EVOLUTION OF THE GRES D'ANNOT BASIN, S.W. ALPS

Thesis submitted in accordance with the requirements of the University of Liverpool for the degree of Doctor of Philosophy by

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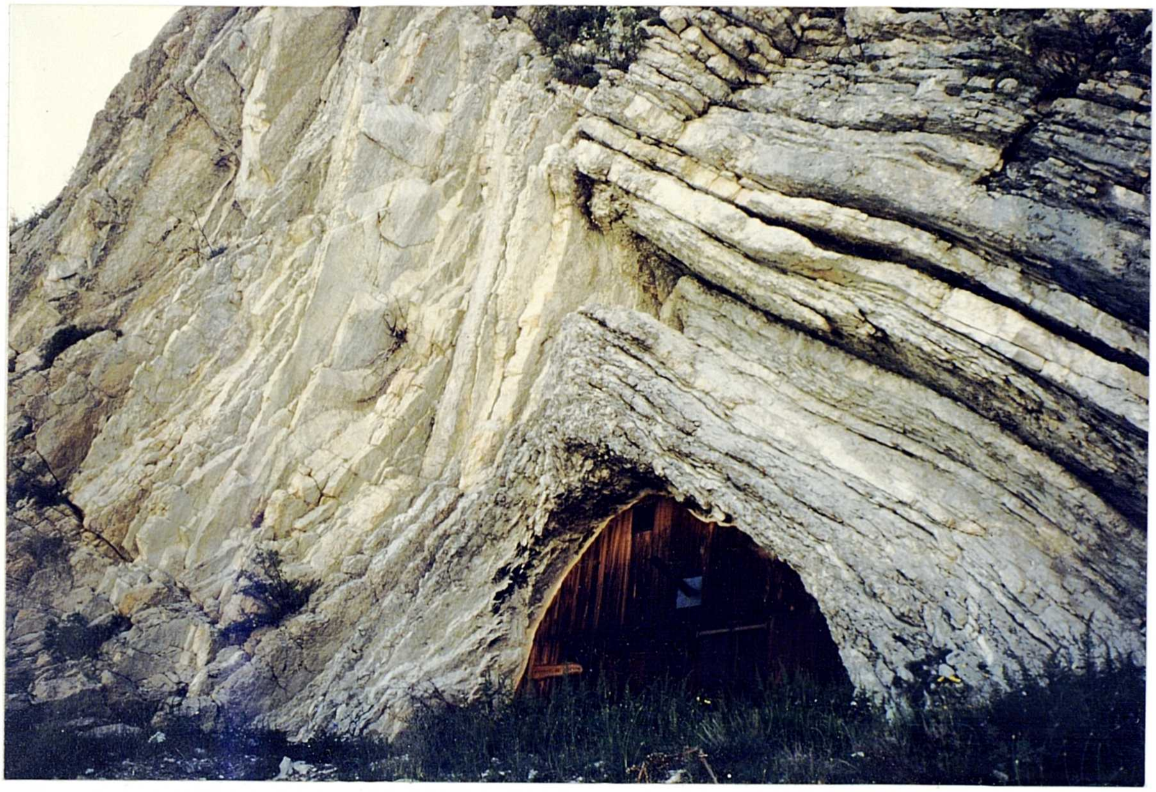

Toi qui viens de si loin
Lorsque tu vois mon toît,
Viens tâter de la main
La flamme de mon bois
Et viens rire avec moi,
Et viens boire avec moi,
Et parle-moi de toi....

André Peyron, La Valette.

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ABSTRACT

The Eocene Gres d'Annot Formation was deposited in a complex foreland basin overlying older, Mesozoic and Permian extensional basins. During the Tertiary, Haute Provence experienced Pyrenean compression from the south (Mid Cretaceous to Eocene) and Alpine compression from the east (Early Eocene onwards).

Regional uplift above E-W trending Pyrenean structures is recorded by a significant unconformity at the base of the Tertiary succession in Haute Provence. During the Eocene, NW-SE ridges developed above thrusts and kink zones at the onset of Alpine deformation. These interacted with the still active Pyrenean structures, producing a complex surface topography which divided the region into small, sometimes isolated drainage basins.

Alluvial fans, comprising the Poudingues d'Argens, fringed the rising structures. In most places, the conglomerate and marlstone sequences were derived entirely from nearby Upper Cretaceous limestone hills. However, the outcrops adjacent to the Barrot massif contain locally derived clasts of Triassic and Permian quartzite and mudstone, showing that this part of the Mesozoic cover has not moved with respect to the Permian massif since its first uplift in the Early Eocene: both are allochthonous. The Barrot massif is an inverted Permian basin. Once it was uplifted, high-level thrusts piled up behind it (Tinee Nappes) and much of the subsequent displacement rooted underneath it.

During the early stages of the Nummulitic transgression the character of palaeosols in the Poudingues d'Argens changed: immature, pyritic, gley soils developed in the wet, lagoonal lowlands due to the influence of salty pore waters. Pseudogley soils persisted on the drier slopes and highlands. The alluvial fans became fan deltas as they were progressively drowned. Shattered clasts and olistostromes in the lower marine facies indicate local syn-depositional seismicity. Within many areas, pulses of deformation caused the fan deltas to prograde, but uplift above individual structures was competing against regional subsidence induced by flexural loading of the European foreland. Eventually, the sea won! The basin deepened rapidly.

The onset of sandy turbidite deposition was diachronous across the region. The turbidity currents were sourced from Pyrenean basement culminations to the south, then flowed along the Alpine foreland basin. Alpine thrusts were partially pinned by the Pyrenean uplifts, and the resulting basin floor topography controlled the sediment dispersal. Sand distribution, palaeocurrents and vertical sequences cannot be predicted using any conventional submarine fan model. Facies relate more to their proximity to palaeoslopes.

In the Oligocene, rifting to the south caused the Pyrenean culminations to subside, and the Alpine nappes were unleashed into the basin.

The evolving palaeotopography is defined throughout the Eocene by sediment thickness and facies changes, together with multiple unconformities and associated onlap and/or truncation. These features constrain the deeper structure and its development. By constructing closely spaced, sequentially restored structural cross sections, it is possible to model the tectonic and sedimentary evolution of the Gres d'Annot basin.

CHAPTER 1: INTRODUCTION

1.1. Geological setting of the study region.

The study region, parts of Haute Provence and Les Alpes Maritimes, (fig.1.1) lies in the Alpine-Himalayan chain, a vast orogenic belt which extends from Morocco to China. The mountain belt formed as a result of the closure of oceanic basins, like Tethys, to the south of the Eurasian plate (Dewey et al, 1973). S.E. France lies at the junction of three parts of this system: the Alps, the Pyrenees and the Apennines. These can, to an extent, be treated as individual mountain belts independent of each other, because they were formed by collision between different continental plates and microplates. Therefore, the timing of deformation in each mountain range and the directions of shortening, in, for example, the foreland thrust belts, differ.

The subject of this thesis is the evolution of the Eocene Gres d'Annot turbidite basin. The basin developed on European continental crust, in the foreland to both the Alpine and Pyrenean orogens.

1.1.1. The Alps.

The Western Alps can be considered simply as the telescoped remains of the Mesozoic continental margins of Europe and Adria, with remnants of several basins and microcontinents sandwiched in between. On a large scale, these are stacked so that units originally lying further south and east generally occupy more internal and higher positions in the orogenic wedge.

From foreland to hinterland, the main units present are:

i) **Helvetic - Dauphinois - Ultra-Dauphinois zone.**

This zone represents the basement and cover of the European continental margin. Triassic and Jurassic rifting of Pangea resulted in considerable extension of the continental basement, but none of the basins within this zone progressed to the oceanic spreading stage.

Deep levels of European crust are exposed in the **external** basement massifs: for example, Pelvoux and Argentera. The Gres d'Annot basin developed in the southern sector of this zone. This zone is referred to as the **External Alps.**

ii) Valais - Sub-Brianconnais.

This zone represents a basin near the edge of the European continent in which stretching and subsidence was most extreme. Locally within the Valais sector, a limited amount of oceanic crust was generated.

iii) Brianconnais.

The Brianconnais zone represents a sliver of continental basement which suffered less extension than the Valais Zone and consequently remained a relative high during the Mesozoic. As a result, it is characterised by a thinner Mesozoic succession of shallower marine facies throughout.

iv) Piemontais - Ligurian.

These units are the remnants of a Mesozoic deep marine basin, much of which was oceanic.

v) Southern Alpine zones (for example, Dent Blanche, Sezia).
The continental margin sequences and the basement derived from the southern continent of Adria.

Units ii)-v) are referred to as the **Internal Alps.** In the study region, the internal zone thrust sheets contain Sub-Brianconnais, minor Brianconnais and the Helminthoid flysch, which is usually interpreted to be the upper part of the sedimentary sequence of the Piemontais zone.

The Western Alpine belt is strongly arcuate, swinging through about 180' from Switzerland to Liguria. The study region of Haute Provence lies at one of the sharpest bends, the Castellane Arc, where the Alpine structural trend turns sharply from NNW-SSE and NW-SE to W-E.

The Alpine compressional deformation front propagated into the European foreland for the first time in the Mid Eocene. Since then, the S.W. Alpine foreland has been continually in compression: the convergence of Europe and Adria has been accommodated on structures used repeatedly, in-tandem and out-of-sequence.

1.1.2. The Pyrenees.

The Pyrenees extend from offshore N.W. Spain to Italy, where they are lost in the Apennines! The geographic mountains called the Pyrenees are a very small part of the system: the western sector is still submerged and intra-oceanic; to the east, late rifting and subsidence have modified the orogen; still further east, the Pyrenees have been buried or overprinted by Eo-Alpine events.

The Pyrenees formed during Late Cretaceous-Eocene times, when Iberia, and the assemblage of continental fragments to the east of Spain, collided with Europe. The phases of Mesozoic extension that gave rise to Tethys, and other major oceans further east, produced the Bay of Biscay, but otherwise Iberia and Europe were not greatly separated. Therefore, when extension gave way to compression during the Late Cretaceous, the continental collision was limited in its effect and duration. Even so, inversions in N.W. Europe during the early Tertiary may be related to Pyrenean forces (Roberts, 1987, Inversion Conference communication).

In S.E. France, Pyrenean displacements entered Europe at mid crustal levels. Mesozoic and older faults were reactivated as compressional and strike slip faults. The geometry and intensity of thin-skinned deformation was controlled by the pre-existing fault systems and sedimentary basins. The inversion significantly uplifted large areas of crust, particularly in Haute Provence, but the amount of related shortening was not great.

1.1.3. The Apennines.

The Apennines developed as a result of post-Oligocene plate interactions, which did not involve the European continent. The principal thrust transport direction was towards the north-east. In the Early Oligocene, S.E. France was effectively separated from Apennine related forces by the spreading Ligurian Sea and so did not suffer Apennine related deformation.

1.1.4. The Stratigraphy of the Gres d'Annot basin. (Fig. 1.3).

Remnants of the Eocene Gres d'Annot basin are preserved in isolated outcrop areas in Haute Provence and Les Alpes Maritimes (fig. 1.2). The rather thin Tertiary succession, less than 1500m thick, caps the mountains in these areas.

The Tertiary sediments record the initiation, deepening and subsequent fill of a marine basin (fig. 1.3). The base of the succession is a major unconformity: Eocene strata overlie Cretaceous (Aptian to Maastrichtian) shales or limestones. The Cretaceous sequence is undeformed over much of the study region. Locally, intensely kinked limestones are truncated beneath undeformed Tertiary sediments. Elsewhere, the topography of such fold surfaces was mantled by sediment.

A continental sequence, up to 100m thick, is locally present at the base of the Tertiary sequence: the Poudingues d'Argens Formation. This comprises, for the most part, alluvial fans derived from local palaeohighs of Upper Cretaceous limestone.

The Calcaires Nummulitiques Formation is a laterally extensive marine limestone which records a transgression that was diachronous across the study region (Campredon, 1977). The oldest successions are in the south and east, where the base of the formation is Lutetian; the youngest are in the west and are Priabonian. The transgressive sequence continued through the bioclastic limestones of the Calcaires Nummulitiques Formation transitionally into deeper marine marlstones of the Marnes Bleues Formation. Water depths of 900m have been interpreted from fauna gathered in the upper metres of this formation at Annot (Mougin, 1978).

The top of the Marnes Bleues Formation at each locality is sharply defined by a rapid facies change from marlstone to sandstone turbidites. At many localities, the formation boundary is an angular, non-erosive unconformity, where the Gres d'Annot onlap a bedding surface at the top of the Marnes Bleues. The unconformity is not a time line: the uppermost Marnes Bleues young from the south and east towards the west, reflecting the diachronous arrival of the Gres d'Annot turbidites. The turbidites progressively infilled the significant relief of a moderately deep marine basin from east to west, thus continuing the trend established by the Nummulitic transgression.

The Gres d'Annot Formation is a sandstone-rich turbidite system. The immature quartz-feldspathic sandstones were derived from the south, direct from uplifted basement massifs. No more than 1000m of Gres d'Annot are preserved at any one locality.

The stratigraphic contact at the top of the Gres d'Annot Formation is rarely preserved: it outcrops only on the Col de La Bonnette. Elsewhere, the contact is truncated by the present day land surface or the sole thrust of the Embrun Ubaye (internal Alpine) nappes. At Col de La Bonnette, the upper contact is abrupt, and the Gres d'Annot are overlain by the Schistes a Blocs Formation. The Schistes a Blocs are interpreted to represent a tectono-sedimentary melange of debris shed from the Embrun-Ubaye thrust sheets as they advanced into the Gres d'Annot basin in the early Oligocene.

Much of the Gres d'Annot basin was tectonically buried at this stage. Subsequently it was underthrust, uplifted and partially eroded. Some of the detritus was reworked into the **Oligocene** S.W. Alpine foreland basin north of Digne (Davies, 1986, pers.comm.).

1.2. Brief tectonic history of S.E. France.

1.2.1. Hercynian Events:

The nature of the Hercynian orogeny east of the Massif Central remains largely unknown. However, a strong Hercynian tectonic fabric persists to this day consisting of steep, dominantly NE-SW, and to a lesser extent NW-SE, deep-seated fractures. These have strongly biassed the response of the European crust to subsequent tectonic stresses. For example, both Mesozoic extensional and Pyrenean-Alpine compressional structures were compartmentalised by fractures like the Nimes and Forqualquier faults.

The distribution of Permian sediments is uneven across S.E. France. The thickest succession is exposed in the Barrot Massif and is markedly thinner not far away on the Maures-Esterel massif. It may be implied that the sediment distribution was fault controlled. It has been suggested by Graham (1986, pers. comm.) that the Barrot culmination is the hanging wall of a Permian basin margin fault. The present research has developed this model and shown it to be viable.

1.2.2. Mesozoic Extension.

The distribution of Mesozoic stratigraphy was obtained from the regional review of S.E. France by Debrand-Passard *et al* (1984).

In the Triassic, the major supercontinent of Pangea broke up. Two major episodes of extension, one in the Trias, the other in the Jurassic, gave rise to Tethys and other oceanic basins along the southern margin of Europe.

The extensional fault system that penetrated the European continent in S.E. France gave rise to several Jurassic depocentres elongated NE-SW, parallel to reactivated Hercynian faults. In Provence, the fault bounded Upper Jurassic basins were small, but subsided very rapidly. In some of these of these basins, more than 12km of Mesozoic stratigraphy are preserved. The edge of the European continent does not appear to have been a simple continental margin. The basins are more akin to pull-apart basins formed in a wrench tectonic setting (eg. Crowell, 1974). This model would accord with that of Livermore et al (1986) who propose 900km of dextral strike slip between Africa and Europe in the Triassic!

The Vocontian basin of Haute Provence was a large extensional basin. The lateral variations in thickness of the basin fill, in particular the Lias, strongly suggest that the basin was bounded to the north, south and west by large, listric extensional faults.

In contrast to the NE-SW Jurassic trends, the Lower Cretaceous stratigraphy was controlled by E-W extensional faults.

1.2.3. Europe goes into compression: end Mesozoic to recent.

During the Late Cretaceous, the south of France experienced compression when Iberia and the assemblage of microcontinents to the east collided with the European plate. As the Pyrenean orogen developed, some parts of

S.E. France were uplifted as a result of thrusting and basin inversion and others subsided due to flexural loading.

Pyrenean compression continued through the Eocene. By the Middle Eocene, the Alpine system was well developed: the collision between Adria and the European continental margin had begun and a SW-directed foreland thrust system had developed in Haute Provence. The effects of the Pyrenean and Alpine compression are difficult to separate.

Coincident with the westward propagation of the Alpine deformation front, Haute Provence subsided beneath sea level and the S.W. Alpine foreland basin developed. The evolution of this and , in particular, the Gres d'Annot basin, is the subject of this thesis. It was supplied with sediment from Corsica-Sardinia and other continental fragments to the south. This supply was terminated when Corsica and Sardinia rifted away from the European margin early in the Oligocene. At the same time, the Embrun-Ubaye (internal Alpine) thrust sheets advanced over the region and filled the Gres d'Annot basin. It was simultaneously smothered and starved to death!

The S.W. Alpine foreland basin migrated westwards until it merged with the Rhone graben system, during the Oligocene, and then did not migrate further.

1.3. Recent ideas on the evolution of the Gres d'Annot basin.

1.3.1. Sedimentology and Stratigraphy.

Bodelle (1971) and Campredon (1972) produced excellent theses describing the whole Tertiary sequence in Haute Provence and Les Alpes Maritimes. They studied, in particular, the fauna of the Lower Tertiary Calcaires Nummulitiques and Marnes Bleues Formations. They demonstrated that the whole Tertiary sequence was diachronous, younging westwards. The oldest successions were in the south and east, near the Italian border.

Bouma (1962) developed the concept of the turbidite to describe the Gres d'Annot of Peira Cava. Stanley (1961, etc) set up the turbidite fan model for the Gres d'Annot system, based on studies in the Annot, Col de La Cayolle and Trois Eveches outcrop areas. He later (1975) interpreted the Annot succession to be the fill of a submarine canyon, drawing analogues with the modern canyons of the eastern USA continental margin.

Ivaldi (1974) addressed the problem of provenance for the Gres d'Annot. He used quartz thermoluminescence fingerprinting to conclude that the principal sources were Hercynian basement massifs similar to those of Corsica and Sardinia.

The Institut Francais du Petrole has had a number of students working the Gres d'Annot since 1980. Cremer (1983) made a comparative study of the recent Cap Ferret muddy turbidite fan and the sandstone dominated Gres d'Annot. Ravenne and Beghin (1983) and Cremer (1983) tackled the process sedimentology of the massive, thick sandstone beds and interpreted them as the product of high density turbidity currents derived from slope failures.

Jean (1985) applied and extended the previous research to a detailed study of the Col de La Cayolle outcrop area. She showed that the turbidites infilled a pre-existing trough, but she did not fully consider either the origin of the topography or the response of the sediment gravity flows to it.

Ghibaudo (1985 and research in progress) has carried out a thorough sedimentological study of the Gres d'Annot Formation. It comprises a monumental amount of data gathered by he and his students. He and Jean have mapped conglomerate marker beds within the succession with which they have established a crude internal stratigraphy of the formation. Ghibaudo is in the process of improving this by identifying and matching patterns of sedimentary sequences.

Ghibaudo, Cremer and Jean have all recognised the existence of a local angular, non-erosive unconformity at the base of the Gres d'Annot Formation. The turbidites onlapped pre-existing basin floor slopes, some as steep as 25'. However, none of these workers have looked in detail at what produced the topography or its effect on turbidite deposition.

1.3.2. Structure and Tectonics.

The core of our understanding of the tectonic history of Haute Provence is Goguel's work (1936, 1939). His numerous and detailed structural cross sections established the nature of decollement tectonics in the region. He showed that folds and thrusts detached above Triassic evaporites. He demonstrated that the structural style was dictated primarily by the lithology of each formation.

Fallot (1949) described the Tinee Nappes and interpreted them to be "intracutaneous". In doing so, he drew attention to the existence of another excellent detachment level within the stratigraphy: the Mid Cretaceous shales.

Graham's early publications (1978, 1981) placed an emphasis on gravity sliding and basement wrench tectonic models for the structure of Haute Provence. In recent years, his ideas have changed substantially.

An important turning point in our understanding of Alpine structural geology was the application of knowledge gained in the Rocky Mountains thrust belt after a seminal field trip, attended by D. Elliott. Graham subsequently developed a thin skinned thrust belt model using balanced cross section techniques. This excellent work remains largely unpublished, but his principal cross section is enclosed in this thesis (enclosure1). A small part of the cross section is presented in Vann et al (1986) This cross section implies large Alpine displacements towards the south-west, of 65km.

Graham also established that some cover shortening pre-dated the Eocene transgression. However, he was not able to determine the nature of these early structures, or the amount of shortening involved, using structural data alone. He did not explore the significance of a phase of Pre-Alpine, Pyrenean-Provencal structuring related to north-south compression (Goguel, 1963 and Siddans, 1979). However, E-W structures are recognised in the southern half of the Sub-Alpine chains, overprinted by younger Alpine folding.

Lawson (1987) applied the principals of thin-skinned thrust tectonics to a larger area of Haute Provence, including the Embrun-Ubaye nappes. His thesis was not available during the preparation of this study. However, many discussions in the field and the cross sections he prepared for conferences contributed to the development of the present study. He, like Graham, derived a south-westward thrust transport direction from a statistical analysis of thrust ramp orientations and fold geometries. He showed that structures developed "out-of-sequence" and continued to be active for long periods of time.

1.4. Aims of the project and the research approach.

The ultimate aim of this thesis was to understand the origin and evolution of the Gres d'Annot turbidite basin. A sedimentological and structural study of the basin fill was integrated with a study of the structures developed beneath the basin and of regional structures outside the basin. The problems that were addressed in the process were:

i) The origin of the Castellane Arc. It was important to establish the relative importance of the Pyrenean and Alpine compressional systems for two reasons:

a) The Gres d'Annot turbidites were sourced from the south, perhaps from basement uplifted above Pyrenean structures. However, the Gres d'Annot basin and laterally equivalent deep marine basins formed a more-or-less continuous trough running parallel to the Alpine chain. Therefore, the controls of basin subsidence appeared to relate to the Alpine orogeny.

b) The direction of shortening on structures in Haute Provence was unlikely to have remained constant throughout the development of the Gres d'Annot basin.

Through a detailed understanding of the development of individual structures and the regional teconic history, it was possible to identfy a mechanism for subsidence of the Gres d'Annot basin. It became apparent why a deep marine basin had developed on the European foreland and what caused its sudden death.

ii) The nature and significance of the unconformities at the base of the Tertiary succession and the base of the Gres d'Annot Formation.

These unconformities were used to establish when individual structures developed. By constructing partially restored cross sections, it was possible to determine the tectonic setting of the basin at stages in its development. The geometry of the basin floor topography was defined immediately prior to, and during, the Tertiary transgression and during the deposition of the Gres d'Annot turbidites.

iii) The relationship between turbidite deposition and intra-basinal slopes. The turbidite facies were examined in detail adjacent to the onlap unconformity surface at the base of the Gres d'Annot Formation. A new model for topographic control of turbidite deposition was constructed.

It was necessary to establish to what extent the basin floor topography subdivided the basin into isolated depocentres: do the present day outcrop areas represent the fill of separate depocentres or the eroded remnants of a single turbidite basin?

iv) The level of control exerted by the pre-existing Mesozoic, and older, extensional fault systems. How much did the existing tectonic fabric influence the development of subsequent compressional structures and control the geometry and depth of the Tertiary sedimentary basins?

The principal results of the study are presented in the abstract and conclusions.

1.5. Thesis structure.

Chapter 2 is a detailed analysis of the tectonic evolution of the study region and S.E. France. The development of individual structures is traced through a description of a dense grid of structural cross sections. The origin of the basin floor topography is discussed.

Chapters 3, 4 and 5 describe the Poudingues d'Argens, Calcaires Nummulitiques and Gres d'Annot Formations. The relationship between the sediment facies and the developing basin floor topographic relief is stressed.

The data and interpretations of the preceeding chapters are integrated in Chapter 6. A description of the evolution of the basin floor topography is presented, and the structures responsble for the palaeohighs and lows at each stage are identified. The evolution of the Gres d'Annot basin is discussed in the context of the tectonic history of both the Pyrenean and Alpine orogenic systems.

CHAPTER 2. STRUCTURE AND TECTONIC EVOLUTION OF SOUTH-EAST FRANCE

2.1 Introduction to the chapter

The structural configuration of South-East France is the key to our understanding of Tertiary sedimentary basin evolution in the region. In return, the history of sedimentation within these basins places constraints on the tectonic models. For example, the timing and amount of basement structuring has a strong influence on sedimentary basin dynamics. In the study region, the level of erosion is such that the nature and geometry of Pre-Triassic basement structures are difficult to define. Provenance studies of the Tertiary sediments are used to infer the timing of crystalline basement uplift.

The Eocene-earliest Oligocene Gres d'Annot basin is of particular interest. The results of this research demonstrate that it formed, evolved and was destroyed at a time when the region experienced the compressive forces of both the Alpine and Pyrenean orogens, directed approximately E-W and N-S respectively.

To date, tectonic models for Haute Provence have separated the effects of Pyrenean and Alpine deformation into time windows of Late Cretaceous-Eocene and Oligocene-Recent respectively. The present study examined the relationship of Eocene sediment facies to underlying structures. Already by the Middle Eocene, structures had developed which were induced by a component of east-west, Alpine, compression. The Alpine deformation front progressed into the European foreland earlier than previously supposed, so that during the Eocene, both Pyrenean and Alpine forces operated on S.E. France. Haute Provence was effectively caught in a vice, with Adria advancing westwards and the lateral equivalents of Iberia moving northwards to collide with the European continent.

In the past, all E-W striking structures were thought to be Pyrenean, induced by regional north-south compression (for example, Siddans, 1979). However, E-W structural trends were probably in part produced when Alpine thrust movement was impeded (section 2.4). It is therefore difficult to assess the relative importance of Pyrenean and Alpine orogenic forces in the generation of the Gres d'Annot turbidite basin and the present-day structures. Certain features of the basin itself, for example, the active source area to the south and an apparent component of basement dip towards the south (section D, enclosure 7), suggest the importance of regional Pyrenean forces. The Alpine system, however, dictated the dimensions of the basin and much of the internal topography.

This study concentrated on defining the effect of Alpine stresses and, in doing so, addressed the associated structural curiosities, like the origin of rapid changes in fault and fold vergence around the Castellane Arc and the origin of SW-directed Alpine thrust transport.

2.1.1. Previous Research

Goguel (1962) identified the presence of major decollements, structural discontinuities, within the stratigraphy which characterise the tectonic style of the S.W. Alps. The Mesozoic and Tertiary cover slipped over the Permian and crystalline basement; much of the cover shortening pre-dated the uplift of the Dome de Barrot and Argentera massifs.

Graham (1985 and pers.comm.) defined the present day structural configuration of Haute Provence through a traverse from St.Etienne de Tinee to Chateau Redon and onto the Valensole plateau (enclosure 1). He was the first to construct structural cross sections down to the Triassic decollement which rigourously obeyed the rules of section balancing. The surface data was accurately recorded, using field and published map data. The assumptions he made to construct the section were:

1. All the present day structures were ascribed to an Alpine thrust system, with a single thrust transport direction, towards the south-west.

2. The Mesozoic stratigraphy was not layer-cake, but changed thickness gradually across the region. He did not identify Mesozoic or older faults in the system which may have influenced the Alpine tectonic style.

3. Menard's map (1980) of the depth to Pre-Triassic basement defined the sole thrust of the Mesozoic and Tertiary cover thrust system. Structures beneath this level do not breach this decollement in the study region.

He concluded that the cover sequence was shortened by 65km along the constructed section (50% shortening). He used the unconformity at the base of the Tertiary succession to conclude that part of this deformation pre-dated the formation of the Gres d'Annot basin. Using the structural geometries alone, he was not able to define the nature of this early deformation or the amount of shortening it represented.

Lawson (1987) used similar structural techniques and essentially the same assumptions to analyse the relationship between the deformed European foreland of Haute Provence and the internal thrust sheets of the Embrun-Ubaye nappes. He showed that the deformation sequence was complicated and that deformation did not prograde simply from the hinterland into the foreland. However, he did not date the structures relative to stratigraphic data and assumed that they all related to a SW-directed thrust system.

Graziansky (1982) represents the opposite end of the spectrum of geological opinion and does not accept Graham's estimate of cover shortening. His maps and sections of the Digne area imply that the principal shortening direction was north-south. He concluded that movement on the north-south trending faults was principally strike slip, since they were lateral structures with respect to his interpretation of the shortening direction. Note that most of these faults are shallow-dipping structures which Graham interpreted as west and south-west directed thrusts.

Graziansky and other French authors stress the importance of "fundamental", steep, ancient faults within the Hercynian basement. These are interpreted to have been reactivated during the Mesozoic rifting and Tertiary compressional deformation of the European margin. The Durance valley flows along the strike of one such fault.

2.1.2. Aims of the present study

The aim of the study was to develop a coherent tectonic model for the region that incorporated the strong evidence for both south-westward and southward directed shortening of the Mesozoic and Tertiary cover rocks.

The age of the orogenic structures, both Alpine and Pyrenean, was determined relative to the Eocene sediments preserved in Haute Provence.

An assessment was made of the influence of these structures on the geometry of the Gres d'Annot basin margins and topography within it, at stages throughout its development. A tectonic model for the origin and evolution of the Gres d'Annot basin was developed through a better understanding of the history of activity on all manner of structures in the region. The range of existing strike slip, extensional and compressional tectonic models are economical with the evidence and do not acknowledge the complexities that arise when two orogenic systems interact.

2.1.3. Working methods used in this study

2.1.3.1. Principles.

To develop a tectonic model for this region, it is necessary to establish the sequence of deformation and the changes in shortening direction that have occurred since the foreland went into compression. It is important to ascertain when individual structures were initiated and then to bracket the periods of activity of each fault and fold.

In the study region, unconformities at the base of the Cenomanian, base of the Tertiary and base of the Gres d'Annot Formation define four depositional sequences related to phases of compressional tectonics. The angular relationships across these unconformities, coupled with the distributions of each of the five Tertiary formations, provide information about the palaeotopography and palaeostructure at these times: the palaeotopography can be shown to relate directly to underlying tectonic structures (see discussion of the cross sections, section 2.3.2.).

Therefore, the stratigraphic and sedimentological information may be used to test and constrain the interpreted geometry and lateral continuity of structures in all cross sections.

The base of the Tertiary succession in the study region has been uplifted to heights between 500m and 2600m. Over large areas, the structural dips are slight, less than 10'. The elevation of particular horizons changes rapidly across discrete fault planes and 1km-scale monoclines. The resulting pattern of restricted areas of steeply dipping strata is interpreted to reflect the ramp-flat topography of underlying fault planes. There are also zones of intense folding where horizons are locally elevated by structural thickening of the underlying sequence, but which do not separate zones of different elevation. These are interpreted to form where thrust movement is inhibited. The Chateau Garnier zone is an example (cross section A1, enclosure 3).

2.1.3.2. How the cross sections were constructed.

A grid of structural cross sections was constructed with a section interval of 10km or less (enclosure 0). The sources of data were:

1. Field observations: changes in level of key stratigraphic horizons were mapped, and the nature and origin of folds were examined in detail.

2. Published geological maps (BRGM, listed in the references; 1:50,000 and 1:250,000, in particular).

3. A contour map of the depth to the pre-Triassic basement, constructed by Menard (1980). This map was used with reservations which are discussed

below (section 2.1.3.3)

The data used to determine present-day structure to depth are the dips and elevation of the base-Upper Cretaceous, base-Tertiary and base-Gres d'Annot unconformities and the orientation of Gres d'Annot strata, which defines structuring that post dates deposition in the basin. The base Tertiary unconformity is the best horizon to map: stratigraphic levels above it were more extensively eroded, levels below it are commonly buried. It is easily recognised and well mapped by the French. The published maps were studied to identify the lowest structural level of the base Tertiary in the region.

Establishing the "regional level".

The lowest level of the base Gres d'Annot unconformity is around Annot, approximately 600m above sea level (cross section Y, enclosure 13). The Gres d'Annot turbidites were deposited in approximately 900m of water. They have been uplifted 1.5-4km subsequently. The total uplift could be caused by:

1. Structural thickening by thrusts.

2. Isostatic uplift in response to eroding the Alpine and Pyrenean orogenic load.

3. Eustatic sea level fall.

4. Lithospheric heating. Note the existence of Oligocene volcanics, including lavas, in the St. Antonin outcrop area.

The isostatic and eustatic effects are on a scale greater than 200km and can only produce regional tilts, not the <10km scale uplifts which have been used to define the subsurface tectonic structures. The regional level of the base Tertiary at Annot on undeformed crust is assumed to have been 500m, with a slope up to the north and west. Positive deviations above this regional surface record structural thickening.

This figure assumes that no extensional faults penetrate the basement which would locally lower the base-Tertiary unconformity. The Rouaine fault zone, for example, is extensional with respect to stratigraphy. It is believed to sole out at the base of the Permian succession (section 2.6, fig. 2.39). There is no evidence that this structure is a deep-seated "fundamental" fault. It is interpreted as a lateral ramp, accommodating differential displacements above a major thrust. Therefore, it is not a rift structure: the stratigraphy on either side must be at, or above, the "regional".

Timing of deformation.

The stratigraphic and sedimentological data were used to define palaeo-dip and relative elevation of strata at intermediate times during the early Tertiary. The technique requires that the structural cross sections be sequentially restored. Early Tertiary tectonic activity can be defined at three times:

1. Prior to Tertiary sedimentation (Late Cretaceous-Early Tertiary).

The amount of relative uplift was derived from the thickness of Upper

Cretaceous strata preserved beneath the base-Tertiary unconformity. The thickness varies from 30m at Colmars to 800m below Pierre Gross. Most of the differences are due to erosion of palaeohighs; some of the variations may be depositional. In all cases, they are attributed to early compressional structures (fig. 2.1).

The unconformity is assumed to represent a sub-horizontal peneplain. When the top of the Upper Cretaceous is restored to horizontal, the base of the Upper Cretaceous defines the paleo-structural dip. Thus, the position of palaeo-fault ramps is defined.

2. Activity contemporaneous with the earliest Tertiary sediments.

The palaeoslopes were defined using the facies distribution and thickness changes in the Poudingues d'Argens, Calcaires Nummulitiques and Marnes Bleues Formations (fig. 2.2 and sections 3.5.4, 3.6, 4.4, 4.5, 4.6).

3. Activity prior to deposition of the Gres d'Annot Formation.

The base of the Gres d'Annot Formation is an unconformity, locally an angular one (sections 5.2.1 and 5.5). When the onlapping turbidites are restored to horizontal, the dip of the Marnes Bleues defines the position of palaeo-thrust ramps (fig. 2.25).

The technique has made it possible to distinguish footwall and hanging wall ramps and has shown how structures evolved with time. In particular, the relationship between the Tertiary sediment facies and structurally induced palaeotopography was used to define the geometry of lateral thrust ramps. For example, if an oblique ramp lies below the level of exposure, structure alone does not reveal whether it is a footwall or hanging wall ramp. However, a footwall ramp moves with respect to the sediments above it, a hanging wall ramp does not.

2.1.3.3. Discussion of the assumptions inherent in this and previous interpretations.

The principal assumptions made by this author are:

1. The principal decollement horizons accommodated large amounts of shortening. Therefore, not all the major uplift can be traced down to basement structures, like deep-seated, steep, wrench faults.

2. Stratigraphic thicknesses change gradually or across simple growth faults. Significant parts of each cross section are filled with duplicated Mesozoic stratigraphy. Large variations in stratigraphic thickness are interpreted as the result of structural thickening and not as depositional becke and schwellen.

3. Large-scale basement slopes may be derived from Menard's (1980) map of the depth to Pre-Triassic basement. It was not considered accurate enough to determine the true thickness of deformed Mesozoic strata.

Menard's map covers the whole of S.E.France and, unlike his later work (Menard, 1984), the data do not include gravity anomalies. The data are seismic refraction (and limited reflection data) profiles and borehole depths. The data points were not defined, but no reflection seismic profiles have been run or boreholes drilled in the study region. It is probable that the depth contours were extrapolated from surface structure east of the Digne thrust. Therefore, the map was not used to construct individual structures in the study region.

Certain assumptions made by other authors are rejected or only accepted with reservation:

1. Both Graham (1985) and Lawson (1987) assumed a consistent southwestward directed shortening. They did not recognise that many of the faults were probably initiated as southward directed thrusts. As a result, they overestimated the SW displacement on these faults. The converse is true of Graciansky (1982) Tempier (1979) and Hamiti (1987, pers. comm.), who do not recognise a significant westward component of displacement on thrusts. In this thesis, the importance of both thrust transport directions is acknowledged.

2. The Triassic decollement was assumed to be effective across the study region (Graham, 1985). However, in parts of Haute Provence major displacements occurred on thrusts below the Triassic, for example, the Barrot thrust (section 2.6). Folding and tilting over the Permian duplex impeded movement on the Triassic decollement. There has been no significant displacement on that detachment in the hanging wall of the Barrot thrust since the mid Eocene. It is possible that the Triassic decollement is breached. Certainly, in Provence, basement thrusts breach this detachment: Hercynian granites are emplaced on Jurassic north of Marseilles. In addition, the decollement may have been breached by Mesozoic extensional faults.

When all is said and done, there remain several problems which cannot be resolved unambiguously. These are:

i) There are four effective detachment levels: base Permian, mid Triassic, mid Jurassic and mid Cretaceous. Consequently, a culmination seen at surface could be developed over a ramp at several alternative levels (fig. 2.3).

ii) A given basement uplift may be produced by a large displacement, low-angle fault or by a steep fault with less displacement. There is limited information available on basement fault dips.

iii) The structure in the Mesozoic stratigraphy prior to compression can be defined in general terms. However the details of the structure are not known; a restored section cannot be constructed with certainty. In particular there may have been more topography on the mid-Triassic unconformity than shown on the sections.

iv) Widespread out-of-sequence movement occurred on structures at all levels. Therefore, simplistic rules of thrust sequences cannot be applied.

2.2. The structural styles exhibited in Haute Provence and the Maritime Alps

2.2.1. Structural Domains

Encl. 15+16

The principal structures in the region are shown on the maps of fig. 2.18. The principal structural domains are:

1. The external crystalline basement massifs of Pelvoux, Argentera and Maures Esterel: these comprise European continental crust.
2. The Castellane Arc, defined by an abrupt change in the strike of thrust faults and fold axes: from N-S or NW-SE, to E-W adjacent to the Maures-Esterel massif.
3. Provence, where large areas of relatively undeformed Mesozoic and Tertiary strata are bounded by relatively narrow "mobile zones", defined by E-W striking thrusts or NE-SW strike slip faults.

In this study, some attention is paid to the origin of the E-W structural trend seen particularly in the southern part of the Castellane Arc. The trend was established in the Late Cretaceous, and the structures were active throughout the Eocene into the earliest Oligocene. Their interpretation has a bearing on the tectonic setting of the Upper Eocene Gres d'Annot turbidite basin. The results of this study indicate that not all the E-W structuring was induced by northward directed Pyrenean compression (section 2.4). Alpine structures were initiated in the European foreland in the mid Eocene, prior to, and syn, the development of the sedimentary basin.

The effects of the Pyrenean orogeny were not felt in S.E.France after the Sanoisian. Alpine deformation continued and intensified through the

Miocene, such that the early Tertiary E-W structures were tightened or overprinted. To establish the nature of the early structures, it is helpful to go to Provence where the Alpine overprinting is less severe and subsurface data are available (boreholes and reflection seismic). Here, there is the opportunity to study Pyrenean structures and the control of compressonal structures by a pre-existing Mesozoic extensional fault system (section 2.5).

2.2.2. Mechanical properties of the stratigraphy.

Fig. 2.4 illustrates the response of each lithology to Tertiary compression. The structural styles depend on lithology and are therefore reviewed in terms of each formation.

2.2.2.1. Decollements.

i) Evaporites within the Triassic: Keuper gypsum and cargneules and possibly halite at depth, plus to a limited extent, argillaceous facies in both the Keuper and Rhaetian. Cargneule (Carniola, in Italy) describes a lithofacies: matrix supported, holey breccias, comprising for the most part Triassic dolomite fragments in a calcareous/gypsiferous matrix/cement. It appears to be an extremely effective fault lubricant.

ii) Terres Noires (Callovian, Oxfordian): thick, black-grey shales, 300m thick in the east to 700m near Barles.

iii) Aptian-Albian shales: These form the highest decollement in the stratigraphy which was the effective roof thrust of the Haute Provence duplex. To the east of Barreme, it is rarely breached by thrust faults.

2.2.2.2. The Remaining Jurassic-Lower Cretaceous stratigraphy

Stratigraphic levels below the Tithonian limestone are rarely exposed in the study area. However, the **Lower Jurassic** succession comprises well bedded alternations of limestone/marlstone/(shale), similar to the Lower Cretaceous lithologies. This and the construction of the structural cross sections imply that they deform in a similar manner: shortening was accommodated primarily on discrete thrusts.

The **Tithonian limestone (Upper Jurassic)** is a relatively thin (70-150m thick), massive limestone. It is mechanically competent and has thick, incompetent bedded shales above and below it and so, in general, shortens by folding rather than thrusting. Spectacular kink folds are observed in the Entraunes valley (fig. 2.5): the geometry of these large scale folds is illustrated in cross section X. These accommodate approximately 50% shortening. They may be interpreted as tip, or sticking folds, formed behind an obstacle to slip on the Triassic decollement. Such an obstacle would be produced by folding the decollement over a basement culmination (fig. 2.6).

The well-bedded limestones and marlstones of the **Upper Cretaceous** are undeformed over large areas of the study region: they were largely transported passively above a thrust duplex between the Keuper and Aptian decollements: the Haute Provence duplex of Graham (1985). Thrusts of kilometre-scale displacement cut sharply through the succession in places, for example, below Mourre Gros in the Trois Eveches outcrop area (cross sections D and B). Elsewhere, shortening is accommodated by kink folding (fig. 2.7, 8), small scale thrusts in limestone beds (fig. 2.9a,b) and decimetre-scale shear zones in shale intervals.

The kink folds range from 1-50m wavelength. Fig. 2.10 and cross section A1 illustrate the range of fold geometry: they form conjugate sets, but one set is commonly dominant (S.W. vergent). Single kink bands are never seen: they live in packs, forming zones of approximately 1 kilometre width.

In some cases, the zones mark changes in structural level and relate to underlying structures, particularly hanging wall ramps. Examples of this are the Melina-Aurent and Ravin de la Frache kink zones (fig. 2.11 and cross section A1 and B).

Other kink zones appear to have developed independently, representing only a zone of shortening by folding. In these latter cases, the uplift is restricted to the buckled area. An example of this is the Chateau Garnier-Argens fold, modified only slightly by a late thrust (cross sections A0, B1 and Enclosure 2.2). These are interpreted to form where the Mid Cretaceous decollement ceased to be effective. Where it stuck, shortening was transferred into the Upper Cretaceous, firstly as kink bands. This style of deformation was probably energetically favoured as a result of the high anisotropy and constant bed thickness of the Upper Cretaceous.

2.2.2.3. Cretaceous-Tertiary boundary and deformation of the Calcaires Nummulitiques Formation.

This is not a major tectonic boundary, but locally where the Calcaires Nummulitiques overlie kinked Upper Cretaceous limestones, there is a distinct structural discontinuity.

The Calcaires Nummulitiques Formation comprises a single massive, competent limestone 10-70m thick; the Upper Cretaceous limestone beds are 5-20cm thick and separated by centimetre scale intervals of marlstone. Interbed slip was thus facilitated in the Upper Cretaceous and flexural slip folds were common as a result. Kink bands of a few metres wavelength could not propagate through the Calcaires Nummulitiques Formation: normal faults in the Calcaires Nummulitiques accommodated the uplift across the kink zone (fig. 2.12). It is not so obvious how the limestone took up the shortening, but in areas where the Upper Cretaceous is severely kinked, the Calcaires Nummulitiques are extensively recrystallised and a spaced pressure solution cleavage may be observed (fig. 2.12b). Small, metre-scale thrust and reverse faults are seen. The central portions of these kink zones are always eroded to levels below the Tertiary boundary, so it is probable that larger scale thrusts did exist and were eroded. However, it is feasible that shortening was almost all accommodated by volume loss by pressure solution.

100m-kilometre scale thrusts are observed to cut the Calcaires Nummulitiques Formation, but only where thrusts have cut straight upwards through unfolded Upper Cretaceous limestones. The best example is a duplex of Calcaires Nummulitiques at the Tete de l'Adrech, in the Trois Eveches outcrop area (fig. 5.27; section 5.5.2; cross sections D,D').

In this part of the Trois Eveches outcrop area, bedding surfaces in the Calcaires Nummulitiques are exposed: they reveal a dense network of tension gash arrays, whose orientations are compatible with regional N-S and NE-SW shortening (fig. 2.13).

2.2.1.4. Marnes Bleues

These monotonous Teriary marlstones are mechanically homogeneous and isotropic. Deformation was accommodated for the most part on discrete faults and narrow shear zones, characterised by penetrative cleavage, calcite veining and 1 cm-scale kink folds. Fig. 2.14 shows one of a set of 1m wide shear zones which accommodate deformation over a lateral ramp at the southern tip of the Trois Eveches outcrop area. This competent behaviour was almost certainly made possible by early, thorough calcite cementation of the sediment. This appears to be the case at Chalufy where a mass of Marnes Bleues is onlapped by Gres d'Annot turbidites: the bedding parallel, 30' topographic slope was stable for a considerable length of time, implying that the marl was already lithified.

A bedding parallel, penetrative cleavage is developed in the Marnes Bleues of most outcrop areas, except Annot. Its intensity increases to the north and is most intense in the Col de la Cayolle outcrop area. In part, it is developed very early, because it was kinked in the hanging wall of the Chalufy slide/thrust that pre-dates deposition of the Gres d'Annot turbidites. It may also result from sediment overburden, but the Gres d'Annot were probably never much thicker than 1000m. The pattern of intensity corresponds with the outcrop of the Embrun-Ubaye thrust sheets. Therefore, the very intense cleavage probably defines the areas which were overlain by these thrusts.

Discrete thrusts are often very hard to detect: examples can be studied between Mourre Simanche and Col de la Vachiere, in the Trois Eveches outcrop area. Thin (<1cm thick) calcite veins mark the fault zone. An intense SW vergent cleavage developed in proximity to faults.

A sub-vertical NW-SE striking cleavage is present in the Marnes Bleues beneath the steeply onlapping Gres d'Annot at Le Ruch. Its orientation is consistent with SW vergent kinks and thrusts in the area.There are two models for this, summarised in fig. 2.15:

i) The marlstone topography is the surface relief at the compressional front of a sedimentary slide.

ii) The slope is the steep limb of a 10 metre-scale, S/SE-vergent kink, one of many that construct the Aurent-Melina monocline.

Uplift across the Aurent kink zone began prior to deposition of the Tertiary sequence; the site of maximum uplift is <1km from the onlap outcrop. There would be little room to develop a large sedimentary slide. Therefore, the second model is favoured, but slope failure may have contributed to the formation of the cleavage at the base of the structural slope.

2.2.1.5. Gres d'Annot

Discrete faults, even at centimetre scale, show that this turbidite succession generally deformed in a brittle manner. At Denjuan (Trois Eveches outcrop area), thrust structures were active during deposition of the turbidites (fig. 5.27, section 5.4.4.1.). A single fault is traced to within metres of the palaeo-sea bed. Near the sediment surface, ptygmatic folds were formed, with rounded, unfractured hinges (fig. 2.16). These are interpreted to have developed in unconsolidated sediment, by interparticulate flow. The thrust and folds at Denjuan imply a considerable amount of displacement occurred before the next turbidite infilled the topography: approximately 50% shortening is observed in this lateral/oblique section through the structure.

Similar style, but much larger, 10-50m wavelength folds are observed at Les Gastres (Annot outcrop area; section 5.6.1.4.; Enclosure 5.8) and on the summit of Sangraure (Trois Eveches outcrop area; cross section D). These too are interpreted to have formed in unconsolidated sediment.

2.2.3. The boundary conditions

The SW Alpine foreland was deformed in response to Pyrenean and Alpine compressional forces. The whole crust was compressed: the Mesozoic and Tertiary cover sequence were shortened by at least 30%; the elevation of the Pre-Triassic basement implies that there has been only a limited amount of basement shortening at any great depth in the crust. Therefore, the majority of the deformation was thin skinned. It took the form of:

i) south-vergent structures, backthrusting of the cover in response to a force exerted from the south towards the north (Pyrenean).

ii) south-west and west vergent structures, foreland propagating and shortening with respect to the Alpine forces.

The deformation enters Haute Provence at several levels in the crust, each marking a boundary between different structural styles:

i) Intra-Mesozoic horizons, particularly Mid-Cretaceous shales.

ii) On the principal detachment level in the Triassic evaporites near the base of the Mesozoic succession: most shortening occurs above this.

iii) <5km into the Pre-Triassic basement, probably at the base of the

Permian sedimentary sequence.

iv) Mid crustal level. This was responsible for the inversion of deep-seated Mesozoic-age and older extensional and wrench faults. It is the level at which Pyrenean and Alpine displacements may be transferred under north-west Europe, surfacing as inversion structures such as the Purbeck Disturbance and Hog's Back in South England.

2.2.4. Resulting style of structures

The foreland structures are predominantly brittle. The Mesozoic and Tertiary successions are shortened in a thrust system, with multiple decollements, developed in the mechanically weak Mesozoic shales and evaporites. The faults ramp towards the surface in the more competent limestones, marlstones and sandstones. Bedding parallel slip is facilitated throughout the stratigraphy by interbedded shales. Flexural slip folds and kinks developed, particularly in the fault ramp zones. Axial planar cleavage is rare.

The structural style of Haute Provence is composite. Many of the complexities arise from its position in the foreland to both the Pyrenean and Alpine orogens. In consequence it is not easy to establish the shortening directions. Graham defined a single direction of shortening using:

i) The bow and arrow rule (Elliott, 1976). The "arrow" of the bow, defined by the Castellane Arc, points to the SW.

ii) Steep NE-SW striking faults in the region are primarily strike slip faults, evidenced by slickensided surfaces and small scale shear zones in and near the fault zone. They may be interpreted as lateral tear faults in a south-west directed thrust system.

Lawson (1987) also argued for a SW shortening direction, using evidence from the northern part of the Castellane Arc. He plotted the orientation of all mappable thrust ramps, and the data supported a SW shortening direction.

However, in the southern part of the Castellane Arc, structures are consistently south-vergent, and the thrust planes and fold axes strike east-west (fig. 2.17). Faults like the Castellard thrust (fig. 2.32) may be traced from east-west striking thrusts of constant and large stratigraphic separation, into NE dipping fault planes with rapidly decreasing stratigraphic separation from south to north. The ramp displacement pattern is consistent with a dominantly southward shortening direction, plus a minor amount of late west or southwest displacement, on what were initially lateral ramps (section 2.3.2.12). This is the basis for Graciansky's interpretation (1982) of dominant southward directed shortening in Haute Provence.

It is more realistic to accept that the crust was shortened in at least two directions. In the Barreme area, several SW displacement thrusts tip into east-west striking, south-vergent folds. These are unconformably overlain by Calcaires Nummulitiques limestones and are therefore Eocene or older in age. East-west oriented structures were invariably initiated prior to deposition of Eocene sediments, whatever their later history. The SW-vergent structures are both Eocene and younger.

The following mechanisms combined to produce the present-day structure of

Haute Provence:

1. A Late Cretaceous - Eocene, south directed thrust and fold system.

2. S.W. directed Middle Eocene - Recent thrust system, comprising:

i) a duplex of intra-Mesozoic thrusts, bounded by the Triassic and Mid Cretaceous decollements, to the east of Thorame Basse;

ii) an imbricate stack to the west, between Barreme and Digne.

There is a corresponding change in the length of individual thrust sheets: in the imbricate stack to the west of Barreme and south of Annot, each thrust horse is short. This and the breaching of the Mid Cretaçeous roof thrust to the west of Thorame Basse are interpreted to reflect a decrease in efficiency of the decollement levels, in particular the Mid Cretaceous shales. Sequential restoration of cross sections imply that this horizon was folded, to the west of Thorame Basse, during the early Tertiary phase of activity on south and south-west/west vergent structures. Later thrust displacements could not be transferred through the uneven topography of the Mid Cretaceous shales, so the thrusts ramped to surface.

3. Folds, both ramp related and buckle folds.

4. Extensional and strike slip faults.

Many faults active during the Tertiary have extensional and lateral displacements with respect to stratigraphy. They are interpreted to sole out on the major detachments within, and at the base of, the Mesozoic or at the base Permian.

Some were produced when a late thrust sliced, out-of-sequence, through already folded stratigraphy. The remaining truly extensional faults fall into two catagories: i) culmination collapse structures; ii) those formed to accommodate uplift. Faults with significant components of lateral slip, like the Rouaine fault zone, are interpreted as lateral and oblique tear faults of the thrust systems (section 2.6). There is no evidence for crustal extension or wrenching anywhere in Haute Provence during the Tertiary.

5. Change of movement direction on some thrusts.

In Graham's section (enclosure 1) and cross section X (enclosure 12), the thrust that surfaces at Col du Defend appears to have a large amount of SW displacement. The same fault appears on cross sections to the south (B, E and Y, enclosures 5, 9 and 13). However, the maps (fig. 2.32) show that the amount of apparent SW shortening diminishes rapidly to the north, and the fault appears to tip near La Javie. Traced to the south, the fault at surface turns through almost 90° to strike E-W. Here, the apparent southward displacement is large along the whole of its length. Therefore, this fault is interpreted to have initiated as a south vergent structure: the NNW-SSE oriented ramp was a lateral ramp. A limited amount of late, out-of-sequence displacement towards the southwest produced the present structural geometry. The model is discussed in section 2.3.2.12.

6. Syn-sedimentary tectonic activity.

Evidence of syn-depositional earthquakes supports the hypothesis that the Tertiary sediments were deposited while the basin was compressed. Tectonic shock had a variety of effects:

- Wet sediment rapidly dewatered, for example in the Gres d'Annot at Annot (section 5.4.4).
- Unconsolidated sediment flowed into fractures in a brittle substrate. Bioclastic limestones of the Calcaires Nummulitiques Formation fill tensional cracks in deformed Upper Cretaceous limestones (section 4.5.2).
- Sediment failed down over-steepened slopes. The resulting structures include small-scale brittle faults; sedimentary slides, slumps and olistostromes, as seen at the base of the Calcaires Nummulitiques Formation at Barreme and Cairas (section 4.5.2) and at the top of the Marnes Bleues Formation at Braux (section 5.5.5); small-large scale folds, for example, at Denjuan (fig. 5.27) and Les Gastres (encl. 5.6), in the Gres d'Annot. The folds were produced by the process of inter-particulate flow, as a result of deforming unconsolidated, wet sandstone. These folds are the only "ductile" tectonic structures observed in the region.

7. Out-of-sequence activity on almost all the structures in the study region.

Faults and folds are shown to have initiated prior to and during Tertiary sedimentation. The same folds are tightened at a later stage and some were cut by late thrusts which came along the same detachment, for example at Chateau Garnier (cross section A1, enclosure 3) and Argens (cross section B and B1, enclosure 5 and 6). These thrusts have significant d~~s~~isplacement postdating all Tertiary sedimentation.

8. Reactivation of ancient, deep-seated faults and basin inversion.

Outside the study region, in Provence, there is good evidence that Hercynian basement uplift was achieved by inversion of steep deep seated faults. Inherited Hercynian and Mesozoic faults were reactivated in compression: those oriented NE-SW became strike slip faults which compartmentalised the compressional structures. For example, the NE-SW Forcalquier fault is the eastern boundary of the Provencal structural domain.

Within Haute Provence, the basement structures are rarely seen at surface and there is no seismic coverage of the region, so the inversion hypothesis is less easy to test. There are three lines of evidence for inversion in Haute Provence:

i) A very large part of the northern sector of the study region was uplifted and eroded prior to deposition of the Tertiary sediments. The boundary of the uplift runs east-west along the Beauvezer-Thorame Haute valleys. It is sharp and narrow. The narrow ramp, the extent and uniformity of the uplifted area suggest that the compressional structure responsible was a steep and deep seated fault, probably an inverted extensional fault (see, for example, cross section Z, enclosure 14; and section 2.3.2.1).

ii) The sedimentary facies changes that define the margins of the Upper Jurassic Vocontian basin are preserved in the thrust sheets around the Castellane Arc of Haute Provence.

The platform limestone facies and the basinal interbedded limestones and marlstones of the Upper Jurassic are separated by a narrow zone of slope facies. It is probable that the Upper Jurassic basin margin, and hence the sediment facies and thickness, were controlled by extensional faults. The lateral transition between the thin, shelf facies and thick, basinal facies is marked by a series of anticlines and thrusts that were deeply eroded prior to the deposition of the Calcaires Nummulitiques. These are well illustrated on the 1:250,000 Nice and Gap geological maps (BRGM, 1979 and 1980, respectively), south of the St. Antonin outcrop area and around the Barreme syncline.

These anticlines may be interpreted as the result of inverting the Mesozoic basin margin in response to Pyrenean and early Alpine compression (compare with the erosion of the Hogs Back anticline, South England).

iii) The distribution of Permian sediments also suggests fault controlled subsidence. This generation of faults probably controlled the geometry of basement uplifts. In particular, the Barrot massif may be modelled as a Permian basin inversion (section 2.6). Inversion of a steep extensional fault would account for the steep structural dips of the hanging wall ramp.

Fig. 2.18 summarises all the features described in the preceding section.

2.3. Northern Sector of the Castellane Arc.

2.3.1. Introduction

The sector (fig. 2.19) is bounded to the south by the Rouaine fault zone, to the west by the Digne thrust and to the north by the basement massif of Pelvoux. It describes the crust and cover sequence of the European foreland and does not include the Embrun-Ubaye thrust sheets. For details of structures in the latter, the reader is referred to Lawson (1987) and Kerckhove (1969).

This sector features mostly SW/W vergent structures, plus faults believed to have significant southward displacement accommodated on them. Note here that thrusts consistently ramp up through the stratigraphy towards the south, probably because many were initiated as southward directed "Pyrenean" structures and others branched onto pre-existing Pyrenean structures (see discussion, section 2.4). Lateral ramps to the SW and W directed thrust system, with associated shear zones and strike slip faults are very important structural elements. Extensional faults are minor components of the system, accommodating uplift and gravity collapse of compressional structures.

This sector has received the most attention of the three in this study. The dense network of structural cross sections presented here were necessary to understand the origin and evolution of the Gres d'Annot basin floor topography (section 5.5 and 5.6). By constructing them, it has also been possible for the first time to trace individual structures across the region, establish the links between them. This exercise has highlighted some important structural features of the region, concerning both their geometry and evolution. The most significant results are outlined here and discussed in the following sections.

1. The initiation and development of many structures have been dated with respect to the Tertiary sediments in the region.

For example, the earliest Tertiary sediments imply that Pre-Triassic basement was uplifted much earlier than previously supposed. A tectonic model has been developed for the region which incorporates this evidence (section 2.6).

2. Graham (1983, pers.comm.) concluded that zones of intensely kinked stratigraphy were invariably related to underlying thrust hanging wall ramps. However, this study has shown that certain kink zones are buckle folds and were not related to changes in structural level. The best examples are the Chateau Garnier and Argens folds (cross sections A1, B1, B, Y): they are laterally discontinuous and form an en-echelon array of kinked strata (section 2.2.2.2.).

3. The importance of lateral and oblique structures in this region is demonstrated by the distribution of Tertiary facies.

The relationship between the sediments and the structurally induced palaeotopography provides information about the nature and geometries of the lateral ramps in the system (section 2.1.3.). The different interpretations are highlighted by comparing cross section D with Lawson's equivalent lateral cross section (1987). He has interpreted south-dipping lateral footwall ramps. This study has shown that thrusts consistently climb towards the south, thus maintaining compatability with the southward directed thrust system.

2.3.2. The structural geometry and tectonic evolution of the Northern Castellane Arc derived from the structural cross sections

An understanding of the tectonic evolution of structures in this sector may be arrived at through a discussion of each cross section drawn. They will not be described in strict geographical order; instead, each will be used to highlight a feature of the tectonic history. However, the western part of the sector will be considered first because it has been studied in most detail. The continuity of certain structures will become apparent in the process. The location of each section is shown on enclosure 0.

The reader is advised to move to a room with plenty of space! The principle structures and their relationship to each other are shown in the maps of enclosure 15 and 16. The cross sections are best understood when laid out together. Figures 2.20, 2.21 and 2.22 are two composite sections which show the evolution of structures to the north and south of the Beauvezer/Thorame Basse valleys. Fig. 2.22 is drawn perpendicular to the Alpine shortening direction, sub-parallel to the Pyrenean shortening direction. At each stage in the text, the relevant cross sections will be listed.

2.3.2.1. The Beauvezer cross section: A1. (enclosures 3 and 4; associated cross sections are A, X, B1, B and D).

This NE-SW cross section is described first because it best illustrates the way in which all the cross sections were constructed. As well as the present day structure (encl. 3), a section restored to the assumed Palaeocene peneplain is presented (encl. 4). The inferred sequential evolution is shown in fig. 2.20, which summarises features from sections A, A1 and X.

At the NE limit of the cross section, the thrust contact at the base of the Embrun-Ubaye thrust sheets is seen to enter slicing horizontal through folded Lower and Mid Cretaceous strata then ramping slightly into the Tertiary succession. These outcrops will be discussed in the context of further evidence in cross section A.

Restoration of the Base Tertiary unconformity to horizontal (enclosure-, fig. 2.20a) shows the following features:

i) The Upper Cretaceous varies considerably in thickness.

ii) Angular relationships exposed around Chateau Garnier show that the variation is erosional and can be related to compressional structures.

iii) At Le Grand Coyer to the south, the undeformed Upper Cretaceous succession is approximately 850m thick. Nowhere to the north of Beauvezer is the section thicker than 500 (fig. 2.23). Therefore, a regional loss of 350m+ is superimposed on local thinning over specific Mesozoic structures. Allowing for slight regional variations in thickness of sediment deposited (the limestone facies were not deposited as bioherms; the sediment simply infilled topography), a palaeotopographic high >300m relief existed to the north or an equivalent thickness of stratigraphy has been eroded from the whole section.

The Upper Cretaceous does not regain thickness anywhere to the north. The southern limit of this regional thinning appears to have an east-west trend. The regional scale of the uplift implies that a basement involved structure was responsible (fig. 2.22). A steep, reverse fault in the basement would produce the regional uplift and narrow ramp zone observed. The hypothesis cannot be tested directly, but supporting evidence lies to the west, along strike of the ramp zone, where the Lias thins abruptly southward across the inferred deep-seated structure. Therefore, the structure is interpreted as a Mesozoic extensional fault, reactivated in compression during the Tertiary.

Large amounts of sediment were removed from the region, but some was preserved at breaks in slope and at the base of the palaeo-regional slope: the Poudingues d'Argens, for example, at Peyresq.

The Upper Cretaceous thickness variation was used to construct the palaeo-structure. Assuming that the kilometre scale wavelength uplifts were produced by thrusting in the Mid Cretaceous shales and kink folding in the Upper Cretaceous, the position of the base Mid Cretaceous should be a constant depth beneath the unconformity. Consequently, the amount of structural thickening of the Mid Cretaceous can be constrained. The process is illustrated in fig. 2.20a.

The assumption that all the shortening during this phase of deformation was taken up within the Mid Cretaceous is shown to be valid at Colmars. Here, the thinnest Upper Cretaceous (<30m thick) demonstrably overlies equivalently thickened Mid Cretaceous shales. This hypothesis is supported by the highly sheared state of the mid Cretaceous shales above Thorame Basse and is consistent with in-sequence thrust propagation from higher to lower decollements at any one point in the system.

