Architecture of
deep-marine confined sandstone bodies,
Eocene-Oligocene Grès d'Annot Formation,
SE France

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Architecture of deep-marine confined sandstone bodies, Eocene- Oligocene Grès d'Annot Formation, SE France

Abstract

The Tertiary upper Eocene-lower Oligocene Grès d'Annot Formation of southeast France essentially is a sand-rich deep-marine turbidite system deposited in a foreland basin with a structurally complex basin-floor topography, where local basinal highs exceeded 400 m over 8 km. Overall, the Grès d'Annot Formation lacks features typical of submarine fans. Deposition of turbidite sandstones and other sediment gravity flow deposits first occurred within basin topographic lows as a passive fill which progressively buried these features.

Within the southern part of the Grès d'Annot Formation outcrops two separate basin-floor systems are identified on the basis of palaeocurrent dispersal patterns and sandstone onlaps against the basin floor topography. These two systems are the eastern basin-floor system, and the western basin-floor system.

The eastern basin-floor system includes the Grès d'Annot Formation outcrops from Peira Cava, Contes and Menton. The oldest part of the sandstone succession in the eastern system is located at Peira Cava where deposition took place at the base of a local submarine slope as relatively sand-rich deposits which tend to shale out into the more distal parts of the basin. The sandstone succession in the Contes and Menton areas show an upsection change from amalgamated and non-amalgamated sandstone packets to essentially amalgamated sandstones.

The western basin-floor system includes the St Antonin, Entrevaux, Annot and Grand Coyer outcrops of the Grès d'Annot Formation. The St Antonin section comprises three members, each showing an upsection change from thin-bedded to thick bedded turbidites to debris-flow conglomerates, to thin bedded sandstones interpreted as distal shelf/upper-slope storm deposits. The Entrevaux succession shows an upsection change from thin-bedded fine-grained to thicker bedded and coarser grained turbidites. The Annot sandstone succession shows an upsection change from essentially non-amalgamated sandstones to amalgamated sandstone packets and interbedded thin-bedded relatively fine-grained turbidites. The Grand Coyer succession comprises amalgamated turbidite sandstone packets with interbedded sandstone and mudstone packets with small-scale channel development within some of the sandstone/mudstone packets. The western basin-floor system is interpreted as an overall progradation and aggradation of a sand-rich submarine ramp/delta slope into the deeper parts of the western basin floor system.

Comparisons of the Grès d'Annot Formation with the deep-marine sandstone reservoirs of the Palaeogene of the northern North Sea, show a similarity of confinement of sandstones and sandstone lobes and may provide a useful comparative system.
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Introduction and geological setting

1.1 Geological setting

The study area for this thesis is located in the Alpes Maritimes and Haute Provence region of southeast France (Figure 1.1), and encompasses the southwesterly margin of the arcuate western Alpine mountain chain, part of the Alpine-Himalayan orogen extending from Morocco to China. The mountain belt formed as a result of subduction of Tethyan oceanic crust and subsequent collision between the Eurasian and African plates, together with the intervening microplates, driven by the relative northward motion of the African plate. In the western Mediterranean region, the Iberian and Adria-Apulian micro-plates were accreted between the Eurasian and African plates, to form three distinct mountain belts in the Mediterranean region: the Pyrenees, Alps and Apennines.

This study is based on the Tertiary Grès d'Annot Formation which is predominantly composed of deep-marine siliciclastics deposited in the foreland basin to the Pyrenean-Provençal and Alpine mountain belts. The formation is uppermost Eocene-Lower Oligocene in age, although the upper age limit has not been accurately dated. The formation is the youngest of four formations, which together represent the infilling and deepening of the basin through time. Shallowing-upward sequences within the Grès d'Annot Formation are preserved around the margin of the basin, particularly towards the south.
Figure 1.1. Geological map of southeast France showing Tertiary sedimentary outcrops.
Chapter 1: Introduction

1.1.1 Plate tectonic framework

This section is in two parts, firstly a plate scale overview of the interaction of the African and European plates in the Palaeogene, and secondly a more detailed description of the plate tectonics for the western Mediterranean.

In the western Mediterranean region, during the Palaeogene, the two major interacting lithospheric plates were the African and European plates. Dewey et al. (1989) produced kinematic maps based on magnetic anomalies to show the relative movements of Africa with respect to a stationary European plate and these are shown in Figure 1.2.

From 175 Ma to about 118 Ma, the African plate motion relative to Europe was essentially one of sinistral strike-slip (Dewey et al. 1989) which corresponded to the increasing separation of the African from the American plate and the formation of the Central Atlantic Ocean. Between 118-84 Ma, the African plate changed its motion to a more northeasterly compressional regime against the European plate. Also, during this time interval the separation of Iberia from North America was initiated. In the earliest Palaeocene, at around 66.7 Ma, convergence of Africa and Europe slowed and motion became more erratic. In the early Eocene (55.7 Ma), northward plate motion resumed, until the late Miocene, when the motion of Africa was more northwesterly, a motion which remains prevalent.

The first occurrence of collision-related compressional deformation in the western Mediterranean area was in Cretaceous (pre-Senomian) times when folds were developed in the Devouly region of France as E-W trending fold axes in rocks up to Cenomanian age, and which are overlain unconformably by Senonian aged limestone (Flandrin 1966, Deblemas and Lemoine 1970). During the Upper Cretaceous to Lower Tertiary, the Iberian continental plate separated from Newfoundland as a consequence of rifting between Africa and Central America which resulted in the formation of the Central Atlantic Ocean (Figure 1.3 a & b, after Dercourt et al. 1986).
Figure 1.2: Positions of Africa with respect to a stable Europe, from 175 Ma to Present. African plate positions shown at different magnetic anomalies (after Dewey et al. 1969).
Plate configurations at 80, 65, and 35 Ma for the western Mediterranean. Modified after Dercourt et al. (1986).

Hilton 1994
Plate configurations at 20, 10, and 0 Ma for the western Mediterranean. Modified after Dercourt et al. (1986).
Chapter 1: Introduction

As Africa and Iberia moved in a north easterly direction, the Iberian plate was pinned between the African and European plates (Figure 1.3c, Figure 1.4), resulting in the suturing of an ocean whose subduction zone passed from the southern side of the Bay of Biscay, along the north side of the Pyrenees and onto the south side of Corsica and Sardinia (Figure 1.4). At this time a NE-SW trending, southward-dipping, subduction zone existed at Corsica and Sardinia, during which time a wedge of oceanic crust (part of the Balagne Nappe) was obducted onto Corsica. This collision-obduction event was completed by Eocene to earliest Oligocene times. Overall, this plate motion produced sinistral strike-slip compressional forces in the Pyrenean-Provençal mountain range (Siddans 1979, Dercourt et al. 1986), and the uplift which supplied sediments for the Grès d'Annot Formation.

The strike-slip closure of the narrow Pyrenean seaway, between Iberia and Europe, resulted in the formation of the Pyrenean-Provençal mountain range and, towards the north of this Provençal mountain belt, the southwest Alpine Foreland Basin was formed (Figure 1.4), with the mountains to the south providing a source area for the Grès d'Annot Formation.

In the earliest Miocene (Figure 1.3d), extensional tectonics became significant throughout the western Mediterranean; Corsica and Sardinia began to separate from the French mainland, although at some time in their rotational history between Miocene times and the present, they moved independently (Bayer et al. 1973, Barrus 1984, Dewey et al. 1989). Also Menorca and Mallorca began to rotate away from Iberia as discrete microplates (ibid.). This rotation of Corsica and Sardinia towards the Apulian plate resulted first in southeastward subduction of oceanic crust under the Apulian plate and eventual continent-continent collision between the Corsica-Sardinian and Apulian plates to produce the Penninic thrusts and formation of the Apennine mountain belt.

From 18 Ma to the present (Figures 1.3 e, f), the relative motion of Africa with respect to Europe has been convergent. Corsica and Sardinia completed their rotation of about 37°...
Figure 1.4. Palaeo-reconstruction of the western Mediterranean at 38 Ma (after Dewey et al. 1989), showing the extent of the Pyrenean-Provençal mountain belt. Corsica, Sardinia, and the Balearic Isles have been restored to their positions at this time.
Chapter 1: Introduction


1.1.2 - Late Cretaceous - early Tertiary Alpine Flysch

The palaeogeography of Alpine flysch has been the focus of much research e.g. (Studer 1827, Faupl 1978, Trumpy 1980, 1982). There is a consensus that the Alpine flysch accumulated immediately north of the northward subducting Tethyan oceanic crust on the thinned continental crust of the European plate in a foreland basin caused by the Sardinia-Corsica volcanic arc loading the lithosphere. The onset of subduction has been put at between 118-84 Ma (Dewey et al. 1989) so that the change from an arc related setting to a foreland basin post-dated this. Numerous studies have shown that the flysch accumulated in a variety of tectonically active basins. The oldest Alpine flysch was deposited during Aptian-Albian times and continued up to Tertiary Oligocene times (Table 1.1), although the first development of major deep-marine clastic successions occurred from Cenomanian times onwards.

Throughout this thesis, the adopted time divisions and stratigraphic nomenclature is that of Matter et al. (1980), Homewood (1983), and Harland et al. (1989). In summary, the principal flysch units accumulated as follows:

During the **Cenomanian to Conician** (97.0-86.6 Ma), the Simme Flysch of the French and Swiss Pre-alps accumulated, together with the Lombardian Flysch of the Southern Alps, and Reiselberger Sandstones of the Rhenodanubian Flysch in the Austrian Alps.

During the **Conician to Maastrichtian** (86.6-65.0 Ma) there was widespread deposition of the Helminthoid Flysch in the Southern Italian Alps, and the Lombardian Flysch continued to accumulate on the northern margin of Adria.
### Table 1.1: Stratigraphic summary of Alpine Flysch units (modified after Homewood 1983).

<table>
<thead>
<tr>
<th>Series</th>
<th>Stage</th>
<th>Flysch Units</th>
<th>Groups</th>
<th>Formations</th>
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<td>CENOMANIAN</td>
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<td>ALBIAN</td>
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<td><strong>ANNOUIS, CHAMPSAUR</strong></td>
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<td><strong>AIGUILLES D'ARVES</strong></td>
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<td><strong>TARENTAISE</strong></td>
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<td><strong>BRIANÇONNAIS &amp; SUB BRIAN</strong></td>
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<td><strong>AUTAPIE, PARPAILLON</strong></td>
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<td>(HELMINTHOID FLYSCH)</td>
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<tr>
<td><strong>VAL D'ILLIEZ, ALTDORF</strong></td>
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<td><strong>TAVEYANNE, HELVETICS</strong></td>
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<tr>
<td><strong>ULTRAHELVETIC</strong></td>
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<td><strong>SARDONA</strong></td>
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<td><strong>NIESIN</strong></td>
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<td><strong>NORTH PENNINIC MÉLANGE</strong></td>
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<td><strong>PRATTIGAU</strong></td>
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<td><strong>BRIANÇONNAIS &amp; SUB BRIAN</strong></td>
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<td><strong>VOIRONS GURNIGEL</strong></td>
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<td><strong>HELMINTHOID</strong></td>
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<tr>
<td><strong>GETS</strong></td>
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<td><strong>FEURSTATT MÉLANGE</strong></td>
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<td><strong>RHENODANUBIAN</strong></td>
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<td><strong>GREIFENSTEIN</strong></td>
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<td><strong>ZEMENTMERGEL</strong></td>
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<tr>
<td><strong>REISELBERGER</strong></td>
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<tr>
<td><strong>VERSPALA</strong></td>
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<tr>
<td><strong>GOSAU</strong></td>
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<tr>
<td><strong>LOMBARDIAN</strong></td>
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</table>

Dates from Harland et al. (1989).
Chapter 1: Introduction

From Maastrichtian to Palaeocene time (65.0-56.5 Ma), western Alpine successions accumulated, including the sandy Günigel and Schlieren Flysch units, the Autapie and Parpaillon Flysch of the French Alps and the last stages of the Helminthoid Flysch. During Eocene times (56.5-35.4 Ma), deposition in the deeper oceanic basin had ceased with continental slope facies migrating northwards and westwards over trench fill facies and, flysch of the Briançonnais, sub-Briançonnais, Dauphiné and Ultrahelvetics were deposited. The Eocene to Oligocene (56.5-23.3 Ma) witnessed the final phase of flysch sedimentation within the Alps, with sediments derived from the collided southern margin of the closed western Tethys (Homewood & Caron 1982); though the Corsica - Sardinia landmass, to the southwest, was an active source for the Grès d'Annot Flysch (Stanley & Mutti 1968). Other flysch which was deposited in this final collisional stage includes the Grès de Champsaur, the Taveyanne Sandstones and the Altdorf Sandstones of Switzerland. Today “flysch” deposits still accumulate, associated with the on-going Alpine orogeny, as the Mediterranean turbidite systems such as the Rhone and Petite Rhone submarine fans.

1.1.3 Extent of the Tertiary Foreland Basin

The Tertiary Foreland basin in Alpes Maritimes and Haute Provence extends over an area of approximately 6,000 km² (Figure 1.5), although the Tertiary fill does not outcrop across the entire area. The basin is presently bound to the south by the Mediterranean Sea, and to the north by the Hercynian crystalline basement exposed as the Argentera-Mercantour Massif. The eastern side of the foreland basin is bounded by nappes containing Helminthoid Flysch and the western side of the basin is defined by younger alluvial-fluvial Miocene-Pliocene sediments of the Valensole Basin. During deposition of the Grès d’Annot Formation, the Argentia-Mercantour Massif probably was buried by a younger cover and therefore was not an active sediment source area (Ivaldi 1974); it has been interpreted as a younger “pop-up” basement (Fry 1989) i.e. basin inversion structure. To the north of the foreland basin, the contemporaneous Champsaur Basin was being infilled with flysch deposits derived from a
Figure 1.5. Location of the Tertiary outcrops in the foreland basin of SE France.
northwesterly emergent landmass. The southern margin of the foreland basin in Alpes Maritimes and Haute Provence is interpreted to have been along the northern margin of Corsica which, when post-Oligocene rotation has been corrected for, is interpreted as being an emergent landmass and a source area to supply the turbiditic and related sandstones (Stanley & Mutti 1968, Ivaldi 1974).

1.2 Stratigraphy of the Grès d’Annot Foreland basin

The Grès d’Annot foreland basin is located in Alpes Maritimes and Haute Provence region of southeast France (Figure 1.4). The stratigraphy of the basin is relatively simple and shows a succession of Tertiary sediments (Figures 1.6 & 1.7), which represent both the infilling and subsequent deepening of the basin through time. Remnants of the Tertiary Foreland basin sediments are preserved in isolated outcrop areas (Figure 1.5) which tend to form mountain tops. The entire succession is relatively thin and reaches a maximum thickness of less than 1500 metres (Apps 1987). The sediments are fully marine except for the basal Poudingues d’Argens Formation which is alluvial.

The contact of the Tertiary succession with the underlying sediments is marked by a major unconformity, the rocks below the unconformity being Cretaceous-Aptian to Maastrichtian in age (Campredon 1977), and comprising interbedded limestones and shales. The Cretaceous rocks are relatively undeformed, but contain local gentle, open, folds, and are generally blanketed by the Tertiary sediments.

1.2.1 Poudingues d’Argens Formation

The Poudingues d’Argens formation is the oldest in the Tertiary succession. Precise ages are uncertain, but an age range of earliest Eocene to Lutetian has been suggested (Bureau de
### Chapter 1: Introduction

<table>
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<th>Epoch</th>
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<td>Chattian</td>
<td>23.3</td>
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<td>35.4</td>
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<td>Marnes Bleues</td>
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<td>74.0</td>
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<td>Campanian</td>
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<td>Santonian</td>
<td>86.6</td>
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<td>Coniacian</td>
<td>88.5</td>
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<td>limestone exposed along</td>
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<td>90.4</td>
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<td>Aptian</td>
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<td>Barremian</td>
<td>131.8</td>
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Figure 1.6. Stratigraphic column for the Tertiary foreland basin of south east France. Dates from Harland et al. (1989), and ages of formations after B.R.G.M. Entrevaux 1:50000 sheet memoir 945 (1987). Broken lines represent uncertain age.
Figure 1.7. Representative stratigraphic log for the Tertiary Foreland Basin (modified after Ravenne et al. 1987).
Chapter 1: Introduction

Recherches Geologiques et Minières 1980). It is dominated by fluviatile conglomerates with interbedded horizons of mud, silt and sand. The formation reaches thicknesses up to 150 m (Bureau de Recherches Geologiques et Minières 1980) and occurs as alluvial fans derived from palaeotopographic highs of upper Cretaceous limestone, and hence has a limited distribution within the basin. Palaeovalleys are cut into the underlying Cretaceous limestones and are preserved, for example around the town of Peyresque located 9 km NNW of Annecy. Also present within the upper parts of the formation are palaeo-soils, which locally pass into the basal Calcaires Nummulitiques Formation without any obvious break in sedimentation (Bureau de Recherches Geologiques et Minières 1980).

1.2.2 Calcaires Nummulitiques Formation

The Calcaires Nummulitiques Formation forms a laterally extensive and diachronous limestone unit recording a transgression across the foreland basin from the southeast and towards the northwest (Campredon 1977). The older sediments of the formation are found in the south and east of the basin, and are Lutetian in age whilst the youngest occur in the north and west and are Priabonian in age (Campredon 1977). The formation reaches a maximum thickness of 120 m, and is divided into two members: the older Mort d'Homme and the overlying Scafferels members. The Mort d'Homme member reaches thicknesses in the order of 30-50 m and forms a massive bioclastic calcarenite with nummulites and lamellibranchs. It also contains reworked clasts of the Poudingues d'Argens Formation with many of the clasts showing lithophaga borings characteristic of upper shoreface and beach deposits (Apps 1987). The Scafferels member is a fine-grained bioclastic silt/mud limestone which forms a transitional unit with the overlying Marnes Bleues Formation, it reaches thicknesses of 10-80 m and is rich in nummulites and lamellibranchs.
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1.2.3 Marnes Bleues Formation

The Marnes Bleues Formation is dominated by marls, typically composed of between 60-70% calcium carbonate (Ravenne et al. 1987, also Hilton this study), with localised thin (<10 cm thick) calcarenite beds in some areas, and reaches thicknesses in excess of 200 m. There is an abundant microfauna within the formation, which gives an age range of Priabonian to Lower Oligocene (Bureau de Recherche Géologiques et Minières 1980). The formation has a grey/blue colouration, but towards the top there is a light brown unit of limited aerial extent, named the Marnes Brune by Ravenne et al. (1987): it has clastic material present in the form of silt-grade detrital quartz. This youngest unit within the formation represents a shift to a more sandy facies, indicating the onset of deposition of the Grès d'Annot Formation in the deep-marine foreland basin.

1.2.4 Grès d'Annot Formation

The Grès d'Annot Formation is predominantly a sand-rich, sheet-like deep-marine system dominated by beds deposited from turbidity currents and associated sediment gravity flows. The name of the formation is derived from the town of Annot where spectacular 350 m high sandstone cliffs dominate this area. Local names for the sandstones have been used, such as the Grès de Peïra Cava by Bouma (1959, 1962). For the purposes of this study, all sandstones within the study area with same age as those in Annot are referred to as the Grès d'Annot Formation. The main outcrops of the Grès d'Annot Formation are located in the areas of: Annot, St Antonin, Menton, Contes, Peïra Cava, Grand Coyer, Col de la Cayolle and Trois Evêchés (Figure 1.5).

The Grès d'Annot Formation attains a maximum preserved thickness of about 1000 m, and contains only local amounts of shallow-marine deposits, for example as seen at St. Antonin and Col de la Cayolle (Stanley 1980 and Sinclair 1993 respectively) (Figure 1.5).
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turbidite beds range in thickness from a few millimetres to >10 m and are generally capped by silty mudstones. The formation is sand rich with an average sandstone/mudstone ratio of 3:2. The sandstones are subarkosic to arkosic in composition, and predominantly contain rock fragments including gneissic basement material, granites, reworked Helminthoid Flysch clasts (Ivaldi 1974), and andesitic volcanics. The sandstones, to a lesser extent contain reworked turbidite beds especially as shale clasts, clasts of the underlying Marnes Bleues and Calcaires Nummulitiques Formations, and small amounts of basaltic material.

As well as sandstone turbidites, there are two distinct conglomerate horizons which are composed of pebble clasts in a matrix of sand-grade material, the clasts being generally matrix-supported, i.e., as pebbly sandstones. These pebble beds occur at the Col de la Cayolle and Lac d'Allos outcrops, and attain a maximum thickness of approximately 10 m. The pebbly sandstones provide useful marker horizons for correlation within the two areas (Stanley et al. 1978).

The contact of the Grès d’Annot Formation with the underlying Marnes Bleues Formation may be gradational, erosional or onlapping. Dating of this contact has been made by Mougin (1978) using foraminifera in the marls immediately below the sandstone contacts, and gives a Priabonian age. Towards the top of the succession, the formation passes into a fine-grained unit named the “Marnes Brune supérieure” by Ravenne et al. (1987), which probably represents the end of deposition of the typical Grès d’Annot facies associations.

The upper stratigraphic contact of the formation is rarely preserved, and is generally represented by the present-day erosion surface. In some areas, however, the Schistes-à-blocs unit is present above (Trois Evêchés, Peña Cava, and Col de la Cayolle). At the Trois Evêchés outcrops, it can be observed to erode into the underlying sandstone formation. In other areas the Nappes de Embrunais-Ubaye cuts across the top of the formation.
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1.2.5 Schistes-à-bloc unit

The Schistes-à-bloc unit is the youngest of the Palaeocene series in the foreland basin of Alpes Maritimes and Haute Provence. The Schistes-à-bloc represents a tectono-sedimentary unit formed at the front of the advancing internal nappes, namely the Nappes de Embrunais-Ubaye (Wazi et al. 1985). The Schistes-à-bloc is a sedimentary breccia composed of angular blocs of 10-100 cm in size, set in a pelitic matrix also containing a "micro-breccia"(Wazi et al. 1985). Stratification within the unit is generally absent but where present is poorly marked. The unit comprises Palaeocene sediments, including marls of the Marnes Bleues Formation, transitional facies of this and the overlying Grès d'Annot Formation, and a fine-grained facies of the Grès d'Annot Formation. Minor amounts of nummulitic limestone and reworked Helminthoid flysch also occurs. The Schistes-à-bloc has an erosional surface with the underlying Grès d'Annot Formation, and represents erosion from gravity sliding resulting in the formation of olistostromes.

1.2.6 Nappes de Embrunais-Ubaye

The overlying Nappes de Embrunais-Ubaye are located to the east of the study area and comprise tectonically interleaved Autapie and Helminthoid Flysch, and elements of a Mesozoic-Palaeogene succession termed the sub-Briançonnais which shows a tectonic transport direction towards southwest (Fry 1989). Major (>100 km) southwest thrusting has effected the Embrunais-Ubaye terrain, with thrusting from depth causing considerable crustal shortening (Fry 1989).
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1.3 Structure

The structure of the Alpine mountain belt is complex and remains controversial (Graham 1978, Siddans 1979, Fry 1989). A brief account of the tectonic history of the study area is presented below:

The earliest period of deformation in the Mesozoic parts of the sub-Alpine mountain chains affected Cretaceous rocks up to Cenomanian age. These rocks show tight upright east-west trending folds that are overlain unconformably by Senonian limestones (88.5 Ma-65 Ma) (Siddans 1979). In southeast France, essentially north-south compressional forces resulted from Iberia colliding with Europe to produce the Pyrenean-Provençal mountain range (Figure 1.4). This north-south compression produced east-west trending fold axes overlain by Nummulitic limestones of Priabonian age. Some E-W thrusts of this age exist and belong to the Pyrenean-Provençal deformation phase (Lemoine 1972) (Figure 1.8). The Alpine deformation front impinged on Provence from the east as early as the late Eocene (Siddans 1979, Apps 1987). Faults produced during the Pyrenean-Provencal period of deformation were reactivated to accommodate southwest Alpine displacements, as well as producing arcuate fold and fault structures, and nappe emplacement (Figure 1.8). Extensive Alpine deformation is interpreted by Siddans (1979) to have occurred in the foreland basin during late Oligocene to post Miocene times, with Miocene rocks overthrust by older rocks. The Nappes de Embrunais-Ubaye and the Digne thrust sheet were emplaced during these times. Figure 1.8 shows the main nappes of the Digne and Embrunais-Ubaye regions, with the Digne nappes emplaced later in the Pliocene (Siddans 1979). The formation of these nappes has been ascribed to gravity sliding (Lemoine 1973), but Fry (1989) suggested that the Embrunais-Ubaye Nappes have a geometry indicative of thrusting from depth. The pattern of fold axial traces (Figure 1.8) shows that from Nice to Castellane and Die, the axial traces vary from north-south to east-west. Around Nice, Castellane, and Die, there is an arcuate pattern of folds. The formation of these arcuate patterns has been explained as a response of the cover-
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Figure 1.8. Sketch map of structural trends in southeastern France (redrawn from Graham 1978), showing structures produced during the Pyrenean-Provençal and Alpine phases of deformation.
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rocks to sinistral displacements along NE-SW trending transcurrent faults in the basement (Graham 1978), although precise dates for the folding remain unresolved.

Outcrops of the Grès d'Annot Formation are preserved in gentle dipping synclines formed from Alpine compression which may have corresponded to earlier formed topographic lows on the sea floor, and which acted as depocentres during deposition of the Grès d'Annot Formation.

1.4 Previous research

The Tertiary Foreland basin in the Alpes Maritimes and Haute Provence region of southeast France has been the focus of much research on sedimentation and tectonics. One of the earliest accounts of the Grès d'Annot Formation was performed by Gras (1840) based on a survey of the rocks in the Annot area. Other early work includes that of Kuenen et al. (1957) who documented the flysch deposits in the French and Italian Maritimes Alps. Gubler's (1958) study addressed the problem of possible source areas for the Tertiary formations in the French Alps, including the Grès d'Annot flysch, concluding that this was derived from local sources such as the Maures-Esterel Massif located towards the south of the foreland basin.

Arnold Bouma also made significant contributions to the understanding of the Grès d'Annot Formation where, based on detailed measurements in 1061 beds, he defined the five fold (T1–5) divisions of a turbidite bed based on the Grès d'Annot Formation at Petra Cava (Bouma 1959, 1962). To define the Bouma sequence, he used a detailed graphical approach to sedimentary logging.

Daniel Stanley has contributed much to the understanding of the Grès d'Annot Flysch, for example, by providing a detailed account of lateral variations of mineral composition and texture, and depositional processes and possible source areas for the sandstones (Stanley Hilton 1994
1961). Later work by Stanley (1963, 1964, 1965) examined in detail the petrography, particularly the heavy mineral distribution, within turbidite beds (e.g., Stanley 1963), and spatial distribution of heavy minerals within the basin (e.g., Stanley 1964, 1965). He concluded that heavy mineral assemblages vary over considerably short distances, and he stressed the importance of local supply in the accumulation of flysch (ibid.).

In a synthesis paper, Stanley & Bouma (1964) interpreted the Grès d'Annot Formation at Annot as a north-south trending submarine canyon-valley/channel. Later, Stanley (1967) studied outcrop areas at Annot, Contes, and Menton and compared them with the Gully Submarine Canyon on the continental margin off Nova Scotia, Canada. Stanley (1967) noted similarities between their geometry, stratification, texture and sedimentary assemblages, and concluded that these outcrop areas were ancient examples of submarine canyons. Stanley (1974) interpreted the Annot sandstones from Annot, Contes and Menton as being chiefly transported by grain flow and fluidised flow, with transport by debris flows and turbidity current flow to be a minor component at these localities.

The provenance of the Grès d'Annot Formation has been the focus of much research. Ivaldi (1974) used quartz thermoluminescence to fingerprint source rocks in the area, and he concluded that the material was sourced from Hercynian basement massifs similar to those on Corsica and Sardinia, and the Maures-Esterel Massif, and from overthrust Helminthoid Flysch located towards the north of the study area. He also confirmed that the Argentera-Mercantour Massif, located towards the north, was not a sediment source during the deposition of the Grès d'Annot Formation and was, therefore, probably totally submerged at that time (ibid.).

Stanley et al. (1978) also suggested that there is a continuum from sediment slumping and sand-flow to turbidity-current flow. He further suggested that the Annot, Contes, and Menton outcrops of the Grès d'Annot Formation show a more proximal facies than those seen in other
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sub-basins located towards north, and that the outcrops were most likely to be submarine valleys (ibid.).

Cremer (1983) gave an account of the Grès d'Annot Formation from the Annot, Col de la Cayolle and Chalufy - Trois Evêchés outcrop areas, and compared it with the Cap Ferret muddy turbidite fan located in the Bay of Biscay. Cremer (1983) described facies changes seen in vertical section from the Grès d'Annot Formation as arising from lateral migration of channels.

With the advent of the submarine fan models (Normark 1970, Mutti & Ricci Lucchi 1972), many interpretations of the Grès d'Annot Formation outcrops have attempted to fit the features seen in this formation into the fan model. For example, Bouma & Coleman (1985) interpreted the Peïra Cava turbidite system as the remnants of laterally migrating channel fills and overbank deposits.

Apps (1987) produced a detailed account of the tectono-sedimentary history of the Tertiary fill and evolution of the Grès d'Annot foreland basin. Apps (1987) also examined the onlap surfaces of the Grès d'Annot Formation in detail, and modelled the tectonic and sedimentary evolution of the Grès d'Annot basin, showing that the evolving palaeotopography was initiated as early as the Eocene, thereby controlling the depocentres, sediment thickness and facies changes, and being associated with multiple internal unconformities.

Ravenne et al. (1987) compiled data from previous local field surveys on the Grès d'Annot Formation undertaken by the Institut Français du Pétrole and the École Nationale Supérieure du Pétrole et des Moteurs. This synthesis emphasised features such as sedimentary onlaps, and basin fills, comparing them with features observed on seismic sections (ibid.).
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1.5 Research aims

The aims of this research are as follows:

(i) Document the facies within the Grès d'Annot Formation deep-marine system and interpret their depositional processes.

(ii) Document the three-dimensional architecture of the Grès d'Annot Formation deep-marine sheet-like system, on scales of metres to hundreds of metres by using detailed sedimentary logs, line drawings taken from photographs, and photo-montages.

(iii) Produce a palaeo-reconstruction for the Grès d'Annot formation deep-marine system during the deposition of the formation and, hence, derive a model that can explain the lack of typical submarine fan features. There is, to date, no detailed depositional model of this type in the literature.

(iv) Attempt to correlate between major outcrop areas and hence improve the understanding of the Grès d'Annot system.

(v) Investigate possible flow reflection/deflection of turbidity currents approaching palaeoslopes.

(vi) Use vitrinite reflectance to determine the burial history of the Grès d'Annot Formation deep-marine system.
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1.6 Thesis structure

The thesis is divided into seven chapters. Chapter 1 introduces the study area and the aims of this study. Chapter 2 describes the facies observed in the Grès d'Annot Formation and an interpretation of their depositional processes is presented. Chapters 3, 4, and 5 describe the sandstone architecture and facies of the various study areas, and attempt to stratigraphically link-up the isolated outcrop areas. Chapter 3 includes the Grès d'Annot Formation outcrops at St Antonin, Entrevaux and Annot. Chapter 4 gives a detailed account of the Grès d'Annot Formation from the Grand Coyer and Montagne de Chalufy outcrop areas. Chapter 5 includes Grès d'Annot Formation outcrops at Peïra Cava, Contes, and Menton. Chapter 6 summarises the sub-basin fills within the larger foreland basin and their relationship to each other, and compares the Grès d'Annot Formation with subsurface and outcrop analogues. Chapter 7 presents and discusses the vitrinite reflectance data from the Grès d'Annot Formation, and briefly documents the sandstones petrography.

1.7 Logistics

The study of the Grès d'Annot Formation was carried out over three field seasons. The first was for 24 days in September 1990. The second and third field seasons were for a duration of four months each, between the beginning of June until the end of September in 1991 and 1992.

Most localities of the Annot sandstones are accessible by car and within easy walking distance. Localities in the Grand Coyer and Chalufy areas require strenuous efforts to reach them, with walks of 1 to 2 hours duration. In the Grand Coyer region a forest track, accessible by car, leads you into the northern part of the outcrop; the forest track can be found by heading south out of Allos Colmars towards Villars Colmars and turning immediately left on leaving the town. The northern outcrops in the Annot area (Tête du Ruch area) can be reached
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by a forestry track via the village of Méailles providing permission has been obtained from
the Office National de Forestier.

Laboratory work included the measurement of vitrinite reflectance from prepared
carbonaceous samples obtained from turbidites. A cursory thin section study was also
performed.
Chapter 2: Sedimentary facies

2.1 Introduction

2.1.1 Facies classification schemes

One of the most commonly used facies classification schemes in the literature is that of Mutti & Ricci Lucchi (1972, 1974, 1975, 1978, Mutti 1977); their classification scheme comprises seven facies distinguished on the basis of grain size, bed thickness, amount of shale, bed amalgamations and sedimentary structures. Specific facies models for resedimented conglomerates of turbidite associations have been presented by Walker (1975), in a model based on the presence or absence of a preferred clast fabric, stratification, and inverse and/or normal graded bedding. A classification scheme for fine-grained deep-water deposits has been presented by Stow & Piper (1984) which is based on modern and ancient deposits. Other schemes exist such as that of Piper (1974), and Carter (1975).

A more recent classification scheme for deep-marine sediments was set up by Pickering et al. (1986, 1989), based on the earlier facies classification scheme of Mutti & Ricci Lucchi (1972). The classification scheme of Pickering et al. (1986, 1989) was developed to integrate data both from the modern and ancient record, and to include facies which do not fit into the Mutti & Ricci Lucchi (1972, 1978) classification scheme. Another, more recent, classification of sediment gravity flow deposits has been presented by Ghibaudo (1992), but which utilises prenouns rather than letters and numbers. All the schemes described above may be used for the field description of deep-water sediments and sedimentary rocks.
Chapter 2: Sedimentary facies

Deep-water deposits may also be defined in the terms of "architectural elements" (cf. fluvial models of Miall 1985). A classification scheme based on facies associations and three-dimensional geometry (including orientation) for deep-marine sediments in the modern and ancient record, is presented by Pickering et al. (1995), based on suggestions made by Miall (1989).

2.2 Facies classification for the Grès d'Annot Formation

For convenience, in this study the facies classification scheme of Pickering et al. (1986, 1989) is adopted (Table 2.1, Figure 2.1). In the classification scheme of Pickering et al. (1986, 1989), seven facies classes are defined largely on: (a) texture of gravely, sandy and silty divisions of beds, (b) relative thickness of mud interbeds or caps, and (c) internal organisation for Facies Class F and composition of Facies Class G. Facies Classes A-E are divided into disorganised and organised facies groups, each facies group is subdivided into facies (Table 2.1, Figure 2.1).

In this study, not all facies described by Pickering et al. (1986, 1989) are present in the Grès d'Annot Formation, and are therefore not described here. Facies which are present in the thesis study area are specifically described using field examples from the Grès d'Annot Formation. The description of these facies differs in some cases from the descriptions presented in the published classification scheme of Pickering et al. (1986, 1989).
<table>
<thead>
<tr>
<th>Class</th>
<th>Group</th>
<th>Facies</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>Group Facies</td>
<td>Gres d'Annot Formation. (Modified from Pickering et al. 1986)</td>
</tr>
<tr>
<td>A</td>
<td>Gravels, muddy gravels, gravelly muds, pebbly sands, ≥5% gravel</td>
<td>Table 2.1 List of facies classes, groups and facies that are found in the Grès d'Annot Formation. (Modified from Pickering et al. 1986)</td>
</tr>
<tr>
<td>A1</td>
<td>Disorganised gravels, muddy gravels, gravelly muds and pebbly sands</td>
<td>Table 2.1 List of facies classes, groups and facies that are found in the Grès d'Annot Formation. (Modified from Pickering et al. 1986)</td>
</tr>
<tr>
<td>A1.1</td>
<td>Disorganised gravel</td>
<td>Table 2.1 List of facies classes, groups and facies that are found in the Grès d'Annot Formation. (Modified from Pickering et al. 1986)</td>
</tr>
<tr>
<td>A1.4</td>
<td>Disorganised pebbly sand</td>
<td>Table 2.1 List of facies classes, groups and facies that are found in the Grès d'Annot Formation. (Modified from Pickering et al. 1986)</td>
</tr>
<tr>
<td>A2</td>
<td>organised gravels and pebbly sands</td>
<td>Table 2.1 List of facies classes, groups and facies that are found in the Grès d'Annot Formation. (Modified from Pickering et al. 1986)</td>
</tr>
<tr>
<td>A2.2</td>
<td>Inversely graded gravel</td>
<td>Table 2.1 List of facies classes, groups and facies that are found in the Grès d'Annot Formation. (Modified from Pickering et al. 1986)</td>
</tr>
<tr>
<td>A2.3</td>
<td>Normally graded gravel</td>
<td>Table 2.1 List of facies classes, groups and facies that are found in the Grès d'Annot Formation. (Modified from Pickering et al. 1986)</td>
</tr>
<tr>
<td>A2.5</td>
<td>stratified pebbly sand</td>
<td>Table 2.1 List of facies classes, groups and facies that are found in the Grès d'Annot Formation. (Modified from Pickering et al. 1986)</td>
</tr>
<tr>
<td>A2.7</td>
<td>Normally graded pebbly sand</td>
<td>Table 2.1 List of facies classes, groups and facies that are found in the Grès d'Annot Formation. (Modified from Pickering et al. 1986)</td>
</tr>
<tr>
<td>A2.8</td>
<td>Graded-stratified pebbly sand</td>
<td>Table 2.1 List of facies classes, groups and facies that are found in the Grès d'Annot Formation. (Modified from Pickering et al. 1986)</td>
</tr>
<tr>
<td>B</td>
<td>Sands, ≥80% sand grade, &lt;5% pebble grade</td>
<td>Table 2.1 List of facies classes, groups and facies that are found in the Grès d'Annot Formation. (Modified from Pickering et al. 1986)</td>
</tr>
<tr>
<td>B1</td>
<td>Disorganised sands</td>
<td>Table 2.1 List of facies classes, groups and facies that are found in the Grès d'Annot Formation. (Modified from Pickering et al. 1986)</td>
</tr>
<tr>
<td>B1.1</td>
<td>Thick/medium-bedded, disorganised sands</td>
<td>Table 2.1 List of facies classes, groups and facies that are found in the Grès d'Annot Formation. (Modified from Pickering et al. 1986)</td>
</tr>
<tr>
<td>B2</td>
<td>Organised sands</td>
<td>Table 2.1 List of facies classes, groups and facies that are found in the Grès d'Annot Formation. (Modified from Pickering et al. 1986)</td>
</tr>
<tr>
<td>B2.1</td>
<td>Parallel-stratified sands</td>
<td>Table 2.1 List of facies classes, groups and facies that are found in the Grès d'Annot Formation. (Modified from Pickering et al. 1986)</td>
</tr>
<tr>
<td>B2.2</td>
<td>Cross-stratified sands</td>
<td>Table 2.1 List of facies classes, groups and facies that are found in the Grès d'Annot Formation. (Modified from Pickering et al. 1986)</td>
</tr>
<tr>
<td>C</td>
<td>sand-mud couplets and muddy sands, 20-80% sand grade, &lt;80% mud grade</td>
<td>Table 2.1 List of facies classes, groups and facies that are found in the Grès d'Annot Formation. (Modified from Pickering et al. 1986)</td>
</tr>
<tr>
<td>C1</td>
<td>disorganised muddy sands</td>
<td>Table 2.1 List of facies classes, groups and facies that are found in the Grès d'Annot Formation. (Modified from Pickering et al. 1986)</td>
</tr>
<tr>
<td>C1.1</td>
<td>Poorly sorted muddy sands</td>
<td>Table 2.1 List of facies classes, groups and facies that are found in the Grès d'Annot Formation. (Modified from Pickering et al. 1986)</td>
</tr>
<tr>
<td>C2</td>
<td>Organised sand-mud couplets</td>
<td>Table 2.1 List of facies classes, groups and facies that are found in the Grès d'Annot Formation. (Modified from Pickering et al. 1986)</td>
</tr>
<tr>
<td>C2.1</td>
<td>Very thick/thick-bedded sand-mud couplets</td>
<td>Table 2.1 List of facies classes, groups and facies that are found in the Grès d'Annot Formation. (Modified from Pickering et al. 1986)</td>
</tr>
<tr>
<td>C2.2</td>
<td>Medium bedded sand-mud couplets</td>
<td>Table 2.1 List of facies classes, groups and facies that are found in the Grès d'Annot Formation. (Modified from Pickering et al. 1986)</td>
</tr>
<tr>
<td>C2.3</td>
<td>thin bedded sand-mud couplets</td>
<td>Table 2.1 List of facies classes, groups and facies that are found in the Grès d'Annot Formation. (Modified from Pickering et al. 1986)</td>
</tr>
<tr>
<td>C2.4</td>
<td>Very thick/thick-bedded, mud dominated sand-mud couplets</td>
<td>Table 2.1 List of facies classes, groups and facies that are found in the Grès d'Annot Formation. (Modified from Pickering et al. 1986)</td>
</tr>
<tr>
<td>D</td>
<td>Silts, silty muds, and silt-mud couplets, &gt;80% mud, ≥40% silt, 0-20% sand</td>
<td>Table 2.1 List of facies classes, groups and facies that are found in the Grès d'Annot Formation. (Modified from Pickering et al. 1986)</td>
</tr>
<tr>
<td>D2</td>
<td>organised silts and muddy silts</td>
<td>Table 2.1 List of facies classes, groups and facies that are found in the Grès d'Annot Formation. (Modified from Pickering et al. 1986)</td>
</tr>
<tr>
<td>D2.3</td>
<td>Thin regular silt and mud laminae</td>
<td>Table 2.1 List of facies classes, groups and facies that are found in the Grès d'Annot Formation. (Modified from Pickering et al. 1986)</td>
</tr>
<tr>
<td>F</td>
<td>Chaotic deposits</td>
<td>Table 2.1 List of facies classes, groups and facies that are found in the Grès d'Annot Formation. (Modified from Pickering et al. 1986)</td>
</tr>
<tr>
<td>F2</td>
<td>Contorted and disturbed strata</td>
<td>Table 2.1 List of facies classes, groups and facies that are found in the Grès d'Annot Formation. (Modified from Pickering et al. 1986)</td>
</tr>
<tr>
<td>F2.1</td>
<td>Coherent folded and contorted strata</td>
<td>Table 2.1 List of facies classes, groups and facies that are found in the Grès d'Annot Formation. (Modified from Pickering et al. 1986)</td>
</tr>
</tbody>
</table>
**Figure 2.1. Facies classification scheme of Pickering et al. (1986, 1989).**
Chapter 2: Sedimentary facies

2.2.1 Facies Class A

Facies Class A forms the coarsest grained deposits of the Grès d'Annot Formation, and occurs throughout the sub-basins studied, and is most abundant in the St Antonin area. Facies within this class contain > 5% pebble-grade or coarser material (Pickering et al. 1986, 1989). A mud matrix may be present in Facies A1.4, indicating a poorly sorted sediment, typical of the depositional process. Clasts usually comprise sub-rounded to rounded granitic, gneissic or rhyolitic material, with maximum observed clast sizes up to 0.5 m by 0.5 m. Intraclasts up to 0.5 m by 2 m are also observed within some beds.

2.2.1.1 Facies Group A1-disorganised conglomerates

2.2.1.1(a) Facies A1.1-disorganised conglomerates

Bed thicknesses are typically from 0.3-4 m, and beds may be laterally continuous in outcrop, or lens-like over distances of 100-400 m. Grain size is from granule to coarse pebble grade, maximum observed clast size is up to 0.5 m in diameter. The base of beds may be planar or erode into the underlying sediment (Figure 2.2a). Deposits are invariably clast supported with a matrix of muddy or clean sandstone which may pass vertically up into a matrix-supported conglomerate of Facies A1.4. This facies may form the basal parts of composite beds also showing Facies B1.1, B2.1 and C2.1.

Transport process: high-concentration turbidity currents or sandy debris flows.
Depositional process: flow freezing with hindered settling on decreasing bottom slopes due to intergranular friction and cohesion.

Hilton 1994
Figure 2.2 a and b. a) Facies A1.1 disorganised gravel with clasts up to 30 cm in diameter, St Antonin. b) Facies A1.1 disorganised gravel with clasts up to 2 cm in diameter, Contes.
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2.2.1.1(b) Facies A1.4-disorganised pebbly sandstone

This facies comprises an apparently random dispersion of granule-grade and coarser clasts in a sand matrix (Figure 2.3a); mud may be present within the sand matrix giving the sandstone a dirty appearance (Figure 2.3b). Such mud-rich sandstones may have accumulated from poorly sorted sediment dispersions. Some beds of Facies A1.4, contain large reworked clasts of intrabasinal material, such as marl and sandstones up to 0.7 m by 0.3 m in size, with angular to subangular clasts, with rounding of clasts possibly indicating a degree of traction transport (Figure 2.3c and 2.3d). Bed thicknesses up to 10 m are recorded, but mostly beds are generally in the order of 1-4 m thick. Beds may show marked lateral thickness variations when traced across outcrop and some are lenticular over distances of 200-300 m (Enclosure 5.5D). The base of beds are essentially planar, although in some cases local shallow scouring is observed. The tops of beds may have an irregular surface, probably due to thickness variations in the original flow. Grain sizes within beds may change laterally between sandy and muddy sediments over distances of 5 m, giving a gradation between Facies A1.4 and Facies C1.1.

