully turbulent high-density flow is also capable of supporting and depositing coarse units. The internal structures observed such as imbrication, cross-stratification, inverse and normal grading result from various combinations of the processes active during deposition from high-density turbidity currents and debris flows e.g. traction sedimentation, deposition from suspension and frictional intergranular freezing.

4.5.3.3 Spatial and temporal variations

69 samples from the Ahmetler section were analysed for nannoplankton by C. Müller (pers. comm., 1992). With an average sampling rate of just under one sample for every 10m a far more detailed nannoplankton biostratigraphy has been set up than had been attempted by previous workers (Fig. 4.18; Appendices 5 and 3a). Other sections in the Manavgat basin were not sampled in so much detail due partly to the lack of such good exposure as that along the Ahmetler section, and because of time constraints. Because in general the exposure of vertical successions is so good in the Manavgat basin, a far more detailed study of spatial and temporal variations is possible here than in the Köprü or Aksu basins.

Analyses of the benthic/planktic foraminiferal ratio of nearly 40 marl samples from the Ahmetler section are shown in figure 4.27. The trend has been interpreted as indicating a shallowing-upwards sequence beginning as soon as shallow-water carbonate deposition had ceased (Latest Burdigalian- Early Langhian) and continuing throughout the Miocene. No evidence of emergence was observed at the top of the section (i.e. no wave ripples, desiccation cracks, evidence of continental or
Figure 4.27  Plot of the benthic to planktic foraminiferal ratio for the Ahmetler section in the north of the Manavgat basin, showing a shallowing upwards trend.
estuarine conditions), so it is assumed that the depositional setting remained below wave base even at its shallowest.

Interpretation of this shallowing upwards trend is not without its difficulties however. As noted in section 3.6.4.1, the marls in the Geceleme and Çakallar formations consist both of horizons containing shallow-water detritus, which is thought to have been transported and others which contain little shallow-water debris. The Karpuzçay Formation has been interpreted as the product of turbiditic and debris flow activity and therefore, by their very nature contain reworked components. Some of the scatter in benthic/planktic ratios may, therefore, be explained, at least in part, by transport of shallow-water material including benthic foraminifera. Despite this scatter, however, a trend is clearly visible and it seems likely that this reflects a real shallowing upwards of the basin. The palaeobathymetry indicated by the benthic foraminifera, however, only provides an estimate of the minimum water depth (table 3.12). The ichnofacies identified in the central part of the Ahmetler section indicate slightly deeper water than the benthic foraminifera, but the absence of burrowing traces in the lower part of the section means that the palaeobathymetry for the first 200m of this section can only be estimated. The Late Miocene abundance of Chondrites traces may indicate either a stable water depth or a decreasing one. The increasingly dominant nature of the conglomerates and debris flows particularly in the Messinian part of the section indicates the increasing proximity of the source area and this may well have been caused by a relative sea level fall.

The sudden influx of sand detritus in the Ahmetler section occurs over a sampling period which failed to produce the characteristic nannoplankton markers to identify NN7 and NN8. The close proximity of this part of the succession to samples containing nannoplankton characteristic of NN9 indicates that the sand input probably occurred in the Early Tortonian. Correlation of the timing of the sand input with other sections in the Manavgat basin is hampered by this ambiguity, but it seems likely that the change in provenance was a relatively synchronous event affecting all parts of the basin.
Nannoplankton characteristic of zone NN10 were not identified in the Ahmetler section. This may indicate a period of non-deposition, although given that the dominant deposits from NN9 onwards are coarse clastics, this seems unlikely. It is possible that though deposited, all traces of NN10 in the Ahmetler section were subsequently removed by the erosive power of NN11a deposits. However, the most likely scenario is that NN10 lies within a sampling gap which, in this part of the section is about 100m.

None of the other sections in the Manavgat basin show anything like the concentration of coarse-grained, high-energy deposits such as those seen in the Ahmetler section. The absence of debris flows containing huge detached blocks anywhere other than the Ahmetler section is an obvious expression of this. In part this is due to the spectacularly continuous nature of the exposure seen along the road to Akseki, but it must also be admitted that some of the difference is due to lateral variation. It seems likely that the region south of Ahmetler was some sort of a broad channelised system, away from which to both east and west the deposits fined (Fig. 4.29). This channelised system may represent the highest energy deposits of a fan-delta system, similar, but on a larger scale to that seen in the Aksu basin, concentrated in a relatively restricted north-south trending area. The patchy distribution of the Çakallar Formation calcirudites and reef talus is concentrated in the same area as the coarse-grained Karpuzçay Formation in the Manavgat basin. This systematic distribution of coarse material throughout the Middle-Upper Miocene is discussed further in chapter 7 (section 7.5.1.2) where a structural hypothesis is suggested for this geographical relationship through time.

There is an increase in sedimentation rate from the Geceleme Formation planktic foraminiferal marls through the Late Serravallian and into the Early Tortonian Karpuzçay Formation in both the Ahmetler and the Alarahan sections (Fig. 4.28). The more limited data set from the Oymapinar section to the NW shows no such increase. Despite this, sedimentation rate information (Fig. 4.28) and field observations (Fig. 4.29) suggest that there is a correlation between high sedimentation rate and the occurrence of coarse, conglomeratic Karpuzçay Formation.
Ahmetler

Figure 4.28 Miocene sedimentation rates for the Ahmetler, Oymapinar and Alarahan sections in the Manavgat basin (see Appendix 6).
Figure 4.29  Correlation of four sections from the Manavgat basin.
4.5.3.4 Palaeocurrent analysis

Rose diagrams for the Karpuzçay Formation in the Manavgat basin are displayed in figure 4.30. The prevailing palaeocurrent direction for the formation is towards the south west although south and south east directions are also sometimes indicated. There appears to be little disparity between the palaeocurrent directions measured along the Ahmetler section and those from other parts of the basin and this probably indicates that the Karpuzçay Formation was deposited from a series of channels along its northern margin, the most high-energy of which was located in the Ahmetler area. The north and west directed cross-sets from a channel system in the far south east of the basin near Akdam, are unusual and may represent a sediment source to the east or south east. This system does not appear to have been a powerful one since there is no indication of further penetration of the west and north directed currents into the main part of the Manavgat basin. No difference in composition of the sandstones was noted.

4.7 Summary interpretation and reclassification of the Karpuzçay Formation

Study of the Karpuzçay Formation in the Aksu and Köprü basins suggests that the interbedded sandstones and siltstones found here can be interpreted mainly in terms of low- and high-density turbidity currents. The exposure mapped as Karpuzçay Formation in the Manavgat basin however, is dominated by clast-supported and matrix-supported conglomerates. These have been interpreted in terms of high density turbidity currents and debris flow processes. The age of this coarse-grained sequence is significantly younger (Tortonian-Messinian) than the turbidites in the Aksu and Köprü basin (mainly Serravallian and Lower Tortonian). To reflect these differences, the Karpuzçay Formation has been sub-divided into two groups: the Beskonak Member, comprising the Aksu and Köprü basin sandstone and siltstone dominated succession, and the Taskesigi Member comprising the coarse, conglomeratic sequence exposed in the Manavgat basin. A summary of the features, type section localities and synonyms is given in Table 2.3.
Figure 4.30  Rose diagrams indicating the palaeocurrent directions in the Karpuzça Formation of the Manavgat basin.
Figure 4.31 is a schematic representation of the possible development of the three basins from Late Burdigalian to Messinian times as deduced from age (Figs. 4.2 and 4.29) and palaeocurrent data (Figs. 4.10, 4.16 and 4.30) for the Karpuzçay Formation. In this diagram it is suggested that the finer-grained deposits of the Karpuzçay Formation represent distal turbidites sourced mainly from the north. The conglomeratic succession in the Aksu basin has been interpreted as an ephemeral coarse-grained fan-delta sourced from the west. A similar interpretation may also apply to the Manavgat basin Taskesigi Member, although its north-south linear geometry suggests that it may represent the channel system of a fan-delta active throughout the Mid-Late Miocene from which finer-grained turbidity currents were also sourced.

4.8 Correlation of the Karpuzçay Formation across the Isparta Angle

Nannoplankton dating of sections across the Isparta Angle by C. Müller has allowed the correlation of seven logged successions which contain Karpuzçay Formation. This correlation is displayed in figure 4.32. The following general observations can be made, always bearing in mind the biostratigraphic uncertainty inherent in redeposited rocks:

- There is a basinward shift in facies from shallow-water carbonates (Oymapinar Formation) to deeper water deposits (Karpuzçay or Geceleme Formations) during Latest Burdigalian-Langhian times (Fig. 4.32).
- This basinward shift in facies is observed throughout the Isparta Angle, but it appears to have started earlier in the south of the Manavgat basin than elsewhere (Fig. 4.32).
- Exposure of the Geceleme Formation is geographically restricted to the eastern part of the Manavgat basin. Elsewhere, coeval fine-grained Beskonak Member turbidites are found (Fig. 4.29).
- Rapid coarsening of the succession in the Serravallian-Tortonian occurred in the eastern part of the Manavgat basin (Taskesigi Member). This is associated with an increase in sedimentation rate.
Burdigalian turbiditic flow inferred down the east side of the Aksu and Köprü basins from palaeocurrent and nannoplankton evidence. Coeval shallow-water carbonates forming on the west side of the Aksu and Köprü basins and in the east of the Manavgat basin.

Karpuzçay Formation deposited throughout the area on top of Burdigalian and Langhian shallow-water carbonates. Axial flow (north to south) is documented by tool marks in the turbidites in the Aksu and Köprü basins. Imbrication from a conglomeratic sequence in the Aksu basin suggests the ephemeral development of a fan-delta on the western margin. In the Manavgat basin, planktic foraminiferan marls were deposited in the north and east, whilst finer-grained turbidites flowed from the Köprü basin, over the Kikkavak fault. A small source of material is inferred to have been in the south east of the basin.

No Karpuzçay Formation younger than Lower Tortonian is preserved in the Aksu and Köprü basins. A thick succession of Beskonak Group turbidites is found in the Manavgat basin, with a core of coarse conglomeratic units (Taskesigi Group) orientated north-south in the east of the basin. The eastern source of material may still have been active at this time.

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Figure 4.31  Schematic representation of the deposition of the Karpuzçay Formation throughout the Miocene.
Figure 4.32  North-west to south-east correlation diagram of Miocene sections across the Isparta Angle.
Figure 4.33 Miocene subsidence curves for the Ahmetler, Oymapinar and Alarahan sections in the Manavgat basin plotted with sedimentation rates along the bottom. (Diagram prepared using software written by J. Turner.)
These points will be discussed further in chapters 7 and 8 in terms of their tectonic and eustatic sea level implications for the Isparta Angle.

4.9 Conclusions

- The Karpuzçay Formation has been subdivided into the Beskonak and Taskesigi Members.
- The Beskonak Member has been interpreted as having been deposited mainly by low- and high-density turbidity currents.
- Some storm generated currents may have deposited and reworked material in the Aksu basin, but there is little evidence to suggest that this process had a widespread effect on the Karpuzçay Formation rocks preserved.
- The coarse-grained units in the Taskesigi Member have been interpreted as having been deposited mainly by high-density turbidity currents and debris flow processes.
- Low-density turbidity currents are also thought to have deposited the finer-grained sediments in the Taskesigi Member.
- The Taskesigi Member contains beds which document an extremely high energy of deposition.
- Coarse-grained successions in the Aksu and Manavgat basin may represent deposition on the shelfal part of a fan-delta.
- Major source of turbidite material is in the north for all basins.
- There is no evidence to suggest that western and eastern margins of the Aksu and Köprü basins were important sediment sources.
- The Kirkkavak fault had little topographic expression during the Lower Miocene allowing the Köprü basin to overfill in the south east.
- A rapid basinward shift in facies from shallow-water carbonates to deeper water foraminiferal marls occurred in the Manavgat basin near the Burdigalian-Langhian boundary.
- A decrease in water depth has been deduced from benthic/planktic ratios in the Manavgat basin.
- A shallowing up into the Tortonian is inferred for the Aksu and Köprü basins during the transition into the Aksu Formation.
Chapter 5

FAN-DELTA SEDIMENTATION, PROCESSES AND CONTROLS: 
THE AKSU AND KIZILDAĞ FORMATIONS.

5.1 Context

Sedimentary data from alluvial, coastal and subaqueous environments is discussed and interpreted in terms of fan-delta sedimentation. This concept integrates sediment transport, deposition and reworking in each of these environments so that basin-wide sedimentation processes can be studied. This results in information on palaeogeography, climate, sea level change, the mechanisms of basin fill and the tectonic process active during deposition and preservation.

5.2 Organisation of this chapter

Following a brief introduction to the temporal and spatial distribution of the sediments discussed in this chapter (5.3) a review of previous work on the conglomerate-dominated successions (Aksu and Kizildag Formations) is given in section 5.4. Section 5.5 contains descriptions of all the sub-facies observed within these successions. These have been divided into four groups (A-D) according to their sub-facies associations. A paragraph on the interpretation of the sub-facies association is given at the end of each sub-section. Palaeocurrent and provenance data is presented for the Aksu, Köprü and Manavgat basins (section 5.6). This is followed by an overview of the sedimentary system as a whole and the classification of the Fan-delta type (sections 5.7 and 5.8). Finally, section 5.9 discusses the controls on fan-delta development, e.g. climate, tectonics and eustacy and is followed by the conclusions (section 5.10).
5.3 Temporal and Spatial distribution

The sequences studied here are predominantly coarse sandstones and conglomerates. They are concentrated in northern and western parts of the Köprü and Aksu basins (Fig. 5.1) where they are Tortonian in age (Aksu Formation) and constitute the basal Miocene sediments in the far north and east of the Manavgat basin (Kizildag Formation). Figure 5.1 also indicates the key localities of the fan-delta sections mentioned on the text.

5.4 Previous work

5.4.1 The conglomerate successions

The basal formation in the classification of Miocene and Pliocene sediments in the area between Alanya and the Köprü basin devised by Blumenthal (1951) is called "les Conglomérats de base". Undated, these conglomerates were classified by their stratigraphic position beneath the Lower Miocene reefs (Calcaire récifal bordier; Fig. 5.2) and were only recognised along the northern margin of the Manavgat basin. Monod (1972) renamed these Manavgat conglomerates "les Conglomérats de Sevinç" after a small village to the south east of Yaylaalan, for his map of the Taurides south of Beysehir (1972; Fig. 5.2). In his thesis five years later however he refers to these conglomerates as "les Conglomérats de Tepekli" (Fig. 5.2). Monod (1977) describes these conglomerates as a formation of greatly variable thickness, up to 1000m thick, with clast-types representing "all the facies of the autochthonous and allochthonous formations of the Taurides". He notes that this succession almost always forms the basal transgressive horizon during the Miocene and suggests that the thickness variation is due to infill of an Oligo-Miocene erosion surface with palaeo-valleys developed orthogonally to the strike of the Tauride chain. He goes on to point out that the absence of this formation on the top of the Bey Daglari and Lycian Nappes (Poisson, 1974; Graciansky, 1972) in the west of the area indicates "a fundamental difference between the two arms of the Isparta Angle".
Figure 5.1 Geographic areas covered by Aksu Conglomerate Formation and the localities mentioned in the text. (Modified after Akay, 1985)
Figure 5.2 Table showing the various names and inferred ages of the basal conglomerate succession in the Manavgat basin (Kizildag Formation). For each box, the south of the basin is at the bottom and the north at the top. Thus, a south-north diachronous change from conglomerates to shallow marine carbonates is suggested as part of this study.
The Köprü basin conglomerates were undifferentiated from the turbidite succession (Karpuzçay Formation) by Monod (1972; 1977). Dumont (1974) studied the north eastern limb of the Köprü basin, named the conglomerates found around the village of Kesme in the Köprü basin (Fig. 5.1), "les Conglomérats de Kesme". He noted that conglomerates generally made up 90% of the formation and that these horizons are often laterally discontinuous. Interbedded with these highly heterogeneous, rounded conglomerates he observed lenses of reef limestone and marl, or sandstone horizons rich in oysters and gastropods. The marls in this area failed to yield age-diagnostic foraminifera. However, by studying a similar succession to the east of the Kirkkavak fault, south of the village of Dumanli (Fig. 5.1), Dumont (1974) obtained a Tortonian age (chapter 3; section 3.6.3.3) and made the assumption that the conglomerates on both sides of the Kirkkavak fault are age equivalent.

Akay and Uysal (1985) was the first person to study the Neogene sediments of the entire study area (i.e. Aksu, Köprü and Manavgat basins). He classified all the conglomerate-dominated successions in both the Upper and Lower Miocene as Aksu Formation (Fig. 5.2; Tables 2.1 and 2.4). Akay et al. (1985) defined this formation as consisting of terrigenous conglomerate-siltstone successions, marine conglomerate-sandstone successions and lenses of reef limestone. His generalised chronostratigraphic section across the area (Fig. 2.4) indicates that in his opinion this formation varies in age from Upper Tortonian in the Aksu basin, to Upper Oligocene near Serik in the south of the Köprü basin (Fig. 5.1). A TPAO report (Akay and Uysal, 1985) submitted on the subject of these basins reveals however, that very little data on the age of these conglomerates actually exists. The majority of the nannoplankton analyses on samples from the north of the Aksu basin yielded NN10-11 (Tortonian-Lower Messinian). A few however, produced NN3 and NN4 zones. Akay suggested that most of these are reworked. Gutnic et al. (1979) also quote a Tortonian age for the conglomerates onlapping onto the northern margin of the Aksu basin (Kapikaya). In conjunction with other workers, including Poisson, Akay and Uysal (1985) mapped the study area and differentiated between the conglomerate-dominated successions on the west side of the Köprü basin and the turbidite-
dominated (Karpuzçayı Formation) succession on the east side of the basin.

5.4.2 *Lacustrine limestone*

Monod (1972; 1977) observed and named the thin (20m) sequence of lacustrine limestones found at Kepezbelenli beyond the northern margin of the Manavgat basin, the Calcaire de Kepez. Overlain by the Conglomérats de Tepekli/Sevinç (see above) and Calcaire d'Oymapinar (Fig. 5.2) he identified this limestone as part of the Miocene succession. Although Monod (1977) reported abundant plant traces including roots and leaf debris and rare *Charophytes*, pollen analysis failed to produce an age for this unit. No other worker has subsequently documented this limestone (Fig. 5.2).

5.5 *Facies description*

The sub-facies described below have been sub-divided, by their sub-facies association into 4 groups (A-D). A table showing the sub-facies discussed in each group, their localities (with reference to Fig. 5.1) and the formation in which they were found is shown at the beginning of each sub-facies association section. A summary table containing information concerning each sub-facies is also provided.

5.5.1 *Sub-facies association A*

A list of the sub-facies discussed in this section, their localities with reference to figure 5.1 and the formations in which they are found is given in table 5.1.

5.5.1.1 Clast-supported angular conglomerate.

*Description*

This sub-facies was only observed in the far north of the Aksu basin, onlapping at a high angle the Lycian remnant (Kapikaya) which forms the northern margin of the basin (Fig. 5.1). Gutnic *et al.* (1979) suggest that the sediments banked up against Kapikaya (Fig. 5.3) are Tortonian in
Figure 5.3 Diagram showing the onlapping relationship of the angular clast-supported conglomerate onlapping onto the Lycian Kapikaya ridge in the north of the Aksu basin (Modified after Gutnic et al., 1979).

Figure 5.4 Photograph of some of the large clasts in the angular clast-supported conglomerate at Kapikaya in the north of the Aksu basin.
Table 5.1 List of the sub-facies in sub-facies association A, their localities and formations.

<table>
<thead>
<tr>
<th>Sub-facies</th>
<th>Localities</th>
<th>Formation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Clast-supported angular conglomerate</td>
<td>Kapikaya</td>
<td>Aksu</td>
</tr>
<tr>
<td>Clast-supported conglomerates and coarse sandstones</td>
<td>Natiflar, Aspendos, Yaylaalan, Bozburun, Kapikaya, Kesme, Dumanli</td>
<td>Kizildag</td>
</tr>
<tr>
<td>Matrix-supported conglomerates</td>
<td>Yaylaalan, Natiflar</td>
<td>Kizildag</td>
</tr>
<tr>
<td>Laminated sandstones</td>
<td>South Kargi, Yaylaalan, Aspendos, Saburlar, Dumanli</td>
<td>Kizildag</td>
</tr>
<tr>
<td>Calcretes and associated fine sediments</td>
<td>S Kargi, Aspendos, Kepezbeleni, Yaylaalan, Kapikaya</td>
<td>Kizildag, Aksu</td>
</tr>
</tbody>
</table>

This is corroborated by Akay and Uysal (1985) who dated the conglomerate and sandstone sequences to the south of Kapikaya into which the angular conglomerate passes, as Tortonian. This sub-facies is laterally very restricted, hugging the margin of Kapikaya and perhaps extending only a few tens of metres from it. There is a clear wedge-shaped geometry associated with the exposure of the angular conglomerates such that the deposit thins to nothing <100m to the south. Clasts within this conglomerate consist of huge (up to 3m in diameter) blocks of Mesozoic limestone identical to that of which the Kapikaya is formed (Fig. 5.4) and a mass of smaller angular blocks of basalt and radiolarite. The fabric is clast supported and dramatically poorly sorted.
A consistent matrix is virtually absent, but smaller angular fragments fill some interstices. The thickness of the unit as a whole is difficult to estimate, as the base of the succession cannot be seen, but it must be >50m.

Interpretation
The nature of the poorly sorted angular clast-supported conglomerate with its limited lateral extent and wedge-shaped geometry banked up against the Kapikaya ridge, suggest formation as a talus deposit. Its similarity to other talus deposits particularly the reef limestone talus seen at Akseki road (section 3.6.2.3) and the modern talus deposited along the Kirkkavak fault on the eastern margin of the Köprü basin (Fig 5.1) is marked. The geographical association with the Kapikaya ridge and the similarity of the limestone of which it is made with respect to the large limestone blocks in the conglomerate, indicate that Kapikaya was a high at the time of deposition and a source of material. There is abundant evidence of faulting within the Kapikaya and Gutnic et al., (1979) suggest that it itself was emplaced from the NW along a low angle thrust plane in the final stages of the Lycian Nappe movement (Fig. 5.3; chapter 7). There is also evidence of higher-angle faulting on the south face of the Kapikaya in the form of slickenside striated surfaces. Poisson (pers. com., 1995) suggested that these faults represent a later, Pre-Tortonian faulting event. It is suggested however, that the talus described above is fault generated, so that some of the faulting may be Tortonian in age. The lateral facies change to calcretes, sandstones (section 5.5.1.2) and conglomerates (section 5.5.1.5) suggests that the talus was deposited in a subaerial environment.

5.5.1.2 Clast-supported conglomerates and coarse sandstones

Description
Clast-supported conglomerates and coarse sandstones constitute 60-70% of all the Aksu and Kizildag Formation exposures. Beds are often characterised by limited lateral extent and erosive lower boundaries. This sub-facies is generally interbedded with finer-grained sandstones and calcretes, but in the southern central part of the study area near the village of Natiflar (Fig. 5.1) it can be seen directly overlying basement
rocks (Fig. 5.5). The upper surface of these units is generally fairly flat and at Yaylaalan this surface is often overlain by red marls containing caliche.

Bed thickness varies from 50cm to several metres. In rare cases conglomerates fine up to coarse or medium sands, but the vast majority are massive and only moderately sorted. Less frequently, large scale planar cross-bedding and imbrication such as that seen at Aspendos (Fig. 5.6) have been recorded from these clast-supported conglomerates. Clasts are generally angular to moderately rounded dependent on clast composition (section 4.6.2.2 for discussion). The composition of the conglomerate clasts is extremely heterogeneous comprising Mesozoic limestone, chert, igneous material, siltstone and sandstone (Appendix 4a, for full list of clast-types found here). Shallow water limestone clasts derived from the Miocene are almost entirely absent. They occur exceptionally at the very base of the Saburlar section overlying bored and oyster-encrusted Alanya Massif basement. Provenance data from these and other conglomerates is discussed in more detail in section 5.6 below.

The matrix is variable in both quantity and grainsize, but red, angular micro-conglomerates dominate with compositions similar to those of the clasts. In some cases a clear bimodal grainsize distribution is observed with coarse often imbricated conglomerates surrounded by a micro-conglomerate-silt grade matrix.

**Interpretation**

The general absence of marine fauna, the red colouration and the association of these conglomerates with red muds and calcrete horizons (section 5.5.1.5) suggest a continental environment of deposition. It seems likely however, that the conglomerates and sandstones described here represent more than one facies. Collinson (1986) described two sorts of traction deposited conglomerates common in coarse-grained alluvial fan environments: sheet or stream floods and stream channel conglomerates. In the former, conglomerates and gravels initially develop a sheet-like morphology under upper flow regime conditions. Later, the flow splits up into small channels dissecting the deposited sheet and reworking it. This may result in well sorted sands and
Figure 5.5  Photograph and field sketch of channelised red conglomerates directly overlying basement melange near the village of Natiflar on the Mesozoic "promontory" between the Asku and Köprü basins.
Figure 5.6 Log and photograph of the continental section at Aspendos with rather thinly bedded sandstones and conglomerates interbedded with quite well developed caliche. Imbrication data from this section is also displayed. Overlying these red beds is an exposure of the shallow marine echinoid-scaphopod facies which also contains well developed rhodoliths (Fig. 3.18). Some of the scaphopods are strongly aligned in small parallel laminated sandstone horizons. This is overlain by marly limestone and then increasingly sandy turbidites with a gastropod-rich layer at the base.
conglomerates with lenticular bed forms and locally scoured bases. Cross-beding and lamination may also develop.

Stream channel conglomerates are markedly laterally impersistent. They are deposited during the waning flood when the flow reworks less well sorted deposits within the channel. Vigorous grain transport either along the channel floor or on the top of a longitudinal bar results in the development both of stratified and unstratified bodies which are often lenticular and commonly display traction structures such as imbrication and cross-bedding. These features are typical of braided stream environments (Tucker, 1981). Table 5.2 is a list of the criteria used by Steel and Wilson (1976) for distinguishing between stream flood and braided stream deposits.

Table 5.2 Criteria for distinguishing between streamflood and braided stream deposits (Modified after Steel and Wilson, 1976).

<table>
<thead>
<tr>
<th>Criteria</th>
<th>Streamflood deposits</th>
<th>Braided stream deposits</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sedimentation unit</td>
<td>Conglomerate bed usually overlain by sandstone beds</td>
<td>Conglomerate pass vertically or laterally into sandstones. Thin impersistent siltstones sometimes present</td>
</tr>
<tr>
<td>Sorting</td>
<td>Conglomerates usually poorly sorted, but normally with pebble and cobble frameworks</td>
<td>Conglomerates usually fine-grained and well sorted</td>
</tr>
<tr>
<td>Structures</td>
<td>Conglomerates often planar cross-stratified, sometimes on a very large scale (set thickness &gt; 1.5m). Sandstones cross-stratified or flat-bedded</td>
<td>Abundance of trough cross-stratification in sandstones and conglomerates. Sandstones are also flat-bedded</td>
</tr>
<tr>
<td>Clasts</td>
<td>Extraformational clasts</td>
<td>Intraformational clasts are common</td>
</tr>
<tr>
<td>Basal erosion</td>
<td>Marked basal erosion</td>
<td>Abundance of concave-up erosion surfaces</td>
</tr>
<tr>
<td>Geometry of units</td>
<td>Laterally impersistent along depositional strike; often filling large channels</td>
<td>Always laterally impersistent, usually filling smaller channels than the stream-flood deposits</td>
</tr>
</tbody>
</table>
The development of cross-stratification is dependent on the availability of space (i.e. depth; Rust, 1975; Church and Gilbert, 1975) and thus has been recognised as a depth indicator (Kraus, 1984). Hein and Walker (1977) suggest that it forms during rapid decrease in fluid and sediment discharge across a bar and that this therefore is evidence for flashy, ephemeral flow. The steepness of this cross-stratification depends on the angle of repose of a gravel sheet foreset margin, the growth of which is governed by the flow symmetry, fluid discharge and sediment discharge (Hein and Walker, 1977).

It seems likely therefore that some of the conglomerates and sandstones described constitute either formation as a sheet flow or as channelised deposits, with traction processes dominating in a semi-arid environment with ephemeral flow. Stream/flood power must have been great to transport the quantity of coarse bedload that has been preserved. However, many of the massive, structureless conglomerates may be the product of mass flows similar to those described by McCallum and Robertson (in press) from the Pliocene of Cyprus. These are discussed further in the next section.

5.5.1.3 Matrix-supported conglomerates

Description
Interbedded with clast-supported conglomerates, sandstones, calcrete, siltstones and mudstones, the matrix-supported conglomerate units are laterally discontinuous on a scale of 100m or less. Within a single outcrop little scour is associated with the lower surface of the unit and the upper surface is generally flat to irregular.

Matrix-supported conglomeratic units are uncommon in the Aksu and Kizildag Formations. Most of the exposure of this sub-facies can be found near the base of the Lower Miocene succession on top of the Mesozoic "promontory" between the Aksu and Köprü basins (Fig. 5.1). Individual horizons are 50cm to several metres thick and vary greatly in appearance from one unit to the next. At Aspendos (Fig. 5.1), red siltstone with widely spaced rounded clasts up to 16cm in diameter are found interbedded with clast-supported fabrics (Fig. 5.6; section 5.5.1.2). At
Yaylaalan (Fig. 5.1) the grainsize of the clasts is much larger (up to 0.75m) and the fabric is only just matrix-supported. The matrix in these rocks is dominantly micro-conglomerate. Other structures associated with these units include crude fining up sequences and sandstone caps on matrix-supported conglomerate units seen near the village of Natiflar (Figs. 5.1 and 5.5).

**Interpretation**

The colour, and association with calcrete horizons (section 5.5.1.5) and conglomerates interpreted as braided stream or sheet flood deposits suggest that these matrix-supported conglomerates were deposited in a subaerial environment. The fabric and general lack of internal organisation of these units suggest that they are the product of subaerial debris flows. Massive, clast-supported conglomerates with coarse matrix were probably deposited by non-cohesive mass flow processes (Lowe, 1982) and these resemble the sandy mass flows documented by McCallum and Robertson (in press) from the Pliocene of Cyprus. Poole (1992) reported similar conglomerates from the Pleistocene of the same area and to the west of the study area, Hayward (1982) documented sandy mass flows from the Mid-Late Miocene of the Kas basin. The similarity of these conglomerates with those already described above indicates that there is a continuum of mass-flow processes which lead to a variety of different fabrics and structures. Indeed it is possible that imbrication observed in clast-supported conglomerates was produced by sandy mass-flows with density modified grain-flow processes active.

Hooke (1967) suggested that mass flows occur preferentially on the upper and middle reaches of alluvial fans. The greater abundance of debris flows overlying the basement potentially indicates a higher energy of deposition at the base of Miocene sedimentation, possible related to early subsidence of the basin. This idea is discussed further in chapter 7 in the light of structural evidence.

Nemec and Steel (1984) indicate that the lack of cohesive structures is indicative of non-cohesive flow which may well be related to the small quantity of mud and clay. It is interesting to note that even when matrix-supported debris-flows do occur in this environment, the matrix is
generally silt grade or coarser. Harvey (1984) noted that the lack of debris flows is typical of semi arid fans where soil cover is poor due to lack of abundant vegetation. This statement should be modified in recognition of the important role non-cohesive debris-flows play in transporting coarse-grained material, e.g. the lack of soil and fine material typical of a poorly vegetated semi-arid environment results in the deposition of few fine-grained, matrix-supported debris flows.

The crude fining up sequences in these units are probably caused by settling of larger clasts through the matrix during flow (Lowe, 1979). Lowe (1982) and other authors (e.g. Rodine and Johnson, 1976) have suggested that where in some cases a partially clast-supported fabric is observed and the quantity of the matrix is small, the largest clasts are not actually suspended during flow. Instead the clay-water matrix provides some buoyant lift reducing the effective weight of the clasts and lubricating them allowing them to intermittently roll, bounce and slide down slope. This type of deposit is often associated, as at Yaylaalan with small clay percentages (Winn and Dott, 1977).

*Distinguishing subaerial and subaqueous mass-flow deposits*

Mass-flow mechanisms have been discussed in Chapter 4 (sections 4.5.1.2 and 4.5.3.2) with reference to turbidite and large scale debris flow deposition in subaqueous environments. However, subaerial mass-flows were not covered in chapter 4 and for this reason they are touched on here.

Although in most cases it was possible to identify whether the mass-flow deposits observed in the field were subaerially or subaqueously deposited from their sub-facies associations, and their diagnostic components and structures, in other cases this was difficult or the relationships ambiguous (e.g. Bucak, Bucakköyü and Kizildag sections; Fig. 5.1). Nemec and Steel (1984) suggested a number of criteria that should be helpful in distinguishing the two environments. These are summarised in table 5.3 below and in figures 5.7 and 4.26.
Table 5.3  The differences between subaqueous and subaerial debris flows.

<table>
<thead>
<tr>
<th><strong>SUBAERIAL DEBRIS FLOWS</strong></th>
<th><strong>SUBAQUEOUS DEBRIS FLOWS</strong></th>
</tr>
</thead>
<tbody>
<tr>
<td>Mud-rich matrix-supported - clast-supported</td>
<td>Bimodal mud-rich conglomerates due to mixing with marine or lacustrine muds (Larsen and Steel, 1978; Nemec et al., 1984)</td>
</tr>
<tr>
<td>Usually ungraded and represent plug-flow deposition (Johnson, 1970)</td>
<td>Better internal organisation with well developed grading (Inverse/normal/both).</td>
</tr>
<tr>
<td>Inverse grading restricted to basal few cm (Gloppen and Steel, 1981)</td>
<td></td>
</tr>
<tr>
<td>Signs of turbulent flow are generally transient</td>
<td>More often fully turbulent, but surges are still common.</td>
</tr>
<tr>
<td>Upward fining sandstone caps sometimes with erosive bases and stratified may result from turbulent fluidal flow or heavily sediment-laden stream flow following the debris flow. May also represent intersurge deposits</td>
<td>Sandy caps here are far more common. This is due to the tendency of subaqueous flows to evolve towards high density turbidity currents (Lowe, 1982; Nemec and Steel, 1984)</td>
</tr>
<tr>
<td>Stream flow reworking may also produce interflow caps of tightly packed conglomerates.</td>
<td>Resedimentation processes common off active slump scars and unstable debris flow noses, density modified grain flows to high density turbidity currents due to partial selective liquefaction and later a series of continuous surging flows. Particularly in the lower reaches of fan-delta slopes debris flows are associated with sand-granule layers of stuff very similar to their matrix because of these processes (Postma, 1984; Nemec et al., 1984)</td>
</tr>
<tr>
<td>Composite units may result from rapidly surging flows</td>
<td>Composite units may result from rapidly surging flows</td>
</tr>
<tr>
<td>Where abundant units are variably channelised, clast-supported textures, crudely stratified, imbricated and have well stratified sandstone caps, this suggests depositional systems conducive to more watery flows (e.g. wet fans)</td>
<td></td>
</tr>
<tr>
<td>Tend to terminate upwards with finer-grained tightly packed conglomerates.</td>
<td>Many show marked upward increase in matrix content particularly near top (Nemec et al., 1980; Kelling and Holroyd, 1979)</td>
</tr>
<tr>
<td></td>
<td>On passing into water, debris flow may reduce its thickness and distally pass into lobes. This is particularly true of low viscosity debris flows and results in gravel lenses with little basal scour, and possible loading interbedded with marine or lacustrine fines.</td>
</tr>
</tbody>
</table>
Figure 5.7  Some of the typical features of subaerial mass-flow deposits (Nemec and Steel, 1984).
Description
Interbedded with both finer and coarser reddened sediments, the sandstones are commonly associated with a calcrete cap. Although occasionally showing scour and channel-shaped geometries, they are more often flat bedded and are in some cases slightly more laterally persistent than the conglomerates and sandstone facies described above. Wedging out geometries can still be seen on a scale of 100 m or more.

