SEDIMENTOLOGY OF JURASSIC SYN-RIFT
RESEDIMENTED CARBONATE SANDBODIES

by

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A thesis submitted to the University of Bristol in accordance with the requirements of the degree of Doctor of Philosophy in the Faculty of Science, Department of Geology.

September 1989
VOLUME CONTAINS CLEAR OVERLAYS
OVERLAYS SCANNED SEPERATELY AND OVER THE RELEVANT PAGE.
CONTAINS PULLOUTS
Abstract

This thesis discusses the sedimentology of three contrasting Jurassic carbonate sand turbidite systems from Southern Europe: the Cutri Formation (Bathonian) of Mallorca; the Vajont Limestone (Bajocian-Callovian) of northern Italy; and the Peniche sequence (Toarcian-Aalenian) of the Brenha Formation of western Portugal.

These sandbodies all formed in syn-rift extensional settings which imposed a primary morpho-tectonic control on both the source platform and depositional basin morphology. The three sandbodies in question display varying geometries and architectures and are discussed in terms of the palaeogeographic, tectonic and eustatic controls that governed their individual development; as well as being used to test the recently developed apron model against that of the submarine fan. In this context oolitic carbonate aprons associated with palaeowindward and palaeoleeward platform margins have been distinguished.

The Cutri Formation is interpreted as an oolitic base-of-slope apron, that displays a minor single syn-rift thinning upward megacycle (retrogradational) trend indicative of subsidence out-pacing sedimentation. The apron correlates with a eustatic sea-level drawdown and was characterised by infrequently laterally correlatable, oolitic turbidite units separated by hemipelagic interbeds. This sandbody is relatively sand-poor in nature, and is interpreted as being sourced from a palaeowindward platform margin.

The Vajont Limestone is re-interpreted as an aggraded oolitic apron from its original interpretation as a sub-marine fan. The apron is composed of stacked oolitic grainstone turbidites and is locally up to 800m thick. It is interpreted as being sourced from a stable ‘keep up’ palaeoleeward platform margin, where dominant off-bank sand transport led to development of line-sourced oolitic turbidites, which were actively aggraded by on-going basin subsidence. Statistics were used to demonstrate a random turbidite sequence which enhances the apron interpretation.

The Peniche sequence is reconfirmed as a carbonate-siliciclastic fan, its facies development conforming to a siliciclastic sand-rich fan model. Statistical analysis indicates a non-random (cyclic) turbidite sequence, thereby enhancing the fan interpretation. The fan occurs as a localised development within fine-grained basinal facies and correlates stratigraphically with a eustatic sea-level drawdown. Interpreted as being sourced from a palaeoleeward margin, the sequence progradates from outer fan lobes to a thick, multi-storey braided channel complex.

These syn-rift resedimented carbonate sandbodies have the potential to be stratigraphically associated with basinal source rocks and therefore may be viewed as prospective hydrocarbon reservoir facies.
Acknowledgements

The project was jointly sponsored by the Natural Environment Research Council (NERC) and London & Scottish Marine Oil PLC (LASMO), whose extra financial support made the project, fieldwork and conference attendance possible. The project was conceived from an idea by Dr. Paul Wright, the author's supervisor; LASMO's interests lying in the possible application of the project to the Vega Field, Sicily. The author's contacts in LASMO, Keith Morris and Steve Mills, are thanked for their support. Paul Wright's assistance and advice throughout is acknowledged in appreciation.

David Prescott of Oxford University, whose thesis is on the Jurassic geology of Mallorca, is especially thanked for numerous discussions on the Jurassic of Mallorca, for helping the author to search the Sierra de Levante for outcrops of the Cutri Formation, for logging help on some sections, and for "allowing" the author to work on the most interesting aspect of Mallorcan geology. Thanks to Toni Simo (University of Wisconsin) and Antonio Barnolas (Spanish Geological Survey) for helpful discussions and an enjoyable few days in the field. Special thanks go to the author's mother, Gill, a brilliant field assistant, who fended off a gamekeeper with a gun and ferocious black Mallorquin dogs, and willingly clambered over mountains from dawn to dusk. She appears several times as scale in Chapter 3.

Massimo Sarti, formerly of Ferrara University and now at Cosenza, is thanked for his logistical support on the author's first field season in Italy, and for subsequent personal communications in Chapter 4. The field assistance of Ghassem Gheissary (Engineering Department, Cambridge University) in acting as the author's chauffeur while in Italy was much appreciated.

Dr Paul Wright collected the samples of the Peniche sequence and his photographs are used in Chapter 5. His rapid turn-around of the manuscript drafts was much appreciated.

Pedro Ruiz-Ortiz and José Miguel Molina of Jaén University are thanked for a memorable and enjoyable field trip to the Betics, the results of this research have not been written up in this thesis.

Discussions with Harry Cook, Henry Mullins, Al Hine, Robert Bourrouilh, Paul Crevello, Monty Hampton, Larry Doyle and Yezekeel Druckmann, helped to clarify many of the author's ideas and aided the progress of the project.

Technical support at Bristol University from Ian Avent and Gary Webber for their thin-section preparation and Simon Powell in photography are acknowledged in appreciation. At Liverpool University, many thanks to Dr Jim Marshall for providing use of the isotope laboratory and especially to Hilary Attenborough, who trained the author to process samples for isotope analysis. The fluorescence microscope was used courtesy of Dr Patterson in Zoology. Advice on statistical analysis was sought from Adrian Romilly, a lecturer in Mathematics at Bristol.

The colour photographs in the text were all printed by Boots Ltd.; the black and white photographs were processed in the Geology Department at Bristol University. Colour Processing Laboratories Ltd. of Bristol printed several colour prints from slides. The plates were all set out by the author's father, Graham, and were reproduced by Copy Colour of Swindon with the help of Bernard Ryan. Figures were photocopied courtesy of Tomura Hinchley Ltd, Devizes.
Finally, the author would like to acknowledge the many geologist friends and colleagues she has made while undertaking this project. Special thanks to the author’s parents, Gill and Graham, sisters Imogen and Melissa, and to Ghassem for their support, help and encouragement throughout, especially in the last few hectic weeks.

Lastly, and most importantly, to Gill Rollings, formerly of Electrical Engineering and now of the Department of Computer Science at Bristol, who word processed the thesis so efficiently and professionally and to Joan Powell of the Mechanical Engineering Department who very kindly took over while Gill was on holiday.
MEMORANDUM

The accompanying dissertation entitled "Sedimentology of Jurassic Syn-Rift
Resedimented Carbonate Sandbodies" is submitted in support of an application
for the Degree of Doctor of Philosophy in Geology at the University of Bristol.

The dissertation is based on independent work carried out by the candidate.
All contributions from others have been duly acknowledged within the dissertation.
This work has not been submitted for any other degree or diploma at this
University or any other Institution.

I hereby declare that the above statements are true.

F V ABBOTS

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(from Sabat 1986)

Fig.4 A) Locality map for the Amoixa area of the Sierra
de Son Amoixa
B) Structural and geological map of the area shown in A
(from Sabat 1986)

Fig.5 Eustatic sea-level curve for the Middle Jurassic.
(from Haq et al 1989) (back pocket)

Plate 1 View of the Puig Cutri ridge, looking to the east
(back pocket)

Plate 2 View of the Puig Cutri ridge, showing the position of
the logs on Appendix Fig.2.
CHAPTER 1
INTRODUCTION

1.1 Aims and Thesis Strategies

Aims

The aims of the project were threefold: firstly, to use ancient resedimented carbonate sand sequences to test the carbonate apron model against the more established classic submarine fan models. The carbonate apron model was in its infancy when the project commenced, and at that time resedimented carbonate sand turbidite systems were being described as fans (which to some extent is still a valid comment at the time of writing) even though work from the Bahamas had indicated that carbonate fans were absent from Bahamian troughs. The later publication of the synthesized apron model concept by Mullins & Cook [1986], produced a firm model to test against the established fan model.

The second aim was to investigate specific palaeogeographic, tectonic and eustatic controls governing resedimented sandbody geometry and architecture. This was considered to be of primary importance due to the contrasting geometries and architectures displayed within ancient carbonate sandbodies described in the literature.

The specific role of these controls and their interaction and effect on resedimented sandbody morphology, had not been previously discussed with respect to the ancient record, even though many ideas had been generated from studies of the modern Bahamas.

The third aim was to investigate the sandbodies sedimentologically and diagenetically with a view to assessing the potential of such facies in general as hydrocarbon reservoirs.

Themes

The volume of literature on resedimented carbonates is infinitely small when compared to that available for their siliciclastic counterparts. However there does appear to be a fundamental difference between the turbidite depositional systems of siliciclastic and carbonate regimes. While siliciclastic turbidite systems tend to form fans, carbonate turbidite systems tend to form aprons, this fundamental difference being due to the primary mode of sediment entry into the basin. Whereas siliciclastic fans are fed by a single major sediment point source, such as a canyon, linear carbonate platform
margins preferentially shed material along their entire length, creating an apron of resedimented facies that parallels the length of the source margin [e.g. Mullins & Cook 1986].

The theoretical difference between fans and aprons is that the long-term point source of fans, favours the development of an ordered network of channels and lobes, which form characteristic turbidite cycles: thickening- (and coarsening-) upward on lobes and thinning- (and fining-) upward in channels. In contrast, the line source with numerous minor entry points characteristic of aprons, gives rise to a random distribution of facies at the toe of slope, which amalgamate to form a slope parallel wedge, that shows proximal to distal trends perpendicular to the margin. Apron facies are not purely restricted to deep-water carbonate environments, both modern and ancient siliciclastic aprons have been documented in the literature. Nevertheless, over the last twenty years, virtually all siliciclastic turbidite systems have been described as a priori fans, as have many carbonate turbidite systems (for details see Section 1.4).

In order to test the concepts of fans and aprons, as displayed in both siliciclastic and carbonate turbidite regimes, ancient carbonate sandbodies that were formed entirely from the resedimentation of unconsolidated carbonate sands were specifically chosen for study. These were selected, in preference to resedimented carbonate deposits dominated by more catastrophic, cohesive mass-flow deposits, for a number of reasons:

(1) Although of lower density than quartz, carbonate sand grains in the form of ooids and peloids pay the closest resemblance both in size-range and shape to siliciclastic sands forming modern and ancient fan sequences. Ooids in particular take the form of perfect hydrodynamic spheres, thereby being directly analogous to artificial spheres of sand size used in hydrodynamic turbidity current experiments [e.g. Bagnold 1954] and sediment fluid mechanics theory [e.g. Gibbs et al. 1971; Middleton & Southard 1978]. Therefore if carbonate turbidites do form fans there is every reason to suspect similar facies developments to those seen in siliciclastic fans. For these reasons also, later sedimentological comparisons with siliciclastic turbidite systems are validated.

(2) During the Holocene-present, carbonate sands have been produced on Bahamian platforms and transported off-bank onto adjacent slopes and basins. The chosen ancient turbidite systems are all composed of similar carbonate sands, which makes a stronger case for using concepts from, and drawing analogies with, Bahamian platforms and their slope and basin facies.
Work on the Bahamas has shown that there are fundamental differences between slope and basin deposits associated with the windward (facing the prevailing wind direction) or leeward (opposing the prevailing wind direction) orientation of the source platform margin. It was suspected that these differences could also be distinguished in ancient carbonate sandbodies and that palaeowindward and palaeoleeward type margins could be differentiated.

Shallow water carbonate sands are produced within a narrow sea-level range, and their production rates are highly dependent upon, and extremely sensitive to, relative changes in sea-level. Tectonically or eustatically induced changes of sea-level of even minor importance can therefore radically alter carbonate sand production rates, and hence their rates of resedimentation to the deep. The effects of relative sea-level changes should therefore be preserved in resedimented carbonate sandbodies, as a record of the evolution of the platform margin.

With these aims in mind, three Lower-Middle Jurassic carbonate sand turbidite systems were selected for study from southern Europe. They were specifically chosen since they were all composed of oolitic-peloidal sands and were all resedimented into structurally controlled extensional basins, that formed in response to the rifting that preceded opening of the Atlantic and the Western Tethyan, Ligurian Ocean. Despite having these similarities in common, the three sandbodies in question also display widely contrasting geometries and architectures. Since sandbody nature is so variable, it is important to investigate the influence of palaeogeography, tectonism and eustacy and the result of interactions between these controls, that combined to produce the individual sandbody architectures displayed in these examples.

This is especially important since these sandbodies may form laterally extensive, major stratigraphic accumulations of grainstone turbidites; that may be laterally and vertically associated with basinal sediments, that could act as potential source or sealant facies. This type of carbonate sandbody may therefore be regarded as having potential in the geological record as hydrocarbon reservoirs.

Resedimented carbonate facies form the extensive hydrocarbon reservoirs of the mid-Cretaceous Poza Rica Trend of Mexico [Enos 1977a; Viniegra 1981] and smaller reservoirs within the Permian Delaware Basin of western Texas [Hobson et al. 1985]. The sandbodies studied in this project were of Jurassic age, when carbonate sands were
calcitic [Sandberg 1975 and 1983; MacKenzie & Piggot 1981; Wilkinson 1985] and therefore had less diagenetic potential than if they had been megastable aragonite [e.g. Scholle 1977; Mullins & Cook 1986]. However porosity in resedimented carbonate reservoirs (such as those mentioned on the previous page) does tend to be secondary porosity created by late stage dolomitization and/or dissolution. It must also be stressed that the sandbodies in question are viewed in this light only with respect to their theoretical reservoir potential as concerns sandbody architecture, geometry and potential for late stage diagenesis and not as they actually exist in the geological record.

1.2 Introduction to the studied formations and their localities

The three Jurassic localities selected display palaeogeographical E-W separation, but similar palaeolatitudes and are as follows:
(1) the Cutri Formation of the Sierra de Levante, Mallorca;
(2) the Vajont Limestone of the Southern Alps, northern Italy;
(3) the Peniche sequence of the Brenha Formation of the Peniche peninsular, western Portugal. (see Fig. 1-1).

(1) The Cutri Formation (Bathonian) of Mallorca formed on the northern rifted margin of the Ligurian Ocean during its syn-rift phase of development (Figs. 1-2 and 1-3. The sandbody is the thinnest of the three examples, being an 80m thick retrograde succession, of approximately equal proportions of major oolitic turbidite units and interbedded hemipelagic units. The Cutri Formation facies were resedimented into a structurally controlled graben basin with a complex internal block configuration, the facies of which now outcrop in the mountains of the Sierra de Levante of Mallorca. The basin was initiated by rifting in the Domerian and at the time of oolite resedimentation was bordered to the north-west by a submerged pelagic plateau now represented by the present-day Sierra Norte of Mallorca. The coeval Middle Jurassic source platform lay to the east, a lateral representation of which occurs on Menorca, with a small fragment of the source platform outcropping in the far east of the Sierra de Levante. Briefly described by Alvaro et al. [1984], this sandbody had received no previous detailed sedimentological study. Its particular research value lay in the availability of good lateral as well as vertical outcrop exposure, which allowed the construction of extremely detailed log profiles.
(2) The Vajont Limestone (Bajocian–Callovian) of northern Italy formed on the southern margin of the Ligurian Ocean during its syn-rift phase and had the most easterly palaeogeographic position of the three localities (Figs. 1-2 and 1-3). The Vajont Limestone displays the greatest longevity and greatest vertical thickness development of the three formations. The sandbody is aggraded, and forms a sequence up to 800m thick of stacked oolitic turbidites. It was deposited in the Belluno Trough, a 50km wide structural graben that formed in the Hettangian between the stable Friuli source platform to the east and the Trento horst platform to the west, which was a submerged and isolated pelagic plateau at the time of oolite resedimentation. Outcrops of the Vajont Limestone occur around the towns of Vittorio-Veneto and Belluno, just as the Southern Alps rise up from the Po plain. The turbidite sequence was interpreted as an oolitic fan by Bosellini et al. [1981a and b] before the work on carbonate aprons (summarized by Cook [1983] and Mullins & Cook [1986]) was published, warranting a re-evaluation of the turbidite system in the light of this important work. Extensive vertical outcrop of this section is available in sections perpendicular to the margin, but good lateral outcrop detail is lacking.

(3) The Peniche sequence of the Brenha Formation (early Toarcian–Aalenian) of Portugal formed in the Lusitanian Basin, on the eastern Iberian margin of the North Atlantic during its syn-rift development stage. As such, this carbonate sandbody had the most westerly palaeogeographic position of the three localities. Outcrops are areally restricted to coastal sections around the Peniche peninsular. The sandbody is a 350 m thick prograde succession of resedimented peloidal-oolite sediment, which contains an important siliciclastic sediment fraction. The sediment was sourced from an isolated, offshore, basement horst that formed part of a ridge of uplifted Hercynian basement, which formed the western boundary of the Lusitanian Basin. The Iberian landmass lay to the east of the Lusitanian Basin (Figs. 1-2 and 1-3). The Peniche sequence was interpreted as a carbonate fan by Wright & Wilson [1984] and seemed to show the development of specific fan characteristics. The potential development of true carbonate submarine fan facies was therefore important to investigate in the light of the development of the apron model.
Figure 1-1

General location map to show relative geographical positions of the three localities.
1 = Peniche sequence, Portugal;
2 = Cutri Formation, Mallorca;
3 = Vajont Limestone, Italy.
1.3 Lower to Middle Jurassic regional setting

The following two sections discuss the regional evolution of Western Tethys (Section 1.3.1) and the western Iberian North Atlantic margins (Section 1.3.2) from the Upper Triassic to the Callovian, to set the regional tectonic and palaeogeographic framework that governed basin formation and deposition of the resedimented carbonate turbidite systems. Both the Cutri Formation and the Vajont Limestone were formed in Western Tethys, where basin development was due to extension preceding sea-floor spreading in the Ligurian Ocean; while the Peniche sequence developed in response to extension in the North Atlantic.

1.3.1 Triassic-Callovian evolution of Western Tethys and the Central Atlantic

The concept of a vast east-west seaway between Africa and Eurasia, occupying areas from the Alps through to the Himalayas, was envisaged nearly a hundred years ago by Suess [1893] and was later named Tethys (after the daughter of Oceanus). The term Tethys is now reserved for the Mesozoic sea, the pre-Jurassic Tethys being referred to as Palaeotethys, the post-Tertiary Tethyan "remnant" being the present-day Mediterranean [Laubscher & Bernoulli 1977; Sonnenfeld 1981].

The complex configuration of the Mediterranean region reflects the post-Upper Triassic history of the region, which was dominated by extension from the Upper Triassic to Early Cretaceous and then compression which commenced in the Late Cretaceous and culminated in the Alpine Orogeny [Dewey et al. 1973; Biju-Duval et al. 1977; Dercourt et al. 1986]. The Mesozoic sea of the present day Mediterranean area is often referred to as Western Tethys, since the ocean that formed in this region was a narrow westwards extending branch of the main Tethys Ocean, which lay to the east. This ocean branch is referred to as the Ligurian Ocean, and it eventually served to connect Tethys in the east to the opening Atlantic in the west [Biju-Duval et al. 1977; Lemoine 1982; Dercourt et al. 1986].

The evolution of the Jurassic Western Tethyan realm and the Central Atlantic was controlled by plate movements between the large stable cratons of Europe and Africa, as America initially rifted and then drifted away from Europe and Africa [Hsu 1971; Dewey et al. 1973; Biju-Duval et al. 1977; Savostin et al. 1986; Tricart et al. 1988]. The early evolution of Western Tethys was closely associated with that of the Central Atlantic (which started its evolution as a western segment of Tethys, only becoming a
part of the Atlantic system in the mid-Cretaceous) opening of the latter necessarily requiring translational movement between Europe and Africa [Dewey et al. 1973]. Throughout the Jurassic, Africa moved left-laterally with respect to Europe, causing the narrowing of Eastern Tethys by subduction along its northern margin (Figs. 1-2 and 1-3) and extension in Western Tethys and the Central Atlantic [Dewey et al. 1973; Biju-Duval et al. 1977; Lemoine 1982; Dercourt et al. 1986]. This westward extension created the Ligurian Ocean, which commenced sea-floor spreading simultaneously with the Central Atlantic, in the late Bathonian-Callovian [Savostin et al. 1986; Lemoine 1983; Tricart & Lemoine 1988] (Fig. 1-3). Lemoine [1982] divided the post-Triassic history of the Mediterranean into distinct phases associated with the opening and subsequent closure of Ligurian Ocean.

(1) Pre-rift phase - this is the Triassic-Early Jurassic phase that precedes extensional block-faulting and is of variable longevity over Western Tethys.

(2) Rift phase - is taken from the time of the onset of block-faulting up until the commencement of sea-floor spreading in the Ligurian Ocean in the latest Bathonian-Callovian. It was during this phase that all three resedimented oolite sandbodies were deposited.

(3) Drift phase - characterizes the Callovian to Early Cretaceous, when the Ligurian Ocean was undergoing extension by sea-floor spreading, prior to the onset of closure in the Late Cretaceous, which culminated in the Alpine orogeny.

Only the pre-rift, rift and the early drift phases that characterize the Lower-Middle Jurassic will therefore be discussed further.

Pre-rift

In the Triassic the Mediterranean-Alpine region was a united landmass, formed by the Hercynian orogeny, and by this time was of low relief, having reached an advanced stage of peneplanation by the end of the Permian [Sander 1970; Trmpy 1971]. Tethys was confined to the east between the Caucasus-Iran and Arabia-India and was already some 2000-4000km wide, extending by northern subduction and southern distension and rifting [Biju-Duval et al. 1977; Sonnenfeld 1981; Dercourt et al. 1986]. In Middle Triassic times, Tethys started to transgress westwards, in response to downwarping of the present Atlantic margins as pre-drift extension commenced. This transgression
To show the palaeogeography of western Tethys during the Pliensbachian (190 million years ago) (from Dercourt et al. [1986]).

Relative positions of the locality areas are marked on these maps, thus:
1 = Peniche sequence, Portugal (note Dercourt has the area marked as land)
2 = Cutri Formation, Mallorca
3 = Vajont Oolite Formation, Italy

Key:
OM = Oran Mesta
AL = Alboran
K = Kabylia
STI = Stilo
SI = Sila
LN = Lago Negro
TR = Tridentin
L = Lombard
B = Brianonnais
VL = Vallais
LAA = Lower Austro-alpine
MAA = Middle Austro-alpine
UAA = Upper Austro-alpine
RH = Rhodope
WP = Western Pontides
DL = Dalmatia
HKA = High Karst
PI = Pindus
PL = Pelagonia
KIR = Kirsehir

orange = land
dark blue = oceanic crust
medium blue = thin continental crust
light blue = thick continental crust
black triangles = breccia
gave rise to the development of increasingly marine conditions, frequently recorded by thick evaporites preceding rapidly accumulating shallow water carbonates [Hallam 1971; Bernoulli 1972; Ogg et al. 1983]. By Middle Triassic times, a shallow epeiric sea existed over the greater part of the European platform [Biju-Duval et al. 1977] and carbonate platform facies were widespread and persisted well into the Lower Jurassic [Bernoulli & Jenkyns 1974]. Growth of these platforms was enhanced by strong subsidence [Bernoulli 1972] that was balanced by increased production rates (up to 100mm/10^3 years) forming widespread, extensive carbonate platform sequences, hundreds to thousands of metres thick.

**Early Rift**

The Upper Triassic marks the commencement of pre-drift extension and rifting that was to shape Western Tethys and the Atlantic margins for the next 35 million years. In the Late Triassic to Early Jurassic, a widespread belt of doming and rifting divided the former Pangaea into Gondwanaland (in the south) and Laurasia (in the north) along the Gulf of Mexico-Central Atlantic-Western Tethys line [Dewey et al. 1973; Savostin et al. 1986]. Tensional features were widespread on both sides of the Atlantic and throughout Tethys [Horvath & Channell 1977]. In Western Tethys, early continental distension, often accompanied by basaltic volcanism, was widespread [Hallam 1971; Laubscher & Bernoulli 1977; Sonnenfeld 1981]. This activity preceded the initiation of continental separation and the formation of the continental margins between Africa and North America, at 190 million years ago [Dewey et al. 1977; Savostin et al. 1986]. An early major activation phase occurred 200-220 million years ago and gave rise to the Cassipore dyke swarm in French Guiana, which correlates with dyke swarms in Liberia [Sadowski 1987].

**Syn-rift**

The initial separation of Africa and North America at 190 million years ago corresponds to the onset of rifting and fault-related subsidence of many areas of Western Tethys, in the Pliensbachian [Trmpy 1971; Bernoulli & Jenkyns 1974; Savostin et al. 1986]. Not all areas were affected at this time, as break-up was diachronous [Bernoulli & Jenkyns 1974]. For example, break-up occurred in the earliest Hettangian in the Eastern Alps, Hungary, Poland and Czechoslovakia [Bernoulli & Jenkyns 1974] and the Southern Alps [Bosellini & Winterer 1981]; while it was of Carixian age in Mallorca [Alvaro et al. 1984; Prescott 1988], and the Betics of Spain [Vera 1988],
Figure 1-3

To show the palaeogeography of western Tethys during the Callovian (155 million years ago) (from Dercourt et al. [1986]) (for Key, see Fig. 1-2).
Tunisia and the Moroccan Northern Rif; and Domerian in the Moroccan High Atlas and Western Sicily [Bernoulli & Jenkyns 1974].

Prior to rifting, shallow-water carbonate platforms continued to flourish over Western Tethys, characterizing the Hettangian and Sinemurian of many regions, for example: the Southern Alps [Bosellini et al. 1981b], the Northern Pyrenees [Sander 1970], the Betics [Vera 1988], the Moroccan Rif [Warme et al. 1988], the Italian peninsular and the Austrian Alps [Sander 1970; Bernoulli & Jenkyns 1974]. Some Hettangian platform sequences may be characterized by syn-sedimentary faulting and fissural mafic volcanism (e.g. the Ragusa zone of Sicily [Pattacca et al. 1979]) or basaltic and tuff intercalations (e.g. the Northern Pyrenees [Sander 1970]) and Mallorca (see Section 3.1.3). Neptunian dykes and sills that penetrate carbonate platform sequences are widespread throughout the Early Jurassic, as indicators of platform extension and fracture prior to rifting and subsidence [Wendt 1971; Jenkyns 1971a; Bernoulli & Jenkyns 1974; Molina 1987].

Continued extension over Western Tethys led to widespread faulting that broke the extensive carbonate platforms into a complex series of fault-bounded tilt-blocks and basins (Fig. 1-4), which underwent differential subsidence as extension proceeded [e.g. Bernoulli & Jenkyns 1974; Lemoine 1982; Winterer & Bosellini 1981; Funk et al. 1987; Lemoine & Trmpy 1987; Tricart & Lemoine 1988]. This facies trend is seen on both the northern and southern continental margins of Western Tethys and is characterized by the development of distinctive, depth (and current) dependent carbonate facies [Aubouin 1963; Hsu 1976; Jenkyns 1970, 1971, 1972, 1974; Bernoulli & Jenkyns 1974; Bosellini & Winterer 1975; Bosellini & Winterer 1981; Jenkyns 1981; Jenkyns & Winterer 1982; Baumgartner 1982 and 1987].

The narrow and deepening seaway between Africa and Europe, created by this extension, was the Ligurian Ocean [Bernoulli & Lemoine 1980; Bosellini & Winterer 1981; Lemoine 1982] (Fig. 1-3). The appearance of the first oceanic crust in the Ligurian Ocean, and therefore the commencement of drift between Africa and Europe, did not occur until the late Bathonian or early Callovian [Lemoine 1983; Baumgartner 1987; Tricart & Lemoine 1988], this event corresponding to the onset of drift in the Central Atlantic [Pittman & Talwani 1972; Dewey et al. 1973; Savostin et al. 1986]. During the preceding syn-rift phase (190-150 million years ago), Africa had moved about 250km left-laterally with respect to Europe, creating in its wake the extension observed over Western Tethys [Savostin et al. 1986; Dercourt et al. 1986].
Figure 1-4

Schematic diagram to show the development of complex block and basin topography developed during the rifting phase in Western Tethys, prior to the onset of drift in the Ligurian Ocean. The formation of faults with NW-SE and NE-SW orientations may reflect the recurring influence of ancient basement structures (from Tricart et al. [1988]).
The evolution of the Ligurian Ocean was governed by extensional tectonics and long-term major strike-slip and transform lateral motions [Bourbon et al. 1977; Biju-Duval et al. 1977; Laubscher & Bernoulli 1977; Lemoine 1980 and 1982; Weissart & Bernoulli 1985; Funk et al. 1987; Tricart et al. 1988]. In this tectonic regime [e.g. Reading 1985], the general downward tract of some fault-blocks was interrupted by episodic stand-still and uplift [Kelts 1981; Jenkyns 1981; Tricart et al. 1988].

Kelts [1981] recognised comparative phases of development between the Tethyan realm and the Gulf of California: the syn-rift Late Liassic to Middle Jurassic of Western Tethys is comparable to the submerged Californian Borderland region. In this area, a complex basin and swell topography has developed, involving transcurrent, fault-bounded, lozenge-shaped basins, characterized by differential sedimentation rates (Fig. 1-4). The second phase of Tethyan evolution, the syn-drift Callovian to Early Cretaceous phase, is compared to the present day Gulf of California. Block displacements became localized and restricted to the spreading axis as sea-floor spreading commenced (Fig. 1-3), with concomitant stabilization of the continental margins. The compressionary phase of Tethyan Alpine range formation in the Tertiary has yet to be matched in the Gulf of California region.

It is likely that the deep-seated Ligurian transform faults played a role in determining the zones of later Alpine thrust tectonics. Lemoine [1982] suggests that Liassic faults bounding tilt blocks, may have been reworked into thrust or strike-slip faults during the Alpine orogeny, this also being suggested for other Tethyan regions by Watts & Garrison [1986]; Teale & Young [1987] and Warme et al. [1988].

1.3.2 Triassic–Callovian evolution of western Iberian North Atlantic margins

The Triassic to latest Callovian also marks a distinctive early syn-rift period in the evolution of the North Atlantic margin basins off Iberia [Wilson 1988; Wilson et al. in press], which was the setting for the Peniche sequence of the Brenha Formation. Rifting in this region was synchronous with the onset of extension over the Western Tethyan realm [Graciansky et al. 1979] in the Late Triassic, initiating a complex series of proto-Atlantic margin basins [Jansa & Wade 1975; Hay et al. 1982; Masson & Miles 1986; Wilson et al. in press]. Sequences developed in the Grand Banks, Newfoundland [Jansa & Wade 1975; Tankard & Welsink 1987] show a strong lithological similarity to those developed in the Lusitanian Basin of Portugal [Wilson 1988; Wilson et al. in press; Watkinson in press].
Similar to earlier Triassic facies development in Western Tethys [Bernoulli 1972], a Carnian-Hettangian early rift phase is characterized by a progression from continental red beds, to marine evaporites, and then to carbonates (refer to Fig. 5-2 for details) in the Early and Middle Jurassic as an epeiric sea was established [Watkinson in press]. The Lusitanian Basin underwent extension from the Late Triassic to Late Jurassic [Watkinson in press], drift not occurring until the Albian in this section of the North Atlantic, north of the Newfoundland-Azores fracture zone (NAFZ) [Masson & Miles 1986]. This extended rift phase, when compared to that of the Ligurian Ocean (Upper Triassic-Callovian, e.g. Lemoine [1982]), is consistent with synchronous rifting and Bajocian sea-floor spreading in the region to the south of the NAFZ [Klitgord & Schouten 1986], and the propagation of the rift zone northwards up into the Celtic and Labrador Seas [Watkinson in press].

The Early-Middle Jurassic marks an episode of reduced extension rate when compared to Late Triassic and middle Oxfordian-Berrian extension rates [Wilson in press]. This episode is considered to have been controlled by thermal relaxation of the lithosphere that followed the Late Triassic rifting and preceded renewed late Jurassic rifting [Wilson in press, Watkinson in press]. These Upper Triassic-Callovian syn-rift facies accumulated in graben and half-graben basins and have relatively simple facies geometries that blanket the rift topography [Tankard & Welsink 1978; Watkinson in press] and are not as diverse as those found in Western Tethys (for details see Section 5.1).

1.3.3 Characteristic facies of the Tethyan Lower and Middle Jurassic

The topography created by extensional rifting created a wide depth range of carbonate depositional environments characterized by some of the most diverse carbonate facies in the geological record. Shallow water carbonate facies continued to flourish or develop on blocks that remained within shallow sea-level depths. These shallow-water facies were strongly affected by tectonic and eustatically induced sea-level fluctuations. They formed source platforms for carbonate sediments that were transported into the basins as turbidites and other gravity deposits [e.g. Bosellini et al. 1981a and b; Halliwell-Hazlett 1988; Bosellini 1989].
Seamount facies

With increased deepening and isolation, a diverse array of "seamount" facies was developed [Jenkyns 1971b; Bernouilli & Jenkyns 1974] (Fig. 1-5). Condensed facies often characterize these settings and current activity played an important role in facies development [Jenkyns 1971a and b; 1974]. **Hardgrounds** characterize non-depositional phases on isolated, current-swept seamounts and slopes; they most frequently directly overlie platform facies, representing the initial post rift sediments on the submerged blocks. For this reason, they are especially characteristic of Domerian-Toarcian over much of Western Tethys [Bernoulli & Jenkyns 1974; Jenkyns 1971a; Prescott 1988]. Hardgrounds are the most condensed of all Tethyan facies, a few centimetres may represent several million years. They are characteristically goethite-rich (although this mineral may be altered to haematite, e.g. Prescott [1988]), with ferromanganese crusts and nodules, and are rich in ammonites and other pelagic fauna. Similar goethite ferromanganese pavements are formed on the modern Blake Bahama Plateau at depths of between 400-800m [Hawkins 1969; Wendt 1974] as well as from high-standing rises in the Gulf of California [Kelts 1981].

*Encrinitic sandwaves* of crinoidal biosparite are classic Tethyan seamount facies, typically occurring stratigraphically placed between the platform sequence and more "basinal" facies [Jenkyns 1971b]. They are strongly depth controlled and may be tidally or current driven. Recent analogues have been described from the Cobb Seamount [Farrow & Durant 1985] where crinoidal-brachiopod sands accumulate at depths of 80-180m, and also from Galicia Bank, where crinoids are found associated with rippled, shell-hash sands [Dupeuple et al. 1976].

*Nodular limestones*, the classic Ammonitico Rosso facies of Aubouin [1964], are one of the most widespread and distinctive Tethyan facies [Jenkyns 1974; Hsu 1976] (Fig. 1-5). They are representative of low sedimentation rates on oxygenated current-swept slopes and seamounts and are, as their names suggests, characteristically red in colour and rich in ammonites. The nodules form by early submarine diagenesis at the sediment-water interface [Jenkyns 1974; Mullins 1980a], and were thought to characterize deep-water sedimentation between A\text{Ly} (aragonite lysocline) and the ACD (aragonite compensation depth) [Bosellini & Winterer 1975; Winterer & Bosellini 1981]. However, these facies have been found associated with stromatolites and oncolitic fabrics, indicating their formation at shallower palaeodepths within the photic zone [Massari 1983].
Figure 1-5

Diagrammatic palaeogeographic evolution of the Tethyan continental margin during the Triassic and Jurassic. Block-faulting and differential subsidence gave rise to seamount and basin topography on which coeval condensed and expanded sequences were deposited (from Bernoulli & Jenkyns [1974]).
Nodular facies are described from modern Bahamian slopes where a downslope trend from hardgrounds to nodular ooze to unlithified ooze correlates with a decrease in current velocity [Mullins et al. 1984]. The presence of ocean currents and low sedimentation rates seem to be the most important factors governing nodular limestone formation. Similar facies also occur on the present-day Mediterranean ridge under conditions of heightened salinity, which led Jenkyns [1981] to propose that some Tethyan nodular facies may have been formed in restricted basins with increased salinity.