Transport process: high- to very high-concentration turbidity currents, mud-rich sandy debris flows, or as slide deposits, possibly moving on relatively steep intrabasinal slopes, with some degree of traction transport for larger intrabasinal clasts which show rounding.
Depositional process: rapid deposition of all grain sizes together under hindered settling due to high intergranular friction as flows decelerate.

2.2.1.2 Facies Group A2-organised gravels and pebbly sands

2.2.1.2(a) Facies A2.2-inversely graded conglomerate

Beds of Facies A2.2 up to 1 m thick occur predominantly in the St Antonin area, in some beds the inverse grading tends to be stepwise rather than gradational (Figure 2.4a). Beds appear
Figure 2.3 a and b. a) Facies A1.4 disorganised pebbly sand, Contes. b) Facies A1.4 disorganised pebbly sand with clasts up to 20 cm in diameter; boot for scale, Col de la Cayolle.
Figure 2.3 c and d. c) Facies A1.4 disorganised pebbly sand, Peira Cava. d) Facies A1.4 disorganised pebbly sand, photo from same locality as Figure 2.3c; note large clasts of reworked muds, tape is 30 cm long.
Figure 2.4 a and b. a) Facies A2.2 inversely graded gravel with erosive base, way-up towards top right, St Antonin. a) Facies A2.3 normally graded gravel, way-up towards top left, Contes.
continuous over outcrop and may show local erosion at their base. The top of some beds may pass vertically into coarse- to medium-grained sandstones forming composite facies, which may have been deposited during the same event or represent deposition from a subsequent turbidity current.

**Transport process:** high-concentration turbidity currents  
**Depositional process:** rapid deposition from a concentrated traction carpet/dispersion near the base of the bed due to increased intergranular friction. The inverse grading is caused by intense grain interaction due to strong dispersive pressure.

2.2.1.2(b) Facies A2.3-normal graded conglomerate

The facies generally forms beds up to 1 m thick with clasts up to coarse pebble grade (Figure 2.4b), and which may form the basal part of Facies C2.1 or grade vertically up into Facies A2.7. The facies may be found infilling large scours at the base of turbidites or at the base of beds that are planar. Normal grading may be stepwise with the absence of a particular grain size; a medium/coarse conglomerate may pass abruptly up into a coarse-/medium-grained sandstone. The absence of a particular grain size may indicate sediment bypassing and deposition in a more distal part of the basin.

**Transport process:** high-concentration turbidity currents.  
**Depositional process:** grain-by-grain deposition from concentrated sediment suspension.

2.2.1.2(c) Facies A2.5-stratified pebbly sandstone

Facies A2.5 occurs in beds up to 2 m thick, with clast sizes up to medium-/pebble-grade material (Figure 2.5a). Parallel-stratified conglomerate bands up to 0.2 m thick alternating with
Figure 2.5 a and b. a) Facies A2.5 stratified pebbly sand, Contes. b) Facies A2.7 normally graded pebbly sand infilling a scour, Contes.
Figure 2.5 c and d. c) Facies 2.7 normally graded pebbly sand at the base of a bed infilling scours, Contes. d) Facies A2.8 graded-stratified pebbly sand, Contes.
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medium- and coarse-grained sandstones. Pebble clasts within the conglomerate bands are essentially matrix supported, although towards the centre of the pebble bands there may be clast-supported layers.

**Transport process**: high-concentration turbidity currents.

**Depositional process**: grain-by-grain deposition from concentrated sediment dispersion, possibly associated with traction transport as a bed load.

2.2.1.2(d) Facies A2.7—normally graded pebbly sandstone

Facies A2.7 occurs in beds up to 1 m thick, with maximum basal grain sizes to medium- / coarse-grade pebbles. The base of beds are either planar or erosive with the graded pebbly sandstone infilling erosion features (Figures 2.5b and c). Where the base of this facies is erosive, bed amalgamation is typical. This facies may form the basal unit of thicker beds which then pass vertically into either Facies C2.1 or B2.1 (Figure 2.5c).

**Transport process**: high concentration turbidity currents.

**Depositional process**: grain-by-grain deposition from high-concentration sediment dispersion, with no significant traction transport.

2.2.1.2(e) Facies A2.8—normal graded stratified pebbly sandstone

Facies A2.8 occurs in the Grès d'Annot Formation in beds up to 3 m thick, with a maximum grain size of medium pebble-grade material. Many beds show an overall finning-upward from pebble-grade material, the matrix may also show a finning upward trend from very coarse- to medium- / fine-grained sandstone (Figure 2.5d). Pebble grade material may form discrete matrix-supported pebble bands up to 0.2 m thick; pebble bands at a higher level in the bed may
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record a larger grain size than pebbly bands located towards the base. Where observed, the grain size of pebble bands towards the top of a bed may be up to 1 cm in diameter, while the grain size of pebble bands below may only be up to 0.5 cm in diameter; this may reflect a short lived increase in flow velocity during the overall waning of turbidity current. The sandstone matrix contains stratification as alternations between very coarse, coarse, and medium-grained sandstones, on a <5 cm scale (Figure 2.5d). This parallel stratification of the sandstone matrix is similar to the parallel stratification observed in Facies B2.1.

Transport process: high-concentration turbidity currents, which become more dilute with time with respect to a fixed point on the sea floor.
Depositional process: grain-by grain deposition from suspension, with initial deposition involving little traction transport. At higher levels in the deposit, grains underwent traction transport as a bedload to produce parallel-stratification.

2.2.2 Facies Class B

Facies Class B comprises sandstone beds with <20% mud and silt matrix <0.0625 mm in diameter, and <5% pebble grade material. This class is divided into disorganised and organised facies groups.

2.2.2.1 Facies Group B1-disorganised sandstones

Disorganised sandstones have been described from deep-marine deposits (e.g., Stauffer 1967, Mutti & Ricci Lucchi 1972, Carter & Lindqvist 1975, Hiscott 1980, Lowe 1982, amongst others), and are interpreted as forming from rapid en-masse deposition from a single surge-type flow (Middleton & Hampton 1976). More recently, Kneller & Branneney (1994) have suggested that these deposits may form by deposition involving the gradual aggradation beneath sustained...
steady or near-steady flows involving a flow boundary that is dominated by hindered settling. This latter explanation does not require the instantaneous deposition of beds up to many metres thick. Distinguishing between both processes in the rock record is not possible.

2.2.2.1(a) Facies B1.1-thick/medium-bedded disorganised sandstones

Beds of Facies B1.1 occurs in beds up to 5 m thick, and with grain sizes of very coarse/medium-grained sandstones, rare pebble grade material and mud clasts in some cases. The bounding surfaces to beds are essentially planar, typically with basal scours and shallow erosion features. Grading is generally absent, but where present, it is subtle with grading from coarse- to medium-grained sandstone. Fluid escape features commonly occur as dish structures and with rare pipe structures (Figure 2.6b).

Transport process: high-concentration turbidity currents
Depositional process: rapid mass deposition due to intergranular friction in a concentrated dispersion near the bed, or gradual aggradation beneath sustained steady or near-steady flows involving a flow boundary that is dominated by hindered settling.

2.2.2.2 Facies Group B2-organised sandstones

Facies Group B2 includes those sandstones which show sedimentary structures that are not readily explicable as part of the Bouma sequence. This facies group shows features of sub-facies B1 and B2 of Mutti & Ricci Lucchi (1972).
Figure 2.6 a and b. a) Facies B1.1 disorganised sand infilling a large scour with a pebble lag at the base, note dish structures midway up the bed, the section is ~3.5 m high, Annot. b) Facies B1.1 disorganised sand with dish structures, Peira Cava.
2.2.2.2(a) Facies B2.1-parallel-stratified sandstones

Facies B2.1 in the Grès d’Annot Formation occur as beds up to 3 m thick, with parallel stratification in the order of 5-10 cm thick comprising inversely graded units (Figure 2.7a and c). Grain sizes in the inversely graded units range from medium- to coarse-grained sandstone with rare granule-grade material in the coarser grained portion. These parallel-stratified units can be traced over 10 m in a proximal-distal direction. Within these parallel-stratified beds, local scours up to 15 cm deep and 25 cm long occur, and these scours may show a draped partial infill also with small-scale cross-stratified sandstone infill (Figure 2.7b). The small-scale cross-stratified infill of such scours is a result of traction-transported grains cascading into the scour, because it was a site for the preferential deposition for traction-transported sediment on the seafloor.

Similar parallel-stratified sandstones were first described by Hiscott & Middleton (1979) from the Arenig Tourelle Formation, Quebec, and interpreted as the deposits of traction carpets. Lowe (1982) assimilated this idea into depositional models for high-density turbidity currents with parallel-stratified sandstones and ascribed them to his division S9. The formation of deep-marine parallel-stratified sandstones from traction carpets has been discussed by Hiscott (1994), suggesting that spaced stratification may not be a structure produced beneath steady flows, but may record strongly fluctuating hydrodynamic conditions and vigorous burst and sweep cycles associated with large turbidity currents.

**Transport process:** high-concentration turbidity currents.

**Depositional Process:** freezing of successively generated traction carpets at the base of the flow. Structureless divisions record rapid grain-by-grain fall-out from suspension or freezing of a thicker, unsorted layer. Local small-scale erosion features may form from surging in the flow resulting in erosion of the underlying sediment and subsequent infilling.
a) Facies B2.1 parallel-stratified sand, field of view is ~2 m wide, Annot.

b) Facies B2.1 parallel-stratified sand with small scours (<15 cm deep and 25 cm long), scours infilled with coarser material than surrounding bed, detail from Figure 2.7 a.

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2.2.2.2(b) Facies B2.2-cross-stratified sandstones

Facies B2.2 occurs in units up to 0.5 m thick with grain sizes ranging from medium- to very coarse-grained sandstones, with granule-grade material occurring in some beds. Cross-stratification is picked out by grain-size alternations, and foresets dipping up to 15°. Cross-sets appear either angular or asymptotic to their base. Cross-stratification may occur at the top of turbidite beds as multiple sets which pinch out laterally (Figure 2.8a and Enclosure 5.7D), or as beds with a tabular geometry over the extent of outcrop up to 20 m (Figure 2.8b, Enclosure 3.6 Ga). Cross-stratification may also occur as isolated, regularly-spaced bedforms up to 0.3 m thick with bedform spacing up to 4.3 m (Enclosure 3.7E). The grain size of these bedforms is generally coarse- to very coarse/granule-grade (Figures 2.8d and e), and these bedforms may rest on mudstone and be overlain either by mudstone or a rippled fine-grained sandstone which is tentatively interpreted as forming from the same depositional episode that produced the cross-stratified bedform. These bedforms may have formed at the base of the essentially non-depositing turbidity, current either as tractional reworked sediments that may or may not have been derived from the essentially non-depositing turbidity current.

Transport process: bed-load transport beneath dilute turbidity currents, strong bottom or traction transport beneath an essentially non-depositing high-concentration turbidity current.
Depositional process: either by avalanching or intermittent suspension over the crests of medium- to large-scale bedforms, or in to scours.

2.2.3 Facies Class C

Facies Class C comprises sandstone-mudstone couplets, with between 20-80% sandstone grade material and <80% mudstone grade material. Generally beds can be described using the facies classification of Bouma (1962), although some facies which occur within beds of Facies Class C are best described using the facies scheme of Pickering et al. (1986, 1989), such as the clast-
Figure 2.8 a, b and c. a) Facies B2.2 cross-stratified sand occurring as two separate bed sets towards the top of a turbidite, Grand Coyer. b) Facies B2.2 cross-stratified sand, Annot. c) Facies B2.2 cross-stratified sand, St Antonin.
Figure 2.8 d and e. d) Facies B2.2 cross-stratified sand forming a discrete bedform, Peïra Cava. e) Facies B2.2 cross-stratified sand, when traced laterally the bed thins and pinches out, Peïra Cava.
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supported conglomeratic bases to beds and the structureless sandstones which can pass vertically up into graded sandstones. Facies Class C is divided into disorganised and organised sandstone facies groups.

2.2.3.1 Facies Group C1-disorganised muddy sandstones

2.2.3.1(a) Facies C1.1-poorly sorted muddy sandstones

Facies C1.1 occurs in beds up to 1 m thick, generally with <20% mudstone occurring within the sandstone. Beds are poorly sorted, with grain sizes up to very coarse-grained sandstone and rare granule-grade material. Reworked mudstone, siltstone and fine-grained sandstone clasts are common up to 20 cm thick and 50 cm long. The base to beds are essentially planar, with their tops having an irregular surface which is invariably draped by clean sandstones of Facies Group C2 (Figure 2.9a). This facies and the associated sandstone facies was described by Stanley (1982) and interpreted as welded slump-sandstone couplets, with the two facies being part of the same depositional event as no draping mudstones separate both facies. This facies is common in the southern part of the Peira Cava sub-basin and absent in the northern part of the sub-basin. The southern part of the Peira Cava sub-basin is in a more proximal position in relation to the intrabasinal slope, and probably due to the high density of the sediment gravity flow which deposited the facies, it appears to have been incapable of reaching the more distal parts of the sub-basin.

Transport process: relatively mud-rich, high-density turbidity currents or fluid sand-mud debris flow travelling down relatively steep intra-basinal slopes.
Depositional process: rapid en-masse deposition due to increased intergranular friction and/or cohesion.
Figure 2.9 a and b. a) Facies C1.1 poorly sorted muddy sand located below lens cap; note reworked clasts of siltstone towards the right, Peira Cava. b) Facies C2.1 very thick-bedded sandstone, van for scale; top of bed can be seen below vegetation towards top left, Contes.
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2.2.3.2 Facies Group C2-organised sandstone-mudstone couplets

Facies Group C2 comprises moderately well sorted sandstone-mudstone couplets, in which beds may be amalgamated. Generally beds of Facies Group C2 show partial Bouma sequences. This facies group is divided up into four facies, with three of the facies based on the bed thicknesses defined by Ingram (1954) (Facies C2.1, C2.2, C2.3), and a fourth based on thick to very thick-bedded, mudstone-dominated sandstone-mudstone couplets with evidence of flow reversal during deposition (Facies C2.4).

The Bouma (1962) sequence summarised on Table 2.2, is a good model for medium-grained turbidites deposited from decelerating low-concentration turbidity currents. The Bouma sequence is incorporated into the facies classification scheme of Pickering et al. (1986, 1989), and is also used in the description of facies used in this thesis. The Bouma sequence, however, does not adequately describe beds emplaced by high-concentration turbidity currents or fine-grained turbidites. Facies Group C2 incorporates the deposits of both high- and low-concentration turbidity currents, deposits of which may show complete, or more commonly partial Bouma sequences. Beds of Facies C2.1 generally begin with Bouma Ta division, those of Facies C2.2 with Bouma Tb division, and those of Facies C2.3 with Bouma Tc division.

2.2.3.2(a) Facies C2.1

Facies C2.1 occurs in beds up to 10 m thick (Figure 2.9b), although beds are generally <5 m thick. The beds are laterally continuous with thinning and thickening observed in some beds over distances of ~700 m. The bases may be erosive to depths of <1 m or essentially planar with flutes and grooves.

Beds of this facies essentially comprise the Bouma Ta and Tb divisions (Figure 2.10a), and may comprise divisions Tc, Td, and Te, although complete Bouma sequences are common. Within
<table>
<thead>
<tr>
<th>GRAIN SIZE</th>
<th>BOUMA (1962) DIVISIONS</th>
<th>INTERPRETATION</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mud</td>
<td>E Laminated to homogeneous mud</td>
<td>Deposition from low-density tail of turbidity current ± settling of pelagic or hemipelagic particles</td>
</tr>
<tr>
<td>Silt</td>
<td>D Upper mud/silt laminae</td>
<td>Shear sorting of grains &amp; flocs</td>
</tr>
<tr>
<td>Sand</td>
<td>C Ripples, climbing ripples, wavy or convolute laminae</td>
<td>Lower part of lower flow regime of Simons et al (1965)</td>
</tr>
<tr>
<td></td>
<td>B Plane laminae</td>
<td>Upper flow regime plane bed</td>
</tr>
<tr>
<td>Coarse Sand</td>
<td>A Structureless or graded sand to granule</td>
<td>Rapid deposition with no traction transport, possible quick (liquefied) bed</td>
</tr>
</tbody>
</table>

Figure 2.10 a, b and c. a) Facies C2.1 thick-bedded sandstone with bouma Tₐ and Tₜ divisions, Annot. b) Facies C2.1 thick-bedded sandstone with mudstone, bed is ~ 2 m thick; note mud clast horizons within Tₐ division, Peïra Cava. c) Facies C2.1 with detail of mud clast concentrations between Tₐ and Tₜ divisions, Annot.
Figure 2.10 d and e. d) Facies C2.1 with a large reworked sandstone clast (Facies F2.1), at base of bed, Contes. e) Facies C2.1 detail of repetitions of Bouma divisions, from bottom of photo upwards: \( T_b, T_d, T_c, T_d, T_e \). Peira Cava.
Figure 2.11 a and b. a) Facies C2.1 very thick bedded sandstone showing Ta and Tb divisions towards top of bed; also note fluid escape pipes and convolution, Contes. b) Facies C2.1, bed amalgamation, Annot.
the thicker beds of Facies B2.1 there may be Bouma T_a divisions present, giving a composite bed comprising Facies B2.1 and C2.1. Reworked mudstones may occur within the Bouma T_a and T_b divisions (Figures 2.10b and 2.10c, respectively), as well as clasts of sandstone material up to 1 m in diameter (Figure 2.10d). The upper parts of Facies C2.1 beds may show repetitions of Bouma T_c and T_d divisions (Figure 2.10e), indicating that some turbidity currents were not simple waning flows, and that some may have been surging.

Dewatering fabrics, such as dish structures and rarely fluid-escape pipes also occur (Figure 2.11a), as well as convolute lamination. Facies C2.1 may be separated by mudstones or amalgamated with scoured bases (Figure 2.5c).

2.2.3.2(b) Facies C2.2

Facies C2.2 occur as beds up to 10-30 cm thick, and they are characteristically tabular, or rarely amalgamated. They essentially comprise Bouma T_b, T_c, T_d and T_e divisions, with rare T_a division (Figure 2.12a). Maximum grain size observed in this facies is medium-grained sandstone. Beds of this facies may comprise entirely of a single Bouma T_b, T_c, or T_d division, or more commonly as a combination of all three. Repetitions of Bouma T_b, T_c, and T_d division are common from the Peña Cava sub-basin suggesting fluctuations in flow velocity of the depositing turbidity currents (Figure 2.12b).

2.2.3.2(c) Facies C2.3

Facies C2.3 comprises beds <10 cm thick which are essentially fine-grained and comprise Bouma T_c and T_d divisions (Figures 2.13a and 2.13b). Beds of grain sizes greater than fine-grade are rare, and where present are structureless. Beds are mostly laterally continuous over...
Figure 2.12 a and b. a) Facies C2.2 medium bedded sand-mud couplets dominated by Tab divisions, Annot. b) Facies C2.2 with repetitions of Tacd divisions, Annot.
Figure 2.13 a and b. a) Facies C2.3 thin-bedded sand-mud couplets, Grand Coyer. b) Facies C2.3, detail of a) showing ripple cross-lamination in thicker beds.
Figure 2.14 a and b. a) Facies C2.4 thick-bedded, mud dominated sand-mud couplet, note "chicken wire texture"; alternations of parallel-stratified and ripple cross-laminations are evident, also note possible current reversals, Contes. b) Facies D2.3 thin regular silt and mud laminae with beds of Facies 2.3, St Antonin.
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outcrop for distances of ~50 m, although thinning over distances of 100 m are observed with some beds showing pinch outs.

2.2.3.2(d) Facies C2.4

Facies C2.4 attains thicknesses up to 1 m, with sandstone portions up to 0.5 m thick, and contains grain sizes up to medium-grained sandstone. The sandstone part of beds comprise alternations of Bouma Tb and Tc divisions. In some beds, the Tc divisions clearly reveal current reversals, indicating turbidity-current flow-reflections (Figure 2.14a). Flow reflections in turbidites were originally recorded by Hiscott & Pickering (1984), and Pickering & Hiscott (1985). In this study area they have been identified from the basal part of the Contes sub-basin sandstone succession (Enclosure 5.8C and log B).

**Transport process:** sediments of Facies C2.1 are transported high-concentration turbidity currents, sediments Facies C2.3 transported by relatively dilute turbidity currents; with sediments of Facies C2.2 are interpreted to be transported by turbidity currents intermediate in character to those that transported sediment of Facies C2.1 and C2.3. Sediments of Facies C2.4, which show flow reflection, may be transported by high- or low concentration turbidity currents.

**Depositional process:** grain-by-grain deposition followed by burial, and/or tractional transport by bedload forming Bouma divisions Tb and Tc.
Chapter 2: Sedimentary facies

2.2.4 Class D

Class D comprises silts, silty muds, and silt-mud couplets, with $>80\%$ mud, and $\geq 40\%$ silt, with between 0-20% sand. Within the study area of this thesis, only Facies D2.3 is represented.

2.2.4.1 Facies Group D2-organised siltstones and muddy siltstones

2.2.4.1(a) Facies D2.3-thin regular siltstone and mudstone laminae

Facies D2.3 is rare in the study area. This facies generally occurs in association with beds of Facies C2.3 (Figure 2.14b) as siltstone laminae and bands up to 5 mm thick occurring within mudstones. These siltstone laminae and bands are laterally continuous over 4 m, although some show a lenticular nature over distances of 0.25 m.

**Transport process:** low-concentration turbidity currents or weak bottom currents, possibly generated by storm events.

**Depositional process:** slow uniform deposition from suspension, with shear sorting of silt grains and clay flocules in the viscous sublayer of the turbidity current, as described by Stow & Bowen (1980).
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2.2.5 Class F-chaotic deposits

2.2.5.1 Facies Group F2-contorted/disturbed strata

2.2.5.1(a) Facies F2.1-coherent folded and contorted strata

Facies F2.1 includes folded and deformed strata on all scales from single beds (Figure 2.15a), to large-scale deformation of strata on a sub-basin scale (Enclosure 3.3A-F). In the study area, Facies F2.1 is common.

Facies F2.1 can be related to deposition of sediments on intrabasinal palaeoslopes and subsequent failure due to the unstable site of deposition. Folded strata may occur at the base of thick turbidites ascribed to Facies C2.1, these are thought to have formed as a result of erosion by the passage of turbidity currents depositing a turbidite which encases the folded strata (Figure 2.15b and 2.10d).

Transport process: gravity-induced sliding and slumping due to deposition of sediments on relatively steep intrabasinal slopes and/or seismic activity.
Depositional process: slides or slumps freeze on reaching lower gradients, e.g., bottom slopes.
Figure 2.15 a and b. a) Facies F2.1 coherent folded and contorted strata, Annot. b) Facies F2.1 coherent folded and contorted strata located towards the base of a turbidite, Peira Cava.
Chapter 3

Sandstone architecture and facies from the Annot sub-basin and surrounding areas

3.1 Introduction and Location

The Annot sub-basin (so called as it is one of several depositional centres within what was an extensive marine basin) is located in the Haute Provence region of southeast France. Its relationship with the surrounding Tertiary basins of Haute Provence and Alpes Maritimes is shown on Figure 3.1, and a detailed map of the Annot sub-basin is shown on Figure 3.2. The basin covers an area of 110 km², with the sandstones covering an area of 54 km². The Tertiary sediments of the Annot sub-basin form part of the Southwest Alpine Foreland Basin fill of southeast France and consist of a succession of sediments which record a change from continental to shallow-marine and then a gradual deepening of the basin to a deep-marine setting. The stratigraphy of the Tertiary Foreland basin fill is described in Section 1.2, with this study concentrating on the Grès d'Annot Formation.

The Annot basin is located north of the junction of the D908 road with the R202, at the commune of Les Scaffarels, from where there are spectacular views of Annot sandstone cliffs. The town of Annot is located on the southwest side of the sandstone outcrops. The cliffs comprise 10 to 80 m thick packets of amalgamated sandstones which are characteristic of sandstones from the Annot sub-basin, but are not representative of the deposits occurring over most of the outcrops of the Grès d'Annot Formation.
Figure 3.1. Geological map of the Alpes Maritimes and Haute Provence regions of Southeast France, the Annot sub-basin is marked (re-drawn from BRGM 1980, Sheet 40 and 45 1:250 000).
Figure 3.2. Geological map of the Annot sub-basin with a summary of palaeocurrent trends for the Grès d'Annot Formation (redrawn from BRGM 1980 sheet 945, 1:50 000)
Chapter 3: Sandstone architecture and Facies from the Annot area

3.1.1 Previous research on the Annot sub-basin

The sandstones at Annot have been the focus of much research in the past and were documented as early as the mid nineteenth century by Gras (1840). The sandstones from Annot were formally named the Grès d'Annot by Boussac (1912), and the sandstones from Annot were originally interpreted to be of shallow marine origin. The name Grès d'Annot Formation is now given to sandstones and mudstones of similar age located in the Alpes Maritimes and Hautes Provence regions of Southeast France, although local names such as the Grès de Petra Cava are still used (Bouma 1962).

Stanley (1961) carried out a detailed study on the Grès d'Annot Formation, including a description of the resistate mineralogy of the sandstones, from which he recognised three spatially distinct provenance groups. Also, Stanley (1967) undertook palaeocurrent analysis, sandstone grain size distributions, and together with the mineralogy, concluded that there was more than one sediment source. Stanley (1967), also proposed that the Grès d'Annot Formation accumulated in submarine channels.

Stanley & Bouma (1964) further emphasised the channel interpretation at Annot by making observations on the geometry of the sandstone bodies, and concluding that there must have been a north-south trending depression which acted as a sand trap. This depression was interpreted to have formed by down-warping of the underlying Marnes Bleues Formation and also because of erosion of a submarine valley or channel. They documented two sections: a > 200 m thick succession of very thick and poorly stratified non flysch-like ("fluxoturbidites") sandstone beds on the west of the Coulomp Valley, and a < 100 m thick succession of sandstone beds with typical turbiditic or flysch-like characteristics on the east of the Coulomp Valley. They interpreted the two sections to be lateral equivalents. The outcrops on the west of the Coulomp Valley were interpreted as the infill of a north-south submarine valley or channel, with classic turbidites (ibid.). Outcrops on the east of the Coulomp Valley represent
deposition on the less deformed slopes adjacent to the depression, where more sheet-like deposits accumulated.

Stanley (1967) developed a model for canyon sedimentation and established criteria for recognising types of submarine valley deposits in the ancient record. The study was based on observations of submarine canyons off Nova Scotia, and of ancient deposits from the French Maritimes Alps in the Annot, Contes and Menton areas. Stanley (1967) classified the outcrops from Annot, Contes, and Menton as submarine valley fills and differentiated them from submarine canyons, that cut into continental slopes by defining the former as involving broader erosional features in any variety of sea-floor depression, a distinction which now seems too arbitrary. The sandstones at Annot, Contes and Menton represent the infill of depressions cut into marls, with thinner, finer sediments deposited on the margins of these depressions.

Stanley & Unrug (1972) discussed the features of submarine channel deposits, as typically "fluxoturbidites" and other indicators of slope, and base-of-slope environments using examples from modern and ancient deep-marine settings. These Tertiary examples were from the Grès d'Annot Formation of southeast France. The Annot, Contes, and Menton outcrops were interpreted as shoestring-shaped bodies in ancient submarine canyons and valleys. These so-called submarine valley deposits at Annot, Contes, and Menton can be traced basinward to the deposits in the Trois Evêchés and Grand Coyer areas where the submarine slope/channel deposits thicken downslope, concomitant with a decrease in gradient at the base of slope (Figure 3.1).

Stanley \textit{et al.} (1978) and Stanley (1982) describe canyon axis, canyon wall, and tributary canyon and fan valley, together with interchannel facies of the Grès d'Annot Formation in the Annot, Contes, and Menton areas. Down-current sediment dispersal patterns were also discussed from these channelised environments to unconfined outer margin environments located towards the north of the Annot, Contes, and Menton outcrop areas \textit{(ibid.).}
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In the Annot basin, the sandstones interpreted as canyon-wall and canyon-axis facies are represented by structureless and amalgamated sandstones, with the canyon wall deposits being thinner, non-amalgamated, sandstones and shales. The canyon sandstones are interpreted as the deposits from high-concentration flows initiated from slope failure as sediment slides and from slurry flow processes (e.g., grain flow).

More recently Apps (1987), interpreted the Annot basin as structurally confined at its northern margin, with a palaeoslope dipping southwards over at least 2 km. The basin was also confined at the west by a north-south ridge produced by the Argens thrust, and to the northeast by the Melina-Aurent "kink zone", the surface expression of a southwestward propagating thrust in the basement rocks below. The confined basin caused turbidity currents to lose energy by momentum changes and dump large amounts of sediments near to the base of slope. The relative lack of mud and shale was interpreted to be a result of turbidity currents surmounting the northern topographic high taking the fine fraction with them deeper in to the basin.

3.1.2 Structure of the Annot area

The structure of the Annot sub-basin is complex and reflects the evolving structure of the Southwest Alpine Foreland Basin. The structures within the Annot area originated from the Pyrenean-Provençal and Alpine orogenies (Siddans 1979, Apps 1987, Fry 1989) (Figure 3.3).

The Tertiary sediments in the Annot sub-basin form a relatively gentle syncline with an axis trending NNW-SSE. The syncline, mapped by the Bureau de Recherches Géologiques et Minières (1980) is based on the sediments that underlie the Grès d’Annot Formation. The sandstones at Annot, however, do not form a syncline, but instead dip towards the west or south (see Figure 3.3), as seen on a structural cross-section (Figure 3.4). The sandstones show a progressive onlap onto the marls towards the west, with sandstone beds in the east of the
Figure 3.3. Simplified structural map of the Annot area showing major faults and fold trends, location of basement thrusts based on Apps (1987).
Figure 3.4 Cross section of the Annot sub-basin showing the relationship of the Grès d'Annot Formation with the underlying Marnes Bleues Formation.
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basin dipping more steeply than those in the west, interpreted to be the result of Miocene folding producing a structural bench in the sandstones.

Towards south, the Annot sub-basin is bounded by the Rouaine Fault Zone (includes the Braux and St Benoit Faults). The fault zone was active prior to sandstone deposition. Apps (1987) interprets the fault to be a lateral ramp accommodating differential displacements above a major thrust. Ravenne et al. (1987), however, interpret the fault as an extensional structure related to a post Pyrenean-Provençal extensional phase, with the evidence for extension from other normal faults located along the Var River valley. A reconstruction of the Rouaine Fault Zone is shown on Figure 3.16a and b, with the faults producing sediment thickness and facies variations in the Calcaires Nummulitiques and Marnes Bleues Formations. Pairis (1971) noted a shallow-marine fauna on the footwall including large unbroken oysters, and that the Marnes Bleues Formation thickens dramatically on the hanging wall side. Apps (1987) interpreted the fault zone to represent the deepest part of the basin during sandstone deposition, and noted that no sandstones are preserved today, inferred as the result of inversion of the fault zone and associated erosion. Towards the west, the Annot Basin is bounded by the Puy de Rent-La Colle St-Michel anticline and a north-south trending thrust interpreted to have been active during deposition of the Tertiary sediments (Apps 1987). The northern margin to the basin is defined by the 1-km-wide Melina kink zone (Apps 1987) which trends northwest-southeast, and continues towards the Rouaine Fault Zone (Figure 3.12). Apps (1987) interpreted the Melina kink zone to have been active during deposition of the Marnes Bleues and Grès d'Annot Formations, creating a barrier to sandstone deposition. The kink zone is interpreted to have formed above a basement thrust (labelled BII, see Figure 3.3) and resulted in a slope which dipped towards the southwest and controlled sedimentation. Evidence for this slope and its influence on sedimentation is given in Sections 3.2.4 and 3.3.2. The eastern margin of the basin is less easy to define, as there are no faults or folds, but the Barrot Dome, an uplifted area of basement of mainly Permian to Lower Jurassic sediments, now forms a relative topographic high.
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3.2 Sandstone architecture and facies

This section gives a detailed description of the exposed sections from the Annot sub-basin. The location of each section described is shown on Figure 3.5. Each section has a corresponding large-format enclosure including a photo-montage of the relevant section, generally with an accompanying line interpretation. Measured sections through the exposures are also included on the enclosure, together with palaeocurrent directions; and close up details of features of interest. Here, a description of the facies is presented, their vertical and lateral extent and their relationship with neighbouring sections. The nature of the contact of the sandstones with the underlying marls, where exposed, is also described. Each description is followed by an interpretation and reconstruction of the sandstones during deposition.

3.2.1 St Benoit

The St Benoit sections are located on the eastern side of the Grès d'Annot basin close to the town of St Benoit (see Figure 3.6). The outcrops are shown on sections A and B on Enclosure 3.1 and are named the Crête de la Barre and St Benoit sections, respectively. These sections can be correlated on the basis of physically tracing sandstone beds between sections. The St Benoit and Crête de la Barre sections show similar facies. The Crête de la Barre section correlates with the uppermost part of the St Benoit outcrops. Both sections show moderately good exposures of sandstones though most of it remains inaccessible due to near vertical slopes and dense vegetation.

This part of the basin is structurally complex with the St Benoit Fault (which forms part of the Rouaine Fault-Zone) passing along and partly underneath the Crête de la Barre section. The St Benoit Fault can be shown to have been active at some time during the deposition of the Tertiary sediments, with the fault cutting through the Calcaires Nummulitiques Formation.
Figure 3.5. Geological map of the Annot area with location of sections as described in the text and shown on Enclosures.
Figure 3.6. Base map of the St Benoit area with palaeocurrents from the Grès d'Annot Formation, apparent onlap directions are marked. Reconstructed palaeoslope in the Marnes Bleues Formation is marked.
(see below), and possibly the Marnes Bleues Formation, but with no obvious faulting of the Grès d’Annot Formation (Pairis 1971, Elliott et al. 1985).

3.2.1.1 St Benoit section

3.2.1.1a Description

The St Benoit section (Enclosure 3.1 Plate B) trends WNW and shows moderately good exposures of laterally continuous sandstones above marls. The contact between the Grès d’Annot and Marnes Bleues Formations is exposed at two localities, on the east side representing the oldest Grès d’Annot Formation in the Annot sub-basin, and also on the west against the palaeotopography created by the St Benoit Fault. The section is estimated to be approximately 300 m thick from the base in the east to the top of the Crête de la Barre section. The eastern contact of the sandstones on marls, shows a low-angle apparent onlap towards the west. The basal Grès d’Annot is represented by a 2 m thick unit of thin (<5 cm) fine-grained sandstones and mudstones, above which there is an 80 cm thick unit of folded sandstone (Plate E-Enclosure 3.1) which shows a consistent vergence towards the east. The remainder of this lower unit consists of sandstones up to 3 m thick of Facies Group C2.2 (see log D and Plate D Enclosure 3.1) with Bouma T_a and T_b divisions present. Dish structures occur in the T_a division. The beds in the lower unit are essentially non-amalgamated and show no large-scale erosion. Above this sandstone unit, there is a break in exposure for approximately 90 m, covered by a densely forested plateau, and above this there is a 100 m thick exposure of sandstones. The sandstone beds range in thickness from 0.1-5 m, with an average of 1 m, and are ascribed to Facies Group C2.2. They comprise Bouma T_a and T_b divisions and less common T_c divisions. Thin sandstone/mudstone couplets interbedded with the thicker sandstones are present; log D1 is a section taken from mid-way up the upper sandstone unit.
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The lower sandstone exposure is restricted to the eastern side of the section and dips towards the west, and cannot be traced farther westwards as it is lost in vegetation cover. Above the lower sandstone unit on the west side of the section, a continuous exposure of sandstones extends to the top of the section. Thin-bedded sandstones present at the base on the east side are replaced by an outcrop of marl in the west. At the locality where Plate F was taken, a steep slope in the marls is inferred because when the sandstones on the east are traced towards the west, marls are encountered and sandstones are not seen.

Below the marls, limestones of the Calcaires Nummulitiques Formation (henceforth referred to as the Nummulitic limestone) crop out for a distance of 200 m. The Nummulitic limestones rest unconformably on Cretaceous limestones, with the overlying marls draping both lithologies. Along this contact, a breccia containing clasts of Cretaceous and Nummulitic limestone occurs, with a matrix of marls and fine-grained carbonate sediment derived from the limestones. The contact between the marls and limestones is interpreted as a fault scarp trending 028-208 with an apparent dip of 70° to the east. Movement on the fault occurred after deposition of the Nummulitic limestone and may have been active during deposition of the marls. The marls show a considerable increase in thickness on the downthrown side of the fault, from approximately 70 m thick on the footwall to approximately 300 m thick on the hanging wall over a distance of <100 m.

Palaeocurrent data from the St Benoit section is limited to grooves at the base of turbidites. These data are shown on Figures 3.2 and 3.4 and show a general southeast-northwest trend.

3.2.1.1b Interpretation

The St Benoit section is a succession of sandstones deposited mainly from turbidity currents in a basin with complex topography. Most of the turbidity currents were associated with no significant erosion at the base of their flows, resulting in essentially non-amalgamated...
sandstones. However, bed amalgamation is observed in the sandstone unit shown on
Enclosure 3.1F: when individual beds are traced westwards towards the palaeoslope they
become amalgamated. The amalgamation of beds in close proximity to this palaeoslope is
interpreted to be a result of increased turbulence of turbidity currents caused by the
interaction of the turbidity currents with the palaeoslope. Individual sandstone beds can be
correlated for distances up to 1.5 km without any observable change in thickness and are only
seen to thin or terminate where they approach a palaeoslope.

The St. Benoit Fault trends NE-SW and has been interpreted to represent a splay of the
Rouaine Fault-Zone (Pairis 1971) which is located 2 km southwest of St Benoit. The Rouaine
Fault-Zone extends for 30 km towards northeast, and extends 10 km towards the southwest
where it appears to die out. Movement on the St Benoit Fault occurred during the deposition
of the Calcaires Nummulitiques and Marnes Bleues Formations. Evidence of this
syndepositional movement can be seen in the thickness changes that occur in the above
formations when traced across this fault, i.e., the marls on the footwall side of the fault reach
40 m in thickness, while on the hanging wall side they are up to 300 m thick, this change
occurs over a distance of < 100 m. The Nummulitic limestones also show an increase in
thickness on the hanging wall side from ~20m to ~100 m. The increase in thickness within
both formations across the fault indicates that the fault was active prior to deposition of the
Nummulitic limestone; the Nummulitic limestone is displaced by the fault down to the
southeast by 400 m, but the Grès d’Annot Formation is not displaced, suggesting that the St
Benoit Fault moved before deposition of the sandstones and sometime after deposition of the
Nummulitic limestone. The marls draped the fault to create a scarp trending NE-SW on the
easterly-dipping slope, but because of poor exposure at the intersection of the fault with the
marls it is not possible to determine if the fault broke the surface. The fault scarp trends NE-
SW and dips at 70° towards the southeast. There is a breccia present on the fault scarp
composed of blocks of Nummulitic and Cretaceous limestones set in a fine clay-rich matrix
which contains nummulites of the same age as the Nummulitic limestones, further supporting
the contention that the fault was active during deposition of these limestones. Limestone
pebbles up to 5 cm diameter occur in the marls, showing that the fault scarp underwent erosion during deposition of the marls. Cumulative movement on the fault is calculated at 400 m displacement of down-throw; also, Pairis (1971) deduced that the fault had a sinistral strike-slip component, but this relative motion may have been produced later than the essentially normal movement during Alpine compression.

3.2.2 Braux

3.2.2a Description

The Braux section (Enclosure 3.2.A, Figures 3.5 and 3.6) comprises a 530 m long and 130 m thick exposure of Grès d’Annot and Marnes Bleues Formations. Sandstone beds of the Grès d’Annot Formation onlap the marls over the entire length of section. The onlap surface dips ~5° towards the west with the sandstones onlapping at an angle of ~20° in a westerly direction, therefore the true dip of the onlap would have been in the order of up to 15°. This onlap surface can be correlated north-westwards to the Coulomp section over 3 km (Enclosure 3.2A), northeasterly along the St Benoit section for 1 km (Enclosure 3.1.A), and westerly along the base of the Les Scaffarels section for 1 km (Enclosure 3.6.A).

The basal Grès d’Annot Formation succession in the Braux section (Log B, Enclosure 3.2) consists of a packet of sandstone and shale 13 m thick, overlain by a marl horizon 14 m thick, and which preserves no original marl bedding because of intense tectonic shearing. Above this unit, another packet of medium/coarse-grained sandstone 68 m thick crops out. The lower 80 m of sandstone succession at Braux can be traced for over 300 m towards the palaeoslope developed in the marls without any observable change in thickness, where the beds terminate abruptly against it.
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The sandstone succession shows a broad coarsening-and-thickening-upward sequence. The lower part of the succession (2-15 m) consists of thin sandstone beds (0.5-60 cm) of Facies C2.2 and C2.3, with beds showing rare ripples and some parallel laminations of the Bouma Tb division in the thicker beds (>10 cm). Sandstones in this lower part range from fine-to medium-grained. The upper part of the succession (31-97 m) is dominated by Facies C2.1 and C2.2, with Facies C2.3 as a minor component. As seen on log B (Enclosure 3.2), the upper part of the section contains two sandstone packets of Facies C2.2 separated by 7 m of Facies C2.3, with some beds of Facies C2.3 showing syn-sedimentary folding. The sandstones show features indicative of deposition from turbidity currents, including Tabc divisions. Large (1 by 0.3 m) shale clasts commonly occur in beds up to 4 m thick of Facies C2.1, with smaller (<20 cm) shale-clast concentrations towards the tops of some beds (Enclosure 3.2.C). Bed thickness ranges from <10 cm up to 3 m with the average being 80 cm thick. Maximum grain size observed is in very coarse-grained sandstone, with most beds being medium-grained. A small proportion of beds show repetitions of structureless medium-grained sandstones and parallel-laminated sandstones on a 5 cm scale (Enclosure 3.2.D shows five of these repetitions in a 33 cm thick bed); one of the parallel-laminated divisions shows flame structures with a sense of overturning consistent with flow towards northeast. The base of some turbidite sands have many grooves which show a cross-cutting and divergent nature of up to 20° (Enclosure 3.2E, log B), although the general trend is north-south. Flutes on the base of beds generally show sediment transport direction towards the north, but in one bed transport was towards the south. The reworked shales within the sandstones have a light brown-blue colouration similar to the underlying marls and, therefore, are inferred to have been eroded from these sediments. Bedding within the marls is conformable (with the marl-sandstone contact) and shows no evidence of significant erosion.
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3.2.2b Interpretation

The Braux section comprises turbidite sandstones that onlap onto an inclined surface formed by the underlying marls. Presuming that the turbidites were deposited on an essentially horizontal surface, then it is possible to estimate the palaeoslope during deposition of the sandstones. The sandstone beds have an orientation of 20°/309, and the onlap between the marls and sandstones is oriented at 9°/263. From this relationship, it can be deduced that the local palaeoslope orientation was 12°/128, i.e., ESE. The wide dispersion of palaeocurrents suggests that the palaeoslope was relatively smooth and consistently orientated. On the base of beds, there are cross-cutting grooves, which are interpreted as forming from meandering, and deflecting turbidity currents. Southward-directed flutes are interpreted as having formed from either upslope failure of previously deposited beds, creating secondary turbidity currents travelling down the opposing palaeoslope, or by the deflection of turbidity currents against the complex palaeotopography in this region (see Section 3.6). Syn-sedimentary folding within Facies C2.3 is interpreted to represent sediment instability on the palaeoslope. The numerous shale clasts found within the turbidite beds are interpreted to represent erosional remnants from turbidity currents excavating into previously deposited semi-consolidated/cohesive slope muds, with their incorporation into the flow, and deposition in a down-current direction.

The interbedded marl horizon 16 m up-section, was interpreted by Apps (1987) as a slide deposit of marls transported downslope from west to east onto the sandstones at Braux. The onlap at Les Gassés (Enclosure 3.3.A), located towards the west and upslope from Braux, is interpreted as a slide scar produced from this sedimentary sliding of the marls (ibid.). The lower sandstone body in the marls dips 30° towards the east as compared to 25° for the rest of the section. One explanation for this discrepancy is that the lower sandstone represents sediment which has moved down-slope in a rotational motion, due to sediment instability on the palaeoslope.
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3.2.3 Coulomp Valley

3.2.3a Description

The Coulomp Valley exposures of the Grès d'Annot Formation form a 4.5 km long cliff section (Figure 3.5, Enclosure 3.3A) showing a spectacular 350 m thick exposure of amalgamated sandstone beds and interbedded thin packets of thin-bedded sandstone and mudstone couplets over the entire length of section (Enclosure 3.3A). The section shows limited accessible exposures of the contact between the sandstones and the underlying marls, although the contact between the two formations may be observed from a distance and be approximately located on Enclosure 3.3.A.