Maximum bed thickness for these red sandstones recorded during this study was 30 cm and generally they were 10-20 cm thick. Structures within the beds varied greatly with well developed cross-bedding observed in the Yaylaalan section and cross-lamination at South Kargi. Parallel- and cross-laminations are common in the thinner units of the Saburlar section, while cross-sets characterise the thicker units which also tend to be the most laterally impersistent. Massive sandstones with no structures or grading were observed in the Aspendos section. This is rather exceptional in the context of outcrops elsewhere in the study area as normal grading is common.

Interpretation
Once again, despite the facies association with caliche horizons, red colouration and lack of marine fossils which suggests a ubiquitous subaerial environment of deposition, the variety of structures in this group almost certainly indicates deposition by more than one mechanism. It is suggested that channelised sandstones showing cross-bedding were deposited by the migration of braided stream bars (Collinson, 1986). These are generally uncommon however and the dominant sharp-bedded sandstone units with parallel and cross laminations interbedded with siltstone and developed caliche horizons may have been deposited by over bank flows (Steel and Aasheim, 1978; Tunbridge, 1981). Towards the top of the continental part of the Saburlar section which is progressively more muddy, the sandstone units may represent sheet floods distal to the active fan channel (Steel and Aasheim, 1978; Tunbridge, 1981; Hubert and Hyde, 1982).
sandstones resemble those described by McCallum (1989) from the Kakkaristra Formation in southern Cyprus.

5.5.1.5 Calcretes and associated fine-grained sediments

Description
Calcrete observed in the study area occurs in three broad morphological groups:

- as laterally continuous layers;
- as semi-continuous lenses and
- as more isolated nodules or pipes (Fig. 5.8).

Occasionally a continuum between these morphological states is visible most commonly between the lenses and the laterally continuous layers. Calcrete is intimately associated with the finest grained sediments, generally red mudstones and siltstones which are often bioturbated. Where most abundant, as at South Kargi (Fig. 5.1), calcrete horizons are interbedded with sandstones and siltstones and rare fine-grained conglomerate. Coarser conglomerate is generally absent.

Although abundant throughout the study area, laterally continuous calcrete layers are rarely thicker than 3-4cm, considerably thinner than the 2-3m reported by Allen (1974) and Leeder (1975). Siltstone, fine sandstone or mudstone layers containing lenses of calcrete can be up to 20cm thick, but in some cases it is not clear whether all the lenses are in situ. The dimensions of the vertical pipe structures vary from 2cm to rather irregular structures up to 10cm in length (Fig. 5.8). It is perhaps worthy of note that these vertical structures are rarely found in mudstone where the best continuous and semi-continuous calcrete layers are developed. Instead, they are found in coarser-grained siltstone. Root traces are very uncommon, but can be found at the Kırkkavak breach locality near Dumanli in the north of the Köprü basin (Fig. 5.1). Figure 5.9 shows rootlets picked out in green in medium-grained sandstone.
Figure 5.8 Morphologies of calcrete. a) Laterally continuous layers (South Kargi, central Aksu basin); b) nodular calcrete (Yaylaalan, northern margin of the Manavgat basin.)
Figure 5.9  Evidence of roots in a medium sandstone at the Kirkkavak breach locality near Dumanli, North Kopru basin.
Interpretation

Calcite pedogenesis is a common form of mineralisation in red sequences (Collinson, 1986) indicating soil formation in an arid to semi-arid climate (Butzer, 1964). It has been suggested that vertical pipes, sheets and more isolated nodules may compare with modern rhizoliths (e.g. Klappa, 1980). Mature soil profiles require about 10,000 yrs and imply that for periods of this sort of duration little or no deposition took place, the water table was deep and the alluvial plain well drained (Leeder, 1975). In general however, none of the even continuous calcrite horizons showed a really mature soil profile i.e. substantial thickness and laminar structures (Allen, 1974). This immaturity probably indicates that periods of low sediment supply were not long enough to allow soil maturation and this is born out by the association of calcretes with coarse clastics even when best developed. Soil development is generally thought to be due to a period of incision and terrace development (cf. Tandon and Narayan, 1981) although it may also be caused by avulsion away from the site of pedogenetic activity. No evidence of terrace development was observed and it is suggested that as the argument for the immature development of the soils is dependent on high sediment influx, avulsion may be the primary control.

The paucity of evidence of plant material bears out the lack of vegetation, suggested as one possible reason for the lack of soils and associated fine-grained material (Harvey, 1984).

5.5.1.6 Interpretation of sub-facies association A

A summary of the description and interpretation of the sub-facies discussed above is given in table 5.4. The sub-facies in group A have all been interpreted as being deposited under continental conditions. The successions containing the facies described above are dominated by clast-supported conglomerates and coarse sandstones with lenticular geometries whose structures and fabric indicate deposition by mass flow and braided-stream processes. Immature caliche is observed associated with finer sediments, indicating palaeosol development when sediment influx and drainage conditions were suitable. The association of these sub-facies suggest that they were deposited as part of an alluvial fan.
Table 5.4  Summary of the features of sub-facies association A.

<table>
<thead>
<tr>
<th>Name</th>
<th>Lower boundary</th>
<th>Upper boundary</th>
<th>Lateral variation</th>
<th>Thickness</th>
<th>Sorting</th>
<th>Current structures</th>
<th>Matrix</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Clast-supported angular conglomerate</td>
<td>unconformable</td>
<td>?</td>
<td>wedges out over 100m</td>
<td>&gt;50m total</td>
<td>very poor</td>
<td>planar cross-bedding, rare channel structures, imbrication</td>
<td>little</td>
<td>fault talus</td>
</tr>
<tr>
<td>Clast-supported conglomerate and coarse sandstone</td>
<td>erosive</td>
<td>planar</td>
<td>thins to nothing over several hundred meters</td>
<td>0.5-2m/bed</td>
<td>massive-moderately sorted</td>
<td>planar cross-bedding</td>
<td>red, silt</td>
<td>sheet flood and channelised braided stream deposits.</td>
</tr>
<tr>
<td>Matrix-supported conglomerates</td>
<td>possibly undulating, but not very erosive</td>
<td>planar-irregular</td>
<td>wedges out over 100m</td>
<td>0.5-several meters</td>
<td>poorly sorted to crudely stratified</td>
<td>crude fining upward sequences</td>
<td>red, silt</td>
<td>debris flow</td>
</tr>
<tr>
<td>Laminated sandstones</td>
<td>generally planar, occasional scour and channel shapes</td>
<td>planar</td>
<td>laterally persistent on a scale of &gt;100m</td>
<td>up to 30cm</td>
<td>cross-bedding, cross-lamination, parallel laminations, often normally graded</td>
<td>moderately sorted</td>
<td></td>
<td>braided stream bars, overbank flows, sheet floods.</td>
</tr>
<tr>
<td>Calcretes and associated fine-grained sediment</td>
<td>gradational</td>
<td>gradational-sharp</td>
<td>rarely laterally continuous</td>
<td>&lt;3-4cm</td>
<td>continuous layers, lenses, nodules, pipes</td>
<td></td>
<td>mud-siltstone</td>
<td>paleosol</td>
</tr>
</tbody>
</table>
system (Nemec and Steel, 1984). Within the alluvial conglomerate association, braided stream processes are thought to be the generally the most active (Nemec and Steel, 1984). Evidence reported here suggests that there is a continuum between traction deposition in streams and mass flow deposition, leading to a difficulty in identifying the dominant depositional process.

Historically, alluvial fans and fan-deltas are associated with tectonically active basin margins (e.g. Bull, 1977 on Death Valley). Here, coarse-grained successions resemble the alluvial sequences seen in the Humbolt Range, Nevada, at present the subject of a Ph.D. study by M. Stewart at Leeds University. Maizels and McBean (1990) suggest that the controls on alluvial fan development are primarily precipitation regime, base level changes and network expansion together with tectonic, pedogenic and fluvial processes.

Humid conditions would result in more persistent river flow resulting in meander development (Jackson, 1978; Arch; 1983; Bridge, 1985). It would also promote more extensive vegetation increasing slope stability and therefore reducing both the quantity and grainsize of the sediments supplied to the fan. Thus the preserved record of humid climatic conditions would be meandering channels through relatively fine-grained bank material (Maizels and McBean, 1990). By contrast, arid conditions would result in flashy ephemeral flow accommodated in large distributary channel systems (Maizels and McBean, 1990). Sub-facies association A described above, resembles sequences documented by Maizels and McBean (1990) in the upper part of a Cenozoic alluvial fan under semi-arid climatic conditions and the fanglomerates of the Quaternary of Cyprus (Poole, 1990). They have therefore been interpreted as having been deposited under semi-arid conditions.

The dominantly coarse grainsize suggests, both that the sequences discussed here are proximal with respect to the source and that the hinterland was tectonically active and therefore able to maintain coarse clastic dominance. Fining upwards sequences typical of lobe abandonment and of prograding alluvial-fan sequences are rare, occurring potentially only at Saburlar and Bozburun Dag. Possible
reasons for the absence of such sequences are that either active tectonic
subsidence or rising eustatic sea level or a balance between the two results
in relative sea level keeping pace with progradation and either
maintaining the gradient of deposition or steepening it.

5.5.2 Sub-facies association B

A list of the sub-facies discussed in this section, their localities with reference to figure 5.1 and the formations in which they are found is given in table 5.5.

Table 5.5 List of the sub-facies in sub-facies association B, their localities and formations.

<table>
<thead>
<tr>
<th>Sub-facies</th>
<th>Localities</th>
<th>Formation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Organic-rich facies</td>
<td>Bozburen</td>
<td>Kizildag</td>
</tr>
<tr>
<td></td>
<td>Yaylaalan</td>
<td></td>
</tr>
<tr>
<td>Green clays</td>
<td>Bozburen</td>
<td>Kizildag</td>
</tr>
<tr>
<td></td>
<td>Yaylaalan</td>
<td></td>
</tr>
<tr>
<td>Laminated limestone and lime mud</td>
<td>Kepezlenli</td>
<td>Calcaire de Kepez (Kizildag)</td>
</tr>
<tr>
<td></td>
<td>Yaylaalan</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Bucakkoyu</td>
<td></td>
</tr>
<tr>
<td>Gastropod siltstones</td>
<td>Kapikaya</td>
<td>Aksu</td>
</tr>
<tr>
<td>Limestone breccia</td>
<td>Kesme gorge</td>
<td>Aksu</td>
</tr>
<tr>
<td></td>
<td>Köprü bridge</td>
<td></td>
</tr>
</tbody>
</table>

5.5.2.1 Organic rich horizons

Description
Organic-rich horizons occur very infrequently and are generally merely thin concentrations of redeposited detrital plant material. However in two localities, Bozburen Dag and Yaylaalan, thicker organic-rich horizons are developed. In both cases the organic-rich horizons are associated with a sequence of coarse continental clastics described in section 5.5.1 above overlain by green clays discussed below in section 5.5.2.2 (Figs. 5.10 and 5.11). At Yaylaalan the organic horizon is poorly exposed and is not thought to be laterally extensive (e.g. on a scale of 10-15m). At Bozburen
Figure 5.10 Log of the Bozburun coal section showing the interbedded marine and alluvial facies with cross-sets indicating that the current direction for the conglomerates was to the south and east.
Figure 5.11  Log of the Yaylaalan transition from continental conglomerates to fully marine shallow-water limestones. Photograph of the bioturbated limemud part of the succession.
Dag however, two organic-rich layers have been mined by local inhabitants for 10 years and they can be traced for several hundred metres along strike. To the west they eventually disappear underneath the huge pile of coarse clastics that makes up the Bozburun Dag recumbent fold, while to the east they are faulted out by a high angle reverse fault (chapter 7).

The organic-rich layers at Yaylaalan and Bozburun Dag differ considerably from each other. At Yaylaalan, the layer is 60cm thick and consists of a dark, clay-rich, poorly lithified horizon which contains no identifiable organic or clastic components. At Bozburun Dag, by contrast, the two organic-rich horizons are 75cm to 1m thick and are much better indurated. A polished section revealed, in reflected light, that they consist of concentrated organic fragments (dominantly fusain) and strands with abundant fine-grained silt and biogenic detritus, much of which was broken. Rare whole, but small *Potamid* gastropods were identified preserved in primary aragonite, but flattened along the plane of fissility. No vitrinite was identified.

**Interpretation**

The environment required to preserve organic-rich material is still a matter of debate (e.g. sapropel formation in the Mediterranean; Ryan and Cita, 1977; Calvert, 1987; Lucci et al., 1994). At Bozburun Dag, however, there is evidence from the lack of bioturbation and the preservation of primary aragonite, of an anoxic environment, at least periodically, allowing the build up of non-decomposed organic material. The microscopic structure of these horizons reveal that the organics are not entirely *in situ* although some of them may be. Influx of terrigenous and biogenic detritus led to less concentrated organic-rich layers and it is along these that the bedding-parallel fissility at Bozburun Dag has developed. The Yaylaalan organic-rich horizon is perhaps best interpreted as an oil shale. It is patently immature and was probably never deeply buried.

The lack of vitrinite in both of the samples collected implies that either the depth of burial, or the time of burial, or both, were not sufficient to mature these horizons to high quality coal, although it is clearly good enough to maintain a local mining industry. The current height of the
Bozburun Dag peak above the coal seams is in the order of 1.5km and it is therefore safe to assume that it must have been buried to at least this depth at the time the nappe formed, during the Aksu Phase of compressional deformation at the end of the Messinian (e.g. 5.5Ma; chapter 7).

5.5.2.2 Green clays

Description
These occur in two localities, Yaylaalan and Bozburun Dag (Fig. 5.1) and in both cases they are intimately associated with organic-rich layers (Figs. 5.10 and 5.11). At both localities the clays are underlain by thick successions of continental conglomerates and sandstones and at Bozburun Dag they are overlain by a similar succession (Fig. 5.10). At Yaylaalan however, green clays and an associated organic-rich horizon are overlain by non-marine limestones (Fig. 5.11 and section 5.6.2.3, below). The exposure of this facies is poor and does not allow lateral facies changes to be observed, but they are not thought to be laterally extensive.

These sticky, green and structureless clays contain no fauna at Yaylaalan. At Bozburun however some horizons contain abundant *Potamid* sp. gastropods which XRD analysis showed to have preserved their original primary aragonite mineralogy. All three washed clay samples revealed rare benthic and planctic foraminifera including *Orbulina*. The thickness of these horizons varies from a few centimetres up to 3m at Yaylaalan.

Interpretation
Potamid gastropods are well known for their lagoonal affinities (e.g. Plaziat, 1991; 1993) and their setting here, in fine-grained clays associated with anoxic sediments resembles that described by Plaziat (1984) from the Eocene of Corbières region of southern France. Periodic marine incursions possibly due to storms, washed foraminifera into this low energy environment and may well have altered salinities and oxygen levels. The structureless nature of the clays and the abundance of Potamids at Bozburun suggest that bioturbation was active and this indicates that anoxia was only periodically present, allowing the
accumulation of organic-rich horizons described above. The presence of *Orbulina* indicates that the succession at Bozburun is Langhian, or younger (Berggren *et al.*, 1985; Bizon *et al.*, 1974). The potential for dating the primary aragonite preserved in these gastropods using Sr dating techniques is limited by the probability that some of the $^{87}\text{Sr}/^{86}\text{Sr}$ signature is derived from a continental source, thus invalidating the use of the Sr sea water curve (e.g. McKenzie, 1988, Müller *et al.*, 1990, Elderfield, 1994).

5.5.2.3 Laminated limestone and lime mud

*Description*

In the Miocene laminated limestone and lime mud only occur in two localities. Both of these are on the northern margin of the Manavgat basin, at Yaylaalan and Kepezbelenli (Figs. 5.1 and 5.12). The Kepezbelenli locality was documented by Monod (1977) and he named this limestone "le Calcaire de Kepez", (section 5.3.2.2 above). Monod did not study the sediments far enough west to incorporate the similar limestone at Yaylaalan into his thesis.

Kepezbelenli, the thicker succession of the two, spans 17m of vertical stratigraphy and comprises a variety of different carbonate textures. The carbonates directly overlie Mesozoic radiolarian chert and are overlain by coarse conglomerates (Fig. 5.13). Mapping (Monod, 1972) shows that these carbonates are of limited lateral extent (Fig. 5.12). The limestone contains abundant plant debris and rare Charophytes. Pollen analysis carried out by Monod (1977) failed to produce age-diagnostic species. The succession at Yaylaalan (Fig. 5.11) is significantly different from that at Kepezbelenli. A thick pile of continental conglomerates and sandstones with less frequent interbedded finer-grained siltstones and variably developed caliche are overlain by interbedded sticky, green clays, fine sandstones and one organic-rich layer (sections 5.5.2.1 and 5.5.2.2). This is overlain by 2m of carbonate and is itself overlain by fossiliferous shallow water limestone (chapter 3).

The dominant structuree of the limestone at Kepezbelenli is millimetre to centimetre, sometimes irregular, algal laminations in variable beige to
Figure 5.12 Location map of the Kepezbeleni section on the northern margin of the Manavgat basin showing the limited lateral extent of the basal lacustrine limestone and its relationship to the other Miocene sediments in the area (after Monod, 1977).
Figure 5.13 Log of the Kepezbelenli section on the northern margin of the Manavgat basin (Fig. 5.1).
brown carbonate, rich in plant fragments. Large numbers of holes less than a millimetre across can be seen throughout the rock. This facies also occurs as clasts in a breccia bound with algal strands in the same sequence. Similar carbonate, but with a different texture occurs towards the top of the section at Kepezbelenli. Here, beige-brown carbonate forms concentric rings around casts of tube structures assumed to be plant stems. White powdery lime mud occurs both at Kepezbelenli and at Yaylaalan, but in the latter locality it is heavily burrowed with a network of large branching structures (Fig. 5.11). Laminated limestones similar to those described at Kepezbelenli underlie this burrowed horizon, but the laminations are contorted. Beds either side of this deformed layer do not show deformation structures.

Interpretation

A lacustrine setting is inferred from the absence of marine fauna, the presence of algal laminations and the abundance of plant material. Quaternary tufas occur in abundance in the study area and are currently the subject of part of a Ph.D. project undertaken by C. Glover at Edinburgh University. She defines tufa as a cold water deposit which is controlled by algal growth in carbonate-rich waters producing a generally porous texture. Algal strands around plant stems at Kepezbelenli, closely resemble freshwater tufa textures in Quaternary deposits of the area. These have been interpreted as forming in association with fresh-water springs. (Glover, 1995). Lime mud horizons generally indicate slightly deeper water conditions in small lakes associated with the pools (C. Glover, pers. com., 1994). The proximity of the fully marine carbonates in the Yaylaalan forces the question of whether the lake in which these carbonates were deposited was influenced by marine conditions. It was hoped that the presence of brackish-water ostracods would shed light on this question, but no ostracods of any sort were detected. There is no evidence of marine fossil lags indicating periodic inundation by sea water in either locality, but as laminated algal structures can form in brackish and hypersaline water (e.g. Burdur Lake; Price and Scott, 1991) marine incursions cannot be ruled out particularly at Yaylaalan.

The contorted laminated limestones have been interpreted as slump structures formed whilst the algal laminations were only partially
cemented. This may have been triggered merely by slope steepening or by tectonic activity. The brecciation seen at Kepezbelenli may also have been generated by one of these two mechanisms and subsequently bound by recurrent algal colonisation.

5.5.2.4 Gastropod siltstones

Description
This facies is only found in one locality, in the far north of the Aksu basin, to the south of Kapikaya (Fig. 5.1). It consists of a 10-20m thick sequence of homogenous-looking siltstones some of which are very rich in a monospecific gastropod. No lateral variations in bed thickness of the siltstones was observed, but exposure is limited to a track cutting only half a kilometre in length. Basal facies relations here were not exposed, but the siltstones pass upwards into an increasingly coarse-grained succession of sandstones and conglomerates. Interbedded with these in the finer horizons immature caliche is developed.

The siltstones are bedded in a 5-10 cm scale and show few sedimentary structures. Horizons containing abundant gastropods were sampled and resemble the freshwater gastropod, *Viviparus*. (Further investigation into the identity of this gastropod is being undertaken in conjunction with a specialist at Université d'Orsay, France.) Previously, mammal vertebra have been found in this succession by Akay (A. Poisson, pers. com., 1994), none were discovered during this study however.

Interpretation
If, as suggested by the initial identification, the gastropod found here is a freshwater species, taken in conjunction with the mammal vertebra this suggests that these silts were deposited in a fresh water, lake environment. During silt deposition, conglomerate incursions were few, but these increased with time and the transition from lacustrine facies to alluvial facies suggests that either infill or dessication of the lake led to this transition. No evidence of dessication was observed and considering the proximity of Kapikaya and fault generated talus (section 5.5.1.1), infill of lacustrine accommodation space due to avulsion of alluvial channels is more likely to have caused the facies change.

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Dating of this succession has proved extremely difficult. Both Akay and Poisson attempted to find diagnostic ostracods within these siltstones, but failed. Gutnic et al. (1979) and Akay and Uysal (1985) give a Tortonian age for the conglomerate sandstone sequence overlying these siltstones and it is therefore assumed that in the absence of an unconformity surface between the two, the siltstones are Middle-Late Miocene in age.

5.5.2.5 Limestone breccia

Description
This facies occurs only in the north of the Köprü basin and is best exposed at the Kesme gorge locality and at the Köprü bridge (Fig. 5.1). It unconformably overlies folded and faulted basement limestone in laterally continuous sheets (Fig. 5.14) and is itself overlain by a sequence of matrix supported conglomerates and sandstones described in section 5.6.4.1 and 5.6.4.3 above. This facies has not been previously recognised probably because of its similarity of the appearance in weathered exposure to basement limestone (Fig. 5.14). In 1974, when Dumont was writing his thesis on the area, he noted that the road south from Egirdir stops at the village of Kesme and it is quite possible therefore that he never saw the exposures created by the dirt track that now runs south from Kesme, through the Kesme gorge to Beskonak in the central part of the Köprü basin.

The age of all the Neogene sediments in the North of the Köprü basin is still a largely unsolved problem. The overlying polymict conglomerates are thought to be Miocene in age for the following reasons:

- Similar conglomerates on the east side of the Kirkkavak fault have been dated as Tortonian (Dumont, 1974);
- A polymict conglomerate-filled "breach" in the Kirkkavak fault exists just to the north of the Kesme gorge area (Fig. 5.16) which contains palaeocurrents which indicate that the transport direction into the north of the Köprü basin is likely to have been at least in part from the north east, through this conduit (Fig. 5.36).
Figure 5.14  Photograph of limestone breccia at Kesme (north Köprü basin) unconformably overlying Mesozoic carbonates.

Figure 5.15  Interbedded limestone breccia and polymict conglomerate at Kesme gorge, north Köprü basin.
This "breach" is now exposed along the scarp of the Kirkkavak fault indicating that it pre-dates the last movement on the fault. The last large scale uplifting event of the fault was probably in the latest Messinian during the Aksu Phase (chapter 7). If this is the case then the polymict conglomerates are pre-Messinian in age. It is suggested that the underlying limestone breccias can be included in the discussion of the Miocene sedimentary system because towards the top of the breccias they are interbedded with the overlying polymict conglomeratic succession. This is illustrated in figures 5.15 and 5.17 which is a log of this transition section 100m to the north of the Kesme Gorge.

The limestone breccia sub-facies is distinctive comprising of 30cm - 1.5m beds of moderately sorted angular breccia whose clasts are entirely limestone of various sorts (Fig. 5.17). Within the clast-supported fabric no structures are visible that indicate traction deposition. Intraclast space is filled with white carbonate silt, similar to that described in section 5.5.2.3 as lime mud.

Interpretation

The clast-supported fabric and angular nature of the clasts suggests that this breccia may have been deposited as slope talus. The fine-grained carbonate matrix indicates that deposition occurred in subaqueous conditions. In the rest of the study area, angular limestone clasts are incredibly rare and relative roundness has not proved a useful tool in deducing distance from the source area. In this case however, the angularity of the limestone clasts indicates that deposition must have occurred very proximally relative to the source and that very little reworking took place subsequently. It is for this reason that it is suggested that the interbedded succession of limestone breccia and polymict conglomerates shown in figures 5.15 and 5.17 is not the result of reworking and redeposition of an older, potentially Mesozoic limestone breccia, but rather primary deposition from different sources at the same time.

Although the angular clast-supported conglomerate at Kapikaya (section 5.5.1.1) and the limestone breccia discussed above, differ markedly in terms of their composition, matrix, sorting and inferred environment of
Figure 5.16 "Breach" in the Kirkkavak fault to the west of Dumanli in the north of the Köprü basin. Deformed Triassic limestone walls are filled with polymict conglomerate thought to be of Miocene age. Note television tower for scale.
Figure 5.17 Log of the interbedding of limestone breccia to polymict conglomerate at the Kesme gorge locality, North Köprü basin. Photographs show the different appearance of the two rock types.
deposition it is thought that they formed by similar processes, e.g. the mass wasting of steep slopes ultimately generated by tectonic movement. Table 5.6 lists the differences between the two breccias. It is suggested that the limestone breccia is better sorted and far more laterally continuous because it represents more distal talus deposition within a subaqueous environment permitting transport of finer material over larger distances.

Table 5.6 The differences between the limestone breccia at Kesme and the mega-breccia at Kapikaya.

<table>
<thead>
<tr>
<th></th>
<th>Limestone breccia</th>
<th>Mega-breccia</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>monomict</td>
<td>polymict</td>
</tr>
<tr>
<td></td>
<td>moderately sorted</td>
<td>poorly sorted</td>
</tr>
<tr>
<td></td>
<td>white matrix</td>
<td>no matrix</td>
</tr>
<tr>
<td>subaqueous environment of deposition</td>
<td>subaerial environment of deposition</td>
<td></td>
</tr>
</tbody>
</table>

5.4.2.6 Interpretation of sub-facies association B

A summary of the description and interpretation of the sub-facies discussed above is given in table 5.7. The sub-facies in group B have all been interpreted as having been deposited in subaqueous lacustrine or lagoonal environments, but in close association with sub-facies interpreted as alluvial deposits. This suggests that lakes and lagoons were at least episodically present during both Lower (Kizildag Formation) and Upper Miocene (Aksu Formation) accumulation of conglomerate-rich successions. In general these facies represent low-energy environments with variable salinities and oxygen contents. The fauna, where present is specific to this sort of environment although there is evidence of periodic marine influence in some of the localities and general close proximity to marine facies (Figs. 5.11 and 5.13, ). Figure 5.18 is a generalised model of how these facies may have accumulated between pulses of coarse clastic sedimentation and an ever encroaching sea.
Table 5.7  Summary of the features of sub-facies association B.

<table>
<thead>
<tr>
<th>Name</th>
<th>Lower boundary</th>
<th>Upper boundary</th>
<th>Lateral variation</th>
<th>Thickness</th>
<th>Sorting/composition</th>
<th>Structures</th>
<th>Matrix</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Organic-rich horizons</td>
<td>planar</td>
<td>planar/eroded into by overlying sediments</td>
<td>continuous &gt;0.5km (Bozburun Dag) 10-15m (Yaylaalan)</td>
<td>0.6-1m</td>
<td>poorly sorted fragments, also contains gastropods and terrigenous material</td>
<td></td>
<td>immature coal</td>
<td></td>
</tr>
<tr>
<td>Green clays</td>
<td>non-erosive</td>
<td>planar/eroded into by overlying sediments</td>
<td>not laterally persistent over 100's meters</td>
<td>0.05-3m</td>
<td>Potamid gastropods, Orbulina</td>
<td>structureless</td>
<td>lagoonal clay</td>
<td></td>
</tr>
<tr>
<td>Laminated limestone and lime mud</td>
<td>unconformity (Kepez/belenli), flat (Yaylaalan)</td>
<td>eroded into by overlying conglomerates (Kepez/belenli), flat (Yaylaalan)</td>
<td>laterally persistent &lt;300m</td>
<td>2-5m</td>
<td>charophytes, algae, plant stems (Kepez/belenli)</td>
<td>burrowing, folded (Yaylaalan)</td>
<td>lacustrine limestone</td>
<td></td>
</tr>
<tr>
<td>Gastropod siltstones</td>
<td>not observed</td>
<td>gradational</td>
<td>&lt;0.5km</td>
<td>5-10cm</td>
<td>moderately sorted, Viviparus-like gastropods, mammal bones</td>
<td>silt</td>
<td>lacustrine silts</td>
<td></td>
</tr>
<tr>
<td>Limestone breccia</td>
<td>unconformable</td>
<td>planar</td>
<td>laterally persistent</td>
<td>0.3-1.5m</td>
<td>entirely Mesozoic limestone clasts</td>
<td>angular, clast-supported fabric</td>
<td>white carbonate silt</td>
<td>slope talus deposited in a lake</td>
</tr>
</tbody>
</table>
Figure 5.18 Generalised model of the depositional environments and processes active in the formation of the lagoonal and lacustrine sediments observed in the study area. No precise scale or orientation is intended.
5.5.3 Sub-facies association C

A list of the sub-facies discussed in this section, their localities with reference to figure 5.1 and the formations in which they are found is given in table 5.8.

**Table 5.8** List of the sub-facies in sub-facies association C, their localities and formations.

<table>
<thead>
<tr>
<th>Sub-facies</th>
<th>Localities</th>
<th>Formation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Echinoid-scaphopod facies</td>
<td>Yesilbag</td>
<td>Tepekli</td>
</tr>
<tr>
<td></td>
<td>Aspendos</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Deniztepesi</td>
<td></td>
</tr>
<tr>
<td>Porites bafflestones</td>
<td>S. Kargi</td>
<td>Tepekli</td>
</tr>
<tr>
<td></td>
<td>Alarahan</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Kesme</td>
<td></td>
</tr>
<tr>
<td>High-angle cross-bedded conglomerates</td>
<td>S. Kargi</td>
<td>Tepekli</td>
</tr>
<tr>
<td></td>
<td>Alarahan</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Bozburun</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Kargi baraj</td>
<td></td>
</tr>
</tbody>
</table>

5.5.3.1 Echinoid-scaphopod facies

**Description**

This facies has been described in detail in chapter 3 section 3.6.3.1 from the view of its large biogenic carbonate component. It is discussed again here because of its implications for the interaction between terrigenous detritus, the formation of biogenic carbonate, fresh water input and marine generated wave action. To save repetition only a brief review is given here.

This facies occurs either just below reef limestone and just above continental red beds as at Aspendos (Figs. 5.1 and 5.6), or in lateral continuity with a framework reef structure as seen at Yesilbag and South Kargi. It consists of coarse micro-conglomerate with abundant whole and fragmented scaphopods, echinoids, large benthic foraminifera and algal rhodoliths up to 5cm in diameter. These rhodoliths encrust rounded pebbles, most commonly veined grey Mesozoic limestone (Fig. 3.17).
They are irregular in shape and individual layers do not encircle the clast, but can be traced out to their terminations, which are abrupt and often ragged. Although algal encrustation is visible on many of the micro-conglomerate clasts, the largest rhodoliths are developed on clasts significantly larger than average.

Random orientation of scaphopods is the norm in this facies, but millimetre scale, graded laminations with scaphopods parallel to these are occasionally seen (Fig. 3.18). The scaphopod most commonly found here is *Dentalium* (Upper Cretaceous-Recent) with its prominent longitudinal ribs. Most of the specimens found were fragments.

**Interpretation**

This facies is vitally important in terms of interpreting these shoreface facies because it is the closest facies to a true beach. The environment envisaged is a very shallow water one (probably only a few meters deep) indicated by the intimate association with reef framestones (e.g. at Yesilbag), and an environment of high-energy (Orzag-Sperber et al. 1977), where constant movement, probably wave action, shift and roll over the sands allowing the formation of rhodoliths. Deposition of coarse material is by periodic fluvial input rather than a constant rain of sediment and this may have generated laminations. Algal rhodoliths encrust the free surfaces of clasts protruding from sediment, colonising other surfaces when currents reorientated the pebbles.

5.5.3.2 Porites bafflestones

**Description**

The best examples of this facies occur at South Kargi, Kesme and Yesilbag (Fig. 5.1) and have already been described in chapter 3 (section 3.6.1.2). They are discussed again here in the light of the interactive nature of their deposition between fluvially sourced terrigenous detritus and marine influences on the biogenic building of the bafflestone itself. Again, to avoid repetition only a brief summary of the description is given here.
These rocks are interbedded with coral framestones, coarse conglomerates with high angle cross-sets and minor amounts of unstructured clastics including fine-grained siltstones and sandstones with caliche development. At South Kargi for instance, five separate reef horizons are interbedded with coarse alluvial clastics (Fig. 5.19). As shown by Fig. 5.20 they do not persist laterally over 100m. Vertical coral sticks of Porites sp. can be seen to have produced a well spaced framework in which terrigenous sediment is deposited. Planar changes in grain size of the terrigenous components can be seen (e.g. bedding planes), but there is no apparent effect on the coral sticks, which pass through these boundaries uninterrupted.

Interpretation
In the geological record it is not possible to specify which of the many non-preserved environmental controls may have influenced the monospecific colonisation of Porites. However, it has been observed that Porites is a coral which is able to thrive in conditions of low salinity, low temperature and reduced water circulation (Marshall and Orr 1931, Manton 1935, Wells 1954 and Scoffin and Stoddart 1978). It is also associated with higher sediment flux than most other corals in the Miocene (e.g. Martin et al., 1989). The presence of these bafflestones at the boundary between continental and marine sediments bears out the hypothesis that these may have been fluvially influenced. The nature of the sedimentary beds through which the coral sticks pass uninterrupted indicates that these boundaries were formed while the coral heads protrude from them. This has three important implications:

- Sea level was higher than sedimentary boundaries at the time of bed deposition suggesting that accommodation space remained (Fig. 5.21).
- The ongoing input of terrigenous material and the continued uninterrupted growth of coral sticks suggests that relative sea level was rising at the time of deposition (section 3.8.1 for discussion of rate of relative sea level rise deduced from this).
- The coral framework may well have behaved firstly as a sediment trap and later as a barrier to further fluvially generated sediment basinward of the reef. Evidence for this can be seen at South Kargi where basinward of the Porites bafflestones (as determined from the palaeocurrents) Tarbellastrea framestones, which are essentially sediment free and finer-grained echinoid-scaphopod facies are deposited (Fig. 5.20).
Figure 5.19  Photomontage showing the five reef-horizons at South Kargi, central Aksu basin (Fig. 5.1), interbedded with coarse red alluvial clastics.
Figure 5.20 Correlation of the logged sections at South Kargi, central Aksu basin, and a model showing the sub-facies relationships deduced from these logs and palaeocurrent data, with reference to basinal and alluvial processes. (Key to symbols can be found on Fig. 3.7.)
Corals colonise and grow up to sea level

Sand input into reef, but insufficient to drown corals

Sea level rise allows corals to grow vertically using sand to support them

Conglomerate input into reef, but insufficient to drown coral

Sand input fills accommodation space and drowns coral

Figure 5.21  Cartoon of rising sea level versus sediment input in Porites bafflestone environments.
High angle, cross-bedded conglomerates

Description
High-angle (~25°) cross-bedded conglomerates are preserved at Alarahan and South Kargi (Fig. 5.20) and are interbedded with reef framestones. At Bozburun Dag they are interbedded with the lagoonal clays and organic-rich horizons (Fig. 5.10) and the angle of repose of the cross-sets is distinctly lower than it is at the other two localities (Fig. 5.22). Periodic interbedded red, caliche-rich, fine sandstones can also be seen in close vertical proximity to these conglomerates at South Kargi.