The restored cross section shows the existence of two separate zones of thickening (labelled 1A and 1B on fig. 2.20). The 1A duplex involved only the Mid Cretaceous, resulting in uplift, but not shortening, of the Upper Cretaceous. The Upper Cretaceous is folded and shortened above 1B, which suggests that not all the displacement could be transferred along the decollement. Thrust propagation was probably hindered by the Col du Defend lateral ramp to the west of La Batie (section 2.3.2.12, cross section X).

In reality, the erosion surface was not a perfect peneplain.
In this section there was some remnant topography, suggested by the thinning of Calcaires Nummulitiques over the crest of the duplex 1A. The stages of development of structure during the Poudingues d'Argens and Calcaires Nummulitiques Formations are not so well constrained in this section. The Calcaires Nummulitiques is thin over the Mid Cretaceous duplex 1A: the result of continued shortening on these thrusts or remnant topography.

Bedding in the Calcaires Nummulitiques onlaps slightly eroded Upper Cretaceous limestones at Boules, on the margin of the La Vallette kinked and thrust zone, showing that this structure existed at the time. This topographic high <u>may</u> also have been in part generated by the initial development of a deeper thrust (3a), but this cannot be tested in this section. It is probable that Poudingues d'Argens were deposited in the depressions around the La Valette structure, but the Tertiary sediments have been eroded from all but the southern tip of the structure. A small outcrop of continental sediments exists there, but palaeocurrent indicators were not found, so their origin cannot be defined.

The Chateau Garnier folds were probably initiated at this stage, forming behind the La Batie folds when continued shortening within the Mid Cretaceous could not be transferred forwards.

Activity on the remaining structures is constrained by comparison with other cross sections: Thrust 2 in cross section A; thrust 3a in sections B1, B and Z.

2.3.2.2. La Valette cross section: A. (Enclosure 2)

This section was drawn to constrain the geometry and development of the structures beneath extensional faults observed in the Ravin de Juan. It shows the same features as cross section A, but provides additional information. The cross section also illustrates the thrust contact at the base of Sub-Brianconnais and Helminthoid Flysch nappes.

The Marnes Bleues Formation thins approaching Chasse. This is interpreted as the response to a palaeohigh above a Lower Cretaceous thrust sheet 2.

The significant normal faults that exist in this section are interpreted to accommodate the uplift associated with raising thrust 2 over the footwall of the Lower Jurassic thrust 3a. The faults do not cut the Beauvezer section because the extensional displacement is taken up on normal faults associated with the oblique footwall ramps of the Mid Cretaceous thrust duplex 1A.

The normal faults and, by inference thrust 2, were active during the deposition of the Gres d'Annot: sandstone bodies are concentrated in the hanging wall. It appears that the position of thrust 2 was controlled by the location of thrust sheet 3A's hanging wall ramp. Consequently, displacement on thrust 3b probably pre-dated the deposition of the Gres d'Annot at this site.

The NE edge of this cross section and A1 provide information about the timing of emplacement of the internal thrust sheets. The Schistes a Blocs are in tectonic contact with the Gres d'Annot Formation, forming a thin basal thrust sheet at the base of the Embrun-Ubaye nappes. They are not considered part of the basin fill in the Trois Eveches outcrop area since they were transported some distance to their present site on a thrust that decapitated the top of the Gres d'Annot Formation. The turbidite succession does not contain any detritus eroded from the internal thrust sheets, which implies that they were some distance from this part of the basin during the Late Eocene, or separated from the depocentre by a topographic barrier, produced by thrusting in front of the Embrun-Ubaye nappes.

There is evidence for the second hypothesis. The sole thrust of the Embrun-Ubaye thrusts slices through folds and thrusts in the Mesozoic and Tertiary strata of the foreland. As it propagated through the already deformed foreland, pre-existing folds were tightened in front of the thrust tip.

Northeast of Chasse, the Gres d'Annot dip steeply to the NE. The sole thrust of the Embrun-Ubaye nappes, which is subhorizontal, cuts down through stratigraphic section in the transport direction. Related subhorizontal faults in the Gres d'Annot are also extensional with respect to stratigraphy. It is probable that all the structures are compressional, developed as the Embrun-Ubaye sole thrust propagated through an already tilted foreland sequence.

2.3.2.3. How the structure and evolution of the Argens outcrop area were derived using the western portion of cross sections B, B1, E and Y and cross sections E1 and Z.

Figure 2.21 shows the sequential development of structures in the southern part of the Northern Castellane Arc. No one cross section could be constructed without reference to adjacent sections and the lateral section Z. Each section gave a small part of the story, but section B1 was the most important.

<u>How the geometry of individual structures was defined.</u>

In the margin of enclosure 6, cross section B1, the topography of the base Tertiary is illustrated in 6 sketches. To establish which structures were responsible for the shape and elevation of this surface, the structures that were seen at surface were placed on each section; their effect was then subtracted to show the remnant topography produced by deep seated structures.

i) **Shortening above the Mid Cretaceous decollement** was responsible for much of the structuring of the Upper Cretaceous and Tertiary. For example, in section B1, the whole of the Argens monocline is developed above the decollement. The way in which shortening is taken up varies across the area. In the south (section E), a sizeable Mid Cretaceous duplex is required to construct a huge Upper Cretaceous anticline. To the north, this shortening is transferred to the SW along the Mid Cretaceous decollement and surfaces as a thrust near La Mure Argens.

ii) **A zone of kinked Upper Cretaceous strata** appears on each section, trending NNW-SSE overall. The best exposure of these folds occurs in the Verdon valley south of Thorame Haute. Their geometry is shown in cross section E1. The syncline of a large scale, asymmetric, west-vergent fold pair -the Argens folds- may be studied in detail above the village of Argens and traced into an open monocline to the south along two valleys that link the villages of Argens and Allons. The relationship of the Tertiary sediments to the Upper Cretaceous structures may be studied in these valleys.

iii) In section B1, the base Tertiary is elevated over a region 2km wide by an amount equivalent to the thickness of the Upper Triassic and Lower Jurassic stratigraphy. The same feature appears on all the sections to the south, suggesting that underneath it there is a 5km displacement thrust which duplicates that part of the stratigraphy.

This model should be compared with that of Graham (encl.1), who ascribed a large percentage of the observed uplift to a thrust which doubled the thickness of the Upper Cretaceous. Outcrop in the Verdon valley reveals that the level of the Lower Cretaceous is much higher than predicted by Graham's model. Therefore, the structure responsible for the uplift lies beneath the Lower Cretaceous.

iv) In cross section B1 and B the base Tertiary is down-folded over the hanging wall ramp of thrust 3b. In the southern sections this monocline disappears, interpreted to be the result of tilting the hanging wall ramp above another structure: the edge of the footwall ramp to the Col du Defend/Castellard thrust.

This accounts for Graham having wished to place the shortening at a high level. His cross section did not contain a SW-vergent monocline. Therefore, he placed a thrust in the Upper Cretaceous which ramped to surface, allowing the hanging wall ramp to be eroded.

In summary, three structural elements interacted to produce the observed relief on the base Tertiary unconformity:

A) A thrust duplex in the Mid Cretaceous, with associated kink folding in the Upper Cretaceous.

B) A thrust in the Lower Jurassic with displacement in the region of 5km. Both the footwall and hanging wall ramps lie in the cross sections.

C) The footwall ramp of the Castellard and associated thrusts. This structure was a lateral ramp to the principal southward displacement on the faults and a frontal ramp to the subsequent SW directed displacement.

Cross section B1 contains only two of these elements and therefore is the easiest cross section to interpret. The present geometry of thrust 3b was defined using this cross section and a careful iterative analysis of the data in all the cross sections. The NE-SW cross sections showed that in the axis of the Argens syncline the base Tertiary was only elevated by one structure: thrust 3b. Therefore, a lateral cross section (Z) was constructed parallel to the axis of the Argens syncline. The level of the base Tertiary defined the lateral continuity of the thrust sheet and predicted which stratigraphic levels were duplicated in each of the NE-SW cross sections.

The resulting cross section Z showed that the thrust ramped to the south and rapidly to the north. Hanging wall or foot wall ramps? The structural data did not distinguish between the alternative solutions. The sedimentological data implied that the position of palaeohighs was fixed with respect to the sediment surface. Therefore, both ramps are hanging wall ramps. These and the hanging wall ramps in the NE-SW cross sections constrain the direction of thrust transport: towards the south west. If it had been southwards or westwards, then one of the two ramps in cross section Z would be a footwall ramp.

Dating the development of each structure.

Phase 1: The Upper Cretaceous succession here is a uniform thickness of 550m, where the full sequence is exposed. This is thin by comparison with the Upper Cretaceous at Grand Coyer. The difference is attributed to erosion above the broad basement uplift, prior to and during deposition of the earliest Tertiary sediments (section 2.3.2.1; fig. 2.23 and cross sections D, Z). This uplift was the main source of conglomerate to the Poudingues d'Argens alluvial fans at Peyresq, Le Ruch and possibly Allons.

Phase 2: In the northern Argens outcrop area, the facies of the Poudingues d'Argens Formation are for the most part fine grained. There were no major hills supplying detritus to the area. However, it is clear that the Argens anticline formed a positive feature. Poudingues d'Argens were not deposited on the steep and now overturned limb of the fold. Where the fold is overturned (cross section B and B1) a patchy breccia is preserved, interpreted as palaeo-scree derived from the earliest expression of the steep limb of the Argens anticline. The few palaeocurrent data in the Poudingues d'Argens in this syncline show that the carbonate detritus was supplied from the east and south-east, supporting the hypothesis that the Mid Cretaceous duplex and Upper Cretaceous folds were developing.

Phase 3: The Calcaires Nummulitiques at Argens are conglomeratic at the base. The interpreted fan delta prograded southwards, supplied from exposed terrain immediately to the north. The highest present-day structural elevation of thrust sheet 3b is in the Thorame Basse valley, to the north of Argens. It is quite feasible that the culmination formed a palaeohigh in the Eocene that protruded above sea level and was fringed by fan deltas.

In cross section E1, the Poudingues d'Argens overstep slightly eroded Cretaceous strata towards the east, directly above the interpreted hanging wall ramp of thrust 3b. This supports the hypothesis that thrust sheet 3b had developed by the Mid Eocene. The nature of palaeosols at the transition from continental to shallow marine sedimentation demonstrate that Allons subsided faster than Argens itself (section 3.4.2.). Regional subsidence was locally countered by shortening on the Jurassic thrust sheet.

Phase 1-3 (?): The age of the Castellard thrust footwall ramp, cannot be determined from any of the cross sections drawn. Evidence presented in section 2.3.2.12 (part 4) suggests that the principal displacement on the fault was southwards and probably related to Pyrenean forces. This dates most of the thrust displacement as early Tertiary (Late Cretaceous-Eocene). Sedimentary evidence to support this is found in the St. Antonin syncline which accommodates some of the displacement to the east where the Castellard thrust tips. The Eocene sediments were deposited in an actively growing syncline.

In the line of the cross sections constructed, the Poudingue d'Argens and conglomeratic Calcaires Nummulitiques at Rouaine are the only possible remnants of Tertiary sediment that may have recorded the uplift associated with this southward movement.

2.3.2.4. Summary of information presented in the whole of cross section B1.

The Argens monocline was initiated during the deposition of the Poudingues d'Argens. In this section, it formed above the hanging wall anticline of a Mid Cretaceous thrust. Subsequently, a second thrust (1d) propagated from behind, breaching the Upper Cretaceous: the Argens fold tightened and was decapitated by the late thrust. Thrust 1d appears to have ramped up out of the folded decollement above the footwall ramp of the Jurassic thrust: therefore, it postdates all the movement on thrust 3b.

The elevation of the Argens syncline implies that Triassic and Lower Jurassic are repeated in thrust 3b. Fan deltaic conglomerate bodies in the Calcaires Nummulitiques date the initation of the structure as Lutetian. **It can be correlated with the Jurassic thrust 3a seen in cross sections A, A1 and X.** The displacements are similar and the two thrusts never appear in the same NE-SW cross section. Cross section D shows the lateral ramp that must link them in the Beauvezer valley. The direction of thrust transport towards the south-west, as defined by cross sections B1 and Z, is consistent with this hypothesis.

The very steep normal fault observed to the west of the Argens syncline has a low displacement and may sole into the Mid Cretaceous shales. There is no obvious reason for its existence!

2.3.2.5. Verdon valley cross section: E1

This short NE-SW cross section constrains the changes in structural level implied by the Upper Cretaceous kinks in the Verdon valley. It also illustrates the geometry of the footwall ramp of the Castellard thrust.

The Upper Cretaceous kink folds define a large kilometre scale anticline with a complex form. The uplift was ascribed to a small Mid Cretaceous thrust (=1c); the upper part of the footwall ramp and first stage of the hanging wall ramp of thrust 3b. The unconformity at the base of the Tertiary and the absence of Poudingues d'Argens on the steep limb of the fold imply that all these structures were early.

2.3.2.6. The western portion of the Allons cross sections: E and Y.

In cross section E, the base Tertiary unconformity is raised to almost 2000m along Le Puy de Rent. The elevation is interpreted to be produced by the hanging wall/footwall ramps combination seen in cross section E1. Here, not one, but a duplex of small displacement thrusts are interpreted to complete the picture.

To the south, in cross section Y, an equal amount of Mid Cretaceous shortening is observed, but the thrusts breached the decollement and propagated sharply through the Upper Cretaceous. The Calcaires Nummulitiques is demonstrably displaced across some of these faults. However, the age of the rest cannot be determined. There is no Poudingues d'Argens which implies that little, if any, displacemnt was early.

Cross section Y illustrates the footwall ramps to the large displacement Castellard thrusts and interprets the presence of a second footwall ramp. Fig. 2.32 and 33 show the geometry of the thrust sheets and how they evolve towards the north. The sketch cross sections are interpreted to be lateral sections through southward directed thrusts: this accounts for the apparently rapid variations in displacement.

2.3.2.7. Argens-Allons syncline cross section Z (Enclosure 14; associated cross sections A, A1, B1, B, E and Y)

This NW-SE cross section is discussed briefly in section 2.3.2.3. It was constructed along the axis of the Argens syncline. With the exception of the regional basement slope (7' down to the south), the elevation of the base Tertiary is ascribed to thrust sheet 3b in the Lower Jurassic, plus a slight thickening of the Mid Cretaceous shales. The cross section shows:

i) The basement uplift of approximately 300m responsible for erosion of Upper Cretaceous over the the whole region north of the Beauvezer and Thorame Basse valleys. The foot of the gentle palaeoslope was near the village of Argens, which explains the absence of coarse member facies in the Poudingues d'Argens Formation to the north of the village.

ii) The Lower Jurassic stratigraphy thickens gradually towards the north, in accord with a regional trend defined on the 1:250,000 Gap map. There is nothing to suggest that syn-sedimentary faults were responsible.

iii) The geometry of the Jurassic thrust sheet 3b.
The high structural level of the Mid Cretaceous in the Thorame Basse

valley defines the maximum thickness of the 3b thrust sheet. The reappearance of Upper Cretaceous in the valley at Chateau Garnier defines a steep structural dip to the north, interpreted as the lateral ramp of thrust sheet 3b. To the south, both the present and palaeo structural dips describe the shallow slope of the hanging wall ramp below.

iv) North of the Col du Villaron (Chateau Garnier), the base Tertiary climbs steeply, from 1200m to 1650m+. There are two interpretations of this, both based on cross sections drawn by Lawson (1985, Pers. comm.; 1987, unpublished thesis). In the latter, the Triassic decollement is folded over a basement high. This implies that the steep structural dip in cross section Z represents the lateral/oblique margin of that basement uplift. Lawson's earlier interpretion related the surface slope to the Lower Jurassic footwall ramp of a large displacement thrust.

The earlier model defined a footwall ramp with a component of climb to the north: if present, it is the only one in the study region. The later model explains the exposure of basement in a small fenster at Barles. If the first model is accepted, the steep surface structural dip, reflects a steep dip on the hanging wall ramp which implies that the compressional basement fault is steep and may therefore be a reactivated normal fault.

v) At the southern end of the cross section, the Calcaires Nummulitiques rises sharply over the footwall ramp of the Cremon thrust.

2.3.2.8. Cross section Y and Y1. (Enclosure 13; associated cross sections are E, B, D and W.)

The western part of cross section Y was dealt with in section 2.3.2.7. The eastern part describes the Annot syncline, at it lowest structural level, and the structural elements that define its eastern boundary:

1. The hanging wall ramp of a far-travelled basement thrust sheet.

The base Tertiary unconformity is elevated to 2300m at the eastern edge of this cross section. It is raised to a uniformly high level over a large part of the study region: over most of the Grand Coyer outcrop area and everywhere to the north-east. It drops rapidly between Mont St. Honorat and Daluis, to 1000m adjacent to the Rouaine fault zone. As illustrated in cross section Y, it dips steeply, then more gradually, down to the west across the Annot outcrop.

Menard's depth to Pre-Triassic basement map places the basement at a high level beneath the Col de La Cayolle outcrop area and then predicts a steep basement slope beneath the Annot outcrop area. There is no direct evidence for this basement culmination, but it accounts for the uplift while minimising the cover shortening required to achieve these dizzy heights!

The geometry of this basement thrust is defined by this and other NE-SW and NW-SE cross sections: E, B, D and W. The amount of basement shortening it represents cannot be determined by structural methods alone. The Tertiary sediments imply a long history of basement uplift which has enormous implications for basement shortening and the tectonic history of the whole region (see also section 2.6).

The latest stage in the development of this structure post dates deposition of the Tertiary sediments; it is defined by cross section Y and B. The early history of the structure will become clear in sections 2.3.2.10 and 11, describing cross sections W and D, but note that no Poudingues d'Argens Formation outcrops in the structural low at Annot.

At Annot, slightly south of this cross section, the base Gres d'Annot is sub-horizontal, but the Gres d'Annot beds downlap westwards at an angle of 10-20'. The turbidite sediment surface was originally horizontal, and the turbidites onlapped to the west, onto the uneroded uppermost sediments of the Marnes Bleues Formation.

There are two mechanisms for producing this angular unconformity and the final structural dips. The models are illustrated in fig. 2.25 (and feature in the summary diagrams of fig. 2.21): the first applies at this locality; the second accounts for the downlapping turbidites along the south face of Le Grand Coyer (cross section B; section 2.3.2.9.).

If cross section Y is partially restored, the Gres d'Annot turbidites onlap a palaeoslope of Marnes Bleues tilted above the footwall ramp to the Cremon thrust. This dates activity on the thrust as no later than the end of Marnes Bleues deposition. This agrees with the interpretation of the fault as a southward directed, "Pyrenean" thrust, arrived at independently (section 2.3.2.12.).

A new, thin thrust sheet developed beneath the Triassic decollement (BIII), under the Mesozoic footwall ramp: the old ramp was tilted above the hanging wall ramp of the basement thrust. The Marnes Bleues strata reflect the new dip of the old footwall ramp; the Gres d'Annot strata were tilted parallel to the new hanging wall ramp. In this way, the cross section may be accurately constructed.

Fig. 2.38 and 2.39 illustrates the evolution of the basement structures and will be discussed in section 2.6.

2. The Melina-Aurent kink zone.

The zone of kinked Upper Cretaceous strata can be easily traced across the bare, overgrazed hillsides, from Daluis, north west to Peyresq where it splits into two low amplitude structural highs on either side of the Tertiary outcrop. The large scale monocline described by the kink zone is correlated across the Rouaine fault with the downward-facing Tete du Pibossan fold (fig. 2.25b). which implies relatively small amounts of lateral displacement on the steep fault zone. This correlation suggests that the kinked Upper Cretaceous limestones reflect shortening of the whole Mesozoic succession, the displacement entering on the Triassic decollement and ramping smartly to surface as a large asymmetric fold.

The position of monocline is defined in cross sections B, E, Y, Y1 and D. Note that the axis does not run quite parallel to the strike of the present day basement hanging wall ramp, giving rise to a complex base Tertiary topography.

The evolution of the structure.

Cross section B shows the Calcaires Nummulitiques folded by the Aurent structure. However, the low angle unconformity at the base of the Tertiary demonstrate an early topographic expression of the fold. This is best illustrated in cross section Y1, where the Poudingues d'Argens rests directly on Mid Cretaceous (Barres de Martignac) and eroded kink folds in the Upper Cretaceous. The Tertiary unconformity cuts down towards the south, because the Melina-Aurent monocline was superimposed on the hanging wall ramp of Proto-Barrot, BI.

The model is based on independent evidence for an early Tertiary topographic expression of Barrot. The continental and transgressive sequences fringing Barrot contain clasts of locally derived Permian and Lower Triassic (discussed in section 3.4., 3.5.6. and 4.4.6.).

The Melina-Aurent structure had a long history of activity, as evidenced by;

(i) Erosion of Upper Cretaceous by the base Tertiary unconformity;

(ii) Facies and thickness changes in the Calcaires Nummulitiques and Marnes Bleues (between Mort de l'Homme , Méailles and Annot);

(iii) Possible contribution of this structure to the topographic confinement of the Annot Syncline during the deposition of the Gres d'Annot;

(iv) 100m-scale kink folds affecting the whole Tertiary sequence.

There is one important regional consequence of developing the Melina-Aurent kink zone prior to and during Tertiary sedimentation. By analogy with the Tete de Pibossan fold, the monocline deforms the mid Cretaceous decollement. All the cross sections show that large thrust displacements were transferred into the foreland at this decollement level. Once the kink zone had developed, however, the mid Cretaceous was bent, and further slip at that level was prevented.

This is a problem pertinent to the Tinee Nappes which developed post-Tertiary deposition. Before it formed displacement could pass smoothly along the Mid Cretaceous decollement, and all the cross sections demonstrate that there was significant movement along that level. Once the kink zone had developed, however, the resulting folding of the Mid Cretaceous would prevent further slip on that decollement at this location. This means that the roof thust of the Haute Provence duplex was deactivated. Subsequently, thrusts in the cover NE of Aurent had to ramp to surface before reaching the kink zone. This includes tens of kilometres of shortening on the Tinee Nappes, which all post-date Gres d'Annot deposition. The related, large displacement thrust through the Upper Cretaceous has been entirely eroded.

2.3.2.9. Peyresq Cross section B (Enclosure 5)

This cross section shows many features described in previous sections. All the structural elements have been introduced with the exception of the thrust out, SW-vergent monocline at the north-east end of the cross section. It is a section in which the Upper Cretaceous is most deformed with a particularly complex array of kink folds and small scale thrusts at Colle St. Michel.

1. Western part of the section: thrust 3b and Argens thrust.

The elevation of the base Tertiary defined the position of the hanging wall ramp and thickness of repeated strata in thrust sheet 3b.

The Argens thrust was initiated late, where the Mid Cretaceous decollement was downfolded over the basement ramp. Much of the complex folding in the Verdon valley and up to the Col de st. Michel is probably related to this thrust. The fault dissected pre-existing folds (associated with the Argens monocline) and produced new kinks and small scale scale accommodation thrusts.

In addition, there are E-W folds possibly related to the early basement involved uplift north of Beauvezer and to the first thrusting along the Mid Cretaceous decollement. Together, these processes uplifted the area to the north which was the source for the Poudingues d'Argens of Peyresq (section D). Some of the kink folds were not eroded, but infilled with Poudingues d'Argens. In many cases, a kink band in the Upper Cretaceous passes up into a normal fault in the Tertiary: some of these were active during the early Tertiary, controlling the position of major channel fills in the Poudingues d'Argens (section 3.4.1.).

2. The structures that bound the Grand Coyer outcrop area.

Onlaps and facies associations of the Gres d'Annot in this area define the pre-Gres d'Annot displacement of two structures in this area: the Cairas monocline and the basement thrust sheet BII. Cremer (1983) demonstrated that the Gres d'Annot here were deposited in an elongate trough. The results of this research show that the trough was defined to the NE by the Cairas monocline, to the SW by the footwall ramp of the basement thrust sheet. The Marnes Bleues thin markedly onto the Cairas monocline implying that it formed a positive feature on the sea bed.

The Gres d'Annot now downlap westwards across Grand Coyer. Restoring the Gres d'Annot to horizontal reveals a NE-dipping slope in the Marnes Bleues. There is no repetition of the Mesozoic beneath that area. This slope coincides with footwall ramp of an interpreted basement thrust sheet. Subsequent displacement has moved the hanging wall ramp onto a footwall flat, tilting the Gres d'Annot into their present orientation (fig. 2.24).

2.3.2.10. La Lance cross section W (Enclosure 11; associated cross sections Y, E, B, D)

This cross section is lateral with respect to the SW directed thrusting. It contains critical evidence to date the basement thrust sheet BII-III (cross sections Y, E, B, D). In the process, it demonstrates some important principles of cross section construction.

The Base Tertiary in the SE half of the section is uniformly uplifted on a basement hanging wall flat to levels close to 2300m. At the southern edge of the section, at Rocher du Carton, the Upper Cretaceous sequence is at its thickest anywhere, at 900m. The base Tertiary unconformity cuts down stratigraphy overall towards the N/NE. The deepest level of Pre-Tertiary erosion occurs at Colmars where the Upper Cretaceous sequence is less than 50m thick. A less eroded palaeohigh occurs in the middle of the section and coincides with the present day topographic and structural high.

The nature and evolution of the structures responsible for uplift were examined in the following way:

The base Tertiary unconformity was restored to horizontal, thus revealing the structural relief in the Early Eocene. The Upper Cretaceous at Colmars was eroded above the regional basement uplift and Mid Cretaceous duplex 1a. These structures are illustrated better in cross section A1 and Z and discussed in sections 2.3.2.1 and .7.

It is not possible to determine the age of the structural high at Cairas from structural analysis alone. Elsewhere, the lack of angular discordance on both Tertiary unconformities gives the (misleading) impression that the structure was post depositional. The Calcaires Nummulitiques facies at Cairas demonstrated tectonic activity during the transgression. The evidence for near surface seismicity, kink folds developing during sedimentation and olistostromes is discussed in section 4.5.2.

The evidence points to a consistant story of progressive deformation of the Upper Cretaceous in response to shortening and uplift on a major structure beneath. At the edge of cross section W, an interpretation of the palaeotopography is presented which shows the small scale kink folds constructing a large scale slope dipping down to the NW: the same slope observed at present day.

Poudingues d'Argens were not deposited in the Grand Coyer outcrop area and do not occour at Annot which is in front of the same structural high. Therefore, the structure was initiated during the very first stages of the Tertiary transgression and grew during deposition of the Calcaires Nummulitiques: the relief may not have increased much since.

The NE-SW cross sections show that Mesozoic structures are not responsible for the uplift. Therefore, this cross section clearly defines a ramp in the Pre-Triassic basement: hanging wall or footwall? This section (Cairas down to Colmars) and cross sections Y, E and D (the northern and eastern boundary of the Annot outcrop area) illustrate different aspects of the same structure. In all cases, the present and past structural highs and basement ramp slopes have remained in the same position with respect to the sediment surface since the earliest stages of the Tertiary transgression. Therefore, the sections describe the geometry of hanging wall ramps on a footwall flat: they strike NW-SE and NE-SW. A south-westerly thrust transport direction is consistent with these data and the shortening recorded in many of the Mesozoic thrusts.

The section does not extend far enough to show that the base Tertiary drops rapidly to the south of Mont St. Honorat, from 2300m-900m in <5km. The sketch section of fig. 2.39 is the continuation of cross section W and shows a small outcrop of Poudingues d'Argens resting on only 100m of Upper

Cretaceous. In the valley, the Poudingues d'Argens lie directly on Mid Cretaceous. The partial restorations interpret the unconformity cutting down towards a proto-Barrot to the south.

A summary of the model for the geometry and evolution of the basement structures will be presented in section 2.6.

This cross section also illustrates a steep lateral ramp in the sole thrust of the Embrun-Ubaye Nappe: it surfaces from the Mid Cretaceous decollement. Its position in the Beauvezer valley may have been controlled by heterogeneity in the foreland cover along the line of lateral ramps in the foreland thrust belt: for example, the basement thrust sheets BI, II and III, plus the Jurassic thrust sheet 3a.

2.3.2.11. Cross section D, D1, D2 (Enclosure 7 and 8; associated cross sections A, A1, X, B1, B, E, Y, Y1)

This staggered cross section was constructed through the whole study region to illustrate the lateral geometry of the principally SW directed thrust sheets and their relationship to one another. The section has revealed much about the evolution of the foreland structures, which is diagramatically summarised in fig. 2.22.

Faults were correlated between the tie lines with NE-SW cross sections: present day and past thrust ramps were located using the multiple unconformities and sedimentological data. The fault traces were drawn consistent with this and all other cross section data.

The principal features are:

i) All thrusts climb **southwards** through the cover sequence, generally at a low angle, with ramp-flat geometries.

ii) Southward-climbing thrusts are also identified in the basement. These two features together have the effect of making the northern part of this cross section, in particular, look like a S or SE directed thrust system.

iii) The basement thrust sheet BII is well illustrated, defining a lozenge shaped uplift bounded by two hanging wall ramps. The northern ramp has a shallow dip that slopes across the Beauvezer valley. The other slope defines a large component of the southerly plunge of the Annot structural low.

iv) If the effect of the basement thrust sheet BII is subtracted, a steep SE-dipping basement slope remains to the south of the

Beauvezer valley. This is interpreted to be the isostatic response of the crust to loading by the Alpine orogen.

Three unconformities can be traced in this section:

1. Intra Cretaceous, at the base of the Cenomanian.

This is observed only in the northern part of the cross section. It truncates folded stratigraphy and progressively cuts down towards the north, so that at Dormillouse, the Upper Cretaceous overlie Tithonian limestones. Restoring the unconformity to horizontal (fig. 2.22) reveals two monoclines between Dormillouse and Montagne de la Blanche. These could have been produced by the inversion of extensional faults of Mesozoic age.

There is a tectonic contact at this level in the Bleone valley. It may be an unconformity modified by a small amount of thrusting.

2. At the base of the Tertiary.

The unconformity was drawn flat for the second stage of the section restoration (fig. 2.22).

i) The Upper Cretaceous is regionally thinned, to less than 500m, everywhere north of the Beauvezer valley.

ii) Superimposed on the regional thinning, structuring of the Mesozoic created local highs. The Upper Cretaceous is completely absent below Crete de la Blanche. The uplift may be interpreted to represent early shortening on one of the thrusts exposed at the present day: it duplicated stratigraphy above the Terres Noires as the thrust ramped up to the Mid Cretaceous decollement.

For some distance to the south, the thrust is a flat on the Mid Cretaceous decollement, and, consequently, it produced no uplift. It climbed stratigraphy again in the 1A duplex which produced more uplift: above it, Upper Cretaceous was reduced to less than 50m.

Across the Beauvezer valley, the large displacement demonstrated by the 1A duplex is not seen. **It is probable that the thrust continued to climb southwards and surfaced above the valley.** The inferred emergent thrust has been completely eroded (fig. 2.23).

The Tertiary sediments support this hypothesis. There is a sedimentological requirement for a range of hills to the north of Peyresq, to supply detritus to the fans of the Poudingues d'Argens and lower Calcaires Nummulitiques. Remnant topography was enhanced by early activity on thrust 3a during the transgression. It remained as a positive feature, forming the northerly barrier to turbidity currents flowing into the Annot palaeo-trough (fig. 2.22 and encls. 6. .).

The base Tertiary unconformity has been structured post-Tertiary deposition, as evidenced by thrusts which ramp through the Tertiary stratigraphy. However, it cannot be assumed that all the structural dips are late. As illustrated in the discussion of cross section W (section 2.3.2.10.), the facies of the Calcaires Nummulitiques reveal the existence of palaeoslopes.

The only palaeoslope studied in detail in this cross section is that which dipped southward from Peyresq down to Annot. The Calcaires Nummulitiques thicken into the palaeo-low. Fan deltaic and patch reef environments are exposed at Mort de L'Homme. These fringed the palaeo-plateau to the north around Peyresq. Refer to section 4.4. for a detailed discussion of the relationship of the limestone facies to the interpreted topography.

The structures responsible for the present and past topographic slopes are: the hanging wall ramp of the basement thrusts BII and III and the Melina-Aurent kink zone. The kink zone was initiated prior to the deposition of Poudingues d'Argens and continued to be active throughout the history of Tertiary sedimentation. The crest was eroded and continental sediments infilled some of the small scale fold and fault topography. Localised occurrence of cracked clasts in the conglomerates of the lower Calcaires Nummulitiques are indicative of continued tectonic activity (section 4.2.2.4.). The Marnes Bleues and the lowermost thin Gres d'Annot turbidites were tilted prior to the deposition of most of the Gres d'Annot Formation.

The large scale topographic slope from north down to the south was significant during deposition of the Gres d'Annot turbidites. Section 5.5.3. discusses the multiple evidence for an effective barrier to sediment gravity flows to the north of Annot.

3. <u>The base of the Gres d'Annot.</u>

The base of the Gres d'Annot Formation is a disconformity along much of the Trois Eveches outcrop area. Along the short lengths of angular unconformity, the turbidites onlap the Marnes Bleues. North of the Beauvezer valley, a component of the onlap or downlap direction is always to the south. The strata dip too steeply towards the north-west to reveal any component of dip on the formation in other directions. At Le Ruch, south of the Beauvezer valley, the turbidites onlap towards the N/NE.

The dip of the Gres d'Annot defines the latest structuring of the foreland. It was removed by restoring the displacement on the late thrusts to bring the bedding in the Gres d'Annot back to horizontal. The restored unconformity then defined the structures that produced the basin floor topography which was infilled by the turbidites. For example, north of Beauvezer, it climbed structural level towards the south-east across a series of monoclines. Each monocline corresponds to a thrust footwall ramp.

The Mid Cretaceous where continuous is an effective decollement. In places (eg. the Bleone valley), early structures caused local uplift and the shales were eroded. Thrusts propagating along this decollement were forced to ramp to surface at these sticking point. The thrusts surfaced prior to deposition of the Gres d'Annot and created relief on the sea floor.

According to Ghibaudo (1985), the consistent southwards onlap of the Gres d'Annot along the Trois Eveches implies that the northern end was originally 3km deeper than the southern end. The base of the Gres d'Annot is now at a roughly constant altitude along the length of the massif. In Ghibaudo's model, this requires that the north end of the basin subsided a great deal and then bounced back to its original height. An alternative model is suggested in which southward propagating thrusts caused the depocentre to migrate southwards (section 6.6.5). This avoids the need for the northern end of the massif to behave like a yoyo.

2.3.2.12. Col de la Cayolloe - Tartonne cross section: X

This cross section is the only one which runs through the whole study region. It was extended into the Barreme outcrop area with the aim of understanding the tectonic relationship between the Gres d'Annot basin and the marginal syncline at Barreme. The Col de la Cayolle outcrop area is covered by the NE end of this section and the three short parallel cross sections. The central portion of the cross section illustrates somewhat different aspects of structures that have aleady been described in reference to other cross sections.

For the purpose of this discussion, the section is considered in four parts:

1. From Crete du Content to Sommet du Chateauvieux

The north-eastern limit of cross section X is the last outcrop of Tertiary stratigraphy to the SW of the Argentera massif. On sections X1 and X2, the boundary of the Col de La Cayolle outcrop area is marked by steep extensional faults that downthrow to the SW, off Argentera. These branch southwards and the displacement decreases rapidly.

The base Tertiary unconformity in the Col de La Cayolle outcrop area is at a high level, above 2000m. Where the stratigraphy is folded up over the Argentera massif, the outcrop shows only a single thickness of Mesozoic in the line of cross section X. I have assumed that the Mesozoic succession is not repeated at all under the Col de La Cayolle outcrop area and so on the constructed section, the base Tertiary is raised above a basement culmination. This hypothesis is supported by Menard's map of the depth to

Pre-Triassic basement.

Cross section X runs to the north-west of the interpreted branch lines of the Tinee culmination of Mesozoic thrust sheets (Graham, 1981). Graham's cross section (encl. 1) cuts through the Tinee nappes over which the Tertiary stratigraphy is elevated above 3000m. The displacements observed in the Tinee nappes must be transferred via lateral ramps into structures behind and infront of the Col de La Cayolle area (fig. 2.26). Some of the displacement passed forwards on the Triassic decollement, forming a small duplex of Lower Jurassic thrusts beneath Col des Champs (part 2 of this section). Part of the displacement was laterally transferred onto the roof thrust in the Mid Cretaceous shales and the thrusts surfaced to the south-west of the Trois Eveches outcrop area. The rest was probably accommodated by shortening to the north-east of the Tinee valley, but the evidence necessary to test this hypothesis has since been eroded.

The area to the north of the Tinee nappes is unusual in that it is dissected by steep, planar, NW-SE and N-S oriented normal faults and by lower angle extensional faults. Depositional thickness variations show that some of these were active during the deposition of the Tertiary sequence (fig. 2.27).

Many of the faults do not have associated roll-over folds in their hanging wall. They are therefore planar to significant depths in the crust, at least down to the Triassic decollement; some cut basement. The evidence for this lies along strike of the faults to the south where they curve through almost 90' and cut the margin of the Barrot massif.

One exception to this is a fault exposed in the south face of Crete de La

Blanche. The fault is steep near the surface where it ramps through the Gres d'Annot. The turbidites dip steeply (>45') into the fault plane defining a tight roll-over fold. The Calcaires Nummulitiques below are not folded, but are fractured by conjugate faults. The roll-over fold geometry implies that the decollement for this major extensional fault is at a high stratigraphic level, predicted to be the base of the Upper Cretaceous. This is compatible with the Mid Cretaceous shales being a thrust fault decollement across the whole study region.

The Estrop fault is also interpreted to decolle in the Mid Cretaceous shales. The fault has a very shallow dip, but is apparently planar because there is no associated hanging wall roll-over fold.

Along Crete de La Blanche (X1) and in the lower Gialorgues valley (X2), the Gres d'Annot are displaced by single widely spaced faults, each with several hundred metre throws: almost all the faults are synthetic to the large displacement faults bounding Argentera. In contrast, the Calcaires Nummulitiques formation is fractured by numerous, closely spaced, 10-100m scale faults which are both synthetic and antithetic to the master faults.

Evidence for the timing of the extension

Where the exposure is good, for example along some of Crete de la Blanche, it is apparent that the small faults in the Calcaires Nummulitiques branch onto a single, planar fault higher in the stratigraphy and that displacement is conserved. In these cases the normal faults post-date deposition of the Gres d'Annot.

Further up the Gialorgues valley, the thickness of the Marnes Bleues

Formation changes abruptly. Figure 2.27 is a sequential restoration of cross section X1 which illustrates that the Upper Cretaceous, Calcaires Nummulitiques and Marnes Bleues thicken in the hanging wall of extensional faults. Therefore, some faults, including the low-angle Estrop fault, were active during the Late Cretaceous and early Tertiary, **prior to the uplift of Argentera.** The reconstructions demonstrate that the topographic relief associated with these faults was infilled prior to the deposition of the Gres d'Annot.

Two mechanisms have been suggested for the NE-SW extension:

i) the faults accommodate the uplift of the Argentera massif.

ii) the faults represent gravity sliding off the Argentera and Tinee nappe culmination.

The large displacement faults bounding Argentera to the north-west seem clearly related to the uplift of the massif. Some of the faulting in the Gialorgues valley may also be due to the same machnism.

Graham's cross section (encl. 1) shows the Tinee nappes and illustrates the culmination collapse hypothesis. It is demonstrably unfeasible: the displacement on the extensional fault that detaches at the base of the Upper Cretaceous is constrained by the offsets seen in the Trois Hommes-Roche du Pigeon section. Overall, post-Gres d'Annot displacement is less than 100m! The cut off nose of the Ciavalet thrust should be seen, transported a short distance along the fault (fig. 2.28), but it is not. Larger displacements cannot be transferred westwards into the fault by slip along the base of the Upper Cretaceous, because movement down the ramp would have created a steep roll-over fold in the Gres d'Annot, similar to that exposed on the south face of the Crete de La Blanche.

A new interpretation of the structure and evolution of the area is shown in fig. 2.29 and described below:

i) Extensional faults were active as early as the Late Cretaceous: approximately 300m throw is required on the Trois Hommes fault. Earlier activity cannot be ruled out.

ii) Significant extension occurred during before and probably during deposition of the Calcaires Nummulitiques and Marnes Bleues. The marlstones thicken to 400m within the graben from 50m elsewhere. Field observations indicate that the western margin of this graben was a low angle, east dipping extensional fault; the geometry of the fault has not been precisely defined. The eastern margin comprised a set of fault terraces across which the Calcaires Nummulitiques (illustrated on antithetic faults in section through Crete de la Blanche) and Marnes Bleues thickness increased in steps. Roll-over of the Gorgias terrace was probably due to curvature of the underlying fault.

iii) The Tinee culmination developed by SW-directed thrusting, with a basal decollement in the Trias evaporites. The first ramp through the Mesozoic occurred where the propagating detachment was blocked by the pre-existing extensional fault: the Ciavalet nappe formed. The extensional fault was rotated by the resulting hanging wall anticline, to give the local eastwards dip shown by the 1:50,000 map. Some of the earlier faults were partially inverted and small-scale compressional structures were generated in the Tertiary strata.

iv) Later, steep faults cut through the whole sequence including the Gres d'Annot.

There is independent evidence that the Tinee nappes developed largely after deposition of the Tertiary succession:

- The Tinee nappes significantly thicken the Mesozoic stratigraphy. If they had formed prior to deposition of the Gres d'Annot, there would have been a substantial topographic high on the south-east margin of the Col de la Cayolle area. There is no sedimentary evidence for such a palaeo-high.

- The Gres d'Annot strata are tilted up onto the Tinee culmination: greater than 500m structural relief is observed in the preserved Gres d'Annot. Therefore, at least 95% of the Tinee nappe development postdates the Gres d'Annot.

- There is some evidence for a slight palaeo-high during Tertiary sedimentation. Poudinques d'Argens are preserved only in the vicinity of the culmination. Locally, the Marnes Bleues Formation thins to zero: on the east face of Tete du Meric, Gres d'Annot were deposited directly on Calcaires Nummulitiques. 5-10% of the development of the Tinee culmination may have been early.

The Tinee nappe displacements do not root beneath the Argentera massif: they were transported over the top of Argentera along the base-Triassic decollement. In order for this detachment to be effective, it could not have had much structural relief, so the Tinee nappes developed prior to the principal phase of uplift of the Argentera massif.

The late, steep extensional faults are interpreted to accommodate the uplift of the Tinee nappes and subsequently the Argentera massif. The extensional faults that were active during early Tertiary sedimentation cannot be related to these structures. There are two hypotheses for their existence:

i) Extension as a result of gravity sliding off the back of the early Tertiary expression of the basement involved structural high, BII-III (north-west continuation of Barrot). Fig. 2.30 illustrates this model.

ii) Extension as a continuation of the Mesozoic tectonic stresses. This is unlikely in view of the widespread evidence for regional compression, in both N-S and NE-SW directions, during deposition of the lower Tertiary.

Therefore, the first hypothesis is favoured. It is entirely compatible with the independent evidence for early basement involvement in the foreland thrust and fold belt.

The extensional faults exposed along the Crete de La Blanche illustrate a problem encountered when attempting to interpret the contact between the Gres d'Annot and the Marnes Bleues Formations. The exposure is not 100% and the contact itself is obscured by grass along the length of the hillside.

Where exposure is good, the faults cutting the Calcaires Nummulitiques can be seen to branch onto large faults in the Gres d'Annot. However, from a distance, some extensional fault contacts between Gres d'Annot and the lower Tertiary resemble angular, onlapping unconformities (eg. east of Cime de la Blanche). Mis-identification of these by Ghibaudo (1985), led him to the conclusion that the Gres d'Annot onlap the steep, faulted boundary of Argentera, whose uplift therefore pre-dated their deposition.

This hypothesis is not tenable because:

i) The "onlap" contacts are, in reality, faults.

ii) More than 20km of thrust displacement on the Tinee Nappes, which post-date the Gres d'Annot, passed over the top of Argentera.

2. Chateauvieux to La Lance

The two late, planar, steep, normal faults lie immediately to the south-west of Chateauvieux: they are laterally continuous across the whole of the outcrop area, striking N-S and NW-SE. The W-dipping fault branches onto the hinterland dipping fault. If the uplift of Argentera and other thrust culminations to the north-east were entirely responsible for the extensional faults, SW-dipping faults would be the master faults. Here, it is the "antithetic" fault which continues to the Triassic decollement, implying regional extension of the cover.

Foreland of these large displacement faults, compressional structures are dominant.

Observations.

The Tithonian limestone folds are accurately drawn from continuous outcrop along the north-south Entraunes valley (fig. 2.8a). They have kink fold geometry and small scale folds construct the kilometre scale structures. Their form as drawn in this cross section differs from the simple asymmetric folds drawn by Lawson (1987): this has a bearing on the interpretation of their origin.

The axial planes of these folds may be traced up into the Col des Champs where they cut cross section X. They are steep and swing round from E-W striking in the Entraunes Valley to NE-SW in the Verdon Valley. The folds may be traced up through the stratigraphy to the Middle Cretaceous, where they are truncated by sub-horizontal, undulating thrust faults. The lower thrusts imbricate slivers of Lower Cretaceous with Mid Cretaceous shales. Some, at least, branch onto the sub-horizontal tectonic contact at the base of the Upper Cretaceous; others may descend to lower decollements, but the Tithonian is not breached in this line of section. This observation is contrary to the cross section (C) drawn by Lawson (1987).

The Upper Cretaceous limestones for the most part parallel the sharp, tectonic contact at the base of the Upper Crataceous. Few hundred metre wide, intensely kinked zones exist, for example on the SE face of Tete du Lac where the axial planes dip at a low angle to the north-east. This zone comprises a south-west vergent monocline of chevron folds. These structures all detach on the base of the Upper Cretaceous and no equivalent structures exist directly beneath.

Interpretations.

The folds in the Tithonian and Lower Cretaceous developement in part pre-dated the thrusting at a higher level. The structures then developed in tandem, so that the thrust planes were folded by further development of the folds beneath. Lawson (1987) interpreted the Tithonian folds as tip folds, on the basis of an intense cleavage in the Terres Noires confined to the cores of the anticlines: his representation of the fold shape is suggestive of tip folds. However, the outcrop shows more box shaped kink folds which are suggestive of simple buckle folds. An example of such a fold is the Tete de Pibossan recumbent fold. Alternatively, the Tithonian limestone may be folded over a small duplex of Lower Jurassic thrusts, where the Terres Noires form an effective roof thrust.

Overall, the Tithonian datum descends to the south-west, over the interpreted hanging wall ramp of the Pre-Triassic basement thrust (BII-III). The hanging wall cut-off line of the basement thrust is interpreted to lie just in front of the Cairas monocline on this cross section. The zone of intense folding does not extend in front of this line. The presence of this area of more intense shortening and the arcuate fold trends are interpreted to relate to the geometry of the early basement uplift and the consequent deformation of the Triassic decollement.

The presence of Permian mudstone clasts in locally sourced alluvial fan deposits of the Poudingues d'Argens at Quatre Canton and Saussas suggest that basement was uplifted and eroded (section 3.4.4. and 5): to the SW of the Col de La Cayolle outcrop area. The basement high probably extended as a low high to the north-west at this stage. Alpine slip towards the south-west on the Triassic decollement was impeded by the proto-Barrot basement high and its continuation to the northwest. Consequently, the cover "bunched up" behind the breaks and the highs in the decollement surface, the geometry of which therefore controlled the structures in the cover (fig. 2.8b; section 2.6). Later displacement on the basement thrust, towards the south-west, uplifted the earlier formed folds and thrusts.

The cover structures continued to develop after deposition of the Gres d'Annot, because the turbidites themselves have a considerable amount of structural relief and are folded (eg. above the Cairas monocline). However, there is direct evidence for palaeohighs in the area of this intense deformation. The Marnes Bleues thin from the Lance valley towards Tete de Nonciere. The base of the Gres d'Annot Formation is an angular unconformity where it is exposed in the east face of Tete du Moulin de Betrand: a southward component of onlap is seen. The observed angle of onlap is shallow, but then outcrop section is probably oblique to the palaeoslope.

The styles of deformation above and below the base of the Upper Cretaceous are entirely different in this part of the cross section: intense folding beneath, discrete thrusts and narrow kink zones above. The contact is a hanging wall flat along the whole of the exposure. Although the fault remains in the Mid Cretaceous shales along its length in this section, the small thrusts that branch onto it truncate structures beneath. The base of the Upper Cretaceous is the roof thrust to much of the shortening on thrusts in the Haute Provence duplex.

Movement on structures within the duplex did not occur strictly "in sequence". Structures developed over long periods of time, and were reactivated with the same sense of displacement several times since the end Cretaceous: out-of-sequence and in-tandem activity is characteristic of all structures at all structural levels (well demonstrated in both this thesis and that of Lawson, 1987). Therefore, as the Upper Cretaceous was folded over thrust ramps and culminations, it became difficult to transfer displacements on the Mid Cretaceous decollement from structures continuing to develop behind. New thrusts propagated from any sticking points and sliced through the folded and thrust stratigraphy, often cutting down stratigraphy. These thrusts were themselves deformed as the folds continued to develop and tilted towards the foreland above the basement thrust ramp.

The complex structural relationships reflect the long history of tectonic activity on structures in and below the cover.

There is no evidence in the study region that the base of the Upper Cretaceous accommodated large, truely extensional displacements, as suggested by Graham (1981). If there was significant culmination collapse from the Argentera massif, for example, it was accommodated by extensional faults that cut down to the Triassic decollement or below.

The outcrop of Gres d'Annot between Sommet du Lausson and Tete du Moulin de Bertrand was studied in some detail. In this area, Ghibaudo (1985) suggested that the Gres d'Annot onlap a steep sided palaeo-hill below Tete du Lac. The results of the new field work indicate a different interpretation.

Observations.

The Gres d'Annot within 100m of the contact are intensely jointed; further away few joints are seen. To the SE, the Upper Cretaceous, Calcaires Nummulitiques and Marnes Bleues are severely deformed within 250m of the contact. Fig. 2.31c shows the tight, N-S folds in Calcaires Nummulitiques and Cretaceous. At the contact zone, the Marnes Bleues is intensely sheared and cleaved (figs. 2.31d, e). It contains vertical shear zones which all have a dextral shear sense. Near to the contact, there is a SW vergent, strongly kinked monocline in the Upper Cretaceous.

Interpretation.

The contact between the Gres d'Annot and Marnes Bleues is a steep fault, not an original sedimentary contact (fig. 2.31b). The anticline seen to the SE of the fault, below Tete du Lac, does not continue across the fault. The apparent onlap of the Gres d'Annot onto the anticline is purely a perspective effect. The location and geometry of the fault is shown on fig. 2.31a.

3. From Tete de Nonciere to La Batie.

This part of the section crosses the area which is covered by all the cross sections.

It cuts obliquely through the hanging wall of thrust sheet 3a which is uplifted by basement thrusts BII-III. Although the two structures were initiated at approximately the same time, during the transgression (see discussion of cross section A1, B1 and D). This cross section shows that the thrusts above and below the Triassic decollement are not linked.

The Poudingues d'Argens preserved to the north-east of Chammatte is interpreted to derive from the initial uplift associated with the La Valette thrust and kink zone (cross section A1).

The northern limit of Tertiary outcrop in the Villaron syncline shows several isoclinal folds and thrusts. This zone of intense shortening probably developed here because, not far to the west, the Upper Cretaceous had been significantly thinned by erosion over pre-Nummulitiques structural highs. Locally, the Mid Cretaceous shales were also eroded. Slip along the Mid Cretaceous decollement was difficult where it was folded or eroded, and the overlying sequence was weakened by deformation and thinning. Consequently, later thrusts ramped to surface in this area.

4. Tete du Belap to Les Sauzeries Haute (Barreme syncline).

This part of the section cuts the northern part of the Tartonne-Castellane-Sommet de Cremon area. This is the margin of the study region and a brief study of the geometry and evolution of structures there has revealed much about the tectonic evolution of Haute Provence.

The principal structures in the area (fig. 2.32a,b) swing round from NNW-SSE to E-W following the regional trends. The E-W striking segments of the main faults have greater stratigraphic offsets which are consistently of south-vergent, compressional sense. In many places they bring Lower Jurassic and Triassic strata to surface.

The most prominent NNW-SSE striking faults have laterally variable stratigraphic offsets which decrease northwards away from the bend in the fault. They are generally west-vergent, compressional, structures, but locally they are associated with extensional faults: for example, the St. Lions structure (Graham, pers.comm., 1983).

Such considerations led Gracianski (1982) to conclude that the NNW-SSE faults were strike slip, lateral ramps to south directed thrusts, locally modified by diapirism. In contrast, Graham (pers.comm.; enclosure 1) considered them to be frontal ramps to WSW/SW directed thrusting, perhaps led by the conclusive evidence elsewhere for movement in that direction.

The difference in assumed movement direction makes a considerable difference to the inferred thrust displacement, which is much greater if the movement was towards the WSW.

Further clues are provided by the 1982 Digne 1:50,000 geological map and section:

1. The thrust on the eastern margin of the Barreme syncline limits areas with different palaeo-erosion levels. To the east, the Tertiary sediments overlie Upper Cretaceous; to the west, the Upper Cretaceous is everywhere absent and the base Tertiary unconformity cuts down to Mid-Lower Cretaceous and Jurassic. In part, the change in palaeo-erosion level may be related to individual structures at surface, but it is also on a larger scale and so probably results from basement involved faulting, possibly inversion of a large Mesozoic or older fault.

2. The pattern of imbrication of Mid Cretaceous around Barreme indicates that there was a considerable amount of WSW or SW directed movement along that decollement. These structures are truncated by the base Tertiary unconformity so they are pre-Priabonian. This is compatible with cross sections A, A1, X which require early displacement at Mid Cretaceous level under, for example, Colmars. Later, continued WSW/SW directed movement controlled Tertiary sedimentation in the Barreme syncline (Evans, 1987); these sediments were subsequently folded and overthrust.

A new model is proposed which fits these observations better than previous simpler models (sketch cross sections of fig. 2.33 and 34) illustrating the evolution of structures and their relationship to the Tertiary sediments in the area).

Evolution of the area.

1. Mesozoic: Extension on NNW-SSE faults probably continued throughout most of the Mesozoic.

2. Palaeocene: Basement involved uplift, probably by inversion of a west-dipping, steep extensional fault. The Upper Cretaceous was eroded from the uplifted block west of Barreme.

South directed thrusting, ramping from the Triassic detachment to the surface, utilising the NNW-SSE faults as lateral ramps. This is interpreted as backthrusting, in response to Pyrenean compression from the south.

Approximately WSW directed thrusting on the Mid Cretaceous detachment, part of the Alpine system, entered the area from the east. It is not clear which system was responsible for the proto-Digne thrust, whose E-W tip folds are eroded and overstepped by Calcaires Nummulitiques.

3. Eocene: Erosion of structural highs and deposition of the Poudinques d'Argens as alluvial fans fringing the highs.

Continued WSW directed thrusting, propagating onto progressively lower levels down to the Triassic decollement. The NNW-SSE faults which previously were lateral ramps, were reutilised as frontal ramps, resulting in tilting up of the hanging wall to the east of the St. Andre reservoir.

The carbuncular complexity of the St. Lions structure contains successive elements of extension,inversion, strike slip, thrusting and diapirism. Unravelling it is beyond the scope of this thesis!

2.4. The Southern Sector of the Castellane Arc and the generation of the Castellane Arc.

This sector (fig. 2.35) is bounded to the north west by the Rouaine fault zone and its continuation beyond Castellane, to the south by the Maures-Esterel massifs, and to the east by the Arc de Nice. The structural elements are:

- South and north vergent folds and thrusts. The south vergent structures predominate.
- North-south striking extensional faults.
- Small displacement, short, steep NE-SW and NW-SE strike slip faults.
- Minor diapirism, associated with symmetrically vergent compressional structures.

Cross section C (fig. 2.38) illustrates some of the features of the compressional structures. The geological map (fig. 2.35) shows that the large displacement thrust sheet exposed north of Vence, in the Var valley, breaks up westwards into many short length, discontinuous thrust sheets. The extensional faults define narrow, elongate graben. Eocene and Oligocene sediments are preserved in the troughs. The faults are overlain by unbroken Miocene deposits, or overthrust by southward directed thrusts.

There is ample sedimentological evidence for the timing of activity on all the structures in this sector. Giannerini (1980-1981) demonstrated that continental Upper Eocene and Oligocene sedimentation was controlled by the active development of the graben:

- Mega-breccias were shed from the footwalls and preserved in, for example, the Broves graben.

- River channels were confined in, and flowed parallel to, the axes of the graben. The rivers flowed from south to north.

- Progressive unconformities exist in the hanging wall sedimentary sequences immediately adjacent to the fault plane, demonstrating episodic movement on the faults throughout the depositional history.

Stampian sediments are lacustrine, implying that by this time the faults were inactive and that the drainage routes to the north were blocked. This is interpreted to be the result of thrusts overriding the northern end of the graben.

The St.Antonin, Roquesteron and Les Lattes synclines are filled with Eocene and Oligocene sediments that clearly reflect pre- and syn-sedimentary development of the east-west compressional structures. The base Tertiary unconformity cuts down to the base of the Lower Cretaceous locally above anticline axes. The base of the Gres de St.Antonin unconformably overlies a small displacement thrust at Ascros. Successive members of the formation infilled a progressively tightening syncline. Palaeocurrent readings in these sandstones show that currents, for example tides, flowed east-west along the axis of the syncline.

Giannerini (1980-1981) concluded that the extensional faults represented regional east-west extension during the Eocene and Oligocene. The faults were part of the system of graben crossing Europe from the Rhone to the Rhine. In his model, the north-south compression postdated the extension: it was Alpine and synchronous with ~~the followed~~ Miocene sedimentation to the south of Castellane. Giannerini does not acknowledge the equally good evidence for early Tertiary compressional tectonic activity. The evidence lay outside his study region.

Other authors (for example, Goquel, 1963) noted the early north-south compression and interpreted it as part of the Pyrenean foreland system. This model is in part accepted. Note that the structures are predominantly <u>backthrusts</u> and folds with respect to the Pyrenean system. This was probably controlled by inherited structural trends and basement topography:

The Mesozoic extensional faults that defined the margin of the Vocontian basin dipped to the north. When the foreland went into compression in the Upper Cretaceous, these faults were easily reactivated in the reverse sense and thus established the dominance of south vergent structures.

Goguel's model fails to account for the contemporaneous east-west extension, although it is not incompatible with north-south compression. I propose an alternative model (fig. 2.36) which accounts for Eocene and early Oligocene contemporaneous compression and extension. The model draws together the northern and southern sectors of the Castellane Arc and accounts for the abrupt change in structural trends at Castellane. It implies that Haute Provence was caught in a wedge, lying in the foreland of both the Pyrenean and Alpine orogens.

The resulting structural style reflects a combination of controlling factors:

1. Inherited Mesozoic and possibly older structural trends and facies changes across them.
2. Backthrusting of the cover with respect to Pyrenean forces from the

south towards the north.

3. Pyrenean compressional structures at depth which induced basement uplift and deformed the already irregular Triassic decollement.
4. Alpine thrusting towards the WSW\SW.

Throughout the Mesozoic, the Maures-Esterel massif defined the margin of the Vocontian basin and did not subside below platform sea depths. The slopes into the deeper seas to the north were probably defined by north dipping extensional faults.

The facies of Upper Cretaceous sediments show that Maures-Esterel massif, Corsica and Sardinia formed a continuous uplifted isthmus at this time. This may be interpreted as the result of Pyrenean foreland shortening. I propose that Corsica and Sardinia and their lateral extension were thrust over the basement of the southern margin of Maures-Esterel. Activity on such thrusts would cause the uplift required to maintain Corsica and Sardinia as a source of sediment throughout the Eocene. Evidence for such basement thrusts are Klippen at Six-Fours in Provence which preserve basement above Lower Mesozoic strata. Thrusts propagated beneath the original horsts, so that Maures-Esterel remained high relative to the study region.

It has been demonstrated that the Alpine deformation front had propagated into the region as early as the Middle Eocene, producing WSW and SW vergent folds and thrusts. To the north, the thrusts propagated and slipped with ease on the Triassic decollement. Further south, thrusts propagating on this detachment encountered the steep slopes of basement thrust and extensional fault ramps: they snagged on the massif buttress.

The southern sector of the Castellane Arc may be interpreted as a zone of differential shear, between the area of easy slip to the north and the pinned cover to the south-east. The shear stresses were partitioned into south directed thrusting, east-west extension and NE-SW strike slip faults.

The Rouaine fault zone formed at the boundary of this zone of differential shear and thus accommodated a small amount of lateral slip.

Palaeo-magnetic studies have shown that Corsica and Sardinia began to rotate away from Europe during the Lower Oligocene. In the proposed model, the basement thrusts were reactivated as extensional faults. New topographic lows opened up between Corsica-Sardinia and Haute Provence: the Gres d'Annot basin was starved of sediment and progressively, the obstruction to south-west directed thrusting subsided. The extensional faults ceased to be active and were unconformably overlain by Miocene sediment. Alpine thrusting continued unhindered.

2.5. The implications of structures observed in Provence

The following observations are based on the work of Tempier (1987), a study of Marseilles 1:250,000 geological map and reflection seismic data (Hossack, pers.comm.). Provence, in this discussion, is bounded by the Forcalquier fault, the Rhone graben and Mont Ventoux. The structural elements are illustrated on cross sections of Tempier (1987; fig. 2.37):

- Basement thrusts that breach the Triassic decollement, exposed in the south of Provence.
- Tertiary compressional structures that strike east-west and are equally north and south vergent.

- Large areas of relatively undeformed cover sequence, separated by intensely deformed zones a few kilometres wide.

This phenomenon is explained by a close examination of the Mesozoic stratigraphy in relation to these "mobile zones". Rapid changes in the thickness of Mesozoic stratigraphy are apparent on Menard's depth to Pre-Triassic basement map. The thrust and folded zones occur where the Mesozoic is thinnest; the relatively undeformed blocks have Mesozoic successions greater than 11 km thick in places. The thickness of the Mesozoic stratigraphy is related to fundamental extensional faults (Hossack, pers. comm. based on examination of proprietry reflection seismic data, also examined by this author). These faults probably sole out at mid crustal levels.

As a result, there is more relief on the Triassic decollement than in Haute Provence where few major faults control the Mesozoic stratigraphy. The orientation of these faults has dictated the strike and vergence of the Pyrenean-Alpine compressional structures: the result of positive inversion of the fundamental faults. This inversion requires that displacements were transferred from the Pyrenees and the Alps at mid crustal levels. They ramped up the pre-existing faults on to the Triassic detachment and formed narrow, thin skinned thrust belts on the palaeo-horsts.

2.6. The influence of basement involved thrusting on the Gres d'Annot basin.

2.6.1. The Barrot thrust.

The Dome de Barrot is an uplifted massif of "basement", a thick sequence of Permian continental sediments. Goguel (1962, 63) and Graham (pers. comm.) considered that the uplift of the massif postdated most, if not all, of the cover shortening. This is based on the following structural evidence:

i) Structures in the cover are folded up over the culmination. .

ii) The cover to the SW of Barrot has been shortened by 20-30km. Goguel considered that all this movement went through the Triassic above Barrot. Since it could not have passed along that decollement once it had been raised up and locally eroded over the dome, the uplift must post-date the displacement.

iii) The topographic expression of the massif is only 15-20km across. If the limits of the massif relate to the hanging wall and footwall ramps of a basement thrust, the displacement on it should be less than 10-15km. This figure is less than the total displacement to the SW, but it is compatible with the amount of post-Miocene displacement (7km in the Chateau Redon dome).

This hypothesis is not compatible with the sedimentological evidence which indicates that the massif was uplifted and deeply eroded by the time of deposition of the Poudingues d'Argens, well before most of the SW displacement. In addition, the cover sequence, now fringing the basement massif, has not moved relative to it since that time.