The succession is dominated by amalgamated sandstones which range in thickness between 1-6 m. Observations on grain size and facies are not possible due to inaccessibility of the sections, except for the exposures at the Gastres onlap shown on Enclosure 3.3.D (described later), and also in sections 3.2.6 and 3.2.7 where the exposures are more accessible. Plate B (Enclosure 3.2) is a close-up of the succession showing packets of amalgamated sandstone beds and interbedded horizons of thin-packeted sandstone/mudstone couplets.

3.2.3b Sandstone Architecture

Packets of amalgamated sandstones and thin interbedded horizons of thin sandstone/mudstone couplets can be traced over the entire section for 4.5 km (Enclosure 3.3A). Individual sandstones beds can only be traced for a maximum lateral distance of approximately 1 km. The amalgamated nature of the sandstones make it difficult to trace beds between exposures along the Coulomp Valley. Beds are best picked out along weathered sections where preferential weathering of less competent and finer grained horizons has
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emphasised individual beds. Observations using binoculars reveals local pinchout of some
thinner beds, generally less than 3 m in thickness.

The amalgamated sandstone packets can be traced across the 4.5 km section along the
Coulomp Valley and range in thickness from 30-80 m, and over this distance show no
observable thinning along strike. The thin packets of thin sandstone/mudstone couplets are
correlateable across the 4.5 km of section and range in thickness from 1-5 m.

The base of the succession in this area is seen at the Les Gastres onlap (Enclosure 3.3D). The
sandstones show an apparent southerly onlap onto a northward-dipping slope developed in
the marls. The sandstone beds are amalgamated, but with beds close to the onlap separated by
shale material. The palaeoslope is calculated to be 8°/038. Thin-bedded sandstones parallel
this slope with the thicker amalgamated sandstones onlapping it and even draping the local
slope. Palaeocurrent readings taken from grooves on the base of beds give a SE-NW
orientation for transport. Beds show dish structures and reworked clasts of shale. Bedding
within the marls can be traced towards the onlap were it is truncated at the slope indicating
that the slope formed by erosion of the marls.

At approximately 420 m up-section in the Annot succession, there is a unit of deformed
sandstones which shows remnants of bedding, and which can be traced across the entire
length of section to the Les Scaffarels section (Enclosure 3.6). This deformed layer is
estimated to be approximately 60 m thick at maximum, and is represented by large-scale
synsedimentary folds and faulting. Enclosure 3.6C and 3.6E, show a close-up view of these
folds with the fold on Enclosure 3.6E showing a vergence towards the south. Enclosure 3.6C
shows an open fold with small-scale sedimentary faults offsetting the bedding. Enclosure
3.6E shows a larger and tighter fold in the sandstones. Between these two folds, there are
surfaces within the slide unit which dip towards the south at an apparent angle of 30° to
bedding and are interpreted to represent rotational slide or slump surfaces which
accommodated movement of the sandstones. The slide horizon has been interpreted by Apps
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(1987) to represent downslope failure towards the south of sandstones deposited on an upslope northern-margin developed in the marls, although the evidence she put forward for this slope is located at the Tête du Ruch outcrops 3 km farther north (shown in detail on Enclosures 3.5.1 and 3.5.2 and described in Section 3.2.5).

The evidence used by Apps (1987) to infer a major basin-wide slope has been shown in this study to represent only a minor, small-scale, topographic feature. This research suggests a major southerly-dipping slope, and is described in Section 3.2.4 and Section 3.4 and Enclosure 3.4, where sandstone onlaps a southerly-dipping slope in the marls. Other indirect evidence for this southerly-dipping palaeoslope include a detailed contour map constructed for the present-day sandstone-marl contact (Figure 3.14, Section 3.3). On the contour map, there is an area where the contours for the sandstone-marl contact are closely spaced as compared to the widely spaced contours for the rest of the Coulomp Valley section, which is interpreted to represent a local change in slope gradient, and which towards the north gave way to a more gentle southerly-dipping slope, as seen at Argenton (Enclosure 3.4). Possible triggering mechanisms for the sedimentary slide include that put forward by Apps (1987) of seismic activity caused by earthquakes in the region. Other possibilities include uplift of sediments at the northern margin of the basin causing previously deposited sediments to become unstable and fail downslope on the newly created southward-dipping sea-floor. Apps (1987) interpreted this slope in the northern margin of the sub-basin to be a result of thrust-tip ramping of northeasterly derived basement thrusts which she identified from the Annot area.
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3.2.4 Argenton-Chabrières

The Argenton and Chabrières sections are located on the western side of the Coulomp Valley, and ~1 km northward of the Coulomp Valley sandstone cliffs.

3.2.4.1 Argenton

3.2.4.1a Description

The section at Argenton (Figure 3.7, Enclosure 3.4.A) trends NNW-SSE and comprises a 50 m high and 300 m long cliff exposure of Grès d’Annot and Marnes Bleues Formations. The contact between the two formations is represented by an onlap surface which dips towards the south at an apparent angle of 11°, the sandstone beds also dip at an apparent angle of 8° towards the south. Bedding orientation in the basal sandstone/mudstone couplets is 26° towards south. The onlap surface is parallel to the bedding in the marls and, therefore, represents a smooth surface with no obvious erosion into the underlying marls.

The sandstone succession in the basal 16 m is composed of thin (<5 cm) sandstone/mudstone couplets of Facies C2.3 with thicker sandstone beds (<20 cm) located at the southern end of the section, and which onlap against the marls (Enclosure 3.4.A). The thin sandstone/mudstone couplets parallel the onlap surface in comparison to the thicker sandstones further up-section which are at an angle of 3° to the onlap. At the northern end of the section, a lateral transition from Facies C2.2 into Facies C2.3 occurs in two beds which thin towards the north until they reach the point shown on Plate C (Enclosure 3.2). Where the two beds become part of the basal thin sandstone/mudstone couplets and parallel the onlap surface.
Figure 3.7. Geological base map for the Argenton area of the Annot sub-basin with reconstructed palaeoslopes and palaeocurrent data.
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The thick-bedded sandstones in the upper part of the section appear at 16 m from the base of the log (Enclosure 3.2). These beds range in thickness from 0.2-1.3 m, and are invariably graded, with a grain size range of fine sand to pebble-grade material, and are ascribed to Facies C2.2. Sedimentary structures are rare; mudstone clasts are present in the Tq division. Bed amalgamation occurs to a limited extent, and some beds are separated by up to 50 cm thick packets of thin (<5 cm) sandstone/mudstone couplets. One such unit is light blue in colour due to its high marl content. Palaeocurrent data give a transport direction towards the north (Figure 3.5).

At 26 m up-section from the base of the Grès d’Annot (Enclosure 3.2.B, Figure 3.8), there are two coarse-grained units which have cross-bedding present and emphasised by mud-clast concentrations and grain size changes parallel to the cross-bedding, and showing transport towards the northwest. Figure 3.8 shows a close up of these two cross-bedded units, they occur at the same horizon and are most probably related: both are approximately 1.3 m thick and are laterally separated by 50 m thick unit of parallel-bedded sandstones. These cross-bedded sandstones are poorly exposed, the southern unit is seen to lens out northwards, while the northerly unit lenses out southwards. The wavelength of these bedforms could therefore not be measured due to lack of sufficiently oriented and accessible exposures.

3.2.4.1b Interpretation

The section at Argenton is interpreted to represent turbidite sandstones deposited from sandy turbidity currents which travelled onto a southerly dipping palaeoslope developed in the marls. This palaeoslope is calculated to have dipped at 8° in a direction of 245 during deposition of the sandstones. The abrupt facies change seen in the sandstones of thin fine-grained sandstone/mudstone couplets to thicker, graded sandstones is interpreted to be a consequence of the palaeoslope influencing sedimentation from the turbidity currents as they approached the palaeoslope. Towards the palaeoslope, beds become progressively thinner and
Figure 3.8. Detail of cross-stratified bedforms located at the northern extent of the Argenton section. Cross-stratification is defined by concentrations of mudclasts and grain-size variations of coarse-grained and pebbly sandstone.
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finer grained, until eventually they parallel the slope as thin deposits, and show a transition from a medium-grained sandstone to a thin sandstone/mudstone couplet (Plate C, Enclosure 3.2). The basal 16 m of the Grès d'Annot succession comprises thin sandstone/mudstone couplets, interpreted as the distal equivalents of thicker turbidite beds located to the south of this locality, and which underwent thinning as they approached the onlap to form a thin-bedded sandstone (Section 3.6).

3.2.4.1c Mesotopographic bedforms

Mesotopographic bedforms such as cross-stratification have been recorded in the literature by numerous authors (Winn & Dott 1977, Piper et al. 1985, Hughes Clarke et al. 1990, Vicente-Bravo & Robles 1994). Mesotopographic bedforms are sedimentary structures which have a lateral extent of 0.1-1000 m and vertical height or depth of between 0.1-100 m (Mutti & Normark 1987. Winn & Dott (1977) describe cross-bedded gravels from Chile which show similarities to the bedforms described above and are interpreted as having formed by traction currents travelling south to north along the sea-floor, reworking and winnowing previously deposited sediment. The cross-stratification seen in the Argenton section contains pebble grade material and mud clasts are up to 10 cm in length, indicating current velocities in the order of 0.2-0.4 ms\(^{-1}\) as derived from a velocity-grain size graph (Sundborg 1967). Transportation could either have been provided by the tail-end of a turbidity current, or strong bottom currents. The section shows a break in slope from a relatively horizontal to an inclined slope developed in the marls with an apparent dip direction towards south. This local southerly dipping slope may have opposed northwardly travelling turbidity currents, and this may be a controlling factor for the formation of these structures.
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3.2.4.2 Chabrières

3.2.4.2a Description

The Chabrières section (Enclosure 3.4 D) is located to the south of the Argenton section and is poorly exposed and mostly inaccessible. Observations, however, have been made using binoculars and photography. The section has an approximate north-south strike, and consists of an erosive contact between the Grès d'Annot and Marnes Bleues Formations. The contact is in the form of a scallop-shaped depression developed in the marls which has been infilled with amalgamated turbidite sandstones characteristic of the Coulomp section 1 km to the south of this locality.

3.2.4.2b Interpretation

Erosion of the marls at the Chabrières section has occurred either as a result of syn-sedimentary sliding of the marls due to sediment instability of a pre-existing slope developed in the marls, or as a result of erosion of the marls by turbidity currents.

3.2.5. Tête du Ruch outcrops

The Tête du Ruch outcrops (Figure 3.5 and 3.9) are located at the northernmost part of the Grès d'Annot Formation in the Annot region. This area consists of two sections: the Tête du Ruch (Enclosures 3.2.5.1/2/3) and North Tête du Ruch outcrops (Enclosure 3.2.5.4) which can be correlated with the Tête du Ruch section.
Figure 3.9. Geological map of the Tête du Ruch area showing palaeocurrent data sandstones dips and "channel" orientation.
3.2.5.1 Tête du Ruch section

The Tête du Ruch section (Enclosure 3.5.1) trends SSE-NNW and shows an almost continuous 300 m long and 70 m thick exposure of sandstones and marls. The contact between the sandstones and marls can be traced continuously over the entire section. The section is divided into two units; the upper and lower sandstone bodies, which are shown in detail on Enclosures 3.5.3 and 3.5.2 respectively, and described in Sections 3.2.5.3 and 3.2.5.2. The two sandstone bodies are separated by a 12 m thick unit of thin (<30 cm) fine-grained sandstone/mudstone couplets (Facies C2.3). The upper sandstone body has a continuous outcrop over the length of the section, while the lower sandstone body has a limited extent over the length of section. The lower sandstone body is well exposed on the north side of the section, and can be seen to onlap onto a topographic feature developed in the marls. When traced towards the south, the lower sandstone body is lost in vegetation cover. Farther south, the lower sandstone body is absent, and instead, there is the contact between the upper sandstone body and underlying marl. The lower sandstone body is interpreted to represent the infill of a depression on the sea-floor, which onlaped onto a scallop-shaped slope. The lower sandstone body is described in detail Section 3.2.5.2 below.

3.2.5.2 Tête du Ruch lower sandstone

3.2.5.2a Description

Along strike to south the lower sandstone body, marls outcrop and are overlain by the upper sandstone body without the presence of the intervening lower sandstone body. The southern margin of the lower sandstone body is interpreted to onlap a slope in the marls in a manner similar to the northern onlap of this sandstone body. The geometry of the lower sandstone body is therefore interpreted to be channel shaped. This gives us an extrapolated channel length of 190 m although the section is oblique to the strike of the channel and a true width of
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136 m is calculated using basic trigonometry (Figure 3.9). The lower sandstone body is separated from the upper sandstone body by a 12 m thick unit of thin (<25 cm) fine-grained sandstone/mudstone couplets, shown in detail on the upper half of log B (Enclosure 3.5.2).

The lower sandstone body comprises beds <1 m thick with the maximum observed grain size of coarse-grained sandstone. These beds can be divided into two facies groups, thin fine-grained sandstone/mudstone couplets <5 cm thick, of Facies C2.2, and medium-grained beds which range in thickness from 5-100 cm and are ascribed to Facies C2.2 and C2.1. Beds of Facies C2.2 and C2.1 are predominantly medium-grained, and are characterised by upper flow regime parallel-lamination with primary current lineation, and subordinate lower flow regime current-ripple lamination. Log B (Enclosure 3.5.2) is a measured section of the beds in the lower sandstone body. Many beds show parallel laminations throughout, while others contain ripple lamination which invariably occurs towards the top of beds, but may occur within the parallel-laminated horizons. Erosion by turbidity currents is evident, by the locally abundant mud clasts in beds. The mud clasts preferentially occur where individual beds approach the onlap, when beds are traced away from the onlap the mud clasts become smaller and less abundant until they eventually die out. When traced along outcrop some sandstone beds pinch and swell with rare beds showing a lensoid shape (Enclosure 3.5.3C).

When traced towards the north, the sandstone beds onlap a steeply-dipping slope developed in the marls. The beds show thinning towards the slope and then parallel the slope. Figure 3.10a and b show how one of the beds changes along section when traced towards the slope and approaches the onlap. The example shown exemplifies bed thinning characteristics towards the slope. An individual bed approaching the slope retains its integrity with respect to thickness and grain size until the base of the bed actually reaches the change of slope where it onlaps and thins rapidly to leave only a thin, fine-grained bed as a veneer. Instead of the bed pinching out completely, it continues parallel to the slope as a thin, fine-grained sandstone for up to 5 m where it eventually pinches out. Enclosure 3.5.3D shows an example of beds approaching the slope, where they thin and then parallel it. These sandstones and mudstones
Figure 3.10. (a) Schematic diagram of turbidite beds approaching slope and onlapping, (b) summary logs showing internal facies organisation as beds approach the slope and onlap.
form a 50 cm thick thinning unit traced up the slope towards the north (Enclosure 3.5.3E and F).

The transition of marl to turbidite deposition at the base of the lower sandstone body, is seen as a change in colour from pale grey/blue to a light brown/grey, this transitional unit is approximately 100 cm thick with thin (<2 cm) fine-grained sandstones which parallel the underlying slope. Bedding is visible in the marls because of subtle grain size variations and changes in colour, and shows a low-angle truncation against the contact with the light brown/grey mudstone and thin sandstones. The orientation of the palaeoslope is obtained from bedding readings taken from sandstones which parallel it, and dips at 40°/207. Sandstones from the lower sandstone body which do not parallel the palaeoslope dip at 199°/199. Rotating these sandstones to horizontal gives a palaeoslope dip of 229°/219 (Figures 3.11 and 3.12). The relationship of the palaeochannel with the dominant marl slope in this part of the basin is described in Section 3.2.5.5.

Palaeocurrent data on Log B from the lower sandstone body on Enclosure 3.5.3 and Figure 3.9 show a varied distribution. Palaeocurrent readings taken from directional features such as flutes and ripples show a bimodal transport direction. Grooves trend northeast-southwest, with flutes oriented towards the southwest. Ripple orientations taken from these beds show a wide dispersion, but with a dominant orientation towards NNW.

The sandstone-marl contact in the this area can be contoured and the strike and dip calculated to give an overall slope orientation for the marls at the time of deposition of the sandstones. Figure 3.14 is a contour map for the present-day sandstone-marl contact constructed using a 1:10,000 scale map. The sandstone-marl contact dips at 229°/199 while the sandstones above the contact dip 229°/234, rotating the sandstones to horizontal gives a sandstone-marl contact orientation of 139°/128 (Figure 3.12).
Figure 3.11. Schematic diagram showing Tête du Ruch 'channel' architecture at the onlap after infilling and before tilting.
Figure 3.12. Marl-slope reconstruction for the Tête du Ruch outcrop area, reconstructed "channel" orientation is marked.
3.2.5.2b Interpretation

The lower sandstone body is interpreted to represent the infilling of a seafloor depression with sandstones deposited from low-concentration turbidity currents. The depression in the marls formed through erosion of the seafloor, either by turbidity currents or from sliding of the marls (the possibilities will be discussed in Section 3.4).

The sandstones are dominated by parallel-laminations with some ripple cross-lamination (Bouma Tb and Tc, respectively) which predominantly occurs at the top of beds or rarely within the parallel-laminated horizons. The alternations of parallel-laminations and ripples within individual turbidites is interpreted to represent fluctuations in flow velocity of the turbidity current as it passed a given point on the sea floor. Discrete mud clast horizons at the margins of beds where they onlap the slope are interpreted to represent erosion by the turbidity current as they encountered the palaeoslope. Turbidity currents in an unconfined setting expand in a radial direction from the source, providing that slopes are not to great (Kneller et al. 1991). Where the flow is confined in a particular direction, however, it will not expand radially but will undergo deflection and reflection against the topographic high and erosion may take place where the flow meets this high.

The lower sandstone body represents the infilling of a channel-shaped depression on the southeasterly dipping reconstructed marl palaeoslope with the channel trending parallel to the maximum dip of this palaeoslope. Palaeocurrent data from the base of turbidites records a trend of SW-NE orientated grooves, the orientation of which parallels the strike of the reconstructed palaeoslope and is not in general accordance with the overall orientation of palaeocurrent data which is south-to-north. This discrepancy of palaeocurrent orientation is interpreted to represent the deflection and reflection of northwardly directed turbidity currents towards northeast. The orientation of ripples from the lower sandstone body show a wide distribution which may reflect the complex interplay of northwardly directed turbidity
currents against an opposing southeasterly dipping palaeoslope with the added complex of a superimposed local topographic channel-like feature.

The previously described paralleling of the palaeoslope by thin fine-grained sandstones is shown in detail on Figure 3.11. The thin sandstones are the lateral equivalents of the thicker sandstones found in the axis of the channel-like feature and are interpreted to represent deposition at the margins of the turbidity current as it climbed the slope. Turbidity currents when travelling upslope will undergo a deceleration in velocity. As a result, the flow competence will decrease and only the finer grained fraction of the flow will be carried further up the slope, until only suspended sediment remains to be deposited as a drape of silt and mud on the slope. Subsequent turbidity currents infilling and passing over this depression built up a thickness of fines on the slope.

3.2.5.3 Tête du Ruch upper sandstone

3.2.5.3a Description

The upper sandstone body is characterised by beds in the order of 1-2 m thick, and they show features indicative of deposition from high- and low-concentration turbidity currents. The upper sandstone body is separated from the lower by a 12 m thick unit of mudstone with thin-bedded fine-grained sandstones which occur as thin (<50 cm) packets. The base of the upper sandstone comprises <30 cm thick medium- /fine-grained sandstones forming a 1 m thick unit (Enclosure 3.5.3). These deposits pass up into 3 m thick sandstones, the maximum grain size observed in these beds is pebble grade material.

Enclosure 3.5.3 shows the upper sandstone body in detail, and shows wedging of the beds towards the south. The wedging unit is approximately 5 m thick and comprises thinner sandstone beds up to 1 m thick; these thinner beds also wedge out towards the south.
beds show Bouma T_a graded bases which contain reworked clasts of shale and silt, and also show T_b divisions, together with distinct mud clast and pebble horizons which occur at all levels within the beds. Palaeocurrent data in the upper sandstone body indicates transport towards the northwest.

3.2.5.3b Interpretation

Deposition is essentially from sandy turbidity currents. Wedging of the beds seen on the north side of the section is interpreted to represent the influence of a topographic high on the deposition of sand from turbidity currents. This wedging unit is located close to the dominant palaeoslope and also overlies the channel-shaped lower sandstone body. Together, these factors may have controlled deposition from the turbidites, resulting in the formation of a composite mound-shaped bedform on the sea floor. Within the beds which make up the mound, there are discrete pebble horizons, interpreted to have accumulated from traction transport processes, probably within the turbidity current, and may represent flow-surge or flow unsteadiness as a result of interacting with the palaeoslope.

3.2.5.4 North Tête du Ruch

3.2.5.4a Description

The section trends ESE-WNW and shows a 270 m long exposure of Grès d'Annot Formation overlying the Marnes Bleues Formation. The contact is essentially planar, apart from at the WNW end where there is a scalloped surface in the marls. The marls are draped by a 2-3 m thick unit of thin, fine-grained, sandstone/mudstone couplets over the entire length of section, and overlain by amalgamated sandstone beds up to 2 m thick. On the WNW side where the scalloped surface is located, there is a drape of sandstone/mudstone couplets which preserve
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the depression in the marls, and which was infilled by sandstones in beds up to 50 cm thick, some showing cross-stratification (Enclosure 3.5.4B, Log C).

3.2.4b Interpretation

The depression in the marls is interpreted to represent an erosion surface formed in a similar manner to the depression located at the base of the lower sandstone body on the Tête du Ruch section (Enclosures 3.5.4.1, 3.5.4.2), and may even be an upslope continuation of the same structure. This is discussed further in Section 3.2.5.5.

3.2.5.5 Conclusions on the architecture of the Tête du Ruch area

The Tête du Ruch outcrops are shown in detail on Figure 3.9 with palaeocurrent data, sandstone dip and strike, and orientation of "palaeo-channel". A detailed stratigraphic map of the sub-surface contact between the Marnes Bleues and Grès d'Annot Formations in the Annot area has been constructed using a detailed 1:10,000 topographic map (Figure 3.14). Unfortunately, due to folding of the Grès d'Annot Formation after deposition, the stratum contours shown on the map do not correspond to the topography of the area at the time of deposition. To calculate the topography of the area during sandstone deposition it is necessary to rotate the sandstones back to horizontal using a stereographic net. The general dip and strike of the marls can then be calculated using the contour values shown on Figure 3.14. The above procedure was applied to the outcrops in the Tête du Ruch area (Figure 3.12). The palaeo-slope in the northwest of the area shows a general dip of 6° towards southeast and in the southeast a dip of 13° towards southeast. The "channel" trend is 132°-312°; the northern margin of the "channel" dips at 22° southwest. The southern "channel" margin is interpreted to dip in the opposite direction towards the north. The calculated "channel" axis orientation of 132°-312° is approximately perpendicular to the strike of the palaeo-slope, with the "channel"
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trending along the maximum dip of the slope. The northwestern extent of this depression in the marls is not known, but it may relate to the small depression in the marls on the North Tête du Ruch section (Enclosure 3.5.4).

The Tête du Ruch area during deposition of the sandstones was characterised by a submarine slope developed in the marls, which dipped towards the southeast at an angle of between $6^\circ$-$13^\circ$. Superimposed on this slope were small-scale topographic features up to 140 m wide and 30 m deep. These topographic features influenced deposition of passing turbidity currents, resulting in diverse palaeocurrent orientation and the formation of topographic bedforms in this area.

Large-scale topographic features in modern deep-marine environments have been documented and, to a lesser extent, smaller scale features similar to those from the Tête du Ruch area have also been recorded. One example of modern small scale seafloor erosional features is that described by Field & Clarke (1979) from the southern Californian Borderlands, obtained from high resolution side-scan sonar images on submarine slopes surrounding San Nicholas Island in the Pacific Ocean west of Los Angeles. Gullies up to 300 m deep were recorded; they \textit{(ibid.)} also noted smaller scale gullies in the order of 50-100 m deep, with some as shallow as 10 m. These gullies occur on slopes which range in gradient from $2^\circ$-$16^\circ$, and with the sides of the gullies sometimes exceeding $20^\circ$. Evidence suggests that the gullies formed from slumping and sliding of slope sediments. A similar process is postulated for the formation of the gully at Tête du Ruch and also for other similar erosion features seen in the Annot sub-basin.
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3.2.6 Les Scaffarels

The Les Scaffarels section (Figure 3.2 and 3.13, and Enclosure 3.6) is located at the southern end of the Grès d'Annot outcrops in the Annot Basin close to the junction of Route Napoleon-N202 and D908, and is visible along the Gorges de la Galange ~1.5 km south along the N202. The section trends SW-NE and offers a 750 m long exposure with up to a 200 m thick succession of Grès d'Annot Formation sandstones which form the spectacular Les Scaffarels cliffs.

3.2.6a Description

The section is dominated by structureless and massive (1 to 8 m thick) amalgamated sandstones with thin, <5 m thick, interbedded packets of thin, <50 cm thick, sandstone-mudstone couplets. The slide unit described in Section 3.2.3 can also be seen at the base of section on the northeast side, contact with the underlying marls in this section is not seen. The sandstones dip 16°/266.

The succession shows a predominance of amalgamated sandstones which comprise approximately 85% of the section, with the remainder composed of thin sandstone-mudstone couplets. Access to the section is via a footpath which follows the Les Scaffarels thin-bedded unit as marked on the section. The amalgamated sandstones show a variety of sedimentary structures, grain size variations and bed thicknesses. Amalgamation surfaces are picked out by grain size breaks, usually of coarse-granule and medium-grained sandstone material. Other indicators of amalgamation include truncation of sedimentary structures, such as parallel stratification, by erosive contacts. Cross-stratification may be present infilling depressions where the erosion is deep enough to allow cross-sets to form. At a height of 16 m from the base of the log, there is a 4 m thick unit with an erosive base infilled with cross-stratified sandstones. Located 40 m north of log B, there is a similar cross-stratified unit infilling a
Figure 3.13. Geological map for the southern region of the Annot sub-basin. Palaeocurrent data is from the Grès d'Annot Formation and palaeoslope dip and orientation shown is for the Marnes Bleues Formation underlying the Grès d'Annot Formation.
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seafloor depression, which can be traced for 10 m down current in a south-north direction (Enclosure 3.6F). The erosive surface shows an undulatory base with a maximum cut down of 1 m diminishing to 40 cm on the down-current northern side. The erosive depression is infilled with cross-stratified coarse-grained sandstones. Cross-sets are identified by grain size alterations and pebble concentrations. Cross-stratification is also present at the top of some turbidite sandstone beds (Enclosure 3.6G). The cross-stratification forms a 60 cm thick unit consisting of medium-grained sandstones on top of a 3 m thick coarse-grained turbidite sandstone. The underlying bed is dominated by discrete pebble clasts in the lower 2 m and can be traced for 10 m in a south-north section. The top 1 m of the sandstone bed has poorly developed parallel stratification, and discrete pebble clast lenses or nests up to 10 cm thick with sparse mud clast horizons. A feature similar to this is shown on log B (Enclosure 3.6) located 5 m from the base above the thin sandstone-mudstone couplet unit.

Amalgamated sandstone beds range in thickness from 50 cm up to a maximum observed thickness of 10 m, and are commonly in the order of 2-3 m thick. Maximum grain size observed in these sandstones is small cobble grade material which is invariably found at the base of beds and is seen at the base of the large erosional feature shown on Enclosure 3.7. The grain size at the base of beds is more commonly small pebble to granule-grade material. A measured section through these sandstones is shown on Log B-Enclosure 3.6, and includes a thin sandstone-mudstone couplet packet located 4 m up-section. Sedimentary structures include low-angle cross-stratification, parallel-stratified coarse/medium-grained sandstones, graded pebble to coarse-grained sandstone units and high-angle cross-stratification (Facies B2.2). Dish structures are also common, and locally there are armoured mud clasts.

Packets of thin sandstone-mudstone couplets reach maximum thicknesses of 5 m and are shown in detail on Enclosure 3.6C, log C, namely the Les Scaffarels unit. The packet essentially comprises medium-grained sandstones up to 40 cm thick, although some beds may be fine grained towards their top, but grading appears to be rare. Sedimentary structures include upper-flow regime parallel lamination, current-ripple structures which occur either
within or at the top of the sandstone beds, and rare grooves on the base of beds. Some beds contain 3 cm thick coarse/granule-grade sandstone intervals. The sedimentary structures and facies suggest deposition from sandy turbidity currents. Cross-stratification in one of the sandstone-mudstone couplets has been recorded (Enclosure 3.6, Log C) at a height of 4.5 m from the base of Log C. The bed is 30 cm thick and comprises a basal cross-stratified unit with an overlying 10 cm thick parallel stratified unit; between the two units there is a thin lenticular, and structureless interval containing pebble-sized material. This unit is interpreted to have been produced by traction currents. The mudstones between the sandstones have a distinct light grey/blue colouration and have an appearance similar to the underlying marls. They probably represent deposition of marly material during a relatively quiescent period of turbidite deposition. Within these muds, there are lenses of medium- to fine-grained sandstones up to 5 cm thick, some of which preserve poorly-developed ripple lamination and are interpreted to be starved ripples, i.e. formed under conditions of under supply of sediment relative to the competence of the turbidity current.

3.2.6b Sandstone Architecture

The amalgamated sandstone packets in the Les Scaffarels section reach thicknesses up to 60 m. Individual beds within these packets can be traced laterally along section for a maximum of 200 m and show no observable change in thickness. The amalgamated sandstone packets can be traced laterally over the entire length of section and correlate with the sandstone packets seen along the Coulomp Valley (Enclosure 3.3) and Annot town sections (Enclosure 3.7). The thin packets of sandstone-mudstone couplets on the Les Scaffarels section can be traced laterally for up to 400 m, but not continuously over the section. These packets of sandstone-mudstone couplets can correlated along the Coulomp Valley (Enclosure 3.3), but not along the Annot town exposures (Enclosure 3.7).
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On the Les Scaffarels section (Enclosure 3.6) the top four packets of sandstone-mudstone couplets (Plate 3.1) can be traced towards the SW where they are truncated by an erosion surface dipping towards the SW with a dip of 40°. The erosion surface can be traced for 100 m along section and cuts down into the sandstones below to an observable depth of 34 m. The erosional feature is interpreted as a channel margin. This channel-like margin dips towards the southwest and has a gentle concave-up orientation. The up-dip end of the channel margin terminates just above the last sandstone-mudstone couplet packet. At a height of 37 m from the observed axis of the channel, a sandstone bed can be traced across the entire length of section, and therefore is interpreted to define its upper limit in the northeast. The deeper part of the channel is less constrained, but is observed to cut into the amalgamated sandstones just above the Les Scaffarels thin-bedded sandstone-mudstone couplet packet. The Les Scaffarels unit is not seen to cropout in the Annot town exposures and its absence is interpreted to be due to truncation by the channel. The channel in a cross-sectional view is infilled with amalgamated sandstones which on closer inspection can be seen to drape the margins of the channel (Plate 3.1). The proximal-distal fill is described in detail in Section 3.2.7. The precise orientation of the channel is difficult to determine but is interpreted as striking approximately northwest-southeast.

Internal onlapping of turbidite sandstone beds is evident within the Scaffarels section, (Enclosure 3.6, Insets D and E), they occur just above and within the Les Scaffarels sandstone/mudstone packet, here named the Les Scaffarels Unit and show an apparent onlap and thinning towards the southwest of the section.

Inset D (Enclosure 3.6) shows four amalgamated sandstones on the northeast side of the section, and when traced towards the southwest they can be seen to onlap the underlying Les Scaffarels Unit of thin packets of sandstone/mudstone couplets. The sandstones contain dish structures and rare mud clasts. Also, immediately above these onlapping beds and directly below the previously described channel feature on the southwest side, there is local small-scale erosion. Thin sandstones in the underlying Les Scaffarels Unit also show a thinning

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Towards southwest (Enclosure 3.6, Inset E). This internal onlap and thinning is interpreted to represent tilting of the sandstones towards the East during deposition of the Grès d'Annot Formation and/or lateral migration.

3.2.7 Annot town

The Annot town section trends NNW-SSE comprising an approximate 250 m thick succession of Marnes Bleues and Grès d'Annot Formations which is 1600 m long, and located on the west side of the Annot Basin (Figure 3.2, and 3.13). The section shows outcrops of the contact between the sandstones and marls, with onlapping of the sandstones against a palaeoslope developed in the marls. The two outcrops are located at the southern and northern margins of the section (Enclosure 3.7A). The Grès d'Annot Formation in this section is characterised by amalgamated sandstones which dip westwards between 10-16°.

3.2.7a Sandstone Onlaps

The amalgamated sandstones on this section show two well exposed onlaps on to palaeoslopes developed in the marls, and both are easily accessible. Observations of both onlaps, using binoculars, reveals that bedding in the marls is truncated by the palaeoslope, suggesting that these slopes formed by erosion of the marls.

The southern onlap is shown in close (Enclosure 3.7Bi, Bii and Inset B) and reveals sandstone beds onlapping a slope over a horizontal distance of 200 m. The beds essentially comprise amalgamated sandstones which attain thicknesses up to 3 m, and the beds are generally coarse- to medium-grained with pebble bases. Grading is present in most beds, with parallel laminations present towards the top of some units together with isolated pebble lenses. Beds approaching the slope thin rapidly against it (Enclosure 3.7.B). In one case (Enclosure 3.7Bi)
a bed continues parallel to the slope for a distance of 5 m and forms a thin 20 cm thick sandstone unit. Thin turbidites also run parallel to the slope for a distance of 15 m (Enclosure 3.7Bii). Below these thin sandstones there is an isolated sandstone bed which also onlaps against the slope. The marls immediately below the sandstones, where exposed, have a sheared appearance and are interpreted to represent tectonically deformed sediment due to movement of the marls during compaction and folding of the Tertiary sediments.

Accurate measurements of the slope orientation have been made in the field, and by using bedding dip-and-strike values for the sandstones, and rotating them back to horizontal, it has been possible to calculate the slope in the marls at the time of deposition of the sandstones. This has been carried out for the broad sandstone-marl contact in the southern part of the basin, and gives a slope orientation of 10°/245. The sandstones dip 16°/266, which gives a reconstructed southern onlap slope orientation of 19°/100 (Figure 3.13).

The northern onlap (Enclosure 3.7A, 3.7Cl and 3.7Cii) comprises sandstones similar in facies to those on the southern side of the section. The sandstone beds show a southward onlap on to a slope developed in the marls; again observations using binoculars reveal that bedding in the marls is truncated by this slope, indicating that it was formed by erosive processes. This erosion surface appears shallower than the one on the southern margin due to the obliquity of the section with regards to the slope orientation. Returning this slope to its orientation during deposition of the sandstones gives a dip and dip-direction of 20°/056 (Figure 3.13). The strike of this slope is similar to that calculated for the slope at La Coste (Enclosure 3.8) and described in Section 3.2.8. Below the onlapping amalgamated sandstones, there is a 1.5 m thick unit of thin (<6 cm thick) sandstones separated by marls with rare parallel laminations preserved (Enclosure 3.7Cii), and some beds show a lenticular nature. The sandstones are parallel to the slope and syn-sedimentary folding of the beds is evident, and is interpreted to be a result of sliding of the marls on a steep slope.
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Evidence from truncation of marl bedding against the palaeoslope indicates that erosion has occurred either as a result of slope movement by sliding or by erosion from the passage of turbidity currents. Later deposition of thin fine-grained sandstones on the palaeoslope occurred, and the slope had a maximum dip of 20°.

3.2.7b Sandstone packet internal geometry

Generally the sandstones on the Annot town section are poorly exposed and inaccessible (Enclosure 3.7A). The maximum extent bed contacts can be traced laterally for is approximately 300 m, and over such distances the bed contacts appear parallel sided, but on the south side of the section there are irregular surfaces which cut down in to the underlying sandstones (Enclosure 3.7A, 3.7D, 3.7E).

A major erosion feature is shown on Enclosure 3.7D, together with a close-up field sketch depicting a large-scale flute-like erosional feature in a south-north proximal-distal orientation. It cuts down to a depth of 20 m into the underlying amalgamated graded sandstones with a pebble lag at the base with clast up to 15 cm in diameter. The sandstones are predominantly medium grained with poorly developed parallel stratified tops; armoured mud clasts are also seen at the base of some beds. The cut-down is infilled with large 20 m high cosets of cross-stratified sandstone which dip at an angle between 15-20°, but with sigmoidal profiles and with pebble stringers marking the foresets of the infill. The majority of the infilling sand is medium to coarse grained becoming pebbly towards the base, the cross-stratified sandstones pass up into medium-grained, parallel-stratified sandstone up to 2 m thick. This cut-down occurs approximately 10 m above the infill of the cut-down shown on Enclosure 3.7E.

The other major erosion feature (Enclosure 3.7E) comprises a 20 m thick and 200 m long section of amalgamated sandstones exposed in a south-north proximal-distal section. This section is interpreted to represent the infill of the channel-like feature described in Section
3.2.6 (Enclosure 3.6A). This section is located at the Chambre and Jardin du Roi locations, and affords good views of the sandstone geometries, allowing detailed observations to be made. Enclosure 3.7 E shows a major erosion surface which can be traced along the entire length of section for a distance of 200 m with a maximum cut down to a depth of 8 m; other smaller scale cut downs are also shown on this section located below the major erosion surface (Enclosure 3.7E, 3.7F and 3.6D). They are interpreted to represent the fill of broad and shallow channels.

The major erosion surface on the south side of the section cuts down into the underlying sandstones in a step-like manner (Enclosure 3.7E), each step-down corresponds to the position of a single bed. At the centre of the section, the erosion surface can be seen to step up by 2 m and then cut back down again 3 m on the northern side of the section. The basal fill of the erosion surface consists of a lag deposit comprising pebble-sized clasts of extra-basinal material up to 15 cm in diameter set in a matrix of coarse- and very coarse-grained sandstones (Enclosure 3.7Eii and 3.7Eiii). Mud clasts are common, some of which are armoured, and also there are rare marl clasts. The topography created by this erosion is infilled with medium-to coarse-grained sandstones comprising planar- and cross-lamination, with the cross-stratification developed on the proximal side of the cut-down. The stratification is marked by variations in grain size, with discrete pebble horizons and coarse-grained sands marking foresets. The stratification can be traced down into the erosional features forming a cross-stratified unit. Along one of these dipping pebble horizons, there is a small scour 15 cm long and 5 cm deep, with a fill of pebble-grade material. Dish structures occur in the vicinity of the cross-stratification and are inclined at an angle up to 15° to the horizontal, parallel to the dipping cross-sets, towards the centre of the erosion feature the dish structures have a horizontal attitude (Enclosure 3.7Eiii). On the northern margin of the section, the erosion surface can be traced laterally for 60 m without any observable cut down. The cut-down on the southern side of the Jardin du Roi section is approximately 35 m long and 8 m deep and is asymmetric in cross section. The erosion surface can be traced towards the north where it cuts...
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down to a depth of 11 m. The asymmetry is flute-like in cross-section, with the steep side up-current.

3.2.7c Interpretation of major erosional features

Infilling of the erosional features is interpreted to have been by sediment deposition from traction transportation at the base of a turbidity current resulting in distinct parallel-stratification draping and infilling the sea-floor depressions.

A possible origin for the formation of the cut-downs is scouring of the sediments below the turbidity currents. The cut-downs were then infilled either by sediment from the same turbidity current or by subsequent flows. Comparable erosional features from modern ocean basins have been recorded by Hughes Clark et al. (1990) from the Eastern Valley of the Laurentian Fan, and ascribed to the 1929 Grand Banks turbidity current. Features include a proximal gravel-wave bedform, a distal macro-dune bedform and sea-floor erosion up to 60 m in depth (ibid.). The large-scale erosion features described in the Annot area are interpreted as forming in a similar manner to those described by Hughes Clark et al. (1990).

The large-scale cross-stratification is interpreted to be as a result of infill in a topographic low on the seafloor and not from migrating positive-relief bedforms. The first sediment to be deposited in the depression was pebble-grade material (Enclosure 3.7E).

The rest of the section on Enclosure 3.7E, below the major erosion surface described above, comprises amalgamated sandstone beds up to 3 m thick with maximum grain-sizes of small pebble-grade material. The sandstones show sedimentary structures indicative of traction transport processes from high-density turbidity currents. The sandstone beds generally have erosive bases and show a grain-size range from pebble to medium-/fine-grained sandstone, with an average of medium-/course-grained sandstone. The beds show poor grading or appear
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structureless and commonly have pebble-grade material at their base which may be loaded (Enclosure 3.7Evi). These pebble bases are useful aids to identifying otherwise concealed amalgamation surfaces. The sandstones do not show typical Bouma sequences, with only the $T_\alpha$ division being common, and are more akin to high-density turbidity currents. Stratified pebble horizons are present at the bases of some of these sandstone units and are ascribed to Facies B2.1 (Facies $S_2$ of Lowe 1982). Parallel stratification within these sandstones is common (Enclosure 3.7Eiv). Each stratified unit comprises alternations of medium- and coarse-grained sandstones on a scale of 5 cm, which are interpreted to have formed as traction carpets below high-density turbidity currents. Within these parallel-stratified sandstones, there are local scour-and-fills (Enclosure 3.7Ev) and asymmetric step-like erosion surfaces (Enclosure 3.7Ei). The shallow symmetrical scours are infilled with granule to fine pebble-grade sandstones and are up to 6 cm deep and 30 cm long. The step-like erosion surfaces occur over a lateral distance of 3 m, with each step down being approximately 8 cm deep, the base of the step down has 0.5 cm thick pebble/granule-size material which is then draped by cross-stratified coarse-grained sandstones of Facies B2.1 forming cross-stratification over the erosion surface until the feature is entirely draped. These small-scale erosion features may represent flow surging or unsteadiness within a turbidity current as it passed a point on the sea floor; the increase in velocity over a small time eroded previously deposited sediment, resulting in the formation of a small-scale topographic low on the seafloor which then became infilled. The infilling sediment is coarser grained than the surrounding sediment. A large proportion of the beds show normal grading, with pebble to granule-grade material at their base to coarse- to medium-grained sandstone at their top, and is interpreted as the Bouma $T_\alpha$ division, equivalent to facies group $S_3$ of Lowe (1982). Other sedimentary features such as dish structures are common in the $T_\alpha$ division. Cross-stratification occurs at the tops of some sandstone beds, and is interpreted to represent reworking of previously deposited sandstones and is ascribed to Facies B2.2 (Facies $T_\gamma$ of Lowe 1982).

Deep scouring also occurs at the base of these beds (Enclosure 3.7E). These features range in depth between 0.5 and 2 m, Enclosure 3.7Ev shows the infill of one of the scours which is
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approximately 2 m deep, at its margin with the underlying bed pebble lags can be seen draping down in to the scour.

3.2.7d Summary

The sections described on Insets D, E, and F (Enclosure 3.7) all occur within the channel feature described in Section 3.2.6 and are interpreted to represent the sediment infill of this channel. The sandstone beds in these three sections are laterally discontinuous and show evidence for by-pass of sediment in the form of large-scale and small-scale erosion features, not seen on other sections from the Annot sub-basin.

3.2.8 La Coste

3.2.8a Description

The La Coste section (Figure 3.2, Enclosure 3.8) trends WSW-ENE and comprises a 200 m thick succession of amalgamated sandstones separated in the middle by a 20 to 30 m thick interval of non-amalgamated sandstones of <1 m thick. The sandstones show an apparent onlap towards the west-southwest onto a slope developed in the underlying Marnes Bleues Formation which dips eastward with an apparent angle of 8°. The contact between the sandstones and marl is as indicated by truncation of bedding planes within the marls along the onlap surface.

The only accessible part of the section is located at the contact of the sandstones where they onlap the underlying slope in the marls. At the base of one of the amalgamated sandstones, there are large flutes developed which cut down into the underlying marls. There are also some thin-bedded sandstones below (<0.1 m, Facies C2.3) which lie parallel to the onlap
The truncation of bedding in the marls at the onlap surface over the entire slope (Enclosure 3.8.A) is interpreted to be the result of wide-scale erosion of the marls either by turbidity currents cutting down into the marls as they pass over the slope and/or syn-sedimentary sliding of the marl palaeoslope prior to the deposition of the overlying sandstones. Bedding readings from within the sandstones, and at the onlap surface, indicate that the slope during deposition dipped 10°/046. After erosion, turbidity currents passing over the marls first deposited thin sandstones (<0.1 m) parallel to the slope, and these were then onlapped and eroded into by the overlying, thick-bedded amalgamated sandstones. The non-amalgamated unit half-way up the section is interpreted to represent a change from predominantly sand-rich to muddy turbidity currents.

3.2.9 Fugeret

3.2.9a Description

The Fugeret section (Figure 3.2, Enclosure 3.9.A) comprises a 100 m thick succession of sandstones, mainly as amalgamated beds. The section trends NW-SE and is well exposed on the south side of the valley, though for the most part the exposures remain inaccessible. On the north side of the valley, however, access to the outcrops is possible and a log (Enclosure 3.9.C) was measured through the lower part of the succession from the base contact with the marls to 40 m up-section. In the lower 20 m, the sandstone beds range in thickness between
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0.3-5 m and are ascribed to Facies C2.1 with Bouma T_bed divisions absent; the maximum grain size observed in these beds is coarse grained sandstone. The beds are separated by silty muds up to 1m thick which contain thin fine-grained sandstones less than 5 cm thick. The upper half of the log comprises amalgamated sandstones showing facies transitions of Facies A2.5-C2.1 (Bouma T_a only). Beds range in thickness from 0.5-6 m, containing pebble sized material invariably found in the basal part of the beds.