_Pelagic oncosids and ooids_ have been described from Tethyan shallow-water seamount areas by Massari [1983]; Massari & Dieni [1983] and Jenkyns [1972] and are frequently associated with stromatolitic domes and found associated with ammonitico rosso facies. They indicate that shallow palaeodepths, within the photic zone existed over some pelagic plateaux.

_Coquinas_ (a term used to describe the accumulation of shells by current action) of pelagic organisms may occur on submerged plateaux. Typical of the Tethyan Middle Jurassic are accumulations of current-winnowed pelagic bivalves known as _Posidonia_ or _Bositra_ [e.g. Kuhry et al. 1976; Sturani 1971; Massari 1983] (see Section 3.3.1.2). Coquinas of ammonites and belemnites have also been described [e.g. Blendinger 1988].

_Basinal facies_

The basin or trough sediments are typically thick, expanded successions of bedded grey pelagic (biomicrite) limestone, often with chert, interbedded with marls and calcareous turbidites [Bernoulli & Jenkyns 1974; Ogg et al. 1983]. Slumped and resedimented slope facies may occur, along with debris flows, resedimented pelagic facies and fault-derived breccias, that reflect continuing extension and block faulting [Bernoulli 1972]. Locally, and especially characteristic of the Toarcian during early basin evolution, some basins were restricted and a combination of the influx of fertile waters giving rise to high non-calcareous plankton productivity and eustatic sea-level rise favoured the deposition of organic-rich shales [e.g. Jenkyns & Clayton 1980; Jenkyns et al. 1985].

Middle Jurassic basinal facies throughout Tethys are characterized by the presence of small pelagic bivalves known as _Posidonia_ or _Bositra_ [e.g. Arkell 1956; Jefferies & Minton 1965; Kuhry et al. 1976], which became "replaced" in upper Kimmeridgian-Tithonian basinal sediments by the pelagic crinoid _Saccocoma_ [e.g. Alvaro et al. 1984;
Bosellini et al. 1981a; Winterer & Bosellini 1981), which were in turn "replaced" by the appearance of nanoplankton in the late Tithonian.

Radiolarites and cherts characterize the Callovian-Oxfordian throughout Western Tethys [Hsu 1976; Jenkyns & Winterer 1982; de Wever & Caby 1981; Ogg et al. 1983; Baumgartner 1987] and their presence indicates the onset of drift in the Ligurian Ocean. The abundance of siliceous limestones and cherts at this time is considered to reflect a change in palaeoceanography: the onset of drift resulted in enhanced subsidence rates, and a corresponding eustatic sea-level rise [Jenkyns & Winterer 1982]. There was increased radiolarian productivity due to the influx of nutrient rich waters as a connection was made between the Atlantic and Tethys [Hsu 1976]. At this time, the CCD (carbonate compensation depth) was considered to be shallow and many of these facies, especially those that are carbonate free, are considered to have been deposited below the CCD [e.g. Winterer & Bosellini 1981]. There is a widespread middle and late Callovian and early Oxfordian hiatus over Tethys, but where rare sediments are found they are generally indicative of reducing conditions. This period corresponds to an episode of rapid eustatic sea-level rise [Haq et al. 1987 and 1989; Hallam 1988] as a result of an early episode of fast sea-floor spreading in the Central Atlantic and Ligurian Ocean [Savostin et al. 1986].

The Late Tithonian saw the blanketing of both deep basin and submerged plateaux by a fine white pelagic limestone, rich in calpionellids, called the Maiolica [Hsu 1976]. This facies blankets most western Tethyan Jurassic sequences at this time [e.g. Hsu 1976; Winterer & Bosellini 1981; Weissart 1981] as well as being found in the Atlantic and East Indies [Hallam 1971a; Bernoulli 1971]. The widespread nature of this pelagic carbonate facies is in direct contrast to the Callovian, and indicates a depression of the CCD [Berger & Winterer 1974]. This period of time was also marked by closure of the Vardar Ocean [which connected Western and Eastern Tethys] and no Maiolica facies are found in Eastern Tethys [Ogg et al. 1983]. This oceanographic change in circulation in Western Tethys and the Atlantic, combined with a net sea-level fall and the evolution of nanoplankton, served to transfer the carbonate budget for the first time, from shallower carbonate shelves into the deep sea, thus causing marked depression of the CCD [Bosellini & Winterer 1975].
1.4 Classification and distinction of submarine fans and aprons

This Section briefly reviews the models currently available in the literature for the interpretation of both siliciclastic and carbonate turbidite systems. Section 1.4.1 covers submarine fans, while Section 1.4.2 covers aprons and 1.4.3 mentions some other models associated with turbidite emplacement.

1.4.1 Submarine Fans

Modern and ancient siliciclastic submarine fans develop in a wide variety of tectonic and geological settings, throughout the geological record [e.g. Nelson & Nilsen 1984; Bouma et al. 1985; Shanmugam & Moiola 1988]. Theoretically, submarine fans are fed by a single major feeder canyon which acts as a point source for all sediment distributed to the fan, as shown by the models of Normark [1970], Mutti & Ricci-Lucchi [1972 and 1974], Mutti & Ghibaudo [1972], Ricci-Lucchi [1975], Mutti [1977], Normark [1978] and Walker [1978]. Therefore unlike aprons, fans are isolated sediment bodies that develop seaward of a major sediment source [Stow 1985a] (Fig. 1-6). Modern and ancient fans show a great variability in shape and size [Barnes & Normark 1985] for example, while the sand-rich Navy Fan is only 15km long, in contrast the mud-rich Bengal Fan is 3000km long.

Fans are characterized by channel and lobe sequences, progradation resulting in thickening and coarsening upward sequences of unchannelised turbidites on lobes; and thinning and fining upward sequences in channels, as a consequence of channel abandonment [e.g. Walker 1978; Mutti & Normark 1987; Shanmugam & Moiola 1988] (Fig. 1-6, A). The validity of thickening-upward lobe cycles has been discussed by Ricci-Lucchi & Valmori [1980] and Hiscott [1981], the validity of fining-upward channel sequences being discussed by Martini & Sagri [1977] and this topic is the subject of discussion in Chapter 4, Section 4.3.2.2.

Many classification schemes have been proposed for submarine fans over the last twenty years or so. Mutti [1979] proposed a classification based on the efficiency of a fan to transport sand over long distances. This was based on the concept that mud-rich fans (such as the Mississippi Fan [Bouma et al. 1985]) are efficient at transporting sand over long distances, whereas sand-rich fans are inefficient, and tend to deposit their sediment load near source (such as the Navy Fan [Normark 1970 and 1978]).
Figure 1-6

A) Model of fan deposition, relating facies fan morphology and depositional environments (from Walker [1978]).

B) Schematic comparison of mature passive margin fans (left) with active-margin and immature passive margin fans (right) (from Shanmugam & Moiola [1988]).

C) Idealized stratigraphic sections through prograde successions in mature passive margin fans (left) and active margin fans (right). Facies nomenclature (letters A to F) after Mutti & Ricci-Lucchi [1975] (from Shanmugam & Moiola [1988]).
### A

**FEEDER CHANNEL**

- **SLOPE INTO BASIN**
- **CONGLOMERATES**: INVERSE-TO-NORMALLY GRADED
- **GRADED-BED**
- **GRADED-STRATIFIED**
- **TERRACES**
- **TERMINAL FAN**
- **UPPER FAN LOBES**
- **UPPER FAN**
- **BRANCHED**
- **THIN BEDDED TURBIDITES ON LEVEE**
- **PEBBLY SSTS.**
- **MASSIVE SSTS**
- **SLUMPS**
- **BEDDING**
- **TURBIDITES**
- **CLASSICAL TURBIDITES**
- **BASIN PLAIN**
- **LOWER FAN**
- **NEW SUPRAFAN LOBE**

### B

**PASSIVE MARGIN**

- **UPPER FAN**
- **MIDDLE FAN**
- **LOWER FAN**
- **ABANDONED CHANNELS**

**ACTIVE MARGIN**

- **DEPOSITIONAL LOBES**
- **BASIN PLAIN**

### C

#### PASSIVE MARGIN

- **Channel-Fill**
- **Overbank**
- **Middle Fan**
- **Sheet Sand**
- **Lower Fan**
- **Basin Plain**

#### ACTIVE MARGIN

- **Channel Fill**
- **Middle Fan**
- **Lobe**
- **Lower Fan**
- **Basin Plain**
Sand-rich fans are characterized by extensive braided mid-fan deposits and a poor development of basin plain and fan fringe deposits [Normark 1970; Link & Nilsen 1980; Link & Welton 1982; Gken & Kelling 1983; Busby-Spera 1985]. Mud-rich fans are characterized by leved fan valleys, sand deposition occurring on channel floors throughout the length of the fan, with channels possessing greater stability (Fig. 1-6, B). Depositional lobes commonly do not develop, lower fan deposits being sheet sands and channels [Shanmugam et al. 1988] and basin plain deposits are typically well developed. Confusion occurred in the literature concerning the usage of the term sand-efficient for mud-rich fans and sand-inefficient for sand-rich fans; and also by the assumption by Mutti [1979] that sand-rich fans had attached lobes and mud-rich fans had detached lobes [Shanmugam & Moiola 1985a].

A classification based on fan shape, sectioned fans into radial fans and elongate fans [Nelson 1983; Nelson & Nilsen 1984], sand-rich fans tending to be small and mud-rich fans tending to be larger and elongate. However, there were many contradictions to this concept, both in the modern and ancient record, and a classification based on fan geometry was often difficult to apply in the ancient record.

Shanmugam & Moiola [1988] recently proposed a four fold classification based on the tectonic setting of the fan system: immature passive margin (North Sea) type; mature passive margin (Atlantic) type; active margin (Pacific) type; and mixed setting type. Small, sand-rich fans characterize immature passive margin and active margin settings, while large mud-rich fans characterize mature passive margin settings (Fig. 1-6, B and C).

Carbonate turbidite systems have been described as fans by Cook & Egbert [1981], Bosellini et al. [1981a and b], Ruiz-Ortiz [1983], Wright & Wilson [1984], Watts & Garrison [1986], Watts [1987], Bernoulli et al. [1988] and Cooper [1989]. In the author’s opinion, convincing evidence for the development of carbonate fans, rather than aprons, has only been provided by the examples described by Cook & Egbert [1981]; Wright & Wilson [1984] and Bernoulli et al. [1988].
1.4.2 Aprons

1.4.2.1 Carbonate apron models

The first apron model to be developed was that of the debris sheet [Cook et al. 1972] (Fig. 1-7, A), which models major, episodic collapse facies associated with platform margins. This model introduced the concept of mass, unchannelized, sheet-flows. Debris sheets are relatively rare but spectacular events in the ancient and modern record; for example the debris sheet that covers the entire floor of Exuma Sound, Bahamas [Crevello & Schlager 1980] and ancient examples such as Cook's Devonian examples from western Canada [Cook et al. 1972; Cook 1983] and the examples of Playford [1980], Davies [1977] and Enos [1977a]. This model was however, too restrictive to demonstrate most features observed in carbonate basin margin sequences.

Cook was the first to propose two carbonate apron models, based on ancient resedimented carbonate sequences (for summaries, see Cook et al. [1983]; Cook & Mullins [1983]; Cook & Taylor [1977]; and Cook & Enos [1977]) first in 1982 and then in Cook [1983]. These models and the apron concept were supported by work in modern Bahamian basins [Mullins & Neumann 1979; Mullins 1983b; Mullins et al. 1984; Schlager & Chermak 1979] and were published as the important paper of Mullins & Cook [1986]. Two end-member models were recognized: (1) the Slope Apron (Fig. 1-7, B); and (2) the Base-of-Slope Apron (Fig. 1-7, C), which are in many ways analogous to the earlier depositional margin and by-pass margin slope models of McIlreath & James [1978].

Slope Apron

In this model (Fig. 1-7, B), resedimented facies may be traced right up to the platform margin, without a steep upper slope by-pass zone. This type of apron is characteristic of margins with smooth, gentle down-to-basin gradients of <4 [Cook 1983; Mullins & Cook 1986]. In this model, randomly distributed mega-breccias, debris flows and turbidites are deposited as broad sheets along the platform margin.

Base-of-slope Apron

In this model, an upper steep slope exists at the platform margin of >4 (frequently much steeper). As a result of this steep upper slope, coarse sediment by-passes the
Figure 1-7

Schematic block diagram models for A) Debris Sheet; B) Slope Apron; and C) Base-of-slope Apron.
DEBRIS SHEET MODEL

(from Cook et al. 1972)

SLOPE APRON MODEL

(from Cook 1983)
slope through a system of gullies, to form a coarse sediment wedge at the base of slope (Fig. 1-7, C). In this model, the upper by-pass slope is distinguished by well-bedded periplatform ooze, cut by small gullies that may be filled with coarse debris. Slumps, slides and truncation surfaces may also be present [Mullins & Van Buren 1981]. Most Bahamian slope facies conform to this model [e.g. Mullins & Neumann 1979; Hine et al. 1981; Davis 1983; Mullins et al. 1984; Schlager & Ginsburg 1981; Schlager & Chermak 1979] (see Section 2.5).

Erosional margins

With steep slope angles and steep high margins, erosional escarpments develop, characterised by mechanical defacement and the accumulation of talus blocks at the base of slope. Although not included in the models of Cook [1983] and Mullins & Cook [1986], they were included in the classifications of Read [1982 and 1985] and are described in more detail in Section 2.5.2.

Carbonate apron facies and their associations

Carbonate apron facies have been divided into inner (proximal) and outer (distal) apron facies belts [Mullins et al. 1984] (Fig. 1-8, B): the inner apron is distinguished by thick mud-supported debris flows and coarse-grained turbidites, the outer apron by thinner clast-supported debris flows and thinner turbidites showing Bouma divisions [Mullins et al. 1984; Mullins & Cook 1986]. Inner apron facies are likely to show evidence of broad, shallow channelization, not the narrow, deeply incised channels typical of clastic fan inner fan facies. A proposed idealised stratigraphic sequence for a prograding base-of-slope carbonate apron is shown in Fig. 1-8, B. Basinal facies are characterized by well-bedded peri-platform ooze (Facies G) interbedded with thin, fine-grained, base-cut-out turbidites (Facies D); the outer apron by unchannelized classic turbidites (Facies C) interbedded with clast-supported conglomeratic turbidites and debris-flows (Facies A) and peri-platform ooze; the inner apron of thick coarse-grained turbidites (Facies A) and thick mud-supported debris flows (Facies F), interbedded with peri-platform ooze [Mullins & Cook 1986; Mullins et al. 1984].

Ancient carbonate aprons (barring the extensive work of Cook mentioned previously) have also been described by Mcllreath [1977]; Kepper [1981]; Jordan [1981]; Hurst & Suryl [1983]; Colacacchi & Baldanza [1986]; Gawthorpe [1986a]; Watts & Garrison [1986]; Halliwell-Hazlett [1988]; Blendinger [1988]; Eberli [1987 and 1988].
Figure 1-8

A) Idealized submarine fan progradational model. Note overall thickening-upward sequence, with smaller scale thickening and coarsening-upward cycles (C-U) and thinning and fining-upward cycles (F-U). CT = classic turbidites; MS = massive sandstone; PS = pebbly sandstone; CGL = conglomerate; DF = debris flow; SL = slumps (from Walker [1978]).

B) Idealized base-of-slope carbonate apron progradational trend. GSDF = grain supported debris flows; MSDF = mud supported debris flows. Letters refer to facies mentioned in the text. (Diagram from Mullins & Cook [1986]).
1.4.2.2 Siliciclastic Aprons

Siliciclastic apron models were proposed at around the same time that the carbonate apron model was being developed. Pickering [1983] defined a fault scarp apron, based on fault-controlled clastic wedges in the California basins and other strike-slip dominated basins [e.g. Field & Clarke 1979; Reading 1980; Norris 1978], later describing such a sequence from the Jurassic of Scotland [Pickering 1984]. Stow et al. [1983/84], Stow [1985a and 1986] proposed a two-fold classification of clastic aprons into slope aprons and faulted slope aprons.

Clastic Slope Aprons

Stow [1985a and 1986] and Stow et al. [1983/4 and 1985] distinguish two forms of slope apron, for high and low energy settings. In the former case, the slope is smooth and contourite drifts develop near the base of slope; in the latter, the slope may be gullied and slump-scarred, with sediment lobes, debris-flow masses and slumps at the foot of slope. Larger feeder canyons may cut across the slope at intervals. Modern examples have been described from the Nova Scotian shelf by King & Young [1977] and Stow [1981]. Numerous small canyons cross the slope and funnel coarse sediment to isolated lobes on the lower slope. Deep currents are active, at depths of 2-4km and contourites are interbedded with the turbidite succession [Stow 1979].

Clastic Faulted Slope Aprons

These aprons are characterized by a thick fault-scarp wedge of sediment that accumulates at the foot of slope. Slides, slumps and short-lived, shallow channels are widespread. Sediment facies vary due to periodic tectonic activity, and faults may develop perpendicular to the margin, creating lateral facies changes. These faults may serve as sites for long-term canyons and channels that funnel sediment out into the basin. Proximal facies may include rock-falls and debris flows, but these die out rapidly away from the fault zone. Intermittent tectonic activity affects the vertical arrangement of facies: thinning and fining upward cycles may characterize each phase of tectonism [e.g. Surlyk 1978; Stow 1985b]. Fault controlled slope aprons characterize strike-slip margins and early rifted margins [Stow 1985a and 1986].

A modern faulted slope apron is the Sardinia apron of the Tyrrhenian Sea [Wezel et al. 1981]. Stow [1985b] reinterpreted the Jurassic Brae oil field reservoir as a faulted slope
apron. The whole sandbody forms a 300m thick thinning and fining upward megacycle, recording apron retrogradation, within which smaller scale megacycles (50-150m) are distinguished, indicating episodic but decreasing fault activity [Stow 1986]. Fault-scarp associated conglomerates and breccias pass basinwards through pebbly sandstones, sandstones, to more mud-dominated facies. Interdigitation and some channeling of these facies occurs, along with marked lateral changes, due to en- eschelon off-set faults, which may have acted as sediment conduits. Similar faulted slope apron systems displaying thinning-upward megacycles have been described from the Jurassic of East Greenland [Surlyk 1978] and from a carbonate faulted slope apron from the Jurassic of the Eastern Alps, Switzerland [Eberli 1987 and 1988].

1.4.3 Some other turbidite models

The other main carbonate slope model is that of the carbonate ramp of Ahr [1973] Ramps have gentle slopes of <1 , on which shallow, nearshore wave-agitated facies pass downslope (without a marked break in slope) into deeper-water, low energy facies [Read 1985]. Homoclinal ramps have uniform, gentle slopes, classically displayed in the Persian Gulf [Purser 1973] and from the Ordovician of Virginia [Read 1980]. Slope turbidites are rare or not well developed in this model. Distally steepened ramps have characteristics of both rimmed shelves and homoclinal ramps: a shelf-slope break occurs some distance seawards of the high-energy shoal belt, and deep-ramp mud blankets may occur seawards of the shoal complex. This type of ramp often forms when a shelf becomes drowned [Read 1985], and gravity flows are more numerous than on homoclinal ramps.

Other siliciclastic turbidite models include fan-deltas, which develop as the subaqueous extensions of alluvial fans [e.g. Pickering 1982; Stow 1985a; Prior & Bornhold 1986]; and delta-fed submarine ramps, which form from multiple sediment entry sources along delta-fronts [e.g. Heller & Dickinson 1985].
1.5 Thesis Structure

This introductory chapter has outlined the aims of the thesis, has set it into a regional Tethyan framework and introduces the concepts of turbidite models, which will be applied in later chapters.

Chapter two reviews carbonate production and sedimentation on modern Bahamian platform margins, and their adjacent slopes and basin floors. This chapter illustrates how the Bahamas can serve as a conceptual model for illustrating specific controls that govern rates of carbonate production and transport into the adjacent basins.

Chapters three to five describe the sedimentology of the three oolitic turbidite systems, which are each interpreted in terms of available models. The role of specific palaeogeographic, tectonic and eustatic controls that govern development of the individual sandbody architecture are discussed in a section at the end of each chapter. Chapter three, which is the largest, considers the Cutri Formation of Mallorca; chapter four the Vajont Limestone of northern Italy; and chapter five the Peniche sequence of the Brenha Formation of western Portugal.

Chapter six briefly discusses some diagenetic aspects of the three sandbodies discussed in the previous chapters, using standard petrographic techniques, cathodoluminescence, ultra-violet microscopy.

Finally, chapter seven compares, contrasts and summarizes the results of the previous chapters, discussing implications of the findings, with an emphasis on the potential of resedimented carbonate sandbodies in general to serve as hydrocarbon reservoirs.
CHAPTER 2

REVIEW OF PLATFORM MARGIN AND TROUGH SEDIMENTATION IN THE BAHAMAS

2.1 Introduction

Over the last ten years, considerable progress has been made in understanding the controls governing carbonate production on modern Bahamian platforms. This has revealed a fundamental control on platform margin sedimentation, by the geographical orientation of the platform margins with respect to the prevailing wind direction and energy flux. Bank margin orientation controls the degree of off-bank transport of carbonate sands and muds, resulting in important differences in the nature of the associated slope and basin facies. Modern Bahamian platform margin and slope to basin sedimentation and its controls are reviewed in this chapter, in order to provide a modern 'standard' for later ancient resedimented carbonate sandbody interpretations.

2.2 Introduction to Bahamian Physiography

Major modern carbonate producing zones lie between latitudes of 30°N and S [Wilson 1975]. Modern carbonate sands (reefs are not dealt with in this review) occur in the Bahamas [e.g Purdy 1963]; the West Florida continental slope [Doyle & Holmes 1985]; the Persian Gulf [Purser 1973]; Campeche Bank, Yucatan [Logan et al. 1970]; Belize [James & Ginsburg 1979] and small oceanic islands, shelves and atolls, such as Bermuda [Garret & Scoffin 1977]. Of all these areas however, the Bahama Banks provide the largest and best developed carbonate sand system in the world. The Bahama platform consists of a series of shallow water, flat-topped carbonate banks separated by inter- or intra-platformal deep water troughs and seaways (Fig. 2-1), and provide a "natural laboratory" for study of platform carbonate sands and associated deep-water trough environments.

Situated along the North American continental margin, the Bahama islands extend 1500km north to south, from Little Bahama Bank which lies off West Palm Beach, Florida, to Navidad Bank, lying off the Dominican Republic; and approximately 450km east to west, from the western edge of Great Bahama Bank to Cat Island, which lies east of Exuma Sound. There are sixteen banks, which collectively support fourteen major islands and thousands of small rock cays [Hine et al. 1985]. The bank tops are
Figure 2-1

Location map of the Bahama Banks showing the location platforms and troughs mentioned in the text (after Hine [1983a]).
Map of Northern Bahamas

(after Hine 1977 & 1983)
flat and covered by shallow seas rarely more than ten metres deep [Mullins 1983b], compared to the intervening troughs where water depths range from 800m in the northern straits of Florida to over 4,000m at the mouth of the Northeast Providence Channel, and off Abaco and Eleuthera Islands which border the open Atlantic Ocean [Schlager & Ginsburg 1981]. Down to basin slopes are highly variable, ranging from 1° off the northwest corners of Great Bahama Bank and Little Bahama Bank, to 30°-60° along the Bahama Escarpment [Mullins & Neumann 1979; Freeman-Lynde et al. 1981]. Schlager & Ginsburg [1981] mapped the present distribution of slopes, basins, canyons and abyssal plains, where sediments are mixtures of carbonate muds and oozes, and turbidites formed both from bank-derived sands and reworked slope facies.

A great deal of research has been undertaken on modern Bahamian sediments and environments since the 1950s, and as such the Bahamas has tended to become the universal analogue for ancient carbonates, becoming a widely accepted standard [Hine 1983a; Schlager & Ginsburg 1981]. However the Bahama platforms are extremely long-lived and are currently at an advanced stage of growth, with very few ancient platforms matching the Bahamas in longevity. Indeed neritic conditions similar to those of today, have persisted in the Bahamas area since the Jurassic [Schlager & Ginsburg 1981]. Sediment facies in the troughs are influenced by the fact that the Bahamas are super-mature, and that the troughs have unusually high and steep flanks [Schlager & Ginsburg 1981; Hooke & Schlager 1981]. Shallow water facies patterns are also strongly shaped by sea-level history in the Holocene, where rapid sea-level rise means that the banks are "lagging behind" present sea-level [Hine et al. 1981; Hine 1983b; Dominguez et al. 1988]. Therefore, some caution is advised when directly comparing ancient carbonate regimes with those of the Bahamas.

2.3 Origin and Geological History of the Bahama Platform

The Bahamas platform is considered to be underlain by continental crust, and the pattern of platform and troughs is thought to be inherited from an Early Jurassic fault-block topography [Mullins & Lynts 1977; Mullins & Neumann 1979]. The concept of an underlying tectonic control on Bahamian physiography was proposed earlier by Talwani [1960] and Ball [1967], the most recent model proposing control by left-lateral wrench tectonics [Mullins & Sheridan 1983; Mullins 1983a].
In this model, constrained by the fact that the strike of any wrench fault would parallel oceanic fracture zones, the N-NW structural grain of the northern Bahamas is interpreted as being controlled by en-echelon folds. As such, Tongue of the Ocean and Exuma Sound may be controlled by folded rather than faulted basement. Superimposed on these folds are E-NE striking normal faults (forming Northeast Providence Channel), W-SW striking synthetic strike-slip faults, with N-NE striking antithetic strike-slip faults [Mullins 1983a]. However, timing of the tectonic activity is still debatable: the structures controlling Bahamian physiography may be inherited from Jurassic rifting along a major transform fault, with recurrent faulting during the Cretaceous to Tertiary, produced as a result of evolving stress systems or by subduction with Cuba. Alternatively, it could be controlled by left-lateral wrenching during the middle to Late Cretaceous, by oblique subduction between Cuba and the Bahamian Platform [Mullins 1983a].

From 150 to 180 million years ago, the facies trend was from clastics to restricted carbonates to normal marine carbonates [Schlee et al. 1979], which are conformably overlain by Late Jurassic to Early Cretaceous dolomite carbonates and anhydrites. These latter sediments correlate over an area of 150,000 km² from the Bahamas to Florida and Cuba [Sheridan et al. 1981] and represent development of a carbonate "megabank" over the region [Schlager & Ginsburg 1981]. The early Cretaceous saw partial drowning and segmentation of the megabank, which correlates with widespread drowning of other Atlantic-Caribbean platforms at the same time [Schlager 1981]. Pelagic mud drilled from Northeast Providence Channel shows that the troughs had been established by mid-Cretaceous times [Schlager & Ginsburg 1981].

Initially there was little relief between the troughs and banks [Hooke & Schlager 1980], but this was accentuated by platform upbuilding as the platforms subsided. The pronounced relief between the troughs and banks seen today is due to a combination of platform upbuilding and erosion by turbidity currents in the troughs [Ball 1967; Mullins & Lynts 1977; Schlager & Ginsburg 1980; Hooke & Schlager 1981]. During the Late Cretaceous to Holocene, 3000m of carbonate sediment accumulated on the platforms, compared to only 300m of coeval trough sediments [Goodall & Garman 1969]. There is evidence for minor post-break up down-faulting around Abaco Knoll [Mullins 1976], with Cenozoic faulting documented by Mullins & van Buren [1981].

An abrupt change occurs during the Late Plio-Pleistocene [Beach & Ginsburg 1980], which coincides with the onset of glaciation in the northern hemisphere [Schlager &
Ginsburg 1981]. The reef limestone-rimmed windward and leeward platform margins, with internal platform sediments, changed from being dominated by skeletal carbonates to non-skeletal grains (peloids and ooids), which dominate the Holocene. This change from atoll-like platforms with deep lagoons, to the flat-topped platforms of today, was due to a change in rate of sea-level rise, or its relative rise [Schlager & Ginsburg 1981]. Oolitic aeolianites developed preferentially along windward margins in the Pleistocene and form the present-day Bahamian islands (Fig. 2-1) [Beach & Ginsburg 1980].

More recently, the geology of the Bahamas has been shown to be more complex than previously thought. Ocean drilling results indicate that certain bank margins have retreated and prograded tens of kilometres in the past 100 million years [Scientific Party 1985; Austin et al. 1986]. Eberli & Ginsburg [1987, 1988 and 1989] demonstrate that the Great Bahama Bank was created by the "welding" of three smaller platforms by a combination of early aggraded trough infill and subsequent progradation of the platform margin. In the mid-Cretaceous, a north-south fault-bounded depression, known as the Straits of Andros, separated Andros Bank to the east from Bimini Bank to the west. Initially the Straits of Andros had similar dimensions to the Tongue of the Ocean, the early phase of rift infill preceded margin progradation and led to a shallowing of the depression. Progradation commenced when the slopes were 500m high and/or had angles of <5° [Eberli & Ginsburg 1989] and progradational episodes progressively infilled the seaway from east to west, to form the united Great Bahama Bank.

Progradational pulses occurred as the result of rise and highstand of sea-level, when sediment production was high and off-bank transport of carbonate sand was increased (see Section 2.3.5). During this time, overproduction of sediment on the platform resulted in lateral outbuilding, which outpaced vertical aggradation on the platform more than ten-fold. In early Tertiary times, an embayment formed by folding within Bimini Bank and this was also infilled from east to west. This research demonstrates how tectonic deformation can affect large carbonate platforms despite their passive continental margin setting, as well as showing how the levelling of an initial structural depression was controlled by sea-level fluctuations.

Comparisons with the Tethyan Mesozoic

By analogy with Tethyan carbonate platforms, the Bahamas show similarities by having a similar underlying horst and graben tectonic control; however compared to Tethyan
examples, the Bahamas show greater longevity. The initial trend from clastics to restricted carbonates to marine carbonates is mirrored in many passive margin sequences, including the Permo-Mesozoic of the Mediterranean-Tethyan realm (see Section 1.3.1) and Atlantic margin basins (for example, the Lusitanian Basin, see Section 5.1.3 and also Wilson [1988] and Wilson et al. [in press]). The Bahamian megabank stage lasted 50-60 million years, with the platform-trough stage existing for 100 million years before the present. By comparison, in the Austro-Alpine Mesozoic, neritic carbonates persisted for 10 million years, with a platform trough stage of 22 million years [Schlager & Ginsburg 1981]. The work of Eberli & Ginsburg [1987, 1988 and 1989] shows some analogies with the Tethyan Jurassic, where rift topographies were initially infilled and smoothed by carbonate resedimentation (see Section 4.4.1).

2.4 Bahamian Platform Margin Carbonate Sandbodies

2.4.1 Characteristics and classification of Bahamian platform margins

The Bahama Banks lie within the trade wind belt, with wind and wave power coming dominantly from the east [Purdy 1963; Hine & Neumann 1977; Mullins & Neumann 1979; Hine 1983] (Fig. 2-1). Easterly facing margins (NE, E and SE) are termed windward, since they face the prevailing wind, while westerly facing margins (NW, W and SW) are termed leeward, as they oppose the prevailing wind direction [Hine & Neumann 1977; Hine et al. 1981b; Tucker 1985]. Where tides are amplified by embayments or topographic constrictions, strong tidal currents mask or overprint the regional windward or leeward influence [Mullins & Neumann 1979; Hine 1983a; Hine et al. 1985].

Windward margins absorb high-energy, deep-water wave energy and the diversity of bank margin sandbodies is due to the level, duration and magnitude of the physical energy flux across the platform edge [for example, Hine 1977; Hine et al. 1981a and b; Hine & Mullins 1983]. This means that along open windward margins, sands migrate onto the bank top (Fig. 2-2, A); the bank-ward orientated, normal to flow oolitic sandwaves of Lily Bank (northern Little Bahama Bank) being a prime example [Hine 1977, 1983b]. Conversely, along open leeward margins, sandwaves are orientated off-bank and sands are actively being transported to the bank-edge (Fig. 2-2, B) for example along the western margins of Little and Great Bahama Bank [Hine et al. 1981a and b; Hine 1983b].
Rock ridges or islands act as physical barriers to water exchange on and off the platforms, and such margins are termed protected [Mullins & Neumann 1979; Hine et al. 1981b]. Along such windward margins, the barriers block the bankward transport of carbonate sand and protect the bank interiors, directing the energy back seawards and thus augmenting off-bank sand transport [Mullins & Neumann 1979; Hine et al. 1981b]. Along protected leeward margins, the energy barriers prevent the vigorous off-bank sand transport that occurs when these margins are open [Hine et al. 1981]. Tide-dominated margins cause the windward/leeward effects to be masked [Hine & Mullins 1983]. Sandbodies along these margins are orientated parallel to flow, and net sand transport is bankwards [Schlager & Chermak 1979; Hine et al. 1981].

Bahamian margins are therefore most simply divided into the five categories discussed above: (1) windward open (e.g. northern margin of Little Bahama Bank); (2) windward protected (e.g. Berry Islands and the south-east margin of Little Bahama Bank); (3) leeward open (e.g. eastern margin of Great Bahama Bank); (4) leeward protected (e.g. south-east corner of Little Bahama Bank, Great Abaco Island and the south-westerly facing margin of Great Bahama Bank [Hine et al. 1981]; (5) tide dominated (e.g. the south-west margin of Little Bahama Bank between Grand Bahama Island and Great Abaco Island, and the Cul-de-Sac of Tongue of the Ocean [Hine et al. 1981b; Hine 1983]) (for their location see Fig. 2-1). (Mullins & Neumann [1979] proposed an earlier, more complicated classification of bank-margin types than the one given here.) Island dominated windward margins export more sand to the deep than protected leeward margins, so the order of "export potential", from most to least, is (1) open leeward; (2) protected windward; (3) protected leeward; and (4) open windward.

Bank margin orientation and the presence or absence of islands can therefore be seen to have a profound control on associated margin sandbodies. This is of fundamental importance since the direction and magnitude of off-bank sediment transport (and oceanic circulation) have the greatest effect on the type of deep bank margin type that is produced [Mullins & Neumann 1979]. These factors in turn control the amount of shallow water sediment available for downslope transport by gravity processes: along shallow windward margins where dominant sand transport is directed onbank, sands are not an important sediment source to the deep. Conversely, along shallow leeward margins (where not blocked by islands) there is a net off-bank transport of sediment and the deep bank margins and basins contain greater quantities of sand grade sediment [Mullins & Neumann 1979; Hine et al. 1981b] (Fig. 2-3).
Figure 2-2

Diagram to show sand transport directions on leeward and windward margins and the resulting nature of their slope deposits (adapted to show the nature of slope deposits, from Mullins & Cook [1986]).
LEEWARD

off-bank sand transport

WINDWARD

on-bank sand transport
2.4.2 Sedimentation on windward platform margins

The nature of platform margin carbonate sands along windward platform margins is illustrated by the following examples. The Lily Bank oolite shoal (Fig. 2-1), located on the northern edge of Little Bahama Bank, is an example of a modern active oolite shoal migrating bankwards on an open windward margin [Hine 1977 and 1983b].

Lily Bank is located adjacent to two broad re-entrants within the margin, which allow active exchange of water on and off the bank top [Hine et al. 1985]. The active sand shoal forms in <2m of water, and the grains move in response to daily tides. Seawards and surrounding Lily Bank are extensive areas of relict, intermittently active bedforms which lie in water depths of 4-6m [Hine 1983b]. These are stabilized by calcareous algae and sea grass. The active shoal zone consists of a number of bankward orientated, spill-over lobes, which extend out bankwards into the lagoon. These formed during storms, when on-bank sand transport is accelerated [Hine 1983; Hine et al. 1981b]. Between the lobes are symmetrical sandwaves which respond to both ebb and flood tides [Hine 1977; Hine et al. 1985].

Joulter's Cay oolite shoal is located along a less active, tide-dominated embayment of the open windward margin north of Andros Island, on Great Bahama Bank [Harris 1983]. The shoal is a vast stabilized sand flat, fringed on the ocean facing border by a narrow mobile sand rim (1-2km wide and 25km long) of oolite grainstone, which is cut by tidal channels. Bankward of the grainstone rim where ooids are forming, the sand-flats are characterized by ooid packstones and fine-peloidal packstones, which have formed as a result of ooids being mixed with bank-interior fine-peloidal packstone by burrowing [Hine et al. 1985].