This shows that the post-Poudingues d'Argens displacement did not come over the top of Barrot, but it could have come underneath it. Therefore, an alternative model is proposed in which the massif is a far-travelled thrust sheet, and the shortening in front of it roots onto that thrust (cross section C; figs. 2.38,39).

The evidence for early uplift of the massif is as follows:

i) The massif was fringed by alluvial fans (Poudingues d'Argens) which contain clasts of Permian and Lower Triassic (section 3.4.4. and 5).

ii) During the Nummulitic transgression, the massif was also fringed by a siliciclastic shoreline (4.4.6.). Elsewhere, the transgressive sequence comprises carbonate detritus.

iii) The base Tertiary unconformity cuts down stratigraphy towards the massif.

The outcrops, which today surround the Barrot dome, also surrounded the palaeohigh of Proto-Barrot at the time of deposition. Therefore, the cover has not moved significantly with respect to Barrot since the early Eocene.

2.6.2. The Model

Figure 2.39 illustrates the stages in the evolution of the basement structures and their relationship to shortening in the Mesozoic and Tertiary cover.

During the Late Cretaceous - Early Eocene, prior to, and during, deposition of the Poudingues d'Argens Formation, the Pyrenean deformation front shot into SE France at mid crustal levels. The displacements were pinned at least in part, at the southern margin of Pelvoux. Deep seated Mesozoic and older faults began to invert to accommodate the shortening.

It is likely that the Barrot massif represents an inverted Permian sedimentary basin (Goguel, 1962, 63), evidenced by the enormous thickness of Permian shales over Barrot, whereas in the other basement outcrops, Permian is either thin (Maures-Esterel, Argentera) or absent (Barles and Pelvoux).

Inversion of the basin began in the Early Tertiary and continued throughout the deposition of the Tertiary sequence. The suggested mechanism is shown on fig. 2.39: the basin inversion was accommodated on a number of faults, defining separate thrust sheets (BI-BIII). Successive movement of these sheets produced a complex, evolving surface topography.

During the early stages of basement involved shortening, displacements were transferred over the NW extension of Barrot on the horizontal, then gently undulose Triassic detachment. This detachment was steeply folded over the E-W footwall of the earliest inversion and Alpine WSW/SW directed thrusts were inhibited by this oblique ramp. The resulting differential shear produced arcuate folds onlapped by the Gres d'Annot turbidites.

During Oligocene - Recent times (post-Gres d'Annot deposition) displacements could not be transferred through on the Triassic detachment folded over Barrot, as Alpine shortening continued. Mesozoic thrusts bunched up behind the footwall ramp, producing the Tinee nappes, for example. The remaining displacements ramped through the basement and onto the Triassic detachment **in front of Barrot** (fig. 2.38 and 39).

Thus, immediately in front of Barrot, a **strain shadow existed which accounts for the undeformed state of most of the Grand Coyer and Annot areas.** To the north of the basement hanging wall ramps, the Tertiary strata are dissected by thrusts. The position of the lateral hanging wall ramp of basement thrust BII-III along the Beauvezer valley probably controlled the position of the footwall lateral ramp in the thrust sheet transporting the Embrun-Ubaye nappes.

Late movement beneath Barrot itself accounts for late tightening of the St. Antonin syncline, for example, and, according to the present model, does account for some of the steep tear faulting along the Rouaine fault zone along the edge of the BI-II lateral ramp.

Barrot is not greatly uplifted now. Therefore, the large post-Gres d'Annot displacements are taken on thrusts that sole out at a shallow level in the Pre-Triassic basement, interpreted as the base Permian.

Argentera was uplifted late: it deforms the rear of the Tinee nappes; steep, extensional faults associated with its uplift postdate all Tertiary sedimentation and the emplacement of the Embrun-Ubaye nappes. The position of the massif may have been controlled by the Permian basin margin faults in the footwall of the Barrot thrust (cross section C; fig. 2.38).

2.7. Conclusions

1. The whole of south-east France experienced Pyrenean forces from the south, from Late Cretaceous to Eocene times. Displacements entered the foreland at mid crustal levels:

- Mesozoic extensional faults, Permian basin margin faults and other Hercynian basement faults were reactivated as compressional and oblique slip faults. Upper Cretaceous strata were eroded from large areas of Haute Provence as a result.

- Barrot was uplifted by inversion of a steep Permian fault. The compressional fault soled out at the base of the Permian.

- The Mesozoic cover responded by:
  - i) backthrusting above the Triassic detachment in Haute Provence, to the east;
  - ii) by both north and south vergent thrusting and folding along narrow deformation zones in Provence, to the west. The difference in behaviour is interpreted to be the result of lateral changes in the geometry of Mesozoic and older fault systems.

The steep NE-SW Forqualquier fault was an effective transfer fault, compartmentalising the deformation styles. Ancient horst blocks, like Maures-Esterel and Pelvoux, formed buttresses that hindered thrust propagation.

2. The Alpine deformation front entered Haute Provence from the east during the Eocene, immediately prior to the Tertiary transgression. Faults

that were initiated as Pyrenean, south directed thrusts were utilised to accommodate SW directed Alpine displacements. Former lateral ramps became frontal ramps and so on. New SW-vergent structures were established, at levels down to the base Permian, at this early stage. On the eastern margin of Barreme, the new thrusts cut through east vergent inversion structures and the resulting structural relationships are complex.

3. The form of the Castellane Arc was established early, in the Mid Eocene as a product of combined Pyrenean and Alpine compressional forces. It almost certainly represents the inversion of the southern margin of the Vocontian basin, since its curvature follows depth related facies changes in the Upper Jurassic limestones.

The east-west structural trends were also generated when the SW directed Alpine thrusts were impeded by the steep basement slope of the uplifted Corsica-Sardinia and Maures-Esterel massif. When that region went into extension in the Lower Oligocene the impedence was removed. The east-west compressional structures continued to develop because the trend was alrady established, but the along strike extension ceased.

4. Alpine displacements were impeded by the early topographic expression of Barrot and the east-west folds in the Entraunes valley were produced as a result. As the Alpine thrusting extended the area of uplifted basement, the Triassic detachment was folded and ceased to be an effective decollement. Thrusts bunched up behind forming the Tinee nappes; most of the Alpine shortening passed beneath Barrot on the base Permian detachment and surfaced onto the Triassic decollement, and into the Mesozoic, to the south-west of Annot. A broad strain shadow of relatively undeformed cover therefore exists in front of Barrot.

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CHAPTER 3: THE POUDINGUES D'ARGENS FORMATION

3.1. Introduction

The Tertiary sediments in Haute Provence represent the initiation, deepening and subsequent infill of a marine basin. The earliest Tertiary sediments are the continental deposits of the Poudingues d'Argens Formation, deposited prior to and during the Tertiary transgression. The formation is only locally developed, in parts of five Tertiary outcrop areas (fig. 3.1).

Where present, it is unconformable on locally folded and eroded Upper Cretaceous limestones and is overlain by the transgressive carbonates of the Calcaires Nummulitiques Formation. The basal unconformity cuts down locally to Lower Cretaceous, for example, at Sausses. A low-angle unconformity is observed at some localities, an example being Mort de L'Homme (fig. 3.2). Elsewhere the basal contact is sharp and for the most part planar, with erosional relief of <2m.

The Poudingues d'Argens Formation is of variable thickness, commonly less than 30m, exceptionally reaching a maximum of 100m, for example, at the eastern margin of the Barreme syncline.

3.1.1. The importance of the Poudingues d'Argens Formation.

The initiation and early stages of evolution of the S.W. Alpine foreland basin were explored through an understanding of the continental sequences preserved at the base of the Tertiary succession and the transition to marine sedimentation in different parts of Haute Provence.

The study of the continental facies associations and the areal development of the Poudingues d'Argens Formation showed that much of the sediment was deposited on alluvial fans (section 3.4.4). The limited occurrence of the Poudingues d'Argens Formation is interpreted to be a depositional pattern and not one of erosional remnants. Sub-aerial deposition was arrested progressively across the region as the Tertiary sea-level rose.

Soil profiles preserved in the Poudingues d'Argens succession were used to reconstruct the palaeotopography across the region and provided evidence for changing relief during this part of the basin history (section 3.5.4.2). The maturity of the palaeosols implied the length of time required to generate the Poudingues d'Argens stratigraphy. This and the limited areal extent of the continental sediments were used to suggest when the Tertiary basin formed.

The distribution of facies across the boundary between the Poudingues d'Argens and Calcaires Nummulitiques Formations suggests that their depositional environments were linked. The streams of the Poudinques d'Argens locally supplied coarse detritus to the adjacent shores of the Nummulitic sea (sections 3.3 and 4.2.1.4). The relationship between the facies of the Poudinques d'Argens and Calcaires Nummulitiques Formations was used to define palaeoshorelines and the nature of the Tertiary transgression at different times, in different places (section 4.1.2).

3.1.2. Previous Research

The Poudingues d'Argens Formation was described by the authors of the booklets which accompany BRGM geological maps of Haute Provence (1:50,000

Allos, Entrevaux, St.Antonin, Digne; 1:250,000 Nice). For his thesis, Bodelle (1971) examined the southern exposures of the Formation in the St. Antonin and Entrevaux synclines. He noted that the contact between the Poudingues d'Argens and Calcaires Nummulitiques Formations was transitional in places, evidenced by brackish water fauna in lagoonal marlstones beneath the massive, bioclastic shallow marine carbonates.

Published descriptions of the deposits contain details of the sedimentary features at outcrop scale and analyses of the sparse fossil content (for example, Campredon, 1977). None of the previous authors have attempted to reconstruct the basin geomorphology or interpret its climatic and tectonic setting.

3.1.3. Principal components of the Poudingues d'Argens Formation.

The sediments of the Poudingues d'Argens Formation are dominantly **carbonate clastics**, derived from Mesozoic, for the most part Upper Cretaceous, limestones. The source areas were *highs within the Tertiary* basin, produced by Pyrenean-Alpine compressional tectonics. The composition of the Poudingues d'Argens varies between outcrop areas (summarised below). Locally, quartz is a significant component which implies that small amounts of Lower Triassic and/or crystalline basement rocks were exposed at surface.

The Poudingues d'Argens Formation comprises water lain sediments deposited in a continental environment. The sediment body geometries and the relationship between the coarse and fine grained facies are characteristic of alluvial and fluvial sediments. There is abundant evidence that pedogenic processes modified the primary sediment facies (section 3.5):

colour mottling, calcrete profiles, and microcodium. Carbon remnants of root fibres and rhizocretions have been identified. The fine sediments are often well bioturbated, and where single burrows are seen, they were apparently produced by fresh water fauna.

The successions are dominated by conglomerate bodies and marlstones. Sandstones and, even rarer, siltstones comprise less than 10% of the sequences logged.

The Poudingues d'Argens were deposited from flowing water or in shallow lakes. Both the hinterland and substrate were relatively pure carbonates, so it is likely that a great deal of the rainfall seeped into the ground and did not appear as significant surface runoff. Streams may have been produced only during storms, in seasons of high rainfall and for short periods of time. This could explain some of the features of the sediment facies, in particular the polarised grain size distribution (conglomerate and marlstone, but few sandstones) and the absence of evidence for persistent traction currents in the conglomerate bodies (section 3.2 and 3.3).

3.1.4. Age of the Poudingues d'Argens Formation.

Bodelle and others have noted that the boundary between the Poudingues d'Argens and Calcaires Nummulitiques Formations is transitional in places. Therefore, it is possible to date uppermost continental sediments using fauna from the first marine sediments of the Calcaires Nummulitiques Formation (Mid-Upper Eocene from east to west across the region).

The continental sediments themselves do not contain a macrofauna. Samples yield only microcodium (fig.3.3) algae, numerous gastropod operculae and very rare undated, fresh water gastropod shells.

The maximum possible age range of the Poudingues d'Argens formation is Maastrichtian to Lutetian, a period of approximately 15 ma. Evidence of facies transitions between the Poudigues d'Argens and the Calcaires Nummulitiques presented later, argue that the Pouingues d'Argens was deposited in a very brief period, immediately prior to and during the early stages of the nummulitic transgression (section 3.5.4.1).

It is true that between throughout the Palaeocene, S.E. France was subaerially exposed. Evidence includes the extensive erosion of Mesozoic cover from the region around Pelvoux and by the thick continental sequences seen in Provence. In Haute Provence, however, the continental deposits are limited. Over most of the region, the Calcaires Nummulitiques formation lies directly on an Upper Cretaceous substrate that was not severely denuded prior to the transgression. In these areas, there is no evidence that the Poudingues d'Argens Formation, or lateral equivalents, were ever deposited and subsequently eroded.

The evidence would suggest that for much of the 15 million years, Haute Provence was a lowland region with little topographic relief. The structural evidence implies no more than 300m uplift, above inverted extensional faults (section 2.2.4). This relief was peneplained, but the detritus was transported out of the study region. This study suggests that a dramatic change occurred in the early Eocene, when compressional structures developed in the foreland to the Alpine and Pyrenean orogeny. The resulting uplift produced a complex topographic relief. Sediment was shed from the structural highs and trapped in small drainage basins (section 3.6).

3.1.5. Aims and methods of the study.

The Poudingues d'Argens was studied in detail in the outcrop areas of Argens and Peyresq, where the successions exhibit a wide range of sedimentary facies. The results were used to make a comparative study of three other outcrop areas, with the aim of defining the nature of the alluvial systems and the source areas.

The field characteristics of the palaeosols were recorded to establish relative rates of sedimentation across and between outcrop areas. Patterns in the soil drainage were noted, and how they changed spatially and with time.

The sum of evidence was used to define areas of uplift and, where possible, the amount of topographic relief and how both may have changed with time. It was possible to infer levels of erosion in catchment areas, certain topographic slopes and the location of breaks in slope. The sedimentological data was added to the structural cross sections. Structurally defined topographic highs can be identified as probable source areas, using the composition of sediment, palaeocurrent data and the relative proximity of deposits to their source.

The quality of outcrop is variable. Sheep graze actively throughout the summer and autumn and this, together with the dry climate, makes the high outcrop (>1300 m) excellent, for example, at Mort de L'Homme and Quatre Cantons. At Peyresq, conglomerates form well exposed and reasonably accessible bluffs and approximately 30% of the fine grained facies are exposed. At Argens, the exposure is not good, in part because the succession has a greater percentage of fine grained facies. Land is ploughed or forested, and outcrop is limited to patches between bushes and a single good road section approximately 0.5km long (AR5, fig. 3.13).

3.2. The facies of the Poudingues d'Argens Formation.

3.2.1. Introduction.

The conglomerate and fine grained facies dominate the Poudingues d'Argens successions. The conglomerate component occurs mainly in discrete, lenticular or planar bodies enclosed in marlstone. The rare sandstones are almost always transitional from, or laterally equivalent to, conglomerate facies.

The conglomerates appear massive at first sight, lacking in sedimentary structures or internal organisation! They were, nevertheless, studied in some detail, and with information about the sediment body geometry, it was possible to define the types of stream which supplied the coarse detritus. In turn, the stream type and the conglomerate body architecture constrained local and regional palaeoslopes and the nature of the alluvial systems (fig. 3.4).

* In the following sections, the sedimentary facies are described in detail. The conglomerates are returned to in section 3.3., where the sedimentary body geometries are discussed. *

3.2.2. Conglomerates (P1).

The conglomerate facies do not vary a great deal and can be divided into two groups, of which the first describes almost all of the conglomerate observed!

P1.1. Cobble conglomerate with moderately packed, rounded, elongate clasts.

The facies comprises moderately sorted, pebble - boulder, clast - (matrix) supported conglomerate. The average average clast size is cobble or, more rarely, pebble grade. The clasts are rounded - well rounded micritic limestone and marlstone, with angular fragments of chert. In certain outcrop areas (noteably, Sausses and Quatre Cantons), a significant proportion of the clasts are small, purple, mudstone pebbles and fragments of yellow sandstone and occasional iron oxide nodules. The matrix is usually m-crs carbonate sandstone, sometimes as fine as siltstone.

The conglomerates are erosively based: the scoured surfaces are overall flat or gently concave upwards, with locally steep relief (see, for example, fig. 3.5). A single deposit may be a few tens to hundreds of centimetres thick; most are amalgamated.

Many of the conglomerates do not display internal structure. The clasts in each erosively based unit are not graded overall, but the maximum average clast size occurs some distance above the base, and the grain size may decrease rapidly at the top of a deposit. The clasts are usually in a stable position, but clast alignment is rare. At the margins of the conglomerate the cobbles and pebbles are imbricated, with the palaeoflow perpendicular to the channel axis (eg. fig. 3.7). Elsewhere within the deposit, either the 'b' or 'a' axes may be parallel to palaeocurrent indicators, like gutter casts.

Cross stratification can be detected, with set heights >50cm. The commonly low angle surfaces are defined by concave upwards layers of different clast size. They extend from top to bottom of the storey and dip perpendicular and obliquely to the principal current direction. In any unit of conglomerate, the moderately - well stratified sediment is transitional from the massive conglomerate. Note that medium sized bedforms (10-50cm set height) are not observed.

These conglomerates form sediment bodies with a variety of geometries, from ribbon to complex and simple sheets, discussed below (section 3.3). These bodies represent 2-70% of the Poudingues d'Argens sediment volume. The figure varies greatly between outcrop areas (section 3.4), but remains roughly constant overall in any one vertical sequence.

Interpretation.

The conglomerates were deposited in broad, shallow channels, from strong currents. The preservation of steep scours, often eroded into unconsolidated conglomerate, and the lack of clast structure implies that these conglomerates were deposited rapidly. The small coarsening upwards sequence in the basal 10cm of some deposits is characteristic of grain size behaviour in the sheared zone at the base of debris flows or grain flows. However, the cross bedding demonstrates the existence of persistent traction currents. The sum of evidence suggests that much of the conglomerate was transported as high density bed load, deposited from a strong current soon after it had eroded a shallow channel.

Many of the limestone and marlstone clasts are very well rounded. The clasts were therefore transported some distance before reaching their final resting place. Some of the rounding occurred during transport from the erosional highs, but the source areas were no more than a few kilometres from the site of deposition (sections 3.4.4 and 3.6). Therefore, much of the attrition occurred when earlier deposits of conglomerate were reworked, as they obviously were in the amalgamated units. In the very strong flood conditions inferred from the coarse nature of the conglomerates, the clasts could be part of several flows before being preserved in the sequence.

The cross stratified conglomerates are interpreted as the depositional surfaces of bedforms within the channels of the same magnitude as the minimum channel depth (which is defined between storey scours, or from storey scour to a surface of non-deposition). They are therefore the lateral and frontal accretion surfaces of barforms which may have produced a braided flow pattern in the channel, particularly at low stage.

The structures observed in the conglomerates will be discussed again in the context of each sediment body geometry (section 3.3). Suffice to say at this stage, the close association of massive and stratified conglomerate implies that the flow discharge and sediment load varied over short periods of time.

P1.2. A poorly sorted breccia of limestone and flint clasts. The clasts, which range in size from 2-15cm, form an open framework with almost no matrix. Calcite partially fills the voids, some of which is laminar and contains microcodium.

The breccia outcrops at the base of the Poudingues d'Argens Formation in the Douroulles syncline and the top of the Argens syncline. In the former, it is thickly developed (Evans, 1987); in the latter, it occurs as a discontinuous sheet no thicker than 2m, covering an area of <1km . Its base is sharp, but not erosive.

Interpretation.

The breccia is interpreted as a scree deposit, locally sourced from steep slopes of Upper Cretaceous limestone. At Argens, the short (500m+) limb of a SW-vergent asymmetric syncline produced the relief (cross sections B and B1; enclosures 6 and 7). The presence of microcodium suggests that the surface of the scree cone was, in part at least, colonised by plants. Any fine grained matrial was winnowed out by surface waters.

3.2.3. Sandstone Facies.

In the outcrop areas studied in detail, sandstone facies are always associated in some way with the conglomerate bodies.

P2.1 Normally graded, massive - laminated sandstone (calcarenite). The 1-35cm thick sandstones are erosively based and laterally extensive, with width:height ratios of >30, often >100. The beds are lenticular, with upper surface relief. They comprise poorly - moderately sorted, granule to fine sandstone. The beds are often grouped in sequences several metres thick, in which the grain size and bed thickness decrease upwards. The beds are usually separated by marlstone.

The sandstones are commonly bioturbated and display features associated with pedogenesis, like colour mottling. The deqree of modification increases upwards in any sequence.

The sandstones can often be traced laterally into the conglomerate facies P1.1. The beds thicken towards the conglomerate and the degree of pedogenic and biogenic alteration correspondingly decreases. In some cases, the sandstones filled small shallow pebble lagged channels, with irregularly eroded bases. THe units are some 10cm thick and 5-15m wide.

Interpretation.

The sharp based, fining upward units and the extensive modification of the sands by animals and soil processes imply that sedimentation was episodic. The lateral extensions or "wings" of the conglomerate body were deposited from a series of decelerating flows. The flows were qenerated during times of flood, when water escaped the confines of the channel and spread over large areas of flood plain and constructed low anqle levees. The small channel deposits are interpreted as those of crevasse splay channels.

P2.2. Very well laminated, moderately - well sorted, coarse - fine sandstone, which are rich in plant fragments. The sandstones overlie and are laterally transitional from conglomerates; they grade upwards into marlstone. Away from the conglomerate, the well laminated sandstones are transitional into facies P2.1. The base of the facies is erosive, increasing in dip down towards the conglomerate body: the surface is the extension of the conglomerate-filled channel itself (fig. 3.13).

The stratification includes parallel lamination, with excellent pcl, and commonly low angle crossbedding, with cosets preserved. The foresets are asymptotic, and the set bases are lined with granules or small intra-formational pebbles (fig. 3.8,9).

Interpretation.

The sandstones were deposited from persistent and fast traction currents. THe low angle stratification may be a preservation phenomena in places, but the bedform may be interpreted as a low relief sandy bar, with no developed slip face. These bars migrated over inactive conglomerate bars, mantling and infilling the remaining channel relief prior to abandonment. The sandstone represents the low stage usage of a previously high energy channel. *They may be the deposits of a single, large decelerating* current, that eroded the channel, filled most of it with conglomerate and then reworked sands in the remaining depression.

P2.3. **A massive, poorly sorted, pebbly - medium sandstone which is locally preserved within conglomerate bodies.** Pockets of sandstone are rapidly gradational from conglomerate and are eroded by the overlying conglomerate storey base. Once, this facies is seen as the cap to, and lateral equivalent of, a small debris-flow sheet conglomerate.

Interpretation.

The lack of sorting and absence of sedimentary structures suggest that these sandstones were deposited from rapidly decelerating flows. The flows were probably the same as those which deposited conglomerate beneath. The absence of any lamination or cross lamination is noteable. It suggests that the sandstone does not represent the repeated use of the channel by smaller currents, between the major flood events.

3.2.4. The fine grained facies.

The percentage of fine grained facies preserved in the Poudingues d'Argens Formation varies between outcrop areas, from <20% at Le Ruch to >80% of the sequence at Argens. The vast majority of the fine grained sediments are marlstones. However, in certain areas, the composition is markedly different. Noteably, in Quatre Cantons and Sausses, a significant proportion of detritus is quartz and grains other than carbonate.

Over much of the region, the sedimentary facies are monotonous, but the nature and extent of pedogenic modification varies wildly! The palaeosols provide important information about the early stages in the basin development and so are the subject of an entire section: 3.5.

F3.1. The dominant sedimentary facies is **homogeneous marlstone of clay - medium silt grade.** It is frequently impure, with varying percentage (0.1-10%) of quartz and silicate mineral fine sand grains, plus rare mineral grains, eg. rutile. These grain are distributed evenly throughout the rock. The sediment colour varies within and between outcrop areas: white, cream or grey, with orange or buff colour mottles, uniform pale buff and, in a few important areas, **purple** (3.5.6).

No sedimentary lamination is observed, but burrows and preserved branching rootlet traces (carbon filaments) are quite common. Bioturbation and pedoturbation are thought to account for the total absence of sedimentary structures in these sediments. Body fossils are rare - absent, but gastropod opercule are seen. Nodular carbonate and massive, laminated and brecciated micrite occur at intervals throughout any sequence.

The facies forms between 10-80% of any vertical section through the Poudingues d'Argens Formation. The marlstones are laterally very extensive in parts of the succession, and tens of metres thick. Several 5-15m thick packages can be traced across a kilometre or so of the Peyresq outcrop (see fig. 3.10).

Interpretation.

These sediments were deposited in low energy environments which covered considerable areas at any one time. The thick sequences demonstrate that the environment of deposition was stable for considerable lengths of time. The fine grained nature of the sediments implies deposition from suspension, probably in water. Some wind-blown sediment may be present, but there are no sedimentary structures to support this hypothesis.

Facies like that described are deposited on flood plains, in an alluvial setting. They are subsequently modified by animals, plant root systems and other soil forming processes. The principal channels of the alluvial system are interpreted as shallow ones which form short lived braid plains rather than a mature braided river course (section 3.3). During floods, it is likely that flow was rarely confined to channels, and so sedimentation rates could be high. The frequency and quantity of sediment supplied to the flood plain, and the preservation potential of the deposits, was dependent on:

i) the density and proximity of active channels;

ii) the migration and avulsion behaviour of those channels.

The facies of the fine grained sediments are not diagnostic of their position with respect to a sediment source. Where large thicknesses are preserved, the locality may be:

i) in the distal parts of an active system, for example a fan;

ii) lateral to the active part of the system;

iii) recording high rates of basin subsidence and flood plain aggradation, with low channel body connectedness.

To distinguish between these alternatives, the succession at Peyresq was studied: ribbon conglomerate bodies associated with the thick sequences of fine grained facies were examined and compared with complex sheet bodies in the same area. The maximum clast size, clast size range and 'bed' thickness do not change when the proportion of fine grained facies increases.

Model (i) predicts a decrease in the channel dimensions as the percentage of fine grained facies increases. This is not observed, so the thick sequences of marlstone are not the distal equivalents of the sheet conglomerate bodies. In sites lateral to the active channel system (model (ii)), the rates of sedimentation should be more irregular, probably with periods of non deposition, during which colour mottling and mature calcretes might be expected to develop. However, the contrary is true: the calcrete and soil profiles are very immature in the thick sequences of fine sediment, ie. sedimentation rates on the floodplain were high. Therefore, model (iii) is favoured, and a mechanism is discussed in section 3.4.1.2.

P3.2. Laminar micrite that closely resemble calcretes (pedogenic micrite), associated with sometimes inversely graded marlstone.

The limestones have many of the features associated with a laminar calcrete. The original texture of the micrite is never clear. However, algal mounds have been observed (fig. 3.11) and rarely, bedding parallel birds-eye structures reveal an originally high porosity filled by vadose cement. The limestones may be associated with colour ripened and mottled sediment (section 3.5.3), but the detailed facies associations differ from those of a normal calcrete profile (section 3.5.1). The laminar carbonate occurs immediately above sediments that have only developed colour mottles and ill defined peds.

The associated marlstones are significantly modified by pedogenic processes, but occasionally they show subtle grain size variations. These may reveal small (10-25cm) coarsening upwards sequences which have abrupt tops often capped with laminar micrites interpreted as calcretes. High spired, ornamented and small, flat, unornamented gastropods are seen (rarely!).

Interpretation.

A close examination of the micritic limestones and rather subtle variations in grain size of the marlstones suggest that the flood plain may have been submerged below water for some lengths of time so that ephemeral lakes were formed. The limestones are interpreted as freshwater limestones. The marlstones represent the infill of the shallow floodplain lakes by small deltas, crevasse lobes (Collinson, 1986). Once filled, the sites become vegetated and may not receive sediment for some time. Therefore, soil profiles develop, here characterised by calcretes.

Successions with interpreted lacustrine deposits (eg. AR1, fig. 3.23) were laid down in regions of low topographic relief as would be found in the distal parts of the alluvial systems or adjacent to, for example, an alluvial fan. In addition, the successions often have stacked, mature soil profiles which imply areas of low topographic relief, where long periods of non deposition punctuated the sedimentation history.

3.3. Conglomerate body geometries, facies associations and the nature of the Poudingues d'Argens channels.

3.3.0. Introduction.

The following terms will be used to describe the sediment bodies of the Poudingues d'Argens Formation:

Sediment bodies are simple if formed from one depositional event, and complex if produced by multiple fills. The individual event fills are called storeys (Friend *et al*, 1979) and are recognised by:

i) Internal erosion surfaces.

ii) Evidence of periods of non deposition within a body, for example, palaeosols or intense bioturbation.

A body is multi-storey if the stacking is vertical (Friend *et al*, 1979) and multilateral if the stacking is horizontal (Stewart, 1981). Where sediment bodies are seen to intersect, they are said to be amalgamated.

The lateral extent of a channel-related sediment body obviously depends on:

i) how effectively the flow is confined to a channel;

ii) the ability of the channel to migrate laterally. These in turn depend on such factors as flow strength, variability and duration, bank stability and sediment load.

The classification of the Poudingues d'Argens conglomerate bodies used here follows that of Friend *et al* (1979), Atkinson (1983) and Hurst (1983), which classify bodies according to their lateral continuity and internal storey geometry. The types of sediment body observed in the Poudingues d'Argens Formation are shown in fig. 3.4. Most of the data for the following discussion was collected from the Peyresq outcrop area. It applies to the sediment bodies observed elsewhere.

3.3.1. The complex sheet conglomerate bodies

The Poudingues d'Argens Formation in most localities is dominated by complex sheet and ribbon conglomerate bodies. They were constructed by channels which shifted rapidly in a random manner, generating an extensive multilateral sheet body, composed of many 1-5m thick storeys, each tens of metres wide. The dimensions of the bodies vary, with thicknesses from 5-30m. and extremely variable widths from tens of metres up to a few kilometres.

The erosion surfaces at the base of these sheets are generally planar, with small and medium scale, concave-upward irregularities (fig. 3.7 and end 3.1B.). These occur where a storey scour cuts more deeply, and so they reflect the different erosive strengths of individual flows. Deep furrows are common, aligned parallel to palaeoflow.

Internal storey scours are often distinct, marked by grain size differences across the boundary. The surfaces are irregular (fig. 3.12), sometimes steep and locally vertical. The maximum storey dimensions do not vary greatly within each body. No vertical trends in grain size have been recognised in the Poudingues d'Argens, either at the scale of the formation (with the exception of Allons), or within any single body.

Each storey consists mainly of conglomerate facies P1.1, with thin lenses of massive sandstone (P2.3) capping, and well laminated sandstone P2.2 infilling remnant relief at the top of some storeys.

A single storey represents the fill of one channel. The nature of the The geometry of the fill of individual storeys is difficult to define from outcrops in the central parts of the sheet conglomerates, because the upper parts of most storeys are truncated. At the edge of the bodies, the depositional form of the conglomerate is preserved at the margin of each storey beneath sandstone and marlstone. The upper surface of the conglomerate fill is concave or convex upwards, dipping towards the channel plug at angles of between 20' and 30' (fig. 3.15).

Once having seen these depositional surfaces, the eye of faith sees cross stratification elsewhere in the body! In fact, it is well defined by layers of different clast size (eg. PQ1, log E,; E l.. 3.1A). It is described in section 3.2.2. and interpreted as the deposit of large, in-channel conglomerate bars. The bar slipfaces dipped down, and oblique to, the palaeocurrent directions defined by gutter clasts. Such barforms probably produced a braided river pattern, particularly at times other than peak flood.

Some sedimentary features of the remaining conglomerate facies should be re-emphasised:

i) the conglomerates are, for the most part, massive with no clast imbrication.

ii) 10-20cm coarsening upward layers exist at the base of many depositional events and the clast size decreases abruptly at the top. Otherwise, there is no grading in these conglomerates, although the clasts are moderately well sorted.

iii) Clast alignment is rarely observed on the base of the conglomerate bodies. Where it is seen, the long 'a' axes are often aligned parallel to flow: a feature which characterises flows with high sediment load densities.

iv) Internal storey scours are commonly steep-vertical over much of the height of a storey (fig. 3.7)

Interpretation of the facies associations.

The prevalance of massive conglomerates imply that a large part of the coarse detritus was rapidly deposited from impersistent flows with high sediment density, such as could be generated by a river in flood. As the flood abated, the river's sediment load and discharge rate lessened. Some of the channels were quickly abandoned; but in others, the flow persisted, and stratified conglomerates were deposited adjacent to, and above, the massive facies. It is probable that the bar form cores were constructed during the peak flow, and lateral and frontal accretion occurred as the flow waned.

Massive and laminated sandstone draped the barforms. Palaeocurrent readings are obtained from well developed pcl and crossbedding in facies P2.2. They demonstrate that flow was highly variable within a single wedge of sandstone, and always at an angle to the palaeocurrents derived from eroded grooves on the base of the sediment body. These data show how the current was diverted around the barforms as it continued to wane.

All the facies seen may be interpreted as the deposits of flows lasting a few hours or days at most. There is no clear evidence that the channels were used repeatedly, and no sediments are preserved which imply that a low stage river existed in any of the channels. The implications of these statements are explored in the following discussion.

The nature of the flows which generated the complex ribbon conglomerate bodies.

The complex sheet conglomerate bodies are the deposits of bedload dominated, braided channel systems. The sheet body geometry demonstrate that the channels were laterally unstable, forming braided multi-channel rivers or a braided channel complex.

The limited facies associations indicate that the channels were ephemeral. Most of the channels were filled during one flood event: mature, far travelled bar forms are not observed. The predominance of coarse conglomerate bodies with little internal structure, associated with thick sequences of marlstone suggests that the channels were incised and filled only during sizeable floods. This could be accounted for by the nature of the carbonate hinterland in the following manner:

The source areas are interpreted as local highs above actively developing compressional structures. The highs were of limited extent, and the rivers did not drain large areas: the streams were short, and therefore peak flood discharge rates decayed rapidly. More importantly, the Upper Cretaceous limestone-marlstone sources were kinked and fractured at and below the surface (Cairas; see section 4.5.2). As a result, the most of the rainfall penetrated the ground surface via fractures and karstic fissures and did not appear as surface run off in places like Peyresq. During storms, streams did develop and they transported conglomerate grade debris from the broken terrains to their present resting places. It is worth noting that the highly episodic sedimentation that characterised the depositional systems of the Poudingues d'Argens may be a feature of other carbonate clastic alluvial systems.

The immaturity of the Poudingues d'Argens channel system would imply that deposition occurred on an alluvial fan or in the proximal parts of a braided river (Miall, 1978; section 3.4.4.). There is no evidence within any body to suggest that a well developed river course existed .

3.3.2. Ribbon conglomerate bodies

Ribbon conglomerate bodies are not common within the Poudingues d'Argens sequences observed. There are two forms:

1. Complex ribbon bodies in which multilateral stacking of storeys dominates, but which show no systematic stacking direction (fig. 3.17,19).

2. Simple ribbon bodies.
   These are commonly smaller than the dimensions of preserved storeys within complex bodies and usually have smaller clast sizes, and sandstone forms a significant percentage of the fill (fig. 3.).

3.3.2.1. Complex ribbons.

There exists a spectrum of geometries between complex conglomerate ribbon bodies and complex sheet bodies. The facies and clast size range within the bodies are similar; the individual preserved storey dimensions are the same, and the maximum body thickness measurements are similar. Therefore, the palaeoflow conditions generating and filling the channels can be assumed to be similar.

Modifying the conclusions of Friend _et al_ (1979), a ribbon body is produced rather than a sheet body if:

i) The channel system was laterally confined.

ii) The bodies represent short lived channel systems or ones which avulsed frequently, so that the channel system migrated a limited distance before abandoning that course. This channel behaviour characterises flash flood regimes.

iii) The alluvial area subsided significantly, or was uplifted relative to some base level of erosion.

Laterally confined channels.

In the outcrops of Poudingues d'Argens studied, there is nothing to suggest that the channels were incised into sediment any more cohesive than that eroded by the channels of the sheet bodies. There are two localities in the Peyresq outcrop area where normal faults were active during deposition of the Poudingues d'Argens and here there is evidence that the channels repeatedly occupied a narrow zone in the hanging wall of the fault (section 3.4.1.2). These channels were, effectively, confined by the topographic low adjacent to the fault.

Flash flood regime.

It is quite probable that the region was characterised by a flash flood flow regime. However, it does not explain why ribbon bodies were generated rather than sheet bodies in the Poudingues d'Argens. It has already been noted, for the complex sheet bodies, that the individual channels were often eroded and filled during single events and that this, and the lack of conglomerate facies sedimentary structures, suggest that the channels did not constitute well established river systems. The flows which produced conglomerate sheet and ribbon bodies were not dissimilar.

Uplift or subsidence of sediment surface relative to the regional base level.

Uplift of a depositional surface relative to the base level of erosion will induce streams to erode deeper into alluvial plain sediments. However, changes of elevation between the alluvial system and base level of erosion would change the local gradients and therefore change the nature of sediment supplied by channels to a given area. No changes of grain size, texture or maturity are observed between the types of body.

The different behaviour of the channels can be explained by differing rates of flood plain aggradation induced by subsidence of an area relative to the base level. When subsidence rates were low, the channels shifted position across a stable sediment surface for long periods of time and generated multilateral sheet bodies. With increased rates of subsidence, the flood plains were accumulating sediment rapidly and channels migrated across an evolving land surface. When they returned to their earlier sites, large amounts of fine grained facies had been deposited over the previous channel fill. Ribbon bodies were therefore preserved.

The ribbon bodies amalgamate upwards with complex sheets as would be predicted by the changing subsidence model. Note that the opposite effect would be produced if the source area were uplifted: the resulting increase in sediment discharge to a given area would affect a change from ribbon body to sheet body geometry with a given rate of subsidence. However, this may be associated with changes in sediment grain size and maturity.

The subsidence model is used to explain the large scale cycles in sediment body geometry at Peyresq (3.4.1.2.).

3.3.2.2. Simple ribbon bodies.

Included in this group are two amalgamated bodies which appear to have one or two storey scours within them. However, each body is of very limited lateral extent and does not warrant a separate classification (Locality AR5, fig. 3.13,14).

Description.

The simple ribbon bodies range upwards in size from tens of centimetres thick and a few metres wide to the dimensions of individual stories within the complex bodies. They occur in parts of the section dominated by fine grained facies deposits.

The larger bodies have simple lenticular geometry, often with low width:thickness ratios. The erosive bases are broad and concave upwards, often decorated with grooves. The larger bodies comprise poorly structured conglomerate facies similar to those seen in the complex bodies. The maximum clast size decreases with the size of the sediment body. The degree of sorting correspondingly decreases, reflected in greater percentages of sandstone matrix. The conglomerates rapidly fine upwards to marlstone within centimetres of the top of the body.

The smaller bodies have high width:thicknes ratios. There is a full spectrum of geometries between these smaller ribbons and simple sheet sandstone bodies bodies with irregularly scoured bases. They are often laterally equivalent. The smaller ribbon bodies have complexely scoured bases (fig. 3.15), with the larger clasts concentrated in the hollows. The palaeocurrent data and cross sections through the small bodies show they were oriented perpendicular to the major channels.

Interpretation.

The bodies represent the fill of very short lived channels. The larger bodies are the same size as storeys within complex bodies. The facies, and therefore the sedimentary processes, were the same in both. The channels which they represent probably formed an anastomosing system which, given time, would migrate to give rise to complex, multilateral ribbon bodies and eventually complex sheet bodies.

The conglomerates of the smaller bodies were deposited from mixed suspended and bedload laden water which covered a wide area (up to 30m across). High velocity zones scoured a braided pattern of very shallow channels, in which were concentrated the bedload of coarse sand, pebbles and cobbles. This gave rise to the lateral segregation of grain size. As the flow strength waned, sand and marl were deposited from suspension across the whole area.

The smaller bodies were formed by smaller channels, laterally and distally equivalent to the major channel systems. Many are interpreted as crevasse splay channels, because they are at high angles to the major channels and are equivalent to sheet sandstones of flood flow origin.

3.4 **Comparative study of the Poudingues d'Argens Formation outcrop areas.**

3.4.1 Introduction

The available outcrop of the Poudingues d'Argens Formation in Haute Provence is too restricted to warrant a sophisticated analysis of the alluvial architecture of the channel systems. However, in several outcrop areas, the exposure of the formation is laterally continuous over several kilometres. In these areas it is possible to study the sediment body geometries in two, and sometimes three, dimensions. Their distribution in vertical sequences can be compared and their lateral extent can be defined.

The type of alluvial system supplying sediment to an area can be ascertained using:

i) The geometry and architecture of sediment bodies.

ii) The facies associations within these bodies and the nature of the fine grained facies associations of the overbank environment.

The identification and correlation of sequences or cycles at several scales has been used to further constrain the alluvial model and define probable controls that operated within the drainage basin and source area (e.g. Steel 1978, Miall 1978, Heward 1978, Atkinson, 1983).

In areas of good outcrop, it is possible to identify consistent changes in the deposits of an alluvial system in the direction of the principal palaeocurrents. The trends provide information about the system as a whole.

The following sections discuss the type of alluvial system represented by the Poudingues d'Argens succession. Where possible, the links between outcrop areas were defined. Thereby, estimates were made of the number of systems, drainage basins and source areas in this study region.* Evans (1987) has extended this study to equivalent outcrop areas to the west.

3.4.2. Le Ruch - Peyresq.

3.4.2.1. Introduction.

Conglomerate facies make up to 90% of any vertical sequence at Le Ruch. Viewed from a distance (fig. 3.16), the succession is seen to comprise amalgamated sheet bodies. The maximum clast sizes are commonly > 50cm, greater than those at Peyresq, where the maximum = 30cm.

At Peyresq, down palaeocurrent from Le Ruch, the succession comprises complex sheet conglomerate bodies, 10-30cm thick of great lateral extent (> 1km) which are separated by fine grained overbank facies and complex ribbon conglomerate bodies. The percentage of conglomerate in a vertical sequence ranges from

*See FIG 3.33

30-70%.

Le Ruch and Peyresq were probably part of the same alluvial system. The outcrop areas are separated by only two kilometres, and palaeocurrent readings at Peyresq imply that Le Ruch was between it and part of the source area.

In the following sections, features of particular interest at Peyresq and Le Ruch are described.

3.4.2.2. Palaeo-depositional surfaces.

The base of the Poudingues d'Argens Formation is a low angle unconformity at both Peyresq and Le Ruch (fig. 3.2). Within the formation, surfaces like the base of the conglomerate bodies, downlap on the unconformity, at angles of 15' or less, towards the south and southwest. These surfaces dip in roughly the same direction as the palaeocurrents measured in Peyresq,that is towards the SW. Along the Mort de L'Homme ridge, the surfaces are storey boundaries which define wedges of conglomerate that thin down towards the unconformity and are asymptotic onto it.

There are two interpretations of the downlapping strata which have important implications for reconstruction of the palaeo-relief during deposition of the Poudingues d'Argens. The conglomerates may have been deposited on a slope, for example, on a fan surface. Alternatively, the beds were originally horizontal, onlapping a tilted unconformity and have been subsequently tilted.

The first model is favoured, because it supports the sedimentological evidence that the conglomerates were deposited from shallow streams on the surface of an alluvial fan (section 3.4.4). The wedges of sediment between the dipping surfaces represent phases of fan progradation. The structural evidence also supports this model: structures that were initiated and eroded early in the Tertiary continued to be active throughout the Tertiary and accentuated original surface dips. This applies to the Melina-Aurent kink zone beneath Peyresq (eg. cross section E). Structural dips were not reversed, as required by the second model (cf. fig. 2.25).

3.4.2.3. **Unusual conglomerate body geometries at Peyresq.**

Most of the complex sheet and ribbon conglomerate bodies have randomly stacked individual storeys. There are two exceptions:

1) **The lowest conglomerate body at PQ1**, east of the village, is a laterally extensive sheet body whose overall dimensions are similar to other bodies at that locality (15m thick, 100m wide). It is unusual because the storey boundaries within it cut from top to bottom of the sediment body and dip consistently towards the west (Encl. 3.2) at a low angle to the body boundaries. In all other bodies, the storey boundaries are closer spaced and

randomly stacked.

The systematic arrangement of the erosive channel bases demonstrates that a single, deep (15m) channel migrated consistantly in one direction. It is unusual for two reasons: (i) its apparent depth; (ii) that the sediment was supplied via a single channel in this case, not the braided systems interpreted elsewhere. The internal facies are the same in both types of body.

Given that a single channel produced the sediment body, then the migration behaviour may be explained, without call to extrinsic controls. The conglomerate facies suggest that the channel was occupied episodically, probably during severe floods. Each successive flow sought the pre-existing topographic low of the old channel. It was able to erode and preferentially scoured into the marlstone bank without removing significant amounts of previously deposited conglomerate (~~fig. 3.~~). As a result, the channel migrated with time, preferentially in the same direction and extended the width of the sediment body.

The existence of a single, large migrating channel at the base of the Poudingues d'Argens Formation, and shallower braided systems above, may reflect an increase in topographic relief. This could represent the build up of an alluvial fan surface, in which case, the change in sediment body geometry may be expected to be gradual which it is not. Alternatively, the changes in topographic relief could be a response to structuring and uplift of the hinterland. The second hypothesis accords with independent structural and sedimentological evidence of syn-sedimentary tectonic activity.

2) **There is direct edvidence for extensional faulting during deposition of the lower part of the succession.** This is best seen on the east face of the Peyresq plateau (fig. 3.18), where, in the hanging wall of a normal fault, complex ribbon conglomerate bodies are vertically amalgamated to form a discrete unit of limited lateral extent (approximately 300m).

Such a relationship is commonly seen in fluvial sediments, where river channels tend to be located in the structurally induced depression in the hanging wall of a normal fault (eg. Leeder and Alexander, 1987).

A similar pattern is seen on a larger scale. To the west of Peyresq, the sequence is dominated by conglomerate (up to 90%),* compared with 50% at the village. The highest concentration of conglomerate bodies occur adjacent to a set of steep, N-S faults with present day extensional geometry. Much fault movement post dates the Tertiary rocks exposed, but the coincidence of faults and conglomerate bodies implies that this region was a preferential site for the channels. It must therefore have been continuously subsiding during deposition of the Poudingues

* FIG 3.19

d'Argens. The outcrop is a direct analogy of theoretical predictions of alluvial architecture developed in the hanging wall of an active half graben (Bridges and Leeder, 1978).

3.4.2.4. **Megasequences observed at Peyresq.**

The succession at Peyresq is interesting because **no overall trend in grain size is detected in the sequence. In a single vertical sequence, coarse member - fine member alternations occur on the scale of 10-30m.** The boundaries between the marlstone and conglomerate associations are erosional or rapidly gradational.

The alluvial architecture diagrams of encls. 3.4 reveal that crude cycles do exist. Marlstones, with complex ribbon bodies, grade upwards into laterally extensive complex ribbon bodies with less marlstone, which give way to a laterally very extensive, complex sheet body. This sequence is repeated. Three extensive sheet bodies can be traced around the whole outcrop area.

Individual storey dimensions are the same whatever the sediment body geometries. The maximum clast size in the conglomerate bodies does not change. Inferred sedimentation rates were continuous, as suggested bythe lack of pedogenic modification, and probably high. The only change in the system is the evolution of the conglomerate body geometry, from ribbon to complex sheet, or vice versa.

In general, progradation (or retreat) of alluvial fans produces coarsening-upwards (or fining upwards) sequences, by stacking vertically progressively more proximal (or distal) parts of the alluvial fan. Since sequences in the Poudingues d'Argens do not have this character, they were probably not produced by progradation and retreat of alluvial fans.

The changes in sediment body geometry could instead be produced by changing the rate of flood plain aggradation. With high rates of fine grained sediment accumulation, ribbon bodies tend to develop. This could be the result of increased subsidence related to tectonic activity adjacent to, and within, the basin.

The model:

In this area, both Alpine and Pyrenean compressional forces were experienced. **The sequence of events post dating an episode of crustal shortening** are interpreted as follows:

There was an immediate, and **regional, subsidence** as an isostatic response to the structural loading. Opposing this, the area **above the growing structure was uplifted.**

i) The alluvial fan responded immediately to the greater rate of subsidence: flood plain aggradation rates increased and ribbon conglomerate bodies were preserved in large thicknesses of overbank fine grained facies.

ii) The uplifted area was weathered and conglomerate clasts were generated and transported onto the alluvial fan. The rate of flood plain aggradation slowed as an equilibrium fan profile was established. Correspondingly, the channels were able to migrate more freely, and so a sheet conglomerate body was produced.

In this way, three features of the Peyresq continental sequence are explained:

i) The systematic and cyclical changes in conglomerate body geometry;

ii) The maintenance of a good supply of large conglomerate clasts throughout deposition of the Poudingues d'Argens Formation;

iii) The great thickness of succession preserved at Peyresq. relative to Le Ruch and Allons. Le Ruch was involved in the localised uplift and Allons may have been progressively shielded from the major supply of detritus by another active structure; for example, a laterally propagating, therefore low relief, fold.

3.4.2. Allons and Argens.

The succession at Allons fines up overall: at the base, sizeable ribbon conglomerate bodies are seen, comprising approximately 50% of the succession. 20m higher in the sequence, the percentage of conglomerate rapidly diminishes, and a **thick sequence of marlstones, with coals and fresh water fauna is eventually transitional into marlstones with brackish water fauna at the base of the Calcaires Nummulitiques Formation.**

To the north, the percentage of conglomerate decreases, until at Argens, thin conglomerates are rare in a sequence dominated by marlstones. Lacustrine carbonates are preserved in the lower part of the formation. Stacked, mature calcrete profiles demonstrate that this locality did not receive sediment continuously.

At Argens, **scree deposits** are preserved at the base of the succession, which demonstrate that steep slopes existed. The slopes did not, however, supply large volumes of sediment. The **palaeoslope is interpreted as the steep limb of a SW-vergent asymmetric fold pair,** the Argens fold (cross section B1). The scree cones are preserved in the core of the syncline, while the Poudingues d'Argens Formation is preserved only in the lower shallow dipping limb of the present-day fold. These two observations imply that the fold existed during the deposition of the Poudingues d'Argens Formation and gave rise to steep SW facing slopes at Argens that were less than 300m high.

The Argens fold pair opens southwards and the lower relief broader hills may have been the source of conglomerate for Allons. It is possible, however, that the fold did not extend so far south at this time, in which case the gravel streams came to Allons via Peyresq, and the low relief hills simply prevented any coarse detritus from reaching Argens.

The **lacustrine sediments at Argens** in the lower part of the succession at Argens are equivalent to the rapidly deposited conglomerate channel fill sequences at Allons. Argens was on the periphery of this channel system.

The lakes developed in an area of poor drainage where water could only be removed by evaporation and downward percolation through the sediment. Inspection of the structural cross sections Z and B, in particular, shows that compressional structures generated topographic highs to the W, N and E of Argens (enclosure 6.1). **Argens was effectively ponded by these highs and the developing alluvial fan to the south at Allons.**

Upwards through the succession the situation is reversed. Well drained mature pseudo gley soil profiles are preserved to the north of Argens (AR1). Subsidence and sedimentation rates increase through the succession to the south at AR5, and gley palaeosol profiles indicate high water table levels. At Allons, a thick fining upwards Poudingues d'Argens succession is overlain by a thick sequence of brackish lagoon sediments (basal Calcaires Nummulitiques Formation). **The interpreted palaeoslope prior to and during the nummulitic transgression was therefore down from north to south.**

The change in surface slope between Argens and Allons prior to, and during, the transgression cannot simply be explained as the effects of regional subsidence. Argens remained elevated while Allons subsided. The model that was used to explain the cyclicity observed in the Peyresq continental succession can be applied here. Regional subsidence was countered by localised uplift above actively developing compressional structures. In this case, the culprit was probably a thrust sheet of Lower Jurassic stratigraphy (thrust 3b; cross section Z; section 2.3.2.7.). **The topographic slope from north to south was the lateral hanging wall ramp of the thrust sheet.** Continued development of the same structure gave rise to hills that supplied the small fan deltas interpreted in the Lower Calcaires Nummulitiques at Argens (section 4.4.3.).

3.4.4. The relationship between the Poudingues d'Argens at Peyresq - Le Ruch and Argens - Allons.

Le Ruch and Peyresq are interpreted to represent the proximal to mid-fan parts of a sizeable alluvial system, probably a fan (section 3.4.6). The total size of the system cannot be determined, but **the extent of the sheet conglomerate bodies implies that braid plains several kilometres wide were active at any one time.**

The outcrops at Allons and Argens may, or may not be linked to the system at Peyresq. They are geographically close and compositionally identical, comprising reworked Mesozoic limestones. The palaeocurrent data from conglomerates at Allons suggests that the coarse detritus may have come from the same source area, so that Allons represents the more distal equivalents of the same system. However, the local palaeotopography suggests that nearby smaller hills could equally well have controlled the facies distribution in the Allons - Argens syncline and even supplied coarse detritus locally.

The interpreted palaeogeography of the area is summarised in encl. 6.1.

3.4.4. Quatre Cantons. (FIG 3.20)

Conglomerate bodies at ~~Argens~~ Quatre Cantons display similar facies associations, but they can be grouped into two compositionally distinctive types. The first contain only fragments of Upper Cretaceous limestone; the second contain clasts from the whole Mesozoic sequence, plus **pebbles derived from Permian mudstones.** (FIG. 3.21 a, d.)

Palaeocurrent indicators are scarce. There is no consistancy and no polarity to be derived from the grooves on the base of some bodies. The channels containing Upper Cretaceous limestone clasts have cross sections which suggest that flow was roughly

E-W; the remainder are elongate N-S. Therefore, it is probable that the local palaeotopography was complex and that two sediment source areas existed, producing two alluvial systems which coalesced at Quatre Cantons.

The Permian mudstones were probably sourced from an early expression of the Barrot massif. If this were the case, it implies that Quatre Cantons has not moved far with respect to Barrot, a hypothesis which has a great impact on tectonic models for the region (section 2.6). There are two reasons why the Permian mudstone clasts are not believed to be derived from far a field:

i) The clasts occur in channel conglomerate bodies as pebble sized clasts and within the matrix fine grained component. Mudstone clasts cannot be transported long distances transport, without disintegrating. This is well demonstrated in present-day Haute Provence, where the bedload of gravel streams flowing off the Barrot massif become carbonate-dominated within a few kilometres of the mudstone outcrop.

ii) If the sediment was deposited on part of a large fan system, then the purple coloured fine grained facies characteristic of Quatre Cantons (section 3.5.6.3) would be observed over a large area at several outcrop localities. Instead, the distinctive colouration is only observed at

Quatre Cantons and Sausses, the two outcrop areas adjacent to the present Barrot structural high.

3.4.5. Sausses (Barres de martignac).

The outcrop is very limited to a small area close to the complex of faults associated with the St. Benoit structure. The barres de Martignac outcrop is a precipitous cliff which can only be studied from a distance, or from extremely close quarters (while hanging on by your teeth). From this vantage point, the Poudingues d'Argens closely resemble the continental sediments at Quatre Cantons, particularly in respect of colour and composition.

Sausses, like Quatre Cantons, is adjacent to Barrot at the present day. The heterolithic source of detritus is interpreted to be the early topographic expressionof Barrot. In section 2.6, it has been demonstrated that significant uplift could be achieved in the early Tertiary without large amounts of shortening, by inverting a steep extensional fault, interpreted to lie at the margin of a small Permian basin.

3.4.6. St Antonin/St Pierre.

The booklet accompanying the Roquesteron 1:50,000 sheet (BRGM, 1978) states that the Poudingues d'Argens at St.Pierre contains clasts of crystalline and metamorphic basement, probably derived from the Maures-Esterel massifs to the south. The nature of the alluvial system is unknown.

3.4.7. The nature of the Poudingues d'Argens alluvial systems.

The Poudingues d'Argens Formation is interesting for three reasons:

1. Most of the sediment is carbonate clastic, derived from a hinterland of well-bedded Upper Cretaceous limestone and marlstone. The source areas were actively being deformed and uplifted during deposition of the Poudingues d'Argens.

2. The sequences are dominated by conglomerate and/or marlstone. Sand grade sediment is relatively rare.

3. The outcrop of Poudingues d'Argens Formation is interpreted as a depositional pattern, not one of erosional remnants. The evidence for this is the nature of the transitional contact between the Poudingues d'Argens and the Calcaires Nummulitiques Formations. Very little of the Poudingues d'Argens was reworked during the transgression (sections 4.2 and 4.3.3.).

The first two points noted above are indicators of the behaviour of the Poudingues d'Argens alluvial systems. They do not define its nature.

However, certain features of the Poudingues d'Argens successions studied can be used to define the alluvial systems which generated them: the limited distribution of the Poudingues d'Argens Formation is one of these. This will be discussed, together with evidence from within the sediments themselves:

i) The geometries of the conglomerate bodies demonstrate that the system was dominated by channelised flow. Debris flow deposits are not common and sheet flood sediments, other than thin pebble-lagged sandstones, are rarely observed.

ii) Most of the conglomerate bodies have **complex sheet geometries.** The individual storeys rarely exhibit a preferential stacking direction. The sheet bodies are laterally extensive, exceeding several kilometres in some cases. Therefore, they are interpreted as the deposits of **multiple braided channels which formed braid plains.**

iii) The facies within the conglomerate bodies suggest that individual channels were not used repeatedly. **Most of the conglomerate was deposited rapidly from high density flows.** Where bar forms existed they are not far travelled or well structured. The channels were almost all **shallow (less than 10m deep): they were ephemeral, filled during single, waning events.**

iv) The successions at most outcrop areas comprise > 40% conglomerate facies, in fact up to 95% of the sequence at Le Ruch. Rust (1979) states that high percentages of conglomerate within a continental sequence imply the existance of high relief between the source area and the site of deposition.

In this case, the predominance of conglomerate and fine grained facies associations may also reflect the nature of the hinterland. The Upper Cretaceous limestones deformed mainly by kinking. This brittle process fractured the limestones. Much of the seasonal rainfall probably drained into the fractured and probably karstic substrate and did not appear as surface run-off. It is likely that streams only developed during severe storms. The disorganised, but coarse grained nature of the conglomerate bodies can be then explained: the flows were fast and furious, able to carry boulders and cobbles long distances, but were very short lived.

v) The fine grained facies of the overbank environment do not contain evaporites and mud cracks are not observed. The palaeosol facies imply that surface temperatures were seasonally high and that the sediment was alternately wet and dry. These palaeosols do not preclude high, <u>seasonal</u> rainfall.

In fact, the large amount of very coarse material deposited from fast flowing streams requires a significant rainfall at regular intervals.

vi) At a few localities (eg. Peyresq, section 3.4.2.2.) surfaces within the succession, such as the base of conglomerate bodies, can be seen to downlap onto the unconformity at the base of the Poud d'Argens formation. These may be interpreted to reflect the surface relief of a depositional system, for example an alluvial fan.

Can the nature of the alluvial systems be deduced from the available evidence?

The limited occurrence of the Poudingues d'Argens Formation and its local distribution fringing the interpreted structural palaeohighs strongly suggests that the principal sites of deposition were alluvial fans. This is supported by the evidence at Le Ruch and Peyresq for primary depositional sufaces which dipped down the local palaeoslope.

The alluvial fans were stream dominated, and the channels which flowed across any fan surface did not evolve to form mature rivers. Many of them were cut and filled in a single, waning, short lived event. Such stream dominated fans are characteristic of drainage areas with intermediate to low relief. They can be contrasted with the mass flow, sheet flood dominated marginal fans of steep fault bounded basins (Collinson, 1986). The development of stream dominated fans is compatible with the Poudingues d'Argens source areas being interpreted as highs above compressional structures. For the most part, the low angle thrust faults and associated folds actively developing at this time (eg. the Mid Cretaceous duplex in the Beauvezer valley; cross section A1 and D) would have generated broad areas of significant uplift. Locally steep slopes may have existed adjacent to inverted extensional faults.

The three dimension form of the fans cannot be precisely defined because outcrop is limited. However, the patchy outcrop of Poudingues d'Argens Formation certainly suggests that there were several fans of limited aerial extent.

3.5 The Pedogenetic Modification of sediments in the Poudingues d'Argens Formation

3.5.1. Introduction

Soils are very sensitive indicators of the environment in which they form. Ancient soil profiles are commonly preserved in the geological record of continental sediments, and their attributes may be used to reconstruct the basin morphology and evolution. For example:

- The thickness of a preserved soil profile and its mineralogy record differences and changes in altitude of the sediment surface relative to the local water table. The nature of each soil provides evidence of palaeoslopes within the basin. A sedimentary sequence containing many profiles records the history of subsidence within the basin.

- The mineralogy of a soil is particularly sensitive to the salinity of pore waters and can be used to locate palaeo-shorelines.

- In the Eocene palaeosols of the Poudingues d'Argens Formation, the host sediments are commonly very pure limestones. The colour of the soil profiles is very sensitive to small amounts of other minerals and can be used directly to identify sources of sediment other than Mesozoic limestones.

The process of transforming a sediment into a soil profile is termed "pedogenesis" (Buurman, 1975). The study of ancient soil profiles is complicated by the effects of "compound pedogenesis" due to soil superimposition in a sedimentary sequence (Bown and Kraus, 1981), including the recent weathering of outcrops. All soils are internally stratified, and where several are superimposed in a sedimentary sequence it may be difficult to identify individual soil profile boundaries. In addition, during burial and uplift of the sediment, diagenetic processes modify and may obscure the original soil characteristics.

3.5.1.1 Brief historical review of research on palaeosols.

In the past, the study of ancient soils, called "palaeosols" (Buurman, 1975), concentrated on the origin of the red colouration in continental sediments and the development of nodular carbonates "calcretes" (Brewer, 1964). Prominent features like organic content and authigenic clays were described. The palaeosol characteristics were used to infer palaeoclimate and rates of floodplain accretion (for example, Allen 1960, 1974, Steel 1974, Leeder 1975, Wright 1980, 1982).

Recently, the techniques used by modern-soil scientists have been carefully applied to palaeosols (e.g. Buurman 1975, Wright 1982, Atkinson 1983). The information has been used in some cases to reconstruct basin histories. For example, Atkinson has successfully deciphered the soil history of the Tremp-Graus Miocene continental basin, Southern Pyrenees, and used the information to define rates of subsidence and phases of tectonic activity in, and adjacent to, the sedimentary basin.

Most authors now use US soil survey terminology (1975), supplemented with genetic terms, like palustrine, gley and pseudogley. In this thesis the terms are defined as they appear in the text.

3.5.1.2 Principal features of palaeosols in general.

Soil processes take place in sediment between the ground surface and the local water table. The downward passage of water through this zone controls the nature and rates of the pedogenic reactions in the sediment. The chemical, physical and biological processes produce a **stratified soil profile**, with zones of humus accummulation, leaching and mineral accumulation. Factors which contribute to the vertical zonation include:

- **Ground water levels,** which fluctuate according to the seasonal rainfall.
- **Plant root systems.** The passage of water through the soil is strongly influenced by their presence, and they therefore control

many pedogenic reactions.

Rapid rates of sedimentation, or conversely, partial erosion of soil, prevents the development or preservation of a full stratified sequence. Sequences often comprise stacked zones of mineral accumulation, without the originally overlying zones of leaching or humus accumulation. This is the case for the palaeosols of the Poudingues d'Argens Formation.

In ancient soils, it is the accumulations of iron minerals, clays and calcium carbonate that can best be described and used to elucidate details of the palaeosol environment in which they formed. The distribution of iron oxides, hydroxides, and sulphides indicate the zones of oxidation and reduction within the soil. These are used to determine ground water levels and distribution of fluids through a soil profile and their acidity. The clay mineralogy is indicative of soil acidity and the quality of drainage through the sediment.

The term applied to pedogenic carbonate observed in the geological record is **calcrete**, which is defined as:

> "an accumulation of predominantly fine grained, low magnesian calcite, formed within the meteoric vadose zone by pedodiagenetic alteration and replacement of any precursor host material" (Klappa, 1980).