Lateral thickness variations are common in this sub-facies. Figure 5.20 shows two logged sections taken from the South Kargi locality which are spaced approximately 100m apart, but which document the same stratigraphic level. The difference in thickness between the cross-bedded units of each section at comparable levels is apparent. A similar feature can be seen on the two walls of the river gorge at Alarahan, separated by approximately 50m.

These conglomerates are coarse-grained and clast-supported. Their thickness ranges from up to 2-10m and they are often associated with erosive channel scours. A few are also imbricated with well-rounded recrystallised carbonate clasts dominating over angular cherts. In general these conglomerates are well sorted although a clear bimodal size distribution exists between the clasts and the micro-conglomerate-grade matrix which has a similar composition. Occasionally biogenic material is found within the conglomerates. These are most commonly coral clasts, but they can include oysters, gastropods, and bivalves of various types.

Interpretation
The association of these conglomerates with both fully marine reef framestones and red beds containing caliche, as can be seen at South Kargi (Figs. 5.19 and 5.20) suggests that these high-angle cross-bedded conglomerates formed at the boundary between continental and marine environments. There are two possible modes of deposition for these conglomerates, Gilbert-type delta foresets and channel mouth bars.
Figure 5.22 Photograph of high-angle cross-bedded conglomerates at the Bozburun Dag pass, north west Kopru basin.
According to Nemec (1990a) these two depositional environments can be distinguished using the criteria displayed in table 5.9.

Table 5.9  Distinguishing features of Gilbert-type delta foresets and channel mouth bars (Nemec, 1990a).

<table>
<thead>
<tr>
<th>Gilbert-type delta foresets</th>
<th>Channel mouth bars</th>
</tr>
</thead>
<tbody>
<tr>
<td>Very thick stratification, generally up to 100s m</td>
<td>Much thinner stratification</td>
</tr>
<tr>
<td>Foresets consist of alternating high and low</td>
<td>Thick laminae (0.4-3cm) that are generally inversely</td>
</tr>
<tr>
<td>density turbidites and debris flows (10-100cm)</td>
<td>graded and resemble grain-flow layers</td>
</tr>
<tr>
<td>Topsets are multiple channel fills</td>
<td>Topsets are parallel-laminated and ripple cross-laminated</td>
</tr>
</tbody>
</table>

Postma (1990) suggests that the control over which of these two delta front systems develops is dependent on stability, spacing, discharge and width of the distributary channels. He maintains that in shallow-water fan-delta system mouth bars will develop on moderate to low gradients which are generally well vegetated braid and coastal plains whilst the classic Gilbert-type delta front develops on steeper gradients characterised by ephemeral discharge. Figure 5.23 shows the 12 major prototype deltas suggested by Postma (1990).

The criteria outlined above suggest that the cross-bedded conglomerates described above may represent more than one delta-type. At Bozburun, the facies relations with organic-rich deposits suggests a well vegetated area, albeit locally, with periods of little coarse clastic input. This points to a relatively shallow gradient and the conglomerates here have been interpreted as having been deposited as mouth-bar-type foresets with a Gilbert-type profile (e.g. number 9; Fig. 5.23). These are similar to those documented by McCallum and Robertson (in press) in the Pliocene of Cyprus. Alarahan and South Kargi, with their higher angle cross-sets and association with immature soils, overbank fines and coarse conglomeratic channels are thought to have been deposited at small scale Gilbert-type delta fronts (e.g. number 3; Fig. 5.23). As in Cyprus (McCallum, 1989), there is little evidence that any of these cross-bedded
<table>
<thead>
<tr>
<th>Feeder system</th>
<th>Type A</th>
<th>Type B</th>
<th>Type C</th>
<th>Type D</th>
</tr>
</thead>
<tbody>
<tr>
<td>SHALLOW WATER DELTAS</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>HJULSTRÖM - TYPE</td>
<td><img src="image1" alt="Diagram" /></td>
<td><img src="image2" alt="Diagram" /></td>
<td><img src="image3" alt="Diagram" /></td>
<td><img src="image4" alt="Diagram" /></td>
</tr>
<tr>
<td>Shoal-water profile</td>
<td>1</td>
<td>2</td>
<td>7</td>
<td>8</td>
</tr>
<tr>
<td>'Classic' Gilbert type</td>
<td><img src="image5" alt="Diagram" /></td>
<td><img src="image6" alt="Diagram" /></td>
<td><img src="image7" alt="Diagram" /></td>
<td><img src="image8" alt="Diagram" /></td>
</tr>
<tr>
<td>GILBERT - TYPE</td>
<td>3</td>
<td>4</td>
<td>9</td>
<td>10</td>
</tr>
<tr>
<td>Debris cones</td>
<td>Gravitationally modified Gilbert type</td>
<td>Delta-fed submarine ramp system</td>
<td>Delta-fed thalweg and lobe system</td>
<td></td>
</tr>
<tr>
<td>DEEP WATER DELTAS</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>MOUTH BAR - TYPE</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>5</td>
<td>6</td>
<td>11</td>
<td>12</td>
<td></td>
</tr>
</tbody>
</table>
conglomerates formed a single, unified delta front as is required by the strict definition of Gilbert-type deltas. Instead, the cross-bedded units are thinner and highly localised.

Dunne and Hempton (1984) suggested that the development of Gilbert-type or mouth-bar-type deltas was dependent on the slope of the basin floor. This was later challenged by Nemec (1990a) who suggested that distributory channel variables outlined above controlled the delta type. On a smaller scale however, it has been shown that the angle of the basin slope is an important factor in determining the structures of the deposits developed and their process of deposition.

In a subaqueous environment, gravity induced processes dominate. Influx of particles to a subaqueous delta slope leads to steepening of the slope because the rate of deposition by fallout from suspension and bedload dumping decreases down slope. Above a certain angle (φi, the angle of initial yield) however, the slope surface becomes unstable and this leads to mass failure, transporting material down the slope face and leaving the previously steepened slope at a more gentle inclination. This angle (φr) is commonly referred to as the angle of repose (Van Burkalow, 1945). Allen (1970) argued that avalanching down slope will occur if the slope (with angle φ), is subjected to a sufficiently large downslope acceleration (a) where:

\[ a = a_{\text{max}} \times \frac{\phi_i - \phi}{\phi_i - \phi_r} \]

in which:

\[ \phi_i > \phi > \phi_r \]

\[ a_{\text{max}} = g (\tan \phi_i - \tan \phi_r) = \text{maximum possible downslope acceleration that will initiate a free-running avalanche.} \]

Allen (op. cit.) shows that the angle of repose (φr) is nearly constant for a given material despite differing avalanche conditions. This suggests the reason for the progradation of constant slope foresets at the front of delta systems as pictured in figure 5.24 (Nemec 1990b, p32).
CONICAL SUBAQUEOUS DELTA

Relative water level changes
Alluvial system
Basin
Aggradation/degradation
Aggradation
Progradation

GILBERT-TYPE DELTA

Relative water level changes
Alluvial system
Basin
Progradation
Aggradation/degradation
Aggradation

Figure 5.24  Schematic diagrams of conical and Gilbert type deltas, where $\beta_f < \beta_c$. (After Nemec, 1990b)
In the study area however, with the exception of the Kargi baraj site, the high-angle cross-bedded conglomerates are not directly associated with basement changes in slope, such as a basin margin or fault. Instead, these conglomerates (e.g. at Alarahan and South Kargi) are associated with reef-framestones. It is possible that the topographic structures produced by coral growth may have been sufficient to produce a gradient change at sea level able to initiate avalanching, from which progradational foresets could develop. Evidence of reef talus is abundant at Alarahan suggesting that basinward steep slopes were already active producing conical aggradational patterns (Fig. 5.24) and though probably too steep for parallel foreset development, these slopes were probably exploited and either reduced in inclination by constant avalanching (Allen, 1970) or by wave action (Nemec, 1990b) of which there is evidence in the form of echinoid-scaphopod facies deposits at South Kargi (Fig. 5.20) until they attained the critical slope. Subsequently, cascading sedimentation down the front of the slope produced prograding cross-sets similar to those described by Postma and Roep (1985) and Colella et al. (1987).

As has already been discussed above in section 5.5.3.2 concerning Porites bafflestones, the controls on this sort of clastic-carbonate interaction seems to be dependent on both sediment flux and relative sea level change. As is shown at Alarahan in figure 5.25, for example, following deposition of cross-bedded conglomerates, relative sea level rise and low sediment influx allowed the colonisation of the frontal foreset by Poritid reefs. This sort of interrelationship contains a potential problem. Reef colonisation enhances the preservation potential of alluvial/delta front sediments and it would be easy to exaggerate their importance in terms of the depositional system.

Lack of transition facies in any one section may be due to avulsion and rapid colonisation by reefs which protect underlying sediments from wave erosion thus increasing preservation potential. Thus, the best section preserved are those which were not active clastic channel systems at the time of the rapid relative sea level rise.
Figure 5.25  Field sketch of high-angle cross-bedded conglomerates and reef limestones at Alarahan. The cliff drawn here is exposed on the left bank of the river (Alara Çay) beneath the Selçuk castle. Reef numbers are with reference to the section logged at this locality (Fig. 3.4).
<table>
<thead>
<tr>
<th>Name</th>
<th>Lower boundary</th>
<th>Upper boundary</th>
<th>Lateral variation</th>
<th>Thickness</th>
<th>Sorting/composition</th>
<th>Structures</th>
<th>Matrix</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Echinoid-scaphopod facies</td>
<td>faintly erosive</td>
<td>channelised/planar</td>
<td>changes to domal coral framestone/ Porites bafflestone</td>
<td>0.01-2m</td>
<td>rhodoliths, scaphopods, echinoids, grit and pebbles</td>
<td>alignment of scaphopods</td>
<td>grit-siltstone</td>
<td>wave-reeled beach gravel.</td>
</tr>
<tr>
<td>Porites bafflestones</td>
<td>gradational</td>
<td>sharply erosive</td>
<td>varies to echinoid-scaphopod facies</td>
<td>~3-5m</td>
<td>Porites sticks, with horizons of conglomerate and gritstone</td>
<td></td>
<td>silt-grit</td>
<td>fluvial sediment trap at the marine-freshwater interface</td>
</tr>
<tr>
<td>High-angle cross-bedded conglomerates</td>
<td>planar-erosive and channel shaped</td>
<td>sometimes convex up</td>
<td>lateral thickness variations common on a scale of 50m</td>
<td>2-10m</td>
<td>well sorted, rare biogenic clasts</td>
<td>high angle (~25°) cross-sets, imbrication</td>
<td>grit</td>
<td>Gilbert-type delta foresets</td>
</tr>
</tbody>
</table>
A summary of the description and interpretation of the sub-facies discussed above is given in table 5.10. All the sub-facies in group C have been interpreted as having been deposited at the continental-marine interface. Shoreline sediments are the most sensitive recorders of relative sea level change. Throughout the study area, wherever these sediments are preserved they record a rise in relative sea level. They also record basinal processes and the presence of the echinoid-scaphopod facies suggests that wave action was an important processes in reworking delta front sediments. Note that overall there is a transition from a continental environment of deposition to a marine environment (Oymapinar Limestone and Karpuzçay Formation at the Kargi baraj locality, Fig. 5.1). In detail however, there is abundant evidence of the small-scale prograding nature of the shoreline at least in vertical section. Figure 5.26 is a generalised model for the formation of the shoreline facies described above.

### 5.5.4 Sub-facies association D

**Table 5.11 List of the sub-facies in sub-facies association C, their localities and formations.**

<table>
<thead>
<tr>
<th>Sub-facies</th>
<th>Localities</th>
<th>Formation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Clast-supported conglomerates</td>
<td>Bucakkoyu</td>
<td>Tepekli</td>
</tr>
<tr>
<td></td>
<td>Altinkaya</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Ballibucak</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Kesme</td>
<td></td>
</tr>
<tr>
<td>Matrix-supported conglomerates</td>
<td>Kargi baraj</td>
<td>Taskesigî</td>
</tr>
<tr>
<td>Fossiliferous and cross-bedded sandstones</td>
<td>Deniztepesi</td>
<td>Tepekli</td>
</tr>
<tr>
<td></td>
<td>Saburlar</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Kesme</td>
<td></td>
</tr>
</tbody>
</table>
Figure 5.26 Model of the fan-delta shoreline facies preserved in the study area.
5.5.4.1 Clast-supported conglomerates

Description
These conglomerates constitute a thick succession (i.e. ~500-1000m) of laterally continuous conglomerates and sandstones. In two places (e.g. at Altinkaya and Bucakköyü; Fig. 5.1) they pass up into shallow-water carbonate successions consisting of small patches of reef framestones and calcarenites. At Altinkaya, they are interbedded with these carbonates on a scale of 10-50cm.

The interbedded conglomerates and sandstones of this facies are well bedded and parallel sided (Fig. 5.27). Clasts are generally rounded and they are distinctively well sorted. Clast fabrics such as imbrication are rare, but do occur. Matrix for the conglomerates is compositionally similar to that of the interbedded sandstones and is generally fine sand grade or coarser. The cement is the ubiquitous carbonate which characterises the cements of all the rocks under study, but in these rocks, perhaps because of the overwhelming predominance of carbonate, this results in “fairy chimney” weathering patterns of palaeokarstic origin (e.g. columnar structures up to 7m in height) between the Köprü Bridge and Ballibucak.

Interpretation
Both the association of these successions with shallow-marine carbonates and the absence of any evidence of subaerial deposition suggest that these conglomerates and sandstones were deposited subaqueously. There is no evidence of the bathymetry, but as the sequence passes up into shallow-marine carbonates, this suggests that they were deposited in an inner-shelf environment. The fairy chimneys are one of the many palaeo-karst features that proliferate in the Köprü basin and are making building the hydro-electric dam planned for the Beskonak area impossible (Degirmenci, 1992).
Figure 5.27 Photograph of the parallel bedded conglomerates at Bucakkoyu, west Kopru basin.

Figure 5.29 Cross-bedded sandstones and conglomerates north of Yesilbag, north-east Kopru basin (Fig. 5.1), Aksu Formation.
5.5.4.2 Matrix-supported conglomerates

Description

The most spectacular examples of these lithotypes occur in the Taskesigi Member (Karpuzçay Formation) of the Manavgat basin (Table 2.6) and have already been described in chapter 4 (section 4.5.3.1). At the Kargi baraj site (Aksu basin, Fig. 5.1) a similar matrix-supported conglomerate directly onlaps an undulating and now steeply dipping unconformity in Jurassic limestone. This horizon is both localised and varies greatly in thickness. Towards the road tunnel (at present under construction), a sequence of matrix supported conglomerates between 15 and 20m thick is visible. This decreases to a single flow <1m thick just to the north of the dam itself and passes laterally into oyster and coral-rich Lower-Middle Miocene limestones (section 3.6.3.4). Overlying these matrix-supported conglomerates is a series of cross-bedded, imbricated, clast-supported conglomerates (described above), which are localised above the matrix-supported conglomerates.

The Jurassic basement onto which this succession onlaps is bored and encrusted. Bored pebbles of this same limestone are found plentifully within the conglomerate itself along with abundant Porites corals and oysters. The sequence is made up of layers with varying amounts of grey silty matrix from a fully matrix-supported texture, to angular micro-conglomerates which are clast-supported. All are markedly poorly sorted.

Interpretation

The shallow-marine bioclastic debris, the bored and encrusted basement and the lateral facies change to shallow-water reef environments indicates a submarine environment of deposition. The present, nearly vertical angle of dip of the basement has, almost certainly, been exaggerated by later east-west tectonic shortening during the Aksu Phase at the end of the Miocene (chapter 7), but the interbedded angular clast-supported micro-conglomerates resemble talus deposits and suggest that the slope on this basement was also steep during deposition. The sequence as a whole, including the overlying cross-bedded conglomerates and lateral facies change to reef framestones is remarkably similar to that described by Barrier (1984) from the palaeofault scarp of the Capo
Figure 5.28 Outcrop sketch of the most proximal (back-edge) part of an underwater conical delta abutting against a paleofault scarp of the Capo dell'Armi horst, Messinian Strait, Italy. (Modified from Nemec, 1990; originally from Barrier, 1984). The facies associations shown here closely resemble those observed in the Kargi baraj area, Aksu basin.
dell'Armi horst, Messinian Strait, Italy, which he interpreted as the most proximal 'back-edge' of an underwater conical delta (Fig. 5.28). There is however no evidence of a fault along this margin of the Aksu basin.

5.5.4.3 Fossiliferous and cross-bedded sandstones

Description
Because of their high biogenic carbonate content, these sediments have been described in detail in chapter 3, and only a brief summary of the salient points is given here. Fossiliferous sandstones are common throughout the study area although they vary greatly from place to place. At Deniztepesi they are interbedded with medium to coarse conglomerates, and in situ coralline limestone. At Kesme by contrast gastropod-rich sandstones are interbedded with cross-bedded sandstones, conglomerates and Montastrea framestones. Well developed cross-bedded sandstones also occur in the lower part of the section at Saburlar, associated with oyster-rich beds and conglomerate horizons.

This sub-facies resembles the echinoid-scaphopod facies rocks in some instances although large algal rhodoliths are generally absent (e.g. Deniztepesi). Also the fossil assemblage tends to be much higher diversity including coral, bivalves, gastropods and oysters as well as abundant algal fragments and proportionally fewer scaphopods and echinoids. Almost all fossil specimens located in these sediments were fragmented. In some more rare cases however, the fossil content may be monospecific constituting a gastropod or oyster lag. These can be seen in the north of the Köprü basin near Yesilbag and Kasimlar and are described in more detail in chapter 3 (section 3.6.3.3) under gastropod grainstone/wackestones. Beds are generally 10-20cm in thickness, with cross-bedding being rather trough-shaped (Fig. 5.29). Rare water escape structure were also observed in this section.

Interpretation
These fossiliferous sandstones are thought to represent sediments which through gravity-driven processes (possibly triggered by storm induced flooding) swept up shells from shallow marine settings, depositing them further off-shore. These were later winnowed by currents to produce the rather concentrated biogenic lags now preserved. Similar sediments have
been documented by McCallum (1989) in the Athalassa Formation of southern Cyprus. It is thought that this sub-facies is of slightly deeper water origin than scaphopod-echinoid facies because they do not show any obvious signs of wave reworking. Neither would storm flooding of continental areas produce the later winnowing. The association of these sediments with cross-bedded sandstones could indicate constant working by unidirectional currents, sweeping the shelf.

5.5.4.4 Interpretation of sub-facies association D

A summary of the description and interpretation of the sub-facies discussed above is given in table 5.12. Water depths are not clearly indicated by the sub-facies in group D although they were certainly all deposited in a marine environment (Table 5.12). There is little evidence to suggest very deep water and the lateral and vertical facies transitions to shallow-water reef carbonates which can be observed for all three of the sub-facies described here, strongly indicates relatively shallow water, at least periodically. It is suggested therefore that these sub-facies were deposited in a shallow-marine shelf environment fed by rivers and reworked by wave and storm processes.

5.6 Provenance and Palaeocurrent data

Interpretation of the abundances of clast types in the conglomerates within the study area is difficult, mainly because so much of the hinterland is dominated by similar successions of Mesozoic carbonate, radiolarian chert and sandstones. A more detailed study involving detailed work in the hinterland itself might have produced a more informative interpretation of the 53 conglomerate analyses measured in Aksu conglomerate successions and displayed in Appendix 4a. This was beyond the scope of this project however. A few simple trends have been deduced from manipulation of conglomerate analysis data and these are discussed below.
<table>
<thead>
<tr>
<th>Name</th>
<th>Lower boundary</th>
<th>Upper boundary</th>
<th>Lateral variation</th>
<th>Thickness</th>
<th>Sorting/composition</th>
<th>Structures</th>
<th>Matrix</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Clast-supported conglomerates and sandstones</td>
<td>planar</td>
<td>planar</td>
<td>laterally persistent</td>
<td>0.3-1m</td>
<td>well sorted</td>
<td>rare imbrication</td>
<td>fine sandstone-microconglomerates</td>
<td>upper shell sands and conglomerates</td>
</tr>
<tr>
<td>matrix-supported conglomerates</td>
<td>unconformity</td>
<td>planar-slightly undulating</td>
<td>thickness changes from 30m to 0m over 1km. Changes to reef carbonates</td>
<td>0-30m</td>
<td>very poorly sorted, coral oysters, Mesozoic limestone clasts (bored)</td>
<td>chaotic</td>
<td>siltstone</td>
<td>debris flow</td>
</tr>
<tr>
<td>Fossiliferous and cross-bedded sandstones</td>
<td>planar-faintly scoured</td>
<td>planar</td>
<td>?</td>
<td>10-20cm</td>
<td>generally well sorted, bioclasts may be significantly larger</td>
<td>rare monospecific oyster/gastropod lags</td>
<td>coarse-fine sandstone</td>
<td>re-deposited, winnowed shelf sands</td>
</tr>
</tbody>
</table>
Figure 5.30 Provenance data from the basal conglomerates in the Manavgat basin (Appendix 4a). The location of conglomerates containing Alanya Massif clasts are all closely associated with present day exposure. Alanya Massif material is not present in the more northerly conglomerates indicating localised derivation. Igneous clasts derived from the Antalya Complex show a similar, but inverse relationship. Miocene limestone clasts do not appear to be geographically controlled suggesting that shallow-water carbonate was abundant throughout the area at this time.
5.6.1 The Manavgat basin

Figure 5.30 shows the location of conglomerates analysed in the Manavgat basin, excluding those that are part of the Middle and Upper Miocene sequence described in chapter 4. This diagram shows that reworking of basement was extremely localised. For instance, the abundance of igneous clasts in the southern part of the basin, where conglomerates overlie Alanya Massif is limited, whilst clasts of the Alanya massif limestone and schist are much less abundant in the north western part of the basin, where the conglomerates overlie Mesozoic limestone and Antalya Complex (primarily limestone, basalt and radiolarite).

In conjunction with the palaeocurrent data displayed in figure 5.31, the following interpretations can be made:

- Very little material from the Alanya Massif, Beysehir-Hoyran-Hadim Nappes and Anamas-Akseki Platform (Fig. 1.3) was transported over the Alanya Massif to the area near Alarahan and Saburlar in the south of the basin. The source for these conglomerates was probably east of the basin;
- Remarkably little Alanya Massif material was transported westwards towards Yaylaalan and Kizildag. These conglomerates were obviously sourced to some extent from the east, north of the Alanya Massif, but also from the north.

Put together, these conclusions indicate that the Alanya Massif formed a palaeo-high during the Lower Miocene, around which conglomerate-dominated successions were deposited. This is confirmed, by the absence of lower Miocene conglomerates overlying the Alanya Massif in the north of the basin (Fig. 4.29) and the direct colonisation of Burdigalian-Langhian reefs. The identification of Miocene algal limestone clasts throughout the Lower Miocene of the basin is somewhat problematic. Their presence indicates that shallow-water marine limestones were forming before those preserved (e.g. earlier in the Burdigalian). Almost all the conglomerates observed in the Manavgat basin below the Burdigalian-Langhian reefs (Oymapinar Limestone) have been interpreted as continental deposits, the exception being at Saburlar. The presence of Miocene marine clasts in these essentially continental
Figure 5.31  Palaeocurrent data from the Lower conglomerate succession in the Manavgat basin. The variety of directions is typical of fan-delta palaeocurrents. However, a few important trends can be discerned. First, the majority of the data from the northern margin of the basin indicates that it was sourced from the east and not directly from the north. In the south of the basin the palaeocurrents indicate that there was both a southerly and westerly component to the flow direction.
Burdigalian (or older) conglomerates indicates that there must have been at least one "transgression" over parts of this area of which these clasts are the only remaining remnant.

5.6.2 The Aksu and Köprü basins

Much less can be deduced from conglomerates analysed in the Aksu and Köprü basins. In both basins, Miocene algal limestone clasts are only found in conglomerates closely associated with preserved reef structures (Fig. 5.32), e.g. at Altinkaya, Dumanli and Bucakköyü in the Köprü basin and in the Bucak junction conglomerates in the Aksu basin. Figure 5.33 shows the distribution of radiolarian chert clasts in the Köprü basin. The abundance difference between the north east Köprü basin conglomerates and the rest of the Köprü basin conglomerates is mirrored in other clast types (e.g. grey limestone clasts, appendix 4a). This suggests that the north Köprü conglomerates were deposited by a different system from the west Köprü conglomerates and eroded a hinterland with different proportions of similar components e.g. radiolarian chert and grey limestone.

Figure 5.34 shows palaeocurrent data collected for the Aksu conglomerates in the Aksu and Köprü basins. In the west of the Köprü basin and including the sub-basins located on the Mesozoic carbonate "promontory" between the Köprü and Aksu basins (section 5.5.1.2), the broad flow direction seems to have been to the south. Variations in flow direction are not uncommon in alluvial fans, due to avulsion (e.g. Postma, 1990) and the palaeocurrents from the north east of the Köprü basin appear to have a bimodal direction distribution: broadly to the south and east-west. The location of the "breach" in the Kirkkavak fault, now filled with these conglomerates (Fig. 5.16) in the far north west of the Köprü basin and its associated southerly palaeocurrents (Fig. 5.34) suggests that part of the source of material was through this conduit although it is not at all clear what palaeogeographic role the Kirkkavak fault played at this time. East-west directed currents to the south west of this "breach" between Kasimlar and Kesme may indicate an interplay between material derived from the breach and material derived from the west. This seems particularly likely since the conglomerates in the Kasimlar-Kesme area are interbedded with shallow-water carbonates
Figure 5.32  Triangular diagram of typical conglomerates from the Köprü and Aksu basins. All the Miocene clasts come from conglomerates associated with in situ Miocene reefs.

Figure 5.33  Bar chart of radiolarian chert distribution through out the Köprü basin. This distribution is mirrored by other clast types and suggests that the north eastern fan-delta system was a separate system from that along the western margin.
Fig 5.34 Palaeocurrent data from the Aksu conglomerates in the Aksu and Köprü basins.
including reef framestones, suggesting close proximity to subaerial conditions and alluvial sedimentation.

The Aksu basin palaeocurrents were all measured in a small area around village of Kargi and the new dam site (Fig 5.1). Much of the outcrop from which these measurements were made is now submerged or only accessible by boat! The overwhelming indication is of north east directed flow, away from the passive basin margin to the west.

5.7 General interpretation of the sedimentary system.

Table 5.13 summarises the interpretation of individual sub-facies and the sub-facies associations as a whole. The close spatial and temporal association of alluvial, lacustrine, lagoonal, coastal and shelf facies within these conglomerate dominated successions suggests that they represent deposition and preservation of the products of a fan delta system. The Aksu and Kizildag Formations have been sub-divided into formation groups as a result of the sub-facies associations described above. The subdivision is shown in table 5.14.

The study of clastic sediments within the framework of the fan-delta concept is a relatively recent one. Defined as alluvial fans that prograde into a standing body of water from an adjacent highland by Holmes (1965) and McGowen (1970) the impetus for studying fan-deltas began primarily in 1975 when Walker first proposed four models for deep water conglomerates. Rust (1979) suggested that there was little significant difference between the internal structures of fan delta facies deposited in a marine environment and their terrestrial counterparts, alluvial fans, save for minor reworking by waves. It is now widely accepted however that fan-deltas are distinct depositional systems (Nemec and Steel, 1988) and that even single subaerial flows are considerably transformed when passing into water (Nemec and Steel, 1984). Coarse fan-deltas have become the focus of increasing interest and research over the last few years because they are sensitive recorders of tectonic, climatic and base level conditions as well as being excellent potential reservoirs for hydrocarbons and coal (Colella and Prior, 1990).
<table>
<thead>
<tr>
<th>Sub-facies association</th>
<th>Sub-facies association</th>
<th>Interpretation of sub-facies</th>
<th>Interpretation of sub-facies association</th>
</tr>
</thead>
<tbody>
<tr>
<td>Clast-supported angular conglomerate</td>
<td>fault talus</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Clast-supported conglomerates and coarse sandstones</td>
<td>sheet flood and channelised braided stream deposits.</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Matrix-supported conglomerates</td>
<td>debris flow</td>
<td></td>
<td>ALLUVIAL</td>
</tr>
<tr>
<td>Laminated sandstones</td>
<td>braided stream bars, over bank flows, sheet floods</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Calcretes and associated fine sediments</td>
<td>palaeosol</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Organic-rich facies</td>
<td>immature coal</td>
<td></td>
<td>LAGOONAL / LACUSTRINE</td>
</tr>
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<td>Green clays</td>
<td>lagoonal clay</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Laminated limestone and lime mud</td>
<td>lacustrine limestone</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Gastropod siltstones</td>
<td>lacustrine silts</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Limestone breccia</td>
<td>slope talus deposited in a lake</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Echinoid-scaphopod facies</td>
<td>wave-reecked beach gravel.</td>
<td></td>
<td>SHORELINE</td>
</tr>
<tr>
<td>Porites bafflestones</td>
<td>fluvial sediment trap at the marine-fresh-water interface</td>
<td></td>
<td></td>
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<tr>
<td>High-angle cross-beded conglomerates</td>
<td>mouth-bar-type foresets</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Clast-supported conglomerates</td>
<td>upper shelf sands and conglomerates</td>
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<td>SHELFL</td>
</tr>
<tr>
<td>Matrix-supported conglomerates</td>
<td>debris flow</td>
<td></td>
<td></td>
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<tr>
<td>Fossiliferous and cross-beded sandstones</td>
<td>redeposited, winnowed shelf sandstones</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Table 5.14  Sub-division of the Aksu and Kizildag Formations (Akay and Uysal, 1985; Akay et al., 1985) into formation Members.

<table>
<thead>
<tr>
<th>Environment</th>
<th>Sub-environment</th>
<th>Aksu Formation</th>
<th>Kizildag Formation</th>
</tr>
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<tbody>
<tr>
<td>Continental</td>
<td>Alluvial</td>
<td></td>
<td>Kargi</td>
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<tr>
<td></td>
<td>Lagoonal and</td>
<td>Kapikaya</td>
<td>Calcaire de Kepez</td>
</tr>
<tr>
<td></td>
<td>Lacustrine</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Marine</td>
<td>Shoreline</td>
<td>Kesme</td>
<td>Tepekli</td>
</tr>
<tr>
<td></td>
<td>Shelf</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

5.8 Classification of the fan delta systems

The extensive development of shelf facies indicates that, contrary to the original model of fan-deltas as fans passing directly into deep water (e.g. in fjords), the subaqueous parts of the fan-deltas studied were developed in relatively shallow water (Etheridge and Westcott, 1984; Elliott, 1986; Fraser and Suttner, 1986; Postma, 1990). They can therefore be classified as shallow-water deltas (Postma, 1990), which includes both marine and lacustrine environments, unlike the exclusively marine classification of shelf-type deltas (Etheridge and Westcott, 1984). Corner's classification (in Nemec, 1990) "steep general slope fan-deltas" suggests Gilbert-type foresets on a scale an order of magnitude larger than those seen here. The difficulty in classifying these fans may be the result of the restricted geographical space (i.e. the narrow shape of the basin) preventing full morphological development.

Figure 5.35 shows the geographic areas covered by the different Formation Members. From the palaeocurrent (Figs. 5.31 and 5.34) and provenance data discussed above 7 separate fan-delta systems can be tentatively identified. Not all of these were active at the same time and no single fan-delta system persisted throughout the Miocene.

5.9 Controls on fan-delta development

Relative sea level changes are hugely more important in the development of the shelf facies of a fan-delta
Figure 5.35 Geographic areas covered by the different formation members in the Aksu Conglomerate Formation.
level change and basinal controls such as wave reworking produces the wide range of environmentally informative deposits described above.

A modified version of the conceptual framework for delta systems suggested by Elliott (1986) is presented in Figure 5.36a. The arrows indicate the direction of thought required for the study of modern deltas. In studying ancient deltas this process is reversed in order to deduce information on the climate, tectonic activity and sea level change. Figure 5.36b shows this conceptual framework with the information gained from study of the facies inserted. These three extra-basinal controls are discussed below.

5.9.1 Climate

The following information from the sub-facies discussed above has palaeoclimatic implications:

- The abundance of calcrete in finer-grained alluvial sediments;
- The paucity of preserved plant debris except in exceptional cases (e.g. coal at Bozburun Dag);
- Some of the conglomerate and sandstone geometries and structures (e.g. channel-mouth-bars) have been interpreted as indicating flashy, ephemeral flow (section 5.5.1.2);
- The alluvial facies described here resembles modern sequences from the Red Sea area (Hayward, 1982);
- Ancient successions of alluvial facies are also similar to those studied here (e.g. Quaternary of Cyprus, Poole, 1992).

A semi-arid climate during the Miocene has been interpreted, not dissimilar from the current climatic regime affecting south coastal Turkey. This is in agreement with Maizels and McBean (1990) who suggested that a humid environment is unlikely to have generated many of the features listed above (section 5.5.1.6).
Figure 5.36  a) The conceptual framework for the comparative study of both modern and ancient delta systems (modified from Postma, 1990; originally from Elliot, 1986).  b) The reversed framework showing the evidence gained from facies analysis of the fan-delta deposits in the study area.
5.9.2 Tectonics

The most two most important pieces of evidence of tectonic activity displayed by the Aksu Formation Conglomerates are firstly, the thickness of the preserved succession, and secondly, the consistently very coarse grainsize.

Wescott and Etheridge (1980) suggested that the thickness of fan-delta deposits depends upon a complex interaction of mountain-front uplift, sediment supply and basin subsidence. It is suggested here that relative sea level change can also play a vital role. Degirmenci (1992) estimated a minimum thickness of 440m for the west Köprü basin conglomerates from borehole data which did not hit basement in the centre of the Köprü basin. The altitude of the basin rises sharply towards the west however, being roughly 1000m higher near Altinkaya (Fig. 5.1) than it was in the centre of the basin close to the contact with the Karpuzçay turbidites. The tortuous track that climbs up to this village with its isolated Lycian remains is lined from bottom to top with continuous, horizontally bedded continental and then marine conglomerates (Fig. 5.27). A slightly more accurate estimate of the minimum thickness of this conglomerate succession is therefore in the order of 1.5km.

All the successions described above, whatever the environment of deposition, are dominated by cobble conglomerates rather than micro-conglomerate or sandstone. Such continual coarse conglomerate sedimentation suggests the sustaining of a high gradient of deposition. There are three potential processes which could maintain high gradients:

- uplift of the basin margins;
- subsidence of the basin itself;
- relative sea-level fall.

Given that these three processes are all interdependent, the discussion below attempts to isolate the relative activity of each.