In contrast, the northern (windward) margin of Great Bahama Bank is a sediment barren, bare rock surface [Hine & Neumann 1977; Hine et al. 1981b]. Due to its low topography, rapid eustatic sea-level rise caused the shoreline to transgress the 20km wide bank in only 150 years. Reefs never developed, both due to the rapid inundation of the bank top and to the lack of protection afforded from the cold bank top waters. What little sediment that was produced was transported bankwards, to leave the bare, rocky substrate (Fig. 2-2, A). Sporadic, sparse bedforms, now relict and under 20m of water characterise this margin and are only active during high-energy events. Since this margin was swept clean of sediment, it was unable to build vertically with sea-level and remains sediment barren, up-building lagging behind sea-level rise.
Figure 2-3

A) Schematic cross-section through the windward margin of northern Little Bahama Bank. Note the onbank direction of sand transport of oolitic sands.

B) Schematic cross-section through the leeward margin of western Little Bahama Bank. Note marked accumulation of bank margin sands due to the off-bank directed sediment flux.

C) Growth history of an open leeward margin in relation to sea-level rise.

(N.B. All diagrams after Hine and Neumann [1977] reproduced from Hine et al [1985])
WINNOWING, CEMENTATION MODIFIED, SUBDUED BEDFORMS

NORMAL CONDITIONS

NEWLY BURIED CEMENTED HORIZON

STORM CONDITIONS

SAND TRANSPORT TO BANK MARGIN

BEDFORM DEVELOPMENT

SAND TRANSPORT TO DEEP

OFFBANK TRANSPORT TO DEEP

SAND TRANSPORT

PRESENT

RISING SEA LEVEL

DEVELOPED FRINGING REEF

7,000 YRS. B.P.

RISING SEA LEVEL

10,600 YRS. B.P.
The southern margin of Little Bahama Bank (fronting Grand Bahama Island) faces south-east and is a protected windward margin, as it faces strong south-east winds and waves [Hine et al. 1981b]. Vigorous off-bank sand transport occurs during storms, due to return flows induced by Grand Bahama Island. Laterally indistinct sand lobes are perched on the bank edge, opposite chutes in the marginal escarpment, which actively funnel sands to the deep [Hubbard et al. 1976; Hine 1983a].

2.4.3 Sedimentation on leeward platform margins

Leeward margins face towards the west away from the dominant wind/wave energy flux, and commonly have an unobstructed exchange of water on and off the bank tops [Hine & Neumann 1977; Hine et al. 1981a and b; Hine 1983b]. The net transport of sand is off-bank over the bank margin and into deeper water [Hine & Neumann 1977; Hine et al. 1981a]. Large scale sandwaves such as those orientated off-bank along the eastern margins of Little and Great Bahama Banks, migrate only in response to storms, normal wave and tidal activity merely winnowing and sorting the carbonate sands [Hine & Neumann 1977; Hine et al. 1981b]. The sandwaves are though to have formed during the early stages of bank-top flooding in the Holocene [Hine 1983b] when sand movement was more vigorous and pervasive than under present higher sea-level stands. This off-bank transport system has brought large volumes of sand to the bank edge, where they form thick accumulations (15-20m) that cover the reefs that formed earlier during the initial Holocene sea-level rise [Hine & Neumann 1977; Schlager & Ginsburg 1981; Hine et al. 1981a; Hine 1983b] (Fig. 2-3, B and C).

Seismic profiles of these sandbodies show that they contain a vertically stacked sequence of parallel reflecting horizons [Hine & Neumann 1977; Hine et al. 1981a]. These horizons are thought to represent submarine cemented layers that formed during non-storm conditions, since when these horizons outcrop, large tabular plates of grainstone cemented by submarine aragonite druse are formed [Hine et al. 1981a]. These cemented layers form at the surface when the grains are relatively stable and where winnowing effectively removes fine-grained material, increasing permeability and stimulating cementation (Fig. 2-2, B). During storms, sands were carried to the bank edge, covering this cemented, hardened surface. It is thought however, that cemented and uncemented units probably represent periods of non-storm and storm activity, rather than any one storm event, due to there being about 16 cemented horizons compared to the potentially large number of hurricanes (3200) affecting the bank-top during the Holocene. Tropical storms or hurricanes have affected this
portion of the Bahamas at a rate of 8 per 10 years, making them geologically speaking extremely significant and frequent events [Hine 1977; Hine et al. 1981b].

The sediments along these leeward margins are generally grainstones and packstones, consisting of mostly micritized ooids, superficial ooids, peloids and composite grains, all generally significantly bored and micritized, and contain up to 50% skeletal grains (mainly Halimeda) [Hine 1983b; Hine et al. 1985]. Much of north-western Little Bahama Bank and northern Great Bahama Bank are covered with large scale sedimentary bedforms (maximum size 10km long x 5m high x 1km apart [Hine 1983]) with sparse to dense covers of sea-grasses and calcareous green algae, showing that they are no longer as active as earlier in the Holocene [Hine 1983b].

Cat Island platform is a small leeward bank on the eastern fringe of Great Bahama Bank (Fig. 2-1), which shows a contrasting morphology to leeward margin of Little Bahama Bank (see previous Section). The margin sandbodies are only 4m thick and are relict, lying under 20-30m of water [Dominguez et al. 1988]. Due to the presence of deep water reentrants, subtidal sand shoals form bathymetric highs which strike parallel to the margin. The sediments are highly micritized and virtually mud-free. There is a paucity of bank-derived sands along the adjacent deep water slope in Exuma Sound [Crevello et al. 1984], contrasting that of the open leeward margins of Little Bahama Bank [Hine et al. 1981a]. The nature of this leeward platform margin was controlled by rapid eustatic rise and the inability of platform sedimentation to catch-up (see Section 2.4.5).

2.4.4 Sedimentation on tide-dominated margins

These margins form where basin shape amplifies tidal currents [Mullins & Neumann 1979]. Windward and leeward effects are diminished along these margins, and tidal processes quickly mask the effects of storms [Hine 1983a; Hine et al. 1985]. Along the south-west margin of Little Bahama Bank and in the Cul-de-Sac of Tongue of the Ocean and Exuma Sound, bank margins are characterized by long linear sandbodies and intervening channels, oriented normal to the margin edge. These are formed of oolite sands and are covered by active sandwaves. Off-bank sand transport would appear to be limited along these margins [Hine et al. 1985]. However, Hoskin et al. [1986] documented off-bank transport of bioclastic sands in response to storms, through bedload chutes in the rocky margin that lies off south-west Little Bahama Bank. These chutes are 1m wide and 400m long and average 175 chutes per kilometre along strike.
There is substantial off-bank transport of platform muds, which form a muddy apron at the base of a steep by-pass slope. Coarse sediments cut across the muddy apron depositing fingers of coarse carbonate sand and gravel out onto the slope.

2.4.5 Response of bank margins to Holocene sea-level rise and its implications

As already mentioned in previous sections, the effects of Holocene sea-level rise on Bahamian margin sedimentation have been substantial. The Bahamian carbonate platforms classically demonstrate how minor sea-level fluctuations of only a few metres considerably affect facies belt distribution, carbonate sand production rates and rates of off-bank sand transport [Hine 1977; Hine et al. 1981a and b; Hine 1983a and b; Mullins 1983b; Hine & Steinmetz 1983; Dominguez et al. 1988].

Between 3,000-4,700 B.P., the bank tops were flooded with a minimum of 2m (4,700 years B.P.) and a maximum of 6m (3,000 years B.P.) of water [Hine 1983b]. Facies developed were water depth (i.e. energy flux) dependent. When water depths were quite shallow (<1m), wave and tidal activity were minimal, there was a reduction in the sedimentation rate and winnowing occurred of the pellet and skeletal-rich muds. The major bedforms and sheets of oolitic grainstone formed during the period of time when water depth was between 2-5m and rate of sea-level rise had slowed, thus prolonging high bottom turbulence [Hine 1983b; Hine et al. 1985]. As the seas increased in depth by the last 2-4m to reach their present level, the intensity and duration of sediment motion was reduced, gradually allowing benthic organisms to colonize and stabilize the sea-floor. Today, constant sand movement occurs only in areas that have build vertically with sea-level rise and are tidally driven, like Cat Cay and Lily Bank ooid shoals [Hine 1977, 1981a and b; and 1983b].

From this information, important and subtle sea-level depths can be seen to control both on-bank carbonate production rates and off-bank transport rates to the deep (Fig. 2-2, C). Sand transport may be reduced during both high-stands and low-stands of sea-level, for different reasons. During high-stands, when >10m of water covers the bank top, sand transport is greatly reduced, the bedforms become relict [Hine 1983b] and only storm suspended fines are transported off-bank [Neumann & Land 1975; Boardman 1978; Mullins et al. 1984]. Off-bank sand transport is ineffective in water depths of <1m [Hine 1983b], or (obviously) if the platform becomes subaerially exposed [Mullins 1983b and d]. Efficient carbonate sand production and offshore transport are most effective when a slow eustatic sea-level rise maintains water depths of 2-5m over
an actively up-building platform [Hine & Neumann 1977; Hine 1983b], allowing efficient sand transport to the bank margins. Sediment budget studies on the Bahamas have shown that during times of sea-level rise there is an "overproduction" of shallow-water sediment by a factor of 1.5 to 3 times [Neumann & Land 1975]. As the Holocene Bahamian history has shown, it is the rate of sea-level rise that is the crucial factor, since very rapid rises over short periods may exceed the carbonate growth potential [Kendall & Schlager 1981], causing the platform to "lag behind".

A short pulse of rapid rise 8,000–10,000 years ago (of 6m per 1,000 years) caused the incipient drowning of Cay Sal Bank, which was of low elevation. It remains partially drowned today, being under too great a water depth to "catch up" [Hein & Steinmetz 1983]. Similarly, Cat Island platform (also see Section 2.3.3) was flooded at the same time (8,100 years ago), submerging the shelf in 500 years at a flooding rate estimated at 4m per 1,000 years [Dominguez et al. 1988]. After initial start up and development of a reefal rim, off-bank sand transport was initiated, which then inundated the reef. As sea-level continued to rise, off-bank transport became insufficient to maintain margin upbuilding and relict sandbodies of a degraded nature now occur under water depths of 20–30m. A slower net rise, as exemplified by the later flooding of the leeward margin of Little Bahama Bank [Hine & Neumann 1977; Hine et al. 1981a] favours the accumulation of sands at the bank margin, and active off-bank sand transport.

This 'selective drowning' of specific platforms implies that carbonate platforms do not drown synchronously over widespread areas, as commonly assumed [Kendall & Schlager 1981; Bova & Read 1987]. Rather, different portions of platforms respond independently, controlled by their size and initial elevation (which may vary tectonically); the isolated banks tending to respond as independent blocks [Dominguez et al. 1988].

Sand-shedding to the deep appears to be periodic [Schlager et al. 1976; Schlager & Chermak 1979] and is most effective upon initial sea-level rise and on lowering prior to exposure [Hine 1983a; Mullins 1983d]. Sea-level effects on the platform should be mirrored on the adjacent slopes, which record the changes in off-bank sediment flux [e.g. Schlager & Ginsburg 1981; Mullins 1985] caused by eustatic variations. Resedimented carbonate sandbodies therefore preserve a record of relative sea-level changes and their effect on the source platform.
2.5 Bahamian Troughs - their slope and basin characteristics and facies

2.5.1 Trough morphology

The deep, narrow troughs that dissect the carbonate platforms are the most spectacular geomorphological features of the Bahamas. Although tectonics controlled trough location (Section 2.2) it is not believed to have contributed significantly to the present day relief [Hooke & Schlager 1980; Schlager & Ginsburg 1981]. The troughs which may be over 1km deep, remained relatively unexplored until the 1970's, when they were seismically profiled, piston cored and explored with the submersible ALVIN [e.g. Mullins & Neumann 1979; Mullins et al. 1979; Schlager & Chermak 1979; Crevello & Schlager 1981; Mullins et al. 1982 and 1984; Mullins & van Buren 1981; Schlager & Ginsburg 1981; Freeman-Lynne et al. 1981]. Earlier piston core work in the large basins, the Columbus Basin [Bornhold & Pilkey 1971] and Hispanola-Caicos Basin [Bennets & Pilkey 1976] as well as the narrow Tongue of the Ocean [Rustnak & Nesteroff 1964] remain valuable works on the geometry of individual carbonate turbidite layers and are discussed in Section 4.3.4.

Some troughs are flat-floored, U-shaped basins, for example Exuma Sound [Crevello & Schlager 1981], Southern Tongue of the Ocean [Schlager & Chermak 1979] and Columbus Basin [Bornhold & Pilkey 1971], whereas others are deeply incised V-shaped, canyon-like troughs such as the Providence Channels and the northern end of Tongue of the Ocean [Schlager & Ginsburg 1981] (Fig 2-1). Both types have gulley-riddled slopes, the slope angle of the U-shaped basins tends to be less, turbidites being deposited on the lower slopes and basin floor, whereas the steeper V-shaped canyon troughs act as by-pass slopes [Schlager et al. 1976; Schlager & Chermak 1979; Hooke & Schlager 1980; Schlager & Ginsburg 1981].

2.5.1.1 The nature of carbonate slopes

Schlager & Camber [1986] measured five hundred slope profiles along carbonate platforms and siliciclastic continental margins and demonstrated that modern carbonate slopes are complex features that are morphologically distinct from their siliciclastic counterparts [also Schlager 1989]. This is due to the fact that carbonate shelf-slope breaks are characterized by in situ carbonate sediment production and are subject to early diagenesis [Hine & Mullins 1983; Mullins 1983b; Austin et al. 1986], which favours up-building at the platform margin. The precise nature of the shelf margin is
controlled by tectonism, antecedent topography, physical energy flux (tides, waves, storms, currents) and sea-level oscillations [Hine et al. 1981b].

Due to their potential for platform margin up-building and early diagenesis (which reduces slope failure), carbonate slopes can be significantly steeper than their siliciclastic counterparts [Schlager & Ginsburg 1981; Schlager & Camber 1986; Schlager 1989] (Fig. 2-4, A). Indeed, in 90% of all cases, carbonate slope angles exceed those of siliciclastic slopes [Schlager 1989]. Carbonate slope angles increase with height, siliciclastic slope angles do not and carbonate slopes tend to have a concave profile, siliciclastic slopes being straight or slightly convex [Schlager & Camber 1986; Schlager 1989]. Carbonate slope sediments lithify faster than clay or nanno-ooze under burial and sea-floor lithification occurs when carbonate sediments are not continually buried [Schlager & James 1978; Mullins et al. 1980; Mullins 1983b] also becoming more extensive as slopes steepen [Schlager & Camber 1986]. There is also some evidence to suggest that carbonate sediments have a higher shear strength than siliciclastic muds, even in an un lithified state [Schlager 1989]. Dissolution may affect deep carbonate basin slopes and floors [e.g. Berger & Winterer 1974], siliciclastic slopes not being affected at all by this process.

On submarine slopes, deposition is controlled by slumping and other gravity flow transport mechanisms [e.g. Nardin et al. 1979]. An increase in slope angle shifts the activity of the flow from depositional to erosional [Schlager & Camber 1986], flow power being proportional to the 1.5 power of the slope [Allen 1968; Schlager 1989]. On the most gentle slopes, sediment is trapped on the slope itself and sedimentation rates decrease with distance from source. As the slope steepens, the depocentre migrates away from the slope and on to the basin floor. Schlager & Ginsburg [1981] and Schlager & Camber [1986] proposed a three-fold classification of carbonate slopes based on the principle of increased slope angle with height, the sediment budget changing from accretionary to by-passing to erosional as the depocentre migrates in response to increased angularity of the slope (Fig. 2-4, B). The Bahamas platforms are at such an advanced stage that nearly all the slopes are of by-pass or erosional types (Fig. 2-4, A).

There are four major geomorphic zones that occur on the transition from platform margin to basin floor, common to all margin types:

(1) Marginal escarpments rim the banks as precipitous slopes extending from 20-40m of water down to 150-180m [Mullins 1983b]. These escarpments originate due to
vertical accretion as a result of Pleistocene sea-level fluctuations [see also James & Ginsburg 1979 - Belize]. Peri-platform sands and talus may characterize the base of the escarpment.

(2) The upper gullied slope (4 - 15°) commonly extends between depths of 300-500m, and is cut by numerous gullies, which tend to be evenly spaced along the slope [Mullins et al. 1984]. These gullies are thought to form as a function of slope angle by turbidite erosion, larger ones may have been initiated along submarine slide scars [Mullins 1983b]. They range in size from a few tens to hundreds of metres wide and 20-150m deep [Schlager & Chermak 1979]. Hardground facies, peri-platform ooze and nodular ooze are characteristic upper slope facies (Fig. 2-5).

(3) The lower slope is broad, smooth and gentle (0.5 - 2°) and characterized by the deposition of gravity deposits (turbidites and debris flows) as slope parallel aprons [Mullins & Cook 1986].

(4) The basin floor is typically smooth and flat, with well bedded pelagic ooze and distal turbidites [Schlager & Ginsburg 1981].
Figure 2-4

A) Slope profiles of Bahamian platform margins, illustrating the increase of slope angle with relief (from Schlager & Ginsburg [1981]).

B) Slope profiles and processes for accretionary, bypass and erosional slopes (from Schlager & Ginsburg [1981]).
Straits
EROSION/ON
Turbidity currents
Contour currents
Tongue
Exuma
Eleuthera
Eleuthera E
BY-PASSING
Mud, hardgrounds, gullies w. sand
Turbidites onlap
ACCRETION
Slumps, gravity flows
Erosion
By-passing
Accretion
2.5.1.2 Introduction to lower slope and basin facies

Lower slope facies are predominantly gravity deposits interbedded with peri-platform ooze [e.g. Schlager & Chermak 1979; Mullins et al. 1984]. Current reworked sediment drifts and lithoherm facies also occur and these are discussed in context in Section 2.4.3. Facies distributions for the northern Bahamas are shown in Fig. 2-5.

Peri-platform Ooze

The term peri-platform ooze was introduced by Schlager & James [1978] for the fine-grained carbonate sediments that accumulate along peripheral margins of carbonate platforms. They may be found on any part of the modern carbonate slope, but are most important on the upper gullied slopes, becoming volumetrically less important where they are "diluted" by gravity flows [Schlager & Chermak 1979; Mullins & Neumann 1979; Mullins 1983].

Modern peri-platform ooze consists of aragonite, calcite and magnesian calcite [Boardman et al. 1986], which distinguishes them from deep sea pelagic oozes, which are all calcite, and shallow water sediments, which are essentially aragonite and magnesian calcite [Mullins 1983b]. Boardman [1978] demonstrated that in Northwest Providence Channel, 80+% of the peri-platform sediments are bank derived, as a result of sediment over-production on the surrounding platforms [Neumann & Land 1975], this sediment being put into suspension by storms and carried off-bank by tidal currents [Boardman 1978; Hine et al. 1981a]. These facies may be found up to 120km from open ocean bank margins [Mullins 1983b].

Modern oozes are commonly homogeneous and bioturbated [Schlager & Chermak 1979; Mullins 1983] but lithified equivalents on the walls of erosional canyons, such as Great Abaco Canyon [Mullins et al. 1982] are characteristically well bedded, as are more ancient examples, such as the Middle Cambrian Cooks Brook Formation, Newfoundland [McIlreath & James 1978]. It seems that this "bedding" is diagenetic and is controlled by cyclic variations in diagenetic potential produced by climatic changes and corresponding oscillations of sea-level. Kier & Pilkey [1971] suggested that the cyclic variations in aragonite and magnesian calcite are controlled by increased input of these minerals during relative high-stands of sea-level. Droxl er et al. [1983] demonstrated a correlation between aragonite cycles and oxygen isotope curves, relating the cycles to
Figure 2-5

Sediment facies distribution map for deep water areas of the Northern Bahamas (from Mullins [1983b]).
climatic variations tied to glacial rhythms of the earth's climate. Boardman et al. [1986] consider that the peri-platform ooze records Quaternary sea-level fluctuations.

**Gravity flows**

Gravity flows are volumetrically important facies of the lower slope and basin floor [Rustnall & Nesteroff 1964; Schlager & Chermak 1979; Crevello & Schlager 1980; Mullins et al. 1984; Mullins & Cook 1986], although they are also found lining slope gulley axes [e.g. Schlager & Chermak 1979]. Sand-silt grade carbonate turbidites are the dominant facies, followed by debris flows [Mullins 1983b]. Gravity flows are fed to the lower slopes and basin floor via numerous gullies, to form a basin margin parallel apron of coalescing turbidites [e.g. Mullins & Cook 1986] which show broad proximality-distality trends. (Individual turbidite morphology is dealt with in Section 4.3.4). A major debris flow covers the entire floor of Exuma Sound to a depth of 2-3m [Crevello & Schlager 1981] and represents catastrophic mass slope failure (see the aforementioned Section in Chapter 4).

### 2.5.2 Slope and basin facies fronting windward margins

Windward margins are dominated by onbank transport of carbonate sands [Hine & Neumann 1977; Hine et al. 1981b; Hine 1983; Tucker 1985], which causes their slope and basin facies to be deprived of resedimented carbonate sand (Fig. 2-2). Slope sediments are characteristically peri-platform ooze, transported off-bank as suspension loads during storms [Boardman 1978; Hine et al. 1981b; Heath & Mullins 1984].

Very high-energy, windward slopes front Eleuthera and Cat Islands, seawards of which lies the precipitous Bahama Escarpment and the Atlantic Ocean [Freeman-Lynde et al. 1981; Freeman-Lynde & Ryan 1985]. There appears to be very little off-bank transport at all along this erosional type margin. The slope is steep (15 - 40°) and drops to the abyssal depths (4,500m) of the Atlantic Ocean. The escarpment is sediment barren, and is formed of cemented Cenozoic platform limestones which are extensively jointed below depths of 2650m. The joint faces form cliffs, below which are talus deposits. A second joint set localizes gullies, which dissect the escarpment and are filled with fine sediment and talus blocks. After 2670m, a smooth apron of pelagic sediments replaces the jagged, fractured slope. In the area of Cat Island the Bahama Escarpment has two salients, separated by a submarine canyon, which appears to funnel sediment down to a "fan" (1km³) at the canyon mouth.
The northern margin of Great Bahama Bank faces the open seaway of North-west Providence Channel and is not directly exposed to the effects of the Atlantic [Mullins & Neumann 1979]. The margin is broad and sediment barren, the slope is broad and gentle with an average slope of only 1.5°. Current winnowed, peri-platform sand occur high on the upper-slope, while the slope sediments are predominantly carbonate muds, along with muddy slope breccias. Turbidites were only recovered from a gully axis and distal facies are peri-platform ooze and thin distal turbidites.

The northern margin of Little Bahama Bank faces the Blake Plateau and is exposed to the open Atlantic. The bank edge is a reef/oolite shoal complex [Hine & Neumann 1977]. Beyond the marginal escarpment (150-200m) is a broad, gentle and shallow 4° slope (150m down to basin) [Mullins & Neumann 1979]. Peri-platform sand facies seaward of the marginal escarpment give way to upper slope pelagic sediments, that contain muddy debris flows with slope-derived foram-pteropod biomicrite clasts, this coarse sediment by-passing the slope through gullies.

The most detailed study of a windward margin slope is that of Mullins et al. [1984], who profiled the slope off north-eastern Little Bahama Bank. The slope consists of a steep gullied upper slope (4°) and a gentler lower slope (1-2°). The upper slope is dissected by gullies 50-150m in relief, which terminate along the upper slope/lower slope boundary.

**Upper Slope**

Immediately adjacent to the shelf-edge, in 200-400m of water, are hardground facies that are covered by recent peri-platform ooze. This grades downwards to patchily cemented nodular oozes (400-600m) [Mullins et al. 1980a] which, with depth, pass to un lithified, bioturbated peri-platform oozes (600-900m) (Fig. 2-6). This downslope decrease in submarine cementation correlates with the decreasing strength of contour following bottom currents [Mullins et al. 1980a; Mullins et al. 1984]. The upper slope is essentially dominated by peri-platform ooze which characterizes inter-gully highs. Gravity deposits may fill or line the gullies, as they do in the Cul-de-Sac of Tongue of the Ocean [Schlager & Chermak 1979], but this has not been proven.

**Lower Slope**

The upper slope gullies serve to funnel coarse sediment across the slopes and out onto the basin floor. The numerous gullies act as multiple feeders along a line source [e.g.
Figure 2-6

Near surface sediment map of the windward slope of northern Little Bahama Bank (from Mullins [1984]).
Rustnak & Nesteroff 1964; Schlager & Chermak 1979], depositing a continuous apron of coarse debris along the lower slope that runs parallel to the shelf-edge (Fig. 2-6). Gravity flows (equal proportions of turbidites and debris flows) form 60% of the apron sediments and are interbedded with peri-platform ooze. The turbidites show proximality-distality trends and are clean, normally graded sands and gravels, of between 20-250cm in thickness. The turbidite sediment shows a paucity of shallow-water grains, the clasts are planktonic foram-pteropod biomicrites, with sand-sized planktonic forams and pteropods [Davis 1983]. The debris flows are also slope-derived, with planktonic foram-pteropod biomicrite clasts, and show evolution from mud-supported to clast-supported with distance from source [c.f. Krause & Oldershaw 1979; Mullins & Cook 1986]. The change from mud-supported debris flows and thick, coarse-grained turbidites to thinner clast-supported debris flows and thinner turbidites marks the change from proximal to distal apron facies [Mullins et al. 1984; Mullins & Cook 1986] (Fig. 1-5, B).

2.5.3 Slope basin facies fronting leeward margins

Active off-bank sand transport occurs along the open leeward margins of Little and Great Bahama Banks, feeding bank edge "reservoirs" which accumulate this sediment and periodically shed it to the deep [Hine et al. 1981a]. The base of the marginal escarpment off leeward margins is therefore characterized by peri-platform sand facies of shallow and deep-water allochems, muds and talus blocks [Mullins & Neumann 1979]. Hubbard et al. [1976] observed downslope transport of platform sands through chutes in the escarpment in response to storms.

Off Little and Great Bahama Banks at depths of 600-700m, these resedimented carbonate sands are winnowed and transported northwards by the strong Florida and Antilles currents [Mullins et al. 1980b; Mullins 1983b] (Fig. 2-5). These massive volumes of platform sands and pelagic sediments form sediment drifts, up to 3,000km², 100km long, 60km wide and 600m thick. Bedding is generally massive with downslope pinch-outs, cross-stratification and small scale discontinuities. Heavily bored submarine cemented hardgrounds are also present. North-south oriented lithoherms (lithified bioherms) characterize the lower slopes, which are swept by strong currents. This facies grades seawards to basinal peri-platform ooze and distal turbidites.

Cores taken from the slopes off the protected leeward margin of Grand Bahama Island (in Northwest Providence Channel) show an upper slope of gravity flow slumped
facies, which give way to a lower slope of discontinuous proximal turbidites, which grade seawards to layered pelagic ooze and thin distal turbidites [Mullins & Neumann 1979].

2.6 Summary

It has been demonstrated that carbonate sand transport is controlled by the orientation of the platform margins with respect to prevailing wind directions: sand being transported bankwards along windward margins and in an off-bank direction along leeward margins. The position of bank margin islands or amplified tidal effects may alter this fundamental concept. This control not only applies to the bank margin sediments, but also to their adjacent slopes and basins, open windward margin slope and basin facies being carbonate sand-poor while those of the open leeward margins are carbonate sand-rich.

Carbonate slope and basin facies therefore record the history of the source margin, and its response to changes in relative sea-level. It has been shown that even minor changes in sea-level may have a major effect on platform sedimentation, the effects being most dependent upon the rate of change. Sea-level changes are especially influential in carbonate environments, when even very minor changes in sea-level may affect types and rates of sediment produced, as well as the degree of sediment transport, by subtly altering water depths over the source platform. This is of crucial importance, since effective carbonate sand production and transport occurs when the banks are flooded between depths of only 2-5m.

These fundamental concepts may be applied to the rock-record, where the recognition of palaeoleeward and windward margins may be of major economic importance: both platform margin facies and resedimented facies adjacent to leeward margins having greater potential for extensive, sand-rich reservoirs [e.g. Mullins et al. 1978; Mullins & Cook 1986]. Both margin types also have the potential for their associated resedimented sandbodies to be stratigraphically sealed, both vertically and laterally, by fine grained basinal facies.

It has been demonstrated by several authors that major sediment transport occurs only in response to storm action, normal tide and current activity only winnowing and sorting the sediment. Tropical storms or hurricanes affect the platforms at a rate of 8 per 10 years, making them extremely important and, geologically speaking, very rapid
agents of sand transport. The Bahamas extend north to south from 23°N to 27.5°N and lie to the west of a major ocean, the Atlantic. Comparable palaeogeographic situations may be envisaged for Mesozoic Tethyan carbonate platforms lying at similar latitudes and to the west of the large Tethys Ocean, which are also likely to have been strongly influenced by storms and hurricanes [e.g. Duke 1985]. Fundamental concepts gained from study of the Bahamas and the way the platform margins have responded to wind direction, energy flux and sea-level change, may be used in the interpretation of ancient reworked carbonate sandbodies to elucidate controls on their development.
3.1 General Introduction and Setting

3.1.1 Introduction

A laterally extensive but stratigraphically thin resedimented oolitic sandbody occurs within the middle Jurassic succession of the Sierra de Levante, Mallorca and is known as the Cutri Formation [Alvaro et al. 1984]. The sandbody developed as a linear feature, which would have been some 100km long and probably not more than 50km wide. With a vertical thickness of only 70-100m, it is thin in comparison to the other oolitic turbidite systems recorded in both this thesis and in the literature. The Cutri Formation was also by comparison a relatively short-lived turbidite system, developing over some 8 million years during the Bathonian, although there is some uncertainty as to whether resedimentation continued into the Callovian. The resedimented facies are preceded by the rhythmically bedded hemipelagic limestones of the Cuber Formation, the upper boundary of the sandbody being marked by a transition to the radiolarites of the Puig de Ses Fites Formation.

Mallorca has recently been re-mapped under the MAGNA project (although the maps are as yet unavailable) and the production of a guide-book on the Jurassic geology of Mallorca from this survey by Alvaro et al. [1984] first introduced the author to the Cutri Formation, which was named after its most classic outcrop at Puig Cutri. These authors suggested that the Cutri Formation may represent an example of a carbonate apron system.

The presence of oolite associated with basin sequences in the Sierra de Levante was documented earlier by Bourrouilh [1973], who was the first to interpret it as being resedimented. Publications on the geology and biostratigraphy of Cabrera Island (a south-western extension of the Sierra de Levante) proved exceedingly useful [Arbona et al. 1984/85; Sabat & Santanach 1984 and 1985]. The most valuable work however, was that of Sabat [1986] who 'unravelled' the thrust tectonics of the Sierra de Levante, allowing precise palinspastic corrected correlation between different structural units; as well as producing a more detailed geological map of the area that was used in this study.
3.1.2 Geological Setting

The Balearic Islands are an isolated lineation of five islands in the Western Mediterranean, of which Mallorca is the largest with an area of 3640km$^2$. The five islands (from SE to NW) Formentera, Ibiza, Cabrera, Mallorca, and Menorca (Fig 3-1, A) are the subaerial manifestations of the Balearic ridge, which curves SW to NE over 300km. The ridge consists of continental crust of crystalline Hercynian basement [Banda et al. 1980] and serves to separate the thinned continental crust of the North Balearic Basin (or Valencia Trough) [Hine 1973] from the Miocene oceanic crust of the South Balearic Basin (or Algerian Trough) [Rehault et al. 1984; Burrus 1984]. Seismics show a Mesozoic-Cenozoic sedimentary cover of 7km over Mallorca, with a depth to the Moho of 25km [Banda et al. 1980]. The Balearic ridge represents an extension of the Betic Cordillera of southern Spain [Bourgois et al. 1970; Bourrouilh 1973; Azema et al. 1974; Banda et al. 1981; Vera 1988]. The islands are therefore part of the Alpine mountain chain and are most closely associated with the External Subbetic of southern Spain, where Jurassic-Cretaceous sediments are characteristically of deep-water affinities [Garcia-Hernandez et al. 1980; Ruiz-Ortiz 1983; Vera 1988].

Mallorca is geographically and geologically divided into three regions which trend SW to NE across the island. Two mountainous regions predominantly composed of Mesozoic carbonates, the Sierra Norte (or Serra Nord) in the north and the Sierra de Levante (or Serra Llevant) to the south, are separated by the central plain of Es Pla (Fig. 3-1, B). The Sierra de Levante exposes Middle and Upper Jurassic facies of basinal affinities and this basin, not given a formal name by previous workers [e.g. Alvaro et al. 1984] is herein referred to as the Sierra de Levante basin. It served to separate a platformal source area to the east-southeast from a submerged pelagic plateau or "seamount" to the west-northwest, represented by the present day Sierra Norte.

The Sierra de Levante basin gained basinal status in the Bajocian, but was initiated by rifting in the Domerian [Alvaro et al. 1984; Prescott 1988] which created a complex block and basin topography over the present day Mallorcan region that was controlled by strike-slip tectonics. During the Middle Jurassic, the Sierra Norte and Sierra de Levante evolved as separate palaeostructural units, both affected by complex internal block faulting (Fig. 3-2). While the Sierra Norte acted as an elevated submarine plateau, accumulating condensed facies and "starved" basinal deposits, the Sierra de
Figure 3-1

A) Location of Mallorca (Sierra de Levante in black) in relation to the other Balearic Islands and the Betic Cordillera of Spain.
F = Formentera; I = Ibiza; C = Cabrera;
M = Menorca; B = Barcelona; V = Valencia;
A = Alicante (after Sabat et al. [1988])

B) Morpho-tectonic and geological map of Mallorca, showing outcrops of Jurassic strata.
P = Palma
Levante acted as a depositional basin, periodically receiving carbonate gravity flows from the basinal area to the east (Fig. 3-3).

This Tertiary structural separation of Jurassic palaeogeographic domains into SW-NE trending belts is evidence for these older structural lineaments being reworked during the Alpine orogeny [c.f. Lemoine 1982]. The reactivation of older fault lineaments during Tertiary has been documented from other Tethyan regions, for example by Watts & Garrison (1986); Teale & Young (1987) and Warme (1988). It is also likely that the Jurassic rift pattern itself may also have followed older NE-SW Hercynian basement structural trends. This suggests an original NE-SW direction of Jurassic extensional faults along the present day thrust fronts. This hypothesis correlates well with facies changes observed in the Sierra de Levante, as well as in the Sierra Norte (D. Prescott, personal communication).

3.1.3 Jurassic palaeogeographical evolution of Mallorca

Four distinct episodes of palaeogeographic evolution are noted in the Sierra de Levante, the first two of which are directly comparable with facies development in the Sierra Norte.

(1) Pre-rift

The Upper Triassic-Caryxian is represented by peri-tidal platform carbonates, which towards the top of the sequence are influxed by siliciclastics that herald the onset of platform rifting and subsidence (Fig. 3-2).

(2) Early Rift

The Domerian-Aalenian is characterized by classic Tethyan seamount-type facies, which may be extremely condensed and laterally variable, and which abruptly overlie the platform sequence.

(3) Rift

The early Bajocian-Callovian/Oxfordian succession of the Sierra de Levante is characterized by the establishment of basinal conditions and ongoing subsidence as rift-related extension proceeded. While oolitic turbidites were being deposited in the Sierra
Figure 3-2

Evolution of Mallorca from the Liassic to the Neocomian
(courtesy of D Prescott)
de Levante basin, the Sierra Norte remained in topographic isolation as a pelagic seamount (Fig 3-2 & 3-3).