The distribution of calcrete within a palaeosol is not uniform. The morphology of the authigenic carbonate reflects its position (depth) within the calcrete profile and is also a measure of the maturity of the soil profile. An ideal, complete calcrete profile is illustrated in fig. 3.12 .. An immature calcrete may contain only micrite glaebules, but given time a continuous horizon of laminar micrite develops which is capped with a hard pan of impermeable carbonate.

In the past, authors have used the existence of calcrete in a palaeosol to infer an arid and or semi-arid palaeoclimate. However, the factors which control the nature and rates of calcite precipitation are numerous and interdependent. Interpretated palaeoclimates are invalid if not supported by independent evidence.

3.5.1.3 Principal features of the Poudingues d'Argens palaeosols.

The host sediments of the palaeosols are predominantly carbonates. The percentage of carbonate in apparently unaltered fine grained sediments ranges from 50-95%. The principal sources of carbonate were local Mesozoic limestones. These are pure and do not contain significant amounts of feldspar or iron rich silicates. Other sources of sediment were intraformational, for example, the calcite crystals of microcodium (fig. 3.14).
As a direct consequence of the composition of the host sediments, most of the palaeosols are charactised by **rare clay** minerals,

weak-intermediate intensity **colour mottles** (there are noteable exceptions to this, which will be discussed in section 3.5.6) and well developed **calcrete profiles.**

The palaeosols are grey - pale cream, yellow ochre - pale terra cotta for the most part, with purple colours developed in the outcrop area of Quatre Cantons, in particular (section 3.5.6). Rhizocretions (Klappa, 1980) are common and microcodium is locally an important component of the altererd sediment, having in places, replaced 60% of the rock volume (fig. 3.25b.)

The colour mottling and the nature of the calcrete profiles are the two features of palaeosols in the Poudingues d'Argens Formation which vary markedly between outcrop areas. The following two sections discuss the properties of calcretes and colour mottles that are relevent to the present study.

3.5.1.4. Calcretes.

Calcretes can develop in any host soil. They characterise semi-arid environments, but are not diagnostic of such climates. The original porosity, permeability of the host sediment and its role as a potential source of calcium carbonate influence the rate at which calcretes develop.

The calcrete morphology and rate of carbonate precipitation are controlled by:

- the total carbonate supply through the soil;
- the climate, in particular seasonal variations in surface temperature, and rainfall;
- the presence of organisms, in particular root systems.

Steel (1974) suggested that calcretes form when there is enough water flow to supply carbonate to the subsurface mineral soil, but not so much as to leach the whole profile. Calcrete formation appears to require a seasonal climate, with alternating wet and dry periods (Steel 1974, Wright 1982). Under these circumstances carbonate is brought to the soil and concentrated during periods of rainfall, and actively precipitated during the dry season, when rates of evaporatal and plant transpiration are high. If there is plentiful supply of carbonate, seasonal high rainfalls do not preclude the development of calcrete profiles.

Biological processes contribute to the development of calcretes. Bioturbation and root systems affect the porosity and permeability of the host material. More importantly, root systems modify the chemical environment in their vicinity and invariably enhance the precipitation of calcite.

Calcretes form preferentially on stable, gentle (<25) slopes in low lying areas, surrounded by older limestones (Klappa, 1980). Modern calcretes form at or near the surface, but the location of carbonate accumulation is controlled by the local ground water level and so varies with respect to the ground/atmosphere interface. Marked lateral and vertical variations in the distribution of pedogenic carbonate are observed (Esteban, 1976).

The maturity of a palaeosol reflects the length of time that soil processes have operated in a stable soil profile. If rates of pedogenic reactions, like calcrete formation, could be quantified, it would be possible to define periods of time when little or no sediment was deposited or eroded at a given site. Steel (1974) and Leeder (1975) used calcrete profiles developed in overbank sediments to infer absolute rates of flood plain aggradation. Leeder suggested that 10^3 - 10^4 years might be required to develop mature calcretes, depending on the permeability of the substrate. However, Whiteman (1970) showed that calcrete glaebules can form in periods as short as tens of years, if there is a good supply of calcium carbonate. Carbonate is precipitated as a result of a number of interdependent soil-forming processes, each of which have different reaction kinetics. Therefore, at this stage, it is safe only to make comparative statements about lengths of time required to produce calcrete profiles.

Finally, it should be noted that the formation of a mature calcrete profile modifies the structure of the soil profile and all subsequent pedogenetic processes. In particular, the formation of laminar calcrete produces an impermeable layer within the sediment. As the vertical permeability of the soil decreases, water may collect and stagnate above this horizon. The lateral movement of solutions may be enhanced. Plant root systems can take advantage of the water trapped above permanent ground water table and often develop horizontal root mats. Thus chemical, physical and biological pedogenic processes evolve with time, with the maturity of a calcrete profile.

3.5.1.5. The significance of colour in soil profiles.

Buurman (pers. comm. to Atkinson, 1983) is of the opinion that strong pigmentation in soil proiles is produced in areas where there are distinct seasonal variations in rainfall. This climate also favours the development of calcrete profiles. The chemical processes responsible for pigmentation require that the sediment is at least part of the time saturated with water, ie. below the water table. The fluctating water levels in alluvial, hydromorphic soils give rise to frequent changes in the oxidation state of the profile. During the wet season, silicate minerals break down rapidly in a soil saturated with ground water. During the dry season, iron oxides are precipitated from water passing through the sediment, if the soil profile is well oxygenated. The colour mottles of palaeosols represent the ancient water and air pathways through the sediment (fig. 3.25).

Buurman (1980) subdivided hydromorphic soils into two groups: gley and pseudogley soils. The distribution of iron oxides within the colour mottles is different in each case (fig. 3.~~16~~27):

Gley soils develop where the water table is high with respect to the sediment surface. When the soil is saturated with water, oxidising solutions can only circulate through cracks, burrows and roots. The soil matrix is grey, and the concentration of iron oxides increases towards the centre of the colour mottles.

Pseudogley soils develop when water stagnates at a level above the water table. This can occur after flooding or during a wet season. In the saturated sediment, the distribution of colour is the same as in gley soils. Below the perched water table, the soil is never saturated with water. As fluids pass along roots and burrows iron compounds are leached. As the fuids disperse into the sediment matrix they evaporate and the iron compounds are reprecipitated and oxidised. As a result, the greatest concentration of iron oxide occurs at and near the margin of the principal fluid conduits, and the whole soil matrix is stained red/tan; the cores of the conduits are leached grey.

In the soils of the Poudingues d'Argens Formation laminar calcretes probably formed impermeable layers above which water could be trapped, while vertical permeability was locally maintained along root systems. Thus, the conditions ideal for pseudogley soil formation were generated by the precipitation of pedogenic carbonate. In this way, the distribution and intensity of palaeosol colouration relates directly to the maturity of the calcrete profile.

3.5.1.6. The aims of the Poudingues d'Argens palaeosol study.

i) The development of soil profiles in the Poudingues d'Argens Formation has not been documented fully to date. In this study, the features of the palaeosols were described and facies associations defined. The relationship between soil colour mottling and calcrete maturity was used to characterise each profile.

ii) The nature and maturity of the palaeosols varied between outcrop areas. The variations were used to interpret local palaeotopography and to infer **relative** rates of sedimentation across the study region.

iii) From other lines of evidence, the upper parts of the continental succession are believed to be contemporaneous with the Nummulitic transgression. Evidence was sought, and found, to support this hypothesis. Persistent variations in palaeosol type were observed which are consistent with rising groundwater levels and increasing influence of brackish pore waters upwards, towards the contact between the Poudingues d'Argens and Calcaires Nummulitiques Formations.

iv) The colour of the fine grained facies in particular varies markedly between certain outcrop areas. It was possible to show that the hue of colour mottling was controlled by the type of host sediment. As a result, more than one source area for the Poudingues d'Argens sediment were identified.

v) To reconstruct the early stages in the history of the Gres d'Annot basin, it is desirable to know the age of the earliest sediment in the basin. The derth of fossils in the Poudingues d'Argens Formation make this difficult! Therefore, the maturity of the palaeosols were used to approximate (in a light hearted fashion!) the length of time represented by the Poudingues d'Argens succession. A sequence at Argens was chosen that did not have any significant erosional breaks and the number and maturity of the palaeosol profiles were recorded.

3.5.1.7. The Method of Study.

The palaeosol facies of the Poudingues d'Argens formation were grouped using five criteria:

i) Composition of the host sediment.

ii) Grain size of the host sediment.

iii) Colour of the soil matrix and its associated mottling.

iv) Distribution and characteristics of colour mottles. For example, their size and shape, and whether iron oxides and hydroxides are concentrated towards the inside or outside of the mottles.

v) Presence and nature of nodular, authigenic mineral horizons, in particular, calcium carbonate.

This list is a modified version of Atkinson's (1983) for soil diagnosis. He included clay mineralogy, but the Poudingues d'Argens samples contained so little clay as to make analyses impractical. Instead, the emphasis of the study was placed on the field characteristics of the palaeosols.

The palaeosol intervals were examined in the field, and thin sections were made from samples collected to establish the morphology of the pedogenic carbonate and to identify other authigenic minerals. They were also used to confirm the presence of plant root systems, microcodium and burrows. The relationship of the biogenic structures to the calcrete morphology was defined.

In the palaeosols of the Poudingues d'Argens Formation the distribution of colour varies with the maturity of the calcrete. Both features were used to define a sequence of facies that constitute a complete palaeosol profile (. 3.3).

Colours are defined using the Munsell Colour Chart System.

The palaeosol sequences in the outcrop area of Argens exhibit all the principal pedogenic facies of the Poudingues d'Argens Formation. These are described in the following sections 3.5.2, 3.5.3 and 3.5.4. The sequences of the remaining outcrop areas are then discussed in sections 3.5.5 and 6.

3.5.2. Type 1 Palaeosols (Type section at Argens; AR1, fig. 3.23) and encl. 3.5

3.5.2.1. Introduction.

At Argens, the continental sediments have almost all been extensively modified by pedodiagenesis. Stacked soil profiles of differing maturity are preserved and the variety of facies observed here serve as the standard against which to compare outcrop areas.

The host sediments are **carbonate clastics**, dominantly marlstones (down to fine silt grade), sandstones and rare conglomerates.

There are two distinctive soil types developed at Argens. The first dominates the succession; the second is observed commonly at the top of the formation and comprises a number of incomplete profiles and one complete profiles, that being immediately below the Calcaires Nummulitiques Formation.

In these two sections, the facies are described as they occur within the profile, from bottom to top. It is rare to observe a complete mature profile. Incomplete profiles display superimposed sequences, like facies 1a-b or 1a-d. Profiles are difficult to measure because unmodified host sediment is not often preserved at the base and discrete upper boundaries are also not common. This suggests that sedimentation rates were not high, but that deposition did not often stop completely.

3.5.3.1. Facies 1a (0 - 3m thick in these sections)

Uniform greyish-ochre to pale yellowish-orange (commonly 10 YR 6/5 to 8 YR 6/2), devoid of mottles. Discrete nodules of carbonate are absent, but at microscopic level, patches of micrite are observed with diffuse boundaries. These are most easily observed where the host grain size is coarser than silt because the grain size contrast is apparent. Incipient corrosion of quartz grains and their replacement by micrite is observedin these patches. This can give rise to a blotchy colouration to the rock, and the weathered surface uneven due to differential erosion of the altered sediments.

Interpretation

This is a colour ripened (Ponns and Zonnevald, 1965) sediment, representing the initial stages of pedogenesis. Colour is the result of finely disseminated limonite, altering in places to haematite, as the rock ages in an oxidising environment.

Uniform colour ripening can occur between low and high ground water levels in a hydromorphic soil. The diffuse patches of micrite form peds (Brewer, 1964) within the rock: significant volumes of sediment can be micriticed at this level which then corresponds to the chalky calcrete of Klappa (fig. 3.22) or "accumulation diffuse" (Ruellan, 1967).

This facies occurs at the base of a mature profile, or alone, where it represents an immature soil.

3.5.2.3. Facies 1b (0 - 3.5m thick, but the thicker units are probably compound horizons).

Description

Uniform greyish ochre to pale yellowish-orange, plus discrete nodular carbonate. Nodules are discrete, easily separated from the matrix and range in size from 0.5 - 2cm, often uniform 1.5cm diameter. They comprise 10% of sediment. The nodules are finely crystalline, non-ferroan calcite. Quartz grains within the nodules are highly corroded, replaced by calcite. Irregular voids or "crystallaria", (Brewer 1964) dissect the nodules and are filled with sparry calcite, often ferroan ~~(E . 3.c)~~. The glaebules are often separated from the S-matrix by a circumference of sparry calcite filling circumgranular cracks.

The S-matrix itself has a highly anisotropic fabric with coalescing, diffuse areas of micrite forming most of the S-matrix (E d. 3.5).

Intrepretation Development of disorthic glaebules (Wright 1982);

Their random distribution in the sediment may suggest that they formed at a level below major biogenic activity in the soil (Klappa, 1978); The facies above this shows that roots and burrows cause carbonate to precipitate in discrete localities, not randomly. Having said that, the carbonate nodules probably seeded in bioturbated sediment. The bioturbation took place near surface, and when the sediment was buried, an anisotropy remained.

The crystallaria within the glaebules are sealed curved fissures interpreted by Freytet (1973) as the result of frequent wetting and drying of the soil, within the zone of water table oscillation.

This facies occurs above facies 1a, either as part of a mature soil and calcrete profile, or as the total of an immature profile prevented from development by renewed sedimentation at surface. It corresponds to type 1 calcrete profile of Steel (1974).

3.5.2.4. Facies 1c (1 - 3.5m thick)

Description

Colour mottled sediment, with colours the same as those of facies 1a and 1b: hydrated oxides of iron are responsible for most of the colour. The sediment is bleached a greyish white 10 YR 8/<1) in a number of places: in irregular patches of variable dimensions; in roughly circular patches of up to 5cm diameter (Encl. 3.5); and along both planar and cylindrical features, that are vertically and horizontally elongate. The ochre - terra cotta colouration in the sediment matrix increases in intensity towards the margins of the bleached areas, reaching colour of 5 YR 5/6 (7 R 3/4).

Nodules of calcium carbonate are ubiquitous; they are commonly amalgamated and their distribution is linked to the bleached areas in two ways (fig. 3.2). The nodules are creamy white, for the most part, but their cores are often tinged pink (10 R 5/4) or green.

Low in the section, the vertically elongate zones are most often cylindrical. The round bleached patches are infact cross sections through similar subhorizontal and oblique cylinders. At one locality, however, the vertical zones are planes; the segments are linked by horizontal planes of colourless sediment (top ot section AR1, fig. 3.. 3 encl. 3.3).

The vertical *tubes and circular patches* of uncolourd S-matrix are associated with small patches of microcodium.

Vertically elongate bleached zones and amalgamated nodules are gradually replaced upwards, within a section of this facies, by horizontal planes of amalgamated nodules.

Interpretation

The colour mottle characteristics are those of pseudogley (Buurman, 1980), hydromorphic soils. The genesis of the soil type in this area will be discussed later. Buurman has stated (pes. comm. to Atkinson, 1983) that the white patches in such soils represent the preferential routes taken by water percolating through the palaeosol.

There are therefore two types of conduit in this palaeosol which have characteristic distributions of carbonate precipitate associated with them (fig. 3.24a and b). Type (a) are interpreted as burrows, from their irregular, branching form. The dark brown silicified core is believed to represent the backfill of such burrows with sediment which may have been more clay rich, at least sufficiently different to localise silicification. Carbonate and iron rich solutions passed from the burrow into the surrounding sediment to distances controlled by the rates of evaporation. The concentration of iron minerals and calcium carbonate precipitates at the same distance around the burrow suggests that their concentration was controlled by rates of evaporation (fig. 3.25).

Type (b) are believed to be rhizoliths and rhizocretions (Klappa, 1980). The vertical elongate cylinders taper downwards and bifurate. The association of these cylinders with microcodium is evidence that roots were present (Klappa, 1978b) and formed the principal water conduits. The occurrence of nodular calcuim carbonate in their cores is correlative evidence. Klappa has clearly demonstrated that carbonate precipitates preferentially in the vicinity of plant roots, and he was able to suggest biochemical processes to account for this phenomena. In particular, plant transpiration concentrates solutions in the vicinity of the plant roots, thus increasing the tendancy of minerals to precipitate.

These large roots were the tap roots of plants seeking permanent ground water. When the plant root died, carbonate continued to precipitate preferentially at sites where nodules had seeded. The nodules coalesced and the tube would eventually have become impermeable to water. When it ceased to operate as a conduit of water the rhizocretion developed no further. This part of the soil profile became inactive and carbonate precipitation continued at higher levels. This process will limit both the diameter of the rhizocretions and the concentration of authigenic iron oxides and hydroxides (colour intensity) measured in different sections.

The vertical planes of bleached and nodular soil are thought to be desiccation cracks, but it is puzzling why these are not seen elsewhere in the section. The original sediment may have contained more water initially (perhaps in a higher percentage of swelling clays). Consequently, when it was exposed at surface and dried, the volume reduction was greater.

This facies corresponds to type 3 cornstone of Steel, 1974. Where a single profile can be identified, the thickness of this facies represents the minimum depth to which roots had to penetrate to reach permanent groundwater (plus, of course, the unknown thickness of soil above the mineral accumulation horizon). At Argens, this depth does not exceed 2m.

3.5.2.5. Facies 1d (10cm - 1m thick).

Description

The sub-facies comprises nodular micrite and sparry calcite with only small fragments of severely modified host sediment preserved. In the field, the horizons can be traced in one exposure over 100m and correlated with confidence, over 0.5km.

They are characterised by a blocky, columnar appearance (ncl. 3..3), the columns being 5 - 15cm diameter. The columns, which are creamy white, are defined by fractures, which are often 'healed' with sparry calcite, or thin planes of intensely terra-cotta coloured (eg. 10 YR 6/6) sediment.

The lower part of this sub-facies may be all that developed. It comprises almost totally coalesced, but distinguishable carbonate nodules (3-8cm across). The rims are intensely stained with iron oxide.

Upwards, the nodules coalesce laterally, for the most part, giving the horizons a distinctive horizontal fabric (Encl. 3.3)

Interpretation

Facies 1d is similar to facies 1c, but concentration of biogenic structures is much greater, reflecting the position of facies 1d at higher levels within the soils where roots penetate in all directions to maximise access to temporary water sources.

Correspondingly, intense bioturbation, by, for example, insect larvae is possible because the roots break open the soil and provide the food source. The field evidence implies what one would expect: that the Upper parts of the mineral accumulation 'C' horizon are those most affected by pedoturbation, bioturbation and associated biochemical processes.

The upward increase in carbonate percentage within a calcrete profile is partly explained by the greater proportion of roots. By drawing water from the system the roots increased the mineral concentration in the adjacent pore fluids. It is also a direct consequence of progressively precipitating minerals from downward percolating solutions. In this carbonate rich host environment, the non-biological control of calcite precipition was principally evaporation. Therefore, the depth to which carbonate could be transported before precipitation changed with seasonal rates of evaporation (fig. 3.25).

An interesting feature of this facies are the intensely iron-stained patches of host sediment and the iron oxide coatings to the nodules. The nodules are **non-ferroan** calcite. For some reason, iron was excluded from the carbonate lattice and progressively concentrated in the remaining matrix. The processes of differentiation were unlikely to be mechanical and therefore the iron was transferred in solution (ie. reduced). It is possible that, during the wet season, iron went into solution and remained there longer than the calcium and carbonate ions when the soil dried. An analogy might be drawn with the scale on kettles in hard water areas, where calcium carbonate is one of the first minerals to be precipitated.

The Sf corresponds to type 4 cornstone (Steel 1974).

3.5.2.7. Facies 1e (0 - 20cm thick).

The facies is distinguished from facies 1d by the laminar fabric to the carbonate. The upper few centimetres may be a complexly cemented breccia. The horizon is often colour banded on scale of 2 - 4cm, olive grey (5 YR 6/1) and orange pink (5 YR 7/1). Within the orange-pink bands, the laminae are defined by films of iron stain around lumpy layers of micrite <2mm thick (Encl. 3.3 and fig. 3.26, locality AR20). The olive grey horizons are massive micrite with cavities, sometimes lined with amorphous silica.

The most interesting feature of these laminar carbonates under the microscope are layers of microcodium. The layers can be millimetres thick, comprising three or four strands of great lateral extent

Very rarely, the laminae form complex lobes, where laminae of massive, dense micrite alternate with fenestral micrite (bird's eye texture). The laminae thicken towards the crest of the anticline ~~(fig. 3.25).~~ This is interpreted as an algal stromatolite texture. In this case, the brecciated horizon at the top of the calcrete may have sediment filling some of the fissures.

Interpretation

The facies represents the hardpan at the top of a mature calcrete (eg. Klappa, 1978a). The occurrence of microcodium forming continuous sub-horizontal sheets is quite unusual (Wright, pers. comm., 1985). It suggests that mats of horizontal roots formed above the calcrete of facies 1d.

Microcodium is believed to result from a symbiotic relationship between a species of fungi and roots. The exact processes which produce this beautiful calcite morphology are not understood, but Klappa's analysis (1978a and b) is the most comprehensive to date. The occurrence of root mats above calcretes has been used by Klappa to account for the laminar nature of calcrete hard pans. The nodular carbonate forms an impermeable horizon within the soil above which water collects during the wet season. Plant root systems adapt to utilise this water source (which may be seasonal, or may remain all year, depending on rates of evaporation) and produce horizontal root mats.

The majority of laminar calcretes at Argens exhibit microcodium, which implies that the hard pans were normally formed below ground surface. However, where the algae grew, the hard pan was at, or very near the ground surface, for at least part of its history. Sediment was washed into the fractures of the brecciated upper surface.

3.5.2.7. Significance of the Type 1 palaeosols.

1. The occurrence of pedogenetic carbonate and the development of colour mottling implies that these soils were hydromorphic with seasonal variations in rainfall. Permanent groundwater levels were probably greater than 3 - 4m below the topographic surface. Estimates of the amount of seasonal fluctuation in ground water levels cannot be made.

2. The colour mottle distributions are those of pseudogley soils which are characterised by perched groundwater levels. The development of calcrete within the profile is believed to be, in part at least, responsible for the development of pseudogley soils at Argens, and the model for their development is presented below and in fig. 3.27:

   At the early stages, the fluctuating phreatic groundwater levels induced the precipitation and oxidation of iron minerals which produced the uniform iron staining of 'ripened' soils. Carbonate was precipitated at various levels in the mineral soil, preferentially at the top where biogenic activity was most intense. The decrease in carbonate downwards also simply reflects the downwards decreasing availability of carbonate in solution as evaporation continued. The nodules of carbonate coalesced and gradually reduced the vertical permeability of the soil.

Once a continuous, laterally extensive calcrete horizon had formed, water was trapped above it during periods of high rainfall. From then on, the horizon between the permanent and temporary groundwater levels received water principally via the tap root and burrow systems. Thus, pseudogley conditions were established below the calcrete permeability barrier and were maintained until the conduits were blocked by carbonate precipitate or until the sedimentation rate picked up again at surface.

Above the massive calcrete, the horizontal root mats and the lateral movement of water enhanced the precipitation of laterally continuous laminar calcrete. It is unlikely that water remained ponded for much of the year: the water could be dispersed by evaporation and lateral drainage. The colour mottling and calcrete formation suggest a seasonal climate with periods of high surface temperature. Therefore it is not likely that water would be replenished in the drier seasons.

3. Only one fully mature, complete profile is recorded at Argens. Incomplete profiles are very rarely the result of erosion at AR1. Instead, maturity of the palaeosols was controlled by the sedimentation rates. Sedimentation rates could vary in two ways:

i) Punctuated sedimentation where periods of deposition were separated by periods of non deposition. This is the case most often used to explain palaeosols.

ii) Continuous, but varying rates of sedimentation.

Plants can colonise environments of active deposition (eg. mangrove swamps), and so the presence of rhizocretions does not exclude this model.

In the first case, the palaeosol profile will comprise Sf 1a, 1a-b, through to 1a-e depending how long the period of non deposition lasts. Sf 1d and e can only be established if the ground surface is stable. Accordingly, the development of 1c, the pseudogley soil horizon, **depends** on punctuated sedimentation.

In the second case, the profile will comprise a continuous sequence of 1a or 1b, recording only the passage of a given sediment level through the zone of fluctuating groundwater levels.

The sections at Argens imply that both situatuions occurred, producing sections very similar to those described by Freylet + Plaziet (1983).

3.5.3. Type 2 palaeosols.

3.5.3.1. Introduction.

This soil type is complex and appears to have features associated with both pseudogley and gley soils. There are three facies which occur in different localities whose differences principally reflect differential subsidence and sedimentation rates and the presence of brackish pore waters at critical localities.

3.5.3.2. Facies 2a.

Sediments displaying the features of this subfacies are a sequence approximately 15 metres thick towards the top of the Poudingues d'Argens Formation at AR5 (fig. 3.2°) and a 3.5 m sequence beneath an interpreted lacustrine carbonate at AR1.

In the field, the sequences display no vertical anisotropy. The properties vary little from bottom to top of the sequences, regardless of grain size variations in the host sediment. At AR5, the colour of the matrix charges slightly, becoming greyer upwards, but retaining the faintly mottled appearance.

The matrix is grey-greyish brown (8 yr 6/2 - 5 yr 6/1). Colour is distributed unevenly but distinct mottles are only seen immediately adjacent to carbonate nodules where colours of 10 yr 8/6 - 5 yr 5/6 are observed.

Carbonate nodules are common, with highly irregular although sharp boundaries. They coalesce to form vertical cylinders up to 15 cm long: in each cylinder the dimensions and degree of amalgamation of nodules increases upwards (fig. 3.29).

In thin section, the matrix is highly anisotropic, with diffuse patches and thread-like concentrations of limonite.

Elongate channel voids are common, filled with sparry calcite and filaments of carbonaceous material, associated with dark terracotta staining (En l. 3.3).

Interpretation

The nodular carbonate defines rhizocretions, and the carbonaceous filaments are interpreted as root hairs.

Limonite appears to concentrate along these zones ~~and~~ interpreted as burrows. it is slightly puzzling that ferric oxide is preserved in association with carbon, which should have been oxidised too. However, different reaction kinetics may explain the phenomena. The oxidation state of iron is changed with ease, but the sediment may have subsided below the phreatic water table before the carbon in the matrix was oxidised.

The grey colour of the matrix and the carbonaceous plant remains suggest that the sediment was modified in a predominantly reducing environment (Buurman, 1975). The sediment could not have been dry for any length of time and was probably below the phreatic water level soon after deposition, therefore implying that ground water levels were high in the soil.

For the most part, the limonite and iron oxides are concentrated along root channels and burrows, which were the principal water conduits in the soil. The soils were therefore 'gley' soils, according to the conclusions of Buurman (1980).

At AR1, the gley soil sequence is capped by a laminated, lacustrine limestone. The ground water level in sediments adjacent to a lake are high with respect to topography and therefore the grey palaeosol of AR1 is interpreted as a marginal lacustrine soil: paludine, in the sense of Freytet, 1973).

At AR5, 15 m of this facies shows no variation in concentration and distribution of carbonate nodules. Steady and continuous sedimentation rates are inferred from this thick sequence. The land surface was well colonised with plants and therefore the gley palaeosols were produced within a continuously aggrading marshy floodplain, just as one might expect to find on a lower alluvial plain.

3.5.3.3. Facies 2b.

This facies is identical to facies 2a, with the addition of authigenic pyrite in the matrix. Irregular clusters of pyrite

(1-2 mm size framboids) are observed and diffuse patches of a very finely disseminated black opaque mineral are also interpreted as pyrite (~~Fig. 3.~~). The larger blebs (crystals) display a hint of brassy-silvery sheen under reflected light. The margins of these accumulations are diffuse, but the pyrite 'crystals' concentrate around their runs. Their is no discernible preferred sites of crystallisation of the pyrite: it seems to form diffuse mottles in the same way as limonite does below. the occurrence of pyrite corresponds with very low percentages of limonite in the matrtix. Therefore, the sediment is coloured grey.

The occurrence of this subfacies is limited to the top of the Poudingues d'Argens Formation and is observed at AR5, not AR1.

Interpretation

The presence of authigenic pyrite in pedogenically modified sediment has been discussed by several authors (e.g. Buringh, 1970 and Wright, 1986). The sulphide is invariably derived from sulphate in sea water which is reduced by bacteria in anaerobic pore waters. Therefore, the formation of iron sulphide is favoured in brackish-marine ground waters where there is abundant organic material and associated micro-organisms. Wright (1986) has recently shown that pyrite forms as a very early diagenetic mineral in palaeosols that have been drowned *during a* transgression.

At Argens, the uppermost deposits underly a transgressive sequence of brackish lagoon to shallow marine carbonates. Therefore, it would appear that the alluvial gley palaeosols were forming adjacent to a transgressive marine environment. An alluvial - coastal plain environment model satisfactorily explains the steady, continuous rates of sedimentation, the permanently high phreatic water levels and mixed marine-fresh pore waters.

3.5.3.4. Facies 2c

This facies is observed at the top of the Poudingues d'Argens Formation at AR16c, north of AR5, and involves sediment from the Calcaires Nummulitiques Formation above (a bioclastic sandstone with nummulites). It comprises a sharp topped, amalgamated nodular massive calcrete which superficially looks like subfacies 1d. However, closer examination reveals that the calcrete is deeply cracked, and the fissures are filled with an orange-buff nummulitic limestone (fig. 3.31). Limestone filled tubes penetrate the calcrete: some are irregular (fig. 3.30), others bifocate and taper downwards (fig. 4.17).

The calcrete is locally brecciated with micrite matrix which contains finely disseminated pyrite (seen in thin section). Framboidal pyrite nodules are distributed randomly in the upper 20 cm of the calcrete.

In thin section, there are many voids which are filled with complex cements. There are also many carbonaceous root hairs.

Above the calcrete, the nummulitic limestone is uniformly orange-ochre colour (10 yr 7/4-7) and diffuse nodules of disseminated pyrite are observed.

Interpretation

In this case, the features of the carbonate are interpreted as the superposition of a gley soil profile on an already formed calcrete of facies 1d (fig. 3.3).

The calcrete developed near to surface, evidenced by limonite which concentrates between the nodules of carbonate that are not extensively brecciated. The top soil was removed by erosion (wind?), and then nummmultic limestone was deposited on the complexely brecciated calcrete surface, during a brief marine incursion. This sediment was subsequently exposed, but a high water table was maintained. A swamp environment may have been established. Certainly, the nummulitic sand filled tubes that resemble roots, not burrows. These could be the hollow roots of plants, like mangroves and reeds.

Reducing soil conditions prevailed elsewhere at the level of the calcrete, evidenced by the lack of iron hydroxides and oxides, the presence of carbon fossil root hairs and authigenic pyrite. The only oxydised cones were those within the roots where the chemical microenvironment was maintained as an oyxidising one. Thus a grey soil overprinted the calcrete.

3.5.4. Significance of the palaeosol facies observed at Argens.

The palaeosols at Argens have the characteristics of hydromorphic pseudogley and gley soils. They display stacked soil profiles, some with mature calcretes and well developed colour mottles which imply that sedimentation was episodic; in parts of the sequence sedimentation was continuous, if slow.

The maturity of a palaeosol reflects the length of time that soil processes have operated in a stable soil profile. If rates of pedogenic processes could be quantified, it would be possible to define the periods of time, in any succession, when neither sedimentation, nor erosion occurred. With this information, the formation thickness and a figure for the rate of sedimentation, it would be possible to quantify the length of time locked up in the continental sequence at Argens. However, there are too many variables in the system! Nevertheless, it is worth noting features about the sediment at Argens that suggest it was deposited in a very short time:

1. In 34m of succession at AR1, there are 6 calcrete profiles.

Only two of these have matured to the hardpan calcrete stage. In this section, there are no major erosive breaks.

2. The sediments are loaded with carbonate, and the surface and ground waters flowed from limestone hills. There was a plentiful carbonate supply for calcrete development.

3. The colour mottled sediment and other palaeoclimate indicators, like pollen collected in Provence (Freytet and Plaziat, 1982), suggest that surface temperatures were seasonally high.

The conditions favoured rapid calcrete development. Using Leeder's (1974) figures which are exceedingly conservative, the succession represents less than 100,000 years. There is a significant unconformity at the base of the Poudingues d'Argens Formation and evidence that the upper boundary is transitional with the Calcaires Nummulitiques Formation. Therefore, the first Tertiary sediments were deposited only shortly before the Nummulitic transgression. This conclusion has important implications for the ear~~t~~ly history of the Gres d'Annot basin, which will be discussed in section 3.6 and chapter 6.

The lower parts of each section show little variation across the Argens outcrop area: the palaeosols are ironstained (for the most part mottled), and the permanent water table levels were generally low. With the exception of a paludine soil, the profile features indicate a seasonal climate where most of the soil recieved water via discrete conduits from a seasonal perched water source.

3.5.4.1. Evidence for the timing of the Nummulitic transgression.

Upwards, the sections begin to show differences. At AR5, to the south, sedimentation rates became steady in association with a rise in phreatic water levels, as evidenced by 15 m of uniform gley soil profile. Relative to localities further north like AR1, sedimentation rates were probably higher. Approximately 6 metres more sediment is preserved at AR5, and the equivalent soil profiles further north have quite mature pseudogley calcretes near the top of the section. Therefore, at AR1, in the north, sedimentation was punctuated by periods of non deposition that were not experienced to the south. This relative difference is maintained right to the top of the formation.

The upper 1-2 m of gley soil in the south contain pyrite which is interpreted to record the influence of brackish pore waters on soil development. **The evidence stongly suggests that the upper part of the Poudingues d'Argens Formation developed close to a shoreline and were therefore contemporaneous with the Nummulitic transgression.** To the north, the final pseudogley calcrete profile was modified by gley soil processes and contains fromboidal pyrite. In this case, the pseudogley and gley soil features may have developed at quite different times, and so the northern outcrops cannot be used to demonstrate that the Poudingues d'Argens are transitional into the Calcaires Nummulitiques.

However, the palaeosols may be used to define the palaeo-relief and how it may have changed with time. The palaeoslopes were probably not great: Klappa (1978) stated that mature calcretes develop in areas of low relief (< 25′).

3.5.4.2. Palaeotopography during the Nummulitic transgression.

Palaeocurrent data shows that early in the Poudingues d'Argens succession, there was a component of palaeoslope towards the N/NW. However, upwards through the sequence of palaeosols, the southern areas record continuous rates of sedimentation, while in the northern sequences were punctuated with periods of non deposition. Similarly, gley soils characterise 15m+ of succession in the south, and there is a gradual increase in the influence of sea water up towards the boundary with the Calcaires Nummulitiques Formation. In the north, pseudogley profiles persisted until the sea drowned them. The evidence implies relative uplift in the north (or increased subsidence southwards!).

The apparent change in palaeoslope with time suggests that the southward dipping slope, which need not have been great, was induced contemporaneously with the marine transgression (~~fig.~~

fig.4-30.). The palaeoslope from north down to the south can be correlated with the slope of a thrust sheet hanging wall ramp in Lower Jurassic stratigraphy (cross section Z, enclosure 14). The sedimentological data would suggest that the thrust was initiated during the Nummulitic transgression.

3.5.5. Palaeosol Facies at Peyresq and Le Ruch

3.5.5.1. Introduction.

At these outcrop areas pedodiagenetic modification of the continental sediments is not obvious. Very close examination of the fine grained facies does reveal faint colour mottling and certain of the nodular carbonates have the features characteristic of calcrete rather than ground water early diagenetic carbonate.

The host sediment is carbonate detritus-mud and sand grade particles with 1-10% quartz and rare other minerals. The lack of colour mottling can be explained two ways: high rates of sedimentation associated with high ground water levels; and the very low percentage of iron silicate minerals which could be leached into pore water solutions.

3.5.5.2. The principal features of the pedogenic modification.

a) Abundant carbonaceous fossil rootlets, associated with small, discrete nodules of iron oxide/hydroxide. These are visible in the field and in thin sections (fig. 3.2); where they are associated with channel voids, which are interpreted as root cavities.

b) Very faint colour mottles which are confined to the rims of orthic diffuse carbonate nodules (mauve-pink + pale terracotta). These concentrate in the vicinity of burrows (5yr 7/2 in the centre, 5 yr 5/5 at the rims) and rhizoliths where the carbonate precipitated preferentially. Otherwise, the sediment is uniformly cream-white.

c) Nodular Carbonate.
Much of the original fabric of the host sediment has been replaced by micrite. The micrite forms diffuse nodules, which amalgamate to form discrete, sharp topped massive carbonates sometimes with a thinly divided horizontal fabric, but no discernible compositional laminae. These carbonates frequently replace the tops of small coarsening upward sequences (microdeltaic infill of small ephemeral lakes). In thin sections, there are abundant features related to bioturbation and sediment disruption by roots. Crystallana are common, and sparry calcite fills the voids (often multiple phase cements).

However, no calcrete profiles are developed and none of the nodules resemble the disorthic glaebules described at Argens. It may be that this carbonate was precipitated from ground water, although certainly near the surface, in zones of intense biogenic activity.

d) Authigenic Pyrite.

It is very interesting to note that finely disseminated pyrite is observed almost at the base of the Poudingues d'Argens sequence. Following the arguments developed previously for facies 2b + c at Argens, it is necessary to conclude that the phreatic water was brackish! This has enormous consequences for the palaeoenvironmental interpretation of the alluvial system <u>and</u> its age!!

Pyrite is observed again, at the top of the formation as expected, immediately below the first marine sediments of the Calcaires Nummulitiques Formation.

e) Algal Stromatolites.

Laminated carbonate with fine, undulose laminae defined by alternations of micrite and drusy calcite cements. These form layers up to 3 cm thick, associated with lacustrine carbonate. Their presence does imply that the sediment surface was stable for some lengths of time, and therefore sedimentation was not continuous. However, it is a rare phenomena! (see log through PQ1, '. 3.1).

3.5.5.3. Interpretation of the pedogenic features.

The lack of colour is partly due to lack of appropriate minerals in the host sediment. In addition, all the pedogenic features imply formation in a reducing environment, saturated with water. These are the characteristics of gley soils with high local water tables. However, sedimentation rates may have been very high in which case there would have been no time to establish mature soil profiles, even if the water tables were quite low. The preserved features are then not strictly pedogenetic, instead they may be early diagenetic.

The authigenic pyrite suggests that the whole succession was deposited within a short distance of the sea. The implications of this are discussed in section 3.6.

3.5.6. The Palaeosols at Quatre Cantons

3.5.6.1. Introduction.

The pedogenetic modification of the continental sediments in this outcrop area has two remarkable features:

a) The deep red-purple colour mottling of the fine grained sediments (fig. 3.2 c).

b) The development of microcodium.

The host sediments are unusual in that they comprise a broad range of lithologies. The conglomerate clasts are mostly carbonate, derived from the Mesozoic succession. However, some conglomerates have a rich variety of clast lithologies including dolomite, sandstone and pebbles of purple mudstone (fig. 3.2 ,d). and). These are interpreted as detritus from lower in the Mesozoic succession: Triassic dolomites, lower Triassic quartzites, and the purple mudstone can be easily matched with those of the Permian exposed today in the Dome de Barrot.

The sandstones are commonly an admixture of quartz and carbonate clastic sediment. The fine grained facies have components that are difficult to identify because of grainsize, but the percentage of calcium carbonate ranges from 40-90%. Erosion of the same mudstone that provided the mudstone pebbles would also generate much fine grained detritus, perhaps imparting a primary red hue to the sediment.

There is a clear difference in host rocks of Quatre Cantons and Argens, Peyresq and Le Ruch. The sediment at Quatre Cantons is potentially quite rich in iron minerals:

a) Reworked from the Triassic quartzite (eg. biotite).
b) Haematite associated with the purple mudstones.

3.5.6.2 Principal features of the pedogenic modification.

Palaeosol profiles exist, but have not been measured. Mature calcrete profiles are observed. The palaeosols observed are of facies 1a-e, i.e. they exhibit the characteristics of pseudogley, hydromorphic soils. No gley profiles were noted, although they may exist.

a) Colour Mottles

The fine grained facies are highly coloured and often mottled: the colour intensity increases towards bleached zones associated with rhizoliths, rhicocretions and burrows (facies 1c; fig. 3. c). Thin sections of these rocks show that the colour does not appear to be derived from detrital grains. Its genesis and distribution are interpreted to be pedogenetic features of pseudogley soils

In some cases, the colour is uniformly distributed and then it is less obvious what gives the mudstones colour. However, comparing thin sections with those of the mottled sediment most of the colour is believed to be secondary: pedogenic or early diagenetic.

The presence of pedogenetic haematite may be explained by the occurrence of haematite within the host matrix. Some of this was undoubtedly reduced and leached from the upper parts of the soil profile during the wet season. These solutions would percolate downwards and distribute the iron through the soil. As the soil dried out, haematite particles remained and may have acted as sites on which authigenic haematite could seed and form small nodules, where in a different host sediment limonite may have formed.

The presence of haematite here and not at the other outcrop localities described cannot be explained by different ages of rock or by different palaeoclimates. These Poudingues d'Argens Formation is penecontemporaneous across the rather small study region.

Finally, it should be noted that colour is not restricted to the floodplain facies associations: the matrix of many conglomerates exhibits the same intense colours (fig. 3. 5).

b) Microcodium

The development of Microdium here is spectacular and must say something very specific about the palaeoenvironment local to Quatre Cantons. That is, if the palaeoecology of microcodium were better understood (Klappa, 1978 and Freytet and Plaziat, 1982).

In the rocks of Quatre Cantons large colonies are observed. The most striking association is in conglomerate bodies where microcodium has totally replaced whatever matrix there may have been and has extensively corroded the carbonate clasts

21b)
(fig. 3.

Most of the Microdium is the "corn cob" variety with a distinct axial canal (fig. 3.34e). In the conglomerate bodies, extremely dense lamellar colonies are seen. Microcodium is seen to form subhorizontal layers greater than a few mm thickness.

All publications to date agree that Microcodium develops beneath the sediment surface, in a variety of environments. It is also certain that the crystallisation of the calcite is a biochemical process (Fontes, 1975). At present, most authors believe that symbiotic associations of several micro-organisms (bacteria + fungi + rootlets) are responsible for the morphology of microcodium. It occurrence here only serves to confirm that the groundwater levels were low and that stable soil profiles were established and probably maintained for some time.

3.5.6.3. The Significance of the palaeosol colour mottling.

The development of intense pedogenetic colour in sediments of the Poudingue d'Argens Formation relates directly to the availability of iron silicates and oxides in the host sediment. The early genesis of haematite may relate directly to the abundance of haematite associated with the detrital grains. Its presence produces the dark red-purple hue of particular sediments at Quatre Cantons.

The principal source of primary haematite was almost certainly Permian mudstones. Where were Permian rocks exposed as early as the Eocene? The clue appears in the distribution of sediment with this purple hue. It is strictly limited to two outcrop areas in the study region: Quatre Cantons and Sausses (Barres de Martignac). Both these localities are adjacent to the present exposure of the Dome de Barrot. The potential source is defined as proto-Barrot. Faults at the margin of the Permian basin were inverted in response to Pyrenean and early Alpine compression early in the Eocene (discussion, section 2.6 and fig. 2.39). A large amount of uplift was achieved with relatively little shortening.

3.6. General comments on the nature of the drainage basins and source areas.

3.6.1. The geometry and origin of the palaeotopography required during deposition of the Poudingues d'Argens Formation.

The limited and apparently unconnected nature of the Poudingues d'Argens alluvial fan systems suggest that Haute Provence was divided into small drainage basins surrounded by hills with significant relief. At certain times, parts of the basin were not well drained and lakes formed. The basin was certainly not a simple trough with an axial fluvial system and marginal fans. (FIG 3·33)

The principal source of fan sediment was Upper Cretaceous limestone of facies specific to Haute Provence, derived from local source areas within the developing basin. The development of stream dominated fans imply that the palaeorelief was not extreme: this is characteristic of folded terrains rather than areas with steep fault scarps.

The size of catchment areas cannot be defined from the limited outcrop of Poudingues d'Argens Formation. However, structural mapping has identified areas of tectonic uplift which most likely sourced the Poudingues d'Argens fans. The amount of shortening and the lateral continuity of the thrusts and kink zones during the Early Eocene, have been estimated using partially restored cross sections. For example, an inverted extensional fault of great lateral continuity, plus emergent thrusts of limited areal extent, produced a broard area of uplift along the Beauvezer valley which shed sediment onto the Le Ruch - Peyresq fan.

The fan system of Le Ruch and Peyresq was supplied with coarse boulder conglomerate detritus from beginning to end of its history. The succession does not fine upwards overall, and so the catchment area was not denuded with time. Therefore, the hinterland was actively uplifting during sedimentation of this extensive fan system.

Uplift associated with active compressional structures can account for all the Poudingues d'Argens successions in Haute Provence. The interaction of both Alpine and Pyrenean structures gave rise to a complex three-dimensional topography which divided Haute Provence into several small drainage basins. The integrated structural and sedimentological data were used to reconstruct the palaeogeography of the region during deposition of the Poudingues d'Argens Formation, presented in encl. 6.1 . There are two areas to highlight:

i) the Maures-Esterel massif was already structurally high and eroded to basement, acting as a source for the St. Pierre system.

ii) Permian mudstone and quartz are seen in the Poudingues d'Argens of the outcrop areas fringing the Barrot massif (Quartre Cantons and Sausses). The source area was undoubtedly local which implies that there was already a structural high at Barrot, eroded down to the Permian. The tectonic implications have already been discussed in section 2.6.

3.6.2. Why did the Poudingues d'Argens alluvial fan systems develop and when?

Alluvial fans form when, for whatever reason, the river systems in an area are not in equilibrium with the sediment supply. They are the response of that river system to increased sediment production in the catchment area, or to abrupt changes in surface gradient. Climatic variations and tectonic activity are the two principle controls of sediment supply.

In the case of the Poudingues d'Argens Formation, the alluvial fans developed in response to regional tectonics:

i) Large volumes of Mesozoic terrain were eroded from the S.W. Alps between Maastrichtian and Lutetian times. These terrains were probably drained by large mature river systems, which took large volumes of sediment out of Haute Provence. None of this sediment appears to be stored in Haute Provence.

The interpreted alluvial fans of the Poudingues d'Argens Formation were supplied by ephemeral braid-plain channel systems. None of the sequences can be interpreted as the deposits of a well developed, mature river system. Therefore, Haute Provence was an area of net erosion during the Late Cretaceous and Palaeocene. For long periods of time sediment was transported out of Haute Provence.

Alluvial fans developed in Haute Provence in the Mid Eocene as a direct consequence of deformation. The resulting changes in topography, with new hills, breaks in slope and isolated low areas, dramatically changed the drainage patterns. Rivers could no longer pass across Haute Provence. Fans developed at the breaks in slope.

ii) The relatively short time represented by the alluvial fan systems is confirmed by the palaeosols preserved in the Poudingues d'Argens Formation. They were able to develop rapidly because of the plentiful supply of carbonate. There are no erosional gaps of significance in the sedimentary record, for example, at Argens. The whole succession was probably developed in much less than 100,000 years.

At Peyresq, the pedogenic modification of the overbank facies is minimal. Therefore, it is reasonable to assume that there were no significant periods of non-deposition during the sedimentation history of the alluvial fan at Peyresq. Sedimentation rates were high in this mid fan position. Therefore, succession does not represent long periods of time.

iii) At Peyresq, authigenic pyrite has been observed near the base of the Poudingues d'Argens Formation. The form is framboidal and almost certainly very early diagenetic: it occurs in marlstones where poorly developed soils of grey character. It appears that the whole succession may have been deposited quite close to a shoreline. At the very least, marine or brackish pore water infiltrated the sediment not long after deposition.

These three points suggest that the Poudingues d'Argens represent a small time window between the onset of thrusting in the foreland and the Nummulitic transgression.

3.6.3. Synthesis.

A model is peresented which links the development of alluvial fan systems to the onset of compressional tectonics in the region. The deformation did not begin long before the Nummulitic shoreline extended onto the European foreland. Infact, the subsidence that gave rise to the transgression was probably induced by the same Alpine and Pyrenean orogenic forces that produced compressional structures in the foreland during the Mid

Eocene.

1. The regression at the end of the Cretaceous was a regional one caused by uplift related directly to Pyrenean tectonics. The river systems established at this stage reached equilibrium with the Palaeocene topography. Haute Provence was a region of net erosion. The contemporaneous depocentres were Languedoc and the North Pyrenees.

2. In the Middle Eocene, compressional structures began to develop in Haute Provence in response to Early Alpine and Late Pyrenean tectonics. The structural configuration was complex, and NW-SE and W trending folds induced a surface topography that divided the Provence into small drainage basins.

The pre-existing river systems were not in equilibrium with the emerging topography. New catchment areas were defined and different stream patterns established. Some drainage basins were isolated until streams had eroded through the bounding topography.

Therefore alluvial fans were constructed: river flow was impeded by the new topography, and there was a rapid increase in sediment supply from the juvenile and activity uplifting catchment areas.

4. The systems were preserved because the region was drowned during the nummulitic transgression. If the region had been

given time, river systems would have adjusted to the new topography. The fans would have been eroded and pediment surfaces created. Instead, the sea level rose and the Poudingues d'Argens fans were eventually buried, but not without a struggle! (section 4.2.5) under shallow marine carbonates.

CHAPTER 4: THE CALCAIRES NUMMULITIQUES FORMATION.

4.1. Introduction.

4.1.1. Regional setting.

The Calcaires Nummulitiques Formation forms distinctive prominent ridges or cliffs of grey to yellowish grey bioclastic limestone. It is characterised by the well cemented, massive appearance of the limestone and by the abundance of the foraminifera Nummulites.

The formation records the trangression, during the Eocene, which established marine conditions throughout the foreland basin fringing the Alps. Similar limestones outcrop in the Austrian molasse basin (Kittler and Neumayer, 1983), the Bavarian molasse basin (Bachmann and Koch, 1983) around the western Alps, and across the SW Alps as far west as Digne and Devoluy. The westernmost and southernmost outcrops of this formation mark the limits of the transgression rather than erosional remnants (Fig. 4.1). Equivalent limestones of the same age and slightly older are exposed in thrust slices of Sub-Brianconnais and Brianconnais which originally lay to the east of Haute Provence. Therefore, the Nummulitic Sea extended some distance to the east: how far is difficult to ascertain, but it is likely that much of the Alpine chain was submerged, because it was not a source of sediment at the time.

Over most of the study region, the Calcaires Nummulitiques Formation overlies Upper Cretaceous, above a significant and often angular unconformity (fig. 4.2). To the west and south, the unconformity cuts down to Lower Cretaceous (West Barreme syncline) and Jurassic (Roquesteron, Castellane and Rousillon). To the north, on the southern margin of Pelvoux, the Calcaires Nummulitiques lies directly on Hercynian crystalline basement. In certain areas, a continental sequence, the Poudingues d'Argens Formation underlies the Calcaires Nummulitiques Formation and passes transitionally into it.

For many years, it has been recognised that the age of the Calcaires Nummulitiques Formation varies between outcrop areas (Boussac 1912; Goguel, 1936, 1953; De Lapparent, 1938; Gignoux, 1950). The oldest overlies Sub-Brianconnais facies Mesozoic. On the European foreland of the southwest Alps the oldest is of Lutetian age, in the South East Maritime Alps around Nice. The youngest are Priabonian, in the western exposures. Campredon (1977) summarised the data (Figs. 4.3) and showed that the Nummulitic Sea advanced westwards across the foreland.

The eventual limit of this transgression coincides with a Mesozoic facies boundary around Haute Provence. The transitions from platform to basinal facies during the Mesozoic were controlled by extensional faults. These faults were reactivated in compression, i.e. inverted, during the Eocene, and the Nummulitic coastline appears to have been defined by these zones of uplift (discussed section 6.6.4).

The thickness of the Calcaires Nummulitiques Formation varies over the study region, subtly in most outcrop areas, thinning over a distance of several kilometres from 30m to 5m or less towards interpreted palaeohighs (Fig. 4.4, section D). It is possible to map the thickness trends qualitively. However, the formation often forms the crests of ridges, so its top is present-day erosion level. In such places the original thickness can only be estimated. In some outcrop areas dramatic thickness variations are observed: for example, near Scafferels, it increases from 20m to more than 300m across the Rouaine fault zone.

For the most part, the Calcaires Nummulitiques Formation comprises open marine, bioclastic limestones deposited below storm wave base. However, at some outcrop areas, a basal facies association is observed, comprising brackish-open marine marlstones and conglomerates with clasts perforated by lithophaga borings. The formation is therefore divided into the Mort de l'Homme and Scafferels Members, where the names are the type localities of the basal facies association and the open marine bioclastic limestones, respectively.

4.1.2. The relationship between the Poudingues d'Argens Formation and the Nummulitic transgression

i) The upper parts of the Poudingues d'Argens Formation record a rise in base level, evidenced by the presence of gley soil profiles and an increased rate of deposition of fine grained facies associations. Authigenic framboidal pyrite is observed in the gley palaeosols which indicates that salt water was present in the soil pore waters.

At some localities, particularly Allons, there is no obvious break in sedimentation from marlstones without fauna, to marlstones with brackish fauna, to those with insitu corals.

ii) The conglomerate assemblages of the Mort de l'Homme Member are not simply conglomerates reworked from the Poudingues d'Argens Formation. Coarse conglomeratic detritus was actively supplied into the shallow marine systems. The localities where Mort de l'Homme conglomerates outcrop are coincident with coarse member dominated Poudingues d'Argens Formation over most of the region. This, and the textural and compositional similarity of the conglomerates, imply that the same alluvial systems supplied the Poudingues d'Argens and Mort de l'Homme conglomerates.

4.2. **The Mort de l'Homme Member.**

4.2.3. The occurrence of this member is limited, and its thickness is variable (0-15m). Its distribution is directly related to that of the Poudingues d'Argens Formation. It is thickest overlying the Poudingues d'Argens, and is not present more than 500m from its limits. The facies associations of this member are used to demonstrate that this member is not just the product of reworking the *Poudingues d'Argens Formation during the Nummulitic* transgression. The source areas which supplied the Poudingues d'Argens depositional systems continued to supply those areas for some time as the basin subsided. Sedimentation on these shorelines kept pace with, and at times outpaced, the sea level rise until several metres of shoreline sediments had accumulated. The transgression continued, and subsequent deposits record the shoreface retreat and the deepening of the marine basin.

The following sections describe the standard facies of the Mort de l'Homme Member and interpret the sub-environments of the shoreline which they were deposited. The overall nature of the Nummulitic transgression and the controls of sedimentation are discussed in Section 4.6. The unique occurrence at Cairas of olistostromes and associated debris flow deposits at the top of the Member is described and interpreted in Section 4.5.2.

4.2.2 Conglomerate Facies

The conglomerates are composed almost exclusively of Cretaceous limestone with 1-20% of flints, in a coarse carbonate-sandstone matrix. The appearance of some conglomerates of the Mort de l'Homme is similar to those of the Poudingues d'Argens formation. They are distinguished by the *Lithophaga* borings (fig. 4.5) which perforate some or many of clasts. Also, the matrix of these horizons contains fragments of marine fauna and often whole *Nummulites*. Both these features demonstrate that the conglomerates were deposited in a marine environment. *Lithophaga* borings are characteristic of upper shoreface and beach deposits.

4.2.2.1. Facies C1.1

Sheet conglomerate bodies (thickness up to 7m, lateral extent often 50m) of clast supported, pebble-boulder conglomerates.

These conglomerate bodies have smoothly concave upwards, or flat, erosive bases. The upper contacts are sharply gradational into medium-coarse calcarenite or bioclastic limestone which drape the clast relief. The upper surfaces are flat, or dip gently ($<15^\circ$) in a consistent direction.

The clasts are well rounded micrite or rounded - angular chert, with a moderate sphericity; they are poorly - moderately - well sorted. The matrix (5-30% of sediment) comprises grey, coarse sand grade detritus which is poorly - moderately sorted. Bioclastic fragments are relatively rare.

Lithophaga borings are not common through most of this facies. Bored clasts are concentrated at certain horizons, defining flat and dipping (up to 30°) surfaces within the conglomerate and at their tops. The dipping horizons can be traced from top to bottom of a body in some places (up to 4m height); in others, they are truncated by internal erosion surfaces (AR3 and 25; Enclosure 4.1). Bored clasts are rare elsewhere in this facies. The pebbles do not have borings; the cobbles and small boulders (up to 20cm) are bored all around their circumference; the larger boulders, which are rare, are bored preferentially on their upper surfaces.

These horizons may be overlain by thin intercalations of sandstone which contain a high percentage of bioclastic remnants (much more than the conglomerate matrix). More commonly an internal erosion surface slightly truncates the layer.

Sedimentary structures are not common:

i) There are no basal surface grooves or gutter casts.

ii) Pebble imbrication is rarely observed (fig. 4.6).

iii) Flat lying clasts ('a' axis horizontal) are common in single, thin (up to 20cm) beds and in the pebble horizons within and at the top of the conglomerate bodies. Otherwise, the clasts are randomly, but stably oriented.

Interpretation.

The conglomerates were deposited rapidly from flows with very high concentrations of coarse sediment. After each depositional event, the tops of the conglomerates were reworked by persistent currents in a marine environment. The larger clasts were reworked only episodically, as evidenced by the bored clasts which show that these clasts were stable long enough to be colonised by boring bivalves. The episodicity probably reflected the frequency of storms in the basin.

iv) Rare cross stratification is observed (fig. 4.7) defined by rapid clast size variations. Single, tabular sets up to 2m thick are partially preserved beneath internal erosion surfaces. These sets cannot be traced far down-palaeocurrent. The stratification is at 20-30° to palaeo-horizontal.

Interpretation.

The clasts were deposited on angle-of-repose surfaces which were slip faces to sizeable (up to 2m), two-dimensional bedforms. The bedforms were relatively short lived, implying that the flows were unstable and/or enough (immediately prior to deposition, at least) to carry their sediment load in suspension.

v) The conglomerates are not graded on any scale. Abrupt variations in clast size are observed, such that pockets of pebbles exist in a boulder conglomerate or a thin layer (up to 10cm) of pebbles forms the upper surface of a conglomerate bed.

Interpretation.

The conglomerates were deposited rapidly from a flow in which the sediment load was not well stratified prior to deposition. While these bodies were deposited, the clast size range supplied to each locality did not change.

The conglomerate bodies can be traced down-palaeocurrent for up to 1 km at some outcrop localities (in particular, Argens). The bodies thin down-current, often dividing to produce a vertical sequence of alternating thin conglomerate beds and well laminated, more or less bioclastic sandstone, facies C2.3 (fig. 4.8). Further in the same direction, the conglomerates thin into stringers of single pebbles and rare cobbles (enclosure 4.1), seen at AR18. They also occur as the basal conglomerate of fining-upward carbonate sandstones (facies C2.4).

At Argens, there is distinct upper surface relief on the conglomerate body. This gentle dip is in the direction the body thins, parallel to palaeocurrent directions interpreted from the foresets.

The cross stratified units do not coincide with the occurrence of the dipping bored clast horizons, but they are oriented the same way.

The presence of bored clasts implies that these high angle surfaces did not receive sediment for some length of time, during which the clasts were rolled so that the bivalves colonised all sides of the clasts, but the angle of the surface was not reduced significantly!

These bored clast horizons may represent slide scars, or more likely, an abandoned depositional surface which was not the slipface of a migrating bedform, instead the progradation surface of a small fan delta.

At this locality, the structural evidence (cross section Z, enclosure 14) suggests that, all these surfaces face down palaeoslope.

Overall interpretation of facies C1.1

The presence of *Lithophaga* borings and Tertiary marine bioclasts demonstrate that the conglomerates were deposited in a marine environment. The conglomerates are, for the most part, structureless and thin rapidly in the direction of flow. This implies that they were deposited from a series of rapidly waning flows. The currents were erosive, sometimes scouring channels, and flowed down palaeoslope. The general absence of bored clasts has been used to infer that these conglomerates actively supplied by rivers which constructed fan deltas at the shoreline. The source of the clasts, and the nature of the currents which transported them, is discussed in section 4.2.2.4.

The bored clast horizons are interpreted to represent periods of non-deposition during which larger clasts were colonised by bivalves. The clasts were rolled - the pebbles too often to be colonised, the larger ones more rarely - by episodically strong currents. These currents were probably waves, with enhanced strength during storms. The bored clast horizons are used to define individual events of sedimentation in conglomerate bodies: they are sheet and wedge shaped units of maximum thickness 5-200cm. The time lapse between depositional events varied. It is to a certain extent reflected in the concentration of borings and the thickness of the bored clast horizons. However, the basin currents themselves transported and deposited sandgrade detritus, burying the conglomerates and killing the lithophaga bivalves.

4.2.2.2. Facies C1.2

Simple sheet bodies (thickness up to 5m, lateral extent very variable) of matrix-clast supported pebble-boulder conglomerate. The sheets have sharp, flat basal and upper contacts.

This facies is characterised by:

i) A variable, but often high - up to 50% - matrix content.
ii) Clasts in unstable orientations within the deposit.
iii) Randomly distributed bored clasts within the conglomerate, with a somewhat higher concentration at the top.

The sheets thin in the same direction as those of facies C1.1. There are no palaeocurrent indicators or other sedimentary structures.

Interpretation

The fabric, high matrix content and lack of sedimentary structures imply that these conglomerates were deposited from debris flows. They are probably redeposited facies C1.1.

4.2.2.3. Special case of facies C1.1 and C1.2: fractured clast conglomerate

In certain restricted zones, up to 5 metres across, the conglomerate clasts are fractured, and the fractures are filled with a bioclastic calcarenite matrix (Fig. 4.9).

Some of the clasts are totally shattered, but all parts of the jigsaw remain in close proximity, supported by the sandy matrix. In other cases, a fracture partially dissects the clast and is widest where that clast is in contact with another. Some clasts are unfractured.

The rock could be described as a matrix rich, clast supported breccia. It is laterally equivalent over very short distances to normal, unmodified facies C1.1 and C1.2. Such a breccia has been observed at Le Ruch and Cairas which are near interpreted palaeohighs (section 4.5.2, cross section W).

Interpretation

The preservation of all the constituent parts of the fractured clasts adjacent to one another implies that the fracturing process occurred post deposition of the conglomerate.

The clasts appear to have fractured as a result of impact with adjacent clasts (fig. 4.9). Penetration of the sandy matrix into fractures suggests that the bed was liquified during the fracturing process. The fact that the fragments did not move far from their original positions suggests that liquifaction was very brief.

Merely liquifying the sediment would not have fractured the clasts. Some other process must have occurred to cause the fractures and liquify the matrix.

There is a critical clue these breccias lie above deformed zones in the Cretaceous substratum. At Cairas, the breccias are limited to the area above a growing kink fold. Activity on structures near surface, eg. thrusts beneath the folds, or brittle kinking, may have caused earthquakes whose effects at surface were localised. I suggest that the sediment fluidisation and shattering was induced by a shock wave resulting from an earthquake. At Cairas, this is supported by the observation of intense, sand-filled fracturing in the Cretaceous in and near the kink. Local syndepositional tectonic activity is also indicated by the association of the breccias with olistostromes and debris flows; see section 4.5.2 for a discussion of the features at Cairas in a wider content.

4.2.2.4. Facies C1.3

A laterally extensive, clast supported pebble-boulder conglomerate of variable thickness, from a few pebbles to tens of centimetres. It is characterised by intensely bored clasts; only the clasts less than 5cm across are not perforated. Whereas individual deposits of the other conglomerate facies are lenticular, these horizons of thoroughly bored pebbles can sometimes be traced 50m along outcrop.

Interpretation

The clasts remained at surface for long periods of time and were almost all colonised by bivalves. This facies is interpreted as the result of extensive reworking of facies C1.1 and C1.2 by marine currents, like waves. The clasts were moved from their original sites of deposition, and formally lenticular deposits were distributed over a large area (Clifton, 1973)

These horizons may therefore represent periods of abandoment of, for example a fan delta mouth bar.