There are two main pieces of evidence for relative sea level rise during the deposition of both the Lower and Upper Miocene conglomerate
successions. On a local scale, sea level can be demonstrated to be rising during clastic accumulation by the *Porites* bafflestone sub-facies where coral sticks pass undisturbed through clastic boundaries (section 5.5.3.2; Fig. 5.21). On a more regional scale, where transitions from one environment of deposition to another are visible, all the successions go from being subaerially deposited to being deposited in a marine environment (Fig. 5.35). This suggests that overall, relative sea level rise was more dominant than the prograding of fan-delta sequences. Bearing in mind the constantly coarse material deposited in these thick sequences (e.g. >1500m), this indicates that relative sea level rise was very rapid.

A conservative estimate for the minimum sedimentation rate for the Lower Miocene conglomerates in the west of the Köprü basin is 20cm/1000yrs. Bearing in mind that many of the cobbles in the conglomerates are roughly 20cm in diameter this suggests that 20cm/1000yrs may be heavily underestimated, possibly by an order of magnitude. Discounting subsidence, this minimum sedimentation rate computes to 200m/Ma sea level rise, which is already twice that of the quickest eustatic sea level change during the Miocene (Haq et al., 1988; Fig. 5.38). Given that this estimate may also be an order of magnitude too low, and that relative sea level rise eventually outpaces delta progradation, relative sea level rise must be accepted as a major control in the development of the fan-deltas studied.

How much of this relative sea level rise can be accounted for by subsidence (tectonic or compaction driven) and how much by eustatic processes? The thickness of the sequences once again holds a key to understanding this problem. It seems highly unlikely that a coarse conglomerate succession over 1500m thick could be filling accommodation space entirely generated by pre-existing palaeotopography. It is therefore safe to assume that some of this accommodation space was generated by subsidence of the basin relative to the margins during deposition. Clearly some of this subsidence is likely to have been driven by compaction of those sediments and the ones beneath as a result of the deposition of 1500m of conglomerate. Note however that the grainsize remains coarse throughout the succession. Thus, this is no generally subsiding area which passes from subaerial to
submarine environment becoming more distal to the source in the process. This is an area where the hinterland is continuing to rise, constantly generating topographic highs susceptible to erosion, over and above a relatively rising sea level.

5.9.3 *Eustacy*

From the above discussion it can be seen therefore that some of the relative sea-level rise can be attributed to subsidence due to compaction and it seems likely that if there is evidence of a rising hinterland, the basins themselves may also have been subjected to active tectonic subsidence. The component of sea level rise contributed by eustacy is difficult to assess in detail. The curve produced by Haq *et al.* (1988; Fig. 5.37) indicates several periods of sharp sea level rise in the order of 100m. These periods may account for the formation of those sequences where sea level rise is so clearly documented (e.g. Porites bafflestones, section 5.5.3.2). The gentle eustatic sea level rise suggested by Haq's curve for the Burdigalian may have contributed to the relative sea level rise deduced for the Lower Miocene Aksu conglomerates, but it cannot have been the only component. The same is true for the Upper Miocene conglomerates, though the gradient on the eustatic curve for the Tortonian is somewhat steeper here. The sudden drowning of the reef carbonates and conglomerates at the Burdigalian-Langhian boundary (Fig. 4.32) across the Isparta Angle may have been caused in part by a eustatic sea level rise. A sea level high stand has been documented elsewhere in the Mediterranean at this time (e.g. south-east Turkey, Gürbüz, 1993; Sardinia, Cocozza and Jacobacci, 1975; southern Cyprus, Robertson *et al.*, 1991).

In terms of the larger scale stratigraphy, nannoplankton dating across the basin by C. Müller (pers. comm., 1992 and 1994) suggests that deposition of the Karpuzçay Formation turbidites began in the Serravallian (Fig. 4.24). There is evidence of interbedding of the Lower Miocene Aksu Formation conglomerates and the Karpuzçay Formation in the centre of the Köprü basin where boreholes have been drilled (Degirmenci, 1992). In general however the Serravallian is conglomerate poor, the exception being the cone of coarse conglomerates in the centre of the Aksu basin reported by
TIME SERIES STAGE EUSTATIC CURVES

<table>
<thead>
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<td>6.3</td>
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</tbody>
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Figure 5.37  Eustatic sea level curve for the Miocene (Haq et al., 1988).

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Poisson (1977a). Relative to the conglomerates the turbidites are very fine-grained. The most obvious cause of this sudden dearth of conglomerate would seem to be either a cessation of tectonic uplift or a very rapid sea level rise that cuts off coarse clastic supply and allows more distal systems to feed the basin. Given that the palaeocurrents in both the Aksu and Köprü basins indicate axially derived turbidites (Figs. 4.10 and 4.16) and marginally-derived Lower Miocene conglomerates (Fig. 5.34), rapid relative sea level rise producing a new, more distal sediment source seems the more likely of the two options. The eustatic sea level curve (Haq et al. 1988) indicates a gradual sea level drop throughout the Serravallian.

5.10 Conclusions

♦ The Kizildag and Aksu Formations can be interpreted mainly as deposits of fan-delta systems.

♦ Alluvial, lacustrine, lagoonal, coastal and shallow-marine shelf environments have been identified.

♦ 7 separate fan-delta systems can be tentatively identified as having been active during two distinct time periods, the Lower-Middle Miocene (Burdigalian-Early Serravallian) and Upper Miocene (Tortonian-Lower Messinian).

♦ These are all shallow-shelf type fan-deltas either with well developed mouth-bar units or with small Gilbert-type foresets.

♦ The nature of the facies preserved suggest that these fan-deltas were fluvially dominated. There is clear evidence of wave action however, suggesting that this process was also important in reworking shoreline facies.

♦ The development of reef frameworks may have influenced clastic deposition. High angle foresets in coarse conglomerates are associated with reef buildups and this topography may have produced gradient changes which initiated foreset formation.
A semi-arid climate subject to flash floods and ephemeral flow can be deduced from the alluvial facies.

There is clear evidence of rapid relative sea level rise throughout both the Lower Miocene and Upper Miocene in the shoreline facies.

Relative sea level rise seems to have been the most important control on fan-delta evolution, overprinting the prograding fan-delta signature despite the rapid sedimentation rate.

Some of the relative sea level rise can probably be attributed to compaction, but the majority is thought to be tectonically induced.

Active tectonic uplift of the hinterland caused continual coarse sedimentation.

Eustatic processes, though potentially active are not thought to have had complete control over any of the changes in sedimentary facies seen throughout the Miocene.
6.1 Context

Lack of good marker fossils in neritic sediments is a common dating problem which can be addressed by the use of isotopic dating methods. Consequently, \(^{87}\text{Sr}/^{86}\text{Sr}\) analysis was attempted with a view to dating the deposition of the shallow water carbonates, particularly those at the base of the Miocene section in the study area (Oymapinar Limestone). Dating of these sediments permits correlation of the neritic sediments, discussion of relative sea level change and ultimately a more detailed interpretation of basin history. This study contributes to the biostratigraphic database on Miocene rocks in providing ages for previously undated key sections.

6.2 Introduction to Strontium dating

Because the residence time of Sr in the oceans is long (2.5-5Ma) relative to ocean mixing time (500-1000 yrs, e.g. Beets, 1992; Richter and Turekian, 1993; McArthur, 1994) both the Sr concentration and the Sr isotopic ratio can be assumed to be constant throughout the oceans at any given time. Sr-isotope ratios have been generally increasing with a step-like pattern since the Jurassic (Hodell, 1994) and studies over the last decade have concerned the accurate documentation of these variations particularly in the Cenozoic (DePaolo, 1986; DePaolo and Ingram, 1985; Hess et al., 1986; and Koepnick et al., 1985; Koepnick et al., 1988). More recently workers have increasingly focused on narrower segments of the curve using independent dating methods such as biostratigraphy and magnetostratigraphy, so that the technique can be used for solving stratigraphic problems.

Originally the \(^{87}\text{Sr}/^{86}\text{Sr}\) of the oceans was predicted as increasing monotonically through time due to the decay of \(^{87}\text{Rb}\) to \(^{87}\text{Sr}\) (Wickman, 1948). Early workers attempting to document this change (Peterman et al., 1970; Veizer and Compston, 1974; Burke et al., 1982) quickly realised that Sr
evolution was not monotonic, but developed irregularly with both positive and negative gradients. Causes for the shape of the oceanic Sr ratio curve have subsequently been studied in considerable detail and the following hypotheses have been put forward:

- $^{87}\text{Sr}$ supplied to the oceans by weathering of continental crust via rivers (Brass, 1976);
- $^{87}\text{Sr}$ subtracted from the ocean budget due to hydrothermal exchange of $^{87}\text{Sr}$ depleted basalts at mid-ocean ridges (Brass, 1976);
- Dissolution of carbonates on the sea floor adds old marine Sr whose $^{87}\text{Sr}/^{86}\text{Sr}$ is not very different from that of sea water and generally acts as a buffer (Elderfield and Geiskes, 1982; Palmer and Edmond, 1989; Richter and Liang, 1993).

### 6.3 Application of $^{87}\text{Sr}/^{86}\text{Sr}$ dating technique to the Miocene of the study area and the correction for radiogenic Sr

The main sink for Sr in the oceans is biogenic carbonate (Brass, 1976; Palmer and Elderfield, 1985). As organisms precipitating carbonate shells or tests do not appear to discriminate between the various isotopes of Sr (e.g. 88, 87, 86, and 84) the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio can be assumed to represent the ratio in sea water at the time of formation. Radiogenic $^{87}\text{Sr}$ derived from the decay of $^{87}\text{Rb}$ can be largely ignored in the study of biogenic carbonates as they generally contain relatively low concentrations of Rb relative to Sr providing diagenetic alteration has not occurred (Elderfield, 1986). For example, uncleaned planktic foraminifera contain $6\pm 2$ ppm of Rb, whilst their Sr concentration is in the order of $1150 \pm 50$ ppm (Beets, 1992) making the Rb/Sr < 0.01.

The sediments discussed in this chapter have been described in chapter 3. They include the shallow-water reef limestones and deeper water planktic foraminiferal marls which overlie them.
6.4 Methodology

6.4.1 Sample selection

The use of bulk samples for Sr analysis comprehends the risk of including unknown components within the sample such as clay and silt-sized detrital calcite (Beets, 1992) and, because bulk carbonates are more susceptible to diagenesis (Richter and DePaolo, 1988; Garrison, 1981) it also increases the risk of using altered samples. For this reason a variety of micro- and macro-fauna rather than bulk sediment were selected for $^{87}\text{Sr}/^{86}\text{Sr}$ analysis and these are listed in table 6.1. As not all these fauna construct their shells/tests of carbonate in its most stable form (e.g. low Mg calcite, see Table 6.1) it was essential to carry out careful screening of selected samples for evidence of diagenetic alteration from the less stable forms of carbonate (e.g. aragonite and high Mg-calcite). Even those samples composed of low Mg-calcite fauna were inspected, for visible cementation either on or in the structure of the shell/test.

Screening was undertaken in a number of ways depending on sample preservation. For instance, foraminifera were hand picked from washed and sieved (63µ) marl samples and inspected under a reflected light microscope. Those specimens showing signs of considerable abrasion or whose tests were infilled with a dark mineral thought to be pyrite, or encrusted with specks of a grey mineral thought to be manganese oxide were discarded. Some foraminiferal samples were also inspected using SEM photography. This indicated that the preservation of both benthic and planktic foraminifera was in general quite good (Fig. 6.1), although rare cases of calcite rhombs infilling chambers were observed, (Fig. 6.2). Hodell and Woodruff (1994) point out however that even though conspicuous evidence of diagenetic alteration is lacking, it is often difficult to demonstrate unambiguously that foraminiferal calcite has been completely unaffected by diagenesis.

This problem was clearly illustrated during the cleaning process. Most authors using foraminifera for Sr analysis do not indicate what method of disaggregation of the marl sample was used prior to hand-picking of the foraminifera. After picking the foraminifera however, Miller et al. (1991)
Figure 6.1  SEM photograph of a benthic foraminifera (Florilous bonneam, C. Glover pers. com. 1995) showing excellent preservation of skeletal structure from the Ahmetler section, Manavgat basin.

Figure 6.2  SEM photograph of a foraminifera which has been filled with diagenetic calcite rhombs.
Table 6.1  Table showing the type of carbonates found in samples used for $^{87}\text{Sr}/^{86}\text{Sr}$ analysis and the sediment in which they were preserved.

<table>
<thead>
<tr>
<th>Fauna</th>
<th>original carbonate</th>
<th>Sediment</th>
<th>Preservation of structure</th>
<th>carbonate at time of analysis</th>
</tr>
</thead>
<tbody>
<tr>
<td>Planktic foraminifera</td>
<td>Low Mg calcite</td>
<td>in marls</td>
<td>good</td>
<td>Low Mg calcite</td>
</tr>
<tr>
<td>Benthic foraminifera</td>
<td>Low Mg calcite</td>
<td>in marls</td>
<td>good</td>
<td>Low Mg calcite</td>
</tr>
<tr>
<td>Oyster</td>
<td>Low Mg calcite</td>
<td>in marls and siltstones</td>
<td>generally good</td>
<td>Low Mg calcite</td>
</tr>
<tr>
<td>Echinoid</td>
<td>Low Mg calcite</td>
<td>in calcarenites</td>
<td>good</td>
<td>Low Mg calcite</td>
</tr>
<tr>
<td>Algae</td>
<td>High Mg calcite</td>
<td>in calcarenites</td>
<td>good</td>
<td>? high/low Mg calcite</td>
</tr>
<tr>
<td>Coral</td>
<td>Aragonite</td>
<td>in marl,</td>
<td>good</td>
<td>Aragonite and low Mg calcite</td>
</tr>
<tr>
<td></td>
<td></td>
<td>in sandstones or in reef framework interbedded with sandstones</td>
<td>good</td>
<td>Low Mg calcite</td>
</tr>
<tr>
<td></td>
<td>Gastropod</td>
<td>in marl</td>
<td>good</td>
<td>Aragonite</td>
</tr>
</tbody>
</table>

report cleaning their foraminifera using an ultrasound bath for 2-10s, whilst Oslick et al. (1994) used approximately 3s of ultrasound treatment. Müller (1990) crushed planktic foraminifera prior to washing them several times in an ultrasonic bath. Hodell et al. (1991) in contrast do not report cleaning foraminifera using ultrasound at all. None of these authors report the amplitude and frequency of the vibrations of the ultrasound used. The rapidity with which present-day foraminifera explode when placed in an ultrasound bath depends on the frequency and amplitude of the vibrations used. In general however, they explode within a few seconds (K. Darling pers. com., 1994).

Due to lithification, marl samples containing abundant foraminifera collected for this study were initially disaggregated using an ultrasound bath (FS400b with a frequency of 35-40Khz) for 2 x 3 minutes. Hand-picked foraminifera from these samples had therefore already received ultrasound...
treatment in excess of that which destroys recent foraminifera. It proved possible to expose these foraminifera to further ultrasound treatment without causing widespread disaggregation of individuals and this poses the question of why these foraminifera behaved differently from recent samples. The obvious explanation is that diagenetic alteration has filled voids in the structure and transformed the test itself, thus rendering the specimen more resistant to destruction by ultrasound vibrations. As pointed out earlier however it has proved difficult to unequivocally identify diagenetic alteration in foraminifera using light microscope or SEM techniques (Hodell and Woodruff, 1991). By not testing the resistance of their foraminiferal samples to ultrasound, other authors (e.g. Hodell and Woodruff, 1994; Miller et al., 1991; Beets, 1992; Oslick et al., 1994) have perhaps not explored this problem or used its implications as a method of identifying diagenetic alteration in foraminifera.

Carpenter et al. (1991) suggest that early marine diagenesis does not significantly alter the $^{87}\text{Sr}/^{86}\text{Sr}$, because the cements formed in this environment are in isotopic equilibrium with ambient sea water. Making the assumption that the diagenetic alteration of the foraminifera took place soon after formation and in the absence of guidance from previous workers it was decided that foraminiferal samples resistant to ultrasound treatment would be used and this issue is discussed further in the light of the results obtained (section 6.5.1.2).

Most workers using foraminifera for Sr analysis sampled planktic foraminifera (e.g. Palmer and Elderfield, 1985; Hess et al., 1986; Hodell et al., 1991; Miller and Feigenson, 1991). However Beets (1992) showed that there is no significant difference in the Sr-isotope ratios of benthic and planktic foraminifera. For this reason samples of mixed benthic and planktic foraminifera were used as well as samples containing only planktic foraminifera and samples of the large benthic foraminifera Operculina. Table 6.2 lists the samples selected, the name of their section and the type of fauna chosen.

Coral samples were inspected in a number of different ways. The majority of the corals initially selected as potential samples for Sr analysis were Porites $sp.$ reflecting the relative abundance of this type of coral. Poritids have small
corallites making species distinctions difficult in all samples. In the samples recovered from the study area it was found to be almost impossible to classify *Porites sp.* corals to species level because almost without exception thin sections revealed that the finely ornamented structure on which distinctions are made, were replaced with secondary calcite. A few of these corals were selected for Sr analysis despite the knowledge that they were diagenetically altered as it was hoped that information concerning the timing of diagenesis might be gained. Where other types of coral were selected and identified, thin section study sometimes revealed that only partial diagenetic alteration appeared to have taken place. Some of these samples were subsequently analysed by x-ray diffraction methods (XRD) to ascertain what minerals were present and by comparing the size of the peaks to standards of known aragonite:calcite composition an estimate of the ratio of the two minerals was ascertained. If the quantity of sample was sufficient the calcite and aragonite were then separated using heavy liquids (Appendix 1). Two coral samples were inspected using an energy dispersive spectroscopic microprobe (LINK system). The mean Sr values for these two samples were 0.7% and 0.65% values far closer to the percentages found in living Scleractinid aragonite (approximately 0.8%) than they are for diagenetic low magnesium calcite (0.02%), although this latter figure does depend to some extent on the Sr concentration of the fluids involved. SEM photographs taken of these same samples reveal good preservation of skeletal fibres (Fig. 3.11). There has been some preferential dissolution along organic-rich growth lines, but there is no evidence of general calcitization (C. Cuif pers. com., 1994).

Screening the algae for diagenetic alteration proved to be very difficult. In thin section evidence of widespread replacement by spar was absent and the preservation of the detailed structure of the algae was generally good. Samples were subsequently drilled from algal crusts as in Beets (1992), but no attempt was made to determine whether diagenetic alteration had taken place to low Mg calcite. Rare gastropod samples that were selected for Sr analysis were first analysed using XRD to determine if aragonite was still present. In hand specimen samples retaining primary carbonate were relatively easy to identify since they had a very fragile structure and were faintly pink in colour. This sort of preservation was exclusively associated with fine-grained, clay-rich marl rather than calcarenites or sandstones.
<table>
<thead>
<tr>
<th>Section name</th>
<th>Basin/area</th>
<th>Sample no</th>
<th>Fauna</th>
</tr>
</thead>
<tbody>
<tr>
<td>Alarahan</td>
<td>S Manavgat</td>
<td>4j. Alarahan.1</td>
<td>Porites sp.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>4j. Alarahan.2</td>
<td>Porites sp.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>4j. Alarahan.3</td>
<td>Porites sp.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>4j. Alarahan.4</td>
<td>Porites sp.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>10j. Alarahan.5</td>
<td>Operculina</td>
</tr>
<tr>
<td></td>
<td></td>
<td>80. Alarahan.1</td>
<td>planktic foraminifera</td>
</tr>
<tr>
<td></td>
<td></td>
<td>80. Alarahan.2</td>
<td>planktic foraminifera</td>
</tr>
<tr>
<td></td>
<td></td>
<td>80. Alarahan.3</td>
<td>planktic foraminifera</td>
</tr>
<tr>
<td>Ahmetler</td>
<td>N Manavgat</td>
<td>12j. Ahmetler.1</td>
<td>algae</td>
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<td></td>
<td></td>
<td>16S. 313.6</td>
<td>algae</td>
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<tr>
<td></td>
<td></td>
<td>12j. Ahmetler.2</td>
<td>algae</td>
</tr>
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<td></td>
<td></td>
<td>16S. 313.7</td>
<td>algae</td>
</tr>
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<td>16S. 341.6</td>
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<td>Yaylaalan</td>
<td>N Manavgat</td>
<td>14j. 290.2</td>
<td>echinoid</td>
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<td>14j. 290.5</td>
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<td>mixed foraminifera</td>
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<td>14j. 290.7</td>
<td>mixed foraminifera</td>
</tr>
<tr>
<td>Oymapinar</td>
<td>N Manavgat</td>
<td>13j. Oymapinar.4</td>
<td>echinoid</td>
</tr>
<tr>
<td></td>
<td></td>
<td>13j. Oymapinar.7</td>
<td>mixed foraminifera</td>
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<td>13j. Oymapinar.8</td>
<td>mixed foraminifera</td>
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<tr>
<td>Saburlar</td>
<td>central Manavgat</td>
<td>11j. 295-6.1</td>
<td>Operculina</td>
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<td></td>
<td>11j. 295-6.2</td>
<td>algae</td>
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<td>11j. 295-6.3</td>
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<tr>
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<td>25S. 373.6</td>
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<td></td>
<td>25S. 373.3</td>
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<td></td>
<td></td>
<td>16j. 373.2</td>
<td>mixed foraminifera</td>
</tr>
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<td></td>
<td>16j. 373.1</td>
<td>mixed foraminifera</td>
</tr>
<tr>
<td>Bucakkoyu</td>
<td>south Köprü</td>
<td>17j. 416.6</td>
<td>mixed foraminifera</td>
</tr>
<tr>
<td></td>
<td></td>
<td>18j. 417.4</td>
<td>algae</td>
</tr>
<tr>
<td></td>
<td></td>
<td>18j. 417.2</td>
<td>algae</td>
</tr>
<tr>
<td></td>
<td></td>
<td>18j. 417.5</td>
<td>mixed foraminifera</td>
</tr>
<tr>
<td></td>
<td></td>
<td>18j. 417.6</td>
<td>mixed foraminifera</td>
</tr>
<tr>
<td>Aspendos</td>
<td>south Köprü</td>
<td>17j. 415.1</td>
<td>echinoid</td>
</tr>
<tr>
<td></td>
<td></td>
<td>17j. 415.2</td>
<td>gastropod</td>
</tr>
<tr>
<td></td>
<td></td>
<td>17j. 415.3-a-c</td>
<td>gastropod</td>
</tr>
<tr>
<td></td>
<td></td>
<td>17j. 415.4</td>
<td>mixed foraminifera</td>
</tr>
<tr>
<td>Altinkaya</td>
<td>Central Köprü</td>
<td>19j. 181.2</td>
<td>oyster</td>
</tr>
<tr>
<td></td>
<td></td>
<td>&quot;19j. 181,1a&quot;</td>
<td>oyster</td>
</tr>
<tr>
<td></td>
<td></td>
<td>19j. 180.2</td>
<td>algae</td>
</tr>
<tr>
<td>Ballibucak</td>
<td>Central Köprü</td>
<td>16j. 163.3</td>
<td>algae</td>
</tr>
<tr>
<td></td>
<td></td>
<td>16j. 164.2</td>
<td>oyster</td>
</tr>
<tr>
<td>Kesme</td>
<td>NE Köprü</td>
<td>26j. 399.1b</td>
<td>oyster</td>
</tr>
<tr>
<td>Yesilbag</td>
<td>NE Köprü</td>
<td>27j. Yesil.3</td>
<td>algae</td>
</tr>
<tr>
<td></td>
<td></td>
<td>27j. Yesil.9</td>
<td>oyster</td>
</tr>
<tr>
<td>Dumanli</td>
<td>NE Köprü</td>
<td>22m. 407.1</td>
<td>mixed foraminifera</td>
</tr>
<tr>
<td>Çetmi</td>
<td>NE Köprü</td>
<td>22m. 404.3</td>
<td>oyster</td>
</tr>
<tr>
<td></td>
<td></td>
<td>20j. 421.1</td>
<td>algae</td>
</tr>
<tr>
<td>Kargi baraj</td>
<td>Central Aksu</td>
<td>15M. 1</td>
<td>oyster</td>
</tr>
<tr>
<td></td>
<td></td>
<td>15M. 3</td>
<td>oyster</td>
</tr>
<tr>
<td></td>
<td></td>
<td>15M. 5</td>
<td>oyster</td>
</tr>
<tr>
<td>Sipahiler</td>
<td>NW Köprü</td>
<td>30j. 236.8</td>
<td>algae</td>
</tr>
</tbody>
</table>

252
Oysters from the Kargi baraj section were studied in thin section before being drilled from remaining sample blocks. Good shell structure was apparent in all thin sections and Jones et al (1994) report that they successfully used oysters from the Jurassic to document Sr ratios.

Two samples drilled from well preserved echinoid tests were included amongst those selected for Sr analysis. Although echinoid shells are made of low Mg calcite, their structure is a porous one and easily altered by diagenesis (S. Hesselbo pers. com., 1994).

6.4.2 Cleaning methods

All samples were cleaned in pure ethanol in an ultrasound bath for up to 40 minutes. Foraminiferal samples were inspected regularly and removed from the ultrasound bath once disaggregation of individual foraminifera began. The foraminifera were rinsed with clean ethanol and removed using a pipette and dried in the oven. Other samples were rinsed in clean ethanol and dried on filter paper. Samples were then ground down to powder using a pestle and mortar and weighed to an accuracy of 5 decimal places.

6.4.3 Chemistry

Weighed samples were placed in teflon beakers and 10ml of 2.5M HCL was added to dissolve the sample. After several hours the sample was inspected to ensure that complete dissolution had occurred and centrifuged for about 5 minutes to remove any particulate matter remaining undissolved. The sample was then loaded into a pre-conditioned cation exchange column containing Bio-Rad AG50W-X8, 200-400 mesh resin. 2 x 1ml of 2.5M HCL was used to rinse the sample down into the column, and it was then eluted with 42ml of 2.5M HCL. The Sr was subsequently collected using 10ml of 2.5M HCL in a cleaned teflon beaker and evaporated under a lamp for several hours. When dry, a single drop of concentrated nitric acid was added and left to evaporate. This converts the Sr-chlorides to Sr-nitrates which give better ion yields. A total procedure blank run through this chemistry was found to contain 0.6 ng of Sr. Relative to the Sr content in the samples studied, this is negligible.
6.4.4 Mass Spectrometry

Re and Ta single filaments were constructed using filament supports that had been boiled in RO water and outgassed for 10 minutes at 4.5A in a vacuum <1x10^{-5} torr. Due to the small size of some the samples two methods of loading were used. The larger samples were dissolved in 1 µl of 1M phosphoric acid (H₃PO₄) and loaded on to a Ta filament. Small samples were loaded in the same way onto Re filaments which had already been loaded with 1 µl or less of H₃PO₄ and 1 µl of a tantalum activator which is prepared by dissolving Ta₂O₅ powder in dilute HF. This method is similar to that of Brick (1986). The tantalum activator enhances the ionisation potential of Sr allowing the analysis of smaller samples to a comparable precision to the H₃PO₄ method. All loaded filaments were dried gradually by passing a small current (1-1.5A approx.) through them. This current was then increased until white fumes of phosphoric acid could be seen and the filament glowed dull red. Loaded filaments were then fixed into a 20 sample turret head.

A VG Sector 54-30 mass spectrometer in dynamic multicollection mode was used to measure the Sr isotope abundances. A beam intensity for mass 88 of 1V (±10%) and a filament current of around 2.5A were selected. The measuring cycle used was 84.5, 86, 87, 88 axial masses with the following collector spacings from axial: L3-4 amu; L2-2 amu; L1-1 amu; H1 + 1 amu and H2 + 2 amu. Peak intensities were corrected for zero and Rb interference where necessary. The \(^{87}\text{Sr}/^{86}\text{Sr}\) ratios were corrected for instrumental mass fractionation to a \(^{86}\text{Sr}/^{88}\text{Sr}\) = 0.1194 using an exponential fractionation law. Isotope ratios were collected in 15 blocks of ten cycles and the mean and standard error for each block and the whole population were calculated. Up to 10% of the values falling outside 2σ were rejected and post-rejection means and standard errors were calculated. If necessary the beam was refocused and the aiming intensity restored between blocks. Typical in-run precision of ± 0.00012% is achieved for 150 ratios. Repeat analyses of the NBS 987 standard during and prior to this study give an \(^{87}\text{Sr}/^{86}\text{Sr}\) = 0.710237± 26 (2S.D.).
6.4.5 Calculations

Two recently published Miocene $^{87}\text{Sr}/^{86}\text{Sr}$ sea water curves (Hodell and Woodruff, 1994; Miller et al., 1991) were used to provide the geochronological reference. Although other curves have been published recently, these two were selected because they used different methods to link data points.

Miller et al. (1991) interpreted the spread of Sr ratios for the Miocene on a plot of $^{87}\text{Sr}/^{86}\text{Sr}$ versus age and deduced a linear evolution with a change in gradient at 14.5-14.7 Ma (Fig. 6.3). These two relationships are described as:

For 25.1-14.7 Ma

\[
\text{Age (Ma)} = 11906.91 - [16777.17 \times (^{87}\text{Sr}/^{86}\text{Sr})]
\]

with a correlation coefficient, $r^2 = 0.984$, n = 38

For 14.5-8.3 Ma

\[
\text{Age (Ma)} = 56239.94 - [79319.75 \times (^{87}\text{Sr}/^{86}\text{Sr})]
\]

with a correlation coefficient, $r^2 = 0.603$, n = 30

An NIST 987 Strontium standard value of 0.710252 is reported.

Hodell and Woodruff (1994) pursue a different line. They claimed that the entire Miocene sea water curve could be best described using a high order polynomial function. They specified a ninth order polynomial best fit, but did not report either the equation or the regression coefficient used. Plotting up their data and fitting the highest order polynomial function (5th order) available on the software package (Cricket Graph 1.3.2, see Fig. 6.3), the following relationship was derived.
Figure 6.3 Graph of Sr ratio versus Age showing the two linear regressions relating the data of Miller et al. (1991) and a 5th order polynomial best fit through the data reported by Hodell and Woodruff (1994).
For 10.74-24.14 (Ma)

\[
\frac{^{87}\text{Sr}}{^{86}\text{Sr}} = 0.75044 - 1.2802 \times 10^{-2} \times \text{[Age (Ma)]} + 1.5413 \times 10^{-3} \times \text{[Age (Ma)]}^2 -
9.0592 \times 10^{-5} \times \text{[Age (Ma)]}^3 + 2.5960 \times 10^{-6} \times \text{[Age (Ma)]}^4 -
2.9104 \times 10^{-8} \times \text{[Age (Ma)]}^5
\]

with a correlation coefficient, \( r^2 = 0.985, n = 171 \)

Data was normalised to a constant standard reference material NIST-987 = 0.710235.

Because polynomial equations higher than 4th order are algebraically insoluble, the roots of this 5th order equation were calculated for the measured \( ^{87}\text{Sr} / ^{86}\text{Sr} \) values by iteration.

For calculation purposes all \( ^{87}\text{Sr} / ^{86}\text{Sr} \) were normalised to the relevant standard by the addition/subtraction of the difference between the measured values of NBS-987 reported (e.g. Miller et al., 1991 and Hodell and Woodruff, 1994 respectively) and those measured during and prior to Sr analysis in East Kilbride (NBS-987 = 0.710237 and 2 S.D. = 0.000026, see for instance Miller et al., 1991).

6.5 Results

Appendix 2 lists all the \( ^{87}\text{Sr} / ^{86}\text{Sr} \) values measured and the ages calculated using the equations derived from both Miller et al. (1991) and Hodell and Woodruff (1994). Errors for these ages were also generated using the analytical error.

6.5.1 Assessment of the validity of the ages

6.5.1.1 Comparison of the two Sr sea water curves

Figure 6.4 is a chart showing the ages and errors derived from \( ^{87}\text{Sr} / ^{86}\text{Sr} \) analysis for those samples for which independent nannoplankton zones were available (C. Müller pers. com., 1992 and 1994) and using the boundaries defined by Berggren et al. (1985, Fig. 2.2). All these samples were either planktic foraminiferal samples or mixed benthic and planktic foraminiferal samples, picked from marls overlying shallow water
<table>
<thead>
<tr>
<th>AGE (Ma)</th>
<th>STAGE</th>
<th>NN ZONES</th>
<th>SECTION NAMES AND AGES DERIVED FROM Sr DATING</th>
</tr>
</thead>
<tbody>
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<td>1</td>
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**PLIO-PLEISTOCENE**

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<td>NN1</td>
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</table>

**OLIGOCENE**

![Figure 6.4](image)

Figure 6.4  Chart showing the ages derived for foraminiferal samples by Sr analysis and their NN zones (C. Müller pers. com., 1992 and 1994). Black squares are derived using the straight line fit from Miller et al. 1991. White squares are derived using a 5th order polynomial fit through Hodell and Woodruff’s (1994) data. Shaded areas are the biostratigraphic ages from nannoplankton analysis of the same samples (C. Müller pers. com., 1993; 1994). See text for more details.
carbonates (Table 6.2). Figure 6.4 shows ages calculated using the equations and $^{87}\text{Sr}/^{86}\text{Sr}$ curves from both the Lower Miocene of Miller et al. (1991) and Hodell and Woodruff (1994) and these vary systematically depending on the regression equation used (Appendix 2) with the ages from Hodell and Woodruff (1994) always being slightly younger than those of Miller et al. (1991). This is due to the difference in the data sets used to define each curve (see Fig. 6.3). Much more spectacular than the variation in age however is the variation in the error bars for the two models (Fig. 6.4; Appendix 2). Those calculated using Hodell and Woodruff's polynomial best fit are up to 5 times larger than those calculated using Miller et al.'s simple linear regression. This discrepancy is due to the coincidental proximity of the foraminiferal $^{87}\text{Sr}/^{86}\text{Sr}$ values to a pronounced inflection point on the polynomial curve (Fig. 6.5). Thus, even relatively small errors on the data span a nearly horizontal part of the curve and result in 1 σs for age of up to 4 million years.

Assessment of which of these two curves is the more valid is beyond the scope of this project. The data reported here merely serve to illustrate the large discrepancies between the results generated by each model in particular regions of the Miocene. A better constraint on how the global $^{87}\text{Sr}/^{86}\text{Sr}$ changes with time during the Miocene will only be gained by producing more independently generated curves from different regions, correlating them and attempting to understand what controls the changes in gradient (e.g. Hodell, 1994).

Because the majority of the $^{87}\text{Sr}/^{86}\text{Sr}$ measured lie within error of this point of inflection on the Hodell and Woodruff (1994) polynomial curve, ages calculated using this curve were discounted and only those calculated using the regression line according to Miller et al. (1991) will be discussed further. It should be noted however that the problem really lies in the data sets on which these curves are based. Both papers document a change in slope in this region of the graph (see Fig. 6.3), but the combination of Millers et al.'s apparently older data set and their chosen point of inflection (in an area of no data!) allows the Sr ratios for this study to be calculated using Miller's Lower Miocene curve with the high correlation coefficient rather than his Upper Miocene curve with its very poor correlation coefficient (sections 6.4.5 and 6.5.2.2). Because of this and the assumption that a polynomial
Figure 6.5 Graph of Sr ratio versus Age (Ma) with the two regression lines relating to the Lower Miocene linear best fit of Miller et al. (1991) and the 5th order polynomial fit of Hodell and Woodruff (1994). Sr isotope ratios for foraminiferal samples shown in Fig 6.5 are plotted using the two regressions to calculate the age. Note that the ages calculated using the polynomial best fit are systematically younger than those of the simple linear regression. Insets A and B are details of the area circled showing the error bars calculated using the two regression equations. Note that the equivalent age errors for the polynomial regression will be much larger than those for the simple linear regression.
relationship does not describe the evolution of $^{87}\text{Sr}/^{86}\text{Sr}$ with time, all the age data presented here should be treated with caution.