(4) Drift

The Callovian/Oxfordian-early Cretaceous facies of the Sierra de Levante remained basinal, the resedimented carbonate facies reflecting platform margins stabilization with the onset of drift in the Ligurian Ocean. The Sierra Norte still remained in topographic isolation.

Pre-rift

Until the end of the Lower Jurassic, the geological evolution of Mallorca is similar to that of the external Betics of southern Spain and the Celtic-Iberian aulacogen of north-west Spain [Garcia-Hernandez 1980; Alvaro et al. 1984; Vera 1988]. Continental Bundsandstein fluvial facies are succeeded by Middle Triassic dolomitic intertidal and supratidal Muschelkalk facies that are only exposed in the Sierra Norte. The Upper Triassic Keuper facies occur in the Sierra Norte and Sierra de Levante and act as the regional décollement horizon for later Alpine thrusting. In the Sierra Norte, an outcrop of gypsiferous marls contains intercalations of basalts, with both subaerially erupted and pillow laval forms present [Pomar-Goma 1979]. Their presence indicates early extension on Mallorca, known elsewhere as the Triassic "aborted rifting" stage [Castellarin & Rossi 1981; Lemoine 1982], that preceded the early Jurassic rifting phase of the Central Atlantic and Western Tethyan realm.

The Hettangian-Caryxian shallow water peri-tidal limestones of the Es Barraca Formation [Alvaro et al. 1984] are ubiquitous over the Sierra Norte and Sierra de Levante, forming sequences 200-300m thick. They are abundantly exposed and mark a continuation from the uppermost Triassic dolomitized Felanitx Formation and the dolomiticrites of the lowermost Jurassic Mal Pas Formation [Alvaro et al. 1984]. The sediments represent classic tidal flat facies, displaying shallowing-upward cycles [James 1984] as sediment production buffered subsidence and relative changes in sea-level [e.g. Kendall & Schlager 1981].

In many localities in the Sierra Norte and Sierra de Levante, the top of the platform is characterized by influxes of siliciclastics of between 0.5-2m in vertical thicknesses. Dated as Carixian [Arbona et al. 1984/85; Sabat & Santanach 1985], these gravelly
Figure 3-3

Diagram to show the relationship between coeval facies development in the Sierra Norte and Sierra de Levante in the Jurassic, and their approximate ages. Modified after Alvaro et al. [1984/85]; Prescott [1988]; Toni Simo personal communication; and the author's own observations.
Figure 3-4

Locality map for localities referred to in the text. Those underlined are localities of the Cutri Formation.
Plate 3-1

A) Ammonite encrusted, haematite stained hardground (Toarcian) Sa Tortuga, eastern Sierra d'Artá

B) Lower Ammonitico Rosso (Aalenian), freshly excavated outcrop, southern section Puig Cutri, Sierra d'Artá (Lens cap 5cm)

C) Puig de Ses Fites Formation radiolarite horizon (Callovian? – Oxfordian), Puig de Ses Fites, Sierra de Artá. (Hammer 30cm)

D) Carboneras Formation Saccocoma-rich siliceous turbidites (Kimmeridgian), Puig de Ses Fites, Sierra d'Artá. (Hammer 30cm)
arenites are the only siliciclastic sediments within the entire Jurassic-Cretaceous succession of Mallorca (Plate 2-II, B). Its influx heralds the onset of rifting and the termination of shallow water carbonate sedimentation in the area for the remainder of the Mesozoic. Alvaro et al. [1984] consider the siliciclastic source to be from the north-west; however on evidence from sections in the Sierra de Levante, a SE-E provenance is indicated. At Amoixa in the south-west Sierra de Levante, the quartz-arenite is a pebbly conglomerate, becoming a gravelly sand at Puig Cutri, while in the Sierra Norte [D. Prescott, personal communication] it becomes more marine in nature and finer-grained north-westwards form Andratx to Cala Fornells and Camp de Mar (Fig. 3-4). The provenance of the siliciclastic would therefore appear to be the Alboran-Kabylia massif, which lay to the south and east during the Pliensbachian [Dercourt et al. 1986] (Fig. 1-2,A).

Siliciclastic influxes across carbonate platforms are often correlatable with eustatic lowering of sea-level [Cook et al. 1987]. Haq et al. [1987 & 1989] record such an event at 189 million years ago in the Pliensbachian; however considering its immediate pre-rift stratigraphic position, a tectonic role may also have been important.

Early Rift

The break-up unconformity, recognized by an abrupt change to deeper water facies, spans the Domerian to Lower Toarcian over Mallorca [Arbona et al. 1984/85; Sabat & Santanach 1985; Prescott 1988] and open marine seamount type facies characterize this period and the ensuing Aalenian [Alvaro et al. 1984]. The immediate post-rift sediments are extremely variable over the island, reflecting differential block subsidence. The first facies formed on the submerged blocks are the pink neritic marls of the Sa Moleta Formation, or the red crinoidal grainstones of the Cosconar Formation [Alvaro et al. 1984]. (Fig. 3-3 & 3-5).

The Sa Moleta Formation may reach 30m in thickness in regions of the Sierra Norte [Alvaro et al. 1984] but if present in the Sierra de Levante is typically 0.5m–1m thick (such as at Puig Cutri and Amoixa II. Fig. 3.4). The crinoidal grainstones may contain large belemnites and typically show cross-bedding and are interpreted as encrinitic sandwave complexes. This classic Tethyan facies (Section 1.3.3), as in the west Sicilian Jurassic [Jenkyns 1971b] is discontinuous in its development as it was a strongly depth controlled facies. At Camp de Mar and Cala Fornells in the Sierra Norte, crinoidal grainstones directly overlie the rifted platform (Fig. 3-5). The development of both
Plate 3-II

A) Bathonian-Oxfordian Intermediate Ammonitico Rosso, Son Vidal, Orient, Sierra Norte. The coeval seamount facies of the Sierra Norte, this is equivalent in age to the Cutri Formation resedimented oolite (Bathonian) outcrop of the Sierra de Levante. (Hammer 30cm).

B) Incipient ferromanganese crust forming at the top of the Liassic platform. Cuevas d'Arta, Sierra de Levante.

C) Quartz-arenite (Carixian) from the top of the platform succession, the only siliciclastic Jurassic facies of Mallorca. This horizon heralds or marks the onset of platform drowning. Amoixa II, Son Macia, Manacor, Sierra de Levante. (Lens cap 5cm)

D) Zoophycos trace fossil from the Cutri Formation (Bajocian), Mount Cuber, Sierra Norte.
these facies types characterizes less subsident, initial rift periods, both being described
from other Tethyan regions [Jenkyns 1971b; Sturani 1971]. Since these facies are depth
controlled, in various localities either one or both may be entirely missing (e.g. Cuevas
d'Artá section) or they may occur stratigraphically above or below each other (Fig. 3-
5).

Hardgrounds, when forming part of the local sequence, are representative of most of
the Toarcian [Prescott 1988] reflecting the widespread "drowning" that occurred over
Western Tethys during the Domerian-Toarcian (see Section 1). As their maximum
thickness is 20cm, they represent the most condensed Mallorcan Jurassic facies. The
most spectacular outcrop of the hardground in the Sierra de Levante occurs at Sa
Tortuga, GR(367,925) (Plate 3-I,A) between Capdepera and the Cuevas d'Artá (Fig. 3-
4). The hardground is a distinctive marker horizon, typically encrusting the underlying
pink marls or crinoidal grainstones (Fig. 3-5). Its distribution in the Sierra de Levante
is variable reflecting subtle local variations developed due to differential block
subsidence. In the far east of the Sierra d'Artá at the Cuevas d'Artá (Section 3.3.4.2),
sporadic ferro-manganese nodules occur directly on top of the Liassic platform
limestone sequence (Plate 3-II,B). In the south-west Sierra de Levante at Amoixa, the
crinoidal grainstones contain pebbles of reworked iron-stained crinoidal material and
reworked ammonites encrusted in goetite. At Amoixa II (Figs. 3-4 & Fig. 3-27), the
thin ferro-manganese crust nucleates on the Caryxian quartz-arenites and is overlain by
the pink neritic marls and glauconitic crinoidal grainstones.

The excellent exposure of the hardground at the Llosetta quarry in the Sierra Norte
(Fig. 3-4) enabled Prescott [1988] to make a detailed sedimentological and geochemical
survey. At this locality, the platform surface is uneven, bored and cut by neptunian
dykes, identical in nature to neptunian dykes seen by the author in the Betics of
southern Spain and described by Molina [1987]. A geochemical analysis of the
hardground [Prescott 1988] suggests that submarine volcanic/hydrothermal activity
influenced formation of the hardground. Aalenian ammonitico rosso (the Inferior
Ammonitico Rosso) (Plate 3-I,B) overlies the hardground and, although it is still a
condensed facies, with only 4m representing the Aalenian, it represents the
commencement of slow sedimentation and a deepening at the beginning of the Middle
Jurassic.
Figure 3-5

Diagram to show the diversity and stratigraphic sequence development of early rift facies, in different localities on Mallorca (for their geographical position, see Figure 3-4)
Rift

The Cuber Formation, interbedded marls, wackestones and carbonate mudstones of the Bajocian-Lower Bathonian, abruptly overlie the ammonitico rosso and are the first sediments to drape the basin rift topography in areas of the Sierra Norte and generally throughout the Sierra de Levante. The Cuber Formation varies from 95-20m in thickness over the island, tending to be thicker bedded (centimetres to decimetres) in the Sierra Norte, with abundant Zoophycos being found in the section at Cuber (Plate 3-II,D and Fig. 3-6). In the Sierra de Levante, the Cuber Formation is thinner bedded (on the scale of a few centimetres) and is succeeded in this region only by the resedimented oolite of the Cutri Formation. The hemipelagic interbed units of the Bathonian Cutri Formation are of similar nature to the Cuber Formation and the resedimented oolite influxes represent temporary interruptions to the 'Cuber Formation type' background sedimentation. Together, the Cuber and Cutri Formations serve to plain the initial rift block topography of the Sierra de Levante basin. Coeval facies to the Cutri Formation in the Sierra Norte are the Bathonian-Upper Oxfordian red nodular limestones of the Intermediate Ammonitico Rosso [Alvaro et al. 1984] which are bestdisplayed at Orient (Plate 3-II,A and Figs. 3-3 & 3-6).

In the Sierra de Levante only, the Cutri Formation is overlain by the Puig de Ses Fites Formation radiolarites, indicative of deepening and a reduction of sedimentation rates (Plate 3-I,C). Radiolarites and radiolarian cherts are abundant throughout Western Tethys during the Callovian-Oxfordian (Section 1.3.3). The general lack of ammonites and diagnostic microfauna in the sediments prevents accurate age determination in many areas [Ogg et al. 1983], this being true for Mallorca. This poses a problem in determining the upper age limit for the Cutri Formation. Alvaro et al. [1984] the Puig de Ses Fites Formation as Oxfordian. There seems to have been a hiatus in sedimentation during the Callovian over Mallorca, no fossils of this age being recorded as yet from the entire island [Alvaro et al. 1984; Prescott personal communication 1989; Simo personal communication 1989] or from the island of Cabrera [Arbona et al. 1984/85]. In the Sierra Norte, ammonitico rosso facies span the Bathonian-Oxfordian, but no Callovian ammonites have yet been identified from these facies [Simo personal communication 1989]. Callovian-Oxfordian hiatuses are widespread elsewhere over Western Tethys at this time [e.g. Ogg et al. 1983; Lemoine & TrÜmpy 1987] (see Section 1.3.3).
Figure 3-6

Schematic log section of the Cuber region of the Sierra Norte, to show the occurrence of sea-mount facies coeval to oolite resedimentation in the Sierra de Levante
SCHEMATIC STRATIGRAPHIC SECTION:
CUBER, SIERRA NORTE

BERRIASIAN

TITHONIAN

KIMMERIDGIAN

OXFORDIAN

BATHONIAN

BAJOCIAN

TOARCIAN

Maiolica

Upper Ammonitico Rosso

Aumedra Formation

Intermediate Ammonitico Rosso

Cuber Formation

Es Barraca Formation

F

Hardground
Crinoidal packstone
Quartz arenite

25m

(modified after ALVARO et al. 1984)

© ammonites
● posidonia
△ chert
Radiolarites and radiolarian cherts elsewhere in Western Tethys correlate with a major palaeogeographic and oceanographic change (the onset of drift in the Ligurian Ocean) in the Callovian (Section 1.3.1). It is therefore likely that the Puig de Ses Fites Formation actually represents the Callovian-Oxfordian, not just the Oxfordian as suggested by Alvaro et al. [1984]. To support the former theory, in the Subbetic of mainland Spain (of which Mallorca is an extension), the Iberian Cordillera and south and central Portugal, sedimentation styles changes dramatically in the latest Bathonian and remains 'condensed' throughout the Callovian-Oxfordian [García-Hernández et al. 1980; Ogg et al. 1983; Vera 1988].

This change in the Subbetic is brought about by the development of this region into the connecting seaway between the Ligurian Ocean and the Atlantic [Ogg et al. 1983; Dercourt et al. 1986; Vera 1988]. It is therefore likely that the termination of oolite resedimentation on Mallorca and its replacement by radiolarite facies coincided with this palaeogeographic and oceanographic change imposed in the Callovian.

Drift

The syn-drift Malm succession is characterized by ammonitico rosso facies and calcareous mudstones in the Sierra Norte (Figs. 3-3 & 3-6), while the Sierra de Levante remained a depositional basin that periodically received influxes of bioclastic turbidites and debris flows [Alvaro et al. 1984] from the east. The style of resedimentation is markedly different from that of the Middle Jurassic facies, being characteristic of a prograding accretionary slope with the predominance of slumps, slides, debris flows and turbidites. These facies indicate progressive up-dip collapse from slope material to collapse of a well cemented carbonate platform margin [Barnolas & Simo 1988] as slope sediments prograded over the levelled basin topography. The Kimmeridgian of the Sierra de Levante is represented by Saccocoma-rich silicified, resedimented limestones, intercalated with mudstones and marls [Bourrouilh 1973; Alvaro et al. 1984] (Plate 3-I,D). The Tithonian-Lower Berriasian bioclastic conglomerates, grainstones and slump masses and interbedded radiolarian-rich marls of the Puig d'em Borras Formation [Alvaro et al. 1984] are overlain by the Maiolica white calpionellid marls of the Lower Berriasian (Fig. 3-3). The Maiolica blankets the whole of Western Tethys at this time (see Section 1.3.3).
3.1.4 Structural Geology of the Sierra de Levante and its relationship to the Cutri Formation

An understanding of the structural geology of the Sierra de Levante is fundamental to the 3-D reconstruction of the Cutri Formation sandbody.

The complex thrust tectonics of the Sierra de Levante was resolved by Sabat [1986], who distinguished eight allochthonous units (units 0 to 7), emplaced from the south-east (Fig. 3-7). The predominant structures are north-west orientated thrusts, which involve a sedimentary series from the Upper Triassic to Middle Miocene [Bourrouilh 1973; Alvaro et al. 1984]. The Upper Triassic evaporites and marls (Keuper facies) acted as the regional décollement horizon [Bourrouilh 1973; Azema et al. 1974; Colom 1975; Sabat 1986; Sabat et al. 1988].

Successively higher units of Sabat's [1986] scheme display more proximal sections of the Cutri Formation, containing certain facies associations within particular structural units. As mentioned in section 3.2.1, Lemoine [1982] suggested that Liassic faults bounding individual tilt blocks are often reworked as thrust or strike-slip faults during later orogenic periods. These eight tectonic units may represent individual Jurassic tilt blocks, their bounding faults being reworked as the unit bounding thrusts during the Tertiary.

The set of lower thrusts in the structural pile developed in piggy-back sequence (where new thrusts develop in the footwall of a previously active thrust [c.f. Butler 1982]) whereas the upper thrusts developed out of sequence due to blocking of the piggy-back progression [Sabat 1986]. The thrusts are laterally imbricated to the SE and SW, those in SW being uppermost. The most complete section is found in the north-eastern Sierra de Levante where several outcrops of the Cutri Formation are found (Fig. 3-8). In this area, five thrust sheets form an imbricate fan dipping to the SE (Sabat et al. 1988).

There are two main ramp directions (Fig. 3-7), one trending NW-SE, the other NE-SW, and associated folds have developed, which show the facing direction of their related ramp: trending NE-SW parallel to frontal ramps or NW-SE parallel to the lateral ramps.

Other prominent features are NW-SE orientated faults, which frequently show sub-vertical strike-slip [Bourrouilh 1973] with secondary sets running N-S and E-W subdividing the Sierra de Levante into a series of blocks and destroying the lateral continuity of the thrust sheets [Alvaro et al. 1984]. These faults are either strike-slip
Figure 3-7

Map to show the structural units of the Sierra de Levante of Sabat [1986] from Sabat et al. [1988].
C = Puig Cutri; SF = Puig de Ses Fites; P = Puig d'en Pare;
A = Amoixa region; CA = Cuevas d'Arta region.
Figure 3-8

Simplified cross-sections through the Sierra de Levante thrust sheets, position of sections is marked on 3-5. (1 is the lower-most thrust sheet, 7 the uppermost). Relative positions of localities studied are marked: C = Puig Cutri; PF = Puig de Ses Fites; P = Puig d'en Pare; A = Amoixa region; CA = Cuevas d'Arta region.
or drop faults [Butler 1982], are synchronous with thrusting [Sabat 1986, Sabat et al. 1988] and served to allow differential sliding of blocks to the NW. All structures developed within the Sierra de Levante thrust belt are compatible with a syn-thrust origin [Sabat 1986; Sabat et al. 1988], developing during a long tectonic event ranging from the Oligocene to Middle Miocene.

Localities

The most extensive outcrops of the Cutri Formation occur in the Sierra d'Artá within structural unit 3 of Sabat [1986] and are the closely associated localities of Puig Cutri, Puig de Ses Fites and Puig de Ses Fites II. Structural unit 1 exposes the distal section at Puig d'en Paré. Structural unit 5 exposes four contrasting proximal sections: Amoixa and Amoixa II occurring in the Sierra de Son Amoixa and the Cuevas d'Artá area (the Cuevas d'Artá road section and that of Sa Talaia) in the far eastern Sierra d'Artá. Of all these sections only that of Puig Cutri was described by Alvaro et al. [1984]. A palinspastic restoration of the thrust sheets [Sabat 1986] allows a pre-thrust relationship between the localities to be appreciated (Appendix Fig. 1).

3.1.5 The location of the source platform: Menorca - a fragment of the coeval Middle Jurassic source platform

The Cutri Formation facies show proximal to distal trends in a westerly northwesterly direction, indicating an easterly direction of the source platform. It would be impossible for the oolite to have been sourced from the N, W, and NW due to coeval seamount-type facies developing on submarine plateaux in these directions at the time and there is no evidence for oolitic sediment by-passing of these regions. The hypothesis of Alvaro et al. [1984] that the oolite was sourced from the north is therefore ruled out both on palaeogeographic and sedimentological evidence.

From the work of Bourrouilh [1973] and Bourrouilh & Bourgois [1970], the Middle Jurassic succession of Menorca was different to that of Mallorca. The pre-rift Triassic and Lower Jurassic succession is identical to that of Mallorca, consisting of a thick dolomitized platform carbonate sequence. The Lower and Middle Toarcian initial post-rift sediments are marls containing rhychonellids and ammonites, overlain by a hardground of Upper Toarcian age [Bourrouilh 1973] and therefore similar to that of Mallorca (Section 3.1.2). However from the Aalenian onwards, the succession differs between Mallorca and Menorca. The Aalenian of Menorca is characterized by quartz-
Figure 3-9

Schematic log section of the Middle Jurassic of Menorca.
oolite

MIDDLE JURASSIC (DOGGER)

dolomitized oolite

oolite

LOWE R JURASSIC (LIASSIC)

dolomite

marls + crinoids

marls + rhy chonellids

hardground

TOARC IAN

L (after BOURROUILH 1973)
arenite influxes, these sediments containing plant fragments, spores and pollen, indicating "subaerial deposition ... and proximity to land" [Bourrouilh 1973]. This landmass was almost certainly the Alboran-Kabylia landmass (Fig. 1-2, A & B). The following Middle Jurassic succession comprises intensely dolomitized platform limestones, often observed to be oolitic and containing ooid "ghosts". The ooid nuclei may be Trocholine foraminifera, monocrystalline and polycrystalline quartz, or peloids. Periodic quartz-arenite influxes occur within this platform sequence. A slight deepening and facies restriction occurs in the Upper Jurassic, with pelagic conditions prevailing (as on Mallorca) in the Barremian.

The Middle Jurassic of Menorca clearly represents an oolite producing platform of coeval age to the resedimented oolite of the Cutri Formation of Mallorca. The Menorca region was clearly situated nearer to the Alboran-Kabylia landmass than the platform region that sourced the Cutri Formation due to the association of the quartz-arenite intercalations and numerous quartz ooid nuclei of the Menorca succession. Menorca represents a more landwards and more northerly lateral fragment of the Cutri Formation source platform. A small fragment of the source platform is possibly exposed in the eastern Sierra de Levante (see Section 3.3.4.2) and a close proximality to the source platform is indicated by the succession of Cabrera (see Section 3.3.5).
3.2 Sedimentary facies of the Cutri Formation

3.2.1 Introduction to the geometry and architecture of the Cutri Formation sandbody

The Cutri Formation is characterized by episodic, resedimented oolitic units that punctuate the normal hemipelagic background basin facies, representing a continuum from the Cuber Formation hemipelagic rhythmites. The characteristic sedimentological feature of this sandbody is that the resedimented oolitic units are encased in vertical section by fine-grained, hemipelagic carbonate sediments, indicating the source platform was always submerged and able to produce carbonate mud and peri-platform fines. The facies developed within the resedimented units are the most varied and distinctive of the three Jurassic sandbodies considered, and this is the only sandbody where distinctive debris flows occur as part of the succession.

The geometry of the sandbody is distinctly linear, being a NE-SW feature, showing increased distality trends in a northwesterly direction. It would have been at least 100km long, with major sand deposition occurring in a belt 35km wide (possibly up to 50km wide; possible outcrops further NW may be covered by the Tertiary molasse of Es Pla - Fig. 3-1, B). At the most conservative estimate, development of the sandbody involved the resedimentation of at least 90-100km$^3$ of unconsolidated oolitic-peloidal sands (upper estimates more than doubling this figure to 250km$^3$). Although later sections elaborate the significance of this geometry it is clear that the gross sandbody geometry was clearly not that of a classic submarine fan: "a cone-shaped body located seawards of a major submarine canyon" [Bates & Jackson 1980]. The oolitic turbidites show no evidence for being arranged in fan-type thickening-upward or thinning-upward lobe and channel cycles. Rather, the sedimented facies formed a basin margin parallel wedge of oolite, suggesting a line sourced resedimented sandbody of apron rather than fan affinities (see Section 1.4 for a discussion on fan versus apron geometries).

The stratigraphic trend of the entire Cutri Formation succession is one of an overall thinning-upward of the resedimented oolitic units (Fig. 3-10), indicative of long-term sandbody retrogradation. This trend, characterized by an overall decrease in bed thickness and frequency, involves an evolution from thick amalgamated basal units, through thick sediment sheets, to thinly sedimented sheets with background basinal facies becoming volumetrically more important up section. In simplest terms, this trend reflects a deepening, combined with an increasing distance from source and...
Figure 3-10

Schematic log section of the succession displayed at Puig Cutri, combining the classic post-rift Tethyan 'break-up' sequence displayed in the northern section of Puig Cutri, with the Cutri Formation seen in the southern section of Puig Cutri.

6 = ammonites
~ = Posidonia bivalves
\( \uparrow \) = burrows
\( \triangle \) = chart
SCHEMATIC STRATIGRAPHIC SECTION:
PUIG CUTRI, SIERRA LEVANTE.

- Cutri Formation
- Cuber Formation
- Lower Ammonitico Rosso
- Sa Moleta Formation
- Quartz arenite
- Es Barraca Formation

Layers:
- 
- 
- Bathonian
- Bajocian
- Aalenian
- Toarcian
- Pliensbachian
- Sinemurian

10 m

MIDDLE JURASSIC (DOGGER)
decreasing sediment supply rate and is seen in other resedimented carbonate systems [e.g. Ruiz-Ortiz 1983; Watts & Garrison 1986]. Such retrogradational trends are generally accepted to develop in response to a relative sea-level rise, most simply explained by the landward (sourceward) migration of the zone of sediment supply, in response to transgression onto the platform [e.g. Read 1982; James & Mountjoy 1983; Cook et al. 1983]. A eustatic sea-level rise, that is traditionally inferred to give rise to a retrograding basin trend, is not recorded by Haq et al. [1987 & 1989] or Hallam [1988] for the Bathonian, in fact the reverse; a period of long-term net sea-level fall is recorded by the aforementioned authors (see Section 3.4.3).

The development of large scale thinning-upward trends is a characteristic of resedimented sandbodies described from other rift-infill sequences, both siliciclastic [Surlyk 1978; Stow 1985b; Howell & Vedder 1985] and carbonate [Watts 1986; Eberli 1987 and 1988], many of which characterize the Jurassic. The thinning-upward mega-sequence trend reflects the progressive decrease in relief and tectonic activity as rifting proceeds and is characteristic of tilted fault-block basin sequences. Frequently, the mega-sequence may consist of several internal megacycles, each reflecting a separate phase of tectonic activity [Surlyk 1978; Stow 1985b; Eberli 1987]. The entire Cutri Formation sandbody is interpreted as a rift-infill megasequence, developing its retrogradational trend as rift-related extension and basin subsidence proceeded, prior to the onset of drift. The whole Cutri Formation mega-sequence is of comparable thickness to the megacycles developed within the Jurassic Brae oilfield faulted slope apron of the North Sea, which are 50-150m thick [Stow 1985b].

3.2.2 Lithofacies of the Cutri Formation

A varied range of lithofacies types are displayed in outcrops of the Cutri Formation, these facies being the most diverse and varied of the three Jurassic oolitic sandbodies discussed. The lithofacies types have been broadly divided into three groups, based on sediment component size of the facies: (1) carbonate conglomerates; (2) graded and massive oolitic calcarenites; and (3) fine calcarenites and calcisiltites. These lithofacies groups have been sub-divided into recognized lithofacies as follows:

(1) Carbonate conglomerates (calcirudites)

*Boulder-pebble conglomerates* - which contain clasts of between 10cm to >1m, set in an oolitic-peloidal matrix.

*Massive, belemnite-rich pebble conglomerates* - which contain clasts which never exceed
64mm, along with numerous belemnites, which occur as matrix-supported components in an ungrated oolitic peloidal matrix.

**Clast-supported, grated pebble conglomerate** - graded oolitic-peloidal beds that contain a basal clast-supported division where the clasts never exceed 64mm.

**Graded pebble conglomerate** - graded oolitic-pelodial beds that contain a basal matrix-supported, clast-rich horizon, that forms >30% of the bed thickness, where the clasts never exceed 64mm.

(2) Graded and massive oolitic calcarenites

**Massive oolitic turbidites** - beds which are 90% massive to poorly graded and are composed of indistinctly graded and poorly sorted oolitic-pelodial sand.

**Coarse-tail graded oolitic turbidites** - oolitic-peloidal beds dominated by coarse-tail grading (grading of the coarsest sediment fraction).

**Graded oolitic turbidites** - oolitic peloidal beds dominated by distribution grading.

(3) Fine grained calc-arenites, calc-siltites and other facies

**Thin bedded fine-grained calcarenites and calcisiltites** - thin bedded (1-10cm) ungraded fine sand to silt grade carbonate sediments.

**Laminites** - fine sand to silt grade carbonate sediments that show fine laminations on a millimetre scale (1-10mm).

**Posidonia coquinas** - sediments which show current-winnowed accumulations of the pelagic bivalve *Posidonia*.

These lithofacies groups and their sub-groups are described and discussed in the following sections and are interpreted in terms of the gravity flow models shown in Fig. 3-11 and more specifically in terms of the debris flow model of Fig. 3-12 and the turbidity current models of Fig. 3-13.

3.2.2.1 Carbonate conglomerates (Calcirudites)

Resedimented beds bearing clasts (that make up >25% of the bed) with maximum diameters of between 4cm to >1m, are herein referred to as conglomerates. The clasts are of wackestone or pelagic mudstone, and are set in a matrix of oolitic-peloidal packstone, or rarely grainstone. This group forms by total volume a small, but varied, percentage of the resedimented facies of the Cutri Formation. These facies have not been recorded by Alvaro *et al.* [1984]. These conglomeratic facies can often be correlated in the field with stratigraphic equivalent high-density turbidites.
Figure 3-11

Basic gravity flow models and their characteristics (after Middleton & Hampton [1976].
TURBIDITY CURRENT

RIPPED OR FLAT TOP
RIPPLE DRIFT
MICRO X-LAMINATION
LAMINATION
GRADING
FLUTES, TOOL MARKS ON BASE

FLUIDIZED FLOW

SAND VOLCANOES OR FLAT TOP
CONVOLUTE LAMINATION
FLUID ESCAPE PIPES
DISH STRUCTURE?
POOR GRADING
GROOVES, STRIATIONS, FLAME AND LOAD STRUCTURES

GRAIN FLOW

FLAT TOP NO GRADING?
MASSIVE
FLOW-PARALLEL GRAIN ORIENTATION
REVERSE GRADING NEAR BASE
SCOURS, INJECTION STRUCTURES?

DEBRIS FLOW

IRREGULAR TOP MASSIVE
POOR SORTING, RANDOM FABRIC
POOR GRADING
BASAL SHEAR ZONE
BROAD SCOURS, STRIATIONS?

(after Middleton and Hampton 1976)

FIG 3.11
Boulder-Pebble Conglomerates

This facies is defined as beds that contain clasts of variable sizes, but whose maximum size lies between 10 cm to >1 m, set in an oolitic-peloidal sand-grade matrix.

This is the coarsest conglomerate type that occurs within the Cutri Formation, and is localised to the Puig Cutri and Puig de Ses Fites localities of structural unit 3, where it is thin and localised, never exceeding more than 1-5 m in thickness.

The conglomerate of this association is poorly sorted, with clasts ranging from millimetre-sized sub-angular chips to a sub-rounded clast population of 1-4 cm, along with a variety of sizes and shapes of larger clasts and boulders up to 1 m across. All these clasts and boulders are of pelagic mudstone, or very fine sand grade *Posidonia*-rich wackestones which may frequently show internal parallel lamination (see Plate 3-XVII,B). Many of the large boulders (60-100 cm) are of high sphericity and rounded nature (see Plate 3-XVII,D), some more tabular clasts do occur and show preferred orientations ranging from SE-NW to SSE-NNW. Smaller clasts of pelagic mud may be plastically deformed; either streaked out in the flow direction or showing bent clast terminations (see Plate 3-X,D). Wackestone clasts may show internal lamination deformation. Locally clasts may be observed ‘frozen’ in the flow, in the process of being disaggregated (see Plate 3-XVI,D). A plastically deformed pelagic mud clast displayed in plan view at Puig Cutri indicates flow towards the WSW (Plate 3-X,D).

The conglomerate is matrix supported, the matrix being sand-grade oolitic packstone. Crude normal clast grading is seen, the base may be slightly (and locally) inversely graded. The boulder horizon is overlain by an oolitic packstone graded unit with occasional scattered pelagic mudstone chips 1-5 mm across, that grades up to parallel laminated *Posidonia* wackestones, which are capped by a thin veneer or locally thick unit of pelagic mud.

**Interpretation**

The variable clast size, matrix-supported nature of the bed and the plastically deformed nature of some clasts, all point to a debris flow origin for the boulder conglomerate [e.g. Nardin *et al.* 1979; Krause & Oldershaw 1979; Cook & Taylor 1977; Cook &
Mullins 1983]. This facies type is not of sufficient magnitude to qualify its
classification as a megabreccia after the usage of Mountjoy et al. [1972]; Cook [1979];
Eberli [1987]; Mullins & Cook [1986].

The clasts are all intraformational, slope-derived, pelagic mudstone and wackestones.
The plastic deformation of these clasts indicates the incorporation of
pencontemporaneous semi-lithified slope sediment into the debris flow [c.f. Hurst &
Surlyk 1982], as well as being a classic indicator of debris flow plastic emplacement
mechanisms [Middleton & Hampton 1976; Nardin et al. 1979]. The encorporation of
slope derived material as clasts indicates the importance of early marine lithification,
which is an important feature of carbonate slopes [Schlager & Camber 1986 see Section
2.5.1.1]. Resedimentation of semi-lithified coeval slope material as gravity flows, have
been documented by Cook et al. [1972]; Hopkins [1977]; Swavely [1981]; Cook [1979a &
b] and Cook & Mullins [1983]. The composition of the clasts and their often tabular
nature is suggestive of their origin as upslope slides which remoulded downslope
into debris-flows [c.f. Cook & Taylor 1977; Cook & Mullins 1983]. This process
requires that the shear strength of the slump or slide is exceeded in order to transform
it into a viscous debris flow [Middleton & Hampton 1976; Nardin et al. 1979; Lowe
1979]. The matrix-supported nature of the debris flow is considered by Krause &
Oldershaw [1979] to be indicative of a proximal (near source) debris flow (Fig. 3.12),
which is concordant with an inferred slope origin of the clasts in the flow.

Within this debris flow, the larger boulders would have been supported within the body
of the flow by a variety of mechanisms, including cohesive strength of the matrix,
buoyancy and dispersive pressure [Middleton & Hampton 1976; Hampton 1972 and
1979; Rodine & Johnson 1976; Nardin et al 1979; Lowe 1979, 1982; Pierson 1981;
Kessler & Moorhouse 1984], along with grain to grain contact traction turbulence
within the matrix and fluid pore pressure [c.f. Rodine & Johnson 1976; Enos 1977;
Lowe 1979; Cas 1979]. Since this debris-flow is matrix supported, these large boulders
were not supported within the flow by boulder-boulder clast collisions, and it is likely
to have been the ooid-oolid grain collisions that generated important dispersive
pressures within the flow. Certainly the role of dispersive pressures in creating
frictional strength [Hiscott & James 1985] is indicated by the preferred orientation of
elongate boulders supported in the top of the flow, indicating dispersive pressure from
the smaller debris flow constituents served to orientate these larger clasts within the
flow [c.f. Hampton 1979; Lewis et al. 1980].
Figure 3-12

Model for the development of two-component carbonate gravity flow models consisting of a basal debris flow unit which evolves from mud-supported to grain-supported with increasing distance down-slope.

1 = disorganised; 2 = stratified disorganised; 3 = stratified, normally graded; 4 = stratified, inversely to normally graded.
(from Krause & Oldershaw [1979])
Since the matrix is predominantly "sandy" oolitic material (along with carbonate mud) the suspension competence of the matrix would be greatly increased [Pierson 1981] over that of a pure mud slurry, which would slow settling of the flow. This appears to be as a result of the creation of excess pore pressures which would increase buoyant forces [Hampton 1979], which Pierson [1981] estimated to be responsible for supporting 2/3 of the coarse sediment load within the flow. Grain to grain contact and cohesive strength would have provided the remainder of the forces necessary to support the larger clasts. With increasing clast size, the amount of clast support provided by cohesive strength decreases [Pierson 1981; Kessler & Moorhouse 1984], excess pore pressures and dispersive pressures becoming more important. This excess pore pressure develops from the inability of the matrix fluid to escape from the debris [Lewis et al 1980] and greatly enhances mobility of the flows [Hiscott & James 1985]. Debris flows with a sandy matrix have been termed "sandy debris flows" by workers such as Pierson [1981], Kessler & Moorhouse [1984] and Hurst & Surllyk [1982]. It is the strength and buoyancy of the debris-flow that permits the flow to entrain and support clasts that are more dense than the bulk density of the flow itself, and to support them above the bed [Pierson 1981; Hiscott & James 1985]. In this case, a flow of a metre thick could support large boulders of wackestone and mudstone up to 1m (maximum dimension) or less.