4.2.2.5. Interpreted depositional environment and source of clasts in the Mort de l'Homme Member.

There are three possible sources of Cretaceous limestone clast in the marine environment:

i) Erosion of the Upper Cretaceous Substratum by powerful waves at the coast and in the shoreface erosion zone.
ii) Transport to the shallow marine environment by high energy rivers.
iii) Reworking of Poudingues d'Argens.

The first mechanism is rejected for three reasons:

i) It does not explain the close association of Mort de l'Homme and Poudingues d'Argens conglomerates.
ii) There is no evidence of substantial wave reworking in the conglomerates.
iii) Some large clasts are well rounded but only bored on their top surfaces. The currents and waves active in the shallow marine environment were therefore too weak to turn them over. These

processes are thus unlikely to have been responsible for excavating and rounding the clasts; this must have happened before they entered the sea.

The strong association of Poudingues d'Argens and Mort de l'Homme can be explained in two ways:

i) the transgression caused some reworking of the existing alluvial fans;

ii) the river system that fed sediment in to the Poudingues d'Argens fans continued to supply clasts to the shallow marine environment.

In most cases, horizons of bored clasts are underlain by thicker sequences of unbored clasts whose facies indicate deposition from unidirectional downslope currents. These are interpreted as the product of rapid transport by rivers into the marine margin during storm floods. These events were separated by long periods of time when the top of each new pulse of sediment was gently reworked and bored, producing a well-packed conglomerate layer with a horizontal fabric, composed of bored clasts (fig. 4.11, facies C1.3). The original matrix was winnowed, but replaced with more bioclast-rich sand.

The conglomerates between the bored horizons are dominated by the facies of <u>river channel mouth bars:</u>

(i) poorly structured conglomerates deposited rapidly from high sediment concentration flows, particularly when the river is in flood;

(ii) resedimented mass flow conglomerates which are frequently produced when the mouth bar sedimentation rates are high and the depositional slopes produced are unstable.

The coarse bed-load streams that might generate mouth bars deposits such as these seen in the Mort de l'Homme Member are known to have existed: they generated the Poudingues d'Argens Formation. Therefore, it is probable that the same, or similar rivers continued to supply of sediment into the shallow marine environment. The Mort de l'Homme conglomerates were therefore deposited on fan deltas (Kleinspehn et al. 1984, Hayward, 1985, De Freytet and Molenaar, 1984), produced as the transgression progressively drowned active alluvial fans (see fig. 4.30 and 4.31).

4.2.3. Limestones C2

These are sand grade or recrystallised limestones with variable percentage of bioclasts.

4.2.3.1. C2.1 A massive, bioclastic limestone of limited lateral extent (a few metres up to 15m). This limestone is rich in corals, both solitary and colonial. Coralline algae and bryozoa are extremely common, and a wide variety of unbroken shelly fauna is present. The algae can be seen to encrust small grains, forming Rhodoliths.

Interpretation

The good preservation of coral and coralline algae suggests that these limestones represent the site of reefs. The restricted development of this facies implies they were patch reefs.

4.2.3.2. C2.2 A laterally extensive (10s of metres) bioclastic limestone. It has a massive appearance, often with a weakly developed spaced (5 to 25cm) pressure solution cleavage.

It contains much detrital carbonate: grains of micrite, rare pebbles and cobbles of Cretaceous Limestone and a large percentage of bioclast material.

The remaining fauna is mostly fragmented - the sediment is rich in fragments of coralline algae, corals and bryozoa.

Nummulites abound and these elongated fossils are often horizontally aligned, in some places imbricated (Fig. 4.12). In the photo, the imbricated fossils form horizontal layers 1cm thick; adjacent layers often have opposing imbrication fabric. Otherwise the nummulites are randomly distributed or clustered in small elongate patches a few cms across.

In some localities (noteably PQ4), the limestones display small (in the order of 2 metres) stacked, coarsening upward cycles (PQ4; fig. 4.13). These units are rare, because the original fabric has often been replaced, for example, by micrite.

Interpretation

The broken shell fauna and the alignment of nummulites imply that these limestones have been reworked in a shallow marine environment, most probably by waves. The lack of similar features through much of the limestone may be explained by bioturbation, for which there is evidence: the patches of nummulites may represent burrows. The significance of the sequences cannot be determined from this facies alone.

4.2.3.3 Subfacies C2.3

These carbonate clastics are well sorted, fine to very coarse sand grade detritus with rare pebble to cobble clasts of limestone, comprising grains of reworked Cretaceous limestone and bioclasts. Bioclasts are rare-common; where present, nummulites is the commonest species. The sediments are also rich in organic debris (fragments up to 1cm length, commonly a few mms). The carbonates are very well laminated, with excellent pcl developed on many surfaces. Parallel and cross lamination and cross bedding (the cross stratification is commonly low angle, up to 15°) are observed fig. 4.14. Tabular and trough cross sets up to 50cm set height are preserved, and cosets are observed (log AR5, enclosure 4.1).

Interpretation

These sediments were deposited from fast, steady palaeocurrents, in which moderately sized bedforms were created and maintained for some time. The palaeocurrents are variable, but in general oppose those of the conglomerate bodies with which this facies is commonly associated. The origin of the currents responsible for this facies are not known (Longshore?). It is not possible to discern larger depositional forms (bars) to which such bedforms may have related.

4.2.3.4 Subfacies C2.4

This facies comprises granule - fine sand grade carbonate clastics which are well bedded. The beds are sheet-like within an exposure and their thicknesses range from 10cm-3 metres. The basal contacts are flat and erosive; the upper contacts are sharply gradational or truncated by the base of another deposit. Immediately above the base, a pebble-cobble conglomerate lag may be present. Above this, the bed is normally graded from very coarse-fine sand. The beds are commonly amalgamated; thinner beds are separated by marlstone with thin sandy laminae. No sedimentary structures are observed.

The sediments are composed primarily of micrite grains (reworked Cretaceous) plus a variable percentage of bioclastic debris. Nummulites are rare. At the base of one bed a large, 2m long, fragment of colonial coral is preserved on its side. Shale (marl) clasts do occur and are concentrated in the basal conglomerate lag. The common constituent of this lag are Cretaceous limestone clasts: both well bored and uncolonised varieties.

Thalassinoides burrows bioturbate the fine grained sediment between these limestones (fig. 4.15).

Interpretation

The sediments were deposited from discrete waning flows, which were probably storm generated, geostrophic currents. The presence of mud clasts at the base of the beds implies that these flows did not travel long distance (with time the flow becomes stratified with less dense components (e.g. mud) towards the top of the flow).

These beds are the distal and lateral equivalents of some of the conglomerate bodies.

4.2.4. Fine-grained Facies: C3

The fine grained facies of the Mort de l'Homme Member are almost entirely concentrated at the base of the succession (see logs AR25 and AR3 on enclosure 4.1). They form laterally extensive sheets. At Argens, these can be seen to pinch out up the inferred palaeoslope.

There are two principal facies and where both are present, the open marine marlstone (C3.2) overlies the brackish marine marlstone (C3.1). The contacts are gradational.

4.2.4.1. C3.1 Marlstones with bioclastic debris of granule grade and/or insitu brackish fauna and tiny nummulites (1mm diameter!). The fauna include gastropods and simple, large bivalves.

Within these marlstones, thin coals (1-10cm) or horizons extremely rich in organic debris occur. These horizons comprise thin (up to 1mm) laminae of organic material alternating with marlstone. These horizons are also rich in unbroken fauna, particularly gastropods.

At one locality (AR16c), these sediments are slightly colour mottled, with gley profile type mottling, (fig. 4.16; see section 3.5.3.4). They are also bioturbated and vertical, long, tapering, carbonate sand-filled (fig. 4.17) features are believed to be the fill of hollow roots (cf. these developed at the top of the Poudingues d'Argens - AR18) like those of mangroves. Bedding planes at intervals of 50cm are defined by finer grained marlstone. These sediments were deposited from shallow brackish-(open) marine water under low energy conditions. The sedimentary environment was colonised by plants, such that soil profiles were preserved (colour mottling and organic rich horizons). These plants had hollow roots which implies that the substrate was somewhat anoxic. From this facies alone, it is not possible to say whether the environment was a lagoon or extensive coastal marshes colonised by mangroves. Such environments are flooded episodically by sea water during storms which may account for the observed grain size fluctuations.

4.2.4.2. C3.2 Marlstones with large colonies of coral preserved in life position. These micritic sediments also contain unbroken bivalves (fig. 4.18).

This facies is observed at one locality AR7 (Allons) and occurs above the facies C3.1 and immediately beneath the erosive base of the Scafferels Member. Here, no conglomerates are developed in the Mort de l'Homme Member.

The corals imply that for some time the seas were shallow and clear. It is a puzzle what substrate they grew on, since they occur above a brackish water marlstone. However, it is probable that sedimentation rates increased such that coral growth was stifled and the corals were preserved in marl. The fine grained nature of the sediments and the preservation of delicate coral structures and bivalves imply that deposition occurred in low energy marine waters: a back reef lagoon, for example.

4.2.5. Facies Associations of the Mort de l'Homme Member and the Interpreted Depositional Environments.

The member comprises sedimentary facies which represent quite different subenvironments of a shallow marine and shoreline setting. For examples of this see logs on enclosures 4.2 and 4.3.

Not all the facies are lateral or distal equivalents of each other: thick (1-20m) developments of fine grained facies are only observed at the base of the succession in certain outcrop areas (for example, Argens, Allons, St. Antonin). They are not laterally equivalent to other facies of this member. Rather, they appear to be equivalent to the correlatable white marlstone, developed at the top of the Poudingues d'Argens Formation at Peyresq and Le Ruch in particular. Elsewhere, there is no equivalent facies and no development of the Mort de l'Homme Member: the erosive base of the Scafferels member lies directly on Upper Cretaceous limestones.

The remaining conglomerate and limestone facies may overly the fine grained sediments, with a somewhat erosive or very sharp contact between them: at Argens and Allons, for example. At Le Ruch, conglomerates with bored clast horizons directly overly the white distal flood plain marls.

The conglomerates and limestones represent time equivalent sub-environments of deposition. The shoreline depositional systems are interpreted as fan deltas, where the clasts are derived from erosion of the same palaeohighs which supplied the Poudingues d'Argens alluvial systems, plus some additional sources created as a result of renewed tectonic activity within the developing foreland basin.

It is difficult to reconstruct the exact geometries of the depositional systems from the available outcrop, but at Argens in particular, a good section, parallel to many palaeocurrents, reveals the equivalence of facies and the presence of large scale primary depositional surfaces, which prograded southwards.

The conglomerate facies (C1.1 and 1.2) are interbedded with well laminated and low angle cross bedded limestones C2.3. Distally equivalent sections are somewhat thicker and comprise interbedded turbiditic limestones C2.4 and well laminated cross stratified limestones (C2.3).

At Le Ruch a similar pattern emerges, with the addition of patch reefs C2.1, which are best developed adjacent to the conglomerate bodies and somewhat down palaeocurrent of them; i.e. where the water was deeper and less charged with sediment.

The very bioclast-rich, well bioturbated limestones, with evidence of wave reworking (imbricated nummulites), C2.2 are observed where no conglomerates exist, and are therefore interpreted as the facies of normal basin sedimentation, close to shore, away from river mouths. Their existence, without pebbles, demonstrates that the marine basin processes did not redistribute the conglomerate clasts far from the river channel mouth.

This facies contains much debris derived from the patch reefs (corals, algae, bryozoa), eroded by the waves. In some cases the limestone may be reef talus; in other cases, the presence of oysters suggests that the environment of depositional was shallower, perhaps inshore or along-shore of a similar reef.

The conglomerate bodies are interpreted as mouth bar deposits of ephemeral channel complexes similar to those observed in the Poudingues d'Argens formation. At Argens, the large dipping surfaces within such a body are interpreted as progradation surfaces of such a mouth bar complex, or fan delta (the surfaces dip to the south away from the interpreted source areas).

In the Poudingues d'Argens Formation, the channelised conglomerate associations are believed to be the deposits of flows with very high concentrations of sediment. The nature of the limestone/marlstone hinterland dictated that surface run off only occurred during exceptionally high rainfall. The channels which supplied the fan deltas would therefore have been active only during periods of torrential and sustained rainfall. They flowed into the basin, probably at high velocity scouring shallow channels close to the shoreline and rapidly dumping large, unsorted volumes of conglomerate on the mouth bar. In the proximal mouth bar regions, traction currents could maintain bedforms over short distances. The resulting deposits constitute facies C1.1. The mouth bar depositional slopes were steep and episodically failed to produce the debris flow - turbidity current deposits (C1.2 → C2.4). The sand grade material was presumably initially deposited from suspension further out on the mouth bar. It was reworked by basin currents, which redistributed this detritus, some of it back onto the proximal mouth bar (perhaps during storms by strong wave currents or longshore currents) to form the well laminated facies C2.3.

The small coarsening upward sequences observed in facies C2.2 probably reflect the progradation of river supplied sand grade detritus into the basin during single or several closely spacing flood events. Occasionally, pebbles and small cobbles would be transported to these regions adjacent to the active mouth bar complexes.

It is not possible to detect larger scale repeated cycles in the Mort de l'Homme member. A single cycle is observed everywhere; the succession is invariably:

i) distal floodplain or coastal plain and shallow low energy marine deposits; succeeded by:

ii) conglomerates and limestones, rich in detritus derived from channel complexes which fed directly into a marine basin. The lateral and distal equivalents of these sediments are patch reefs, and a variety of open marine bioclastic limestones.

The contact of the Mort de l'Homme member with the Scafferels member is sharp and somewhat erosive. It is probable that the fan deltas abruptly ceased to be active sedimentary environments and were rapidly drowned.

The succession records a continuous sea level rise which was interrupted by one period of shoreline progradation.

4.3 The Scafferels member.

This association of limestones is developed across the whole study region, comprising for the most part the monotonous, massive Nummulitic limestone which typify this formation. The base of the Scafferels member is always somewhat erosive and is planar over most of the region (fig. 4.19, at Annot). It is interpreted to be everywhere shoreface.

Over much of the region, it coincides with the base of the Calcaires Nummulitiques formation: the contact with the Cretaceous substratum is often flat (fig. 4.19). Over long distances, it coincides with the same bed in the Cretaceous limestone with minor perturbations (fig. 4.20) In certain areas, a marked angular unconformity is observed (fig. 4. , Ch am.) and in detail, the basal Calcaires Nummulitiques contact is irregular, preserving the eroded topography with, for example, a palaeocliff at the edge of a wave-cut platform (fig. 4.22). Here, clasts of Cretaceous may be preserved in the first overlying nummulitic limestones, but they are not bored.

4.3.1. Limestones C4

4.3.1.1. C4.1. Massive limestones which can be divided into units up to 1m thick, according to the percentage of fauna in the limestone. The original textures are not completely preserved: the bioclasts are supported in micrite, with detrital grains of quartz and glauconite. The percentage of detrital minerals is variable and over much of the region is never greater than 5%. However, at certain important localities, noteably Quatre Cantons and Sausses 60-70% quartz have been recorded, in these cases at the base of the member.

The faunal content is variable; many of the bioclasts are fragmented, although parts of the limestone contain large, unbroken echinoids (fig. 4.23), bivalves and corals, bryozoa and coralline algae. These limestones form discrete patches up to a few metres across and although recrystallisation has obliterated the original contacts, these patches are thought to be reworked. The commonest fossil species are Nummulites and reef corals, other foraminifera include Discocyclina and globegerinids.

There are two subdivisions of this facies:

C4.1.1 Sharp concave up and convex upward surfaces define units of maximum thickness 4 metres and of 10-30m lateral extent. With the exception of common, large (up to 15mm) Nummulites, all bioclasts are fragmented fossils. The limestones are otherwise massive and hard.

The limestones were deposited in high energy shallow marine environment from erosive currents which constructed large, positive sedimentary features on the sea floor, perhaps longshore bars. The lack of internal structures may be the result of diagenetic modification or bioturbation.

C4.1.2 Massive limestones with no sharp planes, simply variations both vertically and laterally in the percentage and components of the bioclasts (fig. 4.24). It is this subfacies which contains the largest fragments of reef talus.

These limestones are almost certainly thoroughly bioturbated. The majority are resedimented limestones, with quite large blocks of reef talus in some parts. It is not possible to define how most of the limestones were deposited or the water depths in which they were preserved, because most of the fauna (algae, etc) is reworked. They were probably transported, during storms, into water depths below the shoreface erosion zone.

This facies constitutes most of Calcaires Nummulitiques Formation and can form monotonous successions up to 40m thick. The structureless limestones overly the lenticular bedded facies, if both are present, and therefore probably represent a deeper marine environment.

4.3.1.2. C4.2 Erosively based, rapidly gradational topped, well bedded (5-40cm) resedimented limestones.

Individual beds are laterally extensive (10s metres) and tabular. Each bed is massive - normally graded (where original grain size is preserved) from granule-fine sand grade, well sorted limestone. The beds are internally structureless or extremely well laminated: parallel lamination throughout, or parallel with undulose lamination in the upper 20cm or less. The undulose laminae are low angle (up to 15°) and dip in all directions. Within these horizons low angle scours are observed: the relief is partially draped and then filled by well laminated limestone. Primary current lineation has been detected on some surfaces. The half wavelength of the concave upwards surfaces is up to 2-3m and the amplitude is up to 10cm making these very low relief features.

The bioclast component is broken shells, coralline algae and whole nummulites, av. 2-3mm diameter. The percentage of bioclasts is extremely variable. The limestone is rich in organic material which defines the delicate laminae. A coal horizon up to 2cm thick and continuous over 100m is observed at AR7 (fig. 4.25).

The beds are amalgamated or alternate with thin marlstones (silt-clay grade) which are fissile, with less bioclastic debris supported in lime mud. Vertical burrows are observed.

These limestones were deposited from several waning flows, and reworked by persistent traction currents. The upper undulose lamination is interpreted as hummocky cross stratification (fig. 4.26). Therefore, the medium and fine grained particles were sedimented from combined flow, almost certainly waves plus unidirectional storm generated, geostrophic currents. The background sediments show no evidence of reworking by basin currents and therefore deposition occurred below fair weather wave base, above storm wave base.

This facies is well developed at Argens where the palaeocurrents interpreted from pcl, are aligned parallel to the interpreted tectonic dip.

4.3.2. Fine Grained Facies.

Marlstone, comprising micrite (original lime mud) and bioclastic material. The species diversity decreases upwards (Besson et al, 1970) through the Scafferels member. This marlstone is preserved only between the beds of the turbidite deposites (C5.2) and towards the top of the formation.

The preservation of such marlstone implies low energy conditions, where deposition from suspension is the dominant process. This and the progressive loss of benthic foraminifera species and maintenance of globigerinids imply that the Scafferels succession records the deepening of the marine environment: quiet offshore sedimentation succeeds continual reworking of sediment in the shoreface.

4.3.3. Basal Conglomerate Horizon developed in certain Outcrop Areas

The base of the Scafferels member is always somewhat erosive. At a few localities, a very well sorted, clast supported, cobble conglomerate lines the flat erosion surface (fig. 4.27). The clasts are well rounded and most importantly all the cobble sized clasts are thoroughly bored. The matrix is a nummulite rich, bioclastic limestone.

The clasts were bored and subsequently reworked to achieve the moderate sphericity and excellent rounding. The bored clast horizon occurs only where the Scafferels member overlies conglomerate facies of the Mort de l'Homme Member and adjacent to those regions, notably up the predicted palaeoslope. The source of the clasts is therefore interpreted to be the Mort de l'Homme conglomerates, ie. ultimately, the palaeohigh catchment areas which remained emergent during the early stages of the transgression.

4.3.4. Facies Associations of the Scafferels Member and the Palaeoenvironmental Implications.

In the sequences observed all three limestone facies may be represented, or any single one, or any combination of two!

As the sea level eventually rose to drown the Mort de l' Homme fan deltaic conglomerates, the clasts were colonised by lithophaga. The sea level continued to rise steadily, and the wave action carried some clasts upslope on to the palaeohighs. In the process, the pebbles were reworked, rounded and redistributed to form a thin, continuous horizon of bored clasts on top of the shoreface erosion surface (fig. 4.27).

The deposits immediately above this surface are the lenticular limestones (bars) of facies C4.1.1, if they are present in the succession. It is difficult to precisely define the depths of water in which these units formed. I have observed them only at certain outcrop areas, (e.g. Allons), and believe they may be the deeper water time equivalents of the fan delta facies associations in the Mort de l' Homme Member. Accordingly, they represent the phase of progradation and were therefore deposited above the shoreface erosion zone. This is a tentative conclusion: unfortunately the outcrop is not good enough to develop the hypothesis.

The storm generated deposits occur further south, down the interpreted palaeoslope and are thought to record the same progradational phase, this time recorded is somewhat deeper water, but above storm wave base. Sediment accumulated in what was, at all other times, a zone of net erosion.

The massive, unstructured limestones of C4.1.2 overlie the lenticular units above a sharp, erosive contact. It is not possible to estimate the amount of sediment eroded at this surface, but it may be the second SFE surface passing through as the sea level rose to drown the fan deltas. In many outcrop areas, this facies constitutes the whole Calcaires Nummulitiques succession (e.g. A100).

At Cairas, well bedded, but massive storm flow deposits constitute 100% of the Scafferels member. The beds are 10-20cm on average and none greater than 40cm are observed. A few centimetres of fissile marlstone is preserved between each bed. No thick massive limestones of facies C4.1.2 are seen. This may have been the result of low sediment productivity rates. In addition, the sea level may have risen rapidly over a region of subdued topography. A deep water environment was established, but the low slopes may have inhibited the formation and evolution of density flows. Therefore, for much of the time coarse sediment was stored in shallower water depths, while at Cairas lime mud accumulated in situ and by sedimentation from suspension.

At many localities, e.g. A100, the massive limestone facies C4.1.2 grade directly into the Marnes Bleues Formation. The percentage and diversity of fauna decreases upwards and the percentage of micrite, originally lime mud increases. There are no very distinct sharp boundaries within the massive Scafferels facies. For this reason, it is most probable that the massive limestone facies sedimentation occurred, via a number of processes, offshore beyond the shoreface erosion zone.

4.4. Examination of the Transgressive and Progradational Sequences Observed at Different Localities Across the Study Region.

The section is written around enclosures 4.1, 2 and 3. Most of the detail is contained in the diagrams and logs, but important features will be outlined in the text, taking localities in order of geographical location from north to south, then west to east.

4.4.1 Dormillouse Outcrop Area (10 - 30m)

4.4.1.1. Seyne.

There is no Poudingues d'Argens Formation. The basal metre of the Calcaires Nummulitiques Formation (fig. 4.28) is very poorly sorted pebble-boulder conglomerate with angular-subrounded clasts of Cretaceous limestone, plus small, well rounded pebbles of sandstone (1%). None of the clasts are bored. Above this is massive limestone with discrete beds of the conglomerate all less than 1m thick.

The thickness of the formation thins dramatically as the unconformity beneath the Calcaires Nummulitiques Formation steps down to Middle Cretaceous limestones. The angular unconformity is located above a thrust duplex in the Mesozoic strata (which in part post-dates the Tertiary succession).

4.4.1.2. Le Lauzet-Ubaye.

The Calcaires Nummulitiques Formation overlies black Upper Jurassic well bedded limestones. The nummulitic limestones themselves are almost indistinguishable from the Jurassic, but are not bedded. The limestones are extremely hard and brittle. Thin sections demonstrate that the limestone is totally recrystallised.

Interpretation

The conglomerates at Seyne were almost certainly rapidly deposited from debris flows in some unknown depth of marine water. The limestone facies is otherwise C4.1.2, and therefore the succession was probably deposited offshore. The unconformity at the base of the Calcaires Nummulitiques Formation cuts down the Mesozoic stratigraphy rapidly in places related to particular structures. Significant erosion occurred prior to the transgression; the sea level rose rapidly over the outcrop area. Almost coincident with the sea level rise debris flows brought sizeable volumes of coarse sediment into the area. The source of the clasts is unknown, but probably from nearby maybe submarine relief created above active structures in the Dormillouse region.

Conclusion:

During the period of rapid subsidence associated with the transgression, the area experienced significant compression. The subsidence may have been the result of crustal loading by thrusting in adjacent areas.

4.4.2. La Grand Croix Outcrop Area.

There are variations in the thickness of Nummulitic limestone and the facies represented. The unconformity relationships and the facies can be interpreted to represent a transgression across an area with significant topographic relief. It is illustrated by Fig. 4.29 (see also figs. 4.21 and 4.22).

There is no evidence to suggest that tectonic activity was contemporaneous with the transgression at this locality. Significant erosion had occurred in places (Villars Colmars), but the topographic relief remained, with palaeohighs above intensely kink-folded Cretaceous strata. At Cabane des Auberons, a normal fault produced a few feet of surface topography which was filled by thicker Calcaires Nummulitiques.

4.4.3. Argens - Allons - Mge de Chamatte.

The details of the sedimentary features of the Mort de l'Homme sequence are summarised in a set of logs (enclosure 4.1).

The evolution of the fan-delta system and its relationship to the developing structures beneath it are summarised on fig. 4.30. The important features to note are:

The Calcaires Nummulitiques successions developed in the Argens - Allons - Chammatte outcrop area cannot be interpreted as the drowning of a passive topography. At Argens, the Mort de l'Homme conglomerate facies are derived from a source that did not supply sediment to this area during deposition of the Poudingues d'Argens formation. The new relief was created to the north and is interpreted to have developed above thrust sheet 3b. The palaeoslope was the lateral hanging wall ramp of this thrust sheet (see section Z).

Coarse detritus was supplied to Allons during the early stages of the Poudingues d'Argens Formation development. This supply was not maintained, and from then on Allons became a site of active subsidence with accumulation of fine grained sediment.

4.4.4. Le Grand Coyer - Le Ruch-Annot

The succession recorded at Le Ruch and Cairas strongly imply that tectonic actively induced changes in the topographic relief and thereby fluctuations in sediment supply to the sites of deposition. The relief envisaged is summarised in Fig. 4.31, where Annot is interpreted as the area of maximum subsidence and Le Grand Coyer as the source of coarse detritus to Le Ruch.

Important features to note include:

- the reduced conglomerate sequence at R2. At the present day, this site lies above a 10m scale kink fold in the Upper Cretaceous (section D).

Therefore, the Melina-Aurent kink zone developed prior to the deposition of the conglomerate. This small scale SW vergent fold diminishes rapidly to the E and W and therefore would have only influenced deposition over a small area.

- The site of patch reefs coincides with a larger scale zone of deformed Cretaceous the lateral extension of the Melina structure (section D). Again, it is feasible that this structure existed during the deposition of both Poudingues d'Argens and Calcaires Nummulitiques Formation. This larger scale of deformation fixed the position of breaks in topographic slope; in this case, defining a plateau region (fig. 4.31 and cross section D).
- The uppermost conglomerate of outcrop R14 (closest to the steep limb of the Melina structure) is of the smashed clast breccia subfacies (4.2.2.3). It is probably not coincidence that this occurs at the contact between Mort de l'Homme and Scafferels members. It would be in accord with the previously stated hypothesis that the subsidence (relative sea level rise) which produced the facies transition was related to intra-basinal tectonic activity (see also section 3.4.2.4).

4.4.5. Peyresq

The successions measured at Peyresq (encl. 4.2) display rapid facies changes across a distance of only 2km. The succession thickens to the west, where both the thickest conglomerate bodies and the thickest Scafferels Member limestones are observed.

The rapid lateral facies variations in the Mort de l'Homme Member reflect the localised input of conglomerate detritus at the river channel mouths. These conglomerates formed mouth bars with significant positive relief in places: they were not extensively reworked by basin processes, and so lateral thickness changes were preserved.

The overall trends in succession thickness and the predominance of conglomerate facies to the west of Peyresq imply that there was differential subsidence across the area. Ths is almost certainly a component of the regional trend but it may be accentuated by displacement on the normal faults observed to the west of Peyresq. (Refer to section 3.4.2.2, where similar displacements were used to explain the concentration of Poudingues d'Argens conglomerate bodies to the west of Peyresq village). It is also important to note that if a regional tilt towards the south west existed across Peyresq, sediment could not be transported further in that direction because of the counter slopes of the Cordeil high (enclosure 6.2).

4.4.6. Melina - Sausses - Quatre Cantons

These outcrop areas surround the present day Barrot Massif. They share some interesting features.

i) Melina

This small outcrop is interesting because the base Tertiary unconformity is well down in the Upper Cretaceous limestones, showing that the unconformity cuts down towards the margin of Barrot.

The outcrop comprises a basal conglomerate preserved in the lower limb of a 10m scale kink fold, which would appear to belong to the Poudingues d'Argens Formation. The clasts are Upper Cretaceous limestones and flints. The remaining sediments are a limestone arenite and a small amount of bioclastic limestone.

ii) Sausses

The outcrop of Poudingues d'Argens and Calcaires Nummulitiques lies unconformably on Middle Cretaceous at the Barres de Martignac and on Upper Cretaceous above Sausses. The Poudingues d'Argens sediments are purple-cream coloured. The basal Calcaires Nummulitiques is a pebble conglomerate to pebbly sandstone, 2-5m thick. The clasts comprise well rounded quartz sandstone pebbles, flints and a minor proportion of Cretaceous limestone. Small (1-2cm), well rounded, reworked ironstone nodules are observed, similar to those observed insitu in the Middle Cretaceous strata. The matrix and associated arenites comprise up to 50% quartz grains.

Above this, bioclastic limestones, rich in fragmented shelly fauna and nummulites, are observed (facies C4.1.2), which fine upwards into the Marnes Bleues formation.

There are no palaeocurrent indicators. It should be noted that this outcrop lies immediately adjacent to the steep and complex Daluis fault zone, which is the NW margin of the Barrot culmination.

iii) Quatre Cantons

The base of the Calcaires Nummulitiques Formation is a flat, erosive contact, above which approximately 10-20m of well sorted, coarse grained arenite are preserved. The sediment comprises 30-70% quartz grains, together with a recrystallised micrite and spar limestone with rare pebbles of Upper Cretaceous limestone. Fauna are rare in this facies (C4.1.1) consisting of Nummulites and other foraminifera.

Upwards, the quartz grains disappear rapidly to leave a massive bioclastic limestone (C4.1.2), which forms the rest of the Calcaires Nummulitiques succession (total thickness 40m).

There are two points to highlight from this locality:

- the absence of conglomerate bodies within the Calcaires Nummulitiques formation.
- the presence of **large quantities of quartz** in the basal faces of the Calcaires Nummulitiques Formation.

The first implies that the source of conglomeratic detritus, which fed the Roudingues d'Argens depositional systems, was denuded. To the west, topographic relief was repeatedly rejuvenated in response to compressive deformation. Here, there was no longer a supply of Upper Cretaceous limestone clasts, and the source of quartz grains was quite rapidly submerged. It would be reasonable to concluded that the topographic relief was passively drowned during the transgression.

The second demonstrates that a source of quartz existed which continued to suply detritus to this shoreline during the early stages of the transgression. At St. Antonin, closer to the

Maures Esterel Massif, the Calcaires Nummulitiques comprises pure limestones. Therefore the source of quartz must have been closer to Quatre Cantons. The Cretaceous and Jurassic strata could not provide such large quantities of quartz, therefore, Lower Triassic quantities must have been exposed at surface. Once again, the evidence suggests that Barrot was elevated adjacent to Quatre Cantons and that it had been eroded in places down to the Permian.

4.5. The Evidence for Tectonic Activity During the Deposition of the Calcaires Nummulitiques Formation and its Control on Sedimentation.

4.5.1. Introduction.

The influence of tectonic activity during the initiation of the Tertiary marine basin should be considered on several scales:

i) The scale of the basin itself

Was tectonically induced flexural subsidence responsible, in part or totally, for the transgression? If it was a regional sea level rise, were the sea level changes a response to Alpine tectonics (sl), or is there evidence for a eustatic sea level rise during the Eocene?

These questions are to be discussed in Chapter 6. They are critical to our understanding of the evolution of the Gres d'Annot basin.

ii) At the scale of topography within the basin

In this section, mesoscale evidence will be discussed which implies that topography was actively growing during the Nummulitic transgression. Surface relief can in some cases be directly related to particular faults and folds. Features at this scale include the progradation phase recorded in the nummultic transgressive succession and the St. Benoit fault, which is believed to have formed a sizeable palaeocliff during the transgression (Pairis, 1971).

iii) The scale of individual facies

There are exceptional facies within the Calcaires Nummulitiques Formation which can be used to demonstrate that earthquakes did occur in the region during the Nummulitic transgression:

a) The fractured, but insitu, clasts of conglomerate facies C1.1 and C1.2.

b) Olistostromes, debris flows and isolated, large (10s of metres) olistoliths of Cretaceous limestone recorded in the Grand Coyer outcrop area. They occur at one correlatable horizon across the area. The significance of these facies is discussed in section 4.5.2.

The exceptional facies corroborate the hypothesis that topographic relief within the basin was evolving as a direct result of tectonic activity, not in response to climatic changes or geomorphic processes alone.

In the following sections, the significance of the sedimentary facies associations of the Calcaires Nummulitiques succession will be discussed. Two outcrop localities will be used as case studies of the effect of tectonic activity on the sedimentary record of the Nummulitic trangression.

4.5.2. The Cairas Case History.

All the features are illustrated on enclosure 4.4.

4.5.2.1. Facies particular to the Cairas locality.

At one locality in the Grand Coyer outcrop area, the Calcaries Nummulitiques facies associations and their relationship with the Cretaceous substratum are quite remarkable! There are two facies, which are intimately related, and have not been described in the preceeding sections.

C7.1 A poorly to very poorly sorted pebble-boulder conglomerate (breccia) in which the maximum clast size observed was 20 metres across. The clasts are angular-(rounded) and composed of Cretaceous micrites. Locally fractured clasts are observed (identical to those of subfacies C1.1 and 1.2). The matrix of nummulitic limestone comprises 1-5% of the facies.

It can be termed an olistostrome. The distribution of this facies is very irregular and it grades laterally into facies C6.2.

C7.2 A poorly sorted pebble-boulder conglomerate, in which the average clast size is 5-20cm. The clasts are rounded-well rounded and composed of Cretaceous micrite. The conglomerate is matrix supported in nummulitic limestone. The rock is brown overall. It can be traced laterally over several hundreds of metres and grades eventually into bioclastic limestone of facies C2.2.

Both these facies are laterally discontinuous in the Grand Coyer outcrop area. They occur at a single correlatable level across the area. Over much of the area all that is seen at this level within the Calcaires Nummulitiques are single enormous olistoliths of Cretaceous limestone.

Both facies are interpreted as the product of mass flows: the first deposited from a flow transitional from slide to a debris flow; the second deposited from a cohesive debris flow.

4.5.2.2. Features related to syn-depositional tectonic activity.

The contact between the Calcaires Nummulitiques formation and the Cretaceous Substratum is an angular unconformity across the area. Over most of the area the unconformity is planar: locally parallel to bedding in the Cretaceous, over large areas slightly unconformable (angular discordance 15°). The unconformity planes off the tops of folds within the Cretaceous. However, at Cairas, the unconformity has a great deal of relief (in the order of 100m). Figures in enclosure 4.4 shows the angular relationships between bedding in the Cretaceous, bedding in the Calcaires Nummulitiques and the unconformity surface.

The field drawings of these photographs also show the facies assocations of the Calcaires Nummulitiques Formation (see also the summary log), and how they relate to the unconformity surface.

The first point to note about the unconformity surface at Cairas is that it approximately parallels bedding in the Cretaceous substratum. The kink folds have not been planed off by erosion; nor have steps in the relief caused by extensional faults. There are local scours at the base of the Calcaires Nummulitiques and the unconformity steps down through the Cretaceous towards the top of the steep fold limb.

A close examination of the Cretaceous substratum, near the unconformity, in that steep fold limb reveals an exciting phenomena. The Cretaceous rocks have shear zones several tens of centimetres wide and millimetre size tension cracks. These are filled with nummulitic limestone! At the hand specimen scale, the processes by which the mesoscale kink fold was generated were largely brittle. It is concluded that this fold developed at the surface, which would account for the brittle deformation processes. The fold was exposed subaerially for a limited time: fault gouge was eroded in places and small karsts were initiated, the cavities developing parallel to bedding planes and along joints and other fractures. The sea level rose and nummulitic limestone filled the cavities. The fold was not exposed for long enough to be significantly eroded.

This outcrop provides a clue to the origin of the olistostromes and extensive debris flows: high level tectonic activity. Brittle deformation of the land surface must have been accompanied by earthquakes. The presence of large isolated olistoliths deposited on nummulitic limestones over large areas at one correlatable level implies that these earthquakes were significant. Here those earthquakes induced sizeable mass flows and created topographic relief. Elsewhere, that relief was probably the catchment area for fluvial systems and fan deltas.

The sequence of events has been interpreted from the facies associations and their distribution. It is fully explained in the enclosure.

4.5.3 The St. Benoit Fault

This fault lies a few kilometres to the east of Annot. It is a NNE-SSW normal fault, downthrowing to the southeast. Activity on this fault throughout the deposition of the Tertiary sequence is evidenced by its strong control on sediment facies and thickness. Pairis (1971), suggested that the fault formed a substantial submarine cliff during the deposition of the Calcaires Numulitiques .

In the footwall e.g. at A100 (Fig. 4.24), the formation is 40m thick. The downthrown, hanging wall sequence at Pont St. Joseph (Fig. 4.32) is about 300m thick. The intraformational unconformity seen half way up this sequence demonstrates the syndepositional growth of this structure, as does the growth fold at Rouaine (Fig. 4.32a)

In addition, the sediment character changes markedly across the fault. Both the hanging- and footwall sequences are massive bioclastic limestone (facies C4.1.2). The footwall sequence contains some pebbles of Cretaceous limestone and a shallow marine fauna including large oysters, mostly unbroken, and Discocyclinids up to 1m diameter! (Pairis, pers. comm.). The unbroken fossils indicate that they were not transported far from their growth positions.

Adjacent to the fault on the downthrown side, there is a poorly exposed breccia of Cretaceous clasts in a matrix of nummulitic limestone. Pairis (1971) interpreted this as a scree breccia; on re-examination, I consider it to be a post-depositional tectonic breccia.

There is, therefore, strong evidence for dip-slip movement on the fault. Pairis (1971) suggested that there is a sinistral strike-slip component; this would be compatible with its relation to the main fault from which it splays, but cannot be proven.

4.5.5. The Progradational Phase Exhibited in the Calcaires Nummulitiques Succession.

Where the Poudingues d'Argens Formation passes gradationally into the Mort de l'Homme member, conglomerates reworked in the marine environment never lie directly above the alluvial conglomerates of the Poudingues d'Argens succession. Several metres of fine grained facies always separates them. These comprise distal floodplain - lacustrine marlstones and micrites, brackish lagoonal or coastal marsh marlstones and coals and shallow open marine marlstones with corals preserved in life position.

The thicknesses of the fine grained facies associations varies between outcrop areas. At Allons, for example, up to 12 metres of brackish - shallow marine marlstones are preserved beneath a shoreface erosion surface (i.e. much more may have been deposited). At north Argens (AR16, AR3), quite thin (1-3m preserved) brackish marine marlstones overlies mature grey soil profiles.

In all cases, during the early stages of the transgression sedimentation occurred in low energy environments of low topographic relief. The rates of sediment accumulation were not constant across the region: the varied sediment thicknesses imply differential subsidence rates within the basin.

Nowhere in the exposed sedimentary record were these early Nummulitic shorelines supplied by active fan deltas. This could be explained in two ways:

i) The transgression resulted in the superposition of facies from the coastline environments on distal alluvial plain on top of proximal alluvial fan channel complexes.

ii) The transgression coincided with (? was caused by) a phase of regional compressive tectonic activity which induced subsidence over large areas of the region. This resulted in an increase of flood plain aggradation on the alluvial plains. In consequence, fluvial channels were laterally confined (mechanism explained in section 3.4.2.4). Their outlets into the Nummulitic sea were narrow, and therefore the fan deltas were restricted. Evidence of their existence may have been simply eroded.

Whichever is true, if the transgression had continued smoothly the sediment succession would record the shoreface retreat, and shallow marine limestones would be overlain by their offshore equivalents.

Instead, the Mort de l'Homme member records the incursion of large volumes of conglomerate into the shallow marine environment. These conglomerates have been shown to derive from the continental environment.

At Argens, these conglomerates can be seen to thin downpalaeocurrent from the north (NW) towards the south (SE). The thickest conglomerates occur where the fine grained brackish water facies associations are thinnest (bedding planes within the marlstones very gently onlap a palaoesol horizon towards the north). At this outcrop area, these thick conglomerate bodies occur above a Poudingues d'Argens succession that contains only a few, small ribbon conglomerate bodies.

Therefore, at Argens coarse sediment was supplied from a new catchment area to the north (NW) of the outcrop area.

The fan delta dominated shorelines were actively supplied for some length of time. (For example, 20m of associated facies preserved at Le Ruch). These fan delta deposits are overlain by Scafferels member limestones above an erosional surface interpreted as the SFE surface. These limestones contain no evidence to suggest that fan deltas existed at a shoreline further up palaeoslope. In addition, Calcaires Nummulitiques Formation is ubiquitous (if very thin at some localities!) across the whole region where Tertiary succession outcrops. Therefore, the sea level rose steadily until all the topography was submerged. At this stage, the relief on the remaining emergent islands was subdued and therefore no fan deltas existed.

The presence of the extremely well bored clasts at the SFE surface implies that large areas of channel mouth bar complex were abandoned rapidly and finally, so that the bivalve colonies did not get buried by more detritus from the river mouth.

The progradation of fan delta shorelines was short lived; it began and ended rapidly. New catchment areas supplied the alluvial to marine systems and previously existing ones supplied very coarse boulder detritus to the shoreline. The fan delta facies associations have a very restricted distribution. It is probable that a phase of localised tectonic activity rejuvenated some catchment areas and created new high relief areas which shed coarse detritus into the marine basin. This phase could be related to mechanism (ii) (above) explaining the thick fine grained facies associations. The deformation induced an immediate isostatic response: thrust uplift in some areas, subsidence in others. As the system achieved isostatic equilibrium, floodplain sedimentation rates slowed in the subsided areas, and the river channel complexes were free to migrate. In this way, channel mouth bar complexes extended along large areas of shoreline and continued to be active until the topographic relief was reduced.

Each phase of compressive deformation within the basin created topography within the basin. However, the compression was related to progressive emplacement of the Alpine orogenic wedge onto the European continent, and so with each phase of deformation the foreland basin as a whole subsided further, until the region was drowned by the Nummulitiques sea.

4.6. The Nature and Controls of the Nummulitic Transgression.

4.6.1. Introduction.

The diversity of Calcaires Nummulitiques successions demonstrates that quite different shorelines existed in different areas during the diachronous Nummulitic transgression. Over much of the region, the sea level rose over a peneplained Cretaceous substratum. However, it has been shown that the region was tectonically active and that a significant topographic relief existed which was episodically rejeuvenated during the transgression. In this section, the relationship between the Poudingues d'Argens and the Calcaires Nummulitiques formation will be explored. The principal features and distinguishing features of successions from different parts of the study region will be summarised and compared. In this way, the palaeotopography can be defined, and the areas of evolving topography can be identified, for different stages before and during the Nummulitic transgression. The changing relief was controlled primarily by compressional tectonics. It is often possible to identify the individual structures which produced particular elements of the surface relief. The nature of the transgression will then be discussed, for example the changes in rates of relative sea level rise, and the relative importance of factors which controlled it will be assessed.

4.6.2. The Evidence used to Reconstruct the Palaeotopography during the Nummulitic Transgression

The palaeotopography can be derived from a number of lines of evidence, many of which are related, all of which give consistant and compatible results.

4.6.2.1 The presence of angular unconformities beneath the Poudingues d'Argens and Calcaires Nummulitiques Formations

Angular unconformities beneath the Tertiary succession are restricted in extent and are located around and above zones of intensely deformed Cretaceous limestone. (The deformation was compressive, and so the deformed areas were topographic highs). Although some erosion of these zones has occurred, the thickness difference between shortened strata and undeformed starta is not always that great. So, put another way, the topographic highs were still elevated when the Nummulitic sea transgressed over those areas. Nowhere in the study area are extensive solution cavity systems (karsts) observed in the Upper Cretaceous limestones beneath the Tertiary unconformity. They do exist, but are restricted to a few fissures filled with Nummulitic limestone where the Cretaceous is highly fractured: for example, Cairas. To the west of the study region, in Provence, huge cave systems developed in similar Mesozoic limestones and extensive deposits of bauxite are quarried from these limestones. Both features are evidence of long periods of sub-aerial exposure and extensive alteration of the substrate.

Haute Provence was exposed from the Upper Cretaceous to middle Eocene times, so why did karst surfaces and cave systems not develop? The paucity of solution weathering in the deformed zones is particularly worthy of thought!

Two conclusions may be drawn:

i) during the Palaeocene the landscape was peneplained down to the local base level throughout the area,

ii) the intensely kinked zones were only produced shortly before the Nummulitic transgression submerged the topography. This would explain why the erosion was limited, leaving remnant palaeohighs, and why these highly fractured rocks were not subject to extensive karstification.

4.6.2.2. The Palaeocurrent data from the base of and within the conglomerate bodies of the Poudingues d'Argens Formation and the Mort de l'Homme Member

The palaeocurrent data are not numerous, but they have been used to broadly identify catchment areas. The conglomerate bodies mark the position of the principal palaeochannels and therefore the palaeocurrent indications give the slope orientation for that outcrop area.

It would appear that the Mort de l'Homme conglomerates were sourced from similar catchment areas (encl. 6.2). The limited data suggest that the present structural highs of the study region may have been palaeohighs during the Eocene and supplied coarse detritus to nearby depocentres within the basin.

4.6.2.3. The distribution of the Poudingues d'Argens Formation and the facies of the Calcaires Nummulitiques Formation

The Poudingues d'Argens Formation and most of the facies of the Mort de l'Homme member are derived from high relief, continental catchment areas within the developing foreland basin. The occurrence and distribution of facies associations are used to locate palaeoslopes and palaeo-highs and lows.

In section 3.7.1, it was shown that the Poudingues d'Argens Formation was penecontemporaneous with the Nummulitic transgression. Extending the discussion, it is possible to demonstrate that the outcrop pattern of the Poudingues d'Argens Formation is a depositional feature and not one of erosional remanants. The argument is as follows: the Mort de l'Homme member is best developed above the Poudingues d'Argens Formation and thins rapidly to nothing away from those areas. The thickest development of the Mort de l'Homme conglomerates is coincident with the most conglomerate rich Poudingues d'Argens Formation exposure, at Le Ruch. It is reasonable to infer that the input of coarse detritus, and the resulting deposits, were localised. Therefore, the original distribution of Pondingues d'Argens Formation was limited; ie no Poudingues d'Argens Formation was ever deposited at for example, Annot, and eroded prior to the deposition of the Calcaires Nummulitiques Formation.

The limited distribution of the Poudingues d'Argens Formation and the Mort de l'Homme member can be used together with the palaeocurrent data to better define palaeohighs and palaeolows.

It would appear that areas like Annot were not sources of detritus for the Poudingues d'Argens. Instead, they were probably relative lows and were first areas to be submerged during the Calcaires Nummulitiques transgression. It would explain why the whole Calcaires Nummulitiques succession at Annot and Montagne de Chammatte, for example, comprises Scafferels member bioclastic and micritic limestones with no record of shoreline sedimentation. During the phase of progradation recorded in the Mort de l'Homme member (preserved higher up the palaeoslope) these areas must have remained below the SFE zone, because no change in sedimentary facies is observed in these outcrop areas.

The palaeocurrent data suggests that the Grand coyer outcrop area may have been the source of sediment for the Poudingues d'Argens Formation and Mort de l'Homme conglomerates. The Calcaires Nummulitiques facies associations at Cairas certainly suggest that the area was the source of fan delta conglomerates. The earliest deposits of Calcaires Nummulitiques are very shallow marine limestones which do not occur over the whole area. Some areas remained subaerially exposed at this stage. These limestones are overlain by debris from deposits and olistostromes which imply that high relief, unstable topography existed very close to, or within this area. The same relative highs could have been the source of conglomerate detritus to Le Ruch and Peyresq fan deltas of the Mort de l'Homme member (Fig. 4.31).

4.6.2.4 Variations in thickness of the Calcaires Nummulitiques Formation across the study region.

The thickness of the formation depends on the relationship between subsidence or uplift and the rate of sediment supply to the locality. In the case of Calcaires Nummulitiques Formation the Mort de l'Homme Member complicates the issue because the sediment rate of sediment supply varies rapidly both in time and space from one area to another: large volumes of sediment were supplied to the marine environment from continental areas of uplift within the basin. However, the thickness of the Scafferels member alone does reflect the palaeorelief that existed when the sea rose to drown the whole study region: the thickest Scafferels Member is developed on palaeolows; the thinnest on palaeohighs. See Fig. 4.30 4.31 for diagrammatic explanation.

4.6.3. Refinement of the Palaeotopographic model

There is not much that can be done in this context, but two points emerge from the data:

i) Very steep slopes were probably relatively rare. Mass flow deposits are rare; scree deposits have been interpreted at one locality, ARIO, at the base of the Poudingues d'Argens Formation. In comparison, to the west at Barreme extensive and thick scree cones have been recognised, again at the base of the Poudingues d'Argens Formation, by Evans (1987).

The debris flow deposits and olistostromes observed at, for example, Cairas, demonstrate that steep, unstable slopes were generated locally within the foreland basin during the Nummulitic transgression. Once again, Evans has observed similar "allochthonous breccias" at Barreme wthin the Calcaires Nummulitiques Formation.

ii) The occurrence of alluvial fan deposits of the Poudingues d'Argens Formation between interpreted palaeohighs (like Le Grand Coyer) and lows (for example, Annot) is difficult to interpret if a constant palaeoslope from high to low is envisaged. What prevented continental material being transported straight into the palaeolows? Rivers were probably induced to deposit their load adjacent to a break in slope between the source areas and the predicted lowest areas (Fig. 4.31a).

The Mort de l'Homme member is developed in the same area and patch reefs are observed near the limit of the Mort de l'Homme exposures. It is possible that these reefs mark the second break in slope (Fig. 4.31b).

This topographic profile could also favour the development of thick lagoonal/coastal plain successions, as the sea first transgressed across the low relief plateau (Fig.4.31c).

4.6.4. Conclusions

The variety of Calcaires Nummulitiques successions demonstrate that different types of shoreline existed at different times during the transgression. In places the sea rose across a passive, peneplained land surface in others over significant topography. A pulse of deformation occured during the transgresion, rejuvenating the source areas, and a brief progradation of fan delta dominated shorelines is recorded which interrupted the inevitable sea level rise.

It can be demonstrated that the rejuvenation of topography was induced by compressional tectonics within the developing foreland basin. The palaeohighs, lows and slopes can be directly related to the underlying structures.

At the larger scale of the basin itself, the nummulitic transgression was probably not due to eustatic sea level rise.

There is a intimate link between the development of topographic relief in the study region, in response to tectonic activity, and the timing of the transgression and subsequent development of a marine basin.

It has been shown that the Pondingues d'Argens Formation was deposited shortly prior to the rise of sea level in each outcrop area. The sources of coarse detritus for the Poudingues d'Argens depositional systems were the same as those which fed the Mort de l'Homme fan delta mouth bars.

Most probably the transgression was the consequence of subsidence of the margin of the European continent due to tectonic loading. This load consisted of:

(i) the main Alpine orogenic wedge;

(ii) thrusts ahead of this, within the foreland basin.

Activity on such thrusts generated topographic highs within the subsiding basin, which were intra-basinal sediment sources. Cycles within the Poudingues d'Argens and Calcaires Nummulitiques Formation suggest that there were distinct pulses of deformation. The overall rate of subsidence outpaced the growth of these topographic highs, so that with time the whole basin was drowned.

CHAPTER 5: THE GRES D'ANNOT FORMATION.

5.1 Introduction.

The Upper Eocene Gres d'Annot formation comprises distinctive sandstone turbidites and other mass flow deposits, which outcrop around the Argentera massif in Haute Provence and the Alpes Maritimes (fig. 1.2)

At Col de la Cayolle, the Gres d'Annot are stratigraphically overlain by the Schistes a Blocs. Elsewhere, the formation is the youngest observed: it caps the mountains or is structurally overlain by thrust sheets and it is rare to see a complete Gres d'Annot succession. The preserved stratigraphic thickness of the Gres d'Annot does not exceed 1000m anywhere.

The contact between the Marnes Bleues Formation and the Gres d'Annot Formation is an abrupt one: in places, there is a rapid transition from blue-grey marlstones, via a few metres of brown silts with thinly bedded turbidites, to thicker bedded sediment gravity flow deposits. At many localities, the Gres d'Annot lie above a significant, non-erosive unconformity (fig. 5.1).

5.1.1. Previous Research.

Sedimentological studies of the Gres d'Annot began early this century: Boussac (1912) provided the foundations upon which later studies of the Tertiary strata are based. Much of the research on the Gres d'Annot has been done in the context of regional studies by, among others, Keunen *et* al (1957), Gubler (1958), Bodelle (1971) and Campredon (1977).

Faure -Muret (1955) interpreted the sandstones as turbidites and used the term "flysch" to describe the facies associations. Keunen (keunen et al, 1957) developed the turbidite hypothesis and addressed what were seen as the major palaeogeographic problems, namely:

1. The Gres d'Annot outcrops are now bounded to the north by mountains, to the south by the Mediterranean;

2. The petrography of the sandstones does not reflect that of the present-day mountains to the north;

3. There is a general decrease in sandstone/shale ratios from south to north. He used gravity data to postulate that a mountain range had existed to the south which is now submerged beneath the Mediterranean sea. This hypothesis was extended by Stanley and Mutti (1968) With some modifications, the theory is still valid.

For some time, subsequent authors were particularly concerned with the petrology and provenance of the Alpine "flysch".

Stanley (1961) and Bouma (1962) focussed their attention on the facies of the Gres d'Annot and provided a fresh insight into the origin and nature of the sandstones and the sediment gravity flows which supplied them. They completed specific theses on the sedimentology of the Gres d'Annot, using data from different outcrops of the formation. Their methods were quite different, but they came to complimentary conclusions.

Stanley (1961, 1967, 1975) covered a large area, studying the impressive sandstones at Annot and the outcrop areas to the north of Annot. He interpreted Annot as the site of a submarine canyon, which fed detritus from a southerly source to an open fan system in the north. Bouma (1962), concentrated on the sequence of sedimentary structures in individual beds at Peira Cava and defined the classic Bouma sequence of turbidite facies.

Only recently have researchers begun to relate Gres d'Annot sedimentation to the structural setting of the study area. The I.F.P. (Conort and Odishou, 1978), working at Peira Cava, suggested that the flow of turbidity currents was controlled by a complex basin floor topography defined by en-echelon folds. They interpreted the structural setting for these folds as dominantly strike slip. In this thesis, similar topography to the west is interpreted in terms of compressional tectonics.

Bouma and Coleman (1985) referred to the Gres d'Annot outcrop areas as separate sub-basins. They presented evidence from Peira Cava for palaeotopographic slopes which opposed the flow of turbidity currents into the area.

Jean et al (1985) demonstrated that there was significant basin floor topography which was filled passively by the Gres d'Annot gravity flow deposits, but they make no attempt to determin the origin of the slopes. Their attention, in common with most sedimentologists, is confined to the internal stratigraphy of the Gres d'Annot formation and the gross palaeogeography of the Gres d'Annot basin.

5.1.2. Aims and approach of the current study.

This study of the Gres d'Annot Formation was designed to elucidate the origin of the basin floor topography and its effect on the deposition of sand from sediment gravity flows. To this end, both sedimentological and structural data were used. The results were then integrated with the original sedimentological data of others, in particular, Ghibaudo. Unfortunately, the Ghibaudo's data remain largely unpublished, with the exception of Ghibaudo, in Elliott et al (1985). In almost all cases, field evidence relating to the previous research has been critically re-examined.

The outcrop and, where possible, the extent of the angular unconformity at the base of the Gres d'Annot Formation were mapped (fig. 5.1). The origin of the palaeoslope in each case is interpreted from a structural and sedimentological study of the underlying strata.

A detailed examination was made of the sediment facies adjacent to a selection of the palaeoslopes. These were compared with data collected at various distances from the slope. In this way, the effect of palaeotopography on sandstone body distribution was assessed at a variety of scales.

The sedimentary facies distributions, interpreted together with the palaeocurrent data and potential sediment source areas, are shown to be related to the overall basin geometry and, most importantly,to its internal topography. Structural and sedimentological models have been derived for the Annot and southern Trois Eveches outcrop areas. The approach has been tested in the Col de la Cayolle outcrop area, and the data from other parts of the basin can now be considered in the light of this study.

5.2. Geological framework of the Gres d'Annot Formation.

5.2.1. The Nature of the Angular Unconformity at the base of the Gres d'Annot.

5.2.1.1. Principal features of the onlap surface.

The map (fig. 5.1) shows the occurrence of an angular unconformity and the direction of onlap at the base of the Gres d'Annot Formation. It has been noted by several authors (eg. Cremer, 1983; Ghibaudo, 1985), but neither the origin of the palaeotopography, nor its effect on the deposition of the Gres d'Annot turbidites have been discussed previously. Selected outcrops of the onlap surface were re-examined in detail, to discover the cause and effect of the palaeoslope in each case (section 5.5). Structural and sedimentological data were integrated to produce a model for the basin slopes which could be applied throughout the Gres d'Annot basin and perhaps be relevant to turbidite basins elsewhere.

The contact between the Gres d'Annot and the Marnes Bleues Formations, where unfaulted, is always parallel to bedding in the Marnes Bleues. Over most of the area, bedding in the Gres d'Annot is parallel to the formation boundary. However, in certain areas the Gres d'Annot onlaps the contact, forming angular unconformity, where the angular discordance is locally as great as 25'. By inference, where the contact is parallel to beds above and below the contact, it is a disconformity. The unconformity is illustrated by fig. 5.2, a photo of Chalufy, at the southern end of the Trois Eveches outcrop area.

The areas of angular unconformity are restricted: single lengths of onlap surface outcrop are no greater than 1km. Along the Trois Eveches outcrop area there are discrete lengths of onlap surface, separated by long sections of parallel disconformity (fig. 5.3a). The direction of onlap is consistant towards the south along the entire length of the exposure from Seyne to Chalufy, a distance of some 30km (fig. 5.3b). Borrowing terminology from the thrust tectonic literature, the basal Gres d'Annot surface can be said to have a **ramp-flat** geometry: the ramp lengths are most often short in comparison to the flat sections; the ramp angles are <30.

5.2.1.2. Recognition of an original unconformity surface.

In the field, the nature of formation boundaries requires close scrutiny: great care has been taken to differentiate between stratigraphic and tectonic contacts. Tectonic surfaces have been mis-identified as stratigraphic formation boundaries by other authors. One case is described in the structural chapter (section 2.3.2.12, part 2). Original stratigraphic contacts can only be identified by close examination of the formation boundary. These cannot be recognised from a distance, nor can they be inferred when the contact itself is not exposed.

A stratigraphic contact is recognised by the Gres d'Annot sedimentary facies in the vicinity of the palaeoslope. The diagnostic features include sedimentary structures associated with down-slope failure of sediment that was deposited temporarily on the steep unconformity slope (for example, slump folds and dewatering structures). The turbidity currents were deviated by the slopes. This was recorded in divergent palaeocurrent indicators and by bedforms which could have been generated by reflected flows or complex waves. Such phenomena are described in section 5.5. If there are no sedimentary features related to turbidite deposition adjacent to a steep topographic slope, then the contact should be viewed with suspicion: it is almost certainly tectonic!

A tectonic contact between the Gres d'Annot and Marnes Bleues Formations is recognised by:

1. A well developed tectonic fabric in both formations (Tete du Lac; section 2.3.2.12; fig. 2.31d,e).
2. Clear continuity between the formation boundary and a fault developed above or below it, which demonstrates that the contact itself is a fault surface.

5.2.1.3. Interpretation of the Unconformity.

The unconformity represents the palaeotopography in the marine basin at the time of Gres d'Annot deposition. The Gres d'Annot sediments onlapped the surface relief, which in places was very steep.

The proposed interpretation of the unconformity surface includes the assumption that bedding in the Gres d'Annot sandstones was horizontal at the time of deposition. There are two lines of evidence to support the assumption:

1. The angular discordance is as much as 25' locally. Only cohesive debris flow deposits can have such steep depositional surfaces. The Gres d'Annot were not deposited from debris flows. Depositional lobes may have been constructed by the turbidity currents, but these are low relief features.

2. The Gres d'Annot dips do not change away from the angular unconformity. It is too much coincidence to suppose that the strata were deposited with downlapping surfaces and then folded after deposition in such a way as to produce a uniform, often horizontal dip in the Gres d'Annot.

The Origin of the Palaeoslopes

The steep slopes map out adjacent to present-day structural highs, often coinciding with footwall and hanging wall ramps in the underlying compressional structures (for example, cross section D; enclosure 7). The structures themselves are therefore interpreted as responsible for the palaeotopography within the Gres d'Annot basin. The "ramp-flat" description of the unconformity surface was therefore quite valid since the unconformity records the position of thrust ramps and flats at depth.

Small and medium scale relief (30-100m relief, over 1km) is produced by sliding and slumping of the upper Marnes Bleues. For example, at Chalufy, the turbidites onlap the front of what is interpreted as a slid wedge of marlstone. At Braux (section 5.5.5. and encl. 5.7), thin turbidites infill a slide scar at the top of the Marnes Bleues.

When did the palaeoslopes develop relative to deposition of the Gres d'Annot turbidites?

The unconformity surfaces were examined for evidence of hard ground development, like phosphate nodules and borings, but none was found. However, it is rare to observe a fresh specimen of Marnes Bleues at the unconformity surface: water flows preferentially along the contact between the formations, and therefore the uppermost Marnes Bleues are commonly severely weathered.

Other lines of evidence were sought.

1. The geometry of the onlapping sediments indicates that most of the relief was created prior to deposition of the turbidites. The early turbidites are not progressively rotated during deposition: the unconformity is a simple one. Only at some localities in the Annot syncline are the very first few thin turbidites tilted together with the Marnes Bleues.

2. The Gres d'Annot display remarkably little evidence of syn-sedimentary deformation. The slump folds observed within the Gres d'Annot can be adequately explained by their position adjacent to a pre-existing steep slope. They were deposited during a period of relative tectonic quiescence in the foreland (section 5.4.2).

The evidence suggest that the basin floor topography was not significantly altered during the deposition of the Gres d'Annot. The exceptions to this include a small amount of relief above an emergent thrust tip (Denjuan, section 5.4.4.1). Some of the local topography could be generated while turbidites were deposited elsewhere in the same outcrop area. For example, the slid mass of Marnes Bleues at Chalufy was prevented from slipping into the deepest part of the basin by a local break in slope and so its age relative to the turbidites further north, down the palaeoslope cannot be defined.

It is not possible to state with more certainty when the structural relief was generated. Suffice to say that steep slopes in the Marnes Bleues were probably stable for considerable lengths of time. Two possible explanations for the stability of such high angle slopes were considered: the cohesive strength of uncemented marls, and lithification of the Marnes Bleues prior to their deformation and the creation topographic relief. The first hypothesis was rejected because the uncemented sediment would deform slowly by creep, and the surface relief would be modified with time. This did occur on a small scale at Les Gastres (section 5.5.4.). The composition of the Marnes Bleues, with 40-60% calcium carbonate, implies that early lithification of the marls was possible.

5.2.2. Age of the Gres d'Annot Formation and correlation within it.

5.2.2.1. The Use of Fauna.

The Gres d'Annot Formation has a negligible faunal content, and the rare microfaunal specimens (Nummulitids, Discocyclines and corroded pelagic foraminifera) are usually reworked from older formations. Therefore, it is not possible to construct a biostratigraphy within the formation, and the age of the Gres d'Annot remains a subject for lively debate.