6.5.1.2 Foraminiferal samples

The stratigraphic positions of the foraminiferal samples in individual sections generally correlates with the ages generated using Miller et al.'s (1991) method and are within error of the nannoplankton zones identified for these samples. This suggests that late diagenetic processes did not significantly alter the original Sr sea water ratio incorporated into the foraminiferal test (Beets and De Ruig, 1992). Carpenter et al. (1991) showed that early diagenetic cementation, if active, may not have altered the $^{87}\text{Sr}/^{86}\text{Sr}$ to any great extent as secondary calcite and aragonite are in isotopic equilibrium with ambient sea water. It seems likely therefore that the resistance of some foraminiferal samples to ultrasound treatment is due to early diagenetic alteration, which has not significantly altered their Sr composition.

Loss of the primary Sr isotopic signature does occur however, when sediments are subject to open-system behaviour (Quinn et al., 1991; Scoffin 1987) i.e. input of extrinsic Sr with a different $^{87}\text{Sr}/^{86}\text{Sr}$ from that of the sediments. This may have occurred in the two sections where the ages do not match stratigraphic positions i.e. at Deniztepesi and Yaylaalan (Fig. 6.4 and Appendix 2). This is discussed further in section 6.6.

6.5.1.3 Macrofossil samples

The lack of correlation between the ages generated by $^{87}\text{Sr}/^{86}\text{Sr}$ dating and the relative stratigraphic positions of the Poritid coral samples in the Alarahan section (see Table 6.4) illustrates the alteration of the Sr isotopic composition on transformation from aragonite to low Mg-calcite (Appendix 1, XRD data).

The algal samples from the Ahmetler section show a similar lack of correlation (table 6.5) also suggesting a diagenetic influence and that transformation from high Mg-calcite to low Mg-calcite has occurred despite there being no visual evidence for it.
Table 6.4  Measured values of $^{87}\text{Sr}/^{86}\text{Sr}$ and the ages calculated using the Lower Miocene regression from Miller et al. (1991) for the Porites sp. samples in the Alarahan section. The results are arranged in stratigraphic order, e.g. lowest sample at the bottom of the table.

<table>
<thead>
<tr>
<th>Sample number</th>
<th>2 S.E. (ppm)</th>
<th>Dynamic $^{87}\text{Sr}/^{86}\text{Sr}$ measured</th>
<th>Dynamic $^{87}\text{Sr}/^{86}\text{Sr}$ normalised to Miller et al. (1991)</th>
<th>Age (Ma) for 25.1-14.7 Ma from Miller et al. (1991)</th>
</tr>
</thead>
<tbody>
<tr>
<td>4j.Alarahan.4</td>
<td>24</td>
<td>0.708613</td>
<td>0.708628</td>
<td>18.13</td>
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<tr>
<td>4j.Alarahan.3</td>
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<td>0.708365</td>
<td>0.708380</td>
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<td>4j.Alarahan.2</td>
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<tr>
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<td>0.708357</td>
<td>0.708372</td>
<td>22.43</td>
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</table>

Table 6.5  Measured values of $^{87}\text{Sr}/^{86}\text{Sr}$ and the ages calculated using the Lower Miocene regression from Miller et al. (1991) for algal samples in the Ahmetler section. The results are arranged in stratigraphic order, e.g. lowest sample at the bottom of the table.

<table>
<thead>
<tr>
<th>Sample number</th>
<th>2(S.D.)</th>
<th>Dynamic $^{87}\text{Sr}/^{86}\text{Sr}$ measured</th>
<th>Dynamic $^{87}\text{Sr}/^{86}\text{Sr}$ normalised to Miller et al. (1991)</th>
<th>Age (Ma) for 25.1-14.7 Ma from Miller et al. (1991)</th>
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</thead>
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<tr>
<td>16S.313.7</td>
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<td>0.708703</td>
<td>16.879</td>
</tr>
<tr>
<td>12j.Ahmetler.2</td>
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<td>0.708736</td>
<td>0.708751</td>
<td>16.074</td>
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</table>

Sr ages generated for the three oysters samples from the Kargi baraj section (Table 6.6) also failed to show systematic correlation with stratigraphic position. Of the other 7 oyster samples loaded, 4 aborted before completing all 15 blocks and two of these samples showed anomalously high Rb contents as inferred from the Rb interference measured during Sr analysis.

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In this case then it seems that the assumption that radiogenic $^{87}$Sr from the decay of $^{87}$Rb can be ignored due to the relatively low concentrations of Rb in biogenic carbonates (Elderfield, 1986) cannot be sustained probably due to diagenetic alteration and the $^{87}$Sr/$^{86}$Sr measured should be treated with caution (section 6.3). Jones et al. (1994) report successfully using oysters from the Jurassic for Sr dating.

It was not possible to assess the correlation of Sr ages with stratigraphic position for echinoids, operculinids, or gastropods independently as no section had sampled a series of each of these fossils. However, in relation to other fossils, the Sr ages for operculinids and gastropods where XRD showed that primary aragonite had been retained (Appendix 1), indicated that they correlated well with stratigraphic position. Echinoids have a more ambiguous record.

Table 6.6  Measured values of $^{87}$Sr/$^{86}$Sr and the ages calculated using the Lower Miocene regression from Miller et al. (1991) for oyster samples in the Kargi section. The results are arranged in stratigraphic order, e.g. lowest sample at the bottom of the table.

<table>
<thead>
<tr>
<th>Sample number</th>
<th>2(S.D.)</th>
<th>Dynamic $^{87}$Sr/$^{86}$Sr measured</th>
<th>Dynamic $^{87}$Sr/$^{86}$Sr normalised to Miller et al. (1991)</th>
<th>Age (Ma) for 25.1-14.7 Ma from Miller et al. (1991)</th>
<th>Age (Ma) for 14.5-8.3 Ma from Miller et al. (1991)</th>
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6.5.2 Ages of the shallow water carbonates and the transgression surface

Prior to this work, only one of the shallow-water carbonate successions had been directly dated (Oymapinar section; Poisson, pers. com. 1992). The ages produced by $^{87}$Sr/$^{86}$Sr analysis for these neritic sediments therefore provide valuable information about the timing of transgression (i.e. the lower surface of the shallow-water carbonates) rather than the age of the transition to
planktic foraminiferal marls (chapter 4). $^{87}\text{Sr}/^{86}\text{Sr}$ ratios from these marls were used to confirm the validity of the results by comparison with biostratigraphic ages. Furthermore, this work has allowed the study of previously undated sections. Particularly important is the Bucakköyü section in the west of the Köprü basin, where shallow-water carbonates are found within a thick succession (~1.5km) of conglomerates. The results of this study are discussed in the light of their structural implications in chapter 7.

6.5.2.1 Lower-Middle Miocene

The ages derived from Sr dating of the shallow water carbonates, where not automatically discounted due to evidence of diagenetic alteration, are shown in figure 6.6. The samples from the Manavgat basin show a younging trend from south east to north west spanning the Burdigalian to Earliest Langhian (Berggren et al., 1985) or NN2 - NN5. In accordance with the log correlation shown in figure 4.24 this suggests that the Mesozoic basement was transgressed from the south and that this transgression took place during the Burdigalian. Samples from the western side of the Köprü basin, though located further north than those of the Manavgat basin span a similar time (NN2-NN4; Fig. 6.6). This has been interpreted as indicating that the Köprü basin already had significant basinal topography during the Early Miocene allowing an earlier marine influence than on the more peneplaned Manavgat area. This is in agreement with sedimentological data discussed in chapter 5.

The Sr ages for the foraminiferal samples show a similar, though less exaggerated pattern of diachroneity to that of the shallow water carbonate samples, such that sections where early shallow water carbonate has been preserved also contain early evidence of deeper water marls. Although no sections which did not contain shallow water carbonate were sampled and analysed, it seems probable that deposition outside the Manavgat basin was controlled by local structure and tectonics. This is discussed further in chapter 7.
<table>
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<tr>
<td>6</td>
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**PLIO-PLEISTOCENE**

- MESSINIAN
- TORTONIAN
- SERRAVALLIAN
- LANGHIAN
- BURDIGALIAN
- AQUITANIAN

**OLIGOCENE**

- AQUITANIAN
- BURDIGALIAN
- LANGHIAN
- LOMBARDOIAN
- MESSINIAN

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Figure 6.6  Chart showing the ages derived from non-foraminiferal samples by Sr analysis using the linear fit from Miller et al. 1991.
6.5.2.2 Upper Miocene

Only three of the 54 samples measured had a $^{87}\text{Sr}/^{86}\text{Sr}$ equivalent to Upper Serravallian or younger. The imbalance is partially due to a sampling bias in that the Lower Miocene reefs are not only more abundant but also better preserved than the Upper Miocene reefs in the study area. It is possible however that diagenetic alteration resulted in lower $^{87}\text{Sr}/^{86}\text{Sr}$ in some Upper Miocene samples (see discussion below). Calculating the equivalent ages of the younger samples in some cases proved problematic for the following reasons.

- The regression coefficient for the Upper Miocene given by Miller et al. (1991) is very low ($r^2 = 0.603$). There is little confidence therefore that this evolution line fits the data points.
- The polynomial fit of Hodell and Woodruff (1994) is not valid beyond the influence of their data points. Thus samples with $^{87}\text{Sr}/^{86}\text{Sr} > 0.708884$ (i.e. younger than 10.74 Ma) cannot be calculated.
- The convergence of Africa and Eurasia resulted in the isolation and desiccation of the Mediterranean at the end of the Miocene. It is not clear when global Sr curves become invalid for the Mediterranean (Müller et al., 1990).

6.6 Diagenesis

In seven out of the 11 foraminiferal samples where independent nannoplankton zones were identified, the Sr ages obtained, though within error, are on the old side (Fig. 6.4). The Sr values of diagenetically altered corals in the Alarahan section produced ages of between 22.5 and 18.1 using the regression of Miller et al. (1991) or 21-17.5 according to Hodell and Woodruff (1994) (table 6.4), significantly older than expected (e.g. Upper Burdigalian). All Sr curves for the Lower-Middle Miocene (e.g. Miller et al., 1991; Hodell et al., 1991; Oslick et al., 1994; Hodell and Woodruff, 1994) predict that diagenetic alteration in a marine environment occurring substantially later than formation would result in a higher $^{87}\text{Sr}/^{86}\text{Sr}$ value (see Fig. 6.3). It would seem reasonable to infer therefore that the low $^{87}\text{Sr}/^{86}\text{Sr}$ values of the Alarahan coral samples are not entirely the product of diagenetic alteration in a fully marine environment.
Rapid stabilisation of primary aragonite and high Mg-calcite to low Mg-calcite occurs in the freshwater phreatic (5000 yrs) and vadose (10000-200000 yrs) zones whilst stabilisation in marine phreatic environments where minerals have stayed in contact with trapped formation waters can take up to 3 million years (Scoffin, 1987). Freshwater \(^{87}\text{Sr}/^{86}\text{Sr}\) ratios are influenced by the geology of the basement through and over which it passes. When looking at the \(^{87}\text{Sr}/^{86}\text{Sr}\) of Messinian evaporites in the Mediterranean, Müller et al. (1990) assumed that the \(^{87}\text{Sr}/^{86}\text{Sr}\) composition of the rivers emptying into the Mediterranean is the same today as it was during the Miocene because the source area has not changed much since then (Albarede and Michard, 1987). Today, the mean \(^{87}\text{Sr}/^{86}\text{Sr}\) value of river water from the Nile is 0.7060 (Brass, 1976; Albarede and Michard, 1987) much lower than both the present day sea water \(^{87}\text{Sr}/^{86}\text{Sr}\) value and the range of values for the Miocene (0.708-0.709) and this is attributed to the hinterland basement being largely Mesozoic carbonate and mafic rocks. A hinterland dominated by heavily karstified Mesozoic limestone (Degirmenci, 1992) and mafic rocks surrounds the study area (Fig. 1.4) suggesting that run off in this region is also likely to have low \(^{87}\text{Sr}/^{86}\text{Sr}\) values. Thus, diagenetic alteration influenced by freshwater may account for the anomalously low \(^{87}\text{Sr}/^{86}\text{Sr}\) of the coral samples.

The ratio of Sr in fresh water to sea water is 1:10 (Müller, 1990). Thus high concentrations of fresh water would be required to influence the Sr ratios in a marine environment. Even in very marginal reef carbonate settings with high, but variable volumes of run-off, the \(^{87}\text{Sr}/^{86}\text{Sr}\) does not appear to be influenced by freshwater \(^{87}\text{Sr}/^{86}\text{Sr}\) values (S. Tudhope and R. Ellam pers. com., 1994) although it is possible that in this case the run-off had a similar \(^{87}\text{Sr}/^{86}\text{Sr}\) to sea water. It therefore seems likely that the diagenetic event that led to the changing of the Sr signature occurred in a non-marine environment e.g. in an open system (Quinn et al., 1991; Scoffin, 1987)

Although it is possible that the entire \(^{87}\text{Sr}/^{86}\text{Sr}\) signature preserved in these corals is due to the most recent period of exposure to an open fresh water system (e.g. ~Middle Pliocene, C. Glover pers. com., 1994) it is possible that it occurred during the Late Miocene when a combination of the Messinian draw down (Hsü et al., 1973) and the Aksu phase compressional event
(Poisson, 1977) is likely to have left most, if not all of the study area above sea level.

6.8 Conclusions

- A diachronous transgression from south to north identified using nannoplankton biostratigraphy of planktic foraminiferal marls (Geceleme Formation) affected the Manavgat basin during the Burdigalian (chapter 4). This is confirmed by $^{87}\text{Sr}/^{86}\text{Sr}$ dating of the neritic sediments (Oymapinar Limestone).

- A Lower-Middle Burdigalian age was produced for a previously undated key section in the west of the Köprü basin.

- Deposition of shallow-water carbonates occurred earlier in the western central part of the Köprü basin than in the north of the Manavgat basin.

- Fresh water diagenetic alteration of samples may result in significantly lowering the $^{87}\text{Sr}/^{86}\text{Sr}$ due to the domination of the hinterland by Mesozoic carbonate and mafic rocks.

- The accuracy of calculated ages for the Mid Miocene are difficult to assess due to a change in the gradient of the $^{87}\text{Sr}/^{86}\text{Sr}$ curve. Further detailed study of the global Sr curve at this time is needed both to aid the use of this technique as a chronological tool and in understanding the controls over such changes in gradient.

- The low gradient of the global $^{87}\text{Sr}/^{86}\text{Sr}$ curve in the Upper Miocene means that error bars on the ages for this time period are large. In the Mediterranean however Late Miocene isolation from the global oceans means that the Sr curve is not valid. Further Sr analysis on mid-Mediterranean samples is needed to pin down when fluvial input to the Mediterranean first began to affect the $^{87}\text{Sr}/^{86}\text{Sr}$. Comparison between the east and west Mediterranean would add to the understanding of the relative independence of intra-Mediterranean basins during the Messinian.
Foraminifera which are resistant to ultrasound treatment have probably suffered diagenetic alteration.
Chapter 7

STRUCTURE AND TECTONIC CONTEXT

7.1 Context

Brittle and ductile structures measured in the field area are considered along with inferred structural information from sedimentary facies and thickness variations. Observations of this kind have been used by other authors to support various models for the evolution of the Isparta Angle during the Miocene. Stress-field analysis from this information is not possible however without taking into account the strong N-S structural grain of basement rocks (e.g. Antalya Complex, section 1.4). Development of Miocene basins must also be seen in terms of the wider picture of tectonic processes active in the Eastern Mediterranean at the time, e.g. westward expulsion of the Anatolian block and subduction along the Hellenic trench.

7.2 Organisation of this chapter

Section 7.3 is a brief summary of the structural setting of the Isparta Angle which has been discussed at greater length in chapter 1 (section 1.4). This is followed by a résumé of the previous work which focused on the structural development of the area. Section 7.5 presents the data collected during field seasons as part of this study. Deformational structures, both from direct observation, and inferred indirectly from other field evidence, are discussed and interpreted. This section has been divided into two parts in order to treat the Manavgat basin separately from the Aksu and Köprü basins. This was thought necessary because of the difference in orientation, both of the present-day basin margins, and the structures within them. Published models of the evolution of the Isparta Angle during the Miocene and Plio-Quaternary are considered in section 7.6 in the light of the data presented here, and a new model is proposed.
7.3 Introduction

The Miocene sedimentary basins under study here are contained within in a zone known as the Isparta Angle ("Coubure d’Isparta"; Blumenthal, 1963) which sits at the junction between the Hellenide and Tauride arcuate orogenic belts and is bounded on either side by Mesozoic allochthonous units (e.g. Lycian Nappes, Beysehir-Hoyran-Hadim Nappes; Fig. 7.1). Tightening of this angle occurred during the Miocene and at the Miocene-Pliocene boundary producing compressive deformational structures in Miocene sediments. Within this compressional context, very few extensional features obviously relating to the formation of the Aksu, Köprü and Manavgat basins are apparent. Extension can be inferred however, from sedimentary thickness variations in the Manavgat basin and small micro-faults, at, or near, the base of Miocene sections (see below). Reactivation of Miocene and older structures is a constant source of complexity in an area which has suffered multiple tectonic events dating back to the Triassic (Robertson and Woodcock, 1982; Senel, 1984). It is within this historical framework that structures observed and measured in Miocene sediments from the study area are considered and discussed in terms of the basin evolution.

7.4 Previous work

The Palaeozoic-Recent history of the study area has been the subject of research over the last 20 years and has in some instances occasioned vigorous debate (for synthesis see Robertson and Dixon, 1984). Much of this centres on the origin and emplacement (mode and direction) of the Antalya Complex, which forms part of the basement to the Miocene basins within the Isparta Angle (Figs. 7.1 and 7.2, e.g. Lefevre, 1967; Brunn et al., 1971; Dumont et al., 1972; Brunn, 1974; Poisson, 1977; Monod, 1977; Ricou et al., 1979; Hayward, 1984; Robertson and Woodcock, 1980; 1982; 1984; Poisson, 1984; Ricou et al., 1984; Waldron, 1984; Marcoux et al., 1989; Robertson, 1993). A more detailed outline of the debate surrounding the Antalya Complex is given in chapter 1 (section 1.4), but for the purposes of this chapter, a brief summary is provided here.
Figure 7.1 Map of a) the north-eastern Mediterranean showing the tectonic units mentioned in the text and b) the Isparta Angle region and surrounding units.
Figure 7.2  A more detailed map of the Isparta Angle showing the north-south structural lineaments within the Antalya Complex and the location of the palaeomagnetic sites of Morris and Robertson (1993). Modified from Waldron (1984).
The Antalya Complex is divided into five north-south trending tectonic zones (Fig. 7.2; Woodcock and Robertson, 1981a; b) which, taken as a whole, record the initiation, construction and later disruption of part of a small Mesozoic-Cainozoic oceanic basin by a combination of thrust and strike-slip tectonics (Robertson and Woodcock, 1980b; Woodcock and Robertson, 1981). Poisson et al. (1983) showed that the Antalya Complex was emplaced prior to the Oligocene in the central part of the study area (south-west of the Köprü basin). By contrast, Hayward (1982) demonstrated that the Antalya Complex in the south-west of the Isparta Angle was an uplifted landmass shedding sediments in the Early Miocene and was thrust westward, near to its present position, onto the Bey Daglari (Fig. 7.2) during the Mid-Miocene.

Extensive palaeomagnetic studies of both eastern and western arms of the Isparta Angle and the Miocene sediment between them, have been carried out over the last decade. These show that the eastern limb, consisting of the Beysehir-Hoyran-Hadim (BHH) nappes and part of the Anamas-Akseki platform (Fig. 7.1) has rotated clockwise by 40° during the Late Eocene-Oligocene (Kissel et al., 1993). The western limb of the Isparta Angle (e.g. the Bey Daglari; Kissel and Poisson, 1987), the south western segment of the Antalya Complex and the northern part of the Anamas-Akseki platform (Morris and Robertson, 1993; Figs. 7.1 and 7.2) was rotated anticlockwise by 30°, after Burdigalian-Langhian times (Morris and Robertson). Kissel and Poisson (1986) stated that there is no evidence for the rotation of Neogene sediments within the Isparta Angle during the last 15 Ma (e.g. since Langhian). They interpreted this apparent paradox by decoupling the Lycian Nappe system from the Akseki-Beysehir Taurides using a décollement horizon at the base of the Antalya Complex to take up the rotation of the Bey Daglari. Their model, shown in figure 7.3, indicates that the formation of the Bey Daglari anticline was caused during anticlockwise rotation, by the movement of the Lycian Nappes towards the SE (Kissel et al., 1993). A 30° anticlockwise rotation of the Bey Daglari is supported by Morris and Robertson (1993). Their data from the south-western segment of the Antalya Complex however, indicated that this area was rotated along with the Bey Daglari. This implies that a décollement horizon at the base of the Antalya Complex is unlikely to have taken up all the rotation. One possibility is that rotation
Figure 7.3  Schematic model for the formation of the Isparta Angle from Kissel et al. (1993). BHH nappes = Beysehir-Hoyran-Hadim nappes; A.C.t.b. = Antalya Complex basal thrust; A. t.b. = Akseki thrust belt; A.t. = Aksu thrust.
was accommodated across a more diffuse zone within the Antalya Complex, potentially utilising the older north-south tectonic lineaments. Morris and Robertson (1993) document a remagnetisation event which occurred during the Early-Middle Miocene and suggest that this is likely to have resulted from the migration of orogenic fluids ahead of the advancing Lycian Nappes.

Studies of field evidence within Neogene sediments of the Isparta Angle have also occupied various authors. Hayward (1982; 1984) followed up earlier work (e.g. Gutnic and Poisson, 1970; Poisson, 1977; Woodcock and Robertson, 1977a; 1977b) on the NE-SW trending Miocene basin to the west of the Bey Daglari (Figs. 7.1 and 7.2). This basin was interpreted as a foreland basin succession related to the south-eastward thrusting of the Lycian Nappes (Hayward, 1982; 1984). Hayward (1982) also studied the small Miocene basin on the south-eastern margin of the Bey Daglari (Fig. 7.2), which he similarly interpreted as having been formed by load induced flexure.

The N-S orientation of the Bey Daglari anticline and the prominent high angle N-S striking reverse faults in the centre of the Isparta Angle (e.g. the Kirkkavak fault, Pinargözü fault zone, Aksu thrust; Figs. 7.5 and 7.22) led Poisson (1977a and b) to define an east-west compressional event which he named the Aksu Phase, and dated as Latest Miocene. The front of this west-vergent system was considered to be situated in the middle of the Bey Daglari (Poisson, 1977; Ricou et al., 1977; Marcoux et al., 1989). The cause of the west vergence of this event has long been thought (e.g. Dumont et al., 1979) to be closely related to the westward expulsion of the Anatolian block (McKenzie, 1972; 1978; Sengör and Yılmaz, 1981; Sengör et al., 1985) which began with the collision at the Bitlis Suture zone (Sengör and Yılmaz, 1981; Dewey et al., 1986) at around 13 Ma (Le Pichon and Angelier, 1979; Le Pichon et al., 1993).

Dupoux (1983) examined the field evidence for the Aksu Phase in the Köprü basin and noted that populations of micro-faults were orientated both NNW-SSE and NE-SW. He interpreted the NW-SE trends as the products of rotation from an original N-S orientation due to strong sinistral shear along NNW-SSE orientated faults. Dupoux (1983)
considered that the Aksu Phase was the last compressive phase to have affected the Isparta Angle.

More recently, this compressional phase has been re-examined by Frizon de Lamotte et al. (1995). These authors compiled previously published data (Dumont, 1979; Dumont et al., 1979; Dupoux, 1983, Akay and Uysal, 1985; Marcoux et al., 1989, Auboug et al., unpublished data) and presented new generational slickenside data from the study area suggesting that contrary to Dupoux's (1983) conclusion, the Latest Miocene deformation can be divided into two distinct events with different orientations:

- First, E-W compression resulting in broadly N-S striking reverse faults in the centre of the Isparta Angle (the Aksu Phase);
- Later, south-directed transport (the Susuz Dag Phase).

Frizon de Lamotte et al. (1995) propose that rather than forming either due to movement of the Lycian Nappes (Kissel, 1993), or as a result of the Aksu Phase east-west compression (Poisson, 1977), the Bey Daglari anticline developed as a regional lateral culmination on a south-vergent blind thrust (Fig. 7.4), parallel to the thrust front visible off-shore along the Florence Rise (Fig. 7.1; Sage and Letouzey, 1990). Investigation of the Florence Rise was undertaken by a "Training through Research" Cruise in 1991. Results from this study, suggest that the interpretation of the structures visible along the Florence Rise may be more complicated (Woodside et al., 1992) than initially suggested by Sage and Letouzey (1990).

The Dariören basin is the north-eastern extension of the Lycian flexural basin at the apex of the Isparta Angle (Fig. 7.2) Here, an Aquitanian-Burdigalian succession was thought to have been terminated by south-westward thrusting and folding of the Davras Dag carbonate platform to the north (Poisson, 1977; Akbulut, 1977). The age of this deformation is thought to be prior to the Aksu Phase (i.e. pre-Messinian).

Much interest has been concentrated on the active N-S extension along the coast of western Turkey (McKenzie, 1972; 1978; Jackson et al., 1982; Jackson and McKenzie, 1984, 1988; Papadopoulos et al., 1986; Lyon Caen et
Figure 7.4 Block diagram illustrating the geometry of a lateral culmination fold on a south-vergent blind thrust as envisaged by Frizon de Lamotte et al. (1995). Diagram modified from Frizon de Lamotte et al. (in press). Originally from Schirmer (1988); Frizon de Lamotte et al. (1991).
al., 1988; Eyidogan, 1988). Berckemer (1977), Le Pichon and Angelier (1979), Le Pichon (1981) and Angelier et al. (1982) have suggested that the extension may be governed mainly by extensional forces acting on the leading edge of the upper plate along the Hellenic subduction zone. This extensional system has also been modelled, taking into account the dextral strike-slip movement related to the North Anatolian Fault (NAF) and the westward escape of the Anatolian block (McKenzie, 1978; Jackson and McKenzie, 1984; 1988; Ekström and England, 1989; Taymaz et al., 1991; Zanchi et al., 1993). The Isparta Angle lies at the boundary between the zone dominated by extension behind the Hellenic Arc and that governed by expulsion-driven compression (Dumont et al., 1979). Neotectonic studies of the area have therefore focused on the formation of the Plio-Quaternary graben systems in the northern lakeland area of the region (e.g. the Aci area, Angelier et al., 1981; Burdur, Price and Scott, 1994) and in assessing the mechanics and implications of formation of these structures (Jolivet et al., 1994).

7.5 Deformational structures

7.5.1 The Manavgat basin

7.5.1.1 Basement and basal Miocene structures

The basement of the eastern part of the Manavgat basin is entirely composed of the Alanya Massif (Fig 7.1). This consists of Permian meta-carbonate, micaceous schists and blueschists (Ozgül, 1983; Okay and Ozgül, 1984). It is thought to have been emplaced onto the Antalya Complex to the north before the Early Eocene (Monod, 1977; Okay and Ozgül, 1984). Outcrops of the Alanya Massif are highly fractured and veined, but unless faults penetrating the Alanya Massif also offset the overlying Miocene sediment, identification and dating of fault groups within the basement could not be attempted. Faults affecting both the Alanya Massif and the Miocene succession were occasionally found along the northern margin of the basin, e.g. 1km from the Akseki road locality, but it was in all cases impossible to be sure whether all overlying Miocene sediment was affected, or just the Lower and Middle Miocene successions.
No Miocene sediment overlies the fault along which the Alanya Massif was emplaced onto the Antalya Complex.

The Antalya Complex forms the boundary of the Manavgat basin for less than 2km. Its relationship with Miocene sediment can be observed along the road north from Yaylaalan (Fig. 7.5) where it is unconformably overlain by a thick succession of Kargi Member conglomerates (see chapter 5, table 5.1). Palaeocurrent evidence from these conglomerates and the absence of Miocene sediments prior to reef formation in the Late Burdigalian-Langhian on the northern part of the Alanya Massif, indicates that during the Lower Miocene the Anatalya Complex was palaeogeographically low relative to the Alanya Massif to the south (see Chapter 5 section 5.3.3). Although there is no clear evidence for fault-generated palaeotopography between the Antalya Complex and Alanya Massif, this possibility cannot be ignored.

The most northern section of the margin of the Manavgat basin is composed of part of the Anamas-Akseki platform carbonates (Fig. 7.1). Access to this part of the northern margin of the basin is extremely difficult, and although the boundary was observed from a distance (e.g. Kizildag and Karabucak; Fig. 7.5), it was not studied in detail. Here again, the basement is overlain by a thick succession of fan-delta conglomerates (see chapter 5) and it is probable that active faulting along, or close to, the present-day margin helped generate the large volume of conglomerate preserved.

7.5.1.2 Evidence of brittle fracture in Miocene sediments

Faulting

Micro-faults within the Miocene sediments of the Manavgat basins are prolific. In the poorly lithified sediments of the Karpuzçay and Geceleme formations however, fault surfaces are generally not well preserved. The better lithified Lower Miocene formations including the Oymapinar limestone and Çakallar Formation contain well exposed micro-scale faults, often with both striations and crystal lineations developed on fault planes.
Figure 7.5 Localities referred to in the text of chapter 7.
The fault populations collected from this area can be divided into two groups based on their stratigraphic position and orientational characteristics. Figure 7.6 is a map of the Manavgat basin with stereonets of fault data measured in the Oymapinar limestone at the Akseki road locality and the Geceleme and Karpuzçay Formations along the road to Akseki from the coast. The normal faults measured from the Oymapinar limestone are well clustered and have slightly lower dips to the planes than those from the overlying softer formations (e.g. Geceleme and Karpuzçay; Fig. 7.6b).

This angular difference may be the result of more than one period of extensional faulting (Fig. 7.7), although lithological control over the angle of faulting cannot be ruled out. There is however sedimentary evidence to support an early faulting event affecting the Lower-Mid Miocene successions i.e. the Oymapinar limestone, in the form of fault talus at the Akseki road locality (see chapter 3 section 3.6.2.3). The rapid transition from shallow-water carbonate formation to deeper water planktic foraminiferal marls and its correlation with talus and coarse calcirudite deposits (the Çakallar Formation) also suggest active tectonics at the Burdigalian-Langhian boundary (sections 3.6.2.3, 4.8, 5.9). Further evidence for a Burdigalian-Langhian extensional event is presented below (see fractures and sedimentary evidence for faulting).

In detail, however, the faults measured from the Oymapinar limestone at the Akseki road locality have a more complicated history than simple normal faulting. Although all the faults measured here had normal offsets, where slickensides were preserved they indicated movement with reverse and sinistral components (Fig. 7.8). Similar faults are seen in the central part of the Manavgat basin in the Oymapinar limestone near the village of Halitigalar (Fig. 7.5). As with the normal-offset faults in the Karpuzçay and Geceleme Formations, reverse-offset faults in these less well-lithified formations generally have poorly preserved fault surfaces and very few slickensides. Reverse-offset faults are more rare than normal offset faults however, but when they are seen they have the same NW-SE strike and high angle associated with normal faults in these formations (Fig. 7.6). Strike slip components on these faults are
Figure 7.6  Stereoplots of the poles to the planes of faults in the Manavgat basin: a) faults with normal offsets in the Oymapinar limestone at the Akseki road locality; b) faults with normal offsets in the Geceleme and Karpuzcay Formations along the road to Akseki; c) planes of faults with strikeslip slickensides showing the orientations of faults with sinistral and dextral displacements.
Figure 7.7  

a) Normal faulting event in the Early Miocene leads to talus formation and forms high-angle normal faults. 

b) Continued extension on these fault causes back-rotation producing a shallower angle. 

c) Subsequent faulting forms new higher angle faults in the overlying sandstones and marls.
Photographs showing the normal offsets and transpressional slickensides on the faults at the Akseki Road locality.
occasionally possible to recognise, but are impossible to assess or measure without slickenside information. Given that there is no evidence of pure reverse faulting however, it is assumed that most of the reverse faults observed in the Middle to Late Miocene sediments were transpressional. Figure 7.9 shows a 2-d model of the changing fault-types through time from the Lower Miocene to Plio-Quaternary.

Figure 7.6C shows strike-slip faults measured largely from Oymapinar limestone exposures around the basin. A clear pattern of NW-SE striking faults associated with sinistral movement and NE-SW striking faults associated with dextral displacement can be discerned. The similarity of strike between the normal offset faults and sinistral strike-slip faults produces an explanation for the reactivation seen on faults at the Akseki road locality.

The youngest sediments affected by this transpressional faulting are those at the top of the Karpuzçay Formation. The youngest so far documented occur in the Ahmetler section and have been dated as nanofossil zone 11b (Flecker et al., 1995). Overlying the Karpuzçay Formation is an erosional unconformity at the base of the Pliocene succession the oldest of which are thought to be N19 (Akay and Uysal, 1985; Akay et al., 1985; C. Glover, pers. com., 1995). It is therefore assumed that the age of the transpressional faulting is Latest Messinian to Early Pliocene (e.g. Poisson, 1977; Frizon de Lamotte et al., 1995). The unconformity can therefore be assumed to have formed between the Latest Messinian and the Early Pliocene and this confirms Poisson's (1977a) age for the Aksu Phase of compression. There is no evidence to suggest conclusively however, that reverse faulting only occurred at the Mio-Pliocene boundary.

Normal faults affecting Upper Miocene sediments are of higher angle than those in the older Miocene strata (Fig. 7.6). Similarly orientated faults also affect Pliocene sediments (C. Glover, pers. com., 1995), but it is not clear when, in the Plio-Quaternary they were formed. The three possibilities are as follows:

- Shortly after the compressional phase at the end of the Messinian when the Pliocene basin was formed;
Figure 7.9  

a) Normal faulting event in the Early Miocene leads to talus formation and forms high-angle normal faults. b) Continued extension on these faults causes back-rotation producing a shallower angle. c) The Aksu phase at the end of the Miocene causes reactivation of normal faults with transpressional displacements, but this is not sufficient to completely reverse the previous normal displacement. High-angle reverse faults are formed in Middle to Late Miocene sediment. d) Subsequent extension results in the formations of new higher angle faults in the overlying sandstones and marls. It is not clear whether many of these are reactivated transpressional faults. There is no evidence that the Early Miocene faults are reactivated during the Plio-Quaternary extensional phase.
During the Late Pliocene, when tilting of Pliocene sediments occurred (C. Glover pers. com., 1995);

During an extensional event thought to have taken place during the Quaternary which formed the graben structures in the north of the area, (e.g. Burdur Graben; Price and Scott, 1994).

It is beyond the scope of this project to speculate as to which of these events may have caused the high angle normal faults in the Upper Miocene sediments, but the reader is referred to a Ph.D. thesis currently being undertaken by C. Glover for further information on the evolution of the Pliocene basin in this area.