The graded oolitic packstones immediately overlying the debris flow are interpreted as being deposited from the turbiditic tail of the debris flow. The association of turbidites with debris flows have been documented from both ancient and modern debris flows [e.g. Krause & Oldershaw 1979; Crevello & Schlager 1981]. The oolitic packstones are interpreted as representing $T_{ab}$-type lower density suspension-traction sediments. The finely, but strongly, laminated wackestones at the very top of the unit are interpreted as forming by successive fall-out from the fast flowing residual turbidite cloud, forming high-velocity plane lamination in these finer sediments [Aalto 1976; Hiscott & Middleton 1979; Lowe 1982]. The pelagic mud veneer is interpreted as forming from $T_e$ type suspension fall-out of the finest material in the turbidite cloud [c.f. Bouma 1962].

Massive, belemnite-rich pebble conglomerate

This facies is defined as massive, sand-grade oolitic-peloidal beds that contain clasts which never exceed an upper size limit of 64mm, scattered throughout the bed. The
presence of numerous belemnites within the flow is a characteristic feature of this facies.

This facies occurs as a single 3.5m thick unit at Amoixa (Section 3.3.4.1) (Plate 3-III,A) where it overlies condensed nodular wackestones (Plate 3-III,B). Weathering has completely obliterated internal texture at outcrop revealing no indication of the underlying nature of the flow.

The conglomerate unit has a thin inversely graded horizon at the base, where it contains stratified, poorly sorted and coarse sand size grade (500-750µm) peloids and ooids. The unit is otherwise massive and ungraded, apart from the top 10cm. The clasts are scattered throughout the flow, are all rounded to sub-rounded and usually do not exceed 4cm across (Plate 3-III,C & E). These clasts are predominantly wackestones, usually containing high proportions of Posidonia filaments (Plate 3-III,D), along with rarer clasts of mudstones with scattered ooids and pelagic mud clasts. The poorly sorted matrix is oolitic grainstone (locally packstone), sediment components ranging between medium sand size to very coarse sand sized (500µm - 100µm) ooids and peloids, the peloids frequently being over-sized (i.e. larger than the matrix ooids). Both in hand specimen and thin section, the clasts show pronounced pressure solution contacts with the surrounding ooids and peloids of the matrix (Plate 3-III).

The numerous belemnites within this unit show a crude preferred sub-horizontal to horizontal orientation within the flow.

Interpretation

The planar upper and lower boundaries, sheet-like nature and random distribution of clasts within the flow fit in with descriptions of carbonate debris flow deposits by Cook et al. [1972], Enos & Moore [1983], Cook [1983a] and Cook & Mullins [1983]. Since the matrix is 'sandy' (i.e. ooids and peloids and not mud), this unit classifies as a sandy debris flow [c.f. Surlyk 1973] and as a fine-grained debris flow (after Hampton [1975] and Shanmugam & Benedict [1978]) since the constituent clasts never exceed pebble grade.

The flat base implies emplacement by laminar flow [Middleton & Hampton 1976], the narrow inversely graded basal horizon indicating traction and shear induced by the load.
Plate 3-III

A) Basal debris flow, Amoixa.

B) Nodular wackestones underlying basal debris flow.

C) Belemnite-rich pebble conglomerate debris flow. Oolite-peloidal matrix showing over-sized peloids (arrow) and rounded *Posidonia* rich wackestone clasts. Basal conglomerate unit 1 (Fig. 3-26) (polished hand-specimen).

D) Stained thin section microphotograph of typical wackestone debris flow clast, containing long *Posidonia* filaments, and radiolaria. Matrix ooid, bottom right-hand corner. (Width of field of view 1.75mm)

E) Clast-rich matrix supported debris flow, containing belemnites (arrow - this belemnite has been sampled for isotope analysis). Basal conglomerate unit 1 (polished hand-specimen).

F) Clast-supported graded conglomerate (debris flow), showing pressure solution clast contacts (polished hand-specimen).
imposed by the overlying flow [Lowe 1982]. The poor sorting and suspension of small clasts throughout the body of the unit indicate deposition by rapid frictional freezing [Postma 1986]. This lack of continuous size grading implies that dispersive pressures, formed by a high degree of grain collisions, along with the buoyancy effect of the interstitial fluid, were important fluid support mechanisms [Cas 1979]. The matrix-supported nature of the flow indicates clast-clast collisions did not play an important role in flow mechanics. Minor grain collisions and traction between the matrix ooids and peloids were probably the most important flow support mechanisms.

This added matrix effect on dispersive pressures within sandy debris flows is not taken into account by the usual debris flow models [e.g. Krause & Oldershaw 1979; Eberli 1987] where a mud matrix is taken as standard. Hence, this debris flow does not readily fit into one of Krause & Oldershaw's classification models (Fig. 3-12). The occurrence of large quantities of belemnites and Posidonia-rich wackestone clasts which in this unit indicates the incorporation of slope derived material into the flow, which is discussed on context in Section 3.3.4.1.

Clast-supported, graded pebble conglomerate

This facies is defined as graded oolitic peloidal units that contain a basal clast-supported layer where the clasts never exceed 64mm.

This is a rare facies, only found as one bed at Amoixa (Section 3.3.4.1). A lower thin conglomerate horizon 20cm thick is abruptly overlain by laminated wackestones containing scattered ooids. Hardly any matrix is present and the clasts have been heavily compacted, so that pressure-solution contacts form between adjacent clasts (Plate 3-III,F). The clasts generally range in size from 0.5-4cm, and are fine-grained, Posidonia-rich, light yellow/dark brown wackestones, which show internal plastic deformation; and muddy wackestones with scattered ooids (Plate 3-III,F arrow). This horizon is overlain by 30cm of peloidal laminated Posidonia-rich wackestone.

**Interpretation**

This unit conforms to the normally graded, clast-supported debris-flow models of Krause & Oldershaw (Fig. 3.12, No.4), which implies it was derived through downslope evolution from a mud-supported debris-flow. The clasts are compacted to such a
degree that the clast-rich horizon may be classified as a stylobreccia [c.f. Steiger & Jansa 1984]. The clast-rich nature of this horizon implies efficient flow separation between the clasts and finer sediment fraction, but need not imply that matrix support was lacking during transport [e.g. Enos & Moore 1983]. Matrix volume can be reduced considerably by post-depositional dewatering [Enos & Sawatsky 1981] and fine matrix is also susceptible to preferential removal during stylolite formation between clasts. The clast composition indicates a slope derivation and the fine-grained overlying sediment cap is interpreted as being deposited by the turbidite cloud overriding the emplacing debris flow. The lamination is interpreted as lower density high-velocity plane lamination formed by suspension fall-out and reworking by the fast flowing tail current [c.f. Lowe 1982].

Graded pebble conglomerate

This facies is defined as graded oolitic-peloidal beds that contain a basal clast-rich interval that forms >30% of the bed thickness. The clasts may be of variable size but never exceed 64mm.

Only found at Amoixa (section 3.3.4.1), these beds are 2-3m thick and contain clasts ranging from 1cm up to small cobble size (64mm) set in a matrix of peloidal oolite packstone. The clasts may be inversely graded at the base, clast size maximising and then grading normally to 2mm sized clasts at the very top of the clast-rich horizon. The wackestone and mudstone clasts tend to be stratified in horizontal to sub-horizontal layers and stringers towards the top of this zone, and are overlain by oolitic peloidal packstones with oversized peloids that show coarse-tail grading. With distribution grading characteristically developing towards the very top of the bed.

Interpretation

This association classifies as a stratified, inversely-normally graded debris flow of Krause & Oldershaw [1979] (Fig. 3-12 No. 3) and as such is likely to have evolved from an up-slope disorganized debris flow. The inversely to normally graded clast-rich horizon is analogous to Lowe's [1982] R_2 and R_3 divisions (respectively), for high-density turbidites of gravel grade, the basal R_2 layer forming by traction in the flow base [Aalto 1976; Lowe 1982], the normally graded R_3 layer by suspension fall-out and traction, as velocity dropped below that necessary to maintain the traction carpet. Once the clasts were deposited, a residual sandy high-density turbidity current rapidly
deposited the overlying oolitic-peloidal sand as traction-suspension fall-out waves that show coarse-tail grading. This facies shows features of being transitional from debris flows to high-density turbidites, although the volume and size of clasts warrants its classification as a conglomerate.

3.2.2.2 Graded oolitic calcarenites

Calcarenites formed in basinal deposition settings which are graded, are termed calciturbidites [eg. Eberli 1987; Colacicchi & Baldanza 1986], the presence of graded bedding qualifying their interpretation as turbidites after the scheme of Bouma [1962]. The majority of turbidites within the Cutri Formation however, conform to Lowe's divisions for high-density (sand-rich) turbidites [Lowe 1982] rather than the traditional Bouma T_a-e classic turbidite sequence (Fi. 3-13). Bed thicknesses range in size from a few centimetres to three metres, and oolitic turbidites form the greater percentage of vertical section in all localities. The turbidite sediments involved are oolitic-peloidal sands, which range in size from granule grade to very fine sand grade (2000µm-125µm).

These oolitic turbidites show several characteristic features:

(1) The turbidite facies are composed of oolitic-peloidal packstone, rarely oolitic-peloidal grainstone. Occasional cement-filled intergranular "pores" may be present within the packstone (Plate 3-IV,B).

(2) The oolitic packstones and grainstones are usually strongly bimodal, characterized by the occurrence of oversized peloids within the oolitic sediment (Plate 3-IV,B). These peloids exceed the equivalent oolite size grade by one to four size divisions (i.e. 500µm oolitic sediment may contain 2000µm peloids). These larger oversized peloids are also preferentially concentrated in the base of the flow, where they may make up the greatest percentage of the sediment fraction and they also pick out the coarse-tail graded nature of the turbidites. This is a form of grading produced whereby only the coarsest percentiles in the matrix show grading [Middleton 1967; Middleton & Hampton 1976]. Which in this case is picked out by variations in both the size and sediment fraction density of oversized peloids in the oolite "matrix". Although the peloids effectively pick out coarse-tail grading, crude distribution grading (normal grading) between coarse and medium sand grade (750-375µm) may also be seen.
Figure 3-13

Turbidity current models for coarse-grained sandy and gravelly turbidites; medium-grained mixed sediment turbidites; and fine-grained turbidites.
COARSE-GRAINED TURBIDITE

MEDIUM-GRAINED "CLASSIC" TURBIDITE

FINE-GRAINED TURBIDITE

(after LOWE, 1982)

(after BOUMA, 1962)

(after PIPER, 1978 and STOW and SHANMUGAM, 1980)

FIG 3.13
(3) There is efficient separation of very fine to fine sand grade peloidal material, which is preferentially concentrated in a laminated cap at the top of the graded granule to medium sand grade turbidite bed. This gives the turbidites a strongly bimodal profile, due to this two-fold flow segregation of sediment component grain sizes.

Pelagic bivalve filaments (*Posidonia*) are very efficiently flow sorted and are only found within the very fine sand to fine sand (95µm-187µm) turbidite fraction, where they are often concentrated in parallel laminae.

(4) There is a marked paucity of bioclastic components within the sediment fraction, which may occur both as ooid nuclei and free sediment components. Trocholine foraminifera occur predominantly as ooid nuclei but may also be 'uncoated'. Echinoderm fragments are less common sediment components, occurring as ooid nuclei but usually as free sediment components. They only develop syn-taxial overgrowths in grainstones (Plate 3-IV,C), these features not developing in packstones (Plate 3-IV,D). The most abundant bioclasts are the *Posidonia* bivalves, concentrated in the sandy turbidite fine-grained caps.

(5) Many of the turbidites have incorporated small (millimetre size to 2cm) slope-derived, wackestone and pelagic mud rip-up clasts into their basal flow division (Plate 3-V,B). This process of entrainment of slope material as clasts within the turbidite occurs due to the impact of fast-flowing turbidites onto the seafloor [Mutti & Normark 1987] and is a feature of high-density turbidite [Kano & Takeuchi 1989]. These rip-up clasts pick up flow amalgamation surfaces within many oolitic units.

(6) Extremely rarely, detrital quartz grains may form ooid nuclei (the author has observed only five such ooids) and in one case only, a muscovite lath was observed within an ooid nucleus. The presumed origin for this quartz was the Alboran-Kabylia landmass, situated to the east of Mallorca during the Middle Jurassic (Fig. 1-2,A & B). Most frequently, however, ooid nuclei are peloids of presumed faecal origin.

(7) Bouma Tc division cross-bedding and convolute lamination are never observed within the major turbidite units (although this may conceivably be due to the weathered aspect of the outcrop). The general absence of these features is interpreted as being due to emplacement by high-density turbidites [Middleton & Neal 1989] but also seems to be a feature of many other resedimented carbonate turbidites. Bouma [1962] observed to division sedimentary structures in up to 60% of siliciclastic turbidite beds.
Plate 3-IV

A) Bimodal oolitic packstone turbidite, showing over-sized peloids and radial ooids. Pressure solution contact of intraformational mudstone clast with matrix and ooids. Unit E, southern section Puig Cutri. (Stained thin section, field of view 3mm across)

B) Strongly bimodal oolitic-peloidal packstone, with large over-sized peloids. Pressure solution contacts occur between over-sized peloids (arrow). Some occasional inter-granular "voids" are matrix free, and filled with a single generation of blocky (ferroan-calcite) sparry cement, some of these show geopetal micrite at the base of the "pore" (top left hand corner). (Unstained thin section, field of view 3mm across)

C) Well sorted oolitic grainstone turbidite, showing well developed echinoderm syn-taxial overgrowth. Note blocky, single generation of sparry cement and calcitized, dolomite rhombs (arrow), presumed ferroan originally, with limonite residual blebs. Turbidites unit 6, Amoixa. (Stained thin section, field of view 3mm across)

D) Well sorted oolitic packstone, showing well developed mechanical and chemical compaction features. Note lack of syntaxial overgrowth on echinoderm fragment. Lower S2 traction layer Unit E (Section 3.2.3). (Stained thin section, field of view 3mm across)
Basal sedimentary structures such as well-developed groove and flute casts, so often beautifully exposed in siliciclastic turbidite systems, are rarely observed. This is partially due to the nature of outcrop which very rarely exposes the undersides of beds. The poor development of basal sedimentary structures or their absence does, however, seem to be a feature characteristic of calciturbidites [e.g. Colacicchi & Baldanza 1986; Eberli 1987; H. Cook & M Hampton personal communication 1987]. It is as yet unclear whether this is due to the nature of the fine background sediments into which the turbidites erode or due to the nature of the composition of the turbidites. It is likely to be due to a hydrodynamic density-related feature of carbonate turbidites [M. Hampton personal communication 1987].

Secondary silicification (chert) is found associated with the turbidite units (as well as in intervening hemipelagic units). Chert layers or meganodules typically occur within narrow zones of a few centimetres, and are generally between 1-5cm thick, and occurring preferentially in the coarse basal turbidite horizon when they directly overlie a hemipelagic unit (Plate 3-V,B & C). Chert mini-nodules of a few millimetres or less in diameter are typically scattered throughout the turbidites, usually partially replacing large peloids. It is these mini-nodules of chalcedony that Bourrouilh [1973] identified as detrital quartz. Since free detrital quartz grains are absent from the succession, it is presumed that siliceous radiolaria debris in the intervening hemipelagic sediments and within the packstone matrix was the main source of silica for the chert. This would be in agreement with the postulated source of chert silica from other carbonate turbidite systems [for example Bustillo & Ruiz-Ortiz 1987; Dombrowski 1987; Eberli 1987]. Silicification appears to have commenced early on in the diagenesis of the sediments.
Massive to poorly graded oolitic turbidites

This facies is defined as oolitic-peloidal beds which are 90% massive and indistinctly graded, but which may be inversely graded at the base and distribution graded at the very top.

This facies is restricted in its occurrence to Amoixa (section 3.3.4.1). The oolitic-peloidal beds are up to 2.8m thick and comprise indistinctly graded and poorly sorted granule-medium sand grade (350-2000µm) grainstones. Oversized peloids make up the 1000µm-2000µm fraction and may be concentrated, over a few centimetres above a thin inversely graded basal oolitic horizon of 1-2cm. The very top few centimetres of the bed may show distribution grading to medium sand grade (350-500µm). A mudstone drape (T_e division) is absent.

Interpretation

Similar thick, massive-poorly sorted sandstone units may have been interpreted in the past as grain flow deposits (Fig. 3-11) but are now recognized as as high-density turbidites, for example MacDonald [1986], Hein [1982], Wueller & James [1989] and Spelletti et al. [1989]. The association of massive-poorly sorted turbidites with classical turbidites is common, and according to Lowe [1982] are manifestations of rapid deposition from high-density turbidity currents by rapid suspension fall-out of sediment. The slightly inversely graded base is the result of basal traction and shear imposed by the body of the flow [c.f. Lowe 1982]. Grain suspension within the flow was likely to have been sustained by the upward flow of pore fluid, combined with additional lift (dispersive pressure) caused by grain-grain collisions between the oords and peloids [c.f. Bagnold 1954; Postma 1986; Kano & Takeuchi 1989]]. Clean sand-rich flows such as these would be expected to have been deposited more rapidly than their mud-rich counterparts, due to the greater permeability of the mud-free sediment involved.

Coarse tail graded oolitic turbidites

This facies is derived as graded oolitic peloidal sediment where only the coarsest sediment fraction displays gradation [c.f. Middleton 1967]. In this facies it is the oversized peloids which display the coarse-tail grading. They may contain clasts in their basal divisions, but these never exceed 64mm (usually between 3mm - 2cm), always form >20% of the total bed thickness. This distinguishes this facies from the
graded pebble conglomerates. Bases may be typically sharp and planar, or slightly erosive (Plate 3-V, A). This facies is the most abundant resedimented facies. The turbidites are typically 50cm-4m thick and may be amalgamated or occur as distinct units. This facies is strongly bimodal and characterized by the occurrence of abundant oversized peloids. At the base of the flow, oversized peloids may exceed the oolite grade by up to five size divisions, i.e. 2500µm peloids may occur along with oolite of 500µm grade. In the very basal division, the sediment may consists almost entirely of very coarse sand to granule grade (1000µm-2500µm) peloids. Inverse grading of these peloids locally occurs within the basal few centimetres (<3cm). Rip-up wackestone and mudstone clasts may occur in this zone, but do not exceed 3cm, typically being 3mm-1cm in size. If present, this clast rich zone (which never exceeds >20% of the total bed thickness) may show inverse grading and then normal grading of the clasts. These turbidites typically contain coarse to very coarse sand grade (1000-500µm) oolitic sediment near the base, indistinctly grading to coarse-medium sand grade (500µm-250µm) sediment near the very top. Oversized peloids show a greater size difference compared to the component oolite in the basal portion of these turbidites, as well as forming a greater percentage of the total sediment fraction.

Stratified peloidal-rich horizons are characteristic features of these turbidites, and are overlain by more massive divisions with scattered oversized peloids. This pattern is often repeated up through the bed, the stratified peloid horizons and overlying massive divisions showing a gradual reduction in size of oversized peloids compared to the oolite fraction which shows little or no size change. The oolite fraction usually becomes slightly finer towards the top of the bed, and in these upper regions the peloid fraction may no longer be oversized with respect to the oolite.

Thin parallel laminated peloidal packstone or wackestone units may abruptly and distinctly overlie the major oolite-rich divisions, which may in turn be overlain by pelagic mud drapes however one or both of these divisions may be absent.

**Interpretation**

The sediments of this facies are interpreted as being rapidly deposited by "sandy" high-density turbidites (after the model of Lowe [1982], Fig. 3-13). Their packstone nature indicates sediment concentration was high during their emplacement, and the beds were rapidly deposited trapping finer sediment within the deposited flow. This is a characteristic feature of high-density turbidite deposition [Middleton & Neal 1989].
A) Sheet-like base of unit C, overlying parallel laminated massive wackestones and fissile pelagic. Southern section, Puig Cutri. (Hammer 30cm)

B) Basal chert layer and chert nodule, base of unit I, log 2 (Appendix Fig. 2). Southern section, Puig Cutri. (Hammer 30cm)

C) Basal inverse to normally grade clast-rich R₂-R₃ horizon of uppermost unit H, log 6, southern section Puig Cutri. Note concentration of over-sized peloids at base of photograph, rounded nature of clasts, mini-nodular chert, and abrupt transition to overlying clast-free turbidite. (Lens cap 5cm)

D) Thick bedded siliceous wackestone interbeds, central section, Puig Cutri. (Hammer 30cm)
beds deposited from more concentrated suspension tending to be both less well graded and contain far more interstitial fires. The basal clast-rich horizon (if present) conforms to the R₂ and R₃ divisions of Lowe: the R₂ division showing inverse clast grading, the R₃ division showing normal (distribution) clast grading.

The stratified peloid horizons conform to the S₁ and S₂ (traction sedimentation) divisions of Lowe [1982] while the normally graded to massive, unstratified horizons conform to the S₃ (suspension sedimentation) division. The repetition of successively finer, oversized peloid and often oolite grade S₂ and S₃ divisions, is indicative of surging high-density flow, successive sediment waves depositing lower density sediment out of suspension [Luthi 1981; Lowe 1982]. The parallel laminated packstone/wackestone cap is interpreted as being deposited by the residual, fast flowing turbidite tail [c.f. Lowe 1982], while the pelagic mudstone drape represents long-term suspension sedimentation. A detailed profile through a single high-density oolitic turbidite is discussed in section 3.2.3 which discusses in more detail both the sedimentological and hydrodynamic depositional processes involved in their formation.

Graded oolitic turbidites

This facies is defined as oolitic pelodial beds that are dominated by distribution grading (where the entire sediment fraction grades throughout the bed [Middleton & Hampton 1976], although basal divisions often retain coarse-tail grading. These turbidites are thinner than the coarse-tail graded turbidites; being <40cm thick. The base of the turbidites is planar, as is the top. These beds show a clearer development of distribution grading from basal coarse to medium sand grade (750 - 350µm) oolitic-peloidal sediment up to sediment of fine to very fine sand grade (187-125µm). If present, oversized peloids do not vary significantly with respect to the component oolite fraction, typically showing a difference of only one size grade. For example, a basal 350µm oolite sand may contain 500µm oversized peloids. Often the peloids are of the same grade as the component oolite. There is a paucity of mudstone rip-up clasts, rare occurrences are sporadic and always <4mm across. The only internal sedimentary structures observed other than grading, is parallel (amination, which may develop towards the top of the bed. A pelagic mud-drape (Te division) may cap these beds.
Interpretation

These terbidites, which are still "sand-rich", represent lower density Tab type sandy turbidites. As such they are representative of "distal" major high-density turbidites. The basal graded layer represents the Bauma Ta division, and forms the majority of the bed. The parallel lamination is analagous to the Bouma Tb division (Fig. 3-13).

3.2.2.3 Fine calcarenites, calcisiltites and other facies

Thin bedded fine calcarenites and calcisiltites

This facies takes the form of medium to dark grey, thinly bedded (10-1cm) fine sand to silt grade peloidal packstones and wackestones, which may be interbedded with massive or fissile pelagic mudstone (Plate 3-V, A & D; 3-VI,C). The packstones and wackestones are well sorted, occasionally crudely graded, and either massive or parallel laminated. They are characteristically *Posidonia* and radiolaria rich, the *Posidonia* filaments aligned with the lamination. Occasionally a few scattered ooids may occur within the laminae of these sediments (Plate 3-VI, A). There is an absence of microfauna; burrowing does occur, but is not important and does not noticeably disrupt bedding. There is a strong emanation of hydrogen sulphide upon breakage or crushing of these rocks and an association with chert is common.

Interpretation

The dark colour and absence of calcareous benthos is common to other deep water fine-grained carbonate facies (e.g. Evans & Kendall 1977; Cook 1983a; Watts 1987) and indicates a poorly oxygenated dysaerobic environment (Byers 1977; Yurewicz 1977). Ichnofauna were not prolific; the paucity of vertical burrows and absence of bioturbation destroying original sedimentary fabrics is evidence for the development of anaerobic conditions in the sediment column (Yurewicz 1977). The emanation of hydrogen sulphide upon crushing indicates bacterial sulphate reduction was occurring under anoxic conditions (Goldhaber & Kaplan 1974).

This facies represents the 'normal' background sedimentation of the basin, most of which is interpreted as being hemipelagic after Scholle et al. (1983). True hemipelagic sediment is defined as downslope resedimented material delivered by dilute turbidite suspension, or settling out from bottom currents; whereas true pelagic sediments are
derived from the slow settling out of biochemically produced material from the water column. The term hemipelagic is used by most carbonate workers to include a mixed assemblage of fine slope derived and pelagic material [e.g. Cook 1983a; Cook & Mullins 1983] and it is in this context that the term is used in this case.

The massive and fissile structureless mudstones are interpreted as pelagic mud, while the fine packstones and wackestones were bank-derived and could have a number of origins: some undoubtedly originated from dilute, low density turbidity currents, or settled out form low-density suspensions of bank-derived fines, carried offbank in response to storms [e.g. Mullins & Neumann 1979]. The predominance of ungraded beds favours the latter interpretation over the former, although some fine graded layers undoubtedly represent deposition by dilute turbidity currents. These thin graded layers show distribution grading from fine to very fine sand grade (187-95µm) sediment to pelagic mud, and are likely to represent dilute turbidites of Bouma Tde affinities, represented by the fine-grained turbidity current model of Stow and Shanmugam [1980] (Fig. 3-13). The predominance of massive, well sorted, parallel laminated beds may indicate a gentle reworking by bottom currents [Bouma 1973] or they may represent high-velocity lags deposited from the tails of high-density turbidites (see Section 3.2.3). Within fine-grained facies, there is a problem in distinguishing deposits of dilute turbidity currents, hemipelagic sediments and sediments formed by current action [Stow & Lovell 1979]. The wackestones and packstones probably entered the basin via dilute turbidites as well as storm suspensates, and subsequently may have been gently modified by bottom currents.

Fine laminates

These dark sediments are very fine sand to silt and mud grade, and are finely laminated on a millimetre scale (1-10mm) (Plate 3-VII,A, B & C). Laminations may be slightly parallel or inclined. The laminated sediments are well-sorted microskeletal peloidal packstones and wackestones, containing radiolaria and Posidonia, the constituent grains being aligned parallel to the lamination (Plate 3- VIII, D). These layers of fine sand to silt grade sediment may alternate with equally thin layers of pelagic mud. Burrowing may occasionally disturb these fine laminations (Plate 3- VIII,C); large horizontal Planolites type burrows also occur in this facies (see Section 3.3.3.1 and Plate XII, O). Macrofauna are characteristically absent, while large chert nodules are present and locally abundant.
Plate 3-VI

A) Occasional scattered ooids within *Posidonia*-rich laminae of hemipelagic interbed unit. May represent dilute turbidites, or storm suspensate periplatform sediments. Post unit Q, southern section Puig Cutri. (Stained thin section, field of view 3mm across)

B) *Posidonia*-rich laminated turbidite cap, showing alignment of *Posidonia* filaments, preferentially deposited convex-up, set in fine peloidal packstone sediment. Also note radiolaria (white-beige spheres). Unit H, southern section Puig Cutri. (Stained thin section, field of view 3mm across)

C) Hemipelagic fine peloidal wackestone interbed facies, with fragments of *Posidonia*. Southwestern Puig de Ses Fites section. (Stained thin section, field of view 3mm across)

D) Hemipelagic interbed facies of units S-T (uppermost Cutri Formation). Laminated nature of fine peloidal sediment, and winnowed pelagic bivalve fragment laminae. These facies are interpreted as contourites. Upper Cutri Formation, Quarry section, southern section Puig Cutri. (Stained thin section, field of view 3mm across)
**Interpretation**

The multiple layering, well-sorted nature and absence of grading distinguish these fine sediments from distal turbidites. They were interpreted as contourites by Alvaro *et al.* [1984] and are analogous to descriptions of other ancient carbonate contourite deposits described in the literature [e.g. Yurewicz 1977; Cook 1983; Eberli 1987]. Good sorting, fine parallel and inclined laminations and microscopic grain orientation are all features of sediments interpreted as contourites [Bouma 1973; Pickering *et al.* 1986]. Most sediments that form contourites originate from turbidity currents that brought the sediment into the basin [Heezen & Hollister 1964] and were subsequently reworked by bottom currents [Stow & Lovell 1979].

The fine sand and silt grade platform-derived sediments may have entered the basin both as dilute turbidites and as storm suspended fines, which were subsequently gently reworked into contourites. The gentle bottom currents may have contributed to the oxygen content of the water mass, providing enough oxygen to support the very limited trace fossil assemblage. Both 'sandy' and 'muddy' contourite types [Pickering *et al.* 1986] are distinguished in the Cutri Formation, the contourites tending to become 'muddier' up-section (see Section 3.3.3.1). *Posidonia* coquinas

A coquina is defined as a wave or current formed accumulation of shells, and these distinctive facies are characterized by concentrations of the valves of the small pelagic bivalve *Posidonia*. These bivalves were thin-shelled, nektoplanktonic Pectinaceae bivalves [Jefferies & Minton 1965]. Up to 1cm long and very thin-shelled (60-70µm) they are of characteristically uniform thickness along their length. They are geographically widespread in basinal facies of the Tethyan Jurassic [Arkell 1956]. The *Posidonia Bositra buchi* (Romer) (or *Posidonia alpina* (Gras)) ranges from the Toarcian-Kimmeridgian boundary [Kuhry 1975] and it is likely to be the species of *Posidonia* concentrated in these coquinas.

As a pelagic organism, their concentration marks episodes of reduced or hiatus sedimentation and/or a concentration and winnowing by ocean currents and are described elsewhere in the Tethyan Jurassic by Sturani [1967], Kuhry *et al.* [1976], Bernoulli & Renz [1970] and Steiger & Jansa [1984]. These facies are discussed in context in Section 3.3.1.2 and shown in Plate 3-XV, E & F.
Plate 3-VII

A) Contourites underlying unit E, southern section Puig Cutri.

B) Unit E, showing positions of sub-units 1, 2, 3, underlying contourites and unit H. (Figure for scale)

C) Polished hand specimen from the outcrop of photograph A. Note fine laminae, occasionally interrupted by burrows (centre left) burrow mottling (upper third) and chert nodule (centre right).

D) Acetate peel of sample in photograph C. (Field of view 3mm). Note silicified radiolaria (beige spheres), pelagic bivalve (centre, far right) and fine peloids (red spheres).
3.2.2.4 Facies Associations

This section briefly describes and introduces the facies associations recognized within the Cutri Formation, that are dealt with in detail in Section 3.3. The lithofacies types are grouped into five facies associations, which form the mega-sequence rift-infill succession of the Cutri Formation: (1) Conglomerate association; (2) Thick bedded and amalgamated high-density turbidite association; (3) Thin bedded high-density turbidite association; (4) Thin bedded turbidite association; (5) Hemipelagic unit association.

Conglomerate association

This association is restricted to the basal part of the succession being especially well developed at Amoxia (Section 3.3.4 & Fig. 3-25), and is also observed at Puig Cutri, Puig de Ses Fites and Puig de Ses Fites II (Section 3.3.1 & Fig. 3-22). This association occurs along with the thick bedded and amalgamated turbidite association. The conglomerates are associated with distinct horizons within the succession, that are laterally correlatable along strike. These conglomeratic, debris-flow events may be traced laterally or palaeobasinwards into their 'distal' equivalent high-density turbidites originating from the same event (Fig. 3-22).

Thick bedded and amalgamated high-density turbidite association

This association is restricted to the basal part of the successions at Amoxia, Puig Cutri, Puig de Ses Fites and Puig de Ses Fites II and is intimately associated with the conglomerate association. This thick, amalgamated turbidites may form basal units in the mega-sequence up to 20 m thick, which consist of up to 25 recognisable amalgamated flows (Appendix Fig. 2; Fig 3-22). More frequently, these amalgamated units range between 15-3 m, and consist of 10-3 high-density turbidite events. These basal units are laterally extensive and correlatable along palaeo-strike over a distance of up to 70 km, individual "events" within the units being correlatable between sections locally and over this distance.

Thin high-density turbidite association

This association characterizes the medial sections at Puig Cutri, Puig de Ses Fites and Puig de Ses Fites II (e.g. Appendix Fig. 2; & Fig 3-22); the upper part of the tc incomplete succession at Amoxia (Fig. 3.25) and the section at Puig d’en Pare (Fig.3-fd
24). These high-density turbidites are interpreted to form as the product of single major turbidite events, and are encased in hemipelagic background sediments. These turbidites are typically between 1-3m thick, and usually show full development of the Lowe [1982] high-density turbidite model sequence (see Section 3.2.3).

Thin bedded turbidite association

This association is developed at the top of the mega-sequence at Puig Cutri (Appendix Fig. 2), Puig de Ses Fites (Fig. 3-22) and Puig de Ses Fites II (Fig. 3-22); and towards the top of the logged sequence at Puig d'en Pare (Fig. 3-24). It is characterized by thin, distribution graded oolitic turbidites, 50cm-3cm thick, usually occurring as isolated turbidite 'events' within a succession dominated by fine-grained hemipelagic facies.

Hemipelagic unit association

This association is found throughout most of the succession at all localities, becoming volumetrically more important up-section. This association includes both of the fine-grained facies of Section 3.3.3.2, which occur as oolite unit hemipelagic interbed facies. The sedimentological nature of these interbeds is laterally correlatable; for example, contourites appear at the same stratigraphic horizon between different outcrops of the mega-sequence.

The development of this association represents episodes of non-resedimentation of oolite and record the back-ground sedimentation of the basin. When occurring between major oolitic units, they mark episodic abandonment of the turbidite system, which became draped with peri-platform fines and pelagic mud. The transition from oolitic units to the hemipelagic units is abrupt and basin-wide in extent (Appendix Fig. 2; 3-22; 3-26), indicating an allocyclic control on facies development [Surlyk & Hurst 1984; Stow 1985a; Larue 1985; Spalletti et al. 1985].
3.2.3 Hydrodynamics and sedimentology of a major high-density oolitic turbidite

Introduction

A fresh exposure due to blasting, of a turbidite referred to as unit E (Plate 3-VII, B) in the southernmost outcrop of the southern section at Puig Cutri (at the location of log 1, Appendix Fig. 2, see Section 3.3.1.1) provided the only exposure where fine internal sedimentary structures could be photographed (Plates 3-VIII and 3-IX). The detailed profile produced (Fig. 3-14) allowed the interpretation and discussion of the hydrodynamics affecting the sedimentation of high-density oolitic turbidites, by analogies with identical features developed in high-density siliciclastic clastic turbidites as proposed by Lowe [1982] (Fig. 3-15). As far as the author is aware, this example is the first detailed sedimentologic description of any single carbonate turbidite in terms of hydrodynamics, and is also the first documented example to be described in detail, of a high-density carbonate turbidite that conforms to Lowe's [1982] model.

Unit E consists of a two metre thick resedimented oolitic unit, underlain by contourite facies (Plate 3-VII, B) and overlain by dark, fissile, shaley very fine wackestones; the latter representing the final suspension phase of turbidite resedimentation, equivalent to the Bouma Te division (Figs. 3-14 & 3-13). Interpreted as being deposited from a single high-density, surging turbidity current [Luthi 1981; Lowe 1982] and unit E displays a complete turbidite sequence. High-density turbidites of this calibre are generally held to be produced by major 'catastrophic' turbidite generating events, such as earthquakes or mass slope-failure [Nelson & Nilsen 1984; Mutti et al. 1984].