Nannoplankton have been used to spot date some samples from the Col de la Cayolle outcrop area (Muller, in the thesis of Jean, 1985.). The samples contained Chiasmolithus, Coccolithus, Ericsonia and Discoaster and they were dated as Upper Eocene or Lowermost Oligocene.

The base of the succession is dated using fauna collected from the uppermost Marnes Bleues strata. These dates must carefully interpreted because the base of the Gres d'Annot is erosive in places. Despite this, the results are consistent: for the most part, the Gres d'Annot Formation is Late Eocene (Priabonian) in age, with a range from Middle Eocene - earliest Oligocene.

The data demonstrate that **the arrival of the first siliciclastic deposits is diachronous** across the study region, with the oldest in the south and east and the youngest in the west. Fig. 4.3 (Campredon 1977) summarises the available data from an east-west transect south of Argentera. The results are compatible with the regional palaeocurrent data: the sediment gravity flows travelled from a southerly source area towards the north and west, progressively infilling the basin.

This diachronous trend continues that observed in the older Helminthoid flysch and other Alpine turbidite sequences, which are preserved in thrust sheets that overlie the Gres d'Annot and restore to sites of deposition east of the Argentera massif.

Cremer (1983) quotes an age of Early Oligocene for the uppermost horizons of Gres d'Annot, but does not list the species described. There are regional considerations which support this date for the cessation of sedimentation in the basin: Sannoisian (Lowermost Oligocene) sediments to the west (Barreme and Digne) contain elements of the Embrun-Ubaye thrust sheets. Therefore, these structural units were emplaced above the Gres d'Annot in the Early Oligocene, which would imply that the sedimentary basin ceased to exist at this time.

5.2.2.2. Correlation within the Gres d'Annot succession.

To date, a technique for correlating sections of Gres d'Annot using the sediments themselves has not been perfected. Ghibaudo (in Elliott et al, 1985) and Jean (1985) have mapped "key beds" in the sequence and have identified two such horizons in the Col de la Cayolle and three in the Trois Eveches outcrop areas.

These beds have distinct compositions (section 5.7.1.3) and facies which allow them to be easily recognised in a Gres d'Annot section. They comprise matrix supported, cobble conglomerates, with little to no internal organisation. It should be noted that individual beds display rapid variations of thickness, thinning to zero in places; they are therefore difficult to trace between outcrops.

The conglomerates are interpreted by Ghibaudo and Jean as debris flows of great lateral extent. Their particular clast composition implies that the sediment was sourced from terrains that only supplied detritus into the deep basin on rare occasions. The debris flows may have been initiated by seismic shock (Mutti et al, 1984). As such, the deposits may be considered as time lines within the Gres d'Annot succession.

Ghibaudo and Jean have measured a number of complete sections through the Gres d'Annot in both the Col de la Cayolle and Trois Eveches outcrop areas and have spent many months in the field tracing units as well as individual beds. They are convinced that these horizons can be correlated across the individual outcrop areas. From their findings, one can deduce that these conglomerates were deposited across areas larger than 30 by 40km. They are therefore valuable marker beds within a single outcrop area. Unfortunately, they cannot be correlated with confidence between outcrop areas.

5.2.3. Dimensions of the Gres d'Annot basin

5.2.3.1. Depth of the basin.

The Gres d'Annot comprises the deposits of sediment gravity flows. True pelagic sediments are not observed, and there is no evidence that persistent traction currents operated in this part of the sedimentary basin. The lack of sedimentary structures indicative of reworking by waves demonstrates that the sandstones were deposited below storm wave base. The sedimentary facies do not imply a maximum possible depth for the basin.

The trace fossil and faunal assemblages show that the basin was open marine. The variety of trace fossils implies that food was not in short supply and well oxygenated. Therefore, the basin was not very deep.

Mougin (1978) collected a mesobathyal faunal assemblage from the strata of Marnes Bleues immediately below the Gres d'Annot Formation, at Annot. She assigned a depth of 900m to the basin at this stage, but gave no depth range for the assemblage. The figure of 900m seems somewhat arbitary, but a depth of a few hundred metres is compatable with the regional ~~and~~ considerations, for example the inferred slope of the basin margins.

5.2.3.2. Lateral equivalents of the Gres d'Annot Formation which indicate the probable extent of the turbidite basin.

The Gres d'Annot Formation includes sand-rich formations from nine major outcrop areas. Equivalent age sandstone successions, with similar sediment facies, occur to the north: the Gres de Champsaur, Aiguilles d'Arves and Gres du Taveyannaz. The distinction between them is principally geographical.

The foreland basin at the time was therefore extensive: greater than 150km long and 80km wide. Its margins cannot be precisely defined.

The Shorelines

The western limit of the Gres d'Annot basin was to the west of Barreme. There, the Gres de Ville Formation is equivalent in age to the youngest Gres d'Annot successions. The complex ripple crosslamination preserved in the thin sandstones of this formation imply depths of deposition shallower than storm wave base.

The southerly shoreline was near St. Antonin, where the sandstone facies are the proximal, shallow marine - fan-delta, estuarine equivalents of the Gres d'Annot Formation (Apps, in prep.).

The eastern boundary of the Gres d'Annot basin cannot be identified. Over much of the region, it is structurally buried beneath thrust sheets of Sub-Brianconnais and Brianconnais Mesozoic-Tertiary sediments. In the south, from Annot to Ventimiglia, the character of the Gres d'Annot changes from west to east, but nowhere are the facies indicative of a shallow marine environment. It is most probable that the eastern boundary was delineated by the advancing Alpine orogenic wedge.

The thrust sheets which overlie the part of the study area contain Late Cretaceous-Tertiary siliciclastic deep marine successions which are superficially similar to the Gres d'Annot. Examples of these are seen at Lac d'Allos, in the Col de La Cayolle outcrop area.

5.3. The facies associations observed in the Gres d'Annot Formation.

5.3.1. Introduction.

The Gres d'Annot Formation is dominated by sandstone which forms more than 70% of the succession, up to 95% in the southern outcrop areas of Annot, Contes and Menton. The remainder of the formation comprises interbedded sandstones and siltstones with rare shales. Conglomerate bodies are uncommon and they appear at random in the succession.

The coarse sandstones are poorly sorted; sorting improves in finer grained sediment. If granules are present in the sample, the grain size distribution tends to be bimodal about the granule-sandstone grain size boundary. Matrix is rare to absent in coarse sandstones, more important in the siltstones.

The sandstones have been interpreted as the deposits of sediment gravity flows, for the most part turbidites (eg. Fauret-Muret, 1955; Keunen et al, 1957; Bouma, 1961; Ghibaudo, 1985). The sandstone fraction of the Gres d'Annot Formation is divided into three facies associations, GdAI, II and III. Following the facies description below, these are placed in the context of the local and regional palaeotopography within the Gres d'Annot basin (sections 5.4. and 5.5.).

GdAI are thinly bedded turbidites interbedded with siltstones and shales. GdAII includes all sequences of thicker bedded (30cm-2m scale), but rarely amalgamated, sandstones separated by thin siltstones. Spectacular sandstone bodies, GdAIII, exist in all the outcrop areas, but are prevalant in the south of the region. The bodies are from a few metres up to 100m thick, and may be a single thick bed or several amalgamated sandstones.

The ratio of facies associations does vary across the study region, with higher proportions of the well bedded associations, GdAI and II in the northern outcrop areas (80%) and large, massive sandstone bodies in the south. The variations are not gradual and may reflect compartmentalisation of the basin as much as an overall proximal - distal relationship between the outcrops (section 5.6). Across any area, the amount of sandstone does not change markedly.

The facies associations are laterally continuous over distances of 100s of metres, up to several kilometres. Vertical sections through the turbidite stratigraphy show that the boundaries between the facies associations are most often gradational, reflecting simply an increase (or decrease) in bed thickness and degree of sandstone amalgamation. Single thick sandstone beds often fill shallow, broad channels. Other than this, channels are rare or absent: only one of any size (100m deep, 1km wide) has been recognised in the field, at Denjuan, in the Southern Trois Eveches outcrop area (section 5.3.5.1. and fig. 5.21).

5.3.2. Thinly bedded sandstone and siltstone facies association: GdAI.

This facies association forms approximately 40% of the formation in the northern outcrop areas and much less in the south. Centimetre-decimetre scale sandstones alternate with siltstones and, more rarely, shales.

The sandstones are laterally continuous over the traceable lengths of outcrop (<100m in this facies) and do not change thickness, except where adjacent to the base Gres d'Annot angular unconformity. The sandstones comprise individual beds; they are never amalgamated. Fig. 5.4 and 5.5 show the appearance of this facies on a large scale and in outcrop.

All grades of sandstone are observed, but medium and fine grained sand are dominant. They are, for the most part, "base-absent" turbidites (tb/c-td/e), although the coarser sandstone have thin ta divisions. The upper contact of sandstone with siltstone or shale is usually marked by a sharp grain size break, sometimes with a preserved ripple form.

This facies association is interbedded with the other facies associations and is not seen in any outcrop as the lateral equivalent of the thicker bedded facies. Units of thinly bedded sandstones and siltstones are only once, in the experience of this author, truncated by a thick, channelised sandstone body. Throughout the successions, the thick sandstone bodies are separated by GdAI, sometimes tens of metres thick (fig. 5.4a). In the Col de la Cayolle outcrop area, one of these units (20 metres thick) can be correlated across the whole area and serves as a marker horizon (Ghibaudo, 1985 and Jean, 1985).

The association occurs throughout the succession, but it is often observed at the base of the Gres d'Annot Formation. It may only comprise the first metre or so of section, but it is very rare to observe coarse, thick sandstone bodies in direct contact with the Marnes Bleues strata.

Interpretation.

The sandstone facies may be interpreted as those deposited from low density turbidity currents, or the dilute tail or lateral extension of a larger turbidity current.

The sharp tops and the preservation of ripple forms implies that the finer grain size sediment bypassed that deposition site or was subsequently eroded.

There is no evidence for the age or origin of the micro-faults. Their close spacing suggests that they formed in uncemented sandstone: they may have formed in response to slight failure of the sediment down slope (section 5.5.1). They are particularly well displayed in these sandstones, because the laminae are well defined by mica and organic material. Discrete fractures may have formed in the coarser sandstones, but their presence is not easily detected.

The facies association is only once observed to be laterally equivalent to a sharply defined, medium scale, channel fill. This and the lateral continuity of individual beds and units demonstrate that the facies were not deposited on channel levees or between major distributary channels. Instead, they almost certainly define periods of time when the basin was supplied with least sediment. The rarity of shale or marlstone horizons implies that the basin was richly supplied with siliciclastic sediment during the deposition of the GdA formation.

5.3.3. Thick (decimetres - 3 metres), well-bedded sandstones: GdAII.

5.3.3.1. Principal characteristics of the facies association.

The characteristic features of the facies association are the thickness and close spacing of sandstone beds and the preservation of sedimentary structures which record the properties of the sediment gravity flows, for example, flute casts, grading, parallel- and cross-lamination and crossbedding. The well bedded sandstones are commonly separated by siltstone, but are sometimes amalgamated.

Individual beds are laterally extensive and of uniform thickness across most observable outcrop (100s of metres). Exceptions to this are observed adjacent to basin floor slopes (section 5.5.).

The sandstones are mostly characterised by a thick ta division of very coarse - medium sandstone and tb, with little or no tc division. The facies sequences are ta-d, ta-c-d, ta-b-c-d, more rarely tb-c-d. Shale clasts are often present, commonly concentrated at the top of the massive sandstone. Plant fragments and mica grains are concentrated in the finer grained sediment at the top of the bed and they define the laminae.

The facies are illustrated in the photos of fig. 5.8.

This facies association forms thick sediment bodies (10s of metres thick) which are tabular at the scale of outcrop. Ghibaudo (1985, pers.comm.) and Jean (1985) have logged many sections through the outcrop areas of Col de la Cayolle, Trois Eveches and Grand Coyer and demonstrate that these bodies and individual beds are lenticular over distances of kilometres.

However, the precise geometries cannot be defined because correlation between outcrops is still speculative. Abrupt lateral terminations are rare, seen only at the basin margins and the edges of the few channels observed. For the most part, it can be assumed that the facies laterally grades into thicker or thinner bedded facies associations: this is observed at some localities.

Interpretation:

The coarse grain size, the lack of clay matrix, the dominance of massive sandstone, the presence of normal grading and absence of inverse grading all imply that most of the flows were moderate to high density sandy turbidity currents.

Several important observations of the sediment facies are described and interpreted below, whose significance contributes to the overall interpretation of this facies association.

5.3.3.2. Sedimentary structures in the ta division.

The lower part of the bed may not be devoid of structure. Often a crude layering, 1-5 cm scale, exists: changes in grain size delineate the layers, sometimes defining small graded units (sometimes,but not often inversely graded), which thin upwards through the bed. See fig. 5.9.

Interpretation.

Lowe (1982) observed a similar centimetre scale layering, but inversely graded sediment defined individual layers. He concluded that prior to deposition from a steady high concentration turbidity current, the concentration of grains at the bed increased such that sediment transport was dominated by grain to grain interactions; ie. the basal layer of particles was supported by dispersive pressure, forming a traction carpet. Sediment fall-out into the layer was continuous, and at a critical point the traction carpet froze, depositing an inversely graded sediment layer. The process was then repeated.

In the Gres d'Annot, some layering may have a similar origin, but the absence of inverse grading and presence of normal grading in most cases suggests that turbulence was not suppressed by grain to grain interactions. Instead, the repeated pulses of rapid deposition may correspond to velocity lows in an unsteady flow. It is probable that turbidity current velocities are somewhat irregular, particularly near the head of the flow. The lower portion of a turbidite, the ta division, is deposited from this part of the current. Where the sediment concentrations are high, only small fluctuations in velocity may be necessary to cause pulsed sedimentation. The volume of sediment deposited in each pulse may be expected to decrease with distance from the head of the flow, giving rise to the decrease in layer thickness vertically in one bed.

5.3.3.3. Cross-Bedding.

Single cross-sets, 10-25cm set height, are observed in some sandstones of the gres d'Annot. Only in one case are cosets developed (encl. 5.4, 5.5; Le Ruch).

They occur at the top of the bed, above the ta division. In the example from Le Ruch, the cosets comprise medium-coarse sandstone which is markedly coarser than the sediment immediately below the crossbedding.

At Peira Cava, there is an excellent example, where the morphology of the final sandy sediment surface is preserved beneath a sediment drape of shale and thin sandstones (fig. 5.10). In this case, the crossbedded (20cm set height) sandstone is well sorted and medium grained; it lies above a sharp discontinuity on 30cm of massive, poorly-moderately sorted coarse sandstone. The well defined laminae have maximum dips of 30 and are asymptotic towards the base of their base. They dip in directions at variance with the erosive lineations on the base of the beds: in the case described towards the east (basal grooves oriented north-south).

Interpretation.

These crossbeds represent bedforms (for an introduction to the occurrence of cross bedding in turbidites, see Allen, 1970). In the Peira Cava example, the steep dip of the laminae and the preserved morphology of the bedform displaying both the stoss and lee surfaces imply that the laminae were the foresets of megaripples.

This interpretation does not accord with that of Bouma and Coleman (1985), who interpret the laminae as lateral accretion surfaces forming point bars within small scale channels. In this way, these authors account for the orientation of the laminae at right angles to the basal palaeocurrent indicators. However, a train of at least two megaripples is preserved at the very outcrop they described! The crossbedding is not anywhere near an erosive contact which could be interpreted as a channel margin: the bedform laminae were accreted on migrating assymmetric bedforms and did not accrete from a fixed surface.

There are two problems with the bedform interpretation of the cross stratified sandstones: i) the origin of traction currents large and stable enough to form decimetre scale bedforms and move them distances of several metres; ii) the divergent palaeoflows!

In the Peira Cava example, the bedforms developed above a sharp grain size discontinuity and comprised very well sorted sandstone. This suggests that the current responsible for the bedforms thoroughly reworked the top of the sediment gravity flow. Here, the traction current need not have been related to the earlier flow: other basin processes, like tides or storm waves, may produce strong currents. However, cross stratification is rare and does not have any of the features characteristic of storm wave or tidal current bed forms. In other localities, the crossbedding appears in continuity with the sandstone below. It is most probable that the sediment gravity flows were responsible for all the bedforms observed.

The hypothesis suggested by this author incorporates the divergence of palaeocurrent readings between the base and top of a single bed. At Le Ruch, it can be demonstrated that a significant palaeoslope opposed the turbidity currents arriving from the south. In section 5.5.3., the effect of this slope is discussed in detail (Encl. 5.4, 5.5). Secondary currents developed when the original turbidity current encountered the slope. These flowed at a high angle to the original turbidity current. The resultant current reworked part of the earlier deposits. The sandstones were finally deposited on a variety of often complex bedforms. A similar model can be applied at Peira Cava.

5.3.3.4. Amalgamated sandstone beds.

Individual beds are not commonly amalgamated, but where they coalesce, the boundaries between them are identifiable. Chaotic layers of mixed sand and fragments of shale define the contact ~~(fig. 5.)~~, which are often laterally equivalent to a simple horizon of irregularly shaped and unsorted shale clasts.

The shale fragments within the chaotic layers are sometimes preserved partly attached to the originally continuous shale or siltstone that divided the two beds. Fig. 5.11c shows that the edge of the incipient clast is flicked upwards and that sands from the beds above and below are thoroughly mixed around that edge. Mud clasts in the vicinity are irregularly shaped, some resembling detached flame structures.

Interpretation.

Rip-up clasts may not always be generated when a turbidity current erodes its substrate. In some cases, the vertical pressure exerted by the over-riding current on the beds below may cause rapid compaction, which causes the sediment below the surface to dewater. As the fluid escapes to the surface, overlying coherent shale or siltstone layers are disrupted. The irregular surface, and the sudden ejection of sand and water into the flow at the same point, may give rise complex eddies in the flow above. In this way, the shale is further disrupted, sediment and water are injected below the broken shale/siltstone horizon and coherent slivers may be incorporated into the turbidity current.

5.3.3.5. Shale Clasts.

In the Gres d'Annot Formation sandstones shale clasts are common and occur at different levels within the sandstone and accordingly have different morphology and distribution. In the sandstones of this facies association, shale clasts tend to concentrate at one discrete horizon high in the bed, often at the top of the massive portion of the bed and are well sorted and rounded. The 'a' axes of the clasts are aligned parallel to the palaeoflow derived from grooves and flute casts.

Where sandstones are amalgamated, a layer of rip up clasts may define the boundary between the two original beds. However, where the amalgamation is more thorough, all that may remain as a clue to the process are clusters or "nests" of shale clasts (fig. 5.10b).

Interpretation.

Fragments of mud are less dense than sand grains of equivalent size and have a large surface area. Therefore, in a turbulent suspension such clasts are better supported. Beyond the region where rip-up clasts are incorporated into the flow, the fragments of mud will migrate upwards in the flow to their equilibrium position. Deposits frozen before stratification has occurred are randomly distributed in the lower part of the bed.

Clusters of shale clasts akin to those described above were noted by Bouma and Coleman (1985) in the Peira Cava outcrop area. These nests were also laterally equivalent to clearly amalgamated sandstones. However, these authors went on to suggest that the amalgamated contacts occurred on the crests of channel levees and they interpreted the shale clast nests as levee and overbank deposits.

In all the examples seen, the opposite seems to be true. The facies association GdAII is not associated with channels. Instead, bed amalgamation occurs adjacent to steep basin floor slopes (eg. Chalufy, section 5.5.1; encl. 5.1). In these cases, the clusters of shale clasts occur near the slope where the sandstones were mixed quite thoroughly along much of the original contact.

5.3.3.6. Organic Debris.

Locally, the concentration of plant debris in the upper parts of the sandstones is very high (fig. 5.12). Individual fragments may be 10s of centimetres long, with well preserved biological structures. The plant fragments have a bright surface sheen; they are brittle and very fragile.

Interpretation.

The plant remains were transported to their final site of deposition rapidly, without being repeatedly reworked in well oxygenated waters.

The carbon is highly reflective. I recommend that samples be analysed for organic maturation indicators, particularly vitrinite reflectance. The rocks were not affected by deformation, igneous activity or mineralisation. Therefore, the maximum temperatures recorded in these plant remains almost certainly reflect the burial history of the formation, by sediment and thrust sheets, and heat flow from basement rocks. The methods and the significance of results have been well studied and applied in the oil industry. For this reason, it would be sensible to study the organic maturation indicators in preference to illite crystallinity (current research project, lead by Fry; 1985, pers. comm.) in this study region.

5.3.4. Large sandstone bodies (few metres-50m thick): GdAIII.

These sediment bodies include single sandstone beds greater than approximately three metres thick and composite units. The latter may comprise amalgamated thick beds or many individually thin sandstones. The facies are characterised by a total absence of siltstone and shale. The sedimentary structures most often record post-depositional processes like dewatering and, more rarely, the original sediment transport processes.

The appearance of GdAIII on a large scale is shown in fig. 5.13a,b.

GdAIII bodies occur in all the outcrop areas, forming <30% of the successions in Trois Eveches and Col de la Cayolle, but much more in the areas of Annot, Contes, Menton and Peira Cava. In these areas, it can constitute 95% of the sequence, for example, the Chambre du Roi at Annot. In all cases, the bodies extend laterally, unchanged, for several hundred metres. They are discontinuous over distances of several kilometres.

Only in one place (Denjuan, section 5.3.5.1) does GdAIII represent the fill of a large scale channel. As the sandstone beds thin gradually, siltstones and shales appear in the sequence.

5.3.4.1. Single thick sandstone beds.

Good examples may be studied in the Trois Eveches and Peira Cava outcrop areas. The characteristics of these beds are described below, but not all of the features are necessarily present in one bed.

Description.

Each bed may be 3-20m thick, laterally extensive on a 100m scale, but of variable thickness. The base of the bed is flat or irregular, on a decimetre scale: grooves are not common. The sediments beneath are often undisturbed, but may be disrupted by flame structures (fig. 5.14a), small scale sandstone dykes and decimetre folds. The thick GdAIII bed comprises a lower chaotic layer and an upper thick, massive sandstone.

A coarsening upward layer is preserved in the basal few centimetres of the bed (fig. 5.14b). More rarely, a pebble conglomerate lies directly above the erosive base. Above, there is a chaotic horizon of highly variable thickness (<1-5m) which contains clasts of contorted shale, siltstone, sandstone and angular clasts of thinly bedded GdAI. The clasts are unsorted, ranging in size from pebbles to rare boulders greater than 8m across. A crude clast imbrication may be present.

The remainder of the large bed is sandstone: the contact between the chaotic layer and that sandstone is laterally variable, sharp or gradational. Where erosive, it is concave upwards, grooved and lined with granules. Lineations on a single face are oriented at high angles to one another. Large flames of shale are injected into the sandstone, from disordered conglomerate beneath.

The lower sandstone is poorly - moderately sorted with lenses and diffuse stringers of granules or pebbly sandstone (fig. 5.14c). It coarse tail grades into coarse sandstone, which forms most of the bed (fig. 5.14d). The sandstone is mostly structureless, but the granule stringers may define low angle cross stratification in the lower part of the bed, and the medium sandstone at the top of the bed can be crossbedded (fig. 5.14e).

Planar or curvilinear surfaces characterise the central part of the bed. The concave upward surfaces, linked by sharp anticlines, are spaced at <20cm intervals and become more pronounced and concave towards the top of the bed (fig. 5.15a and b). These are also seen in the amalgamatd units: their interpretation is discussed in section 5.3.4.2. It is rarely possible to discern even subtle grain size variations across these planes. Dish structures and vertical pipes may be present.

The upper surface of the bed is sharp, but often undulose or irregular.

Interpretation.

At any one outcrop, some of these beds appear similar to megabeds , described from the Hecho Group, Southern Pyrenees, the Apennines and Cantabrian mountains (eg. Mutti et al, 1984 and Rupke, 1978). However, there are important differences: they are not laterally extensive and they have the same source as the beds above and below. Therefore, they cannot be described as "mega" and require no exotic mechanism to generate them.

Fig. 5.18 summarises the principal features of single large beds and illustrates the differing interpretations of Stanley (1982) and Cremer (1983). Stanley interpreted examples at Contes and Peira Cava as welded slumps and turbidites. He suggested that tubidity currents were generated from submarine slides which evolved first into slumps and then into turbidity currents. He did not explain why the turbidite was deposited at the same locality, and on top of the slump from which it was supposed to have evolved. The obvious problem is that turbidity currents travel faster than slumps. Following Stanley's logic, it is possible that the turbidity current somewhat post-dated the slump and was in fact generted by back-failure of the slump scar.

Cremer (1983) interpreted them as deposits of high density turbidity currents, using as a basis the model of Lowe (1982). However, Cremer does not explain the significance of the chaotic facies, rich in shale clasts, observed in the Gres d'Annot examples studied. Also, the presence of several thin coarsening upwards layers is of great significance in Lowe's model (interpreted as the deposits of traction carpets): they are not observed in the Peira Cava type deposits. It should be noted, however, that Cremer's model encompasses the complete range of thick, massive sandstone bodies of the Gres d'Annot Formation: this may be his mistake!

The unified appearance of a single unit and the continuity of the matrix to the chaotic facies with the overlying sandstone justifies their interpretation as single event deposits. However, it is clear from the individual facies that the flows were complex and deposited sediment from more than one hydrodynamic regime.

The disordered and ill-sorted conglomerate at the base of the bed has the characteristics of a debris flow deposit. The enormous clasts could be transported in the plug of a debris flow. Elsewhere, the scoured base and imbricated clasts show that the base of the flow was more turbulent, and the conglomerate facies were probably deposited from a poorly organised bed load of a high density turbidity current.

The locally sharp and scoured contact between the conglomerate and the overlying sandstone facies, and the normal grading in that sandstone, imply that the upper part of the bed was deposited from a distinct, fully turbulent current. It flowed above the traction current or debris flow: the eddies were capable of eroding into clasts in the plug below. While the plug and the clasts within it could move, the grooves were randomly reorientated. Often, however, a large percentage of the lineations ~~record~~ align with the regional palaeocurrent direction. Therefore, in this case, the debris flow or traction carpet was deposited while the upper part of the flow remained highly energetic.

Most of the remaining sandstone was deposited from a high density turbidity current. The rare cross stratification high in the bed shows that impersistent traction currents existed. The remaining structures are all interpreted as water escape routes. The undulose and ridged surface of some such beds was also the ~~result of to be the~~ result of sediment dewatering.

To summarise, the evidence suggests that each bed was deposited from single dense, stratified sediment gravity flow. Within that flow were distinct rheology, density and current velocity boundaries. There is evidence that matrix shear strength supported large clasts at the base in parts of the flow. The majority of the sandstone was deposited from a high density turbidity current.

Why did the turbulent part of the flow not overshoot the debris flow? The results of this study suggest that the Gres d'Annot basin floor topography was complex and may have restricted the ability of flows to travel long distances. In extreme cases of confinment as may have existed at Peira Cava (Bouma and Coleman, 1985), the currents were forced to deposit their load in a small area and could not evolve.

5.3.4.2. <u>Amalgamated Units.</u>

These often enormous (up to 50m thick) sandstone bodies are laterally continuous over kilometres. They exhibit a range of component bed thicknesses and number of visible amalgamation planes (fig. 5.13b). There are no discernable trends or cycles of bed thickness or grain size within these amalgamated units. They are the most enigmatic of the Gres d'Annot sediments, in part because the individual flow deposits cannot always be defined. In addition, sedimentary structures are uncommon and provide little evidence for the sediment transport mechanisms.

From a distance, the bodies can be seen to have sharp, planar bases and bedding planes can be traced through the body. In places, these can be defined as amalgamation by a sharp grain size discontinuity (fig. 5.17). Above the sharp bed base, laminae of granule and coarse sandstone alternate over 10-20cm, then rapidly coarse tail grade to coarse sandstone. Most of the sandstone is moderately - well sorted. Lamination is rare, but dish structures are common. Otherwise, the sandstones are massive or riddled with broad, concave upward surfaces as described in the previous section.

At Col de la Cayolle, sharply defined surfaces may represent sediment bedforms (fig. 5.18). Cremer (1983) describes large scale, low angle cross stratification in the Trois Eveches outcrop area.

The top of the sediment bodies are rapidly transitional into GdAII. At Chambre du Roi, the drop in sand percentage is dramatic: the GdAIII is in sharp contact with GdAI.

Interpretation.

Fig. 5.19 shows the range of sandstone bodies included in this category.

The absence of sedimentary structures may be attributed, in part, to fluidisation by sediment dewatering soon after deposition. In some cases, it may also be a primary feature, where sediment was deposited rapidly from a dense, poorly stratified suspension of coarse sand.

Some of the beds show features characteristic of deposition from high density turbidity currents (Lowe, 1982): bases that are erosive, but which do not scour deeply into the substrate; a basal division with small granule filled scours, pebble and granule stringers and low angle, impersistent cross stratification (inversely graded layers are rarely seen); a central part with dish structures that is otherwise massive. It is probable that most of the sandstones were deposited from suspension in moderately dense turbidity currents. Where low angle cross-stratification is present, the sediment in the lower part of the bed spent its final moments, at least, as bed load.

In massive, poorly-moderately sorted sandstones like these, planes within the sediment cannot always be shown to be more than weathering features. The absence of grain size variations across them makes it difficult, if not impossible to ascertain if some are the bases of shallow, ephemeral channels (Cremer, 1983). The closely spaced, broad, concave upward surfaces (fig. 5.15a,b) are interpreted as principal sediment dewatering planes enhanced by differential cementation and subsequent weathering. The anticlines mark the vertical pathways of the water which are truncated where the fluid encountered another discrete, sub-horizontal surface along which water could escape more swiftly. These planes may have been primary sedimentary structures: it is not possible to test this hypothesis.

It is not always possible to discern the deposits of a single flow in these sediment bodies. The sediment may be totally massive and crudely graded. Even facies repetitions or the existence of a chaotic horizon do not unambiguously imply the deposits were amalgamated. Lowe (1982) pointed out that a sediment gravity flow may be unsteady and deposit very complex facies sequences during a single event. This may also be the case if a flow lasted for a few hours. If a sediment gravity flow was initiated by a sediment slide, progressive failure of the slide scar gives rise to a rapid succession of flows which could generate a thick, massive or complex facies sequence.

However, there are many overall graded, granule to medium sandstone, beds, up to 20m thick, that may be interpreted as the deposits of single large flows. These are termed "granule bars" by Cremer (1983). The only evidence to support the term "bar" is the large scale cross-stratification observed by Ravenne. The surfaces appear to be limited to the top of the bed, and the whole bed geometry cannot always be ascertained.

In other cases, undisturbed bedding beneath the body suggests that the first sediments at least were not deposited from large, energetic flows. It is therefore probable that some of the bodies were constructed by frequent, small-medium scale, moderate (and high) density flows, similar to those which deposited sandstones of facies association GdAII.

5.3.5. Conglomerate Facies

There are two types of conglomerate, one which forms discrete laterally extensive marker horizons in the succession, the other which is associated with granule or sandstone bodies.

The second facies is a minor proportion of the total Gres d'Annot. Pebble - cobble conglomerates occur at the base of massive sandstones and often at the boundary between two amalgamated sandstone beds, where they comprise intraformational clasts of shale and siltstones. Their existance can be explained with reference to the sedimentary processes responsible for the sandstone facies. Therefore, they are described and interpreted in the sections describing GdAII and III, as appropriate.

Discrete conlomerate marker beds.

These conglomerates are a minor, but very important and distinctive element of the Gres d'Annot Formation.

Description.

They are moderately - poorly sorted cobble conglomerates, with 17-21% of clasts larger than 1cm, 60% sand and 20% matrix. No grading is observed, although the base of the conglomerates are often obscured beneath rubble and so grading may exist in the lowest few centimetres of the beds. The whole rock is matrix supported in a pelitic matrix. The monomineralic grains are angular - subrounded; the granules and clasts are well rounded and quite spherical for the most part.

The beds are flat based and not erosive: they passively overlie sandstones or shales. The tops of the beds are irregular, often truncated beneath the overlying sandstone turbidite. Elsewhere, the upper surface relief is preserved beneath thinly bedded sandstones and siltstones, and it is clear that the depositional surface was highly irregular. There is no internal organisation, no clast imbrication or cross stratification.

The beds are less than 12m thick and are laterally extensive over distances as great as 20km. The beds thin laterally and may disappear altogether. For example, the upper of two conglomerate marker beds at Col de la Cayolle has a patchy distribution in the northern part of the outcrop area.

Their composition is distinctive, quite different from that of the Gres d'Annot sandstones. This is discussed in section 5.7.1.3. and was taken into account when deciding how the flows responsible for the conglomerates made their way into the basin.

Interpretation.

The poor sorting and the total absence of parallel or cross stratification demonstrates that the sediment was not deposited from traction currents. The apparently non-erosive bases to these beds, the absence of grading, together with the high percentage of matrix suggest that the conglomerates were deposited from coherent debris flows, in which the coarse sediment load was supported, in part at least, by the matrix. The plugs and levées of debris flows ~~and levees~~ produced beds with uneven top surfaces.

Jean (1985a) proposed that the clasts were derived from steep gradient rivers which fed directly into a submarine canyon. There are no direct palaeocurrent indicators, but the provenance of the sediment~~ology~~ any clasts suggests a southeasterly source area. Some of the detritus was apparently reworked from the thrust sheets of the Alpine orogen. These source areas did not shed sediment into this basin at any other time.

It is probable that the debris flows were triggered by seismic shock, related to activity on an Alpine thrust adjacent to the basin. Large scale failure of steep subaerial and submarine slopes generated large volumes of sediment. This mechanism would account for the great variety of clast types and the high energy flow required to distribute sediment over large areas of the Gres d'Annot basin. The conglomerate beds are akin to the megabeds of the Southern Pyrenean Hecho Group described by Seguret et al (1984) and Mutti et al (1984). These too were sourced from areas which otherwise did not supply the turbidite systems. Seguret argues convincingly for a seismic triggering mechanism, produced by displacement on a thrust fault.

5.3.6. Channel-fill sediments in the Gres d'Annot Formation.

Channel-fill sediments are conspicuous by their absence, both in the Gres d'Annot Formation and the Gres de Champsaur (Livera, pers.comm., 1986). Four important examples are described in this section.

Many of the GdAIII sediment bodies are interpreted by Cremer (1983) and Ghibaudo (1985) to be the fill of shallow, broad ephemeral channels, but individual channels were only metres deep and 100s of metres wide. An example (15m thick, 1km width) is easily identified on the SE face of the Montagne de l'Avalanche (fig. 5.20). A single flow formed a braid plain of channels: they were laterally unstable and therefore deposited a sheet sandstone body of great lateral extent. It is not easy to test the hypothesis in many of these sediment bodies. For example, storey boundaries are difficult to identify: the coarse, uniform grain size distribution and dewatering structures mask the original sedimentary contacts within these bodies.

The same facies associations have been interpreted as large scale channel fills by Stanley (1975). There is no evidence to support his arguments: with one exception (Denjuan; see below), the base of the sandstone bodies, although sharp, does not have any significant erosive relief.

5.3.5.1. <u>The Chalufy Channel</u> (1km+ width, 130m apparent depth).

This is the only channel of this large scale observed in the Gres d'Annot Formation. It occurs within 250m of the top of the formation as it is exposed in the Trois Eveches outcrop area (fig. 5. 21).

Observations.

The exposed cross section of the sediment body is shown in fig. 5.25, together with the principal features of the channel. The margins are symmetric and deeply incised into GdAI and II.

A pebble conglomerate of variable preserved thickness, maximum 30m, lines the base of the body. The conglomerate is poorly sorted, with rounded pebbles and cobbles of principally acid igneous rocks; and angular, elongate boulders of thinly bedded sansdstone and siltstone. The latter are concentrated at the high relief contact between the conglomerate and a massive, pebbly sandstone. They are oriented parallel to the contact and define its steep sections. Grooves on the base of the sandstone are in places aligned to bedding within the clasts, *eroding more deeply into the* beds. They are parallel to grooves elsewhere on the base. These are oriented east-west, parallel to palaeocurrent readings outside the body.

The pebbly sandstone grades into coarse sandstone. The remaining fill comprises **horizontally** bedded sandstones: thickly bedded and massive at the base, thinning upwards to moderately thick sandstones which continue to the top of the exposure.

Interpretation.

The sediment body represents a deep channel incised into thinly bedded sandstones and filled over some length of time. The pebble conglomerate was deposited from a debris flow which incorporated material ripped from the margins of the channel and formed a basal lag. The next event was energetic: it eroded down into the debris flow conglomerate and encountered resistant clasts of sediment. Eddies were generated around them that scoured deeply adjacent to the clasts. The remaining channel-fill sediments are no different from those seen outside this channel.

This channel now appears isolated, and such large channels are very rare in the preserved section of Gres d'Annot. However, although it is considered that only a thin sequence of Gres d'Annot ever existed above this level, it is possible that the eroded section contained many more such channels. Where a stratigraphic contact of Gres d'Annot with the overlying Schistes a Blocs is preserved, it is clear that the Gres d'Annot were tilted and, in places, deeply eroded prior to the emplacement of the Schistes a Blocs olistostromes (section 6.6.7). Pulses of deformation and uplift probably affected the basin during the late stages of Gres d'Annot deposition. In places, the system may have responded by incising channels, of which this is perhaps an example.

5.3.4.2. V37 and V2: two small scale channels (less than 100m width and 10+m depth).

V37

The 40m sandstone body is located at the base of the first thickly bedded Gres d'Annot at this locality, above appoximately 30m of GdAI. It lies immediately above a steep normal fault which down-throws to the NW. The fault can be traced through the Calcaires Nummulitiques and Marnes Bleues Formation, but does not pass through the thinly bedded sandstones of the lower Gres d'Annot Formation.

The photo of fig. 5.22 and the log show the extent of the exposure and the facies in the thickest part of the body. Only the lower massive sandstones vary greatly in thickness: eg. the lowest from 15m to 1.1m over a distance of 50m towards the south-east. The high relief basal surface cuts down through some of the thinly bedded sandstones and siltstones that characterise the base of the Gres d'Annot Formation. There are no erosive lineations at the base of the bed, but subsequent beds display grooves and flutes showing palaeoflow from NE to SW. Elsewhere, the directions recorded are consistantly E to W.

Part way up, one bed is highly erosive defining a shallow, 1-2m relief, broad channel at the same site as bed 1. In the basal metre tabular cross-stratification is observed in granule sediment. It is laterally disrupted by dewatering structures. The laminae dip towards the south-west. Above this level, the sandstones vary in thickness only slightly: the whole body can be traced 3km to the southeast.

Interpretation.

The large turbidity current responsible for the first thick sandstone flowed from the west and encountered a sharp drop, at the site of a recently active steep fault, oriented NE-SW, obliquely to its original flow. It flowed over this "weir" and turned to parallel the fault trace in its hanging wall. The uneven topography and its obliquity with respect to the regional palaeoflow direction caused the current to erode a small channel. Hence the palaeocurrent data: parallel to the fault trace, oblique to the regional trends.

This outcrop viewed from a distance can be deceptive. The lower bed is channelised, but much of the thickness of the body may not be the result of erosion; instead, it probably results from the flow having deposited the thickest sandstone in the pre-existing low. A similar hypothesis has been proposed by Farrell (1984) for turbidite sandstone bodies in the Ainsa basin, South Pyrenees, which he interprets to have filled the pre-existing topography of slump scars. These too appear to be channelised, giving rise to the interpretation of Mutti et al (1981).

V2.

The Gres d'Annot is exposed in the hanging wall of a steep, normal fault oriented NE-SW, down-throwing to the SE. This fault was active during the deposition of the Gres d'Annot, as evidenced by an internal unconformity part way up the turbidite sequence (section 5.4.4.2). The lowermost moderate and thickly bedded sandstone beds are lenticular, of very limited lateral extent (less than 10m) with concave-upward bases and flat tops.

The lower contacts are erosive. Fig. 5.23 shows the geometry of the beds with respect to the fault, and the log shows the sandstone facies of the lower part of the section.

Interpretation.

A small amount of throw on the fault caused the first current to turn, flow parallel to the fault and erode a small channel. In this case, the fault down-throws to the SE: therefore, currents flowing from the east encountered a topographic barrier oblique to their flow direction. They were deflected and new eddies were generated which eroded small channels against the slope. The position of the topographic low changed as the first fill left a slight topographic high, which was compensated for by the next deposits. Further faulting induced a repeat of the cycle.

The channels are small, filled by the deposits of a single current; the channel fills have wings of horizontally bedded sandstone. Therefore, the fault-related subsidence and inferred fault displacement was not great.

To conclude, there is <u>no evidence for distributary channels</u> anywhere in the studied outcrop of the Gres d'Annot Formation.

Only one moderately large channel has been identified in the study region (Denjuan). Apart from the basal 50m, the facies associations within the channel are the same as those observed elsewhere in the succession. It is probable that the channel was incised in response to an extrinsic control (in this case, tectonic activity increased the basin slopes) and does not represent the progradation of a sedimentary system.

Some small scale channels were scoured and filled during a single event. Two of these can be shown to relate to topographic discontinuities adjacent to faults; at least one was a shallow, ephemeral channel, perhaps forming a braid plain, filled with the deposits of one large sediment gravity flow.

The lack of channels may be directly related to the nature of the sediment gravity flows. The ability of a flow to erode a channel is directly related to the turbulent energy of the fluid. The sedimentary facies exhibited in the Gres d'Annot Formation imply that the large currents were high density sediment gravity flows: almost certainly turbidity currents, but perhaps approaching cohesionless sandy debris flows. The high concentration of coarse sediment damped turbulence and perhaps suppressed the formation of the highly erosive head of the classic turbidity current. Voila, no channels!

5.4. The role of tectonic activity during deposition of the Gres d'Annot Formation. Part 1: Evidence for syn-sedimentary tectonic activity.

5.4.1. Introduction.

Compressional tectonics controlled the evolution of the Gres d'Annot basin (section 6.6). The effect of tectonic activity prior to, and to a lesser extent during, the deposition of the Gres d'Annot can be observed at all scales.

The gross basin geometry and its maximum depths were dictated by the relative movement of Europe with respect to Apulia and Iberia. The tectonic setting defined the positions of shorelines, the width and slope of the shelf, the position of slope breaks and the angle of the basin margin slope.

The three-dimensional geometry of the structures under the basin define a meso-scale (area of several square kilometres; relief of >100m) basin floor topography. This divided the basin into depocentres some demonstrably isolated from other parts of the basin for at least part of their history. The complex pattern of subsidence and uplift within the depocentres is controlled by the timing of activity on individual structures. The relationship has been discussed in chapter 2 and it is illustrated in the structural cross sections. The origin of the palaeotopography at all scales and its effect on turbidite deposition is considered in depth in section 5.5.

This section considers the effects of syn-sedimentary deformation on the deposition of the Gres d'Annot turbidites.

5.4.2. Evidence that the Gres d'Annot Formation was deposited during a period of relative tectonic quiescence within the basin.

The Gres d'Annot Formation is remarkable for its monotony: therein lies the evidence, albeit negative, for tectonic quiescence during the deposition of the succession.

There are no olistostromes or intraformational megabreccias. There are few slumped horizons, or slid blocks of intra- or extra- formational sediment: those seen can often be related to failure of sediment from a pre-existing basin slope (section 5.5). There is only one large scale internal unconformity mapped in the formation, at Chambre du Roi (fig. 5.24). All these features are observed in basins with strong syn-sedimentary tectonic activity, for example the Tortonian basin of the Southern Betics.

There are features within the Gres d'Annot that are interpreted to be the product of tectonic activity contemporaneous with deposition of the succession. They are few in number and are either small scale phenomena, or they are broad low amplitude features induced by limited movement on deep seated structures.

5.4.3. Evidence for syn-sedimentary tectonic activity external to the Gres d'Annot basin.

5.4.3.1. The nature of the Gres d'Annot Formation.

The thickness and consistently sand rich nature of the Gres d'Annot Formation are important indicators of tectonic activity to the south of the marine basin. For several million years, the basin received a persistent supply of moderately coarse detritus from its southerly source areas. The St. Antonin syncline is an example of the semi-restricted shallow marine basins to the south in which sand was temporarily stored and periodically cleared by major storms and earthquake related slope failures.

The volume of sandstone in the Gres d'Annot succession does not decrease upwards overall: therefore, the supply of sand did not diminish gradually with time. Therefore, a large area of land was exposed and continued to be uplifted. The source areas to the south are interpreted as the culminations of active basement thrusts in the Pyrenean orogenic belt, principally between Maure-Esterel and Corsica-Sardinia.

5.4.3.2. The conglomerate marker beds.

The conglomerate facies described in section 5.3.6. are interpreted to be the extensive deposits of large debris flows. The distinctive clast composition of these conglomerates imply that the source only supplied sediment in special circumstances.

The simplest mechanism is proposed: large volumes of unsorted detritus were released into the basin after rare earthquakes. Labaume and Seguret (1984) argue that earthquakes triggered exceptionally large turbidity currents in the Hecho basin of the South Pyrenees. Other examples are described by Mutti et al (1984) and Rupke (1976). Many of the megabeds described were sourced from areas that did not supply detritus into the basin at other times. It is not possible to determin the nature or exact location of the tectonic activity responsible for the earthquakes adjacent to the study region.

5.4.4. Subtle changes in basin floor slope that affected deposition over wide areas.

5.4.4.1. Annot.

The south face of the Chambre du Roi can be studied from the road at Scafferels (fig. 5.24) where an unconformity within the succession of thick sandstone bodies is visible. The unconformity is only seen in this E-W section; the sediments are conformable in the N-S section of Les Gastres. Therefore, the principal component of tilt was to the west.

In the Les Gastres section, the horizon immediately below the level of the unconformity is disrupted along the whole exposure, but particularly in the south towards the Chambre du Roi (encl. 5.6). Large scale (30m+) recumbent folds with E-W hinge lines are observed, closing to the south in most cases. Segments of sandstone dip southwards at a high angle to the surrounding beds (encl. 5.6). The horizons above and below this level are undeformed.

Interpretation.

The internal low angular unconformity observed at Chambre du Roi occurs immediately above the disrupted horizon seen along Les Gastres. The two phenomena are therefore felt to be linked. If so, the structures observed in the disrupted horizon should have developed in young, almost certainly unconsolidated, sediments. The structural style provides the evidence: the isoclinal folds observed in the thick, massive sandstone are ductile structures. As pointed out previously, there is evidence that the Gres d'Annot were only ever deformed at high structural levels: the late fold structures, often kinks, were produced primarily by brittle processes. Therefore, the folds at Les Gastres were produced soon after deposition, in unconsolidated sediment.

The structures are not related to a specific tectonic structure. This and the nature of the folds suggest that the structures were caused by slumping. The slope down which the whole sediment body failed in this spectacular manner is interpreted to be generated above the hanging wall ramp of the Permian thrust sheet BIII (fig. 2.38, 2.39).

5.4.4.2. Peira Cava.

Conort and Odishou (1978) show that the position of the thickest sandstone bodies migrates towards the west upwards through the Gres d'Annot Formation (fig. 5.25). They interpret the consistant displacement of the depocentre as a response to tilting the basin floor above an active tectonic structure.

5.4.4.3. The Trois Eveches outcrop area: Dormillouse to Chalufy.

A section through the outcrop area was constructed by Ghibaudo (1985; fig. 5.3b). It shows that the Gres d'Annot onlap the Marnes Bleues in sections along the whole length of the outcrop area: the palaeoslope dipped northwards. Therefore, a traverse from north to south along this outcrop area encounters progressively younger sections of Gres d'Annot.

The dip of the Gres d'Annot strata varies along the exposure: the beds dip steeply to the E, on average, in the north and dip gently to the NE at Chalufy in the south. The apparent stratigraphic thickness of the Gres d'Annot is approximately 3000m. The Schistes a Blocs Formation can be traced along the eastern margin of the outcrop area. The contact between it and the Gres d'Annot is tectonic. The thrust plane dips quite gently to the east and appears horizontal in a N-S section. Therefore, it cuts down through Gres d'Annot stratigraphy to the north.

Interpretation.

Refer to cross section (enclosure 7) and discussion in section 6.6.5. It has been suggested (Ghibaudo, 1985) that the beds of the Gres d'Annot were originally all horizontal and parallel to one another, having filled a pre-existing, passive relief of 3000m. The basin was therefore >3000m deep.

The cross-sections available allow for an alternative hypothesis: the site of Gres d'Annot deposition was progressively offset towards the south, in response to uplift at the southern margin of the outcrop area. The model is illustrated in fig. 5.26. The offlap model predicts that the preserved top of the Gres d'Annot Formation is more or less parallel to the sedimentary surface at the final stages of Gres d'Annot deposition. Only a thin amount of Gres d'Annot need be eroded beneath the Schistes a Blocs. No unconformities are required within the part of the Gres d'Annot preserved. This model is compatible with the interpretation of the formation boundary as a tectonised, but essentially stratigraphic contact.

This model explains the field evidence without recourse to the large vertical yo-yo movements proposed by Jean et al (1986): they require uplift, severe erosion, then subsidence to deposit the Schistes a Blocs in the sea!

5.4.4.4. Conclusion.

Which structures were responsible for the slight, regional changes in slope?

Both the Annot and Peira Cava outcrop areas are bounded to the east by south-west and west vergent asymmetric anticlines. These anticlines are interpreted to be thrust hanging wall ramp anticlines. In the case of Annot, the Mesozoic large scale kink folds are the surface expression of a hanging wall ramp in the basement (BIII; fig. 2.39, 2.21). It is likely that movement on these structures at depth during the deposition of the Gres d'Annot induced the subtle, but regionally consistent, changes in basin floor slopes noted in both these outcrop areas.

Dormillouse is part of a culmination, at present structurally higher than Chalufy. Some compressive structures in the area pre-date the Calcaires Nummulitiques Formation. It is probable that they were active throughout the Eocene.

5.4.5. Small scale surface structures within the Gres d'Annot basin.

5.4.5.1. Denjuan: an emergent thrust tip.

The features described below are illustrated and interpreted in fig. 5.27. This locality is at the southern end of the Trois Eveches outcrop area. The basin floor sloped down to the north across the whole outcrop area, so the Gres d'Annot succession exposed here is high in the formation as a whole. From Denjuan to Chalufy, the beds of Gres d'Annot pinch out towards the south-east, onlapping the basin floor topography defined by the Marnes Bleues Formation (section 5.5.).

A lateral section through a thrust is exposed: both the footwall ramp and the hanging wall ramp are seen within the basal 30m of Gres d'Annot section as seen at this location. Fig. 5.27 shows the outcrop of the thrust and details of the exposure in the west face of Denjuan, the interpretation and the proposed model.

To the North-west, left of the footwall ramp in the base of the Gres d'Annot, two thick sandstone beds are seen to thin abruptly immediately above that ramp. To the south-east, a lenticular sandstone bed is observed to terminate in front of the gentle slope of the hanging wall ramp.

A detailed study of the hanging wall ramp, as developed in GdAI at the base of the formation, reveals SW-vergent assymmetric metre-scale folds. Beneath a 90cm thick, massive sandstone two thin sandstones (6cm and 2.5cm thick) are isoclinally to ptygmatically folded. Sedimentary structures, including parallel and cross-lamination are well preserved. The folds are NW-vergent, with NE-plunging hinges. The folds tighten and become more as$ymetric approaching the thrust plane. 30 metres away from the fault, the beds are totally undeformed. Flame structures are observed to penetrate the base of the 90cm thick sandstone some 15cm above.

Interpretation.

A limited number of beds are deformed around the thrust ramps. The first of the sandstones not to be folded are of limited lateral extent: they thin onto the crest of the hanging wall anticline beneath. The implication is that the thrust structure pre-dated their deposition and formed a small-scale topographic high on the already sloping sea bed. The subsequent sediments onlapped the additional relief.

Additional evidence lies in the structural style of the thrusted sediments. The folds observed in the lateral hanging wall ramp of the thrust are interpreted to have formed while the sediment was unconsolidated. The lack of convolute lamination implies that the sediment was coherent; the fold style demonstrates that the sediment was highly ductile. In the region, folds in the Gres d'Annot that post-date its cementation show brittle fabrics, like joints and veins. Therefore, the folds developed a discrete, but only short time after the deposition of those beds.

The geometry and the intensity of the deformation in these thin sandstones relates entirely to the geometry of the hanging wall ramp. Therefore, the folds are not pre-deformation slump structures. Instead, they are the result of flexural slip during formation of thrust ramp anticline. The thicker sandstone above bent: water ejected from the sheared beds below produced the flame structures observed and fluid escape may have been responsible for the lack of structures in the bed.

All the evidence from the deformed beds close to the hanging wall cut-off suggests that the thrust ramp was produced soon after the beds were deposited. The sandstones which onlap the topography of the lateral hanging wall show that the relief produced by the thrust ramping to surface was not great. The thrust can be mapped through the Marnes Bleues to the culmination in the Calcaires Nummulitiques.

5.4.5.2. Normal fault active during the deposition of the Gres d'Annot: V2 and V4.

V2.

Both V2 and V4 are at the southernmost limit of the Gres d'Annot outcrop in the Trois Eveches outcrop area (fig. 5.23 and 5.28). The Gres d'Annot outcrops in the hanging wall of a steep normal fault. The lowermost sandstones are the fills of small channels (section 4.4.6) There is a distinct change in dip of the beds half way up the exposure. The lower beds dip in towards the fault (bedding 210/32NW): immediately above the change in dips, the sub-horizontal sandstones thin abruptly towards the tilted beds: the palaeocurrent directions recorded in these beds are very diverse. The features of the exposure are summarised in the photo and line drawing interpretation of fig. 5.23.

The palaeocurrent readings from the beds infilling the topography above the tilted beds are diverse, implying complex slope patterns in the third dimension.

Interpretation

The fault was active during the deposition of the Gres d'Annot at this locality. Some displacement preceeded sandstone deposition at this site, Further displacement occurred on the fault after 20m of sediment had been deposited, and the beds in the hanging wall were tilted.

V4.

At this locality, sandstone beds cannot be restored across a small displacement fault. As before, some fault displacement is interpreted to have occurred during deposition which accounts for the thicker, "channelised" sandstone in the hanging wall (fig. 5.28).

5.5. The role of tectonic activity. Part 2: The influence of palaeoslopes on the deposition of the Gres d'Annot Formation.

The extent of exposed unconformity between the Gres d'Annot and the Marnes Bleues is displayed on the map of fig. 5.1. The onlap relationship defines the palaeotopography immediately prior to the deposition of the Gres d'Annot at each locality.

Previous authors (Ravenne and Beghin, 1983; Cremer, 1983; Ghibaudo, 1985) have described many of these preserved onlap surfaces. None of them have considered either the origin of these slopes or, more importantly, the sedimentary phenomena directly related to their existence. Both these subjects have been given a high priority in my research.

Five examples of this unconformity are described in the following sections. They were chosen, because the palaeoslopes in each case were oriented differently with respect to the principal palaeocurrent directions in the area. In each case, the nature and origin of the slope are outlined, prior to a full discussion of the effects of the basin margin slope on the sediment gravity flows, their deposits and the post-depositional processes.

<u>5.5.1. Chalufy (enclosure 5.1A, B):</u> palaeoslope oblique to, and sloping in the direction of, the E-W palaeocurrents.

5.5.1.1. Introduction.

The exposure is the west face of Chalufy at the southernmost tip of the

Trois Eveches outcrop area (fig. 5.28), therefore in the upper part of the Gres d'Annot Formation. The angular unconformity at the base of the Gres d'Annot represents a palaeoslope that dipped down to the north.

Four sandstone bodies (A-D) of the associations GdAII and III terminate against the slope. These constitute the outcrop localities: V36, V41, V45 and V4. For each sediment body, the sediment facies immediately adjacent to the slope were compared with those at various distances from the abrupt termination. The variations observed constitute a blueprint of "edge effects" against which other outcrops may be compared.

5.5.1.2. The nature of the palaeoslope.

The onlap surface is 1km long and is therefore defined as a large scale feature. To the north of the angular unconformity, the contact becomes a disconformity; to the south, the outcrop terminates. The apparent angle of onlap is 15' along most of the outcrop, becoming somewhat steeper at the base of the slope (outcrop V36).

It is possible to define the strike of the onlap surface as NE-SW or NNE-SSW. This accords well with independent palaeocurrent evidence. Away from the slope, the palaeocurrent readings are east to west; adjacent to the slope, they are consistantly NE to SW. This too implies that the slope was oriented NE-SW. The palaeoslope was oblique, a maximum of 45', to the unhindered palaeocurrents.

5.5.1.3. The origin of the palaeoslope.

The onlap surface is parallel to bedding within the Marnes Bleues

Formation. The steep bedding dips in the latter represent the upper surface of a complex wedge of marlstone. The base of the wedge is parallel to the bedding below it and is level with the first onlap surface.

The contact is discrete, lined with sparry calcite veins <2cm thick. There are no lineations associated with the contact or these veins. The Marnes Bleues are intensely kinked in the basal metre or so of the hanging wall; above this, a weak penetrative sub-vertical, NW/SE cleavage can be traced through much of the wedge. In the footwall marlstones there are narrow, widely spaced kink bands. The data collected shows all the structures to be NW-vergent (see enclosure 5.1A).

Interpretation.

The bedding surface discontinuities are narrow zones of intensely deformed sediment. They are not unconformities. Deformation is concentrated along these planes: the bulk of the marlstone in both the footwall and hanging wall is undeformed. The small-scale NW-vergent structures measured near the basal contact, close to the wedge tip, imply that the wedge was transported towards the north-west. However, absolute transport directions cannot be ascertained because there are no lineations.

The sediment wedge could have been emplaced by gravity, as a submarine slide, or by tectonic shortening, as a thrust sheet. Both mechanisms are consistent with the regional data. The onlap directions recorded through the Trois Eveches outcrop area imply that a topographic high existed to the south. It is possible that a submarine slide originated in the Beauvezer area and came to a halt at the first break in slope, at Chalufy.

A southwestward thrust transport direction has been established for the Alpine structures in this region (Graham, 1985 and Lawson, 1987). Alpine thrusting began prior to the deposition of the Gres d'Annot (fig. 2.20, 22). Therefore, the slope of Marnes Bleues could be interpreted as the lateral hanging wall ramp of a thrust sheet. The apparent NE-SW orientation of the slope is compatable with both models.

5.5.1.4 The relationship between sandstone deposition and the palaeoslope.

The sandstone bodies which terminate against the steep palaeoslope can all be traced many kilometres northwards, away from the unconformity. Over most of their length, they are monotonous, but within 100m of their onlap, they develop quite different characteristics. In each case, the percentage of sand changes towards the onlap (enclosure 5·1 5·2). Correspondingly, the palaeocurrents change from E-W (the regional direction) to NE-SW (parallel to the slope). A wide variety of sediment body geometries and facies are observed, the principal features of which are described below.

5.5.1.5. <u>Body A: locality V36.</u> (Enclosure 5.1A)

All the features described below are illustrated on encl. 5.1A The sandstone body of type GdAIII terminates abruptly against the palaeoslope. From a distance, ~~that~~ the contact is not apparently a sedimentary one. However, there are relatively small-scale, but very important facies changes that relate entirely to the effect of a steep slope on turbidite deposition.

The beds thin towards the onlap, within 20m of the contact. At the base above.

In the lower part of the body, beds lose thickness by virtue of relief on either the basal or upper surfaces. Thinly bedded sandstones, GdAI, fill any remnant topography adjacent to the slope.

In the well laminated tb facies of some beds, microfaults are observed. The principal displacement occurs on fractures dipping away from the slope. Dish structures, flames and small, soft sediment folds are common within 10m of the onlap.

Interpretation.

The height of the palaeoslope cannot be ascertained, because the top of it has been eroded, but it was limited by the dimensions of the slid block or the displacement on the thrust. Turbidity currents coming from the west first encountered the rear of the wedge. Some currents were diverted at this stage and could have deposited their load some distance from the palaeoslope, leaving thinner and finer sandsone deposits adjacent to the slope (sand avoidance). The shape and orientation of the wedge alone may have caused the currents to turn and flow parallel to the NE-SW slope, thereby increasing the chances of sand deposition close to the slope. (sand concentration). Some currents may have surmounted the topography and separated from the sediment surface producing eddies in the lee of the slope. This would account for the diversity of ripple crosslamination trends.

The interaction between the steep topographic slope and the turbidity currents may have disrupted the flow and increased the turbulence adjacent of the sediment body, no thick sandstones directly encounter the unconformity. Each bed changes laterally to a thin siltstone layer prior to the encounter. Towards the top of the unit, thick sandstones maintain their thickness and wedge directly against the palaeoslope.

The upper beds can be traced to their contact with the onlap surface. The related facies changes are illustrated on the enclosure.

100m from the slope, well bedded GdAII shows no sign of the approaching onlap. 50m from the slope, the contacts between the beds are disrupted. The bed contacts are discontinuous and lined with large irregular, contorted shale clasts. Parallel and crosslamination are less common, replaced by convolute lamination; dewatering and load structures are common. 15m from the slope, the well bedded sandstones are replaced by several metres of massive, coarse sandstone. Planar discontinuous boundaries are the only evidence that it comprises the deposits of more than one event. 10m from the slope, the single bed is sub-divided by closely spaced (10-30cm) horizons of shale clasts, with siltstone layers.

4m from the slope, the layers break up and disappear. Low angle dipping surfaces, associated with vertical fluid escape pipes are seen towards the top of the now homogeneous sandstone. Grooves at the base of the bed are oriented NE-SW, or due N-S. Sandstone is injected upwards into the overlying sandstone. The basal surface of the bed encounters the unconformity and curves up sharply parallel it. The marlstones beneath are undisturbed: the turbidity currents did not scour into the slope: not even small-scale erosive features are observed. The upper surface of the bed is concave upwards, so that the sandstone projects at least one metre above its regional top. The "wing" of the sandstone amalgamates with the beds to the slope. This enhanced erosion, and so sandstones close to the onlap are progressively amalgamated.

The thickness of a turbidity current is several times greater than the thickness of its deposit. When a high density current flowed parallel to the topographic high, it probably rose a little higher up the slope. For both these reasons, part of the coarse load may be deposited high on the slope, draping the topograhy. This sandstone will be amalgamated with subsequent deposits. Sandstone deposited thus would often fail downslope, giving rise to homogenised sandstone, the dipping surfaces within these massive units, dewatering structures and the microfaults seen. The model for the large scale dipping surfaces is shown in the enclosure and discussed in section 5.6.1. on Annot, where there are more and larger examples of the similar features.

5.5.1.6. <u>Body B: V41</u>

The features of this sandstone body are illustrated and interpreted in enclosure 5.3a . The body is lenticular with the thickest (14m) part 110m from the slope; the maximum thickness is almost directly above the termination of the lower sediment body A. Each sandstone bed is lenticular and successive thin beds amalgamate within 40m of the slope.

The original base of the spectacular upper unit can be traced along part of its length. It is deeply erosive, with a pebble conglomerate filling the scour. Grooves on this base are oriented E-W and SE-NW. This unit is disrupted: huge (10-15m), downward closing folds can be seen at both margins of the body. It is not possible to reconstruct all the bedding discontinuities, in part because much of the sandstone was homogenised soon after deposition which obliterated the original sedimentary features. Flame structures are sheared in a consistent direction towards the NW (038SE/38).

Below the thick channel fill, the thinner bedded sandstones are also isoclinally folded. 10m of section is folded (10/045NE hinge line, by necessity approximate!) immediately beneath the channel. Sand has been injected behind the hinge of the fold. The thin siltstones and shales below are cleaved (025ESE/30 cl.). Thinner bedded sandstones at the base of the body are folded about a N-S (24/002N) hinge.

Interpretation.

The lower beds may be lenticular because they infilled a slight topography on the basin floor. However, the upper unit is channellised. The rarity of channels in the Gres d'Annot Formation generally, and the position of this channel close to the onlap imply that its presence is controlled by the palaeoslope. The model: an energetic current struck the back of the Marnes Bleues wedge (by then somewhat subdued) at an oblique angle; it may have lost some of its total energy, but gained turbulence. As it shot over the side or front of the wedge it turned to the SW and scoured a shallow channel along the topographic depression. It was a sizeable current, or several currents followed in close succession, and deposited a thick unit and a considerable amount of sediment up slope.