On the southern side of the valley, the sandstones onlap the marls over a distance of 200 m in an apparent southwesterly direction (Enclosure 3.9A and 3.9B), and although accurate orientations of the onlap surface are unobtainable, they show an apparent dip of 8° towards the southwest. The sandstone beds do not onlap directly on to marls but on thin sandstones and marly muds which parallel the onlap surface. Evidence of synsedimentary deformation along this onlap is seen on Enclosure 3.9B and 3.9D, where convoluted bedding can be seen, bounded by sandstone beds on either side of this unit, which are undeformed and parallel the onlap surface. At the northeastern end of the section (Enclosure 3.9A), there is a scallop-shaped depression 20 m wide and 6 m deep within the marls which is infilled with mudstone, and thin sandstones being lenticular in shape across the depression.

3.2.9b Interpretation

The section records a palaeoslope in the marls with local erosion causing the development of a scallop-shaped depression, either by sedimentary sliding or erosion by turbidity currents, and which was subsequently infilled with thin-bedded, fine-grained turbidites. The dip of the slope developed in the marls during deposition of the sandstones was low enough to allow turbidites to be deposited parallel to it, but steep enough to cause sediment sliding.
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3.3 Sub-basin morphology

The sandstones from the Annot sub-basin are well exposed and show good lateral and vertical continuity at outcrop. They also show the nature of the contact between the sandstones and the underlying marls. The sandstones at Annot have been folded into a synclinal structure, probably during the Miocene when Alpine compression affected the basin (Siddans 1979).

The beds within the basin generally dip towards the west. Beds on the west side dip at approximately 15°, while in the east they dip at between 15°-25° to give the synclinal structure with both fold limbs dipping in the same direction, with no overturning of beds. Beds towards the north of the basin dip towards the southwest at angles up to 28°.

Using stereographic projection techniques, the dips of the sandstones have been rotated back to horizontal along with the dips of sandstone-marl contact to give a reconstructed slope orientation. Compaction of the palaeoslope is interpreted to have been minimal in the marls as it has been calculated that the marls contain between 55% and 60% calcium carbonate (Ravenne et al. 1987, Hilton this study). Also, early cementation is thought to have occurred thereby arresting significant compaction. At some localities it is not possible to measure the sandstone-marl contact orientation, either due to inaccessibility or lack of exposure, but where this was the case a structure contour map for the sandstone-marl contact was constructed (Figure 3.14). Data for contouring are limited to outcrops as no boreholes have been sunk in to the sandstones, therefore contours have been drawn to best fit the data. Details of sandstone dips and sandstone-marl contact orientations are described in detail in Section 3.2. The paucity of subsurface data on the sandstone-marl contact from the Annot sub-basin may, however, lead to a contour map which may have errors present which must be taken into account when a sub-basin reconstruction is produced.

Reconstructed slope orientations of the sandstone-marl contacts around the Annot sub-basin are shown on Figure 3.15. The map shows the positions of the reconstructed slopes within the basin, with their the calculated orientation and dip.
Figure 3.14. Stratum contour map for the present day contact between the Grès d'Annot Formation and underlying Marnes Bleues Formation, height of contours are in metres above present day sea-level.
Figure 3.15. Palaeoslope reconstruction in the Annot sub-basin during deposition of the Grès d'Annot Formation.
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On the southern side of the basin, the slope orientation varies from NNW-SSE to NE-SW with a dip of between 10° and 15° towards the east. This slope is also predicted through observations of amalgamated sandstone-packet terminations and shaling out towards the west. At the very north of the basin the slope is oriented 6° to 13°/135 while the slope on the northeastern side dips 8° to 10°/225. At the southeastern margin of the Tertiary outcrops in the Annot area, the Rouaine Fault Zone controlled sedimentation, (Section 3.1.2, and Figure 3.3), (Pairis 1971, Apps 1987, Ravenne et al. 1987), and created a topographic feature on the seafloor and controlling deposition of the sandstones and older Tertiary sediments (Section 3.3.1). This fault zone produced differential thicknesses in the Tertiary succession, with thicker deposits in the hanging wall of the fault, and is described in Section 3.3.1.

Taking these slopes and topographic features into account, and the vertical thickness of sandstones, a three dimensional reconstruction of the basin prior to deposition of the sandstones has been constructed (Figure 3.16a and b), Figure 3.16 is a view looking towards the northwest with the palaeoslopes and local erosional features indicated (also marked is the interpreted topography created by the Rouaine Fault Zone which appears to die out towards the NE and SW).

In conclusion, the Annot sub-basin was confined on its western margin by an approximate north-south striking, eastward dipping slope, and which towards its northern margin changed orientation to strike NE-SW and dip towards southeast. The northeastern reconstructed basin margin slope is estimated to have been 0-150 m vertically below the easterly dipping slopes on the western side, thus creating a complex northern basin margin which was raised at its western, northern and northeastern regions. The southern and southeastern margins of the basin are interpreted to have been the principal sediment sources for material to enter the basin. Palaeocurrent trends (Figure 3.2) indicate a sediment source from the south to southeast, the main slope dipped NNW (this is also in agreement with the palaeocurrent readings taken from the proximal equivalents of the Grès d'Annot Formation at Entrevaux and St Antonin described in Section 3.5).
Figure 3.16 (a) Reconstruction of the Annot sub-basin prior to the deposition of the Grès Annot Formation. (b) Detailed reconstruction of the Rouaine-Daluis Fault zone prior to deposition of the sandstones: data for this construction is taken from this study, Pairs (1971), Apps (1987), Ravenne et al. (1987), and from the geological maps of the Annot area: Entrevaux sheet 945, 1:50,000 (BRGM 1980), and Gap and Nice 1:250,000 sheets 35 and 45 respectively, (BRGM 1980).
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3.4.3 Architecture of the Grès d'Annot Formation

3.4.1 Introduction

The sandstone succession at Annot is divided into a Lower and Upper Member, on the basis of the occurrence of amalgamated and non-amalgamated sandstone beds, (Section 3.3.4). The non-amalgamated sandstones outcrop on the eastern side of the basin, whereas the amalgamated sandstones outcrop mainly on the western side of the basin separated by the north-south trending Coulomb River Valley.

3.4.2 Lower Member

Individual sandstone beds can be traced without any observable change in thickness for distances up to 2 km in a N-S direction and 1.5 km in an E-W direction, and show that the beds possess an overall sheet-like geometry. Beds approaching onlap surfaces, however, show termination (e.g., see Enclosure 3.2 and Section 3.2.2). Individual beds approaching this slope which dips 12° towards the east, onlap but do not change in thicknesses or internal characteristics until they are less than 1-2 m away from the slope (cf. Enclosure 3.1F). Beds can be traced towards the west of the exposure where individual bedding planes become obscured due to bed amalgamation, in this region, dish structures are common. This feature of amalgamation and loss of individual bedding is interpreted to represent the influence of a steep slope of 50° on the sedimentation from turbidity currents. The slope formed by draping of marls above a normal fault developed in the older Tertiary succession. Figure 3.17a represents a N-S transect through the Annot sub-basin and Figure 3.17b represents an E-W transect showing the major sandstone packets of the Lower and Upper Members and their relationship with the underlying marls. A three dimensional reconstruction for the east of the basin is shown on Figure 3.18a.
Figure 3.17 (a). South-North stratigraphic transect through the Annot sub-basin showing lateral extent of sandstone bodies. Enclosures used to construct this section are marked.
Figure 3.17 (b) East-West stratigraphic transect through the Annot sub-basin showing sandstone body architecture and the relationship with the underlying Marnes Bleues Formation. Enclosures marked on this diagram are described in Section 3.2.
Figure 3.17 (c) NNW-SSE stratigraphic section through the Annot sub-basin showing lateral extent of sandstone bodies and their relationship with the underlying Marnes Bleues Formation. The section is based on data presented in Enclosure 3.7 and Section 3.2.7.
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3.4.3 Upper Member

The Upper Member is characterised by packets of amalgamated sandstones and interbedded <5 m thick packets of thin-bedded sandstone-mudstone couplets, and a large slide horizon in the lower part of this Upper Member. The outcrops on the western side of the basin are well exposed and provide good lateral correlations of sandstone packets, especially along the Coulomp River Valley on the western side (Enclosure 3.3), north of the village of Les Scaffarelles (Enclosure 3.6), and east of Annot town (Enclosure 3.8). Summary stratigraphic sections for the basin including the above mentioned sections are shown on Figures 3.17a, b, and c.

3.4.3.1 Amalgamated sandstone packets

Amalgamated sandstone packets range in thickness from 30-80 m. The amalgamated sandstone packets can be traced for 5 km in a S to N (proximal to distal) direction. Their southern extent at Les Scaffarelles is represented by the present day erosion surface. At the northern margin, the precise extent of the sandstones is unclear, although sub-basin reconstruction for the northern margin indicate an onlap onto a southerly dipping palaeoslope, (Figure 3.17b). In an E-W section (Figure 3.17a), the amalgamated sandstone packets terminate towards the east along the Coulomb Valley, along the present day erosion profile, although the sandstone packets are interpreted to have extended over the entire eastern side of the basin. On the western side, they onlap an E-dipping surface in the marls as deduced by correlation of the sandstone packets around the basin (Figure 3.18a. and b). A 174 m thick vertical section of amalgamated sandstones is seen to onlap on to the marls, and a further 140 m thick vertical section of amalgamated sandstones occurs above these onlapping units. Also, the amalgamated sandstone packets onlap two small erosion surfaces created on the easterly-dipping marl slope at the Annot town exposures (Enclosure 3.7, Section 3.2.7, Figures 3.17c and 3.18c). Within the basin, the slide horizon in the Coulomp Valley and Les Scaffarelles
sections, represents a phase of substantial slope instability possibly caused by failure of sandstones deposited on this slope triggered by earthquakes or by syn-depositional tectonic movement below this slope. This slide horizon can also be seen to outcrop on the Les Scaffarels section (Enclosure 3.6), although the contact with the sediments above and below is poorly visible.

3.4.3.2 Internal architecture of amalgamated packets

Where beds could be traced in a proximal-distal, S-to-N direction, along the Coulomb River Valley section (for distances up to 2 km), the beds appear essentially sheet-like. Detailed photo-montages reveal that some beds have erosive bases, and others wedge out in either a south or north direction over distances between 10 m to 20 m. Also present are large, up to 2 m long and 1 m deep flute-shaped scours at the base of some beds and which contain cross-stratified coarse-grained sandstone infill (Enclosure 3.7 F and G), low-angle cross-stratification is also present at the top of some bed (see Sections, 3.2.3, 3.2.6, 3.2.8, 3.2.7).

3.4.3.3 Large scale erosion features

On the Les Scaffarels section (Section 3.2.6, Enclosure 3.6) a large erosion surface cuts down into the amalgamated sandstones to a depth of 34 m, over a cross-sectional horizontal distance of 150 m, and is interpreted to represent a broad channel base with infilling sandstones. In cross-section, the channel-fill sandstones drape the sides of the channel, while in a transverse section they reveal a complex infill (described in Section 3.2.7), and can be traced for about 250 m in a northerly direction. Approximately 40 m vertically above, another smaller scale flute-shaped erosion feature occurs for 90 m in a longitudinal section and cuts down at least 20 m into the underlying sandstones, and is infilled with amalgamated sandstones Section 3.2.7, Enclosure 3.7, Figure 3.18c).
Figure 3.18 (a) Sand-body architecture from the St Benoit and Braux sections showing the relationship with the underlying marls. (b) Sand-body architecture along the Coulomb Valley section and the relationship with the underlying marls.
c) Annot

Erosional features with infilling sands

"Channel" margin with draping infill

Grès d'Annot Formation
Amalgamated medium-pebbly sandstones

Thin <1-m-thick sandstone/mudstone couplets

Marnes Bleues Formation
Slide horizon of sandstones

Calcaires Nummulitiques Formation
Marls

Cretaceous Limestones
Limestone

Apparent onlap of sandstones

Figure 3.18. (c). Three dimensional reconstruction of the sandstone architecture for the Annot sub-basin at end deposition of the Grès d'Annot Formation. The relationship of the Grès d'Annot sandstone Formation with underlying Marnes Bleues Formation is shown.
3.4.3.4 Thin-bedded sandstone-mudstone couplets

The thin-bedded sandstone-mudstone couplets occur in packets up to 5 m in thickness and are laterally continuous across the entire basin for distances of 4.5 km in a N-S direction along the Coulomp Valley section (Enclosure 3.3), and 2 km in an approximate E-W direction on the Les Scaffarels section (Enclosure 3.6). The amalgamated sandstones above do not cut down into these units. The thickest packet of thin-bedded sandstone-mudstone couplets, the Les Scaffarels Member, contains thin-bedded sandstone turbidites separated by marl-rich mudstone. Logs through these beds show lateral thinning towards the west, and probably represents syn-depositional slope evolution in the western part of the sub-basin, which is also picked out by the westerly onlaps of the overlying amalgamated sandstones.

3.4.4 Types of sandstone-marl contact

There are two types of sandstone-marl contacts: (1) onlap of sandstones onto a slope in the marls with bedding in the marls parallel to the slope and indicating no obvious erosion, and (2) a slope in which bedding in the marls is truncated by the slope indicating erosion of the marls. Areas where erosion of the marls occurs are described in Sections 3.2.4/5/7/8/9. Areas that show no erosion at the sandstone-marl contact are described in Sections 3.2.1/2/3/4. These erosional surfaces have been interpreted as forming by two processes (i) slope failure of the inclined marls, and/or (ii) by erosion as turbidity currents passed over the marls. Some of the erosion surfaces have a crescent shape and have subsequently been infilled with turbidites, (e.g., Section 3.2.5.2) and appears to be aligned parallel to the maximum slope inclination in the marls. The base of these features are not marked either by any channel lag deposits nor do they appear to be erosional by-passing features on the slope. A slope failure origin is favoured here.
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3.5 Vertical sandstone facies variations in the Annot sub-basin

3.5.1 Description

The overall sandstone succession at Annot is summarised in Figure 3.19. The section is divided into two parts, based on the occurrence of amalgamated sandstones in the upper part of the succession which do not show typical Bouma T cde divisions. The lower part of the succession (0-325 m), is characterised by sandstone beds deposited from turbidity currents, commonly showing Bouma T abd divisions with less common T c divisions. Enclosure 3. Dii and Enclosure 3.2B provide a representative section of the facies seen in this lower part. The beds range from decimetres to a maximum observed thickness of 5 m, thicker beds in this lower part of the succession may show dish structures. Mud clasts occur within some of the beds, especially where they onlap the underlying marls (Enclosure 3.2B). The beds in this lower part of the succession show depositional structures indicative of deposition from sandy turbidity currents, some of the thicker beds (2-5 m) being deposited by flows which probably started off as a high concentration and evolved to lower-concentration flows. The beds in this lower part of the succession are essentially sheet-like across their outcrop and rarely show bed amalgamation.

The amalgamated sandstones vary in thickness from 0.5 to 6 m, generally have erosive bases, and show a grain size range from pebble to medium-to/ fine-grained sandstone, with an average of medium-to/coarse-grained. They generally show poor grading or appear
Figure 3.19. Summary log of the Grès d'Annot Formation from the Annot Basin. The log was constructed using measured sections and detailed photography.
structureless, although some beds show medium/coarse-grained parallel-stratified sandstone which reach thicknesses up to 3 m. Loaded pebble bases are also common, and they are used to recognise/define otherwise unidentifiable amalgamation surfaces. The amalgamated sandstones do not show typical Bouma T bcde divisions. The parallel-laminated sandstones described above comprise alternating grain sizes of medium- and coarse-grained sandstones, deposited as traction carpets which show poor inverse grading and belong to Facies B2.1 (Facies S of Lowe 1982). Within these parallel stratified sandstones, there are, local scour-and-fills and step-like stratification horizons (Enclosure 3.7E) interpreted to reflect surging or flow unsteadiness of the turbidity currents during deposition. This resulted in variations between traction sedimentation, traction carpet sedimentation or suspension sedimentation persisting until the high-density flow has evolved to a low-density flow (Lowe 1982). The top of these sandstones may have low-angle cross-stratification up to 50 cm in thickness, which has been ascribed to Facies B2.2 (Facies T of Lowe 1982), and represent traction transport by the residual turbidity current after deposition of the high-density load. A proportion of the beds show normal grading from pebble-to/granular-clast sized at their base, to coarse-to-medium-grained sandstone towards the upper part of their Bouma T division, with dish structures present. The top of these beds have parallel-laminated medium-grained sandstones showing Bouma Tb division. Deep scours in the order of 1-2 m are observed (Enclosure 3.7E), which are infilled with cross-stratified coarse to medium-grained sandstones with rare pebble stringers (Enclosure 3.7E).

3.5.2 Interpretation of up-section changes: aggradation/progradation/retrogradation

Controls on sedimentation in the deep-sea can be divided into three primary but interacting controls (e.g., Stow et al. 1985 and references therein), these being: sediment supply, tectonic setting, and sea-level fluctuations. To understand the controls on the deep-marine basin it is necessary to understand what is happening in the shallow-marine source area. Interpretation is aided when direct correlations from the shallow to the deep-marine basin are available, either
using facies shifts, i.e., litho-stratigraphic or bio-stratigraphic correlations based on macro or micro-fauna.

In the Annot sub-basin, the source area was located towards the south, and remnants of this are interpreted to be found in the St Antonin sub-basin fan-delta system. Direct correlation between these two basins is not possible as bio-stratigraphic correlations have not been obtained and direct litho-stratigraphic correlations are not possible.

The lower half of the succession is represented by turbidites deposited from sandy turbidity currents. Variations in bed thickness in this lower member are interpreted to represent fluctuations in the volume of sediment mobilised to form each turbidity current, which in turn may relate to a change in sediment supply to the transitional zone where sediment is temporarily stored, or differing sediment initiation processes in this zone.

The upper member consist of two distinct litho-facies, these being: up to 80 m thick packets of amalgamated turbidites and <5 m thick packets of thin (<0.5 m) sandstone/mudstone couplet turbidites. The controlling factors for this alternating of facies is poorly understood and postulations such as changes in relative sea-level in the shallow-marine environment and varying intensity/location of tectonic activity in the transitional zone or source area/drainage basin are equally viable. To deposit packets of basinal amalgamated coarse-grained sandstone requires a fall relative base-level, e.g. lowered seal level, to cause the stripping of sands from the shallow-marine source area. Increasing tectonism in the drainage basin may also lead to increased sediment supply to the shallow-marine environment and ultimately to the deep-marine basin. A counter argument can be constructed for the deposition of the packets of thin sandstone/mudstone couplets.
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3.6 Relationship of Annot and surrounding areas

Proximal outcrops to the Annot sandstones, are, by inference from palaeocurrent data, located towards the south of Annot. These proximal outcrops are located in the vicinity of Entrevaux and St Antonin (Figure 3.1).

3.6.1 Stratigraphy of Entrevaux

3.6.1a Description

The Tertiary outcrops at Entrevaux (Figure 3.1) are located along the Var River Valley in the Alpes Maritimes region of southeast France, and are located in an E-W trending syncline with the sandstones dipping steeply to the south with dips up to 40° to 80°. The Tertiary succession at this locality comprises the Poudingue d’Argens, Calcaires Nummulitiques, Marnes Bleues, and Grès d’Annot Formations. Sandstone exposures of the Grès d’Annot Formation are located 4.5 km NNW of St Antonin, and 9 km southeast of the sandstones at Annot. The sandstones are poorly exposed, but a measured section has been taken through part of this succession along a recently excavated track (Figure 3.20).

The section reveals a 160 m thick vertical succession of sandstones from the basal contact with the underlying Marnes Bleues Formation. The base of the section records a gradation from grey/blue coloured marls to brown silty mudstones with thin (<5 cm) fine-grained turbidites in the basal 5 m (Figure 3.20). The lower 0-30 m of succession comprises silty mudstones with thin-bedded fine-grained sandstones which attain a maximum thickness of 63 cm, and are generally <20 cm thick. These thin sandstones show rare grading and poorly developed parallel laminations in the thicker beds. At a height of 30 to 90 m from the base of the succession the section records an increase in the amount of thicker-bedded sandstones, and a decrease in silty mudstones. These sandstones are up to 2.6 m thick and are generally
Figure 3.20. Measured section from the Entrevaux outcrops of the Grès d'Annec Formation, taken from the base of the formation in contact with the underlying marls. The succession records an overall coarsening and thickening-up sequence.
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<1 m thick, they are interpreted to be deposited from sandy turbidity currents. The beds are normally graded (Bouma T<sub>a</sub>), with a maximum grain size of pebble-grade material, which fine-up into medium-grained sandstone showing parallel laminations (Bouma T<sub>b</sub>) and are ascribed to Facies C2.1. Ripple lamination has been recorded in a limited number of beds, and lower-flow regime parallel laminations are common (Bouma T<sub>d</sub>). Beds may be amalgamated, but are generally separated by silty mudstones and packets of thin fine-grained sandstone-mudstone couplets (Facies C2.2, C2.3). The Bouma T<sub>a</sub> divisions contain mud clasts, either dispersed or concentrated along sub horizontal horizons and are generally located in the upper half of the T<sub>a</sub> division.

The upper part of the sandstone succession, 90-160 m, comprises thicker-bedded and coarser-grained turbidites with thin-bedded fine-grained sandstone-mudstone couplets separating beds, with small amounts of bed amalgamation. The beds range in thickness from 0.1 m to 5.2 m, with an average thickness of 1.5 m. Bouma T<sub>abd</sub> divisions are common, with rare T<sub>c</sub> divisions. The maximum observed grain size is pebble-grade material with clasts up to 2 cm in diameter, reworked mud and shales are common with clasts up to 40 cm in length. These are generally found in the thicker beds. Reworked clasts are either distributed within the T<sub>a</sub> division or located towards the top of the division forming discrete horizons. At a height of between 90 m to 100 m, the section records three contorted and brecciated sandstone and mudstone horizons, 2.15 m, 2.20 m, and 0.97 m thick respectively. These sandstones comprise medium- to fine-grained beds up to 10 cm thick, which are brecciated and set in a matrix of silty mudstone. They are interpreted as forming from tectonic deformation as a result of folding. The base of beds show little erosion with rare grooves and flutes which have a random orientation (Figure 3.20). The stratigraphic top of the Grès d'Annot formation is not preserved in this section.

Throughout the section there is an abundance of organic detritus which tends to be concentrated in the T<sub>cd</sub> divisions of turbidites. The organic detritus is in the form of plant
debris which has been preserved as vitrinite-rich deposits, forming thin 10 cm long, and 1 cm thick lenses.

3.6.1b Interpretation

The overall succession at Entrevaux records an up-section change from mudstones with thin-bedded fine-grained sandstones of a distal turbidite facies, to thicker-bedded and coarser-grained turbidite sandstones of a more proximal facies. This change can be interpreted in a number of ways: (1) the change can be interpreted to be a result of submarine progradation of lobe-like sandstone bodies on the sea-floor with a constant sediment supply, or (2) by an increase in sediment supply to the basin as a result of increased tectonics in the drainage basins. It is not possible to say equivocally which of the above two explanations are correct.

3.6.2 Stratigraphy of St Antonin

3.6.2.1 Introduction

The Saint Antonin syncline, comprises a succession of Tertiary sediments (Figure 3.21), the exposures are located 5 km southeast of the exposures at Entrevaux. The syncline axis trends ENE-WSW, preserving a 11 km long and 2.5 km wide area of Tertiary sediments. The Tertiary sediments in this syncline comprise the Poudingues d'Argens, Calcaires Nummulitiques, Mames Bleues, and Grès d'Annot Formations. Outcrops of the Grès d'Annot Formation are the most southerly in the study area, and represents the most proximal exposures of the formation. The syncline is inverted on its northern limb, which dips at ~70° towards north, while the southern limb dips at an angle between 30° to 40° towards north. Folding of these Tertiary sediments has been ascribed to Early Miocene age during the main Alpine deformation (Graham 1978).
Figure 3.21(a). Geological map of the St Antonin syncline with three stratigraphic members and their representative simplified stratigraphic logs (from Stanley 1980 (after Bodelle 1971)).
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The Grès d’Annot Formation from the St Antonin area is markedly different from exposures in other study areas. The section is dominated by coarse-grained matrix- and clast-supported conglomerates with clast sizes up to 1 m in diameter. Other facies including cross-stratification, structureless sandstone and wave-ripples. Other features of interest include an andesite breccia located towards the top of the sequence.

3.6.2.2 Previous research

The Tertiary sediments of the St Antonin syncline have been studied by Bodelle (1971), he recognised three sequences in the Grès d’Annot Formation, named the Lower, Middle, and Upper Formations, he described their composition, grain size, and facies. Later work by Stanley (1980), based on that of Bodelle (1971), included a study on the temporal arrangement of the sandstone and conglomerate units, interpreting the three formations described by Bodelle (1971) as thick coarsening-up successions or megasequences. Stanley (1980) proposed a submarine slope environment for the deposition of the St Antonin conglomerates. Palaeocurrents were predominantly orientated towards northwest, and, sediments were derived from a southern provenance. Clasts of igneous and metamorphic clasts including granitic and ophiolitic material were akin to the outcrops on Corsica and Sardinia (ibid.). He states that the St Antonin conglomerates are contemporaneous and genetically related to the sandstones located to the north and northwest in the Entrevaux and Annot exposures of the Grès d’Annot Formation.

The three megasequences, A, B and C, described, and their relative positions in the basin are shown on Figure 3.21. The three megasequences described by Stanley (1980) are summarised below.
3.6.2.2a Lower Member

The Lower Member comprises a 400 m thick succession of sandstones and conglomerates, the lower 120 m comprise friable, coarse-grained, poorly-sorted sandstones. Sedimentary features include cross-stratification indicating transport towards northwest. Thin shale drapes, asymmetric ripples, and amalgamated sandstones up to 4.5 m thick are common. Erosional features such as cut-and-fill, moderate sized channels and rip up clasts recording migrating channel processes are present. The sandstones display both horizontal and foreset stratification, many of the foresets contain mud drapes. The first major conglomerate is located at ~120 m from the base of section, forming a 5 m thick, sharp based, matrix-supported cobble and boulder mudstone with poorly developed vertical grading of larger clasts. This unit is followed by a thick succession of sandstone, and conglomeratic strata, varying in thickness from 0.5 m to 4 m, some of the conglomeratic beds are sharp based and lenticular and others channelised. The top of this Lower Member is marked by a ~70 m-thick unit of thin sandstone and mudstone.

3.6.2.2b Middle Member

The Middle Member is ~350 m thick, the base is represented by a coarse debris, matrix-supported conglomerate which is marked by the presence of andesitic cobbles and boulders of local derivation. The sandstone strata and sandstone matrix of the conglomerates tend to be darker than those of the Lower Member, this change in colour is related to the presence of ferro-magnesium-rich minerals derived from the andesite. A 50 m-thick unit of alternating laminated siltstones and shales is interbedded between conglomerate beds, these are then overlain by massive structureless amalgamated and graded sandstones and coarse conglomerate lenses and channels. Above this a 100 m thick unit comprising large blocks >1 m, which outcrop northeast of the village of St Antonin. Above this, 150 m of silty micaceous shale, laminated siltstones and fine-grained sandstones are located, the bases of which show
tool-marks with transport towards northwest. This siltstone member is partly truncated by an andesitic flow breccia which marks the top of the Middle Member. The contact of the upper member is erosive into these shales.

3.6.2.2c Upper Member

This is ~200 m thick, and comprises very coarse conglomerates showing both lenticular and channelized characteristics; also higher proportions of andesitic debris are present. Gypsum stringers present in the uppermost shale section are interpreted by Stanley (1980) to indicate the end of the marine regime in this region.

3.6.2.3 Detailed section of the Lower Member

The lower 125 m of section are dominated by sandstones (Figure 3.22). The lower 30 m of section comprises normally graded beds up to 2.6 m thick, with a maximum grain size of small pebble-grade material. Parallel-laminations are present at the top of some beds, rare reworked mud clasts are found within some beds. Beds are generally amalgamated with rare muds separating them. Flutes on the base of one bed immediately above the marls indicates a transport direction towards northwest. At a height of 30 m to 125 m the sediments show a change from normally-graded units to beds which show foreset cross-stratification with cosets up to 60 cm-thick, and parallel-stratified units of sand which may be lenticular or laterally continuous along strike. Structureless and graded sandstone beds are a minor component of this part of the section. The base of beds can be erosive and may contain mud clasts. Mud drapes are present on top of and at the base of some cross-stratified beds and is invariably bioturbated; ophiomorpha burrows are also present. Cross-stratification shows an apparent transport direction towards west and a corrected direction towards northwest. current ripples are rare, but where observed record a transport towards northwest. It is noted that foresets
from towards the base of the lower member.

C. Massive structureless sandstone lenses

D. Cross-stratified sandstone of Facies B2.2

E. Conglomerate and sandstone interbeds.

F. Interbeds of conglomerate and massive sandstone.

G. Opposing figure cross-bedding from the upper part of the Lower Member. Later, part of the Lower Member can be examined from the lower part of the Middle Member.
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show a consistent transport towards the northwest and only occasionally to they differing from this.

At a height of 125 m there is a sudden change in facies from coarse sandstones to matrix and clast supported conglomerates with interbedded sandstones which may show cross-stratification (Figure 3.22). The beds attain thicknesses up to 4 m and clast sizes up to 1 m. The base of beds can be erosive, infilling channel-like features (Figure 3.22C), lenticular or laterally continuous. Inverse grading in matrix-supported conglomerates, normal grading and poorly developed cross-bedding is also present. The conglomerate beds may show a gradation from clast- to matrix-supported to coarse-grained pebbly sandstone defining a finning-up sequence. Thin <50 cm mud horizons are also evident between conglomerate beds.

At a height of approximately 400 m from the base of section the conglomerates overlain by thin-bedded fine- to medium-grained sandstone beds and laminae, the beds range from 0.1 cm to 40 cm thick. The unit is divided into two on the basis of a muddy interval; the lower half comprises thin-bedded sandstones containing incipient parallel lamination (Figure 3.22b). Rare mud clasts and wave-ripples which show bundled up-building are also present. The top half of the section comprises thin-bedded, medium- to fine-grained sandstones, however, we do find rare sandstone beds up to 15 cm thick which contain poorly developed opposing current-ripple directions. Other features include <0.2 cm thick medium- to fine-grained laterally continuous horizons, and <2 cm thick sandstone lenses which are lateral discontinuous over distances of 2 m to 4 m.

3.6.2.4 Interpretation of the Lower, Middle and Upper and Members

To summarise, the Lower Member from the contact with the marls upwards comprises normal-graded sandstone beds up to 0.5 m thick, and are interpreted to have been deposited by low-concentration turbidity currents, which pass up into 2-3 m thick structureless normal-
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graded turbidites. Above this is a 50 m thick succession of parallel-and cross-stratified medium/coarse-grained sandstones, with rare topset preservation. These are interpreted to have been produced by fast moving bottom currents transporting sediment as a tractional bedload. At 125 m up-section a marked change from sandstones to conglomerates occurs, the conglomerates are either clast or matrix supported, with the matrix comprising coarse to granule grade material and reworked clasts of intrabasinal material. Incipient cross- and-parallel-stratification of the clasts is evident with minor interbedded cross-stratified sandstones. The conglomerates are interpreted to be deposited en-masse from high-concentration density/debris flows, rare erosive bases to conglomerate units indicates that some flows were capable of erosion. At a height of 400 m upsection the conglomerates pass up into a 70 m thick thin-bedded sandstone-mudstone unit, the sandstones generally appear lenticular with rare parallel laminations. Opposing ripple cross-lamination is evident within some beds, and rare wave-ripple cross-lamination is recorded. These thin-bedded sandstones are interpreted as storm generated deposits indicating either a fall in relative sea-level or an increase in the height of base level arising from sediment aggradation in the deep-marine basin.

The Lower Member succession is interpreted to represent an overall coarsening and shallowing-up sequence, the Middle and Upper Members are also interpreted in a similar manner. The whole Grès d'Annot Formation succession at St Antonin is interpreted as comprising three distinct sequences representing progradation/aggradation of facies into the deep-marine basin controlled by variations in relative sea-level and basin subsidence, accommodating sediment and allowing sequences to be repeated. Basin subsidence may have resulted from sediment loading causing an increase in the rate of subsidence; this subsidence must have lagged behind the fall in relative sea level, thus allowing the formation of storm deposits. If sediment load-induced subsidence was instantaneous then base level in the deep-marine part of the basin would not reach storm-wave base. Because of the relatively poorly exposed outcrops and the preservation in a synclinal structure, it is not possible to deduce the
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sub-basin geometry and the exact depositional environment for the St Antonin sandstone succession.

The St Antonin conglomerate and sandstone succession has previously been interpreted as being deposited in a proximal subaqueous slope environment in front of an alluvial fan-backed submerged fan-delta complex (Stanley 1980). Stanley (1980) interpreted the conglomerate units as associated with seaward progradation of an alluvial fan, with conglomerate and sandstone deposition from traction and gravity mass flows suggesting a prodeltaic slope environment. The interpretation of the depositional environment for the St Antonin conglomerates and sandstones given by Stanley (1980) is favoured here. For a detailed interpretation of the St Antonin sub-basin see Section 6.1.2.1a(i).

3.7 Summary of the Grès d'Annot Formation in the Annot area

The Grès d'Annot Formation successions in the St Antonin, Entrevaux and Annot sub-basins are interpreted to be genetically linked, this conclusion is based on the similarity in the orientation of palaeocurrents from the three sub-basins and their close proximity to each other. A palaeoreconstruction for the St Antonin, Entrevaux and Annot sub-basins during sandstone deposition is presented on Figure 3.23. A detailed account of the relationships of these three sub-basins described above, with other sub-basins in the study area, is given in Section 6.1.2.
Figure 3.23. Palaeo-reconstruction for the Annot-Entrvau-St Antonin areas during deposition of the Grès d'Annot Formation.
Chapter 4

Sandstone architecture and facies from the Grand Coyer and Chalufy areas

4.1 Introduction

4.1.1 Location

The Grand Coyer sub-basin is located in the Haute Provence region of southeast France (Figure 4.1). The Grand Coyer sub-basin comprises a succession of sediments which record a progressive deepening of the Tertiary foreland basin, and which at present cover an area of 60 km². These sediments are preserved as basin remnants in a northwest-southeast trending synclinal structure. This synclinal structure is interpreted to have been present during deposition of the Tertiary sediments as indicated by thickness variations in the Marnes Bleues Formation (Inglis et al. 1981, Le Varlet & Roy 1983) with thinner marls occurring at the margins of the sub-basin. This is also confirmed from observations of sandstone packet terminations against slopes developed in the underlying marls (this thesis). The Chalufy exposures are 5 km WNW of the Grand Coyer sections (Figure 4.1). The Grand Coyer sub-basin is located approximately 4 km southeast of Allos-Colmars and 14 km north of Annot. The Annot and Grand Coyer sub-basin Tertiary sediments are only 4 km apart at their nearest point, with the Grand Coyer sub-basin due north of the Annot basin (Figure 4.1).

The Tertiary sediments of the Grand Coyer sub-basin are relatively undeformed and well exposed, and afford an excellent opportunity to study the Grès d'Annot Formation. The exposures form steep, up to 300 m high, cliff sections which enable detailed observations on
Figure 4.1. Geological map of the Alpes Maritimes and Haute Provence regions of Southeast France, the Grand Coyer sub-basin and Chaluy section are marked (re-drawn from BRGM 1980 Sheet 40 and 45 1:250 000).
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sandstone packet architecture using detailed photo-montages, line drawings, and sedimentary logs. Sandstone beds and packets onlap slope marls of the underlying Marnes Bleues Formation on the western side of the sub-basin, whereas on the eastern side such a relationship is not preserved and instead a faulted margin to the sub-basin is present. The true lateral extent of the sandstone packets were therefore not possible to determine.

4.1.2 Previous research

For the most part, the Grand Coyer sub-basin is poorly researched, and the only published major study was carried out by Inglis et al. (1981), in which they describe the sediments from the Colmars-les-Alpes and Col de la Cayolle regions, encompassing the Trois Evêchés, Grand Coyer and Col de la Cayolle sub-basins. In the Grand Coyer area, Inglis et al. (1981) noted thickness variations in the Calcaires Nummulitiques, and Marnes Bleues Formations, palaeocurrent trends during deposition of the Grès d’Anot Formation, and sandstone/shale ratios within the sub-basin. Palaeocurrents were recorded towards NNW, and sandstone/shale ratios were in the order of 5:1 in the centre of the basin decreasing to 2:1 towards the margins (ibid.).

Ravenne et al. (1987) produced a schematic NNW-SSE oriented cross-section of the sub-basin, marking on correlateable sandstone units, to produce an interpretative map for the depositional surface of the Grès d’Anot Formation. Sandstone packets onlap towards the west (ibid.). Sinclair (1994) traced sandstone packets from the axis of the sub-basin towards the marginal marl slope over a distance of 1.5 km, noting the thinning of individual sandstone packets and changes in facies as they approached the marl slope and onlapped.
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4.1.3 Enclosure location

For this study field data was collected from the Grand Coyer sub-basin (Enclosures 4.1 to 4.10), which generally include photo-montages of cliff-side exposures with accompanying line-drawing interpretations. Sedimentary logs obtained from the corresponding sections are also presented with their location shown. Enclosure positions from the Grand Coyer sub-basin are marked on Figure 4.2 (see also Sections 4.2 to 4.6).

4.2 Sub-basin stratigraphy

4.2.1 Tertiary foreland basin fill

The detailed stratigraphy of the Tertiary foreland basin of southeast France is described in detail in Section 1.2. The Tertiary sediments in the Grand Coyer sub-basin rest unconformably upon Cretaceous rocks of Aptian to Maastrichtian age (Campredon 1977). The sediments include the Poudingues d'Argens Formation, comprising fluvial conglomerates and sandstones of reworked Cretaceous limestones, which is the earliest of the sub-basin fill and is only present as a localised thin (< 5 m thick) unit found at the base of the Calcaires Nummulitiques Formation. The Calcaires Nummulitiques Formation attains a maximum thickness of 20 m (Inglis et al. 1981), and the Marnes Bleues Formation reaches a maximum thickness of approximately 140 m, located in the axis of the basin and 60 m at the margins (Inglis et al. 1981). The Grès d'Annot Formation has a maximum preserved thickness of approximately 400 m, although its stratigraphic top is not preserved.

A regional stratigraphic study of the Tertiary foreland basin sediments by Inglis et al. (1981) revealed that the Calcaires Nummulitiques Formation shows thickness variations in the order of 0-20 m, interpreted as being related to local structural highs within the area controlling
Figure 4.2. Geological map of the Grand Coyer sub-basin showing Enclosure location and major peaks (redrawn from BRGM Sheet 945 and 35-40, 1: 50 000).
sediment distribution. The Marnes Bleues Formation shows a more consistent variation within the Grand Coyer area, with the marls being thickest in the axis of the basin up to 140 m, and thinner at the margins with thicknesses up to 60 m. This thickness variation in the marls is interpreted to be the result of sediments accumulating in the axis of a pre-existing depression on the seafloor, with a thinner succession at the margins. The decrease in thickness from the axis to the margins is present on both sides of the sub-basin, indicating that the sub-basin was of restricted width. Using the data obtained by Inglis et al. (1981) the across-flow basin width during deposition of the Grès d'Annot Formation was approximately 6-7 km.

4.2.2 Grès d'Annot Formation

The Grès d'Annot Formation in the Grand Coyer sub-basin attains a maximum preserved thickness of 376 m (Figure 4.3). This stratigraphic log is constructed using detailed measured sections, field observations, and measurements taken from photo-montages for inaccessible sections. The log shows the base of the succession in contact with the underlying Marnes Bleues Formation as seen at the Têté de Mouriès exposures (Enclosure 4.1); the top of the succession is shown as sandstone packet H (Enclosure 3.8), which is the youngest preserved part of the succession in the sub-basin. The Schiste à Bloc unit which is preserved in other parts of the region is absent here.

The succession comprises nine distinct medium- to pebble-grade, essentially amalgamated, sandstone packets between 3 and 55 m thick which are invariably separated by packets of sandstone-mudstone couplets, with a maximum of 16.6 m for the thickest sandstone mudstone packet. The sandstone beds within these mudstones are predominantly medium- to fine grained. Beds within the amalgamated sandstone packets may be separated by rare thin (<30 cm) silt and mudstone layers.
Figure 4.3. Stratigraphic log of the Grès d'Annot Formation from the Grand Coyer sub-basin, sandstone bodies are marked (AA→H), and are shown on Enclosures 4.1-4.8. For description and location see Section 4.2.2.
The stratigraphy of the Grès d'Annot Formation in the Grand Coyer sub-basin in this study is divided into two parts based on onlapping and downlapping relationships seen in the sandstones packets. The lower half is named the Basin-Fill Member labelled AA, and the upper half as the Shelfal Member, which prograded northwards into the sub-basin and is interpreted to be genetically related to the sandstones to the south at Annot, Entrevaux, and St Antonin. The relationship of these two members with the sediment fill of the Grand Coyer basin is described in Section 4.6. These two members are separated by a 14 m thick unit occurring at a height of 37 m from the base of the stratigraphic log on Figure 4.3. This unit is divided into three parts; first the lower part which is a disorganised horizon 3.4 m thick, comprising reworked marl clasts up to 1 m in diameter and contorted sandstone beds set in a mudstone matrix. Above, is a 7.6 m thick unit comprising mud with sandstone/mudstone couplets of Facies C2.3, where sandstones up to 5 cm thick occur (Enclosure 4.1.1 F). These sandstones are very fine grained and structureless although some show lower flow regime parallel laminations with rare current ripples. The mud separating these sandstones is marly. The upper part is 3 m thick and comprises bedded marls with no sandstones. The unit as a whole is interpreted to represent a period of relative non-deposition of clastics with the thin sandstone-mudstone couplets being distal equivalents of proximal sandstones or deposits from low-concentration turbidity currents. The bedded marls represent a period of reduced clastic deposition in which pelagic calcite mud accumulated, and/or resedimented marls from farther upslope were deposited.

4.3 Sandstone architecture and facies

In the Grand Coyer sub-basin nine sandstone packets are recognised based on field observations of lithofacies and correlations of sandstone packets. The sandstone packets are identified as AA to H (Figure 4.3). Sandstone packets are separated by eight mudstone/sandstone intervals of various thickness, these enable correlation of the sandstone packets across the sub-basin.
Chapter 4: Sandstone architecture and facies from the Grand Coyer and Chalufy areas

This section is divided into two parts: (i) external and (ii) internal architecture of the sandstone and intervening mudstone/sandstone packets. Not all of the sandstone packets have been graphically logged due to their inaccessibility, therefore internal lithofacies were not documented, but their gross external and internal features are described.

4.3.1 Sandstone and mudstone packet architecture

The external architecture of each sandstone packet is quantified in terms of thickness, and lateral and proximal extent within the sub-basin. Observations of internal packet architecture, and that of the intervening sandstone and mudstone packet architecture is also described.

4.3.1.1 Sandstone packet AA

Sandstone packet AA is labelled to distinguish it from the overlying sandstone packets which have a sub-basinwide extent. Sandstone packet AA has a limited exposure, and is restricted to the northern side of the sub-basin (Enclosures 4.1/4.1.1/4.1.2, see also Figure 4.2). The packet has a maximum observed thickness of 37 m; sandstones within the packet onlap on to a palaeoslope in the marls which dipped 16°NNW during sandstone deposition (Enclosure 4.1 A, 4.1.1 A and B).

Sandstone beds within packet AA reach a maximum thickness of 1.1 m and belong to Facies C2.1 with Bouma Tₐ and Tₜ divisions present. The beds are amalgamated with silty mudstones separating some beds. At the top of some beds, there are medium-grained cross-stratified sandstones up to 15 cm thick (Facies B2.2). Maximum grain size observed in the beds is granule grade, with a mean grain size of medium/ coarse; the thinner (< 0.3 m) beds being fine-grained. Bioturbation is present in some of the beds in the form of branching sand
filled burrows. Sandstones at the very top of the packet show an increase in the amount of bioturbation, and are interpreted to indicate a period of little or no deposition, allowing organisms to thoroughly bioturbate the sandstones before the onset of subsequent turbidite deposition.

The sandstone packet can be traced in a proximal-distal direction towards the north for a distance of 875 m, and in a lateral direction for 1870 m. When the unit is traced northwards from the onlap, synsedimentary deformation is evident. The sandstone packet is deformed, discontinuous, and is faulted to form curved edges within a cut down which is 10 m deep (Enclosure 4.1.1 A). This was subsequently infilled with a contorted mass of sediment, including reworked sandstones, muds and clasts of the Marnes Bleues Formation (Enclosures 4.1.1 E and 4.1.2). At the southern side of the section there is a synsedimentary normal-fault with a displacement of 20 m, downthrown towards the north (Enclosure 4.1.2). The fault postdates the previously described deformation and predates the overlying sandstone packet A. The faulting created a depression on the seafloor which was subsequently excavated by turbidity currents to produce a channel-like scour (Enclosure 4.1.2 B and C) which was subsequently infilled with coarse/pebble-grade sediments of sandstone packet A.