For ease of description and discussion, the unit is subdivided into three sub-units (Fig. 3-14 & Plate 3-VII, B): basal Sub-unit 1, a basal conglomeratic division overlain by a graded oolitic division; medial Sub-unit 2, more massive and stratified and dominated by alternating S2 and S3 divisions [after Lowe 1982]; and uppermost Sub-unit 3, dominated by suspension sedimentation of fine sand grade packstones and wackestones. The unit shows an overall fining-upward trend and features observed are discussed with reference to the model developed by Lowe [1982] for the description of high-density sandy turbidites (Fig. 3-15).
Figure 3-14

Profile through unit E, southern section of Puig Cutri, with hydrodynamic explanations and comments.
dark, fissile, mudstone drape

parallel lamination

fluid escape structures – dish structures

ripples?

minor distribution-graded pulses

distribution grading

stratified tabular wackestone chips

coarse-tail grading

stratified wackestone chips

coarse-tail grading

stratified peloid layers

inversely graded peloids

coarse-tail grading

concentrated stratified peloids

distribution grading of wackestone clasts

deposition by traction within a traction carpet

direct suspension from high-density flow

formation by traction at base of high-density flow

direct suspension from rapid freezing of highly concentrated flow

deposition by traction within a highly concentrated traction carpet

erosion by head and basal shear

long-term suspension of finest fraction

high-velocity plane lamination

high concentration traction sediment from low-density over-riding turbidite tail

rapid deposition of loosely-packed sediment by direct suspension, from slackening high-concentration flow

lag deposition from high-density flow

direct suspension hindered by turbulence from waning high-density flow

successive traction carpet formation followed by rapid direct suspension

deposition within a traction carpet

direct suspension from high-density flow

formation by traction at base of high-density flow

direct suspension from rapid freezing of highly concentrated flow

deposition by traction within a highly concentrated traction carpet

erosion by head and basal shear

Processes

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Sub-unit 1

The basal division of unit E commences with a clast-dominated division, 25cm thick (Plate 3-VIII, A). The clast-dominated conglomeratic division is matrix supported but clast-rich. The clasts are characteristically restricted in size between a few millimetres to 3cm and are sub-rounded to rounded, and of fine-grained slope-derived wackestones and pelagic mud. They are randomly orientated in 3D, but arranged parallel to bedding. The matrix is a coarse, grainy, 500µm (medium sand grade) oolite, with a concentration of large oversized peloids ranging between 750µm-2000µm (very coarse sand to granule grade).

This division is almost planar with its contact with the underlying contourites at this locality. A large contourite clast is observed 'frozen' in the process of being torn-up and becoming plastically deformed at the flow base (Plate 3-X, E). The basal ten centimetres shows inverse clast grading and a concentration of oversized peloids, the remainder of the clast-dominated division showing normal clast grading (Plate XVII, A).

The features of the clast-dominated horizons of sub-unit 1 conform to the R2 and R3 divisions (for gravel and sand grade sediment) of Lowe's [1982] model for high-density, coarse-grained turbidites. Such turbidites are characteristically highly erosive and surging in nature [Aalto 1976; Lowe 1982], capable of eroding fine-grained muddy slope sediments and incorporating them into the body of the flow both as unconsolidated sediment [Middleton & Neal 1989] and as semi-lithified slope derived clasts [Kano & Takeuchi 1989]. Clasts tend to be concentrated towards the base of high-density turbidites since rip-up clasts never move up beyond the point within the flow that would correspond to the maximum point on the flow velocity profile [Kano & Takeuchi 1989] clasts experience lift up into the flow below this velocity maximum, but are moved downwards once they move above this level. Smaller clasts may become concentrated near the flow base, in a highly concentrated traction carpet [Walker 1975, 1977; Aalto 1976], forming the basal inversely-graded clast division, (Lowe's R2 division), dominated by particle traction and shear. The inverse grading and segregation of finer clasts below coarser, typical of the R2 division, is widely explained by dispersive pressure acting to size-sort grains, or by kinetic sieving during "grain-flow" (i.e. grain to grain collisions [e.g. Middleton & Hampton 1976]).

Deposition of the R2 division involves frictional freezing of the traction carpet, which is driven along by shear [Pickering et al. 1986], followed by direct grain by grain suspension fall-out of coarser suspended clast material, forming the overlying normally
Figure 3-15

A) An ideal sequence of divisions deposited by a single high-density turbidity current declining through discrete gravelly and sedimentation waves (from Lowe [1982]).

B) Complex sedimentation unit deposited by surging sandy high density turbidity current (from Lowe [1982]).
Plate 3-VIII

C) $S_2$ traction layers of upper sub-unit 2. Note tabular aspect of clasts and stratified over-sized peloids. (Lens cap 5cm)

B) $S_2$ traction layer of central sub-unit 2, note stratified clasts concentrated over a thin traction horizon. (Lens cap 5cm)

A) Base of sub-unit 1. The inversely graded basal $R_2$ clast division is overlain by normally graded $R_3$ clast division. Note concentration of over-sized peloids in this basal part of the flow, and stratified peloid $S_2$ horizon beneath hammer.
graded clast division, R₃ [Lowe 1982].

Overlying the R₃ division of normally graded clasts is a coarse peloidal-oolitic division, showing multiple parallel lamination of concentrated, oversized, peloids (Plate 3-VIII, A). This is characteristic of deposition from a slightly unsteady, but fully turbulent, sandy high-density turbidite current, where flow interaction with the bed produces traction-sedimentation structures, of Lowe's SI division [Lowe 1982; Postma 1986]. This type of traction-sedimentation parallel stratification is characteristic of thick "proximal" sandstones in resedimented siliciclastic sequences [e.g. Mutti & Ricci-Lucchi 1975; Middleton & Hampton 1976; Aalto 1976; Walker 1978; Wueller & James 1989]. Cross-stratification is reported from the bases of some such proximal units [Walker 1978] and can also be characteristic of Lowe's SI division: however, in this case, dune-like features did not develop, probably due to flow unsteadiness [Lowe 1982].

This parallel-stratified peloid SI division is then transitional to a coarse-tail graded oolitic-peloidal, S₃ division. Coarse-tail grading is indicative of rapid, direct suspension from sandy high-density turbidites [Lowe 1982; Middleton & Neal 1989]. It is related to decreasing competency of the high-density flow, reflecting the diminished role of dispersive pressure generated by the grain to grain collisions in the S₁ division and the increased role of fluid turbulence, especially above the direct sedimentation surface, where fluctuations of grain concentration are more likely to occur [Aalto 1976].

Sub-unit 2

This oolitic-peloidal dominated sub-unit is characterized by repetitive Lowe S₂ and S₃ couplets, indicative of surging pulsations of the waning high-density turbidity current (compare Fig. 3-14 with Fig. 3-15B). The successive surges are picked out by thin traction carpet horizons (S₂) overlain by thicker suspension sedimentation horizons (S₃) deposited upon deceleration of the surges. This repetition of S₂-S₃ divisions is common in rapidly decelerating, or surging high-density turbidite flows (Figs. 3-15B) [Lowe 1982]. Instead of showing a continuous decrease in velocity, surging flows show an oscillating decline, where each surge reflects an abrupt velocity increase (represented by S₂ divisions) followed by a gradual deceleration (represented by S₃ division). The sediment waves in sub-unit 2 characteristically begin deposition by traction carpet sedimentation (S₂ divisions), and are overlain by suspension fall-out, coarse-tail graded
(S₃) divisions. Fluctuations in the rate of suspended load fall-out during these pulsatory waves of sedimentation may result in traction sedimentation at any stage, until the high-density current has deposited the coarser material and has declined to a lower density flow.

The S₂ divisions consist of inversely-graded stratified layers a few centimetres thick, of concentrated oversized peloids, and occasional millimetre-sized sub-rounded to rounded, wackestone and pelagic mudstone chips (Plate 3-VIII, B and C). These inversely-graded layers reflect increased flow unsteadiness. The suspended load becomes progressively concentrated towards the bed, where transport in the bed-load layer is dominated by grain to grain collisions and friction [Bagnold 1956] leading to the formation of a particle layer. This layer is maintained by intergranular dispersive pressure and "fed" by addition of sediment from above. Fall-out from suspension increases clast concentration in the particle layer and results in the formation of a traction carpet, where the grains are fully supported by dispersive pressure [Lowe 1982]. Continued fall-out loads the traction carpet, which eventually freezes. The process can then either repeat itself [Lowe 1982] or the traction carpet deposits are overlain by massive, or graded S₃ divisions (Fig. 3-14). The thickness of S₂ layers is thought to be controlled by the component grainsize diameter which is also considered to control the thickness of the traction carpet [Lowe 1976a], in a similar way that grainsize controls true grain flow [Bagnold 1954; Middleton & Hampton 1976].

Towards the centre of this sub-unit these traction carpet layers are picked out very small wackestone chips (0.5 - 3mm) and form successive S₂ inversely graded, closely spaced layers (Plate 3-VIII, B) which may be separated by S₃ divisions.

The associated S₃ divisions are either massive, or graded. If grading occurs it is coarse-tail peloid grading. The peloids also show a concentration grading in addition to a size coarse-tail grading, being larger and more numerous towards the bases of the S₃ divisions. Hydrodynamically, the transition from the S₂ to the S₃ division represents waning of the surge or pulse which is accompanied by increased rates of suspended load fallout. This allows insufficient time for the development of the bed-load layer and the organisation of a traction carpet, and sedimentation is by rapid, direct suspension [Walker 1978; Middleton & Neal 1989]. The grains accumulate as the rising surface of the static bed coincides with the top of the falling cloud, from a dense cohesionless suspension; a process analogous to the formation of liquefied beds [Middleton & Hampton 1976; Lowe 1982].
Plate 3-IX

A) Sub-unit 3, unit E. Turbidite cap. Note erosive nature of upper parallel laminated division (change of colour to darker grey).

B) Detail of photograph A - top of sub-unit 3. Parallel laminated fine-grained peloidal packstone (T_d) overlain by pelagic mud drape (T_e). (Lens cap 5cm)

C) Detail of lower part of photograph A. Laminations outlined in pencil, in this top S_3 division, show fluid-escape structures disrupting laminations.
The $S_1 - S_3$ sequence for sandy high-density turbidites [Lowe 1982] reflects a decelerating flow evolution, in the same way that the Bouma sequence [Bouma 1962] does for lower density turbidity currents. Bouma divisions reflect the same mechanical evolution with flow deceleration as the $S_1 - S_3$ divisions, but for different grainsize populations; the $T_{b-c}$ divisions are traction dominated as is the $S_1$ division; the $T_d$ division of traction/suspension lamination is similar mechanically to the $S_2$ division, while the $T_e$ division formed solely by massive suspension is analogous to the $S_3$ division.

By the end of deposition of sub-unit 2, virtually all medium and sand-grade sediment had been effectively deposited from the flow. The residual over-riding turbidity current would have been in an unsteady transitional stage between relatively higher and lower density sedimentation states, containing high concentrations of suspended fine-very fine sand grade material, having effectively sorted coarser material and deposited it in sub-units 1 and 2.

**Sub-unit 3**

Sub-unit 3 represents deposition from the residual high density turbidity current, and forms the turbidite cap. The sub-unit commences with two thin distribution-graded sandy turbidite pulses, with very thin basal oolitic divisions grading up to fine-sands (Fig. 3-14). The rest of the sub-unit consists of sixty centimetres of fine-sand grade sediment (187µm-125µm), the basal twenty-five centimetres of which shows fluid-escape features in the form of dish-structures (Plate 3-IX, A), the remainder showing parallel lamination which becomes better developed towards the top of the bed (Plate 3-IX, B).

The first deposits of sub-unit 3 represent the final deposition of the minor amounts of oolitic sand remaining in the residual flow and as such, represent lag deposits from the higher density depositional phase of sub-unit 2. The two distribution graded layers show coarse-grained, very thin oolitic bases rapidly grading up to parallel laminated peloidal packstones. They are interpreted as lags from the turbidite transition from medium-coarse sand-dominated, higher-density flow to fine-sand, lower-density flow conditions, reflecting the increased unsteadiness of the waning turbidity current. They cannot be assigned to a Lowe's division and are more akin to small Bouma $T_{a-b}$ divisions [Bouma 1962] (Fig. 3-13).
The remainder of sub-unit 3 represents the fine-grained, rapidly deposited turbidite cap and can be described as a large $S_3$ division. $S_2$ traction layers do not develop in sub-unit 3, largely because negligible dispersive pressures, developed between grains of fine to very fine sand grade, would have prevented formation of traction carpets [Walker 1978; Lowe 1982]. Rapid sedimentation is indicated by the nature of this portion of sub-unit 3 in which the lower division shows fluid escape structures (Plate 3-IX, C)) and the upper division shows high-velocity plane lamination (Plate 3-IX, B).

Dish-structures are characteristic of consolidated fluidized beds and were first recognized by Stauffer [1967]. 'Dishes' are thin, sub-horizontal, flat to concave-upwards laminations developed in silt and sand units, [Stauffer 1967; Corbett 1972]. They are formed due to pore-fluid expulsion during dewatering and consolidation of rapidly deposited sediments [Lowe & LoPiccolo 1974]. Less permeable horizons within the bed retard the upward moving pore-fluids forcing some of it to migrate horizontally, until upward escape is possible at a "weak" point. As the water migrates, fine-grained particles such as micrite are carried by the moving water and concentrated in pore spaces, resulting in the micritic enriched laminae, which weather back in outcrop (Plate 3-IX, C). Hydrodynamically, units showing fluid-escape structures are characteristic of rapid deposition by dense, cohesionless suspension sedimentation from high-density turbidites [Walker 1978; Pickering et al. 1986; Lowe 1982] and are characteristic features of the lower shear velocities occurring in Lowe's [1982] $S_3$ divisions [Postma 1986] (Fig. 3-15A). Fine grains settling from suspension accumulate directly as the rising surface of the static bed coincides with the falling turbidite cloud [Lowe 1982], this process being almost instantaneous; the resulting deposit is grain-supported and typically lacks traction structures. This is to be expected since at higher suspended load fall-out rates, there is insufficient time for the development of either bed load layer or an organized traction carpet [Walker 1978]. Middleton [1967] showed experimentally that beds deposited rapidly from very highly concentrated flows behaved for short periods after deposition like viscous fluids (i.e. pseudo-plastically). He also demonstrated the churning of sediment by successive waves moving along the interface between the deposited liquefied "quick" bed and the entrained suspended layer moving above the bed, giving rise to rapid sedimentation of relatively loosely packed sediment. Fluid escape structures form from trapped turbidity current fluid being expelled from these higher parts of the bed, formed by the process of rapid, suspension fall-out.
After deposition of the coarse-grained turbidite load in sub-units 1 and 2, a turbulent residual current remained, containing suspended turbidite fines that did not settle with the coarser sediment. These over-riding currents of suspended finer material are typically of lower density than the main body of flow [Middleton & Hampton 1976]. They may range from true low density flows to the high density fine-grained flows [Lowe 1982] developed in this case. These sediments continue to flow by inertia, due to low friction, after the main body of flow has been deposited. Such residual currents are inferred to be able to accelerate downslope as discrete turbidity currents in their own right, a mechanism mechanically analogous to the generation of turbidites by suspended 'clouds' over-riding debris-flows [Hampton 1972; Enos 1977; Krause & Oldershaw 1979]. These fast-flowing lower density 'clouds' or turbidite 'tails', deposit fine sediment, and are capable of reworking, shearing, liquefying or homogenizing the loosely packed underlying sediment [Middleton & Hampton 1976; Lowe 1982] and forming caps of high-velocity plane lamination [Aalto 1976; Hiscott & Middleton 1979; Lowe 1982]. This is the mode of formation envisaged for the plane lamination at the top of sub-unit 3 (Plate 3-IX,B) which is considered to be mechanically analogous to the Bouma Td upper parallel lamination phase, for the high-velocity plane laminations are traction structures, formed by successive fall-out from fast-flowing, residual turbidity currents.

The whole of sub-unit 3, which forms a major part of unit E, therefore represents a major fine-grained S3 division, formed from suspension sedimentation from a relatively lower-density, but still highly concentrated (i.e. a fine-grained, high-density turbidite); it represents deposition of the turbidite entrained layer and tail [Middleton 1966a; Middleton & Hampton 1976].

Sub-unit 3 is abruptly overlain by fine-grained, dark, fissile mudstones, which are inferred as representing a drape of the finest suspended material associated with the flow, characteristic of longer-term suspension sedimentation of the Bouma Te division.

**Correlation of unit E and general discussion**

The detailed log profile of unit E at the locality of Log 1 (Appendix Fig. 2) shows distinctive internal marker horizons, some of which can be traced through other log profiles of unit E undertaken along outcrop strike of the southern section of Puig Cutri (Appendix Plates I & II) (Appendix Fig. 2 and Fig. 3.16) over a distance of 250m.
Figure 3-16

Lateral correlation of the unit E high density turbidite
scaled position of log sections (Appendix fig. 2)

- mudchips (2-3 cm to 1-2 mm)
- peloids (2000 to 500)
- parallel laminated wackestones
The nine profiles (Fig. 3-15) through unit E were constructed independently while profiling the entire vertical succession (and on different days), thus removing subjective error in their profiling. Although thicknesses of individual horizons may vary from profile to profile, the $S_2$ clast and peloid marker horizons are visible within each profile. It should be noted that some log thickness variation may well be due to the varying angles of vertical surface outcrop of unit E along its length. Additionally, all profiles 1-7 were constructed on completely weathered outcrop (Appendix Plate 2), being undertaken by systematic hammering off of samples for inspection, through the length of the profile. Despite this, lateral correlation between profiles is remarkably good, and unit E can be seen to consist of two major surges.

The two major turbidite surges consist of sub-unit 1, then sub-units 2 and 3. These surges followed rapidly one after the other, since no fine-grained material settled out above sub-unit 1 (bar a few centimetres in profile 1) and there is no marked sedimentological break between sub-unit 1 and sub-units 2 and 3. The sub-units therefore seem to mark major pulsationary waves, or surges, of one main gravity flow event, the surges distinctly partitioning the unit into oolite (sub-units 1 and 2) and a finer-grained, peloidal packstone and wackestone turbidite cap (sub-unit 3, Fig. 3-14). Unlike low-density turbidites which show a gradual decrease in velocity, competence and capacity, high-density turbidites show a surging, oscillating decline, each surge characterized by an abrupt velocity increase followed by gradual deceleration. (Lowe 1982; Middleton & Neal 1989)(Fig. 3-15,B).

Unit E is only locally erosive with its contact with the underlying contourites. The contact is generally planar in nature, evidence of erosion being in the form of broad, shallow scours (Plate 3-X, A), observed on the underside of the bed in the vicinity of Log 3 (Appendix Fig. 2), which have an E-W lineation, parallel to dip. Scours of the same form and dimensions have been recorded from high-density siliciclastic turbidites by Kano & Takeuchi [1989]. A tabular, broken, contourite plate, which is undeformed, is observed incorporated into the very base of unit E, indicative of plucking of the underlying substrate by the erosive turbidite (Plate 3-X, B). Down-cutting of over a few centimetres (<4) into the underlying contourite is not observed. The lack of significant basal erosion indicates that unit E had relief above the sea-floor [c.f. Hiscott & James 1984] and was in the form of a turbidite sheet.

On travelling away from source, the turbidite flow develops and divides into three major parts, the head, the body and the tail (Middleton 1966a; Middleton & Hampton
The head is thicker than the rest of the flow and has a characteristic shape and hydraulic behaviour. Coarse sediment is concentrated in the head, as sediment entrained in fluid sweeps forward and upwards through the head and circulates around to the back of the head. Here, eddies separate sediment sizes, coarser sediment falling back into the head, with finer material being entrained in a dilute cloud that trails behind the head. The head is therefore the main region of erosion [Middleton 1966a], being important in the formation of flutes and grooves which are cut by the head, and for erosion of the slope-derived clasts [Kano & Takeuchi 1989]. The body behind the head is of uniform thickness, the tail is the area where the flow thins, becomes more dilute and is dominated by fines. Mixing between the body and the overlying water add to the entrained layer of fines, derived from sorting by eddies in the head. This low-concentration of fine-sediment is dragged along by the current. This layer continues to flow by inertia, due to low friction, after the body has passed by, depositing fine sediment and reworking the upper part of the sediment deposited by the body and tail.

With reference to unit E, sub-unit 1 is characteristic of head and body sediments, sub-unit 2 of the main body of the flow, and sub-unit 3 fluidised deposits being characteristic of the turbidite tail; with the parallel laminated section being characteristic of the deposition from the entrained layer. Unit E therefore preserves an entire depositional sequence from one major sheet-like turbiditic event, that deposited 2m of sediment at least 20km from source.
Plate 3-X

A) Broad, shallow scours on the base of oolite unit E, showing transport direction to the west. These scours resemble those of the same dimensions, recorded from high-density siliciclastic turbidites by Kano & Takeuchi [1989]. (Notebook 20cm long)

B) Base of unit E, showing underlying contourites and a tabular plate of the contourite facies incorporated into the very base of the flow (arrow). (Gillian Abbots, the author's field assistant and mother, for scale)

C) Gently erosive base of unit C (arrow) overlying a thin development of contourite facies (weathering proud, along with the oolite), which overlie a parallel laminated wackestone unit, underlain by fissile shaley pelagic mud. (log 6, Fig.2)

D) Debris flow, showing large, plastically deformed clasts. Note highly weathered nature of the outcrop. Northern southern section Puig Cutri. (Hammer 30cm)

E) Plastically deformed contourite clast in the process of being plucked into the base of unit E. Photograph from detailed log section of unit E. (Lens cap 5cm)
3.3 Outcrop Facies Associations of the Cutri Formation

3.3.1 Facies associations of structural unit 3

3.3.1.1 Puig Cutri, Sierra d'Artá

This classic locality is the only outcrop of the Cutri Formation that was described by Alvaro et al. [1984], Situated between Arta and Capdepera (Fig. 3-4 & Appendix Fig 3) within Sabat's [1986] structural unit 3 (Fig. 3-7), the outcrop of the Cutri Formation forms a N-S trending ridge, 1km long, which terminates to the north at the highest point on the ridge, Puig Cutri itself (Appendix Plates I and II). The outcrop length is fault-constrained by an E-W strike-slip fault to the north and a NE-SW normal fault to the south (Appendix Fig. 3). The ridge is dissected by NE-SW normal faults, which progressively downthrow the section to the south (Fig. 3-17 and Appendix Plate I).

By virtue of these faults, a classic Tethyan break-up sequence is exposed in the north of the section beneath Puig Cutri (Plate 3-XI, A & Fig. 3-10) at the expense of higher levels of the apron sequence; while a complete apron mega-sequence is exposed in the southern section, part of which is shown in Plate 3-XI, B and also Appendix Plates I & II, from the underlying Cuber Formation rhythmites to the overlying Puig de Ses Fites Formation radiolarites (Appendix Fig. 2). For the purposes of this section, the Puig Cutri ridge has been divided into a northern section and a southern section (see Appendix Plate I).

The Puig Cutri ridge displays prominent oolitic units, which weather proud of finer-grained hemipelagic facies to form resistant ledges (Appendix Plate I). The northern section is dominated by a near vertical cliff of massive oolite 30m thick (Plate 5-XI, A), while the southern section consists of several thinner and distinct oolitic units. In the southern section, twenty oolitic units consisting of amalgamated or individual turbidites have been differentiated, and have been assigned the letters A to T from oldest to youngest (Fig. 3-10, Appendix Fig. 2 and Plate 5-XI, B). This southern section forms the reference outcrop for these oolitic units, which are correlatable between the other localities of structural unit 3: Puig de Ses Fites and Puig de Ses Fites II; as well as a locality from structural unit 5, Amoixa (Figs. 3-19; 3-22 & 3-26).
Figure 3-17

Plan view of the fault displacement of the Puig Cutri ridge (modified from Alvara et al. [1984]).
Plate 3-XI

A) Northern section Puig Cutri, viewed from Puig de Ses Fites corner section, view to the SE. This photograph shows positions of the Liassic platform limestones (A); the Lower Ammonitico Rosso and hardground (B); the Cutri Formation rhythmites (C) and the Cutri Formation oolitic apron (D). Note massive cliff of the northern Puig Cutri gulley and its abrupt faulted northern margin.

B) Southern section Puig Cutri, view to the NNE, showing outcrop of oolitic units A to I. Unit E is 2.5m where marked. Note in the bottom right-hand corner of the photograph, a prehistoric stone dwelling of the earliest inhabitants of Mallorca, the Talaiots, c. 3500 B.C.
Southern section - The Cutri Formation mega-sequence

The most southerly outcrop limit of the southern section displays a complete succession through the Cutri Formation sandbody. From base to top, the succession is interpreted as recording a progression from inner apron facies to outer apron and basin plain facies. The facies of these apron belt divisions do not correspond to the facies developed in the apron model of Mullins & Cook (1986)(Fig. 1-8,B) due to the absence of debris flows in the succession. The facies described are interpreted as resulting from deposition within a sandy turbidite depositional apron, dominated by the resedimentation of unconsolidated oolitic sands. A series of six detailed sedimentological profiles were constructed along the ridge throughout the available outcrop exposure (Appendix Fig. 2), the position of these logs being shown on the outcrop in Appendix Plate II. Fine detail internal correlations observed within these units allowed lateral correlation of individual turbiditic events between these profiles, both at this locality and at several other localities (Puig de Ses Fites, Puig de Ses Fites II and Amoixa). Component oolite and peloid grade was noted throughout the length of each unit, grain size being standardized by use of the ATG grain-size scale comparator.

The detailed profile study through the high-density turbidite described in detail in Section 3.2.3 was undertaken at this locality, from the southernmost outcrop of unit E in the vicinity of log 1, Appendix Fig. 2.

Units A to D - Inner apron association

The association of major units A to D in the basal 25-30m of the succession (Appendix Plates I and II; Appendix Fig. 2) accounts for the majority (68%) of the total volume of resedimented oolite within the mega-sequence, and forms 41% of the total vertical succession. The most active phase of oolite resedimentation was therefore rapidly established, most vigorous and most effectively maintained at the onset of oolite resedimentation. This phase of sedimentation is interpreted as representing deposition within the inner apron environment, due to the occurrence of massive and resedimented units, the degree of amalgamation and the paucity of fine-grained hemipelagic facies.

Unit A is the basal unit which overlies the Cuber Formation rhythmites and is the thickest unit developed in the succession. Generally poorly exposed, it is composed
entirely of amalgamated, often coarse-tail graded, pulses of oolitic-peloidal sand; pelitic intervals are never developed within the unit. Occasionally, sporadic mudchips may pick out amalgamation surfaces.

Unit A is interpreted as displaying repeated $S_2$ and $S_3$ divisions of Lowe [1982], traction carpet and suspension fall-out couplets, the stratified peloid horizons (Appendix Fig. 2) representing flow traction and shearing. The massive character of the unit indicates deposition form highly concentrated flows [Postma 1986], the lack of pelitic intervals suggesting rapid deposition of high-density turbidites in an area with strong accretion rates [Spalletti et al. 1989; Wueller & James 1989]. Amalgamated turbidite sequences such as these are often interpreted as forming by aggradation within fan channels [Link & Welton 1982; MacDonald 1986], within braided rather than meandering fan channels [Wueller & James 1989]. Unit A is also analogous to proximal channel, or coarse-grained gulley fill sequences described from clastic aprons [Stow 1985a]. As such, this unit may represent deposition within a proximal (inner) apron channelised area. This unit does not conform exactly to the inner apron facies of Mullins & Cook (Fig. 1-8,B) due to the absence of mud-supported debris flows within the succession.

Units C and D are similar in nature and closely associated. Unit C is locally erosively based (Appendix Fig. 2, log 6) and inversely graded (log 4). Amalgamation surfaces, picked out by mudchip (<2mm) intervals within the amalgamated, high-density flows, may be traced between log profiles. Unit D is simpler than unit C, and it can be laterally correlated to enable its emplacement to be related to a single mass flow event (refer to section 3.3.1.3 & Fig. 3-22).

The unit A-D inner apron facies association is abruptly overlain by an interval of pelagic mud and then a thick hemipelagic unit, that displays the dark, silty laminitte facies of section 3.2.2.3, interpreted as contourites (Plate 5-VII). The occurrence of this facies marks an abrupt transition from an episode of resedimentation to a long episode of apron 'abandonment' or slope-drape (term after Pickering [1982]).

Units E to I - Proximal outer apron association

The development of individual, major high-density oolitic turbidites, along with significant hemipelagic interbeds, is interpreted as representing the proximal outer apron of a resublimed carbonate sand turbidite system. The occurrence of unit E (see
Section 3.2.3) represents the recommencement of sedimentation to the apron, reflecting rejuvenation (term after Spalletti et al. [1989]) of the apron system. The sheet-like, laterally extensive, major high-density turbidite units E, H and I are the depositional product of single, mass flow resedimentation events, encased in hemipelagic facies, which they punctuated as catastrophic events. The flow profiles pick out turbiditic surges within the same depositional event, which are correlatable between other outcrops of the Cutri Formation in structural unit 3.

*Units J to T - Outer apron fringe to basin plain association*

This association, exposed in a small quarry in the vicinity of log 1 (Appendix Fig. 3 and Plate 3-XII, A) is characterized by thin non-cyclic turbidites (5-65cm thick) that infrequently punctuate the muddy contourite background sedimentation (Plate 5-XII, C). The thinness of the turbidites combined with their sporadic occurrence, justifies their interpretation as apron-fringe (term used by the author in an analogous way to its use in the term fan-fringe after Shanmugam & Moiola [1988]), i.e. distal outer apron to basin plain association of Mullins & Cook [1986].

Turbidites form 18% of the exposure at this locality (post unit I to the radiolarite horizon) and although thinner, these turbidites still comprise a major graded oolitic division (i.e. $T_a$ division) which forms the majority of the bed (Plate 3-XII, B). These basal divisions are formed from coarse sand grade 500$\mu$m to occasionally 750$\mu$m oolite and may rarely show coarse-tail grading of peloids. Parallel lamination (of fine sand grade peloidal packstones and wackestones) is the only other internal sedimentary structure observed. In these cases, the oolite distribution grades up into fine sand grade peloidal packstones and wackestones (which may be parallel laminated) and is not abruptly overlain by a distinct and separate cap as in the major turbidite units down-section.

Generally, these turbidites can be related to Bouma sequences (after Bouma [1962] - Fig. 3-13) although in these turbidites $T_c$ divisions are not observed. The normally graded basal division is interpreted as the $T_a$ division of Bouma; strong parallel lamination as a $T_b$ division; weak parallel lamination as a $T_d$ division; and pelagic mud as the $T_e$ division. $T_{abde}$ sequences are observed in turbidites P, Q and T; $T_{abd}$ sequences in L, F, O and R; $T_{ab}$ sequences in K, R, N and M; $T_{abe}$ sequences in S; $T_a$ sequences in K.
Plate 3-XII

A) Compilation of Southern section Puig Cutri quarry section, view to the north, displaying the upper part of the Cutri Formation mega-sequence. Note thin bedded turbidites, of outer apron-basin plain association, and positions of turbidites Q, R, S and T. Unit T is 60cm thick.

B) Unit T, the last oolitic turbidite of the Cutri Formation, showing silicified basal 3cm and the sheet-like non-erosive base of this unit. Top 3cm is pelagic mud. Note dark and bedded nature of underlying hemipelagic unit.

C) Laminated dark siliceous muddy contourites of this locality. (Hammer 30cm) (compare with Plate VII).

C) Planolites-type burrows on the base of a contourite block, outlined in pencil. (Lens cap 5cm)
The Puig de Ses Fites Formation silicified radiolarite horizon (Plate 5-I, C) lies some 5m above turbidite T and marks the termination of the Cutri Formation mega-sequence.

Northern section Puig Cutri

The northern section of Puig Cutri forms the summit of Puig Cutri itself, dominated by a cliff of massive oolite 30m thick (Plate 3-XII, A) described as a channel by Alvaro et al. [1984]. This massive oolite body is fault bounded to the north (Plate 5-XI, A) and south (Appendix Plate I) and is segmented in half by a normal fault (Fig. 3-17) which downthrows the southern half of this outcrop to the south.

This massive unit is erosional into the underlying Cuber Formation (Plate 3-XIII, C) and has steep fault-controlled margins. It is plugged with compacted oolitic packstone; no blocks or boulders are observed and it appears to have been actively infilled with high-density turbidite sands.

The outcrop is characterized by the strong development of regularly spaced, semi-vertical, planar structures (Plate 3-XIII, D) which Alvaro et al. [1984] interpreted as "large scale crude stratification ... the sets being similar to stratification described in the literature by Piper [1970]; Winn & Dott [1977] and Lowe [1982] ... despite the size of the sets being much larger in this case" [Alvaro et al. 1984]. The authors mentioned in the quotation describe small scale traction sedimentation in sand and gravels, which are not at all comparable to these large scale features, which are over 20m long.

No perceptible grain size changes occur across the boundaries of these features and they are inclined (when corrected for tilt) between 80°-88° towards the southwest (233°), appearing to be too steep to be sedimentary structures, and are traceable up into overlying hemipelagic mudstone facies and these structures are interpreted by the author as thrust-related cleavage that parallels the thrust-related faults that preferentially developed within the massive oolitic unit.

Discussion and Interpretation

The Puig Cutri "channel" is interpreted as a U-shaped apron gulley which served to funnel oolitic-material out onto the inner apron. The muddy oolitic sediment was transported as high-density turbidites, the incorporation of slope-derived muds aiding dispersive pressures supporting the grains during transport [Lowe 1986]. The gulley axis
Plate 3-XIII

A) Northern section Puig Cutri, view towards the east. View through Middle Jurassic break-up facies (see Plate 3-XI, A) to the northern section apron gulley on the skyline. (Scale: gulley approximately 25m thick)

B) Amalgamated oolitic packstone turbidites of the gulley-fill association. View to the north-east, down-faulted section of the northern section of Puig Cutri (see Appendix Plate 1 & 2). (Scale: unit approximately 20m thick)

C) Erosional base of northern section gulley (pencil) into the underlying slope facies. Hammer (30cm long) is on the oolite of the gulley-fill.

D) Detail of high-angle planar structures interpreted as cleavage, not large scale stratification. (Figure for scale)
is orientated E-W (Fig. 3-17), cutting downwards in a westerly direction, indicating an eastward source direction.

Similar U-shaped back-filled gullies cutting across the proximal apron have been described by van Buren & Mullins [1983] and Mullins et al. [1984]. Ancient gullies with similar dimensions have been described in the literature. Gullies cut into slope mudstone and similarly plugged with sediment are described from the Devonian Prongs Creek Formation, Yukon, where they are 50-100m deep and 100-200m wide [Cook & Mullins 1983; Mullins & Cook 1986]; others of smaller dimensions 1-10m deep and 20-500m wide, have been described in other ancient apron sequences by Cook [1979; Cook et al. [1972]; Enos [1977] and Enos & Moore [1983]. U-shaped gullies are also common on clastic aprons [e.g. Field & Clarke 1979; Wezel et al. 1981] and may cut across the proximal apron [Stow 1985a].

The position of the gulley, now delineated by NE-SW faults, may have been controlled by contemporary faults, as they often are on modern faulted-slope aprons [Stow 1985a and 1987; Wezel et al. 1981] and ancient apron systems [e.g. Gawthorpe 1986a; Stow 1985b]. It is possible that the faults that mark the present gully boundaries are reworked Jurassic faults that controlled basin floor topography. A major boulder carrying debris flow (Section 3.2.2.1) terminates the gulley fill association, which is then overlain by a thick hemipelagic unit. An identical sequence of events (but on a larger scale) has been documented by Hurst & Surlyk [1982] from the Silurian of North Greenland, where debris flow emplacement was directly related to a tectonic subsidence and tilting event, which restricted carbonate production and led to the deposition of hemipelagic limestones.

3.3.1.2 Puig de Ses Fites, Sierra d'Artá

This locality marks a SW-NE 700m extension of the Puig Cutri ridge (Appendix Fig 3) and is shown in Plate 3-XIV. It is separated from the Puig Cutri ridge by a WNW-ESE trending strike-slip fault and by a similar bounding fault to the NE, these faults allowing movement of the Puig de Ses Fites block further to the NW (Appendix Fig. 1). Normal faults serve to downthrow the section progressively to the NE, hence more complete apron sequences are progressively displayed in this direction.