Subsequently, the mass of sediment on the steep slope failed down hill, towards the NW, and deformed, giving rise to all the soft sediment structures observed. A small amount of tectonic shortening on a shallow structure (perhaps the base of the wedge) may have induced a small earthquake shock which caused the sediment to fail en-masse.

5.5.1.6. Body C: V45. (Enclosure 5.1A)

This sandstone body is moderately deformed. It is bent upwards near the unconformity until it almost parallels the slope of the contact. A complex thrust fault cuts the outcrop and juxtaposes two quite different facies associations.

To the south east, the principal feature of the sandstone body is a channelised sandstone, which comprises a single bed of massive coarse sandstone: it is thickest (8m) some 130m from the onlap. The basal relief of this channel is 2m. In places the contact is sharp; elsewhere, it is chaotic, with flame structures and mixing of sand from either side of the contact.

A few decimetre thick sandstones outcrop beneath the channel fill. A tight, north-closing isoclinal, metre-scale fold deforms part of the lowermost of these. Coarse sand is injected into the core, disrupting the thin sandstones.

To the north of the fault, the bed thickness gradationally, but rapidly, coarsens upwards from the GdAI facies association that separates the sediment bodies B and C. The sandstones in the main part of the body are <50cm thick. The lower beds are intensely deformed. Tight and isoclinal, decimetre scale folds with rounded hinges are cut by small thrusts. The orientation of individual structures is highly variable, but a significant number have NE plunging hinge lines. The beds are discontinuous, riddled with flame structures, small sandstone dykes and massive balls of sandstone. The deformation is concentrated into several horizons, each a few metres thick.

Away from the fault, to the north, it is not possible to trace the zones of deformation. However, within 200m of the fault zone, the sediment body has the familiar, undeformed features of the facies association GdAII.

Interpretation.

The broad fold of the whole body is a post depositional feature, probably the slope of the thrust hanging wall ramp. The thrust fault zone cuts cleanly through the Marnes Bleues, where it dips to the south. If a southwest shortening direction is assumed for this region, then a fault of this orientation is a lateral ramp. Hence, lateral displacements on the fault juxtapose parts of the onlapping body that were not deposited adjacent to one another in any sense.

Much of the deformation seen in the footwall of this fault occurred while the sediment was soft. The deformation was concentrated in flat lying, broad shear zones. The NE-SW hinge lines are consistent with deformation on a lateral thrust ramp. Therefore, the large scale thrust zone and the soft sediment folds and faults are almost certainly related. This implies an early age for at least one thrust in the Gres d'Annot! The rest of the features are comparable with those of sediment body B and can be explained using a similar model.

5.5.1.7. <u>Body D: V4.</u> ([illegible] 5.[illegible])

This sediment body has two interesting features: the onlap towards the south-east and a syn-sedimentary normal fault which dips to the SE. The more complex bed thickness changes in this body and the slight internal unconformity of the thick sandstones onto the underlying GdAI were produced by the interaction of the fault and the palaeoslope.

The Normal Fault.

To the SE of the small normal fault marked by a gully, the lower thick sandstone is lenticular: flat-topped, with a high relief, erosive base. Away from the gully, to the NW, thinly bedded sandstones are the direct lateral equivalent of the channelised sandstone. These thicken rapidly away from the gully, defining an internal unconformity: thickly bedded sandstones onlap a horizon at the top of GdAI facies.

The features of this lenticular sandstone are summarised in enclosure . It is as~~s~~ymmetric, with the deepest scour 5m from the gully. The sandstone thins quite gradually over 40m to the SE. The base is ornamented with bulbous flute casts grouped into larger erosive troughs and other structures produced by turbulent eddies at the base of the flow. The palaeocurrents derived from the flute casts are consistantly towards the NW! The NW margin of the sandstone is not precisely defined. It is marked by chaotic admixture of coarse sandstone, siltstone and shale clasts. A similar mess is observed on the other side of the gully.

Interpretation.

The gully marks the site of a small displacement normal fault, which was active during the deposition of the lower sandstones of body D. Turbidity currents scoured a depression in the hanging wall. Palaeocurrent indicators on the eroded surface are oriented towards the fault itself: ie. down the palaeoslope into the topographic depression above the hanging wall. The flutes are not perpendicular to the cross-section of the body. There is no evidence that the scoured base is an elongate feature. Therefore, the depression may not be a channel, rather a scoured pit.

The sandstones to the NW of the interpreted fault onlap its footwall. The fault dips towards the palaeoslope and does not cut through the unconformity. Therefore, it is interpreted as an antithetic fault, produced in response to slip (probably not great) along the contact between the Gres d'Annot and the Marnes Bleues.

The onlap.

Individual beds thin as they are traced to the slope, so that along most of the length of the unconformity there is a discrete thickness (10-1m) of thin sandstones and silts (GdAI). In detail, the thickness of a bed may increase before thinning abruptly in the last ten metres or so. Beds amalgamate (fig. 5.29): the sandstones loose their coherent laminae structures in favour of wavy lamination; streaks of different grain size replace layers that were once coherent; large flame structures are consistantly sheared towards the NW.

To the NW of the gully, slump structures are common. Some of the folds crests are erosively truncated by subsequent deposits. The folds have NE-SW axes and are NW-vergent or flat lying with closures to the NW. In one bed, 30cm thick, a single set of cross-bedded medium sandstone forms the upper 25cm of the bed. The sediment is better sorted within the cross-bedded unit. On the base of the bed, grooves are oriented 10/072E; above, the cross-stratification dips to the NW (236NW/33).

Interpretation

The cross-bedding is the only example of its kind in the outcrop area. The better sorted sandstone associated with it suggests that the sand was reworked from a pre-existing deposit. The discrepancy between the palaeocurrent indicators at the base and top of the bed is critical to the interpretation of the cross-set. The sandstone was orignally emplaced from the east by a turbidity current. The current encountered the Marnes Bleues slope and was deflected: it flowed to the NW, perpendicular to the slope, over its recent deposits and reworked the top part into bed forms >25cm high. Reflected turbidity currents have been interpreted by Pickering and Hiscott (1985) in Ordovician sediments of Quebec. In this exposure, the slope encountered by the currents is exposed adding weight to the interpretation.

5.5.1.9. The highlights of the Chalufy Onlap exposures.

Each onlapping sediment body has characteristics particular to its relationship with the palaeoslope and the underlying sediment. The simplest response to the slope is exhibited in the sediments of body A. The lenticular sandstones and the channels of body B and C reflect a more complex response of the turbidity currents to the palaeoslope.

All the turbidity currents were affected by the presence of a slope oblique to their original flow path. Sandstone concentrations both increase and decrease in different beds in the immediate vicinity of the onlap. This demonstrates the variability of the currents' interaction with the slope. Many of the currents were diverted and flowed parallel to the slope; at least one was reflected from it. The resulting current was strong and persistent enough to rework medium sand into sizeable bedforms. The channelised sandstones and some of the bed amalgamation are evidence for increased turbulence in some of the larger flows. Sediment was deposited up the slope and amalgamated with subsequent flow deposits.

Post-depositional failure of sediment downhill produced many of the sediment facies preserved today. It gave rise to chaotic facies, bed amalgamation, homogenised sandstone, dewatering strucutres, slumps and other complex soft sediment structures. There is also evidence that some tectonic shortening on occurred only shortly after the sediment was deposited. Slip may have occurred on the unconformity itself, perhaps giving rise to the syn-sedimentary fault within body D.

<u>5.5.2. Le Mourre de Simanche</u> (fig. 5.30). The palaeoslope was lateral to the principal palaeocurrents.

To the north of Chalufy, the Gres d'Annot onlap the Marnes Bleues again at Denjuan, then on the SW face of Le Mourre de Simanche. The outcrop is spectacularly illustrated in fig. 5.30. The length of the unconformity is 1200m; the palaeoslope dipped to the N (NW) at an angle of <15.

5.5.2.1. The origin of the slope.

Several thrusts are mapped which cut through the whole Tertiary sequence. One of these thrusts broke to surface during the deposition of the Gres d'Annot at Denjuan, south of Le Mourre de Simanche (section 5.4.4.1.). This lateral ramp cuts down through the Marnes Bleues to a culmination of thrust sheets in the Calcaires Nummulitiques at Tete de l'Adrech.

Interpretation.

See the model drawn in fig. 5.30. The culmination observed in the Calcaires Nummulitiques developed immediately prior to the deposition of the Gres d'Annot at Mourre de Simanche. It formed a topographic high. The Gres d'Annot onlap the footwall ramp of the structure. In addition, there was almost certainly a deeper broader structure beneath to account for the length of the onlap. The thrust then propagated through to surface at Denjuan, where the syn-sedimentary deformation has been described.

The amount of shortening prior to, and contemporaneous with, the Gres d'Annot cannot be determined from the lateral ramp displacements.

5.5.2.2. The Gres d'Annot sediments themselves.

The onlap is quite simple with a slight internal unconformity. Above this surface, a thick sandstone body lies in sharp contact with a silty shale horizon. The palaeocurrents are consistantly E-W.

Near the base of the onlap a single 10cm thick bed is isoclinally folded. The folds die out 30m west of the onlap. The hinge lines are highly variable, ranging between 20/010 and 26/102: the structures are west vergent.

Interpretation.

The palaeoslope was lateral to the flow direction. Therefore, the turbidity currents flowed passively past the rather gentle incline. The soft sediment folds and the internal unconformity are believed to be related: further shortening on the culmination beneath increased the slope slightly. A certain amount of slip occurred on and near the unconformity surface, producing soft sediment folds locally in the Gres d'Annot as a result of flexural slip. The subsequent deposits onlapped the new palaeoslope.

<u>5.5.3. Le Ruch.</u> Enclosure 5.4, 5.5. Palaeoslope oblique to, and opposing the regional paleocurrents.

5.5.3.1. The nature of the palaeoslope.

At the northern limit of the Annot outcrop area, the Gres d'Annot onlap a south-west dipping palaeosurface ~~(fig.)~~. The slope is steepest in the south at 30' and shallows to <10' in the north. The onlap is continuous over 400m: to the north, the outcrop terminates; to the south, it is truncated against a steep fault. The palaeocurrent data collected elsewhere in the outcrop area shows that the gravity flows brought sediment from the south. This palaeoslope is therefore at a high angle to, and opposing, the flow of these currents.

Along the unconformity, thinly bedded turbidites occur in the upper few metres of the palaeoslope. They are subparallel to bedding in the Marnes Bleues, but thin gradually to the north.

To the NE of the unconformity, there is a major SW-vergent kink fold in the Upper Cretaceous limestones. The Gres d'Annot dip 28' to the south-west, less than the dip of the Cretaceous rocks. Interpretation: The Gres d'Annot are folded about the kink fold, but the lesser dips imply that part of the shortening accommodated by the fold pre-dated the tubidites. Therefore, the Gres d'Annot are interpreted to onlap the steep limb of the early asymmetric fold.

The lowermost turbidites are folded together with the Marnes Bleues; therefore, the slope developed as the last marls were deposited and progressively as the first turbidites arrived in this part of the basin. Note, the structure may be older still, but the pre-Marnes Bleues topography was infilled with marl.

The outcrop divides into two sandstone bodies separated by 12m of GdAI. The lower sandstone body and the GdAI and the very base of the upper body onlap the unconformity. The relationship of the upper body to the basin symmetric.

Interpretation.

The flows arriving from the south did not erode the slope which opposed them: this implies that the currents were not very energetic and probably each flow did not deposit large volumes of sediment.

The almost total absence of massive sandstone, the well sorted, clean texture and preservation of the fine laminae is not in character with Gres d'Annot sandstones in general. The currents working the sediment were apparently persistent and quite strong, capable of developing sizeable angle-of-repose bedforms. The partings within the sandstones are interpreted as the set boundaries of very low relief, long wavelength bed forms.

Palaeocurrents.

The basal erosive lineations of the thinly bedded sandstones in GdA are oriented S-N. 10m from the slope, within the lower sandstone body, the grooves sharply define SW-NE palaeocurrents; adjacent to the slope E-W striations and grooves are common. Against the unconformity, one flute cast records flow towards the NW.

The climbing and plane ripple cross-stratification and the single cross-bed all record palaeocurrents that flowed from east to west. The same orientations are recorded from cross-stratification everywhere at Le Ruch.

slope cannot be ascertained, due to erosion.

5.5.3.2. The lower sandstone body.

The base of the exposed onlap is illustrated in the photos and interpretations of enclosure 5.5.

Sedimentary facies.

In the GdAII sediment body, the sandstones terminate rather abruptly against the slope. However, the bases of the sandstones do not scour into the slope itself: the onlap is **passive.** Against the steepest part of the slope, the beds do not thin at all adjacent to the unconformity. Where the slope was less steep, the sandstones thin and amalgamate over 15m or so as they approach the slope.

Further away from the unconformity, the beds display rapid and complex thickness changes, the sandstones are lenticular. Most often, but not always, it is the upper surface of the beds that show positive relief.

The sandstone is, for the most part, medium grained, moderately to well sorted and clean. Sharp planar partings sub-divide some beds. Most of the sandstone is very well laminated. The partings may divide massive from well laminated sandstone, but more often than not, they separate portions with differently oriented laminae. The divergence is always low angle. A single tabular set, 11cm thick, is observed at the top of one bed. The upper surface of the bed records the relief of the depositional forms, preserved, either below siltstones and shales, or beneath drapes of sandstone which eventually infilled the topography. The forms were broadly

Interpretation.

The sandstones were first emplaced by currents that flowed from the south. The slope, oriented across their path, caused the currents to loose momentum and deposit sediment. Some or part of the currents may have surmounted the topography (in the order of 400m when the onlap began) taking only fine sediment in suspension. The rest were deflected by the slope and flowed from the slope towards the west. In one case, the deflected current reworked the underlying sediment to form 2D megaripples; in others, ripple crosslamination.

It is probable that many of the original sediment gravity currents flowed towards the slope for several hours and interacted with the currents reflected from the slopes. The large wavelength, low amplitude features are interpreted to be the product of internal waves or seiches produced at the upper boundary of such a combined flow. Interested readers should read Luthi (1980) and Wunsch (1975), who have modelled the behaviour of turbidity currents and internal waves respectively.

5.5.3.3. The upper sandstone body. (Enclosure 5.4)

The upper sandstone body may be divided into three sub-sequences:

i) The lower part consists of parallel bedded sandstone which onlaps the Marnes Bleues towards the north.

ii) The middle unit downlaps southward onto the lower beds at an angle of less than 10'. The total positive relief on this feature is ten metres or less.

iii) The upper unit onlaps the top of the middle sequence, infilling the topography.

Within each package of sandstone, individual beds amalgamate towards the tip of each wedge.

The deposits are coarse-medium, moderately-poorly sorted sandstone, rich in plant debris of all sizes. Their bases are erosive, particularly towards the tip of each wedge. The upper portions of the bed may be well laminated, some with well developed tb division and good pcl; others with *thick ripple cross-laminated* tc divisions.

Some *fragments* of the bases of these beds display complex patterns of tension gashes filled with sand from the overlying bed. These cut flute casts and are generally NE-SW, *parallel to the strike* of the palaeoslope, *at a very high angle to the* palaeocurrent indicators. The palaeocurrent data are shown on fig. 5.31: the grooves and flute casts record SE to NW and S to N palaeocurrents. The ripple crosslamination and one crossbed show that the currents depositing the upper part of the beds flowed towards the west.

Interpretation.

High relief medium scale features, such as the downlapping unit, are not seen elsewhere in the Gres d'Annot Formation. The diversion of the current directions from NW at the base of each bed to W at the top of each bed, and the location of this unit adjacent to a steep palaeoslope, indicate that the positive feature was produced by the interaction of turbidity currents with a topographic barrier.

The bottom and top sequences were deposited by currents which were deflected westwards by the slope. The middle, downlapping sequence was deposited by currents which left no deposits elsewhere. It was constructed by the current reflected off the palaeoslope.

The tension cracks reflect north-south extension of the beds before they were consolidated. This was the result of small scale downslope sliding, or slight gravity spreading of the positive sediment feature.

<u>5.5.4. Les Gastres:</u> Enclosure 5.6. An example of small scale topography. The palaeoslope dipped away from the arriving turbidity currents.

5.5.4.1. Features of the onlap outcrop.

The exposed onlap length is 400m. Restoring the Gres d'Annot to horizontal, the palaeoslope, defined by bedding beneath the unconformity, dipped approximately 16 to the north near its base, towards the NE higher up the slope. The differences in dip are slight; therefore the estimates of palaeoslope are necessarily poorly defined.

The base of the palaeoslope is exposed in contact with the Gres d'Annot, where the Marnes Bleues dip to the west. The Gres d'Annot above the unconformity also dip to the west, but somewhat less steeply. The exact upper limit of the onlap does not crop out, but a large, massive sandstone body limits the maximum extent of the slope to 600m.

Thinly bedded sandstones and silty marlstone form the upper metre of the slope sediments. Adjacent to the slope, the sandstones are disrupted and folded.

5.5.4.2. The origin of the palaeoslope.

The base of the unconformity cannot be traced back into the marlstones, due to the thick forest cover! The northward dipping beds of marlstone appear to be limited to the area near the slope. The soft sediment deformation of the sandstones at the top of the slope are evidence of slumping. Therefore, the palaeoslope is interpreted to be the upper surface of a slide or slumped mass of Marnes Bleues which failed soon after the first Gres d'Annot sediments were deposited.

At Braux, there is corrolative evidence for this hypothesis. There, a slid mass of Marnes Bleues was emplaced above the first 4m of Gres d'Annot. Here, the basal contact and the internal deformation of this *mass can be examined (section 4.6.1.5)*. It was emplaced towards the SW.

5.5.4.3. *Sandstone facies distribution* adjacent to the slope.

The beds do not terminate neatly against the slope! Instead, the sandstones above the contact and the marlstones and thinly bedded sandstones below are intimately involved in soft sediment deformation and dewatering structures. Dish structures are common: as at Chalufy, many of them are aligned parallel to the palaeoslope.

The unconformity surface can be examined at the top of the exposure. Here, the lowermost sandstones can be traced to within 3m or so of the slope. At this point, the bulbous load structures on the base show that the sediment beneath was very soggy. The beds pinch and swell, then terminate abruptly, or thin rapidly to discontinuous silty sandstones which turn up and parallel the slope into a throughly disrupted mass of sand and marlstone.

A strong fabric is developed in the Marnes Bleues further up the palaeoslope. Bedding parallel extension occurred along the planes of this conjugate spaced cleavage. This extension is interpreted to represent failure at the back of a slide, probably related to the slump folds seen further down the slope.

The upper package of sandstones passively infill the topography above the slumped mass. The beds terminate abruptly against the slope, but only one significantly erodes its substrate. The erosive steps on the base of this bed are oriented E-W, sub-parallel to the palaeoslope.

The geometry of the beds over a distance of 300m from the onlap surface is illustrated in the enclosure 5.6. The beds have erosive bases and flat tops.

Palaeocurrent data cannot be collected any distance from the unconformity at this outcrop, but regionally, sediment was transported towards the north. Basal erosive lineations are divergent, even on the base of a single bed; most are oriented W-E, with NW-SE and NE-SW readings. Within 10m of the palaeoslope there are no lineations to measure; even the loads are 3-dimensional.

Interpretation.

The slump mass was emplaced only shortly before the deposition of the first thick sandstones: it was yet unstable and failed, along with thick turbidite beds which deformed and dewatered to produce the facies characteristic of this onlap.

Travelling down to the base of the palaeoslope the turbidity currents gained momentum and scoured the underlying sandstones on the sea bed: the eddies were strongest near, but not immediately adjacent, to the slope. The basal grooves do not necessarily imply that the whole current was deflected by the slope. They may have been produced by secondary eddies generated as the flows passed over the slope.

5.5.5. Braux: Enclosure 5.7. An example of small-scale relief superimposed on large-scale topography.

5.5.5.1. Features of the onlap outcrops.

The outcrop features a large scale base-Gres d'Annot unconformity which can be mapped through the Chambre du Roi and above the village of Annot. This unconformity is therefore greater than 3km long, covering an area >5km. The Gres d'Annot onlapped towards the west: the strike of the palaeoslope varied between N-S and NE-SW. The dip of this slope was 5'-10'.

A small scale unconformity, opposing the regional one, is exposed below the D110 at the eastern limit of the Gres d'Annot outcrop. The Gres d'Annot onlapped towards the east (ESE), onto approximately 300m of an 8' palaeoslope.

5.5.5.2. The origin of the palaeoslopes:

The **large scale, low angle unconformity** represents a complex palaeosurface produced by deformation of Mesozoic and Lower Tertiary strata. The NE-dipping palaeosurface interpreted at Braux can be related to roll-over into the Rouaine-Daluis fault. Partially restored structural cross sections show that displacement on the Argens thrust predates the Gres d'Annot (fig. 2.21). Therefore, towards the north and west, away from the Rouaine fault, the topography is predicted to represents the NW-SE striking slope of the strata rising over the footwall ramp of the Argens thrust. At Annot, the two effects are combined: therefore, the palaeoslope dipped due east. The evidence for this hypothesis lies outside this outcrop area. It is discussed in chapter 2.

Subtle changes of topographic relief were brought about by deformation during the deposition of the Gres d'Annot (section 5.4.3.1. and the infilling of topography related to inactive structures, like the Rouaine fault.

The relationship between the **small-scale onlap**, the Gres d'Annot and tectonic structures in its immediate vicinity.

The St. Benoit fault (encl. 5.8) outcrops to the east of the small-scale unconformity. It is a normal fault, with downthrows to the **south and east.** The fault displaces the Calcaires Nummulitiques and Marnes Bleues *Formations*, but not the Gres d'Annot turbidites of the Crete de la Barre. The Tertiary successions are markedly different on either side of the fault. For example, the thickness of Marnes Bleues is 300m in the hanging wall, but only tens of metres thick in the footwall.

At the base of the Gres d'Annot Formation, a 150m thick sandstone body is exposed in the hanging wall above St. Benoit which has no equivalent west of the fault outcrop. Instead, centimetre and decimetre scale sandstones onlap the thin Marnes Bleues sequence towards the St. Benoit Fault. On this side of the fault a wedge of Marnes Bleues outcrops above the first 30m of sandstone: the sandstones at the contact are totally disrupted and deformed; the marlstones are faulted and veined.

Interpretation.

The St. Benoit fault was active during deposition of the Calcaires Nummulitiques and Marnes Bleues (section 4.5.3). Cremer (1983) proposed that a linear topographic low remained above the hanging wall (the south-eastern fault block) at the end of Marnes Bleues deposition: the first turbidity currents eroded a channel along the depression and deposited thick GdAIII sandstone facies.

A similar model is shown in enclosure 5.8, in which the thick sediment body need not be channelised. The base of the body is not exposed and the facies are not diagnostic of a channel fill. The turbidity currents flowed into the low adjacent to the fault: the shape of the topographic depression alone determined the geometry of the sandstone body in its hanging wall. The St. Benoit fault is a splay from the Rouaine-Daluis strike slip fault: it is probable that the fault displacement diminishes rapidly to the north and with it, the thickness of the sandstone body.

In this model, the turbidites to the west of the St. Benoit fault onlap its footwall. The wedge of Marnes Bleues is interpreted to be the rear portion of a small sedimentary slide that was emplaced westwards from higher on the footwall of the fault. The amount of slope required to explain both the onlap and the Marnes Bleues slide affects too small an area to be accounted for by footwall uplift on a single fault (Jackson and McKenzie, 1983). A synthetic fault in the footwall of the St. Benoit fault, to the west of the unconformity would account for the small scale additional rotation. The predicted fault cannot be traced because the outcrop is obscured by the municipal rubbish tip!

5.5.5.3. Sandstone facies distribution adjacent to the slope.

The response of the sediment gravity flows is characterised by: divergent erosive lineations; both gradual and abrupt bed thickness changes; deposition of sandstone up the palaeoslope, bed amalgamation; soft sediment deformation (see log. A37; encl. 5.8).

Interpretation.

There are some features of the palaeocurrent data that require highlighting:

1. Divergent erosive lineations on the base of any one bed. This phenomenon has not been observed at other onlap localities and probably relates to causes in addition to the immediate palaeoslope. Sediment gravity flows arrriving from a southerly south entered the Annot area over the major, steep Rouaine-Daluis fault. At this time, the Rouaine area was uplifted with respect to Annot, so that the currents flowed down the fault plane, hit the break in slope and expanded in the manner shown in the model, enclosure 5.8. The basal grooves and striations record the divergence

of the lobes and clefts at the erosive head of the current away from the base of the slope.

2. The crosslamination data are very limited. The readings imply that the currents maintained their northward flow apparently up the regional slope. However, the readings all come from the lower part of the regional onlap where the currents may still have been influenced by the opposing minor slope. In addition, the palaeoslope of the hanging wall roll-over was gentle. The momentum gained by the currents as they flowed down the steep slope enabled them to continue northwards. (This statament will be qualified and discussed in the next section).

3. *The* southward *directed* flute casts. These are extremely important, but they relate to area-scale processes, and their significance will be discussed in section 5.6.1.

5.5.6. Principles to be drawn from the preceeding examples of the relationship between the palaeoslopes and turbidite deposition.

The orientation and angle of the palaeoslope with respect to the original flow direction were the primary controls on the behaviour of sediment gravity flows in their proximity, dictating changes in their turbulent energy and flow directions. The sedimentary processes were modified, leading to bed amalgamation, deposition of sand part way up the slope and complex patterns of sediment failure. The sediment volume and the energy of individual currents were secondary controls.

Where the currents flowed along or obliquely down the palaeoslope

(Chalufy), the geometry of the sandstone bodies was determined by the precise three-dimensional geometry of the sediment surface at any time: the combined effect of palaeoslopes and differential compaction.

5.6. The nature of the Gres d'Annot basin floor topography.

5.6.1. The Annot outcrop area: a case history of the restriction of sediment gravity flows and the ponding of sediment. Enclosure 5.8, 5.7

5.6.1.1. Introduction.

The Gres d'Annot at Annot outcrop in a structural low bounded by SW and S vergent anticlines and the Rouaine-Daluis strike slip fault zone. The areal extent of the exposure is 10km * 6km, elongated NNW/SSE (Encl. 5.8).

The sandstone bodies exposed in the cliffs of the Chambre du Roi above Annot are spectacular, famous for their thickness and the massive appearance of the sandstone facies within them.

Previous research.

Stanley (1961) studied the sedimentology of the sandstones at Annot in some depth. Based on the results of his thesis from Annot and the outcrop areas to the north, he applied a submarine fan model to the Gres d'Annot Formation (Stanley, 1975). He proposed that the currents flowed from the south into the open, deep marine basin via major canyons, one of which was sited at Annot, and that the northern outcrops were the deposits of the fan itself (fig. 5.32).

The area has been re-assessed by the IFP (course notes of an IFP field guide, 1984) and briefly by Cremer (1983), who both interpreted the sediment body geometries and facies at Annot as the fill of a sizeable channel. Cremer placed the axis of the channel to the east of the present structural low, in the hanging wall of the St.Benoit fault.

The research approach of the present author.

The submarine canyon and fan models proposed by Stanley and Cremer can be tested by integrating stratigraphic, sedimentological and structural data from the Gres d'Annot and, more importantly, the Tertiary and Mesozoic sediments beneath. The nature of the angular, but non-erosive, unconformity at the base of the Gres d'Annot Formation was assessed. The origin of this relief was evaluated using structural data. The relationship between **large scale** basin floor topography and the deposition of the Gres d'Annot was explored.

In the following sections all the data will be presented with the aim of establishing regional and local palaeoslopes. The results will be summed in section 5.6.1.5. to construct a comprehensive model for the Annot outcrop area. Its context in the regional story of basin evolution will be considered in section 6.6.5 and 6.

5.6.1.1. Tertiary stratigraphy and evidence for the initiation of Gres d'Annot deposition in different parts of the Annot outcrop area.

The sediment facies and thickness of the Calcaires Nummulitiques and Marnes Bleues Formations vary across the Annot outcrop area. They are thicker and of deeper water facies to the south (section 4.4.4.). For example, while shallow marine facies were being deposited at Mort de l'Homme, deep water marls were deposited at Scafferels (Besson et al, 1970).

Mougin (1978) made a detailed comparative study of the microfauna of the Marnes Bleues in the Annot outcrop area. Her sections demonstrate that the first Gres d'Annot at Braux are significantly older than those at Le Ruch (fig. 5. 34). The palaeontological data of Besson et al (1978) and Mougin show that the first sandstones in the south and east lie upon older marlstones than those at Le Ruch.

Interpretation.

The distribution of the transgressive facies strongly implies that a south-dipping palaeoslope existed prior to the deposition of the Gres d'Annot (fig. 4.31).

Mougin's data can be interpreted in two ways:

i) to imply that the Marnes Bleues stratigraphy was more severely eroded in the south than in the north. However, there is no evidence for major erosion at the base of the Gres d'Annot Formation: to the contrary, the sandstones onlap a bedding parallel surface at the top of the Marnes

Bleues Formation.

i) the first sandstones were deposited in the south. The implication of this evidence, together with the northward onlap recorded at Le Ruch, is that a topographic slope from north to south persisted through the Gres d'Annot depositional history. This model~~led~~ is supported by the sediment facies within the Gres d'Annot (see below).

5.6.1.3. The Palaeoslopes interpreted from a structural analysis of the Gres d'Annot onlap relationships.

Palaeocurrents indicate that the source of the Gres d'Annot sediment was to the south. Fifteen or so kilometres south-east of Annot, the sandstone facies at St.Antonin were deposited near the contemporaneous shoreline. Therefore, the regional topography rose to the south of Annot. In Annot itself, the large scale, but gentle unconformity mapped through Braux to the village of Annot dipped to the **south-east, then east** (section 5.5.5.). This slope is interpreted to be the hanging wall roll-over to the Rouaine fault, which then, as now, had a component of downthrow to the NW.
plus onlap onto culmination that lay between the Argens-Allons and Annot syncline.

The basin floor slope south-east of Scaffarels therefore was the steep surface of the Rouaine fault, rising sharply towards the south-east onto the footwall of the fault. The regional rise to the south therefore occurred at the margin of the Annot outcrop area, not within it. confusing! Reword! GMA

Partially restored cross-sections through the region (fig. 2.21) imply that some of the shortening on folds and thrusts between Argens and Annot predated the deposition of the Gres d'Annot. Therefore, a NW-SE palaeo-high existed to the west of Annot, giving rise to the easterly component of dip on the large scale unconformity at Annot.

All the exposed unconformities in the Annot outcrop area have been described and interpreted in the previous section (5.5). A large SW vergent fold (the Melina kink zone) is interpreted to be responsible for the the Le Ruch onlap can be traced in outcrop more or less continuously from Le Ruch to the Rouaine fault. It is continuous, large-scale feature all the way to its intersection with the Daluis fault. Unfortunately, the Annot outcrop area has been tilted and eroded, so the relationship between the Gres d'Annot the fold is only seen at Le Ruch. If the extrapolation is made, the Gres d'Annot onlapped to the north-east,then east all the way around the eastern margin of the Annot sub-basin.

The Annot outcrop area was surrounded by topographic slopes onto structural highs, which developed prior to, and during, the deposition of the Gres d'Annot Formation. As a result, the deepest sea lay immediately to the south of Annot, against the Rouaine-Daluis fault. The basin floor topography within the area was complex, but a component of dip down towards the south existed across the length of the outcrop area.

Nowhere, within the outcrop area itself, is there any evidence of a medium to large scale northward slope.

5.6.1.4. The facies associations of sediment bodies in the Annot outcrop area.

Chambre du Roi:

The Chambre du Roi comprises thick sediment bodies of GdAIII, divided by thin sequences of GdAI. Each unit of GdAIII is an amalgamated unit (section 5.3.4.2) up to 50m thick. The impressive sandstone bodies of Les Gastres, north of La Chambre du Roi are illustrated in encl. 5.8 and fig.5.13b). They can be traced across a few kilometres of outcrop with no apparent changes in thickness. The thinly bedded facies between them are correspondingly extensive. The base of each unit is planar, often running parallel to a single thin sandstone over the traceable outcrop. No disruption of underlying sediment, even shales, is observed.

The sandstones theselves are remarkably monotonous: massive, moderately sorted coarse - medium sandstone. Bed boundaries are sometimes defined by pebbly sandstone or granules, but more often than not, tens of metres of massive and structureless sandstone are preserved with little more than evidence of coarse tail grading in places. Structures indicative of the sediment transport processes are rarely seen. Instead, there are dish structures and curvilinear surfaces which record fluidisation and dewatering of the sediment soon after deposition. The principal features of the sediment facies are illustrated in enclosure .

It will be shown below that the sediment facies provide indirect evidence of palaeoslopes within the basin which strongly influenced depositional and post-depositional processes.

1. Where sharp grain size breaks define individual beds, the bedding is decimetre scale bedding. The beds are dominated by massive or graded (coarse tail graded) coarse to medium sandstone, with pebbly or granular sandstone lagging an eroded and loaded lower bed contact. Layered coarse sandstone is sometimes seen at the base of a bed, with rare inverse grading near the base. Other sedimentary structures are not common, but an example of excellent parallel lamination is shown in encl. 5.8.

Interpretation

The deposits of individual flows are not large. the flows which deposited these beds were turbidity currents, most of which were probably high density flows. This evidence may be extrapolated to some of the thicker, homogeneous sandstones described below. The individual flow deposits were subsequently amalgamated, post deposition.

2. Sandstone layers several metres to >10 metres thick may show no discernable grain size breaks or even crude grading. No sedimentary structures exist within these sandstones, save poorly defined or weathered surfaces that may be water escape routes (see below).

Interpretation

This facies records very little of the processes by which sediment was brought to Annot or deposited there. The homogenised appearance of the sandstone is interpreted to be the result of post-depositional processes: almost certainly the result of fluidising a large volume of recently deposited, water-rich, unconsolidated sand. This may have occurred when a large turbidity current thundered over a large mass of recently and rapidly deposited sediment, or more probably occurred in response to failure of a large volume (comprising the deposits of many flows) of sandstone failed downslope.

3. Dish structures are quite common. In many cases the dish structures are aligned along planes that are oblique to the horizontal bedding surface (fig. 5.15). These surfaces dip at various angles with respect to bedding, towards the west. The water escape pipes are difficult to see in the coarse grained sediment, but they appear to be vertical. The dish structures occur throughout metre scale beds and are themselves centimetre scale, increasing in wavelength downwards.

In the 10m+ thick, massive sandstones, dewatering structures are difficult to identify with certainty. More often than not, discrete partings can be seen, but the concentration of fines predicted to develop along dewatering routes cannot be identified. One such case outcrops to the south of Le Ruch (encl. 5.4, 5.5). The partings form slightly concave upward surfaces, continuous over 20cm - 1m, spaced at intervals of 10-30cm throughout 6m of massive coarse sandstone. The surfaces all dip relative to bedding towards the west.

Interpretation.

The large scale bed partings observed south of Le Ruch are interpreted as dewatering structures. In both this case and the examples of dipping dish structures, the dip of the structures with respect to bedding may be interpreted in two ways: the sediment dewatered parallel to a sedimentary fabric in the bed (cross-bedding); or they formed parallel to planes of weakness when rapidly deposited sandstone failed downslope while still unconsolidated. The former theory is discounted. There is no evidence for cross-bedding on the scale of these structures comprising 90% of one bed. There are no grain size variations at a millimetre-centimetre scale.

Sand box experiments (eg. McClay and Ellis, 1987), used to model the structural geometries developed in unconsolidated sand, show that loose, dry sand fails along discrete planes. Wet sand has a greater cohesion and is therefore stronger, but may be expected to respond to deformation in a similar manner. It would be necessary to inhibit the tendency of the grains to move throughout the sediment: rapid failure of sediment down a slope may induce liquifaction and fluidisation which would homogenise the sediment (this process probably accounts for the enormous thicknesses of structureless sandstone in the area).

However, if sliding took place over some length of time, discrete shear planes may form to accommodate bed length extension. At the localities described, the dipping dishes and the slightly concave-upwards partings are interpreted to be oriented parallel to such planes. The zones of weakness were exploited as fluid escape routes. As the fluid escaped, the sand grains packed closer and the sediment froze, thus preserving the structure.

The planes dip to the west. This will be returned to in the overall interpretation.

4. Large, 30m high recumbent folds are seen at Les Gastres (encl. 5.8). The folds close predominantly to the south, with a single asymmetric fold closing to the north. In the same horizon, but further north along Les Gatres, discrete planes cut from top to bottom of the horizon, dipping to the south.

Interpretation.

The planes represent extensional shear planes towards the rear of a slid mass. Their orientation in this section and the dominance of southward fold closures imply that a significant amount of the sediment failed southwards (encl. 5.8). The two-dimensional section of exposure does not allow the direction to be precisely defined, but the polarity is clear!

5. In addition to this large scale disrupted horizon, other small-scale soft sediment folds are observed at the Chambre du Roi. A pebbly sandstone lag at the base of a very thick, massive sandstone displays load structures which are folded: convolute laminated sandstone forms the sharp crested anticlines. When viewed at right angles the folds are seen to be sheath folds: *the fold eyes are seen on the south face of the cliff.* (ie. folds close to the south)

Interpretation.

The sense of shear was equivalent to moving the upper bed to the south or south-west. The sheath folds may have developed during the deposition of the sandstone, in which case the sediment arrived from the north. Instead, the folds are interpreted to have formed as a result of the sandstone shifting **southwards** soon after deposition. The time lapse between sedimentation and failure, and the distance the sediments were displaced, cannot be constrained.

Stanley's slump data (1961) also imply that a significant amount of sediment failed to the south and south-west.

6. Le Ruch and Braux: the outcrop localities of an angular, non-erosive unconformity at the base of the Gres d'Annot Formation.

The facies associations of the onlapping sandstones have been discussed in some detail in the preceeding sections (5.5.3. and 5.5.5.). At Le Ruch, metre scale bedding is observed, but compared with the facies of the Chambre du Roi, the beds at Le Ruch are less often amalgamated and primary sedimentary structures are common.

It was shown that, at Le Ruch, deposition from a variety of sediment gravity flows was controlled by a SW-dipping palaeoslope that opposed the arriving turbidity currents. At Le Ruch, there was a marked discordance between the basal and internal palaeocurrent indicators; complex bed forms and large sediment bodies are interpreted to be the product of sediment reworking by reflected currents and combined flows.

At Braux, the SE and E-dipping palaeoslope exerted less influence on the currents flowing up slope, perhaps because the currents were at their most energetic, having only recently shot down the steep fault scarp into the area. Soft sediment deformation, chaotic facies and thick nests of shale clasts characterise the sandstone facies: the currents were erosive and deposited sediment rapidly, which dewatered and deformed soon after deposition.

Sediment failure on a grander scale must account for the flute casts at Braux that record south flowing turbidity currents. Rapid sedimentation on steep slopes created unstable slopes. Triggered perhaps by small seismic shocks, sediment gravity flows were generated. They flowed down slope to Braux and deposited their load at the base of slope (encl. 5.7).

5.6.1.5. THE ANNOT MODEL

The model is summarised in encl. 5.8. The controlling elements are tectonic structures, most of which immediately pre-date Gres d'Annot deposition, but which were largely inactive during deposition of the turbidites. Complex basin floor topography developed at all scales in response to displacement on the faults and shortening accommodated by folds (and thrusts, not exposed in this outcrop area). Some of these slopes were unstable and collapsed generating additional small scale relief. Across the whole area a significant component of the surface dip was towards the south.

The Annot area was enclosed to the west by a N-S ridge, produced by the Argens thrust and to the NE by a ridge developed above the Aurent kink zone and the Permian thrust sheet (BII). The palaeohighs were submarine throughout the deposition of the Gres d'Annot, but the Annot area was. ~~became~~ ~~was~~ essentially an isolated sub-basin. Sediment gravity flows arriving from the south entered the confined trough on the sea floor via the steep escarpment of the Roauine fault. The currents, particularly the early ones (the deposits of which are all that remain at Annot), encountered a significant slope that opposed their flow direction and were also unable to spread laterally.

Confined in all directions the currents lost energy and dumped a large percentage of its load near the base of the slope. The rapid deposition trapped water in the sediments which escaped soon after. This gave rise to dewatering structures in some cases, but in most, it simply homogenised the already massive sandstone. These modified sediments constitute the puzzling, spectacular facies of the massive sandstone bodies at the Chambre du Roi.

Some of the sediment may have been deposited on steep slopes and failed shortly after deposition, thus triggering a local, within-basin turbidity current. The effects of this are seen at Braux, exemplified by the southward directed palaeocurrents (encl. 5.7).

The major slumping that produced the large scale disrupted horizon at Les Gastres (fig. 5.8) is ~~has been~~ related to activity on structures at the eastern margin of the basin. Shortening accommodated by the Barrot structure and associated high level folds increased the westward component of dip on the basin floor. Large volumes of sediment collapsed. The sloping dishes and other dewatering structures are interpreted to reflect this dip: the sediment collapsed down towards the west. The syn-sedimentary tectonic activity caused the depocentre to migrate westwards through time.

The majority of currents probably surmounted the topographic high to the north, taking the fine fraction of their load in suspension. However, at Le Ruch, there is strong evidence to suggest that part of any current was deflected by the steep slope: the return flow was capable of reworking sediment or maintaining some of its original load. The complexity of bed forms that arose are interpreted to be the product of internal waves and combined flows, formed when the deflected current interacted with the rest of the arriving turbidity current.

5.6.1.6. How this model differs from previous interpretations of the Annot sandstones.

Stanley (1975) interpreted Annot as the site of one feeder submarine canyon (fig. 5.32). He compared the outcrop geometry with the shoe-string geometry of modern submarine valley fills. Large scale slumps observed at Les Gastres and a massive sandstone facies at the Chambre du Roi were interpreted to be the product of sediment failure: such processes dominate the submarine canyon environment. The contact between the Gres d'Annot and Marnes Bleues was interpreted as a major erosion surface.

There are certain inconsistencies in Stanley's interpretation and those of Cremer (1983) and the IFP (1984, field guide to the Annot sandstones):

The base of the Gres d'Annot Formation is not a *major erosion surface. In* his thesis (1961), Stanley realised that the sandstones onlap a folded Marnes Bleues surface, but he chose to disregard this in 1975. The outcrop he uses as evidence for major erosion is the outcrop at Les Gastres that is interpreted in this thesis as a *small scale onlap* of sandstone onto a bedding parallel surface in the Marnes Bleues (section 5.5.4.).

It does not solve the dilemma to make Annot the site of an inner fan channel. As Stanley noted, the massive sandstones form a ribbon body of sorts, surrounded on all sides by Marnes Bleues. The thick GdAIII sediment bodies are nowhere incised into thinner bedded sandstone facies.

Cremer does not present better evidence for erosion along the axis of his interpreted channel. Nor does he attempt to reconstruct the three-dimensional geometry of the channelised sandstone body.

All the models proposed to date predict a basin floor topography that slopes down from south to north, from Annot to Le Ruch and beyond. The slope is deduced by inference: the source of the sediment was to the south of all the outcrop areas; the facies associations at Annot, Contes and Menton are interpreted to be more proximal than those of the northern outcrop areas. Therefore, the basin floor topography sloped from the south down to the north **everywhere.** This logic is flawed: Chalufy is north of Annot, but the palaeocurrents come from the east. Why did no sediment come due north direct from Annot?

5.6.1.7. How the model differs from examples of ponded turbidites interpreted by other authors.

1. There are no thick mudstone capping the turbidites.

Ricci Lucchi and Valmori (1980) indentified tubidite/mudstone couplets in Miocene turbidites in the Apennines. They interpreted the mud to be deposited from the cloud of suspended fines in the upper and rear portions of a turbidity current that had been totally ponded by the confining basin-floor topography.

At Annot, no such mudstone caps exist. The presence of shale clasts demonstrates that argillaceous material was available. In the proposed model, the turbidity current can surmount the topography to the north: the slope is only locally high angle and the opposing slope is only a few kilometres long. The turbidity currents may have rapidly dumped its coarse load near the base of the slope, and then flowed over the hill, carrying with it a large percentage of its fine sediment load.

2. There is no evidence for reflected turbidity currents.

Pickering and Hiscott's model for repeatedly reflected turbidity currents (1984) cannot be applied here.

Return flows were generated, but not turbidity currents. I have predicted that the flows were not completely constrained by the palaeoslopes, which may have limited the possibility of re-creating a turbidity current. However, I doubt the process occurs as these authors have described it (fig. 5.34). The interactions of the arriving and returning flows will be extremely complex. This has been demonstrated at Annot.

3. The model presented is akin to the onlap model of Scott and Tillman (1981; fig. 5.35). They interpreted the Stevens Sandstone in the San Joaquin basin, California, as the fill of a palaeotrough defined by a syncline.

5.6.2. Col de la Cayolle and Trois Eveches.

Both Jean (1985) and Ghibaudo (1985) described in detail numerous sedimentary logs drawn through the complete Gres d'Annot Formation. Logs are correlated with confidence using the marker conglomerate beds (section 5.7.1.3.). Thickness variations demonstrate that thicker sequences occur along the central axis of this outcrop area (fig. 5.36), suggesting that the Gres d'Annot were deposited in a valley on the sea floor, elongated NNW-SSE. The base of the Gres d'Annot is a tectonic contact in many places, which these authors did not always recognise, so their interpreted thickness variations should be treated with some caution.

Post-Gres d'Annot deformation has destroyed many of the original stratigraphic contacts between the Gres d'Annot and the Marnes Bleues (cross section X). An exception to this is exposed on the east face of Montagne de Bertrand. The Gres d'Annot onlap the Marnes Bleues towards the south, at an angle of ca.10', over a distance of 1km. The unconformity is interpreted to relate to the Barrot structural high, BII, the same structure that bounded the eastern margin of the Annot sub-basin.

Elsewhere, it is impossible to precisely relate the sedimentary fill to the original basin floor topography.

Both Ghibaudo and Jean propose that the deposits infilled a pre-existing topographic low that did not develop further during deposition of the turbidites. With time, successive sediment gravity flows were less confined. The depositional system spread into adjacent areas, specifically Trois Eveches. According to Ghibaudo, the extensive debris conglomerates were deposited across the two outcrop areas during single events.

Therefore, there were no complete topographic barriers to flow between these areas.

5.6.3. Grand Coyer.

Cremer (1983) proposed, without supporting evidence, that the coarsest and thickest sandstone bodies were concentrated towards the centre of an elongate trough oriented parallel to the outcrop itself, ie. NW-SE. Thickness variations in the Marnes Bleues formation show that the depression formed prior to the input of siliciclastic material. The relief did not develop further while the turbidites infilled the relative low.

He does not explore the possible origin of the palaeotopography. Alpine structures were probably responsible for the trough (fig. 2.21). Suffice to say, the strike of the predicted palaeoslopes aligns with the trend of SW-vergent Alpine structures.

5.6.4. Peira Cava.

Bouma and Coleman (1985) summed up their recent paper on the Peira Cava turbidites by stating that the outcrop area was a small, sub-basin, elongated north-south, that closed to the north against the Argentera Massif. They present no evidence for this hypothesis, but Bouma (course notes for Chevron, 1985) sites a concentration of slumped sediments near the margins of the outcrop area.

The model is an appealing one in some respects, but it is highly unlikely that the sub-basin closed to the north against Argentera. This basement culmination is interpreted as a late Alpine feature because it deforms all the Tertiary sediments and the stuctures that pass above it.

5.6.5. Concluding comments on the nature of the Gres d'Annot basin floor topography in areas other than Annot.

The data presented by the previous authors is not comprehensive and the hypotheses have not been tested against independent evidence from other parts of the Tertiary sequence, for example. None of them attempt to explain the origin of the palaeotopography.

However, their conclusions are consistent with the model proposed in this thesis for the Annot outcrop area and individual onlap contacts elsewhere in the basin. It would appear that there was a distinct pattern of palaeotopography within the Gres d'Annot basin. In all cases, it is possible to relate the palaeo-highs and relative lows to the geometry of the underlying tectonic structures.

The "proximal-distal" changes in facies associations observed from Annot, Contes and Menton to Peira Cava, Col de la Cayolle and Trois Eveches can also be interpreted in terms of the observed structural controls. The degree of topographic constraint increased southwards, caused by the increased structural complexity. The greater confinement in the south had a great effect on the sediment gravity flows, for example, at Annot, causing them to dump the coarse fraction of their load in the palaeotroughs.

5.7. **Features of the Gres d'Annot basin important to understanding of basin dynamics at this stage in its evolution.**

5.7.1. Provenance of the Gres d'Annot Formation.

There are two questions posed in any discussion of the provenance of the Gres d'Annot Formation. Firstly, was the sediment primarily reworked from ancient clastic sequences or derived directly from crystalline basement? Secondly, is it possible to define the position of the source areas with respect to the Gres d'Annot? In particular, was some of the sediment derived from the advancing Alpine orogenic wedge?

5.7.1.1 Introduction.

The possible sources of siliciclastic detritus for the Gres d'Annot Formation have been investigated a number of ways, using comparative petrographic studies and palaeocurrent analyses. The principal results of these studies are summarised in the following sections.

Interested readers are referred to the original theses of Stanley (1961), Ivaldi (1973) and Jean (1985) which contain much of the raw data used in this summary. They have defined the total mineralogy of the sandstones, exhaustively analysed grain size measurements and approached the problem of sandstone provenance using parameters like quartz thermoluminescence (Ivaldi, 1974) and zircon typology (Jean, 1985). The petrographic studies are excellent in their details of the mineral textures, which indicate the origin of detrital minerals like feldspar.

Jean (1985 a,b) analysed Gres d'Annot samples from the Col de la Cayolle and North Argentera outcrop areas. She has identified the nature of the GdA source areas and provided reasonable, although largely untestable, palaeogeographic reconstructions.

There is much raw data from other Gres d'Annot outcrop areas which could be re-assessed, following her example and the approach of recent authors like Dickinson and Valloni (1980), Valloni and Mezzadri (1984), Maynard (1984) and Suczeck and Ingersoll (1985). However, it is doubtful that the provenance of the Gres d'Annot Formation could be better defined.

5.7.1.2. The petrography of the sandstones and siltstone.

The total mineralogy of the GdA framework grains, matrix and cement was discussed as early as 1947, by Tercier, as one example of Alpine "flysch". Gubler (1958) and subsequently Stanley (1961) compiled data from all the GdA exposures. Jean (1985) added to the dataset, using samples from the Col de la Cayolle and North Argentera outcrop areas.

80% of the sandstones can be termed arkoses (>25% feldspar); the remainder have >5% feldspar and are therefore feldspathic sandstones. The principal framework mineral is quartz; also important are biotite and muscovite. The proportion of lithic grains is small (fig. 5.37). The sandstones have little to no matrix, but do have variable amounts of a carbonate and phyllosilicate cement. The argillaceous matrix becomes significant in the siltstones.

Stanley (1961, 1968) showed that the mineralogy of the Gres d'Annot Formation did not exhibit a consistant change from base to top of a section. In fact, he demonstrated that the mineralogy was more closely related to the grain size distribution of the sample. The variation in mineralogy of a single fining-upward turbidite is as great as that observed throughout the formation.

Quartz and Feldspar.

The grains are invaribly angular-(rounded), being more angular in the small grain size fraction.

The percentages of quartz and feldspar do not vary much within the fine to very coarse sand grade material: 43-61% (av.50%) quartz, 30-50% (av.40%) feldspar. The precise proportions do vary somewhat with the grain size of the sample: thus sandstones from southern outcrop areas have somewhat higher percentage of feldspar, which reflects nothing more than the coarser average grain size of these sediments.

The quartz grains have been the subject of a thermoluminescence study by Ivaldi (1974). The natural luminescence provided the most characteristic spectra, with two peaks of luminescence. The relative intensity of these peaks varied between samples. Quartz grains from one outcrop area displayed similar spectra, but those from adjacent areas were quite different. Outcrop areas were grouped using this parameter alone, and Ivaldi defined three zones in the study region (fig. 5.38), each with a distinct quartz grain assemblage.

The thermoluminescence spectra were more or less successfully matched with those of presently exposed basement rock types. Quartz grains from the southwestern outcrop areas compared favourably with quartz from the crystalline basement and Permo-Triassic of the external basement massifs.

Those from the Gres d'Annot in the north and east had spectra which ressembled detrital quartz in the Helminthoid flysch.

These results can be interpreted a number of ways. Ivaldi (1974) concluded that a significant proportion of the Gres d'Annot sediment was cannibalised from the older Helminthoid flysch successions. This hypothesis is not favoured by Jean (1985) or myself. For several reasons, it is unlikely that the Helminthoid flysch itself was reworked into the Gres d'Annot basin. For example, the Helminthoid flysch contains significant amounts of carbonate detritus; the Gres d'Annot contain almost none. Therefore, it it is more likely that the quartz grains in the two formations were derived from the same or similar source areas!

All samples of Gres d'Annot sandstone contain a high percentage of feldspar; the grains are large and angular. Plagiclase and alkali feldsar are present in approximately equal proportions (relatively more plagiclase in siltstones). The plagiclase is always rich in sodium: albite and oligoclase. The potassium feldspars are orthoclase (sometimes with microperthitic texture), sanidine and microcline (locally very abundant).

The evidence suggests that the feldspar is 1st order (Jean, 1985), not derived from a sedimentary source area. This is contrary to the conclusions of Gubler (1958), Stanley (1961) and Ivaldi (1974), who were of the opinion that most of the sediment was reworked from exposed Permian sandstones and older Alpine flysch, like the Helminthoid flysch.

Phyllosilicates.

Biotite and muscovite are present in significant, moderate quantities, feldspar sand grains that comprise the bulk of the Gres d'Annot.

It is probable that the individual quartz, feldspar and mica Gres d'Annot sandstone grains originated in igneous and metamorphic rocks of the types represented in the lithic grains.

Heavy Minerals.

The assemblages have been studied by Gubler (1958) and Stanley (1961). The zircon typology has been defined by Jean (1985).

The range of minerals represented is quite limited. The populations vary across the study region. In the south, the association is: staurolite, kyanite and garnet. Further north, the dominant minerals are: zircon, rutile, apatite, toumaline and garnet (Jean's analyses did not reveal garnet). North of the Argentera massif, garnet is entirely absent.

Jean analysed the zircon typology of sandstones in the Col de la Cayolle outcrop area. She identified two distinct zircon forms, the first derived from calc-alkaline volcanics, the second from migmatites.

The heavy mineral populations are interesting: the low species diversity implies a maturity of sediment that is not reflected in the light mineral types and texture. Stanley (1965) attributed this paradox to the sediment source, suggesting that the primary source of Gres d'Annot sediment was the Permo-Triassic cover to Maures-Esterel and Argentera massifs.

However, he did not present analyses of the heavy mineral populations in the crystalline basement of these massifs. Without evidence to the contrary, it is possible that the basement rocks have an equally limited population of heavy minerals. The heavy mineral assemblages are not representative of any rock type and cannot be used to finger print the Gres d'Annot sandstones and a possible source area.

The zircon analyses were useful. Both types compared well with those from Hercynian crystalline rocks of Corsica and Sardinia. Moreover, Eocene sandstones from St.Antonin and Permo-Triassic formations exposed in Provence exhibited the same bimodality of zircon types. A major source of sand grade detritus was therefore the crystalline basement of Corsica-Sardinia and/or similar external basement massifs.

Organic Detritus.

In addition to the minerals above, the sandstones are often rich in organic debris. Sizeable fragments are observed, several 10s of centimetres long, in which the original plant structures are well defined (for example, at Le Ruch: fig. 5.12). Some thin beds, a few centimetres thick, comprise >80% of plant remains (for example, at Denjuan and Calufy).

Plant tissue quickly degenerates after a plant has died, if it remains in a well oxygenated environment. Therefore, the presence of well preserved plant debris in sediment is significant. Either the plant debris accumulated in an oxygen-deficient environment, like a swamp; or the sediment was rapidly deposited and buried at its final resting place.

The fauna present in the Tertiary sediments are all indicative of deposition in a "normal" marine basin. Therefore, the organic detritus was not reworked for any length of time in the oxygenated waters of a shallow marine environment. Instead, it was swiftly transferred from the river mouth off the shelf. It can be concluded that rates of sedimentation on the basin floor were high.

Carbonate Minerals.

Calcium carbonate is present in the sandstones only as cement and rare bioclastic fragments. However, Stanley (1961) found that some of the siltstones contained significant percentages of detrital calcium carbonate (10-18%). These samples came from the base of the Gres d'Annot Formation at Lac d'Allos.

At any time in the Upper Eocene, shelf limestones were deposited to the west of the site of Gres d'Annot deposition. Neither the Gres d'Annot sandstones nor the siltstones contain significant volumes of carbonate detritus. Therefore, the carbonate shorelines did not supply sediment to the deep marine environment.

It follows that the shorelines adjacent to the source areas were not sites of carbonate accumulation. The basin was not large enough that different climate or water temperatures were responsible for lower carbonate productivity. It is most likely that siliciclastic sediment was regularly supplied to these shores in sufficient quantities to inhibit the formation of reefs. Thus, life forms were restricted, and their remains in sediment were heavily outnumbered by quartz grains!

Minerals and Lithics that are conspicuous by their absence.

No sample of Gres d'Annot contains fragments of serpentine, spilite or basic igneous rock. Augite, horneblende and sodic amphiboles, like glaucophane, are never observed.

The sources of Gres d'Annot sediment did not include ophiolite rocks (s.l.), or metamorphic rocks with high pressure/low temperature mineral assemblages. The Gres d'Annot were therefore not sourced from an evolved orogenic wedge. This is discussed in section 6.4.

5.7.1.3. The petrography of the conglomerate marker beds.

Their petrography and sediment body geometry distinguish them from other conglomerate facies in the Gres d'Annot Formation.

The clasts are all well rounded and elongate; they vary in size, but most are less than 10cm long. Their morphology leads Jean (1985) to the conclusion that the clasts were transported in steep gradient rivers through distances no greater than 100km, before being fed directly into submarine canyons. They show no evidence of reworking in a shallow marine environment.

Jean studied in detail the petrography of the two horizons (< = 10m thick) in the Col de la Cayolle outcrop area and similar conglomerate sheets that outcrop to the north of the Argentera massif. She observed no difference between them and concluded that they were at least sourced from the same terrains and could be the same horizons.

The matrix comprises illite and montmorillonite, with no calcite. The sand grade material is similar to that of the remaining GdA, except that almost all the quartz grains are monocrystalline. The clasts comprise 70% acid plutonics, lavas and gneiss; 15% fragments of silicious rock, like vein quartz and flint; and 15% sedimentary rocks.

Seventeen populations of zircon were analysed by Jean (1985) which confirmed the calc alkaline chemistry of the clasts. The study revealed a distinction between the plutonic and volcanic samples, and the magma types were defined quite precisely. The zircon typology narrowed the field of possible sources to such an extent that Jean was able to reject all the present basement outcrops and the Permian conglomerates as possible source rocks! The zircons from the granite clasts matched well with those of a Hercynian granite exposed in the central Argentera. However, the present erosion level only just reveals this central granite, and the conglomerates contain no fragments of the extensive muscovite/garnet granite which surrounds it.

The mineralogy of the gneisses is simple and similar rocks outcrop in all the basement massifs. These clasts cannot be used to identify a source area.

The sedimentary clasts include calcite cemented sandstones, none of which contain glauconite; siltstones; and a variety of limestones, two of which dominate the carbonate samples and both of which are distinctive:

1. Nummulitic limestone of middle Eocene age (at the boundary between Lutetian and Bartonian) and of Biarritzian facies A (Campredon, 1977).
2. Mesozoic limestone, of Uppermost Jurassic and Lower Cretaceous age,

which is a platform carbonate facies (Urgonian, s.l.) characterised by its dark colouration, the result of organic material in the oolitic limestone matrix.

The sandstone clasts are unlike those of the GdA formation: some appear to be very angular fragments of Permian sandstones; the rest are similar to clasts found in the Schistes a Blocs Formation, of unknown derivation. The absence of glauconitic sandstone clasts is significant: glauconite is observed in the clasts of conglomerates exposed in the Annot and St.Antonin outcrop areas. Therefore, the conglomerate marker beds were derived from different source areas and were probably not transported via Annot.

The Nummulitic limestone clasts can be correlated well with the equivalent limestones that outcrop in the Italian Maritime Alps. Thus a south-easterly source area may be inferred.

The Mesozoic limestone facies does not outcrop in the Subalpine chain; the nearest outcrops are in Corbieres, in the Northern Pyrenees. Jean (1985) postulated that this particular facies extended east to Corsica and Sardinia. These areas were uplifted progressively and eroded; clasts of the organic-rich limestone were deposited at Ciotat, Marseilles, in the Upper Cretaceous and subsequently in the conglomerates at Col de la Cayolle in the Upper Eocene. This is an interesting hypothesis with implications for the palaeogeography of the region in the Late Mesozoic and Early Tertiary times (section 6.4).

The dark limestone clasts again highlight the differences between the conglomerate marker beds in Col de la Cayolle and the conglomerate facies in Annot and St.Antonin where similar age carbonate clasts are platform facies, but white. These could be derived from sources within the study region.

5.7.1.4. Palaeocurrent data as applied to provenance studies.

Fig. 5.1. displays the regional palaeocurrent data as collected by myself, Bouma (1962), Jean (1985) and Ghibaudo (1985). The significance of variations from the dominant palaeoflow direction is discussed briefly in section 6.4. In this section, the principal directions will be considered as evidence for the origin of the GdA sediment.

In the southern outcrop areas, the palaeocurrents flowed from the south and southeast. Towards the north, in the Col de la Cayolle and North Argentera outcrop areas, the flow direction was from the SE/SSE towards the NW. To the west, the palaeocurrents flowed to the W/NW. The readings are consistant along the length of the Trois Eveches outcrop area.

Interpretation.

Palaeocurrent readings derived from sediment gravity flow deposits define the palaeoslope at any point in the basin. They can be misleading if used as the sole evidence for the origin of the sediment, for two reasons:

1. Later deformation of the region may cause one part of crust to rotate relative to another, so that palaeocurrent readings no longer point back to the source area;

2. Changes in basin slope can cause a sediment gravity flow to turn

through 90' or more during it travels. This is very common where the sediment is derived from the margin of a elongate basin, like those developed at active continental margins. The turbidity current travels down to the base of the steep slope, then turns to flow along the axis of that basin.

In the case of the Gres d'Annot Formation, the palaeocurrent data corroborates the petrographic evidence. The petrography of the formation is constistant with the facies evidence, which suggests that the southerly outcrops could have been the proximal equivalents (not necessarily in time) of the successions to the north. The bulk of the Gres d'Annot were sourced from the similar terrains.

The petrography of the sandstones suggests that the primary sediment source rock was crystalline basement akin to those that outcrop on Corsica and Sardinia. The palaeocurrent data from the southerly, therefore proximal, outcrop areas implies that the sediment was derived from the south or southeast. Therefore, Corsica and Sardinia and their easterly extension (not now exposed) could have been the principal source of sediment to the Gres d'Annot basin.

5.7.1.5. Conclusions.

The mineralogy and grain fabric allow the observer to draw a tight net around the type of terrains which supplied the Gres d'Annot sediment. The primary source rock was **crystalline basement,** comprising granites, calc alkali volcanics and granulites of Hercynian age and external Alpine affinity. Corsica and Sardinia provide the closest analogue of the source area exposed today.

In detail, the evidence suggests that different parts of the basin received sediment from two different basement types. These terrains were probably adjacent massifs to the south and east of the Gres d'Annot basin which have since subsided or been tectonically buried.