Fractures

On the southern limb of the anticline (Fig. 7.10), along the road from the coast to the Selçuk castle of Alarahan a fairly extensive exposure of Çakallar Formation can be seen containing abundant small fractures. These fractures are well developed in the better lithified fine- to medium-grained sandstone horizons and bound small (i.e. 1-2 cm) tilted blocks, similar to extensional fault blocks (Fig. 7.11). The coarser debris flows and calcarenites interbedded with these horizons however, are not fractured in any systematic way at all and the less well lithified marls show weakly developed, non-penetrative fracturing. At the boundary between fractured sandstone and largely non-fractured calcarenite layers the tilted blocks, in rotating backwards, have impinged upon the overlying sandstone layer. Instead of the fracture persisting across the boundary into the overlying calcarenite, there is generally no sign of the fracture, but the relief on the base of the calcarenite mirrors the offset pattern of the fractures, resembling a cast of asymmetrical ripples (Fig. 7.11).

One method of producing this fracture pattern is to invoke early extensional joint formation prior to complete lithification of the calcarenite which therefore behaves in a more ductile-like fashion. At a larger scale, soft sediment deformation in relation to active faulting would suggest syn-sedimentary deformation. In this case the calcarenite would have been deposited as a drape to a normal-fault structure. Here, however, because of the multiple layers of fractured and non-fractured
Figure 7.10  Geological map showing the broadly east-west orientation of the west-vergent folds in the east of the basin.
Figure 7.11  a) Schematic sketch of the fractured and non-fractured layers in the Çakallar Formation at Alarahan; b) Stereonet projection of these fractures showing that they have a bipartate distribution. Those with planes parallel to the π-girdle of the fold (N-S) all have small normal offsets on them and are almost certainly unrelated to the formation of the fold. The similarity between the orientations of these fractures and the normal faults seen in the Lower Miocene Oymapinar Limestone (Fig. 7.6) suggest that these are joints and may have formed at a similar time. However, later Plio-Quaternary extension cannot be ruled out. The small number of fractures whose planes are orientated east-west could theoretically be related to the folding event in the form of axial planar cleavage. However, it is also possible that they are conjugate sets to the extensional joint system.
beds and the clear relationship between jointing and lithological type/grainsize, it is suggested that fracture formation occurred after deposition of the entire sequence and lithification of the sandstones, but prior to complete lithification of the calcarenites and marls.

Figure 7.11 is a stereonet projection of the poles to the planes of the fractures measured at the Alarahan locality. It shows that the fractures have a bipartite distribution. The majority of the fractures including all those which had a noticeable normal offset have planes orientated N-S parallel to the π-girdle of the main west-verging anticline. From their orientation these joints seem unlikely to be related to the formation of the fold. The second much smaller group of fractures which cluster close to the π-girdle in areas where poles to bedding of the fold are absent is much more likely to be related to fold generation. No normal offsets were observed on these fractures and it is possible that they represent axial planar cleavage formed during fold formation (section 7.5.1.3 below).

Sedimentary evidence for faulting

Evidence for Miocene faulting has been deduced from the study of Manavgat basin sediments and discussed mainly in chapter 3 (section 3.8.1). It can be summarised as follows:

- There is an abrupt transition from shallow-water carbonates to deeper-water planktic foraminiferal marls at the Burdigalian-Langhian boundary;
- This boundary is also marked in many sections by coarse angular conglomerate horizons which have been interpreted as fault talus.

Akay's (1985) NE-SW cross-section across the syncline and anticline in the Manavgat basin (Fig. 7.14a) shows the following features of interest:

- The points of inflection at the sides of the folds are all rather steeper than would be predicted from dips elsewhere on the fold (i.e. rather box-fold-like in shape;
- The thickness of almost all of the formations varies across the fold.
Figure 7.12  A structural cross section across the Manavgat basin, hung off the top of the Geceleme Formation. Note that the largest thickness differences are in the Geceleme Formation. This, and the association of the Çakallar Formation which is thought to be at least partly fault derived, indicate that there was a post-reef faulting event which led to the deepening of the basin in two distinct localities, the far north and the far south.
Figure 7.12 shows sections measured in the field during this study at the three inflection points on the sides of the folds (Ahmetler in the extreme north, Halitigalar on the northern margin of the anticline and Çakallar on the southern limb of the anticline) exposed in the basin. These sections are hung from the boundary between the bottom of the Karpuççay Formation and the top of the Geceleme Formation. The thickness changes become apparent at once and can be seen in table 7.1.

Interpretation

The distribution of the Kargi conglomerates suggests that alluvial depositional processes infilled a Pre-Lower Miocene palaeotopography on top of the Alanya Massif. In the north of the basin, the Alanya Massif formed a "high" on top of which no conglomerates were deposited (see chapter 5, section 5.7.1, Fig. 5.3). The altitude of this "high" cannot be estimated, but there is little evidence of it being a source of material for deposition in the basin and it is therefore not thought to have had extensive relief. In central parts of the basin the accumulation of 60-70m of in situ reef carbonate in the Burdigalian and the absence of any deeper-water sediment suggests that, in this region at least, little of the accommodation space generated by the previous palaeotopography remained. The diachronity of the reefs indicates that there was still a south-dipping slope from the Alanya Massif "high", such that a rising sea level resulted in northward transgression (chapter 4 section 4.7 for discussion; section 6.5.2).

The Çakallar Formation has been interpreted as having been generated largely by faulting of nearby reef carbonates and Alanya Massif basement (see chapter 3). In this context their patchy distribution around the basin would reflect the localised faulting at the time. It is interesting to note that the localisation of Çakallar Formation along the limbs of the folds, albeit unevenly, suggest that they were the focus for faulting prior to formation of the fold. This distribution pattern is similar to that documented by Follows and Robertson (1990) in the Miocene of southern Cyprus. The suggestion that the limbs of the fold were also the location of faulting is borne out by the thickness variation of the Geceleme Formation. Here, the far thicker successions on the southern limbs of the
Table 7.1  Thickness variations and a summary of the main rock types making up the Miocene formations folded in the east of the Manavgat basin.

<table>
<thead>
<tr>
<th>Formation</th>
<th>Rock type</th>
<th>Thickness variation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Karpuzçay Formation</td>
<td>Sandstones, conglomerates, siltstones and marls</td>
<td>Top of the formation not seen in the north of the basin and overlain by an erosional unconformity at the base of the Pliocene in the south. At least 4-600m thick</td>
</tr>
<tr>
<td>Geceleme Formation</td>
<td>Planktic foraminiferal marls (100-200m water depth)</td>
<td>Thick successions on the south-facing limbs of the folds. A much thinner succession is seen at Halitigalar on the north facing limb of the fold</td>
</tr>
<tr>
<td>Çakallar Formation</td>
<td>Calcarenites and debris flows containing clasts of shallow water reefal material and Alanya Massif basement, interbedded with sandstones and marls</td>
<td>Patchy distribution e.g. not at Ahmetler, but 5km to the east several hundred meters is exposed. Distributions is concentrated along the limbs of the folds, not at fold noses</td>
</tr>
<tr>
<td>Oymapinar Limestone</td>
<td>Shallow water in situ coral and algal reefs</td>
<td>Relatively constant thickness across the area. Diachronous from south to north (Bizon et al., 1974; Akay and Uysal, 1985; Akay et al., 1985; Flecker et al., 1995)</td>
</tr>
<tr>
<td>Kizildag Formation</td>
<td>Mainly continental conglomerates and sandstones</td>
<td>Pinch out towards the north. Not found at all on the part of the northern margin of the Manavgat basin which is bounded by the Alanya Massif</td>
</tr>
</tbody>
</table>

Folds relative to the northern limb of the anticline indicates that the extensional faulting which resulted in the deposition of the Çakallar Formation produced an asymmetrical half-graben feature in the north of
Figure 7.13 Schematic cross-section across the Manavgat basin at the end of deposition of the Geceleme Formation. The basically asymmetric fault geometries shown here are inferred from the variations in thickness of the Geceleme Formation across the area (see Fig. 7.12).
the basin and another fault-generated deep in the south. Figure 7.13 is a schematic cross-section suggesting the structure of the Manavgat basin at the end of Geceleme Formation deposition.

Summary of the conclusions drawn from the evidence for faulting

- There is substantial evidence to suggest that a broadly NNW-SSE extensional faulting event occurred in Late Burdigalian-Langhian time, prior to the deposition of the Geceleme Formation;
- This probably resulted in the formation of an asymmetrical half-graben in the north of the basin and a fault-generated depocentre in the south;
- Reactivated Early Miocene normal faults with obliquely (sinistral) orientated reverse slickensides, and reverse and strike-slip faults in Upper Miocene sediments suggest that a transpressional event occurred at some point before the Lower Pliocene.
- Higher angle normal faults than those seen in the Oymapinar limestone affect Upper Miocene sediments. Similarly orientated faults also affect Pliocene sediments. It is not clear however which of the three possible post-Miocene extensional faulting events may have generated these faults (C. Glover, pers. com., 1995).

7.5.1.3 Folding of Miocene sediments

Folding in the east of the Manavgat basin is dominated by the open-style, west verging anticline-syncline system shown in Figure 7.10. This exposes Alanya Massif basement at the eastern end of the core of the anticline. Akay and Uysal (1985) indicates that this folding affected all the Miocene strata equally (Fig. 7.14a) and that there was no previous faulting event. Reconstruction of his cross section to its pre-folding dimensions shows however, that this cannot have been the case. Only very pronounced and patchy subsidence with perhaps dramatic lateral variations in compaction could produce the thickness variations required if no faulting had occurred. Figure 7.14b shows the thickness variations of sections on the limbs of the folds, measured from Akay’s (1985) cross-section across the area illustrating the improbable depositional patterns of this model. The reef carbonates in this diagram were used as a horizontal marker on top of which the overlying thicknesses were constructed, because it is possible to assume that the reefs were deposited at or very near sea level, an effectively horizontal plane. The diachroneity of the
Figure 7.14  a) Cross-section across the Manavgat basin taken from Akay and Uysal (1985). This model assumes that no extensional faulting has occurred during deposition of the Miocene succession. b) Schematic reconstruction of the cross-section showing the improbable thickness distributions of the Upper Miocene sediments. It has been assumed that the reef horizon was effectively horizontal at the time of deposition (i.e. at or near the surface of the sea) indicating that the underlying conglomerates filled pre-existing palaeotopography. The diachronieity of the reefs from south to north has not been taken into account. Neither subsidence, nor compaction has been accounted for in the estimation of the thickness variations of the overlying succession, but it is not thought that these processes could entirely account for the distribution pattern seen.
reefs has not been taken into account and would tilt the diagram towards the south, but would not significantly affect the thickness patterns of the overlying sediments.

Although the thickness variations across the area require faulting of the Lower-Middle Miocene sediments, it is not clear what rôle these faults played during the phase of compression which folded the overlying strata. The largest population of faults collected in the Lower Miocene succession (Fig. 7.6) is found on the northern margin of the basin and indicates that the faults are orientated with a NW-SE strike, not directly parallel to the symmetry of the fold which strikes E-W. As discussed above, the extension direction deduced from the joint pattern on the southern limb of the anticline may only represent a snapshot of the distribution of stresses in one small geographical area at a fraction of the total time of faulting. It does however, provide a measure of the extension direction without direct reference to underlying weaknesses in the Alanya Massif, something which can never be precluded when considering larger faults, where the tips are not seen. The joints strike N-S, perpendicular to the fold (Fig. 7.11), but it can be assumed that if the orientation of the faults bounding the inferred horst-graben system was parallel to this, then the thickness variations shown in figure 7.12 would change from east to west rather than north to south. Thus, the constraints imposed by the geometry of the sediments require that at least a component of the fault strike orientation, was E-W and it seems reasonable to suggest that these were probably parallel to the micro-faults measured in both the north and central parts of the basin i.e. striking NW-SE.

How then would faults orientated NW-SE be affected by a N-S orientated direction of compression? It is highly likely that N-S compression would have reactivated the extensional faults with a oblique reverse motion. The sense of shear theoretically in this case, should be dextral (Fig. 7.15), but it is interesting to note that all the slickensides found on the faults at the Akseki road locality on the northern margin of the Manavgat basin are reverse with a sinistral component. One feasible explanation of this (Fig. 7.15) is that an east-west compressional event followed a north-south directed one (exactly the opposite scenario to that suggested by
Figure 7.15 The theoretical orientation of slickensides produced on a NW-SE striking fault when a) $\sigma_1$ is north-south and b) $\sigma_1$ is east-west. Full arrows indicate the slickenside direction likely.
Frizon de Lamotte et al., 1995). Equally probable however, is the possibility that the sinistral sense of shear is an expression of the dominance of an east-west component in a slightly north-south orientated $\sigma_1$ direction.

Reactivation of these NW-SE faults could well have produced the steeper dips associated with outcrop of the Oymapinar limestone particularly along the northern margin of the anticline producing the rather box-like shape of the fold. The relatively shallow dips of the Upper Miocene formations indicate gentle flexure over the top of this and it is possible that this fold structure, though prominent today, actually represents a very small amount of N-S compression. In other words, the folding of Upper Miocene sediments may be the expression of a small scale inversion structure of the underlying extensional faults (Fig. 7.16).

7.5.1.4 Interpretation and basin evolution

The evidence for the structural evolution of the Manavgat basin deduced from sedimentary and structural information is summarised in table 7.2.

It is not possible to preclude the possibility, that the fault along which the Alanya Massif was emplaced onto the Antalya Complex to the north of the Manavgat basin (Monod, 1977; Okay and Ozgül, 1984) was active during the Miocene and influenced the development of the structures (i.e. faulting and folding, see below) on its upper surface to the south. There is no evidence of foredeep-type sedimentation to the north of this fault however (Monod, 1977; Robertson, unpublished data), suggesting that, if active, it had minimal affect on Miocene sedimentation patterns.

The combination of sedimentary evidence in the form of fault-related facies and thickness variations across the basin with the observation of numerous micro-faults affecting the lower part of the Miocene succession has led to the suggestion that a distinct extensional faulting event occurred in the Manavgat basin at this time (section 3.8.1). This led to the generation of an asymmetrical half graben in the north of the basin and a further fault-generated deep to the south of the present-day southern limb of the anticline. The orientation of the extension direction, as
Figure 7.16 Schematic cross section across the Manavgat basin at the end of compressional event in the Latest Messinian-Early Pliocene. Lower Miocene extensional faults are suggested to have been reactivated, probably transpressionally. Overlying softer Geceleme and Karpuzçay Formations may not have been fully lithified, leading to more ductile behaviour. The rather box-fold like shape of the structure may have been inherited from the Lower Miocene horst-graben structure.
Table 7.2  Tabulation of the timing and tectonic events affecting the Manavgat basin. Evidence from the basin is also listed.

<table>
<thead>
<tr>
<th>Time</th>
<th>Evidence from the Manavgat basin</th>
<th>Structural result</th>
<th>Tectonic event</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pre-Early Miocene?</td>
<td>i) Relationship with Antalya Complex (Monod, 1977; Okay and Ozgül, 1985)</td>
<td></td>
<td>Emplacement of the Alanya Massif</td>
</tr>
<tr>
<td></td>
<td>i) Absence of material coming from the north, i.e. barrier to sediment from the Antalya Complex exposure to the north of the Alanya Massif</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>ii) No conglomerate underneath the reefs on the northern margin</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Burdigalian-</td>
<td>i) Faults with normal offsets in the reefs;</td>
<td></td>
<td>NNW-SSE extensional event</td>
</tr>
<tr>
<td>Langhian boundary</td>
<td>ii) Deposition of fault talus at Akseki road;</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>iii) Extensional joint formation prior to complete lithification near Alarahan;</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>i) Patchy distribution of Çakallar formation;</td>
<td></td>
<td>Asymmetrical half-graben in the north of the basin and fault-generated deep in the south.</td>
</tr>
<tr>
<td></td>
<td>ii) Thickness variations in the Geceleme formation;</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>iii) Steep sides of later fold limbs may reflect the presence of faults</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Late Messinian-</td>
<td>i) Transpressional slickensides at Akseki on older normal faults;</td>
<td>Reverse faulting</td>
<td>Aksu Phase of east-west compression (Poisson, 1977;)</td>
</tr>
<tr>
<td>Early Pliocene</td>
<td>i) Gentle folding of Miocene sediments with points of inflexion on the limbs concentrated along earlier graben faults;</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Fold generation</td>
<td></td>
<td>Susuz Dag Phase of north-south compression (Frizon de Lamotte et al., 1995)</td>
</tr>
<tr>
<td>Plio-Quaternary</td>
<td>i) high angle normal offset faults in Late Miocene sediments</td>
<td></td>
<td>Extension</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
deduced from the faults was NE-SW, approximately parallel to the strike of the boundary between the Alanya Massif and the Antalya Complex to the north. However, extensional jointing in sequences not directly influenced by the fabric of the Alanya Massif, suggest that in some places, at a particular time, the extension direction was closer to east-west. Although the joint orientation may have been controlled by local stress patterns over a relatively short period, it is possible that the overall NW-SE trend to the strike of the faults at this time relates to reactivation of older weaknesses in the Alanya Massif basement orientated in a similar direction and possibly generated during northward emplacement.

Figure 7.17 shows the subsidence curve for the Ahmetler section in the Manavgat basin. Fitted to it is a theoretical, uniform stretching subsidence curve with a stretching factor ($\beta$) = 1.08 (Condon, 1988). The data from the Manavgat basin appears to be consistent with a short pulse of extension between 16-17Ma followed by gentle thermal subsidence (J. Turner, pers. com. 1995).

The full discussion of the Aksu Phase and its relationship to the Susuz Dag Phase can be found in section 7.6.2 and 3. In the Manavgat basin however, the following evidence can be used in the discussion:

- Folding occurred after the deposition of the Miocene sequence;
- Transpressional reactivation of faults also occurred after deposition of the Miocene succession although some syn-sedimentary activity cannot be altogether ruled out;
- E-W striking folding indicates a component of north-south compression;
- Slickensides on reactivated extensional faults are sinistral, suggesting that the largest component of compressional stress is likely to have been orientated east-west;

In summary, there is no clear evidence from the data in the Manavgat basin alone that the Aksu Phase and the Susuz Dag Phase can be separated and chronologically ordered (section 7.3; Frizon de Lamotte, et al., 1995).
Figure 7.17  Subsidence curve for the Ahmetler section of the Manavgat basin with a theoretical (uniform stretching) subsidence curve ($\beta = 1.08$) fitted to it (Condon, 1988). Diagram constructed using software written by J. Turner (see Appendix 6).
7.5.2 The Köprü and Aksu basins

The structural evolution of these two basins is far more difficult to assess than that of the Manavgat basin. In part this is because the stratigraphy is less clear (conglomerates are in general difficult to date), but the main reason is that a continuous succession through from basement to Upper Miocene sediments is not exposed, as it is in the Manavgat basin. The two north-south orientated basins (Aksu and Köprü) are discussed together here, because it seems inefficient to deal with them separately, when, in all probability, they were responding to similarly orientated contemporaneous tectonic stress.

7.5.2.1 Basement and basal Miocene structures

The basement surrounding the Aksu and Köprü basins is heterogeneous in terms of its age and rock types. It can be divided into three broad formations: the Bey Daglari Mesozoic limestone, which forms most of the western margin of the Aksu basin; the Antalya Complex which consists of Triassic- Palaeocene deep-sea carbonates, radiolarites and basalts and forms most of the basement promontory dividing the Aksu and Köprü basins; and the Anamas-Akseki platform, which is dominantly Triassic shales and carbonate and forms the eastern border of the Köprü basin (Figs. 7.1b, and 7.2). It was beyond the scope of this project on the Miocene sediments to study in detail the structural features of the basement on which they are found. However, the satellite image of the area (Fig. 1.4) shows clearly the strong north-south grain of the fabric in the basement around the Miocene basins. The age of this grain is difficult to assess given the duration of time over which it may have occurred, but it is thought to have originated during Triassic rifting (Robertson and Woodcock, 1980; 1982; section 7.4). The study of the overlying Miocene sediments suggests that the prominent N-S grain in these younger rocks is the product of reactivation of older weaknesses (Flecker et al., 1995; this study, see below).

Miocene sediments directly overlying basement rocks are only exposed in a few places in the Aksu and Köprü basin area. At one of these, on the Mesozoic "promontory" which divides the two basins (Fig. 7.5) small
pockets of reddened, channelised and imbricated conglomerates directly overlie basement carbonate, basalt and radiolarite (Fig. 5.5, chapter 5 section 5.5.1.2). Dating of these thin sequences has not been attempted during this study, but Akay and Uysal (1985) identified them as Miocene because of their similarity to the Miocene continental successions seen elsewhere in the area (e.g. Kargi and Aspendos; Fig. 5.6). It is possible however, that these conglomerates are Pliocene in age.

Figure 7.18 is a geological map based on that drawn by Akay and Uysal (1985), centred on the southern part of the Mesozoic "promontory" between the Aksu and Köprü basins. Two small linear conglomerate-filled sub-basins can be seen overlying Mesozoic basement, both with a prominent west-verging north-south striking thrusts along their eastern margins. In detail, these Miocene exposures contain abundant micro-faults. The dominant normal-fault orientation is shown in figure 7.19a and strikes NE-SW, at variance with the orientation of both the micro-normal faults and the extensional joints in the Manavgat basin (see section 7.5.1.2 above). It is not possible to be absolutely certain that these faults are Miocene in age, but their association with the preservation of small sub-basins suggest that they may be related to their genesis.

Reverse faults within these Miocene sediments are also found, often in greater abundance than the normal faults. These reverse faults also show a greater scatter of orientations than the normal faults (Fig. 7.19b), but it is interesting to note how few of them are orientated north-south, parallel to the strike of the east-bounding fault. The dip of this bounding fault is difficult to estimate, but there is little evidence of any that the reverse faults in the study area are low angle and it is therefore assumed, in the absence of any contradictory evidence, that these bounding faults are also relatively high angle.

7.5.2.2 Basin bounding faults and faults within Miocene sediments

Large scale, north-south, high-angle faults dominate Miocene exposures in the Aksu and Köprü basins. This fault trend has previously been documented by Poisson (1977) and Akay and Uysal (1985). The basin margin faults (Kirkkavak Fault, Pinargözü fault zone, and Aksu Thrust
Figure 7.18  Geological map of the central region of the study area after Akay and Uysal (1985). The map is centred on the Mesozoic "promontory" which divides the Aksu from the Köprü basin. Note the development of small N-S orientated linear outcrops of conglomerate on top of the basement and the thrusts that mark their eastern boundaries. Characteristically, the strong N-S orientated fabric of the faults is untraceable throughout Pliocene or Quaternary sediments, indicating that the age of formation of these faults (the Aksu Phase, Poisson, 1977) is likely to have been at the Miocene-Pliocene boundary.
Figure 7.19 Stereonet projections of faults measured from the sub-basins on top of the Mesozoic "promontory" which divides the Aksu and Köprü basins: a) normal faults striking NE-SW; b) reverse faults with a variety of orientations, very few of which are north-south, parallel to the basin bounding thrust.
(Fig 7.5) are mirrored by faults within the basins which have smaller offsets (e.g. Kargi Fault).

The Kirkkavak fault on the eastern margin of the Köprü basin has intrigued geologists over several decades. Today it rises steeply from the basin floor, but as already discussed in chapter 4 (see section 4.6.2.4), palaeocurrent data indicate that during the Miocene it probably had little or no topographic expression, such that there is no evidence of it having been a source of material for turbidite deposition. Its lack of topography in the south of the Köprü basin is thought to be the cause of overfilling in this area. Dumont and Kerey (1975) studied the Kirkkavak fault and identified two periods of movement on it during the Miocene. The first was post-Burdigalian dextral strike-slip movement identified near Burmahan. This was followed by post-Tortonian east-west compression, resulting in high angle reverse movement. Active talus fans obscure much of the accessible face of the fault and its relationship with Miocene sediments, but these fans indicate that it has been reactivated more recently than the Late Miocene, probably as part of the Plio-Quaternary extensional events (Glover, 1995). The southern end of the Kirkkavak fault bifurcates close to the most northern tip of the Manavgat basin and then disappears underneath Miocene sediment close to the village of Çardak. Folding in the area south of the most southerly exposure of the Kirkkavak fault may well indicate that it has subsurface expression as a blind fault.

The eastern margins of both Aksu and Köprü basins are generally less clearly delineated than the western margins. The western margin of the Köprü basin for instance is faulted, but not as a single concentrated fault zone like the Kirkkavak fault. The western margin of the Aksu basin is largely unfaulted with sedimentary evidence suggesting a steep, but passive contact between the Miocene and Mesozoic (see chapter 5, section 5.6.4.2). The eastern margin of the Aksu basin however, like the Kirkkavak fault, is a well defined fault lineament striking north-south, which, like others parallel to it, disappear without trace under Pliocene sediment (Fig 7.18, north-western corner). A few kilometres further south, Antalya Complex rocks are overlain by a shallow-water carbonate called the Gebiz limestone (Akay and Uysal, 1985; Akay et al., 1985). The
The age of this limestone is crucial to the accurate termination age of the Aksu Phase of deformation, but it remains tantalisingly difficult to date and its age cannot be pinned down closer than Late Messinian to Early Pliocene (Poisson, pers. com. 1994).

Strike-slip faulting in both the Aksu and Köprü basins is common although the sense of movement on these faults varies from place to place. Figure 7.20a shows the distribution of poles to planes for strike-slip faults from Serravallian conglomerates (Poisson, 1977) near the village of Kargi in the Aksu basin and from the Karpuzçay Formation in the eastern half of the Köprü basin. There is a clear north-south trend in the orientation of the planes. The Kargi measurements come from a well delineated north-south striking fault zone, which though it appears to have little stratigraphic offset (Poisson pers. com., 1994) has 2-300m of topographic expression just to the east of the village of Kargi. Frizon de Lamotte et al. (1995) also collected fault data from this area with differing results (Fig 7.21). In contrast to the clear delineation of the Kargi fault zone, the Köprü basin measurements were taken along cross-sections from the eastern margin bounding Kirkkavak fault to the centre of the basin. The faults were therefore much more dispersed over an area outside the clearly defined Kirkkavak fault zone.

Figure 7.20b is a stereonet projection of poles to normal faults measured in the area to the west of the Kirkkavak fault. From the limited data available, these faults appear to have a bimodal distribution with one set of planes orientated approximately north-south parallel to the strike-slip faults, and another possibly more dominant set striking NE-SW. It is interesting to note the similarity in the strike of these normal faults and those seen affecting the pockets of Miocene conglomerate overlying the Mesozoic promontory displayed in figure 7.19a (section 7.5.2.1 above). The reverse faults also show this same bimodal pattern of distribution (Fig. 7.20c).

The Pinargözü fault zone which marks the western margin of the Köprü basin is, like the other fault zones in the area orientated north-south (Fig. 7.5). Measurements on faults within this zone also have a strong north-south trend (Fig. 7.21a, and b). From the generations of slickensides
Figure 7.20 Stereoplots of fault data from the Aksu and Köprü basins.

a) Strike slip faults with a dominantly north-south strike, parallel to the basin bounding faults and the structural grain of the basement; b) normal faults from the eastern side of the Köprü basin showing a bimodal distribution pattern, broadly north-south striking and a set striking NE-SW; c) Reverse faults and those whose sense of movement was not discerned also showing a similar distribution pattern to that of the normal faults in the Köprü basin.
Figure 7.21  a) Stereonet projections of fault plane data from Kargi and Pinargözü from Frizon de Lamotte et al. (1995); b) Stereonet projections from the Pinargözü fault zone collected during this study.
collected from these faults, Frizon de Lamotte et al. (1995) suggest that the compressional system forming these north-south faults can be divided into two distinct events, the first caused by east-west compression they called the Aksu Phase (Poisson, 1977), the second caused by north-south compression, they called the Susuz Dag Phase. Based on the small data set of sequential slickenside generations on a single fault plane, this study cannot conclusively confirm or disprove this hypothesis. However, given that the data set presented in their paper (Frizon de Lamotte et al., 1995) for the Pinargözü fault zone was itself small (n = 11, Aksu Phase; n = 8, Susuz Dag Phase) it is perhaps justified to mention that the data set generated during this study (n = 12) suggests that the final movement on faults with more than one generation of slickensides is almost always reverse. On north-south orientated faults this indicates that east-west compression is required. This observation does not preclude there having been a later south directed compressional event as most of the evidence comes from the Bey Daglari, but no evidence can be presented here to support this concept.

7.5.2.3 Folding of Miocene sediments

Frizon de Lamotte et al. (1995) suggest in their paper that the Miocene conglomerates found at the top of Bozburun Dag, the 2500m peak between Pinargözü and Ballibucak (Fig. 7.5) on the western margin of the Köprü basin is part of a west-vergent recumbent fold which also folds Mesozoic basement beneath Miocene cover (cross-section in figure 7.22b). Looking east from the Pinargözü pass, it is possible to observe the change in the dips of the Miocene conglomerate over the fold nose in agreement with this hypothesis (Fig. 7.23). The cross-section taken from the Frizon de Lamotte et al. (1995) paper is unfortunately difficult to correlate with the map from which it was supposedly drawn (Fig. 7.22). Note that to the west of Bozburun, the area of Bey Daglari carbonate is not shown on the cross-section and the missing fault from this area is assumed to be that which on the map divides this carbonate from the Antalya Complex further to the west.

Although these are relatively minor errors, even when drawn correctly, this cross-section does not, in detail fit the field evidence from the
Figure 7.22 a) map of the structural configuration in and around the Isparta Angle from Frizon de Lamotte et al. (1995); b) cross-section drawn from the map by Frizon de Lamotte et al.; c) cross-section correcting minor mistakes on cross-section b and adding the east-vergent high angle reverse fault which can be observed in the field.
Figure 7.23 Photograph of the nose of the Bozburun Dag (Fig. 7.5) recumbent fold looking east.
Figure 7.24  Photo-montage looking north from the Pinargözü pass showing the thick pile of conglomerates of the Bozburun Dag fold to the east jamming up against isoclinal folded conglomerates and marls (centre, foreground). In the west the Mesozoic carbonate interpreted as part of Bey Daglari carbonate platform has been thrust eastwards over the isoclinal folded conglomerates.
Pinargözü area. Figure 7.24 is a photomontage and interpretative sketch looking north from the Pinargözü pass, showing that between the thick pile of conglomerates making up the nose of the Bozburun Dag fold and the Bey Daglari carbonate in the west, conglomerates and marls have been deformed into upright isoclinal folds. The boundary between the Bey Daglari carbonates and conglomerates can be observed in a steep gully just beyond the bridge going up towards the pass from Pinargözü. It is clearly exposed as a high angle reverse fault, placing Mesozoic carbonate eastward on top of Miocene conglomerates.

The Bozburun Dag nappe is not the only large scale recumbent fold in the study area. In the south-west of the Köprü basin near Gökçepinar and Akbas (Fig. 7.5) an east verging recumbent fold with a core of Mesozoic limestone can be seen (Fig. 7.25). Akay and Uysal (1985) marked the boundary between this limestone and the Miocene sediments as a fault and it is quite possible that slip did occur along this horizon.

Smaller-scale folding similar to the isoclinal folds is seen elsewhere in the basin, generally associated with well defined fault zones like that at Pinargözü. Figure 7.26b indicates that the folds found in these situations have axial planes that strike parallel to the north-south trend of the fault zone. The turbidites in the eastern half of the Köprü basin are folded in a different style, generally much more openly and often with an overturned limb. The orientation of the axial planes of these folds is shown in figure 7.26a which suggests that they trend NW-SE.

7.5.2.4 Interpretation

Interpretation of the Aksu and Köprü basin structural data is much more difficult than that of the Manavgat basin primarily because the stratigraphy is far less well exposed and dated. It is not possible, for instance to clearly identify an Early Miocene extensional event like that discussed in section 7.5.1.2 above. Because of the almost complete absence of Pliocene sediment in all but the far south of the Aksu basin, only the large north-south trending faults can be seen to have a clear pre-Pliocene age. The micro-faults are almost impossible to date. However, assuming that a tectonic event affecting Miocene sediments in the Manavgat basin
Figure 7.25  Photo-montage and sketch of the east-vergent recumbent fold structure near Akbas, SW Köprü basin
Figure 7.26 Stereonet projections of fold axes from a) the eastern side of the Köprü basin in the Karpuzçay turbidites, and b) from close to well defined north-south striking fault zones at Pinargözü and Kargi.
is likely to have had some expression in the Aksu and Köprü successions it is possible to put together a tentative structural history.

There is no conclusive evidence to suggest that the micro-scale normal faults in the Aksu and Köprü basins and on top of the Mesozoic promontory which divides them are not Plio-Quaternary. However, if they did form during the Miocene, one possibility is that their dominant NE-SW orientation was generated during formation of a pull-apart basin in relation to strike-slip faulting along north-south lineaments such as the Kirkkavak fault. Dumont and Kerey (1975) suggest that the Kirkkavak fault moved dextrally in post-Burdigalian times and this would fit with a theoretical model of the Aksu and Köprü basin area behaving as a pull-apart system. Note that even outside the well defined fault zones most of the micro-scale strike-slip faults trend north-south, parallel to the basin bounding faults (Fig. 7.20a). These may or may not be connected to faults at depth generated by the earlier fabric of the basement and thus these faults may not represent failure in a stress field free from direct basement influence. They do however indicate that most of the strike-slip movement visible on faults post-Early Miocene is taken up on north-south striking fractures.

Dumont and Kerey (1975) suggest that the Post-Tortonian movement on the Kirkkavak fault resulted in high angle reverse faulting due to east-west compression (later named the Aksu Phase; Poisson, 1977). It has never been quite clear how much compression this event actually resulted in. Poisson et al. (1984) suggested that a minimum of 50% shortening had occurred and certainly the formation of large scale recumbent folds, such as the Bozburun Dag fold, would support significant shortening. However, most authors have viewed the results of the east-west compressional event purely in terms of westward transport (e.g. Poisson, 1977; Poisson et al., 1984). This study presents evidence that eastward transport in some areas was by no means insignificant leading to the formation of east-verging recumbent folds (e.g. at Akbas) and high angle east-directed reverse faults (at Pinargözü). This east-directed faulting may well be a back-thrust within a westward transporting system (Fig. 7.22c). The east-vergent fold is not so easily explained however and may suggest that rather than a simple westward
transport system, the Isparta Angle was being tightened by the movement of both arms. Palaeomagnetic studies of Miocene sediments on top of the Bey Daglari on the eastern margin of the Aksu basin indicate a Post-Langhian anticlockwise rotation of 30° (Kissel and Poisson, 1986). Kissel and Poisson (1987) suggested that this rotation was taken up on a décollement beneath the Neogene basins in order to explain the absence of a rotation signature in the post-Langhian sediments (Kissel and Poisson, 1987). It is possible, however, that some of the east-vergent compression resulting from this rotation might have produced the east-vergent structures observed in the field.