The outcrop is steep and difficult to work. No detailed oolite profiles were taken at this locality, but a measured and correlated section is shown in Fig. 3-18, which
Figure 3-18

Lateral profiles and unit correlation along the Puig de Ses Fites ridge. Log sections correspond with arrows on Plate 3-XIV.
Plate 3-XIV

Puig de Ses Fites ridge, view to the SE. Arrows correspond with those on Fig. 3.18 and note the position of log sections of this diagram. Puig de Ses Fites itself, to the far left (NE) of the photograph is 350m, the corner section on the right (SW) 363m.
Figure 3-19

Correlation of Puig Cutri with Puig de Ses Fites
corresponds with Plate 3-XIV. Correlation between this locality and Puig Cutri is shown in Fig. 3-19.

Southwestern section - Posidonia coquina facies

In the vicinity of profile I (Fig. 3-18) on the south-west edge of the outcrop (Plate XV, A) the top of the Cuber Formation rhythmtes are characterized by the development of a pale pink-grey nodular limestone horizon. The development of similar facies is noted from the top of the Cuber Formation at Amoixa, Amoixa II and from the northern section of Puig Cutri. Above the nodular horizon lies a 20m thick succession of fine-grained, thin bedded wackestone and mudstone facies, the basal part of which is characterized by the development of Posidonia-rich facies.

A detailed log was constructed through this succession of Posidonia-rich sediment (Fig. 3-20) which has been assigned suffixes which indicate increased 'contamination' of Posidonia filament accumulations with fine sediments. These Posidonia accumulations are interpreted as coquinas (Section 3.2.2.3) and from purest to most 'contaminated' the sediments form the following facies associations: coquina grainstones (CG) (Plate 3-XV, E & F) and coquina packstones (CP) are sediments where the bivalve filaments form the majority of the sediment, the grainstones being the purest end member form of the association, the packstones containing minor amounts of fine carbonate sediment. Dilute coquina wackestones (DCW) are finely laminated micritic to fine peloidal wackestones containing large numbers of Posidonia in layers. The dilute turbidites (DT) are fine sand grade peloidal packstones which are graded and may contain chaotically oriented Posidonia in their basal divisions (Fig. 3-20). Turbidites (T) represent sporadic distribution graded units which grade from thin oolitic bases and may contain wackestone chips.

In outcrop, the very Posidonia-rich (CG) facies occur as dark, dense, almost black limestones with a characteristic crinkly appearance due to the dense packing of the Posidonia valves. In thin section (Plate 3-XV, E and F) this facies is seen to consists entirely of long, curved, closely packed filaments, arranged concave upwards. Patches of short, broken filaments, which disrupt the laminar packing, are limited to discrete areas of the slide, and were possibly caused by burrowing organisms. Where filament packing is very dense, small amounts of micrite are trapped between individual filaments. The curved nature of the filaments may also create shelter and umbrella "porosity". An early fringe cement nucleates on the Posidonia valves (Plate 3-XV, F)
Figure 3-20

Sedimentological log of the *Posidonia*-rich facies of the south-west section of Puig Cutri.
Plate 3-XV

- photo D
- photo E & F
- photo C

DCW
dilute coquina wackestone
CG
coquina grainstone
CP
coquina packstone
W
weckestone
T
turbidite
NLST
nodular limestone
F
fissile texture
ooids
weckestone chips
chert
ammonites
belemnites
posidonia filaments
which prevents compaction prior to the occurrence of the second generation blocky cement. Micro-stylolitic solution seams are also present.

Discussion

These coquina levels represent accumulations of a pelagic calcitic organism, and in the case of the coquina grainstones, with a negligible amount of associated sediment. Their bedding parallel orientation and dense packing of the filaments, combined with their well-preserved long and unbroken state of preservation, implies a gentle sorting and winnowing by currents [e.g. Kuhry et al. 1976; Steiger & Jansa 1984]. The presence of this facies represents a relative hiatus in sedimentation at this locality, implying that this region of the outcrop was a gentle, current swept area, elevated from neighbouring areas of oolite sedimentation. The presence of sporadic, thin oolitic turbidites within this unit (Fig. 3-20 and Plate 3-XV, C) indicates concomitant oolite resedimentation. The contaminant sediment in the coquina packstones and wackestones was likely to have been derived from low density muddy turbidity currents, which were subsequently rewaned.

Mullins et al. [1979] mention current winnowed sands and sandy turbidites interbedded with nodular horizons and peri-platform ooze from depths of 330-800m along windward margins in the Bahamas. Cook & Taylor [1977] describe an ancient example of a current winnowed bioclastic coquina, interpreted as forming due to winnowing by oceanic contour currents. This assemblage therefore provides further evidence of the activity of basin currents, complimenting that of the contourite facies. Winnowing could also occur in response to bottom currents induced by by-passing turbidites, which could have been associated with and concentrated in the nearby Puig Cutri gulley.

Debris Flow Puig de Ses Fites

A debris flow outcrops on top of the first oolitic unit in the SW end of this outcrop, where it is exposed in plan view. This debris flow is the lateral equivalent of the major boulder-carrying debris flow capping the gulley fill association at Puig Cutri, which is also recorded from Puig de Ses Fites II (see the following section; & Plate 3-XVII,B). The debris flow at this locality is of similar thickness, varying between 70cm-1m, as in these other two localities. The largest boulders occur in this locality, the largest observed being a 1m x 45cm clast of pelagic mud. Some of these boulders
A) View of SW corner section, Puig de Ses Fites, view to NW from Puig Cutri. Note the three prominent oolitic units, of SW corner (profile 1) Fig. 3-18. Prominent thinly bedded units are the *Posidonia*-rich facies of Fig. 3-20. (Scale: lower-most oolitic unit is 7m thick)

B) Siliceous bedded wackestones from beneath the basal oolite, corner section, Puig de Ses Fites. (Hammer 30cm)

C) Oolitic turbidite influxing the *Posidonia*-rich facies (above and below the turbidite). Note shaley facies at base of the photograph.

D) Load casts in fine-grained, laminated interbed facies, NE Puig de Ses Fites section. (Hammer 30cm)

E) Thin section microphotograph of the *Posidonia* coquina grainstone facies. (Field of view 2.5mm)

F) Enlargement of E (field of view 1mm). Note early fringe cement nucleating on the *Posidonia* filaments (arrow).
have a preferred orientation in a NW-SE direction.

Although the outcrop is highly weathered, occasionally interesting features are observed in outcrop. One large wackestone boulder displayed, revealed the presence of *Planolites* type burrows, along with a very small fragment of a brachiopod. Very occasionally, pelagic mudstone clasts within the debris flow are "frozen" in the process of being disaggregated into smaller clasts (Plate 3-XV, D). A huge rafted block of siliceous thin-bedded wackestone is also observed within this debris flow, which is tabular and over 2m long.

**Evidence for Syn-sedimentary Disturbance**

A feature of the basal oolitic unit of the Puig de Ses Fites southern section (Plate 3-XV, A) is the presence of the syn-sedimentary structure shown in Plate 3-XVI, C at the locality of profile 2 Fig. 3-28, & Plate 3-XIII. At this particular locality, the debris flow is locally missing. Siliceous wackestones overlying the oolitic unit appear to have sunk down into a crevice formed within the top of the oolitic unit. This appears to have occurred when the wackestones were semi-lithified, and were able to deform plastically as they compacted down into the crevice. Note the chert picking out the original laminations in the wackestone. This feature may be evidence for syn-sedimentary disturbance, which caused the semi-consolidated underlying oolitic unit to gape and the overlying semi-consolidated wackestones to collapse plastically into the void. It is feasible that such features could form by tectonic shocks such as earthquakes, acting on semi-consolidated sediments. To the author's knowledge, similar features have not been described in the literature. This formation of feature may be related to the emplacement of the laterally overlying debris flow.

**Channel Margin, South-West Corner Section**

An oolitic channel margin is observed at this locality (Plate 3-XV, A and B), which trends due E-W. The oolite occurs in abrupt transition laterally with thick-bedded *Posidonia* wackestones through which the channel margin cuts. Down-section the same stratigraphic horizon is represented by siliceous wackestone interbeds. This small channel is representative of an inner-apron channel association and is plugged with oolitic-peloidal packstone. The channel margin is steep, not grading out into coeval wackestone interbeds; it was therefore an erosional cut and fill feature. This same unit
Plate 3-XVI

A) Channel margin of unit 2, corner section Puig de Ses Fites. (*Chamaerops humilis* (*"Palm trees"*) for scale - 1m)

B) Close up of A showing oolitic channel cutting down through bedded wackestone interbeds. (Hammer 30cm)

C) Siliceous wackestone interbeds collapsing into a gape at the basal oolitic unit, southern section (corner section) Puig de Ses Fites.

D) Disintegration of semi-lithified pelagic mudstone clast within the debris flow, Puig de Ses Fites. (Hammer 30cm)
is traceable along strike for the length of this outcrop and has lateral stratigraphic
equivalents at Puig Cutri and Puig de Ses Fites II.

3.3.1.3 Puig de Ses Fites II, Sierra d'Artá

This section is exposed on the south-east flank of the Puig de Ses Fites range, between
the peaks of Puig Cutri and Puig de Ses Fites (Appendix Fig. 3). The outcrop is
exposed in a NW-SE ridge (Plate XVII, A), exposed within a minor internal thrust slice
within the Puig de Ses Fites structural complex of Sabat's [1986] structural unit 3.
Palinspastically restored, it represents a slightly more easterly slice through the apron,
preserving a more complete section through the apron than its neighbouring localities.
This locality was mapped as Malm & Cretaceous by Sabat [1986] which is an obvious
error.

The section displays fine sedimentological trends consistent with being a more
proximal outcrop than those of Puig Cutri and Puig de Ses Fites. Importantly, the
sedimentological relationships obtained by construction of the log profiles allow
documentation of the evolution of boulder-pebble debris flows at this locality that
probably originated upslope as mass failure slides, into high-density turbidites bearing
millimetre sized chips, at Puig Cutri. In this locality, oolitic units form 55% of the
total formation succession, slightly higher than that of 51% for Puig Cutri; and clast-
rich horizons, many of which are boulder-pebble conglomerates, account for 20% of
the resedimented oolitic units, compared to a maximum of 5% at Puig Cutri. Both
hemipelagic and oolitic units correlate consistently well between this locality and Puig
Cutri (Fig. 3-22) apart from the basal unit A.

A feature of this outcrop is that thick boulder-pebble debris flow horizons within
amalgamated units correlate with sporadic mudchip horizons at Puig Cutri (Fig. 3-22).
Units A to D are more amalgamated in this locality, and contain several boulder-pebble
conglomerate grade debris flows, which contain clasts up to 40 x 25cm (averaging 0.5-
3cm) (Plate 3-XVII, B), which are correlatable with oversized peloid-rich and sporadic
small mudchip (1-4mm) horizons in the units at Puig Cutri (Fig. 3-22, unit C). Unit D
at this locality is capped by a debris flow that contains clasts up to 40 x 20cm, along
with massive pelagic mudstone boulders 1m x 45cm (Plate 3-XVII, D). This unit
correlates with the major boulder-conglomerate debris flow on Puig de Ses Fites, and
that terminating the gulley fill of the northern section of Puig Cutri. This major
debris flow is marked by a high density turbidite containing stratified mudchip
horizons in the southern section of Puig Cutri Unit 4 (Fig. 3-22). Unit E is of similar
Figure 3-21

Lateral section of the Puig de Ses Fites II ridge.
Figure 3-22

A detailed log section from Puig de Ses Fites II, and its stratigraphic correlation with the southern section of Puig Cutri.
nature and thickness, while unit H is thicker at Puig Cutri.

At this locality, unit I consists of a basal high-density turbidite overlain by a boulder-pebble conglomerate debris flow. At Puig Cutri, these sediment surges are picked up by the presence of tiny 1–2mm wackestones and mudstone chips within the basal 50cm of unit I, the remaining sediment representing deposition from surging high-density turbidite pulses that evolved from the debris flow and its over-riding turbidite cloud (Fig. 3–22).

Some of the thin-bedded post unit I turbidite equivalents are 'absent' at this locality, this may simply be due to the nature of the outcrop was very overgrown at this level within the apron.
Plate 3-XVII

A) View of Puig de Ses Fites II locality, showing prominent oolitic units and typical local vegetal cover. (*Chamaerops humilis* (*Palm trees*) for scale - 1m)

B) Debris flow horizon of unit C, Puig de Ses Fites II, showing large parallel laminated tabular intraformational wackestone clast. (Lens cap 5cm)

C) Fine grained parallel laminted wackestone cap of unit E, Puig de Ses Fites II. (Hammer head 12cm)

D) Massive boulder of fine wackestone (1m long) within debris flow at top of unit D, Puig de Ses Fites II. (Hammer 30cm)
3.3.2 Facies associations of structural unit 2

The Middle Jurassic succession of the northern Sierra d'Artá, north of the Puig de Ses Fites area, is characterized by the absence of Cutri Formation resedimented oolite and the occurrence of coeval condensed modular limestone sequences. The development of condensed facies is confined entirely within the northern-most outcrop of Sabat's [1986] structural unit 2 (Fig. 3-7). Red ammonitico rosso outcrops at Sa Caleta, with grey nodular limestone containing Posidonia and chert at S'Atalya Freya [Bourrouilh 1973] and Ermita de Betlem [Alvaro et al. 1984] (Fig. 3-23).

The occurrence of the Cuber Formation and Cutri Formation in this region is confined within unit 3, being recorded from localities well within the limits of the unit and directly on its margin with unit 2 (Fig. 3-2). Unit 2 outcrops further south in the Sierra de Levante, north of San Lorenzo de Descardazar (San Liorenc) where the Cuber and Cutri Formations are present: the outcrop at Puig de Ses Esquerdes (Fig. 3-4) being dominated by an extensive basal unit, similar to that developed in the Puig Cutri-Puig de Ses Fites area of unit 3.

The presence of the northern unit 2 condensed sequences indicates that this area acted as a structural "high" (like the Sierra Norte) during the deposition of the Cutri Formation, a few kilometres to the north-west of the Puig de Ses Fites area.
Figure 3-23

The occurrence of coeval condensed ammonitico rosso facies in structural unit 2 with resedimented oolite in structural unit 3, northern Sierra d'Arta.
3.3.3 Facies associations of structural unit 1

Puig d'en Paré, Sierra d'Artá

This locality, in the western Sierra d'Artá (Fig. 3-4) is exposed on the western face of Puig d'en Paré (Plate 3-XVIII, A) which occurs in Sabat's [1986] structural unit I (see inset at Fig. 3-24). Palinspastically restored, this locality would be in the order of 15km due west from the Puig Cutri area of structural unit 3 (Appendix Fig. 1).

This section, the lower part of which is shown in Fig. 3-24, is interpreted as representing deposition on an outer apron setting (after Mullins & Cook [1986]). The succession displays several features which indicate a more distal setting (including decreased turbidite thickness and degree of amalgamation) to that of the outcrops in structural unit 3:

1. Debris flows are absent from the succession, as are intraformational rip-up clasts within the oolitic turbidites.

2. Massive amalgamated units such as those of the basal unit A-D association at Puig Cutri are never observed; however, amalgamation still occurs on important feature.

3. Both the turbidite units and the hemipelagic units are of reduced vertical extent and alternations between the two are more frequent.

4. Oolitic turbidites still form an important part of the sequence (35%) although their total percentage is reduced.

5. The turbidites are thinner on average; however, thick (>50cm) high-density turbidites occur, but are characterized by a marked reduction and frequency of occurrence of very coarse sand-granule grade peloids in the 1500-2500µm range. Amalgamation of the high-density turbidites is common (beds 8, 11, 15, 17, 18 of Fig. 3-24).

6. Thin oolitic turbidites still contain a significant Ta division, which forms most of the bed, and may also contain oversized peloids in the basal divisions, which never exceed the constituent oolite grade by more than one size division.
Figure 3-24

Detailed sedimentological log profile of the basal section at Puig d'en Pare. Inset is structural map of the mountain, after Sabat [1986]. 15 = Platform dolomites; 12 = Dogger marls; 11 = resedimented oolite; 9 = Malm/Cretaceous.
Outer apron facies tend to be more discontinuous than outer fan lobe sequences [Mullins et al. 1984; Mullins & Cook 1986] and turbidite sheets are seen to vary somewhat in thickness and over short distances (Fig. 3-24). Some of the amalgamated units (i.e. 5, 8 and 11 & Fig. 3-24) are associated with slumped or contorted thin-bedded strata (Plate 3-XVIII, B) and may pinch out laterally (Plate 3-XVIII, D and Fig. 3-24, 5 and 8). These units appear to have infilled hollows or "ponds" created by slumping of the underlying hemipelagic facies and the edge of unit 8 (Plate 3-XVIII, C) is defined by a small syn-sedimentary fault. Linear scours on the base of some of these flows indicate transport to the WSW. Nelson et al. [1985] describe (although on a much larger scale) slumping associated with high sand content "channelised lobes", which pond into the slump depressions.

The presence of small "channel" features are rare on outer fan depositional environments [Walker & Mutti 1973; Mutti & Normark 1987] however carbonate base-of-slope aprons may be characterized by small, shallow "gullies" that act as conduits to transport coarse sediment out onto the basin plain, actively infilling with the sediment they transport [Cook et al. 1972; Cook 1983]. Similar outer apron "gullies" may also form on clastic aprons [Stow 1985a; Stow 1987]. Similar analogies, erosively based "fingers" of coarse sediment extending out onto the basin floor, are documented from the Bahamas by Hine & Mullins [1983] and Hoskin et al. [1986]. In this locality, these units plug shallow depressions created by gentle slumping, which indicates that tectonic events may have preceded oolitic turbidite emplacement. This direct association with disturbed strata is not seen in localities elsewhere.

In carbonate base-of-slope aprons, sand sized packstone and wackestone turbidites displaying $T_{a-b}$ divisions are characteristic facies of outer apron settings [Mullins & Cook 1986], as a result of coarse sediment by-passing the upper slope [Schlager & Chermak 1979]. For this reason, the progressive loss of the $T_a$ division seen in siliciclastic fan systems with distality [Mutti 1977; Mutti & Ricci-Lucchi 1975; MacDonald 1986] is not as marked in these apron systems. In this case, the major high-density oolitic turbidites evolve into thinner high-density turbidites, which in turn evolved into $T_{a(b)}$ turbidites. This is typical of other carbonate turbidites described in the literature, which, although tens of kilometres from source, still display $T_a$ type distinct grading [e.g. Connolly & Ewing 1967; Davies 1968; Bennets & Pilkey 1976] (see also Section 4.3.2.3.).
Cores taken from the outer aprons of Northwest Providence Channel consist of 35% thin turbidites and debris flows, interbedded with peri-platform ooze [Mullins & Neumann 1979], showing an identical percentage of resedimented facies to this section of the Cutri Formation.
Plate 3-XVIII

A) Puig d'en Pare overview, view to the NNE, showing position of log (arrow).

B) Nodular chert base of unit II, underlying wackestone interbed facies, and disturbed strata (right of hammer)

C) Margin of unit 8, note abrupt and steep contact of oolite (left hand side) with disturbed wackestone strata. (Hammer 30cm)

D) Lateral pinch out of oolite unit 9 into thin-bedded wackestone facies to the NW. (Hammer 30cm)
3.3.4 Facies associations of structural unit 5

3.3.4.1 Amoixa, Son Massia, Sierra de Son Amoixa

This locality is situated in the south-eastern Sierra de Levante, in the Sierra de Son Amoixa (Fig. 3-4). The outcrop is exposed as a NNW-SSE running ridge within a NE facing thrust sheet of Sabat's [1986] structural unit 5 (Appendix Fig. 4 & Fig. 3-7). Palinspastic restoration places this locality some 65km SE of the Puig Cutri region of structural unit 3. This locality displays the most proximal inner apron facies of the Cutri Formation. Only the basal 35km of the mega-cycle is exposed, most of which occurs as a near vertical escarpment (Plate 3-XIX). With some climbing expertise, a detailed log profile was constructed (Fig. 3-25). Characteristic sedimentological features of the outcrop at this locality are:

(1) Carbonate conglomerates, interpreted as debris flows from an important facies type of the succession, forming 20% of the exposed succession. In no other locality are debris flows so prevalent or distinctive, and all the debris flow types described in Section 3.2.2.1 (bar the boulder conglomerate) occur at this locality and are shown in Plate 3-III of the aforementioned section.

(2) The succession is dominated by major oolitic turbidites, which form 56% of the total logged succession. Many of these are major turbidites of thicknesses between 1-3m, which comprise 89% of the total turbidite volume of the succession (i.e. 49% of the total succession). Some of these major turbidites are oolitic grainstones rather than the packstones seen elsewhere.

(3) Oolitic unit interbeds in the lower 27m of the succession are characterized by thin $T_{ab}$ type, distribution graded, oolitic to parallel laminated peloidal packstones, rather than the hemipelagic facies seen elsewhere. Contourite wackestones and mudstone facies occur in thick units towards the top of the outcrop (Fig. 3-25).

(4) The resedimented facies from this locality are light yellow in colour (Plate 3-III), while the facies from the localities further west in structural units 3 and 1 are dark grey-brown. This light colour is also a feature of the unit 5 successions from the Cuevas d'Artà region (Section 3.3.4.2 & 3.3.4.3). Gutschick & Sandberg [1983] show a correlation between limestone colour and palaeobathymetry: darker rocks indicating deep basinal environments, while lithologies deposited at shallower depths are lighter
Plate 3-XIX

View of the Amoixa section, to the SW, ridge strikes NNW-SSE, showing position of sedimentological log. Enlargement of this area shown in insert. Note lateral continuity of beds.
coloured. This is an indication that this section was deposited at shallower palaeodeeps than those further west, perhaps on a structurally higher intra-basinal block.

There is an abrupt transition from Cuber Formation hemipelagic rhythmites into distinctive condensed facies, which precede emplacement of the basal major debris flow (Fig. 3-25). These condensed facies reflect a precursor episode of condensed sedimentation, prior to the onset of oolite re sedimentation. This facies development is characterized by very fine grained (95-125μm), light coloured, Posidonia-rich laminated wackestones and mudstones; interbedded with pink to white nodular marl horizons (Plate 3-III, B) which contain chaotically oriented Posidonia filaments. The randomising of the Posidonia is interpreted as being due to the early diagenetic process of nodule formation [Jenkyns 1974; Mullins et al. 1980a]. Towards the base of the debris flow, occasional very thin (<4 cm) oolitic grainstone intervals are found that are rich in Posidonia. This is the only occurrence of Posidonia-rich oolitic grainstones within the entire apron sequence, and they are interpreted as current winnowed lags.

This facies group is abruptly overlain by the basal debris flow (Fig. 3-25) which marks commencement of apron sedimentation. Abrupt facies changes, such as those observed between the Cuber Formation, the condensed facies and the basal debris flow, have been interpreted on other Tethyan margin sequences as reflecting rapid fluctuations of the platform boundary due to tectonics [Bernoulli & Kalin 1984; Steiger & Jansa 1984] and indicate an unstable slope environment. Similar episodes of condensed sedimentation have been documented by Gawthorpe [1987a] to have formed due to a combination of sea-level rise combined with hangingwall subsidence and footwall uplift, producing marked topographic variations on the basin floor.

The basal major debris flow, the belemnite-rich massive conglomerate of Section 3.2.2.1 (Plate 3-III, A and C) is unique in character to this outcrop. The large numbers of belemnites were likely to have been incorporated as the debris flow moved over an up-slope region where reduced sedimentation rates led to a relative concentration of un cemented belemnite guards, in the form of a belemnite death assemblage "coquina". In situ condensed cephalopod coquinas are (as far as the author is aware) absent from the succession in the Sierra de Levante. They have been reported from other Tethyan margins where they are known as cephalopod limestones, for example in the Calcareous Alps [Zankl 1971]; Morocco [Wendt et al. 1984] and Oman [Blendinger 1988], and they are considered to form by current winnowing on shallow pelagic highs.
Figure 3-25

Sedimentological log section of the succession at Amoixa.
The presence of carbonate breccias, conglomerates and debris flows, all variously named in the literature, are major mechanisms of resedimentation from ancient slope sequences [e.g. Krause & Oldershaw 1979; Cook & Taylor 1977; Cook et al. 1972; Cook 1979; McIlreath & James 1978; Nardin et al. 1979]. Their initiation is often related to catastrophic events causing mass flow wasting, such as tectonic activity [Cook et al. 1972; Shanmugam & Benedict 1978; Colacicchi & Baldanza 1986; Eberli 1987], or rapid rise [e.g. Mullins et al. 1986] or fall [e.g. Cook et al. 1987] of sea-level. The presence of debris flows has frequently been directly associated with a base of fault scarp apron setting [e.g. Bernoulli & Renz 1970; Hurst & Surlyk 1982; Pickering 1984; Colacicchi & Baldanza 1986; Eberli 1987 & 1988]. In these cases, the debris flow clasts may contain stratal blocks derived from the effacement of fault-scarps during downfaulting. No such extraformational lithoclasts are observed within the Cutri Formation at this, or any other locality, neither are blocks or clasts of oolitic grainstone or packstone, or other clasts of direct platform affinities. All the clasts are wackestones and mudstones, containing abundant Posidonia and radiolaria (Plate 3-11, D) or pelagic mudstones, and are interpreted as being derived from semi-lithified coeval facies of shelf-slope break and slope. The emplacement of similar fine-grained debris flows (those with clasts <64mm) has been related to tectonic pulses associated with basin subsidence. Pulses of tectonic subsidence during the pre-drift extension of the Sierra de Levante basin are likely to have been important mechanisms of gravity flow initiation.

Correlation

On sedimentological grounds alone, turbidite and debris flow units of this section (which lay some 65km along strike from the Puig Cutri area, and some 10km further east (sourcewards) (using the map of Sabat [1986], Appendix Fig. 1)), are correlated with the sections of structural unit 3 (Fig. 3-26). From the nature of the condensed facies underlying the basal debris flow, sedimentation probably commenced later at this locality than in the sections unit 3. Contourite units are directly correlatable, as are the post unit D oolitic units (Fig. 3-26). Debris flow horizons at this locality correlate with debris flows of different nature at Puig de Ses Fites II (compare with Fig. 3-22) and thin mudchip (1-4mm) horizons at Puig Cutri (Fig. 3-26).
To show the proposed sedimentological correlation of Amoixa with Puig Cutri.
Amoixa II

This locality is situated 2½km NW of Amoixa, within the NE facing ramp of Sabat's structural unit 5 (Appendix Fig. 4). It is separated from the main Amoixa locality by a sinistral NE-SW tear fault [Fornos et al. 1984; Sabat 1986]. The outcrop consists of massive, dark oolitic units, up to 15m thick, separated by very fine-grained, mud-rich, dark hemipelagic units, consisting of fine-sand grade wackestones and pelagic mud. The units may thin laterally in a NW-SE direction, appearing somewhat lenticular and may pinch out laterally in this direction.

These massive oolitic units form vertical cliffs and are interpreted as base-of-slope plugged gulley-fills. The sediments are all oolitic-peloidal packstones and oolitic wackestones, sediments being much more mud-rich at this locality. The thickness and mud-rich nature of these units indicate their very proximal position on the apron [c.f. Mullins & Cook 1986]. The occurrence of this outcrop, so close to that of Amoixa, indicates that an intrabasinal fault must have separated the two areas during the Bathonian, this outcrop being deposited at a deeper palaeodepth (due to its dark colouration) similar to that of the outcrops in structural unit 3.

This massive gulley succession is separated by a NE-SW normal fault from a partial succession shown in Fig. 3-27. (see Appendix Fig. 5) to the south-east. The basal oolitic divisions in this outcrop are very thinly bedded and underlain by nodular facies. This sequence may indicate that the area of the outcrop shown in Fig. 3-27 represents a palaeotopographic high with respect to the immediately adjacent succession, north-west of the fault.
Figure 3-27

Post-rift sequence and partial outcrop of the Cutri Formation south-east of the NE-SW fault, Amoixa II (see Appendix Fig. 4).
Cutri Fm.

- Tectonised nodular limestone
- Thin bedded Posidonia wackestone
- Cuber Fm.

60 m

- Crinoidal grainstone
- Grey marl

6 m

Hardground
Quartz arenite

Liassic platform
3.3.4.2 Cuevas d'Artá, eastern Sierra d'Artá

This area displays two post-rift sections, the Cuevas d'Artá road section and the Sa Talaia section (Plate 3-XX & inset on Fig. 3-28), the latter lying 0.5km to the NW of the former. The Cuevas road section exposes a massive dolomitized oolitic unit [Fornos et al. 1987] which is likely to be a fragment of the Middle Jurassic source platform of the Cutri Formation oolite. The Sa Talaia section, here described for the first time, exposes two thin resedimented units within a bedded and nodular wackestone facies. These resedimented units are considered to be Cutri Formation equivalents, this section being interpreted as the Bathonian upper slope of the Sierra de Levante basin.

The outcrops are displayed within the uppermost thrust sheet of the Sierra d'Artá, unit 5 of Sabat (1986). Palinspastically restored with respect to other localities, this area lay some 21km SE from the Puig Cutri-Puig de Ses Fites area, 35km ESE from Puig d'en Paré, and approximately 75km NNW of the Amoixa area, which lies within the same structural unit.

The Cuevas d'Artá road section, briefly described by Fornos et al. [1987] (Fig. 3-28) and subsequently briefly visited by the author late in 1988, remains a section that requires further investigation, time unfortunately not permitting a detailed investigation during the duration of this project.

The section outcrops along the road commences with well-bedded, undolomitized, Lower Jurassic platform limestones, which are bioclastic and fenestral in character towards the top of the sequence. Rubbly horizons within the sequence may be incipient palaeokarsts indicative of platform exposure prior to subsidence. The contact with the overlying Aalenian ammonitico rosso facies is sometimes characterized by the presence of scattered ferro-manganese nodule encrustations (Plate 3-II, B) indicating incipient hardground formation. The ammonitico rosso (Inferior Ammonitico Rosso) is redder than facies seen elsewhere, and appears somewhat tectonised. It is slightly thicker by about 1 metre than coeval facies further west. The ammonitico rosso is abruptly overlain by 55m of thinly bedded, light grey marly wackestones, interbedded with carbonate mudstones. The wackestones contain Posidonia and occasionally display a semi-nodular fabric.
Plate 3-XX

An overview of the Cuevas d’Arta locality, view towards the NE. Section shows Liassic platform (Puig Negre) in contact with Dogger upper slope facies (Sa Talaia). Note position of interpreted structural contact which separates platform oolite of the Cuevas d’Art from slope facies.
Within this facies, 10m from the top of the platform, is an oolitic packstone bed 80cm thick (Plate 3-XXI, B). This bed has a planar base and top and appears fairly massive, distinct grading not being evident in the field. It is directly overlain by peloidal wackestone containing Posidonia. The following up-section facies do not show any further oolitic influxes, and are locally cherty. The author suggests that the oolitic packstone bed is resedimented and is the thinned lateral equivalent of a resedimented oolitic unit from Sa Talaia (see following Section 3.3.4.3).

The following 145m of succession (Fig. 3-28), which contains the Artá caves (Plate XX), consists of dolomitized limestones, most of which are oolitic. Dolomitization affects 65% of the sequence [Fornos et al. 1987], the oolite is well bedded in layers 20-50cm thick, occasionally reaching 1.5m thick. Cyclic trends are not recognized and chert is absent. Fornos et al. [1987] have defined three lithofacies but have not differentiated packstones from grainstones in their use of the term grainstone.

(1) Oolitic grainstones, which form 40% of the succession. These sediments are composed predominantly of poorly sorted medium to very coarse sand grade (500-1500µm) ooids and peloids. Grapestones, bivalve and echinoderm fragments, and Foraminifera (Textulariidae) forming a small sediment fraction.

(2) Oolitic-bioclastic grainstones contain up to 40% oolite, along with fragments of Coelenterata, bivalves, echinoderms and Foraminifera (Trocholina, Fischerinidae, Nautiloculina and Lituolidae).

(3) Brecciated mudstones, which are predominantly recrystallised and dolomitized.

The microfauna is consistent with a Middle Jurassic to Oxfordian age [Fornos et al. 1987], these sediments could therefore represent the Middle or Upper Jurassic. To the authors knowledge, no similar outcrops occur elsewhere on the island. Fornos et al. [1987] rule out tectonic emplacement of the oolite, and consider the post-ammonitico rosso sequence to represent a shallow shelf environment of sedimentation.

Discussion

Resedimented oolite is a characteristic feature of the Middle Jurassic and not the Upper Jurassic of the Sierra de Levante, yet Fornos et al. [1987] favour an Upper Jurassic (Malm) age for the oolitic platform sequence and state that the upper contact is not exposed. There is however other evidence to suggest that the sequence is of
Figure 3-28

Log section of the Cuevas d'Arta road section, from Fornos et al. [1987]. Inset is geological map after Sabat [1986]. 14 & 15 = Lower Jurassic platform dolomites; 13 = Platform Limestones; 12 = Dogger marls (Middle Jurassic); 7 = undifferentiated Dogger; 8 = platform oolite. A = road section; B = Sa Talaia section.
CUEVAS D'ARTÀ
SECTION

(from FORNOS et al. 1987)
Middle Jurassic age, and that the oolitic platform was already drowned by the Oxfordian. Bourrouilh [1973] records thin ammonitico rosso levels and poorly developed hardgrounds that contain Oxfordian microfauna, on the NE flank of Cap Vermey (the Cuevas d'Artá headland Fig. 3-4). These facies, that indicate platform drowning, overlie the dolomitized platform sequence, which is in lower thrust contact with the Lower Jurassic platform limestone sequence. This platform drowning event correlates with radiolarite deposition over the rest of the Sierra de Levante basin. Kimmeridgian facies from this locality, as reported from Bourrouilh [1973] are Saccocoma-rich resedimented limestones, like those in the rest of the Sierra de Levante basin (e.g. Plate 3-I,D). This evidence constrains the platform to an upper age limit of pre-Oxfordian.

Rapid drowning and deepening upon the onset of drift in the Callovian was the most likely cause for platform termination, widespread hiatus and condensed facies development seen over Mallorca especially during the Callovian and into the Oxfordian. The author therefore considers the platform sequence developed at the Cuevas d'Artá (Plate XX) to be representative of the Bathonian oolitic source platform for the Cutri Formation.

3.3.4.3 Sa Talaia

The section at Sa Talaia (Plate XX), when traced up-section from the Lower Jurassic platform limestones, is shown in Fig. 3-29, B. In contrast to the Cuevas road section, the thick ammonitico rosso limestone is overlain by 6m of facies that are identical to the Cuber Formation at Puig Cutri, namely wackestone and mudstone couplets with abundant Posidonia.

Above this lies a 2m thick sheet-like oolitic unit, along which a series of vertical profiles was constructed (Fig. 3.2.9. and Plate 3-XXI, A). The unit is composed of at least two amalgamated turbidite flows, or two surges from one major flow. The style of resedimentation is identical to that developed at Puig Cutri, namely by high-density oolitic turbidite deposition. A few metres above the oolitic unit is a thin clast-supported conglomerate unit. This is composed of wackestone clasts up to 5cm across (Fig. 3-29), which may be rich in Posidonia, set in a matrix of medium-coarse sand to granule grade oolitic-peloidal packstone. This closely resembles conglomeratic facies developed at Amoixa (Plate 3-III) and both these units are of similar light colour to the Amoixa sequence. From the stratigraphic position of the Sa Talaia oolitic unit, it may
Figure 3-29

Post-rift sequence development at Sa Talaia.
KEY

- peloids
- intraclast chips
- lamination
- Posidonia

Conglomerate
nodular & bedded wackestone
Resedimented Oolite
Cuber Fm.
Ammonitico Rosso

16 m - no exposure
Hardground
Liasic Platform
correlate with the thin oolitic unit exposed in the Cuevas road section (Plate XXI, B).