The absence of large volumes of reworked sediment, the lack of ophiolite clasts (seen in younger clastic successions at Barreme) and basic volcanic detritus imply that the Alpine orogenic wedge did not contribute significant volumes of sediment to the Gres d'Annot basin. Two possible explanations will be considered in section 6.4:

1. The orogenic wedge had a long leading edge which was still below sea level. Sediment derived from the east was trapped in lows on top of its submarine snout.

2. The early stages of the collision caused part of the edge of the European continent to emerge in front of the Adriatic peninsula and form a barrier to sediment derived from the east. The palaeoslopes defined by the palaeocurrent directions imply this was the case. It is otherwise difficult to explain why the turbidity currents should flow westwards in the Trois Eveches outcrop area.

Jean (1985) concluded that the grain size and immaturity of the Gres d'Annot demonstrated that the sediment was 1st order and not reworked from older clastic successions. In fact, few of the presently exposed sedimentary rocks could have supplied sediment to the Gres d'Annot basin. The successions are, for the most part, too fine grained. Those that are coarse enough have mineral assemblages quite unlike those of the Gres d'Annot (eg. calcite in the Helminthoid flysch).

The conglomerate marker horizons have been shown by Jean (1985) to have a distinctive clast assemblage, comrising granites, volcanics and sedimentary rocks not seen elsewhere in the Gres d'Annot. Perhaps most significant are the sandstone clasts similar to those seen in the Schistes a Blocs. The crystalline rock clasts were probably sourced from basement akin to the Argentera massif (ie. external, European basement). Most of the sedimentary rock types also have affinity with the European cover sequence. However, the sandstones may originate from older flysch successions caught up in the Alpine orogenic wedge.

The conglomerate marker beds only outcrop in the north of the Gres d'Annot basin, which suggests that they were not sourced from the south. The uplifted basement terrain was probably to the east-southeast or southeast, close to European continental margin. This area provided very little sediment to the gres d'Annot basin. Likewise, the marker beds appear to be derived in part from the Alpine orogenic wedge, which did not contribute anything to the Gres d'Annot sandstones. It is possible that the marker beds were sourced from submarine slides, from terrain that was not exposed at surface. The submarine slope failure was probably induced by seismic shock, consistent with tectonic activity at the leading edge of the Alpine wedge.

5.7.2. Sequences: are they present in the Gres d'Annot Formation?

5.7.2.1. Historical context of Vertical Sequence Analysis.

Past
~~Recent~~ studies of turbidite successions have concentrated on the sequence of facies observed in a vertical section. Ricci Lucchi (1969) showed that vertical variations in grain size and bed thickness occurred in symmetric and asymmetric cycles, which he termed "megarhythms". Mutti and Ghibaudo (1972) and Mutti and Ricci Lucchi (1972) compared these rhythms with sedimentary facies cycles observed in deltaic sediments and thereby constructed the **submarine fan model**. Each vertical sequence represented a certain sub-environment on the fan (fig. 5.39). For example, thinning and fining upward sequences were interpreted as channel fills; 10 metre scale thickening and coarsening upward sequences were produced by lobe progradation; larger scale cycles recorded the progradation and abandonment of the fan.

The fan model was accepted and extended: vertical sequences were recognised and interpreted in turbidite successions around the world (for example: Hein, 1976; Rupke, 1977; Walker, 1978; Hiscott, 1980; Pickering, 1982, Shanmugam et al, 1985). Cycles were recognised in cores and used to construct hydrocarbon reservoir models for fields like Forties (Carmen and Young, 1981). Vertical sequence analysis of sedimentary facies has been particularly useful in this industry where detailed sedimentary data may only be derived from wells.

The technique of studying vertical sequences of sediment facies is popular, but is often mis-used. Many authors approach turbidite successions assuming that they represent submarine fans and that facies associations must represent laterally equivalent sedimentary environments. Their interpretations are often untestable. For example, the term "lobe" is used where no evidence is presented for the existance of a positive feature on the sea bed (eg. Cazzola, Mutti and Vigna, 1983/4). Equally, thinly bedded sandstone and shale alternations are interpreted as "levee deposits" without showing their lateral equivalence to incised channels (Nilsen and Abbate, 1985).

Sedimentological models are generally improved by comparative studies of equivalent modern and ancient environments. However, this has not been the case with deep sea clastics. The surface features of modern fans are mapped, and seismic surveys reveal the large scale internal structures (Normark, 1978 and numerous examples in "Turbidite Fans and Related Turbidite Systems", 1985); cores are more scarce and rarely penetrate the sediment deposited prior to the Holocene sea level rise. The present day morphology *of fans is misleading, in many cases. They have been blanketed* by post-Holocene hemi-pelagic sediment. Exceptions to this exist off the active strike-slip margin of the Californian coast. An excellent account of the morphology and sediments of the La Jolla fan was provided by Haner (1971).

Turbidite basins exposed at surface are often deformed, dissociated from their original substrate and largely eroded. It is difficult, if not impossible, to reconstruct the three dimensional form of the original basin, the sedimentary system and its constituent parts. There are exceptions, of course: Rupke (1977) defined the geometry of a turbidite fan from excellent exposure in the Cantabrian mountains, Spain. However, the good data in ancient successions often comprises vertical sections along which the sedimentary facies can be studied in detail. It is not always possible to correlate these sections! The data available from modern and ancient fans are therefore not directly comparable.

~~However~~, it has also become apparent in the literature that morphological features on submarine fans interpreted from the ancient sedimentary record are consistantly smaller than those mapped on modern submarine fans (see the papers presented at the Comfan meeting, 1982. These are collected in Geomarine Letters, 1983-4). The scale of most outcrops and the tectonic dissection of many turbidite successions makes it difficult, for example, to recognise channels the size of distributary channels on the Amazon fan (Damuth and Flood, 1985). Where lobes are concerned the situation is more complicated, because geologists do not agree amoungst themselves on the definition of the term: it includes small scale features which are the scale of lobes mapped on the Navy fan, but also encompasses tabular bodies the scale of the basin itself.

The existence of thickening and thinning upward sequences is questionable in many successions (Hiscott's discussion, 1981, of Ghibaudo, 1980). There is no consensus about what constitutes a "sequence": Heller and Dickinson (1985) show a thickening-upward sequence of <1m thickness, comprising 3 beds!

Where sequences do exist, their significance is debatable. Hiscott (1981) demonstrates that it is invalid to use delta lobe progradation as a comparative model for submarine fan lobe construction. Quite so! In my opinion, it is quite absurd to imply that a succession of sediment gravity flows would flow over an area where sand has already been deposited, ie. over a topographic high, when the unbounded flows can sweep into the adjacent areas. Hiscott and others favour aggradational lobe models; they (eg. Mutti and Sonnino, 1981) emphasise that successive turbidites are deposited in topographic lows, giving rise to complex "compensation cycles" which are laterally highly variable (fig. 5.40). This illustrates how the term "lobe" has lost its identity.

Elsewhere, it is apparent that many sequences are laterally very continuous, that no consistant differences in facies associations are noted between outcrop areas. Even these tabular sediment bodies are still referred to as "lobes" by some (Ghibaudo, 1981 and 1985). Sedimentologists are only now beginning to consider the alternative processes which may give rise to basin wide sediment facies cycles: for example, sea level changes (Mutti, 1984). It is also becoming clear that not all turbidite systems are or were fans! This point is developed in section 6.7.

Many sedimentologists (eg. Mutti and Normark, Pers.Comm., 1985) believe that it is time to go back to the drawing board. The problems should be approached in two ways:

-define elements of modern and ancient successions that are comparable, avoiding the use of morphological terms like lobes and levees and paying strict attention to the scale of each element.

-explore the processes which control the turbidite system and model their influence on all its elements. It is probable that the autocyclic controls of the fan models to date are not the most important. Vertical sequence analysis will continue to be an invaluable technique, but only when the results integrated with other data: for example, large scale sediment body mapping, petrographic studies and structural mapping.

5.7.2.2. Sequences and cycles observed at the 10-100 metre scale in the Grès d'Annot

In the sections studied no sequences less than 10m thick were recognised by inspection. It was therefore not worthwhile making a statistical analysis of the bed thickness or grain size variations.

Sequences are actively being sought by Ghibaudo in the northern outcrop areas of the Gres d'Annot. It has not been the purpose of this research, or possible in the time available, to collect and collate the data necessary for a detailed study of the cycles which may exist in the succession.

Cremer (1983) and Ravenne and Beghin (1983) demonstrated that the facies associations (named GdAI, II and III in this thesis) are arranged in symmetrical sequences, 50-100m thick. The field evidence for their conclusions is good. Each Gres d'Annot succession displays several moderately-bedded sandstone bodies of GdAIII, at random intervals through the formation. Their boundaries are usually gradational, both horizontally and vertically. Abrupt lateral terminations are extremely rare: even individual beds are laterally continuous over 100s of metres.

Note this is not the case at Annot, where the massive, thick sandstone bodies overl~~ie~~, and are sharply overlain by, thinly bedded sandstones. There are very few moderate sized beds that are not amalgamated. In part, this relates to the proximity of the area to its postulated source, but it is a special case in other ways, by virtue of the interpreted palaeotopography (section 5.6.1).

Interpretation.

These sequences have been interpreted two ways to date: both the models are sedimentological and do not invoke external controls on the sediment supply.

The first (fig. 5.41; Ravenne and Beghin, 1983) does not account for the lateral variations observed, but it does not preclude them. The model suggests that the turbidity currents were generated from a slope that failed repeatedly: first, small slides which increased in size to a major failure, then progressively decreased in size and frequency until a stable slope angle resulted. This model for slope failure does not appear valid. It is true that a slope may fail more than once before stabilising, but the major failure is the first, and subsequent slides tend to be smaller (Reading, pers.comm. 1984). This sequence of events would produce an asymmetric facies sequence.

The model proposed by Cremer (1983) is shown in fig. 5.42. It suggests that the principal axis of sediment transport and deposition shifts with time, presumably in response to subtle changes in the topography after each event. The sediments formed extremely low relief, broad features on the sea bed. Cremer terms them "lobes" which is acceptable because the model assumes that each depositional event leaves a positive feature on the sea bed. Note that the thickest sandstones are interpreted by Cremer as the fill of very shallow, broad, ephemeral channels.

Of the two, Cremer's model, albeit simplistic, is favoured by this author. It accounts for: the symmetry of the sequences; lateral variation of facies associations over distances of kilometres, not 100s of metres;

complex sequence variations. The hypothesis is based on the sound principal that turbidity currents seek the lowest regions of the basin floor. That is, the sequences are similar to the compensation cycles of Mutti and Sonnino (1981), fig.5.43, though at a large scale.

When Ghibaudo's extensive data set becomes available, it should be possible to map the sediment bodies with greater confidence. The volume of sand in, for example, the thick sandstone bodies, can be calculated from their lateral extent. Changes in sediment body volume could be responsible for the vertical sequence of facies. If this proves to be the case, then extrinsic controls influenced the succession of facies associations and should be modelled. At present, the undoubted importance of extrinsic controls cannot be tested.

5.7.2.3. Variations observed within the Gres d'Annot succession at the scale of the basin fill, across the whole study region.

Over the region as a whole there appears to be no consistant basin-fill megasequence. In the Peira Cava area, the succession broadly thins upwards (Conort and Odishou, 1978). In Col de la Cayolle, Ghibaudo composite log displays thick, massive sandstones at the base; moderate to thinly bedded sandstones in the central part; a thickening upward sequence towards the top of the formation (fig. 5.43). The upper part he interprets to represent a progradation cycle (Ghibaudo, 1985). Elsewhere in the region, thick, massive sandstone bodies occur at random in any one section.

In a single outcrop area, the facies associations observed and the ratio of the different facies associations remain remarkably constant between sections. The same is true when adjacent outcrop areas are compared. For example, the successions observed in the Grand Coyer, Col de la Cayolle (fig. 2.31b) and the Trois Eveches (fig. 5.6) outcrop areas are similar. Jean (1985) has divided the Gres d'Annot outcrops into two groups, linking areas with similar proportions of each facies association. The southern group are dominated by the large, massive sandstone bodies GdAIII; the northern group by the medium thickness beds GdAII.

Interpretation

The Gres d'Annot Formation does not record the existance of a single overriding control on the basin fill: for example, a major sea-level fall (thickening upward progradation cycle), or the progressive denudation of a rapidly uplifted source area (thinning upward succession). Instead, each outcrop area is different, in part because the successions are not contemporaneous. However, the differences may also be attributed to the basin floor morphology which probably divided the basin into separate depocentres. It is likely that in the areas of Col de la Cayolle, Peira Cava and Annot, at least, the topographic relief was the primary control of turbidite deposition (section 5.5 and 5.6).

The areal distribution of facies associations can, to a first approximation, be related to the proximity of each outcrop area to the southerly source areas. Thus, Annot, Puget-Theniers, Contes and Menton were more proximal than the rest, received sediment at more frequent intervals, and so are dominated by thick sandstone bodies.

5.7.3. Fan or no fan?

5.7.3.1. Introduction.

A submarine fan is a sedimentary system comprising a number of identifiable sub-environments, namely: distributary channels and depositional lobes (Normark et al, 1984). The term implies a distinct morphology (plan and surface relief), although highly elongate sedimentary systems are sometimes classed as fans (eg. Bengal Fan; Emmel and Curray, 1984).

The term can only be applied where the sub-environments can be identified and be shown to relate to one another (eg. Walker, 1978; Rupke, 1977). A high level of control by intrinsic processes is implicit in all the fan models. It must be possible to demonstrate that sediment was supplied to a positive sedimentary feature via a system of distributary channels or the more ephemeral braided channels of a suprafan. The fan models are too often applied without presenting the evidence for a spatial relationship between channelised sediment bodies, thinly bedded sandstones and shales and the laterally extensive "lobe" deposits.

5.7.3.2. The lack of evidence at Annot.

i) No radial distribution of palaeocurrent data.

ii) No **gradual** down-palaeocurrent or lateral facies association trends. Facies differences do exist from one outcrop area to another; these rather sudden changes are attributed to the division of the basin into isolated or partially restricted depocentres by the basin floor topography.

Where two outcrop areas were linked, for example Trois Eveches and Col de la Cayolle, the facies associations and sedimentary sequences are very similar. This implies that the sediment gravity flows were able to traverse the whole basin from source area to the most northern outcrop without appreciable loss of energy.

iii) There is a dearth of channelised sandstone bodies.

iv) Individual sandstones and sediment bodies are laterally continuous over many kilometres. There is no evidence to suggest that bodies formed positive sedimentary features, or lobes, on the sea floor. The exceptions to this were described in previous sections: they related entirely to the proximity of steep basin margin slopes (encl. 5.5).

v) Asymmetric sequences, at any scale, have not been identified within the Gres d'Annot successions anywhere in the study region.

Whatever they really signify when present, the absence of asymmetric facies sequences demonstrates that the Gres d'Annot basin fill cannot be interpreted as a series of fan progradation and abandonment cycles!

Cremer (1983) proposed an autocyclic mechanism by which the symmetric cycles may have developed. The model has not been tested to date. The cyclic nature of the sedimentation can be modelled by varying the volume of sediment in individual flows.

Interpretation: The Gres d'Annot Formation does not conform with any fan model.

5.7.4. The model for the Gres d'Annot turbidite system.

The model for the Gres d'Annot is summarised in encl. 6.3. The main points are:

i) There was a significant basin floor relief produced by earlier shortening. The topographic highs overlie structural highs, like hanging wall anticlines.

ii) The effect of this relief was to divide the Gres d'Annot basin into several sub-basins.

iii) The entire sedimentary system was controlled by the relief. The lateral continuity of sediment bodies suggests that, had the turbidites entered an unconstrained system, the "ideal fan" produced would have been larger than the basin. The ponded sub-basins are an order of magnitude smaller, so "ideal fan" processes are probably of little relevance to them.

iv) The vertical sequence of facies at any one point was principally controlled by the density and energy of the individual sediment gravity flows and the time interval between them. Processes such as lobe propagation, channel abandonment, etc., did not control the vertical sequences.

v) Migrating structural highs produced offlapping sequences, for example, between Dormillouse and Chalufy, in the Trois Eveches outcrop area. The original thickness of the Gres d'Annot was probably not much more than 1000m at any point.

5.7.5. Concluding comment on the Gres d'Annot basin as a whole

In the previous section, it has been established that the deposition of sandstone from the Gres d'Annot sediment gravity flows was strongly controlled by a complex basin floor topography, all of which remained submarine throughout the deposition of the Gres d'Annot Formation.

The influence exerted by a palaeoslope depended on:

i) its scale relative to the strength of the sediment gravity flow;

ii) its orientation with respect to the original flow direction; the dip of the palaeoslope.

At all scales, the palaeo-relief was dictated by the geometry of, and displacement on, accomodated on a variety of structures beneath the basin floor itself. The large scale topography divided the basin into separate depocentres for at least part of their depositional history.

CHAPTER 6: THE EVOLUTION OF THE GRES D'ANNOT BASIN.

6.1. Introduction

In the previous chapters, the nature of Tertiary sedimentation in eastern Haute Provence has been discussed and related to a detailed interpretation of the tectonic evolution of the region. The distribution and evolution of the sedimentary facies demonstrates the existence of significant relief within the sedimentary basin at all stages throughout its evolution. The integrated study shows that the palaeotopographic highs and lows are coincident with structural highs and lows: the basin floor relief developed as the underlying Pyrenean and Alpine compressional structures evolved.

In this chapter, the relationship between sedimentation and tectonics at the scale of individual structures and sedimentary environments will be used to develop a model for the origin and evolution of the sedimentary basin itself. The principal problems to address are:

i) The cause of subsidence.

ii) The geometry of the whole basin and the extent to which it was subdivided into separate depocentres by the structural relief.

iii) Sediment supply throughout the history of the basin.

iv) The relationship between sediment supply and subsidence. Why did a deep marine basin develop at all?

v) The end of sedimentation in the basin.

6.2. The nature of the Gres d'Annot marine basin.

The Gres d'Annot basin has been described as:

- an extensional basin, either the remnants of a Mesozoic extensional basin (Graham, pers.comm.) or the fill of early Tertiary graben, produced by indentation tectonics and probably related to the Rhine-Rhone extensional system (Pfiffner, 1986).
- a strike slip basin (Homewood and Caron, 1983 and Ghibaudo, pers. comm.).
- a peripheral basin, foreland to the Alps or the Pyrenees (Elliott et al, 1985).

There is no evidence for crustal extension of Haute Provence during the early Tertiary. Small and medium scale extensional faults existed, but all may be interpreted in the context of a compressional tectonic setting. Faults such as those of the upper Gialorgues valley in the Col de La Cayolle outcrop area produced 100m scale basin floor topography, but were not responsible for subsidence in the basin.

The limits of the Gres d'Annot basin are approximately those of the Mesozoic Vocontian basin. It is this observation that gave rise to the hypothesis that the Tertiary sediments in Haute Provence infilled the remnant topography of an extensional basin. The evidence does not, however, support the hypothesis:

i) The whole region was uplifted at the end of the Upper Cretaceous and significant amounts of Upper Cretaceous strata were eroded from most areas. This is interpreted as a response to compressional tectonic forces, with evidence for partial inversion at the margins and within the Mesozoic basin. However, it is possible that eustatic sea level changes were responsible for the exposure of S.E. France during the Palaeocene.

ii) The Uppermost Cretaceous preserved at Grand Coyer contain corals, echinoids and other fauna that imply the seas were quite shallow in the basin itself. Therefore, any remnant topography between the extensional horst and graben was a few hundred metres at most.

iii) It is feasible that thermal subsidence associated with the Jurassic rifting continued well into the Tertiary. However, the basinal areas subsided a maximum of 2km during the 70 million years of the Cretaceous. Therefore, allowing for decay of the thermal subsidence response, the predicted rate of thermal subsidence during the Eocene would be less than 25m per million years. This figure does not allow for the thermal uplift due to Pyrenean compression and resulting inversion. The 1000m+ of Gres d'Annot turbidites were deposited in less than 10 million years! Therefore, remnant thermal subsidence cannot have been responsible for the subsidence observed in the Gres d'Annot basin.

The most popular model for the Gres d'Annot basin places it in a strike slip tectonic setting, because:

- Adria moved towards the NW or WNW with respect to Europe. Therefore, the European margin adjacent to Haute Provence was an oblique collision zone.
- The Gres d'Annot turbidites were sourced from basement massifs of European continent affinity. If the Gres d'Annot basin is considered only in the context of the Alps, a simple mechanism for basement uplifts within the orogenic foreland is provided by wrench tectonic models.
- Ghibaudo interpreted the apparently rapid subsidence in the Gres

d'Annot basin and the sharply defined margins of the turbidite basin as characteristic of a strike slip pull-apart basin. He interpreted the lack of evidence for syn-sedimentary deformation in the Gres d'Annot turbidites to indicate that the basin did not subside at all after deposition of the Marnes Bleues Formation, and that the actively uplifting source area for the turbidites was tectonically decoupled from the Gres d'Annot basin by deep seated, steep faults. These points are valid, but they arise from limited knowledge of the tectonic setting of Haute Provence or incorrect interpretation of the evidence.

The Alpine compression is expressed in Haute Provence as SW directed thrusting, possibly as a result of strain partitioning (section 6.5). There is no evidence that S.E. France was ever part of a huge wrench zone. The source of the turbidites was generated in response to Pyrenean compressive forces.

This study has not revealed steep strike slip faults defining the margins of the turbidite basin as required by Ghibaudo. Instead, it has shown that compressional structures were responsible for defining the internal geometry of the basin. If the region had been part of a positive flower structure, it would not then have been a region of net subsidence.

The Rouaine fault zone was trans-tensional: small scale pull-apart basin during deposition of the Calcaires Nummulitiques and Marnes Bleues Formations. However, this fault zone existed only to accommodate differential thrust displacements.

If the Gres d'Annot basin formed as the flexural response (eg. Beaumont, 1981, Karner and Watts, 1983, Karner et al, 1983) of the edge of the

European continent to loading, then potential loads must be sought:

i) To the south.

Uplift along the Pyrenean chain began in the Late Cretaceous and continued through the Eocene. This uplift was related to thickening of the continental crust at the southern edge of Europe and the relief certainly represented a significant load to the south of Provence. The surface of the Pre-Triassic basement dips towards the Pyrenees across the study region. However, a surface dip towards the south was countered by inversion and backthrusting of the Mesozoic cover within the Pyrenean foreland.

ii) To the east.

There is evidence for a surface dip down towards the east during the Eocene:

- The Gres d'Annot basin deepened towards the east from an approximately N-S margin at Barreme. The oldest Tertiary marine sediments were deposited in the east: the Calcaires Nummulitiques are Lutetian in Peira Cava and Priabonian in Barreme. Correspondingly, the whole Tertiary sequence is diachronous across the region towards the west.

- The Gres d'Annot turbidites may be correlated with similar facies sediment of equivalent age, preserved around the Alpine Arc: Gres de Champsaur, Aiguilles d'Arves, Gres de Taveyenne. A more or less continuous deep, but narrow trough (much less than 100km wide at any one time) extended for several hundred kilometres parallel to the Alpine Arc.

It is difficult to be precise about the nature of the Alpine load at this time, but the Brianconnais of the Vanoise massif was depressed beneath crust to depths of 35km (Platt and Lister, 1985). Therefore, Adria was loading the edge of the European continent in the western Alps and subduction of the Brianconnais would have generated significant surface relief.

Although local thrusting in Haute Provence was to the SW or WSW, Adria moved towards the NW/WNW with respect to Europe. Therefore, the Gres d'Annot basin developed primarily as a lateral peripheral basin to the Alpine system.

The Gres d'Annot basin may be modelled as a lateral foreland basin, modified by shortening and compressional uplift <u>within</u> the basin. Surface subsidence was induced by Alpine loading; basin floor topography was generated by both Alpine and Pyrenean compressional structures.

6.3. What delineated the Gres d'Annot basin?

<u>To the North</u>: The margin of the Gres d'Annot basin corresponded with the old Mesozoic basin margin. Inversion in the Cenomanian and at the end of the Upper Cretaceous uplifted this margin. It was eroded, as evidenced by the base Tertiary unconformity which oversteps a truncated Mesozoic stratigraphy onto basement at the southern margin of Pelvoux. The erosion itself re-established the old basin margin, because the basement lithologies of the old Pelvoux horst were resistant to erosion, whereas the Jurassic basin sediments were soft shales interbedded with marlstones. Direct analogy may be drawn with the Channel basin, where the block that was upthrown during inversion is now topographically lower due to erosion.

To the West: The margin of the marine basin is defined by loading-induced flexure of the foreland loading. It is quite near to the platform-basin facies transition in Jurassic sediments, which may be accounted for by a difference in the flexural rigidity of the crust across the Mesozoic basin margin. Mesozoic extension thinned the crust beneath Haute Provence and Late Cretaceous inversion did not restore the crust to its original thickness. Therefore, it is probable that the margin of thinned crust formed a hinge in the flexural response of the European continent to loading.

To the South: The old Mesozoic basin margin was modified by Pyrenean and early Alpine thrusting. The Gres d'Annot basin extended only as far as the east-west fold and thrust belt of the Southern Castellane Arc. The St. Antonin syncline contains a Mid-Late Eocene shallow marine sequence, the lateral equivalent of the Gres d'Annot.

To the East: *The basin was not entirely open the east.* However, the Alpine orogen contributed no sediment to the Gres d'Annot basin at any stage in its development. *this may be explained in a number of ways:*

i) The load depressing the European continent was sub-crustal, such that the load had little or no topographic expression. There are thrust sheets containing mantle, for exmple, the Ivrea body, but there is no evidence to suggest that they were emplaced so early in the history of the Alps.

ii) The Alpine wedge tapered very gradually towards Europe, so that sediment was trapped in small basins on the wedge and failed to reach the Gres d'Annot basin.

iii) A deeper trough existed to the east of the Gres d'Annot basin which trapped and diverted sediment along its axis, in a manner similar to the Aleutian trench. This hypothesis is not favoured because remnants of the fill of such a trough are not preserved in any thrust sheets within the Alpine thrust belt. In fact, the sediments young consistently and progressively towards the west. Eocene deposits were only deposited in the Gres d'Annot basin, on Eurpean coninental crust.

iv) A topographic ridge existed at the edge of the European continent. Such a ridge may have formed above a Pre-Triassic basement thrust sheet. Subsequent shortening on the same structure would have uplifted and eroded any sediment dseposited on or behind the obstruction.

v) A dense mantle wedge, like the Ivrea body, may have been emplaced at an early stage in the Alpine collision. In this way, the Alpine orogenic wedge could depress the European continent without generating topographic relief in the wedge itself (Royden and Karner, 1984).

Early Alpine thrusting to the SW/WSW within the foreland depocentre has been demonstrated. There is palaeocurrent evidence for surface slopes down to the west, probably off the front of such a structure. Therefore, it is quite feasible that European crustal thrusts were responsible for diverting Alpine orogenic wedge sediment. This model which allows for the development of such thrust sheets during the evolution of the Gres d'Annot basin also accounts for the progressive shift of the Gres d'Annot basin depocentre towards the west.

6.4. Source areas for the Gres d'Annot basin reviewed in the light of a revised tectonic setting for Corsica-Sardinia.

Observations regarding the supply of sediment to the Gres d'Annot basin:

- The crust began to subside during the Mid Eocene. Until the Priabonian it continued to deepen and achieved depths greater than 1000m, as evidenced by fauna in the Marnes Bleues Formation. The basin subsided for a long time without being filled by sediment: quite unusual for a foreland basin!

- During the earliest stages of basin development, sediment was locally derived from relative highs within the basin. The Poudingues d'Argens Formation primarily contains clasts of basinal Upper Cretaceous limestone and rare Triassic and Permian clasts shed from proto-Barrot. The Calcaires Nummulitiques contain similar detritus, in part reworked from the Poudingues d'Argens clasts; carbonates were generated in situ by fauna in the shallow marine seas. The Marnes Bleues sediment was reworked from adjacent shallow seas.

- Subsequently, the Gres d'Annot sediment gravity flows supplied sediment entirely from southerly source areas. The sharp stratigraphic contact between the Gres d'Annot and the Schistes a Blocs, preserved at Col de La Cayolle, implies that the supply route for the course siliciclastic detritus was rapidly deactivated.

- The Schistes a Blocs were derived from multiple sources, mostly associated with the Alpine internal zone thrust sheets.

Interpretations:

The absence of sediment derived from the Alpine orogen has already been discussed (section 6.3.). Therefore, it remains to explain the timing of sediment supply to the Gres d'Annot basin from the southerly "Pyrenean sources". The evolution of the basin from one starved of sediment to one supplied with huge volumes of immature silciclastic sediment can be accounted for a number of ways:

i) A fall in sea level may result in sediment previously stored in shallow open shelf seas or perched basins being rapidly reworked into deep marine basins.

This is not obviously the case in the Gres d'Annot basin because the sediment grains are so fresh -angular and approximately 40% feldspar- that they could not have been reworked for any length of time in shallow seas. There is little space for an extensive shelf sea south of the Gres d'Annot basin, even when late Alpine displacements are restored.

The sediment may have been rapidly deposited on fan deltas in narrow structurally defined sub-basins, like the St. Antonin syncline. However, even here, the evidence suggests that the sediment was not protected by the shape of the sub-basins from reworking by waves or tides.

ii) It is possible that alluvial fans may have stored sediment in small drainage basins, delineated by structural highs, until a passage out of the basin was eroded. Large volumes of sediment may have passed out of such a basin as the new river sytstems were established.

iii) I propose a tectonic model for the evolution of Corsica, Sardinia and the Maures-Esterel massif which can account for the observed variations in sediment supply to the Gres d'Annot basin (summarised in fig. 6.1 and 2). The model is derived from PhD thesis research of Warburton, Knott and Waters and my own knowledge of the sediment supplied to Eocene sedimentary basins:

End Cretaceous:

Compression from the south resulted in uplift along the geographic Pyrenees, but further west it was not yet associated with major uplift. Corsica and Sardinia were in collision with an island arc and were depressed beneath it (Knott, pers. comm.).

Pyrenean compression resulted in inversion of deep-seated Mesozoic extensional faults and south directed shortening of the Mesozoic cover of Haute Provence and slight uplift of the Maures-Esterel massif. The relief produced small, locally supplied drainage basins within the foreland in which the Poudingues d'Argens were deposited.

Mid Eocene:

Corsica and Sardinia collided with Adria itself. The Tende massif was emplaced onto Corsica and these fragments of continent were thrust northwards, depressing Maures-Esterel in their footwall. There is good evidence in parts of Corsica for significant backthrusting of crust and Mesozoic cover towards Adria (Warburton, 1986). These displacements uplifted the parts of European basement that supplied sediment to the Gres d'Annot turbidite basin.

Pyrenean shortening is not observed in the cover of Haute Provence in the Eocene. Instead, most of the convergence between Adria and Europe was accommodated by backthrusting in the Corsican orogen; the remaining displacements surfaced south of Maures-Esterel. In consequence of the backthrusting, the continued shortening and repeated uplift of the sediment source areas did not bring them closer to the basin margin.

A narrow, deep foreland basin developed on Corsica: it contains sediment similar to the Gres d'Annot, with no grains of high pressure metamorphic minerals. Therefore, at that stage, the Schistes Lustres and associated metamorphosed imbricates of European basement were some distance from the basin.

The conglomerate marker beds within the Gres d'Annot were sourced from areas still further to the east.

Oligocene:

Corsica and Sardinia began to rotate away from the European continent. Crustal stretching caused what had been a basement culmination to become a rift basin (Ligurian Sea). The source of the Gres d'Annot subsided, ending turbidite sedimentation within the basin. Clastic sediments sourced from the remaining high areas of Corsica-Sardinia-Calabria were unable to cross the rift and were diverted southwards, supplying such formations as the Numidian Flysch of N.Africa and S.Italy.

6.5. A brief consideration of thrust transport directions in the S.W. Alps.

The sum of evidence from the Alpine chain implies that Adria moved consistently towards the WNW with respect to Europe. Despite this, the Alpine structures in Haute Provence are consistently SW vergent, from the Eocene to Recent times.

During the Eocene, apparent SW directed movement may be produced by wedging basement northwards beneath the Triassic detachment (Pyrenean) while displacing the Mesozoic cover towards the west (Alpine). The relative movement vector towards the SW is illustrated in fig. 6.2, part 2.

Throughout the history of Alpine thrusting, it is possible that the WNW convergence was partitioned into SW thrusting, plus strike slip along NW-SE striking faults. This mechanism would also account for the north directed thrusting towards the Bavarian foreland: the strike slip faults would strike E-W. (fig. 6.2, Part 1.)

6.6. Basin dynamics: Tectonics and topography within the basin.

6.6.1. Introduction: the regional influences.

The Eocene S.W. Alpine foreland basin developed on already shortened and actively shortening continental crust. Simplistic models of foreland basin subsidence cannot be applied to the Gres d'Annot basin because the region experienced: i) crustal loading both to the east, by the Alpine orogenic wedge and to the south, by the Pyrenean orogen; ii) uplift due to thin skinned shortening and inversion. The pattern of net vertical movement was complex.

This study has established the importance of both the Pyrenean and Alpine compressional tectonic systems in the generation of topography prior to and during the development of the marine basin. The effect of each system has been analysed:

Pyrenean shortening.

In the geographic Pyrenees, the Iberia continent collided with Europe during the Late Cretaceous-Eocene times. To the east, the equivalent north directed compression resulted in the collision of an island arc system with Corsica-Sardinia late in the Cretaceous. By the Mid Eocene, Corsica-Sardinia were colliding with Adria and were thus incorporated into the Alpine (s.s.) orogen.

The Pyrenean displacements entered Haute Provence at mid crustal levels and the convergence was expressed by inversion of pre-existing faults and the initiation of some south vergent structures in the Mesozoic cover.

A component of compression from the south towards the north was exerted until Corsica-Sardinia rotated away from Europe during the Oligocene.

Alpine shortening.

The Alpine deformation front propqgated into Haute Provence during the Mid Eocene immediately prior to and during the Tertiary transgression. Most shortening occurred above the Triassic detachment at this stage; basement uplifts probably represented structures that soled out on the base Permian. The SW and WSW vergent structures continued to develop throughout the sedimentation history in the Gres d'Annot basin. A period of relative tectonic quiescence during the deposition of the Gres d'Annot turbidites was followed by further shortening of the foreland crust prior to the emplacement of the internal zone thrust sheets. Most of this and subsequent displacements passed into the foreland beneath Barrot and surfaced to form the Digne thrust imbricate stack.

This section traces the development of topography within the Gres d'Annot basin, using a series of maps and block diagrams (enclosures 6.1,2,3).

<u>6.6.2. Mid Cretaceous</u> (fig. 2.22)

Compressive structuring began while the region was submerged and continuing to subside in response to the long term thermal effects of Mesozoic extension of the European crust.

During the Cenomanian, south-north displacements entered S.E. France at mid crustal levels and Hercynian/Mesozoic faults were reactivated in compression. The northern margin of the Vocontian basin was partially inverted. Reverse movement on antithetic faults within the northern part of the Mesozoic basin produced south vergent monoclines: on cross sections with the base Upper Cretaceous restored to horizontal, two such monoclines occur at Dormillouse and Crete de La Blanche. A thrust tip in the Terres Noires is believed to be responsible for folding at Eaux Chaudes. The uplifted terrains were locally elevated above sea level and eroded prior to deposition of the Upper Cretaceous limestones.

During the Upper Cretaceous, subsidence was probably in part induced by Pyrenean loading to the south. This would account for the overall increase in thickness of the basinal limestones towards the south of the study region. The tip of the Pyrenean orogenic wedge was to the south of St. Antonin, where the Upper Cretaceous is represented by a reduced sequence of glauconitic sandstones: the Maures-Esterel massif was uplifted at this stage.

6.6.3. End Cretaceous - Early Middle Eocene (Enclosure 6.1)

During this period, the whole of S.E. France was emergent: in parts of Provence, the eastern limit of the North Pyrenean foreland basin was filled to capacity with continental sediment, much of it probably derived from Haute Provence.

The Mesozoic stratigraphy beneath Haute Provence itself was uplifted and the study region became one of net erosion. A significant proportion of the Upper Cretaceous succession was removed. The limited Poudingues d'Argens Formation only represent a small part of the eroded material. Why were any continental sediments preserved at all, or conversely why is there so little sediment preserved?

The Poudingues d'Argens represent alluvial fans, some deposited around palaeohighs at the same time that the marine Calcaires Nummulitiques were being deposited in palaeolows. Probably little of the Poudingues d'Argens predates the onset of the transgression in Haute Provence, because the absence of many wholly mature palaeosols indicates relatively rapid deposition and the upper continental sediments pass transitionally into marine limestones.

Development of the alluvial fans was probably stimulated by a combination of:

i) The change in the dynamics of the river systems caused by the sea level rise itself; Ref required ?.

ii) The re-organisation of the surface topography that occurred during the transgression due to the propagation of Alpine deformation into the area. Previously, the tectonics, and hence the surface topography had been dominated by Pyrenean E-W structures. NW-SE ridges causded by Alpine thrusts were superimposed on this pattern, dividing the area into small drainage basins.

The existence of an angular unconformity at the base of the Tertiary succession in Haute Provence and the distribution of the Poudingues d'Argens Formation point to significant pre-Mid Eocene structuring. This took the form of:

i) Deep-seated movement, possible inversion, of an inferred fault along the Beauvezer valley. The entire area to the north was uplifted and eroded: it was apparently peneplained prior to the initiation of Alpine structures and therefore did not supply sediment to the Poudingues d'Argens depositional systems.

ii) Reverse movement on an inferred fault beneath the Barreme syncline which induced uplift of a very large area to the west of Tartonne. The entire Upper Cretaceous succession was stripped from the hanging wall. Some of the detritus was preserved in the Douroulles syncline and east of Barreme. At Douroulles, the contemporaneous development of E-W and N-S fold axes produced dome and closed trough topography which controlled the distribution of sediment bodies in the Poudingues d'Argens (Evans, 1987).

iii) Shortening of basement, in particular above an interpreted base-Permian detachment, produced a localised topographic high close to the site of the present day Barrot massif which continued northwards to Cairas, as a low relief feature. North-south Pyrenean shortening was probably responsible for the reactivation (inversion) of a Permian basin margin fault.

Much of the sediment shed from this topographic high was not deposited in the study region, but Permian and Triassic detritus is preserved in the structural depressions, at Sausses and Quatre Cantons.

iv) The Mesozoic cover was backthrust towards the south. Poudingues d'Argens were preserved around the lateral tips to E-W structures, for example, at Roquesteron.

v) Thrusts (SW directed) propagated in the Terres Noires and sliced through the pre-existing monocline at Crete de La Blanche, ramping up into the Mid Cretaceous shales. The resulting Mid Cretaceous duplex and the interpreted emergent thrust tip to the south of Beauvezer formed a topographic ridge. This is believed to have been the primary source of Upper Cretaceous detritus for Poudingues d'Argens of the Mort de L'Homme/Peyresq alluvial fan system. Secondary sources of Upper Cretaceous clasts were defined by other thrusts and folds developed above the Triassic detachment.

The fans formed in a drainage basin bounded to the north by the thrust range described above; to the west by a ridge of kinked Upper Cretaceous limestones and underlying thrusts in the Mid Cretaceous shales (the Cordeil anticline and its southerly equivalents); to the east by the lateral ramp of proto-Barrot and the Melina-Aurent kink zone; to the south by the footwall ramp of the south directed Cheiron thrust.

vi) The Argens syncline was defined by low relief topographic ridges, to the west by the Castellard thrust ramp and to the east by the developing Cordeil-Chammatte high.

vii) The eroded kink folds at Belap and Chateau Garnier are evidence of relief developed above the northerly equivalents of the Cordeil-Chammatte high. The sediment from these is not preserved nearby: they were probably already low relief features when the Poudingues d'Argens depositional systems developed.

Several thrusts and thrust tip folds developed beneath and to the west of the Barreme syncline (for example, the Digne tip fold). They were eroded prior to deposition of the Calcaires Nummulititiques: where limestones were locally exposed on the unconformity surface, coral colonies were able to thrive (Evans, 1987), but there is no evidence of remnant topography.

While the basement structures defined the large scale topography, the small scale relief was produced by thrusts and kink folds within the Mesozoic, especially in and above the Mid Cretaceous shales.

6.6.4. Mid Eocene: during the Tertiary transgression. (Encl. 6.2)

As the sea level rose, the Alpine structures continued to develop. Pyrenean displacements became less important as north-south convergence was accommodated by shortening and continental subduction to the south of Maures-Esterel. E-W compressional structures in the southern sector of the Castellane Arc were tightened, but at the same time N-S striking graben developed. The southward shortening and along-strike extension are interpreted to be the result of inhibited SW directed Alpine thrusting. The boundary of the E-W differential shear zone was the Rouaine fault zone. Sinistral wrench movements along this zone generated localised areas of rapid subsidence (pull apart basins) at Rouaine and St. Benoit: these filled with shallow marine carbonates, several hundred metres thick at Rouaine.

Maures-Esterel was uplifted and rivers flowed northwards, confined to the axes of the graben and controlled by the E-W folds, into the developing marine basin. The sea encroached from the south and east. The southern part of the study region was submerged early and as a result accummulated a thick transgressive carbonate sequence.

Some topographic highs remained above sea level for some time and continued to shed coarse conglomeratic sediment: alluvial fans were gradually submerged and fan deltas now fringed the structural highs. For example, thrusts surfacing in the Beauvezer valley fed the Mort de L'Homme-Peyresq fan/fan delta.

Newly initiated SW and WSW directed Alpine thrusts significantly changed the direction of topographic slopes in the region **for the last time:**

i) A new Permian thrust sheet formed which was of limited lateral extent (B2-3; fig. 2.39) and moved to the SW: the thrust fault propagated from the site of the lateral ramp of proto-Barrot, only utilising part of the pre-existing N-S fault.

Earthquakes associated with the initiation of this structure caused huge boulders of Upper Cretaceous limestone to fall from the new ridge into the shallow seas at Cairas (Grand Coyer outcrop area). The thrust ramps, both frontal and lateral, defined major topographic slopes: the W/SW dipping frontal hanging wall ramp formed the eastern boundary of the Annot sub-basin; the footwall ramp marked the western boundary of the Grand Coyer area.

ii) Remnant and developing topography above the Melina-Aurent structures marked a break in slope along which patch reefs developed (Le Ruch). Cracked Cretaceous clasts imply active development of kink folds at surface.

iii) Kink folds actively developed at surface in Upper Cretaceous limestones of the Grand coyer area (Cairas), immediately prior to and during the earliest stages of the transgression. Freshly fractured fold surfaces were preserved beneath Calcaires Nummulitiques and Cretaceous clast conglomerates, with a shallow marine limestone matrix, are preserved in the synclines.

iv) A S/SE dipping slope was generated along the Argens syncline, above the lateral hanging wall of a developing, SW directed, Lower Jurassic thrust sheet 3b (cross section Z). The culmination was emergent in the Thorame Haute valley and supplied small, south-prograding fan deltas.

v) Thrust sheet 3a accommodated equivalent displacement to the north of the Beauvezer valley. This and the La Valette thrust formed low relief features on the sea floor onlapped by the Calcaires Nummulitiques.

6.6.5. Priabonian: Pre- and syn- Gres d'Annot deposition in the study region. (Enclosure 6.3)

From now onwards, topography within the Gres d'Annot basin was generated by continued activity on already established structures.

- Further displacement on the basement thrust, BII, defined a low angle slope above the footwall, onlapped by the Gres d'Annot turbidites.

Displacements on the Triassic detachment were inhibited by the basement high. E-W folds developed in the Entraunes valley, onto which the Gres d'Annot onlapped at Tete du Moulin de Betrand (Col de La Cayolle area). Displacements on thrusts in front of B2-3 were transferred benath the culminatin, on the base Permian.

- Displacement of the Lower Jurassic thrust 3b moved Allons up the footwall ramp to the pre-existing Cheiron thrust. The resulting west-dipping slope was onlapped by turbidites in the Annot outcrop area.

- The Annot sub-basin was now entirely confined, bounded to the east by a basement thrust sheet and the Melina-Aurent high; to the north by the Lower Jurassic thrust 3a and Mid Cretaceous duplex 1a and emergent thrusts; to the west by the Castellard thrust, Lower Jurassic thrust 3b and the still active Cordeil-Chammatte high. To the south lay the Rouaine fault zone, at this time an area of non-deposition, therefore a relative high. Beyond that, lay the low relief Maures-Esterel massif and the actively deforming, rising source area for the Gres d'Annot turbidites.

The Trois Eveches Problem

This has been addressed in section 5.6.2 and 2.3.2.12, but to restate the paradox observed at Dormillouse, in the north of the Trois Eveches outcrop area:

- Deep erosion of Mesozoic stratigraphy pre-dates deposition of the Calcaires Nummulitiques Formation.
- Dormillouse is stratigraphically the base of the Gres d'Annot succession. In Ghibaudo's model (1985, in Elliott et al), the Gres d'Annot basin was passively infilled and was 3000m deep at Dormillouse.
- Dormillouse is now a structural high.

Ghibaudo's model requires that Dormillouse behave like a structural yoyo. He provides no mechanism for the catastrophic subsidence during deposition of the Marnes Bleues. An alternative model is presented and illustrated in fig. 5·26. It explains the observations by using an observed and testable structural evolution:

1. End Mid Cretaceous.

   Pyrenean north-south compression partially inverted the Mesozoic basin. Dormillouse was uplifted over a south-vergent monocline.

2. Pre-Calcaires Nummulitiques.

   Further inversion, plus thin-skinned Alpine thrusting towards the SW, produced a culmination at Montagne de La Blanche, which was eroded.

3. Calcaires Nummulitiques - earliest Gres d'Annot deposition (fig.-.3c).

   Regional foreland basin subsidence continued. Alpine thrusting took the form of simple slip on the Triassic and Terres Noires in the north Trois Eveches area. To the south, the thrusts climbed up and duplicated stratigraphy: the resulting uplift counteracted the foreland susidence.

   The net result was relative subsidence of Dormillouse, and the Gres d'Annot onlapped southwards onto the thrust ramps, between Tete de Auriac and Mourre Gross.

4. Mid-Gres d'Annot deposition.

   The northern part of the Trois Eveches outcrop area was uplifted as it passed over thrust footwall ramps. The early onlapping turbidites were rotated. Reactivation of the Montagne de La Blanche-Denjuan thrust system caused uplift in the south: at Mourre Simanche, the turbidites onlapped a thrust culmination of Calcaires Nummulitiques; at Denjuan, the thrust displacements were demonstrably coeval with the onlapping tubidite deposition.

To conclude, the turbidite depocentre migrated southwards with time and the turbidite system progressively offlapped the northerly high and onlapped developing thrust ramp slopes. The Gres d'Annot basin may never have been 3000m deep!

6.6.6. Post Gres d'Annot.

In some places, where the top of the Gres d'Annot is preserved, it is truncated by a thrust, in others by an erosional unconformity. In both cases it is overlain by the Schistes a Blocs, which in turn is overthrust by the Embrun-Ubaye thrust sheets.

The Schistes a Blocs Formation incorporates a wide range of lithologies with diverse petrographic signatures. Undeformed carbonate rich conglomerates and sandstones are exposed at Cabanes de St. Anne (Chasse Valley, Southern Trois Eveches outcrop area). Partially disrupted to intensely deformed interbedded thin fine, micaceous quartz sandstone, siltstone and shale are observed in the Col de La Cayolle outcrop area. The sediments have all been interpreted as deep marine deposits, although no thorough analysis has been made. The clasts were apparently derived from the Sub-Briançonnais, Briançonnais and Helminthoid Flysch. In some localities, for example, *Pas de Lausson* (Col de la Cayolle) the formation contains olistoliths of Briançonnais and Helminthoid Flysch and it grades upwards into a tectono-sedimentary melange.

The style and intensity of deformation in the Schistes a Blocs varies considerably, but much of it predated lithification of the sediment. In some cases, an intense cleavage has destroyed all primary sedimentary structures, including bedding.

The formation has been interpreted (eg. Kerckhove, 1969) as the debris shed from the advancing internal zone thrust sheets. This hypothesis explains its occurrence only as a thin veneer at the base of the Embrun-Ubaye Nappes, the diverse petrography and the variable state of deformation of the sediment. The sediment was shed into deep seas which implies that the internal zone thrust sheets were emplaced prior to significant uplift of the foreland.

Along the Trois Eveches outcrop area, the Schistes a Blocs are thrust over the Gres d'Annot Formation. The thrust apparently cuts down Gres d'Annot stratigraphy to the north.

To account for the apparent loss of Gres d'Annot Jean (pers.comm.) and others invoke a period of major uplift, tilting and erosion of the foreland Tertiary stratigraphy prior to the emplacement of the internal thrust sheets . However, the new offlap model, presented in this thesis, does not require that a significant amount of Gres d'Annot be removed, either by erosion or thrusting. The apparently missing stratigraphy was simply never deposited!

Wherever the top of the Gres d'Annot is in stratigraphic contact with the Schistes a Blocs, for example, in the Col de La Cayolle area, the contact is absolutely sharp. There are no beds of Gres d'Annot sandstone interbedded with the Schistes a Blocs. The base of the Schistes a Blocs is locally erosive: the depressions may be 10s of metres wide and several metres deep.

The irregular base of the Schistes a Blocs Formation has been interpreted by Jean (1985) as further evidence of significant erosion post-deposition of the Gres d'Annot. The three-dimensional geometry of the surface cannot be mapped. The relief may be interpreted instead to be that of submarine slide scars: slope failure would be induced by only a moderate increase in the local dip of the sea bed. Tightening of pre-existing foreland structures in front of the advancing internal thrust sheets would have produced the required effect.

It would appear from the Col de La Bonnette outcrops that the source of Gres d'Annot sediment was abruptly deactivated. However, study of sandstone rich turbidite systems in the North Sea (Barraclough, pers.comm.) has shown that, although turbidite deposition ends abruptly at each locality, the age of the final turbidite is younger at sites nearer the source. This implies that the turbidite depositional system became progressively smaller: the supply of sediment died more gradually than would be inferred from any single outcrop. This hypothesis cannot be tested in this field area for lack of palaeontological data and because the original top stratigraphic boundary of the Gres d'Annot Formation is rarely preserved.

There is good reason to believe that the Gres d'Annot sediment supply routes were quickly severed. In the early Oligocene, the configuration of continent fragments south of the Gres d'Annot basin began to fragment. They had been unified in the Eocene by Pyrenean and Alpine compressional forces. At the beginning of the Oligocene, ocean subduction to the south, east and west of the Eocene-Mediterranean coast induced back arc rifts to develop between the Maures-Esterel and a continent, composed of Corsica-Sardinia, the Balearics, Calabria, the Petit and Grand Kabylie and internal parts of the Betic and Rif. This is illustrated in fig. 6.4 Sediment from the massifs, that had been shed northwards into the Gres d'Annot basin, was now channelled into the developing rift valleys and made its way south to generate new deep sea turbidite systems, eg. the Numidian flysch, now preserved in North Africa and Southern Italy.

In the study region, it is impossible to date the emplacement of the Embrun-Ubaye Nappes onto the Alpine foreland. However, excellent evidence exists in the Barreme syncline immediately to the west: Sanoisian shoreline sediments contain high pressure minerals and fragments of Helminthoid flysch. These sediments overlie steeply dipping and eroded Cretaceous and Lower Tertiary sediments at the eastern basin margin. However, within a kilometre of the basin margin, they are conformable above the Gres de Ville, the lateral equivalent of the Gres d'Annot (Evans, 1987).

The deformation and erosion of the Barreme basin margin, the associated sea level drop and abrupt change of sedimentary environment, the rapid increase in volume and grain size of sediment input to this basin, the occurrence of minerals reworked from the internal thrust sheets all point to the emplacement of the Embrun-Ubaye Nappes in the Lower Oligocene.

Therefore, little or no time elapsed between the cessation of turbidite sedimentation in the Gres d'Annot basin and the emplacement of the Embrun-Ubaye thrust sheets. It is probable that the event were causally linked: the change in the micro-plate configuration in the western Mediteranean may have removed obstructions to Alpine thrusts. This has been inferred in the Castellane Arc model presented in this thesis: the SW directed thrusts in the foreland were impeded by the uplifted massifs to the south. It is probable that deformation in the Alpine orogenic wedge was concentrated in the internal zones (backthrusting and re-folding and -thrusting) until thrusts at various levels in the Alpine orogen could advance into the foreland.

To summarise the sequence of events:

1. In the Lower Oligocene, the Gres d'Annot ceased to receive sediment as rifting occurred to the south of the basin margin.

2. At approximately the same time, the Alpine deformation system advanced agressively into the S.W. Alpine foreland. Restructuring of the foreland preceded the emplacement of the internal thrust sheets: Lawson (1987) has shown that foreland structures were decapitated by the sole thrust of the Embrun-Ubaye Nappes. This generated some intra-basinal relief, which failed and was eroded by sediment gravity flows initiated at the front of the internal thrust sheets.

3. The Schistes a Blocs were deposited in front of these sheets and over-ridden or incorporated in the thrust wedge. The sole thrust tip was emergent and advancing across a sub-horizontal sediment surface by the time it reached the Trois Eveches outcrop area.

The limit of the furthest advance of the Embrun-Ubaye thrust sheets is probably defined by a distinct change in character of the Marnes Bleues. To the west, it is blue-grey, relatively soft and not cleaved. The the east, it is harder, lighter coloured and has a strong penetrative cleavage. The alteration front, thus the furthest advance of the internal thrust sheets, lies to the east and north of Annot. This could be confirmed relatively rapidly by measuring the vitrinite reflectance of wood fragments in the Gres d'Annot, combined with a diagenetic study, to map variations in the depth and speed of its burial.

The thrust sheets may have been prevented from advancing by thrust ramp topography: for example, the footwall ramp of the early Barrot culmination and the footwall of the Digne thrust.

4. The foreland basin was severely reduced in size, filled as it was with thrust sheets. The depocentre migrated stepwise in the foreland, as a result. Sedimentation continued to the west in the Barreme syncline and shallow marine, then continental conditions were established.

The S.W. Alpine foreland basin was from then onwards richly supplied with sediment from the Alpine orogen, and shallow marine or continental conditions were maintained.

5. Subsequent to the emplacement of the Embrun-Ubaye thrust sheets, the Digne imbricate thrusts developed and advanced over the Miocene land surface. Late movement on thrusts in the crystalline basement uplifted Argentera and caused further uplift of Barrot. Footwall failure in front of basement thrust sheet BII, to form imbricates BIII. tilted the Castellard footwall ramp and caused the Gres d'Annot to downlap in Annot. Probably coeval with the uplift of Argentera, out-of-sequence thrusting decapitated the Chateau Redon dome (Graham, 1983, pers.comm.) in the Pliocene.

6.6.7. Summary

Compressional structuring of S.E. France gave rise to topographic highs within the Vocontian trough as early as the Mid Cretaceous. The whole region emerged above sea level at the end of the Cretaceous: a drainage system was established which transported the material eroded from the developing topography out of Haute Provence. At this stage, the structures were all Pyrenean and aligned east-west, thus enabling rivers to make their way out of the system at the structural culminations.

During the Eocene, the Alpine deformation front propagated into the foreland. The NW-SE and N-S striking structures generated additional topographic relief, and small, closed drainage basins were formed. Rivers were prevented from transporting coarse debris across and out of the region: the Poudingues d'Argens alluvial fans and lacustrine environments were generated.

The earliest Tertiary transgression, recorded by the Calcaires Nummulitiques of the south-east Maritime Alps, is interpreted by this author to represent the initiation of the flexural foreland basin to the eastern Alpine load. It is penecontemporaneous with the development of Alpine thrusts and folds there and to the west. Thus, the basin developed on already shortened and actively shortening continental crust.

For some time, the study region was deprived of sediment and so subsided to deep marine depths. Continued displacement on SW directed thrusts produced a significant basin floor topography which dictated the facies distribution of the shallow marine limestones. The basin was then rapidly filled with siliciclastic turbidites, supplied from the assembled micro-continents to the south. The sediment gravity flows were controlled by the topography in the basin: in places, they were confined in closed troughs on the sea floor; elsewhere, the confinment of high density turbidity currents allowed them to travel further with their coarse sediment load.With time, the turbidite depocentre shifted westwards.

Early in the Oligocene, the Gres d'Annot basin died when the source area rifted from the basin margin and the basin itself was filled with thrust sheets. Deep marine conditions were never re-established in the subsequent S.W. Alpine foreland basins, even though the Provencal basin was deeply dissected by the Oligocene-Miocene Rhone rift system.

6.7. Comments on the nature and importance of the Gres d'Annot sandstone-rich turbidite system.

This study is the first integrated sedimentological and structural study of a sandstone-rich turbidite system **and its substrate.** A new model for the control of turbidite deposition has been proposed which **can be applied and tested.** Like the fan model, the basin topographic control model invokes controls within the area of turbidite deposition itself and does not require reference to the basin margins or source areas which have so often been severely eroded.

The most recent musings on turbidite systems (eg. Mutti, 1986) have recognised that turbidite facies cannot be modelled with reference to sub-environments of deep sea fans alone. It has been realised that the primary control of grain size, bed thickness, the lateral continuity of facies associations, the occurrence of sequences and cycles may be **sediment supply.** The controls of sediment supply have been addressed:

- Sea level changes may dictate whether sediment is stored on continental shelves or transported to the deep marine environments. These authors often invoke eustatic sea level changes (Mutti), although tectonic activity may induce relative sea level changes over large regions.

- How to mobilise the large volumes of sand contained in thick turbidites. Ricci Lucchi (JAPEC course, 1984) proposed that each large volume event may represent the fill of a perched basin, ruptured and emptied as a result of seismic activity.

The hypotheses are good, but largely untestable. The spatial and stratigraphic links between shelf, slope and deep marine sedimentary environments are not good enough in any of the ancient examples cited to date.

The Gres d'Annot and its lateral equivalents remain equally enigmatic in this respect. Biostratigraphic controls are non-existent and fingerprint sedimentary sequences or cycles do not exist in the Gres d'Annot succession. It is difficult to correlate sandstones within one outcrop area and impossible to correlate between outcrop areas. The relative ages of turbidites in the Annot, Col de La Cayolle, Trois Eveches and Grand Coyer outcrop areas cannot be determined. The exposure of sediment deposited on shorelines adjacent to the source area is limited to the St. Antonin syncline: there is no direct evidence that the Gres d'Annot were supplied from this sub-basin.

Therefore, the controls of sediment supply and the hydrodynamics of the sediment gravity flows within the Gres d'Annot basin cannot be tested. Suffice it to say, the eustatic sea level changes recorded by Vail <u>et al</u> (1977) do record a relative sea level drop at the start of the Priabonian to a level slightly higher than present day sea level. Equally, there is a gradual sea level rise throughout the Lutetian to Bartonian: this would correspond well with the transgression in the study region. However, both the Calcaires Nummulitiques and Gres d'Annot Formations are diachronous across the study region: the sea level continues to rise relative to Haute Provence in the Priabonian, evidenced by the Calcaires Nummulitiques and Marnes Bleues Formations of that age in the Barreme syncline (Evans, 1987). In addition Vail and Mitchum record an equally major transgressive-regressive cycle in the period Thanetian-Ypresian of which there is no record in Haute Provence.

The sea level variations recorded in the Lower Tertiary sediments of Haute Provence may be explained with reference to the regional tectonic evolution. The same applies to Provence and the Pyrenees during the Late Cretaceous to Eocene times. In the Southern Pyrenees, shallow marine limestones, akin to the Calcaires Nummulitiques of Haute Provence, demonstrate that the Tertiary transgression occurred in the Palaeocene. This agrees with the tectonic history of the Pyrenees which implies that foreland basins, both north and south of the Pyrenees were initiated at the end of the Cretaceous. In Haute Provence, it has been demonstrated in this thesis that the Alpine thrust system propagated into the European foreland in the Eocene immediately prior to the Tertiary transgression. The transgression may therefore be interpreted to date the initiation of the S.W. Alpine foreland basin.

To return to the Gres d'Annot: the preceeding chapters have amply illustrated what can be done with turbidites systems such as the Gres d'Annot. Without a detailed understanding of the factors controlling the volume and frequency of sediment supply, it has been possible to model the facies distribution of turbidite sandstones in the Gres d'Annot basin.

In section 5.7.3., it was demonstrated that the Gres d'Annot Formation did not constitute the deposits of one or several submarine fans. The sediment gravity flows, for the most part high density turbidity currents, were controlled by a pre-existing and developing basin floor topography. The deposits were laterally continuous: all observed facies changes arose where the currents interacted strongly with the basin slopes. The orientation of the palaeoslope with respect to the direction of flow of the turbidity current dictated the exact nature of the edge effects. To set up and test this topographic control model, it was necessary to fully understand the tectonic history of the basin. Thus, it was possible to identify the structures responsible for basin floor topography.

The model may be applied to other turbidite basins with ease. The methods described in this thesis can be adopted in any field setting and may be adapted to an examination of good seismic reflection data tied to well logs and cores.

Finally, it is worth considering qualitively the questions about the Gres d'Annot that remain unanswered:

i) How much sediment was transported by each turbidity current?

ii) How were large volumes of sediment mobilised to form turbidity currents at regular intervals over a period of approximately 3 million years? The question addresses the nature of the continental environments in the source area and the adjacent shallow marine environments. the surface slopes and the depositional systems along the sediment transport route.

iii) Should any significance be attached to the apparently rapid transition to siliciclastic sedimentation at any one point in the basin?

iv) What exactly happened to the turbidity current as it entered and travelled across the basin?

The first question is in fact the most difficult! The Gres d'Annot facies associations and the percentage of sandstone do not decrease significantly towards the northern limits of the basin. The Gres d'Annot successions in the Peira Cava and Col de La Cayolle outcrop areas are not dissimilar. Yet Peira Cava is 70km nearer the source area. The high density turbidity currents were therefore highly energetic.

This evidence could be used to infer that the basin was "oversupplied" by very energetic and voluminous turbidity currents. The turbidity currents could travel the length of the basin with undiminished vigour, which suggests that they would, if they had been unconstrained, have constructed a fan much larger than the basin allowed.

An alternative suggestion by Barraclough (1987, pers.comm.) is that high density turbidity currents were only able to travel such long distances because they were constrained by basin floor topography. Linear troughs on the sea floor acted as conduits for the turbidity currents. The turbidity currents were prevented from spreading when they hit a break in submarine slope. They were unable to spread and so did not evolve en route to their destination: consequently, deposition was delayed until the current arrived at, and spread into, a sub-basin, like Col de La Cayolle.

Barraclough's hypothesis is an appealing one. It lends strength to the topographic control model; it accounts for the volumes of coarse grained, thickly bedded turbidites in the Gres d'Annot basin because it supposes that individual beds were deposited over limited areas. It does, however, put pay to any simple volume calculations! It is no longer possible to assume that a sandstone observed at Col de La Cayolle had lateral and proximal equivalents elsewhere in the basin.

The second question becomes less of a problem if Barraclough is correct. The volume :f sediment in each turbidity current need not be as great as previously supposed. Tectonic activity created relief in the hinterland, at the coast and within the basin, producing small ponded basins. Repeated filling and emptying of these could have supplied the large, rapid influx of sediment needed to generate the turbidites. Sediment could be cleared from these areas by slope failure or fluidisation caused by earthquakes or exceptional storms. One such perched basin was the St. Antonin basin.

The sharp contact at the base of the Gres d'Annot Formation is diachronous across the Gres d'Annot basin. The depocentre shifted gradually westwards through the Priabonian. At any one point, it is sharp because it is an onlap surface and simply marks the point in time when the basin was filled with turbidites at that localitiy. However, the arrival of the first turbidites sourced from Corsica-Sardinia was almost certainly abrupt, because it coincided with an important tectonic event: the collision of the southerly continental mass with Adria. The tectonic event induced major changes in the sediment transport paths. Similarly, with the onset of rifting at the southern margin of the basin, early in the Oligocene, the southern continental massifs abruptly ceased to provide sediment to the S.W. Alpine foreland basin. This, and the arrival of the Embrun-Ubaye nappes, effectively killed the basin.

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