The fine-grained sandstone-mudstone packet which separates sandstone packets AA and A comprises three parts (Enclosure 4.1.1, Log B). The basal part comprises reworked sandstones, muds and marls, which infill the depressions described above. Where this reworked unit of sediments rests on undeformed portions of sandstone packet AA it is 3.4 m thick with marl clasts up to 1 m in diameter. This reworked unit has a gently undulating topography on its upper surface comprising small depressions which have been infilled by the middle member of this fine grained unit (Enclosure 4.1.1, Photo E). The middle unit is 7.6 m thick and comprises marly mudstones with thin fine-grained sandstones up to 10 cm thick (Enclosure 4.1.1 F). Ripple cross-lamination is present in the thicker sandstones, and give a sediment transport direction towards the north; otherwise, these sandstones are either structureless or parallel laminated. Overlying the middle unit, there is a 3 m thick unit of
bedded marls (Enclosure 4.1.1 D) interpreted to record a period of non-clastic deposition which allowed the accumulation of marls as a hemipelagic sediment. The marls are similar in composition to those of the underlying Marnes Bleues Formation and may have been derived from the surrounding marl slopes of the sub-basin as redeposited material, or may represent primary deposition: marl deposition is believed to have been active during deposition of the Grès d'Annot Formation (Bodelle 1971). There are horizons of clastic beds in the top of the Marnes Bleues Formation from the Barrême sub-basin located towards the west of the study area (ibid.) (Figure 4.1). These clastic units include the Grès de Ville, La Poste, St Lion and Grès de Senez Members, with the latter three members being lateral equivalents to each other. The above four units are also thought to be time equivalents to the Grès d'Annot Formation (Stanley 1961).

4.3.1.2 Sandstone packet A

Sandstone packet A forms an extensive unit which is traceable for approximately 7 km across the basin in a proximal-distal direction, it has a lateral extent of over 1625 m in an east-west direction (Enclosures 4.1, 4.3, 4.4, 4.6, 4.7 and 4.8). On the western side of the sub-basin, the packet is inferred to onlap onto the underlying marls. Sandstone packet A attains a maximum thickness of 65 m, and comprises turbidite sandstones with a maximum thickness of 4.6 m (Log A, Enclosure 4.6) and a mean of 1 m (Log A, Enclosure 4.8). The maximum grain size observed is cobble-grade material. A pebble bed situated 5 m from the base of log C (Enclosure 4.6) is 4.6 m thick and contains exotic, extrabasinal clasts of basalt, and extrabasinal sediments such as chert, together with a large (2 m x 1.5 m) clast of reworked Nummulitic limestone. This bed forms a clast-supported conglomerate with cross-stratification preserved, with foresets up to 2.98 m thick (Facies A2.1). The other beds in packet A show normal grading from a granule-/pebble-grade base to a medium-/coarse-grained top, with rare parallel stratification (Facies A2.7 and B2.1). Discrete pebble horizons occur in some beds. Beds are generally amalgamated, and mud clasts are commonly
concentrated at discrete horizons. Also, there are scours up to 1 m deep and 10 m long (Enclosure 4.8 D, Log B). For the most part, the beds are laterally continuous with rare pinch outs.

The base of sandstone packet A shows both erosional and non-erosional contacts with the underlying fine grained unit (Enclosures 4.1, 4.1.1 and 4.1.2). A scour infill developed as a result of synsedimentary faulting of the underlying units to produce a local depression on the seafloor which was further eroded and then infilled by sediment from turbidity currents (Enclosure 4.1.2 A, B, and C). The infill comprises pebbles up to 0.1 m in size set in a matrix of granule-grade material, forming discrete bands up to 1 m thick. Other clasts present include reworked intraformational sandstone and mudstone from the Grès d'Annot Formation, marls and underlying Nummulitic limestones. A mudstone clast 2.7 m by 0.8 m occurs which shows that some turbidity currents or sandy debris flows had sufficiently high velocities or clast support mechanisms capable of transporting such large clasts. The top of sandstone packet A is planar with an 8 m thick packet of mudstones and thin (< 0.4 m) overlying sandstones.

4.3.1.3 Sandstone packet B

Sandstone packet B forms a 31 m thick unit of sandstones with limited exposure over the subbasin (Enclosures 4.1, 4.3, 4.6 and 4.8). Although the packet is inaccessible to log, field observations reveal common bed amalgamation. The packet has a proximal-distal extent of approximately 7 km, and a lateral extent of at least 1 km. Sandstone packet correlation shows that the packet onlaps towards the west onto a northeastward dipping slope in the marls (see Section 4.3.3). Above sandstone packet B there is a 4.2 m thick interval of mudstone and sandstones with sandstone beds to < 0.3 m thick.
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4.3.1.4 Sandstone packet C

Sandstone packet C is up to 55.6 m thick, and has a proximal-distal SE-NW extent >7 km, and a lateral extent >2.8 km in an ENE-WSW direction (Enclosures 4.1, 4.4, 4.5, 4.6, 4.7, 4.8). The section on Enclosure 4.7 is orientated at 20° with respect to palaeocurrent trend, while Enclosure 4.8 is at 90° to palaeocurrent trend. The precise lateral extent of the packet towards the east of the sub-basin is not known, but towards the west it onlaps the northeasterly dipping slopes developed in the marls. The base of the unit shows no major erosion surfaces.

Sandstone packet C is easily accessible, and sedimentary logs are shown on Log A (Enclosures 4.7) and Log C (Enclosure 4.8). The packet comprises sandstone beds 0.3-5.9 m thick with a mean of 1.5 m. Beds are generally normally graded, ranging from granule/pebble grade at the base to medium-/fine-grained at their tops. The beds are amalgamated, except for those towards the top of the packet which have <0.1 m thick silt/mud partings. The turbidites are ascribed to Facies A2.7, and C2.1 with Bouma T_a and T_b present only. Mud clasts occur as either scattered clasts or as discrete horizons within turbidites, and generally within the graded structureless T_a divisions. Sandstone packet C shows a series of graded beds which appear to show some cyclicality in the form of finning-/thinning-up sequences, (Enclosure 4.8, Log C). The succession shows distinct 10-15 m thick sequences within which sandstone beds become thinner and finer-grained towards the top, and this pattern is repeated at least four times within the packet (Figure 4.8 b). Sandstone packet C shows an overall thinning-and-finining-upward sequence.

The internal architecture of sandstone packet C is more complex than that of packets AA, A, or B. In packet C the lower 30 m of turbidites are essentially sheet-like with rare shallow scars at the base of beds; in contrast, the upper 20 m show marked vertical and lateral thickness variations and sedimentary stacking features. The two sections which show this best are the Grand Coyer and Rocher du Carton sections (Enclosures 4.8 and 4.7).
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Coyer exposures provide a section through the sandstone packets at −90° to palaeocurrent trend; the Rocher du Carton exposures form a section aligned at approximately 20° to palaeocurrent trend. The proximal Rocher du Carton section (Enclosure 4.7 A, B and C) shows the complex nature of the top 20-30 m of sandstone packet C. On close inspection (Enclosure 4.7 B), the upper part of sandstone packet C comprises two distinct facies types: (1) <10% of the upper part, comprises cross-stratified sandstones up to 4 m thick with a maximum horizontal extent of 80 m (Enclosure 3.7 B), and the tops of the bedforms show a sharp contact with the overlying beds; (2) sandstones with a cumulative thickness of 8 m and lengths of up to 87 m, with lens-like shape and a tendency to be concave-up and draping the cross-stratified bedforms. The sandstone beds which constitute these packets are up to 1 m in thickness and are lens-like and concave-up. The grain size of these beds is medium/coarse sands, with some beds showing grading to fine-grained sand towards their tops (Facies C2.1), but others show no grading and are medium-grained throughout, with parallel-lamination. These sandstones are laterally offset and show a consistent direction of stacking within any sandstone unit. Between packets, however, the thinning may be in an opposite direction to that observed in adjacent units (Enclosure 4.7 B). On the Grand Coyer section (Enclosure 4.8) the sandstone packets in the upper 20 m of section do not show the same extent of lensing and internal complexity and may be a result of the section trending at 90° to palaeocurrent. On the western side of the section, however, a lensing sandstone packet is visible which comprises thin (<0.5 m thick) sandstone beds. The far western side of the section shows a small channel-like depression located towards the top of the packet which cuts down into the sediments below and is infilled with sandstone beds, this scour-like feature is 3 m deep and 12 m wide.

The overlying mudstone/sandstone packet comprises mudstones and thin sandstones (<0.5 m thick), with some beds up to 0.8 m in thickness, which have a lenticular geometry, with a maximum lateral extent of 130 m. At the very top of the packet there is a 0.3 m thick sandstone bed which when traced laterally towards the west is folded, with a fold vergence towards the east, indicating that the bed was deposited on a slope which it subsequently slid down towards the east. Figure 4.9 shows the position of this fold in relation to the channel.
margin and is also described in Section 4.3.2.3.2. On the Rocher du Carton section (Enclosure 4.7) the mudstone/sandstone packet also shows lensing of thin sandstone beds. Sedimentary structures from sandstones within the mudstone/sandstone packet, include parallel lamination in predominantly medium-/fine-grained sandstones.

4.3.1.4.a Interpretation of lensing packets

The cross-bedded units (Enclosure 4.7 B) are interpreted to represent tractional bedforms formed either by reworking of previously deposited sediment from strong bottom currents or sea-floor bedforms solely deposited from high velocity turbidity currents which did not deposit all their sediments, and instead, bypassed this part of the sub-basin. Sandstone beds within these packets show a marked angular unconformity with the beds below, and the tops of these beds also show a marked angular disparity with the overlying beds, suggesting erosion on the lee side of a large bedform. Large-scale asymmetric bedforms from the deep-sea have been documented (e.g., Hughes Clarke et al. 1990). These asymmetric gravel bedforms from the modern record are believed to have been formed by the 1929 Grand Banks turbidity current on the Eastern Valley of the Laurentian Fan (ibid.). From the ancient deep-marine record, sedimentary bedforms have also been recorded by Vicente Bravo & Robles (1995), who noted wave-like symmetric to slightly asymmetric sandy-gravel and gravel bedforms.

The lensing sandstone packets which overlie the above described bedforms are interpreted to represent the draping of scours or channel-like features formed on the sea floor. These depressions were subsequently draped with sandstones and this draping produced the lensing. Because of the obliquity of the section to palaeoflow it is not possible to deduce if these structures are channels or simply scour infills, but as no associated levees are seen scours are favoured. Another possibility in favour of scour formation on top of sandstone packet C is that as the lobe-like body builds up on the sea-floor, subsequent turbidity currents instead of
4.3.1.4.b Interpretation of lensing thin sands in mudstone packet

The lensing of the thin sandstone beds in the mudstone packet may represent the formation of insipient depositional channels within the mudstones or stacked deposits of medium to low concentration turbidity currents. The lack of coarse material in these beds may reflect a change of sediment supply in the source area, possibly controlled by a relative sea level rise and thus starving the basin of coarse-grained sediment. The formation and significance of sandstone and mudstone packets in the sub-basin is discussed in Section 4.5.

4.3.1.5 Sandstone packet D

Sandstone packet D is shown on Enclosures 4.7 and 4.8, it attains a maximum thickness of 13.6 m and on Enclosure 4.8 is seen to thin to 0 m towards the west, the base of the packet is irregular with the top relatively planar. It has an observable proximal-distal extent of at least 3 km and a lateral extent of >530 m. The packet comprises turbidite sandstones up to 2 m thick, they are normal-graded from pebbles at their base to medium-/coarse-grained sands at their tops (Facies C2.1, Bouma T_a only), some beds show amalgamation. Beds at the base of the packet are thinner than those at the top with the succession interpreted to represent a coarsening and thickening-up sequence. The beds within the packet show gradual thinning as they approach the downlap/onlap surface towards the west, and some beds show thinning within the packet.

The base of the sandstone packet shows both a planar onlapping surface on the western side and an uneven base on the eastern side, this unevenness is interpreted to be an erosion surface
developed in the underlying mudstone unit, and is clearly visible on Enclosure 4.8 and in detail on Figure 4.5. Sandstone beds at the base of the packet onlap the erosion surface towards the east and are interpreted to downlap/onlap on the western side of the section on to the mudstone packet below, a detailed account is given in Section 4.3.2.3.2 and 4.3.2.1.1. The sandstone packet thins towards the west at a rate of 1 m for every 40 m of horizontal distance. The packet is shown on Enclosure 4.7 A, where it is cut down into by the overlying sandstone packet E.

Above sandstone packet D, the overlying mudstone packet is 2.5 m thick and comprises sandstones up to 0.05 m thick (Facies C2.3), this mudstone packet drapes the sandstone packet, and is present on both Enclosures 4.7 and 4.8. On Enclosure 4.7 this mudstone packet is also cut down into by sandstone packet E.

### 4.3.1.6 Sandstone packet E

Sandstone packet E is shown on Enclosures 4.7 and 4.8, affording good sections through the sandstone packet. The packet attains a maximum thickness of 28.9 m, thinning to a minimum observed thickness of 13 m in a lateral northeast-southwest section. The sandstone packet can be traced for a proximal-distal southwest-northeast distance of 3 km and a lateral distance of 530 m. The top of the packet appears planar, with the base showing an erosional contact with the sandstone and mudstone packets below, this erosion occurs on the western side of Enclosure 4.7 A, and is absent on Enclosure 4.8. On Enclosure 4.8 the base is non-erosive, and sandstone beds at the base of the packet onlap a sea-floor topography developed in the underlying mudstone packet, the topography being inherited from the underlying sandstone packet D. The erosion feature on Enclosure 4.7 B is approximately 100 m in length with a maximum cut down of 9 m, the section is aligned at approximately 20° to palaeocurrent, therefore an accurate calculation of the dimensions is not possible due to the oblique orientation of the section to palaeoflow. It is evident, however, that the erosion surface is
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asymmetric with the proximal side of the erosion being steeper and cutting farther down into the sediments below, with the distal side less steep. This erosion feature is here interpreted to be a mega-flute-like scour formed from the passage of high velocity and high-density turbidity current across the sea floor.

On Enclosure 4.8, sandstone packet E shows a consistent thinning towards the northeast, with sandstone beds at the base of the packet onlapping towards the northeast. It is apparent that deposition of sandstone packet D developed a topography on the sea floor, which influenced the geometry of packet E, causing sandstone beds to onlap the topography created.

The sandstone packet comprises turbidite beds from 0.3 to 4.3 m thick, the beds are invariably normal graded from pebble/granule grade at their base to coarse/medium grade at their tops, mud clasts in discrete horizons also occur in some beds. The beds are ascribed to Facies C2.1, with Bouma T₃ and T₄ only; the majority of beds show amalgamation. Sandstone beds within the packet are essentially sheet-like in aspect and only show terminations where they onlap the topography below. Local scouring is evident at the base of some beds, and towards the top of the packet. On the southwest side of Enclosure 4.8 A, sandstone beds thin towards the northeast. In a vertical section at its thickest point the sandstone packet shows up-section changes in bed thicknesses. The base of the packet comprises thin (<0.3 m) beds which thicken up to 1 m towards the top forming a unit 10.5 m thick, this is overlain by another thickening-up unit 12.3 m thick. The sandstone packet is then overlain by a mudstone/sandstone packet 2.4 m thick. At the top right hand side of the packet on Enclosure 4.8 A channel-like scour is evident, which cuts down to 3 m and is 30 m long in transverse section, and is infilled with sandstone beds up to 0.5 m thick which appear to onlap the sides of the scour.
4.3.1.7 Sandstone packet F

Sandstone packet F is only observed in an ENE-WSW and NE-SW section on Enclosures 4.7 and 4.8 respectively. The packet is inaccessible but observations on geometry can be made from field photo-montage correlations. The packet attains a maximum thickness of 8.2 m on Enclosure 4.7, in a proximal section, and a thickness of 4.8 m on Enclosure 4.8 in a distal position. The two sections are separated by a distance of approximately 1 km. The packet shows no variation in thickness over its lateral extent of 385 m, the top and bottom of the packet are planar and non-erosive. Sandstone beds within the packet appear sheet-like over their extent, with some showing amalgamation but no scouring. Above this packet is a 17.8 m thick mudstone/sandstone couplet packet with beds up to at 0.4 m thick.

The thinning of the sandstone packet in a proximal-distal direction is interpreted to represent downlapping of beds at the base of the packet in a downcurrent distal direction. The sandstone packet is interpreted to pinch-out down current as it is absent on Enclosure 4.1 located 3 km towards the north.

4.3.1.8 Sandstone packet G

Sandstone Packet G has limited exposure within the sub-basin, and is observed on Enclosure 4.8 with a small amount of exposure on Enclosure 4.7. The packet is inaccessible and measurements were taken from photo-montages. It attains a maximum preserved thickness of 50.3 m, and has a maximum observed lateral extent of 442 m, and a proximal-distal extent of >1 km. The top and bottom of the packet are planar, with a constant thickness over its outcrop. Sandstone beds within the packet are essentially sheet-like over their outcrop, with some beds showing amalgamation, other beds are separated by muds and silts. Throughout the sandstone packet, beds appear to show packaging into thinner non-amalgamated and
thicker amalgamated beds. These packets attain thicknesses up to 6-7 m. Above the packet is an 8.3 m thick mudstone/sandstone packet visible on Enclosure 4.8 but not on Enclosure 4.7.

4.3.1.9 Sandstone packet H

This packet has a limited exposure, and is present only on Enclosure 4.8, the exposure is poor, but a thickness of > 33 m has been calculated from photo-observations. The packet comprises beds up to 3 m thick.

4.3.2 Sandstone packet architecture

In the Grand Coyer sub-basin, nine distinct sandstone packets have been identified which are separated by eight mudstone/sandstone packets and the entire vertical succession is shown on Figure 4.3. They attain a total thickness of 374 m, including the interbedded mudstone/sandstone couplet packets. The stratigraphic top of the Grès d'Annot Formation in the sub-basin is not preserved and is represented by the present day erosion surface. The entire sub-basin is not preserved, but there are remnants from which an interpretation of sandstone architecture, stratigraphy, and sub-basin morphology have been made. Exposures of the formation which provide the best outcrops are located on the western side of the basin, while those on the eastern side (see Figure 4.2) provide poor exposures of the outcrops, caused by deformation related to northwest-southeast trending faults which appear to have shattered the sandstones. The lack of continuous good exposures of sandstone packets on the eastern side of the sub-basin prevents a complete understanding of the sub-basin fill but, nevertheless, enables valuable information on sandstone packet architecture and facies.

In this study the term "sandstone packet" is used instead of "lobe". As defined by Mutti & Ricci Lucchi (1972) a lobe is a feature lobate in plan view and mounded in cross-section. For
the most part the sandstone packets in the Grand Coyer sub-basin do not show a mounded cross section over their extent of outcrop, but are laterally continuous. Normark et al. (1993) state that at the scale of outcrop a lobe may be: sheet-like with a basin-wide extent, slightly mounded bodies, that can occur at the terminus of basin margin channels, or confined bodies that fill both structural and erosional depressions. They also suggest that to avoid confusion the term lobe should be prefixed with an appropriate adjective for its geometry, in the case of the sandstone packets from the Grand Coyer sub-basin the prefix sheet-like and confined would be valid, therefore the sandstone packets would be classified as "sheet-like confined lobes".

4.3.2.1 External sandstone packet geometries

The external sandstone packet geometries are defined in terms of their lateral and proximal-distal extent within the sub-basin, exposures within the sub-basin provide both flow perpendicular and flow proximal-distal sections. The external sandstone and mudstone packet architecture is summarised in a flow lateral and flow proximal-distal orientation on Figures 4.6 and 4.7, respectively.

4.3.2.1.1 Lateral continuity

For the most part, the sandstone packets are laterally continuous over the sub-basin, and are only seen to terminate where they onlap against the sub-basin marginal slopes developed in the marls (see Figure 4.6). The only exception to this is seen in sandstone packets D and E on Enclosure 4.8 and summarised on Figure 4.4 a and b. Sandstone packet D thins from 10.2 m to 0 m over a lateral distance of 261 m towards southwest, which gives a rate of thinning of 1 m for every 25 m. Sandstone packet E thins from 24.6 m to 13.8 m over a lateral distance of 374 m towards northeast, which gives a rate of thinning of 1 m for every 31 m.
Figure 4.4a. Schematic summary diagram of Enclosure 4.8 showing sandstone packet architecture and important features.
Figure 4.4b. Schematic summary diagram of Enclosure 4.7 showing sandstone packet architecture and important features.
Figure 4.5. Three-dimensional expanded reconstruction of sandstone packets C, D and E showing architecture and spatial relationship of observed features. The model is based on outcrops described from Enclosures 4.7 and 4.8.
Figure 4.6. Northeast-Southwest stratigraphic transect through the Grand Coyer sub-basin showing sandstone packet distribution in the sub-basin, with onlaps on the marl paleoslope.
Figure 4.7. Southeast-Northwest stratigraphic transect through the Grand Coyer sub-basin showing sandstone packet distribution and architecture, section is aligned parallel to palaeoflow.
Sandstone packet D onlaps and infills a topography developed in the mudstone below (see Figure 4.4 a and Figure 4.5), when packet D is traced towards the southwest, sandstone beds at the base show a consistent onlap/downlap until the packet pinches out. Sandstone beds at the base of sandstone packet E onlap towards the northeast on to the mudstone unit which drapes packet D. This indicates that sandstone packet D formed a topographic high on the sea floor, as well as infilling a previously developed topographic low located beneath. The formation of this topography may be a result of differential compaction of the sandstones in packet D down in to the underlying muds, or may represent mounding of the sandstone packet during deposition forming a lobe (sensu Mutti & Ricci Lucchi 1972). It would appear that the existence of a topographic low in the mudstone packet prior to deposition of sandstone packet D influenced the geometry of this packet, which in turn created a topography which influenced the geometry of sandstone packet E.

4.3.2.1.2 Proximal-distal continuity

For the most part over the extent of the sub-basin, the sandstone packets appear continuous and show no obvious change in thickness, except for where sandstone packet AA onlaps an intra-basinal palaeoslope (see Figure 4.7).

Only sandstone packet F shows a variation in geometry in a proximal-distal direction which is shown on Figure 4.4b and in detail and on Figure 4.7. The packet thins towards the northwest and is interpreted to downlap and pinch-out, as it is not visible on Enclosure 4.1. At the most proximal outcrops in the south of the sub-basin (Enclosure 4.7), the base of sandstone packet A shows sandstone beds downlapping towards the north on to mudstones which overly the marls. This downlapping of the sandstone beds is interpreted to indicate the proximal-distal progradation of the sheet-like confined lobe in the sub-basin.
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4.3.3 Mudstone packet external architecture

The external geometry of the mudstone packets is relatively consistent throughout the axis of the sub-basin. In general the packets show a constant thickness across the sub-basin. The only account of a packet thickening dramatically is when sandstone packet F downlaps and pinches-out and two mudstone packets amalgamate to form a thicker packet.

The mudstone packets, however, do thicken towards the southwest sub-basin margin as sandstone packets onlap the slope and sandstone beds pinch out. It would appear that the muds preferentially accumulated at the margins of the sub-basin on the slopes. This is described in detail in Section 4.3.4.

The mudstone packet above sandstone packet C shows an irregular top surface (Enclosure 4.8 and Figure 4.4 a.) which is infilled with sandstones of packet D. This irregularity could either be the result of erosion of the mudstone packet creating a scour or the infilling of a channel previously developed in the mudstone packet.

4.3.4 Sandstone and mudstone packet internal architecture

This section provides a brief summary of internal architecture of sandstone packets and also separating mudstone packets.

4.3.4.1 Internal architecture of sandstone packets

The majority of the sandstone beds within the packets are essentially sheet-like over the extent of outcrop, although detailed logs correlating sandstone beds over the basin are not obtainable for the most due to the inaccessibility of the sections. From detailed photo-
montages of the exposures it was possible to trace sandstone beds across the sections, and from this no obvious compensation cycles have been seen within the sandstone packets; however, a small proportion of beds are observed to pinch out when traced laterally.

Scours are present within the sandstone packets, with a maximum cut down of 5 m and a maximum lateral extent of 25 m. The scours are infilled with sandstones of the surrounding sandstone packets, mud drapes are not found infilling these structures.

Sandstone bedforms have been recorded from sandstone packet AA, these are up to 0.3 m thick and comprise beds of cross-bedded sandstones and are interpreted to represent either traction current reworking of previously deposited sediment or primary deposition from essentially bypassing turbidity currents. Larger scale sediment bedforms are recorded in sandstone packet C, and are shown on Enclosure 4.7 A and E, and Figure 4.4b. They comprise two cross-bedded sandstone bedforms with a wavelength of ~87 m and an amplitude of ~4-5 m, the section is ~20° oblique to palaeocurrent showing a transport direction towards the northwest. The stoss side of these bedforms is erosive with the beds on the leeward side downlapping the sandstone bed below. Similar structures to these have been recorded by Hughes Clarke et al. (1990) from the 1929 Grand Banks turbidity current on the Eastern valley, Laurentian Fan, and from the ancient deep-marine record, bedforms have been recorded by Vicente Bravo & Robles (1995), who describe wave-like symmetric to slightly asymmetric sandy-gravel and gravel bedforms.

Immediately above these bedforms, a 10 m thick packet of sandstones can be seen to downlap the underlying sandstone beds (Enclosure 4.7 A, B, and E). The apparent dip of these downlapping sandstones at its maximum point is calculated at 10°, which is a high angle for beds deposited from waning turbidity currents. One possibility is that they represent reworked sands from either traction or turbidity currents, or they may represent the infill and draping of a scour located proximally at the east-southeast side of the section. At the very top of sandstone packet C, shown on Enclosure 4.7 A and B, there is a 3 m thick sandstone bed
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within the thinner sandstones which pass vertically up into the mudstone packet. This sandstone bed also appears to downlap towards the west-southwest, and is inclined at an angle of \(-30^\circ\).

4.3.4.2 Internal architecture of mudstone packets

The internal architecture of the mudstone packets is relatively simple and comprises predominantly silty mudstones with thin (<1 m thick) sheet-like sandstones of medium- to fine-grained sand. The sandstones are invariably parallel laminated with current ripples present in some beds.

A small percentage of sandstone beds within these mudstone packets pinch-out in a lateral and proximal-distal direction, other sandstone beds pinch out at both ends forming lens-like beds. The mudstone packet above sandstone packet C shows lensing of sandstone beds, three of which are laterally offset (Enclosure 3.8). They are shown in detail on Figure 4.8a, where they have a lateral extent of 130 m and a maximum thickness of 0.8 m.

Channel-like bodies are also present in the mudstones above sandstone packet C (Enclosure 4.8 A, Figures 4.4a, 4.8a), the channel-like body attains a thickness of \(~2.5\) m, and is infilled with lens-like sandstone beds up to 0.5 m thick with an observed total of six beds infilling the channel. Only the southwestern channel-margin is visible, when traced towards the northeast the channel margin is absent, instead the lens-like sandstones described above appear to be lateral equivalents of the this channel margin infill. This channel-like feature is here interpreted to be an incipient channel developed within the mudstone packet. The channel is essentially depositional as no major erosion surfaces or channel lag deposits have been recorded.
Figure 4.8a and b. a) Schematic diagram of sandstone packet C and overlying mudstone packet showing architecture of sandstone beds within the mudstone packet. b) Schematic log of sandstone packet C and overlying mudstone, constructed from logs and photographs.
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Another channel-like body is seen at the top of sandstone packet E on Enclosure 4.8 A located towards the southeastern side of the section, this cuts down in to the sandstone packet and is infilled with thin sandstones (< 1 m thick). The channel is approximately 1.5 m deep and 17 m long, and it may represent the infill of a scour because of its small size and absence of levees.

Below the base of sandstone packet D on Enclosure 4.8 and in detail on Figure 4.9 on the southwestern side of the section and above sandstone packet C, there is a packet of thin (<0.5 m thick) sandstones with beds which are laterally offset and appear to shingle towards the northeast. These laterally offset sandstones are interpreted to represent the lateral infill of a shallow channel or large scour on the seafloor. The overlying sandstone packet D as described in Section 4.3.2.1.1 and shown in detail on Figures 4.4a and 4.8a, onlaps a slope developed in the mudstone packet, which is directly above these laterally offset sandstones, and are here interpreted to represent the rapid infill of this shallow channel by sediment-rich turbidity currents, this is summarised on Figure 4.9. Channel-margin sediment sliding is also evident on the southwesterly margin of this channel.

4.3.5 Sandstone packet onlaps

In general, the sandstone packets show onlap towards the southwest on a palaeoslope which is estimated as having dipped at ~10° towards the northeast. This slope appears to have extended along the entire southwestern length of the sub-basin during deposition of the Grès d'Annot Formation. Figure 4.10 shows the direction of sandstone packet onlaps. The onlaps of sandstone packets AA, A, B and C are visible in the field, but those of sandstone packets D, E, F, G and H are inferred.

Enclosure 4.2 shows a detailed study of the nature of onlap for some of the sandstone packets in exposures that are oblique to transverse to the actual, true, onlap directions. Below the
Figure 4.9a and b. Evolution of channel formation from Enclosure 4.8. a) Channel formation with lateral accretion of sandstones and channel margin slumping. b) Infilling of channel with thick-bedded turbidites which onlap the channel margins.
Figure 4.10. Geological map of the Grand Coyer sub-basin showing onlap surfaces and sandstone packet terminations. (re-drawn from BRGM Sheet 89 1:50 000).
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summit of Tête de Lanconet, the beds within the sandstone packets dip at $-6^\circ$ towards the northwest, and when the sandstone packets are traced towards the palaeoslope the present-day dip of the sandstone beds decrease, the beds thin, and appear to drape in a tangential manner onto the palaeoslope. Also, some packets show complete internal pinchout against other sandstone packets (i.e., sandstone packet C, Enclosure 4.2).

Using detailed observations, sandstone packets located towards the centre of the sub-basin can be correlated with those at the onlap, Enclosure 4.2 shows these sandstone packets. Detailed correlations of the sandstone packets in this part of the sub-basin enabled some of the sandstone packets (shown on Enclosure 4.2 A) to be identified: the basal three sandstone packets on this section are packets A, B, and C which show an internal onlap on the north-northeast side of the section. The overlying sandstone packet is interpreted as packet D or E, or possibly both combined as packets D and E proved difficult to correlate along-strike.

Enclosure 4.2 B shows two detailed logs through the section also shown on Enclosure 4.2 A. Log A is a section through sandstone packets B and D, and Log B is through packets B, D, and E. The packets measured on Log A appear different to those in Log B. Packets B and D, on Log A, show no obvious vertical trend, whereas on Log B both packets show thickening-and-coarsening-upward, and are thinner than their counterparts on Log A. Log A was taken nearer to the basin axis, while Log B was taken nearer to the sub-basin margin, and therefore, demonstrates the effects of the marginal slope. This thickening-and-coarsening-upwards of beds within sandstone packets on Log B is interpreted to be a result of the overall thinning-and-finning of turbidites as they onlap the slope. A turbidity current may travel some way up the slope to deposit sediment, drape the slope, and then leave a deposit as a wedge or "pinchout". A subsequent turbidity current could travel farther upslope before depositional pinchout, simply because deposition will tend to smooth topography at the toe of the slope (Figure 4.15). The thinner and finer grained sandstone beds when traced towards the palaeoslope show facies changes with beds showing an increase in the amount of upper flow regime parallel-lamination and rare ripples (see Log A and B, Enclosure 4.2).
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The thinning of sandstone packets and overall shallowing of bedding dip towards the palaeoslope, is interpreted to result from the interaction between the slope and turbidity currents leading to sand and silt accumulation in the neighbourhood of the base of the palaeoslope. Sediment can accumulate on the palaeoslope provided that the slope angle is low enough. At higher angles sediment will either not be deposited or, if deposited, on the slope will be subject to gravity induced failure. The critical angle, above which sediment will not be deposited on the slope must be equivalent to or less than the angle of repose of the sediment. Alternatively, sediments may not be deposited on the slope if a turbidity current passed along the strike of the palaeoslope, rather than at a high angle, thereby allowing only the margins of the current to ride up the slope and deposit sediment due to radial flow expansion and/or deflection (Pickering et al. 1992).

Within the Grand Coyer sub-basin there is an intra-sub-basinal palaeoslope which was mapped (Figure 4.10, Enclosures 4.1, 4.1.1, 4.3, 4.4 and 4.5). Sandstone packet AA onlaps against this slope. The palaeoslope dipped at 16° towards 013, i.e., at an angle of 32° to the marginal northwest-southeast trending sub-basin slope. The sandstone beds onlap this slope but do not actually ride up it, probably because the slope was too steep to favour sand accumulation. This local palaeoslope within the sub-basin formed a weir-like obstruction to the passage of turbidity currents over the seafloor, with turbidites onlapping towards the south.

4.3.6 Depositional sub-basin morphology for the Grès d'Annot Formation

The Grès d'Annot Formation is interpreted as having been deposited in a southeast-northwest trending trough-like depression on the sea floor (Elliott et al. 1985), with a gentle slope towards the northwest. Palaeocurrents reveal an overall transport direction towards the
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northwest (Figure 4.11), and sandstone packet onlaps within the sub-basin confirm a southeast-northwest trending seafloor slope on the western side (Section 4.3.3). The eastern side of the sub-basin preserves no onlap relationships for the sandstone packets, but from observations on marl thickness variations (Inglis et al. 1982), it is inferred that there was a palaeoslope on the eastern side. The thickness of the marls in the axis of the basin shows a decrease towards the east which mirrors the marl thickness variation on the western side of the sub-basin, therefore it is inferred that a marginal slope existed on the eastern side, located approximately 2 km from the present sub-basin axis. The existence of a sub-basin margin on the eastern side permits estimation of the sub-basin width, which is here calculated as ≥6 km wide. An estimation of the depth of the sub-basin in relation to the marginal slopes suggests that the topography was in excess of 250 m.

Within the sub-basin an intra-basinal slope has been identified (Figure 4.10 and described in Section 4.3.3). Based on observations of sandstone packet terminations and onlaps, a reconstruction of the sub-basin floor topography is possible (Figure 4.13); a complete sub-basin reconstruction is shown in Section 4.6, its relationship with the other sub-basins is presented in Section 6.1.

4.3.7 Synsedimentary deformation within the sub-basin

Early wet-sediment deformation in the sub-basin occurs in two forms; (i) sediment sliding, and (ii) normal synsedimentary faulting.

4.3.7.1a Synsedimentary sliding of sandstone packets

Synsedimentary wet-sediment deformation within the sub-basin include folding, faulting and sliding of sandstone packets and individual beds. The processes that trigger synsedimentary
Figure 4.11. Geological map of the Grand Coyer sub-basin with palaeocurrent data from the Grès d'Annot Formation, also marked on are sediment slide directions where observed. (re-drawn from BRGM Sheet 945 and35-40, 1: 50 000).
t1-t5 represent the slope/basin floor break at sequential times during deposition of turbidites

Figure 4.12. Schematic diagram showing progressive bed onlaps and pinchouts against an approximate 80° palaeoslope; note how the slope/basin-floor break migrates up dip with the deposition of subsequent turbidites.
Figure 4.13. Reconstructed marl topography from the Grand Coyer sub-basin, local topographic feature are labelled.
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deformation include; earthquakes, sedimentary over-steepening of slopes, extensional forces producing normal faulting, compressive forces producing thrust faulting, and preferential subsidence or uplift in the basin as a whole (e.g., Woodcock 1979, Pickering 1987).

Synsedimentary sliding and other styles of wet-sediment deformation is common in sandstone packet AA (Enclosures 4.1.1, 4.1.2, 4.1.3). When traced along the length, the packet comprises two localised disturbed horizons in the upper part. One of these disturbed horizons (Enclosure 4.1.1 A) is 16 m thick by 22 m wide, and comprises a chaotic mixture of folded sandstone beds, muds and silts, and reworked clasts of marl up to 2 m in diameter. The sandstone beds on either side of the disturbed horizon show a curved surface with respect to the margins of this feature which cuts down into these sandstones to a vertical depth of 16 m. Fractures within beds close to the disturbed horizon also parallel the curved surface. The depression in sandstone packet AA (Enclosure 4.1.3 A) is 8 m deep and infilled with a broadly triangular wedge of reworked marl material with a lateral extent of 30 m; it appears to infill the depression created within the packet. The top of sandstone packet AA is marked by a unit of reworked muds, marls and sandstones with clasts of marl up to 1 m thick, and this unit reaches a maximum thickness of 3.4 m (Enclosure 4.1.1), but in other sections is <1 m thick (Enclosure 4.1.2).

The up-dip extent of the intra-basinal palaeoslope located in the northern part of the sub-basin (Figure 4.10), levels out at a point below the sandstone outcrops from the Rocher du Carton section (Enclosure 4.4.). Where the break in slope occurs, there is a poor exposure of contorted sandstones which underlies sandstone packet A. This contorted horizon is on top of the fine-grained mudstone unit which overlies sandstone packet AA and therefore is younger than the sedimentary slide described above from sandstone packet AA. At the southern margin of the sub-basin, there is synsedimentary folding in sandstones (Enclosure 4.6 B), interpreted as representing a northward basinward sliding of sediments.

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A large-scale open fold developed in sandstone packet C and its overlying mudstone unit can be seen (Enclosure 4.5). The fold appears to have created a topographic high on the seafloor which was subsequently onlapped by turbidites of overlying sandstone packet D.

4.3.7.1b Interpretation

The wet-sediment deformation in sandstone packet AA is interpreted as due to sediment sliding. The overlying disturbed material represents sediment which were mobilised by the failure event and subsequently deposited on top of the sandstone packet and close to the base-of-slope. Sediment that was deposited in close proximity to the slope was unstable and eventually failed. The syndepositional topographic high, described in Section 4.3.5.1a, is interpreted to have formed through compression related up-doming of sediment either by sliding or by differential compaction of the underlying sediments.

4.3.7.2 Normal faulting-Sandstone packet AA

The synsedimentary fault cutting packet AA also cuts through the mudstone packet above, which is itself truncated by the overlying sandstone packet A (Enclosure 4.1.2 A). Therefore the relative timing of fault movement can be deduced, such that the fault is younger than sandstone packet AA and its overlying mudstone packet but older than sandstone packet A. The fault produced a scarp on the sea floor which was subsequently scoured and infilled by sandstone packet A (Enclosure 4.1.2).
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4.4 Montagne de Chalufy onlap-Trois Evêchés sub-basin

4.4.1 Introduction and location

The Montagne de Chalufy onlap is exposed at the southern margin of the Trois Evêchés sub-basin (Figure 4.1) and provides a near continuous exposure of Tertiary foreland basin sediments in a 32 km long and 5 km wide NNW-SSE trending belt. The Tertiary foreland basin succession of the Trois Evêchés sub-basin includes localised exposures of the Poudingues d'Argens Formation, and near continuous exposure of the Calcaires Nummulitiques, Marnes Bleues, and Grès d'Annot Formations.

The upper stratigraphic contact of the Grès d'Annot Formation is rarely preserved, but where observed in the Trois Evêchés sub-basin, it is represented by the present-day erosion surface or, locally, the Schistes à blocs unit, above which the Nappes de Embrunais-Ubaye cut across the top of the formation (as seen at the Trois Evêchés summit). This upper unit represents mélanges associated with Alpine thrusting of which the Nappes de Embrunais-Ubaye are a part (Graham 1978) (see Section 1.3).

Detailed surveys by Le Varlet & Roy (1984) and Inglis et al. (1981), and summarised by Ravenne et al. (1987) on the Trois Evêchés sub-basin, reveal that the base of the Grès d'Annot Formation shows a consistent southerly onlap onto a northward dipping palaeoslope developed in the underlying marls (Figure 4.14). This northward dipping palaeoslope is exposed at various localities along the length of the sub-basin and the Montagne de Chalufy onlap represents the southern culmination of these onlaps (Figure 4.14 and 4.15).
Figure 4.14. NNW-SSE cross-section through the Dormillouse-Trois Évêchés-Grand Coyer region. The diagram shows sandstone correlations across the area; note the consistent onlap towards the south of sandstones against the northwardly dipping marl palaeoslope. (From Ravenne et al. 1987).
Figure 4.15. Detailed geological map of the Montagne de Chalufy exposures from the Trois Evêches sub-basin depicting the Tertiary Formations, BRGM sheet89 1:50 000.
4.4.2 The Montagne de Chalufy onlap

The Montagne de Chalufy exposures form a 4 km long strike-section trending in a NW-SE direction. The total thickness of the Grès d'Annot Formation at this locality is ~350 m. The extent of the section studied was from south of Le Grand Croix and northwards to Sommet de Denjuan. The study emphasises the behaviour of the sandstone beds and packets as they onlap a slope in the marls, which locally approached 26°, but was more typically in the order of 12° at this locality, and with a general dip towards the north (Figure 4.15). The section at Montagne de Chalufy is aligned at ~40° to palaeocurrent.

4.4.2.1 Sandstone packet architecture

Three discrete sandstone packets have been studied which onlap a slope developed in the marls, and are labelled from the base upwards A, B, and C (Enclosures 4.9 and 4.10). Maximum observed sandstone packet thickness is ~30 m. The packets are separated by packets of mudstone/sandstone couplets which attain thicknesses up to 27 m.

4.4.2.1.1 Sandstone Packet A

Sandstone packet A is located at the base of the Chalufy section and reaches a maximum thickness of 30 m. The packet comprises amalgamated and non-amalgamated turbidite beds belonging to Facies C2.1 and C2.2 (Bouma T₁ and T₂ with rare T₃ divisions). The base of the packet rests on mudstones and thin sandstones and shows a planar contact. The underlying mudstones occur in a packet to 4 m thick, and are themselves underlain by marls. Individual sandstone beds can be traced to their point of onlap against the palaeoslope developed in the underlying marls, where the
palaeoslope attained a maximum dip of 26° towards the NNW although allowing for compaction the true dip may have been slightly greater. Detailed logs through the sandstone packets are shown on Enclosure 4.9 A. The oldest bed is designated I and was logged for a lateral distance of 120 m. Log A1 shows a sandstone bed with a major grain-size break in its lower half and with mud clasts concentrated in the upper portion; this lower portion thins towards the onlap where it completely pinches out. On log A4, the same bed shows four stepwise normally graded units, not seen on log A3, and most of the stepwise graded units in the bed thin towards the onlap, except for one of the units which thickens before it thins and onlaps against the slope. Bed II, as labelled on the measured sections, thins as it approaches the onlap and has a concentration of mud clasts in a discrete horizon in the centre of the bed which only appears close to the slope; the basal 0.2 m shows stepwise normal grading. Bed III retains its integrity without any major thinning until it reaches the slope, where dish structures are present within the T₄ division. Some of the beds actually drape along the palaeoslope as shown by Bed IV (Enclosures 4.9 A ii, 4.10 A). The palaeoslope, where beds of sandstone packet A onlap, does not have a mud drape, instead the sandstones are in direct contact with the marls.

The complex internal architecture of beds close to the palaeoslope is interpreted to be a result of the complex interplay between the palaeoslope and turbidity currents leading to the preferential deposition of sediment in proximity to the slope. Stepwise grading within beds is interpreted to represent flow surges at a fixed point on the sea-floor as the margins of turbidity currents probably rode up local parts of the slope creating surges in the flow velocities. Flow instability at a fixed point on the seafloor appears to have become less important away, from the palaeoslope, presumably towards the main part of the turbidity current flow, hence no stepwise grading of beds is seen. The overlying mudstone packet is 23 m thick and comprises mudstones with thin (< 0.2 m) sandstones of Facies C 2.2 and 2.3, with Bouma Tₕd divisions present.
4.4.2.1.2 Sandstone Packet B

Sandstone packet B reaches a maximum thickness of 12 m, although locally this value reflects over thickening by synsedimentary folding and sliding (Enclosure 4.9 A and B). The base of the packet shows both local erosive and non-erosive contacts with the underlying mudstones. This erosion is in the form of shallow scours up to 0.5 m deep and 10 m long. The sandstones comprise Facies C2.1 and C2.2 (Bouma Ta and Tb with rare Tc divisions) with some beds showing amalgamation. Towards the southeast of the sandstone packet, a reverse fault can be seen (Enclosure 4.9 A) which cuts out sandstone packet C. Correlation of mudstone and sandstone packets, shows that sandstone packet B has been juxtaposed against packet C. The main exposure of sandstone packet B shows large-scale synsedimentary folds which are interpreted to represent downslope failure of sediment deposited on the slope. At this locality, the slope during deposition had a dip of ~12° towards the northwest. The overlying mudstone packet is 27 m thick and comprises mudstones with thin (< 0.2 m) sandstones of Facies C 2.2 and 2.3, with Bouma Tcd divisions present.