For about 20m above the conglomerate lies a succession of bedded, often nodular, Posidonia wackestones, above which lies a massive oolite limestone bluff over 10m thick, that is the lateral equivalent of the Cuevas d'Artá oolite. The contact of the oolitic unit on the underlying wackestones is abrupt and disconformable (Plate 3-XXI, C and D). The bedding underneath the contact is significantly deformed, and has been extensively affected by dissolution and weathering (Plate 3-XXI, D). The suspected structural nature of this contact is strengthened by the deformation and shortening seen in the underlying Sa Talaia resedimented oolite unit (Plate 3-XXI, C, arrow) and may therefore indicate a structural contact between the platform oolite and the underlying facies.

Discussion

Although only a reconnaissance survey of this locality was undertaken, important features of the sequence are immediately obvious. Only 0.5km to the NW of the Cuevas road section, there is unequivocal evidence for oolitic gravity deposits, of similar nature (although of reduced occurrence) to those observed elsewhere in the Sierra de Levante. The facies equivalents of the Cuber and Cutri Formations are stratigraphically condensed when compared to sections further west, the "Cutri Formation" being dominated by bedded and nodular wackestones.

The Sa Talaia section is interpreted as representing the upper by-pass slope of the Sierra de Levante basin. Oolitic turbidites mainly by-passed the slope, probably through a system of gullies as on modern by-pass slopes [e.g. Schlager & Chermak 1979; Mullins et al. 1984; Read 1982 and 1985]. The thin-bedded often nodular wackestones and mudstones are interpreted as upper-slope peri-platform sediments and the oolitic turbidite units and conglomerate units as shallow shoe-string upper-slope gulley-fills [c.f. Cook 1983; Read 1985].
Plate 3-XXI

A) Middle Jurassic upper slope facies at Sa Talaia. The resedimented oolitic unit, interpreted as an upper-slope sand stringer gulley fill, is marked with an arrow. Note the nodular aspect of the bedded wackestones towards the top left of the photograph.

B) Resedimented oolitic unit in the Cuevas d'Arta road section over- and underlain by thick bedded *Posidonia* wackestones. This oolitic unit may be the equivalent of the unit shown in (A). (Hammer 30cm)

C) View of Sa Talaia from the NE, looking SW, to show the thrust contact between the massive (platformal) oolite and the tectonically deformed underlying Sa Talaia slope succession (arrow).

D) Close up of contact between massive oolite and slope facies at Sa Talaia, note deformation of underlying facies.
3.3.5 The Jurassic of Cabrera - its relationship and importance to the Cutri Formation of Mallorca

Cabrera Island lies 15km to the SSW of Mallorca, and is a south-westerly prolongation of the Sierra de Levante of Mallorca [Arbona et al. 1984/85; Sabat & Santanach 1985]. The whole island occurs within the uppermost structural unit 7 of Sabat [1986]. The Middle Jurassic succession consists predominantly of nodular limestones and ammonites collected from these condensed facies allowed specific biostratigraphic dates to be placed at specific horizons (Fig. 3-30). Oolitic limestone occurs within the nodular facies from the Lower to Upper Bathonian and is considered to be the equivalent of the Cutri Formation in the Sierra de Levante of Mallorca [Arbona et al. 1985/85]. The Middle Jurassic succession is extensively affected by syn-sedimentary faulting, which is also associated with oolite emplacement [Sabat & Santanach 1984].

The Jurassic Succession of Cabrera

The Lower Jurassic succession is represented by peri-tidal carbonates. The dolomitization front observed is subhorizontal, indicating post-rift dolomitization prior to sedimentation of the pelagic series [Sabat & Santanach 1984] (Fig. 3-31). As on Mallorca, post-break up drowning is marked by red crinoidal grainstones and packstones of seamount affinities, which have been dated as Middle-Upper Carixian [Arbona et al. 1984/85]. Quartz-arenites are intercalated in these facies, indicating foundering occurred slightly earlier in Cabrera. The hardgrounds are dated as Lower Domerian [Arbona et al. 1984/85], being older than the Toarcian hardgrounds of the Sierra de Levante [Prescott 1988]. A 'stratigraphic gap' occurs in the Middle Domerian/Middle Bajocian [Arbona et al. 1984/85], followed by Upper Bajocian ammonite-rich packstones or nodular limestones.

The Lower Bathonian consists of nodular limestones with silicified packstones and wackestones. Slumps are abundant and syn-sedimentary faults affect the sequence [Sabat & Santanach 1984] (Fig. 3-31). The Middle Bathonian is characterized by discontinuous oolitic grainstones 20-35m thick, described as an olistolith [Sabat & Santanach 1984], overlain by thin-bedded oolitic grainstones and wackestones with Posidonia (Fig. 3-30). These facies are succeeded by nodular, bedded limestones of the Middle-Upper Bathonian, which are overlain by a discontinuous, erosively based, carbonate conglomerated, up to 8m thick, described as an olistostrome by Sabat & Santanach [1984]. Nodular packstones and wackestones, which in the upper part
contain Kimmeridgian fauna, overlie the conglomerate. Micrites with *Saccocoma* and then the Maiolica overlie these sediments.

The succession of Cabrera shows important implications for the coeval succession of Mallorca:

(1) Oolite resedimentation occurred from the Lower–Upper Bathonian.

(2) The Callovian 'stratigraphic gap' is most likely to occur within the nodular wackestones overlying the conglomerate, but it cannot be ruled out that the conglomerate is of Callovian–Lower Kimmeridgian age.

(3) Oolite resedimentation is intimately related to syn-sedimentary tectonism, being preceded by a phase of tectonism that created a disconformity, and affected by later syn-depositional faulting [Sabat & Santanach 1984].

(4) Syn-sedimentary faulting, which indicates NNW-SSE extension [Sabat & Santanach 1984] occurs throughout the Middle Jurassic and into the Lower Cretaceous.

From descriptions by Sabat & Santanach [1984] and Arbona *et al.* [1984/85], the interpretation of the Bathonian oolite as an olistolith seems inappropriate and as will be explained, the oolite has characteristics of being gulley-fill deposits. Sabat & Santanach [1984] describe the oolite as being 'conjugate' in nature, indicating that more than one 'olistolith' or gulley-fill unit is present, intercalated within the nodular limestone. The 'olistolith' shows internal graded bedding from oolitic packstones to wackestones with *Posidonia*. With such large blocks as the Cabrera 'olistolith', plastic deformation, intimately connected with base of the block would be expected [e.g. Teale & Young 1987]. Slumping occurs several metres below the oolite and the base of the 'olistolith' is erosional, with soft sediment deformation occurring on its lateral margins [Sabat & Santanach 1984]. The margins of the 'olistolith' are also associated with slumping and syn-sedimentary faulting. Syn-sedimentary internal discordances show soft-sediment deformation of the oolite, which in some cases has led to post-depositional resedimentation (Fig. 3-31, D).

This evidence points to the fact that the oolite was not emplaced as a solid block, but as oolitic sediment, and the 'conjugate olistoliths' are reinterpreted as upper slope gulley-fill deposits. Gulley formation results as a consequence of by-pass slope
Figure 3-30

The Jurassic succession of Cabrera (from Arbona et al. [1984/85])
development [Mullins & Cook 1986] and the nodular facies of the Cabrera Middle Jurassic are characteristic of by-pass upper-slope facies of intergulley areas [e.g. Mullins et al. 1984; Hine & Neumann 1977; Mullins 1983b]. Gullies may originate as slides or slumps and then become enlarged by gravity flow processes [Mullins et al. 1984] and their position may often be controlled by syn-sedimentary faults [e.g. Stow 1985a]. Slumped strata and syn-sedimentary faults are intimately associated with the location of the oolite, a major phase of syn-sedimentary faulting creating an angular disconformity that preceded gulley formation, syn-sedimentary faults control the gulley margins and displace the base of the oolite. Another episode of syn-sedimentary faulting preceded emplacement of the stratigraphically higher conglomerate [Sabat & Santanach 1984].

Summary

The Middle Jurassic succession of Cabrera (Fig. 3-29) is interpreted as representing the upper by-pass slope of the Sierra de Levante basin. The presence of abundant syn-sedimentary faults associated with olistolith block emplacement (i.e. olistolith blocks sensu stricto, as in Fig. 3-30, C), and slumping preceding oolite resedimentation, all indicate tectonic instability [e.g. Nardin et al. 1979; Cook 1983a]. These features are interpreted as resulting from the close proximity of this area to the main fault scarp that differentiated the platform from the Sierra de Levante Basin. The olistolith blocks would have been detached from the fault scarps during fault-induced earthquakes, which cut back the margin [e.g. Teale & Young 1987].

The conjugate olistoliths of Sabat & Santanach [1984] are interpreted as gulley-fills that became plugged with oolitic turbidites, their position being controlled by the position of syn-sedimentary faults and slumps. Similar features of truncation surfaces, disconformities, slide scars and slumps to those seen on Cabrera have been described from the slopes of Little Bahama Bank, where they are related to syn-sedimentary tectonism along the Walkers Cay Fault [Mullins & van Buren 1981]. The condensed and nodular aspect of much of the Cabrera succession characterizes sediment 'starved' inter-gulley areas of modern by-pass slopes, where coarse sediment transits the slope through gullies [Hine & Neumann 1977; McIlreath & James 1978; Schlager & Chermak 1979; Mullins et al. 1984].
Figure 3-31

A) Syn-sedimentary conjugate fault system affecting the Mesozoic series. Es Cap des Morobuti and Sa Punta des Codolar.

B) Conjugate system of normal faults and unconformity near Sa Cova.

C) Discordance of post-rift sediments upon the Liassic platform at Cala Forco and presence of an olistolith block.

D) Oolitic "olistolith" intercalated in the lower part of the post-rift sediments, lateral brecciation and associated slumps. Northern Cabrera Island.

E) Post-rift sediments unconformably overlying Liassic; the platform at Es Cap de Llebelg and the sub-horizontal dolomization front.

(compilation from Sabat & Santanach [1984])
3.3.6 Summary

The Cutri Formation is interpreted as a base-of-slope apron after Cook [1983] and Mullins & Cook [1986] and shows many similarities with clastic slope apron models [Stow 1985a] (Section 1.4.2). Under the platform margin classification scheme of James and Mountjoy [1983], it would classify as a sand-shoal dominated margin, and as the by-pass margin of Read [1985]. The apron was at least 100km long and 40km wide, and was sourced from the east, not the north-west as proposed by Alvaro et al. [1984].

In this model, the outcrops within structural unit 5 display regions in closest proximity to the source platform: the Cuevas d'Artá outcrop representing a fraction of the source platform; and the outcrop at Sa Talaia representing the upper by-pass slope, characterized by a stratigraphically reduced section of bedded and nodular wackestones and rare oolitic shoe-string gulley-fills. Amoixa characterizes the most proximal inner apron setting and is interpreted to have formed at a shallower palaeodepth on a structurally elevated tilt-block. The section at Amoixa II represents an abrupt transition to the deeper Sierra de Levante basin, and represents the muddy proximal base-of-slope, truncated by broad U-shaped oolitic packstone and wackestone plugged gulley-fills.

The sections of the Puig Cutri region of the Sierra d'Artá in structural unit 3 record the retrogradational mega-sequence development of the base-of-slope apron from inner apron to apron-fringe and basin plain; while the succession at Puig d'en Pare in structural unit 1 records sedimentation on the outer base-of-slope apron.

A region that acted as a coeval submarine high bounded the Sierra de Levante basin to the north and is represented by the outcrop of structural unit 2 in the northern Sierra d'Artá. Middle Jurassic facies on Cabrera represent the laterally equivalent (c. some 90km from the Puig Cutri area) starved upper by-pass, gullied slope of the Sierra de Levante basin that was affected throughout the middle Jurassic by syn-sedimentary faults which also controlled the location of gullies.

The palinspastically restored distance between the upper slope at Sa Talaia and the distal section of Puig d'en Paré was some 35km, while the Amoixa localities lay some 75km along strike from Sa Talaia. The confinement of specific facies types within particular structural units is interpreted as reflecting the Middle Jurassic structural control on intra-basin block topography. The apron displays a retrogradational
sequence trend, which is interpreted as a single syn-rift thinning upward mega-sequence [c.f. Surlyk 1978; Stow 1985b; Eberli 1987 and 1988] form as the result of past-rift subsidence.

The oolitic units of the Cutri Formation were emplaced as high-density turbidites and debris flows, which form discrete and laterally correlatable, resedimented oolitic units, that episodically punctuate the background hemipelagic sedimentation of the basin. Seven major episodes of resedimentation are distinguished from the apron sections of structural unit 3 (units A, B, C, D, E, H and I) and each unit may be sedimentologically correlated with units 65km along strike in structural unit 5 (Fig. 3-26).

The sharp transitions observed between the oolitic units and their intervening hemipelagic units indicate an allocyclic control on sandbody architecture [e.g. Mutti 1985; Larue 1985; Cazzola et al. 1985; Spalletti et al. 1989]. The transition form hemipelagic sedimentation to oolite resedimentation represents a progradational rejuvenation sequence (after Larue [1985], Larue & Speed [1983], Spalletti et al. [1989]), attributable to a relative fall of sea-level, by eustatic means [e.g. Cook et al. 1987] or by tectonic rejuvenation of the source area [e.g. Stow et al. 1985; Gawthorpe 1986; de Vries Klein 1984]. The transition from oolite resedimentation to hemipelagic sedimentation is a retrogradational sequence transition [Larue 1985], developed as a result of relative rise of sea-level, by eustatic means [e.g. Pickering 1982; Howell & Vedder 1985; Stow et al. 1985; Damuth & Flood 1985] or by increased tectonic subsidence of the source area [e.g. Hurst & Surlyk 1984; Watts 1987]. The presence of intervening hemipelagic facies indicates the source platform was always submerged and able to produce peri-platform fines and carbonate mud.

Sandbody architecture of the Cutri Formation is interpreted as being allocyclically controlled, by relative sea-level fluctuations; whether these fluctuations are the result of eustatic sea-level variations or tectonically induced changes of sea-level [e.g. Burton et al. 1987] is the subject of discussion in the following Section.
Figure 3-32

Proposed depositional model of the Cutri Formation.
3.4 Controls on Sandbody Architecture

3.4.1 Palaeogeographic Control

During the Bathonian, this area lay at a latitude of some 23°-24°N (by extrapolating the maps of Dercourt et al. [1986]) placing it on or very close to the Tropic of Cancer (23¼°N). It was separated from the extending (not spreading) Ligurian Ocean zone to the east by a linear landmass, the Alboran-Kabylia-Stilo-Brianconnais landmass (Fig. 1-2 & 1-3), which presumably afforded shelter from the effects of the Tethyan ocean to the east. It is this landmass that most likely sourced the Carixian quartz arenite of the area (Section 3.1.3), and the very rare quartz grains occurring as ooid nuclei in the Cutri Formation.

Hurricanes (wind speeds >75mph) may be expected to affect palaeo-latitudes of between 10-45°, with a broader belt inferred for the non-glacial Mesozoic [Duke 1985]. Winds intensify to hurricane status over warm seas [Adam 1975] and with prevailing north-easterly wind blowing across Tethys [Tottman-Parrish & Curtis 1982], Western Tethys was well situated to encounter strong hurricane effects. The region was therefore within the Mesozoic hurricane belt and was likely to be affected by persistent north-east trade winds. Storms as agents of sediment transfer to the bank margins are likely to have been important; however the Alboran landmass probably acted as buffer sheltering the region from the east.

There is also the possibility that summer low pressure cells situated over the Atlantic region gave rise to north-westerly and south-westerly winds that could have affected the area [Tottman-Parrish & Curtis 1982]. There are uncertainties too as to the effects of extension and rifting in the Atlantic and western Tethyan realm on circulation patterns at this time [Tottman-Parrish & Curtis 1982]. With respect to the northeasterly trade winds, the source platform margin would have been leeward, with respect to the low pressure cell winds, it would be windward. Since the margin was open to the west, it was probably more strongly influenced from waves, tides, currents etc. from this direction, rather than from the north-east and may therefore have acted as a windward margin.

A phenomenon of the Cutri Formation oolite is the predominance of oolitic-peloidal sediment and a relative paucity of skeletal components, whether coated or uncoated. This type of sediment is mirrored on the modern Caicos Banks [Lloyd et al. 1987] and
indicates that a low organic-productivity platform was the source for the oolite. On Caicos Bank, ooids form in 'lower energy' areas away from the high-energy shelf margin and are generated by wave and current agitation generated by SE trade winds, rather than tidal energy. Pelletal sandwaves are also formed offshore by storm-generated winds, which often end up as ooid nuclei. Such a 'low-energy' broad shallow epeiric oolitic platform, rather than shelf edge shoals, may well have been produced in the area, by the sheltering of the region by the Alboran landmass.

The slope evolution of the Sierra de Levante basin shows a resemblance to facies development along the open-ocean windward margin of Little Bahama Bank (Section 2.5.2. and Fig. 2-6). As extension proceeded through the Toarcian to Bathonian, a sequential vertical stratigraphic sequence evolved that mirrors facies transitions downslope and north-east of Little Bahama Bank [Mullins 1983b; Mullins et al. 1984]: the passage from hardgrounds to nodular ooze, unlithified peri-platform ooze and then the carbonate apron with depth on the slope of Little Bahama Bank, mirrors the progression from the Toarcian hardground, to the Aalenian ammonitico rosso, to the peri-platform rhythmites of the Bajocian Cuber Formation, and into the Bathonian oolitic apron, seen in the Sierra de Levante of Mallorca (Fig. 3-3).

The geographical situation of Little Bahama Bank shows similarities with the Jurassic palaeogeography of Mallorca: the shallow water carbonate platform of Little Bahama Bank is separated from a large, isolated, submarine plateau, the Blake Plateau, by a deeper trough, these topographic areas being analogous to the source platform, the Sierra Norte and the Sierra de Levante respectively.

If the source platform was predominantly affected by winds and currents from the open sea that lay to the west (Fig. 1-3), rather than being dominated by the north-easterlies blowing across Tethys and over the Alboran-Kabylia landmass, the margin may have been a windward type margin. This may help to explain the nature of facies with the Jurassic Sierra de Levante basin, which were predominantly peri-platform derived fines, episodically interrupted by major resedimentation events. A windward margin setting would result in net onbank transport of sands, and a predominance of mud and fine sand grade facies in the basin (see Section 2.4.2.). A distally steepened platform margin (i.e. steepened by faults) may have additionally favoured the mud-rich nature of the sandbody [c.f. Read 1985].
The presence of contourites may reflect bottom current intensification due to flow restriction through the narrow Sierra de Levante basin, which was a submarine depression bounded to the north-west by the submerged pelagic plateau of the Sierra Norte. Also, contour currents are intensified on the western margins of basins by the coriolis force which may be another reason for the widespread occurrence of the current winnowed facies during the Middle Jurassic.

3.4.2 Tectonic Control

Indications are that tectonics played the major role in controlling sandbody architecture and geometry; both by a primary morphotectonic control of the Sierra de Levante basin, and by secondary control governing resedimentation phases. The tectonic control of the Sierra de Levante basin has already been demonstrated and is a characteristic feature of both rifted passive margins and transform margins [e.g. Mougenot 1983; Stow 1985b; Stow et al. 1985; Crevello 1988]. The Cutri Formation was also relatively short-lived by comparison with the other two resedimented oolite sandbodies studied, which is to be expected from active tectonic settings during rift phases of evolution, where source areas and depositional basins are affected by tectonics [e.g. Mutti 1985; Mutti & Normark 1987; Shanmugam & Moiola 1988].

The Jurassic Sierra de Levante basin underwent continued subsidence from the Toarcian onwards, as a result of the thermal contraction phase of post-rift evolution [e.g. Sleep 1976; Steckler & Watts 1978; Boillot 1981]. Fault activity decreases with the onset of drift [e.g. Bernoulli & Kalin 1984] and the pre-drift Bathonian was therefore likely to be affected by extension related faulting. Syn-sedimentary faults throughout the Middle Jurassic sequence of Cabrera [Sabat & Santanach 1984] and the occurrence of olistolith blocks are evidence that this was indeed the case. The emplacement of olistolith blocks (not the one described as an olistolith (i.e. the gulley fill sequence) but olistolith blocks associated with syn-sedimentary faulting on Cabrera, indicates a fault control on differentiation of platform and slope, and the proximity of Cabrera to the main fault scarp [c.f. Teale & Young 1987].

The retrogradational nature of the sandbody indicates that tectonic related subsidence was dominant, preventing apron progradation, and giving rise to an overall decrease in the volume of sediment gravity flows. This sandbody trend is characteristic of tectonically dominated resedimented sequences from other rift basins [e.g. Stow 1985b; Watts 1987; Eberli 1987]; and may have occurred by episodic, tectonically induced
back-stepping of the platform margin [e.g. Playford 1980; Hurst & Surlyk 1984] by fault-induced earthquakes [e.g. Teale & Young 1987].

In addition to tectonic setting, the magnitude, location and periodicity of syn-sedimentary tectonic activity are significant factors in controlling resedimented sandbody development: basinwide apron/fan abandonment or rejuvenation; channel erosion, filling and abandonment can all be influenced by tectonic activity [Stow et al. 1985; Gawthorpe 1987a; Cazzola et al. 1985; Spalletti et al. 1989]. A gross sandbody architecture characterized by rapid transitions between resedimented packages and finer grained basinal facies are indicative of sediment discharge due to periodic tectonism [e.g. Cazzola et al. 1985; Mutti 1985; Gawthorpe 1987a]. The abrupt transitions observed between the oolite rejuvenation sequences and the retrogradational hemipelagic facies of the Cutri Formation are highly suggestive of a direct tectonic control on resedimentation [e.g. Mutti 1985], especially since some of the oolitic units are capped by debris flows, suggesting tectonically induced subsidence terminated resedimentation [e.g. Hurst & Surlyk 1982]. Direct tectonic initiation of carbonate turbidites composed of shallow-water carbonate sediments has been proposed by Bernoulli & Kalin [1984]; Colacicchi & Baldanza [1986]; Halliwell-Hazlett [1988] and Cooper [1989].

The distinctive style of resedimentation by major high-density oolitic turbidites indicates infrequently initiated and rapidly deposited mass flows [e.g. Wueller & James 1989; Middleton & Neal 1989] resulting from a single major event. Similarly thick, individual high-density turbidites have been interpreted as tectonically induced by Mutti [1985] and Wezel et al. [1981], and such major flows may be termed seismoturbidites, implying a seismically induced collapse of stored sediment [Mutti et al. 1984]. The occurrence of debris flows, all with clasts of semi-lithified coeval slope sediments, is evidence for remobilisation and failure of the slope; episodic tectonic subsidence is a likely cause for such mass-flow events [e.g. Surlyk & Hurst 1982; Gawthorpe 1986; Cook et al. 1979; Cook & Mullins 1983]. The presence of slope derived clasts may indicate that faults steepened the outer part of the platform margin and slope [e.g. Read 1985]; a distally steepened platform margin could also account for the high productivity of peri-platform muds and fine peloidal packstones seen in the apron succession.

Cazzola et al. [1985] describe a siliciclastic sequence (known as the Cengio sandstone member of the Tertiary Piedmont Basin in north-west Italy) characterized by eight
active episodes of resedimentation, separated by mudstone units, where the
resedimented units are interpreted as the product of intermittent, syn-sedimentary
tectonic uplift of the source area, related to periods of strike-slip faulting. Similarly,
Gawthorpe [1987a] interprets coarse-grained carbonates influxing background sediments
as reflecting episodic remobilization of carbonate sediments off the footwall block of a
half-graben system, from the Carboniferous of the Bowland Basin, northern England.

The Cutri Formation shows analogous resedimentation patterns to these two examples.
Episodic tectonic uplift (either in a divergent strike-slip regime [e.g. de Vries Klein
1984; Reading 1980; Kelts 1981] or an extensional regime [e.g. Leeder & Gawthorpe
1987]) in response to tectonic subsidence in the Sierra de Levante basin (hangingwall) is
a likely mechanism for gravity flow initiation of sediments off the source platform
(footwall). Uplift in footwall blocks occurs as a result of rotation of tilt blocks along
listric faults [Gibbs 1982] and/or by rebound upon unloading of the footwall block
during downfaulting of the hangingwall [Jackson & McKenzie 1983].

Episodic resedimentation in the Cutri Formation involved the basinward transportation
of large quantities of unconsolidated shelf-edge oolitic sediment, their presence in the
basin indicating episodic collapse of the platform margin accumulation of oolitic sands
[e.g. Hine 1983a; Mullins et al. 1986; Halliwell-Hazlett 1988]. Tectonic instability or
oversteepening are often considered to be the causes of this mass movement [e.g. Hine
et al. 1981; Hine 1983a; Cook 1983]. Resedimentation of the quantities of
unconsolidated oolitic sands involved in the formation of the Cutri Formation
necessitates the previous formation and accumulation of these carbonate sands at the
platform margin at sufficiently high rates of sedimentation to prevent their lithification
[i.e. Hine & Neumann 1977; Hine et al. 1981a and b].

This condition would be met during a relatively slow net sea-level rise on a subsiding
shelf [Mutti 1985] that would allow the shelf edge reservoirs to become loaded with
oolitic sands [Hine et al. 1981b]. It therefore follows that large carbonate sand gravity
flow bodies have to be derived from a shelfal area that underwent sufficiently long
periods of slow relative sea-level rise (i.e. subsidence/uplift effects, plus or minus
eustatic effects) to accumulate the volume of oolitic sands that were resedimented at
the shelf-edge prior to resedimentation.

Resedimentation necessitates a trigger in the form of a relative fall of sea-level: either
in the form of tectonically induced uplift of the platform margin [e.g. de Vries Klein
1984; Cazzola et al. 1985; Gawthorpe 1987a] or by a eustatic sea-level fall [e.g. Mullins et al. 1986; Cook et al. 1987] to give rise to instability and retrogressive wasting of the shelf-edge sands [e.g. Mutti 1985; Ricci Lucchi 1985b].

Superimposed on an actively subsiding tectonic regime, periods of eustatic sea-level fall would give rise to an episode of slower net sea-level rise, which would encourage the accumulation of oolite at the bank margin. This process could itself produce bank margin instability, or would simply provide reservoirs of oolite sands which were resedimented in response to tectonic initiation. Periods of eustatic sea-level rise would have the opposite effect, producing a faster net sea-level rise, rendering oolite shoals inactive and leading to episodic abandonment of resedimentation in the basin. The possibility of eustatic control is discussed in the following Section.
3.4.3 Eustatic Control

With reference to the eustatic sea-level curve of Haq et al. [1987 & 1989] (Appendix Fig. 5) the stratigraphic occurrence of the Cutri Formation coincides with an episode of long-term sea-level draw down that occurred in the Bathonian (Fig. see Appendix Fig. 5). There is therefore a likelihood of an eustatic control on sandbody development. This episode of eustatic sea-level fall occurs sandwiched between periods of geological time when long-term eustatic sea-level was undergoing net rise, namely the Aalenian-Bajocian and the Callovian-Kimmeridgian.

During the Bathonian, Haq et al. [1987 & 1989] recognize four transgressionary-regressionary eustatic pulses, which form third order cycles 2.2, 2.3 and 2.4 of supercycle LZA-2, and part of cycle 3.1 of supercycle LZA-3. These pulses combined to produce a net eustatic sea-level fall of about 70m from the beginning of the Bathonian (165 million years ago) (Appendix Fig. 5) to the end of the third order cycle 2.4 (at 156 million years ago). The long term sea-level was beginning to show a rise from a lowstand, in third order cycle 3.1. These short term fluctuations within the third order cycles may have played an important role in oolite resedimentation to the Cutri Formation apron.

The long term sea-level fall correlates with a resedimented sandbody that shows a retrogradational sequence, indicating subsidence outpaced sedimentation. The effect of a net sea-level fall on an actively subsiding region would be to counteract subsidence rates by producing a slower rate of net sea-level rise; and conversely, the effect of eustatic sea-level rise would be an enhanced relative rise. These periods of net slower and rapid eustatic sea-level fluctuations may well have controlled resedimentation during development of the Cutri Formation: oolite may have been resedimented during episodes of slower net sea-level rise (i.e. when eustatic sea-level was falling) while hemipelagic sedimentation prevailed and draped the apron when net sea-level rise was rapid (i.e. when eustatic sea-level was rising). This hypothesis is discussed with reference to Fig. 3-33.

A short term eustatic sea-level fall commenced within the Lower Bathonian after the maximum flooding interval of third order cycle 2.2 (163.5 m.y). This interval correlates with global condensed sequences at this time [Vail et al. 1984]. The deposition of localised condensed facies at the top of the Cuber Formation, prior to oolite resedimentation (see Section on Amoixa, 3.3.4.1 & Puig de Ses Fites, 3.3.1.2 )
Figure 3-33

Possible eustatic control on the Cutri formation.
<table>
<thead>
<tr>
<th>SMW</th>
<th>TR</th>
<th>HS</th>
<th>SMW</th>
<th>TR</th>
<th>SMW</th>
<th>HS</th>
<th>LSW</th>
<th>TR</th>
<th>HS</th>
<th>SMW</th>
<th>TR</th>
<th>HS</th>
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</thead>
<tbody>
<tr>
<td>163.5</td>
<td></td>
<td></td>
<td>161.5</td>
<td></td>
<td>159</td>
<td></td>
<td>158.8</td>
<td></td>
<td>158</td>
<td></td>
<td>157</td>
<td></td>
</tr>
</tbody>
</table>

Curve after Haq et al. 1987 (not drawn to scale)

**Note:**
- **oolite units A-D**
- **E, H & L**
- **Q-R**

**Legend:**
- **3rd Order cycles**
- **2nd Order cycles**

**Strata:**
- **Bathonian**
- **Callovian**
may represent this interval.

The commencement of sea-level fall after this highstand event culminates in a lowstand at the end of third order cycle 2.2 (a period of some 2.5 million years), which is characterized by a global shelf margin wedge systems tract (SMW) situation [Haq et al. 1987 & 1989]. This episode of sea-level lowering could have facilitated the resedimentation of the major lower units A to D of the Cutri Formation (see Section 3.3.1.1 & Appendix Fig. 2), which account for the majority of oolite resedimented to the sandbody. Evidence from Cabrera (Section 3.3.5) demonstrates that major oolite resedimentation occurred at the Lower to Middle Bathonian boundary, which correlates with this period.

The ensuing transgression (TR) within third order cycle 2.3 would have given rise to a period of increased relative rise, during which time the major slope-drape episode of interbed D to E contourites (Appendix Fig. 2 & Section 3.3.1.1) occurred. Apron-wide contourites are developed at this stratigraphic horizon (Fig. 3-26 & 3-22) and contourites are classically developed within highstand episodes on modern aprons [e.g. Stow 1985a; Stow et al. 1985] and fan [Stow 1979 and 1981].

Eustatic sea-level fall at the beginning of the third order cycle 2.4 (SMW) and at the end of the same cycle (LSW) may have facilitated resedimentation of the major turbidites of units E, H and I respectively (shown in Appendix Fig. 2).

The ensuing transgression (TR) of third order cycle 3.1 could have produced the abrupt reduction of resedimentation rate seen in the post unit I Cutri Formation mega-sequence (Appendix Fig. 2), and the predominance of hemipelagic facies and contourites above this level.

It is not known whether the Cutri Formation continued up into part of the Callovian; it may be possible that an episode of sea-level fall in the basal Callovian of third order cycle 3.1 may have accounted for resedimentation of units Q, R and S within the upper part of the mega-sequence. After this slight short term fall, sea-level continued to show a long term rise through to the Kimmeridgian (Appendix Fig. 5).

There is strong evidence to suggest that an episode of long term sea-level drawdown controlled longevity of the Cutri Formation apron, resedimentation commencing at the
start of long-term sea-level fall and ceasing with the commencement of long-term sea-level rise. Minor sea-level fluctuations within the third-order cycles may have played an important role in the episodic resedimentation of the oolitic turbidites.

3.4.4 Summary

The development of the Cutri Formation apron was governed by dominant tectonic control and is interpreted as a footwall apron system (after Surlyk [1978]; Stow [1985b]; Eberli [1987]) that formed a single thinning-upward syn-rift mega-sequence. Dominant subsidence resulted in the retrogradational trend of the apron which culminated in platform drowning (see Section 3.3.4.2) marking the final part of the deepening upward sequence, with deposition of the Puig de Ses Fites radiolarites (Appendix Fig. 2).

Superimposed on the tectonic control was an important secondary control both by long term net sea-level drawdown and by third order cycle eustatic sea-level fluctuations. While long term sea-level probably enabled development of the sandbody, its effects were not substantial enough to counteract subsidence sufficiently for progradation or aggradation to occur. However, short term episodes of sea-level fall within the third order cycles served to produce episodes of slower net sea-level rise. During these times, tectonic uplift would further aid the reduction of net sea-level rise over the platform, the latter process being favoured by increased isostatic subsidence induced by the sediment load in the basin [Sleep 1976; MacKenzie 1978; Steckler & Watts 1978; Winterer & Bosellini 1981]. This process may be reflected by the occurrence of the large volume of oolite resedimented in the basal portion of the sequence, where the longest short-term period of sea-level fall occurred (Fig. 3-33).

The reduction in volume of oolite resedimented into the basin may have been affected by a lack of platform recharge time if the retrogradational succession developed by episodic tectonic back-stepping [e.g. Playford 1980; Surlyk & Hurst 1982; Winker & Buffler 1988]. Between tectonic stepbacks, the platform could grow by progradation, but due to net sea-level effects may have been unable to occupy the margin of the previous cycle.

The relatively small volume of carbonate sands resedimented may also be due to the fault-bounded nature of the basin; tectonic uplift or eustatic fall only being able to remove limited amounts of sediment [e.g. Mutti 1985]. It may also be due to the fact
that the margin was a palaeowindward margin, which would not favour the offbank transport of sand. This may also explain the volume fine-carbonate muds and silts that characterize the succession, offbank transport of such sediments characterising windward margins and being favoured during highstand situations [e.g. Boardman et al. 1986] (see Section 2*). Also, progradation is favoured along leeward margins and not windward margins [i.e. Hine et al. 1981b; Eberli & Ginsburg 1987 and 1989], the palaeogeographic situation of the margin may therefore not have favoured progradation.

3.5 Final Conclusions

The Cutri Formation is interpreted as a retrograding base-of-slope (faulted slope) apron system that served to plane the rift topography of the Sierra de Levante basin. Sandbody geometry and architecture was primarily controlled by tectonics, both by a primary morphotectonic control of the basin and platform margin, and by secondary ongoing, fault-related subsidence and related episodic footwall uplift.

The importance of the interaction of long term and short term eustatic sea-level fluctuations, acting in concert with tectonic control, in governing both the occurrence and longevity of the sandbody is demonstrated. The relationship between short-term eustatic sea-level fluctuations superimposed on the subsiding tectonic regime, is shown to be important in controlling the periodicity and volume of sediment transported to the basin. These controls limited off-bank sand transport, resulting in the most carbonate mud-rich resedimented sandbody of the three considered in this thesis. The "mud-rich" and relatively "sand-poor" nature of the succession may have been additionally aided by a probably palaeowindward orientation of the source margin, which would also not have favoured progradation of the system.

The stratigraphic occurrence of the Cutri Formation shows important similarities with hydrocarbon bearing siliciclastic submarine fans and aprons [e.g. Shanmugam & Moiola 1988] in that it correlates with an eustatic drawdown of sea-level. The laterally extensive and continuous resedimented oolite beds are vertically encased by "basinal" hemipelagic facies, and the generation of allocyclically controlled stratigraphically adjacent permeability barriers and potential reservoir sands has important implications. However in this case, the oolitic units are predominantly compacted packstones; they may have originated on the margins as mud-free sediment, becoming choked with
carbonate mud upon resedimentation in the mud-rich "low energy" setting of the Sierra de Levante basin, thus drastically reducing their theoretical potential as reservoir facies. The sandbody as a whole is only 80m thick, both the entire sandbody and its component oolite units (25-1m thick) would be extremely difficult to pick up on seismics at depth.