7.6 Tectonic context

7.6.1 Formation of the Miocene basins

7.6.1.1 Post-Langhian anticlockwise rotation of the Bey Daglari and extension in the Manavgat basin.

The orientation of the extensional faults in the Manavgat basin (NW-SE, Fig. 7.6) are, on the surface, an unlikely direct result of anticlockwise rotation of the Bey Daglari. The cause of this rotation however is perhaps more worth considering. Both Kissel and Poisson (1987), Kissel et al. (1993) and Morris and Robertson (1993) suggest that south-eastward movement of the Lycian Nappes played an important rôle in the recording of the palaeomagnetic signature in the western arm of the Isparta Angle either as the origin of orogenic fluids which caused widespread remagnetisation throughout the permeable rocks of the basement at this time (Morris and Robertson, 1993), and/or as a mechanism for inducing rotation (Fig. 7.3, Kissel et al., 1993). Can then the Lycian Nappes be viewed as an encroaching load causing flexure and faulting in the foreland much like the model put forward for the evolution of the Adana basin to the east (see below, Williams and Unlügenç, 1992; Gürbüz, 1993)? Once again the orientation of the faults in the Manavgat basin are difficult to reconcile with generation during a south-eastward transport of the Lycian Nappes. However, Flecker et al. (1995) suggested that rather than generating new faults related to flexure, Lycian loading of the lithosphere might have induced block faulting in
the foreland which exploited pre-existing lines of basement weakness. There are various pieces of evidence which support this hypothesis:

- There is a west to east diachroneity of basin formation (e.g. the Kas-Dariören flexural basin (Fig. 7.2) contains Aquitanian transgressive limestone; the thick pile of Lower Miocene conglomerates in the Aksu and Köprü basin is certainly Burdigalian in age, although it could be older; the Manavgat basin contains a thinner succession of probably Upper Burdigalian conglomerates. This diachroneity could have been caused by an encroaching nappe pile from the NW;
- The orientation of the extensional faults within the Manavgat basin mirrors the northerly bounding fault with the Antalya Complex suggesting that these faults may have reactivated older weaknesses within the Alanya Massif. This is supported by the evidence that extensional fracture (e.g. localised jointing), when unconnected to basement, is orientated at a different angle (section 7.5.1.2; Fig. 7.11);
- Evidence of recurrent reactivation of N-S faults in the Aksu and Köprü basins from Early Mesozoic to recent times (Robertson et al., 1991; Robertson, 1991).

Fault-defined blocks in the Isparta Angle area may have behaved in a similar style to the broken slat model (Taymaz et al., 1991; Fig. 7.27). This model was originally put forward to explain Quaternary N-S extension along the Aegean coast of western Turkey in the context of a westward moving Anatolia. The Isparta Angle lies along the transition zone where in the model Taymaz et al., (1991) indicate that pinning of the slats to the Anatolian block occurs. Taymaz et al., (1991) did not really consider deformation in this region, but their model predicts dextral strike-slip between individual ENE-WSW trending individual slats (Fig. 7.27). This is not observed today in this transition zone (Price and Scott, 1994). These authors proposed that the boundary between the slats and Anatolia was accommodated by a N-S trending deforming zone across which dextral displacement occurs and within which NE-SW trending fault blocks rotate clockwise (Fig. 7.28).

It is suggested that during the Early Miocene, a not dissimilar tectonic scenario could have been active. Figure 7.29 is a schematic representation of the region from Burdigalian to Tortonian times, when Lycian Nappe emplacement to the SE is thought to have caused the rotation of the Bey Daglari (Kissel et al., 1993; Morris and Robertson, 1993). Kissel and
Figure 7.27  Model of broken rotating slats. Filled circles are screws that attach the slats to the margins of plates 1 and 2. Open circles are the screws which join only the two arms of each broken slat (from Taymaz et al., 1991).

Figure 7.28  a) close up view of the dextral shear direction between slats and between the slat and the plate to which it is attached. b) Modification of the eastern margin of the model by Taymaz et al. (1991) to accommodate the structures observed in the Burdur region. The region of attachment is suggested to be an approximately N-S deforming zone across which dextral shear is taken up and in which NE-SW trending faults rotate.
Figure 7.29  Model of the development of extension in the study area during the Lower-Middle Miocene invoking the formation of a pull-apart basin driven by the rotation of the Bey Daglari and the emplacement of the Lycian Nappes.
Poisson (1987) suggested that this rotation was accommodated by a deep décollement at the base of the Antalya Complex in order to explain the absence of rotation in post-Burdigalian sediments. It is suggested however, that some of the rotation might have been taken up by dextral strike-slip along N-S orientated old lineaments. This strike-slip appears to have been focused within the Isparta Angle as a north-south deforming zone, like that envisaged by Price and Scott (1994) for the Quaternary. Within this dextral shear zone the formation of one or several, small, pull-apart basins potentially explains the formation of accommodation space (e.g. ~1.5km of Lower Miocene conglomerate (Kizildag Formation) in the Köprü basin) within a compressional regime at this time, something that the décollement hypothesis of Kissel and Poisson, (1987) fails to generate.

The Manavgat basin behaved differently from the Aksu and Köprü basins (Fig. 7.29 inset). It is suggested that the orientation of the NW-SE trending normal faults which developed at this time in the Manavgat basin (section 7.5.1.2) are controlled by basement weaknesses in the Alanya Massif trending parallel to its boundary with the Antalya Complex. Dextral shear on the eastern bounding fault of the N-S deforming zone (i.e. the Kirkkavak fault) might be expected to reactivate these. There is little evidence to suggest that the motion on these NW-SE faults was purely normal and significant oblique slip could have occurred whilst still producing the horst-graben structure deduced from the sediment facies and thickness variations.

7.6.1.2 Subduction related extension on the upper plate of the Hellenic-Cyprus system

Another possible cause for extension in the study area is tectonic activity along the Hellenic Arc south of Cyprus. The timing of initiation of northward subduction along this system has long been a matter for debate (e.g. Crete: McKenzie, 1978; Le Pichon and Angelier, 1979; Meulenkamp et al., 1988; Cyprus: Kempler and Ben Avraham, 1987; Eaton and Robertson, 1993). However, the agreement between most recent authors lends credence to an Oligo-Miocene time of initiation (Meulenkamp et al., 1988; Eaton and Robertson, 1993). Robertson and Woodcock (1986)
suggested that the Kyrenia Range in northern Cyprus subsided allowing the transgressive deposition of the Kithrea Flysch at this time. This subsidence can be explained (Robertson and Woodcock, 1986) by extension on the upper plate of a recently initiated northward-dipping (Jackson and McKenzie, 1984; Kempler and Ben Avraham, 1987) subduction zone to the south of Cyprus. The extension of the northern Cyprus Miocene basin across the Cilicia Basin (Fig. 7.1) to Adana is well established (e.g. Evans, 1978; Robertson and Woodcock, 1986; Ünlügenç et al., 1990) and it seems probable that the Antalya area was part of the same system. However, extensional faulting in the Manavgat basin occurs significantly later (e.g. 10 Ma later) than the collapse of the Kyrenia Range and workers in the Adana basin suggest that Early Miocene extensional faults formed in relation to load-induced flexure resulting from renewed thrusting in the Taurides to the north rather than due to supra-subduction zone extension.

Another possibility stems from a suggestion by Poisson (1984) that the Hellenic arc could have extended onto land along the front of the Lycian Nappe system (Fig. 7.30). In connecting with the Cyprean arc, this would place the Isparta Angle in the position of a transform fault along which dextral shear would be predicted (Fig 7.31). Within this context, (i.e. the close connection of the two subduction systems) the supra-subduction zone extension in the Cilician basin assumes a much more credible position in the formation of the Manavgat basin, which represents the northern margin of Cilicia during the Miocene.

7.6.2 Emplacement of the Lycian Nappes and the Aksu Phase

In their model, Kissel et al. (1993) suggest that the north-south Aksu thrust on the eastern margin of the Aksu basin, used by Poisson to date the phase of compression at the end of the Miocene (the Aksu Phase) was generated during the rotation of the Bey Daglari (Fig. 7.3). There is no clear evidence to suggest that compressional activity could not have begun significantly earlier that the Latest Miocene in some areas. In fact the thick successions of coarse grained conglomerates and sandstones deposited in the basin despite a rising sea level (section 5.10) during the whole of the Tortonian suggest that active uplift and erosion of the
Figure 7.30 Early Miocene plate tectonic reconstruction of the connection between the Aegean and Cyprus subduction systems along the Lycian Front (Pousson, 1984). (Diagram from Robertson and Grasso, in press.)
Figure 7.31  Model of the possible rôle of the Isparta Angle as a transform zone linking the Hellenic and Cyprus arcuate subduction systems during the Miocene. Dashed lines with open triangles on them indicate the location and orientation of these systems at the Oligo-Miocene boundary. See text for further explanation.
hinterland of the basins was occurring at this time (chapter 8 for full discussion), alongside subsidence due to pull-apart formation within the Isparta Angle (Fig. 7.29). The Aksu Phase has previously been envisaged as a product of the westward escape of Anatolia and it was for this reason that the dominance of west-vergent structures was explained. The presence of some quite large scale east-vergent systems (e.g. the Akbas syncline, Fig 7.25) supports the idea that the western arm of the Isparta Angle was not entirely passive at this time. The relative dominance of the west-vergent structures over the east-vergent structures may suggest the complex interplay between the initiation and strengthening rôle of westward escape of the Anatolian block during the final phases of Lycian emplacement to the SE and anticlockwise rotation of the Bey Daglari.

Also important, however, is the recognition that at least 30° of the curvature on the Lycian front would have been produced during the Miocene. Using the reconstruction of the pre-rotation front from either Poisson (1984; Fig. 7.30) or Kissel et al., (1993; Fig. 7.3) it seems reasonable to suggest that the thrust direction at the beginning of the Middle Miocene was south or SSE directed rather than the SE thrusting direction which might be deduced from the curvature of Lycian front today. The change in thrusting direction during the Miocene due to rotation of the Lycian front may therefore have been fundamental in changing the orientation of the stress direction acting on a N-S zone of faults within the Isparta Angle. A more southerly directed thrusting direction during the Middle Miocene would initiate strike-slip movement on a N-S suture, while a larger component of eastward compression, likely to have occurred towards the end of the Miocene, would result in less transpression and more orthogonal compression along N-S orientated faults.

7.6.3 Evidence for a later Susuz Dag Phase of southward transport

Although there is clear evidence that a component of the compression at the end of the Miocene was orientated towards the south, none of the data presented here gives unqualified support to the idea that this southward-directed transport event was separate from the east-west compressional event as envisaged by Frizon de Lamotte et al. (1995). The
folding of Upper Miocene sediments in the Manavgat basin is the clearest indication that north-south compression was important at this time, but it is suggested here that this deformation may represent very minor shortening over an already faulted Lower and Middle Miocene succession. The N-S structural grain of the Köprü and Aksu basins may perhaps obscure any E-W orientated structures present in the Miocene succession. Equally, it is possible that southward transport was taken up along a décollement beneath the basins parallel to the Florence Rise as suggested by Frizon de Lamotte et al., (1995) although more recent investigations of the Florence Rise suggest that a simple thrust-front interpretation is problematic. Southward transport along north-south fault zones such as that at Pinargözü has been measured, but the data set did not indicate that the southward movement was either separate from, or later than east-west compressional structures. It is suggested rather that the combination of lessening Lycian movement towards the south-east and strengthening westward movement due to the initiation of expulsion of the Anatolian block (Sengör et al., 1985) resulted yet again in the partitioning of obliquely orientated stress directions causing the reactivation of older weaknesses (Fig. 7.32).

7.7 Conclusions

♦ There is clear evidence of a previously unidentified extensional faulting event in the Late Burdigalian-Langhian which resulted in the development of a horst-graben system in the Manavgat basin;

♦ A theoretical (uniform stretching) subsidence curve fitted to the data in the Ahmetler section suggests that a short period of low-magnitude ($\beta = 1.08$) extensional rifting occurred between 16-17Ma, close to the Burdigalian-Langhian boundary;

♦ Subsequent reactivation of extensional faults in a transpressional (sinistral) sense may have been caused by a dominantly east-west compressional system.

♦ A component of north-south compression must have affected the area in post-Miocene times to form the open folds in the east of the
Upper Miocene-Early Pliocene

Figure 7.32 Schematic model of the development of the Late Miocene-Pliocene deformation. Because of the close proximity of the Lycian Nappes, further emplacement tectonics is likely to have affected the study area. Westward expulsion of Anatolia is also thought to have had a more regional effect. All deformation however, is accommodated along older lineaments.

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Manavgat basin. It is not thought however, that very much compression is required to bring about this folding, particularly if reactivation of underlying faults also occurred at this time.

- A model has been put forward for the evolution of the Miocene basins suggesting that:
  - Rotation of the Bey Daglari was taken up by strike-slip motion on N-S orientated faults during Mid-Late Miocene times rather than on a décollement horizon at the base of the Antalya Nappes;
  - Narrow, deep (~1.5km) sediment depocentres in the Aksu and Köprü basins areas were created by pull-apart formation in a north-south orientated deforming zone undergoing dextral shear;
  - Dominantly west-vergent compressional structures were formed due to initiation of westward transport related to the expulsion of the Anatolian block at the end of the Miocene;
  - More rare eastward-vergent structures may suggest that Lycian emplacement and rotation of the Bey Daglari were still active at the end of the Miocene.

- There is no clear evidence for a separate, north-south compressional event (the Susuz Dag Phase) after the Aksu Phase at the end of the Miocene in the Miocene sediments of the Aksu, Köprü and Manavgat basins.
8.1 Introduction

The Miocene sedimentary rocks of the Aksu, Köprü and Manavgat basins have been described in preceding chapters. The reader is referred to table 8.1, which is a summary chart of the information derived from this study. Chapter 7 examined the structural evidence for the initiation and development of basins within the Isparta Angle during the Miocene and placed them in a tentative tectonic context of Eastern Mediterranean evolution. This chapter compares the evolution of the Isparta Angle with other Eastern Mediterranean Miocene basins and then examines the relative roles of the fundamental processes of tectonics and eustacy in basin development and the controls on their sedimentary fill and architecture.

8.2 Regional comparisons

8.2.1 Kas basin, outboard of the Lycian Nappes, SW Turkey

A Miocene stratigraphy is preserved to the west of the Aksu basin mainly in a narrow NE-SW orientated basin along the Lycian thrust front which runs from Isparta in the north (Dariören basin) to Kas on the south Turkish coast (Figs. 7.1 and 7.2). A transgressive shallow-water limestone (the Karabayır Formation, Poisson and Poignant, 1974) of Aquitanian age is known also as "la barre Aquitanien" because of its strongly linear outcrop pattern along the south-eastern margin of the basin. It is overlain by a thick succession of Burdigalian-Langhian turbidites (Hayward, 1982; Gutnic et al., 1979), which are themselves overlain by Serravallian-Tortonian fan-delta conglomerates in the south (Hayward, 1982). Hayward (1982) interpreted the development of the Dariören-Kas basin succession as being controlled by movement of the Lycian Nappes located directly to the NW. He suggested that the basin formed by flexural
Table 8.1  Summary of sedimentation information from previous chapters.

<table>
<thead>
<tr>
<th>Age</th>
<th>Formation</th>
<th>Köprü and Aksu facies</th>
<th>Manavgat facies</th>
<th>Implications</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tortonian-Messianian</td>
<td>Aksu Formation</td>
<td>Continental, shoreline and shelf conglomerates, sandstones and patch reefs</td>
<td></td>
<td>Relative rising sea level</td>
</tr>
<tr>
<td></td>
<td>Karpuzçay formation</td>
<td>Corridor of very coarse sub-marine debris flows containing large detached blocks. Background sedimentation of sandy turbidites and planktic foraminiferal marls</td>
<td></td>
<td>Shallowing of the basin due to infill. Sedimentation rate increases dramatically here with no evidence of increased subsidence.</td>
</tr>
<tr>
<td>Serravallian-Tortonian</td>
<td>Sand and silt turbidites.</td>
<td></td>
<td>Base of formation abrupt marking a rapid relative sea level rise</td>
<td></td>
</tr>
<tr>
<td>Langhian-Serravallian</td>
<td>Geceleme formation</td>
<td>Planktic foraminiferal marls</td>
<td>Low energy environment of deposition in relatively deep water. Base marks rapid sea level rise.</td>
<td></td>
</tr>
<tr>
<td>Langhian</td>
<td>Çakallar formation</td>
<td>Calcarenites containing shallow-water debris and locally derived basement</td>
<td></td>
<td>Likely to have been fault generated</td>
</tr>
<tr>
<td>Burdigalian-Langhian</td>
<td>Oymapinar limestone</td>
<td>Patch reef facies often associated with conglomerates</td>
<td>Patch and fringing reef facies</td>
<td>Diachronous transgression from south to north visible in Manavgat basin</td>
</tr>
<tr>
<td>?-Burdigalian</td>
<td>Kizildag Formation</td>
<td>Continental-coastal conglomerates</td>
<td>Continental-coastal conglomerates</td>
<td>Relative sea level rise</td>
</tr>
</tbody>
</table>
loading of the lithosphere (Fig. 8.1), allowing Aquitanian transgression followed by passive turbiditic fill. The subsequent transition to fan-delta sedimentation indicates further advance of the nappes (Hayward, 1982). This also led to deformation of the Lower Miocene sequences, particularly clearly visible in the north of the Aksu basin (Gutnic et al., 1987). Hayward (1982) suggested that this later advance of the Lycian Nappes occurred at the Langhian-Serravallian boundary and this is in good agreement with palaeomagnetic evidence (Morris and Robertson, 1993; section 7.4). The time also coincides with the abrupt rapid relative sea level rise, thought to have been controlled by extensional faulting in the Manavgat basin and in the south of the Aksu and Köprü basins.

On the south eastern margin of the Bey Daglari (Fig. 7.2) a small Lower Miocene succession is preserved. This documents regional subsidence of the Bey Daglari in the Oligocene tentatively linked to initial emplacement of the Lycian Nappes to the NW (Hayward and Robertson, 1982). During the Early Miocene, the Antalya Complex to the east formed an elevated landmass from which sediments were shed westwards into a relatively deep water (~500m), narrow, north-south trending basin on the eastern flank of the Bey Daglari (Fig. 8.2; Hayward, 1982; Hayward and Robertson, 1982). The Antalya Complex landmass was rapidly eroded with rivers supplying coarse conglomerates to fan-deltas, which in turn fed a series of small submarine fans. At the same time, fault dissection of the Bey Daglari produced considerable volumes of carbonate clastics (Hayward and Robertson, 1982). The boundary between the Antalya Complex and the Bey Daglari was active during Eocene times when the Antalya Complex underwent extensive strike-slip faulting (Robertson and Woodcock, 1980a). In the Late Miocene sedimentation was abruptly terminated by a phase of westward thrusting of the Antalya Complex into its present position. During this deformation, the Antalya Complex, overrode earlier Miocene sediments forming a basal tectonic melange (Hayward and Robertson, 1982).

Synchronous deepening in the Aksu-Köprü-Manavgat basins with shallowing in the basins on the Bey Daglari can be explained in part by the behaviour of the Lycian Nappes. Hayward tentatively linked changes in sedimentation to periods of emplacement of the nappes (e.g. in the
Figure 8.1 Model of the Miocene emplacement of the Lycian Nappes with respect to the Dariören-Kas flexural basin on the west flank of the Bey Daglari (Hayward, 1982).
Figure 8.2 Palaeogeographic model of the Lower Miocene sedimentary basins on top of the Bey Daglari with the Antalya Complex to the east and the Lycian Nappes to the west (Hayward, 1982)
Oligocene and Late Miocene). It is possible that loading of the lithosphere induced block faulting in the Aksu-Köprü-Manavgat basins and allowed the accumulation of the thick succession of Burdigalian (or older) conglomerates (~1.5km of Kizildag Formation). Some of the east-vergent structures in the study area (chapter 7; section 7.5.2.3) may be attributed to Late Miocene movement on the Lycian Nappes and this may also have contributed to the uplift and erosion which formed the Mio-Pliocene unconformity. It is not clear how, or if the Mid-Miocene eastward thrusting of the Antalya Complex onto the Bey Daglari affected sedimentation in the Aksu-Köprü-Manavgat area. Note however, that it coincides with a period of uplift which causes shallowing in the Manavgat basin and a transition from turbiditic sedimentation to continental conglomerates and shallow-marine reef carbonates in the Aksu and Köprü basins.

8.2.2 Adana basin, north-east corner of Mediterranean, SE Turkey

A similar stratigraphy to that in the Lower-Middle Miocene of the Manavgat basin developed 400km to the east in the Adana basin at the same time. Gürbüz (1993) has shown that transgressive patch reefs characterise the Aquitanian-Burdigalian, slightly earlier than the south to north transgression seen in the Manavgat basin (Burdigalian-Langhian; Fig. 8.3). Williams and Unlügenc (1992) suggest that the marine transgression leading to the formation of patch reefs in the Adana basin was due to extensional faulting, in response to load-induced flexure resulting from thrusting in the Taurides to the north. In the Isparta Angle by contrast, there is very little evidence to suggest tectonic activity within the basin at the time of transgression although there is clear evidence for strong relative sea level rise. It is difficult to exclude the possibility that the Aquitanian-Burdigalian extensional event in the Adana basin masks a regional sea level rise at a slightly later date correlating with that seen in the Antalya area to the west. However, further clarification from the Adana and Mut (situated between Antalya and Adana) basins and in off-shore sediments along the south Turkish coast, would be required to assert this hypothesis more strongly.
Figure 8.3 Stratigraphy, sea level and tectonic interpretation of the Miocene sediments of the Adana basin (after Gürbüz, 1993; G. Williams pers. com., 1993).
Shallow-water carbonates in the Adana basin were drowned in Late Burdigalian-Early Langhian times, causing a sharp transition to deep-marine turbidites in the Adana basin, as seen in the Manavgat basin and in the south of the Aksu and Köprü basins. This rapid relative sea level rise in the Adana area is attributed to tectonically induced subsidence forming a deep, underfilled, flexural foreland basin caused by continued thrusting in the Taurides to the north (Williams and Unlügenç, 1992, Gürbüz, 1993). A serious difficulty arises in the direct application of this model to the evolution of the Isparta Angle, in that there is no evidence of thrusting of the Taurides in the area at this time. Flecker et al. (1995) suggested that in a broader sense however, this model may be relevant to the development of the Isparta Angle in the Miocene. As described in section 7.6.1, the Lycian Nappes on the western limb of the Isparta Angle are thought to have been active at this time (Kissel and Poisson, 1987; Kissel et al., 1993; Hayward, 1980; Morris and Robertson, 1993). Flexural loading from the north-east may therefore have induced block faulting of the foreland which, in turn, exploited pre-existing lines of basement weakness and resulted in rapid relative sea level rise. It is not clear if thrusting in the Taurides north of Adana is genetically related to thrusting of the Lycian Nappes to the north-west of the Isparta Angle, within a regional tectonic context.

Seismic reflection profiles across the Adana basin have been interpreted to indicate that little or no syn-sedimentary faulting occurred during subsequent infilling and shoaling during Langhian and Serravallian times (Williams and Ünlügenç, 1992). This shallowing ultimately led to the deposition of deltaic sequences in the Tortonian and Messinian in the Adana basin (Fig. 8.3). Shallowing in the Manavgat basin, in contrast, was not sufficient to allow deposition of anything other than fully marine sediments. The apparently passive post-rift fill of the Adana sequence is however, mirrored by the deposition of fine grained Karpuzçay Formation in the Aksu and Köprü basins and the planktic foraminiferal marls of the Geceleme Formation in the Manavgat basin.

The Misis Complex to the east of the Adana basin (Fig. 8.4) is an accumulation of Late Oligocene to Early Miocene olistostromes which were emplaced onto Early Miocene-Pliocene sediments (Kelling et
Figure 8.4  Map of Cyprus and the surrounding area showing tectonic units referred to in the text.
A combination of thrusting and strike-slip movement, not dissimilar from the emplacement tectonics of the Antalya Complex (Robertson and Woodcock, 1980a) is envisaged for the Misis Complex during the Late Miocene to Early Pliocene deformation (Kelling et al., 1987). To the north-east, close to the junction between the East Anatolian Fault and the Bitlis thrust (Fig. 8.4), a flexural (Çüngüş-Lice) basin formed during the Miocene (Aktas and Robertson, 1985). Kelling et al. (1987) suggest that overall transpression characterises the Adana area (i.e. the north-eastern segment of the Bitlis-Cyprus sector; Fig. 8.4) during the Neogene-Holocene, whilst, transtension affected the Cyprus area to the south-west. The oblique orientation of the tectonic stress thought to be active in this system accord well with observations of both transpression and transtension in the Isparta Angle area at this time (chapter 7).

8.2.3 Northern Cyprus

In northern Cyprus, patchy lacustrine limestone and lime mud (Flecker et al., 1995) and Upper Oligocene conglomerates are exposed adjacent to the Kyrenia Range (Fig. 8.4), interpreted as a palaeo-subduction zone (Robertson and Woodcock, 1986). These were the first sediments to accumulate after a phase of thrusting (Baroz, 1979). The conglomerates have been interpreted as alluvial fans derived from the future Adana basin to the north-east, that prograded across a broad alluvial plain, now covered by the sea dividing Cyprus from Turkey, the Cilicia basin (Robertson and Woodcock, 1986). A thick succession of turbidites, known as the Kithrea flysch, overlies these conglomerates and palaeocurrent directions also indicate that these were derived from the Adana basin to the north-east (Robertson and Woodcock, 1986; Kelling et al., 1987). The alluvial plain linking Turkey and Cyprus is thought to have collapsed around the Oligo-Miocene boundary leading to a rapid transition to marine turbidites. Dewey et al. (1986) interpreted the Adana-Cilicia basin as a triple-junction incompatibility feature generated by westward expulsion of Turkey (Sengör et al., 1985). Robertson and Woodcock (1986) however, interpreted the cessation of thrusting and collapse of the Kyrenia Range as being due to a jump in the location of subduction from the Kyrenia lineament to near the present-day Cyprus trench, located just
south of Cyprus. Relative to this new, northward-dipping subduction zone, the Kyrenia Range is situated on the upper plate of the subduction zone and thus the subsidence seen in northern Cyprus and the Cilicia basin could be explained by supra-subduction zone extension (Robertson and Woodcock 1986). Anastasakis and Kelling (1991) however, suggest that initiation of the subduction zone south of Cyprus did not occur until the Mid-Miocene. These authors also note evidence of extensional block faulting which is thought to have occurred after the Mid-Miocene thrusting of this area.

The relevance of supra-subduction zone extension in generating subsidence in the Isparta Angle generally and more particularly the Manavgat basin, which was the northern margin of the Cilicia basin during the Miocene has been touched on briefly in chapter 7 (section 7.6.1) is discussed further below in the light of information from Crete (section 8.2.5). Further clarification of the age of faulting of the Misis-Kyrenia thrust belt north-west of Cyprus is important in understanding whether the north and south margins of the Cilicia basin were acting in tandem or were subject to more local tectonic stress patterns.

8.2.4 Southern Cyprus

A significantly different stratigraphy is preserved in southern Cyprus where Oligocene (and older) deep-water pelagic chalks and marls of the Lefkara Formation are overlain by more localised, shallower-water, Miocene pelagic chalks, marls, calcarenites and conglomerates of the Pakhna Formation (Robertson et al., 1991). Eaton and Robertson (1993) argue that this change at the Oligo-Miocene boundary, was caused by uplift related to the initiation of subduction along the Cyprus trench, following the termination of thrusting along the Kyrenia Range to the north.

In addition, reefs developed in the Burdigalian and Tortonian (Follows and Robertson 1990; Follows, 1992; Eaton and Robertson, 1993) in southern Cyprus with a similar Mid-Miocene hiatus of reef growth to that of the Isparta Angle area (section 3.8.1). Follows (1990) tentatively
suggested that this was a result of eustatic sea level rise. Although eustacy may contribute a component of the relative sea level rise associated with drowning of Lower Miocene reefs in the Isparta Angle, the evidence of fault related subsidence at this time is too strong to allow the suggestion that eustatic processes may have completely controlled vertical facies change.

Payne and Robertson (1995) also suggest that the cause of extensional faulting and deepening of the Polis graben, Western Cyprus during the Tortonian might be due to "roll-back" of the Cyprus trench. "Roll-back" (i.e. the downward and backward migration of the subducting slab, which can result in extension in the upper plate; Dewey et al., 1980) as a process controlling extensional faulting and subsidence on the upper plate is discussed further in the light of evidence from Crete (8.2.5).

8.2.5 Crete, north of the Hellenic trench

Middle to Late Miocene basins on the island of Crete have been linked to activity along the Hellenic trench for over 20 years. Early work on Cretan exposures and on the mantle structure beneath the island suggested that basin generation occurred as a result of initiation of subduction along the Hellenic arc (e.g. McKenzie, 1978; Le Pichon and Angelier, 1979). More recently however, Meulenkamp et al. (1988) suggested that the initiation of subduction occurred approximately 26Ma ago, at the Oligo-Miocene boundary, over 10 million years earlier than the age of the oldest sediments in the basins on Crete itself. They suggested therefore, that Late Serravallian to Early Tortonian inception of "roll-back" within the South Aegean subduction system changed the tectonic stress regime and caused the fragmentation of Crete into a series of transtensional basins. In the Late Miocene and continuing throughout the Pliocene-Quaternary, uplift of the Cretan area occurred. This is thought to be related to outward gravity collapse away from the locus of crustal thinning in the Sea of Crete (Meulenkamp et al., 1994).

As argued by Flecker et al. (1995) and in section 7.6.1 the agreement between Meulenkamp et al. (1988) and Eaton and Robertson (1993) in Crete and Cyprus, respectively, lends credence to an Oligo-Miocene time
of initiation of subduction of the South Aegean and Cyprus segments of
the active margin, although evidence exists for the initiation of the NW
Hellenic segment of the trench around 6Ma (Underhill, 1989). The
formation of the Cilicia basin has been credited to supra-subduction
extension relating to the initiation of the active margin south of Cyprus.
However, the transgression and rapid relative sea level rise in the
Burdigalian and Langhian of the Isparta Angle, the northern margin of
the Cilicia basin, post-dates this initiation event possibly by as much as 10
million years.

Another possibility suggested by Flecker et al. (1995), however, is that the
previously subducted slab underneath the Kyrenia lineament, located by
earthquake foci beneath northern Cyprus (Kempler and Ben-Avraham,
1987) may also have had a rôle. Once the compressional forces of the
Africa-Eurasia collisional system have been relieved in the Kyrenia area
by active subduction to the south of Cyprus, a slab lying beneath, but still
attached along the Kyrenia lineament, might have been gravitationally
unstable causing "roll-back" and extension of the upper plate. The 10
million year diachroneity between initiation of extension south and
north of the Cilicia basin remains a problem however.

Equally, the timing of the inferred inception of "roll-back" for both the
Cyprus (Payne and Robertson, 1995) and South Aegean (Meulenkamp et
al., 1988) sections of the trench system, causing extension-related
subsidence, appears to coincide around the Serravallian-Tortonian
boundary, synchronous with the input of coarse clastic material into the
Manavgat basin, the initiation of continental and coastal deposition in
the north of the Aksu and Köprü basins and the sub-deltaic sequence
boundary in the Adana basin (Williams et al., 1993). It is possible that the
uplift to the north of the Isparta Angle, postulated as having caused the
change in depositional mode in southern Turkey, could have resulted
from marginal uplift related to the initiation of "roll-back". If this was
the case however, "roll-back" on the Cyprus trench would have been
behaving effectively synchronously across Cyprus and the Cilicia basin, in
contrast with the diachroneity across the Cilicia basin inferred for
extension related to the sinking of a slab beneath the Kyrenia Range 5-15
million years earlier.
Evidence to support marginal uplift around the Isparta Angle during the Burdigalian-Langhian extensional event is easily found in the form of thick (i.e. up to 1.5km) Burdigalian conglomerate successions. This strongly suggests hinterland uplift and erosion as well as the generation of accommodation space. These deposits have previously been attributed to uplift of the hinterland caused by flexural upwarp of the foreland outboard of the advancing Lycian Nappes.

How likely is "roll-back"-generated extension to have affected the Isparta Angle and have caused the 10Ma diachroneity from south to north? Within such an asymmetrical system as a subduction zone, it is perhaps unreasonable to expect related extensional stress to cause symmetrical rifting. There is little evidence however, of the character of the Pre-Messinian subsidence of the Cilicia basin (Evans et al., 1978). Symmetrical extension-related subsidence is inherently difficult to fit with a picture of diachronous subsidence away from the subduction zone. Asymmetrical stretching models appear more relevant, but are also difficult to apply simply in a meaningful way. Figure 8.5 is a grossly simplified model of the generation of the Cilicia basin due to "roll-back" on the Kyrenia slab. In an asymmetrical model where the effects of extension migrate out from the point of initiation, the application of average rates of "roll-back" from the present day Aegean system (Chase, 1978; Minster and Jordan, 1978) and probable angle of slab descent (Jarrard, 1986) suggest that the effects of extension might be expected to be seen in the south of the Isparta Angle around 10Ma after initiation. Note however, that this system cannot be treated meaningfully in this simple way, as there is no clear information on the type of rifting active at this time. For instance, the Kyrenia lineament itself is thought to have been arcuate (Fig. 8.2) so that alterations in the stress patterns along its length would be expected. Also, heterogeneities in the lithosphere around the Isparta Angle itself are clearly fundamental in controlling the structural development of the area and without considerably more knowledge of the basement beneath the Cilicia basin it would be unwise to go any further than pointing out that "roll-back" on the Kyrenia suture zone may have played a part in evolution of the Isparta Angle.
Assumptions: slab remains attached
*dip of slab is 45°

Slab sinking rate** approximately 1cm/yr

\[=\quad 10\text{m}/1000\text{yrs}\]
\[=\quad 100\text{km}/10\text{Ma}\]

Sedimentation rate of Kithrea Flysch approximately 2km/20Ma

\[=\quad 10\text{cm}/1000\text{yrs}\]

Rate of sinking > Sedimentation rate = UNDERFILLED BASIN

As horizontal (x) and vertical (y) displacements are equal, this model predicts subsidence 100km away occurs 10Ma after initiation of "roll-back"

*Angle of descent of slabs undergoing "roll-back" is between the angle of the slab dip and vertical (Jarrard, 1986). This is in agreement with predictive models of "roll-back" of Dewey (1980) and Molnar and Atwater (1978). Slab of present Aegean subduction system dips at 20-40°.

**Rate of horizontal component of "roll-back" of the Aegean system today is between 2cm/yr (Chase, 1978) and 2.9cm/yr (Minster and Jordan, 1978). Other subductions thought to be undergoing “roll-back” generally have a smaller rate (Jarrard, 1986).

Figure 8.5 Tentative assessment of the ability of "roll-back" on the Kyrenia lineament to create subsidence in the Cilicia basin area.
8.3 The relative importance of tectonics (local and regional) versus eustacy in controlling sedimentation

Figure 8.6 shows the sea level curves generated by sedimentation in the Manavgat and a combined curve for the Aksu and Köprü basins taking sedimentation rate into account, but assuming no tectonic activity. These can be compared with the sea level curve for the Adana basin (Gürbüz, 1993) and a composite curve for Cyprus (Robertson et al., 1991) and the eustatic sea level curve generated by Haq et al. (1988) in figure 8.7. In very broad terms the following points concerning the similarities and discrepancies between the relative sea level curves and the eustatic curve can be discerned.