4.4.2.1.3 Sandstone Packet C

Sandstone packet C is 10 m thick and comprises sandstones of Facies C2.1 and C2.2 (Bouma Ta and Tb with rare Tc divisions) with beds showing amalgamation. The base of the packet shows local scouring into the underlying mudstones. Closer inspection of the scour surface reveals that it has a drape of two, 0.2 m thick, sandstones on its surface (Enclosure 4.10 B). The fine-grained interval above comprises mudstones and sandstones of Facies C 2.2 with the sandstones making up at least 50% of the packet. The packet onlaps the slope in the marls, at which point it is draped by mudstones similar to those of the underlying mudstone packet; here the angle of the slope was ~6°.
towards the northwest, which was low enough to permit the deposition of muds on the slope.

4.4.2.2a Large-scale erosion

Sandstone packet B can be traced northwestwards where it thickens and infills an erosional depression on the seafloor (Enclosure 4.10 A). The erosion surface is ~50 m deep and ~300 m wide with an infill of sandstones onlapping the sides of the cut down.

Work by Apps (1987) revealed that the base of the erosion surface contains a lag deposit of poorly sorted pebble conglomerates and reworked clasts of thin-bedded fine-grained sandstones and siltstones. The reworked clasts of sandstones are up to 1 m in diameter and show preferential erosion of a silty horizons (as clasts) which corresponds to the beds immediately below the erosion surface. These reworked clasts rest above and within the conglomerates, and are overlain by structureless pebbly sandstones, indicating that an erosional topography existed even in the base of this major erosional feature (Apps 1987).

4.4.2.2b Interpretation of erosional feature

The aspect ratio of the erosion surface, 50 m deep and 300 m wide appears to indicate that it is not a channel but rather a feature more akin to a canyon-like or gully-shaped erosional surface on the seafloor, and which allowed sediment to bypass this part of the basin. Here, it is interpreted as a submarine slide infill, probably a cross-section through a bottle-neck slide scar.
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4.5 Controls on sandstone and mudstone packet development

The variables which control the development and evolution of deep-marine systems are well discussed in the literature (Pickering 1982, Stow et al. 1984, Mutti & Normark 1987, Shanmugam & Moiola 1988, Normark et al. 1993, Posamentier & Allen 1993, Reading & Richards 1994). Stow et al. (1984) summarise three primary controls on deep-marine system development, all of which may be mutually interactive: (1) sediment supply and rate, (2) tectonic setting and activity, and (3) sea-level variations. Posamentier & Allen (1993) list the basic controlling parameters on stratal architecture in depositional sequences as: (i) sediment flux; (ii) rate of change of sediment accommodation; and (iii) basin physiography. Local factors within a particular basin will play a dominant rôle in determining the internal organisation of depositional sequences, e.g., the position of a shelf/slope break. Eustasy and seafloor subsidence/uplift, together with sediment supply rates, will determine the timing of sequence bounding surfaces, and sediment flux. Basin physiography will determine the geographic position of stratal architecture between the bounding surfaces (ibid.).

4.5.1 Basin topography and stratal architecture

Topography within a depositional basin exerts a fundamental control on the depositional patterns of turbidite systems (Nelson & Kulm 1973). Specific tectonic parameters that determine fan type are the size and internal geometry of the basin, including gradients of the basin margin and floor (ibid.). In convergent margins, morphologically constricted basins develop which exert a control on the style of sediment accumulation (Pickering et al. 1982, Stow et al. 1984). Such complexity typically prevents radial submarine fans from developing, at least until the topography has been drowned. Posamentier & Allen (1993) also note the importance of topography on controlling the depositional patterns of turbidite systems, not only in the receiving basin but at basin margins where shoreline progradation occurs across a ramp margin or a discrete shelf-slope break.
Mutti & Normark (1987) identify four types of turbidite basin which depend on geographic setting, longevity of the basin, and topography. Using this classification, the sandstones of the Grès d’Annot Formation were deposited in a "Type C basin" (Mutti & Normark 1987), i.e., formed on continental crust with a relatively large and long-lived sediment supply with important structural control. In linear basins axial flows may be prevented from expanding in a radial manner, and therefore travel farther than an unconfined flow. The slopes of the Grand Coyer sub-basin probably acted in such a manner as a result of marginal sub-basin slopes.

In the Grand Coyer sub-basin during deposition of the sands, basin-floor topography appears to have controlled the loci of sand deposition (see Section 4.3.3 and 4.3.4. This confinement of the sands, and their stacking into packets in the sub-basin prevented the formation of classic unconfined and "radial" lobes (sensu Mutti & Ricci Lucchi 1972). Sandstone beds within the packets show no lensing geometry (apart from at basin margins), or evidence for compensation cycles (Mutti & Sonino 1982). The formation of such features is interpreted as having been suppressed by the confining effects of the basin topography. The essentially sheet-like lateral continuity of the sandstone packets on the Grand Coyer sub-basin scale, as opposed to mounded and lensing lobes, is characteristic of deposition of turbidites within a confined basin.

4.5.2 Sediment type

Sediment type or calibre is important in controlling the type of turbidite system that will develop. It is well known in the literature that the sediment calibre controls the style of deposit.

Mutti & Normark (1987) distinguish three types of turbidite system based on sediment type: Type I, highly efficient, sand rich; Type II, coarse-grained sediment; Type III, fine-grained...
sediment. Type I deposits are dominantly composed of unchanneled sandstone lobes, which are correlative with erosional channels, with the bulk of the sand derived from shelf-edge deposits. Mutti & Normark (1987), state that Type I deposits are apparently characteristic of elongate foreland basins (Type C basins, Mutti & Normark 1987) where tectonic activity produces and maintains narrow basins that enhance the distance of transport of sand. Type II deposits form relatively smaller turbidite systems and are characterised by coarser-grained sediment than Type I systems. Type III deposits comprise fine-grained and thin-bedded sediments that may record channel-levee complexes and various types of slope drapes, and is characterised by the lack of associated sandstone lobes.

The Grand Coyer sandstones are classified as belonging to Type I sand-rich deposits. Sand-rich systems are less likely to form submarine channels and levees, but instead they tend to form non-channelised sandstone packets. Conversely, mud-rich systems (Type III, Mutti & Normark 1987) tend to form channel-levee complexes and slope drapes, with sandstone lobes generally lacking.

Reading (1991) and Reading & Richards (1994) classified deep-sea depositional systems by sediment calibre, and also feeder system type, e.g., mud-dominated systems are associated with long-lived channel levee complexes, and increasing grain size tends to be linked to an increase in slope, frequency of flow, and impersistence of channel systems. The classification scheme used by Reading (1991) and Reading & Richards (1994) does not include basins which have a complex basin floor topography which will play an important part in the control on sediment dispersal patterns. The classification scheme does, however, include sand-rich slope aprons and submarine ramps, which are thought to be the main sediment supply routes for the sand-rich sediments of the Grès d'Annot Formation.
4.5.3 Sea level fluctuations

The controls exerted by sea-level change on sedimentation within shallow-marine environments are relatively well understood. This influence, however, on deep-marine sedimentation remains less well understood. There is a consensus that sea-level fluctuations play an important part in the supply of sediment to the deep-marine environment (Posamentier & Allen 1993).

Changes in relative sea-level control the sediment flux to the deep-marine environment. In sand-rich systems at the onset of a sea-level fall, there may be an abrupt increase in deep-water sediment influx, with the sediments being sand-rich, and forming widespread tabular turbidite complexes (Posamentier & Allen 1993). The sandstone packets of the Grand Coyer sub-basin are predominantly laterally continuous across the sub-basin, and tentatively are interpreted to represent deposition during a period of lowered sea-level. With a continuing fall in relative sea-level at its lowest level, the sand/mud ratio of sediment supplied to the basin begins to fall as incised valleys preferentially trap sand-rich fluvial sediments (Posamentier & Allen 1993). With this decrease in sand content, turbidity currents become more efficient at partitioning sediment and may allow channel-levee complexes to develop, with finer-grained sediment accumulation. Such a period of channel-levee development during lowstands prior to the ensuing transgression is inferred to correlate with deposition of the mudstone packets in the Grand Coyer sub-basin. Channel formation is absent from most mudstone packets, but occurs within the mudstone packets above sandstone packets C and E, where the channels that are essentially depositional in character. During a transgression and relative highstand, the sediment flux will be low and relatively little deposition will occur in the basin. Within the sandstone succession at Grand Coyer the variations in facies types may be controlled by sealevel variations, but due to the paucity of data and absence of shallow-marine time equivalents, it is not possible to test the controls that sealevel variations may have had on the facies types observed within the sub-basin.
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4.6 Conclusions

The sediments in the Grand Coyer and Trois Evêchés Sub-basins were deposited within the much larger Tertiary deep-marine foreland basin. The relationship between the Grand Coyer sub-basin and the northernmost part of the Trois Evêchés sub-basin located at Chalufy, during deposition of the Grès d'Annot Formation, is shown in Figure 4.16. A complete Tertiary foreland basin reconstruction is presented in Section 6. Both sub-basins are believed to have been openly connected during sandstone deposition, with the Grand Coyer sub-basin in a more proximal setting to the Montagne de Chalufy sandstone outcrops. Sandstone successions from Chalufy and Grand Coyer also show a similar packeting of sediment.

The confining nature of the Grand Coyer sub-basin (Figure 4.13 and Figure 4.16) is inferred to have played an important rôle in the development of the sandstone architecture in this part of the system. The lateral confinement of the sub-basin prevented radial expansion of turbidity currents as indicated from the palaeocurrents from the Grand Coyer sub-basin, which show a consistent direction towards northwest. Downlapping of individual beds at the base of sandstone packets is interpreted to indicate the prograding of sandstone lobes into the deeper parts of the basin.
Figure 4.16 Summary model for the Grand Coyer and Chaluzy areas during deposition of the Grès d'Annot Formation, showing the possible and likely connection of the two areas.
Chapter 5

Sandstone architecture and facies of the Peïra Cava, Contes and Menton areas

5.1 Introduction

5.1.1 Location

Outcrops of the Grès d'Annot Formation, towards the east of the study area, lie within the Peïra Cava, Contes and Menton sub-basins, located in the Alpes Maritimes region of southeast France approximately 15 km north of Nice (Figure 5.1, in detail on Figure 5.2): the Peïra Cava sub-basin is the most northerly, and the Menton sub-basin the most southerly. The Contes sub-basin is situated approximately 9 km WNW of the Menton sub-basin, and approximately 5 km SSW of the Peïra Cava sub-basin (Figure 5.2). All three sub-basins form local topographic lows in the much larger Tertiary Foreland Basin of southeast France. Tertiary sediments are preserved as basin remnants in gently folded synclines with their axial planes trending approximately NW-SE to NNE-SSW.

5.1.2 Previous research

Early research on the Tertiary Grès d'Annot Formation includes that of Keunen et al. (1957) who documented the flysch deposits in the French and Italian Maritime Alps. A detailed study by Bouma (1959, 1962) on the Grès d'Annot Formation outcrops at Peïra Cava led to the construction of the Bouma sequence, based on a detailed graphical approach to sedimentary
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Figure 5.1. Geological map of the Alpes Maritimes and Haute Provence regions of Southeast France, the Menton, Contes and Peira Cava sub-basins marked on as study area (re-drawn from BRGM 1980 Sheet 40 and 45; 1:250 000).

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Figure 5.2. Simplified geological map of the Alpes Maritimes region of Southeast France, showing the spatial relationship between Menton, Contes and Peira Cava (re-drawn from BRGM 1980 Sheet 40 and 45 1: 250 000).
logging. Stanley (1961) provided a detailed account of lateral variations of mineral composition and texture, sandstone facies distribution, depositional processes and possible source areas for the sandstones. Stanley (1967) compared the outcrops at Annot, Contes and Menton with the Gully Submarine Canyon off the Nova Scotian continental margin, Canada. Similarities between both systems led him (ibid.) to the conclusion that the outcrop areas were ancient examples of submarine canyons. Later, Stanley (1974) interpreted the sandstones from Contes and Menton as chiefly being transported by grain flow, fluidised flow and debris flows, with turbidity current flow as a minor component. Stanley et al. (1978) described downslope sediment dispersal patterns in channelised and unconfined outer continental margin environments, with the outcrops at Contes and Menton being interpreted as confined channelised environments, and those at Peira Cava as outer-continental margin environments.

Bouma & Coleman (1985) studied the Grès d'Annot Formation in the Peira Cava sub-basin and concluded that the succession comprised laterally migrating channel-fills and overbank deposits. These conclusions were based on the lenticular nature of shale nests, large foreset bedding with preserved top and bottom set contacts, and variations of palaeocurrent directions in successive layers: similarities were drawn with observations made on the Deep Sea Drilling Project Leg 96 from the Mississippi Fan. Ravenne et al. (1987), based on previous local field surveys from the Peira Cava sub-basin, summarised sandstone onlaps and the sub-basin fills, comparing them to features observed on seismic sections.

5.1.3 General stratigraphy of the Peira Cava, Contes and Menton areas

5.1.3.1 Pre-Tertiary stratigraphy

The pre-Tertiary sediments in this region range back to Carboniferous age (Westphalian-Stephanian), with younger Permian, Triassic, Jurassic, and Cretaceous aged rocks. Crystalline basement rocks cropout towards the north of the area, and comprise gneisses, amphibolites,
migmatites and granites, forming the Argentera-Mercantour Massif which is essentially Hercynian in age (Figure 5.1).

5.1.3.2 Tertiary foreland basin fill

A detailed account of the Tertiary foreland basin fill is given in Section 1.2. Each of the three sub-basins, which form part of the much larger foreland basin share a similar stratigraphy. The oldest formation in the Tertiary succession is the Poudingues d’Argen Formation which has a limited outcrop and is only seen in the Contes sub-basin. Other formations present in all three sub-basins are the Lutetian-to-Priabonian Calcaires Nummulitiques, Priabonian Marnes Bleues, and upper Eocene-lower Oligocene Grès d’Annot Formations. In the Peira Cava sub-basin, isolated outcrops of the Schiste à Bloc unit occur (Wazi et al. 1985).

The precise age relationships between the Grès d’Annot Formation in all the sub-basins remains poorly understood, due to a lack of high-resolution dating. Relative ages of the Grès d’Annot Formation from sub-basin to sub-basin is inferred based on sub-basin fill patterns which enable a relative age relationship between the Peïra Cava and Contes sub-basins to be determined. The sandstone succession from Peira Cava is interpreted to be relatively older than that preserved in the Contes sub-basin.

5.2 Sandstone architecture and stratigraphy from Peira Cava

5.2.1 Introduction

The Peira Cava sub-basin covers an area of approximately 130 km², and is preserved in a gently folded NNE-SSW trending syncline. The northern extent of the sub-basin is overturned as indicated by inverted Bouma sequences and sole marks in turbidites. This overturning is
interpreted to have resulted from the exhumation of the Argentera-Mercantour Massif causing basin inversion and folding of the sediments as it emerged during post-Oligocene times (Fry 1989). The original depositional margins of the sub-basin are not preserved and are now represented by the present-day erosion profile. In general, the Tertiary sediments are moderately well exposed; with well exposed outcrops along road cuttings where measured sections through the sandstones were made (Figure 5.3).

The oldest formation of the Tertiary succession in the Peïra Cava sub-basin is the Lutetian-Priabonian Calcaires Nummulitiques Formation with thicknesses up to 40-50 m. Overlying this, the Priabonian Marnes Bleues Formation attains thicknesses up to 150-200 m, and the upper Eocene-lower Oligocene Grès d’Annot Formation is the youngest preserved sediment of the sub-basin fill, with thicknesses of >700 m.

5.2.2 Grès d’Annot Formation stratigraphy

The Grès d’Annot Formation in the Peïra Cava sub-basin, preserves a maximum measurable thickness of 767 m in the north (Enclosure 5.2, Log B, and Figure 5.3), and 692 m in the south (Enclosure 5.1, Log A, and Figure 5.3). These two logs are separated by a proximal-distal parallel-to-palaeocurrent distance of approximately 7 km, with Log A in a proximal position relative to log B, thereby permitting comparisons between proximal and distal facies variations.

5.2.2.1 Log A

Log A is a 692 m long section through the Grès d’Annot Formation, starting at the contact with the Marnes Bleues Formation (Enclosure 5.1 A, B and Ci). The lower 37 m comprises Facies C2.3 and C2.2 sandstones as rippled and parallel-laminated beds up to 0.5 m thick, and with grain sizes up to fine-to-medium-grained sand. Debrite-like deposits are common in the lower
Figure 5.3. Geological map of the Peïra Cava sub-basin, with location of measured sections and enclosures (redrawn from BRGM 1968).
Chapter 5: Sandstone architecture and facies from Peirot Cava, Contes and Menton areas

270 m of section and comprise reworked well rounded clasts of marl and mudstone up to 1 m in diameter suggesting a degree of traction transport prior to transport as a debris flow. The remainder of the section comprises predominantly Facies B2.1, C2.1, C2.2, and C2.3. Cross-stratified bedforms are common and are ascribed to Facies B2.2, and occur between 220-250 m from the base of the section (Enclosure 5.1Cii and Ciii). Bed amalgamations occur between the thicker-bedded sandstones (Enclosure 5.1Cviii), otherwise beds are separated by Facies C2.2 and C2.3. Bed amalgamations have led to packeting of the sandstones between distinct mudstone packets up to 9 m thick (Enclosure 5.1Civ). The base of thicker sandstones show scour to depths of 2 m into the underlying sandstones and mudstones over distances of 10's of metres (see Section 5.2.4.2). Mudstone and sandstone rip-up clasts are common, and occur either randomly or as discrete horizons within the Bouma Tₐ division (Enclosure 5.1Cv). Sandstone/mudstone ratios are calculated at 3.27 for Log A.

5.2.2.2 Log B

Log B is a 773 m long vertical section through the Grès d'Annecy Formation from the basal contact with the Marnes Bleues Formation (Enclosure 5.2, Log B). The lower 504 m of section was logged in detail, and the upper 504-773 m of section was not logged in such detail due to a lack of suitable and accessible exposure of the thin-bedded sandstone/mudstone couplets and thin-bedded sandstones (Enclosure 5.2A). The contact with the underlying marls is gradational, with sandstone/mudstone couplets of Facies C2.3 and C2.2, up to 0.5 m thick, dominating the lower 50 m of section. The remainder of the section comprises sandstones of Facies C2.1, C2.2, and C2.3, with Facies C2.2 and C2.3 predominant. Sandstone of Facies B2.1 are only present in the thicker-bedded sandstones, up to 10 m thick, and generally occur within sandstones of Facies C2.1 between Bouma divisions Tₐ and Tₕ. Bed amalgamation is absent and thicker bedded sandstones of Facies C2.1 are separated by silty-mudstone/sandstone couplets of Facies C2.2/2.3. Sandstones and mudstones show packeting to a lesser extent than for the beds on Log A. Scouring is present at the base of thicker sandstones, up to 0.5 m deep,
but with beds generally showing non-erosive bases. Mudstone and sandstone rip-up clasts are present in some turbidites, commonly within the Bouma T₂ division. Incipient incorporation of the underlying sediment occurs with "clasts" showing attachment to the underlying in-situ bed (Enclosure 5.1B). The sandstone/mudstone ratio for the lower part of the section logged in detailed is calculated to be 1.07. Palaeocurrent indicators give a transport direction towards the NNE.

Lateral correlations in the distal parts of the sub-basin proved possible (Enclosure 5.3), though not correlatable to Log B. These logs are located approximately 4 km west of Log B, and display similar facies and sand/mud ratios. Palaeocurrent data from these logs also show a transport direction towards NNE.

5.2.2.3 Comparison between logs A and B

Table 5.1 is a comparison between logs A and B. Both logs are separated by 7 km in an approximate proximal-distal, southwest-northeast direction. The sediments in log B are interpreted as the distal equivalents of the sediments in log A, although no actual visual bed correlation was possible. Both measured sections have similar palaeocurrent trends, suggesting that they were derived from the same source area: no topographic barrier existed between the two localities, and therefore they are inferred to be genetically linked.

Although logs A and B are relatively closely spaced, there are marked changes in facies: the sand/mud ratio for log A is 3.27 and for log B is 1.07, representing a change by c. 75% sand in log-A to 50% sand in log B. This reduction in sand content is interpreted to be a result of beds thinning in a distal direction. This relationship is shown on the histogram plots for bed thickness on Figure 5.4, Log-A histogram plot (Figure 5.4aii) shows a greater proportion of thicker beds than the histogram plot for Log-B (Figure 5.4bii).
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<table>
<thead>
<tr>
<th>Position in sub-basin</th>
<th>LOG A</th>
<th>LOG B</th>
</tr>
</thead>
<tbody>
<tr>
<td>Proximal: ~ 2 km from southern margin</td>
<td>Distal: ~ 9 km from southern margin</td>
<td></td>
</tr>
<tr>
<td>Length of measured section</td>
<td>692 m</td>
<td>773 m</td>
</tr>
<tr>
<td>Sand/mud ratio</td>
<td>3.27</td>
<td>1.07</td>
</tr>
<tr>
<td>Bed amalgamation</td>
<td>Present in thick-bedded sandstones</td>
<td>Absent</td>
</tr>
<tr>
<td>Contact with Marnes Bleues formation</td>
<td>Gradational</td>
<td>Gradational</td>
</tr>
<tr>
<td>Palaeocurrent direction</td>
<td>NNE</td>
<td>NNE</td>
</tr>
<tr>
<td>Scouring</td>
<td>Common, up to 2 m deep</td>
<td>Rare, up to 0.5 m deep</td>
</tr>
<tr>
<td>Packeting of beds</td>
<td>Yes, sands and muds</td>
<td>Yes, sands and muds</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Facies present</th>
</tr>
</thead>
<tbody>
<tr>
<td>A1.2/A1.3</td>
</tr>
<tr>
<td>B1.1</td>
</tr>
<tr>
<td>B2.1</td>
</tr>
<tr>
<td>B2.2</td>
</tr>
<tr>
<td>C2.1</td>
</tr>
<tr>
<td>C2.2</td>
</tr>
<tr>
<td>C2.3</td>
</tr>
</tbody>
</table>

| Max bed thickness | 11.25 m | 10.82 m |
| Min bed thickness | 0.002 m | 0.004 m |
| Mean bed thickness | 0.363 m | 0.231 m |

Table 5.1. Summary comparison of logs A and B, from the Peïra Cava sub-basin. The logs are separated by approximately 7 km in a proximal-distal direction.

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Figure 5.4, Histogram plots for log A and B: (a) all beds for log A, (b) detailed histogram for log A, (c) all beds for log B, (d) detailed histogram for log B.

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The proportion of the various facies types between both logs show a marked contrast. Both logs comprise sandstones of Facies C2.1, C2.2, and C2.3, with Facies B2.1 being less common in log B. The only group which is absent from Log B, but present in Log A, are Facies B2.2 which comprise cross-stratified medium- to coarse-grained sandstones up to 0.6 m thick, occurring as regularly spaced bedforms (Section 5.2.4.1). These Facies B2.2 sandstones are abundant between 220 and 250 m in Log A. Log A also contains debris-flow deposits in the lower 200 m of section, and these are ascribed to Facies A1.2 and A1.4.

Notably, the facies which are only present in Log A, the more proximal site, are those deposited from high-concentration turbidity currents and debris flows. Bed amalgamations are very rare to absent in Log B, but common to Log A. Large-scale scours are present in Log A and absent in Log B. Overall, these facies variations indicate that the energy flow levels in the turbidity currents, as expected, were greater in the proximal settings of the sub-basin, and were capable of eroding and bypassing sediment to more distal parts of the sub-basin. The more proximal log A shows features typical of accelerating (Fr >1) and/or equilibrium flows, whereas log B appears to represent an essentially depositional environment.

5.2.2.4 Interpretation of proximal to distal changes

Proximal-distal changes observed in logs A and B are interpreted as indicating thinning of beds in a distal direction, arising from the transition of high-to low-concentration turbidity currents as they pass from proximal to distal parts of the sub-basin. Since correlations of sandstone beds between logs A and B was not possible, this explanation of the transitions from high to low-concentration flows probably is an over simplification for the factors that control the distribution of facies and associated sand/mud ratios described in Sections 5.2.2.1-3.
5.2.3 Sandstone architecture

Exposures of the Grès d'Annot Formation from the Peïra Cava sub-basin are relatively poor, except for sections exposed along road cuttings where logs A and B were measured. The lack of exposure did not facilitate detailed proximal-distal correlations over distances greater than 1 km. Lateral correlations in the distal part of the sub-basin, however were made and are shown on Enclosure 5.3. Small-scale observations on sandstone architecture are possible from exposures in the southern part of the sub-basin.

5.2.3.1 Cime de Sourcas

The Cime de Sourcas exposures are located towards the northwestern margin of the sub-basin (Figure 5.3, Enclosure 5.3), where exposure is relatively poor, and with outcrops tending to show only the thicker bedded sandstones. Thinner-bedded sandstones of Facies C2.2 and C2.3 are relatively poorly exposed. Five sedimentary logs have been measured, allowing sandstone beds to be correlated over a lateral distance of 2.15 km.

Sandstone facies in this section comprise Facies C2.1, C2.2 and C2.3, with Facies B2.1 as a minor component, and occurring with thicker-bedded sandstones of Facies C2.1 (between Bouma T_a and T_b divisions). Bed thicknesses range from <0.01 m up to 10.09 m, and with a maximum grain size of pebble-grade material. Palaeocurrent data indicate transport was toward the NNE. The section records packeting of the thick-bedded sandstones.

The lateral sandstone architecture of the Cime de Sourcas reveals a predominantly sheet-like geometry for the majority of sandstone beds across the section. Other beds, however, as marked on Enclosure 5.3, show changes in thickness when traced laterally. This thinning and thickening of beds occurs both in an east-west and west-east direction: two beds from the section show thinning in both directions (marked on the logs as beds CS/X and CS/5,
respectively). Facies transitions within individual beds is evident, and includes variations in mud clast abundance, and the erosive nature at the base of beds (see Bed CS/Y, Enclosure 5.3).

5.2.3.2 Southern sub-basin architecture

Observations of proximal-distal, and lateral variations in sandstone architecture in the southern part of the sub-basin are limited due to the lack of good continuous exposures and the general inaccessibility of outcrops. At inaccessible outcrops, photo-montages were used as a substitute for detailed logging, thereby permitting at least some summary of the sandstone architecture (Enclosures 5.4, 5.5, 5.6.1 and 5.6.2). At the northern limit of the southern exposures, Log A records a 690 m thick section through the Grès d’Annot Formation (Enclosure 5.1 and 5.4). Log C records a discontinuous 200-m thick section through the formation (Enclosure 5.4), which can be correlated with sandstone packets labelled on Enclosure 5.4. Correlations between Log C and Log A are not possible due to lack of continuous exposure between both logs, but generalised sandstone-packet logs suggest that both individual beds and sandstone packets change in thickness between the logs.

5.2.3.2.1 Sandstone packets A, B, C, and D

Four distinct sandstone packets have been identified as seen in Enclosure 5.4(Photo A and B), and which can also be seen to outcrop over a maximum distance of 625 m in a proximal-distal, south to north direction. The packets appear laterally continuous along section, and are not correlatable with packets in log A.
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5.2.3.2.2 Basal thick-bedded sandstone

At the base-of-section on the eastern side of the southern outcrops (Enclosure 5.4 B and C), a distinct 20 m thick, coarse to pebble-grade sandstone unit occurs, and which is correlated with the thick sandstone unit seen at the base of Log A. The base of the thick sandstone unit infills a depression 3.5 m deep, cut into thin-bedded fine-grained sandstones. The material at the base of this depression comprises deformed marls and mud/silt clasts. The sandstone unit infilling the depression also contains reworked marl, mud and sandstone clasts up to 0.5 m in diameter. Sediment below the depression comprises marly mudstones and thin-bedded sandstones which are equivalent to the thin-bedded sandstone/mudstone couplets at the base of Log A. Palaeocurrents at the base of the sandstone infill indicate a south to north flow orientation.

The sandstone unit shows no obvious amalgamation surfaces or grain-size breaks, and is interpreted to be the deposit of a single event associated with a high-concentration turbidity current. The scoured surface at the base may have been present prior to deposition of the infilling sandstones or it may have been created by the turbidity which deposited the sandstone bed.

5.2.3.2.3 Chaotic horizon

At approximately 10 m above the marl-sandstone contact on the southwestern side of the southern outcrops, there is a chaotic horizon visible (Figure 5.3, Enclosures 5.5 and 5.6.1). The unit attains an estimated thickness in excess of 15 m, and comprises a basal 4 m thick horizon of chaotic contorted sandstones, muds, and marl clasts in a matrix of mud, sand, and pebble clasts (Facies P 2.2). This unit thins to 2.5 m towards SSE where it is lost in cover. The unit is overlain by an 11 m thick sandstone of medium-to-coarse-grained sand, with discrete pebble horizons and pebble nests (Enclosure 5.5).
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The basal contorted horizon is exposed over the entire length of section, and the base shows a planar contact with the underlying mudstone unit (Enclosure 5.5, Photo Dii), the top of the chaotic horizon preserves an uneven contact with the overlying sandstone (Enclosure 5.5, Photo C and Di). At the NNW margin of the section, the chaotic horizon cuts into pre-existing sediment and appears to have injected sediment underneath the pre-existing bed (Enclosure 5.5 C and D). The chaotic horizon comprises intra-basinal sediments of silty mudstone as clasts up to 4 m by 0.5 m, blocks of medium-grained sandstones up to 1.5 m by 0.8 m, reworked marl clasts up to 0.5 m in diameter, and very large well-rounded pebble-grade exotic clasts of granitic/gneissic material. These clasts are set in a matrix of muddy fine- to-medium-grained sandstones. Within this chaotic horizon, concentrations of mud clasts occur (Enclosure 5.5, Photo C), as well as concentrations of pebble-grade exotic clasts. The intra-basinal clasts of silty mudstone and marl are sub-angular to rounded, indicating at least some transport by traction and rolling to abrade the material (Enclosure 5.5, Photo D(iv)).

A sandstone unit which overlies the chaotic horizon is exposed over the entire length of section, and the contact with the underlying chaotic horizon is irregular. The top of the unit is not exposed, but has an estimated minimum thickness in excess of 11 m. The sandstone unit is medium- to coarse-grained (Facies B1.1). Discrete mud-clast horizons occur and define surfaces dipping towards the NNW. Mud-clast pebble nests also occur within the sandstone body (Enclosure 5.5 D and Di). Within the sandstone unit shallowly dipping partings are evident, but there is no obvious grain-size break or concentrations of fines along these partings.

Previous interpretations of this chaotic unit defined it as a "welded slump-turbidite couplet" (Stanley 1982), or a channel-lag deposit formed in a laterally migrating channel (Bouma & Coleman 1985). The interpretation favoured here is essentially that of Stanley (1982), with the chaotic mudstone-sandstone horizon interpreted as a slide/debris flow deposit, and the overlying sandstone as the deposit from an ensuing turbidity current which may have been genetically associated with the preceding slide/debris flow event.
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The chaotic mudstone-sandstone horizon comprises angular to rounded clasts of marl, mudstone, and sandstones. Marl clasts appear well rounded, with the mudstone clasts showing variations from sub-angular to rounded, and the sandstone clasts are angular to sub-rounded. The varying degrees of roundness are interpreted as a result of material differences, the degree of transport the clasts have undergone, and the type of transport each clast has experienced. The well-rounded marl clasts may have originated close to the source of the deposit, while the less well-rounded clasts of mudstone and sandstone may have been incorporated at a later stage in the evolution of the flow. The marls are more competent and are capable of forming slopes in excess of 10° (Apps 1987, Ravenne et al. 1987), whereas the mudstones and sandstones are less competent and will readily break up during transportation. The size of these reworked clasts favours transport by high-velocity currents such as debris flow/gravity-induced sliding down-slope, or by high-concentration turbidity current flow. Within this study, evidence for a northwardly dipping slope located towards the south of this locality is presented in Section 5.3.3.

The large clasts within the slide/debris may have allowed it to retain its high-concentration and hence attain higher velocities on the slope, thereby enabling it to travel faster than any associated turbidity current. Modelling of snow avalanches provide an insight into sediment slides and turbidity currents (Gubler 1989), and demonstrate that the avalanche *sensu stricto* corresponds to a debris flow or grain flow and the associated snow cloud as a turbidity current (Middleton & Hampton 1976). A similar deposit to the one described here has been recorded from the Scotian Slope by Piper et al. (1985), where a seismically triggered downslope slide is overlain by a laterally discontinuous sandy turbidite.

5.2.4 Onlaps

Sandstones from the Grès d’Annot Formation onlap a slope developed in the underlying marls, which cropout at the southern margin of the sub-basin (Figure 5.1, Enclosure 5.6.1B). The
5.2.4.1 La Blanchiera

The Blanchiera onlaps are located at the southern tip of the sandstone exposures (Enclosure 5.6.1A and B), and comprise a 500 m long section trending northwest-southeast. The section preserves marls at the base, above which there is an outcrop containing a 1.5 m thick unit of light-brown mudstones with rare thin sandstones. Above this there are thick-bedded sandstones, up to 3 m thick, with scoured and grooved bases and normal-graded bedding. Sandstone beds onlap towards the west against a slope developed in the marls, which dipped at between 4° and 12° towards the east during sandstone deposition (Enclosure 5.6.1C). The palaeoslope on the southeast side of the section reveals a scalloped surface ~200 m long developed which is developed in, and truncates bedding in the marls. This scalloped surface indicates sediment sliding of the marls or erosion into the marls produced by the passage of turbidity currents over the slope. Sliding of slope sediments are well documented from the modern record, for example in the southern California Borderlands (Clarke 1979, Woodcock 1979).

The La Blanchiera section is relatively inaccessible to log, but observations on individual beds close to the onlap reveal that depositing turbidity currents were affected by the presence of this palaeoslope. Lensing of beds is evident (Enclosure 5.6.1D), and surging of turbidity currents at the base of palaeoslope is indicated by discrete mud-clast horizons, which when traced away from the palaeoslope towards the southeast thin and disappear over a lateral distance of 3 m (Enclosure 5.6.1E and F). The dip of the palaeoslope developed in the marls at the Blanchiera section varies from dips of 6° to 8° on the northwest side to dips of ~10° on the southeast side. This steepening on the southeast side has arisen from erosion of the marls, as indicated by the scalloped surface.
5.2.4.2 Cime du Tournet

The section at Cime du Tournet is located on the western side of the sub-basin outcrops and is approximately 40 m stratigraphically above the palaeoslope seen at La Blanchiera (Figure 5.3, Enclosure 5.6.2 A-D), and ~1.1 km NNW of the onlaps at La Blanchiera. The section is 150 m long and preserves sandstone onlaps against a slope developed in the marls, calculated to have dipped at c. 8° towards the east. Sandstones which onlap this slope show signs of localised shallow scouring to depths of 0.5 m into the marls below. The slope itself is draped by a thin, 0.4 m thick packet of sandstones and mudstones (Enclosure 5.6.2 D), and when traced up the palaeoslope these beds thin and pinch out, and probably represent lateral upslope equivalents of thicker bedded sandstones. Palaeocurrent data indicate a transport direction from south to north. Within individual beds, repetitions of coarse-/granule-grained Bouma Ta- and medium-grained Tb divisions indicate a possible slope control on turbidity current deposition, causing unsteadiness in the turbidity current flow leading to surging and/or fluctuating velocities.

5.2.4.3 Cime de Rocaillon

The section at Cime de Rocaillon trends southeast-northwest, and is approximately 750 m northwest of the onlaps at Cime du Tournet. It is stratigraphically at a similar level to that of the Cime du Tournet onlaps (Enclosure 5.6.2 F), and both sections have been correlated. A sandstone packet (as marked on Enclosure 5.6.2 A and F) provides a reliable correlation horizon between the two sections.

Sandstones onlap against a slope in the marls, calculated to have dipped at approximately 8° towards the east during sandstone deposition. Palaeocurrent data indicate a transport direction
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towards the north. The lower part of the onlap comprises marls with a thin 0.5 m thick drape of thin (<0.3 m thick) sandstones and mudstones, overlain by thicker bedded sandstones, and reworked clasts of marl and mud are present in some beds (Enclosure 5.6.2 H). Towards the western extremity of the onlap, sandstones in contact with the marl-slope show synsedimentary folding. The axes of these folds when corrected for sandstone dip, are oriented approximately parallel to the strike of the slope, favouring downslope sliding of sediment with beds maintaining their integrity (Woodcock 1979, Pickering 1987).

5.2.4.4 Summary of sandstone onlaps

The three localities described above show a consistent onlap towards the west against a slope developed in the underlying marls which dipped at between 4° to 8° towards the east. Palaeocurrents from beds close to the slope, and from measured sections above the slope, show a consistent flow towards the north without any evidence for turbidity current flow reflection. A 200 m thick vertical section of Grès d’Annot Formation succession is observed to onlap the westerly slope in the marls, indicating that there was topography in the sub-basin in excess of 200 m.

Synsedimentary folding of sediments against the slope, indicate either: upslope deposition of sand with subsequent downslope failure, or an increase in the slope angle at some time after deposition of the turbidites resulting in sediment instability and failure. The base of some turbidites show erosion into the slope-drape of thin sandstones and mudstones, to depths up to 0.5 m. A large erosion surface (described in Section 5.2.4.1 at La Blanchiera) is interpreted to represent slope failure. Field & Clark (1979) describe slide-scars from the southern Californian Borderland occurring on slopes as low as 0.2°, and Lewis (1971) describes slumps and slides from continental margins with slopes between 1° to 4°. Slide scars from the slopes described by Field & Clark (1979) down cut to depths as little as 10 m, the erosion surface in the marls described in Section 5.2.4.1 is comparable to those described by Field & Clark (1979).
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The thin sandstone-mudstone slope sediments between the marls and onlapping sandstones do not exceed thicknesses \(>2\) m, which is markedly different to the thin bedded sandstone-mudstone couplets of Facies C2.3/C2.2 observed at the base of logs A and B, and which directly overly the marls. These thin-bedded sandstone-mudstone couplets attain a thickness of 20-30 m, representing a transitional unit from marls to overlying thicker bedded sandstones. This discrepancy in thicknesses between the two sections is interpreted to represent deposition of fine-grained sediments in the axis of the sub-basin, with the lack of fines on the slopes resulting from an over-steep slope preventing sedimentation, or an area in the basin where turbidity current activity was absent due to the position of this locality on the palaeoslope, such that it only received sediment when the sandstone succession reached this level possibly linked to the filling of other sub-basins higher upslope and then overspilling into this sub-basin. The latter proposition could only be tested with better stratigraphic resolution which is beyond the precision of analysis in this thesis.

5.2.5 Scours and Bedforms

Various sedimentary bedforms have been identified from the Peïra Cava sub-basin, including: cross-stratification, cross-stratified regularly spaced bedforms, and scours. Cross-stratification and scouring are common in the proximal parts of the basin and are observed on Logs A and C (Enclosures 5.1 and 5.4, respectively).

5.2.5.1 Scours

Scouring at the base of turbidites, from the Peïra Cava sub-basin occurs predominantly within the southern proximal exposures, and are invariably located at the base of thick-bedded sandstones (Enclosure 5.7 A, B, C and D). The maximum cut-down observed is approximately
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2 m (Enclosure 5.7 A and B), while the proximal-distal extent is 160 m (Enclosure 5.7 D). Where the base of the bed shows erosion into the underlying sediment, there are concentrations of mud clasts and reworked sandstones occurring as part of the scour-infill (Enclosure 5.7 A, B, C and D). The infill of these scours can be complicated, comprising ungraded pebbly sands, normal-graded pebble to medium-grained sands, or cross-stratified pebbly sands. Enclosure 5.7 B depicts eleven logs through a scour-infill over a proximal-distal distance of 80 m. At the deepest part of the scour the sediment infill comprises ungraded pebbly sand with reworked mud and sandstone clasts. In an up-current direction, the scour shallows, and is infilled with parallel and cross-stratified pebbly sands, with an absence of mud clasts. It is evident from Enclosure 5.7 D, which records 160 m of proximal-distal section through a cross-stratified and scour-infill unit, that the scour influenced deposition of the depositing turbidity current, with the scour acting as a site for the preferential deposition of coarse material. The section comprises a normally graded turbidite with a 0.5 m thick cross-stratified unit at the proximal margin of the scour. In a down-current position, the base of the bed scours into the sediment below to a depth of 1 m, the scour-infill comprises non-graded pebbly sandstones with reworked clasts of mudstone and sandstone (interpreted to be derived from the scour), overlain by the normally graded turbidite. Farther down-current, the bed becomes stepwise graded, and in a down-current position the stepwise grading is absent and replaced by a normally graded turbidite. Small-scale scouring to a depth of up to 0.25 m are common (Enclosure 5.7 F).

Interpretation

Large-scale scours formed by gravity currents are well documented in the literature, using sidescan sonar from the modern oceanographic record (Normark 1970, Normark et al. 1979, Hughes Clarke et al. 1990). High resolution observations with deep-towed imaging such as TOBI, still cannot resolve the small-scale scours as described here. In the ancient record, however, small-scale scours have been recorded (Mutti & Normark 1987, Vicente Bravo & Robles 1995, Hilton 1995). Scours described by Mutti & Normark (1987), are in the order of
0.5 m to 10 m across and are found downslope from or close to the terminations of major channels, and are generally infilled with mud drapes, indicating non-deposition from the scour-producing gravity current. Scours from the Peïra Cava sub-basin lack mud drapes but dimensionally are of similar magnitude to scours described by Mutti & Normark (1987). Vicente Bravo & Robbins (1995), describe large-scale flute-like structures from the Albian Black Flysch from northern Spain, which range from 1-5 m deep, and 5-50 m across and appear flute-like in longitudinal section. The infilling sediment records deposition from more than one turbidity current event, and scours from the Peïra Cava outcrops show a single cut and fill history without mud drapes.

5.2.5.2 Cross-stratification and bedforms

Cross-stratification is common in the proximal sandstones of the Peïra Cava sub-basin (Enclosure 5.1, Log A and Enclosure 5.4, Log C), and is absent in the distal part of the succession (Enclosure 5.2, Log B). A concentration of cross-stratified beds occurs between 220 to 250 m upsection on Log A.

Cross-stratification occurs as three different forms: (1) Cross-stratified beds up to 0.7 m thick, with asymptotic foresets, concentrations of mud clasts and coarser grained sands occurring along the foresets (Figure 5.5 a and b), and grain size is up to coarse-granule-grade; (2) Medium-grained cross-stratified units up to 0.5 m thick, occurring at the top of turbidites, generally between the Bouma Tₐ and Tₐ divisions (Facies B2.2) (Enclosure 5.7 D). In a proximal-distal section they may be discontinuous and pass distally into parallel-stratified units (Enclosure 5.7 D); and (3) Cross-stratified coarse to granule-/pebble-grade units up to 0.7 m thick, forming the basal part of a turbidite (Figure 5.5 c, Figure 5.6 a, b, and c). These can also show regular spaced bedforms (Enclosure 5.7, Photo E), with wavelengths of 4.37 m and amplitudes up to 0.45 m, or occur as isolated mounds (Figure 5.6 a). Figure 5.6 c. depicts a 0.65 m thick sandstone bed separated by the enveloping mudstones, and is interpreted to
Figure 5.5a, b and c. a) Laterally continuous cross-stratified bed. b) Cross-stratified unit with local erosion surfaces located towards the top of the bed. c) Cross-stratified bedform, bedform is ~30cm thick.
Figure 5.6a, b and c. Examples of cross-stratified bedforms from Petra Cava.
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represent the deposit from a single event. The basal part of the bed comprises a 0.05 m thick granule-grade discontinuous unit. Above is a 0.45 m thick coarse to very coarse-grained cross-stratified unit which preserves a bedform shape, indicating palaeoflow towards the NNW. This cross-stratified unit is capped by a fine-grained rippled unit up to 0.35 m thick, infilling the topography created by the cross-stratified unit; ripple cross-lamination indicates palaeoflow towards the NNE.

Discussion

The occurrence of cross-stratification between Bouma Ta and Tb divisions, is interpreted to represent traction reworking of previously deposited sediment by the dilute tail-end of high-concentration turbidity currents (Allen 1970). Single cross-stratified beds and composite units (groups 2 and 3 above) are interpreted to have formed at the base of predominantly bypassing high-concentration turbidity currents. Figure 5.6 c. shows a single bed comprising three parts: a thin granule-grade base; a coarse-grained cross-stratified bedform in the middle; overlying fine-grained ripple cross-laminated unit draping the topography created by the bedform. This bed lacks medium-grained sand, and the absence of this grain size is consistent with the bypassing of this grain size population in the turbidity current and its deposition in a more distal setting (Allen 1970).

The fine-grained ripple cross-laminated unit is interpreted as being deposited from the same event since there is no mud between the two deposits, and probably represents deposition from the more dilute parts of the turbidity current. Turbidity currents which by-pass a given point without depositing sediment, or only deposit coarse grain sizes, are associated with high velocities. The absence of this cross-stratified facies type in the distal parts of the basin suggests that turbidity currents were essentially depositing sediment as opposed to bypassing and reworking, with relatively slow flow velocities that were incapable of reworking sands into dune bedforms.
5.3 Grès d’Annot Formation stratigraphy and architecture from Contes sub-basin

5.3.1 Location and stratigraphic context

The Contes sub-basin is located approximately 4 km SSW of the Peïra Cava sub-basin (Figure 5.2), and comprises a Tertiary foreland-basin succession. The sub-basin covers an area of 42 km², and is preserved as a NNW-SSE trending syncline (Figure 5.7). The Tertiary succession comprises a basal Poudingues d’Argens Formation of restricted aerial extent, outcropping at the southern and eastern margins of the sub-basin. Overlying this is the transgressive Calcaires Nummulitiques Formation, overlain by the Marnes Bleues Formation. The Grès d’Annot Formation overlies the marls and is the youngest formation in the sub-basin (Figure 1.6 & 1.7).