**Similarities**

- Burdigalian-Langhian transgression in the Isparta Angle, Cyprus and Adana correlates with the eustatic curve. The Langhian is recognised as a period of high sea level throughout the Eastern Mediterranean (e.g. Cyprus, Follows, 1990; Sardinia, Cocozza and Javobacci, 1975);
- The shape of the eustatic curve suggesting a regression throughout the Serravallian fits with foraminiferal evidence from the Manavgat basin, but the shallowing deduced from this data could be entirely accounted for by infill at the calculated sedimentation rates. This has been taken into account in the drawing of the curves for both the Manavgat and combined Aksu and Köprü basins and results in the curve appearing flat at this time. Gürbüz took sedimentary fill into account when drawing the curve for the Adana basin; Robertson et al., (1991) did not;
- Evidence for relative sea level rise during the Tortonian in the north of the Köprü and Aksu basins correlates with a eustatic rise in sea level. There is also some evidence of a short period of transgression at the beginning of the Tortonian in the Adana basin and in northern Cyprus;

**Differences**

- Oscillation of the eustatic curve during the Langhian cannot be recognised in the sedimentary record of the Isparta Angle. This may be due to unobserved hiatuses in the record;
- The eustatic curve does not indicate the major relative sea level rise deduced for the Isparta Angle, Adana and Cyprus areas around the Langhian-Serravallian boundary.
Figure 8.6 Relative sea level curves for the Manavgat basin and a composite for the Köprü and Aksu basins derived from separate successions in the north and south of the area. The curves have been drawn taking into account the sedimentation rate, the subsidence induced by the sediment pile and the compaction associated. The curve represents changes in relative sea level which cannot be accounted for by these processes.
Figure 8.7  Relative sea level curves for the Manavgat basin and a composite for the Köprü and Aksu basins compared with the curves for the Adana basin (Gürbüz, 1993) and Cyprus (Robertson et al., 1991) and the eustatic sea level curve of Haq et al. (1988). The curves are not drawn to the same horizontal scale. The horizontal lines on the eustatic curve indicate the position of stage boundaries. Not all of these are in agreement with those defined by Berggren et al. (1985; see Figure 2.2).
In the Manavgat basin, foraminiferal evidence suggests that shallowing of the basin continued throughout the Serravallian, Tortonian and Messinian. A similar trend can be seen both in the Adana basin and in Cyprus, but it contrasts with evidence from the northern parts of the Aksu and Köprü basins and the eustatic curve;

The erosion associated with the Mio-Pliocene boundary is not associated with a regression on the eustatic curve. The Messinian drawdown (Hsü et al., 1973) of the Mediterranean is well documented throughout the region and resulted from the isolation of the Mediterranean from the Atlantic.

Only the Lower Miocene transgression and the Serravallian regression can therefore be directly and consistently linked with the eustatic curve produced by Haq et al. (1988). Where a direct correlation cannot be made, there is, in all cases, evidence of local tectonic activity (e.g. the Burdigalian-Langhian faulting event; the Tortonian dissimilarity of sea level curves between north and south of the Isparta Angle suggesting independent movement of small blocks).

### 8.3.1 Closure of Neotethys

The Mediterranean Sea is the last remaining remnant of a Neotethyan Ocean which opened behind rifted fragments of Africa and caused the closure of Paleaotethys during the Late Palaeozoic and Early Mesozoic (e.g. Sengör et al., 1984; Robertson and Dixon, 1984; Üstaömer and Robertson, 1994; Pickett, 1994). Subsequent northward movement of Africa relative to Eurasia resulted in near complete closure of Neotethys. Miocene sediments throughout the Mediterranean document a crucial period in this closure, from a narrow ocean fully integrated with global circulation to a marginal basin periodically closed at both ends. This closure has implications not only for the tectonic style of oceanic subduction and continental collision, but also for the palaeoceanographic regime. In terms of the marine sediments preserved in the Isparta Angle, the dwindling coral diversity supports similar evidence from elsewhere in the Mediterranean (e.g. Cyprus, (Follows, 1989; Spain, Martin et al., 1989; Southern France, Chevalier, 1961) which documents the termination of the seaway connecting it to the Indian Ocean (chapter 3; Chevalier, 1977) as Arabian Margin collided with the Eurasian plate (Angelier et al., 1982).
The development of a mud-dominated accretionary prism (Mediterranean Ridge; Limanov and Ivanov, 1994), active arc volcanism and back-arc extension in the Aegean area contrasts with the limited subduction which occurred along the Cyprus segment (see Robertson and Grasso, in press for review). The Isparta Angle lies at the boundary between these two systems and may well have played a vital rôle in accommodating the tectonic disparity between the two systems (Fig 7.31). The marked change in convergence direction of Africa and Eurasia around 9Ma influenced the kinematics of the tectonic systems in both the Eastern and Western Mediterranean basins (Dewey et al., 1989). The Tortonian uplift of the Isparta Angle and the deformation which characterised the Latest Messinian (Aksu Phase) can, and should be viewed within this context taking into account the effects of other, more local units, also reacting to this change in tectonic stress.

As far as the applicability of the eustatic sea level curve is concerned there is no evidence that the Mediterranean became isolated from the global oceans prior to the Messinian salinity crisis although the exact timing of this is not well documented (Müller et al., 1990; section 6.5.2.2). More important is the effect that closure of an ocean has upon the marginal regions affected by subduction. Within the context of such active tectonics, can the concept of eustacy have any relevance to the Mediterranean? The correlation, albeit inconclusive, of the eustatic and locally derived curves suggests that eustacy probably contributed to the sea level changes documented by the sedimentation, but are only visible when tectonic quiescence can be demonstrated.
CONCLUSIONS

1. Prior to the earliest deposition of continental conglomerates a marked and uneven palaeotopography had been developed. In the Manavgat basin the northern margin of the Alanya Massif formed a high which funnelled sediment westward in the south and along the Antalya Complex (a relative palaeotopographic low) to the north (section 5.7.1). Palaeotopographic features are localised along older sutures which have been interpreted as zones along which closure of Tethyan ocean basins took place (i.e. Antalya Complex-Alanya Massif boundary). This suggests that reactivation of these lineaments may have, at least in part, generated and controlled drainage during and prior to Miocene deposition (table 7.2).

2. Preservation of thick successions of continental fan-delta conglomerates and sandstones throughout the basin (e.g. ~1.5km in the Köprü basin, section 5.9.2; several hundred metres in the north-west of the Manavgat basin, figure 4.24) underlines the initial presence of palaeotopography, but also suggests that relative uplift of the hinterland is occurring to provide the quantity of coarse clastic material deposited (section 5.10). This uplift and the generation of accommodation space within the Isparta Angle at this time has been tentatively linked with south to south-eastward movement of the Lycian Nappe system to the north-west of the area (7.6.1), causing block faulting, pull-apart basin formation within a north-south striking zone.

3. Rising relative sea level during the Burdigalian and Langhian resulted in a clear diachronous transgression from south to north in the Manavgat basin (sections 4.7 and 6.5.2). This relative rise in sea level also affected the Aksu and Köprü basins in areas already undergoing active deposition (i.e. in central and southern parts of both basins; figure 5.31). Fan-delta sedimentation continued unabated on previously established fans, but the successions preserved reflect the relative sea level rise by a transition from purely continental deposits to coastal fan-delta successions (figure 5.31). Patch-reefs developed within these successions
on coarse conglomerate horizons similar to those described from the present day Red Sea (Hayward, 1982b), and the Miocene of southern Spain (Martin et al., 1989) and the Kas basin, to the south-west of the study area (Hayward, 1982a and b). The interaction of rising relative sea level and rapid, episodic sediment influx is clearly documented by the growth of individual corals up through successive clastic horizons (section 5.6.3.2, figure 5.21). Overall, study of these sensitive coastal sediments suggests that the dominant control on sedimentation at this time was relative sea level rise, despite the rapid sedimentation rate (section 5.9.2 and 5.10).

4. A previously unidentified extensional faulting event occurred in Late Langhian-Serravallian times. In the Manavgat basin this is most pronounced in the south-eastern part of the basin where Miocene sediments overlie Alanya Massif basement. Here, NW-SE trending normal faults developed at this time and resulted in the formation of an asymmetrical horst-graben system (section 7.5.1.2). These micro-faults affecting Lower-Middle Miocene sediments, parallel older structural trends within the Alanya Massif and probably represent reactivation along such weaknesses (section 7.5.1.4).

5. Extension-related deepening resulted in the abrupt transition from shallow-water carbonates (Oymapinar Limestone) to planktic foraminiferal marls (Geceleme Formation) in the Manavgat basin. This boundary is marked by localised deposition of fault generated-talus and calcirudites (Çakallar Formation; sections 7.5.1.2 and 4.7). In the Aksu and Köprü basins a similar transition occurred, where shallow water-carbonates were overlain by deeper-water fine-grained turbidites (Karpuzçay Formation). In some places (e.g. Deniztepesi) this boundary is also marked by a coarse talus horizon (section 3.6.2.3 and figure 4.24) indicating that faulting was not restricted to the Manavgat basin at this time.

6. After a short (1Ma) period of extension-generated faulting the Manavgat basin underwent a period of tectonic quiescence which allowed the accumulation of several hundred metres of planktic foraminiferal marls (Geceleme Formation; figure 4.22) as post-rift fill (section 7.1.5.2). In the Aksu and Köprü basins, turbidites sourced from the north flowed
south along north-south orientated older lineaments at this time (figure 4.11). Evidence from the south of the Köprü basin suggests that these lineaments, though focusing sediment dispersal patterns, may not have had a strong topographic relief during turbidite deposition (Flecker et al., 1995; section 4.6.2.4).

7. The Serravallian-Tortonian boundary is marked throughout the Isparta Angle by a change in sedimentation. The hinterland to the Manavgat basin was extended and uplifted dramatically at this time resulting in the deposition of conglomerates, debris flow, sandstones and siltstones (4.6.3.3). There is no correlative signature of tectonic intrabasinal subsidence which continued to shallow with time throughout the Tortonian to Middle Messinian. In the north of the Köprü and Aksu basins the Tortonian and Lower Messinian were characterised by continental and coastal conglomerate deposition (section 5.7.2). Once again, reef carbonate colonisation of conglomerate horizons suggests that rising relative sea level was a dominant control on sedimentation over and above high sediment influx (section 5.9.2). The different behaviour of the north and south of the Isparta Angle is evidence for the block faulted nature of the controlling basement.

8. A period of compression which terminated in the Lower-most Pliocene (the Aksu Phase) produced many of its characteristic west-vergent structures in the Latest Miocene (Poisson, 1977) and formed the angular unconformity which separates the Miocene from the Pliocene successions in the area. Previously unrecognised east-vergent structures can also be found affecting Miocene sediments and underlying Mesozoic basement (e.g. Akbas syncline, section 7.5.2.3). In addition there is evidence of minor folding related to inversion of the horst-graben system in the Manavgat basin at this time (section 7.5.1.3). Earlier interpretations argued that westward escape of the Anatolian block was fundamental in forming the westward transport signature during the Late Miocene (Poisson, 1984; Sengör et al., 1986). It is suggested here (section 7.6.2) however that movement of the Lycian Nappes orientated obliquely to north-south trending lineaments may have dominated the area close to the thrust front, including the Isparta Angle and resulted in some of the apparently east-west compression.
9. Facies models have been constructed for the Miocene sediments within the Isparta Angle. These indicate that the area was dominated by coarse (sandstone-conglomerate) clastic input throughout the Miocene. Deposition of these clastics occurred in a variety of environments from continental, to coastal to shelfal marine. Reef carbonates were deposited in coastal environments when environmental factors including sediment influx and relative sea level rise were favourable for coral colonisation.

10. Increased biostratigraphic and isotopic age resolution has allowed the correlation of Miocene sections across the Isparta Angle (Fig. 4.32). This has led to the modification of the existing stratigraphic framework (Akay and Uysal, 1985).

A final comment

The most important conclusion to come out of this study is that the fundamental control over the evolution of the Isparta Angle during the Miocene was the strong north-south structural grain of the heterogeneous basement. Stress, orientated in various directions, affected this long lived suture zone, but the structural features within it relate first and foremost to basement fabric. This conclusion severely limits the amount of direct interpretation that can be put on microstructural features in terms of the stresses affecting the Isparta Angle as a whole (e.g. Hague, 1993). The placing of the evolution of the three basins studied into the context of the development of the Eastern Mediterranean can therefore only be done in a tentative manner relying heavily on the temporal and spatial correlation of tectonic events outwith the Isparta Angle to those identified from within it. In doing so, a valuable perspective on the size and style of deformation that tectonic processes have on pre-existing suture zones is gained. More study of transitional areas between those which respond in an orthogonal manner to tectonic stress and those which develop oblique structures is required before the less tentative statements can be made.
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APPENDIX 1
XRD analysis on coral samples from the Köprü and Manavgat basins

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<th>XRD no</th>
<th>Sample no</th>
<th>Area</th>
<th>% Aragonite</th>
<th>Low Mg Calcite</th>
<th>High Mg Calcite</th>
<th>Quartz</th>
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<td>RA1</td>
<td>27j.Yes.5</td>
<td>Yesilbag</td>
<td></td>
<td>yes</td>
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<tr>
<td>RA2</td>
<td>10j.Alaran</td>
<td>Alaran</td>
<td></td>
<td>yes</td>
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<tr>
<td>RA3</td>
<td>26j.399.3</td>
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<td></td>
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<tr>
<td>RA4</td>
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<td>Beskonak</td>
<td>50-55</td>
<td>yes</td>
<td></td>
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<tr>
<td>RA5</td>
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<tr>
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<td>26J.217.4c</td>
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<td></td>
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<td>yes</td>
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<tr>
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<td>Karapınar</td>
<td>55</td>
<td>yes</td>
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<td>RA8</td>
<td>22M.406.2</td>
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Standard NBS987
- 126: 0.710227
- 26: 0.000026
- 24: 0.710239
- 2: 0.000013

F & A
- 2: 24
- 2: 26

mean of all 987: 0.710237
2S.D.: 0.000026
mean of my 987: 0.710239
2S.D.: 0.000013

8.52294E-06
### APPENDIX 3.A.1

Nannoplankton from the Ahmetler Section samples, Manavgat basin

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<th>NN number</th>
<th>Age</th>
<th>Species present</th>
<th>Abundance</th>
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### APPENDIX 3.A.2

Nannoplankton from the Aksu, Köprü and Manavgat basins (with the exception of the Ahmetler section). See Appendix 5.2 for stratigraphic logs.

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APPENDIX 3.B
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</tbody>
</table>
**APPENDIX 4.B.1**

Typical values of point counted marls from the Manavgat and Köprü basins

<table>
<thead>
<tr>
<th>sample</th>
<th>locality</th>
<th>carbonate</th>
<th>composite</th>
<th>quartz</th>
<th>organics</th>
<th>mica</th>
<th>heavy minerals</th>
<th>planktic forams</th>
<th>benthic forams</th>
<th>shell fragments</th>
<th>spicules</th>
<th>TOTAL</th>
</tr>
</thead>
<tbody>
<tr>
<td>205.348.3</td>
<td>Ahmetler</td>
<td>274</td>
<td>0</td>
<td>76</td>
<td>42</td>
<td>4</td>
<td>0</td>
<td>1</td>
<td>3</td>
<td>0</td>
<td>0</td>
<td>400</td>
</tr>
<tr>
<td>165.341.3</td>
<td>Ahmetler</td>
<td>534</td>
<td>0</td>
<td>0</td>
<td>36</td>
<td>0</td>
<td>0</td>
<td>23</td>
<td>7</td>
<td>0</td>
<td>0</td>
<td>600</td>
</tr>
<tr>
<td>165.341.6</td>
<td>Ahmetler</td>
<td>371</td>
<td>0</td>
<td>0</td>
<td>19</td>
<td>0</td>
<td>0</td>
<td>9</td>
<td>1</td>
<td>0</td>
<td>0</td>
<td>400</td>
</tr>
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<td>175.343.8</td>
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<td>0</td>
<td>20</td>
<td>0</td>
<td>0</td>
<td>25</td>
<td>5</td>
<td>0</td>
<td>0</td>
<td>400</td>
</tr>
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<td>Ahmetler</td>
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<td>0</td>
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<td>12</td>
<td>1</td>
<td>0</td>
<td>0</td>
<td>300</td>
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<td>0</td>
<td>0</td>
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<td>0</td>
<td>0</td>
<td>14</td>
<td>2</td>
<td>0</td>
<td>0</td>
<td>400</td>
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<td>Ahmetler</td>
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<td>2</td>
<td>5</td>
<td>0</td>
<td>0</td>
<td>20</td>
<td>7</td>
<td>0</td>
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<td>300</td>
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<td>14</td>
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<td>84</td>
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<td>300</td>
</tr>
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<td>Ahmetler</td>
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<td>0</td>
<td>17</td>
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<td>4</td>
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<td>300</td>
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<td>3</td>
<td>2</td>
<td>0</td>
<td>2</td>
<td>2</td>
<td>1</td>
<td>0</td>
<td>0</td>
<td>300</td>
</tr>
<tr>
<td>26J.217.1</td>
<td>Karapinar</td>
<td>241</td>
<td>0</td>
<td>59</td>
<td>18</td>
<td>0</td>
<td>2</td>
<td>0</td>
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<td>0</td>
<td>0</td>
<td>79</td>
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<tr>
<td>26J.217.2</td>
<td>Karapinar</td>
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<td>14</td>
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<td>0</td>
<td>0</td>
<td>3</td>
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<td>2</td>
<td>1</td>
<td>145</td>
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<td>26J.217.11</td>
<td>Karapinar</td>
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<td>34</td>
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<td>0</td>
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<td>1</td>
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<td>77</td>
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</table>
### APPENDIX 4.B.2

Typical point counts from the Karpuzçay Formation sandstones of the Köprü, Aksu and Manavgat basins

<table>
<thead>
<tr>
<th>Sample</th>
<th>1S.301.2</th>
<th>1S.301.3</th>
<th>17S.343.11</th>
<th>17S.343.18</th>
<th>18J.174.2</th>
<th>18J.174.3</th>
<th>22A.255.3</th>
<th>22A.255.1 (%)</th>
<th>22A.257.1 (%)</th>
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<tr>
<td>quartz</td>
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<td>4.5</td>
<td>5</td>
<td>0</td>
<td>128</td>
<td>42</td>
<td>25</td>
<td>8.2</td>
<td>31.4</td>
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<tr>
<td>carbonate</td>
<td>29.6</td>
<td>31.5</td>
<td>55</td>
<td>62</td>
<td>64</td>
<td>98</td>
<td>91</td>
<td>41.1</td>
<td>22.4</td>
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<td>mica</td>
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<td>0</td>
<td>0</td>
<td>0</td>
<td>5</td>
<td>4</td>
<td>0</td>
<td>0.3</td>
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<tr>
<td>organic</td>
<td>10</td>
<td>8.1</td>
<td>18</td>
<td>4</td>
<td>24</td>
<td>17</td>
<td>27</td>
<td>3.2</td>
<td>5.1</td>
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<td>chert</td>
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<td>0.4</td>
<td>0</td>
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<td>7</td>
<td>48</td>
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<td>pyroxene</td>
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<td>sutured qz</td>
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<td>0</td>
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<td>3.2</td>
<td>11.8</td>
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<td>micrite</td>
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<td>5</td>
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<td>3</td>
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<td>plagioclase</td>
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<td>3</td>
<td>1</td>
<td>1</td>
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<td>igneous</td>
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<td>0</td>
<td>0</td>
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<td>4.9</td>
<td>8.3</td>
</tr>
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<td>shell frags</td>
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<td>4.9</td>
<td>7</td>
<td>10</td>
<td>0</td>
<td>6</td>
<td>49</td>
<td>0.9</td>
<td>0.3</td>
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</table>

Planktic

<table>
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<tr>
<th>Component</th>
<th>11.2</th>
<th>15.7</th>
<th>33</th>
<th>5</th>
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<tr>
<td>Benthic</td>
<td>3.2</td>
<td>6.3</td>
<td>19</td>
<td>26</td>
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<tr>
<td>foraminifera</td>
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<td>Bivalve</td>
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<td></td>
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<tr>
<td>algae</td>
<td>3.2</td>
<td>0.4</td>
<td>69</td>
<td>64</td>
</tr>
<tr>
<td>brachiopods</td>
<td>8</td>
<td>6</td>
<td></td>
<td></td>
</tr>
<tr>
<td>echinoid spines</td>
<td>2</td>
<td>0</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Total count</td>
<td>310</td>
<td>304</td>
<td>395</td>
<td>318</td>
</tr>
</tbody>
</table>
APPENDIX 5.1

Akseki Road Section

Key

- Marl
- Siltstone
- Sandstone
- Conglomerate

Sedimentary structures also denote grain size

Sample numbers are coded in the following way:

Date (i.e. 25S = 25th September) . Site number (i.e. 348) . Sample number (i.e. 2)

25S.348.2

The section was logged from bottom to top and the dates therefore follow this trend.

Samples are referred to extensively in Chapter 4 and 6 and in Appendix 2.
Top of log

coal fragments

coal fragments

coal fragments

coal fragments

n.e. coal fragments

185.308.2
APPENDIX 5.2

Stratigraphic location of samples used for nannoplankton analyses and Sr dating (Appendix 2)

For location of sections, see figure 3.1. For key, see figures 3.3 and 3.4.

KARGI BARAJ

1. 15m 284.3
   15m 284.4
   15m 284.5
   Oyster dome, planar bedded with oysters on top, massive, oyster-rich, bored, condensed oysters at top, decrease downwards, Jurassic, bored with palaeotopography

2. 15m 284.1
   15m 284.2
   Oyster dome, tubular, knobbly lms

3. 24J 384.2

ASPENDOS

17J 415.5
17J 415.3a-c
17J 415.2
17J 415.1

- Gastropod-rich marl
- Marly and bioclastic
- Scaphopod-echinoderm facies
- Imbricated x-sets
- Rare caliche channelised
- Channelised base
- Channelised base
ALTINKAYA

SECTION 180 Altinkaya
Scale 1: 5:0

BALLIBUCAK

SECTION 163 and 164 Ballibucak
Scale 1: 150:0

- Bivalves and gastropods
- Large corals, oysters and algae
- Abundant small coral
- Shells and coral fragments in sandy matrix
- Fossil-rich, planar bedded
- Fossil rich
- Wood, oysters and whole bivalves
- Clasts of nummulite-rich lms
- Bivalves and gastropods
- Echinoderms
- Oysters, gastropods, bivalves and algae
- Abundant shells
- With shells
PAGE NUMBERING AS ORIGINAL
sample numbers | comments
---|---
96 Alarahan 3 | bioclastic debris flow
80 Alarahan 2
80 Alarahan 1
78 Alarahan 4 | calcarenite, pectens, echinoderms, rhodoliths, operculina
76 Alarahan 5
R3 104 Alarahan 4 | domalis concentrated at bottom, Purius
| x-bedded
| channelized
R2a 104 Alarahan 3 | Domantly Purius. Domals at top
R2a Alarahan 3 | Purius with spaced domals
| erosive base
| x-bedded, inflected
R2 104 Alarahan 1 | various domals at base, reef debris at top
| faint x-strat
R2 = reef number
104Alarahan 2 = sample number
Appendix 6

BACKSTRIPPING

The procedure followed in backstripping is outlined below and illustrated using a fictional well, Well X (Figure A6.1).

Data Preparation

First the well to be backstripped is split up into a number of layers, the tops of which can be dated. Each layer is assigned a bulk lithology and the water depth of deposition for sediments at the top and bottom of the layer is estimated (Figure A6.1). Unconformities are treated as periods of zero deposition, and assigned a nominal thickness of 0.1 metres, unless independent estimates for the amount of erosion associated with the unconformity are available. The age, depth and water depth of deposition of the oldest sediments on basement are also required (Figure A6.1).

These layers are then removed, one by one and youngest first, (back stripped) to reconstruct the subsidence history of the well site being investigated.

Decompaction

The simple calculation described above assumes that each layer has remained the same thickness through time. This is clearly not the case. Sedimentary rocks can be thought of as being composed of two parts: solid grains and the void spaces between these grains which are normally filled with water. Porosity is a measure of the proportion of the total rock volume made up of void spaces and is found to vary with depth of burial. For example, the porosity of newly deposited marine muds is typically around 65% whilst the present day porosity of Jurassic Heather Formation marine shales at depths of 4 km in the North Sea is closer to 10%. This reduction in porosity (volume) since deposition represents a decrease in layer thickness of more than 60%. The empirical relationship observed between porosity and depth of burial (Figure A6.2) has led to the general belief that mechanical compaction, due to the weight of the overlying sediments, is the

This appendix is provided by J. Turner as part of his thesis work (Turner, 1996).
major control on porosity reduction. Correcting for the effects of compaction is a major part of backstripping and geohistory calculations.

This observed relationship between porosity and depth of burial, which varies between lithologies (Figure A6.2), is used to correct for the effects of compaction. As each layer is removed the porosity, and so thickness (Figure A6.3a), of each of the remaining layers is recalculated by moving the layer back up the appropriate porosity/depth function (Figure A6.2). This increases the thickness of all the remaining layers (Figure A6.3a). On a plot of depth to basement against time (Figure A6.3b) decompaction increases the amount of basement subsidence calculated for the early part of the basin's subsidence history. The amount of subsidence calculated for later, nearer to the present day, is reduced by a commensurate amount due to the fact that space for sediments being deposited is being generated, in part, by the compaction of the existing sediment pile rather than any additional basement subsidence.

Compaction also has a major impact on calculated sedimentation rates (Figure A6.3c). Sedimentation rates uncorrected for compaction are simply calculated by dividing the observed sediment layer thickness by the length of time it took for that layer to be deposited. Sedimentation rates corrected for compaction, assuming a simple exponential porosity/depth function, can be calculated using the equation:

\[ R = \frac{L(1 - \phi)}{T(1 - \phi_0)} \]  

(van Hinte 1978)

- \( R \) - Sedimentation Rate (mMa\(^{-1}\))
- \( L \) - Layer Thickness (m)
- \( T \) - Time taken for layer to be deposited
- \( \phi \) - Present Day Porosity
- \( \phi_0 \) - Depositional Porosity

This correction increases calculated sedimentation rates over the entire period of deposition (Figure A6.3c) although again these effects are greatest for the most deeply buried sediments. Sediments rates calculated in this way are plotted on Figures 4.28 and 4.23.
Sediment Unloading

By making corrections for the effects of sediment compaction and changing water depths, a picture of the actual subsidence history of well X has been reconstructed (Figure A6.3 a&b). Total subsidence is plotted in Figure 4.33. For tectonic subsidence to be calculated an additional step in the backstripping calculations is necessary.

A number of different forces and processes control basin basement subsidence and the way in which it is recorded in the basin's sedimentary fill. Apart from tectonic subsidence, the main cause of basin basement subsidence is sediment loading. To calculate tectonic subsidence the effect of sediment loading is removed by isostatically unloading the remaining sediment column from the basement (Figure A6.4) as each layer is removed. The resulting plot of water loaded depth to basement against time can then be compared to the predicted tectonic subsidence curves generated by theoretical models. Tectonic subsidence has been calculated for the Ahmetler section and is shown fitted to a theoretical subsidence curve in figure 7.17.

By convention, tectonic subsidence curves are presented as plots of water loaded, rather than completely unloaded, depth to basement. For comparison with theoretical subsidence curves the most important consideration is that both sets of curves are unloaded in the same way. In this study, as has been the practice in most previous studies using well data, Airy (local) isostasy is assumed. Flexural backstripping, which is more realistic mechanically and so desirable, requires additional information concerning variations in the two dimensional structure of the sediment load and the mechanical properties of the lithosphere through time.

Decompaction of Lithologically Mixed layers

Five lithology types have been used for the decompaction calculations in this study. Each layer is only assigned one of these lithology types. Clearly many layers will be mixtures of grain size (sand, silt, shale) and/or composition (clastic, carbonate, volcanic). The most rigorous method for dealing with this would be to treat each stratigraphically dated layer as a composite, made up of

This appendix is provided by J. Turner as part of his thesis work (Turner, 1996).
several separate layers of different compositions. This approach would, however, be very time consuming, both in breaking down the preserved sedimentary section and in computer time for the calculations, and would probably not produce any great change in the calculated subsidence history. A simplifying approach taken in this study was to define each layer's lithology on the basis of the percentage of each lithology type present. This was only possible for clastic units. Where units were a mixture of clastics, carbonates or volcanics the dominant lithology type was used.

3.6.2 Non-mechanical Compaction

The vast majority of decompaction models are based upon the premise that mechanical compaction due to overburden pressure is the only form of porosity reduction. This is not the case as cementation and dissolution can have an important impact on porosity, and therefore compaction, throughout a sediment unit's burial history. For example, an early phase of cementation may bind together individual sediment grains, strengthening the sediment and so retarding mechanical compaction. In these circumstances the present day thickness of that unit will be greater than it would have been had there been no cementation, and the decompacted sediment thickness will be overestimated.

The degree to which mechanical compaction may have been retarded, or enhanced, by cementation or dissolution cannot be assessed without detailed petrological information and, as a result, is almost impossible to estimate in most wells. Although some workers have attempted to introduce mathematical terms into their decompaction calculations to make an allowance for these effects (e.g. Thorne & Watts 1989) these corrections are almost impossible to verify. For this reason, and because mechanical compaction is still the dominant factor in porosity reduction, we have elected to leave them out of our calculations rather than introduce additional complexities.

This appendix is provided by J. Turner as part of his thesis work (Turner, 1996).
Non-exponential porosity-depth Functions

In this study the exponential porosity/depth functions of Sclater & Christie (1980) have been used for the decompaction calculations. These corrections have become the standards which many workers have used in previous studies of basins worldwide.

Their applicability has been challenged on a number of grounds, as has the validity of an exponential rather than linear porosity/depth relationship. Compared to a linear relationship an exponential porosity/depth function will overestimate the amount of porosity reduction, and so compaction, in the early part of a well's burial history. A number of studies have also suggested that the effects of variations in factors like sediment composition or fluid flow mean that porosity/depth relationships for specific lithologies in each basin are different. Furthermore, part of the area over which Sclater & Christie (1980) took their data may have been affected by Cenozoic erosion and overpressure development.

A full assessment of the likely effects of the many different published porosity/depth functions and compaction corrections is beyond the scope of this study. However, since all of the wells in this study have been backstripped and decompacted using identical corrections, any trends in subsidence behaviour through time around the basin will not be the result of inadequacies in the compaction correction.

Overpressure

Overpressure development may lead to localised anomalies in calculated subsidence patterns. Where shales, in particular, are deposited and then buried very rapidly water is trapped and the intraformational pore pressure increases to levels above normal hydrostatic pressure. In these circumstances the pore pressure profile, rather than increasing steadily downhole, proceeds in steps with zones of anomalously high pore pressure corresponding to overpressured shale units. In these units water is unable to escape. This keeps the shale unit's porosity higher than normal, and so retards compaction.

This appendix is provided by J. Turner as part of his thesis work (Turner, 1996).
The present day thickness of an overpressured shale unit will be thicker than it would be under normally pressered conditions because of the presence of the trapped water. If this present day thickness is decompacted using a normal porosity/depth function its depositional thickness will be overestimated. As with most of these possible sources of error, the magnitude of its effect is likely to be relatively minor.

**Water Depth Estimates**

Changes in water depth as well as sediment accumulation preserve the record of basin basement subsidence. Changes in water depth through time can be assessed from a variety of sources of geological information. They are notoriously difficult to assess accurately and uncertainties and errors in water depth estimates are likely to be the largest single source of error in any subsidence analysis study.

Probably the only diagnostic paleobathymetric markers are coals (Bertram & Milton 1989) which are indicative of terrestrial environments. Other sedimentary facies like red beds (non-marine), coarsening upward pure sands (shoreface) and turbidites (deep marine) are strongly indicative of, but not conclusive proof for, particular depositional environments and more importantly water depths. Even evidence from sedimentary structures, of processes like wave or storm action, can be misleading when attempting to determine a particular depth, since the depth to which the processes reach is strongly dependent upon the strength of the tide and prevailing winds. Fossils, macrofossils and particularly microfossils, are particularly useful environmental indicators. Ratios of planktic to benthic and particular assemblages of benthic foraminifera can, by reference to present day faunal distributions, be used to assess paleowater depths.

Often, however, particularly for deep water sediments or where microfossil preservation is poor, water depths are very hard to assess confidently. Good examples of this come from the Kimmeridge Clay of the Late Jurassic and the Lower Cretaceous in the North Sea. The Kimmeridge Clay Formation is a laminated organic mudstone apart from parallel laminations it contains no sedimentary structures but frequently contains mass flow sands with a

This appendix is provided by J. Turner as part of his thesis work (Turner, 1996).
characteristic blocky log character. This is characteristic of fairly deep water, certainly deeper than a shallow shelf, whilst the poorly preserved microfauna support this interpretation. Whether the Kimmeridge Clay represents an outer shelf (175-500 metres), upper bathyal (500-1000 metres) or even lower bathyal/slope (>1000 metres) is far less clear as are the likely variations in water depth it reflects throughout the area.

This appendix is provided by J. Turner as part of his thesis work (Turner, 1996).
Before a well can be backstripped, the preserved stratigraphic succession must be simplified and split up into a number of layers ready for input to the backstripping calculations. The age, depth below sea-bed (or an alternative datum level) and water depth of deposition for the top of each layer are assessed and each layer is assigned a bulk lithology. Although not shown in the diagram above, in this study the water depth of deposition for the base of each layer is also included. The age, depth below sea-bed and water depth of deposition of the oldest sediments sitting on basement are also required.
Figure A6.2 Porosity/Depth Relationships

Plots of log porosity against depth for the four most common lithologies in wells from the Central North Sea (Sclater & Christie 1980). These functions are used in the decompaction calculations that generated the curves shown in Figures 3.4 and 3.5. Mechanical compaction is most intense during the early portion of burial. The effects of compaction vary between lithologies. Although chalk and shale have higher depositional porosities (depth = 0) than sand they rapidly loose this porosity on burial. Sand, on the other hand, retains relatively high porosities until it has been very deeply buried.

\[ \phi = \phi_0 \, e^{-cz} \]

\( \phi \) - porosity
\( \phi_0 \) - porosity at depth = 0
\( c \) - porosity constant
\( z \) - depth

from Sclater & Christie (1980)

Jon Turner, University of Edinburgh (1995)
As each layer is removed all the remaining layers are decompacted and "fluffed out":

**Equation:**

\[
Z_i' = Z_i + \frac{Z_i - Z_{i+1}}{(1 - e^{-cZ_i^l})} \frac{cZ_i^l}{cZ_i^l - cZ_i^l} 
\]

(Sclater & Christie 1980)

Well X - somewhere in the North Sea

Plot of depth to basement against time, uncorrected and decompacted, with minimum and maximum likely water depth estimates. The actual position of the basement at any time should be somewhere in the shaded area between these estimates.

Plot of sedimentation rate against time, corrected and uncorrected for the effects of compaction.

Correcting for the effects of compaction increases the calculated sedimentation rates.
Figure A6.4

Sediment Unloading - Tectonic Subsidence

Tectonic Subsidence (water loaded depth to basement through time) is the component of subsidence due only to the tectonic force driving basin basement subsidence. This is calculated by attempting to remove all the other factors contributing to the subsidence of the basin. This includes sediment compaction and, more importantly, their loading effect. This load is removed isostatically and the depth to basement through time is recalculated assuming that the basin was only ever filled with water. The contribution of sediment loading to total subsidence increases through time as the basin fills, so that the amount of tectonic subsidence is much less than the total subsidence later in the basin's history.