5.3.2 Grès d’Annot Formation stratigraphy and facies

The Grès d’Annot Formation in the Contes sub-basin preserves a maximum estimated thickness of approximately 350 m, and records an up-section change, together with lateral variation in sandstone facies types. Sandstone exposures in this area are relatively poor, but with easily accessible exposures along road cuttings. Large-scale observations on sandstone architecture are limited to inaccessible hillside exposures.

5.3.2.1 Vertical facies variations

Observations of sandstone facies types reveals an up-section change from thin-bedded sandstones and mudstones at the base of the succession (Facies C2.2/C2.3). These pass vertically up into amalgamated, thicker-bedded and packeted sandstones (Facies C2.1). Packets
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Enclosure 5.8A

Log B

Enclosure 5.8

Grès d'Annot Formation
Marnes Bleues Formation
Calcaires Nummulitiques Formation
Cretaceous and older

Figure 5.7 Geological map of the Contes sub-basin; showing: palaeocurrent trends during sandstones deposition, apparent onlaps of sandstones and enclosure location.

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of thinner-bedded sandstones and mudstones separate the sandstone packets (Facies C2.2/C2.3). The upper part of the succession predominantly comprises amalgamated sandstones with <5% mud.

5.3.2.1a Basal thin sandstones and mudstones

Basal thin sandstones and mudstones outcrop at the eastern limit of the Grès d'Annot Formation, with accessible and good exposures along road section D115 (Figure 5.7). These beds cumulatively attain a thickness of at least 30 m. A measured 25 m thick section records alternations of sandstone beds and mudstones, with sandstones up to 0.9 m thick (Facies C2.2, C2.3 and C2.4) (Enclosure 5.8, Log B, Photo C). The sandstones are fine-grained with parallel and ripple cross-laminations. Thicker sandstone beds from this section show alternations of parallel and ripple cross-laminations on a 5 cm scale; rare opposing ripple cross-lamination is evident in some beds, indicating flow reflection and reversals (Enclosure 5.8, Photo D). These thin sandstones and mudstones are in contact with the underlying marls; Photo E (Enclosure 5.8) records these thin sandstones onlapping an erosional surface developed in the marls.

5.3.2.1b Packeted sandstones and thin sandstone-mudstone couplets

Above the basal thin sandstones and mudstones described in Section 5.3.2.1a, packeted sandstones and mudstones occur (Enclosure 5.8 A, Log A). The sandstone packets attain thicknesses up to 55 m, and comprise amalgamated sandstones of thicknesses up to 5 m. The sandstones are normally graded from pebble-grade material at their base to medium-/fine grained tops, with common rip-up mudstone and sandstone clasts. Sedimentary structures include parallel stratified coarse-grained sandstones, and ripple cross-laminated fine-grained sandstones; otherwise beds appear structureless and lack mud caps (Facies C2.1). The finer-grained sandstone-mudstone packets attain thicknesses up to 20 m, comprising medium- to
5.3.2.1c Massive amalgamated sandstone units

Massive amalgamated sandstone units occur in the upper part of the Grès d'Annot Formation succession, outcropping on the central to western part of the sub-basin (Enclosure 5.8, Photos B and F). Exposures of these sandstones are poor and relatively inaccessible, apart from outcrops along road sections. Individual beds attain vertical thicknesses up to 8 m (Figure 5.8 a.), but more commonly up to 3-4 m thick. Maximum observed grain size is large pebble-grade material at the base of beds, although commonly they are normally-graded coarse to granule-grade passing up into medium-/fine-grained sand. Mud and sandstone rip-up clasts are common and up to 0.8 m by 1.2 m in size (Figure 5.8 b.). Sedimentary structures include parallel stratified gravels, normal-graded bedding, parallel-stratified coarse-grained sandstones, and rare ripple cross-lamination. Dewatering structures are common and include dish structures, fluid-escape pipes and bed convolutions. Mudstone caps and fine-grained sandstone horizons are rare.

Amalgamation surfaces vary from relatively planar, with little erosion at their base (Enclosure 5.9 A) to amalgamation surfaces cutting down 1 m or greater, with parallel-stratified or structureless pebble-grade material infilling erosion surfaces (Enclosure 5.9 B and C). An exact thickness measurement for this part of the succession is not possible, but a preserved thickness of 150 m represents the best estimate.
Figure 5.8 a and b. a) Thick-bedded turbidite from the southern part of the Contes sub-basin. b) Large reworked clast of sandstone at the base of a bed of Facies C2.1.
5.3.2.2 Interpretation of vertical facies variations

The up-section facies variations observed in the Contes sub-basin, of a progressive overall change from thin-bedded sandstones and mudstones, to interbedded amalgamated sandstone packets and sandstone-mudstones packets, and finally to predominantly amalgamated sandstones in very thick packets represents an increase in turbidity-current flow competence/capacity allied to a net increase in volume of sediment deposited by turbidity currents above this part of the seafloor. The controlling factor for these upward changes was sediment supply due either to increased tectonic activity in the source area making more sediment available for transfer to the deeper parts of the basin, and/or progradation of the deep-marine system, possibly linked to a falling relative sea level.

5.3.2.3 Lateral facies variations

On the eastern side of the Contes sub-basin, the basal thin-bedded sandstones and mudstones separate thicker-bedded sandstones from the marls. On the western margin of the sub-basin, the thin-bedded sandstones are absent and marls are in direct sedimentary contact with thick-bedded sandstones. One possible explanation for the absence of thin-bedded sandstones on the western margin is that the thin-bedded sandstones were only deposited in the axial part of the sub-basin, and that the western sandstone-marl contact was the sub-basin marginal slope. The presence of this slope may have prevented deposition of thin-bedded sandstones if the turbidity currents were of a low velocity and could not surmount the slope. The same situation probably appertains to the Peïra Cava sub-basin, where a similar interpretation is invoked. This variation in the lateral distribution of fine-grained thin-bedded sandstones is also observed in the Peïra Cava sub-basin, with an abundance of this facies in the sub-basin axis and an absence on the western sub-basin marginal slope.
5.3.3 Onlaps

Within the sub-basin one major onlap surface has been recognised (Enclosure 5.8 A), and is exposed at the northeastern extent of Grès d'Annot Formation outcrops (Figure 5.8). The section is 1.5 km long, trends NNW-SSE, and preserves sandstone beds onlapping towards the south against a palaeoslope in the marls. The sandstones, at present, dip 20° towards southwest, whereas the marls are near horizontal. Rotating the dip of the sandstone back to horizontal, gives a palaeoslope which during sandstone deposition dipped northwards at 8°. Because of the lack of outcrops allowing detailed observations on sandstone onlaps and marl-slope orientation, an estimation of the marl-slope orientation was obtained using strike-lines for the sandstone-marl contact. The extent of this palaeoslope, however, in the sub-basin is not known, and it may represent a localised steep slope developed in the marls.

Approximately 1 km SSE of the steeply dipping marl palaeoslope described above, a small-scale onlap of thin-bedded sandstones and mudstones is exposed over a distance of ~25 m, trending northeast-southwest (Enclosure 5.8 E). The onlap surface truncates bedding in the marls, suggesting active erosion of the marls prior to deposition of the thin sandstones and mudstones. The erosion may have resulted either from sliding of the marls, and/or erosion by turbidity currents.

5.3.4 Sandstone architecture

Because of the relatively poor exposures of sandstones, accurate observations on individual bed continuity proved impossible. Sandstone packets are exposed on hillsides but are inaccessible to log. However, packets of amalgamated sandstones can be visually traced for proximal-lateral distances up to 2 km (Enclosure 5.8 B).
5.4 Sandstone architecture and stratigraphy from Menton

5.4.1 Location and stratigraphic context

The Menton sub-basin is located approximately 6.5 km ESE of the Contes sub-basin and 14 km southeast of the Petra Cava sub-basin (Figure 5.2). The sub-basin covers an area of 11.5 km², and is preserved as a NW-SE trending syncline (Figure 5.8). The Tertiary succession comprises the Calcaires Nummulitiques Formation at the base, the Marnes Bleues Formation above, and then the Grès d'Annot Formation which is the youngest in the sub-basin. Exposures of the sub-basin succession are poor, with the best exposures located along road sections. Here, a brief account of the Grès d'Annot Formation from the Menton sub-basin is presented.

5.4.2 Grès d'Annot Formation stratigraphy and facies

The Grès d'Annot Formation in the Menton sub-basin has an estimated thickness of 300 m. Observations of sandstone architecture, and sandstone-marl contacts were not made because of the lack of good exposures. Three measured sections from the sub-basin are presented on Figure 5.10, and their location in the sub-basin is given on Figure 5.9.

The lowermost units of the Grès d'Annot Formation, on the eastern side of the exposures close to the marl contact, comprise thin beds of fine-grained sandstones and mudstones (Facies C2.3/C2.2). Parallel laminations are common, and ripple cross-lamination is present in thicker beds. Upsection from these thin-bedded sandstones, beds become thicker, up to 5.6 m thick, and they are normally graded from pebble to medium-grained at their base to fine-grained at their tops (Facies C2.1)(see log A, Figure 5.10). Sedimentary structures include parallel-laminations, and ripple cross-laminations; beds are non amalgamated and capped by mudstone and thin sandstone/mudstone couplets (Facies C2.2/C2.3). Log B and Log C (Figure 5.10), measured from the exposures to the west of the sandstone outcrops, are higher up in the
Figure 5.9 Geological map of the Menton sub-basin, showing positions of measured sections.
Figure 5.10. Logs showing representative measured sections from the Menton sub-basin.
sandstone succession than Log A, and record a change in facies types. The succession on log B is stratigraphically below that on Log C, and comprises sandstones up to 5 m thick (Facies C2.1), pebble-grade bases are common and normally grading up to fine sandstones. Beds are either separated by mudstones and thin sandstones or amalgamated. Clasts of reworked sediment are common in thicker beds. Log C (Figure 5.10) corresponds to a position stratigraphically above Log B, facies are similar to those on Log B, with a higher proportion of bed amalgamation and thick pebble bases up to 3 m.

The upsection changes in facies recorded in the Menton sub-basin are similar to those observed in the Contes sub-basin. The sandstone succession at Menton is interpreted to have formed in a similar manner to that of the sandstone succession in the Contes sub-basin.

5.5 Conclusions

The Menton, Contes and Peïra Cava sub-basins are interpreted to represent remnants of what was a much larger basin in the study area. The Peïra Cava and Contes sub-basins are in close proximity, and are interpreted to have been connected, with no topographic barrier between the two, allowing unhindered passage of gravity flows. Palaeocurrents between these two areas show an identical trend of south to north.

The spatial relationship of the Menton sub-basin with the Peïra Cava and Contes sub-basins is more tenuous, as the sub-basin is not in close proximity to the other two. Palaeocurrents from the sandstones at Menton indicate palaeoflow northwards, the Contes sub-basin is located 6 km towards WSW, giving a divergence of 60° to general palaeoflow at Menton. These two basins may represent separate unconnected palaeotopographic lows on the sea floor, or remnants of a much wider topographic low.
Sub-basin morphology in the Menton sub-basin is poorly constrained due to the lack of good exposures. At Peïra Cava and Contes, however, it is better understood and a model for these two adjoining sub-basins is presented on Figure 5.11. At Peïra Cava, on the western margin of the sub-basin, a lateral basin-margin slope was present during sandstone deposition. This slope is interpreted to have been the downslope equivalent to the inferred basin slope on the western margin of the Contes sub-basin (Figure 5.11). Palaeocurrent data at Peïra Cava and Menton show northwards flow, indicating the presence of a marl slope dipping towards the north. The angle of this slope is estimated to have been $<1^\circ$, an environment which favoured the sub-basin axial sediments to accumulate. Superimposed on this gently dipping slope in the Contes Sub-basin was a steeper slope dipping $8^\circ$ northward towards the Peïra Cava sub-basin.

Sandstone facies in the proximal outcrops at Peïra Cava show evidence of flow bypassing in the form of isolated bedforms and the absence of certain grain-size fractions in individual beds, which are interpreted to have been deposited during a single event. Large-scale erosion is also common in the proximal outcrops at Peïra Cava.

Sediment erosion and bypassing, in the deep-marine environment, occurs when the flow velocity of turbidity currents is high and flows are in a supercritical state. Komar (1971) concluded that turbidity currents would be supercritical on slopes $>5^\circ$, and the slope at the northern margin of the Contes sub-basin is estimated to have been $8^\circ$, which is well within the criteria for the formation of supercritical flows. When turbidity currents reach the base of such a slope they are subcritical, and become involved in an hydraulic jump, and undergo an increase in turbulence (Komar 1971). This increase in turbulence can explain scour formation in the proximal sandstones in the Peïra Cava sub-basin. During an hydraulic jump, the flow velocity will decrease to roughly half its pre-jump value (Komar 1970), and one might expect a large volume of sediment to be deposited in the process. However, the competence of a flow may remain high, and only the coarse material in the bedload be deposited. This phenomenon would explain why coarse-grained sand/gravel bedforms were preferentially formed in the proximal Peïra Cava sub-basin sediments.
Figure 5. 11. Palaeo-reconstruction for the Pei'a Cava and Contes sub-basins during sandstone deposition. The two sub-basins are inferred to be openly connected forming a much larger basin.
5.5.1 Depositional model for the Contes and Peïra Cava sub-basins

A depositional model based on field observations of facies distributions and palaeoslope orientations for the Contes and Peïra Cava sub-basins is presented on Figure 5.11. The model shows the two sub-basins during sandstone deposition, at a time when the Peïra Cava sub-basin fill has reached the northwardly dipping intra-basinal slope, within the Contes sub-basin, the basal fine-grained sandstones are also present. The expanded part of the model represents the remainder of the sub-basin fills, removal of which reveals the position of the sub-basin marginal slopes.

The part of the model depicting the Peïra Cava sub-basin shows the inferred proximal to distal changes in the sand/mud ratio, with sandstones becoming less dominant in the distal part of the sub-basin. Also shown at the southern part of the Peïra Cava sub-basin are chaotic horizons located in close proximity to the localised northward dipping palaeoslope, scouring is also shown in the southern part of the sub-basin. The southern most part of the model, which corresponds to the Contes sub-basin, schematically shows the upsection change in sandstone facies from amalgamated and non-amalgamated packets to essentially amalgamated sandstones. A western margin to the Contes and Peïra Cava sub-basins is shown, an eastern extent to these two sub-basins is not known due to the lack of exposure. It is inferred that deposition proper of sandstones in the Contes sub-basin would only occur once the Peïra Cava sub-basin sandstone fill reached a level which corresponded to the top of the localised palaeoslope located towards the northern margin of the Contes sub-basin. The Peïra Cava sub-basin is compared to a base of slope system, possibly akin to a slope apron. The sandstone succession preserved at Contes is interpreted to record the progradation of a submarine sand-rich ramp-like system which prograded over the succession at Peïra Cava.
6.1 Palaeogeographic reconstruction during sandstone deposition

6.1.1 Introduction

The major problem with studying outcrops of ancient deep-marine systems is the relatively poor preservation potential of complete systems (Pickering 1982, Stow 1986). Subsurface ancient deep-marine systems may invariably have the advantage of preserving the entire deep-marine sediments and their shallow marine correlative equivalents. Subsurface deep-marine systems may also preserve submarine feeder channels, canyons, and shelf-slope breaks on seismic records: basin topography and onlap relationships may also be evident and can be used to construct a reliable basin palaeogeography.

The advantage of ancient outcrop over subsurface deep-marine systems is that mesotopographic features such as scours, small-scale channels, and local basin features are readily identifiable. Detailed observations on the internal architecture and facies of sandstone packets are observed, as well as proximal-distal changes over distances up to several tens of kilometres. Such features may not be distinguishable in the relatively widely-spaced wells drilled and seismic lines shot during exploration of subsurface systems.

In contrast to many other outcrops of ancient deep-marine systems, the Grès d'Annot Formation is relatively undeformed, and preserves good proximal-distal exposures of up to 6
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km, and lateral exposures up to 4 km. These good exposures allow relatively large-scale observations of sandstone and mudstone/sandstone packet geometry. The Grès d'Annot Formation exposures also preserve contacts with the underlying basin floor and sub-basin confining palaeoslopes. The basin successions require little tectonic unravelling, thereby enabling accurate measurements of syndepositional slope orientations and dips. Basin topography is rarely preserved in exposures of ancient deep-marine systems, but, the Grès d'Annot Formation in the Tertiary foreland basin of SE France shows outcrops which are comparable to high-resolution seismic sections in subsurface deep-marine systems with the advantage that small-scale detail can be observed.

Exposures of the Grès d'Annot Formation are preserved as isolated areas within the larger Tertiary foreland basin. These isolated exposures are termed "sub-basins" and represent the infilling of topographic lows within the foreland basin. These sub-basins record differing facies distributions. Due to the isolated nature of the sub-basins some of them are not obviously genetically linked (i.e. Annot and Grand Coyer), while other sub-basins can be (i.e. Peïra Cava and Contes).

The Grès d'Annot Formation from within the studied sub-basins appear to lack any inter-basinal correlation horizons, therefore, time equivalence between sub-basins remains poorly constrained. Palaeontological evidence to correlate between sub-basins within the Grès d'Annot Formation using palynomorphs and nannoplankton have not been studied, therefore inferences for sub-basin fill correlations are based on similarities in sedimentary facies and the relative positions of sandstone successions within adjacent sub-basins. The base of the Grès d'Annot Formation from within different sub-basins can be correlated by using the contact between the Marnes Bleues and Grès d'Annot Formations.
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6.1.2 Palaeogeographic reconstruction and sub-basin fills

Two distinct systems within the much larger Tertiary foreland basin are identified. The first system includes the St Antonin, Entrevaux, Annot and Grand Coyer sub-basins, which form the western basin-floor system (Figure 6.1); the second system includes the Menton, Contes and Peïra Cava sub-basins which form the eastern basin-floor system (Figure 6.1). These two systems have been distinguished using palaeocurrent analysis and basin-floor morphology. Palaeocurrents from these two areas appear to deviate around the present day outcrops of the Dome de Barrot (Figure 6.1), which may have been a structural high on the seafloor during deposition of the sandstones. A combined palaeo-reconstruction of the two systems at sequential intervals during basin infill is presented on Figures 6.2 a, b and c. The reconstructions show the relative sub-basin positions during sandstone deposition, with sediment dispersal patterns deduced from palaeocurrents taken from the Grès d'Annot Formation. Sediment supplied to the sub-basins is interpreted to have been derived from a landmass located towards the south of the study area. Work by Stanley (1963, 1964) and Ivaldi (1974) suggest that the sediment source was from a landmass towards the south probably represented by the Corsica-Sardinia massif prior to its rifting and anti-clockwise rotation away from the mainland European continent during Miocene times. The two systems are discussed separately below in Sections 6.1.2.1 and 6.1.2.2, respectively.

Within the Tertiary foreland basin, outcrops of the Grès d'Annot Formation from the Trois Evêchés-Dormillouse and Col de la Cayolle sub-basins, have not been studied in detail, but work by previous authors indicates that they were also sourced from the south and southeast of the foreland basin, with palaeocurrents showing flow towards the north and northwest (Bouma 1959, Stanley 1961, Ivaldi 1974, Elliott et al. 1985). These sub-basins are in a distal position relative to the sub-basins studied in this thesis. The relative positions of these sub-basins to those in the southern part of the foreland basin is shown on Figures 6.3a, b and c, and discussed in Section 6.1.2.3.
Figure 6.1. Geological map of the Alpes Maritimes and Haute Provence regions of Southeast France (re-drawn from BRGM 1980 Sheet 40 and 45 1: 250 000). Local topographic highs and palaeoslopes are marked on. The restored position of the Col de la Cayolle sub-basin is marked, based on Graham (1986) (data for palaeo-highs from this study and Ghibaudo, in Elliott et al. 1985).
Figure 6.2a. Palaeogeographic reconstruction of the Tertiary foreland basin during deposition of the Grès d'Annot Formation; the model depicts the sub-basins at the onset of sandstone deposition. The exact timing of sub-basin fills with respect to each other is not known.
Figure 6.2b. Palaeogeographic reconstruction of the Tertiary foreland basin during deposition of the Grès d'Annot Formation; the model depicts the sub-basins midway during infill. The exact timing of sub-basin fills with respect to each other is not known.
Figure 6.2c. Palaeogeographic reconstruction of the Tertiary foreland basin during end deposition of the Grès d'Annot Formation; the model depicts the sub-basins after complete infill.
6.1.2.1 St Antonin-Grand Coyer sub-basins

The St Antonin, Annot and Grand Coyer sub-basin exposures, with minor exposures at Entrevaux located between St Antonin and Annot, form the western basin floor system. At the northern extent of the Grand Coyer sub-basin the Chalufy exposures are located, forming the southern extent of the Trois Évêchés sub-basin exposures.

6.1.2.1a Summary interpretations of sub-basin fills

6.1.2.1a (i) St Antonin-Entrevaux

The St Antonin exposures are the most proximal in the western basin-floor system, and are preserved in an east-west trending overturned syncline with the limbs of the syncline dipping north. Bodelle (1971) suggested that the sandstone succession is up to 1000 m thick, comprising three distinct members, the lower 400 m thick member of the succession has been studied in detail, and a summary of the facies and features from all three members is presented in Table 6.1 (and for detail see Section 3.7.2). An accurate interpretation of the succession is difficult to ascertain because of the limited lateral extent of the sections and poor exposure of the basal contact between the sandstone succession and underlying marls.

The St Antonin conglomerate and sandstone succession has previously been interpreted as being deposited in a proximal subaqueous slope environment in front of an alluvial fan-backed submerged fan-delta complex (Stanley 1980). Benthic and pelagic foraminifera suggest that the slope remained submarine throughout conglomerate and sandstone deposition (Bodelle 1971). Stanley (1980) interpreted the conglomerate units as associated with the seaward progradation of an alluvial fan, with conglomerate and sandstone deposition from traction and gravity mass flows, suggesting a prodeltaic slope environment, with the slope dipping northwest (ibid.).

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Figure 6.3 a, b, and c. Conjectural palaeogeographic reconstruction of sub-basin fills in the Tertiary foreland basin of SE France. a) early sedimentation, sub-basins towards south of foreland basin are infilled first. b) southern sub-basins infilled and systems prograde to deeper parts of foreland basin towards north. c) final stage of foreland basin infill, sub-basins are located.
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#### Table 6.1 Comparison table for sub-basins of the western basin-floor system

<table>
<thead>
<tr>
<th>Preserved Succession</th>
<th>St Antonin</th>
<th>Entrevaux</th>
<th>Annot</th>
<th>Grand Coyer</th>
<th>Chalufy</th>
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<td>coarsening and thickening-up sequence</td>
<td>coarsening and thickening-up, lower non-amalgamated, upper amalgamated</td>
<td>packeted amalgamated sandstones and sandstone-mudstones</td>
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<td>NNW-N-NNE</td>
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<td>NW-WNW</td>
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<td>vfs-granule/pebbly</td>
<td>vfs-cobbles</td>
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---

Table 6.1 Comparison table for sub-basins of the western basin-floor system

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interpretation of the depositional environment for the St Antonin conglomerates and sandstones given by Stanley (1980) is favoured here.

During sandstone deposition, the Entrevaux section was separated from the St Antonin outcrops by ~20 km in a northwest-southeast direction. The 155 m thick vertical sandstone succession at Entrevaux records an overall coarsening-and-thickening-up sequence from the base upwards above the marls. The Entrevaux section is interpreted to represent deposition in the more distal parts of the basin than the St Antonin section, and shows a sediment source from the sandstone succession at St Antonin. The exact stratigraphic relationship between St Antonin and Entrevaux is poorly constrained, but the succession at Entrevaux is here interpreted to represent deposition at the base of the pro-delta slope, with the up-section facies changes at Entrevaux representing progradation of the pro-delta slope at St Antonin.

6.1.2.1a (ii) Annot

The Grès d'Annot Formation exposures at Annot preserve a 690 m thick vertical section through the formation and represents an overall coarsening-and-thickening-upward succession. The sandstone succession at Annot is divided into two members: a Lower and Upper Member. The Lower Member comprises turbidites in beds up to 6 m thick which are non-amalgamated and have a sheet-like geometry. The Upper Member comprises amalgamated sandstone packets up to 80 m thick with interbedded sandstone-mudstone packets up to 5 m thick, both comprising turbidite sandstones. The Annot sub-basin characteristics are summarised in Table 6.1. Within the Upper Member, at a height of 510 m from the base of the sandstone succession, there is a 34 m deep cut-down into amalgamated sandstone packets, interpreted as a mega-flute being eroded by high-concentration gravity flows (Enclosures 3.6 and 3.7, Sections 3.2.6 and 3.27). The entire Annot sandstone succession is interpreted to reflect an overall increase in sand-rich turbidity currents, from essentially depositional and non-erosional, to depositional and erosional, bypassing, flows. This up-section increase in sand content and bulk mean grain
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size is interpreted here to represent progradation of the deep-marine system into the Annot sub-basin, which was supplied with sediment from the pro-delta slope at St Antonin (Figure 6.2a, b, and c).

6.1.2.1a (iii) Grand Coyer

The Grand Coyer sub-basin is located approximately 4 km north of the Annot sub-basin and preserves a 374 m thick succession of Grès d'Annot Formation, the characteristics of which are summarised in Table 6.1. The succession is typified by sandstone and sandstone-mudstone packets; the sandstone packets attain thicknesses up to 65 m thick, and sandstone-mudstone packets up to 25 m thick. Turbidite beds within the sandstone packets are invariably amalgamated with rare mudstone caps. Scouring up to 10 m deep is present at the base of some sandstone packets, and shallow scouring up to 3 m deep within sandstone packets is common. The sub-basin forms an elongated trough trending SE-NW and bounded on the SW by a slope developed in the marls which dipped at ~8° NE, against which the Grès d'Annot Formation onlaps. A marl slope is inferred to have been present on the NE side, though sandstone onlaps are not preserved. Palaeocurrent data indicates sediment transport towards the NW. The packeting of sediments in the Grand Coyer sandstone succession is interpreted to represent fluctuations in the amount of sediment supplied to the sub-basin from the proximal equivalents at Annot and St Antonin. An increase in sediment supply from the source area which resulted in the progradation of sand-rich packets, here termed sheet-like confined lobes. Two of these sheet-like confined lobes downlap towards the NW, indicating progradation from the SE (Enclosures 4.6, 4.7, and 4.8).

The Chalufy exposures are located ~4 km WNW of the sandstone exposures at the northwestern extent of the Grand Coyer sub-basin (for detail, see Section 4.4, Table 6.1). The Chalufy section comprises the southern extent of the NNW-SSE trending Trois Evêchés sub-basin and is characterised by packets of sandstones and sandstone-mudstone couplets.
6.1.2.1b Genetic relationships of St Antonin, Annot, and Grand Coyer sub-basins

A restoration of the Annot and St Antonin areas during sandstone deposition (Upper Eocene-Lower Oligocene) indicates that the two areas were separated by ~25 km in a NW-SE direction, and that the outcrops at Entrevaux were located 14 km SE of Annot. The sediments that accumulated between St Antonin and Annot are poorly preserved and are represented by the succession at Entrevaux, but because of the paucity of data between St Antonin and Annot, their exact relationship remains unclear. The Annot sub-basin sandstones are interpreted to be distal equivalents of the pro-delta slope sediments at St Antonin.

The principal difference between the St Antonin and Annot sandstone successions is in the general absence of conglomerate clasts in the Annot succession (Table 6.1). Stanley (1980) noted that the most coarse debris may have been trapped in depressions such as perched basins on the slope between the St Antonin and Annot sub-basins. It is likely that coarse sediment remobilised at the pro-delta slope-top would be deposited either farther down the pro-delta slope or at the base of the pro-delta slope where gradients were low enough to allow deposition of the coarser fraction from sediment gravity flows. A general slope dipping towards the north and northwest is inferred from palaeocurrent dispersal patterns, but the exact angle of this slope is not known. Assuming a slope in the order of 1°-2° situated below the pro-delta slope, then any turbidity currents reaching the base of slope would probably be in a state of autosuspension (Southard & Mackintosh 1981), and hence would be able to bypass these slopes and travel farther into the basin to deposit sediment at Annot.

The relative positions of St Antonin, Entrevaux and Annot, and temporal sub-basin fill patterns are depicted on Figures 6.2a, b and c. Figure 6.2a shows early sub-basin fill at Entrevaux and Annot. At this stage, sedimentation in the Grand Coyer sub-basin is interpreted to be restricted to thin-bedded sandstones and mudstones accumulating in the axial part of the sub-basin.
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Figure 6.2b shows deposition of the Upper Sandstone Member at Annot, and interpreted sandstone deposition proper at Grand Coyer. Figure 6.2c depicts the St Antonin and Annot sub-basins when they were completely infilled with the pro-delta sediments being deposited in the deeper parts of the basin.

The exact stratigraphic relationship between the Annot and Grand Coyer sub-basin is difficult to ascertain due to the absence of sandstone outcrops between the two areas and gentle folding of sediments. Sandstone packets within the Grand Coyer sub-basin downlap towards the northwest, indicating that sandstone packets prograded into the Grand Coyer sub-basin, and were supplied with sediment from the Annot and St Antonin sub-basins in the south and southeast. It is likely, based on stratigraphic positions of sandstone successions from Annot and Grand Coyer, that the Upper Member of the Annot succession is essentially time-equivalent to the lower parts of the Grand Coyer sandstone succession in the southern part of the Sub-basin.

The Grand Coyer sandstone exposures at the northern extent of the sub-basin are interpreted to have been openly connected to the outcrops at the Montagne de Chalufy exposures, though no definite marker horizon can be traced between these two sub-basins. The connectivity of the two sub-basins is inferred as they are in close proximity and separated by ~4 km, palaeocurrent data from the two areas also share a similar orientation which is towards northwest suggesting sediment transport was from southeast to northwest, and hence to a more distal part of the same system.
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6.1.2.2 Menton-Contes-Peïra Cava sub-basins

The Menton, Contes, and Peïra Cava sub-basins are located towards the east of the study area, and form part of the eastern basin floor system (Figure 6.1). Palaeocurrents from the Grès d'Annot Formation, from within the sub-basins described above, record a sediment dispersal pattern towards the north. Based on stratigraphic relationships with that of the Contes sub-basin the Peïra Cava sub-basin fill succession is interpreted to be the oldest part of the Grès d'Annot Formation in this eastern system. Palaeo-reconstructions of these sub-basin fills, during different stages of infill, are presented on Figures 6.2a, b and c.

6.1.2.2a Summary interpretations of sub-basin fills

6.1.2.2a (i) Peïra Cava

The Peïra Cava sub-basin is located 4 km NNE of the Contes sub-basin and 14 km NW of the Menton sub-basin (Figure 6.1). Within the Peïra Cava sub-basin two measured sections have been taken and are represented by Logs A and B (Enclosures 5.1 and 5.2 respectively), Log A represents a proximal section from the southern part of the sub-basin with Log B taken from a more distal section located 7 km north of Log A. Both logs are summarised in Table 5.1 (Section 5.2.2). Both sections comprise turbidite sandstones and other sediment gravity flow deposits, and both share a similar facies at the base of section, comprising predominantly Facies C2.2 and C2.3. The marked difference between the two section is the sandstone/mudstone ratio which is 3.27 from Log A and 1.07 from Log B. Apart from the base of Logs A and B, there are no obvious overall up-section facies changes, however, packeting of sandstones is evident in both sections.

The Grès d’Annot Formation succession from the Peïra Cava sub-basin is interpreted to represent a base-of-slope sub-basin fill with proximal to distal changes indicating deposition of predominantly sand-rich turbidites in the proximal parts of the sub-basin, and relatively sand-
deficient turbidites in the distal parts of the sub-basin. The packeting of sandstones is interpreted to represent periods of increased sediment supply to the sub-basin.

6.1.2.2a (ii) Contes

The Contes sub-basin is located 4 km SSW of the Peïra Cava sub-basin and 6.5 km WNW of the Menton sub-basin (Figure 6.1). The Contes sub-basin comprises a 350 m thick Grès d'Annot Formation succession, which records an up-section change from basal thin-bedded turbidite sandstones to packeted sandstones and sandstone-mudstone couplets, to amalgamated sandstones in the upper part of the succession (Table 6.2 and Section 5.3.2). The up-section changes in the sandstone succession are interpreted to indicate an increase in sediment flux to the sub-basin arising from an increase sediment supply in the source area to produce a progradation of the deep-marine system into the Contes sub-basin. At the northern margin of the sub-basin, sandstone beds onlap against a northward dipping palaeoslope developed in the marls.

6.1.2.2a (iii) Menton

The Menton sub-basin is located 6.5 km ESE of the Contes sub-basin and 14 km SE of the Peïra Cava sub-basin (Figure 6.1). The sub-basin preserves an approximate 300 m thick Grès d'Annot Formation succession comprising sandstone turbidites. The succession records an up-section change from basal thin sandstone-mudstone couplets to progressively thicker and coarser-grained turbidites with increased amounts of bed amalgamation and increased sandstone-mudstone ratios (Table 6.2). The up-section changes in the sandstone succession are similar to those from the Contes sub-basin, and are interpreted as forming in a similar manner (see above).
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Table 6.2 Comparison table for sub-basins of the eastern basin-floor system

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6.1.2.2b Genetic relationship of Menton-Contes-Peïra Cava system

The Peïra Cava sub-basin sandstones are interpreted to be the oldest within the eastern basin-floor system. Sediments were supplied to the Peïra Cava sub-basin via the Contes sub-basin, with little or no sandstone deposition within the Contes sub-basin at this stage (Figure 6.2a). The Peïra Cava sandstone succession is interpreted to have been deposited close to a base of slope, located between the Peïra Cava and Contes sub-basins. Sandstone deposition within the Contes sub-basin could have taken place when the intra-basinal slope was completely covered with sediment, although fine-grained sediment may have been deposited in the axis of the sub-basin (Figures 6.2b and c). The similarities in sandstone successions and the close proximity of the Contes and Menton sub-basins suggest that they shared a similar fill history.

6.1.2.3 Northern sub-basins

The northern sub-basins include the Trois Evêchés and Col de la Cayolle outcrop areas located north of the present study area (Figure 6.1). The Trois Evêchés and Col de la Cayolle sub-basins have only been briefly studied in this thesis, but a summary is presented below on work carried out by Ghibaudo, summarised in Elliott et al. (1985) and by Ravenne et al. (1987).

6.1.2.3a Summary interpretations of northern sub-basins

6.1.2.3a (i) Trois Evêchés sub-basin

The Trois Evêchés sub-basin remnant exists as a 34 km long, NNW-SSE trending, 10 km wide linear belt, preserving a total restored vertical sandstone succession of 2900 m of Grès d'Annot Formation. At Dormillouse, in the northern part of the sub-basin the succession thins towards
the south at Montagne de Chalufy where it attained a restored thickness of ~400 m. Observations along the sandstone-marl contact, record a consistent apparent sandstone onlap towards the south against a slope developed in the underlying marls, which is calculated to have dipped northwards. Calculations of slope angles, based on these restored section of Ravenne et al. (1987), indicates an average dip of ~5° towards the north with local slopes up to ~15° located towards the centre of the outcrop area between the Trois Evêchés and Tête Noire summits (Figure 4.14, Section 4). Elliott et al. (1985), note that this palaeoslope dipped towards the NNE, and that palaeocurrents from within the Grès d'Annot Formation indicate turbidity currents travelling towards the west and northwest. This consistent onlap of sandstone packets towards the south and the stratigraphic correlations made by Ravenne et al. (1987) (Figure 4.14, Section 4) imply that the oldest part of the Grès d'Annot Formation is located in the northern part of the sub-basin, and that the sediments become progressively younger as they onlapped towards the south.

The Grès d'Annot Formation in the Trois Evêchés sub-basin comprises non-channelised sandstone bodies alternating with subordinate amounts of fine-grained and thin-bedded sandstone deposits (Elliott et al. 1985). The sandstone bodies attain thicknesses between 10-80 m and the fine-grained thin-bedded deposits attain thicknesses <25 m (ibid.). The Grès d'Annot Formation from the sub-basin, below the Trois Evêchés summit, preserves a 1000 m thick succession interpreted as aggradational lobe deposits alternating with interlobe deposits in an outer fan setting (Ghibaudo, in Elliott et al. 1985). At the top of the succession, there is 150 m thick unit of non-channelised, coarse-grained and pebbly sandstone, interpreted as the proximal part of a suprafan lobe with rare pebbly channel-fill deposits at the top of the succession representing fan progradation. Within the Trois Evêchés sub-basin succession three key beds comprise coarse matrix-supported conglomerates up to 10 m thick, which can be traced along section for distances up to 6 km in a SSE-NNW direction.

Observations and interpretations on the sandstone facies and sandstone bodies made by Elliott et al. (1985), and correlation of sandstone bodies by Ravenne et al. (1987), including the
southerly onlaps of sandstone bodies, suggests that the Grès d'Annot Formation succession from the Trois Evêchés sub-basin represents an onlapping base-of-slope fan/apron system.

6.1.2.3a (ii) Col de la Cayolle sub-basin

The Col de la Cayolle sub-basin is located ~15 km east of the southern half of the Trois Evêchés sub-basin and covers an area of ~200 km², preserving a southeast-northwest aligned sub-basin which was at least 18 km wide (Elliott et al. 1985). Apps (1987) suggested that the present-day position of the Col de la Cayolle sub-basin should be moved 20 km towards the ENE to allow for later alpine thrust movements below the sub-basin and which propagated towards the WSW.

The Col de la Cayolle sub-basin shows a 1200 m thick succession of Grès d'Annot Formation, with palaeocurrents indicating a transport direction towards the NW and NNW. The lower part of the succession shows palaeoflow towards the north and northwest, while higher up in the succession the palaeocurrent direction is more varied (Ghibaudo, in Elliott et al. 1985). The deepest part of the sub-basin formed a narrow NNW-SSE trending linear depression ~9 km wide and focused about 500 m of vertical sandstone succession in a local depocentre confined in a down-current direction (Ghibaudo, in Elliott et al. 1985).

Comparison of the Trois Evêchés and Col de la Cayolle sub-basins indicate that the morphological barrier was eventually breached to connect the sub-basins and allow turbidity currents to spread out in a west and northwest direction (ibid.). The 1200 m thick sandstone succession is interpreted to represent a sand-rich submarine-fan system, comprising two distinct large-scale progradational sequences separated by a fine-grained unit indicating basin-wide starvation. The lower sequence is 800 m thick and represents an overall coarsening-upward trend, with lobe facies associations in the lower part and local channel-fill deposits in the upper part representing a proximal supra-fan setting (ibid.). The upper sequence is 400 m thick with
lobe deposition in the lower half, followed by a succession of pebbly sandstones and coarse-grained sandstones, interpreted as supra-fan deposits and large-scale channel fill deposits. As mentioned above, within the sandstone succession at Col de la Cayolle three distinct marker horizons have been recognised (Elliott et al. 1985, Ravenne et al. 1987), two of these marker horizons comprising matrix-supported conglomerates similar in composition to those described from the Trois Evêchés sub-basin. These two conglomerate horizons are separated by 100 m of sandstone succession and are located in the lower part of the succession. The third marker horizon comprises a fine-grained and thin-bedded 20 m thick unit which separates the two sequences described above, and is located in the upper half of the succession.

6.1.2.3b Relationship of Trois Evêchés and Col de la Cayolle sub-basins with southern sub-basins

A palaeogeographic link between the Col de la Cayolle and Trois Evêchés sub-basins was proposed by Elliott et al. (1985) based on the observation that both sandstone successions from the sub-basins described above show similarities in the conglomeratic debris flow units. Regional facies and palaeocurrent analysis also show a systematic down-current transition from the more proximal Col de la Cayolle succession of channelised and non-channelised sandstone bodies, to the distal Trois Evêchés succession, in the Trois Evêchés area, where the deposits are essentially non-channelled sandstone bodies (ibid.). A reconstruction of the Col de la Cayolle and Trois Evêchés sub-basins and the sub-basins in the southern part of the basin are presented on Figures 6.3a, b, and c which depict sediment dispersal patterns and interpreted infill patterns.

In this thesis, a distinction between the northern and southern sub-basins is made, with the divide between the northern and southern basins represented by an approximate east-west trending line that passes between the northern extent of the Grand Coyer sub-basin and the southern margin of the Trois Evêchés sub-basin close to the Montagne de Chalufy sandstone.
exposures. The positioning of this dividing line marks the start of the northwardly dipping palaeoslope identified at the base of the Grès d'Annot Formation from the Trois Evêchés sub-basin; south of this line no major sandstone onlaps towards the south were recorded. Direct stratigraphic correlations between the northern and southern sub-basins is impossible as no marker horizons between the two areas have been identified. Palaeocurrent indicators from the Grès d'Annot Formation throughout the basin indicate sediment flow was towards the north and northwest with sediments derived from a landmass at the southern margin of the basin. Petrographic studies utilising artificial and natural cathodoluminescence show that the sediments were derived from the Corsica-Sardinian massif which was located south of the present day basin (Ivaldi 1974).

The palaeo-reconstruction of the Tertiary Foreland Basin presented in Figures 6.3a), b), and c) depict the sediment fill histories of the various sub-basins at subsequent times. At the onset of sandstone deposition, the eastern and western basin-floor systems received sediment first, with deposition in the Petra Cava sub-basin in the eastern basin floor system and deposition in the St Antonin, Entrevaux and Annot sub-basins (Figure 6.3a). Turbidity currents which were of sufficiently high competence and capacity may have been able to bypass the western system and travel down to the deeper parts of the basin, i.e., the Trois Evêchés sub-basin. Sedimentation in the Col de la Cayolle and Trois Evêchés sub-basins was only able to commence once the Grès d'Annot Formation infilled the western system and prograded towards the intra-basinal break in slope (Figure 6.3b), the position of which is represented by the east-west line which divides the basin in to the northern and southern halves. The palaeo-reconstruction presented in Figure 6.3b suggests deposition in the northern part of the basin, but this is speculative. The relationship between the western basin-floor system with the northern part of the basin (Col de la Cayolle and Trois Evêchés sub-basins) is described below.

The precise nature of the relationship between the northern and southern halves of the basin cannot be properly understood until accurate age dating and correlation of the sandstone successions, using either magneto-stratigraphy or micropalaeontology, is utilised throughout
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the basin. A broad interpretation of the basin fill is presented below, although the exact timing of sedimentation in various sub-basins is not accurately known.

The southernmost sediments, from the St Antonin sub-basin, have been ascribed an age of latest Eocene to earliest Oligocene for the Lower Sandstone Member of the sandstone succession based on biostratigraphic analysis (Globigerina gortani zone), with a lower Oligocene age for the remainder of the succession (Bodelle 1971). This date for the base of the succession corresponds to the oldest parts of the Grès d'Annot Formation, and suggests that earliest sedimentation occurred at least in the southern part of the basin. Dating of the most northern outcrops of the Grès d'Annot Formation from the Trois Evêchés sub-basin at Dormillouse, as to date were not made. The western basin floor system is interpreted to have prograded northwards infilling the topography at St Antonin, Annot and Grand Coyer with sandstone packets downlapping towards the north in the Grand Coyer sub-basin. Sedimentation in the northern part of the Trois Evêchés sub-basin may have occurred once sedimentation in the Grand Coyer sub-basin commenced.

Once sedimentation reached the Grand Coyer sub-basin, subsequent turbidity currents would have encountered the 5° northerly-dipping palaeoslope developed in the marls. Turbidity currents travelling down this slope may have been in a state close to autosuspension and, therefore tend not to deposit all their load on the slope, but instead deposition probably occurred at the base of this slope. This phenomenon would explain why there is a consistent onlap of turbidites towards the south as opposed to sediments draping the slope. In essence, the palaeoslope acted as a transfer zone for turbidity currents, enabling them to reach the deeper parts of the basin located ~30 km northwards of the Grand Coyer sub-basin in the Dormillouse area. The Col de la Cayolle sub-basin received sediment once the western basin-floor system had prograded past the Grand Coyer sub-basin, with the Col de la Cayolle sub-basin supplying sediment to the Trois Evêchés sub-basin once the topographic barrier between the two was drowned with sediment. Once sediments in the Trois Evêchés sub-basin reached the top of the
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slope, the eastern and western basin-floor systems could then prograde across the northern parts of the basin.

6.2 Synthesis and models for sandstone architecture

In conclusion, the following features of the Grès d’Annot system are considered as characteristic of these deep-marine deposits and, therefore, they should be integrated into any generalised depositional model:

- Inherited seafloor topography exerted the primary control on the location and overall shape of depocentres for the accumulation of sand, i.e. the establishment of sub-basins.

- Syndepositional tectonics do not appear to have exerted a major control on sand deposition, but are recognised as having been important in increasing local seafloor gradients, leading to oversteepening and mass flow (e.g. sediment slides).

- The sedimentary structures associated with all grain sizes are consistent with turbidity currents being invoked as the principal sediment transport and depositional process, with relatively minor debris flows, sediment reworking (to isolated megaripple and dune bedforms) and slide processes. Assuming that individual beds are of basin wide extent, and using the present-day (preserved) sub-basin dimensions, typical sediment volumes associated with the sandy turbidity currents were in the range 0.03-0.12 km³ (for beds of 2 m and 0.5 m thick in the Annot sub-basin ca. 6 x 10 km).

- Water depths are very poorly constrained: there are no shallow-marine sedimentary structures; the faunas merely indicate upper bathyal water depths, and basin-wide considerations (topographic arguments) favour water depths of at least ca. 500 m.

- Sedimentation was controlled by abrupt, metronome-like, changes in the sediment flux to result in the vertical alternation of sand-prone (between 8-80 m thick) and mud-prone (between 1-17 m thick) intervals, typically of sub-basin-wide extent.
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- In general, the packets of coarser grained clastics show a high continuity relative to the dimensions of the sub-basins, i.e., over distances <10 km.
- Sand packets typically show amalgamation of medium-/very coarse-grained, thick-/very thick-bedded \( T_{ab} \) deposits. Large-scale erosional surfaces cutting down to 15 m over ca. 200 m laterally occur within the sand-rich packets.
- The mud packets are typically silt-rich, with isolated thin-/medium-bedded fine-/medium-grained sands.
- Erosion at the base of sand packets into the mud packets is up to 50 m over ca. 300 m laterally - a total of four examples being identified throughout the entire formation with only one showing both margins. Despite a channel-like geometry associated with such sand packets, no convincing levee and overbank environments could be recognised. These features are interpreted as having formed entirely by erosion and then passive infill.
- Sediment dispersal patterns, together with petrographic data, suggest that only one side of the basin, in the up-current direction (to the south) acted as a major sediment source. There are no unequivocal preserved submarine canyons associated with the basin (or sub-basins) and, therefore, the nature of the feeder system/s remains obscure. Locally, the remnants of contemporaneous fan-delta deposits suggest that much of the deep-marine deposits may have accumulated as fan-delta-to-slope basin/basin-slope systems.
- The Grès d’Annot Formation cannot be interpreted as a classic radial submarine fan system (e.g., non-radial palaeocurrent pattern, lack of convincing channel-levee-overbank complexes, and the absence of any logical distribution of channel and non-channel environments), but rather as a basin-slope and slope-basin system in which sub-basins may have been connected by relatively narrow corridors.

Two possible end-member options exist for explaining the pattern of sand accumulation: (1) a terraced and compartmentalised slope in which sub-basins (depositional units) were disconnected and separated by submarine ridges over which turbidity currents spilled (overspill and flow-stripping); (2) a terraced and compartmentalised slope in which sub-basins were connected by relatively narrow corridors. In the latter scenario, sediment gravity flows would still experience...
some overspill and confinement during travel between sub-basins. Although key marker beds were identified within individual sub-basins, this proved impossible between the various sub-basins. The occurrence of well-defined onlap surfaces (up to at least 400 m high over lateral distance of ca. 5 km) showing onlap of the sands onto the marls in proximal-distal and axial-lateral directions, up- and down-current, support a reconstruction in which the depocentres must have been at least somewhat separated. However, the thickness of the sub-basin fills (up to ca. 770 m) suggests that depocentres remained stable for long time intervals and, by inference, that their sediment supply routes probably remained relatively stable. Circumstantial evidence leads to option (2) as the favoured candidate.

A schematic summary model which can account for the sandstone body architecture observed within the Grès d'Annot Formation is shown in Figures 6.4a & 4b (see also Table 6.1 and Table 6.2). These models include the overall geometry and internal characteristics of the sandstone, and mudstone/sandstone packets, their thickness variation, onlap characteristics of sandstone packets, erosional features and proximal-distal/lateral facies changes observed from the Grès d'Annot Formation in the various sub-basins.

The type of feeder system and the bulk mean sediment grain size (sediment calibre) play a major role in the nature of sediment transport and depositional processes, together with the pattern of sediments deposited. The three main types of feeder system include: (i) submarine-fan point source; (ii) multiple sourced ramps, and (iii) slope apron linear source (Pickering 1982, Reading & Richards 1994). Pickering (1982), Stow (1984) and Reading (1991) classified deep-sea depositional systems by sediment calibre and feeder system type; sediment calibre was divided up into mud, sand and gravel dominated systems. In confined basins, which are to some extent comparable to at least parts of the Grès d'Annot Formation system, the radial expansion of a decelerating turbidity currents is suppressed but not inhibited and, hence, produce relatively elongate sedimentary bodies - a feature of the sub-basins within this system.
Figure 6.4a. Schematic summary of sandstone architectural features from this study.
Figure 6.4b. Schematic summary of sandstone architectural features observed in sandstone and sandstone-mudstone packets; in a lateral section.
Figure 6.5. Schematic sequence for Monterey Fan growth. Diagrams A to E show the control of topographic features in fan development (from Normark 1985).
On a cautionary note, modern submarine fans may show complex within-system compartmentalisation of deposits and if they were preserved as ancient (fragmented) deposits could lead to the erroneous conclusion that the remnant sections represent discrete sub-basins that were not part of a larger scale submarine fan. Normark (1985), for example, discussed how large-scale morphological features in the basin control fan development and growth, citing examples from the Monterey and Navy Fans. In the Monterey Fan, five stages of fan growth have been identified which, in part, appear to have been controlled by topography (Normark 1985) (Figure 6.5).

The following sections describe aspects of other ancient deep-marine systems that show features in common with the Grès d'Annot Formation.

6.3 Comparative studies

At present there are few documented modern examples of sand-rich confined deep-marine turbidite systems described in the literature. The examples of sand-rich confined deep-marine turbidite systems described below are taken from ancient systems which either occur as outcrops, or as subsurface systems from areas where petroleum exploration has produced sufficient data to delineate a confined basin setting.

6.3.1 Tertiary Palaeogene of the Northern and Central North Sea

The Tertiary Palaeogene of the Northern and Central North Sea have provided successful hydrocarbon reservoirs since exploration in this area commenced in 1967 (Hartog Jager et al. 1993). The Central and Northern North Sea areas comprise platforms of continental crust dissected by grabens which were filled by sediments (Figure 6.6a), the main grabens include the North and South Viking Grabens, the Central Graben divided into the West and East Central
Figure 6.6a. Structural map of the UK Central and Northern North Sea (from Knox and Cordey 1992).
Figure 6.6b: Lithostratigraphic nomenclature scheme of the UK Central and Northern North Sea (from Knox and Corney 1992)
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Graben at its northern extent, and the Inner and Outer Moray Firth and Shetland Basins (Figure 6.6a). The grabens in the Central and Northern North Sea areas originated from extensional tectonism during rifting of the North Atlantic during Jurassic times (Coward 1990). Sediment was derived from the northwest and west, where an active uplift of 1-3 km of the UK mainland occurred, with a tilt towards southeast (Hartog Jager et al. 1993).

The Palaeogene of the Central and Northern North Sea comprise sediments of Palaeocene to mid-Miocene in age, these are divided into formations which can be correlated across the region (Figure 6.6b). These formations essentially comprise shales with interbedded sandstone intervals, the sandstone intervals contain facies and facies-associations characteristic of deep-marine deposits, generally forming submarine fans in confined settings (Knox & Cordey 1992).

Examples of Palaeogene sandstone reservoirs which are laterally confined with similar facies characteristics from the Grès d'Annot Formation of Southeast France are presented below. Examples are taken from the Everest trend and Fleming area of the East Central graben, and from the Gryphon Oil Field in the Beryl Embayment of the northern part of the South Viking graben.

6.3.1.1 Gryphon Oil Field

The Gryphon Oil Field is located in Block 9/18b on the western edge of the Crawford Ridge in the southern part of the Beryl Embayment, located in the northern part of the South Viking Graben (Figure 6.7a & b). The underlying structure of the Gryphon Field comprises NNE-SSW trending graben-margin fault system cross cut by a NW-SE Hercynian trend, which has been a dominant control on all subsequent sedimentation and the most significant factor in the interpretation of the area (Newman et al. 1993). Two sandstone fan systems have been identified, belonging to the Balder Formation and Frigg Formation, respectively. These two
Figure 6.7a. Location of the Gryphon oil field within the South Viking Graben (from Newman et al. 1993).
formations have been interpreted using the systems tract nomenclature of Vail et al. (1990) to be lowstand systems tracts, where a lowstand systems tract is a linkage of contemporaneous depositional systems that occur during a relative lowstand in sea level (Mitchum 1977). The Balder Formation is interpreted as representing a lowstand fan and the Frigg Formation a lowstand wedge/channel levee complex (Newman et al. 1993).

Sandstones from the Balder Formation were mainly deposited from high-density turbidity currents which carried sand across the Beryl embayment and deposited them against the Crawford Ridge as a lowstand fan (Figure 6.8a). The topography of the Crawford Ridge fault-system controlled transportation and deposition of these sand lobes which attain thicknesses up to ~120 m, comprising amalgamated sandstones (Figure 6.8b, Figure 6.9). Subsequent deposition of sandstone packets within the Balder Formation was controlled by the original topography of the area and this relief was developed over earlier sand bodies after initial compaction of claystones with deposition along the flanks of older sandstone bodies (Newman et al. 1993).

The analogy drawn from the Balder Formation, within the Beryl Embayment, with the Grès d'Annot Formation is made by Newman et al. (1993), with the basin floor topography in the Annot sub-basin controlling rapid lateral pinch-out of sandstone bodies. Newman et al. (1993) describe the effect that differential compaction of claystones around sandstone bodies have in controlling deposition of subsequent sandstone bodies; in the Grès d'Annot Formation, the underlying marls were relatively more lithified than mudstones of the Balder Formation and therefore less likely to compact to any great extent. The interbedded mudstone intervals within the Grès d'Annot Formation are relatively thin and comprise sandstones as well as mudstones, therefore compaction was relatively small and did not play an important part in controlling subsequent sandstone bodies as with the Balder Formation. During deposition of the Frigg Formation the Crawford Ridge formed a less significant bathymetric feature as the Balder Formation previously infilled the topography (Figure 6.8b)(Newman et al. 1993). Sediment input to the Frigg Formation was clay rich as relative sea-level began to rise, resulting in
Figure 6.7b. Stratigraphy of the Gryphon oil field from Triassic to Recent; reservoir zonation of the Gryphon oil field is shown on the right hand side (from Newman et al. 1993).
Figure 6.8. a) Reconstruction of latest Paleocene-earliest Eocene depositional environment-Block 9/18b area. b) Seismic line showing lowstand systems tracts-Block 9/18b, note pinchout and lateral thinning towards NW of lowstand fan and lowstand wedge deposits respectively (from Newman et al. 1993).
Basement faults controlled the topography of the sea floor and guided fluidised sand flows into place.

During the Thulean volcanic episode aeolian transported tuffs were deposited as a blanket covering the area.

Subsequent sands were deposited on the flanks of the older bodies.

Compaction on the flanks of the sand bodies lead to widespread failure and the formation of slump troughs. Injection and deformation occurred within both the clay and sand sections.

Thick sands were deposited in the slump hollows.

After compaction and burial to 6000', sands attain full mounded character within the overall fan mound.

Oil-gas condensates
Dry well

Figure 6.9. Schematic evolution of the Gryphon oil field, no scale implied. Note lateral stacking of sandstone bodies (from Newman et al. 1993).
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channel/levee-complex development and deposition of thin, discontinuous sandstones and hence a non-viable reservoir (ibid.).

6.3.1.2 Everest and Fleming areas

Two examples of lateral confinement and sandstone pinch out against basin structural highs are taken from the Everest and Fleming areas described by O'Connor & Walker (1993). The Fleming and Everest areas, occur within the Everest trend against the western flank of the Jaeren High, located at the southern margin of the South Viking Graben (Figure 6.10a and b). Clastic sediment supply to this area was from the north and west in the platform areas of the East Shetlands and Scottish Highlands (Figure 6.10b) (O'Connor & Walker 1993). The Everest trend Palaeocene stratigraphy comprise the Ekofisk, Maureen, Andrew, Heimdal, Lista, and Sele Formations, with the Forties Sandstone Member included in the Sele Formation (ibid.).

The first example is from the Maureen Formation in the Fleming area, where four major sandstone lobes are identified (Figure 6.11a and b). The Maureen Formation sandstone lobes are divided into an upper and lower part, with the reservoir facies in the lower part comprising massive amalgamated sandstones interbedded with claystones with some cross-stratification of sandstones. The upper part comprises sandstones with an increased proportion of cross-stratification with interbedded claystones, the cross-stratification is interpreted to represent an increase in traction-current deposition and reworking (O'Connor & Walker 1993). The sandstone lobes in the Fleming area are 8-12 km wide and are partially offset (Figure 6.11b), the lobes have an elongate sheet-like geometry which thins rapidly towards the basin margin against the Jaeren High, producing an asymmetric lenticular form in the east (ibid.). Away from this margin a slab-like geometry is maintained towards west in the centre of the basin. The Jaeren High acted as a structural baffle, indicated by the rapid thinning of turbidite deposits, some pinch-outs occur over distances <1 km, as indicated from observations from a borehole located ~1 km from a borehole which penetrated the lobes (O'Connor & Walker 1993).
Figure 6.10. a) Regional setting with structure of the Central and South Viking Grabens showing location of Everest, Fleming area and Sleipner Øst fields. b) Palaeogeography of the Montrose Group; arrows indicate major transport routes of submarine fan sequences (from O'Connor and Walker 1993).
Figure 6.11. a) Fleming area Paleocene cross-section showing rapid up-dip pinch-out into Lista shales. North-south closure is provided by structural roll-over. Note stratigraphic pinch-outs against the Jaeren High towards east. b) Block diagram of the Fleming area showing rapid dumping of Maureen Fan sandstones at the base of the high. (from O'Connor and Walker 1993).
Figure 6.12 Upper and Lower Forties lobe distinguished by extinction of diverse agglutinants and gamma spike marker as chronostratigraphic horizons. Note lateral pinch-out of lower and upper lobes towards right of figure (from O'Connor and Walker 1993).
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The Forties Sandstone Member is absent in the Fleming area, but in the Everest area two lobate sandstone bodies are identified, which pinch-out laterally towards east, with the upper lobe pinching out farther east (Figure 6.12) (O'Connor & Walker 1993). The sandstone facies of these lobes include amalgamated sandstones, which are less abundant towards east, also present are structureless fine- to medium-grained sandstones with interbedded mudstones, these sandstones show typical Bouma Ta Tb and Tc divisions. Sediment slides are common and involve heterolithic sandstones, siltstones and mudstones (ibid.).

Both of these examples from the Everest trend show similarities with sandstone packet terminations from the Grès d'Annot Formation in the Annot and Grand Coyer sub-basins. In the Grand Coyer sub-basin lateral pinch-out of sandstone packets is observed, individual sandstone packets can be traced from the centre of the sub-basin towards the marginal slope on the southeast side of the sub-basin (Enclosure 4.2, Section 4). At the marginal slope on the southeast side of the Grand Coyer sub-basin, individual sandstone packets, which in the centre of the basin essentially comprise amalgamated sandstones, become much thinner and non-amalgamated. Within the Forties Sandstone Member in the Everest area described above, a similar situation developed where amalgamated sandstones are less common towards the east, i.e., against the marginal palaeo-high formed by the underlying Jaeren High. In the Annot sub-basin, sandstone packets onlap against an easterly dipping palaeoslope, but no such thinning of sandstone packets is observed, they simply onlap against the slope developed in the marls. This palaeoslope in the Annot sub-basin dipped at ~12°, while the palaeoslope developed in the Grand Coyer sub-basin is calculated to have dipped at ~8° during sandstone deposition. The angle of dip of marginal palaeo-slopes is interpreted to be an important factor in controlling the rate of lateral pinch-out of sandstone lobes. In essence, the shallower the palaeoslope, the greater the distance of lateral pinch-out and the greater the likelihood that turbidity currents could surmount and deposit sediment upon the palaeoslope.
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6.3.2 California Borderlands

The Stevens Sandstone is a deep-marine turbidite sandstone system of Miocene age, located at the southern margin of the San Joaquin Valley, California (Figure 6.13). Sediment deposition was in a <30 km wide basin, with deposition essentially from sediment gravity flows sourced from the east and south, and which entered the basin via channels (canyons) eroded into slopes. The sediment gravity flows moved through these slope channels to form a series of laterally coalescing submarine fans which prograded basinward (Scott & Tillman 1981). The sandstone lobes essentially comprise beds 2.5 cm to 0.5-0.75 m thick, with beds either separated by shale or amalgamated, with thicker sandstone beds and bodies amalgamated. Four discrete cycles of sandstone deposition are identified along the eastern margin of the basin, comprising middle- and inner-fan distributary channel systems (cf. Mutti & Ricci Lucchi 1972). Turbidites flowed out of the distributary channel systems of the mid-fan environment on the east flank of the basin westwards into structurally deformed portions of the basin (Figure 6.14) (Scott & Tillman 1981).

In the western and central parts of the basin, during upper Miocene times, there were two persistent depositional patterns which were structurally controlled (Scott & Tillman 1981). The first depositional pattern was formed from turbidity currents which were either diverted or blocked by topographic features on the basin floor, these topographic highs comprise anticlinal highs normal to sediment transport direction. Sediment deposits accumulated to form essentially lobe-shaped sand bodies, with lobes coalescing and stacking vertically along the flanks of these sea-floor highs. Scott & Tillman (1981) produced an onlap model to explain these features (Figure 6.15a). In the western portion of the basin, deposition was confined by basin topography to the synclinal lows immediately after entering the basin, and a confinement model was used to explain the external channel-like morphology of the sandstone deposits, although they do not necessarily exhibit facies associations or the structureless character of channels commonly ascribed to channels in fan models (Figure 6.15b) (Scott & Tillman 1981).
Figure 6.13 Location map of the Great Valley of California, area of Stevens Sandstone production at the southern end of the San Joaquin Valley is marked as study area (from Scott and Tillman 1981, after MacPherson 1971).
Figure 6.14. Upper Miocene palaeogeographic reconstruction of the San Joaquin Basin, showing multiple sediment source areas along the margins of the basin (from Webb 1977).
Figure 6.15. a) "Onlap model" for the Stevens Sandstone, sandbodies onlap and stack vertically against an anticline. Arrow indicates sediment dispersal. b) "Confinement model" for the Stevens Sandstone, sandbodies are deposited in synclinal lows between adjacent anticlines, external sandstone body geometry is "channel-like" (from Scott and Tillman 1981).
Chapter 6: Grès d'Annot Formation summary and analogues

The onlap and confinement models put forward by Scott & Tillman (1981) to explain depositional patterns in the Stevens Sandstone can be applied to sandstones of the Grès d'Annot Formation. In the Annot sub-basin, both the confinement and onlap model apply to the sediments of the basin. The lower part of the sandstone succession at Annot is interpreted to have been confined at its northern margin by a palaeoslope which dipped towards the south, both the lower and upper part of the succession was laterally confined at least on the western margin of the sub-basin and was probably confined on the eastern margin of the sub-basin although no data is present to support such a contention. Sandstones from the Grand Coyer, Peïra Cava, and Contes areas also show a lateral confinement, which prevented the formation of radial submarine fan lobes, instead sandstone lobes are elongate as a result of the influence of the sub-basin geometry.

6.3.3 Cengio turbidite system, Italy

The Cengio Member is an Upper Oligocene-Lower Miocene small turbidite system deposited within a small, 6.4 km long by 4.8 km wide, southwest-northeast trending, fault-controlled submarine depression as part of the Piedmont Basin of Northwestern Italy (Figures 6.16a and b). The Cengio Member comprises submarine fan deposits represented by depositional sandstone lobes, with interbedded mudstone units which attains a thickness of 170 m, supplied from a fixed sediment source (Cazzola et al. 1985). The Cengio Member comprises two distinct sandstone units, with the lower sandstone unit poorly exposed. The upper sandstone unit is 110 m thick, and comprises eight sandstone bodies 5-25 m thick, comprising thick bedded and structureless to, medium to-granular-grade beds, amalgamated or separated by mudstone, conglomeratic sandstones, and thin- to medium-bedded medium-grained, poorly graded beds (Figure 6.17). The sandstone bodies were deposited by southerly derived turbidity currents, deflected towards the northeast against a northwest basin margin slope which dipped between 2.5-5° towards the southeast, occurring within the SW-NE trending submarine depression.
Figure 6.16. a) Outcrop area of the Tertiary Piedmont Basin and location of the study area and Cengio systems. b) Diagramatic stratigraphic cross-section of the stratigraphically lower part of the Tertiary Piedmont sequence in the study area. c) Outcrop area of the Cengio system, showing palaeocurrent directions and location of measured sections (from Cazzola et al. 198).
Figure 6.17. Detailed stratigraphic cross-section of the Cengio sandstone lobes, location of measured sections is shown in Figure 6.16c. (from Cazzola et al. 198.).

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(Figure 6.16c). Sandstone bodies, within the upper sandstone unit onlap against this northwesterly margin (Figure 6.17).

The Cengio Member shows similar characteristics to the Grès d'Annot Formation from the Annot and Grand Coyer sub-basins, which comprise distinct sandstone bodies separated by mudstone/sandstone bodies, though these packets are 8-80 m thick and probably reflects larger-volume turbidity currents. The sub-basin margin slopes from Annot and Grand Coyer are between 8°-12°. Palaeocurrents from the Annot sub-basin indicate palaeoflow towards the north, the proximal equivalent deposits at St Antonin indicate palaeoflow towards the northwest. This discrepancy in palaeocurrents between both areas is interpreted to be a result of the confining topography on the western basin margin at Annot, deflecting northwestward directed flows from St Antonin. This phenomenon is also observed in the Cengio Member, where southerly derived turbidity currents were deflected towards the northeast by a southeasterly dipping palaeoslope.

6.4 Concluding comment

The Tertiary Grès d'Annot Formation of Southeast France is a sand-rich deep-marine system deposited in a basin with a complex basin-floor topography subdivided into local topographic highs and lows. The topographic lows acted as sites for the preferential sediment deposition from sediment gravity flows, mainly turbidity currents, debris flows and slide deposits. The deep-marine system appears to have lacked integrated typical deep-marine fan-like features such as: distinct environmental segregation into canyon, channel/channel-levee-overbank and onto fan deposits. This lack of typical submarine fan features may be due to the type of feeder system supplying sediment to the deep-marine basin. For the Grès d'Annot Formation no point source for sediment supply is observed, and a multiple sourced feeder system is inferred, akin to that described from the Eocene Tyee Formation, Western Oregon (Heller & Dickinson 1985). In sand-rich systems which have a point source and no complex basin topography, typical
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submarine fan features develop, such as channel-levee overbank complexes, and radial fan-like development, such as the modern Redondo fan of Southern California (Haner 1971).
Chapter 7: Vitrinite reflectance and sandstone petrology

7.1 Vitrinite reflectance

7.1.1 Introduction

Vitrinite forms one of the three major maceral groups of coalified organic matter, together with exinite and inertinite. Macerals are essentially coalified plant remains whose form and/or structure is still preserved in the bituminous-coal stage, and are also in part degradation products of plant origins (Stach et al. 1975).

The maceral group vitrinite originates from the mouldering and peatification of the lignin and cellulose of plant walls, leading to the formation of compounds which contain the elements carbon, oxygen, hydrogen and nitrogen. The relative proportions of these elements in vitrinite denote the rank of the coal, with the rank increasing as the carbon content increases and the hydrogen content decreasing (Figure 7.1). A simple indicator of the rank of coal is given by the reflectivity of vitrinite, which is also dependent on the chemical composition of the coal. In general, the reflectivity of vitrinite increases with coal rank. The reflectivity of vitrinite is controlled by the depth of burial of the vitrinite and the heat flow in the sedimentary basin. An increase in the depth of burial is usually accompanied by an increase in heat, which will cause an increase in the reflectivity of vitrinite. An increase in pressure (related to folding and faulting of sediments) will also cause the reflectivity of vitrinite to increase. Time is an important variable on the coalification process where temperatures are high enough, and is outlined by

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Figure 7.1. Relationship between vitrinite reflectance, different chemical rank, and coal rank (from Stach et al. 1976).
Teichmüller & Teichmüller (1966). They describe two boreholes from coal-bearing sediments from the upper Miocene of the Gulf Coast of Louisiana which was buried to a depth of 5440 m, and reached temperatures of 140°C for 17 Ma, and another example from the Carboniferous of NW Germany buried to a similar depth and temperature of ~5000 m and ~140°C for 270 Ma. High-volatile coals are found in the first case, and low-volatile coals in the second, indicating the importance of time in forming low-volatile coals. With increasing temperature and time, the reflectance of vitrinite in oil increases from values of about 0.2%Ro in near surface sediments to values greater than 2.5%Ro in deeply buried anthracites.

7.1.2 Occurrence of organic material in turbidites

Generally, large-scale deposits of coals originate from sedimentary environments associated with swamps and river deltas; the deposits are formed in situ, although reworked coal deposits can occur (Stach et al. 1975). Subsidence permit preservation of vast thicknesses of coal-bearing sediments (Stach et al. 1975).

In the deep-marine environment, deposits of vitrinite within turbidite beds must have originated as terrestrial material. Samples of vitrinite from the Grès d’Annot Formation indicate an origin in temperate climates, as indicated from large samples of organic material up to 4 cm by 2 cm in size which record growth rings. For organic matter, in the form of woody material, to reach the deep-marine environment it would be in a waterlogged state and non-buoyant, thus allowing it to be transported as bedload by turbidity currents/debris flows.

Vitrinite in deep-marine sediments has been recorded from the Upper Eocene-Lower Oligocene Taveyanne Sandstones in the Helvetic zone of the Swiss Alps (Kisch 1980), although it is not a common component of deep-marine sediments. Organic debris in the Grès d’Annot Formation generally occurs in the Bouma Tb/Tc divisions of turbidites forming lenses up to 1 cm thick and 15 cm long (Figure 7.2a), or rarely within debris flow deposits, as large fragments of fossilised
Figure 7.2 a and b. a) Lens of coalified material within a turbidite bed, Grand Coyer. b) Large, 2 m long, coalified tree trunk within an andesite debris flow, St Antonin.
Chapter 7: Vitrinite reflectance and sandstone petrology

branches up to 2 m long (Figure 7.2b). The organic debris may also occur as dispersed fine particulate matter within the fine-grained sandstone Bouma Td division.

7.1.3 Analytical techniques

Vitrinite reflectance measurements (%Ro), were made on samples of vitrinite, samples were crushed by hand and sieved between 0.5 mm and 1 mm mesh. The 0.5-1 mm fraction was retained and set into a two-part epoxy resin block. The resin blocks were then polished with a 400# abrasive and then sanded down to give an initial surface using wet and dry 1200# paper. The resin blocks were then polished on a lead-lap lined with a fine-grained polishing paper with a 15 μm diamond paste and using liquid paraffin lubricant. The same procedure, on the lead lap, was carried out for 6 μm and 1 μm diamond paste to produce a polished block.

Measurements of vitrinite reflectance were made on a Nikon reflected-light polarising microscope, with a green 546-nm-wavelength fibre optic light source, and measured using a Nikon P-1 photometer interfaced to an IBM-personal computer or compatible. An oil immersion objective lens of x40 magnification and oil of refractive index 1.518 was used to measure the reflectivity of the vitrinite samples. The measuring diameter of the photometer was 5 μm. The photometer was calibrated using either Yttrium-alumina garnet or Spinel as standards whose reflectance values, using green 546-nm-wavelength light and immersion oil of refractive index 1.518, are 0.917 and 0.413 %Ro respectively. All measurements of %Ro were made on random samples, with rotation of the stage to obtain the maximum reflectance. On average, thirty points were measured for each sample block and the arithmetic mean calculated (Table 7.1).
<table>
<thead>
<tr>
<th>Location/Sample No.</th>
<th>%Ro mean</th>
<th>%Ro min</th>
<th>%Ro max</th>
<th>Std. dev.</th>
<th>Points-n</th>
</tr>
</thead>
<tbody>
<tr>
<td>St Antonin VTR 200</td>
<td>1.206</td>
<td>1.096</td>
<td>1.312</td>
<td>0.067</td>
<td>30</td>
</tr>
<tr>
<td>Entrevaux VTR 201</td>
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<td>0.403</td>
<td>0.660</td>
<td>0.015</td>
<td>30</td>
</tr>
<tr>
<td>VTR 202</td>
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<td>0.423</td>
<td>0.521</td>
<td>0.024</td>
<td>30</td>
</tr>
<tr>
<td>VTR 203</td>
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<td>0.364</td>
<td>0.471</td>
<td>0.024</td>
<td>30</td>
</tr>
<tr>
<td>Rouaine VTR 208</td>
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<td>0.288</td>
<td>0.398</td>
<td>0.023</td>
<td>30</td>
</tr>
<tr>
<td>VTR 209</td>
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<td>0.173</td>
<td>0.366</td>
<td>0.062</td>
<td>30</td>
</tr>
<tr>
<td>Annot VTR 217</td>
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<td>0.488</td>
<td>0.603</td>
<td>0.025</td>
<td>30</td>
</tr>
<tr>
<td>VTR 218</td>
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<td>0.244</td>
<td>0.475</td>
<td>0.046</td>
<td>30</td>
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<tr>
<td>VTR 219</td>
<td>0.484</td>
<td>0.405</td>
<td>0.563</td>
<td>0.037</td>
<td>30</td>
</tr>
<tr>
<td>VTR 220</td>
<td>0.412</td>
<td>0.361</td>
<td>0.482</td>
<td>0.028</td>
<td>30</td>
</tr>
<tr>
<td>VTR 213</td>
<td>0.525</td>
<td>0.445</td>
<td>0.595</td>
<td>0.034</td>
<td>30</td>
</tr>
<tr>
<td>VTR 214</td>
<td>0.498</td>
<td>0.447</td>
<td>0.586</td>
<td>0.031</td>
<td>30</td>
</tr>
<tr>
<td>VTR 216</td>
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<td>0.399</td>
<td>0.564</td>
<td>0.039</td>
<td>30</td>
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<tr>
<td>VTR 210</td>
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<td>0.458</td>
<td>0.626</td>
<td>0.047</td>
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<tr>
<td>VTR 211</td>
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<tr>
<td>VTR 212</td>
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<td>0.389</td>
<td>0.522</td>
<td>0.035</td>
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<td>Contes VTR 204</td>
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</tr>
<tr>
<td>VTR 205</td>
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<td>0.368</td>
<td>0.513</td>
<td>0.029</td>
<td>30</td>
</tr>
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<td>0.381</td>
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<td>0.089</td>
<td>200</td>
</tr>
<tr>
<td>VTR 102</td>
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<td>0.330</td>
<td>0.579</td>
<td>0.043</td>
<td>200</td>
</tr>
<tr>
<td>VTR 103</td>
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<td>0.496</td>
<td>1.024</td>
<td>0.091</td>
<td>200</td>
</tr>
<tr>
<td>VTR 104</td>
<td>0.636</td>
<td>0.525</td>
<td>0.750</td>
<td>0.039</td>
<td>200</td>
</tr>
<tr>
<td>VTR 105</td>
<td>1.320</td>
<td>0.549</td>
<td>1.567</td>
<td>0.125</td>
<td>200</td>
</tr>
<tr>
<td>Grand Coyer VTR 207</td>
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<td>0.932</td>
<td>1.252</td>
<td>0.061</td>
<td>30</td>
</tr>
<tr>
<td>VTR 221</td>
<td>0.761</td>
<td>0.702</td>
<td>0.856</td>
<td>0.038</td>
<td>30</td>
</tr>
<tr>
<td>Chalufy VTR 222</td>
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<td>0.265</td>
<td>0.563</td>
<td>0.085</td>
<td>30</td>
</tr>
<tr>
<td>VTR 223</td>
<td>0.453</td>
<td>0.346</td>
<td>0.530</td>
<td>0.041</td>
<td>30</td>
</tr>
<tr>
<td>Col Cayolle VTR 215</td>
<td>2.396</td>
<td>2.093</td>
<td>2.649</td>
<td>0.116</td>
<td>30</td>
</tr>
</tbody>
</table>

Table 7.1 Vitrinite reflectance data (%Ro), from the Tertiary foreland basin of southeast France.

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7.1.4 Results

The results obtained from the study of vitrinite reflectance are presented in Table 7.1 and Figures 7.4a and b: the location of vitrinite samples are shown in Figure 7.3. The majority of data obtained from samples within the foreland basin have mean maximum-reflectance values between 0.4 %Ro and 0.6 %Ro. The minimum and maximum reflectance values are 0.273 %Ro and 2.396 %Ro. Spatial variations in vitrinite reflectance from the various areas studied are described below.

An accurate calculation of depth of burial for the Grès d'Annot Formation is not possible because the thermal and burial histories of the foreland basin are not known but general inferences about the burial of the sediments can be made, and are discussed in the following sections.

7.1.4(a) Contes- Peyra Cava

There is a marked increase from south to north in %Ro in the Contes and Peñra Cava sub-basins. Reflectance values at Contes are between 0.439 %Ro and 0.491 %Ro, whereas in the southern part of the Peñra Cava sub-basin vitrinite reflectance values range from 0.478 %Ro to 0.786 %Ro. In the northern part of the Peñra Cava sub-basin, sample VTR 105 has a reflectance of 1.320 %Ro. The markedly high reflectance value in the north of the Peñra Cava sub-basin is interpreted to have resulted from deformation and associated elevated pressure and temperature conditions at the northern margin and due to the uplift and unroofing of the Argentra-Mercantour Massif and folding of the overlying sedimentary cover.

Mudstones from the northern Peñra Cava outcrops are more lithified than those located towards the south of the sub-basin, which supports the inference of either higher pressures or temperatures in the northern part of the sub-basin. The difference in the reflectance values
Figure 7.3. Geological map of the Alpes Maritimes and Haute Provence regions of southeast France, showing the location of vitrinite samples from within the foreland basin.

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Figure 7.4a and b. a) Histogram plot of vitrinite reflectance data from the Grès d'Annot Formation. b) Percentile plot of %Ro vitrinite reflectance for data from the Grès d'Annot Formation; note the abundance of data points between 0.4 and 0.6 %Ro.

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between Contes and south of Peña Cava may also indicate northwards increase in pressure, or possibly reflect the relative stratigraphic positions of the Contes vitrinite samples in relation to the southern Peña Cava vitrinite samples. Vitrinite samples from Contes are estimated to be 400 m stratigraphically higher than samples from southern Peña Cava based on comparisons of the preserved sandstone successions from these two sub-basins (Section 5).

7.1.4(b) St Antonin

The St Antonin Grès d’Annot Formation succession is preserved in an east-west trending syncline overturned towards the south; only one vitrinite sample was obtained from this sub-basin. Vitrinite sample VTR 200 from St Antonin has a reflectance of 1.206 %Ro, which in comparison to vitrinite reflectance values from Entrevaux (0.417-0.472%Ro), Rouaine (0.273-0.319%Ro) and Annot (0.403-0.550%Ro) is relatively high. The sample was taken from a large coalified branch ~1 m long, located within an andesite breccia (Figure 7.2b) located ~ 500 m from the base of the sandstone succession. One explanation for the high reflectivity of this sample is that it occurs within an andesite breccia, possibly indicating that the woody material may have been partly carbonised prior to deposition within the breccia by lava flows and hot ash associated with the volcanic activity which produced the andesite breccia.

7.1.4(c) Entrevaux-Rouaine-Annot

Vitrinite samples from Entrevaux have reflectance values ranging from 0.417 %Ro to 0.472 %Ro (Table 7.1), with samples preserved in a steeply dipping limb of a fold. In contrast to the vitrinite sample from St Antonin (which is preserved in an overturned syncline and has a high reflectance), the Entrevaux vitrinite samples are much lower in reflectance. Folding as a cause of the different %Ro can therefore be ruled out as the cause of the high reflectance value for the St Antonin sample.
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Vitrinite samples from Rouaine have low values (0.273 %Ro and 0.319 %Ro), interpreted to be the result of surface weathering producing a lower reflectance value. Vitrinite samples from the Annot sub-basin range from 0.403 %Ro to 0.550 %Ro. Sample VTR 217 was taken from the Calcaires Nummulitiques Formation at the base of the sandstone succession near to the town of Braux, and sample VTR 218 from a sandstone bed within the Marnes Bleues Formation located approximately 150 m above sample VTR 217. The remainder of the samples are taken from within the Grès d'Annot Formation succession from various locations in the sub-basin, and show no major trend (Table 7.1).

7.1.4(d) Grand Coyer-Chalufy-Col de la Cayolle

Two vitrinite samples were obtained from the Grand Coyer sub-basin, VTR 207 and VTR 221, with reflectance values of 1.102 %Ro and 0.761 %Ro, respectively. Sample VTR 207 was taken from close to the base of the Grand Coyer sandstone succession in the axial part of the sub-basin, and sample VTR 221 was taken from the southwestern margin of the sub-basin.

The sandstone succession at Grand Coyer thins as it approaches the southwestern margin of the sub-basin against a palaeoslope and, therefore, any vitrinite sample present above the palaeoslope would have undergone less overburden pressure than the sample from within the axis of the basin located towards the base of the sandstone succession. This could explain why the reflectance values are markedly different.

Vitrinite samples from the Chalufy section yield reflectance values of 0.443 %Ro and 0.454 %Ro, and these values are much lower than the Grand Coyer values but are similar in the values of %Ro obtained from samples at Annot (Table 7.1).
Only one sample was obtained from the Col de la Cayolle sub-basin, located approximately 600 m from the base of the sandstone succession. Sample VTR 215 has a reflectance value of 2.396 %Ro which would place it in the rank of semi-anthracite. The high reflectance of the sample is interpreted to have formed by an increase in pressure and temperature resulting from the emplacement of the overlying Nappes de Embrunais-Ubaye above during Palaeogene times.

### 7.1.5 Conclusions

Measurements of vitrinite reflectance of samples from the Tertiary foreland basin succession vary markedly over short distances, suggesting a complex burial history of the basinal sediments. A precise calculation of the depth of burial for the sediments based on vitrinite reflectance requires a detailed knowledge of the burial history and thermal history of the basin, something that is beyond the scope of this cursory study.

### 7.2 Sandstone petrology

Although a detailed account of the Grès d’Annot Formation sandstone petrology was not a principal part of this research, some selected thin sections were described and field observations on clast petrography made. Observations on sandstone petrography made by previous workers is also summarised below.

#### 7.2.1 Previous work

One of the first major studies of the Grès d’Annot Formation petrology was carried out Gubler (1955), who described the mineralogy of the arkosic sandstones and conglomerate clasts, especially from St Antonin. Within the conglomerates four major clast types were identified: (1)
granite clasts, of which three types were distinguished (microperthite/biotite/microcline granite, microgranite with micropegmatite, and aplites with biotite); (2) two types of rhyolite clasts (devitrified flow-banded biotite rhyolite, and rhyolite with layers of bipyramidal quartz); (3) metamorphic rocks (with mica schists, two mica gneiss); and (4) andesites (the latter being particularly abundant at St Antonin).

Within the arkoses, Gubler (1955) recorded angular clasts of feldspar including: microcline, microperthite, and orthoclase. Lithic rock fragments include rhyolite with biotite, quartzites, micropegmatites, microgranites, and pelitic schists. Within the matrix of the sandstones, muscovite mica is a common component, together with calcite cement in some samples (ibid.).

Stanley (1963) studied the vertical petrographic changes in the Grès d'Annot Formation, and concluded that the variations in mineral distribution was due to selective sorting of grains and clasts during transport as a result of size, shape and density. In this study, the mineralogy of the mudstones was also deduced as comprising chlorite and sericite, angular quartz grains, and disseminated organic carbon and pyrite (op. cit.). Stanley (1963) recorded a decrease in the amount of garnet, from the base upwards in individual beds; also there was an increase in the abundance of rutile and zircon from the base upwards, although these grains were smaller and more dense.

A detailed study on the distribution and lateral variability of heavy minerals in the Grès d'Annot Formation was undertaken by Stanley (1964), from which he deduced that four heavy mineral associations occur in the southern French Alps: (1) Southern zone, defined by staurolite-kyanite-garnet; (2) Intermediate zone, defined by staurolite-kyanite-resistates (zircon, tourmaline, rutile); (3) Northern zone, garnet-resistates-apatite; and (4) Dôme de Remollon-Lac d'Allos area, defined by resistates and apatite. The distribution of the above four associations are shown in Figure 7.5. Stanley (1964) concluded that the heavy mineral associations represented different source areas supplying different parts of the foreland basin areas.
Figure 7.5 Lateral petrographic variability of heavy mineral associations in the Annot sandstones and related time-equivalent formations. (redrawn from Stanley 1985).
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In the southern non-flysch portion of the Grès d'Annot Formation, sediments were derived from crystalline rocks of the Maures-Esterel Massifs and their Permian sedimentary cover. Another source area located towards the WNW of the basin, comprising a sedimentary series in a region between the Pelvoux Massif and the Dôme de Barles, and a source area derived from the east in the northern flysch portion of the succession (Stanley 1964).

7.2.2 Results from this study

The sandstone petrology is based on a cursory thin section study. Clasts within the sandstones are generally subangular to subrounded, and comprise a variety of types, including polycrystalline quartz clasts, gneissic material with quartz and feldspar and varying amounts of muscovite, and chert fragments. Also noted are secondary alteration minerals associated with feldspars, including possibly sericite and chlorite.

Field observations on conglomeratic clasts show both intrabasinal and extrabasinal clasts. Intrabasinal clasts include large reworked pieces of the Marnes Bleues and Calcaires Nummulitiques Formations, where clasts attain sizes up to 1 m in diameter. Extrabasinal clast include large, up to 0.5 m diameter granitic/gneissic clasts, basalt clasts, chert, and limestone clasts.

7.2.3 Conclusion

Sediments forming the Grès d'Annot Formation essentially comprise granitic, gneissic, and andesitic material, with minor amounts of reworked older sediment comprising chert, sandstone and limestone. Andesitic material is most abundant in sandstone exposures from the St Antonin area, where andesite debris flows are common in the Middle and Upper Members. and The Grès d'Annot Formation lacks any major amounts of metamorphic rock fragments which
suggested that the sediments were not derived from an orogenic mountain belt. The provenance of the Grès d'Annott Formation has been studied in detail by Ivaldi (1974), using quartz thermoluminescence to fingerprint the possible source areas for the formation. Ivaldi (1974), concluded that sediments were sourced from Hercynian basement massifs similar to those on Corsica and Sardinia, and the Maures-Esterel Massif located towards the south of the study area. Sediments were also derived from overthrust Helminthoid Flysch in the northern parts of the study area (ibid.). Studies on the basement rocks of the Argentera-Mercantour Massif, which now outcrops towards the north of the study area, precludes this as a possible source area during sandstone deposition suggesting that the massif remained submerged during sandstone deposition (ibid.). The conclusion made from this cursory study and that of other workers (Gubler 1955, Stanley 1963, 1964, and Ivaldi 1974), suggests that sediments were essentially derived from basement material, with a prominent input of volcanic material in the form of andesites, which are believed to be contemporaneous with sandstone deposition.
References


Graz, S. 1840. Statistique minéralogique du département des Basse-Alpes ou description géologiques des terrains qui constituent ce département. Prudhomme, Grenoble, 1, 224.


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PAIRIS, J. L. 1971. Effets de la tectonique en <coins> sur la marge orientale du synclinal 

PICKERING, K. T. P. 1982. The shape of deep-water siliciclastic systems: A discussion. Geo-
Marine Letters, 2, 41-46.

PICKERING, K. T. 1987. Wet-sediment deformation in the Upper Ordovician Point 
Leamington Formation: an active thrust-imbricate system during sedimentation, Notre 
Dame Bay, north-central Newfoundland. In: Jones, M. E. and Preston, R. M. F. (eds), 
Deformation of sediments and sedimentary rocks. Special Publication of the 
Geological Society London 29, 213-239.

PICKERING, K. T. AND HISCOTT, R. N. 1985. Contained (reflected) turbidity currents from 
the Middle Ordovician Cloridorme Formation, Quebec, Canada: an alternative to the 
antidune hypothesis. Sedimentology, 32, 373-394.

PICKERING, K. T., CLARK, J. D., SMITH, R. D. A., HISCOTT, R. N., RICCI LUCCHI, F. AND 
problems for sand-prone deep-water systems. In: PICKERING, K. T., HISCOTT, R. N., 
environments: Architectural styles in turbidite systems, 000-000. Chapman and Hall 
(London).

facies, processes and models: a review and classification scheme for modern and 

416. Unwin Hyman (London).

PICKERING, K. T., UNDERWOOD, M. B. AND TAIKA, A. 1992. Open ocean to trench 
turbidity-current flow in the Nankai Trough: Flow collapse and reflection. Geology, 
20, 1099-1102.

PIPER, D. J. W., FARRE, J. A. AND SHOR, A. N. 1985. Late Quaternary slumps and debris 

POSAMENTIER, H. W. AND ALLEN, C. P. 1993. Variability of the sequence stratigraphic 

technique dans le bassin marin Eocene superieur-Oligocene des Alpes du sud. Revue 
de l'Institut Francais du Petrole, 42, 529-553.


classified by grain size and feeder system. American Association of Petroleum


